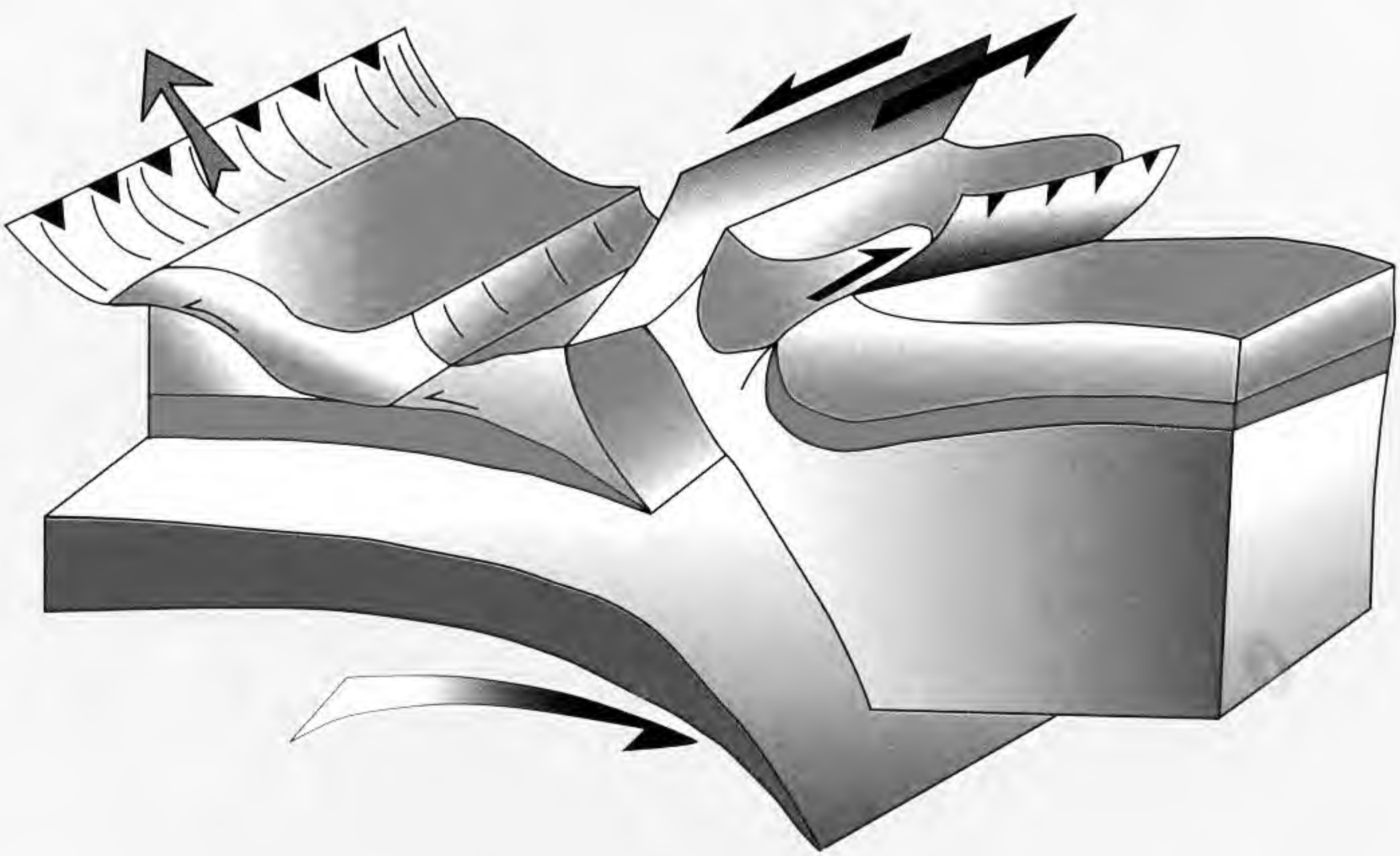
Peri-Tethys Memoir 2

Structure and Prospects of Alpine Basins and Forelands

Edited by
Peter A. ZIEGLER
& Frank HORVÀTH



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The Alpine orogen. Block diagram outlining the role of strain-partitioning in areas of oblique convergence or collision (see paper by Ziegler & Roure, p. 21).

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MÉMOIRES DU MUSÉUM NATIONAL D'HISTOIRE NATURELLE

TOME 170
GÉOLOGIE

Peri-Tethys Memoir 2

Structure and Prospects of Alpine Basins and Forelands

edited by

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CONTENTS

PRÉFACE/*PREFACE*

La Téthys s'est installée au sein de la Pangée à partir du Trias. On la retrouve aujourd'hui entière ou en lambeaux en Amérique centrale, dans l'Atlantique, dans les chaînes méditerranéennes et moyennes orientales, l'Himalaya et les montagnes indonésiennes. Elle débute par des distensions permo-triasiques affectant de vastes surfaces pangéennes, puis l'accrétion se localise au cœur de ce domaine distendu et ouvre une voie d'eau au Jurassique supérieur entre la Laurasia et le Gondwana. Le plus souvent, la distension entre ces mégacontinents se fait à la faveur d'accrétions océaniques génératrices de plaques peu nombreuses aux limites simples, à l'image de l'Atlantique actuel. Ultérieurement, des collisions, des subductions, des obductions font progressivement disparaître cet océan, hormis dans l'Atlantique entre Afrique du Nord-Ouest et États-Unis où il perdure.

On examine dans ce volume une partie seulement du domaine téthysien, celle comprise entre l'Europe d'une part et l'Afrique, l'Arabie et le Moyen-Orient, d'autre part. Là, la complexité est forte car les plaques européennes et africaines ne se séparent pas simplement ; des microcratons s'égrènent entre les deux mégacratons ; de nombreuses plaques existent, ce qui crée un seuil lithosphérique méditerranéen qui, lors de la collision des mégacontinents égéens, crée les chaînes fort complexes analysées dans ce mémoire. De tels seuils sont connus à l'ouest et à l'est de la Téthys dans des domaines limités, ce sont le seuil des Caraïbes entre les Amériques et le seuil indonésien entre les bassins océaniques, pacifiques et indiens.

Les chaînes présentées dans ce mémoire sont analysées à partir des domaines géologiques de surface et de subsurface obtenus en usant de techniques et de méthodes récentes et des concepts actuels, mais aussi à partir de données géophysiques nombreuses ; la plupart d'entre elles avaient jusqu'à présent valeur patrimoniale pour les entreprises industrielles qui les avaient acquises, d'autres ont été très récemment acquises par des programmes internationaux. Peter A. ZIEGLER et Frank HORVÀTH ont coordonné dès leur conception ces diverses monographies sans jamais contraindre les auteurs à se conformer à un cadre inutilement rigide. Les bassins flexuraux à l'avant des chaînes et les bassins molassiques à l'arrière des chaînes sont particulièrement étudiés et renouvelés ; chaque lecteur peut lui-même établir de fructueuses comparaisons.

The Tethys Ocean set itself in place at the centre of Pangea from the Trias. Remnants can be found even today either complete or in outliers in Central America, in the Atlantic, in the Mediterranean and Middle Eastern ranges, the Himalayas and the mountains of Indonesia. It began when Permo-Triassic distension affected vast areas of Pangea, followed by accretion concentrated at the heart of this extended region and, in the Upper Triassic, opened up a limb of water between Laurasia and Gondwana. In the most frequent of cases, distension between these two supercontinents occurred with the aid of oceanic accretions that generated a small number of plates with straightforward boundaries, like the present Atlantic. Later on collisions, subductions and obductions progressively closed up this ocean, obliterated except in parts of the Atlantic between North-West Africa and the USA where it persists.

This book focuses on just one part of the Tethyan realm between Europe on the one hand and Africa, Arabia and the Middle East on the other. The situation here is highly complicated because the boundary between two major plates – the European and African – is far from simple. Strings of small plates formed between the two super-cratons to constitute a Mediterranean lithospheric sill which at the time of the Aegean collision of the supercontinents created the extremely complex mountain belts analyzed in this work. Such sills are known in restricted areas to the west and east of Tethys: the Caribbean sill between North and South America and the Indonesian sill between the great basins of the Pacific and Indian Oceans.

The mountain chains featured in this work are examined from the starting point of surface and subsurface geological domains found using modern techniques and methods and current concepts, and also from a large amount of geophysical data, most of which up to the present formed part of the heritage of the industrial companies that had acquired them; other data have been acquired very recently through international research programmes. Peter A. ZIEGLER and Frank HORVÀTH have coordinated these diverse monographs since their inception without restricting the authors to compliance with an unnecessarily strict framework. Particular focus is placed on updating notions on flexural basins in front of the chains and molasse basins behind the chains; readers can hence make their own fruitful comparisons.

Les chaînes péri-méditerranéennes et les chaînes associées résultent de l'évolution de talus et de marges téthysiennes ; la genèse des bassins flexuraux dans les avant-pays est une réponse des cratons et des microcratons du seuil lithosphérique méditerranéen à l'épaississement crustal.

Les distensions anté-téthysiennes, celles associées à l'accrétion, se font sentir loin sur les cratons bordiers bien au-delà des domaines qui s'infléchiront sous les chaînes et deviendront des bassins flexuraux. Ensuite, lors des étapes de la collision des mégacratons et des microcratons constitutifs du seuil lithosphérique méditerranéen, des subsidences reprennent le long de failles fragiles et ductiles anciennes, le long de coulissements. Ces multiples conséquences de l'évolution de la Téthys de sa naissance à sa fermeture sur les plates-formes cratoniques bordières sont au cœur du programme international « *Peri-Tethys Programme* ».

Une série de mémoires jalonnent ces travaux et ceux conduits depuis plusieurs années. Le mémoire Péri-Téthys n° 1 était consacré aux méthodes d'études des plates-formes. Les prochains seront consacrés aux résultats acquis sur ces plates-formes bordières de la Téthys et des chaînes qui en sont issues.

Nous remercions le Muséum national d'Histoire naturelle pour avoir accepté d'inclure cette série de travaux dans sa collection renommée de monographies scientifiques.

Ces mémoires s'inscrivent dans l'ensemble des études téthysiennes réalisées par des équipes auxquelles sont associés ou qu'animeront les coordonateurs et participants de ces programmes. Les travaux antérieurs de références sont :

The chains around the Mediterranean and those associated with them result from the changes that occurred in the continental slope and Tethyan margins as the ocean evolved. The formation of the flexural basins in the foreland is a reaction of the cratons and microcratons of the Mediterranean lithospheric sill to crustal thickening.

The effects of the pre-Tethyan distensions, associated with accretion, were felt far away on the bordering cratons well beyond the regions which were to bend underneath the chains to become flexural basins. Subsequently, in the collision phases between supercratons and the microcratons that were to constitute the Mediterranean lithospheric sill, subsidence events resumed along the ancient weak and ductile faults, along strike-slip faults. These multiple repercussions of the evolution of the Tethys, from its first opening to its closure onto the bordering cratonic platforms, are the crux of the International Peri-Tethys Programme.

A whole series of reports stand out as landmarks in this research and work that has now been under way for several years. The collection Peri-Tethys No 1 has been devoted to methods of investigating the platforms. Subsequent ones will bear on the results obtained on these Tethyan margin platforms and the mountain ranges that resulted from them.

We thank the Muséum national d'Histoire naturelle (Paris) for having included this series of works among its collection of monographs of excellent composition.

These collections of work fit in well with the whole body of research on the Tethys performed by research teams of which the coordinators and participants of these programmes are either members or leaders. Previous works of reference are as follows:

ZIEGLER, P. A., 1990. — *Geological Atlas of Western and Central Europe*, Shell Internationale Petroleum Maatschappij, distributed by Geological Society, London, Bath, 239 pp., 5 maps.

DERCOURT, J., RICOU, L.-E. & VRIELYNCK, B. (eds), 1993. — *Tethys Palaeogeographical Maps*, Gauthier-Villars & Beicip, Paris, distributed by Commission de la Carte Géologique du Monde, 77 rue Claude Bernard, Paris, 307 pp., 14 maps.

ROURE, F. (ed), 1993. — *Peri-Tethyan Platforms*, Technip, Paris, 275 pp.

NAIRN, A., RICOU, L.-E. & VRIELYNCK, B. (eds), 1996. — *The Ocean Basins and Margins*, vol. 8, The Tethys Ocean Plenum, New York & London, 530 pp.

Jean DERCOURT & Maurizio GAETANI

FOREWORD

The Peri-Tethys Memoir 2 developed out of the symposium on "Structure and Prospects of Alpine Basins and Forelands", held during the *American Association of Petroleum Geologists* International Conference and Exhibition October 17-20th, 1993 in The Hague, The Netherlands. This two and one half days symposium was convocated by Peter A. Ziegler and Frank Horvàth; it aimed at discussing the structure, evolution and hydrocarbon habitat of the different segments of the Alpine-Mediterranean fold-and-thrust belts and associated foreland and back-arc basins. The program consisted of 33 papers covering the Alpine system from the Gibraltar Arc to the Black Sea and the frontal thrust belt of the East Taurus. Abstracts of this symposium were published in the American Association of Petroleum Geologists Bulletin, 77 (9), 1993. Participation of several speakers from Eastern Europe and the former Soviet Union in this international symposium was sponsored by *Shell Internationale Petroleum Mij. B.V.*

In the course of this symposium it was realized that papers presented provided a unique insight into the data accumulated, mainly by the petroleum industry, during its exploration for oil and gas in the different fold-and-thrust belts of the Alpine system and its associated basins. As this information is vital to the understanding of the evolution of the Alpine system and the Peri-Tethyan platforms, speakers were canvassed already during the symposium for a possible contribution of their papers to the publication in a proceedings volume. In general the response was positive and often immediate.

The next step was the task of the conveners to find an organization which was willing to sponsor publication of such a volume. After the *American Association of Petroleum Geologists* had declined sponsorship, the conveners were informed July 13th 1994 that the Executive Committee of the international Peri-Tethys Program, headed by Prof. Dr. Jean Dercourt and Dr. Bernard Tissot, had decided to sponsor publication of the symposium proceedings as Peri-Tethys Memoir 2, to be edited by P. A. Ziegler and F. Horvàth. With this, collection of manuscripts from contributors commenced. Unfortunately, a number of papers presented during the symposium were withdrawn as authors were assigned to different tasks in their organization or had changed to an other company. Nevertheless, the editors were able to assemble 24 papers which provide a fairly comprehensive coverage of the Alpine-Mediterranean system of fold-and-thrust belts and basins. This list includes some papers which were not presented during the symposium but which were invited at a later stage to cover areas of particular interest.

Papers included in this memoir come from the petroleum industry, from national research agencies and from universities. These papers are in so far heterogeneous as the aspects emphasized by the different authors vary considerably. However, all papers address the structural and stratigraphic evolution of the respective area and its hydrocarbon habitat. All papers included in this volume are based on recent compilations and integrations of surface — and sub-surface geological and geophysical data, acquired in the context of hydrocarbon exploration and/or scientific research programs aiming at understanding the architectures and origin of a specific orogen or basin. Much of the data presented in this volume, particularly on East-European areas, was hitherto difficult to access, partly due to its confidentiality and partly due to linguistic difficulties. In respect of the latter, major and often very time consuming editorial efforts were required.

The Peri-Tethys Memoir 2 is organized into an introductory and five regional chapters. In the introductory chapter, the architecture of the Alpine orogen and its hydrocarbon systems are reviewed. In addition, a set of palaeogeographic maps is presented, retracing the Mesozoic opening of Tethys and its Alpine closure, resulting in the suturing of Africa-Arabia and Europe. The following chapters aim at providing the reader with a summary of the structural and stratigraphic evolution, architecture and hydrocarbon potential of selected parts of the Alpine-Mediterranean orogenic belt and its associated basins.

All papers were reviewed by the senior editor and most of them by one or more peer or outside reviewers; thanks are extended to these reviewers for their efforts. The lay-out of this memoir was taken care of by Dr Frank Horvàth, Péter Szafián, László Lenkey and Orsolya Magyari at the Loránd Eötvös University, Budapest. Editorial expenses were covered by the Peri-Tethys Program.

The editors would like to thank all contributing authors for their commitment to this project and their sponsoring organizations for releasing the respective papers for publication. Special thanks go to *EXXON Exploration Company* for absorbing the reproduction costs of the palaeogeographic maps, given in the paper by P. O. Yilmaz *et al.*, and to *Elf-Aquitaine Exploration and Production Company* for financially supporting colour reproduction of text figures given in the paper by M. Le Vot *et al.*

This volume could not have been produced without close cooperation between academic and industrial Earth scientists. Such cooperation forms the basis of the on-going Peri-Tethys Program, the sponsors of which are thanked for having agreed to the publication of this volume.

Peter A. ZIEGLER & Frank HORVÀTH
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Architecture and petroleum systems of the Alpine orogen and associated basins

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ABSTRACT

The Alpine orogen extends from Gibraltar to the Black Sea and consists of an interlinking system of fold-and-thrust belts and associated foreland and back-arc basins. Although these all evolved in response to convergence of the Africa-Arabian and European cratons, and coeval closure of the Tethys oceanic basins, they differ widely in their architecture and evolutionary history in which such aspects as orthogonal or oblique collision, the intensity of collisional coupling of the evolving orogen with its foreland, and the alternation of compression, transpression and transtension, or even extension, played a significant role.

Recently recorded deep seismic profiles image the crustal architecture of some segments of the Alpine orogen. Progressive, gentle orogenward downflexing of the foreland crust, accommodating foredeep basins, as well as localized crustal roots are evident beneath the axial parts of the Pyrenees and the Alps. However, in the northern and eastern Carpathians, in Languedoc-Provence or in the Betic Cordillera, the Moho remains horizontal or even shallows toward the internal zones of these orogens. This is thought to be related to syn-orogenic back-arc extension, as seen in the Pannonian basin and the Tyrrhenian Sea, or to the opening of the oceanic Algero-Provençal Basin. Seismic tomography images deep lithospheric roots in different segments of the Alpine orogen. Lithospheric slabs appear to be still attached to the underthrusted Apulian-Ionian, East-Mediterranean and Moesian crusts beneath the southern Apennines-Calabrian and Aegean arcs and along the south-eastern salient of the Carpathians, respectively.

Effective source rocks are widely distributed within the Alpine orogen and in associated basins. They are alternatively localized in pre-rift, syn-rift or passive margin sequences, the age of which is highly variable. In addition, syn-orogenic foreland basin sequences can host important source-rocks and are frequently the locus of biogenic gas generation. Examples of syn-orogenic oil source rocks are the Oligocene Menilite shales of the Carpathian domain. The Po Plain and the Adriatic foreland basin contain major biogenic gas reserves.

Contrasted thermal regimes and successive episodes of sedimentary and tectonic burial account for the great diversity of petroleum systems identified in the Alpine orogen and associated basins. Each hydrocarbon province is characterized by a very distinct scenario for the timing of source-

Ziegler, P. A. & Roure, F., 1996. — Architecture and petroleum systems of the Alpine orogen and associated basins. *In*: Ziegler, P. A. & Horvàth, F. (eds), Peri-Tethys Memoir 2: Structure and Prospects of Alpine Basins and Forelands. *Mém. Mus. natn. Hist. nat.*, **170**: 15-45. Paris ISBN: 2-85653-507-0.

rock maturation and petroleum expulsion, for hydrocarbon migration from effective kitchens to potential trapping domains, and for the preservation of hydrocarbon accumulations.

INTRODUCTION

The Alpine orogen of the Mediterranean area consists of a system of interlinking fold-and-thrust belts and associated foreland and back-arc basins. These differ in the age of their development and deformation, their tectonic setting and architecture. The evolution of all these features is intimately linked with the Mesozoic break up of Permo-Triassic Pangea, culminating in the step-wise opening of oceanic basins forming the Western Tethys, and their closure during the Alpine orogenic cycle.

For a long time, geologists have identified within the Alpine allochthons remnants of the Tethyan passive margins as well as ophiolites, interpreted as remnants of coeval oceanic basins. On the basis of these interpretations the former plate boundaries between Europe, Apulia and Africa were defined (Biju-Duval et al., 1977; Bernoulli and Lemoine, 1980; Dercourt et al., 1986; Favre and Stampfli, 1992). Oceanographic surveys and off-shore drilling have shed light on the origin of the various Mediterranean sub-basins. These are variably floored by remnants of the former Tethys Ocean (Ionian and Libyan seas), newly formed oceanic crusts (Algero-Provençal basin and Tyrrhenian Sea), extended continental lithosphere (Valencia trough, Aegean Sea) and thick continental lithosphere (Adriatic Sea).

Available geological and geophysical data sets, including recently recorded deep seismic profiles, provide a 3D-image of the present crustal architecture of the Alpine orogen. These data permit palinspastic restoration of former oceanic and continental domains and, by applying plate tectonic concepts, a simulation of the plate kinematics which underlay the evolution of the Alpine orogen. Construction of regional palaeogeographic-palaeotectonic maps has considerably advanced our understanding of the evolution of the Alpine orogen and of the sedimentary basins which are associated with it. Yet, remaining uncertainties, for instance about the configuration of the Tethys embayment at the end of the Variscan orogeny (how far to the East had collisional coupling between Africa-Arabia and Europe progressed and was the northeastern margin of Africa-Arabia indeed fringed by an orogen), the timing of opening and closure of the various Mesozoic oceanic basins forming the Western Tethys and the derivation and dimension of some of the tectono-stratigraphic units which are involved in the Alpine orogen, account for major differences between reconstructions proposed by, for instance, Ziegler (1988, 1994a), Stampfli et al. (1991), Dercourt et al. (1993) and Yilmaz et al. (this volume). It is the objective of the on-going Peri-Tethys program to resolve some of these outstanding questions.

As an introduction to this volume, in which selected case history studies on basins and fold-and-thrust belts and petroleum provinces are discussed, this paper aims at providing an overview of the crustal architecture and evolution of the entire Alpine orogen and its associated basins. The role played by Tethyan syn- and post-rift and Alpine syn-orogenic series in the development of petroleum systems will be discussed with reference to the regional examples documented in the following chapters.

ALPINE OROGENS AND ASSOCIATED BASINS

The complex, arcuate geometry of the Alpine orogen was preconditioned by the rift-induced configuration of its forelands, the pattern of oceanic basins and intervening microcontinental blocks and the kinematics of their interaction during the Alpine convergence of Africa-Arabia and Eurasia. The major elements of the Alpine orogen evolved by closure of oceanic basins of variable size and age; such sutures are characterized by internal ophiolitic zones and major nappes. However, a number of fold belts developed by inversion of intra-continental rift zones (e.g. Atlas, Celtiberian range, Provence-Languedoc, Dauphiné, Dobrogea, Crimea). A transitional feature between these two end members are the Pyrenees which evolved out of an inter-continental transform rift zone, characterized by localized mantle denudation.

Remnants of Mesozoic Tethys Versus Neogene Oceanic Crust

According to geophysical data and palinspastic reconstructions, remnants of the Mesozoic Tethys Ocean are still preserved in the Ionian Sea and probably also in the northern parts of the Eastern Mediterranean. Both domains are at present still being subducted beneath the Calabrian and the Cretan and Cyprus arcs, respectively (Fig. 1). Thick Mesozoic and Cenozoic sediments prohibit direct sampling and dating of the oceanic crust in these areas, which could be as old as Permian (first deep marine sediments on the Pelagian block and pelagic series of the Hawasina nappes in Oman; Stampfli et al., 1991; Stampfli, 1996), or as young as Cretaceous (i.e. coeval with the onset of the rotation of Apulia with respect to Africa and rifting of the Syrte Basin in Libya). In the Eastern Mediterranean, the area occupied by oceanic crust is uncertain (Makris et al., 1983; Sage and Letouzey, 1990) and its age is still debated; similar to most of the ophiolitic units accreted in the Alpine orogen, the oceanic part of the East-Mediterranean basin is probably of mainly Triassic to Jurassic age. However, lack of hard information on the nature and age of the East-Mediterranean crust provides for major uncertainties and differences in palaeo-reconstructions retracing the opening of the Western Tethys and its closure during the Alpine orogenic cycle (Ziegler, 1988; Stampfli, 1996; Dercourt et al., 1993; Yilmaz et al, this volume).

In contrast, the oceanic crust of the Western Mediterranean is Neogene in age. Opening of the oceanic Algero-Provençal Basin is dated as Burdigalian by the transition from syn-rift to thermal post-rift subsidence of its margins and by palaeomagnetic data (Vially and Trémolières, this volume). Oligocene-earliest Miocene rifting in the domain of the Gulf of Lyons and the Valencia Trough, culminating in opening of the Provençal Basin, was contemporaneous with northwest-dipping subduction of the Alboran-Ligurian-Piemont Ocean and thus initiated in a back-arc setting (Maillard and Mauffret, 1993). However, sea-floor spreading in the Algero-Provençal Basin cannot be related to back-arc extension (Ziegler, 1994b). On the other hand, Late Miocene and younger rifting in eastern Sardinia and in Tuscany (Keller et al., 1994; Spadini et al., 1995), as well as magnetic anomalies and results of deep-sea drilling, date opening of limited oceanic domains in the Tyrrhenian Sea as Pliocene and Quaternary (Wezel, 1985). This young oceanic basin developed in a back-arc setting with respects to the Apennine orogen; rifting and opening of the Tyrrhenian Basin behind and above the west-dipping Apennine subduction zone was contemporaneous with continued subduction of the Apulian-Ionian crust (Serri et al., 1993).

Back-arc extension, even if it did not always culminate in opening of oceanic basins, as e.g. in the Aegean Sea (Jolivet et al., 1994) and the Pannonian Basins (Royden and Horváth, 1988; Tari et al., 1992), causes rapid subsidence of formerly elevated compressional structures (negative inversion). Under such conditions, pre-existing compressional detachment faults can be tensionally reactivated, as seen in the Oligocene-early Miocene evolution of the the Languedoc coast (Vially and Trémolières, this volume) and in the Miocene development of the Danube Basin (Tari, this volume).

Ophiolitic Sutures

Unlike the West- and East-Mediterranean basins, which both still comprise undeformed oceanic crust, the Alpine orogen is characterized by several systems of ophiolitic bodies; these represent tectonized and obducted remnants of former Mesozoic oceanic basins, presently localized in often narrow suture zones within the allochthon. As outcropping ophiolitic nappes represent only fragments of these oceanic basins, they do not necessarily record the onset of sea-floor spreading in the respective basins, the timing of which must be

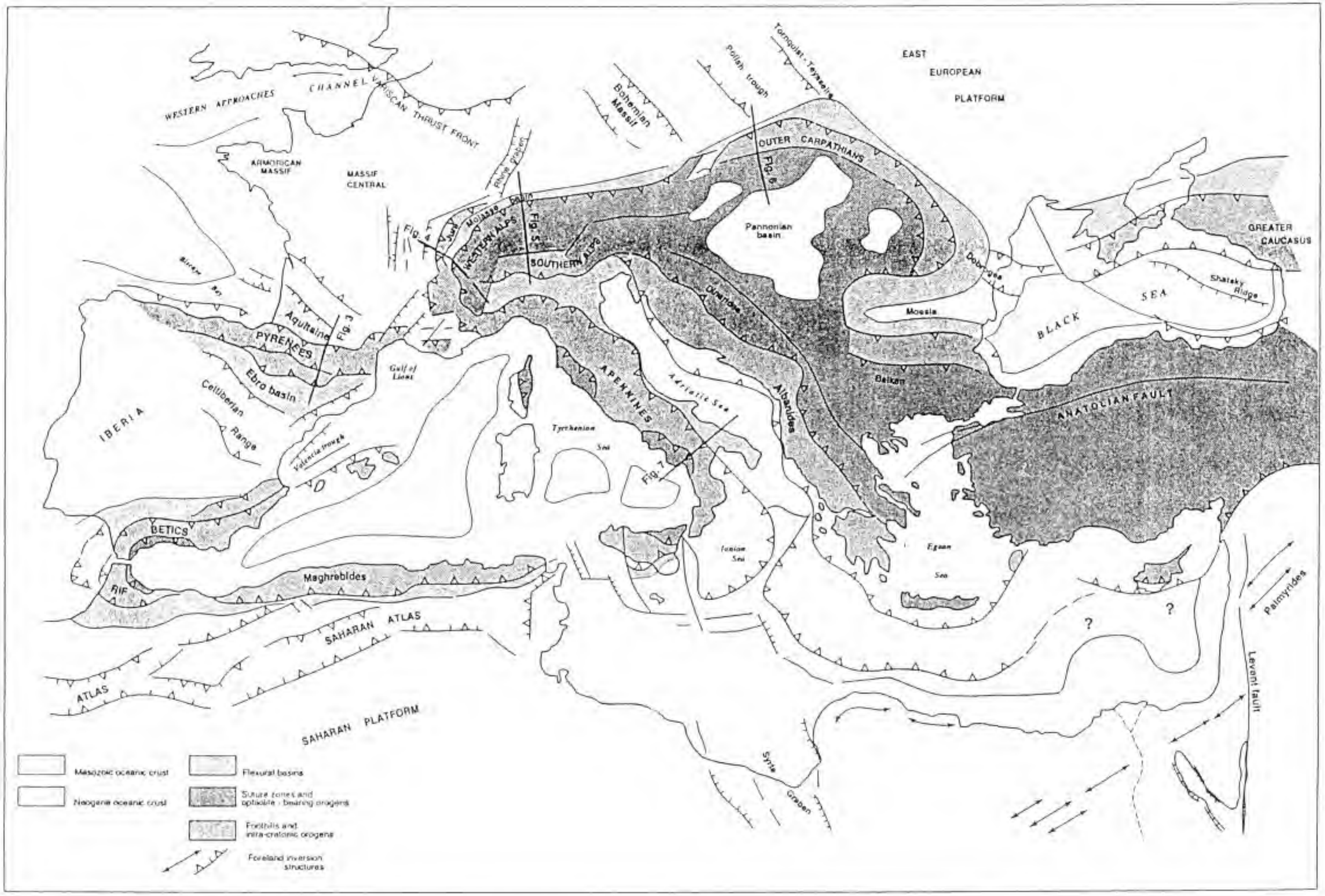

FIG. 1. Present distribution of oceanic sutures and relict oceanic domains within the Alpine orogen.

derived from the sedimentary record of the offsetting passive margin prisms, now involved in the Alpine nappes. Nevertheless, these ophiolitic suture zones relate to distinct segments of the former Tethys and record its step-wise opening (Fig. 1).

During the Late Permian-Early Jurassic initial break-up phase of Pangea, the Hallstatt-Meliata-Mures, Vardar and Sub-Pelagonian and possibly also the East-Mediterranean oceanic basins opened. To the east, this system of oceanic basins finds its continuation in the Izmir-Ankara-Erzincan and the Taurides ophiolitic belts. Opening of this system of oceanic basins resulted in partial separation of the Italo-Dinarides-Anatolia bock from Europe. With the Middle Jurassic development of a discrete transform/divergent plate boundary between Gondwana and Laurentia, opening of the Central Atlantic entailed a change in the opening kinematics of the Western Tethys; these were dominated during the Late Jurassic-Early Cretaceous by a major sinistral translation between Africa-Arabia and Europe, inducing progressive transtensional opening of the Alboran-Ligurian-Piemont-South Penninic Ocean. This was accompanied by decoupling of the Apulian terrane from Africa-Arabia and its complete isolation . At the same time new subduction systems developed in the Vardar-Hallstatt system of oceanic basins, governing their gradual closure. Early Cretaceous opening of the North Atlantic was paired with opening of the Bay of Biscay and of the North Penninic Valais Trough, which may equate to the intra-Carpathian Magura-Piennidic zone (Ziegler, 1988; Dercourt et al., 1993).

Some elements of the Alpine orogenic system, such as the Pyrenees and the Greater Caucasus, lack true ophiolitic sutures, despite the fact that they developed by closure of pre-existing smaller or larger oceanic troughs. In such fold belts ultramafic rocks may occur locally as narrow tectonic slices.

Closure of the different segments of the oceanic Tethys basins during the Alpine orogenic cycle was diachronous, also along the trace of the individual basins For example, in the Central Alps, the South Penninic Ocean was closed during the early Paleocene whereas in the Western Alps, final closure of the Piemont-Ligurian Ocean occurred only during the late Eocene (Ziegler et al., this volume).

Foreland Flexure and Residual Crustal Roots

Deep seismic profiling, including reflection and wide-angle surveys, gravity data, seismic tomography and modelling have provided a better insight into the crustal and lithospheric architecture of the various segments of the Alpine orogen.

Palaeomagnetic data, the inventory of seafloor magnetic anomalies and palinspastic reconstructions of the different segments of the Alpine orogen indicate that Late Mesozoic and Paleogene convergence of Africa-Arabia and Europe amounted to hundreds of kilometres and that it was accommodated by the subduction of equivalent amounts of oceanic and partly also continental lithospheric material (de Jong et al., 1993). This concept is supported by presence of long lithospheric slabs, penetrating the asthenosphere, which are imaged by the distribution of earthquakes and by seismic tomography, for instance beneath the active Calabrian and Aegean arcs as well as in the western Mediterranean (Spakman, 1990; Wortel et al., 1990; Spakman et al., 1993). Alternatively, major slab detachments probably occurred beneath the Alps, the Dinarides and the Carpathians during or soon after the Cretaceous-Paleogene episodes of intense shortening (von Blankenburg and Davies, 1995).

In contrast to major subduction zones, the Pyrenees are characterized by a limited crustal root only; this is in agreement with the relatively small lithospheric contraction during the late Senonian-Paleogene Pyrenean orogeny (about 110 km). Similarly, about 120 km of Oligocene to Recent subduction of the European continental lithosphere beneath the Alps, accounts for their present crustal root and a corresponding new subduction slab (de Jong et al., 1993; ECORS Pyrenees team, 1988; Choukroune et al., 1989; Frei et al., 1989; Roure et al., 1989, 1990; Pfiffner et al., 1988; Schmid et al., 1996). In both cases, conjugate foreland basins developed on either side of the orogen, although seismic and gravimetric data attest for strong asymmetries at depth (Bayer et al., 1989). Whereas in the Pyrenees the Iberian infra-continental mantle was progressively subducted northward, the European lithosphere was underthrusted to the southeast and south beneath the Western and Central Alps.

Other segments of the Alpine chain, such as the Languedoc-Provence, the Western and Eastern Carpathians and also parts of the Apennines, are characterized by limited crustal roots and a Moho which progressively shallows toward the internal zone of these orogens, corresponding to the Gulf of Lions, the Pannonian Basin and the Tyrrhenian Sea, respectively (Figs. 2 and 3; Tomek, 1993; Szafián et al., 1995). Frequently interpreted as a new Moho, this geometry of the crust-mantle boundary probably results from the extensional collapse of the internal parts of these belts, involving extensional reactivation of pre-existing compressional detachment horizons (thrust faults) and progressive denudation of the lower crust (metamorphic core complexes). Under post-orogenic conditions, this can entail rapid uplift and erosion of the external parts of the orogen and unflexing of the foreland lithosphere, as seen, for instance, in the northern Carpathians.

Foredeep basins flanking the Apennines and the southeastern Carpathians are extremely deep and contain up to 10 km of Miocene to Pliocene sediments. In contrast, the eastern Pyrenean and the West-Alpine forelands are almost devoid of syn-flexural sediments.

The width and depth of a flexural basin largely depends on the thickness and thermal regime of the foreland lithosphere, controlling its rheology, as well as on the loads which are exerted on it by the subduction slab and the overriding orogenic wedge (Watts et al., 1982; Kusznir and Karner, 1985; Deségaulx et al., 1991; Doglioni, 1993). Flexural down-bending of thick continental lithosphere can be accompanied by the development of an array of relatively small tensional, essentially basin-parallel normal faults at upper crustal levels, as evident in the German and Austrian parts of the Molasse Basin (Roeder and Bachmann, Zimmer and Wessely, this volume). If such a foreland crust is weakened by pre-existing faults which can be reactivated during its flexural deformation, strain may be concentrated on a few major faults which can have throws of the order of 1 km and more, as seen in the Ukrainian part of the Carpathian foredeep (Sovchik and Vul, this volume). Rapidly and deeply subsiding foredeep basins are generally characterized by strongly stretched and attenuated continental crust which is thinner and warmer than adjacent parts of the foreland. This accounts for the contrasted geometries of the western and eastern portions of the Aquitaine foredeep, and for the segmentation of the Periadriatic depressions in front of the Apennines or between the Dinarides and the Albanides (Royden et al., 1987; Deségaulx et al., 1991; Doglioni, 1993).

In some foreland basins, intra-plate compressional structures play an important role and, by their development, can either impede the development of a flexural basin or can cause partial or total destruction of a pre-existing flexural basin. Such structures can develop by inversion of pre-existing tensional basins, in which case they may involve the basement (i.e. Provence, Dauphiné; Roure and Colletta, this volume). Alternatively, activation of sedimentary detachment horizons within the foreland basin can cause the development of thin-skinned compressional structures, as seen in the Prerifaine chains of Morocco (Zizi, this volume), the South-Alpine Lombardian Basin (Ziegler et al., this volume), the Jura Mountains (Philippe et al, this volume) and also in the Apennine foredeep (Anelli et al., this volume).

Oblique Versus Orthogonal Collision and Strain Partitioning

Although Africa-Arabia converged with Eurasia since Senonian times in a north-south directed counter-clockwise rotational mode, sinistral motions between them, related to the opening of the Atlantic Ocean, decreased gradually and ceased at the transition from the Paleocene to the Eocene in conjunction with opening of the Norwegian-Greenland Sea. However, during the Oligocene and Miocene a dextral component is evident in their convergence pattern; this translatory movement probably persisted into Recent times, as indicated by earthquake focal mechanisms and neotectonic deformations (Ziegler, 1988, 1990).

During the Late Jurassic-Early Cretaceous opening phases of the Central Atlantic, sinistral motion between Africa-Arabia and Europe gov-

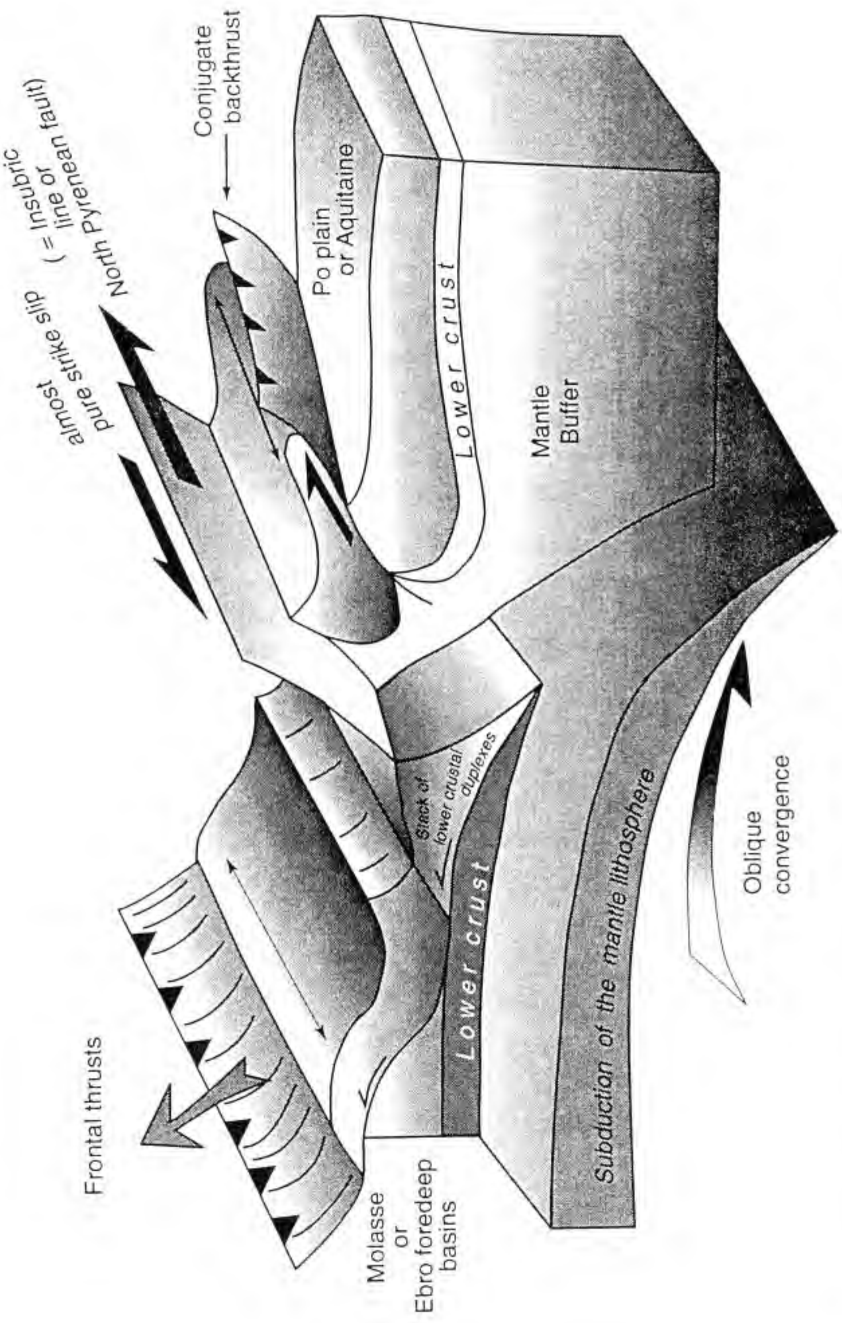

FIG. 2. Block diagram outlining the role of strain-partitioning in areas of oblique convergence or collision.

erned the closure of the Hallstatt-Vardar ocean system, rotation of the Italo-Dinarides-Anatolia block and its progressive incorporation into the Dinarides-Hellenides-Pontides orogenic system. With the Senonian onset of northward drift of Africa-Arabia and progressive opening of the Atlantic, these sinistral motions gradually decreased whereas increasing space constraints in the Western Tethys caused rapid westward propagation of subduction zones into the the domain of the Ligurian-Alboran Ocean. For instance, in the Alps, the Cretaceous orogenic cycle was characterized by northwest-directed mass transport whereas the Paleogene orogeny was dominated by north directed mass transport (Schmid et al., 1996). Paleogene progressive closure of the Tethys and increasing collisional coupling of the evolving orogen with its forelands, lateral block escapes and oblique motions played an increasingly important role. Suturing of Iberia to Europe was accompanied by the development of the left-lateral North Pyrenean Fault. Eastward directed Oligo-Miocene mass transport from the Alpine into the Carpathian domain, as a consequence of full-scale collision of the Adriatic indenter with Europe (Ratschbacher et al., 1991), was accompanied by left-lateral motions along the North Carpathian Pienniny Klippen belt and incipient right-lateral motion along the South Carpathian foothills (Fig. 1; Laubscher, 1992a; Ellouz and Roca, 1994). Miocene-Pliocene development of a system of right-lateral strike slip fault systems, including the South Atlas fracture system, the Insubric line of the Southern Alps, the intra-Dinarides Peri-Adriatic and the North Anatolian fault systems, may be partly related to the dextral translation of Africa-Arabia and Europe and partly to lateral mass redistribution in response to the massive indentation of Arabian and the Adriatic block (Ziegler, 1988).

However, as in still presently active transpressional orogens, such as the South Caribbean belt in Eastern Venezuela, or along the San Andreas Fault west of the San Joachin Valley, a strong strain partitioning occurred within most of the Alpine orogenic belts (Laubscher, 1992a; Passalacqua et al., 1995). Thus, the overall northwest- or northeast-trending collision zone between major plates was ultimately confined to the subducted lithosphere in the footwall, and to the hanging-wall mantel indenter (Fig. 2). In contrast, frontal accretion characterized the conjugate external thrust belts on both sides of the orogenic wedge, with major thrusts always paralleling the active plate boundary (see e.g. Gilmour and Mäkel, Le Vot et al., Philippe et al, this volume). Apparently, the oblique convergence component was largely absorbed within the orogenic wedge by strike-slip motions along, at shallow levels, sub-vertical faults. Such faults can strike normal to the overall convergence direction as e.g. the Insubric line of the Central Alps, or at a high-angle as e.g. the intra-Dinarides Peri-Adriatic line. Their orientation is largely controlled by the geometry of rigid indenters, such as the Apulian platform.

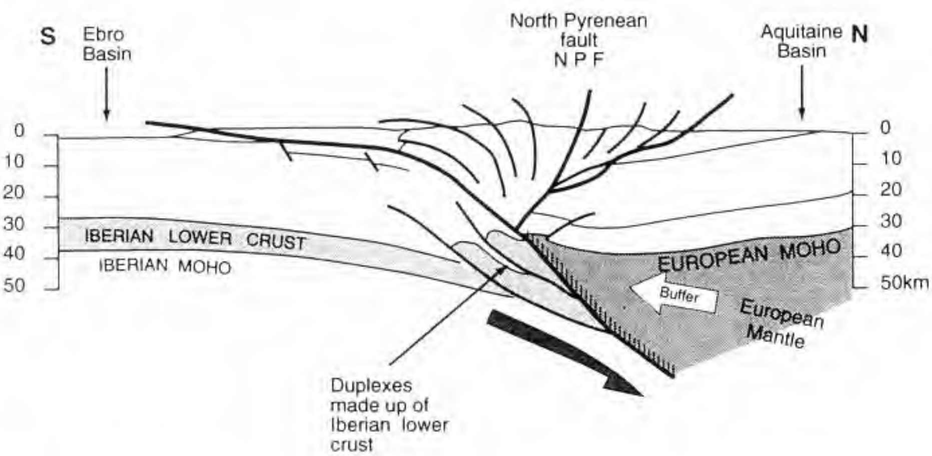

FIG. 3. Crustal section across the Pyrenees

RIFTING AND DEVELOPMENT OF PASSIVE MARGINS

Tethyan rifting initiated soon after the consolidation of the Variscan orogen at the end of the Westphalian (von Raumer and Neubauer, 1993). Development of the widespread, wrench- and rift-induced Permo-Carboniferous successor basins of Western and Central Europe Europe can be related to such processes as changes in the convergence pattern of Gondwana and Laurussia during the Allegenian phase of the Appalachian-Mauretanides orogen, to back-arc extension in response to roll-back of the Variscan subduction slab and its ultimate detachment from the lithosphere and to the onset of regional extension, controlling the break-up of Pangea (Ziegler, 1990, 1993).

In the Western Tethys domain, major crustal extension commenced, however, only after the Appalachian-Variscan suture between Gondwana and Laurussia had become inactive at the transition to the Late Permian (Ziegler, 1989; Stampfli and Pillevuit, 1993). Rifting activity accelerated during the Triassic, propagated westward and interfered in the Atlantic domain with the southward propagating Arctic-North Atlantic rift system (Ziegler, 1988). Step-wise opening of the different oceanic basins of the Western Tethys, resulting in the development of passive margins, as summarized above, is retraced in the palaeogeographic/palaeotectonic maps of Yilmaz et al. (this volume). We recall here only, that probably most of the Tethyan passive margins developed during the Triassic and the Middle Jurassic.

Variscan Inheritance and Localization of Tethys Rifts

The Triassic-Jurassic Gulf of Mexico-Central Atlantic-Western Tethys rift/transform system represents one of the major break-up axes of Permo-Triassic Pangea. Significantly, this break-up axis coincides to a large extent with the Appalachian-Variscan suture of Gondwana and Laurussia (Ziegler, 1990, 1993). Rheological considerations suggest that the orogenically destabilized lithosphere of this suture was considerably weaker than that of the flanking cratons (Cloetingh and Banda, 1992) and, as such, preconditioned the localization of this break-up axis.

In Western and Central Europe and the Western Tethys domain, the lithosphere was further weakened during the Stephanian-Autunian collapse of the orogenically thickened Variscan crust. Wrench- and extensional faulting, resulting in the uplift of core-complexes, was associated with the synkinematic intrusion of granites and the extrusion of alkaline and calc-alkaline magmas. This was presumably accompanied by the detachment of the subducted lithospheric slab(s), at least partial delamination of the deep lithospheric roots of the Variscan orogen and its corresponding uplift. Moreover, transtensional uplift of core-complexes and erosional unroofing of the crust was paralleled by upwelling of the asthenosphere and the interaction of mafic melts with the crust. With the termination of wrench activity at the transition to the Late Permian, cooling of the thermally destabilized lithosphere controlled crustal subsidence and regional transgressions. In time, the crust/mantle boundary re-equilibrated regionally at depths of about 35-40 km (Ziegler, 1990; Ziegler et al., 1995; Costa and Rey, 1995).

In the European Alpine foreland there is ample evidence of repeated Mesozoic reactivation of Permo-Carboniferous crustal discontinuities, in part guiding the localization of rifted structures, such as the Polish Trough (Ziegler, 1990). Deep reflection-seismic profiles show that Variscan compressional structures were only rarely reactivated during the Mesozoic rifting phases in the distal parts of the Alpine forelands.

For instance, the ECORS seismic profiles image beneath the conjugate Ebro and Aquitaine forelands of the Pyrenees south-verging Variscan structures (Servey Geologic de Catalunya, 1993). These were transected by Permo-Carboniferous wrench faults and apparently did not materially contribute towards the localization of Mesozoic extensional faults. In contrast, reactivation of south-verging Hercynian structures probably accounts for coaxial Alpine deformations along the southern flank of the Pyrenees (Deségaulx et al., 1990). On the other hand, the Mesozoic Bay of

Biscay rift zone appears to be superimposed on a major Permo-Carboniferous wrench zone.

Although the architecture of those parts of the Variscan orogen which were overprinted by Alpine deformations is still poorly known (von Raumer and Neubauer, 1993), it is evident that also these areas were affected by intense Permo-Carboniferous wrench tectonics and associated magmatism. It is likely that also in these areas reactivation of Permo-Carboniferous crustal discontinuities played a significant role in the localization of the Tethys rift systems.

Late Carboniferous coal-measures contained in Variscan foreland and successor basins, as well as coal-measures and lacustrine shale of the Stephanian-Autunian wrench-induced troughs, provide potential source-rocks in the pre-rift sedimentary sequences of the European foreland. Depending on their burial beneath the Tethyan passive margin sequence, or beneath syn-orogenic flexural sequences and the Alpine allochthon, such source-rocks can have locally preserved part of their petroleum potential and thus can contribute to effective petroleum systems, as for instance in the Jura Mountains and in the subthrust play of the Polish Carpathians (Bessereau et al., this volume).

Rifting and Development of Passive Margins

As discussed above, rifting activity in the Tethys domain spanned Permian to Early Cretaceous times and culminated in the step-wise opening of its constituent oceanic basins. Late Permian, Triassic and Early Jurassic rifting activity is well documented in the different parts of the Western Tethys (Stampfli and Pillevuit, 1993; Stampfli, 1996). During the Triassic, rifting activity propagated westwards and affected very wide areas around the future zones of crustal separation (Ziegler, 1988). In time, rifting activity concentrated on zones of future crustal separation, as seen for instance in the Southern Alps (Bertotti et al., 1993). From Mid-Jurassic to Early Cretaceous times, rift and wrench activity in the Western Tethys was governed by the sinistral translation of Africa-Arabia relative to Europe, in response to progressive opening of the Atlantic Ocean.

Earliest passive margins were associated with the opening of the Hallstatt-Vardar system of oceanic basins. Mid-Jurassic opening of the Alboran-Ligurian-Piemont-South Penninic ocean resulted in the development of a new set of passive margins. Late Jurassic-Early Cretaceous opening of the Bay of Biscay, the North Penninic and the Magura basins led to the development of yet an other system of passive margins.

Development of the different Tethys rift systems entailed a renewed destabilization of the asthenosphere-lithosphere system. Crustal extension was accompanied by variable levels of rift magmatism. As there is no obvious evidence for hot-spot activity, crustal extension was presumably of a "passive" nature, driven by far field stresses governing the break-up of Pangea (Ziegler, 1993, 1995a). The availability of pre-existing crustal discontinuities, which could be tensionally reactivated, favoured simple shear crustal extension and the development of upper and lower plate margins (Favre and Stampfli, 1992). Locally crustal stretching involved the activation of intra-sedimentary detachment levels; for instance, in the basin of Southeastern France, low-angle listric extensional faults, soling out in Carboniferous coal-measures or Triassic evaporites, account for decoupling of the sedimentary cover structures from the basement. Subsidence of grabens and half-grabens was accompanied by uplift of the rift shoulders and the development of rift-flank basins (Favre and Stampfli, 1992). Upon achievement of crustal separation, marginal graben systems became inactive and were incorporated into the newly formed passive margins.

The duration of the rifting stage of the different Tethys extensional systems is highly variable. For example, the Aquitaine-Bay of Biscay basin records some 130 Ma of intermittent rifting activity prior to its Mid-Aptian transtensional opening (Deségaulx and Brunet, 1990), whereas the Atlas troughs, after about 60 Ma of rifting activity, became inactive in conjunction with crustal separation in the Central Atlantic.

During much of Late Permian to Early Cretaceous times, large parts of the Tethys domain were dominated by carbonate shelves. Syn-sedimentary tectonics accounted for rapid lateral facies and thickness changes. Partly reefal or dolomitic shallow water carbonates were restricted to the rift flanks, the crests of tilted blocks and to little extended platforms, whereas shaly or cherty deeper water carbonates and shales were deposited, partly under anoxic conditions, in rapidly subsiding grabens and in sediment starved lagoons.

Triassic, Jurassic and Early Cretaceous syn-rift source rocks provide for an number of effective petroleum systems (Table 1).

How Passive Were the Tethys Margins

The present day Central Atlantic margins remained in a passive setting for 180 Ma. Also the Arabian Shelf had undergone a passive margin evolution for some 175 Ma before it became collisionally coupled during the Senonian with the evolving Zagros orogen. In contrast the passive margins of the Western Tethys had a relatively short life span. For instance, the eastern margin of the Italo-Dinarides block was incorporated into the Dinarides orogen about 100 Ma after the Vardar Ocean had opened (Frasheri et al., this volume). The Austroalpine margin, facing the South Penninic Ocean, remained in a passive setting for some 60 Ma before it was converted into an active margin during the Early Cretaceous. Intra-Senonian and Paleocene compressional deformation of the East-Alpine and North-Carpathian forelands occurred about 50 Ma after the Valais and Magura basins had opened (Kovac et al., 1993; Ziegler et al., this volume). On the other hand, the Pyrenean margin was only for some 25 Ma transtensionally "passive" before it was compressionally deformed (Le Vot et al., this volume).

In view of the opening kinematics of the Atlantic Ocean and of the Tethyan system of oceanic basins, controlling also the interaction of the different blocks delimited by the latter, many of the evolving Tethyan passive margins were repeatedly tectonically destabilized. This had repercussions on their thermal subsidence pattern and the resulting development of passive margin sedimentary prisms. Examples of tectonic destabilization of Tethyan passive margins are:

(1) Late Jurassic and earliest Cretaceous wrench activity in the Bohemian massif and southward adjacent areas in conjunction with a stress reorganization in the North Sea rift system (Ziegler, 1990)

stratigraphic age	basin type	source-rock type	petroleum provinces
Neogene	flexural foreland	diluted III biogenic gas	E. Molasse, Carpathians Peri-Adriatic Basin
Oligocene	flexural foreland syn-rift	II and III ?	Carpathians, Caucasus E. Molasse Rhine G., Valencia T.
Albian-Turonian	passive margin	II	Apulian & Scythian Platf., Maghrebides
Late Jurassic	passive margin syn-rift	II II-III	Vienna Basin Aquitaine Basin
Middle Jurassis	passive margin	II	Kuban foredeep
Early Jurassic	passive margin syn-rift	II II	Albania, Epirus Sicily, S. Alps
Triassic	syn-rift	I-II	S. Alps, Apennine Siclily, Albanides
Late Carboniferous	flexural foreland	III	Permian Basin
Silurian	passive margin	I-II	S. Anatolian foreland

TABLE 1
Effective source rocks in Alpine Orogen and associated basins

(2) Early Cretaceous opening of the Valais Trough, disrupting the subsidence pattern of the European shelf bordering the South-Penninic trough which had begun to open during Mid-Jurassic times (Stampfli, 1993)
(3) Late Jurassic-Early Cretaceous wrench-induced deformation of the Moroccan shelf during the opening of the Central Atlantic and Alboran-South Penninic oceans (Favre and Stampfli, 1992)
(4) Early Cretaceous rifting in Libya and Egypt. It is likely that also thermal subsidence of the Italo-Dinarides block was influenced by its Early Cretaceous rotation.

The timing of incorporation of the different peri- and intra-Tethyan passive margin prisms into syn-orogenic flexural basins is also highly variable. Whereas during Cretaceous times progressively more internal units of the Italo-Dinarides Block were incorporated into the Hellenic-Dinarides foreland basin, the western shelves of this block were incorporated into the Apennine foredeep only during the Late Oligocene and Miocene (Frasheri et al., Anelli et al., this volume). A special case is presented by the Helvetic shelf of the Central and Eastern Alps which began to subside during the Eocene under the load of the advancing nappes; however, pre- or even syn-collisional stresses exerted on this shelf induced transpressional reactivation of pre-existing crustal discontinuities and profound disruption of its sedimentary cover. For instance, intra-Senonian and Paleocene inversion of the Polish Trough and the Bohemian Massif resulted in partial destruction of the passive margin prism of the East-Alpine and the North- and East-Carpathian forelands (Ziegler, 1990; Ziegler et al., 1995).

Correspondingly, the age, thickness and composition of the different peri- and intra-Tethyan passive margin prisms is highly variable. These passive margin prisms attained maximum thickness prior to their incorporation into flexural basins. Correspondingly, potential source-rocks of the syn-rift and passive margin sequences reached maximum sedimentary burial and started to generate hydrocarbons already prior to their subsequent burial beneath flexural foredeep prisms or the Alpine allochthonous units. In distal foreland areas, which escaped such burial, the syn-rift and passive margin sedimentary series preserved part of their petroleum potential (e.g. Aquitaine Basin, Le Vot, this volume).

The Tethyan passive margin sequences account for the source-rocks, reservoirs and/or seals of some of the most efficient petroleum systems identified within the external parts of the Alpine fold-and-thrust belts and their forelands (Table 1). Examples are the Early Jurassic Posidonia shales of the Albanian Ionian zone (Baudin and Lachkar, 1990; Frasheri et al., this volume) and the Aquitaine basin (Le Vot et al., this volume) and the Late Jurassic shales beneath the Vienna basin (Ladwein et al., 1991). Although numerous Cretaceous black-shale intervals have been reported from the peri-Tethyan passive margins, particularly from the Albian-Cenomanian and Turonian series of the Periadriatic domain, in Aquitaine or in the Carpathians (Le Vot et al., Stefanescu and Baltes, this volume), their efficiency is still questionable as only limited hydrocarbon accumulations could be directly correlated with them.

CONVERGENCE, COLLISION AND SUBDUCTION OF CONTINENTAL LITHOSPHERE

We recapitulate that in the Western Tethys initiation of and activity along subduction zones was controlled during the Late Jurassic-Cretaceous Pangea break-up phase by the sinistral translation of Africa-Arabia and Europe, and during Senonian to Recent times by their Alpine convergence. As the European and Africa-Arabian passive Tethys margins were not parallel, and as the Iberian and the Italo-Dinarides microcontinents acted as indenters during the Alpine cycle, collisional events were diachronous (Tapponnier, 1977). Preservation of remnants of Tethyan oceanic crust in the Ionian Sea and in the Eastern Mediterranean illustrates that continent-to-continent collision has not yet occurred along the entire trace of the Tethys suture (Fig. 1).

The onset of subduction processes can be stratigraphically dated by the sedimentary record of accretion prisms, or petrologically and radiometrically by arc volcanism and HP/LT metamorphism. The collision of an arc trench system with a passive margin, marking the transition from subduction of oceanic to continental lithosphere (B-type to A-type subduction), is reflected by the rapid subsidence of the foreland crust and by the deposition of syn-orogenic flysch series on top of the passive margin sequence. However, it must be realized, that initial collision zones are generally deeply buried beneath orogenic wedges and in part have been subducted to great depths. Only in rare cases will subduction progradation, or perhaps extension, entail the exhumation of an earlier active subduction zone and render it accessible to surface observation (e.g. Tauern window; Froitzheim et al., 1996). These criteria have been applied in the construction of palaeogeographic/palaeotectonic maps retracing the evolution of the Alpine orogen (Ziegler, 1988; Dercourt et al., 1993; Stampfli, 1996; Yilmaz et al, this volume).

Not only the timing of orogenic activity is highly variable in the different segments of the Alpine system but also the amount of crustal shortening, including subduction of oceanic crust and post-collisional subduction of continental lithosphere. In this respect, the intensity of collisional coupling of the evolving orogenic wedge with its foreland, as well as the availability of a thick sedimentary cover which could be detached from the foreland crust, played an important role in the architecture of the different fold-and-thrust belts forming the Mediterranean-Alpine orogen. In the following selected examples are discussed which are further analyzed in the different chapters of this volume.

Pyrenees

The Pyrenees evolved in response to late Senonian-Paleogene transpressional closure of the wedge-shaped inter-continental Bay of Biscay Basin; this was induced by the build-up of far-field compressional stresses related to the convergence of Africa with Europe, causing clock-wise rotation and escape of Iberia. Crustal shortening across the Pyrenees amounts to some 110 km (Roure et al., 1989; Deségaulx et al., 1990).

Their axial zone is characterized by asymmetrically north- and south-verging, thrusted basement blocks. Their northern foreland, the Aquitaine Basin, is characterized by partly reactivated extensional fault bocks and thin skinned thrust sheets which are detached from the basement at the level of Triassic evaporites (Le Vot et al., this volume). The southern Pyrenean external zone is characterized by thin skinned thrust sheets, detached from their basement at the level of Triassic evaporites; the basement cores of these sheets form part of the internal structure of the Pyrenees (Vergés and Munoz, 1990). The Pyrenean orogeny is coeval with orogenic activity in the Betic Cordillera. Paleogene compressional stresses exerted on cratonic Iberia caused reactivation of Permo-Carboniferous and Mesozoic crustal discontinuities, controlling inversion of the Celt-Iberian and Catalonian Coast ranges and upthrusting of the Sierra Guadarrama basement block (Ziegler, 1988; Salas and Casas, 1993; Ziegler et al., 1995).

Commensurate with the amount of crustal shortening achieved across the Pyrenees, they are characterized by a limited north-verging crustal root and no pronounced lithospheric root; they lack a syn-orogenic magmatism (Fig. 3; Spakman, 1990; Servey Geologic de Catalunya, 1993).

Whereas the northwestern, deep foreland basin of the Pyrenees hosts the major Aquitaine hydrocarbon province (Le Vot et al., this volume), its eastern shallower parts and the southern foreland contain no significant hydrocarbon accumulations.

Western Alps

From plate reconstructions, the total crustal shortening achieved across the Western Alps is estimated to amount to some 400 km (Platt et al., 1989). This involved Late Cretaceous activation of the Apulian margin, Late Cretaceous-Paleogene closure of the oceanic Piemont, Eocene closure of the Valais basins (Froitzheim et al., 1996) and sub-

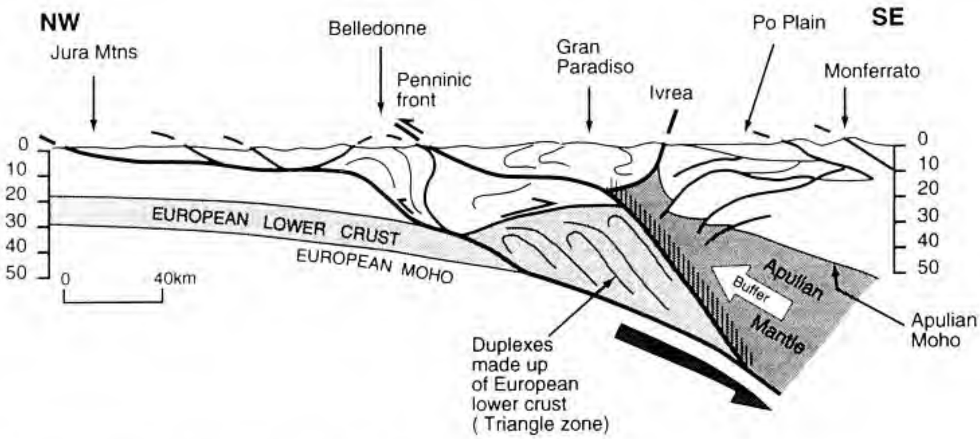

FIG. 4. Crustal section across the Western Alps.

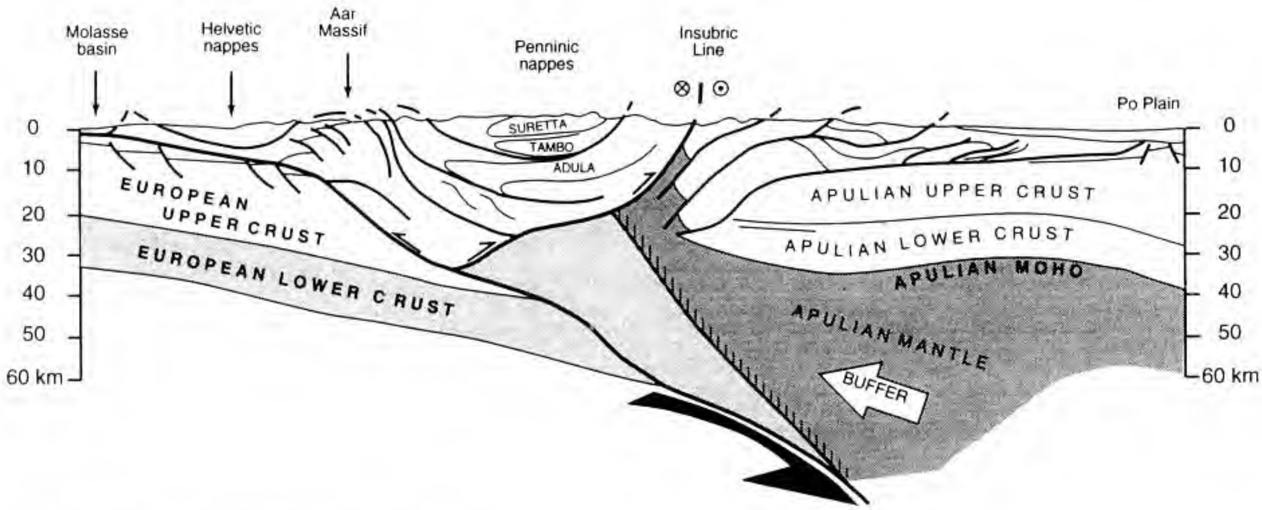

FIG. 5. Crustal section across the Central Alps

sequent imbrication of the European foreland crust, resulting in the uplift of the external Pelvoux, Belledonne and Mont Blanc crystalline massifs (Fig. 4).

Nappes derived from the Apulian margin occupy the northeastern part of the Western Alps (Sesia-Lanzo and Dent Blanche nappes). The internal Schistes Lustrés nappe, containing ophiolites, was derived from the Piemont Ocean. The Valais Trough and the Briançonnais were deformed into a system of west- and east-verging nappes, respectively, which overrode the passive foreland margin, causing late Eocene-Oligocene subsidence of a flexural foreland basin. Subsequent intense coupling of the orogenic wedge with the foreland is evident by compressional reactivation of Mesozoic tensional faults on the distal foreland margin, the step-wise imbrication of the foreland crust and detachment of the passive margin series from their basement. By this process the earlier developed foreland basins was largely destroyed. During late Eocene to Miocene times, the West-Alpine foreland was transected by a system of grabens which form part of the European Cenozoic rift system (Ziegler, 1994b). During Miocene and Pliocene times, Mesozoic and Cenozoic extensional basins in the foreland of the Western Alps were inverted, causing disruption of the Mesozoic proximal passive margin sedimentary prism; moreover thrusting propagate into the domain of the Jura Mtns. (Roure et al., 1990; Roure and Colletta, Philippe et al, this volume).

Commensurate with the dominantly west-vergence of the external Western Alps, their up to 60 km deep crustal root (Laubscher, 1992b) is located under their eastern, internal parts. Moreover, seismic tomography images an about 175 km deep lithospheric root, located beneath the western Po Plain and an apparently detached, east-dipping subduction slab (Spakman, 1990). The Western Alps are devoid of subduction related magmatism. Their evolution reflects increasing collisional coupling of the orogenic wedge with the European foreland in which contemporaneous inversion structures developed as far away as in the Western Approaches and the Celtic Sea (Ziegler, 1990; Ziegler et al., 1995).

The external parts of the Western Alps and their foreland contain no significant hydrocarbon accumulations. Sub-thrust plays are areally restricted due to the limited extent of the foreland beneath the frontal thrusts.

Central Alps

The Central Alps evolved by Late Cretaceous and Paleogene closure of the larger South Penninic and the smaller North Penninic Valais oceanic basins. The total amount of crustal shortening achieved across the Central Alps is in the range of 500 to 550 km (Fig. 5; Schmid et al., 1996; Ziegler et al., this volume).

The Central Alps are characterized by major, north-verging, partly basement cored nappes. The Austroalpine nappes were derived from the southeastern margin of the South Penninic Ocean. The Penninic nappes derive from the North and South Penninic troughs and the Briançonnais block, separating them. The sedimentary Helvetic nappes were derived from the northern, European shelf; their basement core is represented by the Gotthard Massif and the lowermost Penninic nappes. Mio-Pliocene uplift of the external Aar Massif, entailing deformation of the overlaying stack of nappes, was partly contemporaneous with the development of the Jura fold-and-thrust belt. The Southern Alps consist of a south-verging internal stack of basement imbrications and an external, thin-skinned thrust belt.

End Cretaceous to early Paleocene closure of the South Penninic trough was accompanied by large radius deformation of the Helvetic shelf, causing erosion of much of its Cretaceous sedimentary cover. Late Eocene closure of the North Penninic trough was followed by flexural subsidence of the Helvetic shelf under the load of the advancing Austroalpine, Penninic and Helvetic nappes; the Helvetic nappes consist of sediments which were detached from the foreland crust. Post-collisional indentation of Apulia, amounting to some 120 km, caused overthickening of the orogenic wedge. This gave rise to Oligocene back-folding and -thrusting along the Insubric line, Mio-Pliocene step-wise imbrication of the northern and southern foreland crust and ultimately thrust propagation into the conjugate flexural foreland basins, causing their partial destruction. Post-collisional crustal shortening was accompanied by the subduction of continental lithospheric material.

South-directed underthrusting of the European foreland gave rise to the development of a some 60 km deep crustal root beneath the southern part of the Central Alps and a 175 km deep lithospheric root located beneath the northern parts of the Po Plain (Spakman, 1990). Oligocene detachment of an earlier formed subduction slab from the lithosphere was accompanied by the intrusion of partial melts derived from the mantle-lithosphere (von Blankenburg and Davies, 1995).

The South-Alpine external, thin skinned thrust belt evolved out of a Triassic extensional basin, a superimposed passive margin prism and a late Oligocene-mid-Miocene flexural foreland basin; it hosts significant hydrocarbon accumulations (Anelli et al., this volume). The northern, Molasse foreland basin contains a thick syn-orogenic clastic wedge which rests on a relatively thin passive margin sequence. This basin is internally little deformed, extends only some 25 km beneath the Alpine nappes to the frontal basement imbrications of the Aar massif and is limited to the north by the Jura fold-and-thrust belt. The Molasse Basin has a very limited hydrocarbon potential (Ziegler et al., Philippe et al., this volume).

Eastern Alps

In contrast to the Central Alps, the Eastern Alps are characterized by an autochthonous basement which extends from the northern thrust front for at least 60 and perhaps as much as 100 km under the stack of Austroalpine nappes (Zimmer and Wessely, Tari, this volume). The eastward disappearance of major Helvetic nappes, involving European passive margin sediments, is striking. As Penninic nappes play a subordinate role in the architecture of the Eastern Alps, it is assumed that the Valais Trough terminates to the east or merges with the South Penninic Trough. In this context, Froitzheim et al. (1996) propose that the crystalline core of the Tauern window is formed by European crust and, as such, is not equivalent to the Briançonnais, as commonly suggested (Tollmann, 1985).

Evolution of the Eastern Alps involved Late Jurassic closure of the Hallstatt and Cretaceous closure of the Penninic oceans. Cretaceous mass transport was northwesterly directed. Profound late Senonian and Paleocene disruption of the European passive margin shelf, involving transpressional reactivation of Permo-Carboniferous and

FIG. 6. Crustal sections across the Carpathians, imaging the changes from collision to back-arc extension or collapse (after Roure et al., 1996).

Mesozoic crustal discontinuities and ensuing uplift and erosion (Ziegler, 1990), presumably reflects initial intense coupling between the evolving orogenic wedge and its foreland. During the Paleogene the Austroalpine nappes were thrusted over the foreland crust; this was accompanied by the flexural subsidence of the eastern Molasse basin, which is characterized by an array of syn-flexural tensional faults. By early Miocene times, the sedimentary Austroalpine nappes had arrive near their present location. During the emplacement of the East-Alpine stack of nappes, the orogen was apparently mechanically decoupled from the foreland, as indicated by the absence of compressional foreland structures. Paleocene destruction of the Mesozoic passive margin prism, particularly in areas located to the south of the Bohemian massif, is held responsible for the eastward disappearance of the Helvetic nappes.

The small remnant foreland basin of the Eastern Alps hosts several petroleum systems. These are tied to Mesozoic and early Oligocene oil source-rocks and to biogenic gas generated in the Oligocene and early Miocene deeper water clastic series of the foreland basin fill. The subthrust play of the Eastern Alps has to contend with considerable reservoir and structural risks, deep objectives and topographic difficulties. The sedimentary allochthon contains important hydrocarbon accumulations beneath the Vienna Basin; elsewhere its hydrocarbon potential has as yet to be proven (Roeder and Bachmann, Zimmer and Wessely, this volume).

Northern and Eastern Flysch Carpathians

The Northern and Eastern Carpathians consist of a stack of sedimentary nappes, involving Early Cretaceous to Miocene flysch series which are thrusted over the foreland. The latter extends at least some 75 km under this orogenic wedge, which accounts for at least 250 km of shortening (Fig. 6). The ophiolite bearing Pienniny Klippen zone marks the internal boundary of the Flysch Carpathians, in which the involvement of passive margin sediments remains conjectural (Sandulescu, 1984; Roure et al., 1993; Bessereau et al., Sovchik and Vul, Dicea, this volume).

The Flysch Carpathians evolved in response to Mid-Cretaceous and Cenozoic south and westward subduction of the oceanic Magura-Piennide basin and eastward displacement of the intra-Carpathian North Pannonian, Tisza and Dacides blocks; these record Late Jurassic and Early Cretaceous orogenic events related to the closure of the Vardar-Hallstatt Ocean (Sandulescu, 1984; Csontos et al., 1992; Stefanescu and Baltes, this volume).

The passive margin sedimentary prism of the North- and East-Carpathians forelands was disrupted by the late Senonian and Paleocene deep inversion of the Polish Trough (Ziegler, 1990) and the Early Cretaceous inversion of the Dobrogea Trough (Belov et al., 1987). These features appear to link up beneath the Eastern Carpathians. Correspondingly, large parts of the autochthonous foreland beneath the Flysch Carpathians lack a thick passive margin prism which otherwise could have been detached from its basement during the Late Cretaceous and Cenozoic phases of the Carpathian orogeny. Reflection-seismic data show that flexural subsidence of the autochthonous foreland was accompanied by major normal faulting. Moreover, they image mild compressional deformation of the autochthonous basement beneath the internal parts of the Flysch Carpathian accretionary wedge; however, major basement structures of the Aar Massif type are lacking (Roure et al., 1993, Sovchik and Vul, Dicea, this volume).

Roll-back of the subducted oceanic slab and its dehydration is tracked by a chain of mid-Miocene to Pliocene calc-alkaline volcanics, paralleling the Klippen zone (Szabó et al., 1992). Deep reflection profiles through the North-Carpathians image the steeply south dipping European foreland crust (Tomek, 1993). The absence of post-Paleocene compressional foreland structures and the presence of only minor compressional basement structures beneath the Flysch Carpathians suggest that coupling between the orogenic wedge and its foreland was at a low level. Post-Oligocene development of the Flysch Carpathians was accompanied by the collapse of the Pannonian Basin in their hinterland (Royden and Horváth, 1988; Tari et al., 1992).

The Flysch Carpathians and their foreland host important petroleum systems. The most important one is related to the Oligocene Menilite shales; a second potential petroleum system is related to Cretaceous black-shales. Jurassic and even Palaeozoic source-rocks have locally contributed hydrocarbons (Bessereau et al., Brzobohaty et al., Stefanescu and Baltes, Dicea, this volume).

Peri-Adriatic Thrust Belts

The stable Adriatic (Apulian) platform is flanked to the east by the Dinarides-Albanides and to the west by the Apennines. Whereas in the Dinarides-Albanides orogenic activity commenced during the Late Jurassic and persisted variably into Miocene and Pliocene times, the Apennines evolved essentially during Neogene times (Frasheri et al., Anelli et al., this volume).

The west-verging **Albanides** are characterized by thin skinned thrust sheets which are detached from their autochthonous basement at the level of Triassic evaporites. These thrust sheets involve a thick Triassic-Jurassic passive margin sequence, dominated by carbonates, and Cretaceous to Paleogene flysch series which grade to the west into platform carbonates. Mio-Pliocene molasse series are involved in the most external zone. Westward progression of the flysch facies through time describes the gradual advance of the orogenic front and the migration of the flexural foreland basin. The up to 14 km thick Mirdita ophiolite nappe forms the orogenic lid of the Albanides. These ophiolites, which presumably represent the oceanic crust of Sub-Pelagonian trough, were obducted during the Late Jurassic closure of the Vardar system of oceanic basins. The autochthonous foreland crust extends essentially unbroken from the thrust front of the Albanides at least 100 km under their internal nappes

The external, Ionian zone of the Albanides host a major hydrocarbon province which derives its charge from multiple Mesozoic source-rocks (Frasheri et al., this volume).

In contrast to the long lived Albanides, the **Apennines** are a young orogenic belt which evolved only during late Oligocene to Plio-Pleistocene times. The Apennines are characterized by major sedimentary nappes; these involve Triassic to Paleogene syn- and post-rift shallow-water and pelagic series and Oligocene to Miocene flysch, deposited in an eastward migrating foreland basin. The flysch nappes are detached from their substratum. An ophiolitic nappe, or rather an ophiolitic mélange, is only locally preserved along the Tyrrhenian coast (Liguride units). The external flysch nappes override a thick parautochthonous and autochthonous Triassic to Paleogene sedimentary sequence, consisting predominantly of platform and pelagic carbonates (Fig. 7).

Following Paleogene closure of the Alboran-Ligurian-Piemont Ocean, the Apennines evolved in response to Oligocene and younger westward underthrusting and subduction of the Adriatic foreland. The flysch facies mirrors the progressive eastward migration of the flexural foreland basins and the gradual incorporation of its proximal parts into the evolving orogenic wedge (Ricci Lucchi, 1986). The most internal units of the Apennines are formed by Ligurides, representing the sedimentary fill of the Ligurian-Alboran Basin, and the basement involving Calabria-Peloritani nappes which are of uncertain origin. The main elements of the allochthon are derived from the distal parts of the Apulian platform which were separated from its main parts by the deep-water Lagonegro trough. Emplacement of these sedimentary nappes was accompanied by partial detachment of the autochthonous sedimentary cover of the Apulian platform, its imbrication and by in-sequence thrust propagation into the Adriatic Sea (Anelli et al., this volume). There is only little evidence for involvement of the Apulian crust in the orogenic edifice of the Apennines (Ponziani et al., 1995); correspondingly, the orogenic wedge and the foreland crust were largely mechanically decoupled during the Apennine orogeny.

Substantial supra-crustal shortening, evident in the Apennines, was apparently compensated by the subduction of continental lithosphere, giving rise to extensive magmatic activity along the western margin of Italy from mid-Miocene times onward. Steepening and roll-back of the subducted lithospheric slab is held responsible for contemporaneous back-arc rifting and the gradual opening of the Tyrrhenian Sea (Spakman et al., 1993; Serri et al., 1993; Doglioni, 1993).

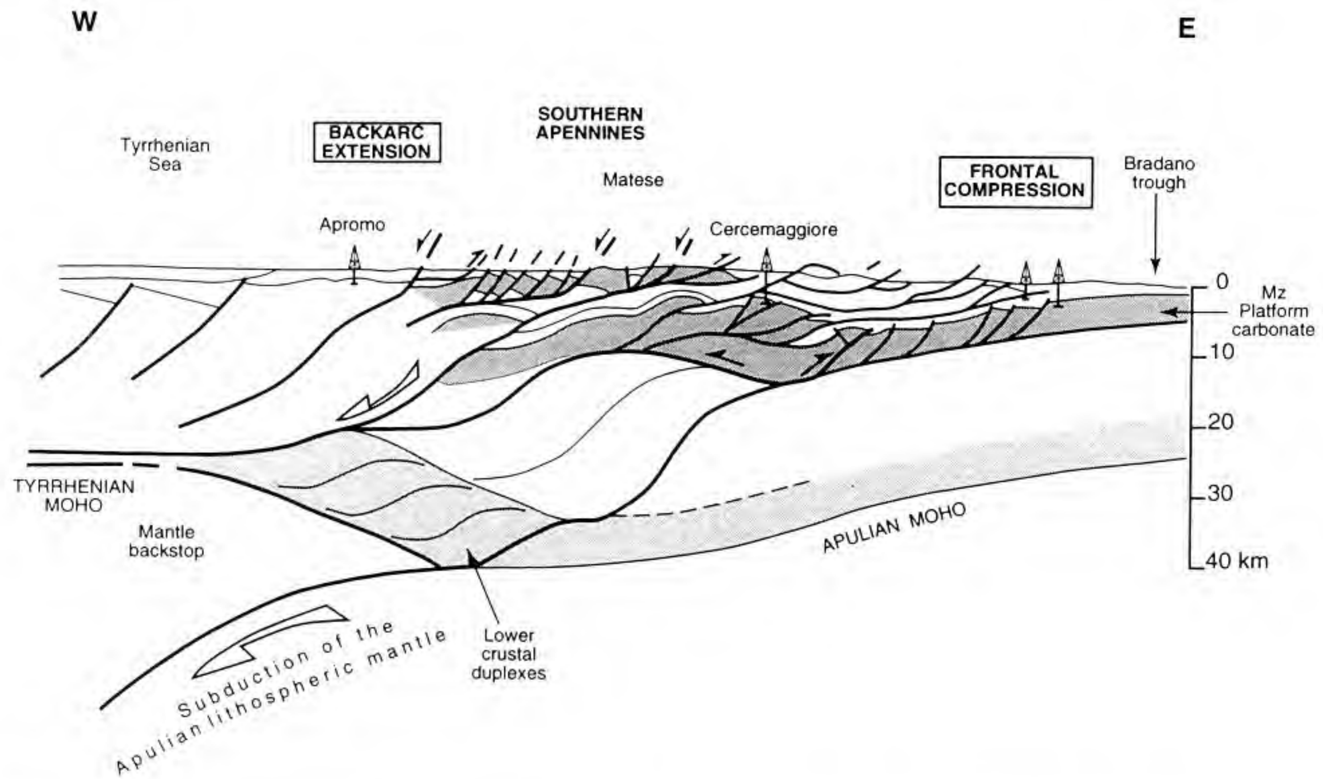

FIG. 7. Crustal section across the Apennines imaging the changes from collision to back-arc extension or collapse (after Roure et al., 1996).

The Apennines and their foreland basin host important petroleum systems which are related to Triassic and Jurassic source-rocks, deposited during the syn-rift stage. Secondary petroleum systems are related to the passive margin sequence and the syn-orogenic flexural basin. The latter contains major amounts of biogenic gas (Anelli et al., this volume).

East-Taurus Foothills Thrust Belt

The foothills of the eastern Taurus orogenic belt consist of a 30 to 40 km wide zone characterized by south-verging thin-skinned thrust imbricates, cored by Albian-Campanian shelf carbonates which were detached from the autochthonous Palaeozoic sedimentary cover of the Arabian Shield. To the north this thrust belt is limited by ophiolite and matamorphic nappes. Its southern foreland is characterized by a broad belt of intraplate compressional structures. This thrust belt developed during the Late Senonian to Eocene collision and of the North-Arabian passive margin with the Taurides arc-trench system (Yilmaz, 1993; Gilmour and Mäkel, this volume).

The North-Arabian passive margin developed during the Late Triassic with the opening of the Tethys Ocean. Its passive margin sequence spans Jurassic to Campanian times; it is interrupted by a major Early Cretaceous hiatus which truncates Jurassic strata to the north. Development of a flexural foreland basin commenced during the late Senonian and was accompanied by the obduction of the ophiolitic nappes and imbrications of the Cretaceous passive margin series. The external thin-skinned thrusted structures are sealed by Paleocene and younger series which were gently folded during the late Eocene. Eocene and Miocene continued thrusting was confined to the internal zones of the orogen and was accompanied by the closure of remnant oceanic basins. Miocene(?) foreland compression caused reactivation pre-existing crustal discontinuities, resulting in the upthrusting of foreland structures at distances of over 100 km to the south of the Alpine trust front.

The external, thrusted structures of the Taurus foothills are charged with hydrocarbons generated by Silurian shales of the autochthonous Arabian foreland, forming part of the Tethyan pre-rift sequence (Gilmour and Mäkel, this volume).

To round-off this pallet of diversified structural styles of the Alpine chains, attention is drawn to the paper by J. Flinch (this volume) who describes the gravitational collapse of the external accretionary prism of the Rif-Betic Cordillera arc which spilled over the Atlantic continental margin of Morocco and Iberia, forming the Gibraltar allochthon.

ALPINE FORELANDS

During the Late Cretaceous and Cenozoic phases of the Alpine orogeny, both the European and the Africa-Arabian forelands record a sequence of intra-plate deformations which caused drastic changes in their structural and palaeogeographic evolution.

Micro-tectonic analyses indicate that the trajectories of principal horizontal compressional stress axes affecting the European and Africa-Arabian platforms changed repeatedly during Late Cretaceous and Cenozoic times (Letouzey and Trémolières, 1980; Bergerat, 1987; Müller, 1987; Blès et al., 1989; Lacombe et al., 1990); this can be related to changes in their convergence pattern. Moreover, from late Eocene onward, gradual development of the Red Sea-Suez-Libyan-Pelagian Shelf and Rhine-Rhône-Valencia-Trans-Atlas rift systems was broadly contemporaneous with compressional intra-plate deformations. This is interpreted as reflecting the interference between stress systems related to the late phases of the Alpine collision and stresses related the gradual assertion of a new tensional kinematic regime, governing a fundamental plate-boundary reorganization, which may ultimately lead to the break-up of the present continent assembly (Ziegler, 1988, 1990, 1994b; Ziegler et al., 1995).

Intra-plate Compression

During the Late Cretaceous and Cenozoic, intra-plate compressional stresses reactivated a broad spectrum of tensional and transtensional Mesozoic and Palaeozoic basins in Europe and in Africa-Arabia, causing their inversion (Fig. 1). At the same time there are indications for broad buckling of the lithosphere, such as Plio-Pleistocene accelerated subsidence of the North Sea basin, uplift of the Fennoscandian Shield and Neogene isolation of the Paris Basin (Kooi et al., 1989; Ziegler, 1990; Ziegler et al., 1995).

In Europe, inversion structures and upthrusted basement blocks occur within a radius of up to 1500 km from the thrust front of the Alpine orogen. Late Cretaceous and Paleogene inversion features are located in the northern and northwestern foreland of the Alps and in the foreland of the Pyrenees, whereas Eocene and younger inversion structures are located in the foreland of the Western Alps. (Ziegler, 1987, 1990; Roure and Colletta, this volume).

In northern Africa, main intra-plate compressional structures evolved by latest Cretaceous and Eocene inversion of the Triassic Atlas rift system. (Casero and Roure, 1994; Zizi, this volume). The Saoura-Ougarta chain is a Palaeozoic rift which was inverted during the Permo-Carboniferous and underwent a second Alpine inversion during the Eocene (Ziegler, 1988).

On the Arabian craton, the Sinai arc, consisting of the Negev and the Palmyrides fold belts, represents a major inversion feature. The Palmyrides developed by inversion of a Permo-Triassic aborted rift. First mild compressional deformations occurred at the end of the Cretaceous; the major Mio-Pliocene phase of basin inversion involved transpressional deformations in response to NNW-SSE directed compression. Inversion of the Palmyra trough was contemporaneous with the thrusting phases in the East-Taurus orogenic belt and early movements along the Dead Sea wrench fault (McBriden et al., 1990; Chaimov et al., 1993). Transpressional deformation of the Negev fold belt is mainly late Senonian in age; minor inversion movements continued into the Miocene (Quennell, 1984; Moustafa and Khalil, 1995).

Basically we recognize two phases of prevailing intra-plate compression. The first phase is related to initial collisional coupling between a nascent orogen and its foreland (e.g. Carpathian and East-Alpine foreland). Such deformations may be controlled by the crustal configuration of the respective passive margin (upper or lower plate, availability of a thick passive margin prism or the lack thereof), the rheological characteristics of the oceanic lithosphere separating the accretionary wedge from the respective passive margin and the rate of their convergence. The second phase of foreland compression is related to lithospheric overthickening of the orogenic wedge and to resulting thrust propagation into the foreland (e.g. West and Central Alpine foreland). Processes governing the coupling and decoupling of an orogenic wedge and its foreland are, however, still poorly understood (Ziegler et al., 1995).

Inversion can have severe repercussions on the hydrocarbon habitat of a basin, mainly by reversing its subsidence pattern, by re-configuration of pre-existing structural traps and profound erosion, resulting in the loss of hydrocarbons to the surface. However, partly inverted basins can host important hydrocarbon provinces such as those of the Lower Saxony and West Netherlands basins (Ziegler, 1990, 1995b).

Cenozoic Rift Systems

Whereas lateral escape of, for instance, the Intra-Carpathian blocks, rotation of microplates, such as the Corsica-Sardinia block, and progressive retreat of subducted slabs in the Tyrrhenian Sea and the Aegean arc controlled the distribution of Neogene extensional structures within the frame of the overall compressional Alpine system of fold-and-thrust belts, late Eocene and younger development of the Rhine-Rhône and the East-African-Gulf of Suez-Libyan-Pelagian Shelf rift systems cannot be directly linked to the evolution of the Alpine orogen (Fig. 1).

The Cenozoic rift system of Western and Central Europe extends from the shores of the North Sea over a distance of some 1100 km into the West-Mediterranean domain; from there an alkaline volcanic chain projects southwestwards across the Alboran Sea, the Rif fold belt and the Atlas ranges to the Atlantic coast and to the Cape Verdes Islands. Including this volcanic chain, the entire rift system has a length of 3000 km. This rift system began to evolve during the middle and late Eocene in the European Alpine foreland and propagated during the Oligocene northward and southward. In the western Mediterranean, crustal extension culminated, after a rifting period of only 7 Ma, in Miocene crustal separation and the opening of the oceanic Algero-Provençal Basin; this involved a counter-clockwise rotation of the Corsica-Sardinia block (Torné et al., Vially and Trémolières, this volume). During Miocene and Plio-Pleistocene times, tectonic and partly also volcanic activity persisted along the different segments of this mega-rift system, albeit under changing regional stress regimes. Development of the Cenozoic rift system of Western and Central Europe was contemporaneous with the Eocene and later phases of the Alpine orogeny, during which the northwestern Alpine foreland was repeatedly subjected to horizontal intra-plate compressional stresses, causing inversion of Mesozoic extensional basins at considerable distances from the Alpine thrust front. The southern elements of the European Cenozoic rift system cross-cuts the Alpine chains of the western Mediterranean domain.

Viewed on a broader scale, evolution of the West and Central European Cenozoic rift system was broadly contemporaneous with the development of the East African-Red Sea, Libyan and Pelagian Shelf rift systems; its Neogene development was paralleled by back-arc extension governing the subsidence of the Pannonian Basin and the Aegean, Tyrrhenian and Alboran seas. As such the Rhine-Rhône-Valencia and the Red Sea-Libyan-Pelagian Shelf rift systems can be considered as forming part of the Neogene Alpine-Mediterranean collapse system (Ziegler, 1988, 1990). In this context, Neogene progressive eastward escape of Apulia and Sicily relative to North African may have contributed to the evolution of the Pelagian rift system (Casero and Roure, 1994).

However, considering the dimensions of the West European-West African and the East-African-Red Sea-Libyan rift systems, it is hardly conceivable that such major rifts developed solely in response to the Alpine collision. More likely, they form part of a new break-up system which may culminate in the disruption of the Alpine plate assembly (Ziegler, 1994b).

The European Cenozoic rift system hosts the Rhine Graben and Valencia Trough hydrocarbon provinces; whereas the former relies exclusively on syn-rift source-rocks, the latter relies on a combination of syn- and pre-rift source-rocks (Ziegler, 1995b, Torné et al, this volume). The major Gulf of Suez hydrocarbon province relies exclusively on Late Cretaceous pre-rift source-rocks (Ziegler, 1995b). The hydrocarbon charge of the Pannonian system of back-arc basins is related to a combination of syn- and pre-rift source-rocks. Uplift of major rift domes, such as those associated with the Rhine Graben and the Red Sea, causes disruption of the pre-rift platform sedimentary sequences and changes the hydrodynamic setting of the remnant rift flank basins (e.g. Paris Basin).

PETROLEUM SYSTEMS OF ALPINE FORELANDS AND EXTERNAL THRUST BELTS

The complex geodynamic and tectonic evolution of the Peri-Tethyan platforms and the subsequent Alpine development of the European-Africa-Arabian plate boundaries have conditioned both the distribution and maturation history of potential source rocks in the conjugate forelands as well as within the Alpine orogen and its successor basins.

Pre-Orogenic Versus Syn-Orogenic Source Rocks

Palaeozoic source-rocks of the pre-rift sequence, preserving part of their initial petroleum potential until the onset of Alpine flexuring and thrusting, play an important role on the Arabian platform (Silurian shales; Gilmour and Mäkel, this volume) and in Western and Central Europe (mainly Carboniferous and Autunian; Ziegler, 1990). Numerous Mesozoic source rocks have been identified within the Mesozoic syn-rift series (mainly Triassic and Early Jurassic) and in the passive margin sequences of the former Tethys (mainly Late Jurassic to late Early Cretaceous). These pre-orogenic sequences account for most of the oil and thermal gas potential of the Alpine orogen and its associated basins.

In addition, organic-rich series (TOC values up to 10%) have been identified in the Oligocene syn-flexural fill of the East-Alpine and Carpathian foredeep (Menilite series); these are coeval with the prolific Maykop source-rocks of the Black Sea, Crimea and the foredeeps of the Great Caucasus and Southern Caspian. Based on oil/source-rock correlations and the occurrence of the oleanane biomarker, even in subthrust Mesozoic reservoirs of the foreland, these Oligocene series are believed to be the most efficient source-rock in the Outer Carpathians and the adjacent East-European foreland, from Poland to Romania (Ulmishek and Klemme, 1990; Koltun, 1992; Ten Haven et al., 1993; Lafargue et al., 1994; Bessereau et al., this volume). Similar facies occur in the German and Austrian Molasse basin where they are the primary source of accumulated oils (Zimmer and Wessely, Roeder and Bachmann, this volume). Oligocene series display also good TOC values in the successor Pannonian basin where they have been sufficiently buried, under a relatively high geothermal gradient, to enter the oil window and account for significant hydrocarbon reserves in Hungary and former Yugoslavia. In addition, Oligocene shales have a good source-potential in the Cenozoic rift basins of Western Europe (Vially and Trémolières, Torné et al., this volume)

The thick Mio-Pliocene terrigenous fill of the Carpathian and Periadriatic foredeeps (Apennines, Albanides) have TOC values averaging 1%. Mostly immature, these Neogene series account for large reserves of bacterially-generated gas, but only for minor oil (Kotarba, 1992; Anelli et al., this volume).

Sedimentary Versus Tectonic Burial of Source-Rocks

For most Alpine foreland fold-and-thrust belt systems, available surface and sub-surface geological and geophysical data provide sufficient constraints to construct reliable regional structural cross-sections, extending from the autochthonous foreland to the frontal parts of the thrust belt. Combined with a good biostratigraphic dating of the syn-orogenic sequences, this permits construction of high quality balanced cross-sections and their step-wise palinspastic restoration, retracing the evolution of the respective area from the onset of deformation to its Present configuration (Bally et al., 1988; Roure et al., 1993; Zoetemeijer et al., 1993).

Forward kinematic modelling permits to retrace the maturation history of potential source-rocks within an evolving Alpine foredeep basins, i.e. during their initial sedimentary burial and their subsequent tectonic burial beneath the advancing allochthonous units. Ultimately, also uplift and erosion, related to tectonic accretion of individual units into the Alpine orogenic wedge, can be simulated. Coupled with palaeo-thermal reconstructions and transformation kinetics of kerogens into hydrocarbons, these new techniques provide an efficient tool to date the timing of source-rock maturation and to evaluate migration pathways between hydrocarbon kitchens and potential traps (Roure and Sassi, 1995).

Distinct source-rocks frequently coexist in a single Alpine foreland basin. However, due to lateral and vertical changes in their distribution, they usually contribute to independent petroleum systems, characterized by unique timing of maturation, hydrocarbon expulsion and migration. Therefore, it is essential to obtain an impression of potential migration pathways between source areas and traps and the timing of trap development, before drilling a prospect, in order to decrease exploration risks in such complex areas as the Alpine orogen.

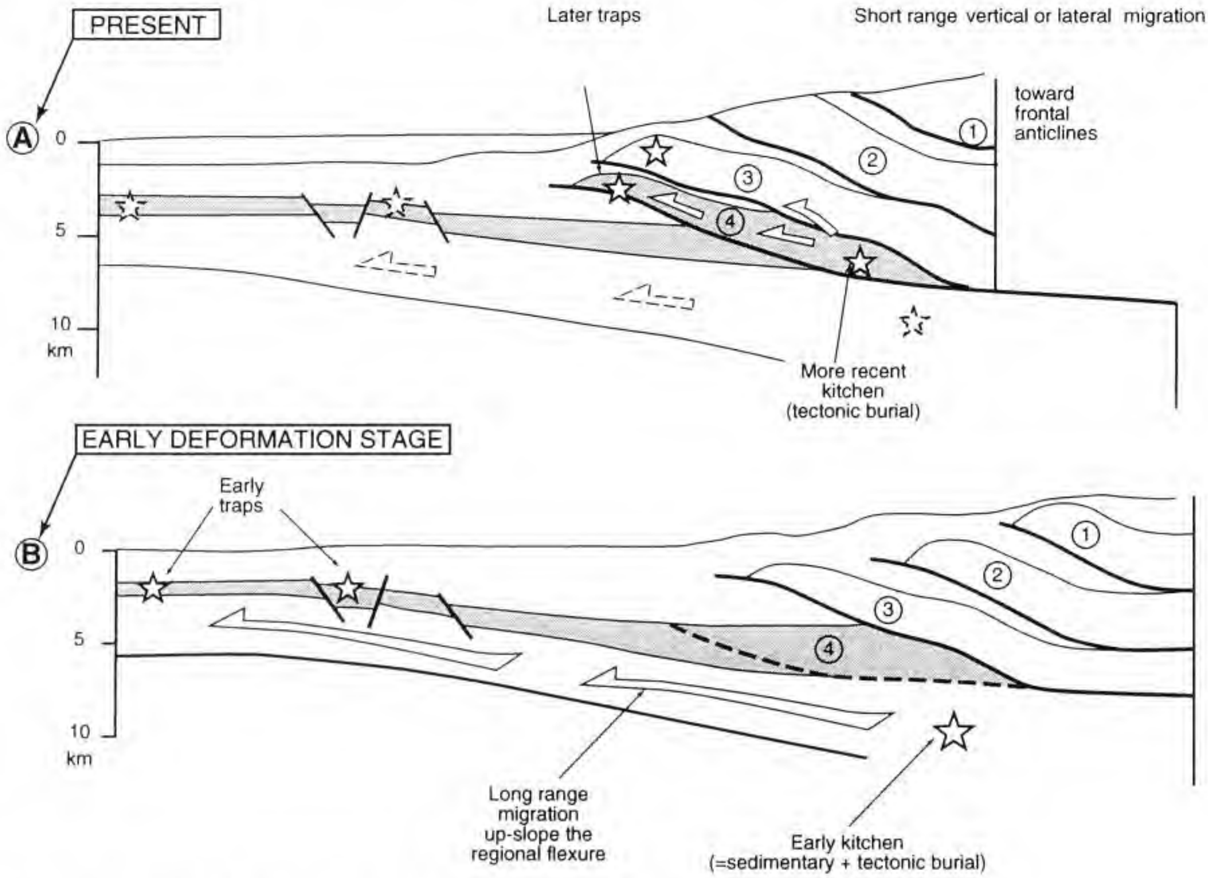

FIG. 8. Evolution of a foreland fold-and-thrust belt system and distribution of effective petroleum systems:
a) Sedimentary burial, early maturation and long-range migration pathways
b) Tectonic burial, late maturation and short range migration pathways.

Long-Distance Lateral Versus Short-Distance Vertical Migration

Calibration and reconstruction of palaeo-temperatures and -hydrodynamics (convective heat and fluid transfers) has hardly been addressed for the external imbricate structures of the different Alpine fold-and-thrust belts. However, distinct episodes of petroleum generation and migration are usually recognized in the Carpathians, Apennines and Albanides, which constitute the most productive petroleum provinces in the European part of the Alpine orogen (Roure and Howell, 1996).

Most frequently, oil is generated early in foreland basins, resulting in long range migration of the first hydrocarbon products from the foredeep source area toward the foreland, up-slope the regional flexure. This buoyancy driven mechanism is even facilitated by regional hydrodynamics, accounting for recharge of meteoric waters in the foothills and discharge in the foreland in the area of a potential forebulge, if present (Fig. 8). Thus, early long-distance updip migration of oils generated by Triassic source-rocks in the Albanian foredeep, charging the Aquila field, located on the eastern Puglia slope in the Adriatic (Anelli et al., this volume), is indicated. Apparently, the Mesozoic reservoirs at Lopushnia oil field in the Ukrainian sub-thrust foreland (Izotova and Popadyuk; Sovchik and Vul, this volume), were laterally charged by the Oligocene series of the Outer Carpathians; these are presently entirely detached from their former substratum and are now accreted to the allochthon (Lafargue et al., 1994).

A residual oil potential is eventually preserved within the allochthonous units or in the under-thrusted foreland, accounting for late phases of maturation and migration. Under such conditions, expelled hydrocarbon can hardly reach the foreland as frontal antiform structures are already well developed and provide traps; instead, hydrocarbons generated during such late stages are preferentially trapped in nearby structures, requiring short distance lateral and/or vertical migration. This concept applies, for instance, for the deeply buried duplexes of the Borislaw-Pokut zone of the Ukrainian Outer Carpathians (Sovchik and Vul, this volume) and the subthrust para-autochthonous plays of the Southern Apennines (Casero et al., 1991; d'Andrea et al., 1993; Anelli et al, this volume).

On the other hand, hydrocarbons generated by source-rocks forming part of the under-thrusted sedimentary cover of the foreland may migrate vertically into the allochthon or into neo-autochthonous basins (Fig. 8). An example of such a "plumbing system" are the oil and gas accumulations in the Neogene fill of the Vienna Basin and in the underlying Austroalpine allochthon; these were charged hydrocarbons generated by autochthonous Late Jurassic basinal shales (Ladwein et al., 1991, Zimmer and Wessely, this volume). Similarly, the structures of the East-Taurus foothills thrust belt were charged by Silurian source-rocks of the autochthonous foreland sedimentary cover (Gilmour and Mäkel, this volume).

CONCLUSIONS

Opening of the different constituent oceanic basins of the Mesozoic Tethys was controlled by a sequence of rifting cycles, the kinematics of which changed with the progressive development of discrete plate boundaries first between Gondwana and Laurasia and later between Eurasia and Laurentia-Greenland. During these opening phases of Tethys and the Atlantic, a sinistral translation between Africa-Arabia and Europe and the interaction with intervening microcontinents, such as the Italo-Dinarides block, is held responsible for the development of first subduction zones and the onset of the Alpine orogenic cycle. With the Senonian onset of convergence of Africa-Arabia and Europe, new subduction zones developed and progressive closure of the different Tethys oceanic basins was followed by multiple and diachronous collisional events. However, the preservation of remnant Tethys oceanic basins in the Ionian Sea and the Eastern Mediterranean shows, that continent-to-continent collision has not yet occurred along the entire Alpine-Mediterranean suture of Africa-Arabia and Europe. Moreover, Neogene continental rifting, lateral block escape and subduction slab roll back has governed the opening of the oceanic Algero-Provençal and the partly oceanic Tyrrhenian basins.

The different fold-and-thrust belts of the Alpine orogenic system display highly variable architectures. These are partly controlled by the intensity of collisional coupling between the respective orogenic wedge and its cratonic foreland (and hinterland), and partly by the availability of a thick passive margin sedimentary wedge, or its absence. Imbricated passive margin wedges characterize, for instance, the external elements of the Apennines, the Dinarides-Albanides and the East-Taurides. In contrast, major nappes, derived from the southern margin of the South Penninic Ocean, directly override the undeformed foreland of the Eastern Alps which extends some 100 km beneath these nappes. In the Central and Western Alps, increasing collisional coupling between the foreland and the evolving orogenic wedge resulted in imbrication of the foreland crust, uplift of external crystalline massifs and deformation of the overlaying stack of nappes. In contrast, the Flysch Carpathians consist of an accretionary wedge which was thrusted over the essentially undeformed autochthonous foreland.

Most of the Alpine fold-and-thrust belts are associated with more or less pronounced flexural foreland basins; however, in some cases these were destroyed by late thrust propagation into the foreland. The Alpine foreland basins display major variations in their their architecture and the facies development of their sedimentary fill. An array of minor syn-flexural normal faults characterizes the

German and Austrian Molasse Basin (Roeder and Bachmann, Zimmer and Wessely, this volume). In contrast, the Carpathian foreland basins are characterized by a limited number of major syn-flexural faults. On the other hand, thrust propagation into the foreland plays an important role in the Apennine foredeep (Anelli et al., this volume). Whereas the Swiss Molasse Basin is characterized by shallow marine and continental clastics, lacking effective reservoir/seal pairs (Ziegler et al, this volume), the Molasse Basin of Austria contains Oligocene and early Miocene deeper water clastics and shales, grading upwards into deltaic series, which provide for several reservoir-seal pairs (Zimmer and Wessely, this volume). Similarly, the flysch-type sediments of the Apennine foreland basins, which grade up into a Pleistocene deltaic complexes, contain multiple effective reservoir/seal pairs. In contrast, the sedimentary fill of the East-Taurus foreland basin consists of marls, evaporites and carbonates (Gilmour and Mäkel, this volume).

Compressional intra-plate deformations characterize both the European and the Africa-Arabia platform. These developed in response to collisional coupling of these forelands with the Alpine orogen. Early collisional foreland deformations occur in the Carpathian and East-Alpine forelands, where they caused partial destruction of the passive margin sedimentary prism and inversion Mesozoic tensional basins and the upthrusting of basement blocks as far north as Denmark and southern Sweden. Late collisional foreland deformations occur, for instance, in the forelands of the Western and Central Alps and the Pyrenees. Examples of thin-skinned foreland fold-an-thrust belts are the Lombardian belt of the Southern Alps (Ziegler et al., this volume), the Jura Mountains of Switzerland and France (Philippe et al, this volume) and the Prerifain chains of Morocco (Zizi, this volume). Their development caused partial destruction of pre-existing syn-orogenic flexural basins. Basement involving up-thrusted blocks characterize the Bohemian Massif in the foreland of the Eastern Alps (Ziegler, 1990). Inversion of major Mesozoic grabens in Europe, in Iberia and on the African-Arabian Platform probably involved the entire crust and possible also the mantle lithosphere (Ziegler et al., 1995).

Eocene and later development of the Rhine-Rhône-Valencia and East Africa-Red Sea-Libyan-Pelagian Shelf rift systems caused disruption of the sedimentary cover of the European and Africa-Arabian platforms, particularly across major thermal domes. This had repercussions on the hydrodynamic conditions in remnant platform basins and their hydrocarbon habitat.

The petroleum systems of the external parts of the Alpine orogen, its foreland and internal basins are highly variable. They can rely either on pre-rift, syn-rift or syn-flexural source-rocks or a combination thereof. Some of these source-rocks attained maturity already during the passive margin or flexural basin stage; others reached maturity only during the emplacement of allochthonous units. Each basin has its own case history.

The petroleum exploration history of the Alpine orogen and its associated basins commenced in the early Nineteenth century. First hydrocarbon exploration efforts were based on surface seeps, concentrated in the shallow, frontal anticlines of the Outer Carpathians of the Ukraine and Romania. In the course of time, improved geophysical methods provided new tools for imaging sub-surface structures. This resulted in new discoveries of oil and biogenic gas in such foreland basins as the Po Plain, the Adria, the German and Austrian Molasse Basin and in the Aquitaine Basin, as well as in the Pannonian and Vienna neo-autochthonous basins. With the development of static corrections and the development of CDP-methods, and ultimately of 3D-seismic, imaging of even structurally complex areas has become possible, thus providing access to complex thrusted structures (Le Vot et al., this volume) and subthrust prospects of the Lopushnia (Carpathians) and Tempa Rossa type (Southern Apennines; Roure and Sassi, 1995; Anelli et al., this volume).

No doubt, future exploration efforts in the external parts of the Alpine orogen an beneath the Neogene fill of the Pannonian Basin will yield further oil and thermal gas discoveries, provided it can be established that efficient petroleum systems are available, have survived the most recent deformations and that only minor hydrocarbon re-migration has occurred.

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Tectonic evolution and paleogeography of Europe

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ABSTRACT

Multiple rifting and suturing events through Phanerozoic times amalgamated Europe as we know it today. Our detailed analysis of the crustal blocks, now forming of Europe, during the Caledonian, Hercynian and Alpine orogenies, allowed us to understand the influence of these events on the hydrocarbon systems of Europe.

To summarize this, we present a series of 11 palaeogeographic maps from Carboniferous to Pliocene times. These maps were produced as part of a project to develop basin-wide models for regional play element distribution in the major hydrocarbon-producing basins of Europe.

Description of the tectonic evolution of Europe can be divided into four main phases which are related to motions between Baltica, North American/Greenland and Gondwana. The first phase culminated in the assembly of Laurussia (Europe and North America/Greenland) during the Early Palaeozoic Caledonian Orogeny; it was followed by the Carboniferous assembly of Pangea (Laurussia and Gondwana) during the Hercynian orogeny. The third phase, involving rifting and separation of these blocks, started in Permian time. The fourth and final phase, that continues today, is the Alpine orogenic cycle which resulted from convergence of Africa and Europe.

INTRODUCTION

We present a series of palaeogeographic maps which summarizes our understanding of the geologic evolution of Europe since Carboniferous times. These maps were produced as part of an Exxon project to develop basin-wide models for regional play element distribution in the major hydrocarbon-producing basins of Europe.

The tectonic evolution of Europe can be divided into four main phases which are related to motions between Baltica, North American/Greenland and Gondwana. The first phase involved the formation of Laurussia (Europe and North America/Greenland) during the Early Palaeozoic Caledonian Orogeny. The second phase was the Carboniferous assembly of Pangea (Laurussia and Gondwana) during the Hercynian orogeny. The

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third phase was dominated by rifting and separation of these blocks, starting in Permian times. The fourth and final phase continues today and corresponds to the Alpine orogenic cycle which results from convergence of Africa and Europe.

For times younger than Jurassic, relative motions of cratonic blocks were determined from sea floor spreading data in the Atlantic. Motions between Europe and Africa were essentially strike slip from 180 to 110 Ma, then swung round to the convergent motion which continues today. Pre-Jurassic relative plate motions were determined from a combination of palaeomagnetic data and geologic data on the timing of tectonic events. An important final constraint was that the derived relative motions were required to produce a pattern that was geologically reasonable, i.e. no convergent and divergent rates that exceed rates known from global post-Jurassic plate motion rates, and relative motion directions that agreed with the tectonic data. This last constraint is particularly important in the Hercynian orogeny, which includes significant amounts of strike slip motions.

PALAEOZOIC CRUSTAL BLOCKS

Palaeozoic crustal blocks, as they were assembled in Permian times at the end the Hercynian orogeny, are shown in Plate 1. Brief descriptions of these blocks are given below.

Baltica

Baltica, also known as the Russian Platform, is the Precambrian core of Europe; it consists of several Archean age blocks that were amalgamated into cratonic Baltica before 1.6 Ga (Zonenshain et al., 1990). Baltica is bounded on the west by the Iapetus suture and on the east by the Ural suture. The northern edge of Baltica we take to be the Timan Belt and its extension along the northern coast of Scandinavia. This boundary was reactivated during the Late Cambrian Fenno-Scandian orogeny. The southern boundary is less well defined. In Early Palaeozoic time, Baltica faced the Tornquist Sea to the (present day) south. The Tornquist Line itself, however, does not mark the edge of Baltica; the edge is further outboard, buried beneath younger cover (Cocks and Fortey, 1982).

Pechora

This is the crust under the Barents Sea and includes Svalbard. Genesis of this area is poorly understood; we assume it to be amalgamated by the end of the Caledonian orogeny, but most of this block probably consists of older crust.

Laurentia

North America and Greenland are Precambrian cratons that make up the Laurentian block. Like Baltica, Laurentia consists of Archean terranes that were amalgamated during Precambrian times, culminating in the Grenville orogeny between 800 and 1000 Ma.

Avalonia

Avalonia consists of southern England, Ireland and the northeastern seaboard of North America. It rifted away from Gondwana during the Early Cambrian (550 Ma) and was sutured to Laurentia during the Caledonian orogeny (first collision at 425 Ma, end of orogeny at 405 Ma; McKerrow, 1988). There is not enough reliable palaeomagnetic data from Avalonia to determine its positions between rifting and collision. A motion path for it was determined by first plotting the positions of Europe, North America and Africa at the start and end of its motion, then interpolating positions between these times so that the motion of Avalonia was continuous.

Armorica

Armorica is the name for the western Iberian Peninsula and also for what is now western and northern France. Armorica is separated from the rest of France and Iberia (the Southern European Block) by a suture zone of Hercynian age. Structural studies within Armorica indicate that it was severely deformed during the Hercynian orogeny by its collision with the Southern European Block (Matte, 1986), which acted as a solid indentor, wrapping Armorica around itself. Maximum compression of Armorica was 500 km and the axis of maximum deformation forms a line of weakness along which the Bay of Biscay developed in Late Cretaceous times.

Southern Europe

Southern Europe consists of northeastern Iberia, the Balearics, southern France, Corsica and Sardinia and probably some of the Palaeozoic crustal elements involved in the Alps. Like Armorica and Avalonia, this block consists of Pan African affinity crust which rifted off Gondwana during the Early Palaeozoic.

Rhenohercynian

We interpret the Rhenohercynian zone as a zone of Caledonian accretion that was the locus of Devonian extension, as evidenced by the occurrence of bimodal volcanics of Devonian age.

Bohemian Massif

The Bohemian Massif consists of a Precambrian terrane which, according to deep seismic data, must be separated from the Brno-Malopolska Block (Suk et al., 1984).

Brno-Malopolska

The Brno-Malopolska Block is composed of Precambrian granites and metamorphic rocks. This unit includes the southern Holy Cross Mountains area (Malopolska Massif).

Moesia

This block is assumed to consist of some Precambrian crust that was accreted to southern Baltica during the Early Palaeozoic.

Tisza

This block includes crust with European affinity (Royden and Baldi, 1988). It rifted from Europe in Jurassic time, then joined Apulia in colliding with Europe during the Alpine orogeny. Tisza's crystalline and Mesozoic rocks outcrop only near its eastern and western terminations.

PALAEOGEOGRAPHIC MAP FORMAT

The palaeogeographic maps presented here were designed to show depositional environments using the following colours:

Dark brown:	highlands, considered to be sediment source areas.
Light brown:	lowlands, or zones of sediment bypass.
Green:	continental, fluvial and lacustrine.
Yellow:	coastal plain, deltaic to inner shelf.
Teal:	neritic to shelfal.
Light blue:	basin and slope
Blue:	abyssal sediments on either thinned continental or oceanic crust.

Pink and red colours distinguish collision- and extension-related igneous rocks. The only sedimentary lithology shown is a chevron pattern for evaporites. All active structures are indicated using standard symbols as shown on the map legends. Dashed lines show some political boundaries, present-day coastlines are in blue and some geographic zones are identified with letter codes. For orientation purposes, some cities are also shown. Maps are plotted on an Albers equal area projection (standard parallels 44° and 67°) with Europe in its present-day position. A 5° present-day latitude/longitude grid is included, as are palaeolatitude lines derived from a compilation of palaeomagnetic data.

PALAEOZOIC PALAEOGEOGRAPHY

Mid-Carboniferous (Namurian, 322 Ma)

The Mid-Carboniferous (Namurian, Serpukhovian) map, Plate 2, illustrates collision of Armorica and the South-European Block near the end of the Hercynian orogeny. This collisional and magmatic episode was largely ensialic, and resulted in emplacement of abundant synorogenic granites, shown in pink. Widespread orogenic deformations occurred across northern Europe and Iberia and in the Armorican, Saxothuringian, Bohemian, Silesian, Massif Central, Ligerian, Corsica-Sardinia and Carnic Alps areas.

Continental clastics were deposited in the Armorican and Saxothuringian basins; linear basins formed within the collision zone. Flysch was deposited in foredeeps on either side of the main orogenic belt. Principal flysch basins are the Cantabrian Basin in the south, and the Rhenish Basin in the north. The Rhenish Basin was eventually filled to capacity, and by Late Carboniferous time (next time slice) marine connections to this basin were severed.

Upper Carboniferous (Westphalian A/B, 306 Ma)

Hercynian deformation continued into the Late Carboniferous. This final phase of crustal shortening is sometimes referred to as the Variscan phase. Plate 3 shows palaeogeography for the Westphalian A/B stage of the Late Carboniferous. Widespread Variscan deformations consist of thrust- and wrench-faulting, folding, post-tectonic granite emplacement and the accumulation of thick continental sediments in the developing foredeeps. Marine connections to the North-European foredeep basin, located along the northern flank of the Hercynian orogenic belt, were cut off as it progressively filled with clastics. In this basin, thick coal measures were deposited during Westphalian times. These provide the source for most of the gas found in the Rotliegendes sandstones of the Southern Permian Basin. Sedimentation on the South Apulian shelf was locally disrupted by Variscan tectonic events.

Lower Permian (Rotliegendes, 254 Ma)

Lower Permian time was dominated by the collapse of the Hercynian mountain ranges and the deposition of thick clastic sequences in the area of their northern foreland basin. The Rotliegendes (Plate 4) clastic reservoir facies was the first sequence to be deposited. Sedimentation was accommodated partly by thermal subsidence and partly by continued subsidence of the relict Variscan foredeep. This basin is known as the Southern Permian Basin. Possible dextral shear between Gondwana and Europe created intracontinental transform systems creating local transtensional and pull-apart basins. Individually, these faults show relatively small displacements. Grabens along the faults filled with continental clastics. Marine shelf sedimentation continued in the Apulian area.

Upper Permian (Zechstein, 251 Ma)

In Upper Permian time (Plate 5) a marine connection was established between the Arctic shelves and the Northern and Southern Permian basins via the Arctic-North Atlantic rift system (Ziegler, 1988). In the extensive Zechstein inland sea, glacio-eustatic cycles controlled the accumulation of alternating carbonate and evaporite deposits. Late Permian fauna suggest communication between the Boreal Zechstein seas and the Tethys seas via Dobrudgea. Apart from providing halokinetically induced structural traps, the Zechstein evaporites are an important seal facies, which seals the Rotliegendes sands. Further to the south, on the Apulian platform, rifting created basins containing both evaporitic and continental deposits.

MESOZOIC CRUSTAL BLOCKS

Crustal blocks involved in the Mesozoic and Cenozoic Alpine-Carpathian deformation are shown on a Present-Day base map in Plate 6. Europe and Africa remained relatively stable through this time, although older structural grain was reactivated during the Alpine orogeny. Europe and Africa are relatively stable plates to which different crustal blocks were amalgamated through Mesozoic and Cenozoic times. Amalgamation is still going on today with active subduction at the Hellenic trench.

Iberia

The Iberian Block behaved independently of Europe and Africa during the Late Cretaceous and Paleogene (Choukroune et al., 1989; Roure et al., 1989). The cratonic core of the Iberian peninsula consists of Precambrian and Palaeozoic rocks which have undergone Hercynian structuring during the Late Palaeozoic (Pin, 1990; Franke and Engel, 1986; Ziegler, 1988, 1990).

Apulia

Apulia played a key role in the Alpine-Carpathian deformations (Channell et al., 1979; Biju-Duval et al., 1977). The Apulian Block extends from Italy through the Pannonian area, the Adriatic, Greece and further east to Turkey. There is no known pre-Hercynian crust in Apulia. It probably formed as accretionary crust during the Hercynian collision of Gondwana with Europe. Apulian crust was extensively affected by subsequent Mesozoic rifting, until Jurassic separation from Africa eventually led to formation of the Eastern Mediterranean oceanic crust. In Tertiary times, Apulia collided with Europe, initiating the Alpine-Carpathian orogen. During this event, the Apulian Block was shortened and partly subducted. East-dipping subduction formed the Dinaric belt along the eastern margin of Apulia, and west-dipping subduction formed the Apennine belt along its western margin (Royden, 1993; Doglioni, 1992; Casero et al., 1990; Moretti and Royden, 1988; Frasheri et al., Anelli et al., this volume).

Tisza and Moesia

Tisza and Moesia were discussed above in the section on Palaeozoic blocks.

Sakarya

Sakarya is a Cimmerian block which rifted off the northern margin of Gondwana during Late Permian to Triassic times and collided with Europe in Late Triassic to Liassic times (Sengor, 1984; Sengor et al., 1984).

Rhodope

Rhodope is part of Europe. It rifted off Europe but never strayed very far before colliding with the European margin during the Meso-Alpine orogeny (Dixon and Dimitriadis, 1984; Sengor, 1984).

MESOZOIC PALAEOGEOGRAPHY

Upper Triassic (Rhaetian, 210 Ma)

Plate 7 shows Upper Triassic palaeogeography. Triassic times were characterized by regional extension with multidirectional systems of grabens being superimposed on Hercynian structural trends. This extension, known as the Tethyan rift event, had a profound influence on hydrocarbon plays in the Apulian and Carpathian regions. Shallow water reefal limestones characterize rifted margins throughout Apulia. Rifting was accompanied by a widespread extrusive and shallow intrusive volcanism, (Dietrich, 1979; Spray et al., 1984), although individual outcrops are not large enough to be shown on Plate 7. Large areas of northern Europe were affected by Triassic extension, controlling the subsidence of many basins (Plate 7). Generally low Triassic sea levels and narrow rift basins led to highly restricted water bodies and the deposition evaporites (e.g. Muschelkalk, Keuper of northern Europe). During the Triassic, numerous marine connections were established between Tethys and the basins of northern Europe, which, since the Lower Triassic, were separated from the Arctic Seas.

On the Apulian block, a distinctive platform-basin palaeogeography was established by Triassic rifting. Grabens were filled with terrigenous sediments, grading upward into pelagic, cherty carbonates. Platforms localized shallow water carbonates (including reefs) and evaporites. In northern Italy, Middle Triassic to Carnian volcanics and massive reefs outcrop in the Dolomites. In the Adriatic and Central Apennine areas, shallow water evaporites (e.g. Burano formation) were deposited, while deep-water clastics (e.g. Riva di Solto formation) characterized the Northern Apennines and Po Basins. Both the Burano and Riva di Solto formations include source-rock facies (Anelli et al., this volume). During Late Triassic times, possible initial opening of the Vardar ocean occurred. This marked the first formation of post-Hercynian oceanic crust in the study area (Spray et al., 1984; Dietrich, 1979).

In northern Apulia, distinct transverse zones, following Hercynian trends, were established. These zones were manifested as a series of flexures which delimited domains of platform/basin geometry.

During Late Triassic times, long-standing south-directed subduction of the Palaeo-Tethys ocean along the northern margin of the Cimmeria blocks terminated with their collision with Europe (Sengor, 1984; Sengor et al., 1984). In Europe, Cimmerian deformations of Late Triassic and Early Jurassic age is documented by unconformities in the Polish Trough, the Northern Po Basin, and in the Italian Dolomites.

Along the African margin, Ladinian to Carnian age extensional tectonics created half-grabens and pull-apart basins in the High Atlas trough. This trend followed an aborted Late Carboniferous to Permian rift (Cousminer and Manspeizer, 1977). During the Late Triassic and Early Jurassic, the newly created grabens were filled with continental and evaporitic sediments. Further west in Morocco, grabens were associated with continental clastics and alkaline volcanism (Wildi, 1983).

Early Jurassic (Toarcian, 179 Ma)

During Early Jurassic times the Tethyan rift system remained active (Biju-Duval et al., 1977; Dewey et al., 1973; Dercourt et al., 1986; Ziegler, 1988, 1990). Plate 8 shows palaeogeography for the Toarcian stage of the Early Jurassic. This was also the time of initiation of opening of the Central Atlantic between North America and Africa; with this a sinistral strike-slip regime was established between Africa and Europe. At this time marine connections were reopened between the Arctic Seas and the Tethys Ocean via the Arctic-North Atlantic rift (Ziegler, 1988). Continued tectonic subsidence of the Tethyan and European rift systems, combined with a eustatic sea level rise, open oceanic circulation patterns and low palaeolatitudes favoured deposition of widespread carbonate platforms, especially on Apulia. In the Briançonnais area (#4 on Plate 8), rifting caused foundering of the older Middle to Upper Triassic platforms, on which shallow water carbonates had been deposited (Rudkiewicz, 1988; Michard and Henry, 1988).

In the Helvetic realm, sedimentation was dominated by carbonates grading into marly sequences (Funk et al., 1987). In Lias and Dogger times, sedimentation consisted of mainly shales (Dauphinois facies). Rapid horizontal facies changes and wide stratigraphic gaps characterize this facies (Masson et al., 1980). Rifting in the Eastern Mediterranean resulted in formation of oceanic crust in the Antalya area (# 2 on Plate 8; Robertson and Dixon, 1984; Yilmaz, 1984).

Along the Cimmerian collision zone (#1 on Plate 8), flysch deposition occurred in Dobrudgea, Crimea and Northern Turkey (Sengor, 1984). Cimmerian orogenic activity terminated in Early Jurassic times, as evidenced by intrusion of Middle Jurassic plutons on both sides of the suture. Effects of this event were also felt in the Polish Trough, and local inversion occurred in the Donets Trough.

The Iberian Meseta was an important source of clastics in the Cantabrian Basin, Lisbon and Cavalla basins and also the Duero Basin (Wildi, 1983; Ziegler, 1988). Scattered extensional magmatism occurred on the Iberian Meseta and also in the Pyrenees area.

Along the African margin, tilting and foundering of fault-blocks occurred mainly during the Sinemurian and thus is slightly older than the time represented on Plate 8 (Favre and Stampfli, 1991).

Middle Jurassic (Bathonian, 158.5 Ma)

The Bathonian map (Plate 9) shows the onset of sea floor spreading in the Central Atlantic. This spreading system continued to the north between Iberia and Africa and then bifurcated into two branches, one between Apulia and Europe and one between Apulian and Africa. Continuation of the northern branch past the Tisza and the Pienniny area into the Black Sea is speculative; this is based on occurrence of rifting in the Magura Trough and

Mecsek Zone and provides a mechanism for the formation of the Transylvanian ophiolites.

In the Helvetic domain, sedimentation patterns indicate progressive starvation and deepening of the basin: Middle Jurassic manganese oozes are succeeded by Late Jurassic radiolarites, Tithonian Calpionellid oozes, slump deposits and transported turbiditic calcarenites. In the Tethys area, a global high-stand led to widespread carbonate platform development. Pelagic sediments were deposited in troughs while shallow water carbonate sedimentation appears to have kept pace with subsidence on the platforms, continuing the distinctive basin/platform topography of Apulia. Carbonate platforms in the Dinaric and Friuli areas supplied large amounts of debris into the Belluno Basin (Massari et al., 1983). Apulian carbonate platforms reached their maximum extent in Cretaceous time.

Late Jurassic-Earliest Cretaceous

We do not have palaeogeographic maps covering Late Jurassic and earliest Cretaceous times. Some of the major tectonic events are briefly summarized here.

Late Jurassic to Early Cretaceous tectonic development of southern Europe was primarily controlled by opening of the Central Atlantic which established a regional sinistral shear between Africa and Europe. Apulia rotated in a counter-clockwise direction, opening the Mediterranean until Mid-Cretaceous times, when it became fixed to Africa (Dewey et al., 1973; Bernoulli and Lemoine, 1980; Biju-Duval et al., 1977; Channell et al., 1979). Several ophiolitic suites (e.g. Liguride, Piedmont, Transylvanian and Vardar) were formed by these spreading events.

A Late Jurassic subduction zone with calc-alkaline volcanism and flysch sedimentation formed along the Dinaric shelf margin of Apulia (Frasheri et al., this volume). Faunal evidence from the Mecsek and Villany-Bihor areas on the Tisza platform and the Briançonnais zone of the Alps (Roux et al., 1988) indicates that these areas were linked until Tithonian times and but were separated afterwards. This indicates decoupling of the Tisza block in Tithonian time. We suggest that Tisza separated from Europe in Tithonian times and subsequently became attached to Apulia before again colliding with Europe in Eocene times.

Lower Cretaceous (Aptian, 112 Ma)

By Mid-Cretaceous times, Atlantic sea floor spreading had propagated northward between Iberia and North America and Iberia started to separate from Europe (Plate 10). Mediterranean sea floor spreading was nearly complete. Motions between Africa and Europe, which were sinistral from Jurassic through Early Cretaceous times, changed in Mid-Cretaceous times to progressively more convergent, with the convergence direction becoming almost normal to the European margin by Eocene times as the Arctic-North Atlantic opened. This convergence established the Alpine orogeny as well as several other, more localized compressional events. One of these was in the Rhodope area, with compressional deformation and flysch deposition along the southern margin of the Moesian Platform. Compression also occurred along the eastern margin of Golija and on the Pelagonian Platform. During Early Cretaceous times, Pelagonia collided with Rhodope and emplacement of nappes took place in the Hellenides and Dinarides.

In Iberia and the Aquitaine Basin, block tilting associated with extensional tectonics as well as halokinetic movements of Triassic evaporites took place during Early Cretaceous times (Le Vot et al., this volume). In the Western Alps, collision started during Cenomanian-Early Senonian time due to subduction of the European margin south or southeast beneath the Apulian margin; this is evidenced by blueschists and eclogites (Debelmas, 1989). The suture is seen in the Canavese slices (schistes lustres) and Sesia zone. High pressure metamorphism associated with the suturing event has been dated at 130 Ma and also 100-80 Ma (Debelmas, 1989). Upper Jurassic -Lower Cretaceous ophiolite-bearing, highly deformed nappes are overlain by Upper Cretaceous relatively undeformed flysch nappes, indicating the beginning of European-African compression in the Albian. Intra-Apulian deformation was localized along pre-existing lines of weakness, principally Permo-Triassic grabens. An east-dipping subduction zone initiated in front of the Golija and Pelagonian platforms. During the Late Cretaceous, major tectonic movements affecting the Austro-Alpine domain, the internal Dinarides.and the Southern Alps are expressed by a transition from flysch sedimentation in the Lombardian and Julian-Slovenian basins to a pelagic setting in the Belluno basin and Trento platform (Massari et al., 1983).

In the Apuseni mountains, Albian thrusting and folding, along with flysch sedimentation, indicates that the Tisza block collided with or was close to Apulia by Middle Cretaceous times (Burchfiel, 1980). By Late Cretaceous times, large strike-slip movements dominate the Tisza block and North Pannonian part of the Apulian block as the two blocks impinged on the Carpathian embayment. This deformation is also expressed in the Eastern Alps by lateral extrusion structures (Ratschbacher et al., 1991). Further to the east, a north-dipping subduction zone, characterized by magmatic activity and back-arc rifting, was established in the present Black Sea (Gorur, 1989).

Counter-clockwise rotation of Iberia relative to Europe resulted in sinistral shear along the North Pyrenean fault zone (Galdeano et al., 1989; Choukroune et al., 1989; Roure et al., 1989). Oceanic crust developed in the Bay of Biscay following early Aptian separation between Galicia Bank and Flemish Cap (Dewey et al., 1973; Ziegler, 1988, 1990). Rifting movements decreased considerably in the East-Iberian Basin during Cretaceous time, with continental to deltaic sandstone deposition. In addition to the Pyrenean area, shear deformation took place in the Cantabrian Mountains of northern Spain and in the Celt-Iberian Range in central Spain. The nature of this shear was predominantly transtensional and was manifested in the form of rifts and pull-apart basins. Post-rift deposition began during late Aptian-Albian time. A thick sequence of shallow water, interbedded clastics and carbonates accumulated on the subsiding Atlantic margin.

The Late Cretaceous was characterized by the same opposed evolution of a shallow carbonate platform in the eastern Iberides and a deep, terrigenous flysch basin in the western Pyrenees. During Mesozoic times, both on the platforms and in the basins, the depositional sequence organization was closely linked to eustatic sea level changes. Local extensional processes generated important modifications in thickness, particularly within the Lower Triassic, Liassic, Kimmeridgian and Aptian series, as well as within the Pyrenean Mid- and Upper Cretaceous deposits. By the early Campanian, sea floor spreading ceased in the Bay of Biscay and Iberia began to converge with Europe. In the eastern Pyrenees, the main deformation was Santonian and Campanian. The Pyrenean collision front propagated westward during late Senonian to Palaeocene time.

CENOZOIC PALAEOGEOGRAPHY

Lower Oligocene (Rupelian, 33.5 Ma)

Plate 11 shows palaeogeography for lower Oligocene times. This was a period of intense tectonic activity following the collision of the Apulian and European blocks. This collision was the result of continued convergence between Africa and Europe. The Apulia-Europe collision was diachronous, starting north of the present-day Adriatic and propagating eastward into the Carpathians and westward toward the Western Alps. Extensive deformation, metamorphism, plutonic activity and deposition of thick flysch and molasse sequences occurred along the entire deformation front.

In the Alps, initial deformation occurred in the Piemont, Briançonnais and Valais zones with later involvement of the Ultrahelvetic and locally the Helvetic domains. Development of the Molasse Basin accompanied this deformation phase (Ziegler et al., this volume). Northward transgression of the Tethys sea led to a progressive onlap of Cenozoic rocks onto the basal Tertiary unconformity (Roeder and Bachmann, this volume). Local positive features in the foreland persisted until Oligocene time when the basin deepened rapidly (Bachmann et al, 1987). Rising sea levels and local restricted circulation provided ideal conditions for the accumulation of very rich source-rocks (e.g. Fish Shales formation in the Molasse Basin).

Alpine deformation was coeval with orogenic activity along the Apennine and Dinaric fronts. Initial deformation of the palaeo-Apennine chain started during Oligocene times with thrusting of the Liguride oceanic and flysch units (#2 on Plate 11). Deformation and flysch sedimentation were controlled by the palaeotectonic framework inherited from Mesozoic tectonics. The shape and interrelations of different basins varied as the structural framework became better defined. The Apennine depocenters shifted from south to north as different structural units deformed. The amount of deformation increases from the north to the south (Bally et al, 1986). This areal distribution of deformation controls trap size and trap integrity (Anelli et al., this volume).

During Eo-Oligocene time, the Apulian promontory pushed the North Pannonian and Tisza blocks into the Pannonian embayment (Plate 11). Together they "escaped" into this embayment, which existed as a gap between the buttresses of the Bohemian Massif to the northwest and the Moesian platform to the southeast. In this process, the North Pannonian and Tisza blocks pushed the flysch, which had been deposited in front of the inner, crystalline part of the Carpathians, over the Carpathian foreland (#1 on Plate 11). The flysch was folded and thrusted, accommodating at least 200 km of shortening. Loading of the foreland and a coincident high stand in sea level provided ideal conditions for the accumulation of widespread, very rich Oligocene source-rocks in the region of the Carpathian fold-and-thrust belt (e.g. Dysodilic and Menilitic shales; see Bessereau et al., Dicea, this volume).

Compressional deformation of the Alps and their foreland was contemporaneous with the evolution of the intracratonic Rhine, Bresse, and Rhône rifts (Ziegler, 1987, 1990; Ziegler and Roure, this volume). A pulse of volcanism may have triggered extension in the Rhine Graben during the late Eocene (#4 on Plate 11). By late Eocene-early Oligocene times, a marine connection was established between the Alpine foredeep and the North Sea via the Rhine Graben in which the Fish Shales formation was deposited, the source-rock for most of the hydrocarbon accumulations in this graben.

Continued convergence of Iberia with Europe caused westward propagation of the Pyrenean deformation front into the Cantabrian region. During the Eocene main Pyrenean deformation phase Iberia was sutured to Europe (Roure et al., 1989; Chouckroune et al., 1989; Ziegler, 1988, 1990).

Tethys ocean crust subduction continued in the Alboran-Corsica/Sardinia region (#5 on Plate 11). Back-arc rifting behind this northwest-dipping subduction zone separated the Balearic Islands, Kabyl, Alboran and Corsica/Sardinia blocks from Europe (Torné et al., Vially and Trémolières, this volume).

Middle Miocene (Serravallian, 10.5 Ma)

Miocene Alpine deformation strongly affected several parts of the orogenic belt (e.g. Central, Western and Southern Alps, Carpathians, Apennines, Plate 12). In the Molasse Basin, with falling sea levels and increasing influx of clastics from the rapidly advancing Alpine orogenic front, the basin shallowed and a thick continental molasse section was deposited (Roeder and Bachmann, this volume).

During Middle Miocene time, back-thrusting of the Southern Alps accompanied deformation along the Dolomites (Doglioni, 1991, 1992; Ziegler et al., this volume). These thrusts involved the Apulian Mesozoic platform and raised the northern Po area (#3 on Plate 12). In the southern Po area, local emergence and evaporite deposition (e.g. Gessosso Solfifera formation) occurred. Active west-dipping subduction in the Apennine region and development of an east-facing foredeep took place, recorded by the Macigno flysch (Anelli et al., this volume). Flysch sedimentation from both the Apennine and Albanian accretionary complexes created the Po/Adriatic basin as a bivergent foredeep area. The main thrusting event in the Apennines occurred during the Miocene. These thrusts utilized Triassic evaporite layers as detachment surfaces (D'Argenio et al., 1980, Boccaletti and Coli, 1982). Tortonian continental and lacustrine sediments, unconformably overlying flysch facies, record this event, thereby constraining the timing of the end of the main deformation phase. On the east side of the Adriatic, subduction of the relict Tethyan ocean continued along the Albanian and Hellenic subduction boundary (#2 on Plate 12).

The Carpathian foredeep expanded over the European margin as the deformation front migrated to the East- and South-Carpathians (#4 on Plate 12). Rapid advance of the thrust front caused the Oligocene source section to be uplifted and maturation of the rich source facies terminated. Therefore, over large areas, Oligocene source-rocks are only mature where they have been structurally buried in the thrust belt or buried by sediments in the very proximal parts of the foredeep (Bessereau et al., Ziegler and Roure, this volume). Volcanism around the Carpathian arc, related to this phase of deformation, began during the late Oligocene and is thought to be related to the subduction of highly attenuated European continental crust beneath the overriding Carpathians. Back-arc extension began in the Pannonian basin during this time. By the end of the Miocene, deformation had stopped in all but the Romanian portion of the Carpathians (#5 on Plate 12).

By early Miocene time, the marine connections through the Rhine, Leine, and Eger grabens were severed as a result of uplift of the Rhenish Massif and Massif Central. Tectonism in these grabens, in which as much as 3 km of continental sediments were deposited, continued, as shown by volcanic activity.

Corsica/Sardinia and the Kabyl blocks were separated from Europe during the early Miocene (23 to 19 Ma; #1 on Plate 12; Vially and Trémolières, this volume). This motion was a primary driver for deformation of the Apennines. Fault blocks formed by rifting in the Valencia Trough contain the largest oil play found to date in Spain (Torné et al., this volume).

The Late Miocene was characterized by compression along the southeastern margin of Iberia in the Prebetic fold belt and in portions of the Balearic Islands. Westward escape of the Alboran Block opened the North Algerian Basin in its wake. The Alboran Block collided with the southeastern margin of Iberia and the northwestern margin of Africa during the late Oligocene/early Miocene. Extensive flysch basins mark this Betic/Rif deformation front (the Numidian flysch of Wildi, 1983; Ziegler, 1988). The Kabyl block escaped southwards and collided with North Africa to form the Tellian mountains of Algeria and Tunisia (Wildi, 1983).

During the late Tortonian, Calabria rifted from the Corsica and Sardinia block, creating the Tyrrhenian Sea as a back-arc rift basin.

Lower Pliocene (3.8 Ma)

During Pliocene time, development of ocean crust in the Tyrrhenian Sea was accompanied by extensive magmatism (Channell and Mareschal, 1989, #1 on Plate 13). This was synchronous with Pliocene Apennine nappe emplacement. Shortening in the Apennines (75 km in the north, 150 km in the south; Bally et al, 1986) was balanced by extension in the Tyrrhenian Sea (Doglioni, 1991). Extension initiated in late Tortonian time in the Apennine hinterland, creating small rift basins filled with Neogene clastics, sitting piggyback-style on the thrust belt (Boccaletti and Coli, 1982). The Apennine foredeep shifted northwards as the Liguride thrust belt was reactivated during the Pliocene. The Pliocene section, 9 km thick, consists of shallow water fluviatile sediments (#4 on Plate 13). Tertiary biogenic gas plays are found in this foredeep (Anelli et al., this volume). The dramatic subsidence in this foredeep can not be explained alone by the topographic load of the Apennine thrust sheets (Royden and Karner, 1984; Royden, 1993).

In the Alps, the most significant deformation is in the Helvetic domain. The deformation front continued to progress to the north, leading to late Miocene and early Pliocene folding and thrusting of the Jura Mountains. Peak deformation was during the latest Pliocene. The thrust decollement in the Jura is located in the Triassic evaporite section which continues to the south under the Molasse Basin and eventually under the Alps. The Molasse Basin was carried passively to the north (as a "piggy-back" basin) during this part of its history (Philippe et al., Ziegler et al., this volume). By the end of Tertiary times an approximately 5 km thick section of synorogenic clastics had accumulated in this basin.

Deformation continued in the outermost East- and South-Carpathians. Extremely rapid subsidence in the East-Carpathian foredeep occurred during Mio-Pliocene time with accumulation of approximately 9 km of coarse clastic molasse sediments. The East-Carpathian foredeep is intensely deformed to the west where it is partially overridden by thrusts of the flysch zone. To the north, this inner zone, which consists mainly of the lower molasse, is deformed by thrusts and folds. Salt appears to act as a detachment surface but is also involved in folding as salt diapirs pierce some of the folds and salt is locally squeezed up along thrust faults. This section contains the giant fields of the Ploesti district (in the area of SCF on Plate 13; Dicea, this volume).

The dramatic subsidence in the East-Carpathian foredeep can not be explained by the modest topographic load of the Carpathian Mountains (Royden, 1993, Doglioni, 1992). Royden (1993) suggests that the Carpathians are an example of a retreating subduction boundary where overall plate convergence is less than the rate of subduction. The deficiency in plate convergence was compensated by extension in the Pannonian back-arc basin. Regional extension continues today as seen in high grade metamorphic rocks of mid-crustal origin which are exposed in Rechnitz window and in Hungary (Tari, 1991).

The Carpathians are tectonically active today with activity largely restricted to the area of the Carpathian bend, also known as the Vrancea seismic zone. Large earthquakes with very deep hypocenters and compressional to strike-slip fault plane solutions typify this area; these may be produced by the leading edge of the (detached) downgoing slab under the Carpathian arc.

CONCLUSIONS

In this study, we utilized the plate tectonic history of Europe to constrain the understanding of sedimentary basin development and the effects of regional scale tectonic events on play elements for major basins. The tectonic framework and palaeogeography were used as constraints on models for basin formation, climate distribution and accommodation space which, in turn, control the distribution of reservoirs, source-rocks, seals and traps.

The structural and stratigraphic framework of Europe is the result of its Phanerozoic tectonic history, involving the amalgamation of crustal blocks. Multiple rifting and collision events created extremely complex mountain systems during the Caledonian, Hercynian, Cimmerian and Alpine orogenies. Basins are diverse, superimposed, have long-lived tectonic histories with complex structuring, and have highly variable play elements. The Hercynian orogen provides the framework for North European hydrocarbon systems. Its collapse sets up the Apulian Mesozoic hydrocarbon system. Alpine deformation and tectonically related extension, in turn, set up the Neogene hydrocarbon systems of the Carpathians, Pannonian Basin and the Apennines (Ziegler and Roure, this volume)

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Enclosures

Enclosure 1 Paleozoic crustal blocks on Permian base map

Enclosure 2 Mid-Carboniferous Namurian (322 Ma) paleogeography

Enclosure 3 Upper Carboniferous Westphalian A/B (306 Ma) paleogeography

Enclosure 4 Lower Permian Rotliegendes (254 Ma) paleogeography

Enclosure 5 Upper Permian Zechstein (251 Ma) paleogeography

Enclosure 6 Mesozoic crustal blocks on present-day base map

Enclosure 7 Upper Triassic Rhaetian (210 Ma) paleogeography

Enclosure 8 Lower Jurassic Toarcian (179 Ma) paleogeography

Enclosure 9 Middle Jurassic Bathonian (158.5 Ma) paleogeography

Enclosure 10 Lower Cretaceous Aptian (122 Ma) paleogeography

Enclosure 11 Lower Oligocene Rupelian (33.5 Ma) paleogeography

Enclosure 12 Middle Miocene Serravalian (10.5 Ma) paleogeography

Enclosure 13 Lower Pliocene (3.8 Ma) paleogeography

Accretion and extensional collapse of the external Western Rif (Northern Morocco)

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ABSTRACT

The frontal part of the Gibraltar Arc consists of allochthonous tectono-sedimentary complexes classically interpreted as gravity driven units or "melange". High quality seismic data along the northwestern Moroccan Atlantic margin and the Rharb Basin, as well as field data in the Western Rif provide a new view of this complex region. The overall type of deformation suggests an accretionary prism involving deep-water sediments, that was emplaced on the attenuated passive margins of Iberia and Africa in response to westward motion of the Alborán domain during the Miocene. The timing of deformation and the age of the sediments involved suggest an accretionary progression towards the external portion of the Arc.

Late Miocene and Pliocene extensional collapse of the unestable accretionary prism controls the structure of the region. The geometry of the extensional system present in the frontal accretionary wedge of the Rif Cordillera is very similar to the one of the Gulf of Mexico.

Data presented here suggests that some units classically interpreted as thrust sheets are in fact mixed extensional-compressional "satellite" basins.

1 REGIONAL SETTING

The Western Mediterranean region consists of a collage of several blocks located between the Euro-Asiatic and the African plates. These intermediate blocks or micro-plates (i.e. Iberia, Alborán, Corsica-Sardinia and Apulia) interacted with each other and with Eurasia and Africa, defining the geodynamics of the region (Dercourt et al., 1986; Andrieux et al., 1989; Dewey et al., 1989; Ziegler, 1987; Favre and Stamfli, 1992).

The Gibraltar Arc is the western limit of the Alpine-Mediterranean system. The Betic Cordillera in southern Spain and the Rif Cordillera in northern Morocco constitute the northern and southern part of the Arc (Fig. 1). The geological units of the Betic and Rif Cordilleras can be subdivided, as most of the Alpine orogens, into an External and an Internal domain.

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This article includes 2 enclosures on a folded sheet.

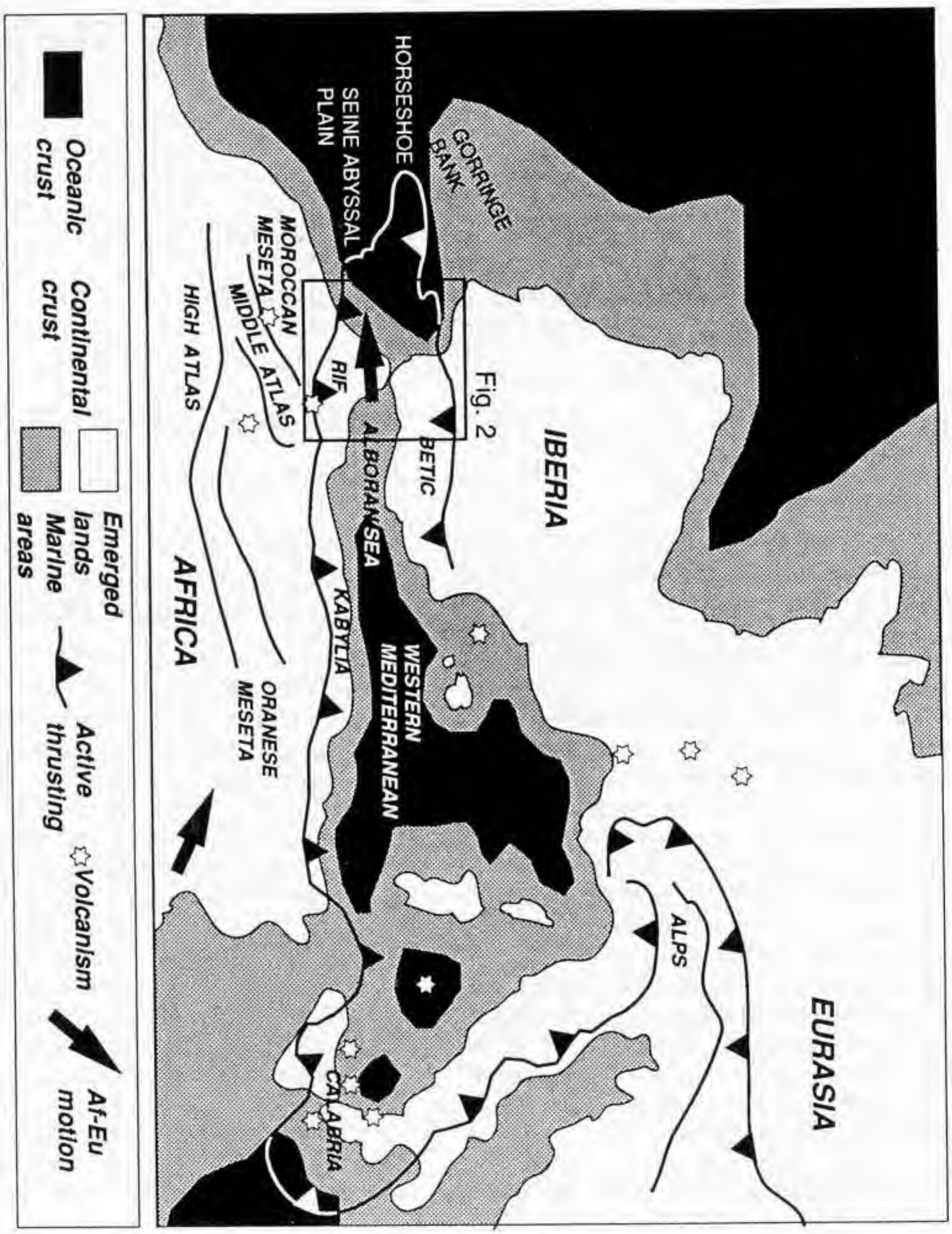

FIG. 1. Location of the study area within the Western Mediterranean Alpine system. Present-day structural sketch, modified after Dercourt et al. (1986) and Dewey et al. (1989).

Classically, the **External domain** of an orogenic belt is represented by non-metamorphic sedimentary successions, characterized by thin-skinned tectonics. The **External domain** of the Betic and Rif Cordilleras is separated into a number of structural units or thrust sheets (García-Hernández et al., 1980; Vera, 1983; Wildi, 1983). Some thrust sheets consist primarily of platform carbonates and other sheets consist mostly of siliciclastic sediments. The different types exhibit widely different styles of decollement tectonics. The **External domain** of the Betic and Rif Cordillera represents, respectively, the south-Iberian and north-African passive margin successions incorporated into the folded belt (Michard, 1976; Vera, 1981; Wildi, 1983; Martín-Algarra, 1987).

The **Internal** or **Alborán domain**, which includes the Internal zones of the Betic and Rif Cordilleras (Fig. 2), differs stratigraphically and structurally from the External zones (Suter, 1965; Fontboté, 1983). The most notable characteristics that distinguish the Internal from the External zones include: Alpine-type Triassic carbonates, Early Alpine (Cretaceous-Paleogene) polyphase compressional deformation and HP/LT metamorphism (Fontboté, 1983; Wildi, 1983; Galindo-Zaldivar et al., 1989; De Jong, 1991). The lowermost unit, the Nevado-Filabrides, is only present in the Betic Cordillera (Fig. 3a). The intermediate Alpujarrides-Sebtides unit contains metamorphic rocks and mantle peridotites (Beni-Bousera and Ronda ultramafics). The upper unit (Ghomarides-Malaguides) overlies both the Sebtides and the Dorsale unit.

Neogene extension associated with the collapse of the Alborán Sea and late inversion and transpressional tectonics severely modified the overall compressional development of the Gibraltar Arc (García-Dueñas et al., 1992; Flinch, 1993). Equivalent structural units of the Betics and the Rif are presently separated by the Alborán Sea. Extensional delamination accounts for dramatical thinning of the Internal domain units, resulting in stratigraphic omission (García-Dueñas and Martínez-Martínez, 1989; García-Dueñas et al., 1992). The structure of the extended Internal domain is similar to core-complexes of the Basin and Range province of the Western United States (Galindo-Zaldívar et al., 1989).

Two generalized regional geological cross-sections were constructed to compare the structure of the Betic and the Rif Cordilleras (Fig. 3). These sections are in part based on reflection seismic data (Blankenship, 1992; Flinch, 1993), well-logs (IGME 1987) and surface geological data (Suter, 1980a, 1980b; Blankenship, 1992; García-Dueñas et al., 1992; Flinch, 1993). They provide a broad approximation of the Gibraltar Arc structure and help to define the main structural problems. At a first look both cross-sections show great similarities. The Rif and the Betic Cordilleras are both characterized by foreland-vergent thin-skinned piggy-back thrusting involving the passive margin succession of the north-African or south-Iberian domains and a main Triassic decollement. Extensional structures affect the Internal domain of both fold and thrust belts, often reactivating previous thrust sheets (negative inversion). The Frontal tectono-sedimentary unit, referred to as Guadalquivir Allochthon in the Betic Cordillera and the Prerifaine Nappe in the Rif, is equivalent to both sides of the Gibraltar Straits.

The main difference between these fold belts is that the Betic is a carbonate dominated margin, while the Rif is a detritic dominated margin. The principal problems of the area concern the role of extension related to the opening of the Alborán Sea, the role of strike-slip faults, the way per which the shortening in the lower part of the passive margin succession is accomodated, the area of origin of the frontal allochthonous tectono-sedimentary complexes and the sequence of thrust emplacement.

2 THE ACCRETIONARY ZONE

The frontal units of the Betic and Rif Cordilleras (i.e. Frontal tectono-sedimentary Complex) consist of highly deformed Triassic, Cretaceous, Paleogene and Neogene strata, which were detached from their original base and thrust over the Mesozoic to Lower Miocene of the foreland. This unit is known as the "Guadalquivir allochthonous units" in the Betic Cordillera which is equiva-

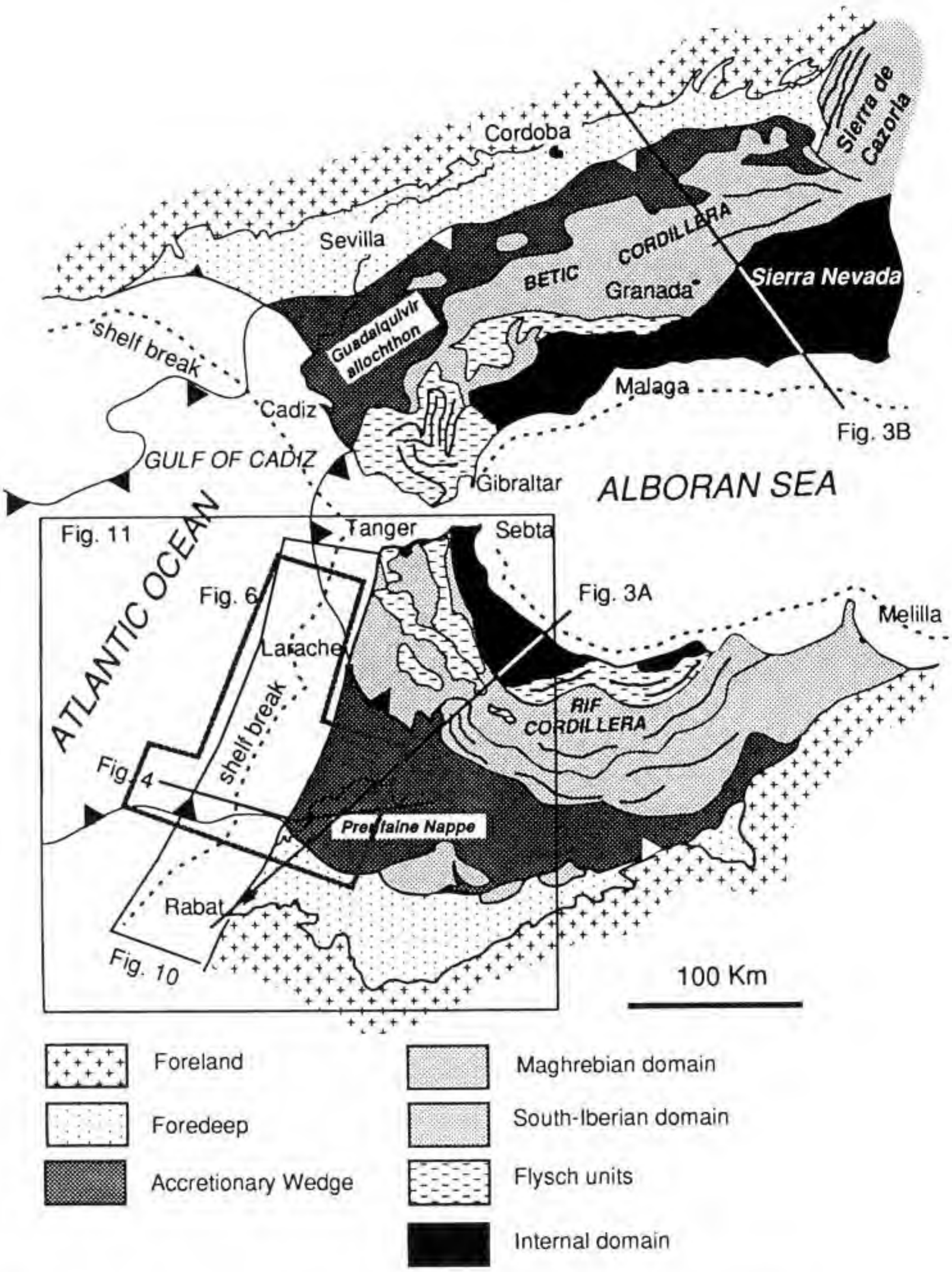

FIG. 2. Tectonic map of the Gibraltar Arc, after Flinch (1993). Location of the study area and following figures.

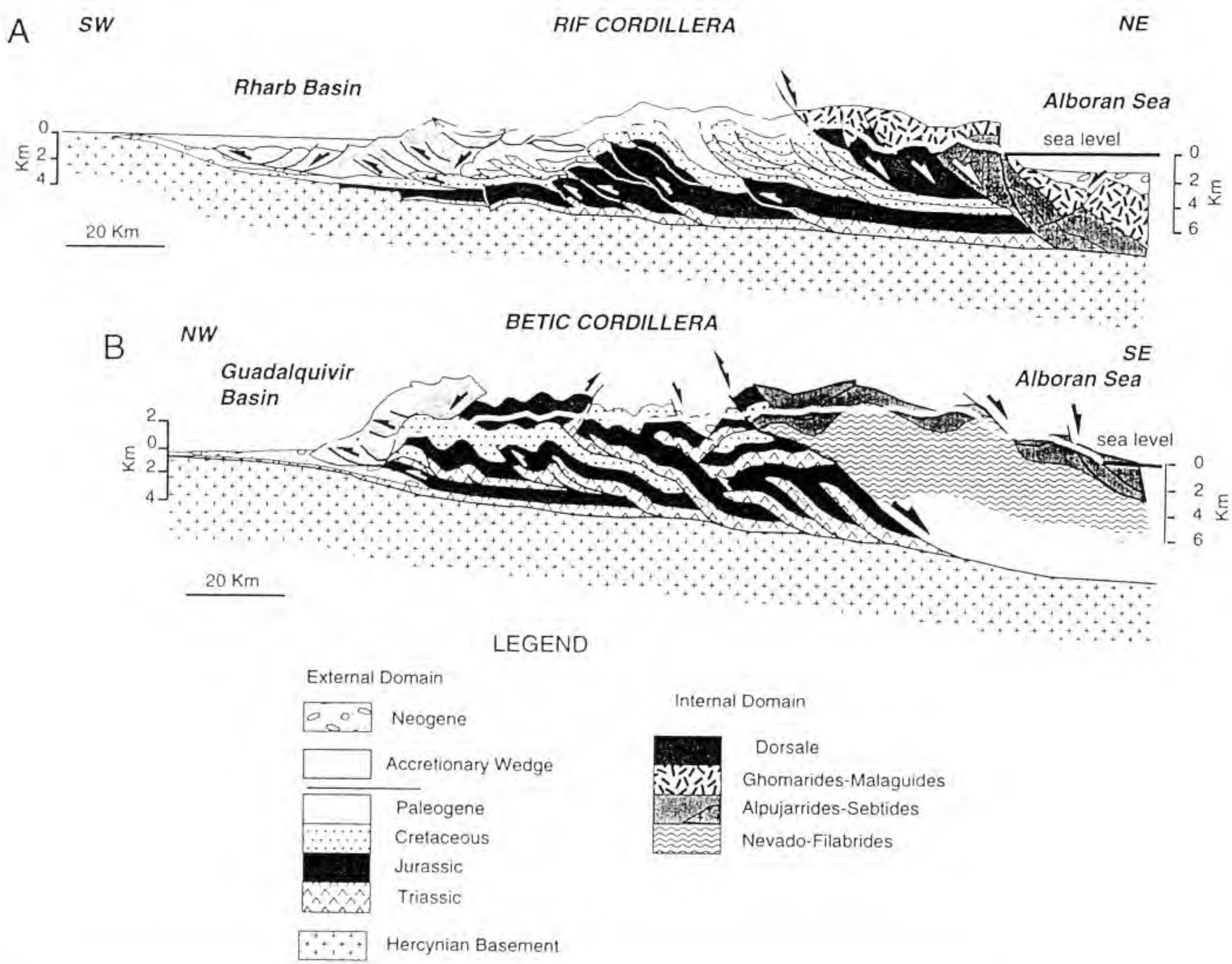

FIG. 3. Cross-sections of the Betic and Rif Cordilleras. Based on subsurface and surface data from Suter (1980a), Blankenship (1992) and García-Dueñas et al. (1992).

lent to the Prerifaine Nappe of the Moroccan Rif. Both the Guadalquivir Allochthon and the Prerifaine Nappe constitute an up to 100 km wide belt (Fig. 2). The frontal allochthonous units of the Betic Cordillera (i.e. the Guadalquivir allochthonous units) extend into the Gulf of Cádiz and farther west to the so-called Horseshoe or "Fer du Cheval", east from Gorringe Bank and the Seine abyssal plain in the Central Atlantic (Fig. 1) (Lajat et al., 1975; Malod and Didon, 1975; Malod and Mougenot, 1979).

Near Gibraltar the inner units of the Arc are bounded by the so called Flysch units, which consists of deep-water shales and turbiditic sandstones.

The Guadalquivir Allochthon incorporates voluminous Triassic shales and evaporites, as well as marls, carbonates and siliciclastics (Perconig, 1960-62). In contrast, the Prerifaine Nappe includes only minor amounts of Triassic evaporites, but mainly younger detritic turbidites or shales (Daguin, 1927; Termier, 1936; Bruderer and Lévy, 1954; Tilloy, 1955a, 1955b, 1955c, 1955d). In the following I will focus on the southern part of the Gibraltar Arc, in the accretionary zone of the Rif Cordillera, classically refered to as the **Prerifaine Zone** (Suter, 1965).

2.1 Stratigraphy

The sedimentary prism of the accretionary wedge can be subdivided into three major tectonostratigraphic units: Supra-Nappe, Nappe and Infra-Nappe. Due to the lack of offshore wells (only a shallow well was drilled offshore Larache, see Fig. 6 for location) stratigraphic information offshore is largely based on correlation with onshore data (Flinch, 1993). Onshore wells were tied to seismic sections and correlated with offshore data. Section of Figure 5 illustrates an example of onshore-offshore correlation.

In the study area few wells have penetrated the sedimentary succession located below the Prerifaine Nappe (Fig. 5). Cretaceous and Lower to Middle Miocene strata unconformably overlie metamorphic and igneous Paleozoic rocks of the Hercynian Basement. Locally a Triassic shaly and evaporitic section was encountered. Seismic data demonstrates that Triassic sediments occupy extensional half-grabens. The Cretaceous and Lower-Middle Miocene section represents the cover of the Morocan Meseta. The lack of Jurassic in this area is explained by some authors as a result of shoulder uplift related to the opening of the Central Atlantic (Favre et al., 1991; Favre and Stampfli, 1992). The sedimentary succession encountered by exploratory wells in the Rharb Basin is similar to the stratigraphic section exposed in the western Moroccan Meseta described by Gigout (1951).

The Prerifaine Nappe consists of Triassic to Miocene sediments that are bounded by stratigraphic and/or tectonic contacts (Bruderer and Lévy, 1954). The stratigraphy for the Prerifaine Nappe is obscured by complex deformation. Resedimentation and the presence of reworked Cretaceous faunas (i.e. ammonites) together with Triassic and Miocene sediments leads to difficult biostratigraphic problems (Feinberg, 1986). Thickness estimates are not easily made. The stratigraphy of the Prerifaine Nappe is based on very few outcrops located in the areas surrounding the Rharb Basin and on exploration wells located within the basin. The following description of the stratigraphy is based on surface data obtained by the SCP (Tilloy, 1955a, 1955b, 1955c, 1955d; Feinberg, 1986) and information from exploration wells that penetrated the Nappe. Table 1 describes the lithology and fossil content of the Prerifaine Nappe. The Neogene planktonic biozonation is derived from Feinberg (1986) and Wernli (1988). Even though there is not a discernable stratigraphic order because of imbrication and reworking, the sediments involved in the Nappe proceed from older to younger.

The Supra-Nappe complex consists of a seaward prograding wedge that ranges in age from Late Miocene to Holocene. However the presence of restricted anoxic environments poor in planktonic faunas, resedimentation and tectonic complications hinder the establishment of a generalized biostratigraphy. The litoral faunas of the peripheral regions of the Rharb Basin are difficult to correlate with the pelagic faunas of the center of the basin.

The Neogene biostratigraphy of the Rharb and Rif areas, was based on Wernli (1988). The Supra-Nappe subsurface stratigraphy of the Rharb Basin based on selected wells is shown in Fig. 5.

AGE	LITHOLOGY	FORAMINIFERA	THICKNESS	BOUNDARIES	COMMENTS
Middle Miocene	Interbedded sandy-marl and sandstone with occasional limestone beds.	*Globorotalia miozea rifensis* *Orbulina suturalis*	200-300 m	Transitional	
Lower Miocene	Interbedded sandstone and marl with occasional marly-limestone levels.	*G. klugeri, G. primordius, G. trilobus* *G. deshincens, G. bisphericus.*	100-300 m	Transitional	
Upper Oligocene	Interbedded marl and turbiditic sandstone with occasional breccia intervals and limestone beds.	*G. angulisuturalis*	150-200 m		Paleontologic problems to define the Oligo-Miocene boundary.
Lower Oligocene		*Lepidocyclina sp. Amphistegina sp* *G. ampliapertura G. evapertura* *G.. gortanii G. sellii G. ampliapertura*		Transitional	
Upper Eocene	Interbedded limestone and sandstone	*G. semmiinvoluta* *G. cocoaensis*	400 m	Transitional	
Middle Eocene	Interbedded marly-siltstone and sandstone with conglomerates and olistostromes	*Hantkenina aragonensis* *Globigerinatheka subconglobata subconglobata* *Globorotalia lehneri*	450-500 m	Transitional	
Lower Eocene	White marl and grey marly-limestone with nodular silex.	*G. caucasica, G. palmerae, G. aragonensis* *G. lensiformis G. marginodentata*	200-300 m		These facies are referred to locally as "marnes blanches a silex"" or" le Numulitique" due to the abundance of Numulites
Upper Paleocene		*G. pseudomenardii, G. velascoensis*			
Lower Paleocene		*G. edita, G. pseudobulloides, G. compressa* *G. elongata, G. triloculinoides, G. daubjergensis* *G. trinitatensis.*		Transitional	
Upper Cretaceous	Magnesium-bearing green marly-shale with planktonic foraminifera	*Trinitella scotti, Plummerita hantkeninoides* *Globotruncana sp.*	?		Triassic evaporites with blocks of sedimentary, volcanic and metamorphic rocks appear to be interbedded with Cretaceous marls
Lower Cretaceous	Albian calci-turbidites consisting of interbedded marl and marly-limestone Ammonite-bearing Neocomian gray marl and white marly limestone	?	?	Tectonic or diapiric	
Triassic	Varicolored and often purple marl with interbedded salt and gypsum. Occasional diabase and pillow-lavas	?	?		Locally blocks of Paleozoic gneiss micaschist or organic rich black schist are intermixed with Triassic

TABLE 1

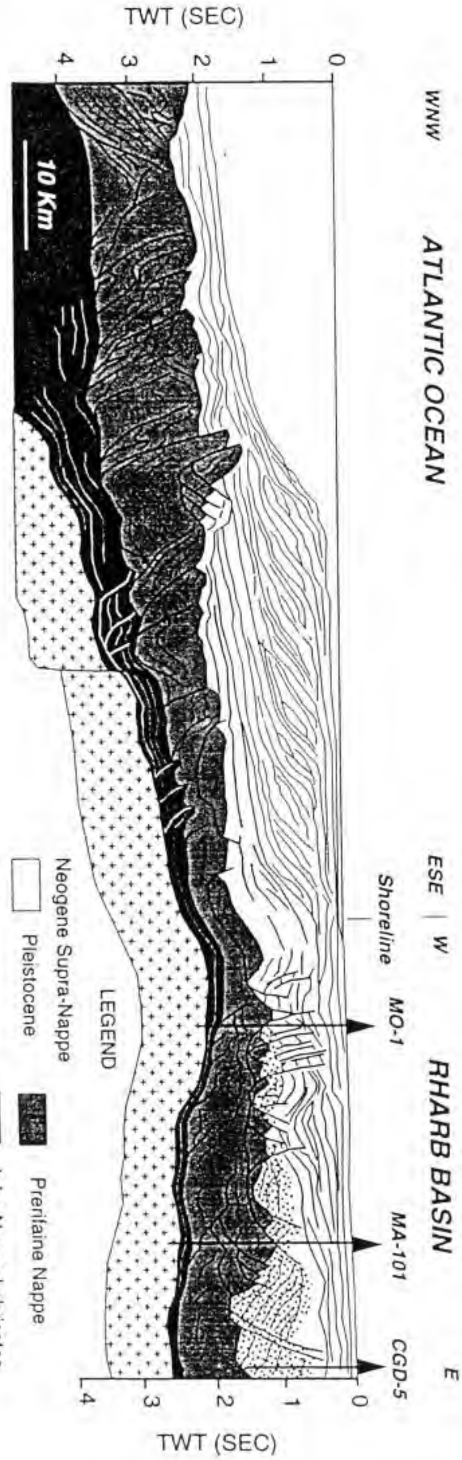

FIG. 4. Cross-section of the Accretionary Wedge in the Western Rif outlining the Offshore-Onshore correlation. Notice that the apparent gentle westward deepening of the wedge is a velocity effect due to the water column.

3 STRUCTURE OF THE EXTERNAL WESTERN RIF

Reflection seismic, well-log and field data were integrated to provide an understandable picture of the external part of the Western Rif. Subsurface data is specially important in this area due to the lack of good surface exposures. A large number of seismic profiles covering the northwestern Atlantic margin of Morocco were used to map the structure of the Prerifaine Nappe in the offshore region. Mapping permits to establish the offshore prolongation of the Rif frontal thrusts, the leading edge of the frontal accretionary wedge and the contact between extensional and compressional provinces (see Fig. 6). A complex set of basins can be outlined within the accretionary zone (Flinch and Bally, 1991).

3.1 Regional Transects

Five NE-SW offshore regional sections, extending from Asilah to Rabat, have been selected to show the structure of the northwestern Moroccan Atlantic margin (Fig. 6 and Enclosure 1). The transects display a variety of structural styles. According to the seismic character and structural significance, a number of structural units are differentiated, from bottom to top:

(1) Acoustic basement: Paleozoic.
(2) Infra-Nappe: Mesozoic and Lower Miocene cover of the Paleozoic basement.
(3) Prerifaine Nappe: Accretionary Wedge.
(4) Supra-Nappe: Upper Miocene and Plio-Pleistocene siliciclastics.

In the following these units will be referred to as: Basement, Infra-Nappe, Nappe and Supra-Nappe. The NE-SW oriented sections (Enclosure 1) traverse the main structural units of the accretionary complex shown on the structural map (Fig. 6). The regional sections will be described proceeding from the southern foreland basin to the northern frontal folded belt. The southern part of the transects shows northward-dipping layered reflectors of the foreland which project under the frontal imbricates of the accretionary complex. These reflectors are occasionally detached from the acoustic basement to form imbricates. The frontal imbricates are characterized by thrust planes which dip steeply to the north. Thrust sheets emanate from a gently northward-dipping basal décollement which separates them from the underlying autochthon. Proceeding northward, the complex is overprinted by northward-dipping normal faults and associated extensional basins with no significant growth. Further north, extensional faults step down and confine thick extensional basins. They constitute the southern part of an extensional system running nearly perpendicular to the plane of the sections. Ridges cored by folded accretionary complex sediments occur in the central region of the extensional system. The southern branch of the extensional system is characterized by northward-dipping normal faults. These faults are connected with conjugate southward-dipping listric normal faults that often constitute the lateral ramps of the extensional system. The central portion of the extensional basin, with its high ridges and deep troughs, is detached from the basal extensional contact. This portion of the margin provides exceptional sections across the extensional system. Proceeding northward, the northern branch of the extensional system cuts thrusts and folds of the underlying accretionary complex. The northern portion of the sections is characterized by north-dipping thrust planes and related ramp anticlines.

According to the data presented here, the study region can be subdivided into several structural domains (Fig. 6): Offshore Tanger-Asilah Compressional Belt, Offshore Larache Extensional Zone, Offshore Rharb Compressional-Extensional Zone, Rharb Basin and Rabat Foreland Basin. To facilitate the presentation of the data, the offshore data will be presented separate from the data of the Rharb Basin, despite their structural affinity.

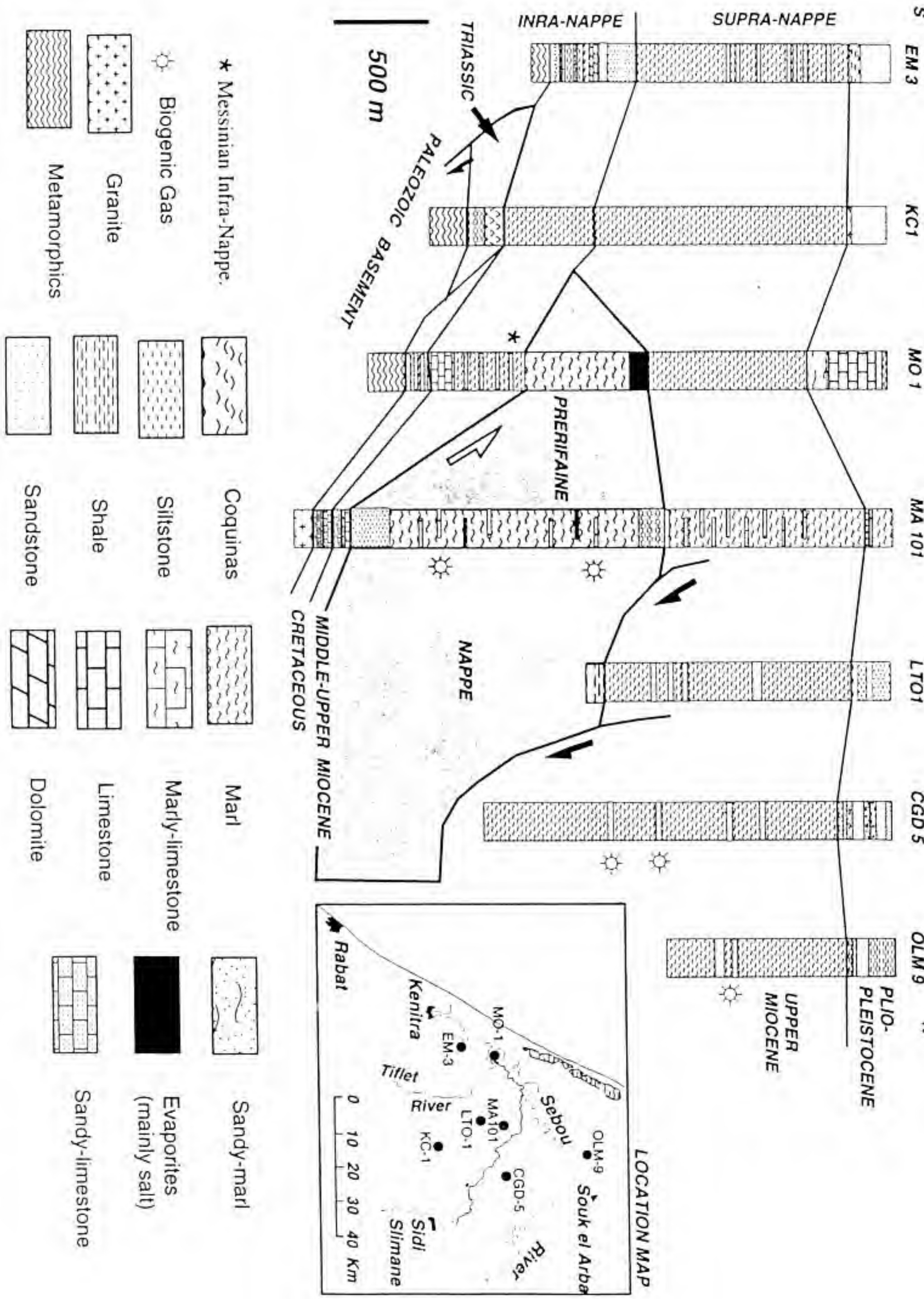

FIG. 5. Stratigraphic correlation of some selected wells of the Rharb Basin. Data from Wernli (1988), Feinberg (1986) and unpublished ONAREP reports.

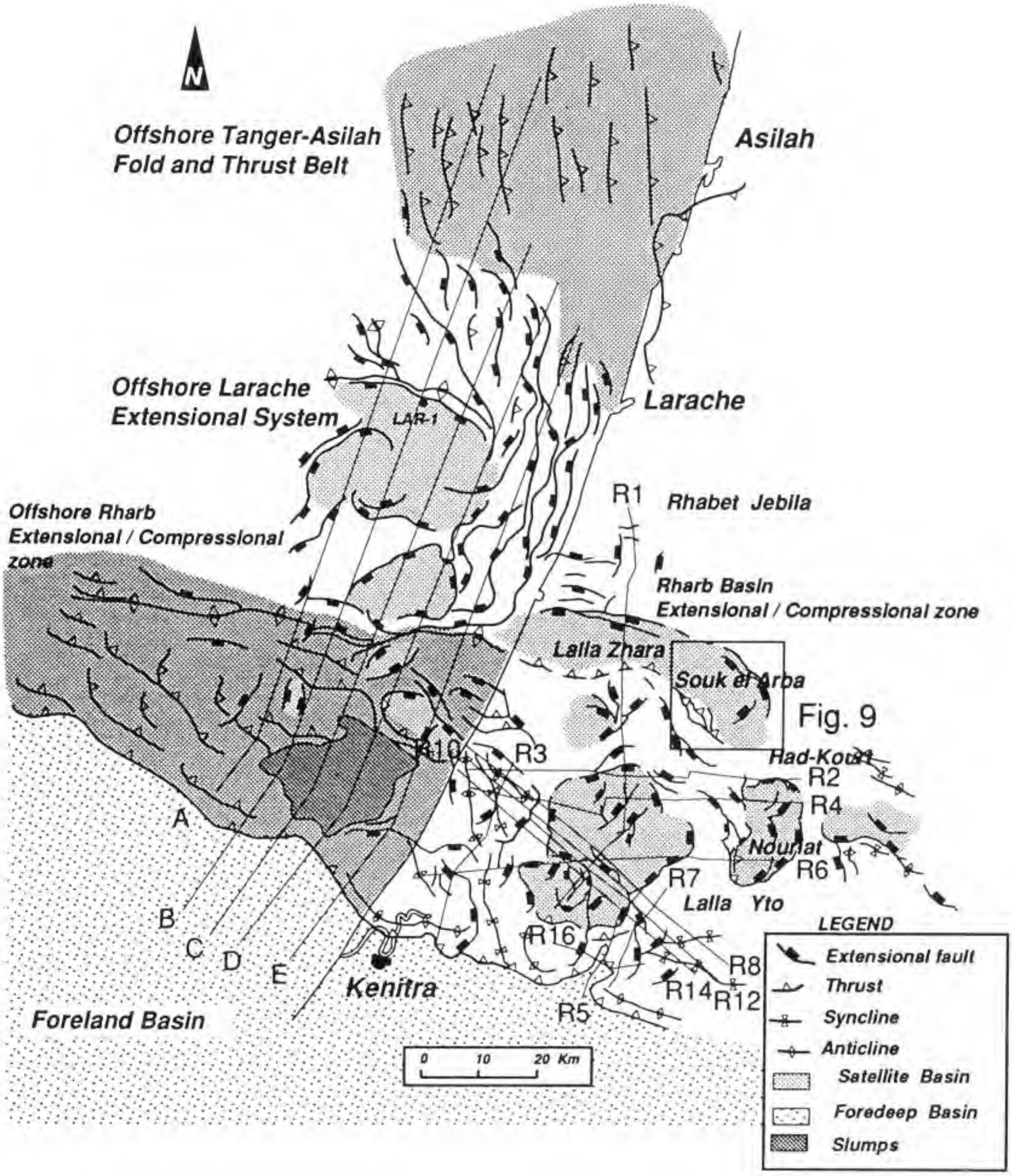

FIG. 6. Structural map of Northwestern Morocco with indication of the major structural domains consider in this study.

3.1.1 Offshore Tanger-Asilah / Fold and Thrust Belt

The northern part of the Moroccan Atlantic margin consists of westerly-vergent folds and thrusts. Fold axes and thrusts strike NNW-SSE, following the general trend of the Rif Cordillera (Fig. 6). They represent the continuation of the structures of the Tanger and Habt units exposed in the western Tanger peninsula (Suter, 1980b). The quality of the seismic data in this area is mediocre, and there are some regions with virtually no useful data.

3.1.2 Offshore Larache / Extensional Zone

The frontal thrusts of the Western Rif Cordillera extend further west into the region offshore Tanger-Asilah. Thrusts and related folds are cross-cut by NW-SE trending SW-dipping low-angle listric normal faults (Enclosure 1). The structure of this region is defined by troughs and ridges. Extensional basins are bounded by several anostomosing listric normal faults that sole out into a basal low-angle detachment which offsets the top of the accretionary wedge (Fig. 6). This network of anostomosing faults results in extensional horses that merge with each other and are superimposed on the accretionary wedge. Normal faults trend N-S, nearly parallel to the present day shoreline, and dip towards the west. E-W oriented sections (see Enclosure 1) show strongly rotated blocks on the hangingwall of the low-angle extensional detachment. In the eastern portion of these sections, growth-faulting results in large Supra-Nappe expansion controlled by westward-dipping normal faults that sole out into the basal detachment. Sediments above the extensional system show significant fault growth. In the central portion of the area, the top of the Nappe attains depths of 3.5 sec (TWT). The basal detachment of the extensional system steps down from 0.5 sec in the east to 3.5 sec in the west. The aparent transport direction of this extensional system is to the west. Nearly E-W oriented shale ridges are present in the central part of the extensional system; these were caused by shale withdrawal induced by extensional displacement.

3.1.3 Offshore Rharb / Frontal Imbricates-Extensional-Compressional Zone

This area is located west of the southern Rharb Basin, between the confluence of the Sebou River and the village of Moulay Bou Selham. High quality seismic data in this region show the details of the frontal part of the accretionary complex and the northernmost portion of the Rif foredeep. The Offshore Rharb area displays a combination of compressional and extensional elements (Fig. 6). In this complex area, NW-SE trending normal faults and occasional NE-SW-oriented normal faults cut NW-SE trending SW-vergent folds and thrusts. To the east of the area, normal faults share the same décollement level as thrust faults, defining toe-thrusts that accommodate the normal fault displacement (see Enclosure 1).

The frontal part of the wedge is characterized by closely spaced NW-SE trending, NE-dipping thrust faults that define a zone of frontal imbricates. Lateral and oblique ramps related to these frontal thrusts are common in the central and western portions of the area (Enclosure 1). The front of the accretionary complex has a NW-SE orientation (Fig. 6).

The structure of this area is well constrained by high quality seismic data. Dip lines are those trending perpendicular to the leading edge of the accretionary complex, that is NE-SW. Strike lines are those trending roughly parallel to the front of the wedge, that is NW-SE.

The most conspicuous features shown by the dip NE-SW sections are:

(a) The wedge-like geometry of the Prerifaine Nappe.
(b) Imbrications within the Infra-Nappe authochthonous succession.
(c) Frontal thrusts within the Nappe.
(d) Extensional faults crosscutting the Nappe.
(e) A northwestward-dipping basement beneath the Nappe.

(f) The northernmost portion of the foredeep.
(g) Frontal slumps

The strike sections show the following structural features:

(a) Steeply-dipping basement-involved faults offsetting the base of the Infra-Nappe units.
(b) Eastward thinning of the Prerifaine-Nappe.
(c) Lateral ramps of the frontal imbricates.
(d) Westward progradation of the Supra-Nappe succession.
(e) Frontal slumps and detached units.

3.1.4 Rharb Basin / Frontal Imbricates-Extensional-Compressional Zone

The topographic Rharb Basin overlaps the western front of the Rif Cordillera and its foreland (Fig. 2). It is bounded to the east and north by the frontal ranges of the Rif Cordillera and to the west by the Atlantic coast. The southern limit is the Paleozoic Moroccan Meseta. The surface expression of the Rharb Basin is a fluvial-alluvial coastal plain drained by the Sebou River. Subsurface data presented in this paper display the structure of the Neogene sedimentary succession of the Rharb Basin and the underlying Prerifaine Nappe.

South of the imbricated zone represented by the Asilah, Habt and Tanger units (Suter, 1980b) most of the structure of the accretionary wedge is controlled by extensional structures (see Fig. 6 and Enclosures 1 and 2). Extensional basins are located between the Rif fold and thrust belt and the leading edge of the accretionary wedge (Prerifaine Nappe). The structure of the Rharb Basin does not consist of well defined half-grabens but it is composed by a complex set of extensional and locally compressional basins (Fig. 6, Enclosure 2). These "satellite"[1] basins are bounded by several anostomosing listric normal faults which sole out into a basal low-angle detachment offseting the top of the accretionary wedge. Often extensional structures in the rear are coeval with conpressional toe-thrusts at the front of these "satellite" basins. The orientation of main faults is variable and random. Eventhough there is controversial data on the timing of development of these basins, most of them are Tortonian-Messinian in age. Growth is limited and most of the supra-nappe sediments are characterized by parallel bedding, which suggests fast extensional collapse. Basal re-deposited shallow-water sandstones and anoxic marls with occasional siltstone beds fill these extensional basins. Anoxia is the result of the extensional topography at the top of the Prerifaine Nappe, which involves highs and lows (Cirac and Peypouquet, 1983). In the cover sequence, rapid facies changes through time suggest also rapid extensional collapse of the Nappe (Flinch, 1993).

In the southern part of the Rharb Basin compressional structures associated with the front of the Prerifaine Nappe are observed (Enclosure 2). They consist of NE-SW trending anticlines and synclines related to SW-vergent imbricates of the Prerifaine Nappe. The southern area is occupied by E-W and N-S trending troughs associated with extensional faults superimposed on top of the Nappe. The structure of the central part of the Rharb Basin consists of E-W, N-S and NE-SW trending extensional troughs bounded by low-angle listric normal faults (Fig. 6). Depocenters in excess of 4500 meters occur southwest from Mechra bel Ksiri. (Dakki, 1992). Three of these basins are located in the contact between the foothills of the Rif and the Rharb Basin, namely: the Lalla-Zhara, Souk el Arba (Fig. 7) and Nouriat Satellite Basins. All these basins are characterized by rearward extension and frontal compression, which defines large scale toe-thrusts. The sedimentary fill of these basins, consists mostly of Tortonian-Messinian marls, with occassional interbedded sandstones and siltstones. Shallow biogenic gas, and locally oil has been recovered from the Aïn Hamra area (see Fig. 7 for location). The presence of Cretaceous source oil at a very shallow level, few hundred meters, suggests a connection between the

[1] The term **satellite basin** is preferred to piggy-back basin because of the presence of normal faults combined or not with thrust faults. The mechanics of these basins is therefore different than conventional thrust-related piggy-back basins in the sense of Ori and Friend (1984).

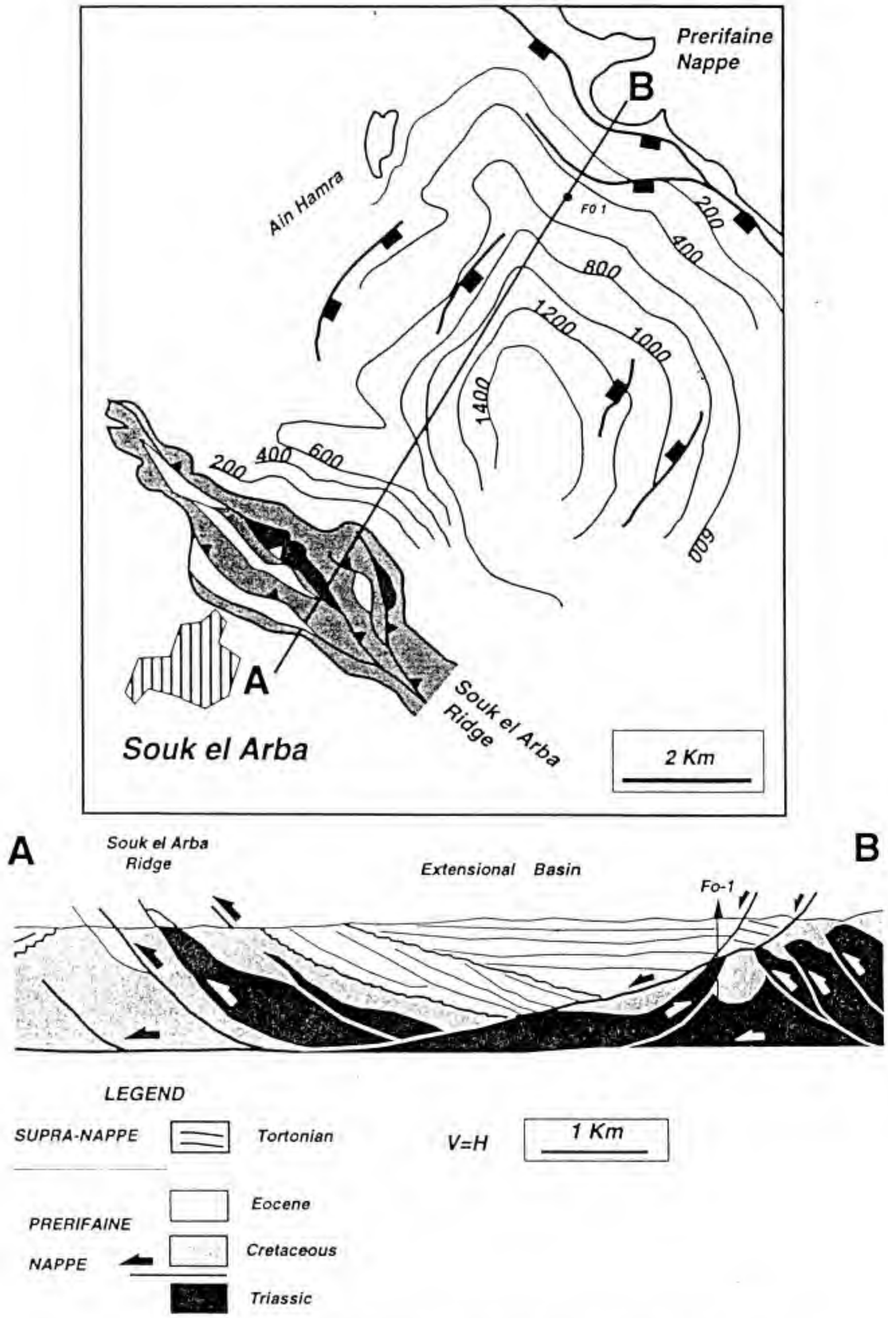

FIG. 7. Structural map and cross-section along the Souk el Arba Satellite Basin.

extensional decollement and the underlying thrust faults within the Nappe.

The structure of the Rharb Basin is illustrated by a series of composite regional seismic lines. Enclosure 2 shows four regional dip sections and eight strike lines. The dip lines are roughly oriented NE-SW, trending nearly perpendicular to the leading edge of the Prerifaine Nappe. The strike lines are oriented perpendicular to the transport direction of the Nappe and parallel to its leading edge. Enclosure 2 displays five northeastward trending sections in the Lalla Yto area. Three E-W oriented sections, extending from the Nouriat region in the east to the Sebou region in the west; four NE-SW trending sections; two in Lalla Yto in the East, one through the central Rharb extending from Rhabet Jebila in the north to Lalla Yto in the south and one along the Sebou coastal area. On the basis of seismic character and structural significance, the same seismic units are recognized as in the offshore sections (Enclosure 1), that is: Basement, Infra-Nappe, Nappe and Supra-Nappe.

The most characteristic features evidenced on the dip lines (odd numbered sections of Enclosure 2) are:

(a) The wedge-like character of the Prerifaine Nappe.
(b) The northward dip of the basement and the infra-nappe succession underneath the Prerifaine Nappe.
(c) Southward vergent imbricates involving infra-nappe sediments.
(d) Extensional faults offsetting the top of the nappe.
(e) Clinoformal patterns in the Supra-Nappe units indicating southwestward progradation.

The most conspicuous features evidenced in the strike lines (even numbered sections of Enclosure 2) are:

(a) The thickness change of the Infra-Nappe unit.
(b) Nearly vertical faults offsetting the Infra-Nappe succession.
(c) Lateral ramps of the extensional system (the oblique orientation of the section reveals lateral ramps).

3.1.5 Onshore-Offshore Rabat / Foredeep

South of the leading edge of the Prerifaine Nappe, the structures consist of NE-dipping Infra-Nappe units that plunges beneath the accretionary wedge. Basement-involving nearly-vertical normal faults disrupt the otherwise continuous Infra-Nappe succession (see Enclosures 1 and 2). Some of these faults account for a thickness change of the infra-nappe succession but do not affect the overlying sediments (Enclosure 2); these faults are related to the Central Atlantic rift system (Flinch, 1993). Other faults do not account for any thickness change of the infra-nappe units but offset the foreland succession, these faults are related to flexural extension induced by tectonic loading of the foreland by the Prerifaine Nappe (Enclosure 1) (Flinch, 1993) as seen also in other foreland basins (Bradley and Kidd, 1991). An angular unconformity which is onlapped by the foreland sequence corresponds to the "basal foredeep unconformity" in the sense of Bally (1989). Locally, sediments derived from the craton, prograding into the foredeep can be recognized in the southern part of the area (see Enclosure 1). This region is located in the offshore prolongation of the Rif foredeep, located east from Rabat (see Figs. 2 and 6 for location).

4 DISCUSSION

The overall composition and type of deformation of the external Betic and Rif domain suggests the involvement of an accretionary prism consisting of deep-water sediments, that was emplaced on the attenuated Iberian and African passive margins in response to westward motion of the Alborán domain during Middle Miocene to Pliocene time.

The timing of deformation and the age of the sediments involved suggest an accretionary progression towards the external portion of the arc. Deformation within the accretionary complex was previously explained by several phases of deformation and by gravitational tectonics (Feinberg, 1976; Vidal, 1977; Feinberg, 1986). In contrast, a continuous accretion model, similar to current models of more conventional accretionary complexes (e.g. Dickinson and Seely, 1979; von Huene 1986), appears to apply here. On the basis of field data and seismic data, a block diagram of the accretionary complex was constructed (Fig. 8). Note that the accretionary wedge is presumably underlain by a normal to transitional continental crust which dips towards the Mediterranean (A-type subduction); this contrast with the more conventional accretionary wedges that are related to subducting oceanic lithosphere (B-type subduction).

The style of extension present in the external part of the Gibraltar Arc is similar to the Gulf of Mexico (Worrall and Snelson, 1989). It consists of listric normal faults rooted into a low-angle extensional detachment composed of overpressured shales and marls. In the offshore Larache area, the transport direction of the extensional system coincides with the geometry of the continental slope, and is nearly parallel to the present day shelf-break.

There is no clear relationship between the frontal compressional zone (Offshore Rabat) and the rear extensional system described above. The lack of seismic resolution and penetration in the lower part of the seismic sections does not allow to see if down-dip thrusts and folds share the same decollement as the up-dip low-angle normal faults. Therephore, it is difficult to demonstrate if extensional displacement is compensated by frontal compression at a regional scale. The most frontal parts of the Gibraltar Arc accretionary wedge may represent compressional belts that are the result of rear extension, as suggested by Platt (1986) for other accretionary wedges. This type of deformation would represent the response of the system to the unstable oversteepened slope (critical taper theory) generated by the stacking of thrust slices within the wedge (Davis et al., 1983).

The three-dimensional diagram presented here has some important implications for the geology of the Betic and Rif Cordilleras. The Rif Cordillera consists of numerous structural units, which classically are referred to as "Nappes" (Suter, 1980b), however many of these units do not have the attributes of thrust sheets. Omission of strata or no duplication of the stratigraphic section are common to these units. A structural map of the Western Rif was put together integrating offshore and onshore subsurface data and field data based mostly on published geologic maps (Flinch, 1993) (Fig. 9). The relationship between the Prerifaine Nappe and the underlying units of the **External domain** can be observed particularlly well in the Had-Kourt/Teroual area (Fig. 10). Seismic data through the area permits to see the relationship between several stacked thrust-sheets (Fig. 11). The data presented here suggest that several geologic units referred to as "unites flottantes" or "Nappes rifaines superieures" (upper thrust sheets) (Wildi, 1983), are in fact satellite basins which overlie the accretionary wedge (Figs. 9, 10 and 11). These basins were deformed at the same time as the transport of the accretionary wedge towards the foreland, in response to the collision of the Alborán allochthonous terrane with the Iberian and African foreland. In the past, some of these Satellite basins were interpreted as out-of-sequence thrusting (Morley, 1992). Instead a model where the structure is the result of the piggy-back emplacement of sequences, that is, the upper units were emplaced first, the later emplacement of the lower units deformed the upper units from underneath. The first unit to be emplaced was the accretionary wedge and the underlying more landward passive margin units were emplaced afterwards (Fig. 11). This lead to widespread structural envelopment of the accretionary wedge (Flinch, 1993).The new concept presented here, significantly simplifies the structural framework of the Rif Cordillera. I postulate that these units may have a similar origin as the Neogene satellite basins of the offshore and onshore Rharb area.

The Ouezzane Unit (Hottinger and Suter in Durand-Delga, 1960-1962) of the external Western Rif is the most obvious example of such a complex of satellite basins. This unit was interpreted as a thrust sheet or "nappe" located on top of the Prerifaine nappes (Suter, 1965; 1980b). In the so-called flysch domain, the Numidian Unit represents the highest structural unit which is located above the underlying imbricates; also this unit may represent

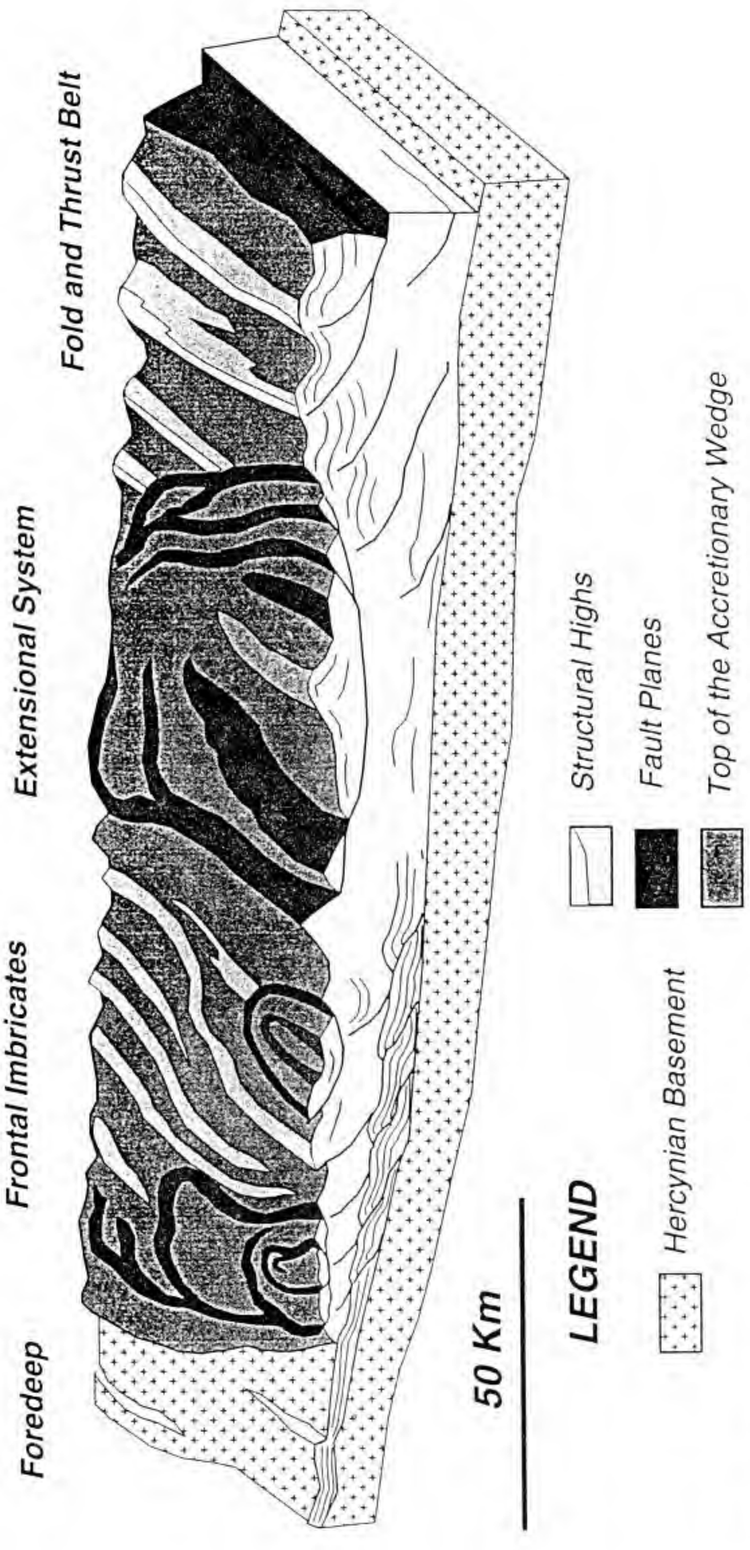

FIG. 8. Three-dimensional block-diagram of the accretionary wedge in the External Western Rif.

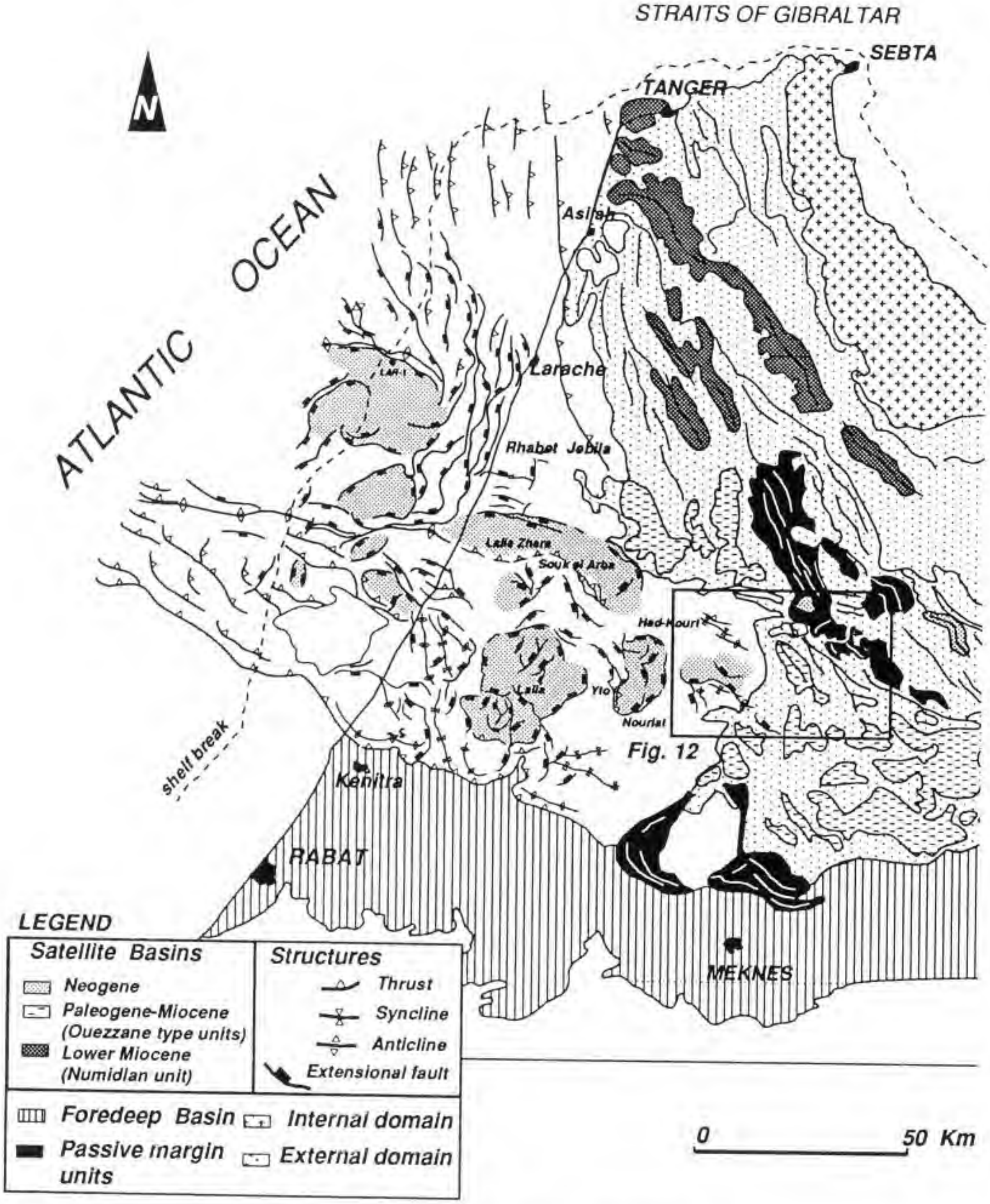

FIG. 9. Structural map of the Western Rif integrating onshore and offshore data, after Flinch (1993).

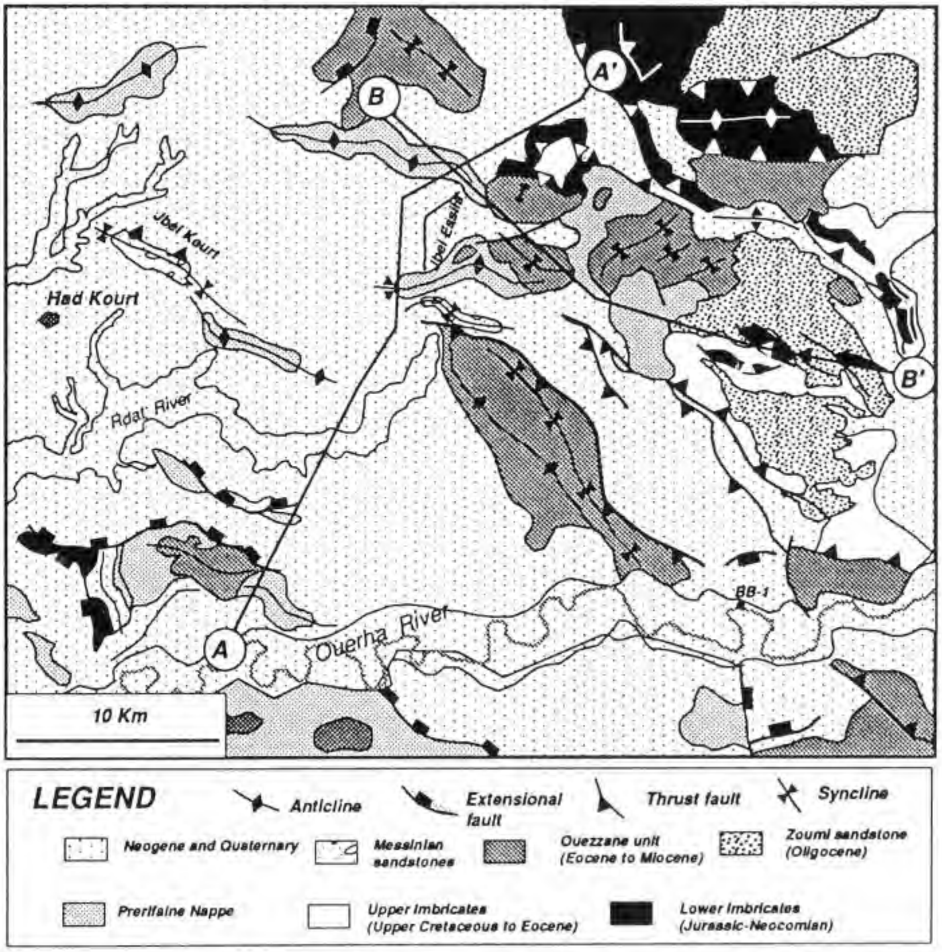

FIG. 10. Structural map of the Had Kourt-Teroual area. Modified from the Service Géologique du Maroc 1:50,000 scale maps of Had Kourt (1984) and Teroual (1990).

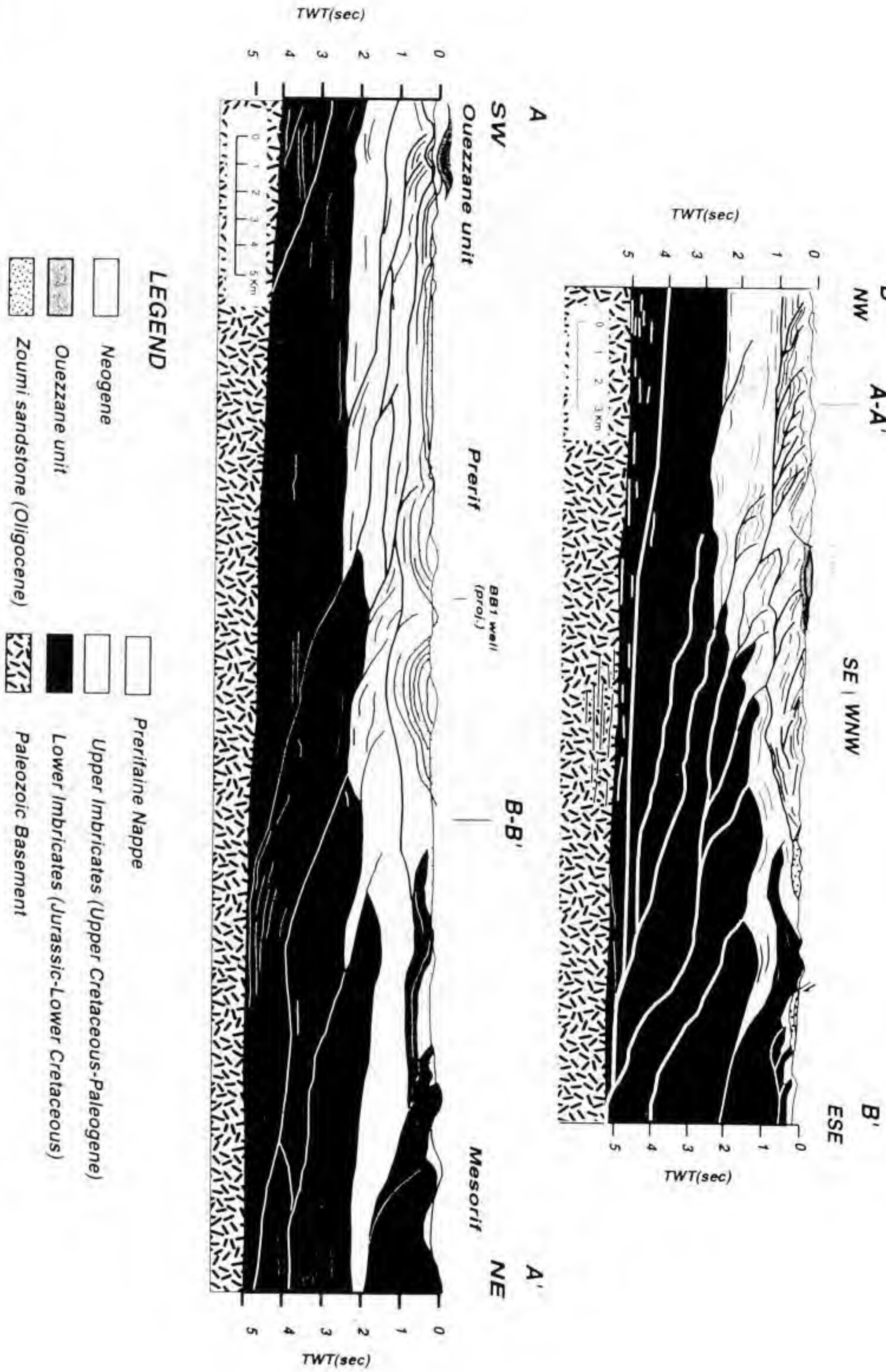

FIG. 11. Line drawings from seismic sections along the Had Kourt - Teroual area.

a set of satellite basins. The Numidian sandstone was deposited on top of imbricates involving previously deposited turbiditic deposits of the Tanger and Ketama Units. I suggest that the Numidian Sandstone represents the turbiditic Satellite basin fill which overlies the Oligocene-Early Miocene accretionary wedge of the Gibraltar Arc. These Satellite Basins appear to be detached from the underlying units by thrust or normal faults, thus simulating an independent thrust sheet. However, unlike real stacked thrust sheets, younger sediments overlie the deeper tectono-stratigraphic unit.

5 CONCLUSIONS

The frontal tectono-sedimentary complexes of the Betic and Rif Cordilleras, the Guadalquivir Allochthon and the Prerifaine Nappe, constitute an accretionary wedge which was superposed on the attenuated passive margin of Iberia and NW Africa during the Miocene phases of the Alpine orogeny.

The structure of the accretionary wedge consists of frontal imbricates, ridges, toe-thrusts and low-angle extensional detachments. The Supra-Nappe sediments involve compressional-extensional and extensional satellite basins trending parallel and perpendicular to the Arc. Satellite basins are not directly related to the opening of the Alborán Sea; instead they are due to oversteepening of the accretionary wedge and gravitational gliding down the continental slope.

Very rapid extensional collapse affected the accretionary wedge during Tortonian and Messinian time. The paleogeographic evolution, involving the superposition of deep-water facies onto shallow-water sediments and the lack of significant growth support this argument.

The style of extension is characterized by low-angle listric normal faults, similar to the Gulf of Mexico. Extensional displacement is compensated by frontal compression. Toe-thrusts are common structures in the frontal part of the Gibraltar Arc. An undisturbed upper part of the prograding Supra-Nappe succession suggests that the accretionary wedge was stabilized during Pleistocene time.

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APPENDIX

In the following I will describe the structure of the seismic line drawings presented in Enclosure 2.

Regional Section R1

This N-S and NE-SW regional section extends from Rhabet Jebila to Lalla Yto. In the northern portion of the transect a series of southward-dipping growth faults offset the top of the Prerifaine Nappe. The region of Rhabet Jebila-Lalla Zhara consist of a northern extensional satellite basin and a southern ridge. The basal decollement is located at 2 seconds recording time. Thrusts connected with this basal detachment are present within the Nappe. The maximum thickness (2.5 sec) (1800 meters) of the Supra-Nappe Neogene is attained in the central part of the section. The southern part of the section (Lalla Yto region) is represented by lateral ramps merging into the basal decollement of the Prerifaine Nappe. Antithetic and synthetic normal faults also offset the Mio-Pliocene boundary. Normal faulting is younger in this region than in the northern Rharb, where the Mio-Pliocene boundary is not offset.

Regional Section R3

This northeastward trending section follows the Sebou River coastal plain and shows the wedge-shaped Prerifaine Nappe. In the southeastern portion of the section, gently northeastward-dipping Infra-Nappe reflectors unconformably overlie the Hercynian basement. The Supra-Nappe prograding units onlap directly the basal foredeep unconformity. Well EM-3, located in the foredeep region just in front of the Nappe, penetrated the whole sedimentary succession, encountering Cretaceous and Tortonian-Messinian Infra-Nappe and Plio-Pleistocene Supra-Nappe. North of the leading edge of the wedge, the Infra-Nappe unit is characterized by a basal zone of imbricated layered reflectors. The structure of the Prerifaine Nappe itself is characterized by northeastward-dipping thrust faults. Extensional faults are superimposed on the Nappe. Supra-Nappe units show a southward prograding clinoformal pattern. Well MO-1 penetrated the Nappe and Infra Nappe units, reaching the Paleozoic basement.

Regional Section R5

This NE-SW oriented section located in Lalla Yto illustrates essentially the same features as on section R3. Again, the geometry of the wedge, the basal imbricates of the Infra-Nappe unit, and the normal faults that cut the Supra-Nappe succession are the most interesting features shown on this profile. On the SE margin of the section, south of the leading edge of the Prerifaine Nappe i.e. in the foredeep, the Supra-Nappe units onlap directly on the northeastward-dipping Infra-Nappe reflectors, thus definnig the basal foredeep unconformity. Well KC-1, located in front of the Nappe, reaches the basement after penetrating Infra-Nappe Triassic, Cretaceous and Middle Miocene.

Regional Section R7

This northeastward trending section is located in the region of Lalla Yto, west of section R5. The section shows the same features as section R7. In the frontal part of the Nappe, extensional and compressional structures are detached at the same décollement level, thus defining toe-thrusts.

Regional Section R2

This section is oriented E-W, extending from the region of Nouriat to the Atlantic Coast. The line shows the subsurface expression of the contact between the Rharb Basin and the frontal ranges of the Rif. The eastern end of the section shows westward-dipping listric normal faults, responsible for the westward-thickening of the Supra-Nappe succession. Most normal faults merge into a low-angle extensional detachment. Rotated blocks develop in the hangingwall of the extensional system. Extension in the Nouriat area occurs mostly during Messinian time. Proceeding westward, the structure of the central Rharb Basin consists of a series of westward-dipping listric normal faults. Lateral ramps, suggesting a transport direction oblique to the plane of the section, are common in the central portion of the section. In the western part of the section (Sebou region), the base of the Prerifaine Nappe is at 3 sec recording time. Infra-Nappe layered reflectors are well imaged. Occasional thrusts are present within the Nappe. The Prerifaine Nappe is not significantly affected by extensional faults in this western area.

Regional Section R4

This E-W trending regional section is parallel to R2 and extends from Nouriat to the Atlantic. Most conspicuous on this section is the westward deepening of the top-of-the-Nappe extensional detachment. Ramps and flats define the geometry of the basal extensional detachment. The extensional décollement deepens down to 2.5 sec in the central part of the section. Supra-Nappe sediments fill the downthrown depressions of the extensional system. Only on the easternmost portion of the section is the Mio-Pliocene boundary offset by normal faults. In the Sebou region the basal decollement of the Prerifaine Nappe is imaged on the profile at 3 sec (TWT). Layered Infra-Nappe reflectors underlie the basal decollement.

Regional Section R6 (see Fig. 4)

This section shows lisia and Africa, defining the geodynamics of the region (Dercourt et al.; 1986; Andrieux et al., 1989; Dewey et al.,1989; Ziegler, 1987; Favre and Stampfli, 1992).

Regional Sections R8, R10, R12, R14

These sections located in the Lalla Yto area are closely spaced (2 to 3 km). They are oriented NW-SE, providing more examples of strike sections across the Rharb Basin. Wells MA-101 and MO-1 were tied into the seismic. The basal décollement steps down from 1.8-2 to 3-3.2 sec recording time. Thickness changes of the Infra-Nappe succession coincide with the location of nearly vertical faults that offset the basement. The top of the basement and the overlying Infra-Nappe succession dip to the northwest. The Prerifaine Nappe is thrusted onto Cretaceous and Middle Miocene Infra-Nappe sediments that overlie the Hercynian basement of the Moroccan Meseta. Lateral ramps of listric normal faults offset the top of the Nappe and the overlying Supra-Nappe succession.

Regional Section R16

This section across the central Lalla Yto area is roughly oriented WNW-ESE. The section is located near the leading edge of the Prerifaine Nappe. A complete Infra-Nappe succession penetrated by the KC-1 well consists of a Triassic half-graben and is unconformably overlain (Post-Rift unconformity) by parallel-bedded horizontal Cretaceous and Miocene sediments. Plio-Pleistocene units show a downlap pattern that suggests west-northwestward progradation.

Enclosures

Enclosure 1 Line-drawings of regional seismic sections, offshore Northwestern Morocco.
Sections A, B, C, D, E: offshore Asilah-Rabat. Sections F, G, H, I: offshore Larache.

Enclosure 2 Line-drawings of regional seismic sections: Rharb Basin, onshore Northwestern Morocco.

Triassic-Jurassic extension and Alpine inversion in Northern Morocco

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ABSTRACT

The Early Mesozoic half-grabens of northern Morocco form part of the regional extensional system which developed in conjunction with the opening of the Western Tethys. During the Tertiary, Alpine collision of the African and European plates, these half-graben system were inverted, giving rise to the uplift of the High and the Middle Atlas mountains. Similar, albeit less spectacular inversion features occur in the Guercif area and in the "Rides Prérifaines" of northern Morocco. Reflection seismic data show that inversion of the Guercif Basin involved the reactivations Mesozoic basement faults. In contrast, the "Rides Prérifaines" correspond to an extensive detachment system which is decoupled from the basement at the level of the Triassic evaporites. The geometry of this complex system of Late Miocene-Pliocene thrust faults and associated lateral ramps was preconditioned by the configuration of the Triassic-Jurassic extensional faults.

Detailed structural and stratigraphic analyses, combining surface geology and seismic data, greatly advanced the understanding of the geological history Northern Morocco and has led a reassessment of its hydrocarbon potential.

INTRODUCTION

The main structural elements of northern Morocco are the Moroccan Meseta, the Rif fold and thrust belt and the northeastern part of the Middle Atlas Mountains (Fig. 1).

The Moroccan Meseta is upheld by the outcropping, peneplained Hercynian basement and its Mesozoic cover. To the North, the Meseta dips gently under the Neogene fill of the Rif foreland basin which is underlain by a thin Mesozoic cover. This basin is subdivided into the Rharb Basin in southwest, the south-central Saiss Basin and the Guercif Basin in the northeast. These basins are filled by an upwards shallowing Late Miocene to Pleistocene sequence, commencing with deep-water pelagic sediments which are followed by alluvial-fluvial and finally lacustrine deposits. In the Rharb Basin, these sediments were deposited in

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This article includes 4 enclosures on 2 folded sheets.

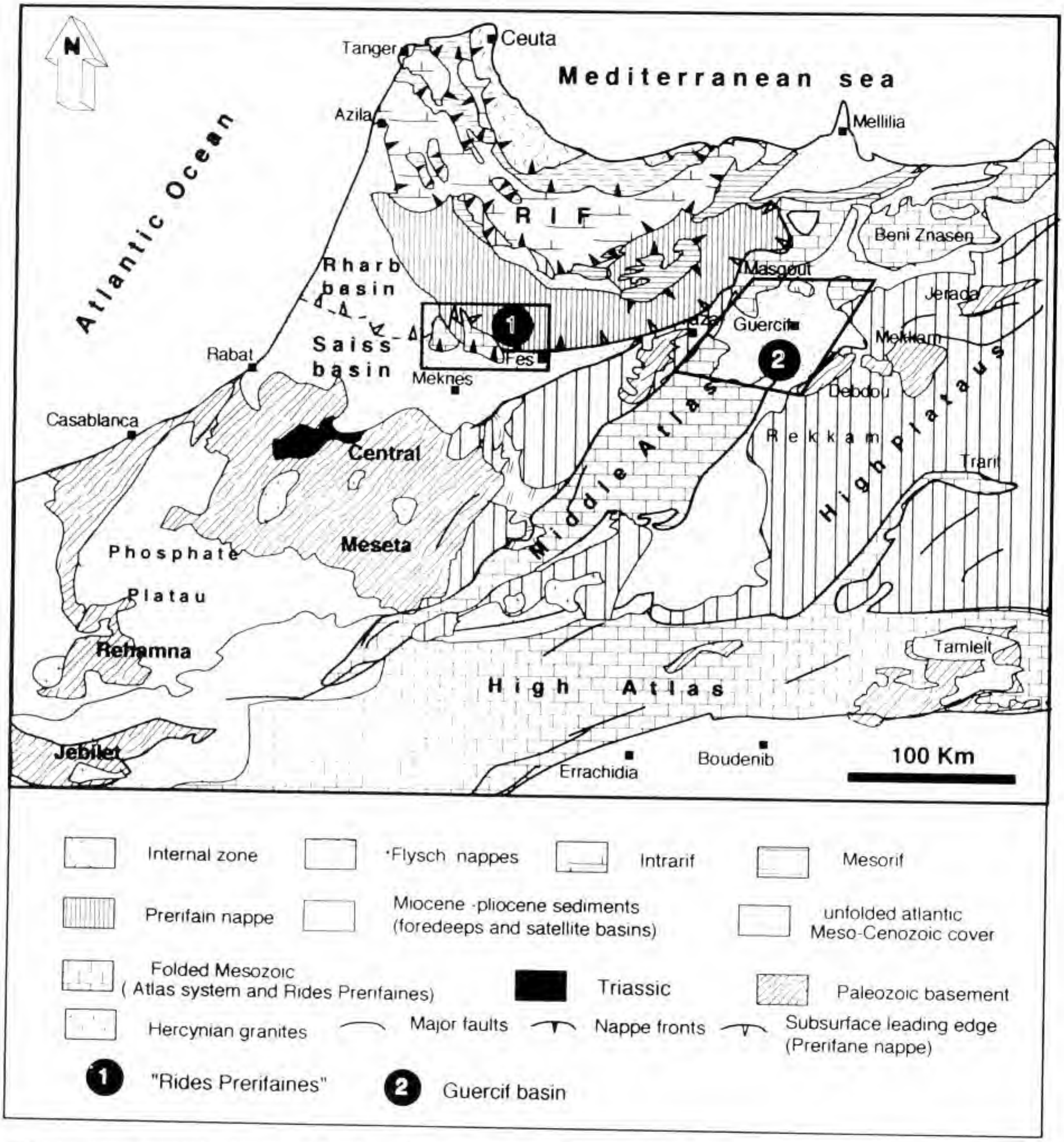


FIG. 1. Major tectonic element of Northern Morocco, showing location of study areas (modified after Michard, 1975)

extensional satellite basins on top of the chaotic "Nappe Prérifaine" (Flinch, 1993, this volume).

The "Rides Prérifaines", translated from the French as the "Fore-Rif ridges", form the mountainous terrain which extends from the Sidi Kacem to Fes. These mountains are upheld by folded and thrust-faulted shallow water Mesozoic carbonates which are capped by Cenozoic sediments and the chaotic Prérifaine nappe. Most authors interpreted the "Rides Prérifaines" as south-west, south and east verging thrust sheets (Sutter, 1980a and 1980b).

The southwest-northeast striking Middle Atlas Mountains are located to the southeast of the Neogene Rif foreland. These mountains developed in response to Late Cretaceous and Paleogene inversion of a system of deep Triassic to Jurassic half-grabens. To the southeast, the Middle Atlas is bordered by the High Plateau, a stable block which is characterized by a relatively thin Mesozoic sedimentary cover, consisting of Triassic-Early Lias redbeds and volcanics and Jurassic carbonates, which rests on peneplained Palaeozoic rocks. The Guercif Basin is a Late Neogene depression which is superimposed on the northeastern part of the inverted Middle Atlas Trough.

STRATIGRAPHIC FRAMEWORK

The Mesozoic and Cenozoic sediments of northwestern Morocco rest unconformably on the peneplained surface of the Hercynian basement. As evident from outcrops on the Moroccan Meseta, this basement consists of deformed Carboniferous, Devonian and Cambrian clastics and carbonates which are intruded by Hercynian granites (Piqué, 1982; Laville and Piqué, 1991). During latest Carboniferous and Permian times, accumulation of continental clastics in small intramontane basins was accompanied by the intrusion of acidic plutons (Cousminer and Manspeizer, 1977; Van Houten, 1977).

During the Triassic and Early Jurassic, the evolution of northern Morocco was dominated by rifting activity which was intimately related to the early phases of the Pangea breakup (Ziegler, 1988; Dercourt et al., 1993). Extensional tectonics, accompanied by the intrusion and extrusion of tholeiitic basalts, controlled the subsidence of the Middle and High Atlas Troughs and the external Rif system of grabens and half-grabens (Beauchamp, 1988; Favre et al., 1991; Laville and Piqué, 1991). Marine transgression entered these grabens during the Late Triassic, giving rise to the accumulation of a thick evaporitic series which laterally grades into continental red beds (Fig. 2). During the Early Jurassic, open marine conditions were established; whereas deeper water shales and carbonates accumulated in the continuously subsiding grabens, carbonate platforms developed on the graben flanks (Favre and Stampfli, 1992). Triassic and Early Jurassic sediments range in thickness between 200 m and 2000 m.

With the early Middle Jurassic onset of sea floor spreading in the Central Atlantic (Emery and Uchupi, 1984), rifting activity ceased in Morocco. However, continued tectonic activity, resulting in the subsidence of small transtensional basins, must be related to the sinistral translation of Africa relative to Europe in response to progressive opening of the Central Atlantic and the Western Tethys (Ziegler, 1988; Laville and Piqué, 1991; Dercourt et al., 1993). During the Middle and Late Jurassic times, the grabens of Morocco were gradually filled in with clastics derived from southern sources and later by platform carbonates, as evident by the stratigraphic record of the Middle Atlas, Guercif and the external Rif basins (Fig. 2).

With the Late Senonian onset of counterclockwise convergence of Africa-Arabia with Eurasia, the Alboran-Kabylia Block began to move westwards with respect to North Africa and started to converge with Iberia and northwestern Africa (Wildi, 1983; Ziegler, 1988, 1990). Paleocene-Early Eocene collision of the Alboran Block with the Moroccan Tethys margin was accompanied by the development of an accretionary wedge, corresponding to the Rif flysch nappes, the subsidence of the Rif foreland basin and inversion of the Middle Atlas Trough.

In the distal parts of the Rif foreland basin, corresponding to the domain of the "Rides Prérifaines", Middle Miocene (Langian-Serravalian) shallow water carbonate and clastic sediments transgressed over truncated Mesozoic strata; in

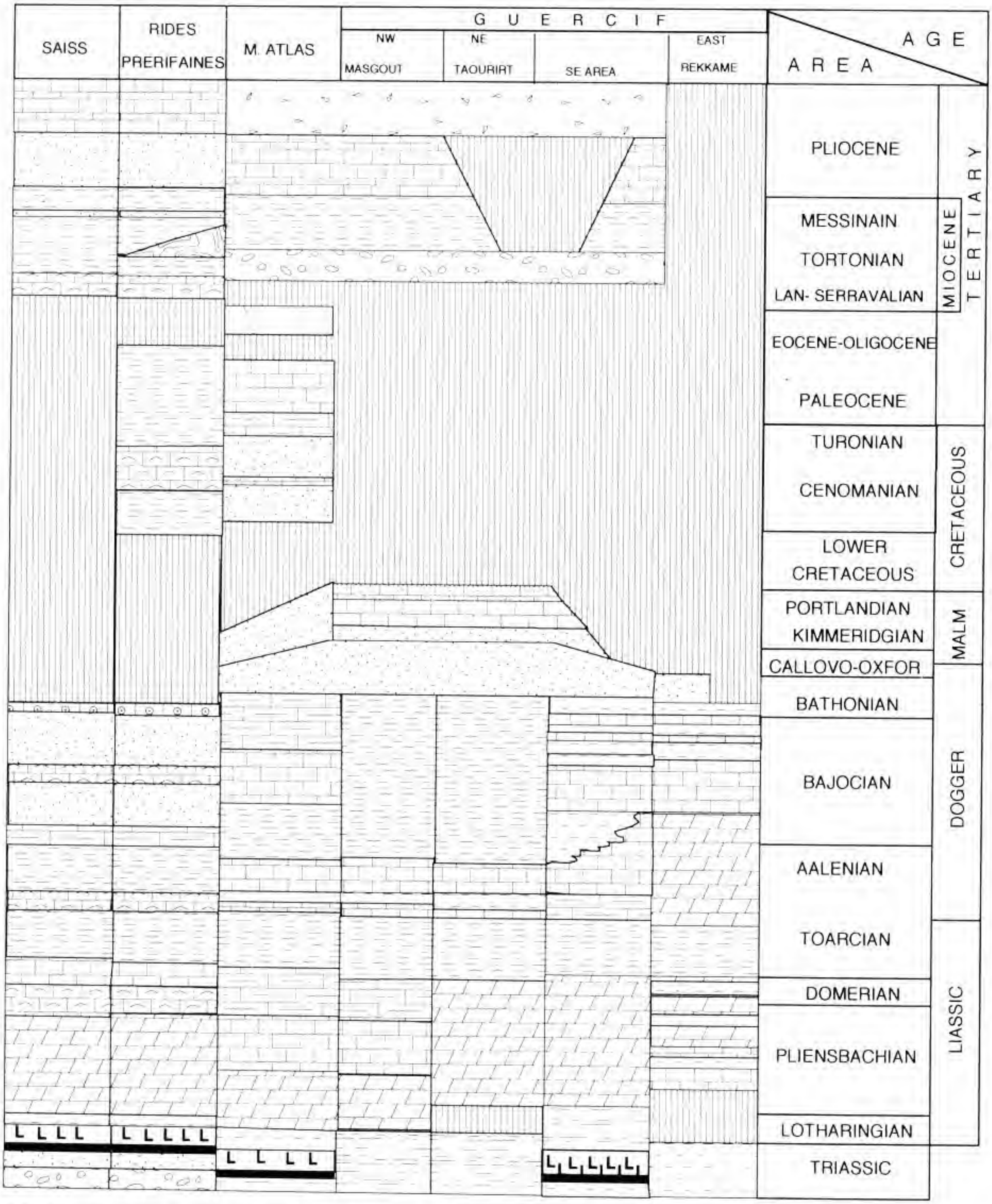


FIG. 2. Chronostratigraphy of post-Hercynian series of Northern Morocco ("Rides Prerifaines", Middle Atlas and Guercif Basin).

turn, these are covered by Tortonian basinal marls. During their accumulation, the Prérifaine nappe was emplaced. This nappe consists of an up 2000 m thick chaotic assemblage of blocks, varying in size from 1 m to 100 m, embedded in Tortonian marls. Most common components are Triassic red beds, evaporites and volcanics. In turn, the Prérifaine nappe is covered by Tortonian-Messinian blue marls which grade upwards into the Early Pliocene epicontinental sands and Late Pliocene lacustrine limestones. Accumulation of this post-nappe sequence, which attains thicknesses of up to 1500 m, was contemporaneous with Messinian-Pliocene foreland compressional phases. Quaternary travertines, conglomerates, and yellowish sands of presumably Villafranchian age, rest disconformably on truncated Pliocene strata.

During these late deformation phases, Triassic salt provided a regional detachment level. The Early Mesozoic rift geometry and the distribution of the Triassic salt played an important role in guiding the geometry of the developing thrust structures.

In the Guercif Basin, Tortonian to Pliocene alluvial, deltaic and coastal sediments, reaching a thickness of up to 2000 m, were deposited under a tensional regime during Tortonian times and under a compressional regime in Pliocene times. This basin is superimposed on the northeastern parts of the inverted Middle Atlas Trough.

This paper integrates the surface geology of the "Rides Prérifaines" and the Guercif Basin with reflection-seismic data imaging the subsurface structures and documenting the larger scale geometry and the areal extend of these features. Our interpretation are based on seismic grids which cover the entire "Rides Prérifaines" and the Guercif Basin. Selected lines of these grids and their interpretation are provided by Enclosures 1 to 4.

STRUCTURE AND EVOLUTION THE "RIDES PRÉRIFAINES"

Surface geological studies of the "Rides Prérifaines" date back to the late 1920's. Daguin (1927), Levy and Tilloy (1952), Durand Delga et al. (1960, 1962) and Sutter (1980a and 1980b) all suggested a compressional origin for the "Rides Prérifaines". On the other hand, Faugère (1978) proposed that these structures resulted from the interactions of two basement-involving strike-slip systems, located along the southern and western margin of the "Rides Prérifaines". According to this author, movements along these faults were transmitted to the Mesozoic series which, due to their decoupling from the basement by the Triassic salt, display more complex structures. All authors agree that the formation of the "Rides Prérifaines" is of Messinian to Plio-Pleistocene age and post-dates the major tectonic phases which controlled the evolution of the Rif fold and thrust-belt.

Ait Brahim and Chotin (1984) carried out a microtectonic study of the "Rides Prérifaines" and identified four compressional phases, characterized by principal horizontal compressional stress trajectories changing from NW-SE during pre- Miocene times, to N-S during the Late Tortonian, to E-W during the Messinian and to NE-SW during the Plio-Pleistocene.

Figure 3 provides a tectonic map of the "Rides Prérifaines", showing their main structural elements and the location of reflection-seismic lines discussed below. The autochthonous foreland of the Saiss plain lies to the south of the "Rides Prérifaines". The northeastern parts of the "Rides Prérifaines" are overridden by the chaotic Prérifaine nappe on which the Late Miocene to Pleistocene post-nappe satellite basins subsided (Flinch, 1993, this volume).

The seismic profiles, given in Enclosures 1 and 2, show that the dominantly NE-SW and E-W trending surface structures are superimposed on Early Mesozoic extensional fault systems. Seismic line P-12 (Enclosure 1) crosses the "Rides Prérifaines" in an east-westerly direction and covers the eastern part of the Rharb Basin, the Bou Draa, Tselfat and Mesrana anticlines and the northern parts of the Nzala des Oudayas structure. This profile shows that the "Rides Prérifaines" consist of two large, superimposed thrust sheets, namely the Prérifaine nappe and the Late Miocene-Pliocene thrust belt which involves the sedimentary fill of the Mesozoic grabens, the "Aquitanian-Burdigalian" (Languian-Serravalian, according to recent palaeontologic studies) foreland basin, as well as

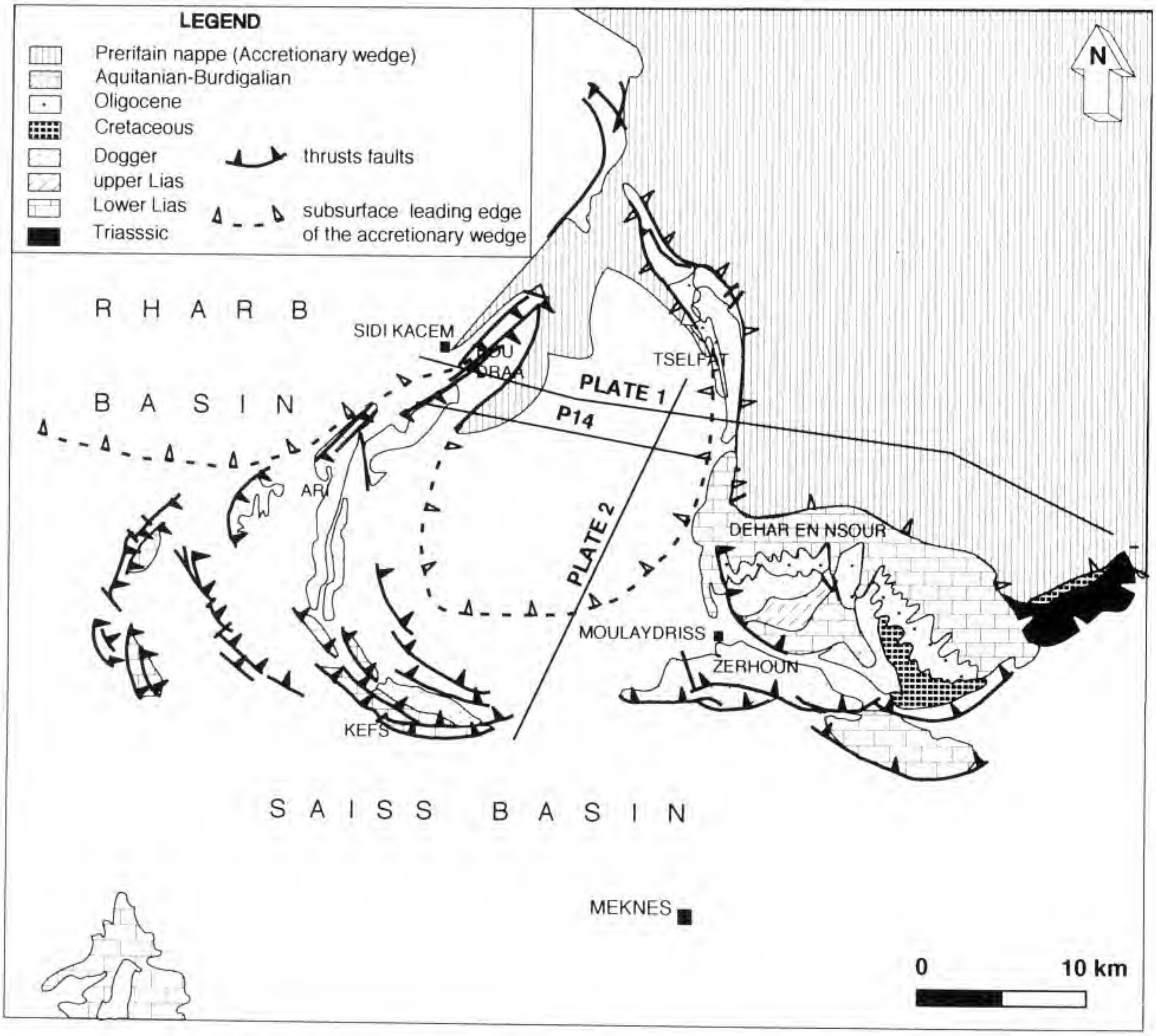


FIG. 3. Tectonic map of "Rides Prérifaines", showing location of seismic lines given in Enclosures 1 and 2 and Fig. 5.

the Prérifaine nappe. This late compressional deformation was contemporaneous with the deposition of the Late Miocene-Pliocene sedimentary series (line P-15, Enclosure 2). The entire sedimentary package, including the allochthonous accretionary wedge of the Prérifaine nappe, is detached from the autochthonous basement at the level of Triassic evaporites.

The thrusted Bou Draa and Tselfat ridges are superimposed on Triassic normal faults, offsetting the top of the basement. The broad and the gentle Mesarna structure, defined at intra-Jurassic levels, is associated with lateral thicknesses changes which are more pronounced at Toarcian and Aaleno-Bajocian levels than in the Domerian; these thickness changes are interpreted as reflecting intra-Jurassic salt movements. However, as the base of the Prérifaine nappe is also deformed, a Neogene growth component can be postulated for the Mesarna structure. To the South of line P-12, where Triassic salts are involved in this structure, unconformities within the Late Mio-Pliocene sediments give good evidences for reactivation of salt movements at the time of the development of the thrusted Bou Draa and Tselfat structures.

The northwest-verging the Bou Draa structure is evidently associated with a sharp increase in thickness of the Triassic-Jurassic sediments across a deep seated basement fault. This fault is part of the Sidi Fili fault system which, decades ago, had been identified by petroleum geologists.

The seismic profile P-15 (Enclosure 2), which

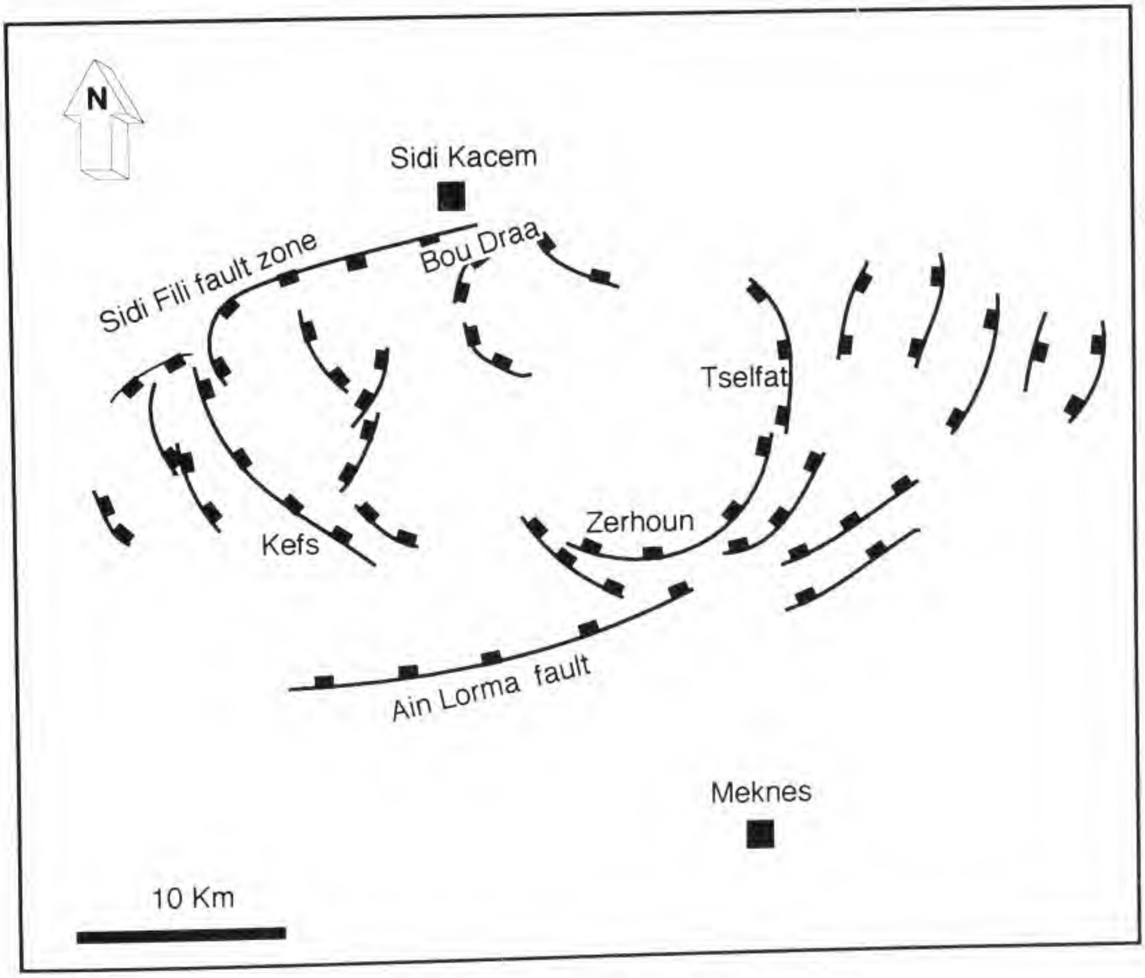
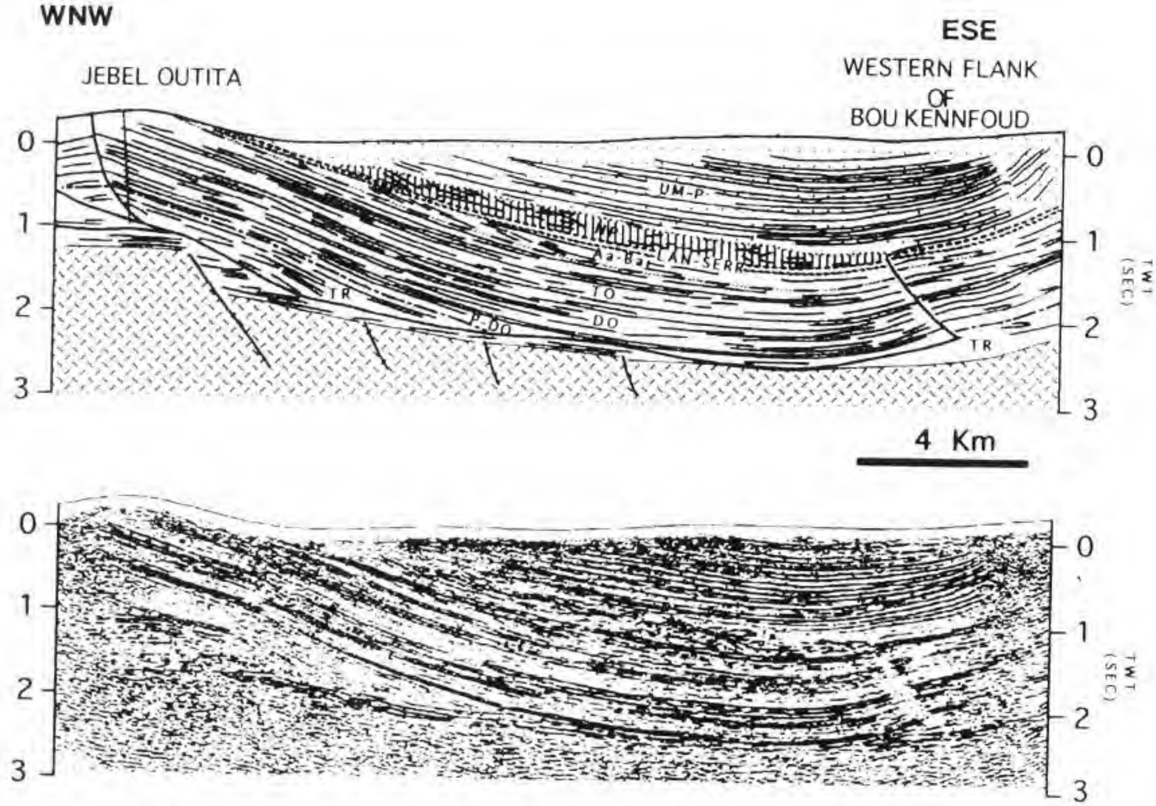


FIG. 4. Triassic-Jurassic normal fault pattern of "Rides Prérifaines", based on reflection-seismic data.



ABREVIATIONS:
TR : TRIASSIC ; P-DO : PREDOMERIAN ; DO : DOMERIAN ; TO : TOARCIAN
Aa-Baj : AALENO- BAJOCIAN ; LAN-SERR : LANGHIAN- SERRAVALIAN
NP : NAPPE PRERIFAINE ; UM-P : UPPER MIOCENE-PLIOCENE.

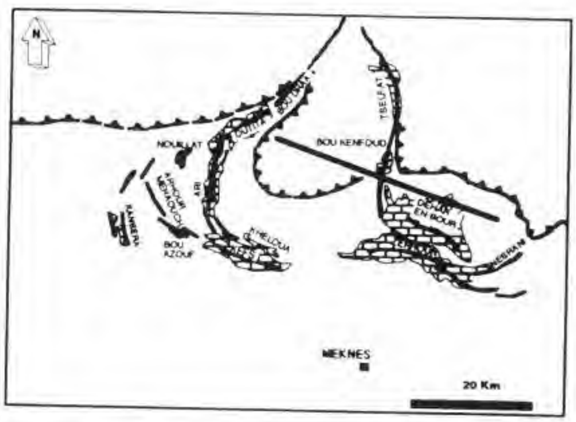


FIG. 5. E-W profile throgh "Rides Prérifaines", illustrating block-faulted basement, sub-horizontal reflectors beneath intra-salt decollement horizon, progradation to the east of the Aaleno-Bajocian sequence, convergence of Late Miocene-Pliocene sequence towards the western flank of the Bou Kennfoud structure.

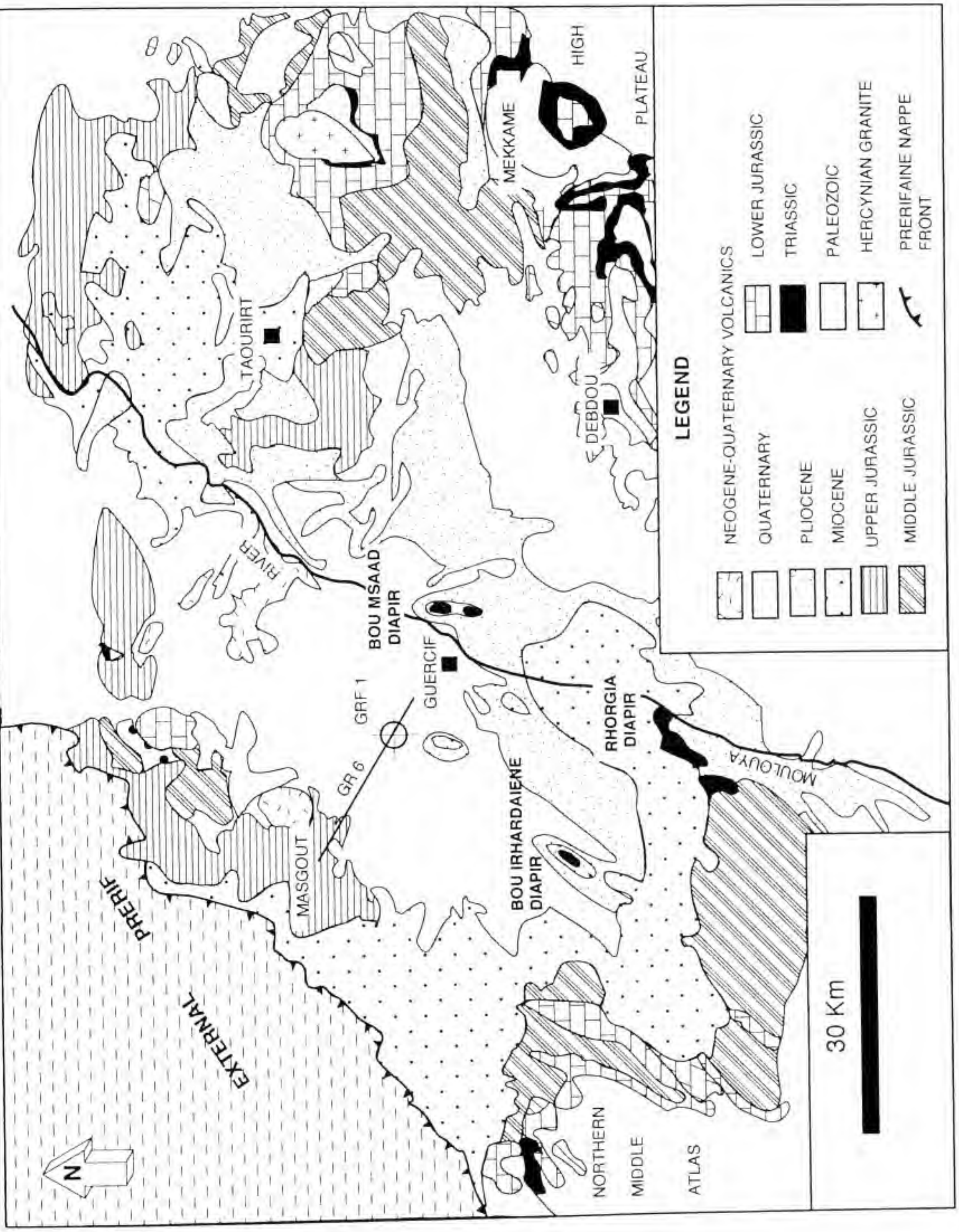


FIG. 6. Geological map of Guercif Basin, showing location of outcropping diapirs and seismic line given in Fig. 7.

crosses the "Rides Prérifaines" in a northeasterly direction, shows that also the southerly verging thrust faults, which carry the Kefs and Kheloua structures, ramp up from the Triassic evaporites through Jurassic strata above a set of normal basement faults. Also the largely salt induced Nzala des Oudaya structures are associated with Triassic normal faults offsetting the basement (Enclosure 1).

The Triassic-Jurassic fault map, given in Fig. 4, is based on an interpretation of the entire seismic grid. It illustrate that the Early Mesozoic fault system consists of west, east and north hading normal faults, the length of which varies between 2 km and 20 km. The Sidi Fili fault trend defines northwestern margin of the Triassic-Early Jurassic "Rides Prérifaines" half graben system.

On seismic lines, the Prérifaine nappe is characterized by discontinuous to chaotic reflectivity, abounding with diffractions (Fig. 5). Its southern termination is clearly evident on the line P-15 (Enclosure 2); this allochthonous body is laterally offset and onlapped by Tortonian series, characterized by parallel and laterally continuous reflectors. The base of the nappe corresponds to the smooth surface which forms the top of the "Aquitanian-Burdigalian" reflectors. This clearly shows that this allochthonous unit was emplaced in post-Middle Miocene and pre-Late Miocene times (Flinch, 1993, this volume), that is, during a relatively short time span. Emplacement of this nappe was accompanied by drowning out of the foreland platform and its rapid subsidence to considerable water depth.

The Late Miocene to Pliocene series, which covers the Prérifaine nappe, is characterized parallel to sub-parallel reflectors which generally converge towards the top of "Rides Prérifaines" anticlines (Enclosure 2). Significant syn-depositional deformation is indicated by the convergence of reflectors and the presence of unconformities within this Mio-Pliocene series, as shown by the growth of the syncline away from the anticlines (Enclosure 2). Moreover, the thrust fault carrying the Bou Draa structure (Enclosure 1) clearly cuts through the Prérifaine nappe and the Late Mio-Pliocene series. Therefore, deformation of the "Rides Prérifaines" clearly post-dates the emplacement of the Prérifaine nappe. This was already evident from surface geology (e.g. Levy and Tilloy, 1952; Sutter, 1980).

The northerly striking structures of "Rides Prérifaines" are generally associated with preexisting salt pillows, some of which are superimposed on normal basement faults. Line P-12 and P-15 show that during the Late Miocene-Pliocene deformation of the "Rides Prérifaines" the Mesozoic and younger series were decoupled from the basement at the level of Triassic salts. Although there is no evidence for the reactivation of normal faults affecting the basement, their intra sedimentary part was clearly reactivated and played an important role in localizing the deformation of structures which now form the "Rides Prérifaines" (Fig. 5). As such, the entire system of the "Rides Prérifaines" must be considered as a thin-skinned thrust belt which partly scooped out the sedimentary fill of the Triassic-Early Jurassic Prérifaine grabens. The thrusted folds forming the western and eastern structures correspond to lateral ramps whereas the southern structures form the frontal ramps of this thin-skinned thrust belt.

EVOLUTION OF THE GUERCIF BASIN

The Guercif Basin contains up to 2000 m of Tortonian to Pliocene sediments and is superimposed on the northeastern parts of the Early Mesozoic Middle Atlas Trough which was inverted during Paleogene times (Figs. 6 and 7).

Development of the Middle Atlas Mountains is generally thought to result from the sinistral transpressional deformation from an Early Mesozoic rifted basin (Mattauer et al., 1977; Jacobshagen, 1988; Fedan, 1988; Boccaletti et al., 1990; Bernini et al., 1994). Choubert and Faure-Muret (1962) visualize Late Eocene, Late Oligocene and Middle Miocene compressional phases whereas du Dresnay (1988) recognized Late Senonian precursor events followed by a major Late Eocene phase of basin inversion.

The tectonic history of the Neogene Guercif Basin was studied by Colletta (1977) who identified a Late Tortonian to Messinian extensional phase, a late Pliocene compressional episode and possibly renewed extension during the Quaternary.

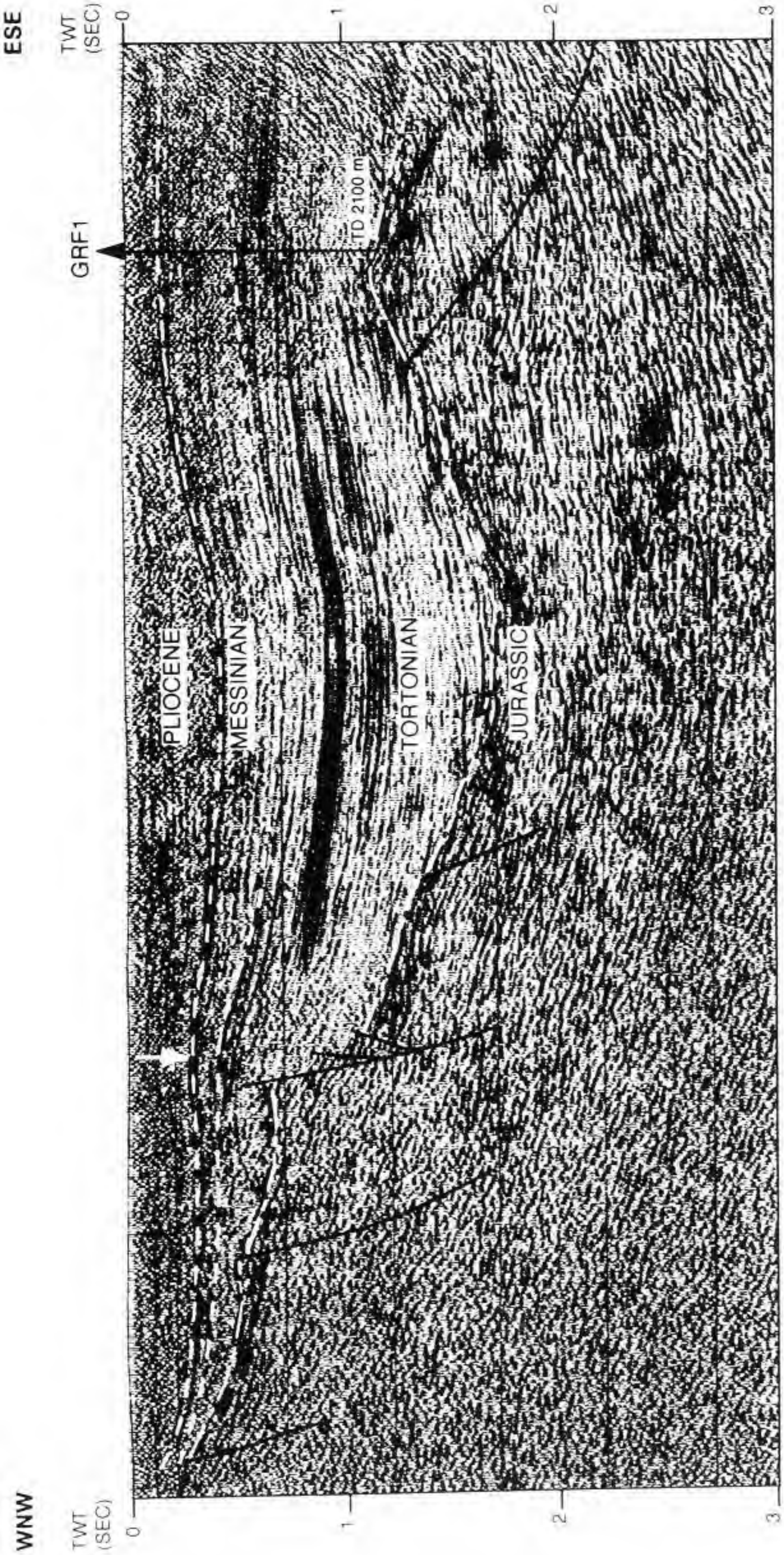


FIG. 7. Reflection-seismic line through Guercif Basin, illustrating Tortonian extension and Pliocene compression. Note minor inversion of Tortonian graben below white arrow. (for location see Fig. 6).

Mokhtari (1990), Boccaletti et al. (1990) and Bernini et al. (1994) all interpreted the Guercif Basin structures in terms of flower structures. Bally (1992) described the structures of the Guercif Basin as mini-inversion structures, involving the reactivation of Mesozoic extensional faults during pre-Tortonian inversion movements, however, without fully compensating their Mesozoic offsets.

The following discussion on the evolution of the Guercif area is based on the analysis of a grid of reflection-seismic profiles which are partly calibrated by wells.

Lack of well control impedes precise dating of the presumably Late Triassic-earliest Jurassic onset of the rifting activity in the northeastern parts of the Middle Atlas Trough (line G-17, Enclosure 3). Strong thickness changes across the fault limiting the Debdou platforms suggests that differential subsidence of the Middle Atlas half-graben persisted into early Late Jurassic times; diverging intra-Jurassic reflectors and a number of unconformities are taken as evidence for syn-sedimentary extension. An unconformity at the base of the Bathonian to Late Jurassic sequence indicates continued tectonic instability of the area.

Some salt-cored anticlines have been mapped in the western Guercif Basin (Fig. 6). Line G-5 (Enclosure 4) gives a good impression of these structures even though the seismic data are not very good. This section crosses the outcropping Rhorgia diapir to the SW and the Bou Msaad diapir to the East. The presence of a third diapir between these two structures is suggested. Note the interpreted thinning of the lower Liassic-Domerian interval towards the western domes and the thickening of the Bathonian-Portlandian section along the southwestern flank of the central salt structure. The thinning of the Early Jurassic intervals can be interpreted as reflecting the pillow phase of the diapirs which was followed by salt evacuation during the Bathonian-Portlandian, the rise of the diapir and the formation of a rim syncline during the Late Jurassic. Therefore, it is concluded that halokinetic movements were initiated during the Early Jurassic, resulting in the formation of salt pillows, and culminated by the end of the Jurassic with the rise of the diapirs and the development of rim synclines.

In the area of the Guercif Basin, a major hiatus spans Cretaceous to Middle Miocene times. However, in the southeastern part of the Middle Atlas, Cenomanian to Maastrichtian sediments are preserved (Fig. 2). Enclosure 3 clearly illustrated that the Mesozoic half graben, which underlies the Neogene Guercif Basin, was only mildly inverted prior to the transgression of the Tortonian and younger series. During these pre-Neogene inversion movements, the basement fault bounding the Debdou platform was reactivated; transpressional movements along this fault gave rise to the development of an anticlinal structure in the hanging-wall block, causing uplift and erosion of its Bathonian to Portlandian sedimentary cover.

Surface geological data from the southeastern Middle Atlas indicates the occurrence of pre-Maastrichtian inversion events (du Dresnay, 1988), which were followed by the commonly accepted Late Eocene event (Michard, 1976; Robillard, 1979; Ziegler, 1988). Therefore, it is reasonable to assume that the main inversion of the northeastern part of the Middle Atlas Trough had also occurred during the Late Senonian to Late Eocene time span. However, the extent to which this area had been covered by Late Cretaceous sediments prior to its inversions is unknown.

Subsidence of the Guercif Basin commenced with the transgression of Tortonian strata over truncated Jurassic series and persisted, under regressive conditions, through Messinian into Pliocene times. Basal conglomerates are followed by lacustrine shale and carbonates which are capped by Pliocene continental sands (Fig. 2).

The seismic profiles across the Guercif Basin, given in Fig. 7, show that its Tortonian subsidence was governed by extensional tectonics. As Messinian and Pliocene strata are not affected by tensional faults, rifting activity must have been of relatively short duration. The reflection configuration of Messinian strata indicates that they were deposited during a tectonically quiescent period. Convergence of Pliocene reflectors over structures, such as the one drilled by the well GRF1 (Fig. 7) and the Safsafat anticline, indicate that a compressional regime dominated the Pliocene evolution of the basin. During the Pliocene partial inversion of the basin, some of the Tortonian extensional faults were apparently compressionally reactivated. During this late compressional phase, the diapirs were reactivated to the extent that they now form outcropping elongated folds.

It is concluded that the area occupied by the Neogene Guercif Basin has undergone a plyphase evolution. Triassic to Early Jurassic crustal extension governed the subsidence of the Middle Atlas Trough. Tectonic instability continued during Middle and Late Jurassic times. Transpressional deformation of the Middle Atlas Trough commenced during the Late Senonian and culminated in its Eocene inversion. Tortonian extension governed the subsidence of the Guercif Basin. Messinian strata were deposited under a quiescent regime. Remobilisation of the Triassic salts under the overburden of thick Neogene sediments gave rise to the development of diapirs and domal structures. These features were modified during the Pliocene compressional phase. Pliocene compressional structures strike NE-SW and N-S.

CONCLUSIONS AND IMPLICATIONS FOR HYDROCARBON EXPLORATION

The Triassic-Jurassic grabens of northern Morocco form an integral part of the Western Tethys and North and Central Atlantic rift system. During the Alpine orogeny these grabens were inverted to various degrees. Eocene inversion of the Middle Atlas rift must be related to collisional coupling between the evolving Rif orogen and its foreland. The emplacement of the Prérifaine nappe at the transition from the Middle to the Late Miocene was accompanied by rapid subsidence of a relatively narrow foreland basin. Late orogenic phases of foreland compression resulted in the destruction of this foreland basin. Eocene inversion structures of the Guercif Basin involve compressional reactivation of tensional basement faults. In contrast, there is no evidence for basement reactivation during the development of the Messinian-Pliocene "Rides Prérifaines"; these are characterized by a major detachment system which soles out in Triassic evaporites, deposited in an extensional basin. South-verging thrusts are associated with east- and west-verging lateral ramps. All structures of the "Rides Prérifaines" are strongly influenced by the configuration of the Triassic-Early Jurassic rifted basin, the sedimentary fill of which was partly scooped out by Messinian-Pliocene thrust faulting.

The "Rides Prérifaines" contain significant hydrocarbon accumulations, contained in Late Neogene structures. During the Messinian-Pliocene deformation of the "Rides Prérifaines", hydrocarbons contained in pre-existing salt induced structural traps, were partly destroyed, thus releasing sizable volumes of oil for the charge of the newly formed structural traps. Early salt induced structures, which retained part of their closure during the late phases of basin deformation, may still contain significant amounts of hydrocarbons.

In the Guercif Basin, four exploratory wells were drilled on Pliocene structures all failed to encounter hydrocarbons. Eocene inversion structures are more likely to contain hydrocarbon in Mesozoic reservoirs, provided Early Jurassic source-rocks have attained maturity, as attested by oil seeps found the Middle Atlas (e.g. Issouka oil seep). Triassic-Jurassic extensional tectonics controlled the distribution of reservoirs, seals, source-rocks in half grabens and migration pathways of hydrocarbon generated. Generally, the footwall is dominated by relatively shallow water facies, such as carbonate build-ups, sand shoals and slope turbidites. The hanging wall is marked by ramp-type margin facies, including sand shoals and reefs.

Palaeotectonic basin analyses, based on an integration of surface and subsurface geological data, as well as structural and seismostratigraphic analyses of reflection seismic data, can greatly advance the understanding of the evolution of foreland basins and their hydrocarbon habitat.

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Enclosures

Enclosure 1 Seismic line P-12, Rides Prérifaines

Enclosure 2 Seismic line P-15, Rides Prérifaines

Enclosure 3 Seismic line G-17, Guercif Basin

Enclosure 4 Seismic line G-5, Guercif Basin

The Valencia Trough: geological and geophysical constraints on basin formation models

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ABSTRACT

The Valencia Trough, located between the Spanish mainland and the Balearic Promontory, is one of the Cenozoic extensional basins of the Western Mediterranean which developed in a region of convergence between the European and African plates. The Valencia Trough began to subside during the late Oligocene-early Miocene along the Mediterranean side of Spain, coeval with a phase of compression which affected its southeastern margin, formed by the Balearic Promontory. This rifting phase was followed by a period of post-rift subsidence. During the Plio-Quaternary, extensional tectonics and minor volcanic activity resumed along the Spanish margin of the Trough. This phase of extension is, however, not recorded in the central parts of the basin. The wealth of geological and geophysical data collected from the Valencia Trough during the last decades provides a precise image of its crustal structure; however, its deep lithospheric configuration is still poorly known.

The Valencia Trough is characterized by a strongly attenuated continental crust, except in its northeasternmost part where oceanic crust is probably present. It is underlain by an anomalous low-velocity uppermost mantle, the lateral extent and thickness of which are still under debate. Kinematic models of basin development have been quite successful in describing some aspects of the central parts of the Trough, though not of the entire basin and particularly not for its southern region. As these models failed to integrate the complex plate interactions which governed the development of the Valencia Trough, they must be regarded as simplistic. The Valencia Trough hosts an oil province which is largely restricted to the Ebro delta.

INTRODUCTION

The Valencia Trough is a NE-SW oriented triangular shaped basin which is located between the Spanish mainland and the northeastern prolongation of the Betic Cordillera, the Balearic Promontory (Fig. 1). The continental crust underlying this trough was consolidated during the Hercynian orogeny. During the Mesozoic break-up of Pangea,

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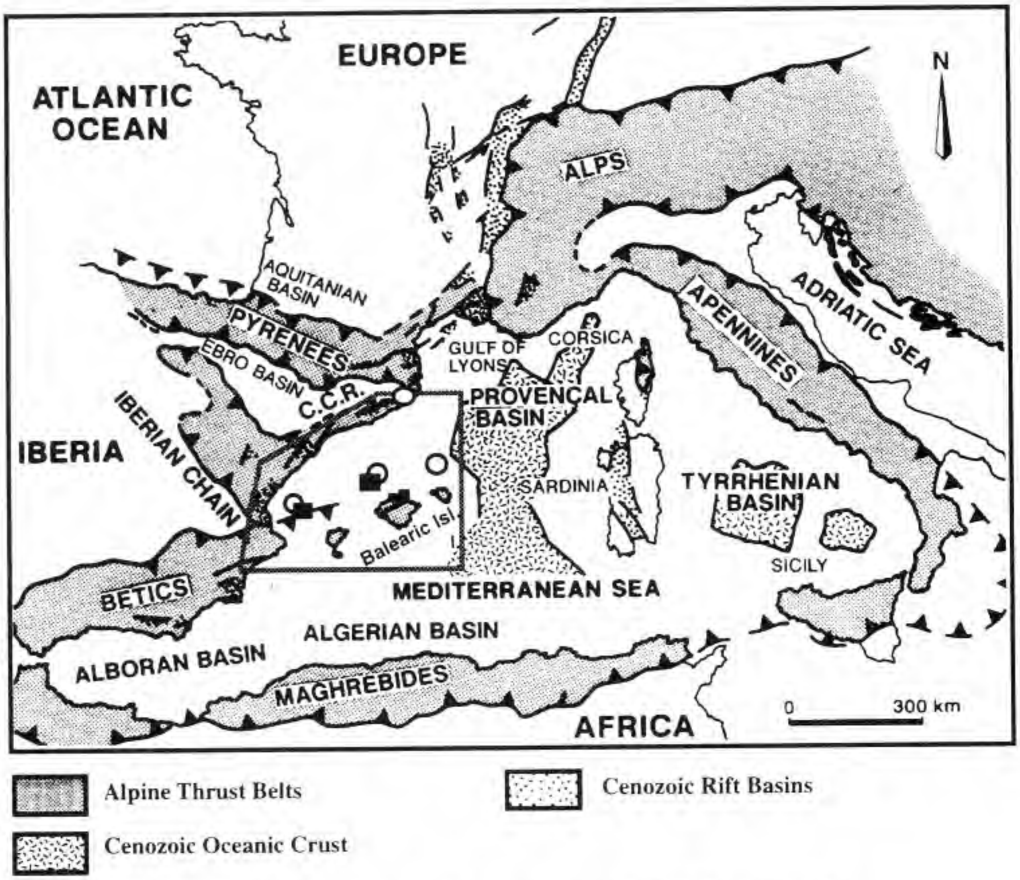

FIG. 1. Location map of the study area showing main geological features of the Western Mediterranean (modified after Banda and Santanach, 1992b). Valencia Trough is outlined by heavy grey line. C.C.R.: Catalan Coastal Ranges. Black squares: calcalkaline magmatism (30-15 Ma); Open circles: alkaline magmatism (15-0 Ma). (location of volcanic outcrops from Martí et al.,1992).

this crust underwent repeated extensional events, controlling the subsidence of graben structures. During the latest Cretaceous and Paleogene, intraplate compressional stresses caused inversion of the Iberian Chain, the Catalan Coastal Ranges and their off-shore equivalents (Fig. 1); during this process, Mesozoic extensional crustal thinning was apparently largely recovered.

The present day structure of the Valencia Trough is thought to result from a late Oligocene-early Miocene rifting event which was coeval with a phase of compression confined to the southeastern margin of the basin, formed by the Balearic Promontory. This extensional phase affected mainly the northwestern part of the basin, as evident by the development of a series of horst and graben structures, bounded by ENE-WSW trending normal faults. Whereas the horsts are held up by a variety of Palaeozoic and Mesozoic rocks, the grabens were filled by late Oligocene-early Miocene shales (Torres and Bois, 1993). In contrast, the Balearic part of the basin is characterized by a series of SE-NW trending thrusts and reverse faults which developed during a Late Oligocene to Middle Miocene compressional phase (Roca and Desegaulx, 1992). Subsequently, these compressional structures were tensionally reactivated, resulting in the development of horst and graben structures, as seen on the Island of Mallorca (Fig. 1). For further details the reader is referred to Banda and Santanach (1992a).

The Valencia Trough has been the target of extensive geological and geophysical investigations by the petroleum industry and academic institutions (Fig. 2). During the 1970´s and 1980´s, the Catalonian shelf was intensely explored for hydrocarbons. The recording of extensive reflection-seismic surveys and the drilling of over 90 wells (Fig. 2a) resulted in the discovery of a number of small and one large oil fields, having cumulative ultimate recoverable reserved of the order of 250 to 300 x 106 bbls of oil. Most of these accumulations are contained in extensional fault blocks and buried hills, upheld by karstified Mesozoic carbonates. These structures are sealed by middle Miocene basinal shales. Hydrocarbon charge is provided by Mesozoic source-rocks as well as by late Oligocene-early Miocene shales which were deposited during the early rifting stage of the Valencia Trough. These data sets provide valuable information on the sedimentary record of the Valencia Trough and its subsidence and thermal evolution.

The first academic geophysical experiments commenced in the early 70´s with the acquisition of two seismic refraction profiles, located North and South of the Island of Mallorca (Hinz, 1972; Gobert et al., 1972), followed up by a seismic refraction/wide-angle reflection experiment along the Balearic Promontory (Banda et al., 1980). During 1988 the VALSIS experiments were carried out to determine the lithospheric configuration of the basin. During a first cruise (VALSIS-I) heat-flow measurements were acquired along four transects crossing the axis of the Trough (Fig. 2b; Foucher et al., 1992). During a second cruise (VALSIS-II), twelve deep multichannel seismic-reflection profiles (CDP), eight wide-aperture multichannel profiles (COP) and six expanded spread profiles (ESP) were recorded (Figs. 2c and 2d). Marine seismic data were complemented by land recording of shots (Gallart et al., 1990). An additional refraction/wide-angle reflection experiment was carried out during the summer of 1989 along three profiles crossing the basin (P-I, P-II, and P-III of Fig. 2d; Dañobeitia et al., 1992). For a detailed discussion of CDP results data the reader is referred to Mauffret et al. (1992), Maillard et al. (1992) and Torné et al. (1992). The ESP results are discussed by Pascal et al. (1992) and Torné et al. (1992) and the wide-angle data by Gallart et al. (1990) and Dañobeitia et al. (1992). COP results were presented by Collier et al. (1994). In 1992, the crustal structure of the Valencia Trough was investigated by coincident steep and wide-angle reflection data (Fig. 2c) under the auspices of the Spanish Estudios Sísmicos de la Corteza Ibérica (ESCI) Program (Gallart et al, 1995; Vidal et al., 1995).

The results of these experiments show that the Valencia Trough is characterized by a strongly attenuated continental crust which is underlain by an anomalous low-velocity upper mantle (7.6 to 8.0 km/s), except in its northeasternmost part where oceanic crust is probably present. In the central part of the basin, the crust has a thickness of about 15-16 km; towards its margins it increases asymmetrically to average values of 20-22 km along the Iberian coast and about 24 km below Mallorca. Along the flanks of the basin, the upper and middle crust are characterized by an almost transparent layer with velocities ranging from 6.1 to 6.4 km/s. In contrast, the lower crust is variably reflective below the flanks of the basin and has velocities in the 6.4 to 6.9 km/s range, whereas it is almost absent under its axial parts.

Information on the configuration of the lithosphere-asthenosphere boundary comes primarily from modelling results. Modelling of the central part of the basin shows that the lithosphere thins towards the axis of the basin to values in the 60 to 65 km range (e.g., Watts and Torné 1992a; Zeyen and Fernàndez, 1994). Moreover, surface-wave studies favour thinning of the lithosphere towards the central part of the basin (Marillier and Mueller, 1985). There is no information, however, on the position of the lithosphere-asthenosphere boundary in the southern parts of the Trough, and also in the North at its transition to the oceanic Provençal basin (Fig. 1).

Direct evidence for volcanic activity comes from outcrops and exploration wells (Lanaja, 1987) and DSDP Site 123 (Ryan et al., 1972). Reflection-seismic and aeromagnetic data provide indirect evidence for additional volcanic centres (e.g., Mauffret, 1976; Maillard et al., 1992; Martí et al., 1992; Galdeano et al., 1974) (Fig. 1). Following Martí et al. (1992), Cenozoic magmatism in the area is characterized by two volcanic cycles which are clearly separated in time and by their petrology and tectonic setting. The first volcanic cycle (late Chattian-early Burdigalian) coincides

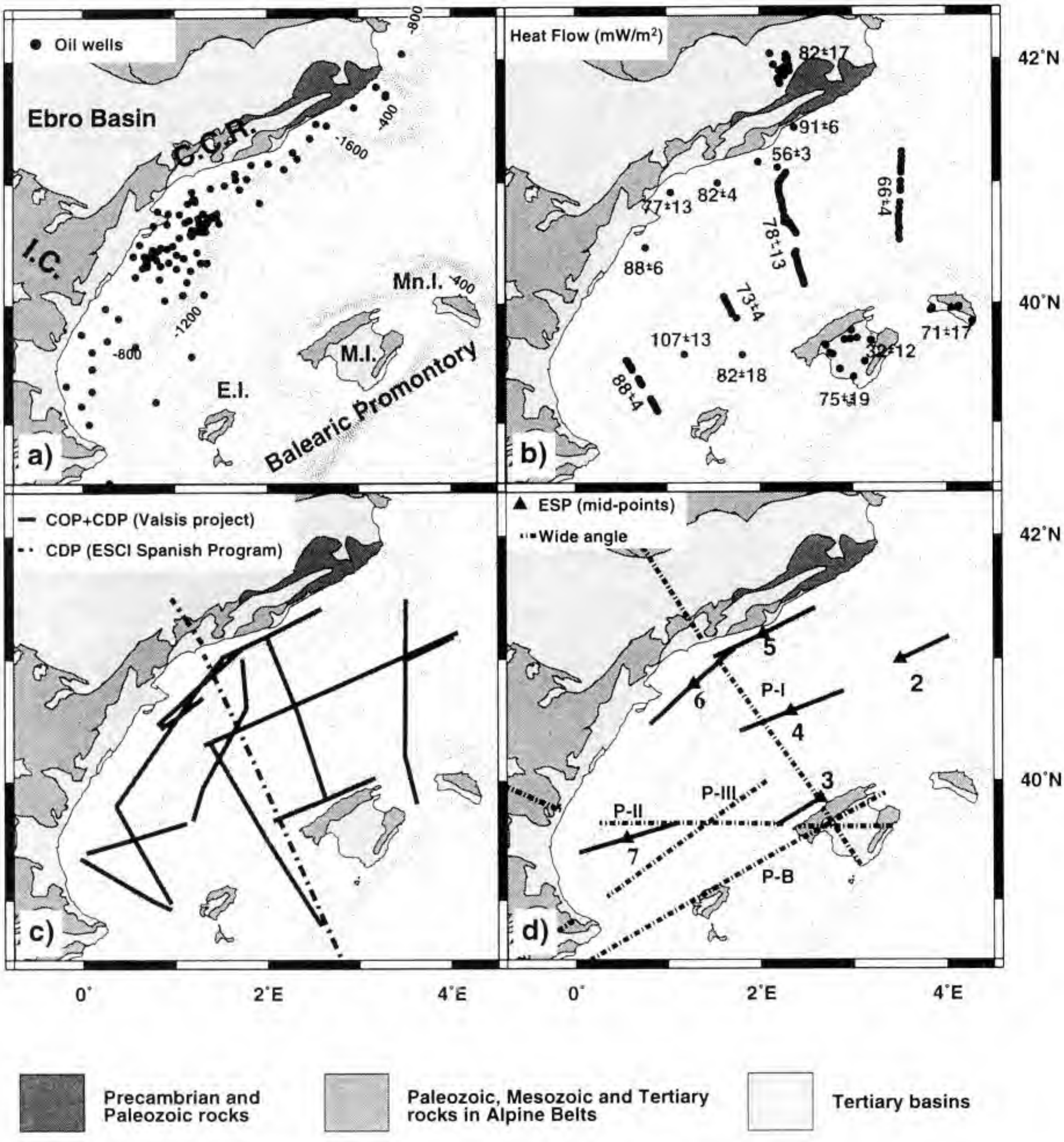

FIG. 2. **a**) Location of hydrocarbon exploration wells along the Ebro platform and Spanish margin. C.C.R.: Catalan Coastal Ranges; I.C.: Iberian Chain; E.I: Island of Eivissa; Mn.I.: Island of Menorca; M.I.: Island of Mallorca.
b) Heat flow determinations in mW/m^2.
c) Track lines of deep multichannel seismic profiling. Thick lines: Valsis-II CDP/COP profiles. Dashed-dotted line: ESCI coincident steep and wide-angle reflection profile.
d) Thick lines: Valsis-II ESP Profiles (triangles show mid-point locations). Dashed-dotted lines: wide-angle/refraction seismic profiles.

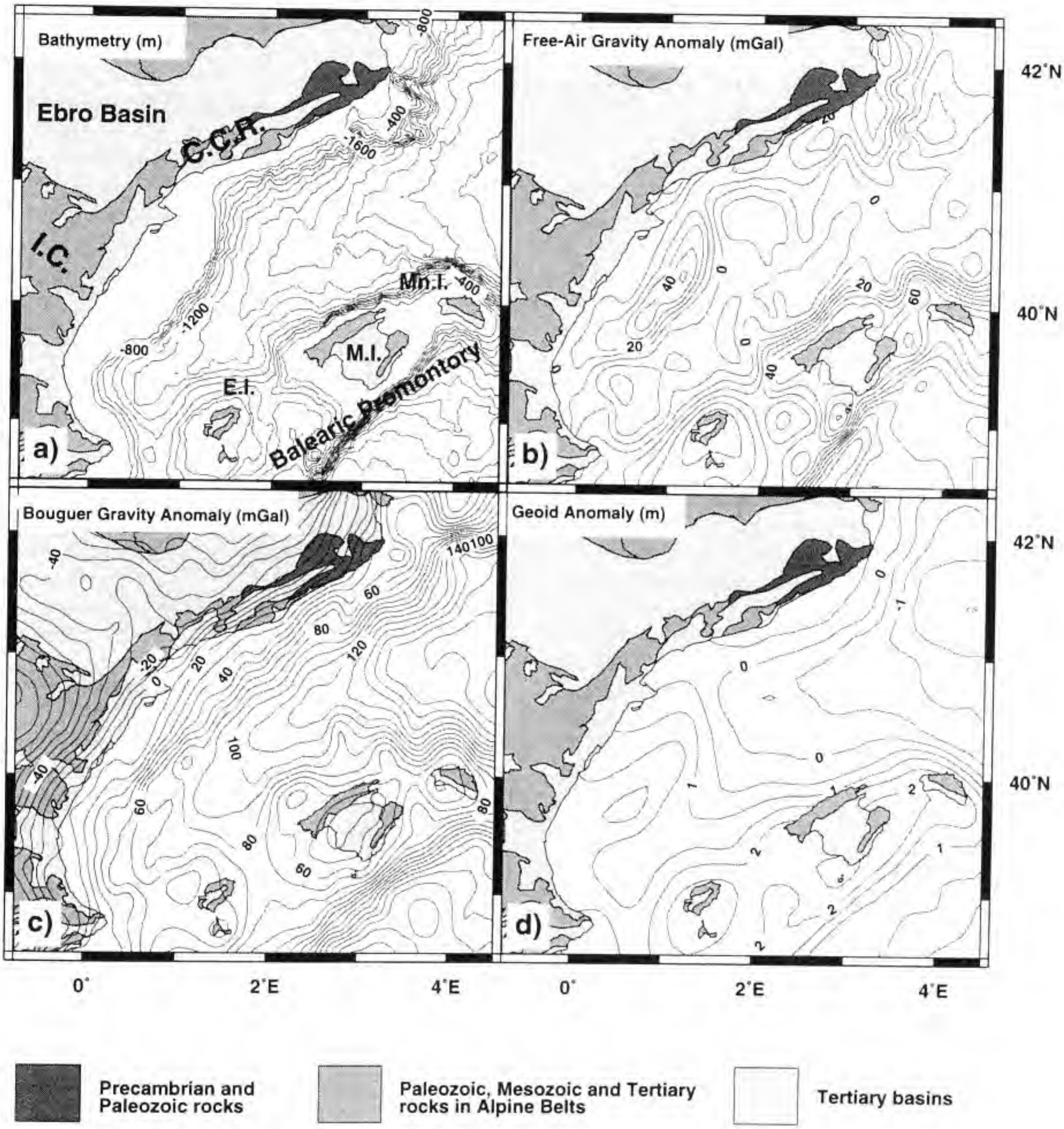

FIG. 3. **a)** Bathymetry of Valencia Trough, 200 m contour intervals.
b) Free-Air gravity anomaly map based on all available marine gravity data. Contour interval 10 mGal. Abbreviations as in Fig. 2.
c) Simplified Bouguer gravity anomaly map, contour interval 10 mGal (after Torné et al., 1992). Data on Iberia are based on Casas et al. (1987) and on Mallorca on IGME (1981).
d) Geoid anomaly map, contour interval 0.5 m. The map is based on GEOMED (e.g., Sevilla, 1992). A regional field based on the OSU91A model to degree and order 12 has been removed from the observed data prior gridding.

with the initial, main phase of rifting and is characterized by calcalkaline andesitic and pyroclastic rocks. The second cycle (Tortonian to Recent) is characterized by poorly differentiated alkali-basalts and was accompanied by a minor phase of extensional tectonics.

Paleomagnetic investigations suggest a 20^{o} early-middle Miocene clockwise rotations of the islands of Mallorca and Menorca relative to the Catalan Coastal Ranges, coinciding with the emplacement of the Betic-Balearic thrust sheets. A further late Miocene-Pliocene rotation of up to 20^{o} may be related to extensional faulting. However, it can not be excluded that part of these rotations are related to whole lithospheric movements (Parés et al., 1992). These authors point out that clockwise rotation in the Balearic Islands is in disagreement with most of the models proposed for the evolution of the Western Mediterranean, and that the total amount of rotation is too large to be accounted for by crustal extension in the Valencia Trough.

In summary, the various studies of the basin, using different geological and geophysical approaches, have resulted in a wide range of geodynamic hypotheses. These range from back-arc extensional mechanisms, related to northwestward subduction of oceanic lithosphere (e.g., Bocaletti and Guazzone, 1974; Mauffret, 1976; Banda and Channel, 1979; Cohen, 1980; Horvath and Berckhemer, 1982) to intra-continental rifting, related to horizontal movements of crustal blocks and seafloor spreading (e.g., Auzende et al., 1973; Rehault et al., 1985) and rift propagation from the Alpine forelands across the Alpine megasuture (Ziegler, 1988, 1992). Recently, Fontboté et al. (1989) and Roca and Desegaulx (1992) postulated that development of the Valencia Trough could be explained by foreland-type mechanisms, whereas Doblas and Oyarzun (1990) proposed that its origin is controlled by a major extensional detachment surface, cutting the crust and lithosphere, and upwelling of the asthenosphere.

In this paper we summarize the present-day crustal and lithospheric configuration of the Valencia Trough and focus on modelling results and difficulties and limitations which have been encountered when applying "classical" extensional models to explain the origin and evolution of this basin.

GEOPHYSICAL OBSERVATIONS

In this section we present a compilation of the various geophysical data sets which provide constraints on modelling the evolution of the Valencia Trough (Figs. 2 and 3). Figure 3a illustrates the triangular shape of this basin and its opening to the Northeast where water depths reach values of up to 2500 m, whereas in its southwestern parts they not exceed 1200 m. The axial trough is flanked by narrow shelves, except in the area of the Ebro delta (Fig. 3a).

Up to the VALSIS-I survey, information on the thermal regime of the area was limited to temperature gradients obtained from on-shore and offshore water and oil wells. These indicate that areas adjacent to the Trough are characterized by large temperature gradient variations, probably related to ground-water circulation (Fernàndez and Banda, 1989; Fernàndez et al., 1990). Similarly, heat-flow values for the eastern part of the Ebro basin (Cabal and Fernàndez, 1995) and the Spanish shelf (Negredo et al., 1995a) are also high variable and range from 55 to 90 mW/m^2. The islands of Mallorca and Menorca are characterized by background heat flow values of about 70-80 mW/m^2 and 75-90 mW/m^2, respectively (Fernàndez and Cabal, 1992). Results of the VALSIS-I survey (Foucher et al., 1992), which are not corrected for the thermal blanketing effect of sediments, indicate for the axial part of the basin heat-flow values varying from 88 mW/m^2 in the Southwest to 66 mW/m^2 in the Northeast at the transition to the Provençal basin, and thus demonstrate a significant decrease in heat-flow values towards deeper waters (Fig. 2b).

Free-Air gravity data (Fig. 3b) show that the axial parts of the Trough are dominated by values around 0 mGal whereas on its flanks maximum values of about 40 mGal are reached. This may suggest that the regional features of the study area are in local isostatic equilibrium; this is in accordance with various modelling results which show that the lithosphere has acquired little or no strength since rifting (e.g. Watts and Torné, 1992b; Zeyen and Fernàndez, 1994). Bouguer gravity data shows that the Trough is associated with a gravity anomaly high of about 100-150 mGal (Fig. 3c). 2D gravity modelling by Watts and Torné (1992a) shows that this gravity high can be explained by the combined effect of mantle and crustal thinning.

Figure 3d gives a map of geoid anomalies derived from GEOMED (e.g., Sevilla, 1992). Since the geoid is more sensitive to deep mass distribution than gravity, this can help in better deciphering the topography of the base of the lithosphere. Figure 3d shows that the Valencia Trough is associated with a relatively broad negative geoid anomaly which increases in magnitude toward the northeastern, whereas both flanks, and particularly the Balearic Promontory, are associated with positive geoid anomalies of up to 2 m. A first evaluation of this geoid anomaly low favours the hypothesis that the lithosphere is thinner under the central part of the Trough.

CRUSTAL AND LITHOSPHERIC STRUCTURE

In the following we discuss the sedimentary fill of the Valencia Trough, its upper and lower crustal configuration and the structure of the lithosphere/asthenosphere boundary.

Sedimentary Record and Basement Structure

Following Soler et al. (1983), the stratigraphic record of the Valencia Trough and surrounding areas is characterized by a major unconformity separating the Palaeozoic-Mesozoic basement from its Oligocene to Quaternary sedimentary cover (Fig. 4). On the Spanish shelf, Hercynian deformed Palaeozoic sedimentary and metamorphic rocks are uncomformably overlain by Permian and Mesozoic series, consisting of carbonate, siliciclastic and evaporite rocks (Torres and Bois, 1993). Rapid lateral thickness variations and facies changes indicate that these sediments accumulated in tensional basins which developed in conjunction with the opening of the Tethys Ocean (e.g. Ziegler, 1988; Fontboté et al., 1989; Maillard et al., 1992). On the islands of Mallorca and Eivissa (Fig. 2a), Middle Jurassic-Late Cretaceous series were developed in a slope and base of slope facies, reflecting their location along the Tethys passive margin (Banda and Santanach, 1992b).

The latest Cretaceous and Paleogene phase of intraplate compression was responsible for the inversion of the Mesozoic basins and uplift of the present-day off-shore parts of the Valencia Trough (Fig. 1). Inversion of the Iberian Chain was accompanied with the development of a series of NW-SE and E-W striking thrust faults (Guimerá, 1984; Guimerá and Alvaro, 1990) whereas the Catalan Coastal Ranges evolved in response to convergent wrench movements along NE-SW striking faults (Anadón et al., 1985). In the off-shore, Paleogene rocks are generally absent, except in the Barcelona graben (Bartrina et al., 1992) where continental or transitional deposits are present. In contrast, widespread Eocene to Oligocene lacustrine carbonates and lignites occur on the island of Mallorca (Ramos-Guerrero et al., 1989).

The Oligocene to Quaternary sedimentary fill of the Valencia Trough can be subdivided in four depositional sequences, separated by unconformities (Fig. 4b; García-Siñeriz et al., 1979; Soler et al., 1983; Anadón et al., 1989; Clavell and Berastegui, 1991).

On the western flank of the Trough, the lower sequence, spanning late Oligocene to Langhian times, consists of basal continental-transitional clastics and shales, which accumulated in fault-bounded shallow basins, and the transpressive early Miocene marine shales and carbonates of the Alcanar formation. The latter was deposited after the first rifting stage (Chattian-Aquitanian) and during the Betics compressional phase (Torres and Bois, 1993) and oversteps the block-faulted relief of the Valencia rift (Fig. 5). On the southeastern flank of the basin, Chattian-Aquitanian sediments consist of shallow-water carbonate and clastic rocks (Rodríguez-Perea, 1984; Anglada and Serra-Kiel, 1986) whereas Burdigalian and Langhian series consist of pelagic and calcareous turbidites that were deposited during the thrust deformation of the Balearic Promontory.

The second sequence spans Serravallian to early Messinian times. On the Balearic Promontory it is represented by carbonates deposited under a a)

b)

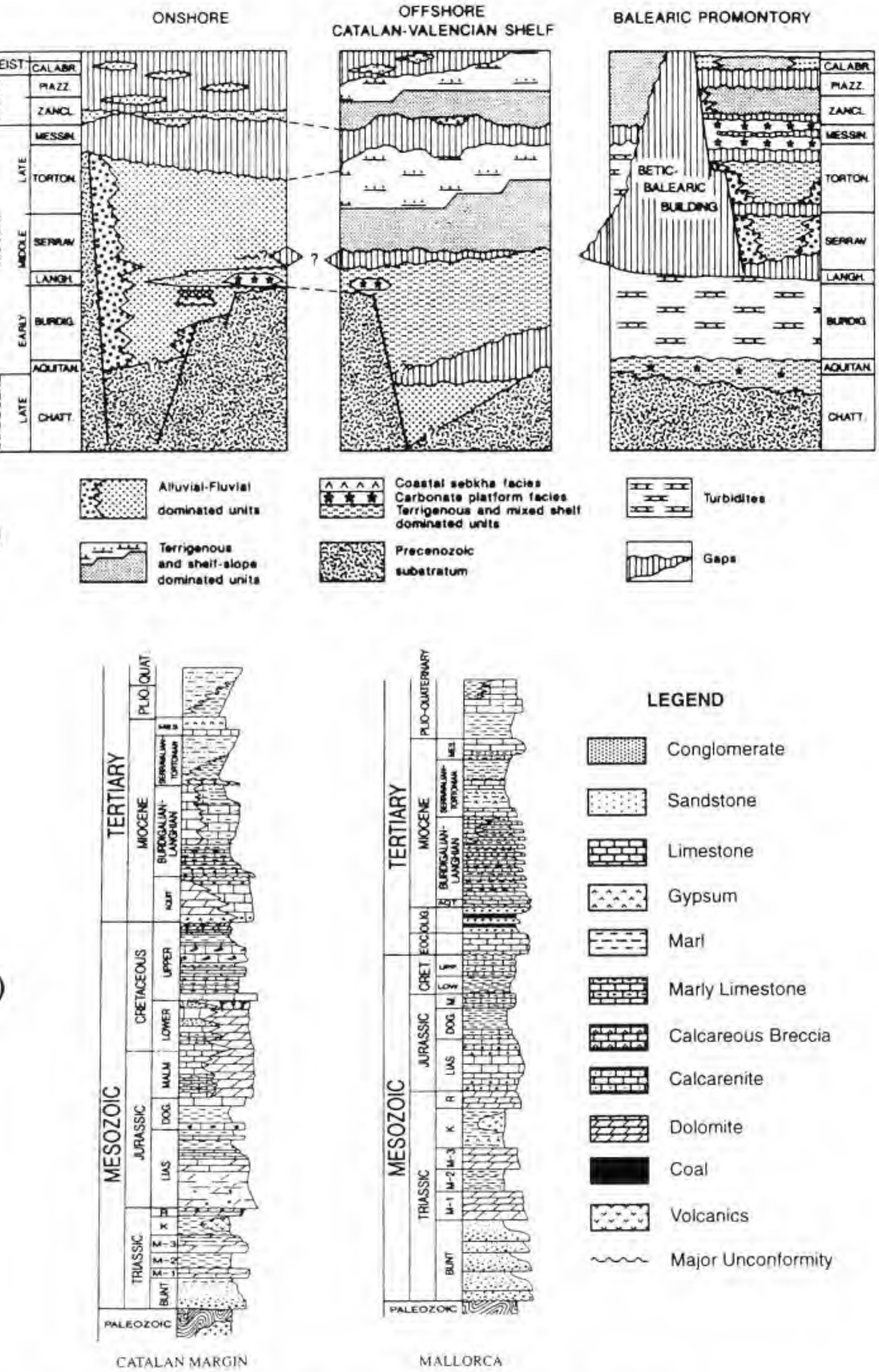

FIG. 4. **a)** Simplified stratigraphic charts of Valencia Trough region (after Banda and Santanach, 1992b). On-shore and off-shore columns after Bartrina et al. (1992).
b) Stratigraphic columns of the Spanish margin and Mallorca (after Torres and Bois, 1993).

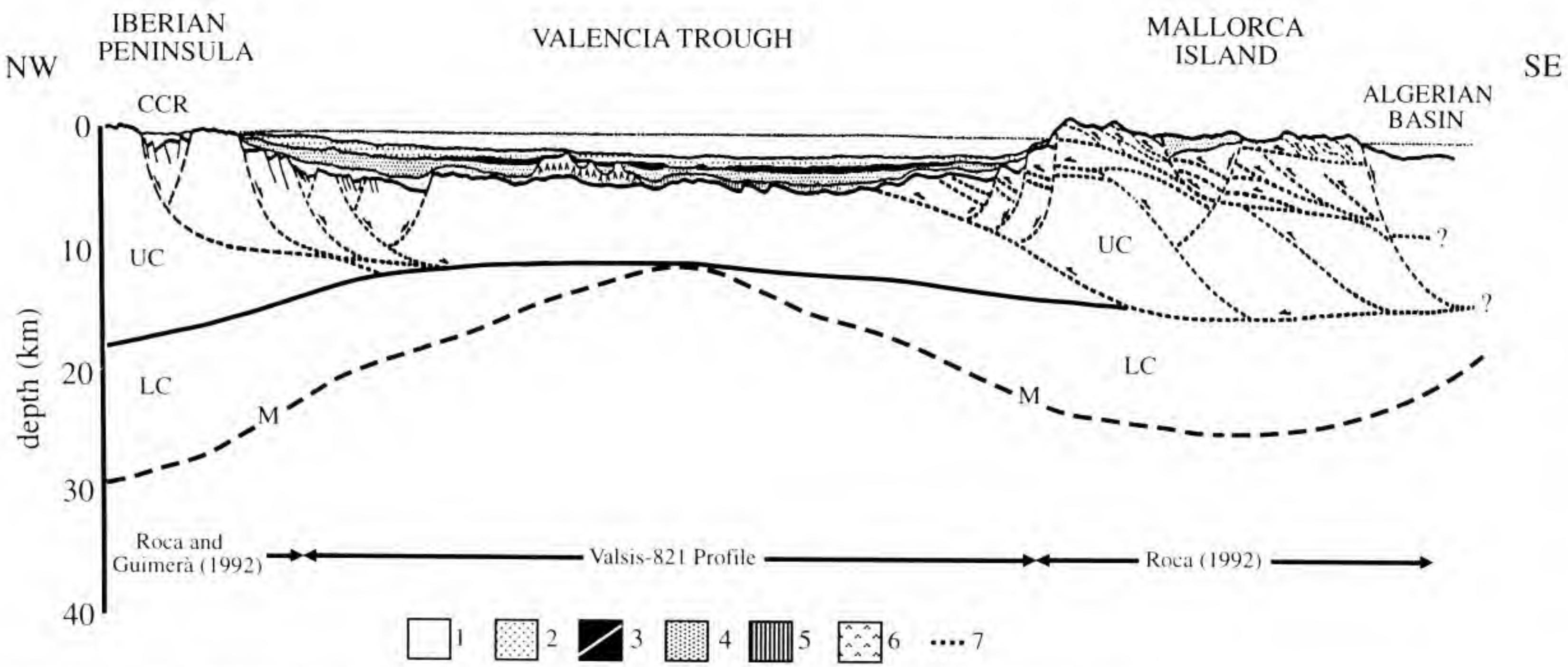

FIG. 5. Regional cross-section through Valencia Trough region along CDP line 821. For location see Fig. 8. (after Torres et al., 1993). UC: Upper crust; LC: Lower crust; M: Moho. 1 = water; 2 = Ebro Group (Pliocene-Quaternary); 3 = Messinian; 4 = Castellon Group (Serravallian-Tortonian); 5 = Lower Neogene (Aquitanian-Langhian); 6 = Volcanics; 7 = Main geodynamics features. Dashed line indicates approximate location of crust-mantle boundary.

tectonically quiet regime, whereas from the Spanish coast, a major deltaic complex, referred as the Castellon Group, prograded into the Valencia Trough, characterized by rapid deepening in response to post-rift thermal subsidence.

During the Messinian, a major drop in sea-level gave rise to the development of high relief unconformities on the continental shelves and the deposition of variably thick halites and minor sulphates in the central and northeastern parts of the Trough (Mulder, 1973; Ziegler, 1988). Sequence three is therefore only represented in the deepest parts of the basin.

The forth sequence, which spans late Messinian-Quaternary times, corresponds to the Ebro Group which, similar to the Castellón Group, consists of a major deltaic complex building out into deeper waters from the Spanish coast. However, it records in coastal and shelf areas a renewed phase of extensional tectonics that was accompanied by the extrusion of alkaline volcanics, forming the Columbretes Islands. Similar extensional faults are no evident in the central parts of the Trough (Banda and Santanach, 1992b).

Figure 6a provides a structural map at the base of the Cenozoic sedimentary fill of the Valencia Trough (Lanaja, 1987; Maillard et al., 1992) and Figure 6b gives an isopach map of the Cenozoic strata. The structure map illustrates the broadly saucer shaped configuration of the Cenozoic Valencia Trough. The isopach map, in combination with the bathymetric map, shows that this basin is clearly partly sediment starved.

Upper and Middle Crust

The available geophysical data permit to image the present day crustal configuration of the Trough (Figs. 7, 8 and 9). Reflection and refraction seismic data show that along both margins the upper-middle crust, almost reflection free, varies in thickness between 6-10 km and has velocities of 6.1-6.2 km/s (Figs. 7 and 9). The overlaying pre-rift Mesozoic carbonates are characterized by velocities ranging between 5.4 to 5.9 km/s (Pascal et al., 1992; Torné et al., 1992; Dañobeitia et al., 1992). Across the axis of the Trough, COP and CDP data allow to trace two different scenarios. In the southern region, seismic data show a series of NW dipping reflectors at upper/middle crustal levels; these are attributed to a Mesozoic basin (e.g. Mauffret et al., 1992; Torné et al., 1992) which may represent the southeastern prolongation of the Iberian Chain. The thickness of the upper/middle crust, without Cenozoic sediments, is in this area of the order of 7-8 km (Fig. 8a and P-III of Fig. 9). In the central parts of the Trough there is no evidence for the presence of thick Mesozoic series (Fig. 8b).

Seismic refraction/wide-angle reflection data also reveal that the upper/middle crust thins from the southwestern off-shore regions in a northeasterly direction towards the central areas of the Trough (P-III of Fig. 9). In contrast, profiles P-I and P-II (Fig. 9) reveal that the thickness of the upper/middle crust does not vary significantly across the Trough; this is in accordance with the ESP data (e.g., Pascal et al., 1992).

Lower Crust

Reflection seismic data show that beneath the Ebro Platform the 6-7 km thick lower crust is characterized by good reflectivity, involving 1-4 km long subhorizontal reflectors (Fig. 7a; Torné et al., 1992; Collier et al., 1994). In contrast, the 9-10 km thick lower crust of the Mallorca margin (P-I of Fig. 9) is variably reflective, showing disrupted reflectors which are difficult to trace along the profile (Collier et al., 1994; Fig. 7b). Considering the shallow water and lack of near-surface low-velocity layers, Collier et al. (1994) favour that disruption of the lower crustal reflectors is genuine and not caused by multiple interference. Towards the axis of the Trough, the reflective lower crust thins very rapidly where it appears to be missing (Torné et al., 1992) or to be reduced to a 1-2 km thick layer (P-I of Fig. 9), and displays a moderate velocity gradient of $0.1\ s^{-1}$ (Dañobeitia et al., 1992). These results are confirmed by coincident steep and wide-angle reflection profiles, which show that lower crustal reflectivity diminishes below the slope-break and vanishes towards the central parts of the Trough (Gallart et al., 1995;

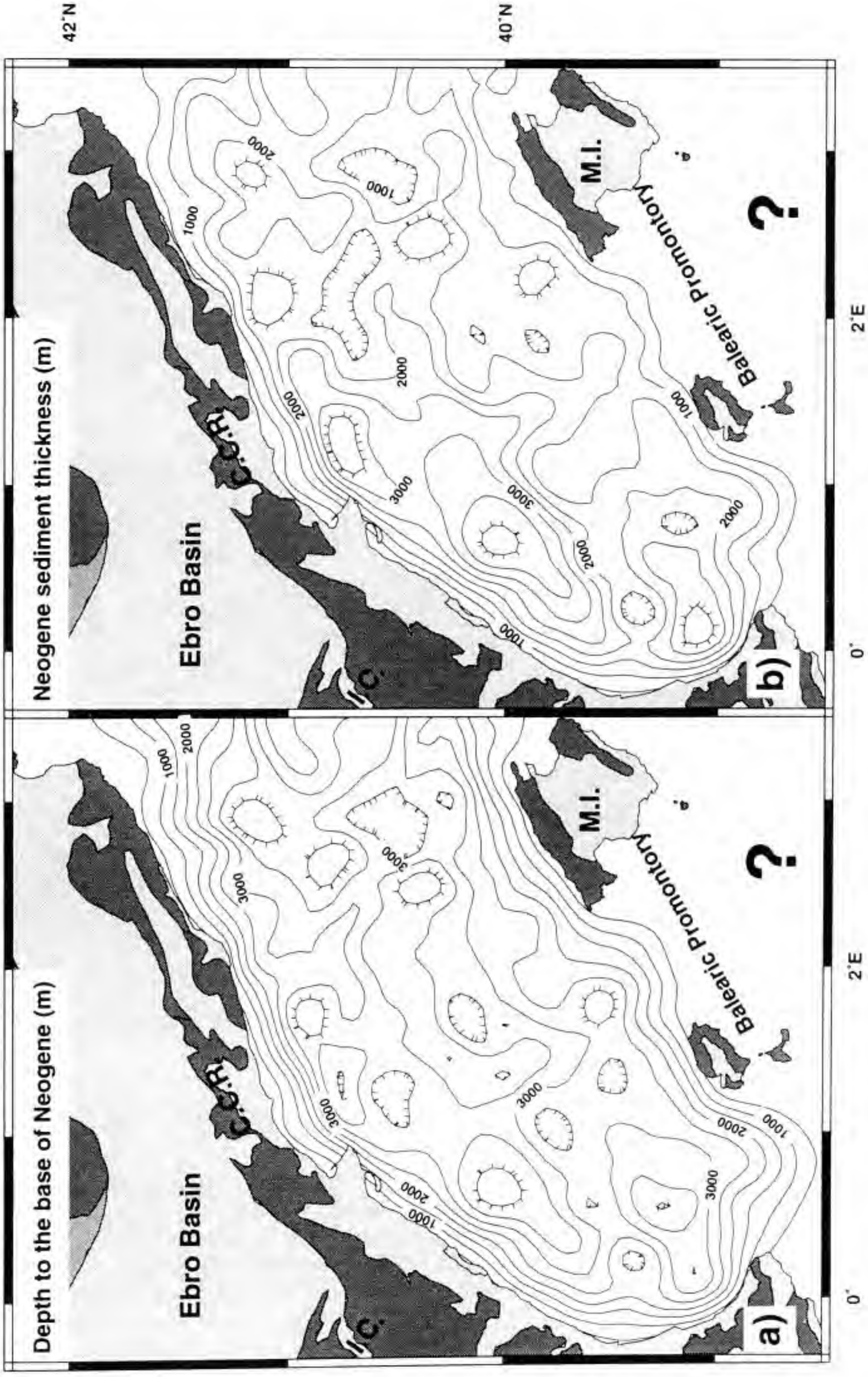

FIG. 6. **a)** Depth to the base of Neogene, contour interval 500 m. **b)** Neogene sediment thickness, contour interval 500 m. Compiled from Maillard et al. (1992) and Lanaja (1987). Abbreviations as in Fig. 2.

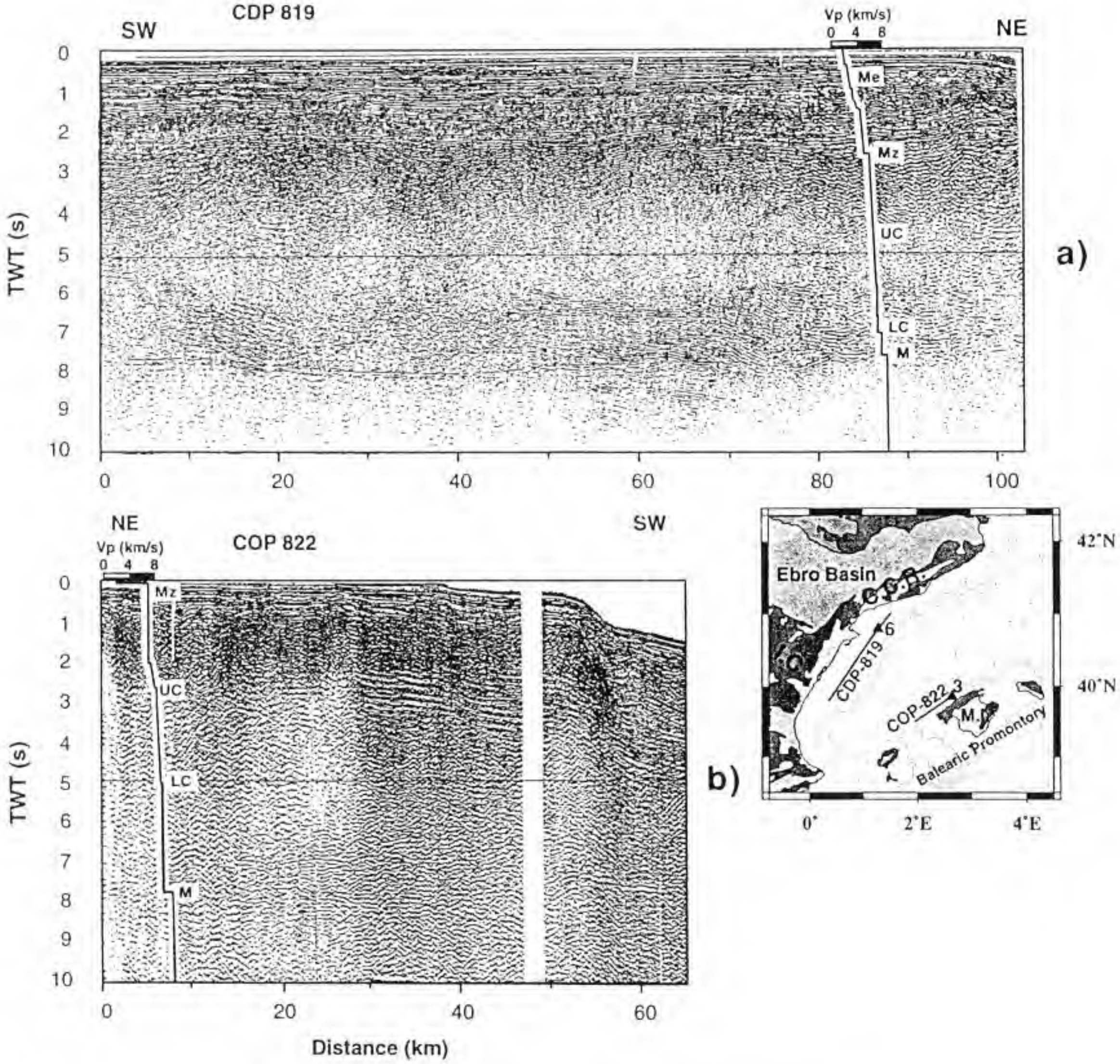

FIG. 7. **a**) Unmigrated CDP line 819 with velocity solution from ESP-6.
b) Unmigrated COP line 822 with velocity solution from ESP-3. No vertical exaggeration at 4.1 km/s. Inset shows location of seismic sections. Triangles give mid-point locations of ESP-6 and 3. Abbreviations as in Fig. 2. COP profiles from Collier et al. (1994), ESP results from Pascal et al. (1992) and Torné et al. (1992).

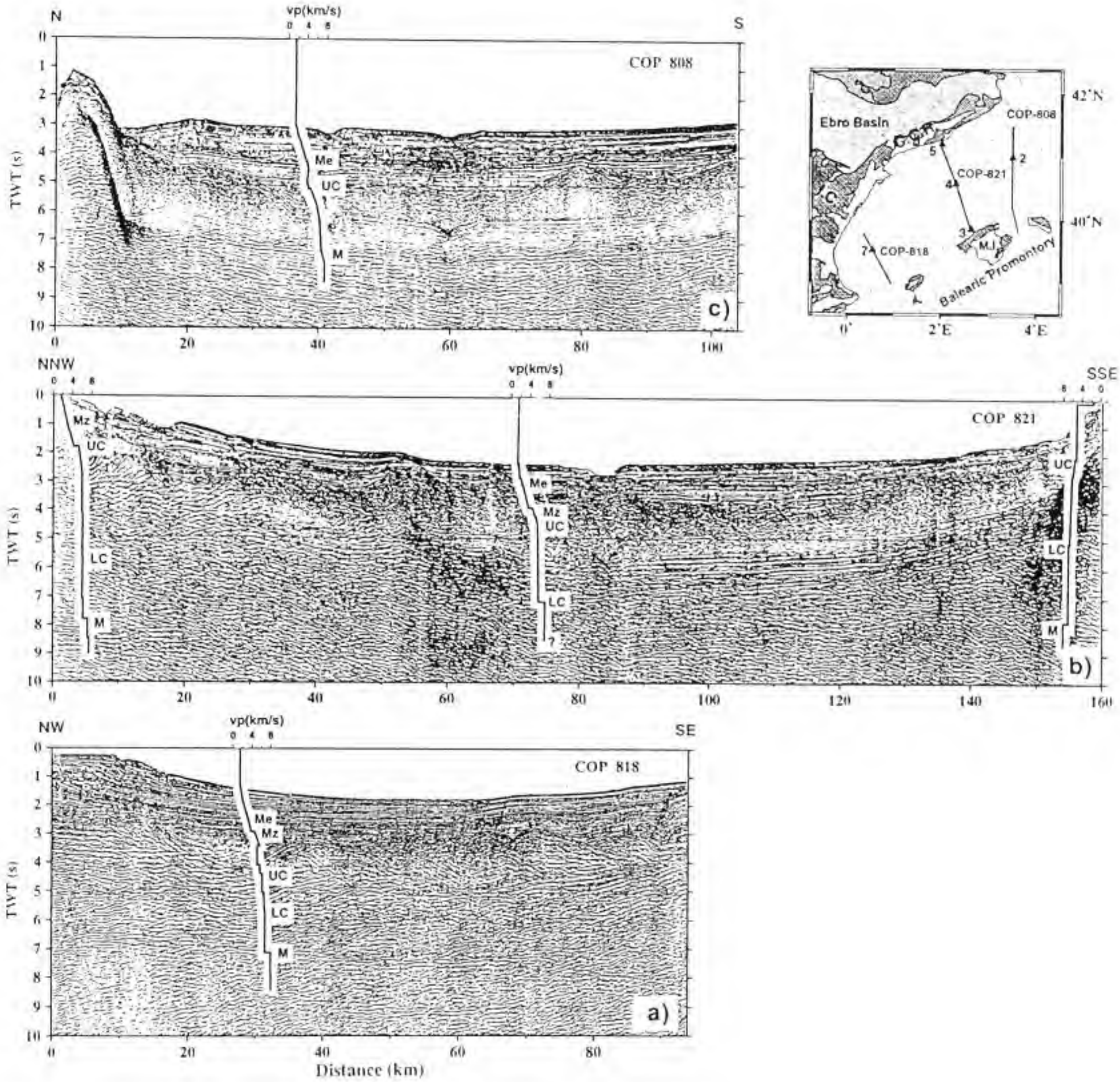

FIG. 8. **a)** Unmigrated COP line 818 with velocity solution from ESP-7. **b)** Unmigrated COP line 821 with velocities solutions from ESP-3, 4 and 5.

c) Unmigrated COP line 808 with velocity solution from ESP-2. No vertical exaggeration at 4.1 km/s. Me: Messinian; Mz: Mesozoic; UC: Upper Crust; LC: Lower Crust; M: Moho. Inset shows location of seismic sections. Triangles give mid-point locations of ESPs. Abbreviations as in Fig. 2. COP profiles from Collier et al. (1994), ESP results from Pascal et al. (1992) and Torné et al. (1992).

Vidal et al., 1995). In the southern parts of the basin, the lower crust is 5 km thick and is characterized by average velocities of 6.8 km/s (Pascal et al., 1992).

Watts et al. (1990) argue that the highly reflective lower crustal layer predates the mid-Tertiary extensional event; as such, underplating processes, related to the Cenozoic extension, could not cause the observed reflectivity of the lower crust. This view is supported by Collier et al. (1994), who based on the analysis of the lower crustal reflectivity patterns in different parts of the Trough, determined that Cenozoic extension significantly weakened or even destroyed the lower crustal reflectivity.

Moho Topography

In the area of the Valencia Trough, the crust/mantle boundary is reasonably well constrained by seismic and gravity data. Figure 10 gives a smoothed depth map of the Moho which is based on an integration of seismic data and gravity modelling results. The Moho raises gradually from a depth of 27-30 km under Iberia to 15-16 km beneath the Trough axis and descends toward the Baleares to a depth of 22-24 km. The Moho shallows from 18-19 km at the southwestern end of the basin to about 13-14 km at its northeasternmost end where it opens into the Provençal Basin. In cross-section, the Trough is slightly asymmetric, having a steeper flank towards the Baleares (Fig. 10).

In off-shore areas, velocities of the uppermost lithospheric mantle range from 7.7 to 7.9 km/s, whereas on-shore Iberia, velocities of 8.1 km/s are recorded. The upper-mantle velocity increases from 7.7-7.8 km/s beneath the axial parts of the Trough to 7.9-8.0 km/s towards the mainland, whereas beneath the Balearic Promontory, they remain in the 7.7-7.8 km/s range.

On COP the Moho can be traced throughout the basin (Figs. 7 and 8; Collier et al., 1994). The reflection signature of the Moho varies beneath the different parts of the basin and can be correlated with differences in the amount of stretching. Cenozoic extension may have modified the reflection character of the Moho.

Lithosphere-Asthenosphere Boundary

Information on the configuration of the lithosphere/asthenosphere boundary comes primarily from 2D modelling of the central parts of the Trough, using gravity and geoid anomalies along a profile extending from the Ebro basin to the South-Balearic basin (Watts and Torné, 1992a; Zeyen and Fernàndez, 1994). Watts and Torné (1992a) pointed out that the two margins of the Trough are characterized by different structural styles; the Spanish margin is a rift-type margin whereas the Balearic margin appears to be a constructional margin which is underlain by a broad region of lithospheric thinning. Zeyen and Fernàndez (1994) conclude that the shallowness of the lithosphere/asthenosphere boundary indicates that the causal rifting event had terminated only very recently, as previously suggested by Morgan and Fernàndez (1992). Surface-wave and tomography studies (Marillier and Mueller, 1985; Spakman, 1990) show that the Trough lies in a region of lithospheric thinning and anomalous low-velocity sublithospheric upper-mantle, characteristic for the western Mediterranean.

BASIN MODELLING

Subsidence Analysis

In an effort to isolate tectonic subsidence from the effects of sedimentary loading during the evolution of the Valencia Trough, backstripping analyses were carried out by several authors (Watts et al., 1990; Bartrina et al., 1992; Roca and Desegaulx, 1992) on the basis of wells located on the western margin of the Trough. The observed tectonic subsidence curves show an exponential

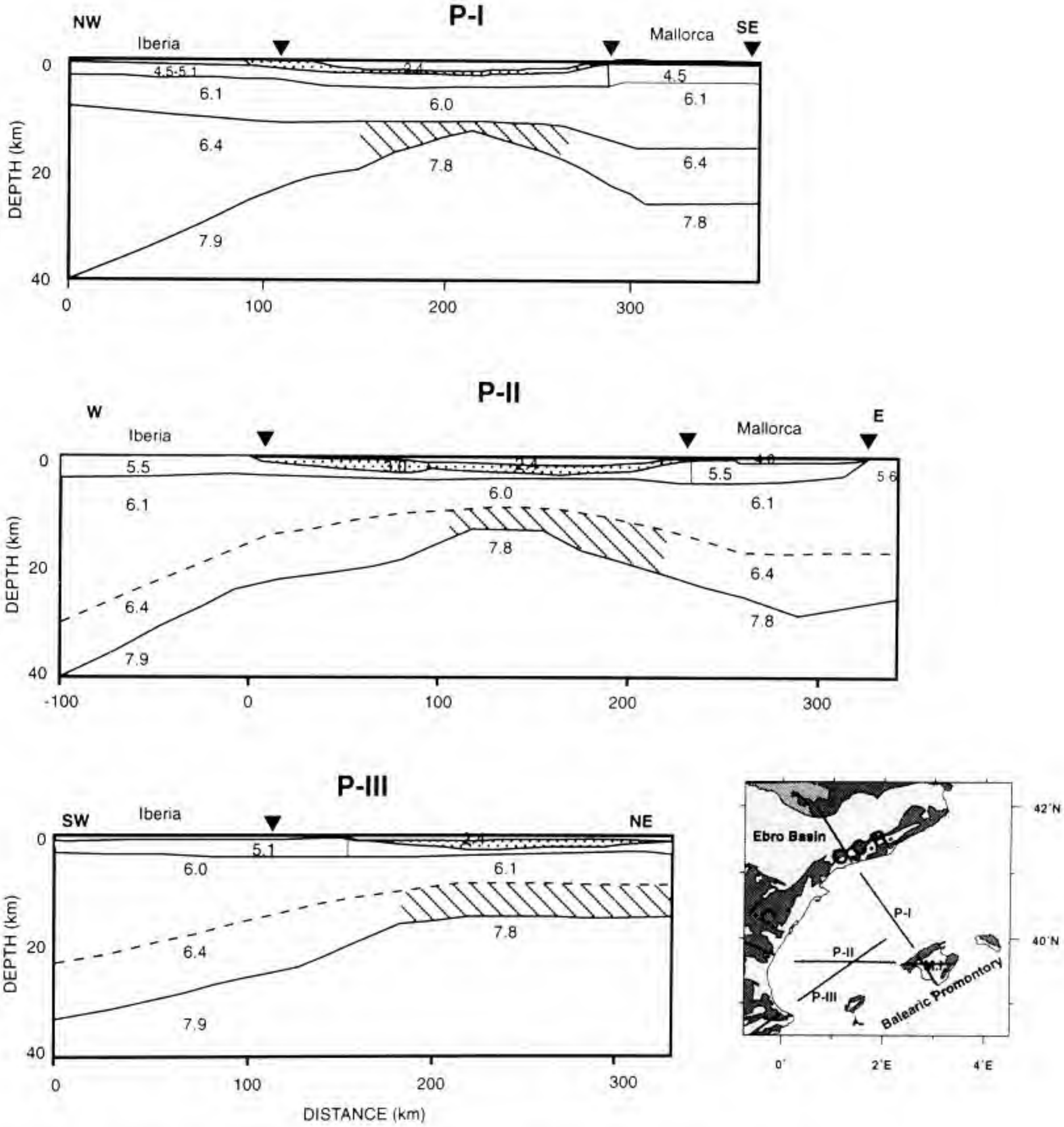

FIG. 9. Velocity-depth crustal models along profiles P-I, P-II, and P-III. Stippled zones indicate Neogene sediments. Dashed lines mark the transition between the upper and lower crust. Inclined stripped pattern denotes a gradient of 0.1 s^{-1} in the lower crust. Triangles indicate shore-line location. Values show P-wave velocities in km/s. Inset gives location of profiles (after Dañobeitia et al., 1992).

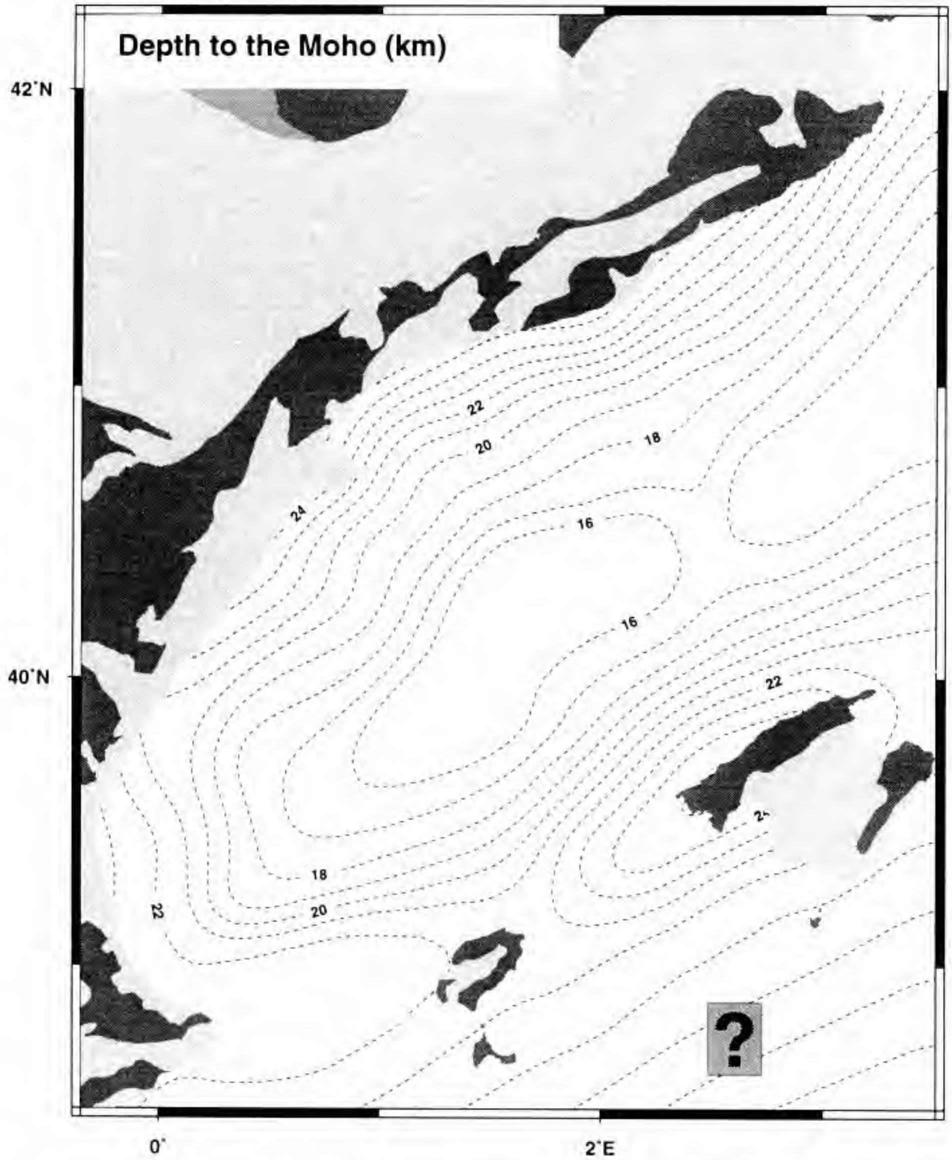

FIG. 10. Smoothed depth to the Moho map, contour interval 1 km. Compiled from results of Banda et al. (1980), Dañobeitia et al. (1992), Gallart et al., (1990), Pascal et al. (1992), Torné et al. (1992) and Zeyen et al. (1985).

trend with total tectonic subsidence values in the range of 1.4 to 1.6 km. 1D subsidence results also show that it is difficult to separate the late Oligocene-early Miocene syn-rift from the post-rift subsidence (Watts et al., 1990; Roca and Desegaulx, 1992). The shape of the subsidence curves must be carefully analyzed, as inaccuracies in palaeo-waterdepth estimates result in significant uncertainties (e.g, Watts et al., 1990; Bartrina et al., 1992).

Spatial variations in the tectonic subsidence along a regional transect considering CDP line 821 were analyzed by Watts and Torné (1992a) and Torres et al. (1993) who concluded that the form of the backstripped curves is similar, irrespective of the elastic thickness (Te), providing that Te values of 5 and 25 km and that corresponding to the 450°C isotherm, are assumed. Such quantitative subsidence analyses confirm the regional broad subsidence of the Valencia Trough and its accentuation to the northeast towards the continent-ocean transition (Fig. 11). Flexural subsidence analyses indicate uplift of the Catalan Coastal Ranges and the Balearic Islands, contemporary with subsidence with the Valencia Trough. Detailed studies, incorporating fine stratigraphic and palaeo-waterdepths analyses, are able to distinguish an initial phase of rapid subsidence (30-15 Ma) followed by a post-rift phase, characterized by lower subsidence rates (Roca and Desegaulx, 1992; Torres and Bois, 1993).

A 3D backstripping analysis of the entire region, carried out by Watts and Torné (1992b) using seismic and well data, confirmed that the Trough corresponds to a broad region of subsidence which is flanked by uplift in the Catalan Coastal Ranges and in the Balearic Promontory. Maximum tectonic subsidence values of 4.2-4.4 km are reached in the NE at the continent-ocean transition (e.g., Pascal et al., 1992), whereas elsewhere tectonic subsidence values range between 1 and 3 km (Fig. 11).

Numerical Modelling

Numerical models applied to the study area are far from complete and self-consistent. Nevertheless, they provide valuable information on the evolution of the Valencia Trough.

Two key features of this basin are difficult to handle with "classical" extensional models, namely the observed crustal and lithospheric asymmetry across and along its axis, and the fact that the initial rifting phase was coeval with compression along the Balearic margin. The majority of the numerical approaches, so far performed, omitted both facts and thus result in the assumption of simplified models. The difficulties encountered in modelling are particularly acute for the southern region of the Trough, where also the surface heat-flow values are consistently higher and the tectonic subsidence less than expected (Watts and Torné, 1992a). Therefore, numerical approaches were applied to the central parts of the Trough and concentrated on determining the pre-rift lithospheric conditions, the duration of the rifting stage, and the amount of lithospheric stretching.

Keeping in mind that the area of the Valencia Trough was uplifted and subjected to erosion during the Paleogene, Morgan and Fernàndez (1992) attempted to evaluate the lithospheric conditions which prevailed in the western and central regions prior to the Oligocene-Miocene rifting. A formulation of lithospheric buoyancy was used to back-calculate the possible pre-extension lithospheric structure, consistent with mid-Tertiary elevation and lithospheric strength constraints. This 1D approach was applied to different lithospheric columns along a profile from the Ebro basin to the centre of the Valencia Trough. For the Spanish margin, stretching factors ranging from 1.45 to 1.87 and an initial crustal thickness between 27 and 35 km were arrived at, whereas for the centre of the Trough these values are 2.9-4 and 26-36 km, respectively. The estimated stretching factors can be reduced by about 10%, assuming erosion of 1 km of pre-rift sediments. Although Morgan and Fernàndez (1992) proposed differential stretching for the Valencia Trough, where $\beta_{crust}>\beta_{mantle}$ in the centre of the Trough but $\beta_{crust}<\beta_{mantle}$ under the Spanish margin, uniform stretching in the centre of the Trough can also fit the model, provided an erosion of about 1.5 km of pre-rift sediments is acceptable.

A second approach, based on a 1D pure-shear stretching model, was explored by Foucher et al. (1992) using as constraints measured heat-flow

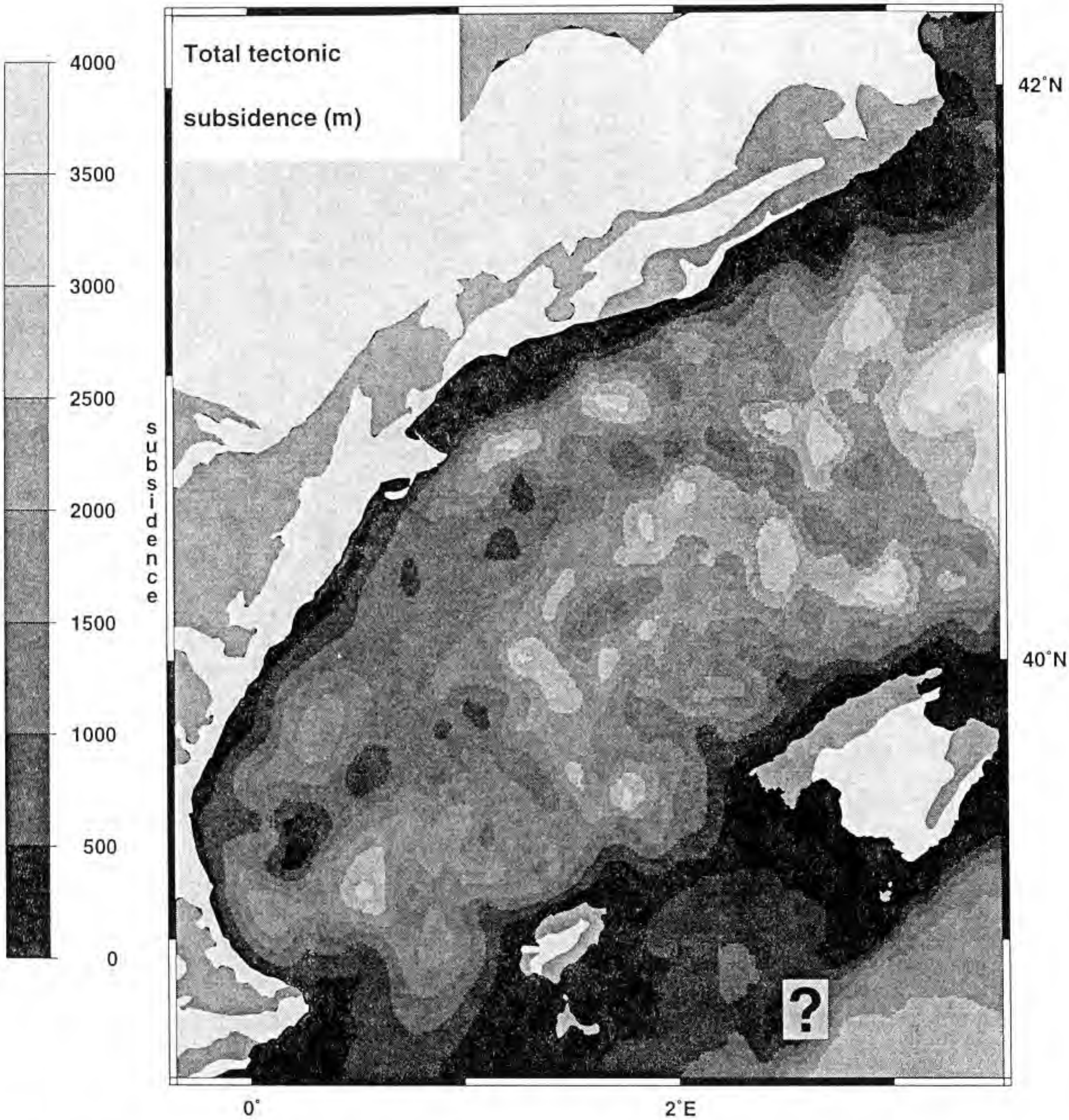

FIG. 11. Tectonic subsidence/uplift map, contour interval 500 m. Based on flexural backstripping of sediment thicknesses given in Fig. 6b. Parameters used: elastic thickness (Te) of 5 km, water, sediment and mantle densities of 1030, 2400 and 3300 kg/m^3, respectively. Modified after Watts and Torné (1992b).

data and bathymetry along three transects located in the central parts of the Trough (for location see Fig. 2b). Their results show that the northeastern half of the Valencia Trough could be successfully explained by a single rifting event lasting from 28 to 22 Ma with crustal stretching factors of 3.5 and 3.3 for the northern transects. However, this approach fails to explain the relatively high heat flow and shallow water depths observed in the southern part of the Trough (transect T4 of Fig. 2b), even when a multi-stage Neogene rifting event with different initial crustal and lithospheric thicknesses is considered. Foucher and co-workers speculate that this may be the result of a recent Pliocene-Quaternary event which would favour a southward propagation of the rift activity.

Fernàndez et al. (1995) attempted to model the southern Valencia Trough, applying a 1D uniform stretching model constrained by thermal and palaeothermal data. Their model takes into account the thermal effects of sedimentation/erosion, compaction, mineral composition and heat production. This model suggests that the SW Valencia Trough underwent a complex geodynamic history, including three Mesozoic rifting events, a Paleogene compressional and an uplift phase, causing erosion of about 5 km of Late Jurassic and Cretaceous sediments, and a late Oligocene-early Miocene extensional phase.

Several 2D numerical models were applied along different transects located in the central region of the Valencia Trough. These models, based on a kinematic approach, are mainly constrained by the basement structure, crustal and mantle stretching factors and surface heat-flow data. The first model, presented by Watts and Torné (1992a), was performed along a transect extending from the Ebro basin to the South Balearic basin, coinciding off-shore with CDP line 821. The total tectonic subsidence obtained from a 2D backstripping analysis and the measured heat flow were compared with that obtained from a 2D non-uniform finite stretching model. The best fit shows a homogeneous stretching with a b factor increasing from 1.4 beneath the Spanish margin to 3.0 at the centre of the Trough. However, on-shore, below the Catalan Coastal Ranges, the observed tectonic uplift requires a broad region of mantle stretching ($\beta_{mantle}=1.4$) extending well beyond the region of crustal stretching. It was assumed that the rifting episode started 24 Ma ago and had a duration between 8 and 16 Ma, and that the initial crustal and lithospheric thickness were 31.2 and 125 km, respectively. A combined geoid and gravity model, where a temperature-dependent mantle density is considered, was used to better constrain the present-day geometry of the lithosphere below the Balearic margin. The model results reveal asymmetry at crustal and subcrustal levels between the flanks of the Trough and, that the region of lithospheric thinning is much broader than the width of the initial rift and extends far beyond the Balearic Promontory.

Torres et al. (1993) modelled the off-shore areas of the Trough along CDP line 821, using a 2D uniform stretching model, considering local isostasy, and the loading and thermal effects of sediments. Unlike Watts and Torné (1992a), it was assumed that the Valencia Trough underwent a first rifting event between 25.2 to 20.0 Ma, which had minor effects on the deep basin and affected mostly the Spanish margin, an that a second rifting event between 15.2-10.2 Ma, coeval with the opening of the Tyrrhenian basin, was responsible for the present-day Moho uplift. However, there is no clear reflection-seismic evidence for this second rifting event. These authors also conclude that, in contrast to the interpretation given by Fontboté et al. (1989) and Roca and Desegaulx (1992), the post mid-Miocene development of the basin does not conform to a flexural foreland-type mechanism.

Finally, Jansen et al. (1993) proposed a 2D necking model along two profiles crossing the central part of the Valencia Trough. Basement subsidence, Moho depth and gravity anomalies were calculated for different elastic thicknesses, duration of rifting and necking depths. Results support a rifting model with a low elastic thickness (5-20 km), an intermediate depth of necking (17-33 km) and a finite duration of the stretching event (16 m.y.), propagating southwestward. Rifting would have started 24 Ma ago in the northeastern profile (Mallorca-Barcelona) and 20 Ma ago in the southwestern profile (Mallorca-Ebro Delta). A Pliocene uplift, caused by additional mantle thinning, is proposed to fit the present-day Moho depth and uplift in the Valencia Trough and the Iberian-Mediterranean margin. Whether this uplift is related with the Plio-Quaternary extensional phase is not evident.

DISCUSSION

The Valencia Trough is one of a several extensional basins located in the western Mediterranean region, which developed under a compressional regime. It differs, however, from other rifted basins of the western Mediterranean, such as the Tyrrhenian and South Balearic basins, in that extension has not progressed to crustal separation. Available geological and geophysical data provide a reasonably well constrained image of its present-day crustal structure. It is now well accepted that the Valencia Trough is characterized by a strongly attenuated continental crust, underlain by an anomalous low-velocity upper mantle, and that its lithospheric structure shows asymmetry both at crustal and sub-crustal levels. There is still some debate, however, among the different authors on the cause and magnitude of extension and on the evolution of the Trough.

A major problem in understanding the evolution of the basin is the poor knowledge on the structure of the lithosphere and configuration of the crust before the late Oligocene extensional phase. It is well established that during the Mesozoic break-up of Pangea the area underwent repeated extensional events and that, during Late Cretaceous and Paleogene times, intraplate compressional stresses caused inversion of most of the extensional Mesozoic basins. However, whether or not the crust recovered its pre-Mesozoic thickness as a consequence of basin inversion is still a matter of debate. Therefore, the first question that has to be solved concerns the effect of Mesozoic extension and Paleogene compression on the lithospheric configuration of the Valencia Trough, prior to the onset of Oligocene-Miocene extension.

The Oligocene to Recent evolution of the Trough is characterized by a first rifting stage (late Oligocene-early Miocene) which can be explained either by a back-arc model (the Valencia Trough evolved in a back-arc position relative to the Kabylian-Calabrian arc) or by southward propagation of the European rift system across the West-Mediterranean fold belts. During the early-middle Miocene, Balearic thrusting counteracted extensional forces and aborted rifting. After locking of the Balearic thrust front, extension resumed during the latest Miocene and persisted into the Pleistocene.

Assuming that the Trough is indeed a rift-type basin and forms part of the Cenozoic rift system of Western and Central Europe, there are still several questions that need to be addressed, such as the initiation, duration and mechanism of the Oligo-early Miocene rifting event, the amount of stretching, and rift-related processes such as lower crustal reflectivity, underplating, Moho rejuvenation, thermal regime, etc.

Timing and Duration of Rifting

Surface geological and reflection seismic data show that crustal extension commenced during Chattian(?)-Aquitanian times with the deposition of the first syn-rift sediments and the development of a block-faulted relief along the western flank of the Valencia Trough. To the South, however, the first syn-rift sediments record an upper-Langhian age. This suggests that the rifting propagated from North to South, or that the southern areas had a higher topographic relief when rift activity commenced. Geological data also show that in the northern and central parts of the Trough the rifting phase persisted into Langhian-Serravallian times (Roca and Desegaulx, 1992). This indicates that the main rifting activity lasted 8-10 Ma. Subsidence analysis do not help too much in deciphering the duration of rifting since, as mentioned above, it is difficult to separate the Oligocene-early Miocene syn-rift from the post-rift phase. This underlies the proposal of different durations of the rifting stage, ranging from 6 m.y. (Foucher et al., 1992) to 8-16 m.y. (Watts et al., 1992a) and even to multi-stage Neogene events (Torres et al., 1993).

Mechanisms of Rifting

Numerical models attempt to establish the mechanisms of lithospheric extension. Two end-member models are available, the pure-shear (McKenzie, 1978) and the simple-shear model

(Wernicke, 1985). In the study area, most of the available geophysical data favours a pure-shear mechanism, since there is no clear evidence in support of a simple-shear model. This does not rule out, however, that simple shear deformation is restricted to the upper/middle crust and is combined with pure shear deformation at deeper levels (Lister et al., 1991), though such a model has not yet been tested. Simple shearing would explain some major upper crustal structural features, such as low angle faults and fault-block rotations observed in the upper/brittle part of the crust. Although the pure-shear model has been successfully applied to off-shore areas, a differential stretching mechanism is required to fit the uplift of on-shore areas. In this context, Cabal and Fernàndez (1995), based on thermal evidences from the easternmost Ebro basin, proposed that during middle Miocene to Recent times thinning of the mantle lithosphere occurred beyond the zone of actual rifting and extended some distance to the West of the Catalan Coastal Ranges.

Stretching Values

The amount of Oligocene and younger lithospheric stretching was obtained mainly from the crustal configuration and subsidence studies. Assuming a pre-rift crustal thickness of 30-35 km, crustal stretching factors, ranging from 1.4-1.55 for the western flank of the Trough and 3.2±0.25 for central areas, and upper-crustal thinning ratios of up to 2.1 were obtained. This value is much larger than upper-crustal extension ratios deduced from structural analysis which give values of up to 1.4-1.5 (Roca and Guimerà, 1992). The observed differences between extension and thinning ratios may be explained by assuming that the thinned crust is partially inherited from the Mesozoic rifting events. On the other hand, Morgan and Fernàndez (1992) suggest that upper crustal stretching factors in the 1.8 to 2.55 range, can be expected if a mid-Tertiary topographic relief between 0 and 500 m is assumed. Finally, it can not be excluded that the observed discrepancies between upper crustal extension by faulting and mid to lower crustal attenuation involved a Neogene destabilization of the Moho and its upwards displacement by magmatic delamination of the lower crust (Ziegler, 1992, 1995).

Rift Related Processes

Collier et al. (1994) established a correlation between the lower crustal/Moho reflectivity and the degree of crustal attenuation and concluded that extension could significantly weaken or even destroy the reflectivity of the lower crust (see also Watts et al., 1990), but enhanced the reflectivity of the Moho. Where the lower crust is very thin, its reflectivity and the Moho reflector are weak or non existent. In the centre of the Trough, the absence of a reflective lower crustal layer and the presence of a single lower crustal reflector (reflector X), which coincides with a P-wave velocity increase from 6.4 to 7.8 km/s, has led to Collier et al. (1994) to propose that this reflector represents the top of a crustal transition zone, composed of crustal and mantle material. Further evidence for intrusion of mantle material into the base of the crust is the observed high-velocity gradient in the lower crustal layer (Dañobeitia et al., 1992). The occurrence of a crust-mantle transition zone was already postulated by Banda et al. (1992) in an effort to explain the observed difference between upper and lower crustal attenuation and the presence of anomalous upper mantle velocities recognized throughout the basin, the thickness of which is of the order of 20 km, as deduced from gravity modelling (Torné and Banda, 1988).

The thermal regime of the Trough, as summarized in Fig. 2b, is not too well constrained by measurements which show considerable variations in the central regions of the Trough. This makes it difficult to decipher whether this area is characterized by intermediate or high heat flow values. Moreover, the background heat flow of the Balearic Promontory is quite similar to that of the Trough axis, except on Menorca where a slight increase is recorded. This regional heat flow pattern is in accord with modelling results which show that the region of lithospheric thinning extends beneath the Balearic Promontory (Watts and Torné, 1992a; Zeyen and Fernàndez, 1994); as such, the latter can not be regarded as the conjugate margin of the Iberian margin.

CONCLUSIONS

Although basin modelling has been quite successful in imaging some aspects of the central parts of the Trough, it was less successful on a basin wide scale and particularly for the southern areas. The underlying reason is, that kinematic models applied do not account for the complex tectonic evolution of the area. This concerns mainly the alternation and nearly contemporaneous extension and compression and the possible southward rift propagation. Some of these aspects have been explained in a step-like manner rather than incorporating all of them into a self-consistent model. Advective heat transport, radiogenic heat production, melt generation and sediment thermal blanketing are not contemplated in most of the models, yet they may be important factors in controlling the evolution of the basin.

On the other hand, lithospheric deformations imposed on these kinematic models are not constrained by constitutive equations. Depending on the initial lithospheric structure, the strain rates obtained from stretching factors and the duration of the rift stage are at odds with the thermo-mechanical behaviour of lithosphere. As pointed out by Negredo et al. (1995b), thermo-mechanical constraints on kinematic models may result in different styles of rifting but can also invalidate a preconceived mode of deformation. Fully dynamic models are more self-consistent since deformation is calculated by coupling constitutive and thermal equations (e.g. Bassi et al., 1993), yet the high non-linearity of the equation renders the results very sensitive to the initial conditions, making it difficult to reproduce the present-day crustal structure of a given extended area.

Therefore, we conclude that current basin modelling techniques do not allow to properly account for the evolution of the Valencia Trough. In addition, even assuming that these techniques are capable to handle the main tectonic aspects, there are still some gaps in our knowledge of the study area such as:

- Is the Valencia Trough an extensional feature on its own? An important aspect when modelling the Trough is whether it should be considered as an "extensional" feature on its own, or should be considered as a "branch" of a much wider extensional region covering the western Mediterranean. If the latter is correct, then the Trough would form part of the Gulf of Lyons, Corsica-Sardinia rift system which was active prior to the late Aquitanian-early Burdigalian opening of the Algero-Provençal Basin (Ziegler, 1992).

- What was the pre-rift lithospheric configuration of the Trough? It is not clear whether or not the Paleogene compressional phase was large enough to restore the thinned Mesozoic lithosphere to its initial conditions. In other words, is the present-day lithospheric structure mainly the result of the Neogene rifting? or is it partially inherited from Mesozoic extension?. The Alpine deformation of the area occupied by the Valencia Trough is still poorly understood. Therefore, interactive forward and backward models must be developed which take into account various degrees of Mesozoic extension and Paleogene inversion and pre-rift erosion.

- Is the present-day lithospheric structure the result of a finite rifting episode or a multi-stage rifting? As pointed out earlier, subsidence analysis do not permit to clearly separate syn-rift and post-rift phases; correspondingly different durations of rift activity have been proposed. This is mainly due to the lack of knowledge of several parameters, which strongly influence backstripping results, such as palaeo-bathymetry, precise timing and amount of erosion, pre-Neogene basement morphology and density variations. The superposition of a thrust loading phase on the southeastern parts of the Trough is an additional complication factor. In this respect, stress-induced lithospheric deflections may also have played an important role (Cloetingh and Kooi, 1992).

- What is the configuration of the lithospheric mantle and the lithosphere-asthenosphere boundary? Information on the structure of the lithospheric mantle and the topography of the base of the lithosphere comes only from thermal modelling and surface-wave studies. There is no information on the deep structure of the lithosphere along the basin axis and its transition to the oceanic Algero-Provençal basin. The lateral extent and thickness of the low-velocity anomalous upper mantle is unknown. Passive and active source seismology in the form of deep seismic soundings and intermediate tomographic inversion schemes may help to outline the topography of the base of the lithosphere and to image the shape of the low-velocity upper mantle layer.

- Was the Moho a passive marker throughout Cenozoic extension? There is some evidence that the geophysically defined crust/mantle boundary was destabilized during Oligocene and younger rifting. The differences observed between the magnitude of crustal extension by faulting and crustal thinning could at least partly be explained by an upward displacement and rejuvenation of the Moho. In this respect, the presence of an anomalous low-velocity uppermost mantle or "transitional" layer could be explained by the presence of partial melts which may have interacted with the lower crust. However, whether the top or the bottom of this anomalous layer correspond to the present crust/mantle boundary, remains to be investigated.

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Geodynamics of the Gulf of Lions: implications for petroleum exploration

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ABSTRACT

The Gulf of Lions Basin forms the northern passive margin of the oceanic Provençal Basin which opened at the transition from the Aquitanian to the Burdigalian, entailing a 25-30° counterclockwise rotation of the Corsica-Sardinia Block. The rifting phase, preceding crustal separation and the onset of sea-floor spreading, spans 6 Ma and commenced during the Late Oligocene. Oceanic crust occupies a some 200 km wide strip in the central parts of the Provençal Basin; this crust is covered by up to 6 km thick Miocene to Pleistocene sediments, including thick Messinian salts involved in diapiric structures.

The Oligo-Miocene rifts of the Gulf of Lions are superimposed on the Late Cretaceous-Early Eocene Provençal fold belt which forms part of the Pyrenean orogen. Crustal separation between Iberia and France was achieved during Mid-Aptian times. The Pyrenean orogen developed in response to convergence of Iberian micro-continent with the southern margin of France during the Alpine collision of Africa-Arabia with Europe. The Provençal fold belt evolved out of a Mesozoic rifted basin which either formed an aborted branch of Pyrenean rift or corresponded to a segment of the latter. It is here proposed that the Central Pyrenean Fault, representing the suture between Iberia and Europe, projects to the southeast of Sardinia. Consequently, we assume that the Corsica-Sardinia block remained attached to Europe during the Late Aptian to Campanian opening of the oceanic Bay of Biscay Basin.

During the Oligo-Early Miocene rifting phase, which preceded the opening of the Provençal Basin, the Pyrenean compressional structures of the Provençal fold belt were tensionally reactivated. Folded Permo-Carboniferous and Mesozoic strata form the pre-rift sequence which is at least partly preserved beneath the syn-rift sediments of the Gulf of Lions grabens. These pre-rift series contain several viable hydrocarbon source-rocks, many of which probably became over-mature for oil generation at the end of the Pyrenean orogeny. In on-shore grabens, the Oligo-Miocene syn-rift series attains thicknesses of 3-4 km and contains lacustrine source-rocks having a limited geographic distribution. The post-rift series is devoid of source-rocks.

Of the 11 exploration wells drilled in the offshore parts of the Gulf of Lions, all of which bottomed in Mesozoic strata or the basement, none

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penetrated the syn-rifted series; all wells were located on structurally high rift flanks and failed to encounter hydrocarbons. The remaining hydrocarbon potential of the Gulf of Lions must be regarded as speculative.

INTRODUCTION

The Gulf of Lions Basin forms the northern passive margin of the oceanic Provençal Basin. Development of the Gulf of Lions Basin began with the Oligo-Aquitanian rifting phase which culminated in crustal separation and the opening of the oceanic Provençal Basin, entailing a 25-30° counter-clock-wise rotation of the Corsica/Sardinia Block away from the European mainland. During this rifting event the Languedoc-Roussillon area was affected by regional extension, causing reactivation of pre-existing NE/SW trending normal faults of the Southeast France Basin and of compressional structures of the latest Cretaceous-Paleogene Pyrenean fold-and-thrust belt (Fig. 1).

Whereas in the framework of plate tectonics a consensus was quickly reached on the origin of major Atlantic-type oceans, the evolution of the Western Mediterranean remained for a long time a subject of debate. This is mainly due to the great complexity of its structural setting (Dercourt et al., 1993; Ziegler, 1988, 1994; Biju-Duval, 1984), the lack of direct information on the pre-rift sequence (boreholes) as well as to the narrowness of its basins in which thick sedimentary series obscure magnetic sea-floor anomalies, thus rendering it difficult to distinguish between thinned continental and oceanic crust.

Under such a scenario, exploration for hydrocarbons commenced in the off-shore parts of the Gulf of Lions during the late 1960's with the acquisition of regional reflection seismic surveys. The impetus for this activity was given by the discovery of small oil accumulations in the on-shore Oligocene Alès and Camargue rifted basins (Gallician field) which indicated the presence a functioning petroleum system. Unfortunately, results of the eleven exploration wells drilled between 1969 (Mistral and Sirocco) and 1985 (Agde Maritime), all of which were targeted at structural highs, were very disappointing in so far as they failed to encounter any hydrocarbons.

OPENING OF THE GULF OF LIONS

Already Argand (1924), as recalled by Durand-Delga (1980) and Olivet (1988), considered the fit of the continental slopes of Catalonia, Languedoc and Provençe on the one hand, and Corsica and Sardinia on the other, as a basic argument in favour of the oceanic nature of the Provençal Basin. This hypothesis implied that the Corsica-Sardinia Block was separated from the continental margin of Southern France after the Pyrenean orogeny.

Rotation of Corsica-Sardinia Block

As early as the 1970's different methods were applied in an effort to determine the position of the Corsica-Sardinia block prior to the opening of the Provençal Basin (Fig. 2). Based on aeromagnetic data, Auzende et al. (1973) attempted to determine transform motions between Corsica, Sardinia and the French mainland. Westphal (1976), emulating Bullard et al (1965), sought the best morphological fit of the shelf edge isobaths. These early reconstructions are tantamount to describing a linear NW/SE translation of the Corsica-Sardinia Block and do not account for its 25° to 30° rotation, indicated by paleomagnetic data.

These discrepancies were emphasized by Edel (1980) who tried to reconcile both hypotheses by considering Corsica and Sardinia as separate blocks. Faced with the objections of geologists (Arthaud and Matte, 1977; Mattauer, 1973; Auzende and Olivet, 1979; Biju-Duval and Montadert, 1977), it is now proposed that Corsica and Sardinia form a single block which rotated during the opening of the Provençal Basins by a maxi-

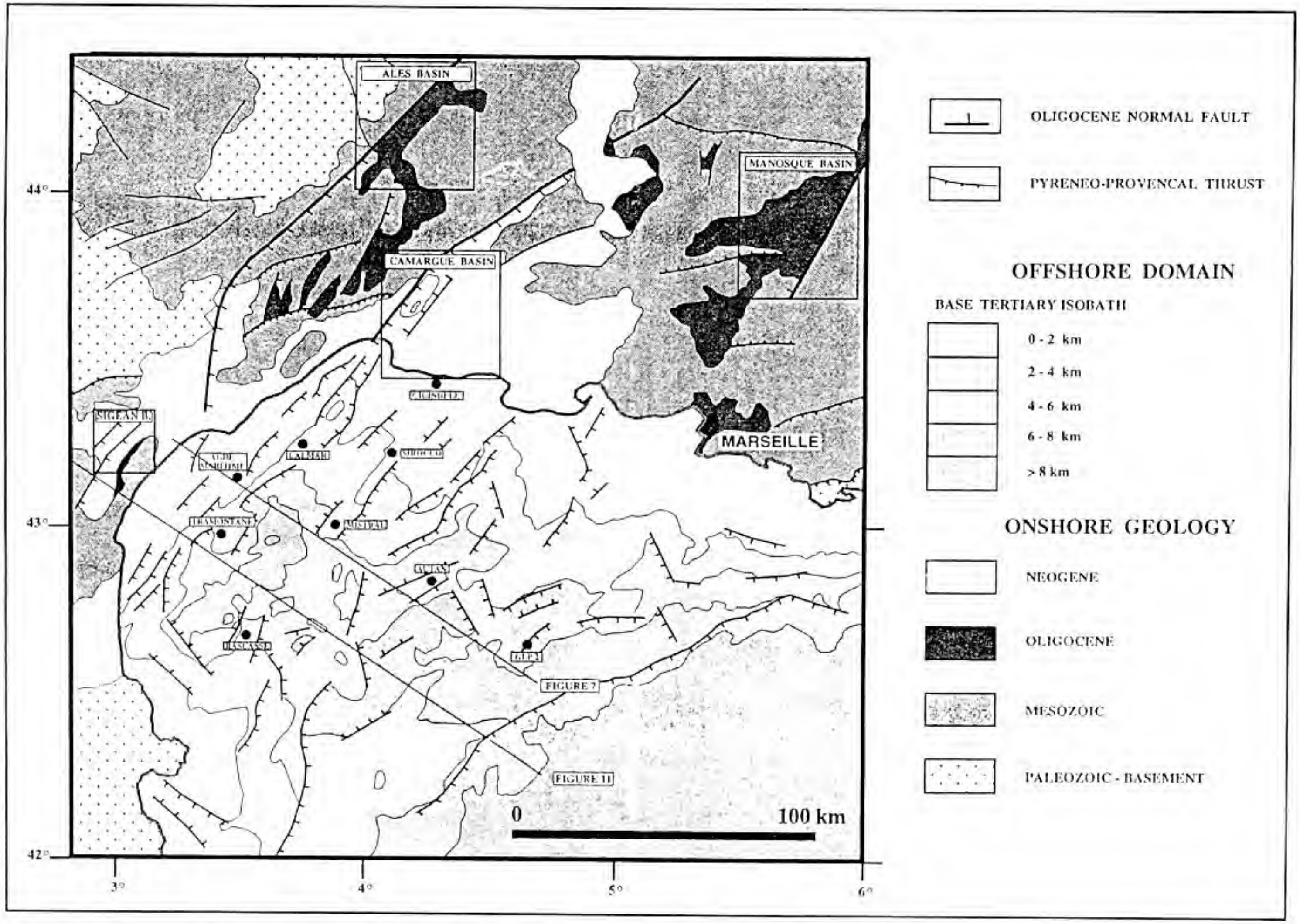

FIG. 1. Structural sketchmap of the Gulf of Lions and surrounding areas, showing location of cross-sections given in Figs. 7 and 11 and basins cited in text. Isobaths of the base-Tertiary modified after Gorini (1993).

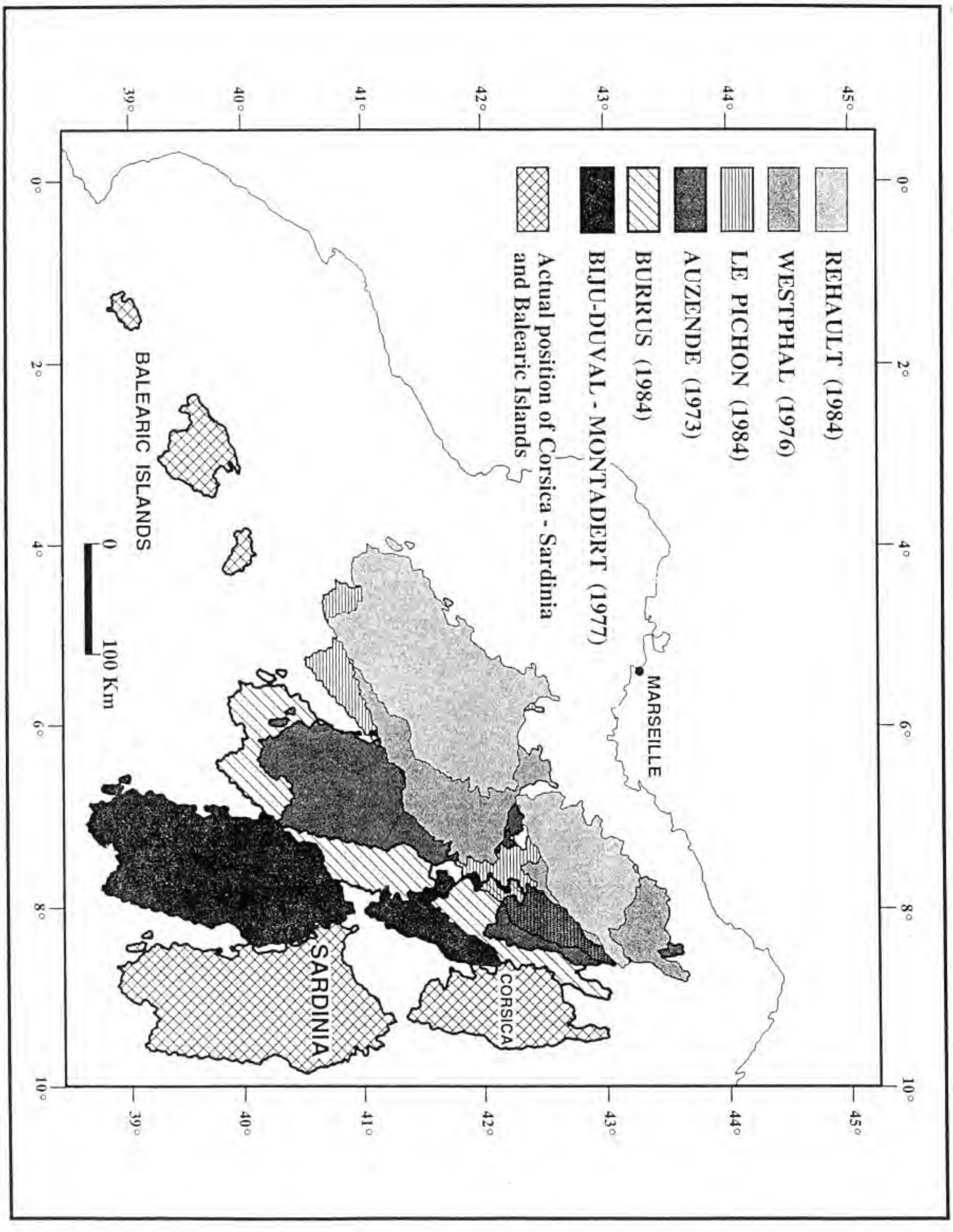

FIG. 2. Possible palaeo-position positions of the Corsica-Sardinia Block (for explanation see text)

mum of 30° about a pole located in the Gulf of Genova. Despite many uncertainties about the position of the boundary between oceanic and continental crust (Fig. 3), all reconstructions raise the problem of an apparent gap between the continental slopes of the Gulf of Lions and Western Sardinia.

After McKenzie (1978) showed that thinning of the continental crust could be achieved by stretching, this model was applied to the Gulf of Lions. New refraction-seismic data, clarifying the nature of the crust, aided in the to development of new kinematic models (Le Pichon, 1984; Burrus, 1984; Le Douaran et al., 1984; Réhault et al, 1984a, 1984b, 1985). Current studies, largely undertaken in conjunction with the Integrated Basin Studies project, integrate seismic and 3D gravimetric data and aim at proposing a new crust stretching model for the Gulf of Lions passive margin in an effort to overcome the apparent gaps in palinspastic reconstructions of the Provençal Basin.

Based on reflection-seismic surveys, Auzende et al. (1971) and Le Pichon and Sibuet (1971) suggested that a connection existed between the rotation of Corsica-Sardinia Block and the evolution of the Oligocene rifts of Western Europe. Palaeomagnetic data and studies on volcanic rocks of Sardinia provided constraints on the age of this rotation. Edel (1980) and Montigny et al. (1981) reached the conclusion that Sardinia underwent a 25 to 30° rotation during about 3 Ma at the transition from the Aquitanian to the Burdigalian, entailing a horizontal displacement of Sardinia by some 300 km.

Geological Constraints

The postulated earliest Miocene separation and rotation of the Corsica-Sardinia Block away from the southern margin of France is fully compatible with the geological data summarized below.

Upon closure of the Provençal Basin, the Palaeozoic structural, metamorphic and igneous record and zonation of the Corsica-Sardinia Block correlates readily with that of the Maures-Esterel area of Southern France. Moreover, the palaeomagnetic record of Permian volcanics supports a 30° rotation of the Corsica-Sardinia Block (Orsini et al., 1980; Lardeaux et al., 1994).

Comparison of the Mesozoic stratigraphic record of Sardinia, southeastern France (Languedoc) and Catalonia has been the subject of numerous papers (Cherchi and Schroeder, 1973, 1976; Chabrier and Fourcade, 1975; Chabrier and Mascle 1984 ; Azéma et al., 1977; Fourcade et al., 1977; Alleman, 1978; Philip and Alleman, 1982). Figure 4 provides palaeogeographic syntheses for Domerian, end Bathonian, and Valanginian times. An NW/SE trending stratigraphic cross-section, extending from the French mainland to Sardinia, datumed at the top of the Cretaceous, is given in Figure 5. It illustrates strong facies analogies between western Sardinia and the Provençal domain. Until the Middle Cretaceous, the Provençal domain and western Sardinia appear to have formed the southern margin of the rapidly subsiding intracratonic Vocontian Trough of southeastern France. In the Provençal domain, as well as in the western Sardinia (Nurra region), the Middle Cretaceous unconformity is clearly evident by more or less pronounced erosion and the deposition of bauxite. The associated uplift and deformations must be related to sinistral motions during the Aptian-Albian separation of Iberia from Europe. First indications for strongly compressional deformations occurred during the Late Cretaceous; these correlate with the early phases of the Pyrenean orogeny and the development of the flexural South Provençal Trough, the axis of which migrated in time progressively northward.

During the Paleogene, the area occupied by the future Gulf of Lions was uplifted and subject to deep erosion. Erosion products were shed northward into the continuously subsiding Eocene foredeep where they were deposited as flysch (Stanley and Mutti, 1968; Ivaldi, 1974; Jean, 1985; Ravenne et al., 1987; Vially, 1994).

Structural data on the rotation of the Corsica-Sardinia Block are much more limited and are largely based on microtectonic analyses (Cherchi and Trémolières, 1984; Letouzey et al., 1982; Letouzey, 1986). These demonstrate a counterclockwise rotation of about 30° of pre-Oligocene markers. Eocene compressional structures in Sardinia have a similar style as those in Provençe and indicate, upon palinspastic restoration, north- to northwestward directed mass transport, thus con-

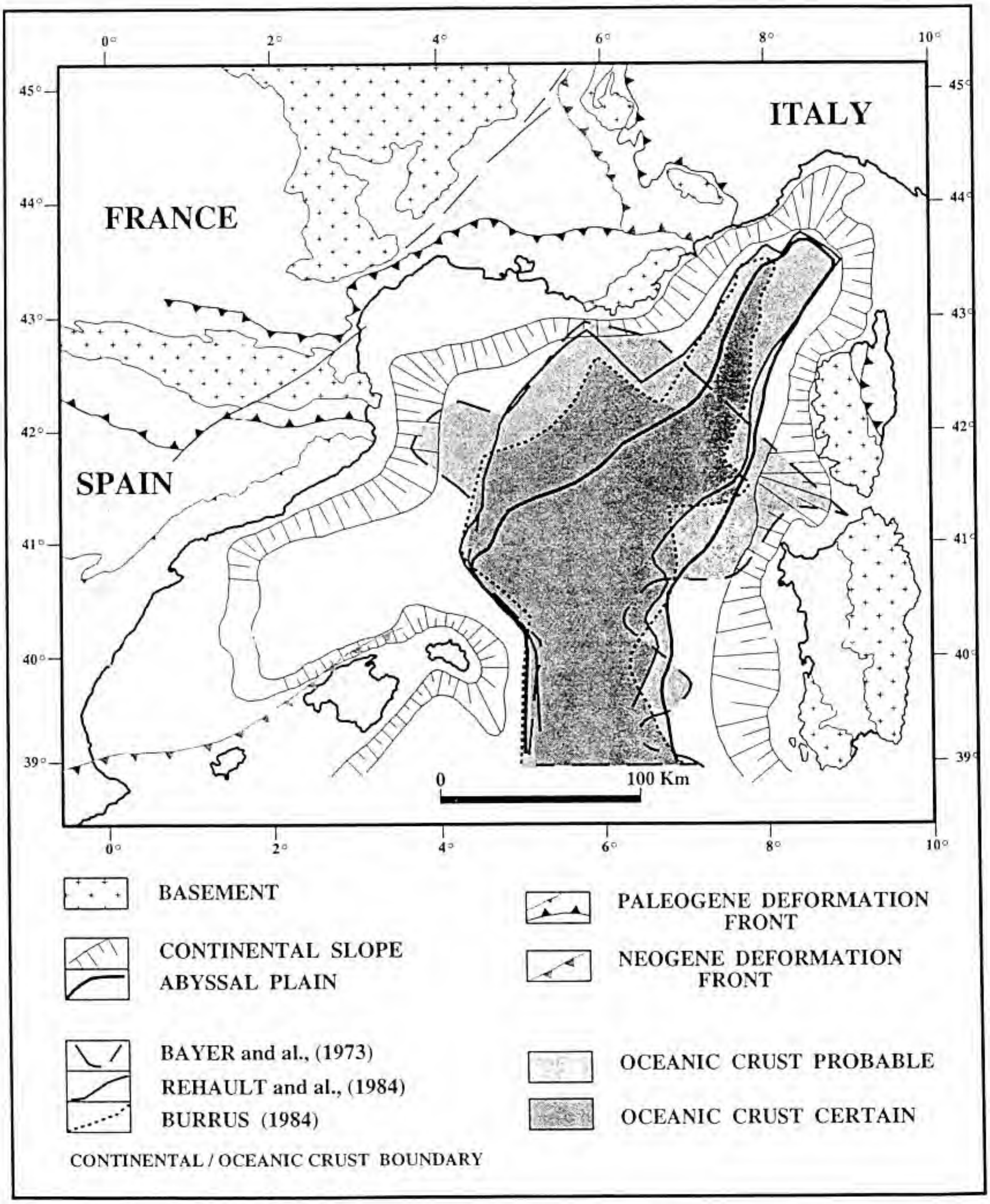

FIG. 3. Outlines of oceanic crust in Provençal Basin according to different authors.

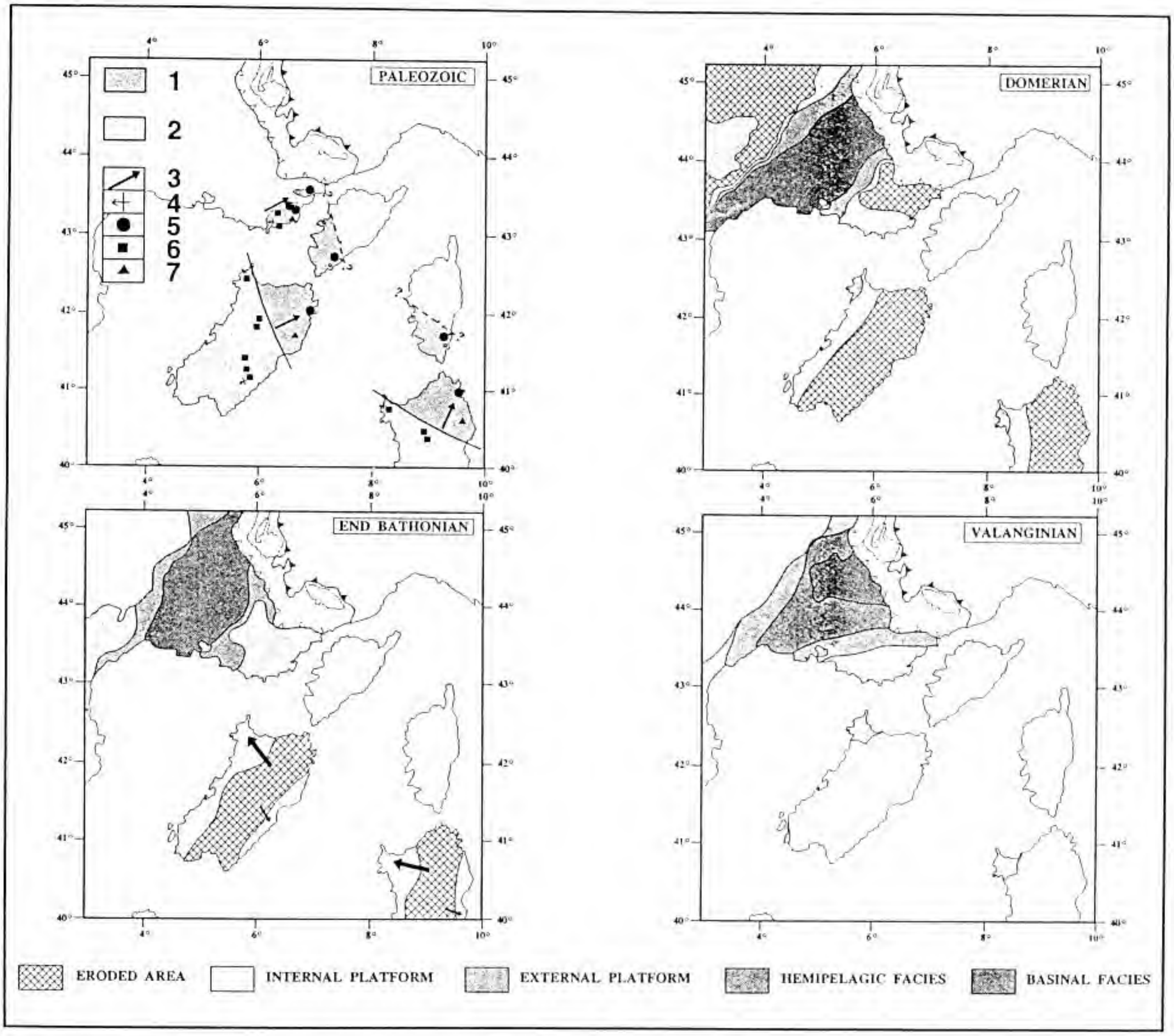

FIG. 4. Paleogeographic sketchmaps of Gulf of Lions area, showing Corsica-Sardinia Block in its pre-drift position.
Paleozoic (after Orsini et al., 1980)
1- metamorphic belt, intermediate pressure; 2- metamorphic belt, intermediate to low pressure; 3- direction of increasing degree of metamorphism; 4- fold axis; 5- granulite facies; 6- metamorphosed continental alkaline basalts; 7- metamorphosed tholeitic basalts.
Mesozoic reconstructions (after Chabrier and Fourcade, 1975; Chabrier and Mascle, 1975; Azéma et al., 1977; Fourcade et al., 1977; Alleman, 1978, Philip and Alleman, 1982; BRGM, 1984).

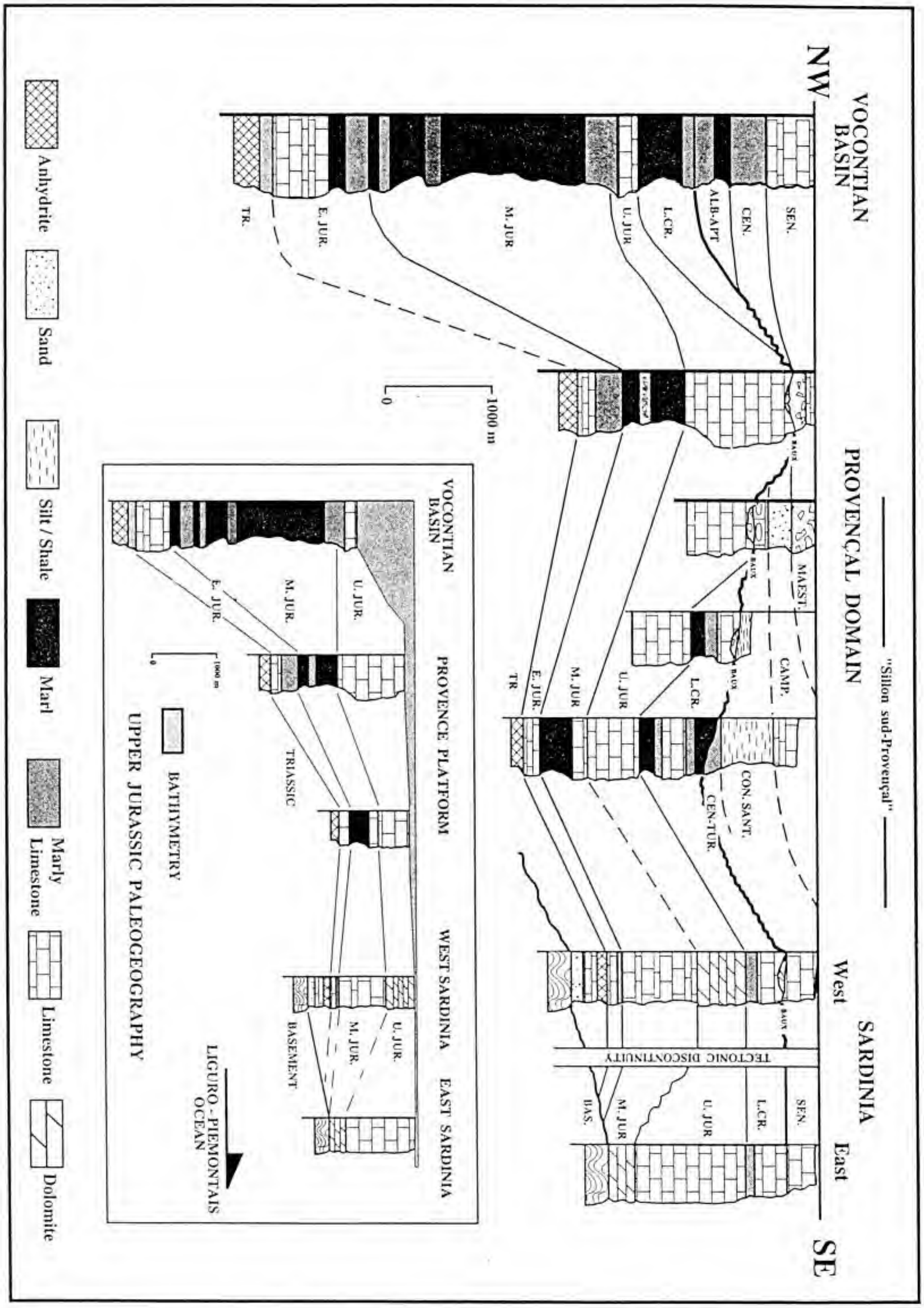

FIG. 5. Mesozoic straticraphic cross-section through Vocontian Basin, Provençal domain (after Aubouin, 1974) and Sardinia, datum top Late Cretaceous. Inset: Late Jurassic basin geometry, showing bathymetry.

firming the rotation of the Corsica-Sardinia Bock (Fig. 6; Chabrier and Fourcade, 1975; Trémolières et al., 1984).

Mechanisms Responsible for the Rotation of the Corsica-Sardinia Block

Mechanisms responsible for the Tertiary evolution of the Gulf of Lions area are clearly related to the Alpine convergence of Africa and Europe (Olivet et al., 1982, 1984; Patriat et al., 1982). However, microtectonics analyses (Letouzey and Trémolières, 1980; Bousquet and Philip, 1981; Cherchi and Trémolières, 1984; Letouzey, 1986; Villegier and Andrieux 1987) and the inventory of sea-floor magnetic anomalies (Savostin et al., 1986) indicate significant variations in their convergence pattern. In this respect, the Oligocene beginning of dextral translation between Africa and Europe may have played a significant role (Fig. 6; Ziegler, 1988).

Kinematic models for the development of the Neogene West-Mediterranean basins, involving an Oligocene change in the polarity of the subduction zone along the eastern margin of the Corsica-Sardinia Block from East-dipping to West-dipping, date back to the early 1970's (Boccaletti and Guazzone, 1972; Auzende et al., 1973; Cocozza and Jacobacci, 1975). Tapponier (1977), based on an analogy with a 'rigid/plastic' deformation model, envisaged development of the Alpine arc system and the Mediterranean basins as being the consequence of a horizontal redistribution of continental masses, induced by the collision of Africa and Europe. This view does not clash with the previous hypotheses and provides a plausible explanation for the opening of the Gulf of Lions under an overall compressional regime (see also Ziegler, 1988, 1994).

Neogene Evolution of the Gulf of Lions

Since the development of numerical models, which relate crustal stretching, thermal perturbation of the lithosphere and tectonic subsidence (McKenzie, 1978; Wernicke, 1985), the Neogene evolution of the Gulf of Lions has been considered as being governed by the dissipation of the lithospheric thermal perturbation which was induced by Oligocene rifting and early Miocene crustal separation (Steckler and Watts, 1978; Bessis, 1986; Burrus, 1989, Kooi and Cloetingh, 1992). In these models, which considered the post-Burdigalian development of the Gulf of Lions Basin to be of the passive margin-type, the Messinian 'salinity crisis' (Cita, 1973) accounts for a major incision. Messinian isolation of the Mediterranean Sea from the world oceans caused a significant lowering of the erosional base-level, inducing in the upper parts of the Gulf of Lions margin deep erosion of Miocene and older strata and in the Provençal Basin deposition of a thick salt series on oceanic crust.

Quantitative subsidence analyses by Bessis (1986) and Burrus (1989) showed that, according to numerical models, the post-rift subsidence of the Gulf of Lions Basin was much greater than implied by its syn-rift subsidence (Fig. 7a). Underlaying reasons may be seen in the initial stretching conditions of the area, as well as in its overall tectonic setting.

Standard models of lithosphere stretching (McKenzie, 1978) assume an initial crustal and lithospheric thickness of 30 and 120 km, respectively. However, in the Gulf of Lions, Oligocene rifting followed on the heel of the Pyrenean orogeny and therefore affected a thickened lithosphere and crust that probably was characterized by a considerable topographic relief. Taking this into account, stretching factors determined from syn- and post-rift subsidence agree more closely (Fig. 7a), particularly for the moderately stretched portions of the margin (stretching factor <1.8). However, a different explanation must be sought for the strongest attenuated, distal parts of the margin.

Although the Gulf of Lions has been interpreted for many years as a classical passive margin, this concept must be revised in view of the continued convergence of Europe and Africa during Neogene times. Compressional stresses can cause lithospheric deformations at wave lengths of over 100 km, resulting in uplift of broad arches and accelerated subsidence of basins (Cloetingh, 1988;

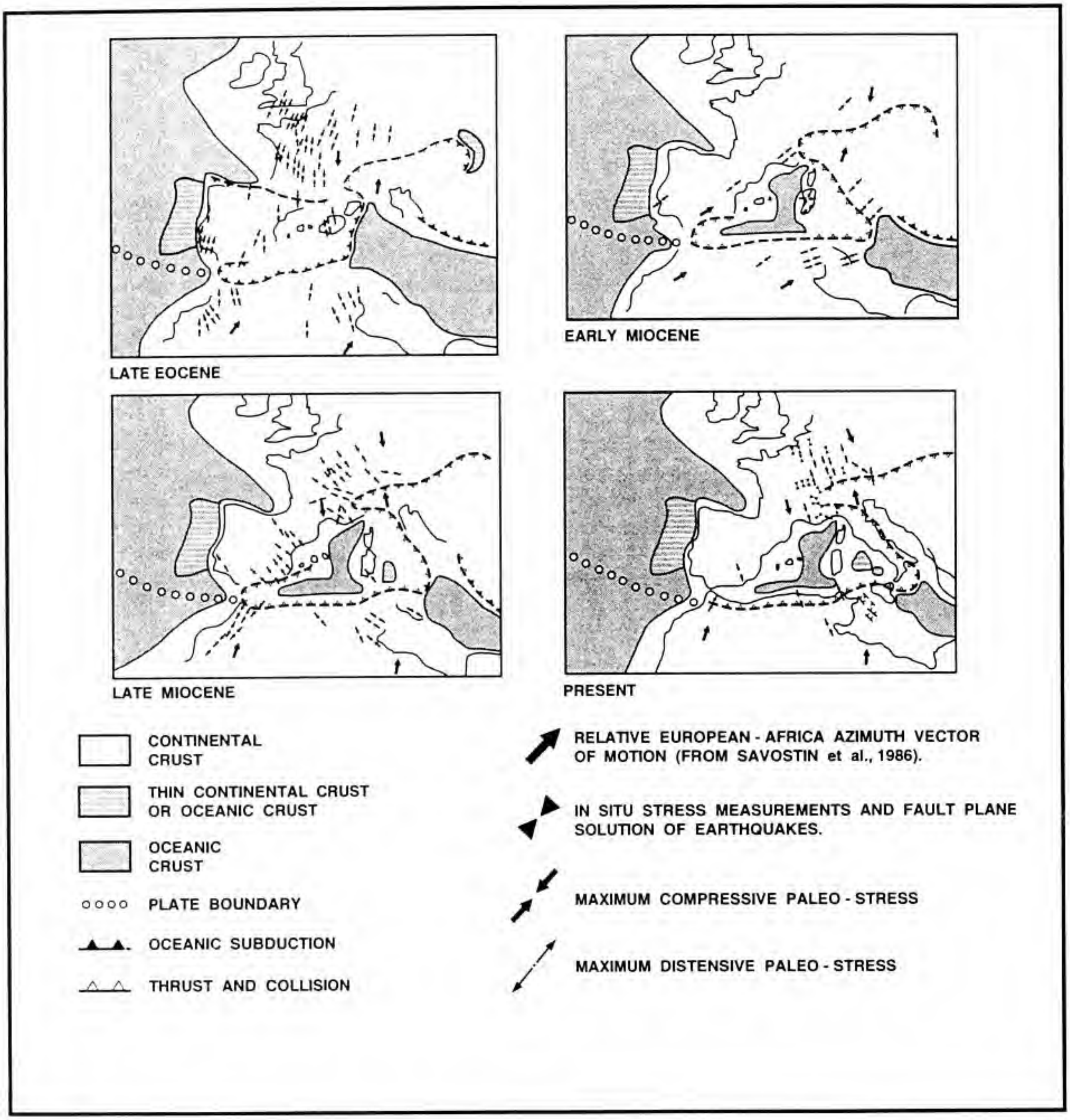

FIG. 6. Cenozoic paleo-stress systems of the West-Mediterranean area (after Letouzey, 1986).

Nikishin et al., 1993). Stress-induced accelerated subsidence of the Provençal Basin and its margins, accompanied by an exceptional rate of sediment supply, may therefore explain the observed 'abnormal' tectonic subsidence of the most extended, distal parts of the Gulf of Lions shelf and of the young, and therefore weak, oceanic lithosphere of the Provençal Basin.

GULF OF LIONS IN THE CONTEXT OF THE PYRENEAN OROGEN

During the Late Cretaceous convergence of Africa-Arabia with Eurasia, the Italo-Dinarid Block began to collide with the southern margin of Europe. At the same time subduction zones propagated westward into the West-Mediterranean domain. Convergence of Iberia with the Europe began during the Santonian and culminated in their Paleogene suturing along the Pyrenean orogen (Ziegler, 1988; Dercourt et al., 1993). The Pyrenean orogeny strongly affected also the area of the Gulf of Lions and adjacent domains.

During the Senonian first compressional deformations gave rise to the development of E-W trending folds in the Languedoc and Provençe area. However, the paroxysmal phase took place during the Eocene. Last compressional deformations were nearly synchronous with the first extensional movements which ultimately culminated in the opening of the oceanic Provençal Basin. Accepting the counter-clockwise rotation of the Corsica-Sardinia Block, we constructed two schematic palinspastic cross-sections in an attempt to clarify the relationship between Pyrenean structures on the Provençal-Languedoc-Gulf of Lions margin and those of Corsica-Sardinia (Fig. 8).

Provençal Transect (Fig. 8a)

This cross-section extends from the Vocontian domain of southeastern France to the ophiolitic nappes of Corsica; the latter represent the most internal units in this transect. Since the position of Corsica at the end of the Pyrenean orogeny is quite well known, this transect is fairly well constrained.

In the Vocontian domain, Late Cretaceous to Eocene compression gave rise to the development of E-W trending folds which are detached from the basement at a Triassic salt layer. Overall, the Vocontian domain represents an inverted extensional basin (Roure and Colletta, this volume) which is located in the foreland of the Provençal Pyrenees. Further South, in Provençe, the tectonic style changes due to a sudden thickness decrease of the Mesozoic series (Provençal Platform) and more intense shortening. Here, Eocene compression caused the development of major north-verging, thin-skinned thrust sheets. However, at Cap Sicié (Toulon region), the Palaeozoic basement is involved in the Pyrenean compressional structures.

After a some 50 km wide zone of no information, corresponding to the shelf of Ligurian Sea, the Palaeozoic basement re-appears in Corsica. According to our interpretation, the Corsican Palaeozoic massifs are allochthonous like all structures, palinspastically speaking, located south of the Cap Sicié thrust. In Corsica, a complex tectonic stack, involving basement and reduced Mesozoic and Eocene sedimentary series, representing the original sedimentary cover of eastern Corsica, is wedged between the parautochthonous Palaeozoic basement and the allochthonous "Schistes lustrés" and ophiolite nappes.

Our transect shows that we have to deal with with a typical orogenic belt which consists of an inverted basin (Vocontian Basin), an external thin-skinned thrust belt (Provençal domain), a more internal basement involving thrust belt and internal nappes consisting of obducted oceanic crust. A similarity with the Western Alps is quite apparent.

Languedoc Transect (Fig. 8b)

This cross-section, which extends from the Palaeozoic basement of the Massif Central to the East coast of Sardinia, is relatively poorly constrained. This stem from the lack of information on the nature of the substratum of the Gulf of Lions

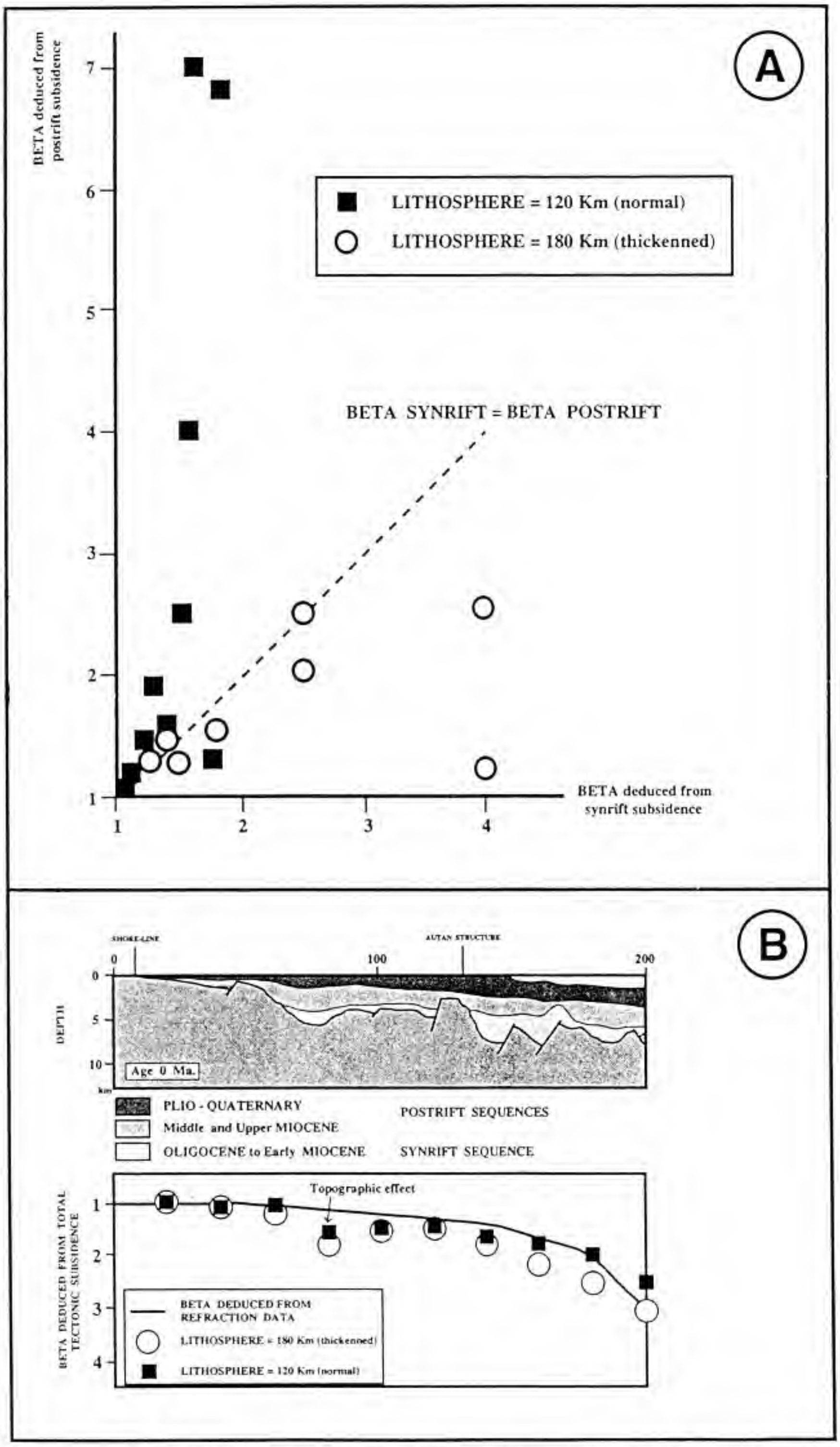

FIG. 7. a) Graph showing stretching factors (Beta) determined from syn-rift (Oligo-Aquitanian) and post-rift (Burdigalian to Present) subsidence. Black squares: after Bessis (1986), white dots: this study.
b) Comparaison between stretching factor determined from total tectonic subsidence (Bessis, 1986 and this study).

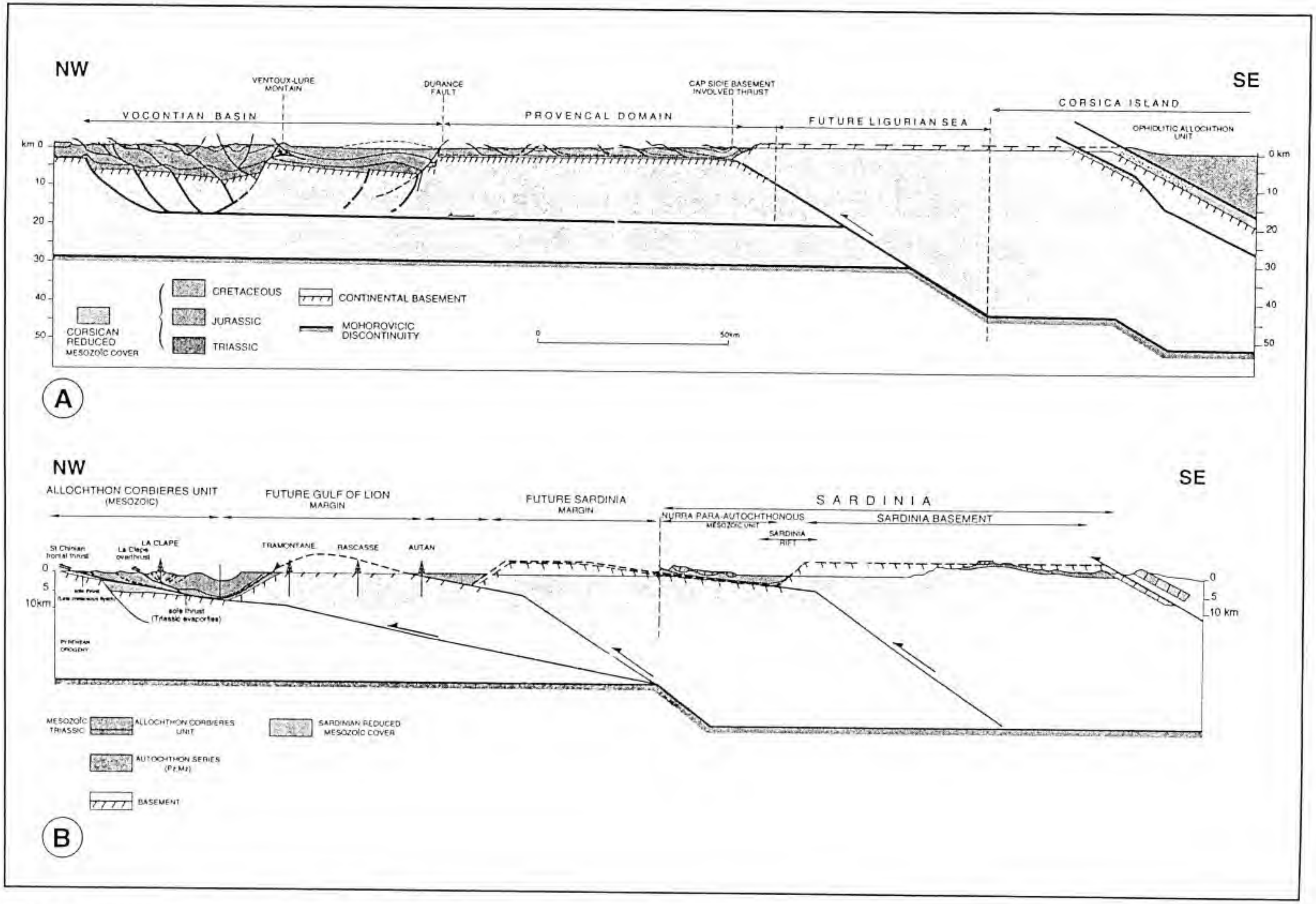

FIG. 8. Conceptual cross-sections through Gulf of Lions at end of the Pyrenean orogeny
a) Cross-section from the Vocontian Basin to Corsica.
b) Cross-section from the Massif Central to Eastern Sardinia.

and uncertainties about the pre-separation position of Sardinia.

The most external structure is the north-verging, thin-skinned Corbière nappe which accounts for about 40 km of shortening (Deville et al., 1994). This nappe may continue southward under the Gulf of Lions. The next, more internal unit, corresponds to the basement high which is defined by the wells Tramontane, Rascasse and Autan. This high is interpreted, in analogy with the Provençal transect, as a ramp-anticline which is carried by the Corbières thrust ramping down into the basement.

After a 60-70 km wide zone of no information, corresponding to the distal parts of the Gulf of Lions and the Sardinian margins, the basement surfaces again in Sardinia where it is covered by north-verging folded and thrusted Mesozoic sediments of the Nurra nappes; these were derived from eastern Sardinia. The north-verging basement imbrications of Sardinia represent the most internal elements of this tectonic edifice.

The very schematic reconstructions given in Fig. 8 raise the question about the relationship between the Corbières nappe and the basement block drilled by the Tramontane, Rascasse and Autan wells and the distribution of Mesozoic sedimentary series beneath the Gulf of Lions. Due to the presence of Mesozoic series in Sardinia, it may be inferred that, prior to the Pyrenean orogeny, Mesozoic sediments had covered the entire area of the Gulf of Lions. During the Paleogene compressional phases, the most internal area were deformed first. On the East coast of Sardinia, due to the absence of a decollement level, Mesozoic strata remained attached to the basement. At a later stage (Fig. 9), the Mesozoic cover of the Gulf of Lion was detached from its basement as the Sardinia basement back-stop advanced northward. At some stage, the Tramontane-Rascasse basement imbrication was activated, resulting in the uplift of a major high in the Gulf of Lions.

Palaeogeographic and structural reconstructions lead us to relate the Corsica-Sardinia Block to a more Alpine than a Pyrenean origin, thus raising the question of the location of the Pyrenean chain beneath the Gulf of Lions. Kinematics models (Olivet et al., 1982) generally assume that the Corsica-Sardinia Block remained attached to the Iberian plate up to Oligocene times. Hence, it was assumed that the boundary between the Iberian micro-continent and Europe projects eastwards through the Gulf of Lions to the Alpine front. However, this model raises a number of structural problems. In Corsica-Sardinia, Eocene compressional structures verge northwards when restoring this block to its pre-drift position. However, the north-vergence of the Corsica-Sardinia Paleogene structures is not compatible with a model which assumes that the Pyrenees extended through the Gulf of Lion as Corsica-Sardinia would be located along the southern margin of such a "Pyrenean" fold belt. The eastward drift of the Iberian micro-continent during Aptian to Early Senonian times, in conjunction with the opening of the North Atlantic, is thoroughly documented. Only the prolongation of the axial zone of the Pyrenean range and the significance of the North Pyrenean Fault, marking the suture between Iberia and Europe, raises problems. This leads us to propose that the Iberia/Europe plate boundary projects from the Pyrenees southeastwards and by-passes the South coast of Sardinia. In such a model, the Languedoc fold belt developed out of an intracratonic rift which formed a branch of the Mesozoic Pyrenean rift system (Fig. 10).

PETROLEUM GEOLOGY OF THE GULF OF LIONS

The Camargue, the Gulf of Lions and its Languedoc margin were explored actively, particularly on-shore, since the 1950's. On the whole, results were disappointing as only three small oil fields were discovered. However, the Gallician (7000 t produced) and the Saint-Jean de Maruéjols oil fields of in the Alès Basin demonstrate that a petroleum system can function in the Oligocene rifted basins of this area. Unfortunately, the eleven off-shore exploration wells were all dry. Yet, it must be pointed out that none of these wells penetrated the Oligocene syn-rift series.

During the last 10 years, a major effort was undertaken to re-assess the hydrocarbon potential of the Gulf of Lions and the geodynamics of its

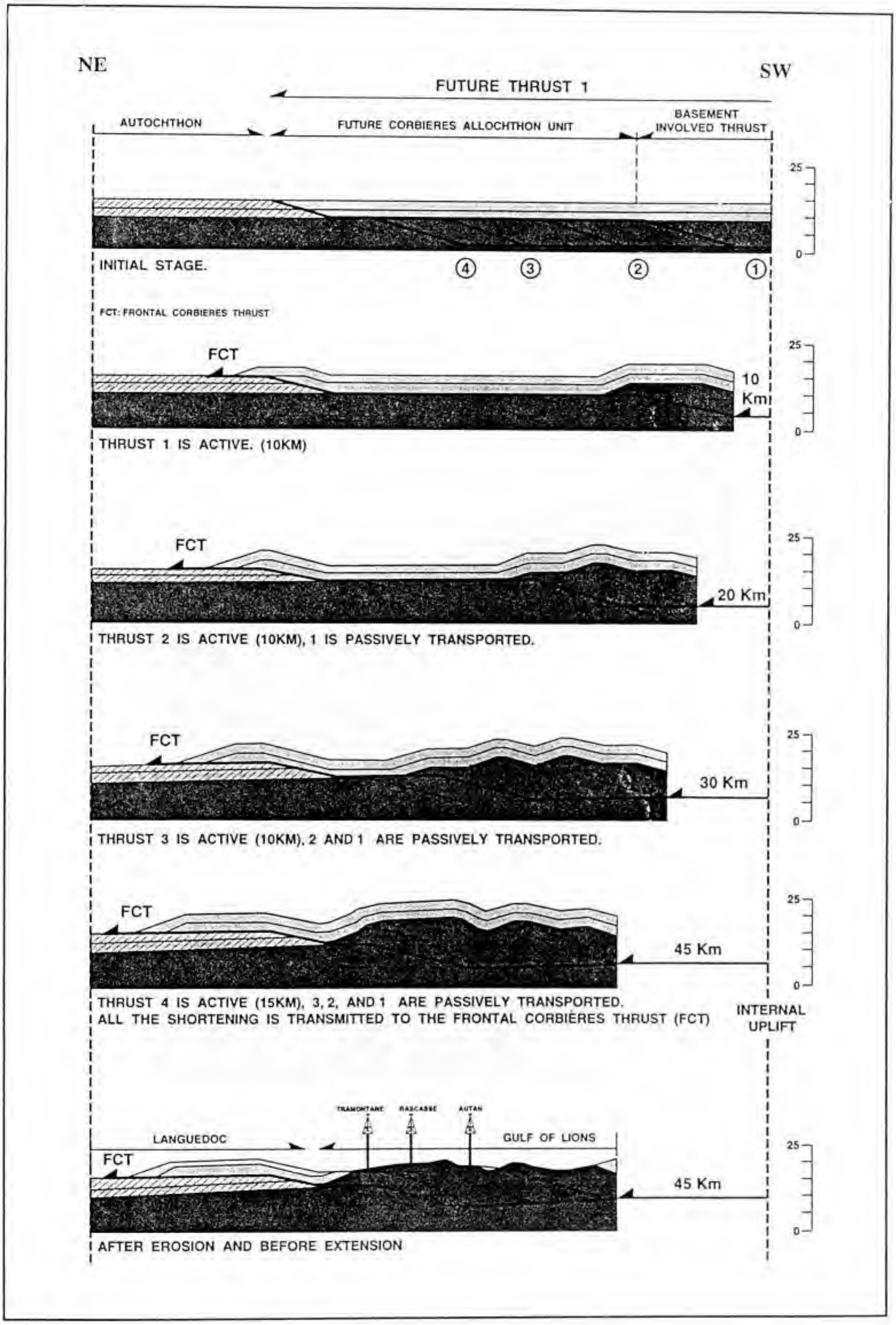

FIG. 9. Cartoons showing develpoment of Corbières allochtonous unit

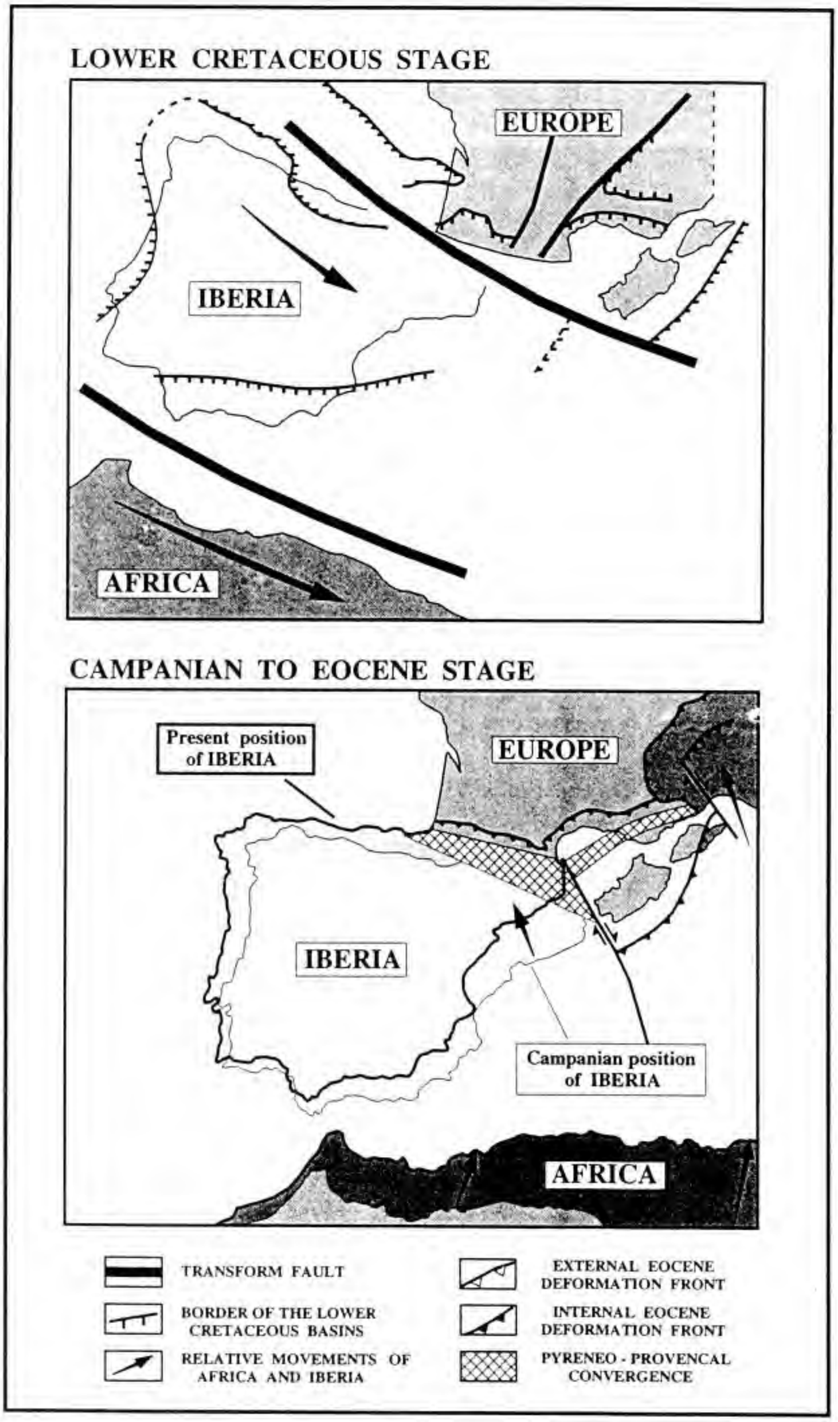

FIG. 10. Plate kinematics of Africa, Iberia, Europe during Early Cretaceous and Campanian to Eocene Pyrenean orogeny (modified after Réhault et al., 1984a)

evolution. In this respect, the availability of a network of regional deep and industry-type reflection-seismic profiles was a great advantage (De Voogd et al., 1991; Guennoc et al., 1994; Gorini et al.,1994; Pascal et al., 1994; Gaulier et al., 1994).

Regional Cross-section

The regional cross-section, given in Fig. 11, extends from the Massif Central through the Gulf of Lions to the edge of the Provençal basin and is complemented by a section through the conjugate western margin of Sardinia. The following four structural domains are recognized:

Onshore, Oligocene extensional basins developed along the Cevennes, Nîmes and Durance faults on a complex Mesozoic substratum (Roure and Colletta, this volume). In the Languedoc, the preferential extensional decollement occurs within Triassic salts which form the sole of the Corbières nappe. Reflection-seismic data reveal highly listric Oligo-Miocene normal faults, rooted in Triassic salts, which do not affect the underlying autochthonous series (Roure et al., 1992, 1994; Mascle et al., 1994; Deville et al., 1994). Although fairly superficial, these normal faults have horizontal throws of the order of 10 km. Therefore, the corresponding amount of crustal thinning must be accommodated further south. Extensional reactivation of pre-existing thrust faults is also thought to be responsible for the development of the large Clape Massif roll-over structure (Gorini et al., 1991). This structure shows evidence of Miocene (Messinian?) compressional reactivation, explaining its orographic expression. Unfortunately, poor seismic resolution at pre-Oligocene levels does not permit to map the southward extension of controlling fault systems in the near off-shore of Gulf of Lions.

On the **proximal parts of the continental shelf**, many NE/SW trending Cenozoic grabens are recognized. The largest of them, the so-called Central Graben, is located between the Tramontane and Rascasse wells, has a width of 25 km and contains up to 3000 m of sediments attributed to the Oligocene-Aquitanian (Mauffret, 1988). The geometry of these grabens is variable and complex but is generally limited by major, southeast-hading listric normal faults. The width and the depth of the Central Graben lead us to speculate that it coincides with the down-ramping of the extensional decollement level from intra-Triassic to intra-crustal levels. This zone correspond to an area where the continental crust is still only moderately attenuated, as indicated by a Moho depth of about 20 to 22 km. The boundary towards the Rascasse horst is very abrupt and is formed by a network of northwest-hading normal faults, having a cumulate throw of 4 km and more.

The **distal parts of the continental shelf**, southeast of the Rascasse horst, are characterized by a series of tilted blocks. In the hanging-wall of these blocks, syn-rift series are relatively thin whereas the total thickness of the Miocene and Plio-Peistocene post-rift series remains fairly constant. Hence, it can be assumed that during the rifting stage this area formed a high which probably had developed already during the Pyrenean compressional phase. To the southeast, the seismic facies of the basement changes drastically across a major southeast-hading normal fault and displays volcanic characteristics. This part of the margin may be either underlain by highly stretched continental crust or may correspond to the transition zone between continental and oceanic crust. In this zone, the Messinian unconformity fades out and gives way to Messinian salts which thicken uniformly seaward. These salts rest on a regionally seaward dipping monocline; their gravitational down-slope gliding gave rise to listric faults affecting the post-Messinian series.

The **conjugate Sardinia margin** is characterized by a structural style closely resembling that of the Gulf of Lions; however, the zone of block-faulting is much narrower and is dominated by east-hading antithetic normal faults (southeast-hading before rotation of Corsica-Sardinia Block). As such, the Corsica-Sardinia margin does not display the typical style of a conjugate margin (opposite polarity of normal faults) but rather forms the prolongation of the Gulf of Lions from which it is now separated by the oceanic Provençal Basin. Syn-rift series crop out only in the eastern parts of the Northwest-Sardinia Basin; its central parts are filled with post-rift sediments and substantial Miocene volcanic flows.

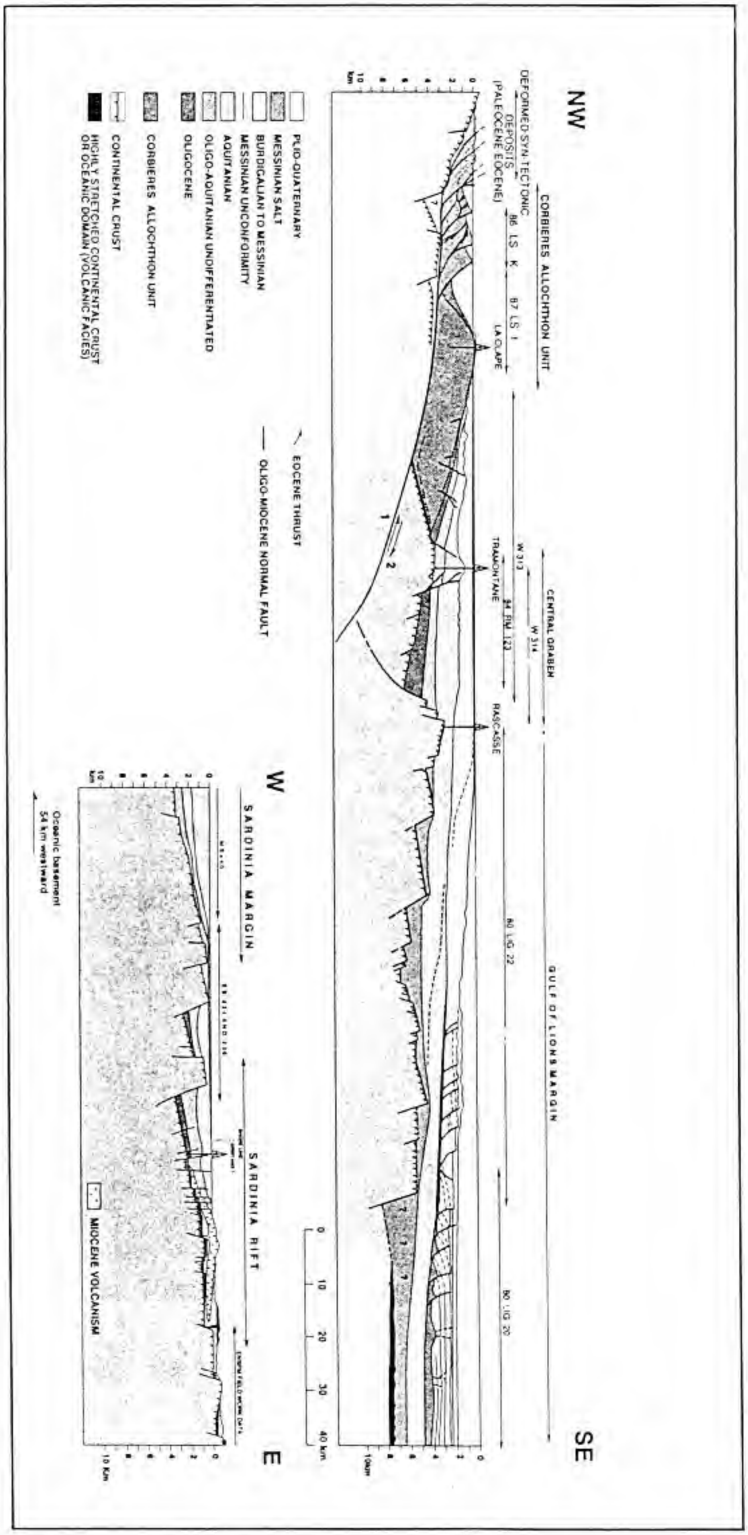

FIG. 11. Cross-section through Gulf of Lions and West-Sardinia margins (modified after after Gorini 1993, Gorini et al., 1991, 1994; Benedicto et al.,1994; Guennoc et al., 1994, Mascle et al., 1994, Séranne et al., 1995).

In recent years, mechanisms of post-orogenic extension were widely discussed (Dewey, 1988). This type of extension, which is related to body forces inherent to orogenically over-thickened crust (Bott, 1993), can give rise to the development of collapse basin and associated tectonic denudation of the deeper parts of a fold belt (Seguret et al., 1989). For the Gulf of Lions, it is difficult to determine the contribution of this mechanism to crustal extension. Yet, it is evident that tensional reactivation of pre-existing compressional faults guided the structuration of the Gulf of Lions (Gorini et al., 1991; Gorini, 1993; Benedicto et al., 1994; Séranne et al., 1995). In this respect, it is noteworthy that not only Pyrenean compressional structures, but also Late Hercynian faults, trending nearly perpendicular to the Oligocene stress direction, were reactivated (Fig. 12).

It can be assumed that in the internal parts of the Pyrenean fold belt thrusts involved all or most of the crust. Their tensional reactivation guided the zone of crustal separation and the opening of the Provençal Basin. Moreover, in areas of basement-involved thrusting, tensional reactivation of thrust-faults caused the development of wider and deeper Oligo-Miocene basins, such as the Vistrenque Trough and the Central Graben. This area is characterized by a moderately thinned crust (stretching factor <1.7), the subsidence and thermal regime of which can be modeled by an uniform stretching model. In contrast, tensional reactivation of the external parts of the Pyrenean orogen involved simple-shear detachment at a supra-crustal level (Triassic salt). Correspondingly, extensional subsidence of the Narbonne and Alès basins was not associated with a localized thermal perturbation of the lithosphere; this explains the lack of their post-rift subsidence.

Potential Reservoirs and Seals

The prediction of potential reservoir/seal pairs in the off-shore parts of the Gulf of Lions Basin relies on the results of the 11 wells drilled and on extrapolations from adjacent, geodynamically related basins. In this respect, we recapitulate that the available off-shore wells failed to encounter syn-rift series and either bottomed in Mesozoic pre-rift series or the basement.

On-shore, the **Oligo-Miocene syn-rift sequence** is confined to narrow grabens, bordered by more or less listric faults. Well data from the Camargue, Alès and Manosque/Forcalquier basins show that the Oligocene series is composed of lacustrine silty marls and limestones and logoonal evaporitic deposits, lacking good reservoir development with significant lateral continuity. The Gallician oil field produced from fractured lacustrine limestones. The absence of good reservoirs, combined with a highly waxy land-plant derived oil, accounts for poor field production.

An other model, applicable to the off-shore parts of the Gulf of Lions Basin, is provided by the Valencia Througt where the syn-rift sequence consists of the lacustrine and marine organic Taraco shales, the Amposta carbonates (Lithotamium, chalk) and conglomeratic scree-slope deposits. However, in the Valencia Trough, principal reservoirs of oil accumulations are formed by karstified Mesozoic carbonates, sealed by Taraco shales and the overlaying Castellon clays (Roca and Delsegaulx, 1992; Torné et al., this volume).

In contrast to the on-shore parts of the Gulf of Lions Basin, off-shore syn-rift series may possibly contain better quality siliciclastic reservoirs due to their proximity to major basement uplifts. This concept is supported by the results of sedimentological studies in the West-Sardinia Basin, on the basis of which a depositional model was developed for potential syn-rift reservoir sands (Fig. 13; Trémolières et al., 1988). Coastal sands, intercalated with carbonates, are developed in the foot- and hanging-wall of the basement-involving Isili block; these sands have porosities of about 30%. Around the Grighine block, the presence of volcanics significantly reduces reservoir qualities. However, in the hanging-wall block, bounded by the Grighine fault, coastal sands, reworked by tides, have good porosity and are relatively clean.; tidal bar sands have porosity between 12 and 30%. Coarse, highly bioturbated carbonate sands, prograding into the basin at the outlet of the transfer corridor between the Isili and Nureci blocks,.reach thicknesses of 50 m. Across a fault, these sand bodies give laterally way to a basinal turbiditic series which is characterized intercalated sandstones and shales; individual sands have thicknesses of the order of 1-2 m.

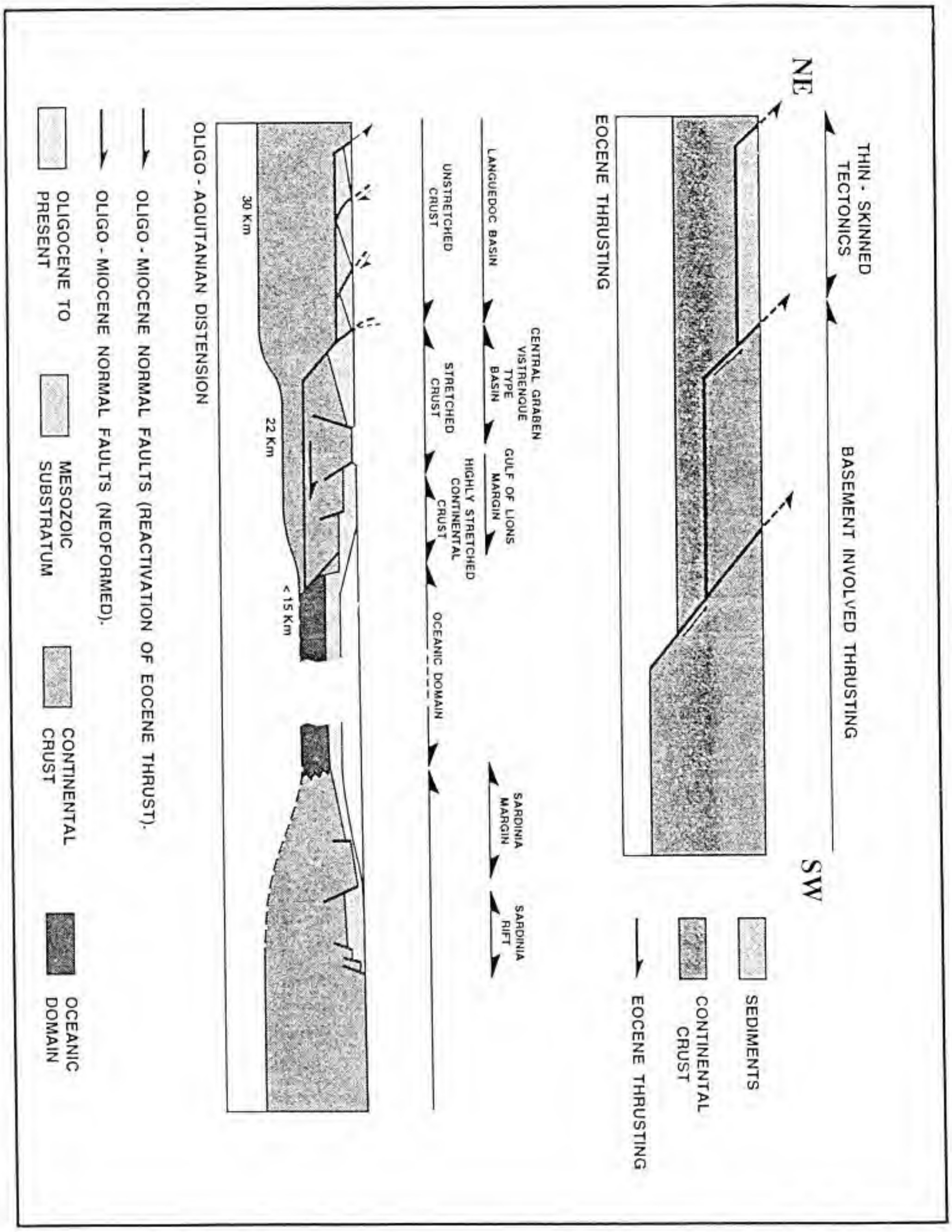

FIG. 12. Extensional model for the Gulf of Lions.

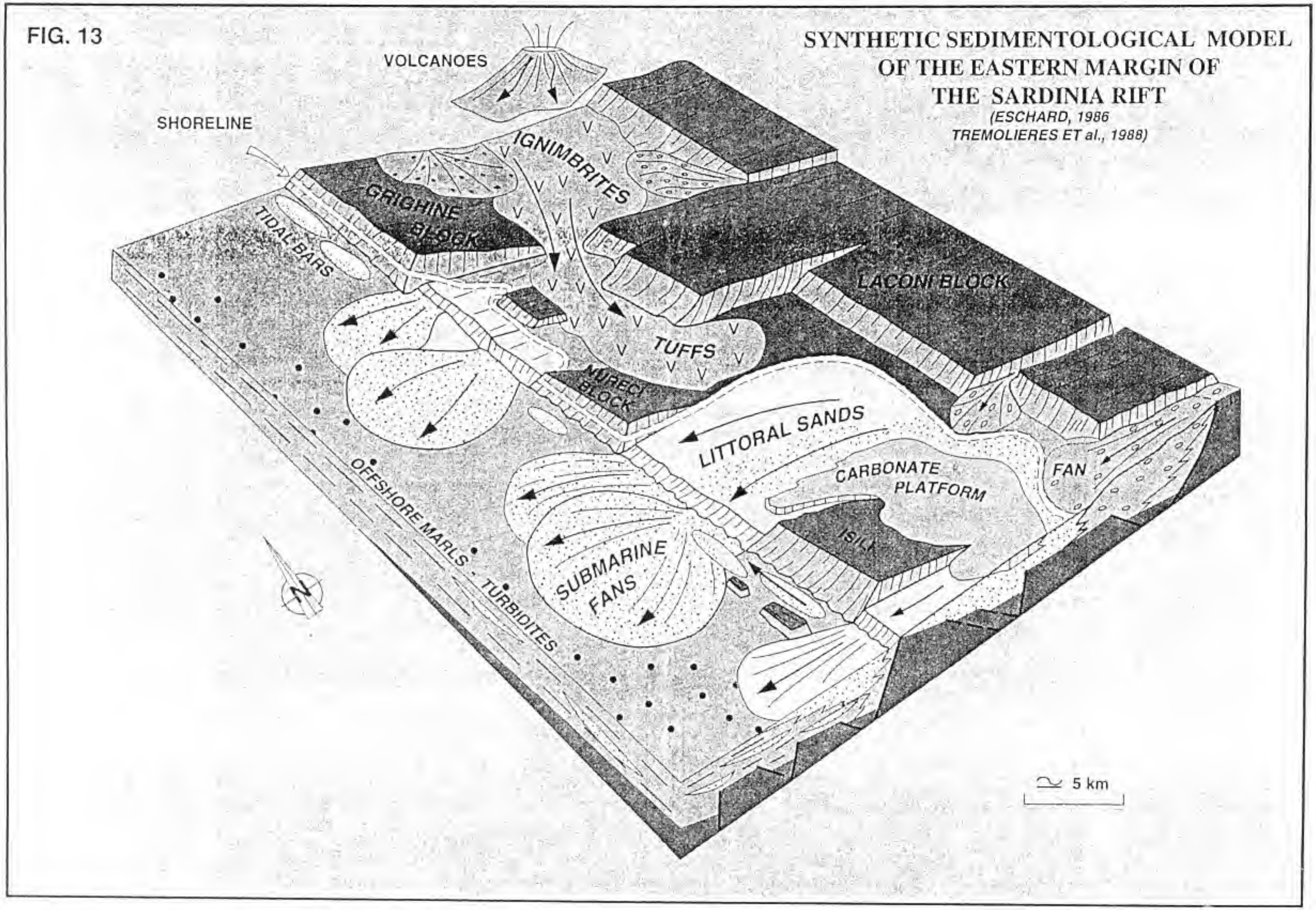

FIG. 13. Deposition model for syn-rift series of the West-Sardinian rift (Eschard, 1986; Trémolièreset al., 1988).

The **post-rift sequence** forms the largest part of the sedimentary fill of the Gulf of Lions and provides a potential seal for the syn-rift series. On the basis of the Messinian unconformity, the entire Miocene to Pleistocene sedimentary package can be subdivided into two first-order depositional sequences (Gorini, 1993). The pre-Messinian sequence commenced with the transgressive basal late Aquitanian-early Burdigalian sub-sequence which accumulated under gradually rising relative sea-levels. It is very thick in palaeo-depression and onlaps the rift topography; in platform areas this sequence is developed in a carbonate platform facies (Tramontane and Rascasse wells). This basal transgressive unit is followed by the deltaic, seaward prograding mid-Burdigalian-late Tortonian sub-sequence which accumulated under rising sea-level conditions, as indicated by well-developed top-sets. In shelf areas, the Messinian rapid drop in sea-level gave rise to a major down-cutting unconformity, whereas in deeper waters, thick salts accumulated in depositional continuity with the preceding unit. On the basis of reflection-seismic data, the Messinian drop in erosional base level was of the order of 1000 m; it included a true sea-level drop and an isostatic rebound component, induced by water unloading (Ziegler, 1988). With the post-Messinian rise in sea-level, normal marine conditions were re-established. The prograding Plio-Pleistocene sequence of the Rhône delta and its associated deep sea fan accumulated under glacio-eustatically oscillating sea-levels.

Potential Source-rocks

Potential source-rocks occur in the Oligocene syn-rift series as well as in Palaeozoic and Mesozoic pre-rift series. Although post-rift sequences are devoid of source-rock development, they may have generated biogenic gas.

Oligocene source rocks were identified in the Alès, Camargue and Manosque/Forcalquier basins where they were deposited under lacustrine to lagoonal conditions (Fig. 14). Rock-Eval analyses (Espitalié et al., 1986) indicate the presence of lacustrine type I and terrigenous type III source-rocks with an input of higher land-plants. Mixtures between type I and III source-rocks reflect very rapid lateral and vertical variation in the sedimentary environment. Lacustrine limestones have the best oil generation potential. Such limestones have a type I TOC content of up to 20% in the Camargue Basin and over 10% in the other basins. Moreover, at immature levels, these source-rocks have Hydrogen Index values higher than 730 (930 in the Camargue Trough), indicating an excellent oil generation potential. Due to the wide thickness variations of Oligocene series, the degree of maturation ranges from immaturity in the Manosque/Forcalquier Basin to the beginning of the oil window in the Alès Basin and to the gas window and overmaturation in the Camargue Through. Intercalated with these lacustrine deposits, more detrital layers can contain coals or lignites (type III), characterized by high TOC values but a relatively low Hydrogen Index, indicating a mediocre to weak oil generation potential (essentially gas). Although such levels are found throughout the Oligocene series, they mainly occur at its base where they attain a higher maturity than the shallower, oil-prone lacustrine deposits.

Stephanian coals are known from the Alès and Lodève basins. These very mature coals offer a weak petroleum potential and are essentially gas prone. **Early Permian (Autunian) lacustrine shales** occur in the Lodève Basin; they contain type I/III organic matter and are thought to have sourced the Gabian oil field (Mascle et al., 1994). Beneath the Gulf of Lions, the geographic distribution of Late Palaeozoic basins is unknown. Furthermore, Permo-Carboniferous series appear to have attained a high degree of organic maturity, partly already during the Permian.

The **Toarcian 'Schistes cartons"** have significant regional distribution and display a good residual petroleum potential. However, these shales attained overmaturity in the South-East Basin and in the Corbière nappe prior to or during the Pyrenean orogeny whereas they are still immature along the southeastern margin of the Massif Central.

Campanian oil shales (Type I/III), occurring in the Alès Basin, offer a good petroleum potential and, at present, have just entered the oil window. Campanian lignites are reported from the Provençal margin. The geographical distribution these source-rocks is, however, unknown.

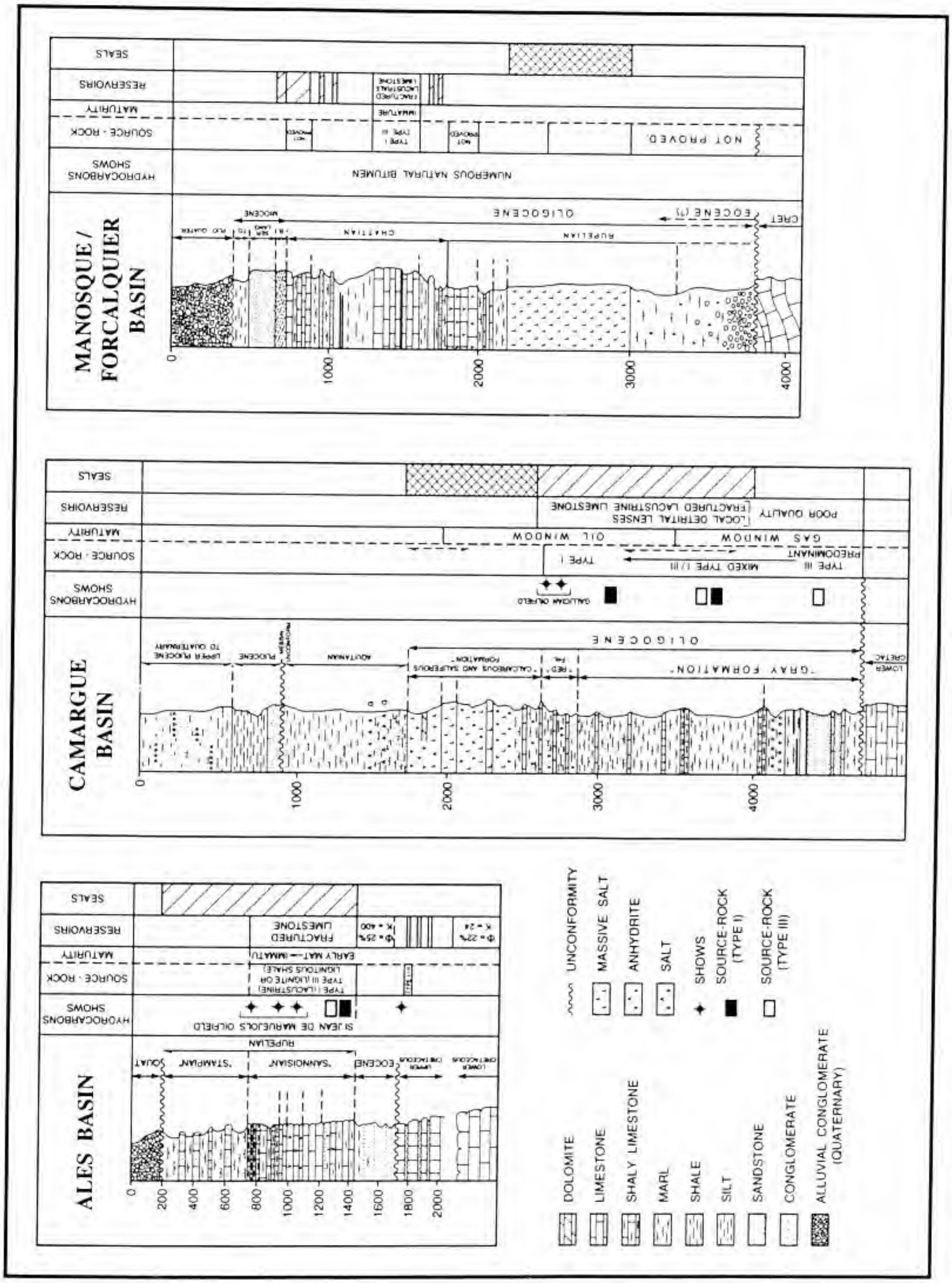

FIG. 14. Summary lithological colums for Alès, Camargue and Manosque/Forcalquier basins

The distribution of Palaeozoic and Mesozoic sediments beneath the Gulf of Lion is unknown. However, even if present, it is likely that they have reached a high level of organic maturity, either due to the depositional thickness of the Mesozoic series and/or due to tectonic overloading during the Pyrenean orogeny. Therefore, the generation potential of Palaeozoic and Mesozoic series must be heavily discounted. On the other hand, it should be noted, that the oil accumulations of the Valencia Trough were charged by source-rocks contained in the Mesozoic series as well as by the Taraco shales (Torné et al., this volume).

REMAINING HYDROCARBON POTENTIAL OF THE GULF OF LIONS

Although eleven exploration wells have been drilled in the Gulf of Lions, the petroleum potential of the syn-rift series, which on-shore contains small oil accumulations, paradoxically has not been tested. The hydrocarbon potential of this large, under-explored off-shore area (<1 well/1000 km^2) still remains to be demonstrated. New concepts developed on the evolution of this basin and the reservoir potential of the syn-rift series may provide some encouragement for further activity.directed towards the evaluation of its still untested Oligocene extensional troughs.

The available geological, geophysical and geochemical data enable us to model the evolution and maturation history of these basin (Fig. 15) Geodynamic considerations lead us to postulate that the the Gulf of Lions Basin evolved in response to uniform extension with stretching factors ranging in its different parts between 1 to 1.8. Based on this assumption, we reconstructed the heat-flow of this basin through time (McKenzie, 1978). Due to a fairly low stretching factor and a high sedimentation rate, the Gulf of Lions Basin is rather cool. The one-dimensional model constructed by means of GENEX software, shows that the oil window is only reached around a depth of 3500 m. Hence, only basins with a depth of 4000 m and more are likely to generate oil and gas, provided source-rocks are present. In such basins, whatever the type of the source-rocks (type I or III), maturation and expulsion of oil and gas commenced during Mid-Miocene to Pliocene times; at present, source-rocks have reached maximum maturity. Clearly, hydrocarbon generation and expulsion post-dates the formation and sealing of traps.

Under such a scenario, consideration must be given to potential reservoir/seal pairs involved in traps having commercially attractive volumes.

Based on the depositional model developed for the syn-rift series of the West-Sardinia Basin, **intra-Oligocene sands** (Fig. 16) may be better developed off-shore, where rift flanks are upheld by basement, than on-shore where on graben flanks Mesozoic carbonates were subjected to erosion. Such sands may occur in the basal transgressive unit and along fault scarps. Laterally these sands may interfinger with basinal shales and carbonates having a source-rock potential. Clearly, reflection-seismic data would have to be of sufficient quality to permit seismostratigraphic analysis of the syn-rift basin fill and the identification of potentially sand-prone facies. Within the syn-rift fill of Oligo-Early Miocene grabens, structural, stratigraphic or combination traps can be anticipated.

Pre-rift sediments may also provide reservoirs (Fig. 16). For instance, in the wells Calmar and Adge, oil shows were recorded in karstified Mesozoic carbonates and Palaeozoic sediments, respectively. In analogy with the Valencia Trough, karstified and fractured Jurassic carbonates can be regarded as viable reservoirs, assuming they are preserved in a down-faulted position beneath the Oligo-Early Miocene syn-rift sequence, providing for hydrocarbon charge and seals. On intermediate fault blocks, such carbonates may be sealed by the early post-rift pro-delta clays.

Bacterial gas can be expected to occur in the Rhône submarine fan delta which is characterized by high sedimentation rates. Unfortunately, so far, no bright or flat spots, gas hydrate reflections or gas chimneys have be pointed out on the available reflection-seismic data.

The remaining hydrocarbon potential of the Gulf of Lions is questionable and difficult to assess for want of a clear understanding of its pre-Oligocene evolution. Although Oligo-Miocene syn-rift sediments can have a source-rock poten-

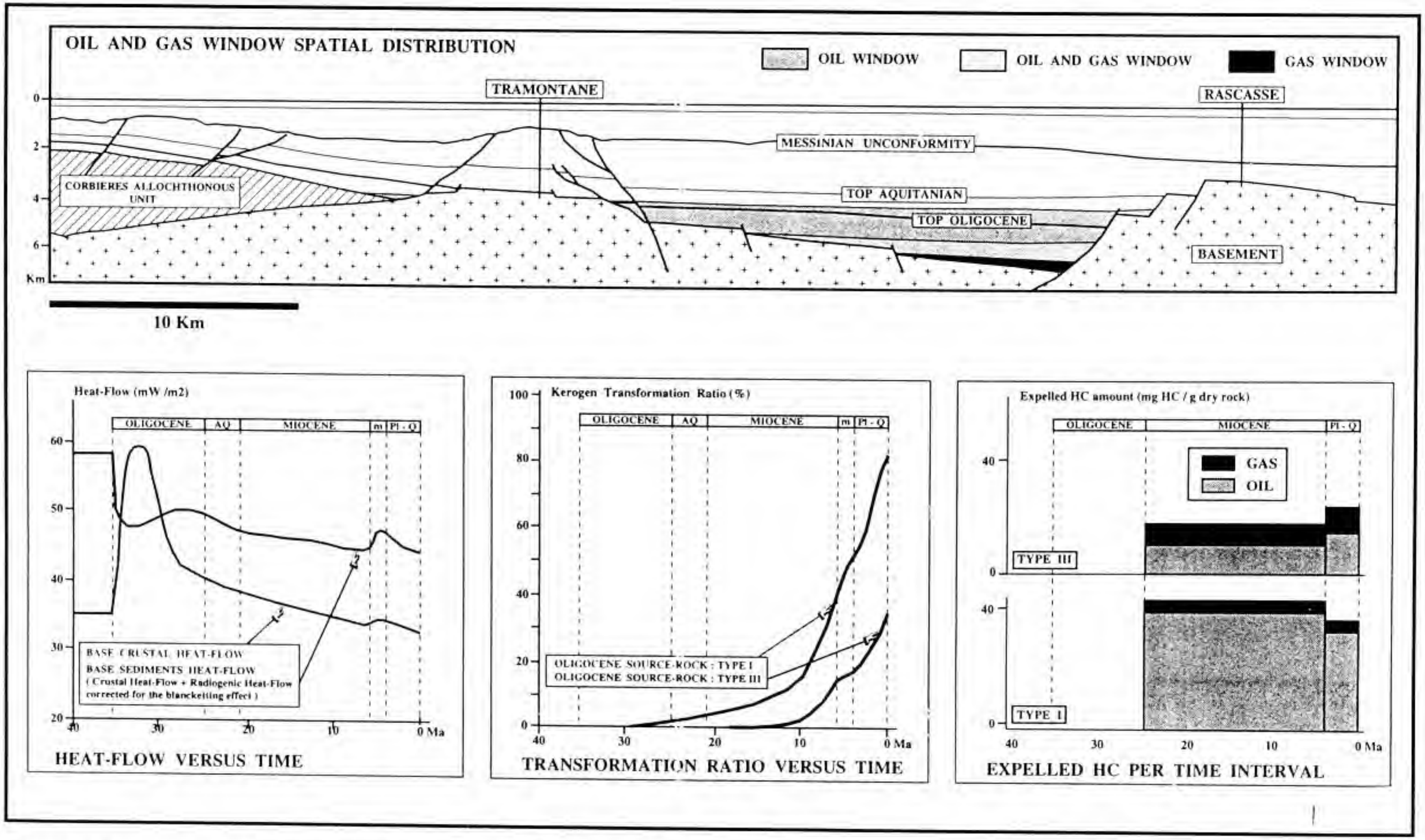

FIG. 15. Evolution of possible hydrocarbon kitchens in off-shore Gulf of Lions

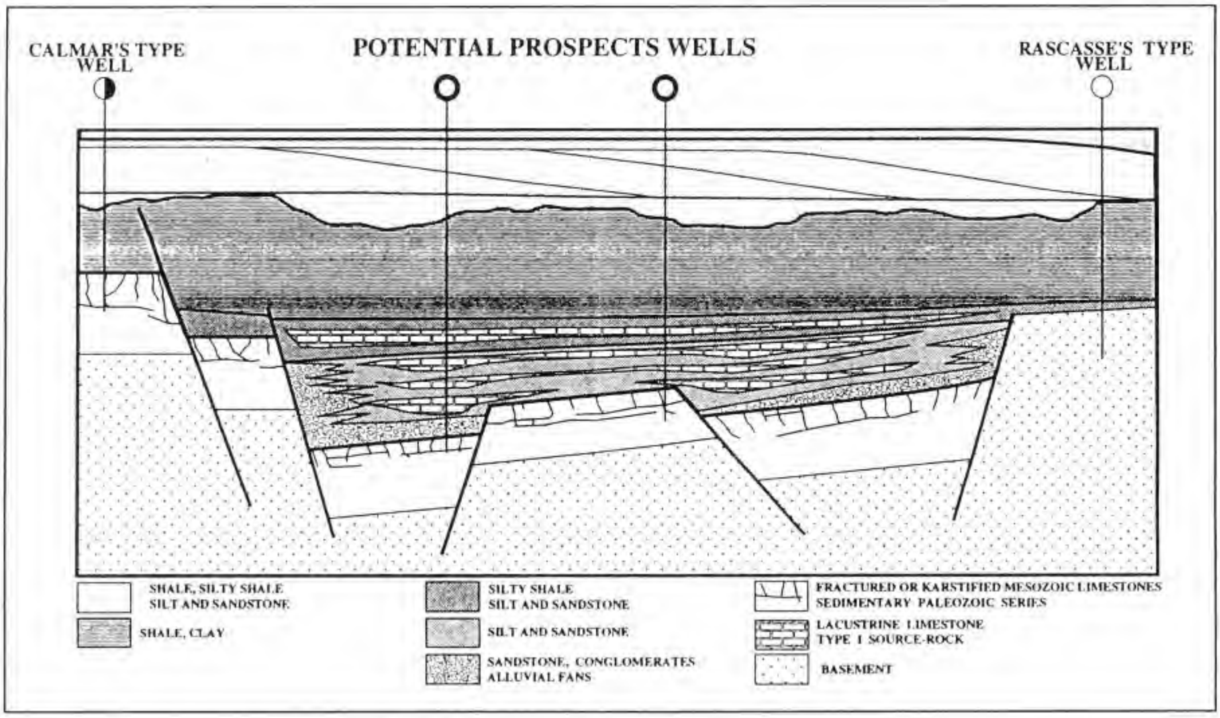

FIG. 16. Play concepts for the Gulf of Lions.

tial, the development of reservoir-seal pairs in the syn-rift series depends largely on the availability of a clastic source in the Gulf of Lions. The distribution of Mesozoic sediment, containing both potential source-rocks and reservoirs, remains an open question. As long as the resolution of reflection-seismic data cannot be improved, further hydrocarbon exploration in the Gulf of Lions.Basin will have to contend with major uncertainties and risks.

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The Aquitaine Basin: oil and gas production in the foreland of the Pyrenean fold-and-thrust belt New exploration perspectives

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ABSTRACT

The Aquitaine Basin of southwestern France lies in the foreland of the Pyrenean fold-and-thrust belt. Since the first gas discovery in 1939, the area has produced as of December 1993 a total of $287 \cdot 109$ m^3 of gas (±10 TCF), $10.2 \cdot 10^6$ t of condensates ($75 \cdot 10^6$ bbl) and $12.3 \cdot 10^6$ t of oil ($\pm 90 \cdot 10^6$ bbl).

The structural evolution of this basin was strongly influenced by early basement tectonics dating from the Variscan and Hercynian orogenies. The subsequent evolution was governed by extensional block faulting and associated salt diapirism during Jurassic and Early Cretaceous times, and by compressional deformations during the Late Cretaceous through the Oligo-Miocene Pyrenean Orogeny. Thus the Aquitaine Basin underwent a complex evolution, both structurally and stratigraphically.

Kimmeridgian and Barremian shales are the most prolific source-rocks. These have variably entered the oil and ultimately the gas generation windows during Early Cretaceous to Paleogene times. Hydrocarbon accumulations are trapped in tilted fault blocks involving Jurassic and Barremian carbonates, which developed during an Early Cretaceous phase of crustal extension, as well as in Jurassic and Barremian erosional reservoir pinch-outs over salt-induced structures which developed at the same time; these features were inverted to various degrees during the Pyrenean orogeny.

In this particularly complex structural setting, conventional 2D seismic provided generally poor resolution. From 1987 to 1993, over 1300 km^2 of 3D seismic were acquired with the dual purposes of field development and oil and gas exploration. These good quality subsurface data have allowed us to define the distribution, geometries and relationship between the different tectono-stratigraphic units as well as to image structures providing potential hydrocarbon traps. As a result, our understanding of the geodynamic evolution of the entire basin, as well as of the dynamics of its petroleum systems, was greatly enhanced. This new understanding has opened new perspectives for oil and gas exploration in the entire Aquitaine Basin.

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This article includes 6 enclosures.

INTRODUCTION

The Aquitaine Basin of Southwestern France forms the northern foreland of the Pyrenean Mountain Belt (Encls. 1a and 1b). During the last 60 years, exploration in the area has led to the discovery of ultimate recoverable reserves amounting to over $350 \cdot 109$ m^3 of gas (12.5 TCF) and $90 \cdot 10^6$ t of oil ($660 \cdot 10^6$ bbl). Thus, the South Aquitaine area represents the largest gas producing and the second larges oil producing province of France.

Exploration in the Aquitaine Basin started in the 1930's and resulted in 1939 in the discovery of the St. Marcet gas field which has recoverable reserved of $8 \cdot 10^9$ m^3 of gas (290 BCF). The discovery well was drilled on a surface anticline. The potential of the area was later on confirmed by the discoveries of the Upper Lacq oil field in 1949, the giant Deep Lacq gas/condensate field in 1951 ($260 \cdot 10^9$ m^3 gas, 9.2 TCF) and the Meillon gas field in 1965 ($65 \cdot 10^9$ m^3 gas, 2.3 TCF). In the same area, several smaller sized fields, such as Ucha, Lacommande, Rousse and Cassourat fields, each having gas reserves in the 3 to $7 \cdot 10^9$ m^3 range (110-250 BCF), were also discovered in the same period. In the 1970's, exploration interests moved northward towards the basin edge, resulting again in the discoveries of five sizeable oil fields, namely Pecorade, Vic Bilh, Lagrave, Castera Lou and Bonrepos-Montastruc (Encl. 1b).

By now two hydrocarbon fairways are recognized. The southern gas trend is associated with the leading edge and the proximal parts of the foreland of the Pyrenean fold-and-thrust belt. The northern oil trend is tied to the distal margin of the foreland basin (Encl. 1b). The distribution of the oil and gas fields also demonstrates the variety of the area's hydrocarbon occurrences (leading edge of the thrust belt and general foreland area).

In the evolution of the Aquitaine Basin Hercynian basement features played a critical role. The Mesozoic and Cenozoic palaeogeographies and tectono-sedimentary units are overprinted on an inherited framework of basement discontinuities (Villien and Matheron, 1989). The complex, post-Hercynian (Triassic and younger) geodynamic evolution of the basin can be summarized as follows (Encl. 1c):

(1) During the Jurassic and Early Cretaceous, regional extension was related to the opening of the Atlantic Ocean (Canérot and Delavaux, 1986; Canérot, 1987; Canérot, 1989; Villien and Matheron, 1989). Throughout this entire period generally WNW-ESE extensional stresses played a controlling role. The paroxysm of extension took place during the Aptian-Albian, resulting in sinistral transcurrent motions between the Iberian and European plates. Rapid subsidence of Early Cretaceous pull-apart basins in the area of the Aquitaine Basin involved transtensional reactivation of major Hercynian fault zones.

(2) During Late Cretaceous to Oligo-Miocene times, regional North-South compression was related to the subduction of the Iberian Plate beneath the European Plate (Roure and Choukroune, 1992). This was accompanied by the uplift of the intracratonic Pyrenean fold-and-thrust belt (Encl. 1a). The Pyrenees are characterized by an upthrusted internal crystalline core which is flanked by the opposite verging northern and southern external fold-and-thrust belts. The Pyrenean diastrophism resulted in a fundamental reversal of the earlier palaeogeographic setting of the Aquitaine Basin.

This paper summarizes the geological domains, the structural styles as well as the geological evolution of the Aquitaine region. It is based on an integration of numerous wells and the available 2D and 3D reflection-seismic data. In the concluding chapter, the main parameters controlling the petroleum systems of the Aquitaine Basin are described.

REGIONAL GEOLOGY

From North to South the area can be subdivided into three distinct geological provinces, namely the North Aquitaine Platform, the Pyrenean Foreland and the Pyrenean Mountain chain (Encls. 1a and 1b, 2a).

North Aquitaine Platform

The North Aquitaine Platform occupies the northern, distal parts of the Aquitaine Basin (Encl. 2). This stable platform shows a moderately complete Mesozoic and Cenozoic stratigraphic sequence which rests unconformably on the Palaeozoic basement (Encl. 1c). Overlying basal Triassic and Liassic evaporites, the sedimentary section is thin and consists mainly of carbonates. Although good closures associated with large amplitude salt domes do exist, the hydrocarbon potential of this zone is poor due to insufficient maturation of the source-rocks, as well as a lack of efficient seals.

Pyrenean Foreland Basin

This province contains all major discoveries in the Aquitaine Basin. Structuration of the foreland was primarily acquired during the Early Cretaceous extensional phase which controlled the subsidence of the Arzacq and Tarbes basins (Encl. 2). These basins, which contain over 5000 m of Barremian to Albian sediments, are flanked by Early Cretaceous platforms and salt ridges. The latter are located along the margins of these basins. The Arzacq and Tarbes basins were partially inverted during the Late Cretaceous and Tertiary phases of the Pyrenean orogeny.

Pyrenean Fold-and-Thrust Belt

The East-West striking Pyrenees extend over a distance of some 400 km from the Mediterranean Sea to the Atlantic Ocean (Encl. 1a). Their internal core, formed by upthrusted and out-cropping basement blocks, is flanked to the South and the North by thin skinned fold-and-thrust belts, involving Mesozoic sediments which are detached from the basement at the level of Triassic-Early Jurassic evaporites. On the southern, Spanish side, compressional deformations are mainly Tertiary in age; shortening related to southward thrusting is of the order of 50 to 70 km. On the northern, French side, most of the shortening is concentrated in the internal zone and is primarily Late Cretaceous in age, as documented by the development of a syntectonic Cenomanian to Maastrichtian foredeep basin (Encl. 2). Shortening associated with Tertiary northward thrusting is minor and is thought to be of the order of 20 km.

GEODYNAMIC EVOLUTION

Palaeozoic

The structural framework of the Palaeozoic basement, which was acquired during the Late Carboniferous Hercynian orogeny, was studied along the margins of the Aquitaine area by Winnock (1971), Autran and Cogné (1980), Cogné and Wright (1980), Paris (1984) and others. Two major sets of basement faults characterize these early deformations; these faults strike N110° and N160° and are associated with conjugate systems oriented at N20° and N50°-70°, respectively (Encl. 1b). These fault trends represent important basement discontinuities which were reactivated during the Permo-Carboniferous Late Hercynian tectonic phases. The Mesozoic and Cenozoic palaeogeographies and the respective tectono-stratigraphic units appear to be largely controlled by these main basement fault systems.

Triassic To Early Liassic

During Triassic to Early Liassic times, the entire Aquitaine Basin was characterized by a high rate of tensional subsidence, accounting for the deposition of a thick, uniform sequence of anhydrites and salts across the entire area (Encl. 1c; Curnelle, 1983). The presence of this evaporitic cushion is a key element for the future geodynamic evolution of the Aquitaine Basin, as this ductile body was easily remobilized during subsequent tectonic movements and/or differential rates of sedimentation (Canerot and Lenoble, 1993).

Middle And Late Jurassic

This period was characterized by the early opening phases of the Atlantic Ocean. The Middle and Late Jurassic evolution of the Aquitaine Basin was controlled by a low rate of extension which was guided by WNW-ESE directed extensional stress systems. As a result, the Jurassic subsidence patterns were primarily controlled by the reactivation of inherited faults which strike close to perpendicular to the main stress direction (essentially the N20° and N50°-70° sets of faults). During the Middle and Late Jurassic, the basin was characterized by a calm carbonate platform which generally deepened westward (Encl. 1c).

The Oxfordian to early Kimmeridgian period corresponded to a phase of differentiation of this platform in response to active crustal extension (Encl. 3a; Canérot, 1987). The basin was subdivided into an inner shelf with evaporitic tendencies to the East (dolomites and limestones of the Meillon and Baysere Formations; Encl. 1c) and a more open marine environment to the West (Oxfordian Ammonite Marls and lower Kimmeridgian shaly limestones of the Lower Cagnotte Formation). Following this phase of differentiation, middle to late Kimmeridgian times corresponded to a period characterized by a stable depositional environment; locally condensation was related to synsedimentary salt tectonics along North-West and North-East trending basement faults. These local events, which are related to the reactivation of earlier extensional faults, were of minor importance.

At the end of the Jurassic, the Late-Cimmerian tectonic phase was accompanied by a general regression. The dolomitic facies of the Mano Dolomite (average thickness 200 m; Encl. 1c) is associated with this regressive cycle. During the late Portlandian, the basin became again separated into two domains along a northerly trend which was already evident during the Oxfordian to early Kimmeridgian tensional phase (Encls. 3a and 3b). The eastern shelf was uplifted and subjected to erosion, as evident by the regional deposition of the Portlandian dissolution breccias of the Garlin Formation (Encl. 3b) and the development of a hiatus spanning end-Portlandian to Barremian times. In the western part of the basin, sedimentation continued under gradually increasing water depth; here a dolomitic environment of deposition, controlled by North-South directions, persisted during Portlandian, Berriasian and Hauterivian times.

Early Cretaceous

The Early Cretaceous geological evolution of the Aquitaine Basin was marked by the paroxysm of general East-West extension which governed the gradual and step-wise opening of the North Atlantic Ocean, the mid-Aptian onset of opening of the Bay of Biscay and the ensueing sinistral motion of Iberia relative to Europe.

In the Aquitaine Basin, a high rate of transtensional deformation is evidenced by important strike-slip motions along the N110° and N160° inherited basement faults. Although the major transcurrent motions took place along the European-Iberian plates boundary, the entire region was affected by important strike slip movements. The Jurassic platform was delaminated along these main fault zones (Canérot, 1989), allowing for the development of the deep, confined, lozange-shaped Arzacq and Tarbes basins which are characterized by NW-SE trending depocentres (Encls. 3b and 3c; Bourrouilh et al., 1995). The sedimentary record of these basins permits to establish three major periods of subsidence associated with this extensional phases:

Barremian to early Aptian

The Early Cretaceous palaeogeography was initiated during this period. Depocentres were characterized by very thick deposits, while the surrounding platforms showed a very low rate of subsidence and sedimentation. Rapid subsidence of the Arzacq and Tarbes basins was accompanied by the main phase of diapiric salt movements along their edges. This involved the migration of Triassic and Liassic evaporites away from the subsiding depocentres towards their margins (Encl. 3b).

Latest Aptian to early mid (?) Albian

A platform-basin configuration was acquired during this period. The Arzacq and Tarbes basins were again characterized by very high subsidence and sedimentation rates. The platform-basin transition zones were marked by the development of a system of patch reefs (Encl. 1b) whereas the platforms proper were characterized by low sedimentation rates.

Late Albian

During the late Albian the palaeogeography of the Aquitaine Basin changed completely and was controlled by important tectonic movements taking place along the Iberian-European plates boundary (Peybernes and Souquet, 1984). As a result, subsiding zones shifted to the South towards this very mobile zone, while a major transgression invaded the entire area. Late Albian sediments overlap the earlier basin margins, thus demonstrating a change in controlling stress systems at this time. In fact, this period corresponded to the transition from the general East-West extension, which had prevailed during the Jurassic and Early Cretaceous, to the northerly directed Pyrenean compression which controlled the evolution of the basin during the Late Cretaceous and most of the Tertiary.

The Early Cretaceous palaeogeographic framework of the area is particularly well preserved in the foreland area where Pyrenean inversion movements were of minor importance. The Arzacq Basin in particular (Encls. 1b and 3c) allows to describe the various tectono-sedimentary units which developed during the Early Cretaceous extension. From the edges to the centre of this basin three main elements can be distinguished:

Salt ridges surround the basin on all sides (Encls. 1b and 2). They developed by migration of the Triassic and Liassic evaporites from the subsiding depocentres towards the basin margins (Encl. 3c). Salt ridges are well expressed on the northern side of the Arzacq Basin (Audignon, Garlin, Antin and Maubourguet salt structures, see Encl. 1b) where their original geometries have been only slightly modified during the Pyrenean compression phases (Encls. 2b to 2d). Along the southern basin margin, at the level of the Grand Rieu palaeohigh (Encls. 1b and 2), the salt has either been completely eroded along the axis of the Late Cretaceous foredeep or has migrated away during the Pyrenean compression. The main phase of salt tectonics took place during the earliest Cretaceous period, as shown by Barremian limestones or early Aptian shales, sealing erosional pinch-outs of Jurassic strata along the salt ridges (Encls. 3a and 3c). Latest salt movements, of minor importance, occurred during late Aptian and Albian times; locally this is demonstated by Late Cretaceous sediments resting directly on the salt domes. Along the edges of the Arzacq Basin, traps for the oil and gas fields were clearly formed during this episode of salt tectonics (Encls. 1b, 2d and 3b).

Early Cretaceous Platforms surround the Early Cretaceous depocentres and are located between the basins and the salt ridges (Encls. 1b and 5a). These platforms are characterized by a continuous but slow rate of sedimentation throughout Barremian to late Albian times. The platform-basin transition was marked by the development of Albo-Aptian patch reefs, which have been recognized in several wells (e.g. Lacq, Morlaas, Boucoue and Theze reefs; Encl. 3c). These reef build-ups permit to date the main extensional phase as late Aptian to mid Albian. The lack of oil exploration successes at the level of these reefs can be explained primarily by the lack of efficient top-seals, but also by the remoteness of these features from the main hydrocarbon generating kitchens (Encl. 1c). On the other hand, Jurassic objectives on the platforms are ideally located, namely in structural continuity and updip from basinal areas in which Barremian and Kimmeridgian shaly limestones, representing the principal source-rocks, have reached maturity for oil and gas generation (Encls. 2c and 2d).

The **Early Cretaceous Arzacq and Tarbes basins** correspond to zones of maximum subsidence. Good quality 2D seismic data permit to evaluate the evolution of the Arzacq Basin (Encl. 5a). Differential extensional subsidence of this basin commenced during the Barremian and early Aptian, as indicated by important thickening of the respective sediments from the adjacent platforms into the basin. Halokinetic movements, occurring during this period, can be directly related to the high subsidence and sedimentation rates characterizing this basin. Latest Aptian through middle Albian times correspond to the main episode basin subsidence, as again evidenced by thickening of the respective sediments into the basin. The lack of contemporaneous major salt tectonics suggests, that most of the salt had already migrated away from the basin centre during the Barremian and early Aptian episode of extensional basin subsidence.

Late Cretaceous

The Pyrenean compressional deformation of the Aquitaine Basin clearly commenced at the beginning of the Late Cretaceous. During this time, deformations were mainly concentrated on the internal zone of the Pyrenees where they correspond to the main phase of basement thrusting and faulting. Deformation of the associated forelands involved the development of important uplifts along East-West trending palaeostructures, such as the Grand Rieu palaeohigh and Meillon gas field monocline (Encl. 1b), as well as important strike-slip movements along transverse striking (N20°, N50°-70° and N160°) inherited fault zones.

In this general structural environment the Late Cretaceous palaeogeography was dominated by two major depositional domains (Encl. 3d).

During Cenomanian to Maastrichtian times, the southern parts of the area were occupied by a **thrust-loaded foredeep basin**, located immediately to the North of the internal core of the Pyrenean mountain chain (Encls. 2c and 2d, 3b). This basin corresponds to an E-W trending asymmetric, narrow flysch trough (Dubois and Seguin, 1978). Near the Late Cretaceous deformation front, clastics derived from the rising Pyrenees, attain thicknesses of over 3000 m and thin out northwards (Encl. 2). This palaeogeographic setting was initiated during the Cenomanian; the axis of this foredeep basin migrated northward from Cenomanian to Maastrichtian times (Encl. 3d). Erosional pinch-outs of the Cenomanian to Santonian deposits are associated with pre-Campanian uplifts along the Meillon gas field monocline. As a result of these uplifts, Early Cretaceous sediments were deeply truncated by this unconformity (Encl. 2d). The distribution of the Campanian Soumoulou breccias, which consists of reworked Cretaceous sediments, is limited to the flanks of such uplifted structures (Encl. 3d).

The flysch dominated foredeep was offset to the North by a wide **stable carbonate platform** which persisted during Cenomanian to Maastrichtian times (Encl. 3d; Dubois and Seguin, 1978). It is important to point out that, along the southern limit of this platform, Cenomanian carbonates rest unconformably on late Albian sediments; this unconformity fades out northwards. Cenomanian to Maastrichtian platform carbonates attain thicknesses in the 250 to 1250 m range. The Upper Lacq and the Lagrave oil fields produce from Late Cretaceous platform carbonates (Encls. 1b and 1c).

Tertiary

After a period of tectonic stability at the Cretaceous-Tertiary transition, during which the on average 100 m thick Danian limestones were deposited over the entire foreland, North-South compressional deformation of the basin resumed. The main tectonic pulses occurred primarily during Ypresian, Eocene and Oligo-Miocene times. These compressional phases were responsible for the structuration of the Pyrenean fold-and-thrust belt as well as for limited inversion of the northern salt ridges (Encls. 2b to 2d). The N 20°, N 50°-70° and N 160o striking transverse fault zones, located in the most external part of the foreland, showed again evidence for important strike-slip movements (e.g. Seron fault zone, Encl. 1b).

Following Late Cretaceous basement faulting, Tertiary compression was characterized by thin skinned tectonics in the external parts of the Pyrenean fold-and-thrust belt. The structural geometry of resulting structures appear to be controlled by the earlier palaeogeographies.

For instance, to the West, in the area of the Sainte-Suzanne salient (Encl. 1b), a thick Early and Late Cretaceous sedimentary section overlays thick Triassic and Liassic evaporites (Encl. 2b). Maximum shortening associated with Tertiary thrusting occurred in this zone and involved the activation of a single decollement level, corresponding to the Triassic and Liassic evaporites. In conjunction with these deformations, the Cretaceous basins and salt domes were inverted (Encls. 2a and 2b). Immediately to the East of the Sainte-Suzanne salient, (Encls. 1b, 2a and 2c), the frontal thrust ramps up laterally along the western margin of the Arzacq Basin in the area of the deep Lacq gas field. Still on the same trend, further to the East, in the area of the Grand Rieu palaeohigh (Encl. 1b), most of the thrust deformations were confined to the South of this basement high which corresponds to the southern palaeo-margin of the Arzacq Basin (Encls. 2a and 2d).

As such, these observations highlight the importance of the different structural inheritances (Hercynian, Jurassic, Early Cretaceous) on the structural style of the area. Basement highs, which delimit the Early Cretaceous basins, acted as buttress zones during the Pyrenean compression (Encl. 1b). As a result of this structural configuration, basal thrusts ramp up section from the Triassic and Liassic evaporites to the basal Late Cretaceous unconformity (Encls. 2b to 2d). To the North, most of the Tertiary deformation thus affected only the Late Cretaceous and Tertiary part of the sedimentary sequence, while Jurassic and Early Cretaceous sediments were not deformed. (e.g. Meillon gas field, see Encls. 2c and 2d).

During Late Cretaceous to Oligocene times, Pyrenean compressional deformations propagated northwards, as documented by a piggy-back sequence of deformation. This evolution is evident within the thrust belt itself, where it images a northward migration of the deformation front from the internal zone to its present leading edge. In the foreland, the different periods of active shortening reflect a similar northward progression of deformation and foredeep migration.

PETROLEUM GEOLOGY

Most of the oil and gas fields which were discovered in the Aquitaine Basin are associated either with Late Cretaceous carbonate platforms (e.g. Upper Lacq and Lagrave oil fields), with Jurassic/Early Cretaceous inherited structures located along the northern, western and southern margins of the Arzacq Basin or with partly inverted Early Cretaceous salt ridges along the northern margin of the Arzacq Basin (Vic Bilh, Pecorade, Castera Lou oil fields) as well as its southern margin (Ucha, Lacommande, Rousse and Cassourat gas fields). The largest fields are the Deep Lacq and the Meillon gas fields; these are located in the transition zone between the Early Cretaceous Platform and the southward adjacent deeper-water basin (Encls. 1b, 1c, 2a, 2c and 2d).

Reservoirs (Encl. 1c)

Jurassic dolomites and Barremian limestones represent the main hydrocarbon reservoirs. Their distribution is closely linked to the Jurassic and earliest Cretaceous palaeogeographies. In accordance with the palaeogeographic provinces described above, two domains can be distinguished (Encls. 3a and 3b).

On the **eastern Jurassic shelf**, reservoirs are represented by the early Kimmeridgian Meillon dolomites (average thickness 200 m), the Portlandian Mano dolomites (150-200 m) and the Garlin Breccias (Encls. 1c and 3b). In the Meillon, Ucha, Lacommande and Rousse trend of structures, these reservoirs are totally or partially gas bearing. Although porosities of Jurassic carbonates are rather poor (2 to 4% matrix porosity for the Mano dolomites and 4 to 8% for the Meillon dolomite), effective permeability is primarily provided by fissures and fractures, allowing for good well productivities.

On the **western Jurassic outer shelf** (Lacq, Pecorade and Vic Bilh fields; Encls. 1b, 3a and 3b), only the Mano dolomites are preserved within the Jurassic sequence (Encl. 3b). Although petrophysical characteristics are better here, they remain in average poor (porosity 2-10%). Production is again primarily associated with intensely fractured reservoirs.

In this same area, upper **Barremian limestones** provide a further reservoir, displaying porosities varying between 10 and 15%. Permeabilities are relatively poor but often enhanced by the intense fracturing.

On the **Late Cretaceous platform**, reservoirs are formed by the 200 to 250 m thick Lower Senonian limestones (Encls. 1c and 3d). The good reservoir characteristics of these limestones (10 to 25% matrix porosity) are closely linked to secondary dolomitization in the vicinity of the main Pyrenean faults. The example of the Lagrave oil field demonstrates that dolomitization decreases some distance away from the Tertiary Seron transcurrent fault system, resulting in lateral reservoir deterioration (Encl. 2a).

Source-Rocks (Encl. 1c)

Important reserves of oil and large amounts of gas have been discovered in the Aquitaine Basin, implying that source-rocks of regional extent are available, have a good hydrocarbon generation potential and have expelled significant quantities of oil and gas. Although some source potential has been recognized in Tertiary, Albian and Liassic shales, these formations have contributed little and the main source-rocks are clearly associated with the Barremian and Kimmeridgian formations. This is confirmed by the geochemical source-rock to oil and source-rock to gas correlations, summarized in Enclosures 4a and 4b.

The marine **Barremian source-rocks** contain type II-III organic matter. Across the basin they enter the oil window on average at depths of -3000 m, while the gas window is reached at –4000 m (Encls. 4c and 4d). Hydrocarbon generation and expulsion started during the late Albian in the Early Cretaceous depocentre of the Arzacq Basin and expanded during the Tertiary to both sides of the basin (Encl. 4c).

The marine **Kimmeridgian source-rocks** appear to have the best petroleum potential. Their organic matter is again primarily of type II-III. TOC values range between 2 and 7% (Espitalié and Drouet, 1992) with values of S2 up to 20 kg/t rock.

Trapping Mechanisms

The traps of the different fields are clearly related to structures inherited from the Early Cretaceous extension (platform-basin transition, see Encls. 1b and 4d) and associated salt tectonics along the margins of the Arzacq and Tarbes basins (erosional pinch-outs of the Jurassic and Barremian reservoirs, Encls. 2a, 2c and 3b). These traps were modified during the Pyrenean orogeny which is responsible for the present structural configuration of the area (Villien and Matheron, 1989). Under such a structural setting, fields are primarily located in the proximal (Lacq, Meillon, Ucha, Lacommande, Rousse fields; Encls. 1b, 2d and 3b) and distal foreland (Vic Bilh, Castera Lou, Lagrave fields; Encls. 1b and 4d) and also within the Pyrenean fold-and-thrust belt (Saucede and Ledeuix fields; Encls. 2b and 4d). In the following, selected examples of accumulations are described and their structural evolution discussed.

Rousse and Lacommande gas fields

These fields are located in the proximal parts of the Pyrenean foreland and are contained in structures which developed in conjunction with Early Cretaceous salt tectonics. Acquisition of the

3D Meillon surveys in 1989-1990 has greatly advanced our understanding of the geometry of the Rousse and Lacommande fields and the geodynamic evolution of these traps which produce from erosionally truncated Jurassic carbonates, sealed by Barremian carbonates, early Aptian shales and Late Cretaceous flysch (Encls. 6a to 6c).

Development of these structures was initiated by Barremian to early Aptian diapirism of the Triassic-Liassic salts along the Grand Rieu palaeohigh, forming the southern margin of the Arzacq Basin, resulting in uplift and erosion of the Jurassic reservoir section over the crest of the evolving diapirs (Encls. 3b, 3c, 4d and 6b). The truncated Jurassic reservoirs were sealed by Barremian carbonates and early Aptian and Albian shales (Encls. 3b and 3c). These Early Cretaceous structures were inverted during the Late Cretaceous compressional phase while the Albian sediments above the Rousse and Lacommande structures, as well as over the salt ridges, were eroded in the area of the Grand Rieu palaeohigh along the syntectonic Late Cretaceous foredeep. During the Tertiary compressional phases, the basal Late Cretaceous unconformity acted as a decollement level along which Late Cretaceous and younger series were thrusted northwards (Encls. 2d and 6c).

The structures containing the Lacommande and Rousse gas fields developed therefore very early on during the geological evolution of the area, but were subsequently repeatedly modified. As Jurassic source-rocks were truncated to the North and South of these structures, a local hydrocarbon charge must be implied. Generation and expulsion of hydrocarbons from these source-rocks commenced during the Eocene and persisted during the Oligo-Miocene period and, thus, clearly post-dated the structuration of these traps (Encl. 4d).

These traps rely on a combination of seals. Their southern flanks are sealed by Late Cretaceous flysch whereas their northern, eastern and western sides are sealed by preserved Early Cretaceous sediments which include over-pressured shaly limestones. Top seals are provided by Barremian carbonates and Late Cretaceous flysch (Encl. 6c).

Meillon gas field

Early Cretaceous salt and extensional tectonics and Late Cretaceous uplift and strike slip movements played an important role in the development of the Meillon structure, the geometry of which is also defined by 3D seismic data (Encls. 6a to 6c). This field is located along the platform-basin transition zone on the southern side of the Arzacq Basin (Encl. 3c). The trap corresponds to a 30 km long monocline which dips at 20° to 30° to the North and is upheld by Jurassic and Barremian carbonates (Haller and Hamon, 1993). This structure is bounded on its southern side by an Early Cretaceous normal fault. The monocline is cut by inherited transverse N20° and N160° striking faults.

Initial development of this structure is related to Barremian to early Aptian salt tectonics along its southern limit. Well data show that truncated Jurassic reservoirs are sealed by Barremian and early Aptian sediments on the northern flank of the Meillon monocline. The normal fault, delimiting the field to the South, is related to latest Aptian-middle Albian extension. Late Cretaceous uplift and strike slip motions in the proximal foreland, associated with the early Pyrenean compression phase, are responsible for the present configuration for the field. Tertiary compression affected, however, only the Late Cretaceous and Tertiary sediments which are thrusted northwards over the deep seated Meillon block.

As a result of its configuration, this trap relies on a combination of seals, namely Barremian tight shales and carbonates on the monocline and latest Aptian through middle Albian over-pressured shaly limestones along its southern, faulted flank.

The Meillon field relies for charge on hydrocarbons generated in the Arzacq Basin depocentre. The evolution of the area, as well as organic geochemical studies, suggest two phases of generation and migration of hydrocarbons from the Arzacq Basin into the Meillon structure. During the Early and Late Cretaceous, the structure was charged with oil. Towards the end of the Cretaceous and during the Paleocene-Eocene the Arzacq kitchen entered the gas window; correspondingly, the oils accumulated in the Meillon structure were partly displaced by gas and partly cracked in situ (Encl. 4c).

Giant Deep Lacq gas field and Upper Lacq oil field

Early Cretaceous salt and extensional tectonics and Late Cretaceous and Tertiary compressional deformations contributed to the development of the anticlinal Lacq structure. The recently acquired 3D seismic survey over the Lacq field had a huge impact on the general understanding of the geometry of this structure, its geological evolution which led to the formation of the trap and the dynamics of its petroleum systems (Encls. 1b and 6d).

The Deep Lacq gas field produced from Late Jurassic and Barremian carbonates along the platform-basin transition on the southwestern side of the Early Cretaceous Arzacq Basin. It is located in a particular structural setting, characterized by three main faults trends, striking N20°, N110° and N160° (see Encls. 1b and 3c). The present, trap providing structure is the result of multi-phase deformations.

During the Oxfordian to Portlandian extensional phase, the Lacq field area was located along the eastern limit of the Jurassic open marine outer shelf (Encls. 3a and 3b). During Barremian times, the area remained stable and eustatic sea-level fluctuations mainly controlled the sedimentary evolution of the gas reservoirs.

Due to its particular setting, the Early Cretaceous evolution of the Lacq structure differs from that of other structures. Salt tectonics were initiated during the early Aptian and culminated during the latest Aptian to Middle Albian and thus coincided with the major extensional subsidence phase of the Arzacq Basin. Early growth of the Lacq salt structure is held responsible for the localization of the Aptian-Albian Lacq reef.

Following minor compressional deformations during the Late Cretaceous, the Lacq field acquired its present structural configuration during a phase of Tertiary compression. Most of the movements were located within the Triassic-Liassic evaporites, causing accentuation the asymmetric geometry of the structure by southward migration of the evaporites. The resulting uplift led to the formation of the Upper Lacq structure. 3D seismic data show that the salt swell, upholding the Jurassic anticlinal feature of the Deep Lacq gas field, has a thickness of about 2000 m. Movements along the Ste. Suzanne thrust fault, which ramps up section along the northwestern side of the Lacq structure, took place at the same time (Encl. 6d).

The Deep Lacq gas/condensate accumulation is trapped in the uppermost Jurassic Mano dolomites, Purbeckian to basal Barremian carbonates (lower Annelides, 310-510 m thick) and upper Barremian limestones (upper Annelides, 40-75 m thick). The intra-Barremian laterolog shales (±50 m) separate the two reservoirs. The deeper Jurassic to basal Barremian reservoirs contain the giant gas/condensate accumulation whereas the upper Annelides limestones contain no-commercial gas and oil lenses.

Recent studies by Connan and Lacrampe-Couloume (1993) show that the Jurassic and lower Barremian reservoirs were charged by hydrocarbons generated from the Kimmeridgian Lons Formation whereas the upper reservoir received its charge from the uppermost Barremian "Calcaires à Annelides" which are located within the oil window in the Lacq structure (autochthonous origin). In contrast, the gas and condensate accumulation contained in the Jurassic carbonates was most likely generated by very mature source-rocks; this is compatible with the maturity of Kimmeridgian source rocks in drainage areas offsetting the Lacq structure (Encl. 4d). Moreover, geochemical characteristics of the fluids reflect intense cracking and interaction of both gas and condensates with anhydrites (H_2S content) in the reservoir (Connan and Lacrampe-Couloume, 1993).

The top seal of the deep Lacq gas field is provided by the "laterolog shales" and by upper Barremian anhydrites whereas over-pressured Albo-Aptian shaly limestones form the lateral seal of this accumulation (Encl. 6d).

The reservoir of the Upper Lacq oil field is formed by Late Cretaceous carbonates (Encls. 1c and 6d). Oil to source-rock correlations show that sourcing is from the uppermost Barremian "Calcaires à Annelides". Migration of oil from this deeper level was associated with the fracking of the lower Aptian shales during the Tertiary.

Lagrave oil field

The Lagrave field is located along the eastern margin of the Early Cretaceous Arzacq Basin on the Late Cretaceous platform in the Pyrenean foreland (Encl. 1b). The Lagrave field is structurally trapped and produces from early Senonian limestones of the Jouansalles Formation. The good petrophysical characteristics of this reservoir are due to secondary dolomitization of the Jouansalles limestones along the Seron Fault zone. The trap of this field developed in response to sinistral strike slip movements along the inherited Early Cretaceous Seron Fault zone during the Tertiary phases of the Pyrenean Orogeny. The top seal for the field is formed by late Senonian shales of the Pé-Marie Formation whereas lateral seals are provided by the syntectonic Ypresian flysch (Encl. 4d).

Oil to source-rock correlations show that hydrocarbon charge was again provided by Barremian and Kimmeridgian shaly limestones.

An analysis of the area shows that the original closure at the level of the Senonian reservoir of the Lagrave structure was formed by compaction-drape over a deep seated fault block, involving the Jurassic carbonates. Hydrocarbon charge to this early structure occurred during the Paleocene. During the Oligo-Miocene this structure was compressionally modified; at the same time fracturing of the Early Cretaceous seals permitted migration of light oil from the Jurassic source-rocks and reservoirs into the Late Cretaceous reservoirs.

VIc Bilh oil field

This field is located in the distal parts of the Pyrenean foreland, along the northeastern margin of the Arzacq Basin, which was affected by Early Cretaceous salt tectonics and limited inversion during the Tertiary Pyrenean phases. The trap of this field is formed by an erosional pinch-out of Jurassic carbonates along the northern salt ridges of the Arzacq Basin (Encls. 1b and 4d). The main halokinetic episode is again pre-Barremian in age, as attested by the transgression of the Barremian and early Aptian sediments over deeply eroded Jurassic carbonates and Triassic and Liassic evaporites at the top of the salt dome (Encl. 4d). The reservoir comprises the Portlandian Mano dolomites and Barremian limestones. The top seal is formed by the early Aptian shales of the Sainte-Suzanne Formation; Albo-Aptian shaly limestones are thought to provide lateral seals.

The oil contained in the Vic Bilh structure was derived from Kimmeridgian and Barremian source-rocks which probably reached maturity during the Oligo-Miocene at the same time as the trap was closed.

CONCLUSIONS

The geology of the northern Pyrenean fold-and-thrust belt is very complex; its structural style is primarily controlled by inherited trends which were repeatedly reactivated during younger tectonic pulses. Structures which developed during the Jurassic and Early Cretaceous extensional phases are generally faulted blocks which were modified to various degrees by salt tectonics. The reconstruction of early formed, Early Cretaceous structures and an understanding of their regional palaeogeographic setting are considered to be essential steps before proceeding further. During the Late Cretaceous to Oligocene Pyrenean Orogeny, pre-existing structures were reactivated to various degrees and at different periods. These polyphase deformations are responsible for the great variety of prospective structures in the Aquitaine Basin.

A similar diversity characterizes the petroleum systems of the Aquitaine Basin. The most prolific source-rock is the Kimmeridgian Lons Formation; however, a significant contribution to hydrocarbons generated and accumulated comes also from Barremian shales. The timing of generation, expulsion and migration phases ranges from Early Cretaceous to Late Tertiary with a progressive evolution closely linked to the local tectonic evolution of each area. To date roughly $330 \cdot 10^6$ tons ($2.4 \cdot 10^9$ bbl) of oil and oil equivalent have been proved up in around fifteen fields.

In such a context the challenge for the petroleum explorationist lies in the evaluation of the remaining potential of very complex and unexplored zones to the South of the northern Pyrenean front. Modern 3D seismic data, combined with detailed surface and subsurface geological studies, proved to be essential for the development of a more comprehensive understanding of this complex area and the definition of prospects. Moreover, a validation of geological models and petroleum systems in the foreland is a key to exploration targeting thrusted zones.

The explored parts of the Pyrenean fold-and-thrust belt and its foreland have been proven to be very prolific. The comprehensive studies, summarized in this paper, show that all key parameters necessary for hydrocarbon entrapment and preservation are present also in the unexplored domains of the thrust belt. This gives reasons for an optimistic outlook for future exploration potentials.

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Enclosures

Encl. 1

a	General structural map and regional cross-section through the Pyrenean Mountain chain
b	Aquitaine Basin, general structural map
c	Aquitaine Basin, stratigraphic chart and Petroleum Systems

Encl. 2

a	South Aquitaine Basin, structural framework and Petroleum Provinces
b	Regional cross-section 1
c	Regional cross-section 2
d	regional cross-section 3

Encl. 3

a	General palaeogeographic map of the Aquitaine Basin at the end of the early Kimmeridgian
b	subcrop map at the base of the Cretaceous showing palaeogeography of the Portlandian as was as the erosion due to salt tectonics anolng the edges of the Arzacq Basin
c	Worsm's eye view at the base of the Cretaceous unconformity
d	Map showing the distribution of the upper Cretaceous formations above the base Upper Cretaceous unconformity

Encl. 4

a	Oil to source-rock correlations in the Aquitaine Basin
b	Gas to source-rock correlations in the Aquitaine Basin
c)	General cross-section through the Arzacq Basin showing timing of generation and migration of hydrocarbons in the area as well as the isomaturation levels
d	Aquitaine Basin: traps associated to oil and gas fields in the fold-and-thrust belt and foreland area

Encl. 5

a	2D seismic line through the Arzacq Basin, time migration
b	2D seismic section through the Rousse and Meillon fields

Encl. 6

a	South Aquitaine, 3D seismic surveys
b	3D Meillon survey, Rousse and Meillon gas fields
c	Dip structural cross-section through the Rousse and Meillon gas fields
d	Dip cross-section through the Upper Lacq oil field and the giant Deep Lacq gas field.

Cenozoic inversion structures in the foreland of the Pyrenees and Alps

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ABSTRACT

Southeastern France forms the foreland of the Late Cretaceous to Paleogene Pyrenean and the Cenozoic Alpine orogens. Unlike other thrust belts, neither the eastern parts of the Pyrenees in Languedoc and Provence nor the Western Alps are associated with major flexural foreland basins. Instead the area is characterised by a complex array of multi-directional compressional foreland structures.

These foreland structures developed in response to Pyrenean and Alpine compressional and transpressional reactivation of pre-existing tensional and transcurrent crustal fault systems which delimit extensional sedimentary basins ranging in age from Permian to Oligo-Miocene. These basins developed during the Permo-Carboniferous collapse of the Hercynian orogen, the Mesozoic Tethyan rifting phase and the Eo-Miocene development of the West-European rift system. Crustal shortening achieved during the inversion of these basins is locally rooted in the basement and possibly translated via a lower crustal detachment level into the respective orogens.

Due to the intensity of these foreland deformations, and partly also due to their superimposition, the Pyrenean and Alpine thrust fronts are poorly defined and in large areas diffuse. Although remnants of a Paleogene flexural foreland basin are locally preserved, this basin, which had presumably limited dimensions, was largely destroyed in conjunction with the inversion of Permian, Mesozoic and Cenozoic tensional structures. In addition, the passive margin sedimentary prism of the Gulf of Lions, which opened in Early Miocene times, conceals part of the Pyrenean fold belt.

In order to gain a better understanding of the evolution and deep architecture of some of the observed foreland structures, surface and sub-surface geological data were integrated and compared with the results of sand-box analogue models that were monitored by X-ray tomography. The kinematics of basin inversion are discussed in an evolutionary framework that is characterised by repeatedly changing stress regimes.

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1 INTRODUCTION

Fold-and-thrust-belt foreland systems are usually characterised by a wide flexural basin (Beaumont, 1981), which develops in the autochthonous foreland in front of a composite tectonic wedge made up of various allochthonous thin-skinned and basement-involved units. Simultaneously, the surface topography of the wedge itself tends to reach an equilibrium, characterised by a specific taper (Dahlen et al., 1984). This is not the case in Provence and Languedoc, as the Pyrenean and Alpine thrust fronts, well expressed in the morphology of the Pyrenees to the West or in the northern part of the Western Alps, become progressively cryptic in the basin of southeastern France (BSEF).

In fact, due to strong Hercynian and Tethyan tectonic inheritance, numerous pre-existing extensional structures were reactivated and inverted during Paleogene and Neogene compressional phases, the orientation of reactivated structures being at any time directly controlled by the direction of the prevailing horizontal compressional stress axis. The occurrence of basement-controlled inversion structures accounts for the abnormally high elevation of isolated structures in the foreland, far from the thrust front.

Inversion tectonics were recognised for a long time in the Pyrenean and Alpine orogens (Lemoine et al., 1981; Mugnier et al., 1987; Gillcrist et al., 1987; Graciansky et al., 1988; Gratier et al., 1989); however, their study was mainly restricted to the allochthon, where the initial relationships between the basement and its sedimentary cover are unfortunately no longer preserved. Because of the great variability in the timing and trends of both extensional and compressional episodes, Languedoc and Provence constitute a famous tectonic province, studied since the early days of the structural geology (Lutaud, 1935, 1957; Goguel, 1947, 1963; Aubouin and Mennessier, 1960; Ellenberger, 1967; Mattauer, 1968; Lemoine, 1972; Aubouin, 1974). It is indeed a key area for the study of foreland inversions, in which very contrasted boundary conditions resulted in quite distinct tectonic features.

This paper aims at documenting the role of inversion in the development of some classical structures of Provence and Languedoc. Analogue models of tectonic inversions will also be confronted with regional surface and subsurface data to provide additional constraints when connecting shallow geometries with the poorly imaged basement architecture.

2 REGIONAL GEOLOGICAL BACKGROUND

Detailed regional syntheses dealing with the geology of the basin of southeastern France (BSEF) were published by Baudrimont and Dubois (1977), Debrand-Passard et al. (1984) and Roure et al. (1992, 1994). Therefore, only major tectono-sedimentary episodes will be discussed in the following.

2.1 Late Hercynian Collapse and Permo-Carboniferous Basins

Recent deep seismic profiling across the Biscay and Aquitaine domains have imaged south-verging Hercynian thrusts and overlying post-nappe Permian basins (Fig. 3a; Choukroune et al., 1990). Simultaneously, new field work, petrographic and microtectonic studies on the fabric and transport direction (kinematics) of Late Paleozoic sedimentary or metamorphic units, forming the southern part of the Massif Central (i.e. Montagne Noire; Burg et al., 1990; Echter and Malavieille, 1990; Van den Driessche and Brun, 1991), have greatly improved our knowledge of the West European Hercynian tectonic edifice. Following the last compressional episodes during the Carboniferous, the Late Hercynian deformations reflect the general collapse of the pre-existing orogen, as evident from two distinct types of structures:

(1) high angle intra-crustal strike-slip faults, trending north or northeast, such as the Sillon Houiller and Cévennes Fault (Fig. 1), are identified in the French Massif Central and in Languedoc (Arthaud and Mattauer, 1969),
(2) eastwards trending normal faults, which controlled the development of Late Carboniferous and Permian extensional basins (Ste. Afrique, Rodez or Lodève in Languedoc; Figs. 1 and 8a; Santouil, 1980; and Le Luc in Provence; Toutin-Morin and Bonijoly, 1992; Figs. 1 and 9a).

Although these Late Paleozoic basins are well exposed along the western (Massif Central) and eastern (crystalline Provence and Belledonne Massif in the Alps) borders of the BSEF, their subsurface extent in Languedoc and Provence remains highly conjectural. Nevertheless, exploration wells have revealed extensive Permian sequences north of the Pic-St. Loup structure in Languedoc (Roure et al., 1988; Figs. 1 and 8a), and Paleozoic sequences occuring beneath the Arc syncline, south of the Ste. Victoire structure in Provence (Tempier, 1987; Biberon, 1988; Figs. 1 and 9a).

Seismic profiles along the western border of the BSEF, have also imaged extensive Carboniferous strata. Recently cored during scientific drillings (GPF program; Giot et al., 1991; LeStrat et al., 1994), these Carboniferous strata, especially the coal measures, constitute potential detachment levels which were activated at least locally during Jurassic and Oligocene extensional phases (Roure et al. 1992, 1994; Bonijoly et al., 1995; Fig. 2).

Finally, exploration wells drilled in the Jura Mountains of France and Switzerland also identified Carboniferous strata in a subthrust position beneath the folded Mesozoic series (Laubscher, 1986; Noack, 1989; Mascle et al., 1994; Philippe, 1994). Reflection seismic profiles on the eastern border of the Jura also revealed inverted Permo-Carboniferous basin (Gorin et al. 1993, Signer and Gorin, 1995).

2.2 Tethyan Rifting and Intra-Mesozoic Detachment Levels

After local evidence of Early Triassic extension, the main Tethyan rifting episode occurred in Late Triassic and Liassic times (Lemoine and Trümpy, 1985; Graciansky et al., 1988; Elmi, 1990; Bergerat and Martin, 1993, 1995). This is demonstrated by the strong subsidence of the BSEF at this time, and by rapid Liassic facies and thickness changes, which were controlled mainly by the activity of major northeast-trending normal faults (Elmi et al., 1991; Giot et al., 1991). Recent reflection seismic profiles confirmed that only few of the Jurassic faults involves the infra-Triassic basement (i.e. the Cévennes, Nîmes and Durance faults, Fig. 1), whereas most of the other structures are detached in Triassic evaporites (Petit et al., 1973; Roure et al., 1988, 1992).

Block faulting ceased during the Middle Jurassic, as attested by the regional Dogger unconformity, which marks the onset of the post-rift thermal subsidence of the BSEF (Figs. 2 and 12).

However, due to the contrasted paleobathymetries, platform conditions prevailed until the Lower Cretaceous (i.e. Aptian Urgonian platform) in Languedoc and Provence in the south, as well as in the Vercors and the Jura Mountains in the north. On the contrary, basinal conditions and deposition of thick, ductile black-shales (the Liassic to Oxfordian "Terres Noires") occurred in the intervening Vocontian Trough (Fig. 13a), where effectively only the Late Jurassic sequence is made up of brittle carbonates (Tithonian deep-water carbonates and breccias; Fig. 2).

Potential décollement horizons occur, apart from Triassic evaporites, in the Jurassic and/or Lower Cretaceous shales. Nevertheless, due to rapid lateral facies and thickness variations, these potential detachments horizons are rather discontinuous, especially at basin-to-platform transition zones. As a result, the most complex structural Cenozoic configurations are found in these areas (Figs. 2 and 13).

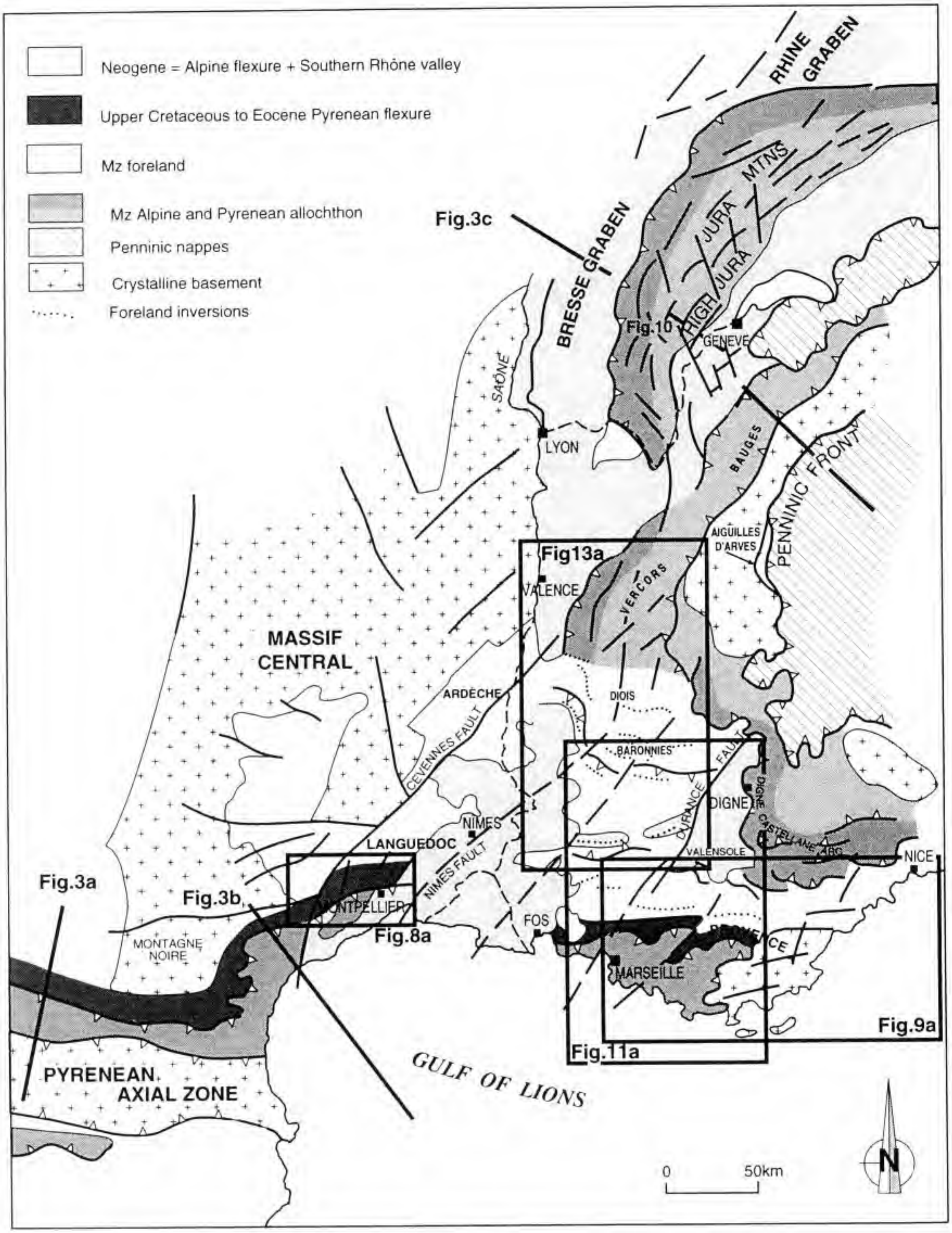

FIG. 1. Structural framework of the basin of southeastern France outlining the Pyrenean and Alpine frontal thrusts.

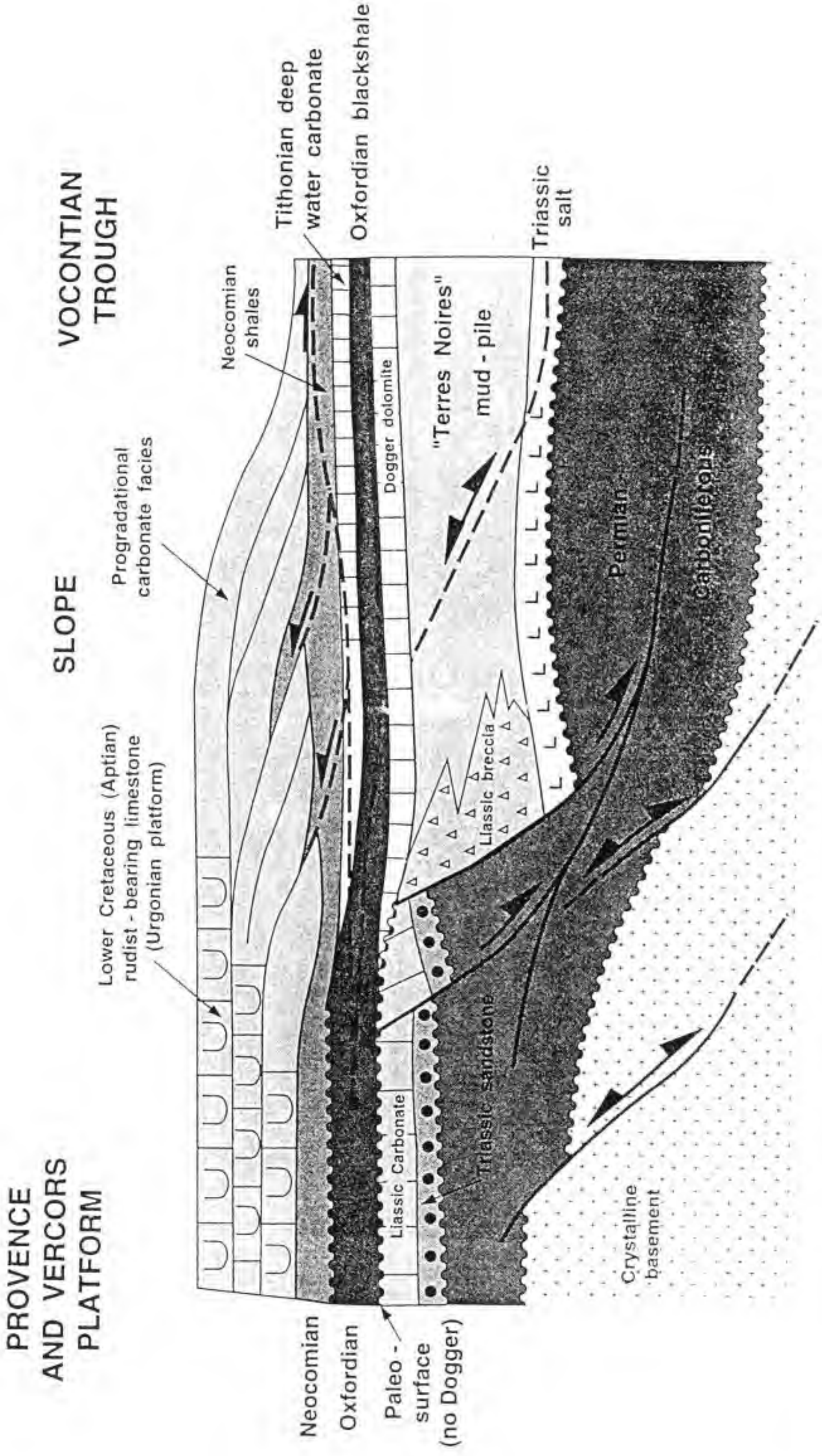

FIG. 2. Lithostratigraphic diagram showing lateral sedimentological variations in the location of potential detachment horizons.

the La Nerte-Ste. Baume structure as representing the easternmost part of the Pyrenean frontal overthrust. Accordingly, all the surface structures identified further north (Alpilles, Luberon, Ste. Victoire, La Lance and Ventoux-Lure east-trending structures; Fig. 11a) are interpreted as basement-controlled inversion features which are connected with the Pyrenean thrust front.

Further east and northwards, i.e. in the Annot area, in the Aiguilles d'Arves or Bauges massifs (Fig. 1), the Alpine allochthon itself also provides evidences for Paleogene deformations and the development of an early flexural basin. The Priabonian succession, which disconformably rest on Cretaceous carbonates, is now entirely incorporated into the Alpine tectonic edifice and comprises shallow-water nummulite-bearing carbonates that grade upwards into pelagic marls and deep-water turbidites. As such, they outline the rapid subsidence of the European foreland in front of advancing nappes (Vially, 1994).

2.3.2 Oligocene Extension and Stress Permutations

Widespread extension occurred in western Europe during Oligocene times, leading to the development of the West European Rift system, extending from the Netherlands and Germany (Rhine Graben) to the Mediterranean (Bresse Graben, Limagnes, Rhône Valley and Camargue).

In the south, rifting progressed to crustal separation, thus inducing the opening of the oceanic Gulf of Lions and consequently, the building up of a classical passive margin during the Neogene, off the Languedoc and Provence coasts of the West Mediterranean Basin (Burrus, 1989; Gorini et al., 1993). During the extensional process, numerous Pyrenean thrusts were reactivated (negative inversion), especially south of St. Chinian (Fig. 3b; Gorini et al., 1991; Roure et al., 1994).

Because of important syn-rift and post-rift subsidence of the area, the former Pyrenean structures are no longer preserved at the surface near the Rhone Delta, thus impeding a direct structural correlation between Provence and Languedoc (Fig. 1).

The paleostress regime remained relatively consistent during the Oligocene, with an overall N110 trend for the minimum principal horizontal stress trajectories (Bergerat, 1985; Villéger and Andrieux, 1987). Detailed microtectonic analyses outline a progressive permutation of the principal stress axes during the transition from the Eocene to the Oligocene (Pyrenean compression with a N0 to N10-trending sigma1 and Oligocene extension, with a N110-trending sigma3). Similarly, a progressive change of extensional stress axes from N110 to N155 is evidenced during the Oligocene in the Marseille Basin (Hippolyte et al., 1993), prior to the onset of the Neogene Alpine deformations.

During Oligocene times, major northeast-trending Late Hercynian basement structures were reactivated (Cévennes, Nîmes and Durance faults; Figs. 1 and 11a). Moreover, the Triassic detachment level was also reactivated as indicated by numerous Oligocene normal faults which are restricted to the Mesozoic sediments and thus have a listric shape (Fig. 1; Roure et al., 1988, 1992); the Alès basin also developed in a piggyback position above the intra-Triassic décollement.

2.3.3 Alpine Compression and its Sedimentary Record in Southeastern France

Alpine compressions occurred during the Neogene, with a progressive change in the orientation of sigma1 (from N20 to N90 or N120; Bergerat, 1985; Villéger and Andrieux, 1987).

The Alpine thrust front is well defined in the Digne and Castellane arcs (Figs.1 and 9a). West of Digne, it is bordered by the autochthonous flexural Miocene molasses of the Valensole Plateau (Dubois and Curnelle, 1978). Similarly, from the Vercors to the Jura Mountains, the Alpine thrust front can be accurately traced and is also bordered by marine sediments deposited in a Miocene flexural basin (Figs. 1 and 3c; Mugnier and Viallon, 1984; Gratier et al., 1989; Guellec et al., 1990a, 1990b). However, in the intervening area, i.e. between the Drôme River to Digne (i.e. in the Vocontian Trough; Figs. 1 and 13a; Goguel, 1963; Gratier et al., 1989), neither the Alpine thrust front

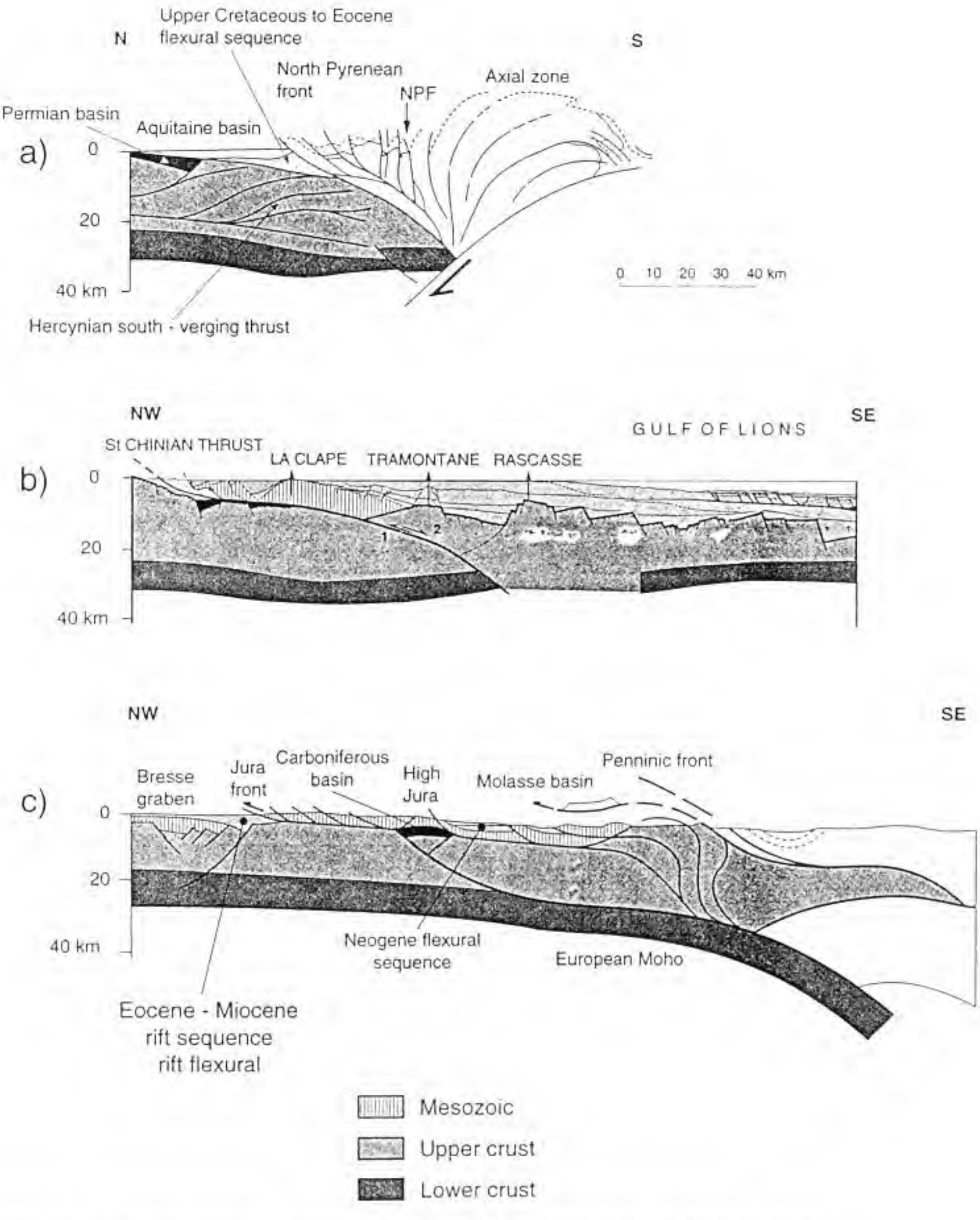

FIG. 3. Crustal sections outlining the contrasted architecture of the European foreland between Alps and Pyrenees (loaction on Fig.1):
a) section across the Aquitaine foreland.
b) section across Languedoc and Gulf of Lions (modified from Vially, in prep.)
c) section across the Alps and Jura Mountains.

2.3 Kinematics and Timing of the Pyrenean and Alpine Deformations

Recent deep reflection seismic profiles imaged the entire crust of the European foreland beneath the North Pyrenean (Choukroune et al., 1989) and Western Alpine (Guellec et al., 1990a, 1990b) thrust fronts, outlining the gentle flexure of the underthrusted lithosphere, and the development of a subaerial accretionary wedge and adjacent flexural basin (Figs. 3 a, c).

Because of its multistage tectonic history, with interferences between Pyrenean or Alpine compressional structures and Oligocene rifting, the BSEF hardly compares with typical foreland basins. Even the Pyrenean and Alpine thrust fronts become diffused in this area, and special attention is required to trace them with some confidence.

2.3.1 Pyrenean Compression and its Sedimentary Record in Languedoc and Provence

Although Late Cretaceous compressional deformations induced in Languedoc and Provence local erosion and deposition of tectonic breccia at the leading edge of early structures (i.e. Ste. Victoire; Lutaud, 1935, 1957; Tempier and Durand, 1981), most of the Pyrenean shortening occurred in Eocene times. However, microtectonic evidences in continental conglomerates in Languedoc indicate that compression persisted during the Lower Oligocene in Languedoc.

The Eocene stress field was remarkably stable in the entire European foreland and was characterising a north-trending maximum horizontal compressional stress trajectory (Bergerat, 1985).

West of the Rhône River, the North Pyrenean thrust front is still relatively easy to trace in Languedoc. Late Cretaceous to Eocene continental sequences progressively onlap the basement of the Massif Central north of the Pyrenean thrust front (Ellenberger, 1967; Figs. 3b and 4). Younging progressively northwards, this flexural sequence is partly underthrust beneath or even accreted to the Pyrenean allochthon (Figs. 3b and 4; Roure et al., 1988).

Eastwards, the Pyrenean thrust front continues into the Montpellier overthrust, which, based on well results, was transported a minimum of 8 km to the north (Andrieux and Mattauer, 1971). Moreover, there is local evidence for Eocene transpressional reactivation of such major northwest trending Late Hercynian structures as the Cévennes Fault in Languedoc (Arthaud and Mattauer, 1969), and the Durance Fault in Provence.

In Provence, however, the conditions are more complex. Despite the widespread occurrence of Paleogene north- and south-verging compressional structures, many authors proposed a complete decoupling of the Mesozoic sedimentary cover from its basement, and very large amounts of shortening (Guieu and Rousset, 1980; Tempier, 1987; Biberon, 1988; Deville et al., 1994). However, as discussed in the following, most structures, even some of those verging towards the north, are better interpreted as basement-controlled inversion features; this concept precludes a general detachment of the Mesozoic sediments from their basement.

In fact, only two north-verging structures can be traced over large distances and have a regional lateral extent (Fig. 9a):

(1) in the south, the Cap Sicié basement overthrust relates to a shallow-dipping detachment that is related to the Pyrenean allochthon,
(2) immediately to the north, La Nerte and Ste. Baume thrusts are two contiguous thin-skinned structures, which only involve Mesozoic sediments, and are thrust over the Late Cretaceous to Eocene continental deposits of the Arc syncline (Aubouin and Chorowicz, 1967; Carrio-Schaffhauser and Gaviglio, 1985). Both, the age of the sedimentary infill and the structural position of the confined Arc basin, compare adequately with the coeval flexural sequence of the limited Pyrenean foreland basin, as already identified in Languedoc (Figs. 1 and 3a).

Because no other continuous north-verging structure occurs farther to the north, we consider

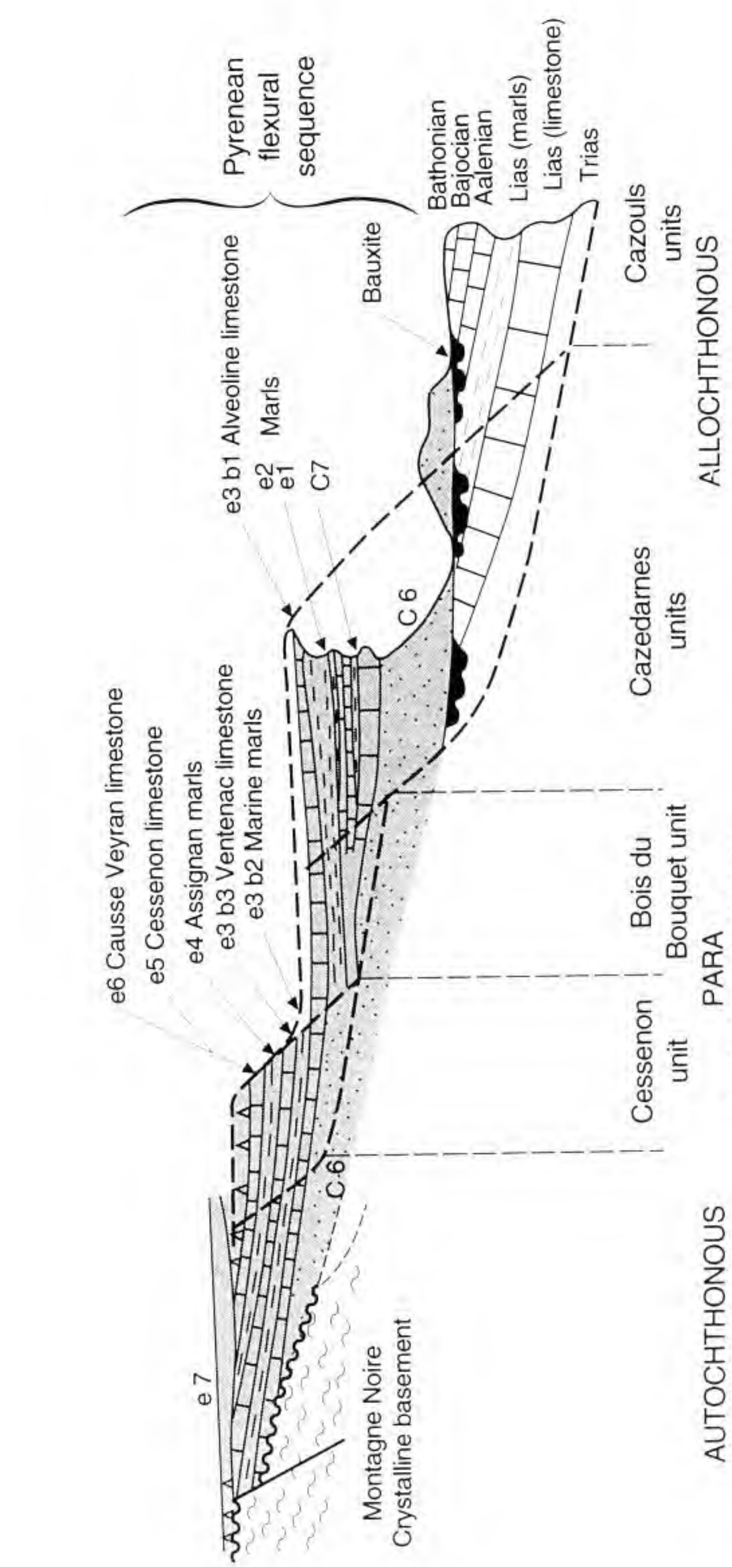

FIG. 4. Schematic diagram outlining the progressive onlaps of the Late Cretaceous to Eocene Pyrenean flexural sequence near St. Chinian.

nor the Miocene flexural basin can readily be identified. This is due to the inversion of the Vocontian Trough, containing a thick Jurassic basinal sequence, during the Pyrenean and Alpine episodes of crustal shortening; the resulting structural relict, that undoubtedly was already initiated during the Late Cretaceous to Eocene Pyrenean orogeny, precluded the development of a Miocene flexural depocentre in this part of the European foreland.

Miocene marine deposits occurring in the Valreas Basin west of the basement-controlled Vocontian inversion structure (Figs. 1 and 13a), are underlain by a stable structural domain, corresponding to the Urgonian carbonate platform, which was unaffected by the foreland inversion movements. There, Miocene molasse deposits can be interpreted either as representing the outermost onlapping sequence of the Neogene foredeep, or preferentially as having accumulated in a post-rift thermal subsidence basin, that is related to Oligocene extensional structures; the latter accounts for the geophysically defined crustal thinning evident in this portion of the foreland (Ménard, 1979; Hirn et al., 1980).

3 SCALED-DOWN MODELS OF BASIN INVERSION

Sand-box experiments are frequently used to study the incidence of various parameters during basin inversion processes (McClay, 1989; Buchanan and McClay, 1991). However, recently developed computerised X-ray tomographic conservative techniques permit, at any time of the ongoing experiment, a better imaging of incremental deformations and the documentation of the spatial architecture of the model in any direction, without having to destroy it (Colletta et al., 1991).

This technique has been applied at the IFP to study the boundary conditions of structural inversion. Thus, during a systematic set of experiments, either the basal friction, the number and the location of ductile interbedded layers, the attitude of the pre-existing basement faults and the orientation of the maximum compressional stress axis, were modified (Sassi et al., 1993). Some of these experiments, especially those dealing with oblique inversion, were successful in modelling the evolution of the Saharan Atlas (Vially et al., 1994).

Below we present some of the results of such model experiments which can be compared with the regional inversion structures occurring in the BSEF, and provide additional constraints to support new hypotheses on the deep architecture of these structures, particularly concerning the basement-cover relationship (Roure et al., 1992, 1994).

3.1 Extensional Listric Growth-fault Structure and its Subsequent Inversion

A first experiment attempted to simulate the structural inversion of a basin controlled by a listric growth fault (Fig. 5; see Roure et al., 1992 for details of the apparatus). The hanging-wall was maintained rigid during the compressional deformation of this model, whereas a Mohr-Coulomb behaviour was assumed for the foot-wall. The geometry of the listric fault, as determined by the shape of the rigid hanging-wall, was therefore fixed during the entire experiment. A total decoupling between hanging-wall and foot-wall was made possible.

During the initial extensional process, an asymmetrical half graben and a conventional rollover structure, with a crestal collapse graben and numerous normal faults, was generated in the foot-wall (Fig. 5). Simultaneously, the topography of the model was preserved horizontal, by progressively filling in the developing depressions with sediments.

In a second stage, incremental shortening was applied to the model in an effort to restore the mobile foot-wall to its initial preextensional position. The resulting geometry indicates that normal faults in the roll-over crestal collapse graben are not reactivated, but act as the roots for newly developing reverse faults during the first stage of shortening.

As discussed below, this mode of deformation can be compared with regional examples along the Durance Fault, which effectively outlines syn-extensional Oligocene growth strata above an

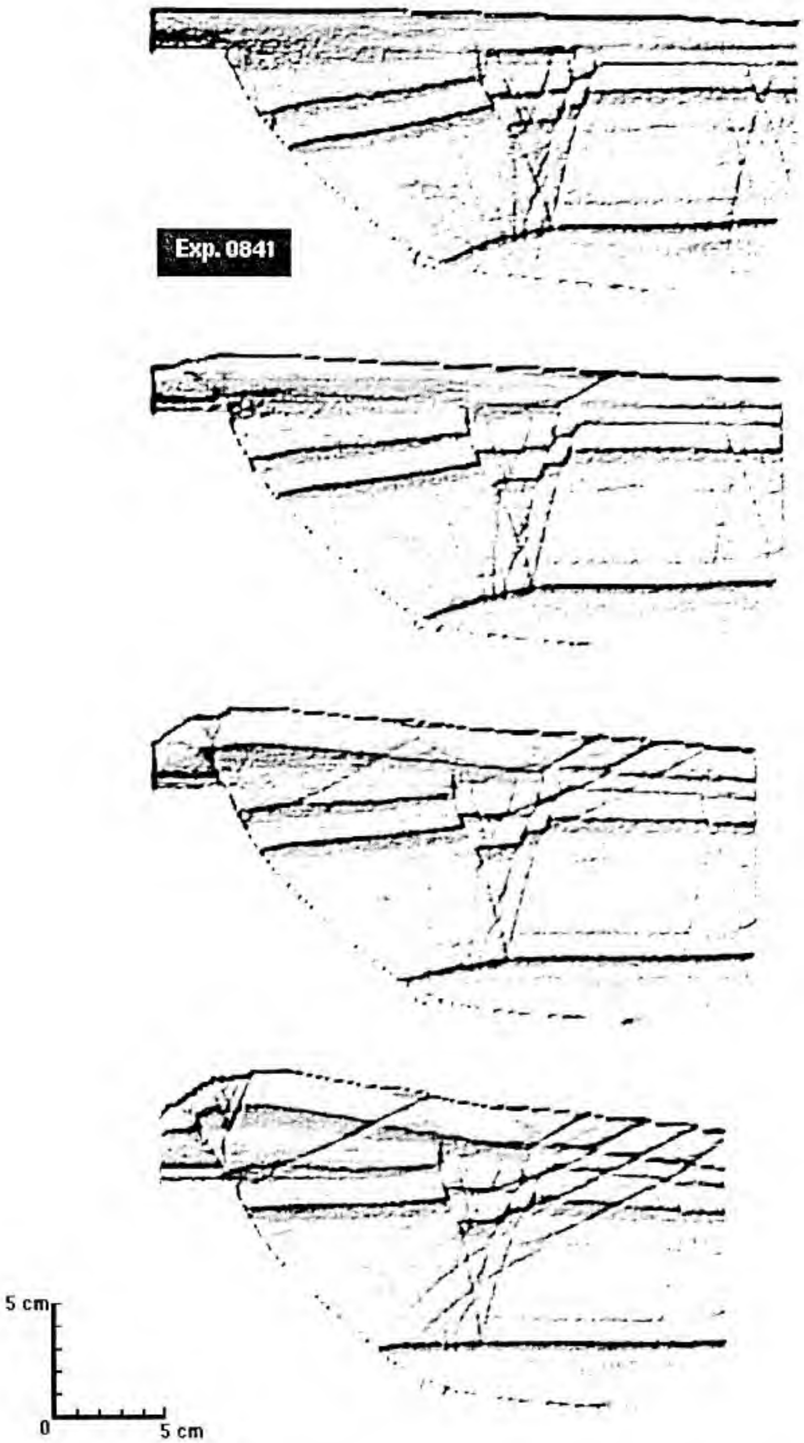

FIG. 5. X-ray images of sand-box experiment simulating the tectonic inversion of a hangingwall block controlled by a normal listric growth fault. Note the nucleation of reverse faults at the tip of normal faults bounding the crestal collapse graben.

intra-Triassic detachment that developed along the trend of a high-angle basement fault (Fig. 11b). Unfortunately, the present apparatus does not permit a coeval reactivation of the basement high-angle fault in the hanging-wall. Nevertheless, as shown by the results presented, additional vertical motion of the basement would mainly result in an overall increase of the footwall culmination, and would have only a minor effect on the internal geometry of the foot-wall block, due to accommodation of these motions along the listric detachment fault.

3.2 Multilayer Sand-cake, Basement Short-cut and Wedging

Another set of experiments aimed at evaluating basin inversion in a multi-layer sand-cake with a pre-existing high-angle basement fault (see Roure et al., 1994; Vially et al., 1994 for a precise description of the apparatus). For this study, a rigid basement was decoupled along a mobile high-angle fault plane, and sediments were deposited horizontally on both the hanging-wall and foot-wall sides of the model. A vertical offset was preserved between alternating brittle and ductile materials in the sedimentary cover of the basement in an attempt to simulate the presence of a pre-existing normal fault in the sediments. Only the shallowest brittle horizon was deposited over the entire model, across the now inactive normal fault, thus simulating the post-rift sequence (Figs. 6 and 7).

In a first set of experiments (Fig. 6), orthogonal compression was applied to the entire foot-wall, permitting simultaneous incremental vertical motions along the basement fault. As a result, the sedimentary infill of the basin was gradually inverted, with deformations being guided by a forced motion along the rigid hanging-wall basement block.

However, as predicted by mechanical studies on fault reactivation (Jaeger and Cook, 1969; Sibson, 1985), the free upper portion of the pre-existing normal fault is not reactivated in the sedimentary cover during this process of orthogonal inversion, but is rather passively transported above a newly created low-angle fault which splays upwards from the rigid basement, thus outlining a typical short-cut geometry (Huygue and Mugnier, 1992; Fig. 6). At the same time, a triangle zone (fish-tail) develops, with the deep short-cut block progressively wedging out the brittle cover of the hangingwall, inducing the development of a backthrust.

By slightly modifying the boundary conditions in the shallow horizons (i.e. the lateral extent or the location at depth of the ductile horizons), it is possible to generate non-cylindrical structures, characterised by shallow thrusts which accommodate main transport out of the basin (same dip attitude as the deep high-angle basement fault) or into the basin (shallow conjugate backthrusts). Double vergence pop-up structures may indeed occur in the transition zone between these two asymmetric domains when they develop along trend of the same pre-existing basement fault (Fig. 7).

In a second set of experiments, oblique compression was applied to the same initial geometry. In these models, horizontal sections display predominantly en-echelons structures in the sedimentary layers, with crestal collapse fractures developing in inversion-related anticlines, obliquely to the trend of the pre-existing basement fault. Under transpressional conditions, reactivation of the shallow and free portion of the pre-existing high-angle normal fault becomes possible, with no systematic development of newly created low-angle faults and short-cuts.

4 LATE HERCYNIAN INHERITANCE AND CENOZOIC INVERSIONS

In the following, we compare structures observed in the BSEF with results of the above discussed analogue experiments. In most cases, surface and subsurface data provide sufficient control on the geometry of structures at the level of the Mesozoic sediments. However, interpretations at the basement level (undeformed or involved), are often more conjectural, particularly where surface

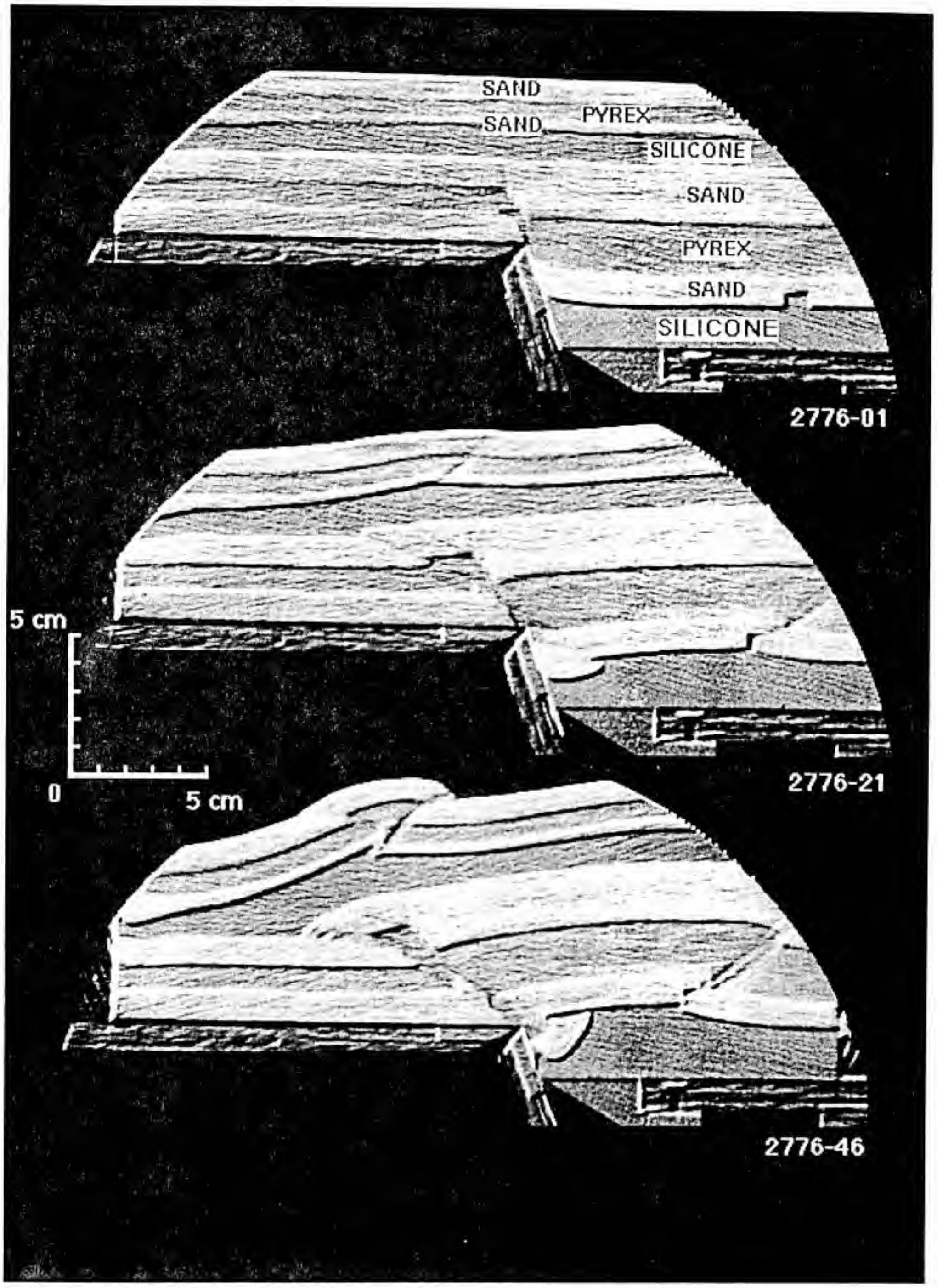

FIG. 6. Incremental deformation of a hangingwall block controlled by a planar basement fault, in a sand-box experiment simulating the tectonic inversion of a multilayer sand-silicone model. Note the short-cut in the basement, and wedging and backthrusting in the sedimentary cover.

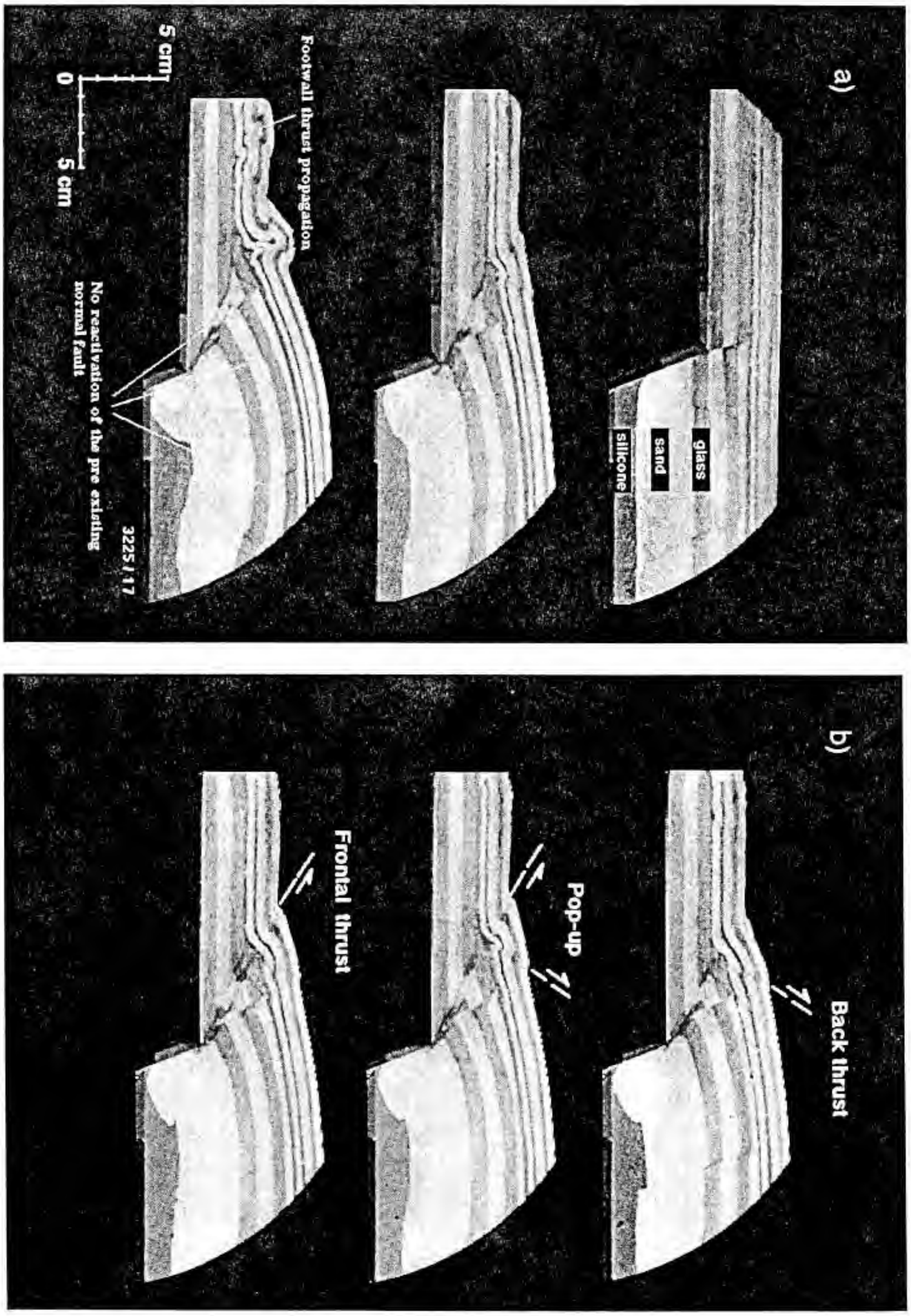

FIG. 7. a) Incremental deformation in a sand-box experiment simulating the tectonic inversion of a multilayer sand-silicone model, with reactivation of planar normal basement faults.
b) Serial cross sections of the deformed model showing the lateral changes from forward-directed thrusting to back-thrusting in the uppermost sedimentary cover, with the occurrence of a pop- up structure in the intermediate domain.

complexities prevented good reflection seismic resolution at depth.

As the major point of discussion concerns the relative allochthony or autochthony of the surface structures, we first tried to restore and balance the sections assuming a minimum amount of horizontal shortening (in-situ deformation above pre-existing basement structure). The subsequent comparisons with the models provided further insight into the coherency of the structural reconstruction, and brought a new approach to identify pre-existing basement structures and buried inverted Paleozoic basins at depth.

4.1 Pyrenean Inversion of Permian Basins

In the European foreland of Languedoc and Provence, several structures strike easterly; however, not all of them are of Pyrenean origin nor are they related to pre-existing Late Paleozoic structures. Some of them display Miocene (Alpine) deformation or are related to the reactivation of Jurassic (Tethyan) structures. This is the case for instance for the Alpilles, Ventoux-Lure and the Luberon massifs in Provence (Figs. 1, 9a and 13a).

Therefore, we have selected here only two structures, one in Languedoc (Pic-St. Loup), and the other one in Provence (Ste. Victoire), for which Pyrenean deformation is attested, and Late Hercynian inheritance can be documented with some confidence.

4.1.1 The Pic-St. Loup Triangle Structure in Languedoc

The Pic-St. Loup structure (Fig. 8) is an east-trending, north-verging asymmetric anticline, which is slightly thrusted over the syntectonic Eocene and Early Oligocene conglomerates of the St. Martin de Londres Basin (Figs. 1 and 8). In the west, the Pic-St.Loup structure is bounded by the northeast-trending Cévennes fault, a pre-existing Late Hercynian basement wrench fault, which was reactivated by left-lateral strike-slip motion during the Eocene deformation, synchronously with the development of the Pic-St. Loup structure (Arthaud and Mattauer, 1969). Eastwards however, the Pic-St. Loup overthrust is limited by the Corconne Fault, a younger, northeast-trending Oligocene normal fault, which is detached in the Triassic evaporites (Roure et al., 1988).

Regional geological studies and recently acquired reflection seismic data provide control on the shallow architecture of this structure. Only slight facies variations are recorded in the surrounding Upper Jurassic sequence, with shallower-water deposits being found south and west of the Pic-St. Loup than to its north and east (Dreyfus and Gottis, 1949). Similarly, bauxites developed in the south, whereas active subsidence still characterised the St. Martin de Londres Basin during the Cretaceous. Liassic platform carbonates outcrop in the core of the structure and were also penetrated by the Pic-St. Loup exploration well, thus implying a local basal detachment of the structure along Triassic evaporites.

Nearby exploration wells identified Permian sediments at depth to the north of the Pic-St. Loup; although these can be traced over a distance along seismic profiles (Roure et al., 1988), their presence is not confirmed south of the Pic-St. Loup. On a regional scale, the Permian strata presumably represent an eastwards extension of the Lodève Basin, which crops out west of the Cévennes Fault and is bordered into the south by the major north-dipping Montagne Noire high-angle basement fault (Figs. 1 and 8; Santouil, 1980). The eastwards extent of this basement fault cannot be confirmed on seismic lines, which are of poor quality beneath the steeply dipping Jurassic limestones of the Pic-St. Loup. Remote sensing studies, however, help to trace it beneath the Mesozoic sedimentary cover. On satellite images it appears as an east-trending lineament which extends beneath the Pic-St. Loup and crosses the Corconne Fault (Figs. 1 and 8; Chorowicz et al., 1991). The lateral offset of this Permian fault across the Cévennes Fault is the best argument for the Pyrenean reactivation of the latter.

Keeping these constraints in mind, a palinspastic restoration of a structural cross section across the Pic- St. Loup/St. Martin de Londres syncline was attempted. Among the various possible solutions (see Roure et al., 1994 for a discussion), the simplest one refers to an in-situ balance

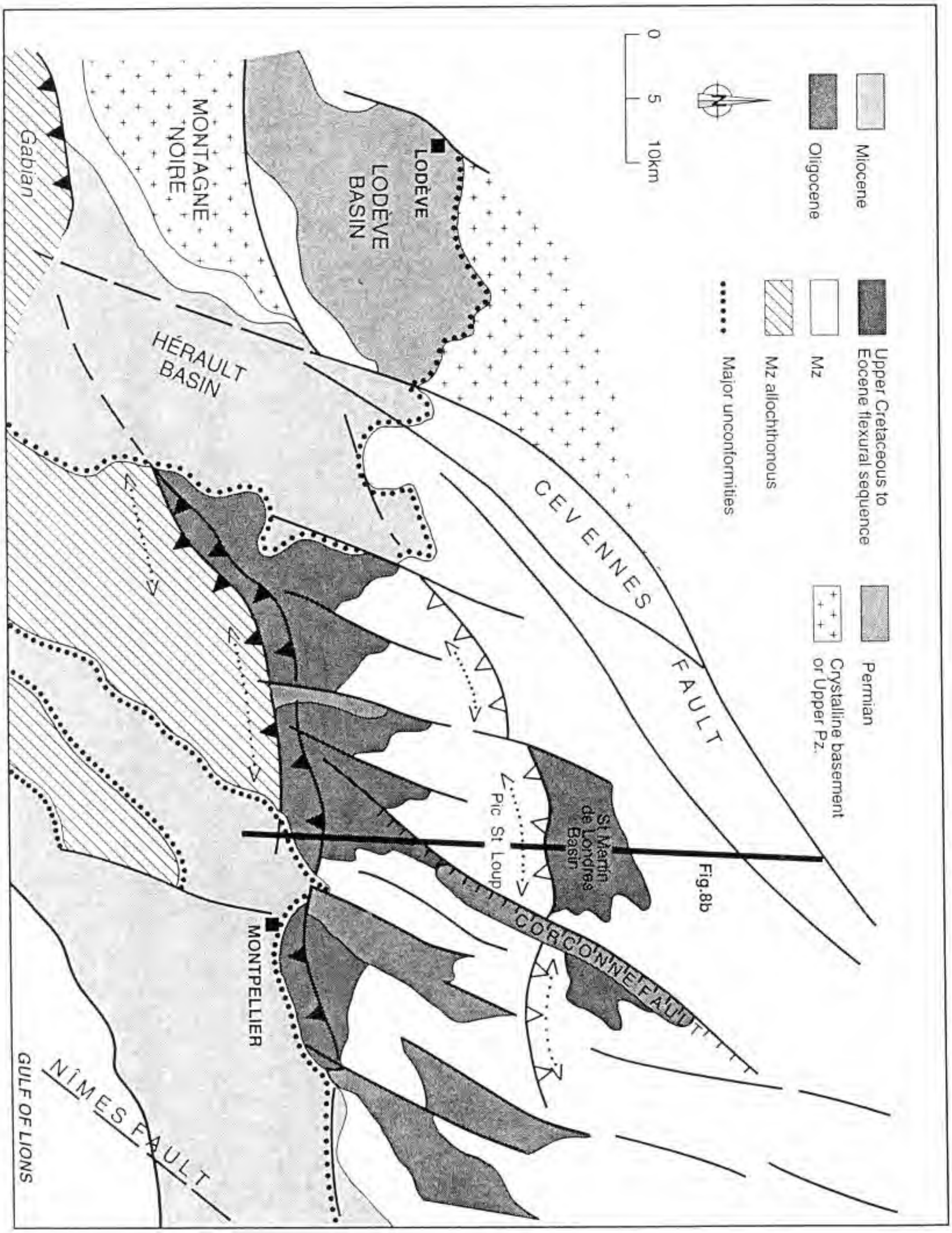

FIG. 8a. Structural map of the Pic-St. Loup / St. Martin de Londres area

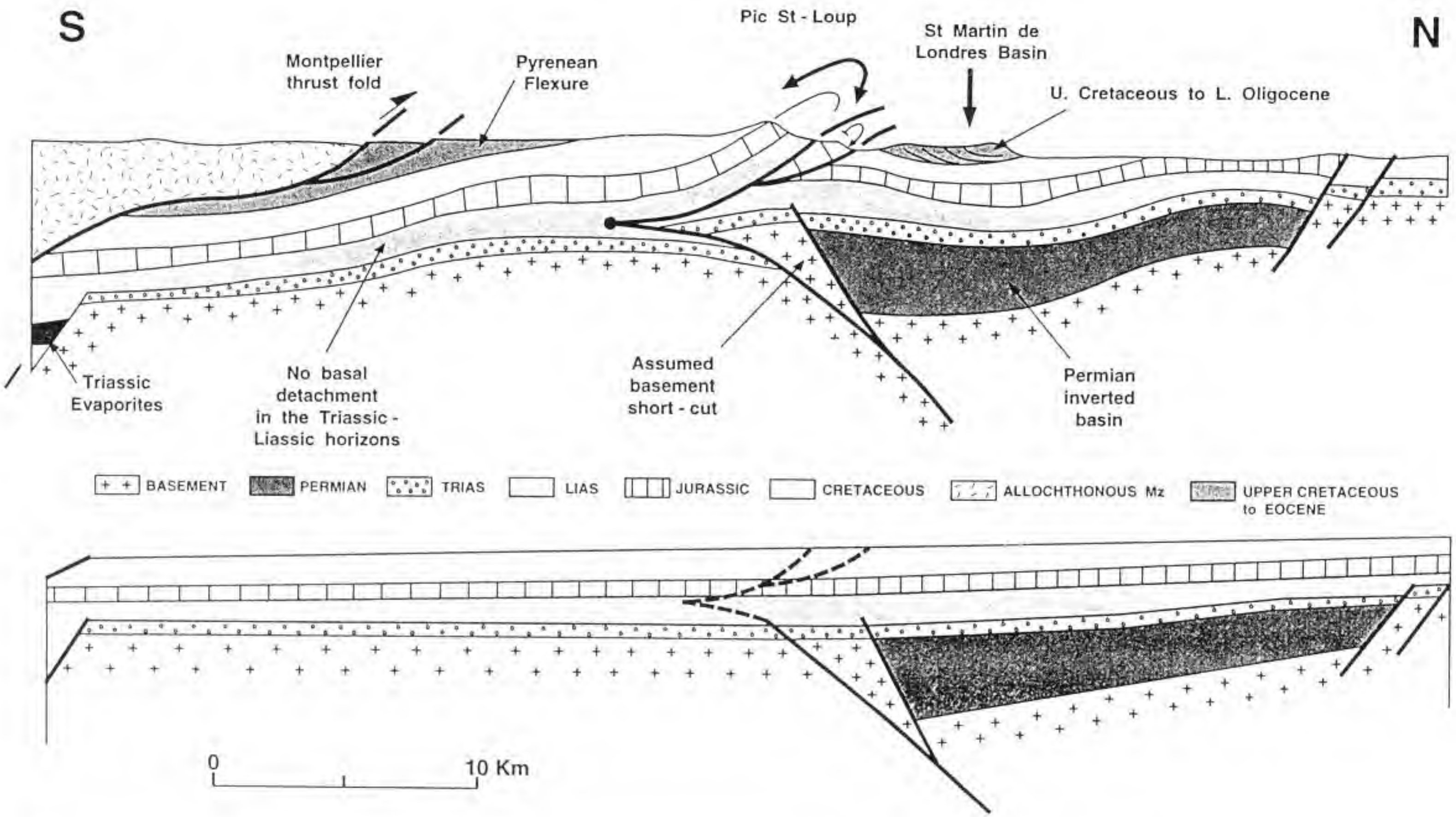

FIG. 8b. Structural and palinspastic sections of the Pic St. Loup structure in Languedoc. Note north-verging back-thrust in Mesozoic platform sequence over adjacent basin, due to the deeper inversion along a north-dipping basement border fault of a thick Permian basin.

between surficial shortening and basement reactivation (Fig. 8). As discussed earlier, the Pyrenean thrust front can be located with good confidence at the front of the Montpellier thrust, and there is no compelling evidence to postulate a continuous detachment of the sedimentary cover between Montpellier and St. Loup structures. Moreover, surface observations along the north-dipping Montagne Noire-Lodève Basin Permian border fault confirm that this structure was reactivated during the Pyrenean orogeny (Santouil, 1980).

A better understanding is obtained of the Pic-St.Loup structure, which is characterised by backthrusting in the sedimentary cover and basement inversion of the substratum, when compared with the analogue models (Fig. 6). In this case, the brittle platform domain in the south is progressively wedged out during the partial inversion of the Permian basin. The occurrence of shallow Triassic ductile layers at the base of the rigid Mesozoic carbonates permits their decoupling from deeper levels and the development of a conventional triangle zone. However, the poor quality of the reflection seismic data beneath the Pic-St. Loup structure does not provide any direct evidence for a low-angle short-cut in the basement, which would have to be assumed according to the sand-box models, if the maximum horizontal compressional stress trajectory was at a high-angle to the trend of the pre-existing normal fault during the inversion process (Figs. 6 and 7).

East of the Corconne Fault, a careful study of the architecture of surface structures also attests for a reactivation of the Permian structure. However, unlike in the Pic-St. Loup area, Mesozoic sediments are here involved in a south-verging overthrust (Fig. 8), which fits with the geometry of the pre-existing basement fault; this is a more common attitude for inverted structures.

4.1.2 The Sainte Victoire Pop-up Structure in Provence

The Ste. Victoire Massif (Fig. 9), an east-trending and about 1000 m high structure, is probably the most famous geological feature in Provence; however, its development is still the subject of controversy (Durand and Tempier, 1962; Corroy et al., 1964; Chorowicz and Ruiz, 1979; Durand and Gieu, 1980; Biberon, 1988). At the surface, the Ste. Victoire structure corresponds to a large anticline, cored by Jurassic platform limestones. In the west, it is thrusted to the south, and overlies thick Late Cretaceous synkinematic proximal breccias, as well as the Late Cretaceous to Eocene synflexural continental sediments of the Arc syncline (Figs. 1 and 9). In the east, the Ste. Victoire Massif gradually acquires a north-verging attitude; its central portions constitute a pop-up structure.

Geophysical data confirm a southwards deepening of the basement, from a 3 km depth north of the Ste. Victoire Massif, down to 4 km beneath the Arc syncline (Biberon, 1988); however, reflection seismic do not allow to determine whether the basement is faulted or only gently flexed beneath the Ste. Victoire anticline. Triassic and Jurassic sequences display no significant lateral thickness change and thus, preclude Mesozoic faulting. On the other hand, more than 1 km thick Permian strata occur along the eastern margin of the Arc syncline, whereas to the north of the Ste. Victoire structure, Triassic strata rest directly on basement; this is taken as indirect evidence for Permian faulting beneath the Ste. Victoire structure.

Within the predominantly brittle Permian and Mesozoic clastic and carbonate strata, Triassic evaporites and Upper Jurassic (Oxfordian) black-shales provide potential décollement levels. In view of the omnipresence of Triassic evaporites, many authors consider the Ste. Victoire structure and even northwards adjacent folds and thrusts as thin-skinned features forming part of the Pyrenean orogenic front (Tempier, 1987; Biberon, 1988).

Alternatively, and keeping the results of sand-box simulations of basement-controlled inversions in mind, most the Ste. Victoire structure can be balanced in situ, assuming it is superimposed on a south-dipping Permian high-angle fault controlling the distribution of Late Paleozoic strata beneath the Arc syncline.

Therefore, it is proposed that structural inversion of the Arc Permian basin induced wedging in the Mesozoic sequence, involving its detachment along intra-Triassic and Oxfordian ductile levels as evident by rapid lateral changes in attitude of surficial thrusts. Scale-down models of tectonic inversion in multilayer sand-cakes illustrate the development of similar pop-up structures in transition zones, with mass transport evolving from out of the basin (forward) to into the basin (backthrusting) (Figs. 6 and 7).

In the case of the Ste. Victoire structure, the amount and location of erosion during and after an initial episode of gentle deformation (i.e. Late Albian-Early Cenomanian, or Maastrichtian to Montian; Durand and Tempier, 1962) probably modified the geometry of the hanging-wall, inducing lateral changes in the boundary conditions, which localised the change from backthrusting to frontal deformation during the Eocene episode of main inversion (Lutaud, 1935, 1957). Also in this case, the postulated occurrence of a low-angle basement short-cut beneath the Ste. Victoire structure remains hypothetical for lack of subsurface control.

In our interpretation, most, if not all, of the shallow deformation observed in the Ste. Victoire structure can be balanced in situ by an equal amount of basement shortening (Fig. 9b). However, a minor northwards translation of the entire Arc syncline above an intra- Triassic detachment cannot be precluded. In this second hypothesis, the role of the basement fault would be to localise the Pyrenean deformation of the detached Mesozoic sediments.

4.2 Alpine Inversion of Late Paleozoic Basins (Subthrust High Jura Architecture)

Also in the Jura Mountains, the incidence of pre-existing Paleozoic features on Cenozoic compressional structures is evidenced, where boreholes and reflection seismic profiles have identified Late Paleozoic basins, deeply buried beneath the allochthonous Mesozoic cover (Laubscher, 1986; Noack, 1989).

The ECORS deep seismic profile and recent petroleum exploration data permit a correlation between the occurrence of basement-controlled inversions at depth, and the abnormally high topographic elevation of the inner parts of the Jura Mountains (i.e. the Grand Cret d'Eau, 1600m high; Guellec et al., 1990b; Roure et al., 1990, 1994; Philippe, 1994), as compared with the low relief of the Molasse Basin to the Southeast (average 500 m), and of the Bresse Basin in the northwest (average 250 m) (Figs. 1, 3 and 10).

Figure 10 gives a structural cross section through the High Jura Mountains, that is based on surface and subsurface data. A recently drilled well (Charmont) demonstrates the occurrence of thick Permo-Carboniferous sediments beneath the deformed Mesozoic strata of the Jura Mountains, which are detached from their substratum at the level of the Triassic evaporites. In the direction of mean transport, the detachment plane rises from the Molasse Basin into the High Jura Mountains and plunges again at their outer margin.

Backthrusting occurs in the Mesozoic series at the western border of this anomalous subthrust configuration (Oyonnax backthrust), and can be attributed to deformation of the Paleozoic substrate and basement. The observed geometric relationships between the Mesozoic allochthon and its substrate suggests inversion of the Permo-Carboniferous basin occurred after the westwards translation of the Mesozoic series, resulting in deformation of the intra-Triassic detachment horizon. Eventually, the west-verging overthrust of the Grand Cret d'Eau structure, along the southeastern margin of the Jura Mountains, can be interpreted as an out-of-sequence reactivated thrust, that formed during the inversion of the Permo-Carboniferous trough. It is proposed that the recent basement-involving shortening, causing deformations of the allochthon, should be locally balanced, rather than involving reactivation of the entire Jura allochthonous and mass-transport in the opposite directions, i.e. to the west near Oyonnax, and to the south-east beneath the Molasse Basin.

Again, sand-box models of basin inversion (Figs. 6 and 7) provide a rational for proposing coherent scenarios for the development of such triangle zones, with basement inversion at depth, and synchronous conjugate backthrusting in the sedimentary cover. However, it is proposed that the structure of the High Jura Mountains results from a 2-phase deformation:

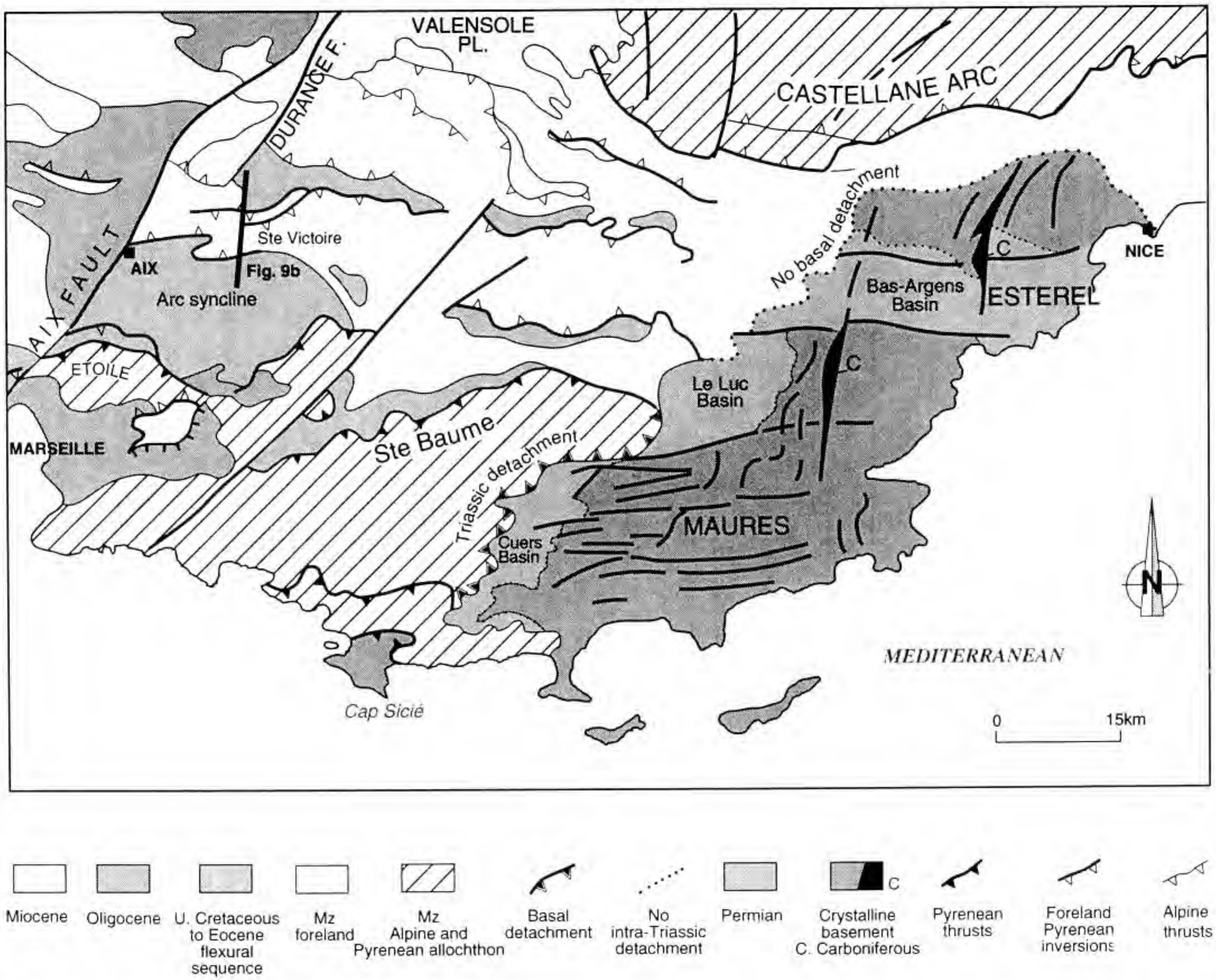

FIG. 9a. Structural map of the Ste. Victoire / Arc syncline area.

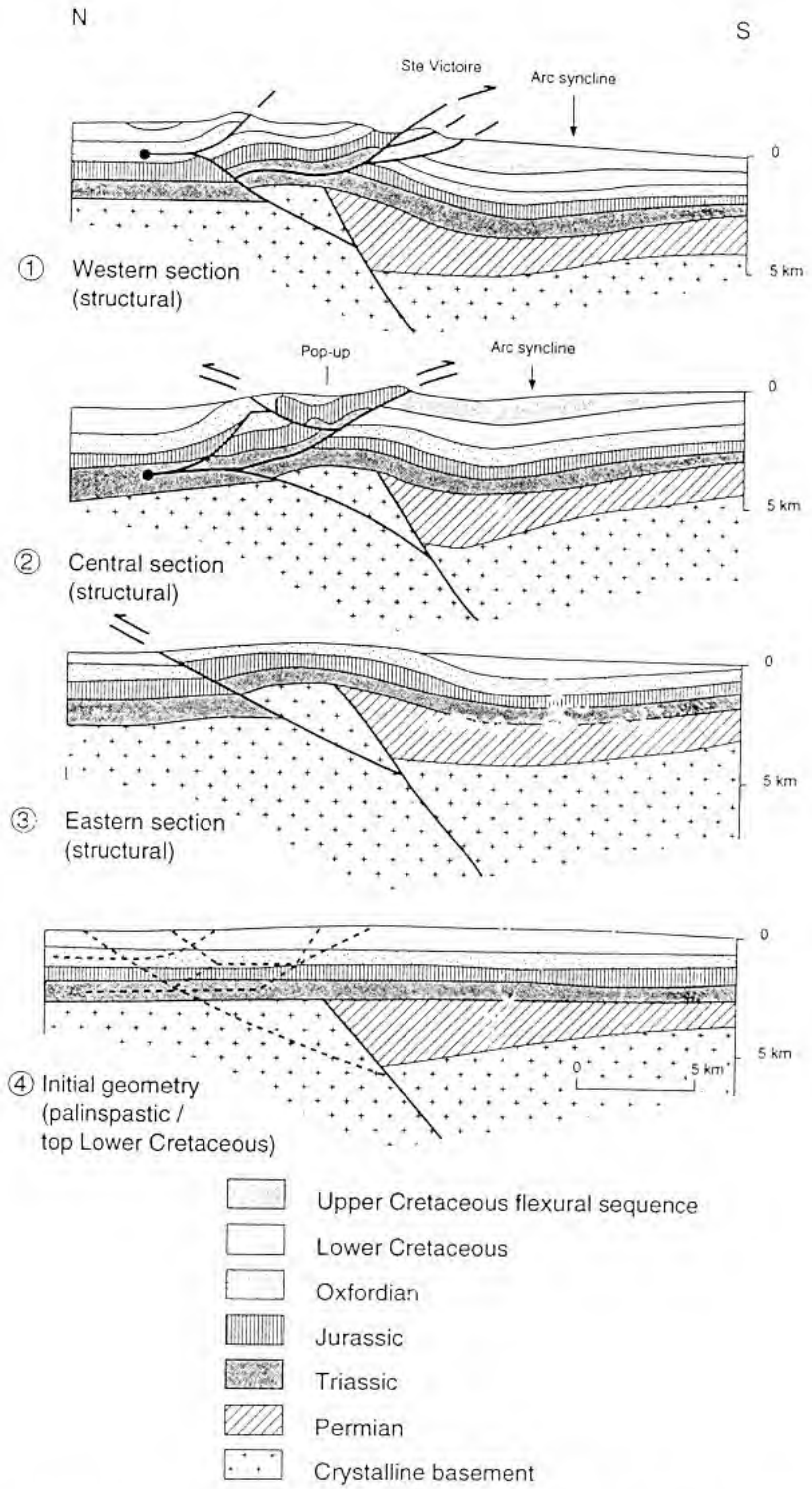

FIG. 9b. Structural and palinspastic serial cross sections of the Ste Victoire structure in Provence.

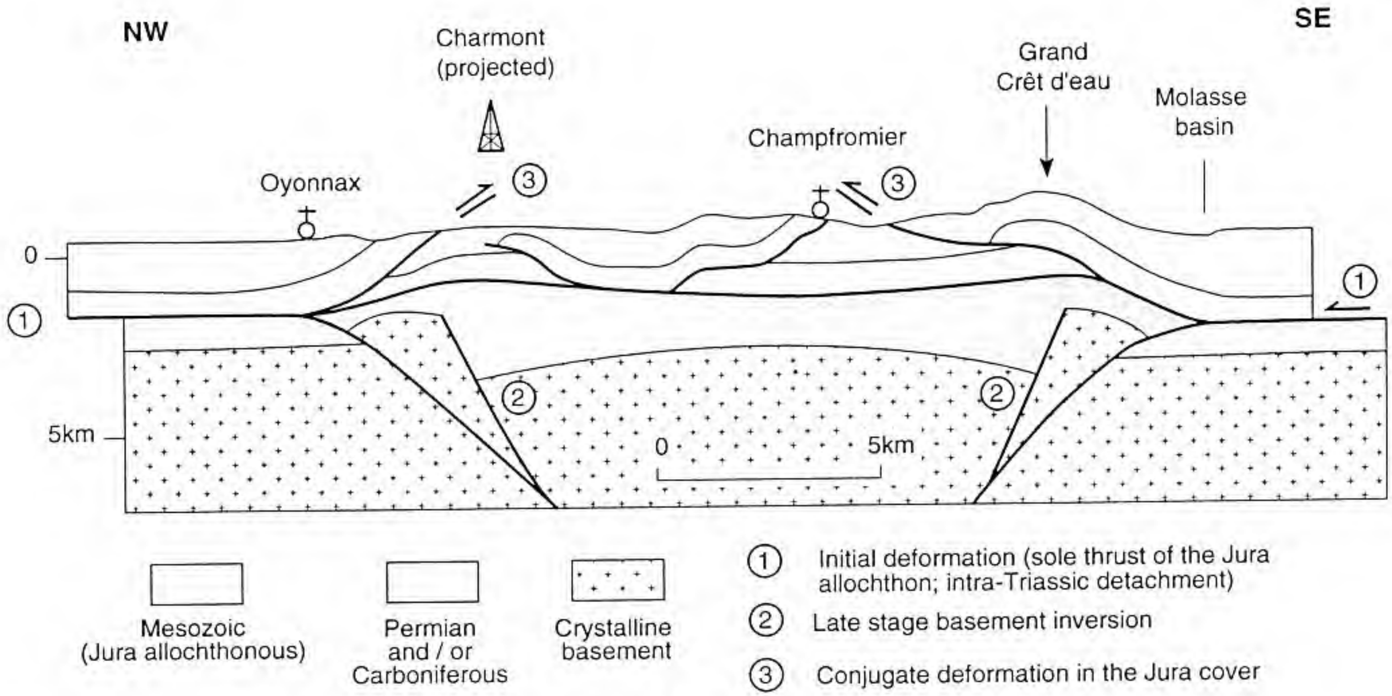

FIG. 10. Subthrust inversion of Late Paleozoic basins beneath the High Jura Mountains (modified from Philippe, 1994).

(1) west-verging detachment folding and thrusting of the Mesozoic sediments during the Miocene deformation phase which stopped during the Pontian (Jura thrust-emplacement over the Bresse Graben),
(2) subthrust basement-involving inversion of Permo-Carboniferous basins, development of the Oyonnax backthrust and reactivation of the Cret d'Eau structure between the Pontian and Present.

5 TETHYAN INHERITANCE AND CENOZOIC INVERSIONS

The role of pre-existing Tethyan rifting on subsequent Alpine deformations is well known in the sub-Alpine or External Crystalline Massifs of the Western Alps (Gillcrist et al., 1987; Mugnier et al., 1987; Graciansky et al., 1988; Gratier et al., 1989; Guellec et al., 1990a, 1990b). However, large amounts of shortening and the frequent complete allochthony of these reactivated structures prevents accurate reconstruction of their initial configuration (i.e. basement-cover relationships). The study of Alpine inversion features in the European foreland, away from the major Alpine thrust front, is facilitated by the quality of the reflection seismic profiles which usually increases towards the autochthon; moreover, reduced amounts of shortening provide better conditions there for studying the geometry of both the basement and the involved sediments.

5.1 Basement Control on Deformation (the Durance Fault)

The Durance Fault is generally regarded as a Late Hercynian structures, due to its northeast strike (Arthaud and Matte, 1975). However, geological and geophysical data permit only to evaluate its Mesozoic and Cenozoic history.

Surface geology, exploration wells and recent reflection seismic surveys provide constraints on regional structural cross sections across the Durance Fault:

(1) east of the fault, the Valensole Plateau is characterised by a reduced Mesozoic platform sequence (Dubois and Curnelle; 1978), which is still attached to its basement due to the lack of Triassic evaporites. In most places, the Cretaceous sequence has been entirely eroded and marine to continental Miocene sediments rest disconformably on Jurassic carbonates (Fig. 11).
(2) a second extensional phase occurred during Late Aptian to Cenomanian times, giving rise to the development of horsts-and-grabens on, the hanging-wall. During the Late Senonian-Eocene Pyrenean the hanging-wall basin was partly inverted and the resulting topography eroded. During Oligocene times the tensional Manosque basin developed. During the Miocene, this basin was incorporated into the Valensole basin. During the Late Miocene to Present, the Manosque basin was inverted in response to Alpine compressional stresses (Fig. 11b).
(3) the Durance Fault clearly involves the basement, but part of its successive vertical or oblique motions were balanced in the foot-wall by lateral escape of the Mesozoic series along a basal intra-Triassic décollement. As a result, the deep architecture of the fault comprises two complementary features: a listric normal fault, flattening in the Triassic and a high-angle fault, rooting in the upper crust.

The Oligocene infill of the Manosque Basin is presently involved in a complex anticlinorium, partially transported towards the east, the culmination of which is located at a higher elevation than the Miocene molasse of the Valensole Plateau in the hanging-wall; this attests for a Neogene, Alpine, inversion episode (Fig. 11b). When restoring the

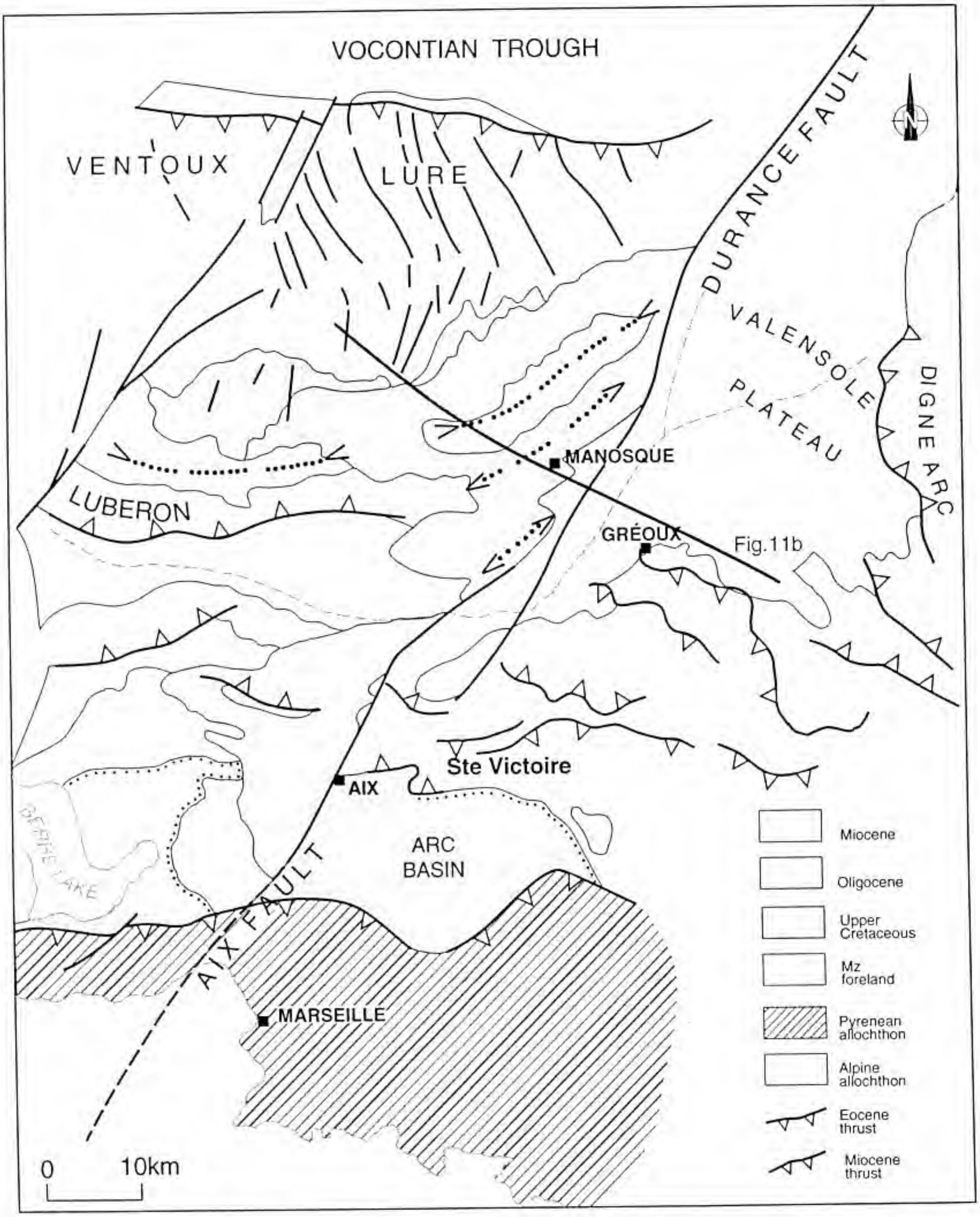

FIG. 11a. Structural map of the Durance Fault / Manosque area

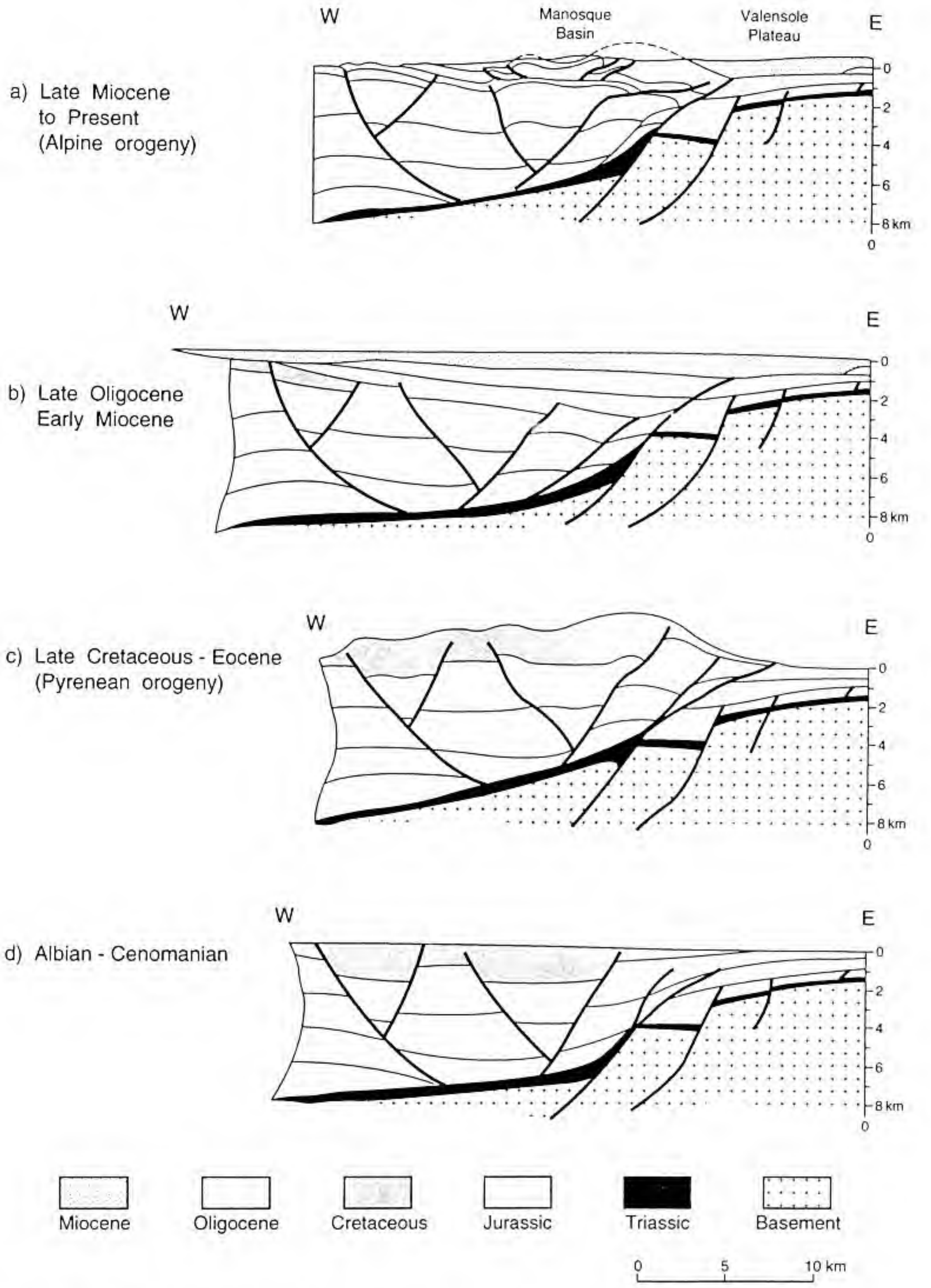

FIG. 11b. Structural and palinspastic sections along the Durance Fault, outlining the Alpine structural inversion of the Manosque Oligocene basin.

section to its pre-Oligocene geometry, the distribution of Albian-Cenomanian series is very peculiar; Lower Cretaceous horizons are only preserved in distal portions of the foot-wall. This is interpreted as evidence of post-Cenomanian erosion, resulting from an earlier episode of tectonic inversion, occurring after the Lower Cretaceous and prior to the Oligocene, presumably during the Late Cretaceous-Eocene Pyrenean orogeny.

By comparing the geological cross-section and its palinspastic restoration with analogue models for compressional/transpressional reactivation of a listric normal fault (Figs. 5 and 11), a better understanding is obtained of the internal deformation of the Oligocene Manosque Basin. Its antiform geometry is controlled by a set of conjugate thrust faults, which are rooted in an intra-Oligocene salt horizon (Fig. 11). When restored to a pre-Alpine geometry, these faults root in the vicinity of Cretaceous grabens, which developed in the crestal part of a regional roll-over structure. As observed by X-ray tomography during incremental deformations of a sand-cake, the localisation of the shallow thrusts is controlled by the pre-existing high-angle tensional faults of the roll-over structure (Roure et al., 1992; Fig. 5).

5.2 Sedimentary Control of Deformation

Apart from the configuration of the basement, lateral thickness and facies changes of the sedimentary cover can control the localisation of inversion deformations. Two examples are discussed below.

5.2.1 Inversion of the Terres Noires Shale Basin

Figure 12 gives a regional structural cross-section and its palinspastic restoration through the Alpine Vercors thrust front and the Eastwards adjacent inverted "Terres Noires" shales basin (for location see Figs 1 and 13). This is one of the best examples to demonstrate the control of thickness and lithofacies changes on the structural style of a compressionally deformed sedimentary package.

From east to west, the thickness of the deformed Mesozoic sequence decreases from more than 8 km in the Vocontian Trough to less than 4 km beneath the Rhône Valley, in the footwall of the Cévennes Fault. Deformation of this basin involved the activation of a sole thrust which ramps up from an intra-Triassic salt layer through Early Liassic carbonates into the Liassic Terres Noires shales and ultimately through the Middle and Late Jurassic carbonates into basal Cretaceous shales.

At the same time, the intra-Triassic detachment constitutes the sole-thrust during the Alpine basin inversion in the east, whereas the Liassic platform carbonates of the external domain are still preserved in the foot-wall beneath the Vercors overthrust. There, two additional shale horizons in the Jurassic (above the Liassic carbonate) and the Neocomian (beneath the Aptian Urgonian Formation), allow for the detachment of the more rigid, mainly brittle platform areas.

The topographic culmination of the inverted "Terres Noires" basin at Aurel is located above the place where the basal detachment ramps up from an intra-Triassic to an intra-Liassic level, and thus corresponds to a ramp anticline involving the entire shale basin. As imaged in the analogue models, a triangle zone develops at the front of the inverted structure; beneath the Rhône Valley, the brittle Urgonian platform carbonates are progressively detached above the Neocomian shales, and thrusted backwards over the basinal allochthon (Fig. 12).

5.2.2 Inversion at Urgonian platform margin (La Lance and Ventoux structures)

La Lance and Ventoux-Lure overthrusts, which strike northwest and east, respectively, are located along the shale-out edge of the Aptian carbonate shelf (Urgonian Formation) of the Provence platform. These structures are thrusted northwards or northeastwards over the margins of the "Terres Noires" shales basin corresponding to the Vocontian Trough (Villéger and Andrieux, 1987; Ford,

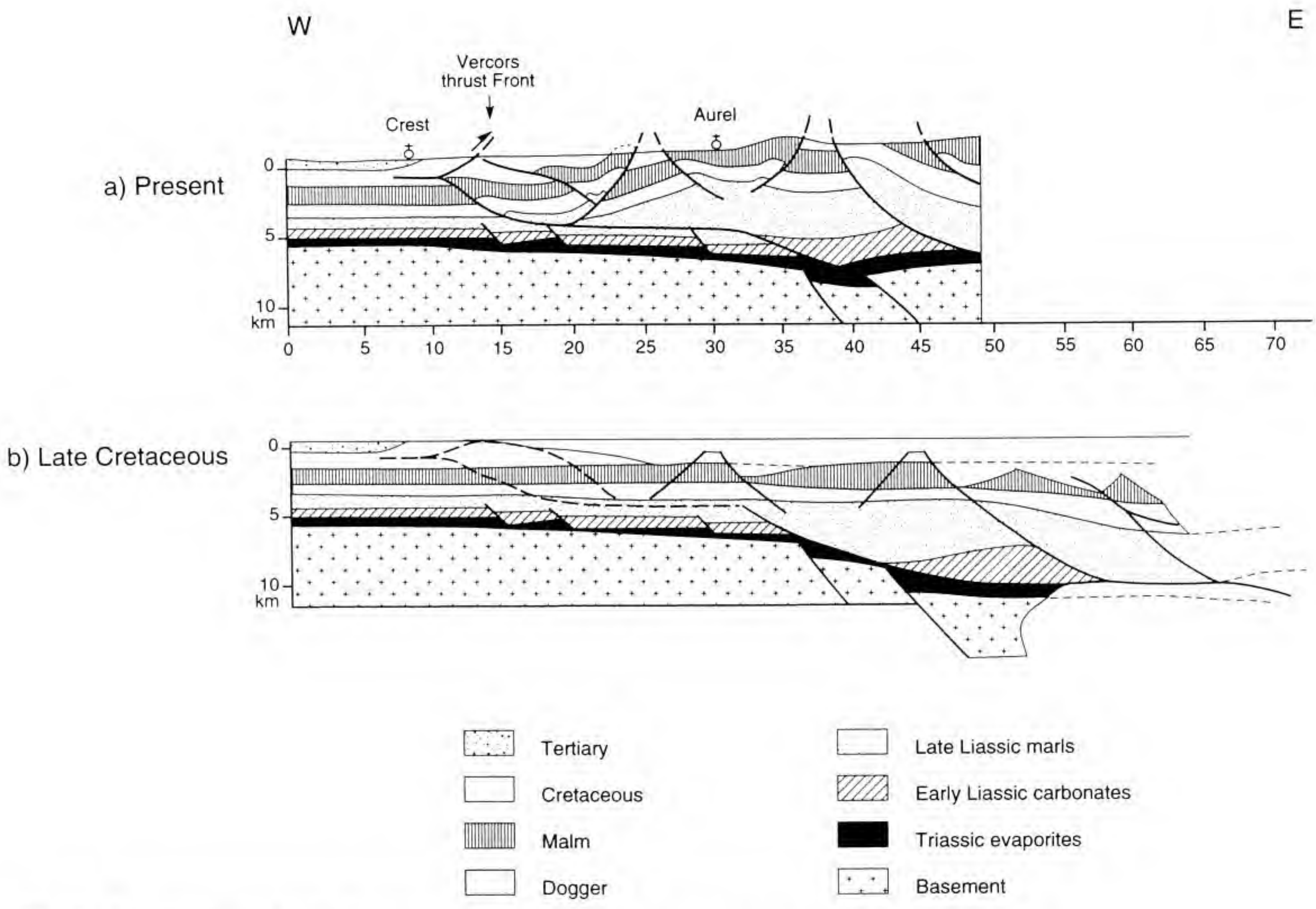

FIG. 12. East-west srtructural and palinspastic sections across the Vercors overthrust parallel to the Drôme River. See location of the section on Fig. 13a.

1993; Ford and Stahel, 1995; Figs. 1 and 13), and form two distinct and isolated culminations near the Rhône Valley; these are located far away from the Alpine thrust front. The highest peak of the Ventoux-Lure trend has an elevation of 1912 m and is referred to as the Giant of the Provence.

Reflection seismic profiles provide control on a major basement involving normal fault beneath La Lance structure, and associated changes in thickness of the Mesozoic sediments (Fig. 13b). The Jurassic sequence is thicker in the north ("Terres Noires"), whereas the Urgonian platform, quite thick beneath the autochthonous Valreas Basin, displays classical northwards progradations (Figs. 2 and 13b). Structurally speaking, the top of the Jurassic sequence culminates in the north in the area of the inverted basin, presently at 1800 m above its position in the stable Valreas Basin platform. This is clearly the result of a post-Cretaceous structural inversion. A minor complicating factor are Triassic evaporite diapirs, which are associated with basement faults, as evident on seismic data near Dieulefit beneath La Lance overthrust (Fig. 13b; Dardeau et al., 1990).

Paleogene and marine Miocene molasse is only preserved in the Valreas Basin, where it rests concordantly on Mesozoic strata. The structural conformity between Miocene and Mesozoic strata on the back-slope of the La Lance structure is the best constraint on its Alpine deformation.

Apart from the basement-involving high-angle fault and the intra-Triassic décollement level observed in the foot-wall of the structure, two additional detachment horizons are evident on seismic records (Figs. 2 and 13b; Roure et al., 1994), namely:

(1) Jurassic shales above the Liassic carbonates in the southern platform domain, in the hanging-wall beneath the Valreas Basin,

(2) and a Neocomian to Aptian shaly sequence in the basinal domain, which constitutes a ductile counterpart (lateral facies transition) for the brittle Urgonian carbonates.

However, similarly to the La Lance structure, it is assumed that the Ventoux Lure and Luberon thrusts are superimposed on deep seated basement faults which were actives during Triassic and Jurassic times (Fig. 13c). Observed thickness changes in the Urgonian platform carbonates are related to differential compaction of the Terres Noires shales over which these carbonates prograded.

As in some analogue experiments, the tectonic inversion of the Vocontian Basin has led to a progressive backthrusting of the platform domain over the thick shale basin. This wedging, and related development of a triangle zone, is indeed mainly controlled by the rheological contrast between the brittle Urgonian carbonate platform and the more plastic shaley basinal sequences; it is, however, greatly facilitated by the occurrence of secondary intra-platform décollement levels.

6 ALPINE INVERSION OF OLIGOCENE STRUCTURES

The latest Miocene episode of oblique inversion recorded along the Durance Fault is the best evidence for Alpine compressional reactivation of Oligocene extensional structures. It resulted in intense folding and thrusting in the Paleogene sequences of the Manosque basin (Fig.11; Roure et al., 1992).

Less obvious Miocene compressional deformations and fault reactivations have been locally described along the Cévennes Fault and in the Oligocene fill of the Alès Basin, west of the Rhône Valley (Fig. 1).

Surprisingly, however, large portions of the so-called West European Rift system were preserved from any direct reactivation during the Neogene, even close to the Alpine front. For instance, the Oligocene and Miocene fill of the Bresse Graben was partly overridden during the Pontian (Mio-Pliocene boundary) by the Jura allochthon; however, no major Oligocene faults appear to have been reactivated at that time within this basin (Mugnier and Viallon, 1984; Bergerat et al., 1990; Guellec et al., 1990a, 1990b; Fig. 3c).

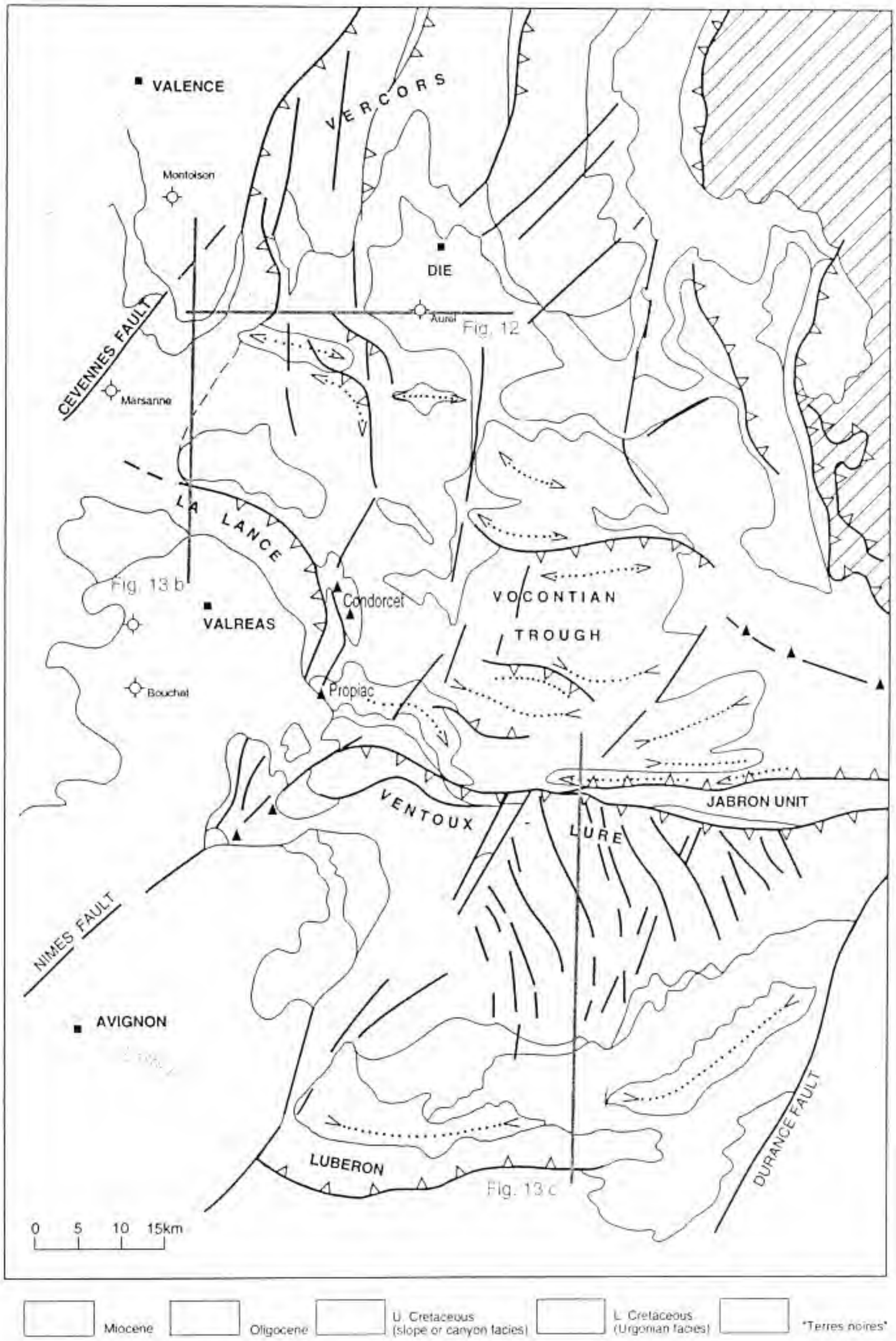

FIG. 13a. Structural map of La Lance / Ventoux-Lure area

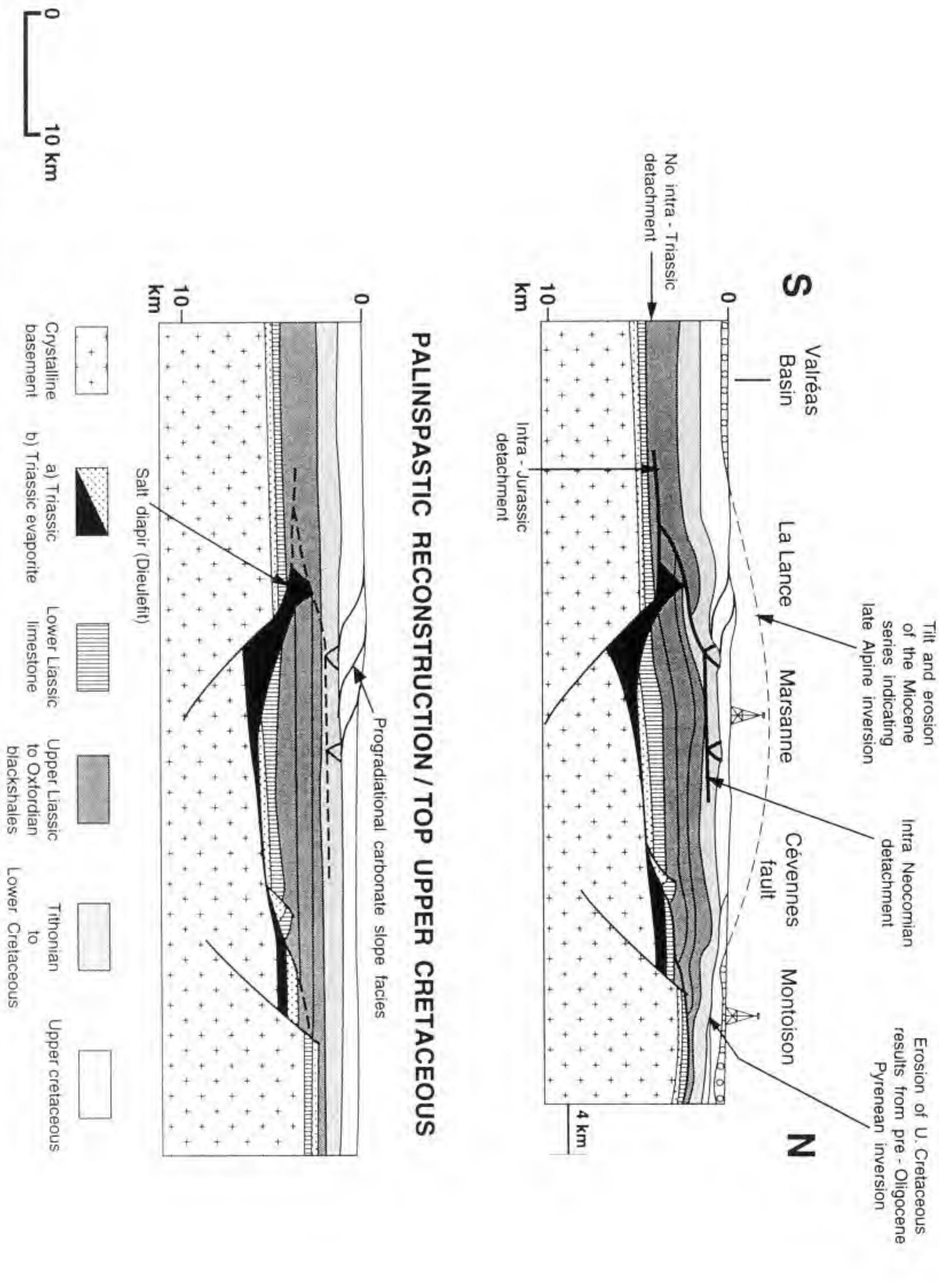

FIG. 13b. Structural and palinspastic sections across La Lance overthrust.

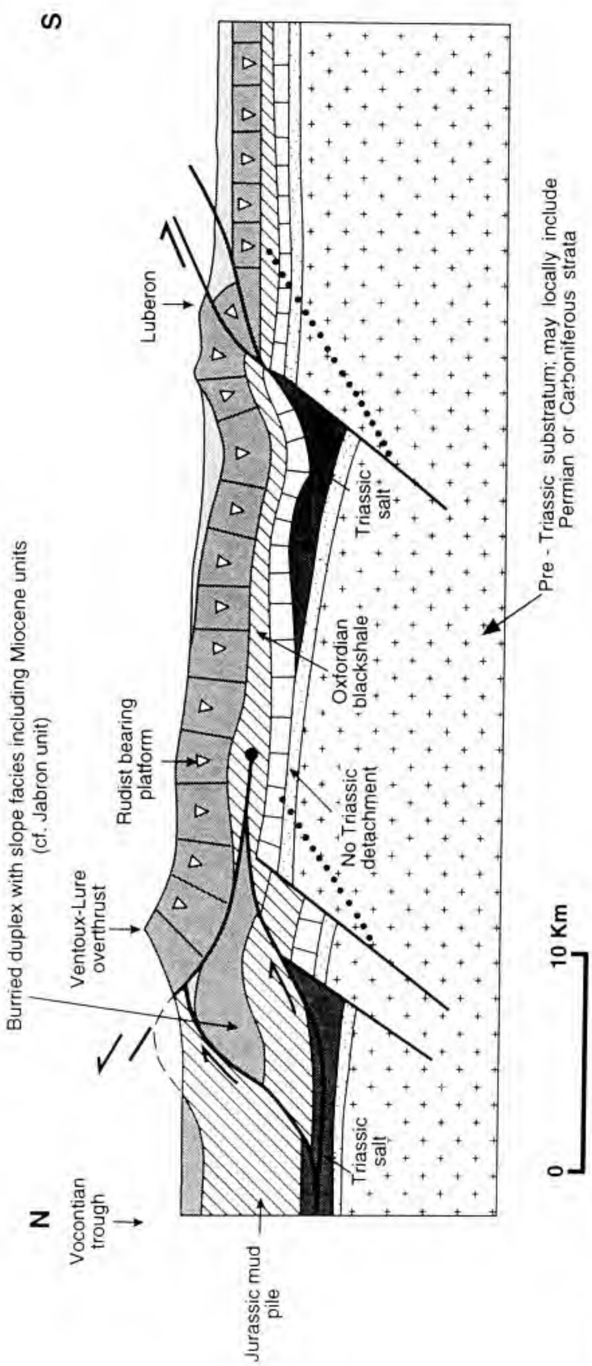

FIG. 13c. Structural and palinspastic sections across Ventoux-Lure overthrust.

Nevertheless, detailed microtectonic analyses outline the effects of the Alpine compression all over the European foreland, especially near the Jura front in the Bourgogne area (Bergerat, 1985; Lacombe et al., 1990). The important uplift and erosion observed in the southern part of the Rhine Graben, as imaged by longitudinal cross-sections of the graben (Sittler, 1965), could eventually account for Late Miocene inversion. Moreover recent strike-slip movements have been recognised on several faults of the Vosges Massif (Villemin and Bergerat, 1987), and a strike-slip mechanism is indicated by earthquakes in the southern part of the Rhine Graben (Bonjer et al., 1984).

Cenozoic foreland inversions are effectively also widespread further in the north, away from the Pyrenean and Alpine fronts, and have been described in the British Channel (Gillcrist et al., 1987), in the North Sea (Glennie and Boegner, 1981; Badley et al., 1989; Huyghe and Mugnier, 1994) as well as in Northern Germany and Poland (Ziegler, 1983, 1989, 1990 and references therein).

7 CONCLUSIONS

Tectonic inversion of pre-existing basement structures is the main mechanism which accounts for the structural complexity of the European foreland in Languedoc and Provence, north and west of the Pyrenean and Alpine thrust fronts, respectively. There, the competition between foreland basin inversions, resulting in localised uplifts, and low flexural subsidence of a thick lithosphere, prevented the development of large Late Cretaceous to Eocene and Miocene flexural foreland basins. Remnants of such basins are confined to relatively small depressions bordering the poorly expressed Pyrenean and Alpine frontal thin-skinned structures (St. Chinian, Montpellier fold, Arc syncline or Valensole Plateau). Oligocene rifting and subsequent thermal subsidence, linked to the development of the Gulf of Lions, provided additional complexities, and prevent any direct structural continuity between Languedoc and Provence compressional structures.

These compressional foreland deformations imply that the European plate as a whole was not rigid during the Pyrenean and Alpine deformations. Instead, the observed basement-involving within-plate deformations account for a progressive activation of very deep detachment levels within the continental lithosphere (presumably within the ductile lower crust), away from the recognised plate boundaries (Ziegler, 1983, 1989; Gillcrist et al., 1987;Roure et al., 1990; 1994).

However, the comparison of surface and subsurface data with sand-box experiments permits us to propose coherent structural interpretations at depth, and thus to link the major shallow complexities observed in the sedimentary cover with high-angle border faults delimiting Permian basins, and in some cases with Mesozoic or Oligocene extensional faults. Nevertheless, unlike in sand-box experiments or numerical models of tectonic inversion (Huygue and Mugnier, 1992; Vially et al., 1994), no clear basement short-cuts could be identified near the edges of reactivated normal faults. This may be due to poor resolution of the reflection seismic data beneath the inverted anticline structures (i.e. Pic-St. Loup and Ste. Victoire frontal structures, in which short-cuts could be expected), or an even greater importance of oblique mass transport than assumed (i.e. in La Lance or Durance). In this context, it is important to note that models predict that oblique compression facilitates the reactivation of pre-existing faults, precluding the development of basement short-cuts.

Due to the coexistence of a large set of trends, and due to progressive stress rotation during Late Cretaceous to Present times, the timing and mode of inversions was different for east-, north- or even northeast-trending structures. Orthogonal and oblique inversions thus coexisted at a regional scale, and multiphase deformations are recognised along a number of basement structures such as the Durance and Cévennes faults.

These new structural concepts may find application in the exploration for hydrocarbon. Most Late Paleozoic basins of southeastern France have a source potential for oil, as indicated by oil recoveries in Languedoc (Gabian trend beneath the intra-Triassic detachment; Barrabé and Schneegans, 1935), or recent oil shows in the subthrust Jura autochthon (Deville et al., 1994). Old mining industry cores also give evidence for the presence of mature source-rocks in the Paleozoic basins of Provence. Moreover, the recent discovery of oil in good quality Triassic reservoirs along the western border fault of the BSEF in the Ardèche (Morte-Mérie GPF scientific well; LeStrat et al., 1994), is apparently related to rich Permian source-rocks.

The structural complexity of the area presents a major exploration risk, particularly since prospects are mainly related to reservoirs located beneath the Triassic detachment surface. Maximum burial was generally reached prior to the inversion episodes; this would favour an initial migration of the hydrocarbon towards early extensional structures (Roure et al., 1994; Guilhaumou et al., 1995). However, inversion induced destruction of pre-existing hydrocarbon accumulations and the re-migration of oil and gas into newly developed structures, or their escape to surface, are further risk factors.

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Structure and evolution of the Central Alps and their northern and southern foreland basins

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ABSTRACT

A combined geological and deep reflection- and refraction-seismic profile crossing the Central Alps helps to unravel the crustal structure of this classical orogenic belt which had been the focus of pioneering geologists since he middle of the 18th century. New insights were gained by integrating the stratigraphic, structural, geochronologic and metamorphic record of the Alpine nappe systems and of the northern and southern foreland basins with new geophysical data on the deep structure of the Alps.

The Central Alps developed in response to Middle and Late Cretaceous dextral oblique partial or complete closure of oceanic basins, which had opened during Middle Jurassic to Early Cretaceous times, and to Paleogene orthogonal full-scale collision of the Apulian block with the European craton. Neogene continued convergence, accompanied by dextral transpression, resulted in thrust-propagation into the forelands and partial destruction of the flexural northern and southern foreland basins.

Across the Central Alps, Cenozoic N-S plate convergence amounting to 500 to 550 km was accompanied by subduction of substantial amounts of continental and oceanic lithospheric material. Following Paleogene collision of the Alpine orogenic wedge with the little attenuated northern foreland, Neogene back-thrusting governed the evolution of its southern parts. Imbrication of the northern and southern foreland crust, resulting in uplift of basement cored external massifs, is a consequence of continued post-collisional crustal shortening and lithospheric overthickening.

The Molasse Basin was displaced together with the Jura Mountain fold-and-thrust belt which represents the northernmost external unit of the Central-Alpine orogen. The Molasse Basin is a remnant of a fore-arc foreland basin. The thin-skinned external South-Alpine thrust belt scooped out an Early Mesozoic rift-induced basin, causing partial destruction of the southern, conjugate retro-arc foreland basin.

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This article includes 1 enclosure.

INTRODUCTION

This paper discusses the structure and evolution of the Central Alps on the basis of a regional geological-geophysical cross section which extends from the Molasse Basin of Eastern Switzerland into the Po Basin near the city of Milano. Supporting structural cross-sections are provided for the eastern and central parts of the Swiss Molasse Basin and the southern margin of the Southern Alps.

The geotransect, given in Enclosure 1, integrates surface and sub-surface geological data with refraction-seismic and deep reflection-seismic data. Geophysical data were acquired in the context of the Swiss National Research Project 20 (NFP-20; Pfiffner et al., 1988, 1996) and during the recording of the European Geotraverse (Blundell et al., 1992). This transect crosses the Central Alps where the external massifs plunge axially to the east-northeast and straddles the western erosional margin of the Austoalpine nappes (Fig. 1). This permits axial projection into the plane of the section of major structural units, including the basement-involving Aar and Gotthard massifs, the supra-crustal Helvetic and Penninic nappes and the orogenic lid, formed by the Austoalpine nappes. Correspondingly, this profile gives also a possible reconstruction for the eroded parts of the Alpine orogen (Schmid et al., 1996a and 1996b).

The Central Alps developed in response to Cretaceous and Cenozoic convergence of Africa-Arabia and cratonic Europe. This involved progressive closure of three oceanic basins which had opened during the Mesozoic break-up of Pangea and the development of the Tethys (Fig. 2). The oldest of these oceanic basins is the Hallstatt-Meliata Ocean which opened during the Middle Triassic along the eastern margin of the continental Apulia terrane (Italo-Dinarid Block); this ocean may have formed part of the Hellenic-Dinarid basin, referred to also as the Vardar Ocean. The second oceanic basin is the South Penninic (Piemont-Ligurian) Ocean which opened during the Middle Jurassic between Apulia and the continental Briançonnais domain. The third oceanic basin is the North Penninic (Valais) Trough which opened during the Early Cretaceous, thus separating the Briançonnais terrane from the Helvetic Shelf; the latter formed the southern continental margin of cratonic Europe.

In Enclosure 1, different signatures are given for continental basement complexes which are attributed to the proximal and distal parts of the European margin, the Middle Penninic Briançonnais terrane and the Austoalpine and South Alpine parts of Apulia. Ophiolitic sequences, corresponding to the floor of the former North Penninic Valais and the South Penninic-Piemont-Ligurian ocean, are highlighted in black. The Hallstatt-Meliata ocean is not involved in the area of the Central Alps, although its Late Jurassic closure did play a significant role in the evolution of the Austroalpine nappes (Stampfli et al., 1991; Froitzheim et al., 1996)

Enclosure 1 illustrates clearly that during the the Alpine orogeny the European and the Apulian margins were intensely deformed and that these deformations were not restricted to their sedimentary cover but involved large-scale imbrications of the basement which propagated far into the foreland. The autochthonous basement of the Molasse Basin extends only some 20 km beneath the external units of the Alps and rises to the surface in the imbricated Aar Massif. The Oligocene to Miocene synorogenic clastic wedge of the Molasse Basin attains a thickness of some 4000 m and is underlain by a relatively thin sequence of Mesozoic shelf series. Late Miocene and Pliocene compressional deformation of the Jura Mountains, attributed to in-sequence thrust propagation into the foreland, caused uplift and erosion of the western and central parts of the Molasse Basin (Laubscher, 1974; see also Philippe et al. and Roure and Colletta, this volume). Exploration for hydrocarbons in the Swiss Molasse Basins has yielded only oil and gas shows and one very small gas accumulation (Brink et al., 1992). In contrast, the southern margin of the Central Alps is characterized by a relatively wide, thin-skinned foreland fold-and-thrust belt involving a thick, southward tapering wedge of Mesozoic and Paleogene series overlain by synorogenic clastics (Cassano et al., 1986). To the north, this thin-skinned thrust belt gives way to a system of major basement imbrications such as the Orobic and Mezzoldo (Colitignone unit) blocks (Laubscher, 1985; Schönborn, 1992; Roeder and Lindsay, 1992). The discovery of major hydrocarbon accu-

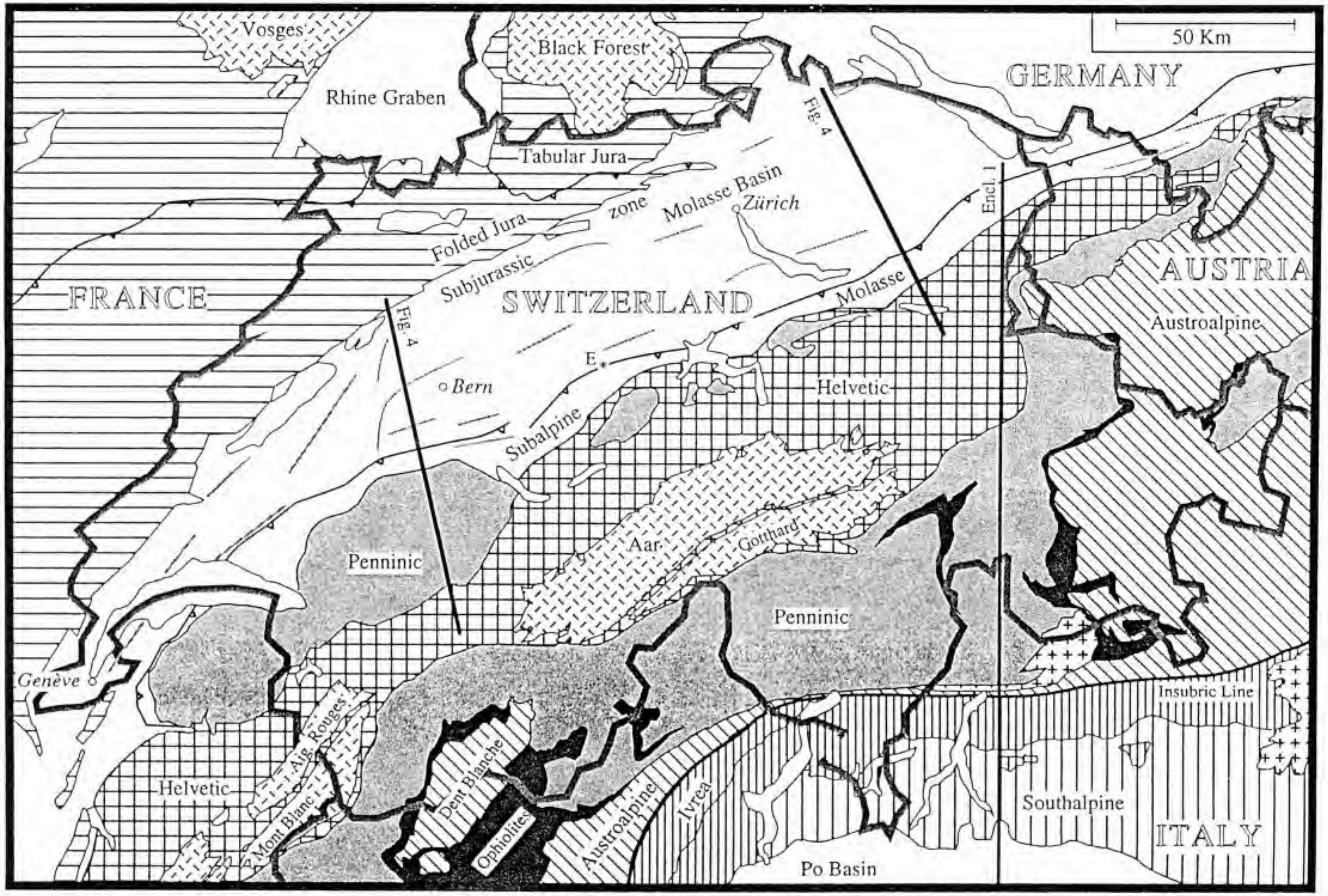


FIG. 1. Tectonic map of the Swiss Alps, showing major structural units and traces of cross-sections given in Fig. 4 and Encl. 1. Gray lines in Molasse Basin: major anticlines. E: Entlebuch gas accumulation. +: Tertiary intrusions. Narrow and wide spaced pattern in South-Alpine domain corresponds to crystalline basement and Mesozoic sediments, respectively.

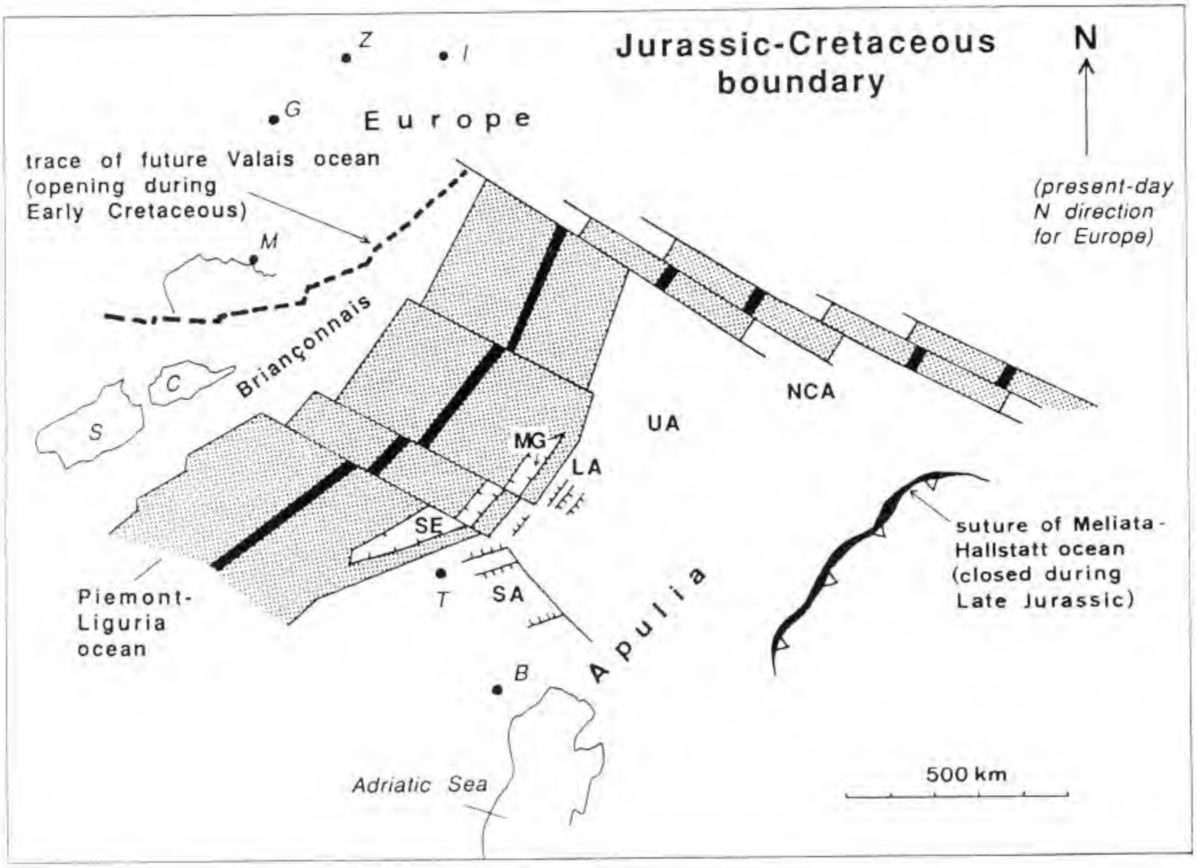


FIG. 2. Palinspastic sketch map of Alpine domain at end-Jurassic times (after Schmid et al., 1996b). LA: Lower Austroalpine domain, MG-Magua extensional allochthon, NCA: Northern Calcareous Alps, SA: South Alpine passice continental margin, SE-Sesia extensional allochthon, UA: Upper Austroalpine domain. Geographic reference: B (Bologna), C (Corsica), G (Geneva), I (Insbruck), M (Marseilles), S (Sardinia), T (Torino), Z (Zürich).

mulations, such as the Malossa gas/condensate field, testifies to the hydrocarbon potential of the South-Alpine external thrust belt (Anelli et al., this volume).

The two cross-sections through the Molasse Basin, given in Fig. 4, are based on industry-type reflection-seismic profiles which are calibrated by wells drilled during the search for hydrocarbons (Stäuble and Pfiffner, 1991; Pfiffner and Erard, 1996). The profiles through the Southern Alps and the adjacent Po Valley Basin, given in Fig. 6, are partly constrained by industry-type reflection-seismic profiles and well data (Schönborn, 1992).

EVOLUTION OF THE CENTRAL ALPINE OROGEN

The crystalline basement of the Alpine area was consolidated during the Variscan orogeny which terminated at the end of the Westphalian (von Raumer and Neubauer, 1993). However, during the terminal Stephanian and Early Permian phases of the Hercynian suturing of Godwana and Laurussia, crustal shortening persisted in the Appalachian orogen; this was accompanied by dextral shear movements between Africa and Europe, causing the collapse of the Variscan orogen and the subsidence of a system of wrench-induced troughs in which thick continental clastics accumulated. Following the Early Permian assembly of Pangea, a fundamental plate boundary reorganization underlies the development of the Tethys and Arctic-North Atlantic rift systems (Ziegler, 1990).

Opening of The Alpine Tethys Segment

During Late Permian and Triassic times, the Tethys rift systems propagated westward and interferred in the North Atlantic domaine with the southward propagating Arctic-North Atlantic rift system. In the East Alpine-Carpathian-Dinarid domain, rifting activity culminated in the intra-Triassic opening of a first system of oceanic basins, namely the Hallstatt-Meliata and Vardar oceans; these were possibly connected (Fig. 2). Following Middle Jurassic development of a discrete transform-divergent plate boundary between Gondwana and Laurasia, progressive opening of the Central Atlantic was accompanied by a sinistral transtensional translation of Africa-Arabia relative to Europe. This led to the opening of a second oceanic basin in the Alpine domain, the Liguria-Piemont-South Penninic oceanic basin, resulting in the isolation of the Apulian (Italo-Dinarid) microcontionent. Opening of the Ligurian-South Penninic Ocean went hand in hand with the gradual closure of the earlier formed Vardar and Hallstatt-Meliata oceans (Fig. 2). Latest Jurassic-earliest Cretaceous collision of the Apulia terrane with the eastern margins of the Vardar and Hallstatt-Meliata oceans and continued sinistral translation between Europe and Africa entailed the onset of counter-clockwise rotation of Apulia. This sequence of events indicates that opening of the Ligurian-South Penninic Ocean was neither spacially nor kinematically related to the opening of the Hallstatt-Meliata and Vardar oceans (Ziegler, 1988, 1990; Dercourt et al., 1993).

Early Cretaceous gradual opening of the North Atlantic and couter-clockwise rotation of Apulia were accompanied by the transtensional opening of a third oceanic basin in the Alpine domain, the North Penninic Valais Trough. The trace along which this youngest oceanic basin opened is shown in Figure 2, giving a latest Jurassic-earliest Cretaceous palinspastic sketch map of the Alpine region. Opening of the Valais Trough entailed separation of the continental Briançonnais terrane from Europe. It is questionalble whether the Briançonnais terrane formed part of the larger Iberian terrane, as postulated by Stampfli (1993), who visualizes a kinematic link between the opening of the Bay of Biscay and the Valais Trough. In this respect, data presente by Vially and Trémolières (this volume) suggest that the Corsica-Sardinia block remained attached to Europe during the Cretaceous opening of the Bay of Biscay and that the suture between Europe and Iberia projects from the Pyrenees to the south of Sardinia (after palinspastic restoration of the Corsica -Sardinia block; see also Ziegler, 1988). The eastern continuation of the Valais Trough is probably found within or near the northern margin of the earlier formed Piemont-Liguria Ocean (Rhenodanubian flysch and Upper Schieferhülle of the Tauern window, Outer Carpathian flysch belt). Correspondingly, the Briançonnais terrane is essentially confined to the Central and Western Alps. Relative movements between the European and Africa-Arabian continents and intervening microplates or terranes, leading to the opening and closing of oceanic basins in the Alpine domain, is discussed in greater detail by Stampfli (1993), Stampfli and Marchant (1996), Froitzheim et al. (1996) and Schmid et al. (1996a and 1996b).

Cretaceous Orogeny

Induced by the Cretaceous counter-clockwise rotation of Apulia, mass transport along its northwestern margin, facing the South Penninic-Piemont-Ligurian Ocean, was directed westwards. In the area of the Austroalpine units of Austria, closure of the Hallstatt-Meliata Ocean had occurred during a first stage in the Early Cretaceous (Neubauer, 1994). During the Cenomanian to early Turonian second stage of the Cretaceous orogeny, a dextral thrust wedge propagated westwards into the Central Alpine domain (see Schmid et al., 1996a and 1996b for a discussion of constraints on timing of orogenic activity along our transect). Subduction processes during both stages are indicated by the occurrence of Cretaceous-aged HP/LT eclogites which must be related to the activation of subduction zones along the former Meliata Ocean as well as along the northwestern margin of Apulia (Froitzheim et al., 1996). Late Cretaceous west-vergent imbrications and penetrative deformations, partly associated with metamorphism, are also observed in the Western Alps (France, Italy) and in the Eastern Alps (Austroalpine nappes), as discussed by Polino et al. (1990), Ring et al. (1989) and Froitzheim et al. (1994). The Austroalpine nappes were emplaced as thin allochthonous flakes onto the South Penninic ophiolites. This Late Cretaceous orogenic activity was accompanied by the shedding of clastics into the gradually closing South Penninic Trough. The Insubric Line marks the boundary between Austroalpine nappes, which are characterized by Cretaceous metamorphism, and the South Alpine domain which lacks such an overprint (Laubscher, 1991). However, in the South-Alpine domain, there is also good evidence for a Late Cretaceous first stage activation of the south-verging, basement involving Orobic and Gallinera foreland thrusts (Schönborn, 1992). These rising ramp anticlines acted as the source of the Turonian to Campanian flysch series which were deposited in the Lombardian Basin, located to the South of the South-Alpine domain (Bichsel and Häring, 1981; Bersezio and Fornaciari, 1987; Wildi, 1988; Bernoulli and Winkler, 1990).

Paleogene Orogeny

In conjunction with the Late Cretaceous and Paleogene step-wise opening of the Arctic-North Atlantic, sinistral motions between Europe and Africa decreased during the latest Cretaceous and Paleogene; with this the rotational movement of Apulia decreased gradually and westward mass transport along its northern margin came to an end. However, in connection with the progressive break-up of Gondwana, Africa-Arabia commenced to converge during the Senonian with Europe in a counter-clockwise rotational mode; this motion persisted during Cenozoic times (Ziegler, 1988, 1990) and controlled the collisional and post-collisional phases of Alpine orogeny.

During the late Senonian, the Austroalpine nappe stack was affected by tensional tectonics. This so-called Ducan-Ela extensional phase is viewed by Froitzheim et al. (1994) as reflecting the gravitational collapse of an overthickened orogenic wedge upon relaxation of the stress systems controlling its development. Exhumation and cooling of the Austroalpine units during the Ducan-Ela phase had severe implications for the subsequent evolution of the Central Alps. During the Cenozoic orogenic phases, the Austroalpine units remained largely undeformed and acted as a relatively rigid orogenic lid (in the sense of Laubscher, 1984), floating on viscously deforming Penninic units.

In our transect, the South Penninic Ocean was not closed before the end of the Cretaceous. The evolving orogen, which during the Late Cretaceous had been confined to the southeastern margin of the Piemont-Liguria Ocean and the Austroalpine-South-Alpine domain, collided in the Central Alpine region during the Paleocene with the southern margin of the Middle Penninic Briançonnais terrane (Figs. 3a and 3b; for timing constraints see Schmid et al., 1996a). However, in the Western Alps, collision of the evolving orogen with the Briançonnais terrane did not occur before the Oligocene, as evident by ophiolitic nappes overriding late Eocene pelagic series (Barféty et al., 1992).

During the Senonian, and particularly during the Paleocene, the European Alpine foreland was subjected to horizontal compressional stresses which gave rise to important intra-plate deformations, including the upthrusting of basement blocks and the inversion of Mesozoic tensional basins as far North as Denmark and the Central North Sea (Ziegler, 1990; Ziegler et al., 1995). In the area of the Central Alps, large parts of the Helvetic Shelf were uplifted at the end of the Cretaceous and subjected to erosion; this is confirmed by latest Cretaceous and Paleocene fission-track data from the Black Forest area (Wagner and van den Houte, 1992). Regional uplift and large radius deformation of the Helvetic Shelf caused the removal of much of its previously deposited Cretaceous cover and truncation and karstification of the Jurassic platform carbonates particularly in the area of the Jura Mountains, the Molasse Basin and the North Helvetic domain (Trümpy, 1980). Although the Paleocene deformation of the Helvetic Shelf of Switzerland was not as intense as further to the East in the area of the Bohemian Massif and the southward adjacent Austrian Molasse Basin (Zimmer and Wessely, this volume), its positive deflection must be related to compressional stresses which were exerted on the Alpine foreland in response to its collisional coupling with the evolving orogen (Ziegler, 1990; Ziegler et al., 1995). However, as by the end of the Cretaceous the Alpine orogenic front was still located along the southern margin of the Briançonnais terrane, it must be assumed that the lithosphere of the Valais Trough had sufficient strength to permit the transmission of large stresses through it and into the European foreland.

Along our transect, subduction of the Briançonnais microcontinent had commenced during the Paleocene and by the early Eocene this terrane was completely subducted together with the oceanic parts of the Valais Tough (Schmid et al., 1996b; Figs. 3a and 3b). By early Eocene times, the southern margin of the European foreland, corresponding to the Adula nappe, started to be overridden by the advancing more internal nappe systems of the Central Alps; subsequently it was subducted to great depth, as indicated by a Tertiary aged eclogite facies metamorphism (Figs. 3c and 3d; for timing of eclogite facies metamorphism in the Alps see Froitzheim et al., 1996). By late Eocene time, the Austroalpine and North Penninic nappes had advanced into the area of the future Gotthard massif which corresponds to the crystalline substratum of the future Helvetic cover-nappes (Fig. 3c). This led to the progressive flexural subsidence of the Helvetic Shelf under the load of the advancing orogenic lid, resulting in the development of a classical flexural foreland basin. By late Eocene time, marine transgressions had advanced northwards across the truncated Mesozoic strata to the southern margin of the present day Molasse Basin (Pfiffner, 1986; Lihou, 1995). Flexural subsidence of this foreland basin was accompanied by the development of an array of relatively small, essentially basin-parallel normal faults (Herb, 1965, 1992). Synsedimentary faulting is indicated by rapid lateral facies and thickness changes of Eocene sediments, containing large slump blocks of carbonates (Menkveld-Gfeller, 1995); this points to a considerable, fault-related relief in the Helvetic facies domain. During the Eocene-Oligocene phases of nappe emplacement onto the European foreland, the latter was apparently mechanically decoupled from the orogen, as there is no evidence for contemporaneous intraplate compressional deformations.

Detachment of the sedimentary cover of the Gotthard massif, resulting in the development of the Helvetic nappes, commenced during the late Eocene; by early Oligocene time, the Helvetic nappes, together with the overlying North Penninic and Austroalpine nappes, had advanced into the area of the future Aar massif (Figs. 3c and 3d). During the Paleocene and Eocene phases of the Alpine orogeny, substantial parts of the crust of the Briançonnais, the North-Penninic realm and the distal parts of the European foreland were subducted. However, the entire upper crustal volume of the more proximal and less attenuated part of the European crust (Gotthard and Lucomagno-Leventina units) was accreted to the orogenic wedge during the Oligocene and later phases. Resulting post-Eocene excessive thickening of the orogenic wedge implies that, following the main collisional event, only lower crustal material was subducted. Overthickening of the orogenic wedge was accompanied by south-directed back-folding north of, and back-thrusting along, the Insubric Line, causing rapid exhumation of the formerly deeply burried supra-crustal units in the Penninic (Lepontine) area during the Oligocene, as well as by thrust propagation into the northern foreland crust, resulting in step-wise imbrication of the Gotthard and Aar massifs and detachment of the Helvetic cover-nappes. In the Southern Alps, late Oligocene dextral transpressive movements along the Insubric Line induced in the area of the Lago Maggiore restraining bend East-West directed compressional deformations (Figs. 3e and 3f; Schumacher et al., 1996).

Oligocene post-collisional overthickening of the Alpine orogenic wedge was associated with the onset of northwestward movement of the rigid Adriatic indenter, south of the Periadriatic line (Schmid et al., 1989). This indenter is composed of stacked Apulian and European lower crustal and mantle material at its western end (Ivrea Zone and Ivrea geophysical body, Fig. 3e). In map view this indentation is associated with dextral strike slip

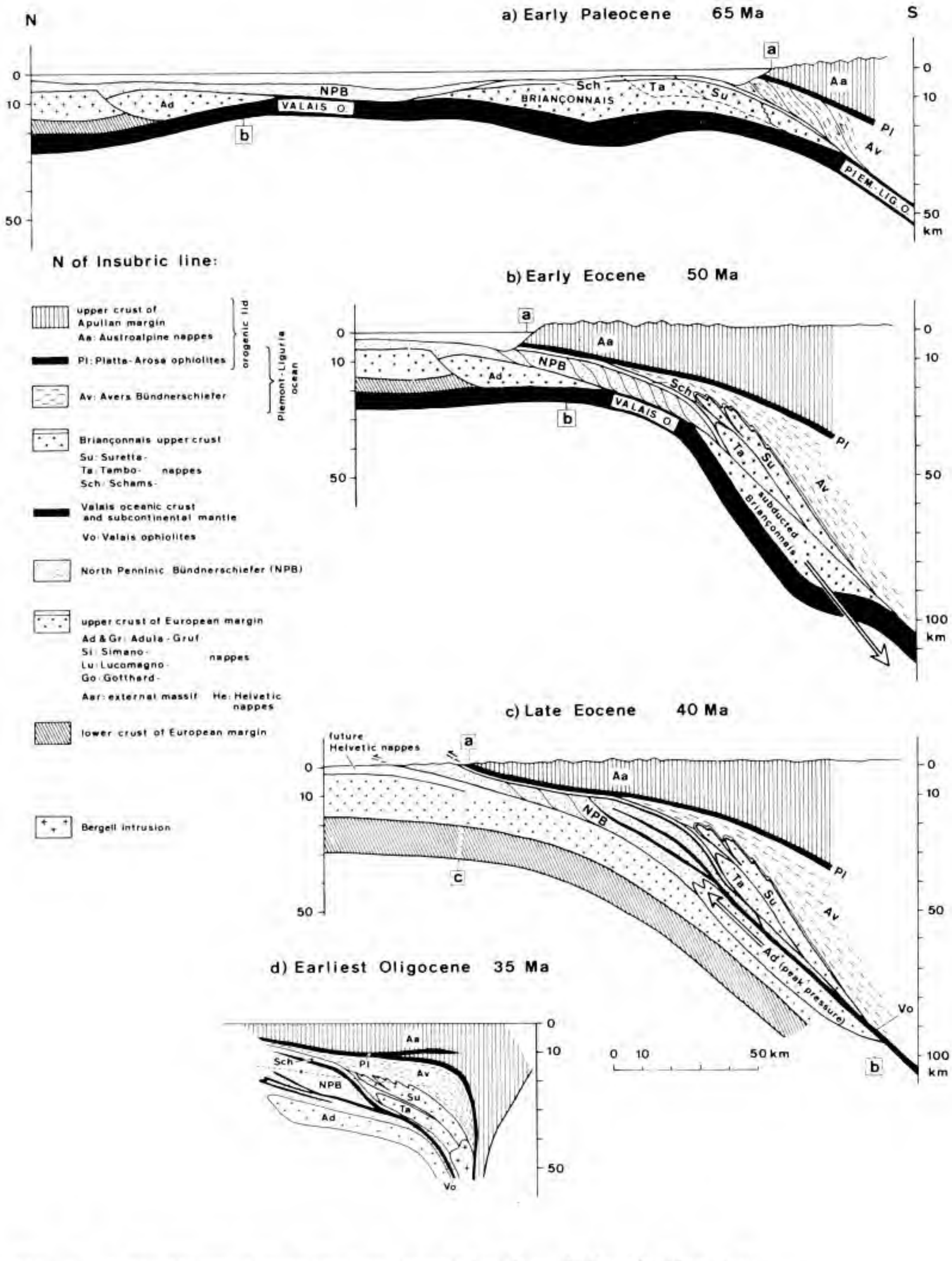
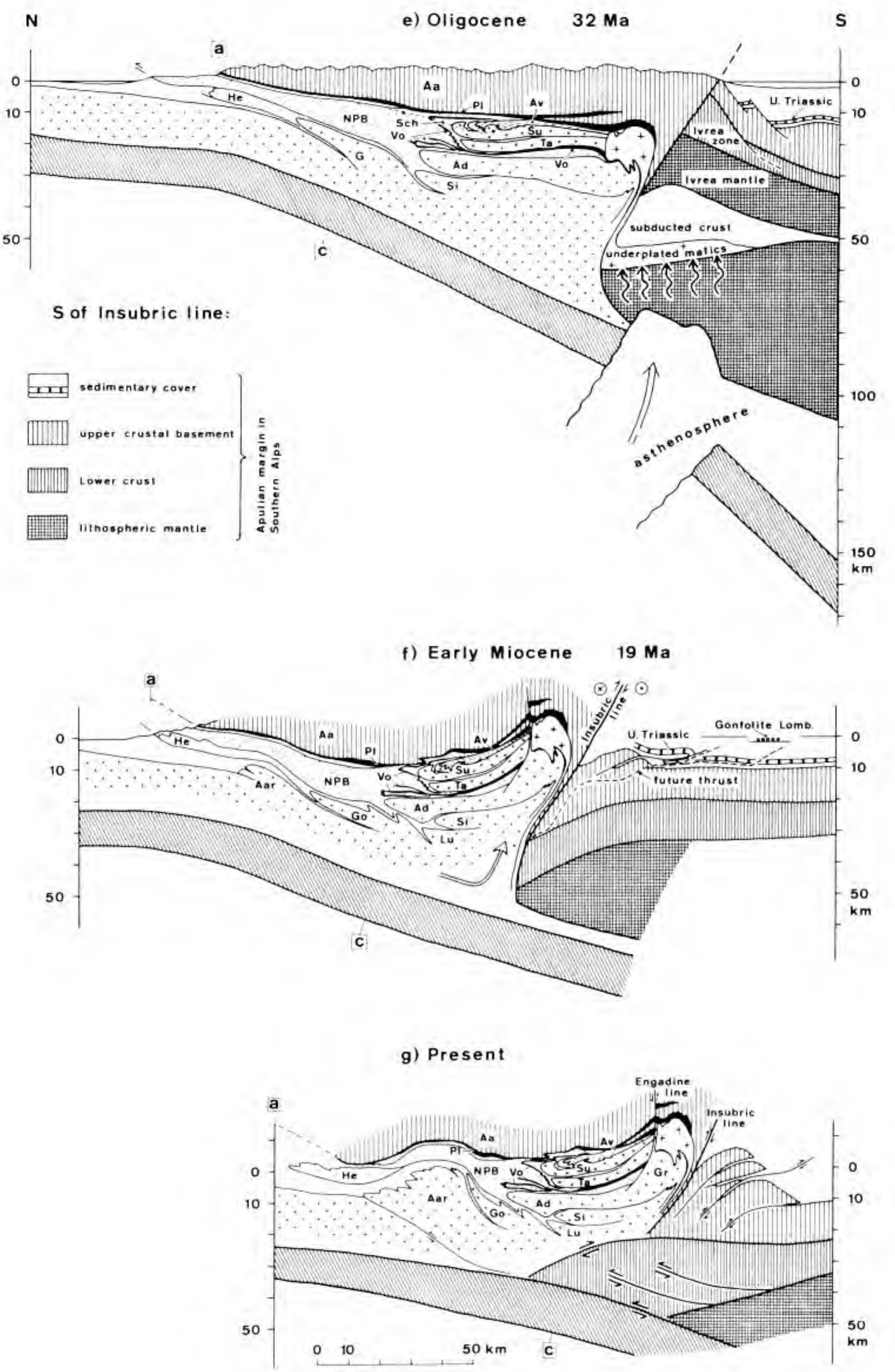


FIG. 3. Retro-deformed cross-sections through the Central Alps showing step-wise evolution of the Alpine orogen (after Schmid et al., 1996b).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upper crustal basement
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lithospheric mantle
Apulian margin in Southern Alps
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movements along the Insubric line. Note that, due to the subsequent dextral strike slip movements, Fig. 3e depicts a section through the Ivrea zone (after Zingg et al., 1990), presently located west of the transect given in Enclosure 1. Only by Miocene times (Fig. 3f) may the present-day section through the Southern Alps be depicted in our transect (Fig. 3f). In profile view, Oligocene differential uplift of the southern Penninic zone can be related to upwards directed material flow in its southern steep belt and its deflection into a North-directed horizontal movement of the Tambo-Suretta pair of nappes (Schmid et al., 1996a and 1996b). This induced spectacular refolding of some of the earlier formed Penninic nappe structures. These deformations were contemporary with the activation of the Glarus thrust along which the Helvetic nappes were transported to the North. During the Oligocene, the northern foreland basin subsided rapidly, expanded northwards into the area of the present-day Molasse Basin and received the detritus of the rising Alps. In the South-Penninic domain, the tonalitic and granodioritic Bergell pluton intruded syntectonically during early Oligocene post-collisional shortening (Rosenberg et al., 1995). These magmas were derived by partial melting of the mantle lithosphere and were contaminated by crustal material during their ascent; their generation is thought to be related to detachment of the subducted lithospheric slab and ensuing upwelling of the asthenosphere to the new base of the lithosphere (von Blankenburg and Davies, 1995). Considering the total amount of Cretaceous to early Oligocene crustal shortening along our Central Alpine transect, this slab may have had a length in excess of 400 km.

Although slab-detachment probably contributed to the rapid early Oligocene uplift of the Alpine orogen (Bott, 1993), thickening of the orogenic wedge due to continued northward convergence of Apulia with cratonic Europe was presumably the dominant mechanism. This process continued during the late Oligocene and Miocene but included now an orogen parallel dextral slip component that is difficult to quantify. However, late Oligocene to Miocene lateral movements along the Insubric Line alone amount to some 50 km. Along the transect given in Enclosure 1, late Oligocene to Recent crustal shortening in a North-South direction is estimated to amount to some 120 km (Schmid et al., 1996b) and resulted in the development of a correspondingly long new subduction slab. Based on seismic tomography, such a slab is at present still attached to the lithosphere of the Alpine orogen (de Jonge et al., 1993) and exerts a negative load on it (Bott, 1993). Of this total amount of late Oligocene to recent shortening along our transect, about 50 km were accommodated by imbrication of the European crust, 55 km by imbrication of the South Alpine basement and 15 km by back-thrusting along the Insubric Line.

Neogene Orogeny

During the late Oligocene and early Miocene, northward transport of the Helvetic nappes continued. By early Miocene time the Glarus thrust had probably broken surface and by mid-Miocene time, Helvetic detritus appeared in the Molasse sediments. Uplift of the Aar Massif along a crustal scale ramp commenced at the end of the Oligocene, persisted into late Miocene and Pliocene times and was probably directly linked to thrust deformation of the Sub-Alpine Molasse (Figs. 3e-g). Crustal shortening in the Aar Massif amounts to about 20 km. Folding of the Jura Mountains, which form the northwestern margin of the western and central Molasse Basin, commenced during the late Miocene (Serravallian/Tortonian, ±11Ma) and persisted into Pliocene and possible into recent times (Laubscher, 1987, 1992; Burkhard, 1990; Philippe et al., this volume). The origin of this external crescent-shaped fold belt, which separates from the Alps near Geneva, is under dispute. Shortening in the Jura Mountains, as derived mainly from surface geological criteria, boreholes and limited reflection-seismic data, decreases from approximately 30 km in its southwestern parts to zero at its northeastern termination. This amount of shortening may be taken up at an intra-Triassic sole thrust which extends from the Jura through the Molasse Basin and ramps down to the basement at the northern margin of the Aar Massif (thin-skinned model of Laubscher, 1961, 1992; Philippe et al., this volume). Alternatively, shortening may be transferred to an intracrustal a sole thrust, incorporating Permo-Carboniferous sediments and the upper parts of the crystalline basement, which extends from the the Jura Mountains through the area of the Molasse Basin beneath the Aar Massif (thick-skinned model; Ziegler, 1982, 1990; Pfiffner and Erard, 1996; Pfiffner, 1995). Burkhard (1990) notes that shortening in the Sub-Alpine Molasse increases northeastwards as shortening in the Jura Mountains decreases in the same direction. He postulates a 7o clock-wise rotation of the Mesozoic and Cenozoic sediments of the Molasse Basin above a basal detachment horizon and along a system of wrench faults. Folding of the Jura Mountains entailed uplift and partial destruction of the Molasse Basin. The degree of uplift of this basin increases towards the southwest as shortening in the Jura Mountains increases. According to both the thin-skinned and the thick-skinned model, the Molasse Basin and the Jura fold-and-thrust belt form part of a major allochthon which represents the most external element of the Central Alpine orogen.

Back-thrusting of the South Penninic nappes over the South Alpine domain along the Insubric Line persisted during the late Oligocene under a dextral transpressive scenario; however, by Miocene times, movements along the Insubric Line were purely dextral (Schmid et al., 1989; Figs. 3e-g). Pebbles and boulders of the Bergell pluton appeared during the latest Oligocene-earliest Miocene in the deeper water Lombardian foreland basin (Gonfolite Lombardia; Giger and Hurford, 1989). In most of the Southern Alps, Tertiary-aged thrusting did not resume before the mid-Burdigalian (Schönborn, 1992). The external, thin-skinned Lombardy thrust belt is sealed by the Messinian unconformity and is covered by up to 2.5 km of latest Miocene and Plio-Pleistocene, only slightly folded clastics. Uplift of the internal parts of the Southern Alps is related to the stacking upper crustal thrust sheets which were detached at a mid-crustal level. The corresponding lower crust and mantle lithosphere forms the Adriatic or Apulian wedge which interfaces with the south-dipping European lower crust beneath the Central Alps (Encl. 1 and Fig. 3g). The geometry of this wedge is, according to the resolution of the geophysical data available, only schematically outlined in Enclosure 1. In fact, this highly reflective wedge may have a more complicated internal structure (Hitz, 1995). However, for material balance reasons, imbrication of lower crustal material within this Adriatic wedge is a corollary of some 46 km Miocene-aged N-S shortening taking place within the South Alpine upper crustal fold- and thrustbelt (Schönborn, 1992; Schmid et al., 1996a and 1996b). Hence, shortening at upper crustal levels south of the Insubric line is kinematically linked to thickening within the Apulian wedge located beneath the southern part of the Central Alps. Formation of this Neogene wedge post-dates back-thrusting along the Insubric line and thrusting of the Helvetic nappes; however, it is contemporaneous with shortening in the external massifs and the Molasse-Jura allochthon (Laubscher, 1991).

The Central Alpine orogen is at present tectonically still active as evident by earthquake activity and an uplift rate of about 1 mm/year. Within the Central Alps, earthquake hypocentres are concentrated in the upper crust whereas in the Molasse Basin they are distributed over the entire crust. Focal mechanisms indicate that the crust of the Molasse Basin is affected by sinistral and dextral shear with the principal horizontal compressional stress trajectories trending NW-SE, compatible with the overall stress fied of Central Europe (Pavoni, 1990; Deichmann and Baer, 1990; Balling and Banda, 1992; Grünthal and Strohmeyer, 1994).

MOLASSE BASIN

The Swiss Molasse Basin is limited to the northwest by the Jura Mountains and to the southeast by the Alps (Fig. 1). Its sedimentary fill consists of a southeastward expanding, up to four kilometres thick wedge of Oligocene and Miocene sandstones, conglomerates and shales, derived from the Alpine orogen, which rests unconformably on truncated Mesozoic carbonates, shales and clastic rocks, ranging in thickness between 1.5 and 3 km. The latter overlay a Variscan basement complex and, more locally, several kilometres thick Permo-Carboniferous clastics contained in fault-bounded, wrench-induced troughs (Fig. 4, Encl. 1).

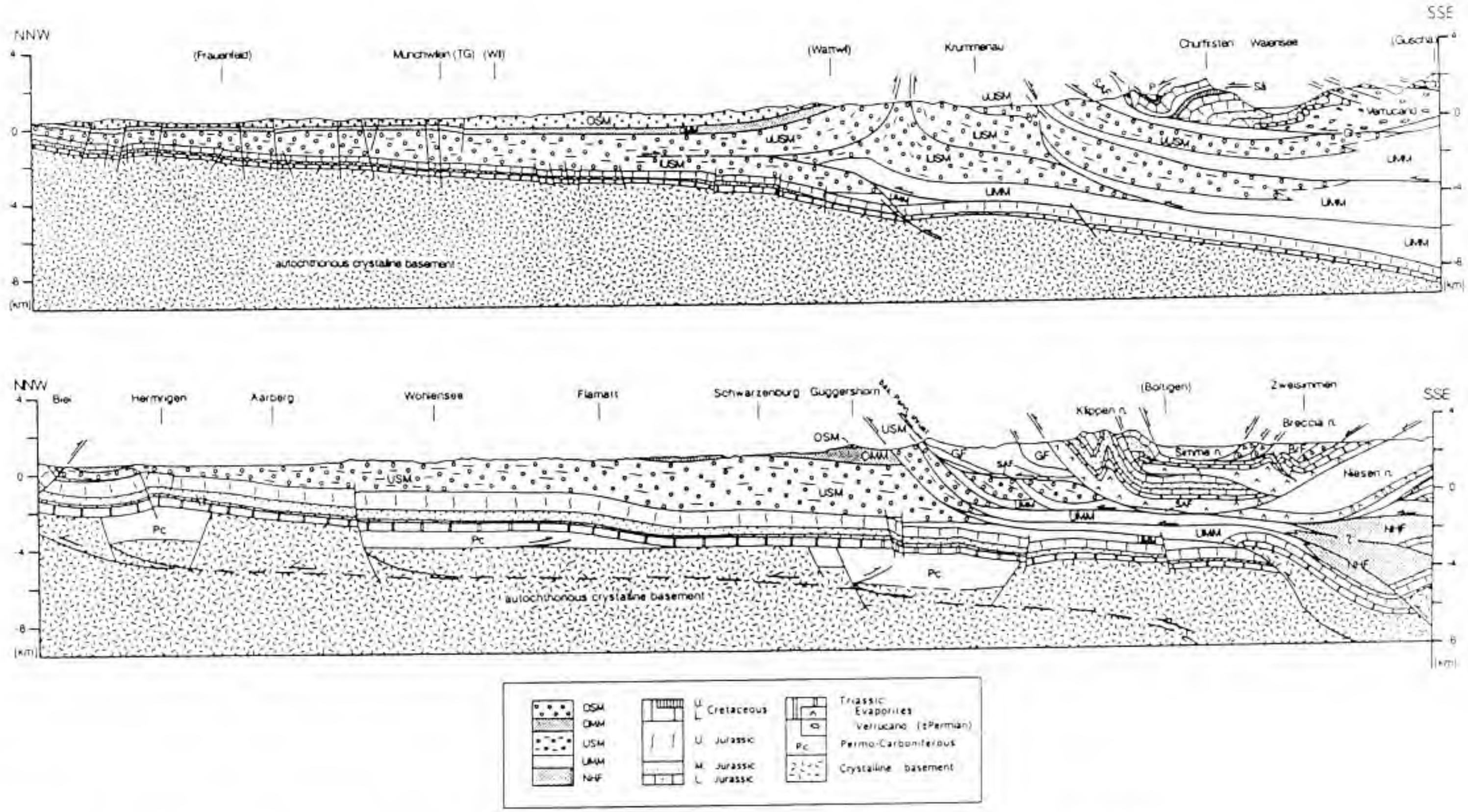

FIG. 4. Structural cross-sections through the eastern (top) and central (bottom) Molasse Basin (modified from Pfiffner and Erard, 1996). The basin fill is cut by numerous faults and thrust faults in the internal part, the Subalpine Molasse. The western cross-section shows fault-bounded Permo-Carboniferous basins, which were partly inverted; the top of the crystalline basement is shown to be allochthonous, in agreement with the Jura Mountains as a thick-skinned fold-and-thrust belt. OSM: Upper Freshwater Molasse, OMM: Upper Marine Molasse, USM: Lower Freshwater Molasse, UMM: Lower Marine Molasse; NHF: North-Helvetic Flysch, SAF: Sub-Alpine Flysch, P: Penninic nappes, GF: Gurnigel Flysch, Gl: Glarus thrust, Sä: Säntis thrust.

The Mesozoic evolution of the area occupied by the Swiss Molasse Basin was dominated by regional tensional stress regimes related to the break-up of the Late Palaeozoic Pangea. Its latest Cretaceous and Tertiary evolution was governed by the development of the Alpine orogen and the folding of the Jura Mountains. The Cenozoic Rhine-Rhône rift system influenced the evolution of the Molasse Basin only marginally (Trümpy, 1980; Ziegler 1990, 1994).

Basin Evolution

The crystalline basement underlying the Swiss Molasse Basin was consolidated during the Variscan orogeny which terminated at the end of the Westphalian. Stephanian-Autunian wrench-faulting gave rise to the subsidence of often narrow and deep fault-bounded troughs in which continental clastics, partly coal-bearing and containing lacustrine shales, accumulated. At the same time the Variscan fold belt was uplifted and deeply eroded. Late Permian, Triassic and Early Jurassic series transgressed, under a regional tensional setting, over this erosional surface from the Northeast and the Southwest and onlapped against the Alemannic High, coinciding partly with the area of the present Aar Massif. This led to the gradual establishment of a broad basin which occupied the area of the future Swiss and German Molasse Basin, extended northwestwards into the Paris Basin and was linked to the North with the Northwest European Basin (Ziegler, 1990; Bachmann et al., 1987). Following Mid-Jurassic crustal separation in the South-Penninic domain, the Alemannic High subsided and a wide carbonate and shale platform occupied during the Middle and Late Jurassic and the Cretaceous the Helvetic Shelf. This broad shelf occupied large parts of southern Germany and extended through the area of the future Swiss Molasse Basin into the Paris Basin. Early Cretaceous tectonic instability of this shelf, reflected by subsidence anomalies (Funk, 1985; Loup, 1992), can be related to the transtensional opening of the Valais Trough and to rifting activity in the North Sea and the Bay of Biscay. Palaeogeographic reconstructions suggest that the entire Helvetic Shelf was once covered by Late Cretaceous carbonate-dominated sediments. During the latest Cretaceous and Paleocene large parts of the Helvetic Shelf were uplifted and mildly deformed in response to compressional stresses exerted on it from the collision zone between the Alpine orogen and the Briançonnais block. This uplift, which reflects broad lithospheric buckling and smaller scale crustal deformations, caused the development of a regional unconformity, which in the area of the Molasse Basin cut deeply into the Cretaceous and Late Jurassic strata, causing karstification of carbonate rocks and leaching of Triassic salts, resulting in the development of salt pillows (Ziegler, 1990). Thrust-loading of the Helvetic Shelf commenced apparently during the Eocene, as evident by the gradual development of a foreland basin in the proximal parts of which syn-orogenic clastics accumulated in deeper waters whereas in its distal parts fluvio-deltaic sands, derived from the foreland, and shallow water carbonates were deposited. Progressive northward displacement of the basin axis was accompanied by overstepping of its northern margin and the development of a system of essentially basin-parallel normal faults (Herb, 1965, 1992; Menkveld-Gfeller, 1995). In the southernmost parts of the Molasse Basin, sedimentation commenced during the latest Eocene to earliest Oligocene, rapidly spread northward during the middle Oligocene and persisted under alternating shallow marine and continental conditions into early late Miocene times.

The main phase of folding of the Jura Mountains spans late Miocene to Pliocene times (Laubscher, 1987, 1992; Kählin, 1993; Bollinger et al., 1993). At the same time the Molasse Basin was uplifted and its sedimentary fill subjected to erosion with the degree of uplift and erosion increasing towards the southwest in tandem with increasing shortening in the Jura Mountains. The seismicity of the Molasse Basin and of the Jura Mountains, as well as geodetic data indicate that crustal shortening is at present still active (Pavoni, 1990; Deichmann and Baer, 1990; Jouanne et al., 1995).

Basin Architecture

Figure 4 gives two reflection-seismically controlled cross-sections though the Swiss Molasse Basin; the eastern section crosses the basin to the east of the Jura Mountains whereas the western section extends from the folded Jura Mountains to the Alps. These sections show that the southeastern parts of the Molasse Basin were imbricated during the uplift of the Are Massif and are partly overridden by sedimentary nappes (Pfiffner and Erard, 1996). Moreover, they illustrate that the Mesozoic and basal Tertiary series of the Molasse Basin are cut by numerous normal faults, some of which were compressionally reactivated at a later stage. However, some normal and wrench faults appear to cut to the surface. Compressionally reactivated faults play an increasing important role in the central and western parts of the Molasse Basin. Some of these structures are related to partial inversion of Permo-Carboniferous troughs (Brink et al, 1992; Gorin et al., 1993; Pfiffner and Erard, 1996). Ramp anticlines, involving Mesozoic carbonates carried to surface by thrusts soling out in Triassic evaporites, play only a significant role to the southwest of Lake Geneva (Gorin et al., 1993; Philippe et al., this volume).

Small-scale extensional faults, cutting up from the basement through the Mesozoic series and dying out in the lower part of the Oligocene sediments, must be related to flexure of the foreland during its thrust-loaded subsidence. This type of faulting is well expressed in the German and Austrian Molasse Basin where reflection-seismic data, calibrated by numerous wells, permit dating of fault activity as ranging from early to late Oligocene with fault activity younging towards the north in conjunction with gradual northward displacement of the basin axis (Bachmann and Müller, 1992; Roeder and Bachmann, Wessely and Zimmer, this volume). On the other hand, faults cutting up from the basement through Mesozoic and Tertiary strata to the surface were presumably active during the main folding phase of the Jura Mountains or may even post-date it. Some of these faults probably form part of wrench systems which accommodated rotation of the Molasse Basin during the deformation of the Jura fold-and-thrust belt (Burkhard, 1990; Brink et al., 1992).

Based on recently released reflection-seismic data, Oligocene normal faults, which were not reactivated in later times, are also evident in the central and western parts of the Swiss Molasse Basin (Brink et al., 1992; Gorin et al., 1993; Pfiffner and Erard, 1996). This raises doubts about the applicability of the thin-skinned distant-push model, proposed by Laubscher (1961, 1974) (see Philippe et al., this volume) for the development of the Jura Mountains. However, it must be kept in mind, that for the French Jura Mountains an initial phase of thin-skinned thrusting, followed by a phase of basement-involving shortening is envisaged (Jouanne et al., 1995). Pfiffner and Erard (1996), following the earlier proposed model of Ziegler (1982, 1990), envisage that during the folding of the Jura Mountains compressional reactivation of Permo-Carboniferous troughs, underlying part of the Jura and the Molasse Basin, was accompanied by the development of an intra-crustal detachment along which the uppermost crust of the Molasse Basin, together with its sedimentary cover, was transported northwestwards. Pfiffner (1995), based on balanced cross-sections through the Jura Mountains, estimates that a minimum of about 2.5 km of crystalline basement and/or Permo-Carboniferous sediments, were incorporated into this thick-skinned detachment. However, the distribution of earthquake hypocentres in the North-Alpine foreland suggest that the whole crust underlyig the Molasse Basin is presently undergoing brittle deformation; moreover, focal mechanisms indicate that these deformations are controlled by northwest directed compressional stresses (Deichmann and Baer, 1990). In contrast to the thin-skinned model, the thick-skinned model accounts fully for the observed uplift of the Molasse Basin.

Decoupling of the Mesozoic and Cenozoic strata from their autochthonous basement has occurred in the French part of the Molasse Basin, is also evident in many parts of the Jura belt (Buxtorf, 1907, 1916; Guellec et al., 1990; Philippe et al., this volume) and has been confirmed by recently released reflection-seismic data from the central Jura (Sommaruga, 1995). The southeastward continuation of such detachments beneath the Molasse Basin and the occurrence of intra-basement thrusts below the Jura is at present debated. Sheared Triassic evaporites have also been encountered in a number of boreholes drilled in the east-central parts of the Molasse Basin, thus attesting to the activation of an intra-sedimentary decoupling layer (Jordan, 1992). On the other hand, locked normal faults suggest that basement involved rotation of the Molasse Basin may have contributed to the shortening observed in the Jura fold and thrust belt. The lack of access to the entire reflection-seismic data base, acquired by the Petroleum Industry in the Molasse Basin, impedes the evaluation of the relative contribution of thin- and thick-skinned deformations to the folding of the Jura Mountains.

Hydrocarbon Habitat

During the exploration for hydrocarbons in the Molasse Basin some 8500 km of reflection-seismic lines were recorded and 33 wells drilled, resulting in the discovery the very small Entlebuch gas accumulation (see Fig. 1; rec. res. 3.6 BCF gas; Lahusen, 1992; Brink et al., 1992; Gunzenhauser and Bodmer, 1993). However, as most of the wells yielded minor oil and gas shows, hydrocarbon charge does not appear to be the primary constraining factor in the hydrocarbon potential of the Molasse Basin.

Coals and lacustrine shales of the Permo-Carboniferous series, contained in wrench-induced, partly inverted troughs, provide non-predictable potential source-rocks; these have reached maturity in most of the area. Early Jurassic organic shales are generally mature for oil generation under the Molasse Basin and enter the gas window near the Alpine deformation front where they are less well developed. The basal Oligocene "Fischschiefer" (Sannoisian), the primary oil source-rock in the Bavarian and Austrian Molasse Basin, may occur only beneath the Alpine nappes where they have probably reached maturity. The Val de Travers tar deposit of the western Jura Mountains indicates that hydrocarbon generation and migration had occurred already prior to the deformation of the Jura fold and thrust belt.

Potential reservoirs are the Triassic Bunter sands and Muschelkalk dolomites which are sealed by salts. Rhaetian sands, sealed by Early Jurassic shales, are poorly developed. Karstified Jurassic carbonates, partly developed in a reefal facies, are only sealed by early Oligocene marine shales in the deepest parts of the basin where they host the Entlebuch gas accumulation. Elsewhere these carbonates are directly overlain by Oligocene sands and therefore are not sealed. The Oligo-Miocene Molasse series lack well defined reservoir-seal pairs.

Remaining prospects in the Molasse Basin are related to sub-salt Triassic reservoirs, charged by Permo-Carboniferous source-rocks, and to Mesozoic carbonates and sands, charged by Early Jurassic source-rocks and sealed by basal Oligocene marine shales. Triassic prospects in the northwestern parts of the basin are difficult to define by reflection seismic data. Similarly, in the southeastern-most parts of the basin, definition of Jurassic carbonate prospects is impeded by the complex overburden of the Alpine nappes and by topographic constraints. In view of these difficulties and past discouraging results, exploration activity in the Molasse Basin recently has been discontinued.

SOUTH-ALPINE THRUST BELT

The arcuate central South-Alpine thrust belt has a width of 80 km and is bounded to the North by the Insubric Line (Fig. 5). Its internal parts consist of stacked, basement-involving thrust sheets, whereas its external parts are characterized by thin-skinned thrust sheets which are detached from the basement at Triassic levels (Fig. 6; Roeder and Lindsey, 1992; Schönborn, 1992). The Po Plain hosts a remnant Oligo-Pliocene foreland basin which is underlain by a thick Paleogene and Mesozoic succession. The external parts of the South-Alpine thrust belt and the Po Basin have been extensively and successfully explored for hydrocarbons (Pieri and Groppi, 1981; Cassano et al., 1986; Anelli et al., this volume).

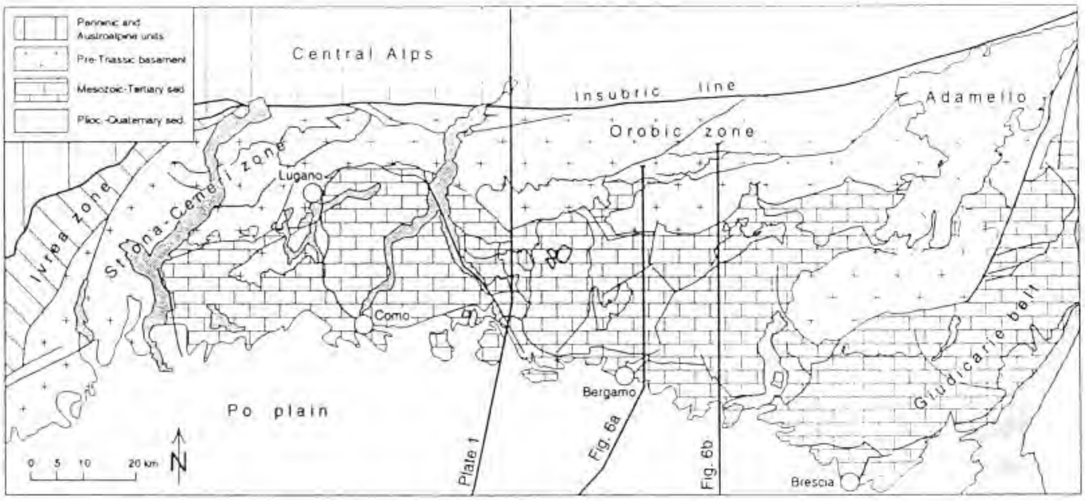

FIG. 5. Tectonic map of Lombardian fold-and-thrust belt, showing traces of cross-sections given in Fig. 6 and Encl. 1 (modified after Schönborn, 1992)

Basin Evolution

Following termination of the Variscan orogeny, the area of the Southern Alps was affected by wrench tectonics (Arthaud and Matte, 1977), controlling the accumulation of up to 1.5 km thick latest Westphalian to Early Permian continental clastics in SW-NE trending transtensional basins and a widespread intrusive and extrusive magmatism (Cassinis and Perotti, 1993). Late Permian continental clastics were deposited on a regional peneplane under tectonically quiescent conditions; these grade upwards into shallow marine Early Triassic sands, carbonates and evaporites (Asserto and Casati, 1966).

During the Middle Triassic, development of a complex pattern of carbonate platforms and intervening anoxic basins was accompanied by transtensional/transpressional tectonics and widespread volcanic activity, probably related to opening of the Hallstatt-Meliata Basin (Stampfli et al., 1990). Tectonic and volcanic activity persisted during the Carnian low-stand in sea-level during which terrigenous clastics were shed into basinal areas from the south (Brusca et al., 1981). By end-Carnian times, the South-Alpine domain was occupied by a uniform evaporitic platform, reflecting renewed tectonic quiescence.

However, rifting activity resumed during the deposition of the Norian Hauptdolomite, as evident by its lateral thickness changes describing the development of northerly trending platforms and intervening basins; amongst the latter, the Lombardy Basin is of special interest as its configuration controlled the geometry of the external South-Alpine thrust belt (Castellarin and Picotti, 1990). During the Rhaetian, up to 2500 m of black shales, capped by shallow water carbonates, were deposited in this basin under rising sea-level conditions (Gnaccolini, 1975); offsetting platforms were characterized by considerably thinner series. Rifting activity intensified during Early Jurassic times, as indicated by the accumulation of up to 4 km of hemipelagic carbonates in basinal areas, containing slump breccias derived from active fault-scarps, and of shallow water carbonates on platforms (Bernoulli, 1964). Erosion on some platforms reflects extensional footwall uplift (Gaetani, 1975). During the Toarcian and Middle Jurassic, rifting activity abated in the central South-Alpine domain and shifted westward towards the margin of the Piemont Trough. In the Lombardy Basin, which by now had subsided below the photic zone, carbonate turbidites, followed by the pelagic Rosso Ammonitico were deposited. The flanking platforms were drowned during the Callovian and Oxfordian. Late Bathonian crustal separation in the Piemont Basin was followed by regional subsidence of the South-Alpine domain in which Late Jurassic series are represented in basinal areas by radiolarites and on palaeo-highs by Rosso Ammonitico-type limestones (Winterer and Bosellini, 1981; Bertotti et al., 1993).

During the latest Jurassic and Neocomian, the area was covered by a blanket of coccolith limestones (Maiolica). After a short break during the early Aptian low-stand in sea-level, sedimentation resumed with the deposition of black shales, grading upwards into hemipelagic marly limestones (Scaglia). late Cenomanian onset of terrigenous flysch influx from northern sources, presumably reflects uplift and erosion of the internal South-Alpine domain to basement levels; this flysch cycle culminated during the early Senonian and lasted until Campanian times. During the late Campanian hemipelagic shaly carbonate deposition resumed (Scaglia) and lasted until late Eocene times (Bichsel and Häring, 1981; Bersezio and Fornaciari, 1987; Bernoulli and Winkler, 1990).

The turbiditic, partly conglomeratic "Gonfolite Lombardia" was derived from northern sources; this syn-orogenic succession ranges in age from late Oligocene to middle Miocene, attains thicknesses of up to 3500 m and shales out towards the south. It was deposited in a typical foreland basin. Deformation of the South-Alpine external thrust belt is Seravallian to Tortonian in age. All thrusts and folds are sealed by the Messinian unconformity which is related to an evaporation-induced draw-down of the Mediterranean sea-level. The up to 2.5 km thick Messinian and Plio-Pleistocene sedimentary fill of the Po Basin essentially post-dates the deformation of the Southern Alps and is affected by gentle folding only (Pieri and Groppi, 1981). Therefore, it can be regarded as the fill of the North-Appenine foreland basin (Gunzenhauser, 1985; Schönborn, 1992; Cassano et al., 1986).

Basin Architecture and Hydrocarbon Habitat

The arcuate geometry of the thin-skinned external thrust belt of the central Southern Alps was preconditioned by the configuration of the Mesozoic Lombardian Basin. The availability of Triassic detachment horizons and the rheological composition of the carbonate dominated Triassic and Jurassic series favoured the development of 10 to 20 km wide thrust sheets and in-sequence thrust propagation (Schönborn, 1992). Deformation of this thrust belt resulted in partial destruction of the Oligo-Miocene flexural foreland basin (Figs. 5 and 6).

Middle and Late Triassic basinal shales are oil-prone source-rocks (Bernasconi and Riva, 1993; Stefani and Burchell, 1993). Fractured, low porosity Late Triassic and Early Jurassic dolomites form the reservoir of the Malossa gas/condensate field which was drilled on a ramp-anticline. The Oligo-Miocene sandy Gonfolite Group contains reservoirs which are charged by hydrocarbons generated from Mesozoic source-rocks. The Pliocene series of the Po Valley contain biogenic gas (see Anelli et al., this volume).

CONCLUSIONS

The Central-Alpine orogenic wedge consists of a stack of nappes which involve continental and oceanic crustal and supra-crustal rocks. This wedge started to develop during Cretaceous oblique subduction of the Hallstatt-Meliata Ocean and of the southeastern parts of the South Penninic-Piemont-Liguria Ocean (Cretaceous orogeny). Final closure of the South Penninic Ocean and the partly oceanic Valais Trough is attributed to the Paleogene second orogenic cycle, post-dating the latest Cretaceous collapse of the Eoalpine orogen. The Cenozoic orogenic phases were governed by collision of the Apulian block with the European craton. Cumulative Tertiary-aged crustal convergence across the Central Alps amounts to some 550 km. Basement cored nappes typically involve only 5 to 10 km of continental crustal material. Moreover, only relatively small volumes of oceanic crustal material were incorporated into the Alpine orogenic wedge. Therefore, sizable volumes of lithospheric material, including oceanic and continental crust, were apparently subducted.

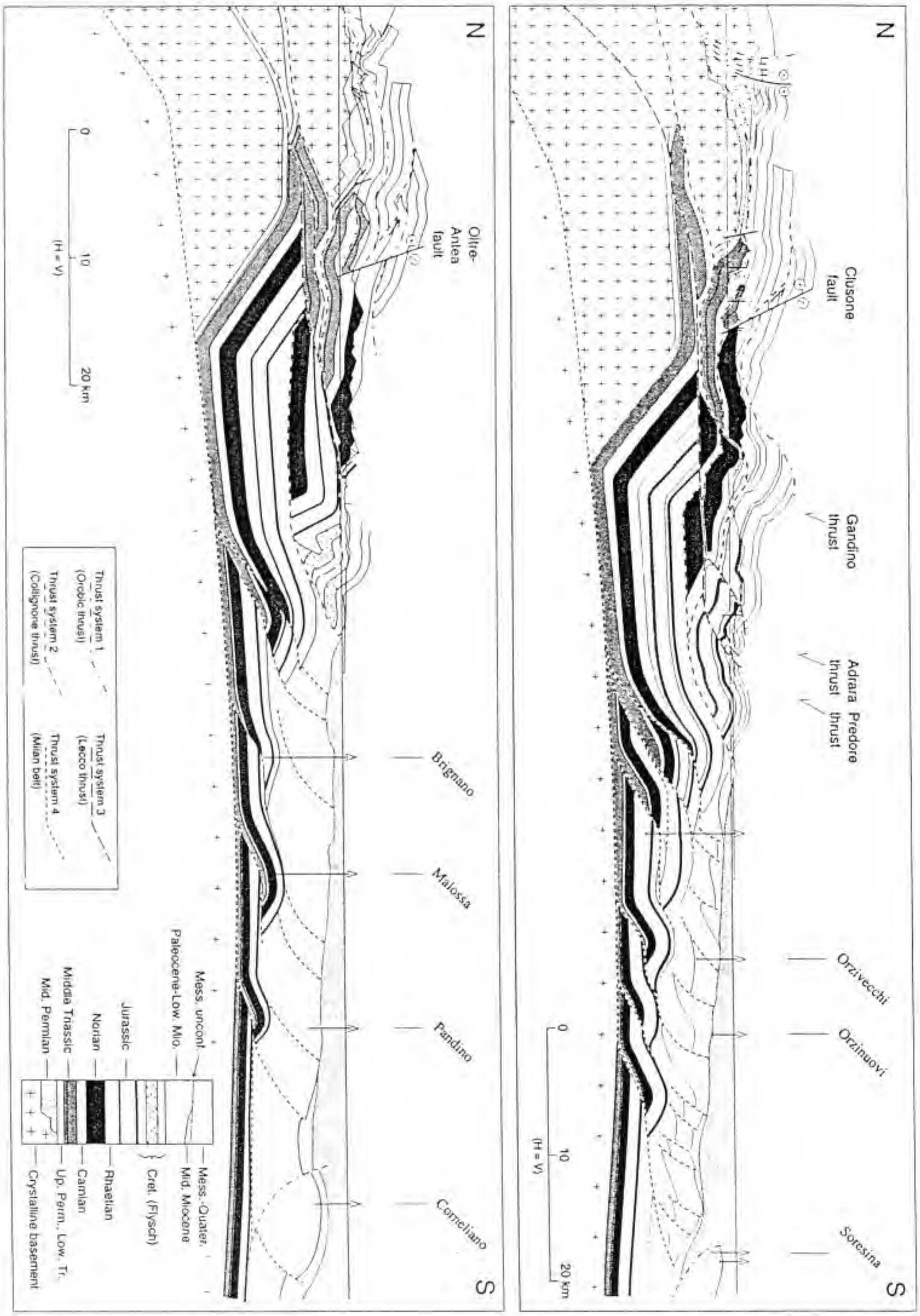

FIG. 6. Structural cross-sections through Lombardian thrust belt (modified after Schönborn, 1992). The pre-Messinian basin fill is affected by folding and thrusting. The crystalline basement is involved in the most internal units.

A Triassic to Early Jurassic rifting cycle preceded Mid-Jurassic opening of the South-Penninic oceanic basin. The North-Penninic Valais Trough subsided in response to Early Cretaceous sinistral translatory movements between Europe and Apulia which accompanied the gradual opening of the Atlantic Ocean. The Alpine orogenic cycle consists of a pre-collisional Cretaceous compressional phase, characterized by a westward directed mass transport, and a Cenozoic collisional phase, marked by northwards and southwards directed mass transport. The controlling factor is seen in the motion of the Africa-Arabia plate relative to the Eurasian plate and their interaction with the intervening Apulian microplate. During the Cenozoic collisional phase, southward directed back-folding and -thrusting, playing a significant role in the architecture of the Central-Alpine orogen, commenced upon incorporation of little attenuated northern foreland crust into the orogenic wedge.

Latest Cretaceous-Paleocene uplift of the northern Alpine foreland is thought to be the expression of compressional stresses which were projected from the orogenic wedge into the foreland, prior to its collision with the Helvetic passive margin, inducing broad lithospheric deflections. This suggests that the lithosphere of the Valais Trough had sufficient strength to permit transmission of large stresses into the European foreland.

The Central Alpine transect illustrates that post-collisional convergence, resulting in substantial lithospheric overthickening, was accompanied by thrusts propagating into the foreland crust, causing its imbrication and the uplift of basement-cored external massifs. Such external massifs, carried by thrusts soling out at mid-crustal levels, occur both on the European and the Apulian margin of the Central-Alpine orogen. Their uplift, combined with the development of the Lombardian and the Molasse-Jura nappe systems, resulted in partial destruction of flexural foreland basins which had developed during earlier phases of nappe obduction. In the context of the overall kinematics of the Alpine orogen, with the European plate dipping south beneath the orogenic wedge, the Molasse Basin evolved as a pro-wedge (in the sense of Willet et al., 1993) or a fore-arc foreland basin (in the sense of Ziegler, 1990), whereas the conjugate South-Alpine Lombardy basin evolved as a retro-wedge or "retro-arc" foreland basin (Willet et al., 1993; Ziegler, 1990, see also Doglioni, 1993). Both basins were partly destroyed by subsequent incorporation into the orogen. Thrust faults propagating into the forelands caused uplift of these basins and partial erosion of their sedimentary fill.

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Enclosure

Encl. 1. Alpine cross-section along the NFP-20-East traverse, integrating geological and geophysical data (from Schmid et al., 1996b).

The Jura fold-and-thrust belt: a kinematic model based on map-balancing

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ABSTRACT

The Neogene thin-skinned Jura fold-and-thrust belt is a crescent shaped feature which branches off from the Western Alps and almost subparallels the deformation front of the Central Alps over a distance of 300 km. Its geometry is largely preconditioned by the distribution of Triassic salts which act as a basal detachment horizon. Regional balanced cross-sections indicate that bulk shortening in the Jura orocline increases from zero at its northeastern termination to about 30-32 km at the southern termination of the Central High Jura. During the deformation of the Jura fold-and-thrust belt, the Molasse Basin, located between it and the Alps, was stably displaced northwestwards by a similar amount above a basal Triassic detachment layer.

Development of the Jura fold-and-thrust belt is probably kinematically related to the uplift of the basement involving external massifs of the Alps which acted as crustal-scale back-stops. Theoretical considerations and analogue models indicate that the initial taper of the undeformed Molasse wedge was equal to the critical taper; in contrast the initial taper of the internal parts of Jura was below the critical taper and thus became the locus of strain concentration. Availability of an effective viscous basal layer allowed for in- and out-of-sequence thrust propagation during the Jura deformation.

Northwestwards displacement of the Jura-Molasse nappe involved radial outward directed mass transport, facilitated by wrench faulting. Southwestwards increasing bulk shortening, accompanied by an apparently 10° clockwise rotation of the detached Molasse Basin, is related to Neogene differential westwards displacement of the Mont Blanc-Aiguilles Rouges Massif relative to the Aar Massif.

INTRODUCTION

The northwest verging, arcuate Jura fold-and-thrust belt is a typical arcuate mountain range which branches off from the Western Alps near Chambéry and extends over a distance of some 300 km to the North of Zürich where it dies out

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This article includes 2 enclosures on a folded sheet.

near the eastern termination of the Lägern anticline. The Jura Mountains, which form the most external part of the West-Central Alpine orogen, have a maximum width of 70 km. They are separated from the Central Alps by the Molasse Basin which corresponds to a typical flexural foreland basin developed during Oligocene and Miocene times. In the western and northern parts of the foreland of the Jura the Bresse and Rhine grabens subsided from Late Eocene to Miocene (Aquitanian). These troughs are linked by the sinistral Rhine-Saône transform zone (Fig. 1a).

The Jura Mountains are upheld by folded and thrust Late Triassic to Middle Cretaceous carbonates and shales. In synclinal areas, Oligo-Miocene clastics are preserved which were deposited in the distal parts of the Molasse Basin and a depression which linked the latter with the Rhine Graben. During the compressional deformation of the Jura and Molasse Basin, spanning Middle Miocene to recent times (the "Jura phase" proper within the Jura fold-belt was active between about 10-6 Ma, according to Laubscher, 1987; 1993; Burkhard, 1990), Middle and Late Triassic evaporites played a major role as detachment horizons between the allochthonous cover and its apparently non-involved substratum, including the Hercynian basement, Permo-Carboniferous clastics contained in troughs the evolution of which is very complex kinematically, and Early Triassic clastics (Fig. 1b).

This paper aims at developing new (in addition to some old) geometric, kinematic and dynamic arguments on Jura development, on the basis of several sets of data:

(1) field observations,
(2) subsurface data (seismic and drillholes),
(3) conventionally accepted mechanical models on fold-and-thrust belts and
(4) analog viscous-brittle models.

Moreover, we attempt to propose a kinematic analysis of the deformation of the Jura thrust belt, based on a balanced palinspastic map constructed from a series of restored regional cross-sections.

Structural Zonation of the Jura Thrust Belt

On the basis of contrasting structural styles, the Jura thrust belt can be subdivided into an internal and an external zone (Chauve et al., 1980):

The Internal or **High Jura** is characterized by large overthrusts, at least in the southern and central parts of the belt, such as the Mont Tendre thrust and the Risoux nappe (see Encl. 1, section n°4 ; Winnock, 1961; Bitterli, 1972), whereas box-folds, in which post-Triassic series are detached from the basement, locally affected by reverse faults and/or back-thrusts (Laubscher, 1965, 1977), predominates in the eastern Jura. The transition to the Molasse Basin is generally sharp where Mesozoic series appear to be little affected by compressional deformations. An exception is the southwestern most part of the Molasse Basin where Mesozoic strata crop out in ramp anticlines, such as the Mont Salève (Guellec et al., 1989, 1990a and 1990b; Wildi and Huggenberger, 1993; Deville et al., 1994).

The **External Jura** comprises four tabular plateaux which are devoid of major compressional structures. These are delimited by an array of narrow, strongly tectonized zones, corresponding to Late Eocene to Oligocene extensional structures which were reactivated during the folding of the Jura by convergent wrench and compressional movements (the so-called "pinçées" in the sense of Glangeaud, 1949; Chauve and Perriaux, 1974). The external deformation front of the Jura is characterized by relatively narrow zones of imbricate thrust sheets. Such thrust sheets override the margin of the Bresse Graben (Lienhardt, 1962; Enay, 1982; Chauve et al., 1988; Philippe, 1991) whereas upright box-folds encroach on the southern margin of the Rhine Graben.

Evolution and Tectonic History

The stratigraphic column given in Fig. 3 outlines the lithostratigraphy of the Jura Mountains and highlights regional and local detachment horizons. Tectonic stress conditions dominating the

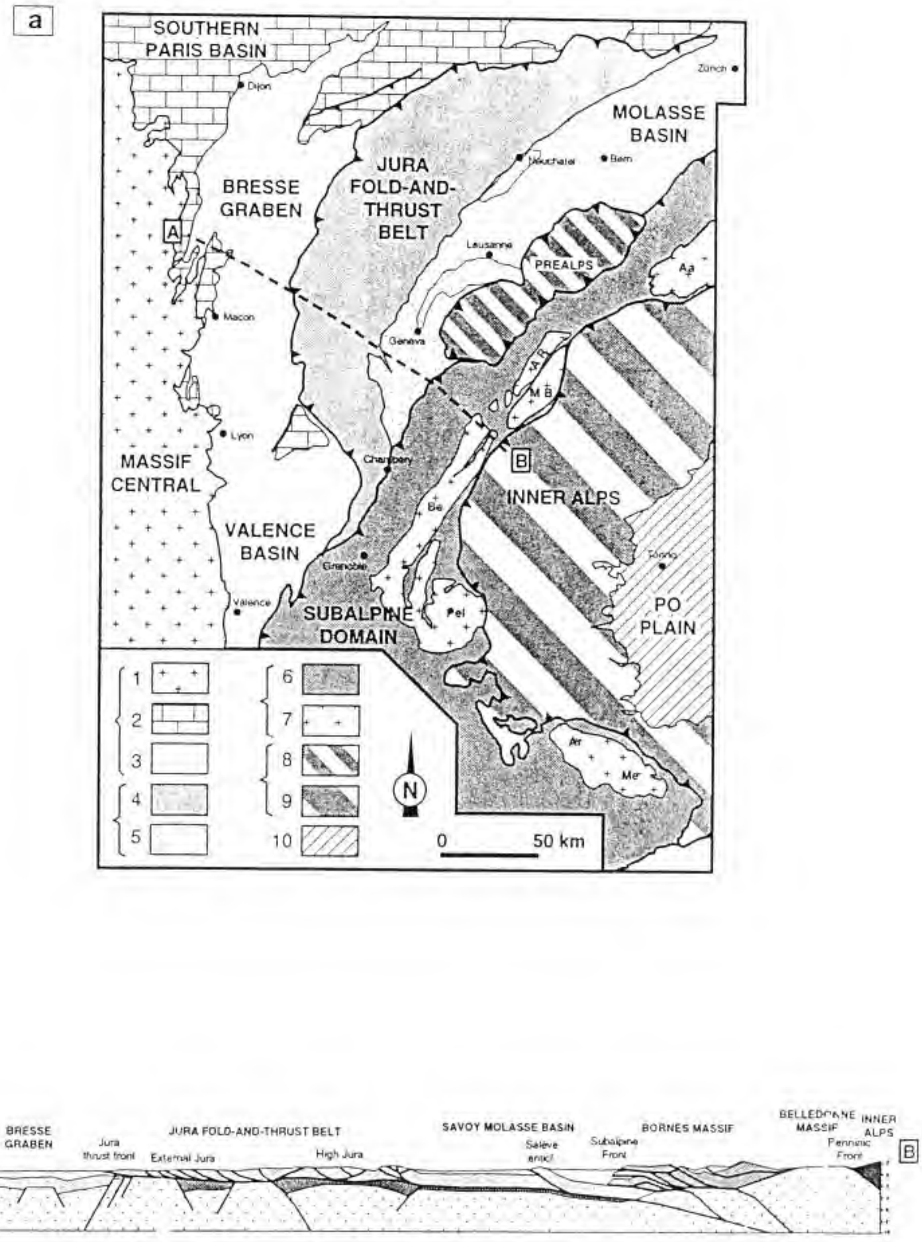


FIG. 1a. Schematic structural map of western Alps and foreland.
1-3: Stable Western European craton (1: basement, 2: Autochtonous Mesozoic sediments, 3: Tertiary sediments); 4-5: Alpine foreland (4: Mesozoic sediments of Jura thrust belt, 5: Neogene molasse); 6-7: Subalpine domain (6: Mesozoic sediments, 7: basement); 8: Internal Alps and Swiss Prealps; 9: Southern Alps.
External Crystalline massifs: Aa: Aar; AR: Aiguilles-Rouges; Be: Belledonne; MB: Mont Blanc.
FIG. 1b. Schematic cross-section through western Alps and their foreland, along the ECORS deep seismic profile (Bergerat et al., 1990; Guellec et al., 1990a; modified).

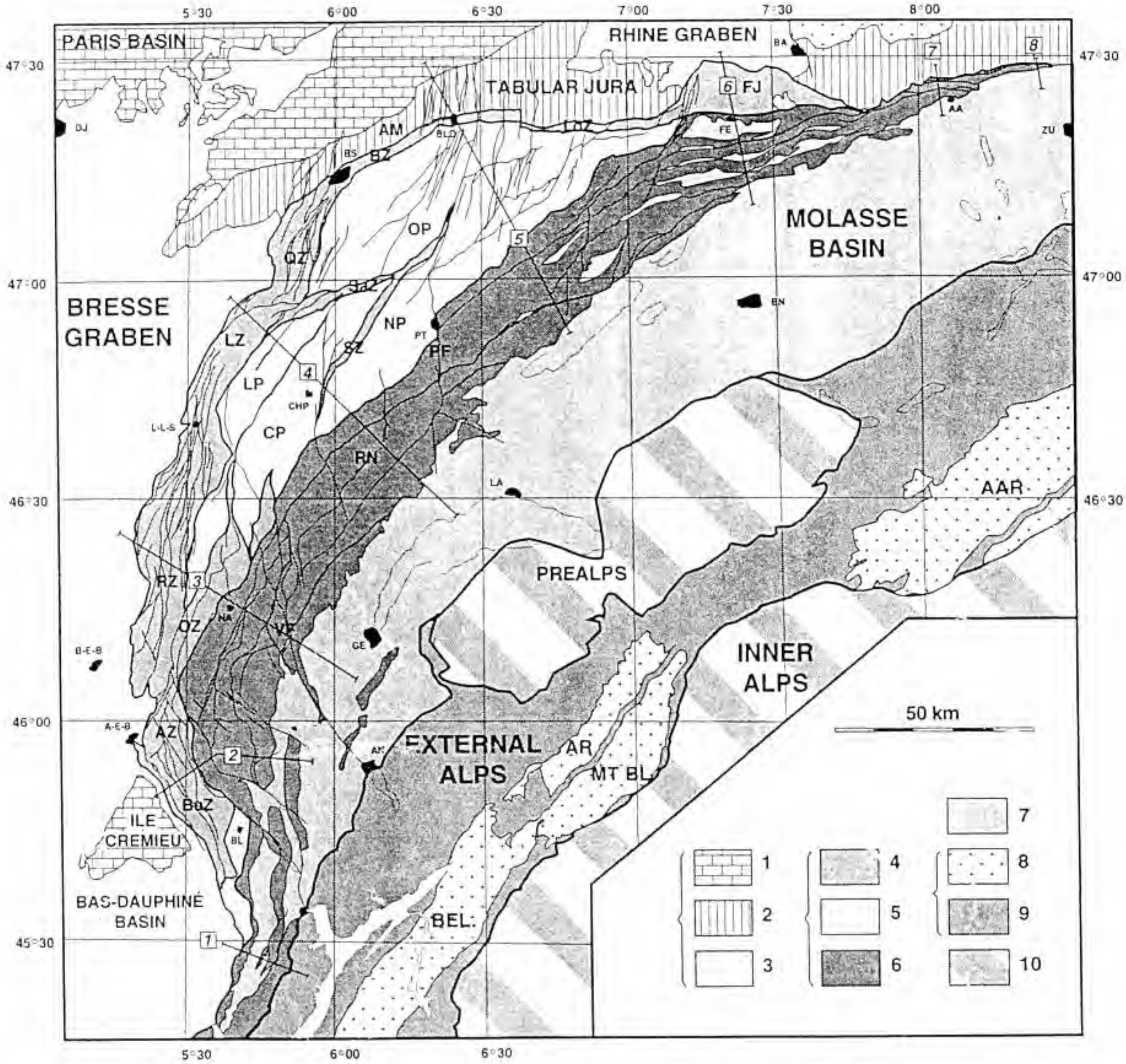


FIG. 2. Tectonic map of the Jura fold-and-thrust belt (Chauve et al., 1980, modified) showing location of the cross-sections.

1-3: Stable domain (1: Autochtonous Mesozoic cover of the Burgundy platform, 2: Para-autochthonous Mesozoic cover of the Avant-Monts zone and Tabular Jura; 3: Tertiary fill of the Bresse and Rhine grabens); 4-6: Jura fold-and-thrust belt (4: Imbricate zones ("Faisceaux"), 5: Plateaus, 6: Internal Jura); 7: Neogene deposits of the Swiss Molasse Basin; 8-9: Subalpine domain (8: Palaeozoic basement, 9: Mesozoic cover); 10: Inner Alps and Swiss Prealps.

Internal Jura: BJ: Basel Jura; RN: Risoux Nappe; VF: Vuache Fault; PF: Pontarlier-Vallorbe Fault.

Imbricate zones ("faisceaux"): AZ: Ambérieu Zone; BZ: Besançon Zone; BuZ: Bugey Zone; FJ: Ferrette Jura; LZ: Lons Zone; LoZ: Lomont Zone; OZ: Orgelet Zone; QZ: Quingey Zone; SaZ: Salins Zone; SZ: Syam Zone.

Plateaux and Tabular Jura: AM: Avant-Monts zone; AP: Ajoie Plateau; CP: Champagnole Plateau; LP: Lons Plateau; NP: Nozeroy Plateau; OP: Ornans Plateau.

External Crystalline massifs: AA: Aar; AR: Aiguilles-Rouges; BEL: Belledonne; MT BL: Mont Blanc.

Cities: AA: Aarau; A-E-B: Ambérieu-en-Bugey; AN: Annecy; BA: Basel; B-E-B: Bourg-en-Bresse; BLD: Baumes-les-Dames; BL: Belley; BN: Bern; BS: Besançon; CHB: Chambéry; CHP: Champagnole; DJ: Dijon; FE: Ferrette; GE: Geneva; LA: Lausanne; L-L-S: Lons-le-Saunier; NA: Nantua; PT: Pontarlier; ZU: Zürich.

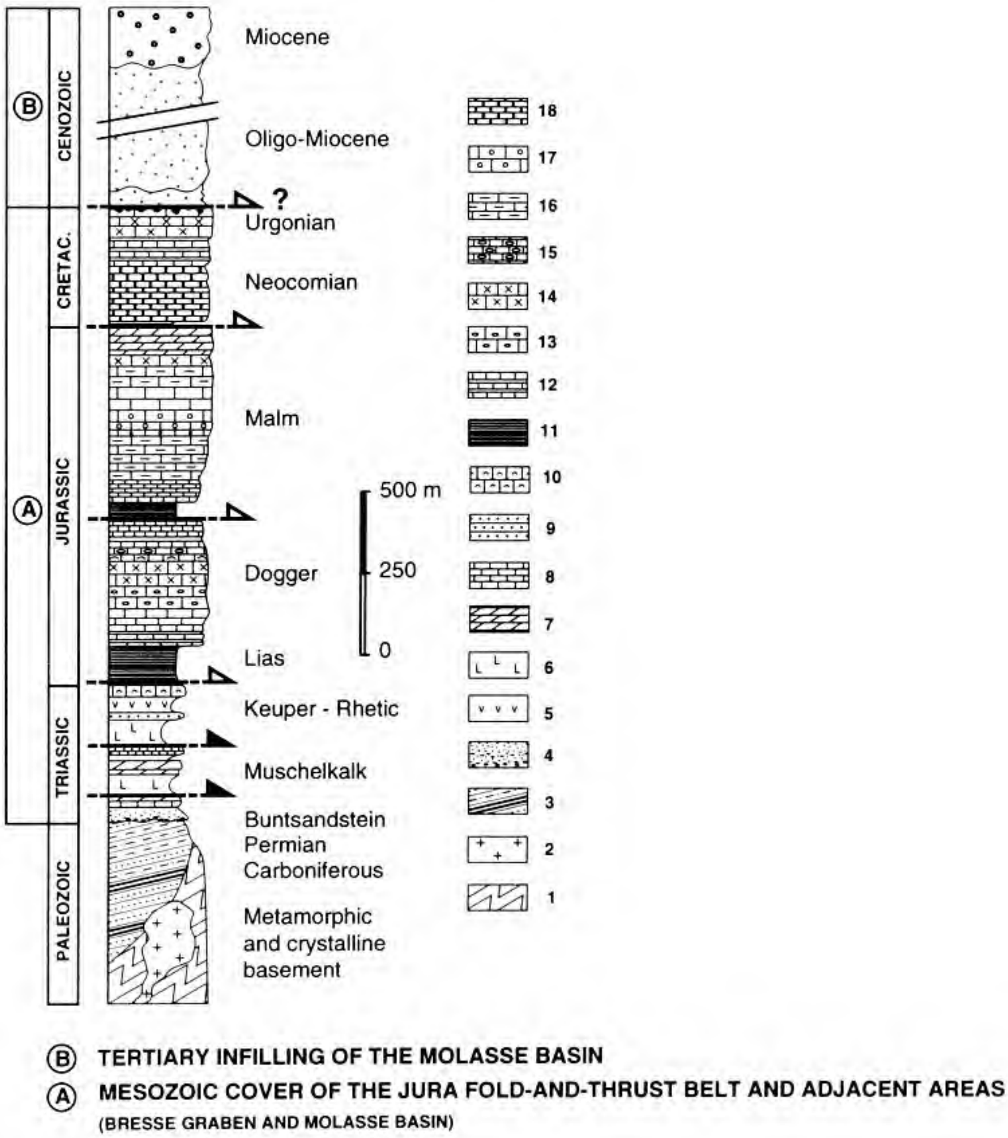


FIG. 3. Summary lithostratigraphic column of Mesozoic and Tertiary series of the Jura and the Molasse basin.
Arrows indicate potential décollement levels (black arrows show the location of the two regional sole thrusts. Eastern Jura: Middle Muschlelkalk salt; Central and Southern Jura: Early Keuper salt).
1: Metamorphic basement; 2: Granites; 3: Coal-measures and black shales; 4: Conglomerates, pebbles and sandstones; 5: Evaporites; 6: Massive salt layers; 7: Dolomites; 8: Limestones; 9: Sandstones; 10: Shelly limestones; 11: Marls; 12: Alternating marls and limestones; 13: Oncolitic limestones; 14: Reef limestones; 15: Cherty limestones; 16: Argilaceous limestones; 17: Nodular limestones; 18: Bioclastic limestones.

area of the Jura Mountains during Late Palaeozoic to Cenozoic times are summarized in Fig. 4.

The Jura Mountains and adjacent areas are underlain by a basement complex, consisting of metamorphic and intrusive rocks, which was consolidated during Carboniferous phases of the Variscan orogeny. This basement mainly crops out in the Massif Central, the Vosges and Black Forest, and in the external massifs of the Alps. In addition, it was reached by numerous boreholes drilled in the Bresse Graben, the external Jura and the Molasse Basin.

During Stephanian-Autunian times, a system of mainly transtensional, narrow and deep, fault-bounded basins subsided in which partly coal-bearing and lacustrine bituminous shales accumulated. Detailed structural analyses of Permo-Carboniferous basins of the Massif Central indicates, that they evolved under changing stress conditions, causing their partial inversion at the transition to the Late Permian (Blès et al., 1989; Ziegler, 1990).

Borehole and reflection seismic lines indicate that such basins underlay also parts of the Jura thrust belt as well as parts of the adjacent Molasse Basin (Fig. 5; Arthaud and Matte, 1977; Laubscher, 1986, 1987; Ziegler, 1990). In northern Switzerland, the main Late Paleozoic trough ("Constance-Frick basin") subparallels the eastern part of the belt. Its southern margin has been reactivated in the Early Tertiary and afterwards acted as loci for the developement of thrusts during the Jura phase (Laubscher, 1986; see Encl. 1, sections n°7 and n°8), thus controlling structural trends and the deformation style of the eastern Jura.

During the Late Permian, the area of the Jura formed part of a northeasterly trending broad depression in which essentially continental series were deposited. These conditions of continental sedimentation prevailed during the Early Triassic, but as evinced in the Swiss Jura and Central High Jura, the Buntsandstein overlie discordantly the Permo-Carboniferous troughs (see Encl. 1, section n°4).

During the Middle Triassic, marine transgressions entered this basin from the northeast as well as from the southwest, giving rise to the accumulation of the Muschelkalk carbonates and evaporites. Late Triassic regressive conditions are indicated by the deposition of the Keuper red beds and salts. During Triassic times the area of sedimentation expanded progressively into the domain of the Paris Basin and links were established with the basin of southeastern France (Debrand-Passard et al., 1984) from which marine transgressions entered the area of the Jura at the onset of the Jurassic. During Early Jurassic times, a broad, shallow marine shelf was established which extended northwards into the Paris Basin and the Northwest European Basin. To the south, this shelf was limited by the so-called Alemannic high, running obliquely across the Aar Massif, which partly separated it from the Tethys shelves. Regional isopach maps and the distribution of Middle and Late Triassic salts (Fig. 6) indicate that the area of the Jura Mountains corresponded during Triassic and Early Jurassic times to a differentially subsiding basin, referred to as the Burgundy Trough. This trough formed part of the regional Triassic-Early Jurassic Arctic-North Atlantic and Tethys rift system. Upon achievement of crustal separation in the Tethys during early Middle Jurassic times, the Burgundy Trough ceased to subside differentially. Middle and Late Jurassic carbonates and shales were deposited on a broad shelf which reached from the Helvetic Tethys margin into the Paris Basin and southern Germany. These tectonically relatively stable shelf conditions apparently persisted throughout Cretaceous times (except probably in the eastern Jura; Laubscher, 1995, pers. comm.), with basin margins being controlled by major fluctuations in relative sea-level (Ziegler, 1990).

During the Early Paleocene, the Late Cretaceous carbonate shelf was destroyed in response to the build-up of tangential compressional stresses, reflecting increasing collisional coupling between the Alpine Orogen and its foreland. Resulting broad lithospheric deformations caused regional uplift of the western Alpine foreland (including the area now occupied by the Molasse Basin and the Jura Mountains) and deep truncation of the Cretaceous and Late Jurassic sedimentary cover. During Late Eocene times the Rhine and Bresse grabens, which form part of the Cenozoic rift system of Western and Central Europe, started to subside while thrust-loaded subsidence of the Helvetic Shelf commenced. From Oligocene to Early Miocene, the evolving flexural Alpine foreland basin expanded northwards. Continued crustal extension in the Rhine and Bresse grabens was

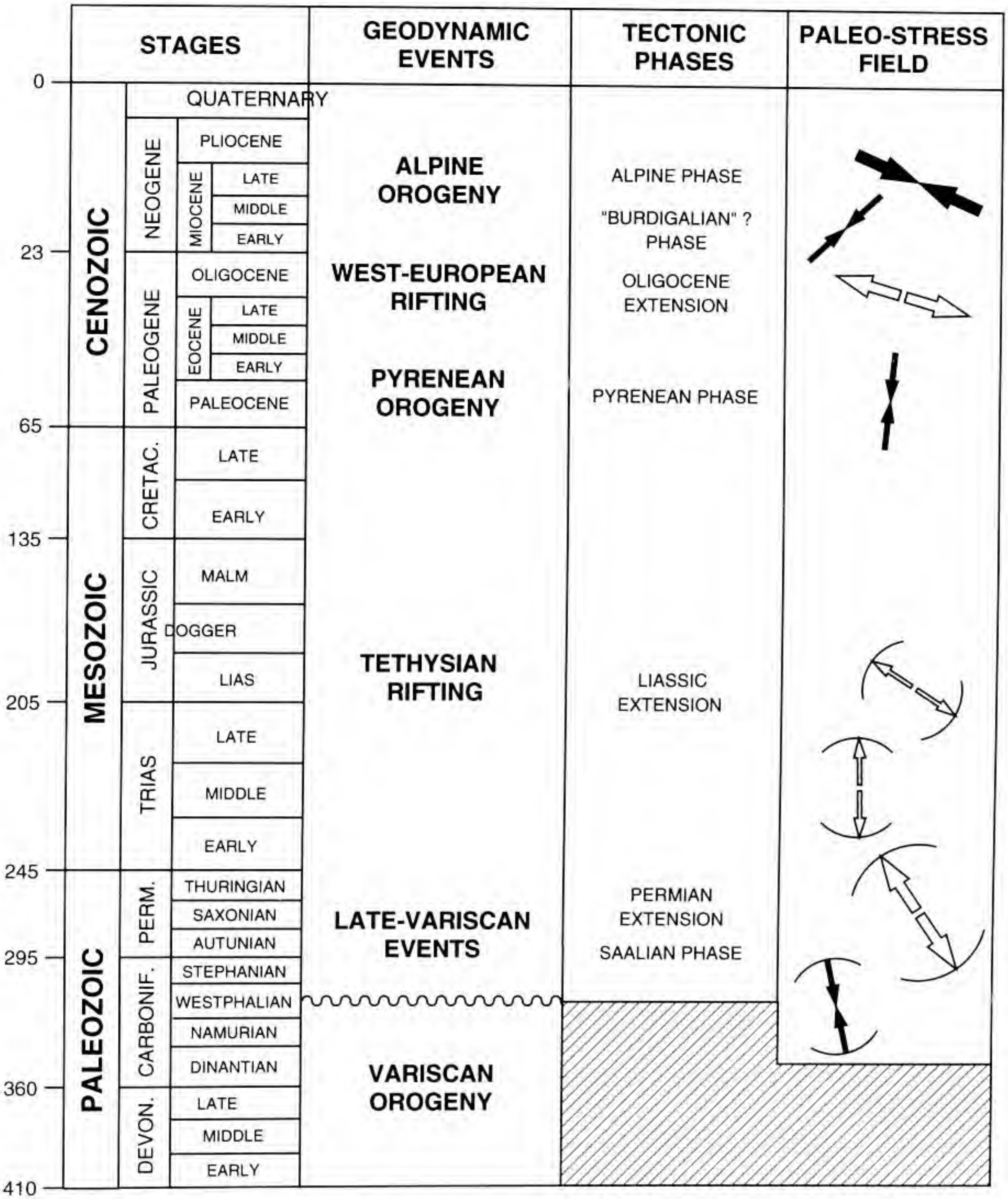


FIG. 4. Tectonic events recorded in the Jura domain.
Black arrows: maximum (s1) principal stress axis; open arrows: minimum (s3) principal stress axis.

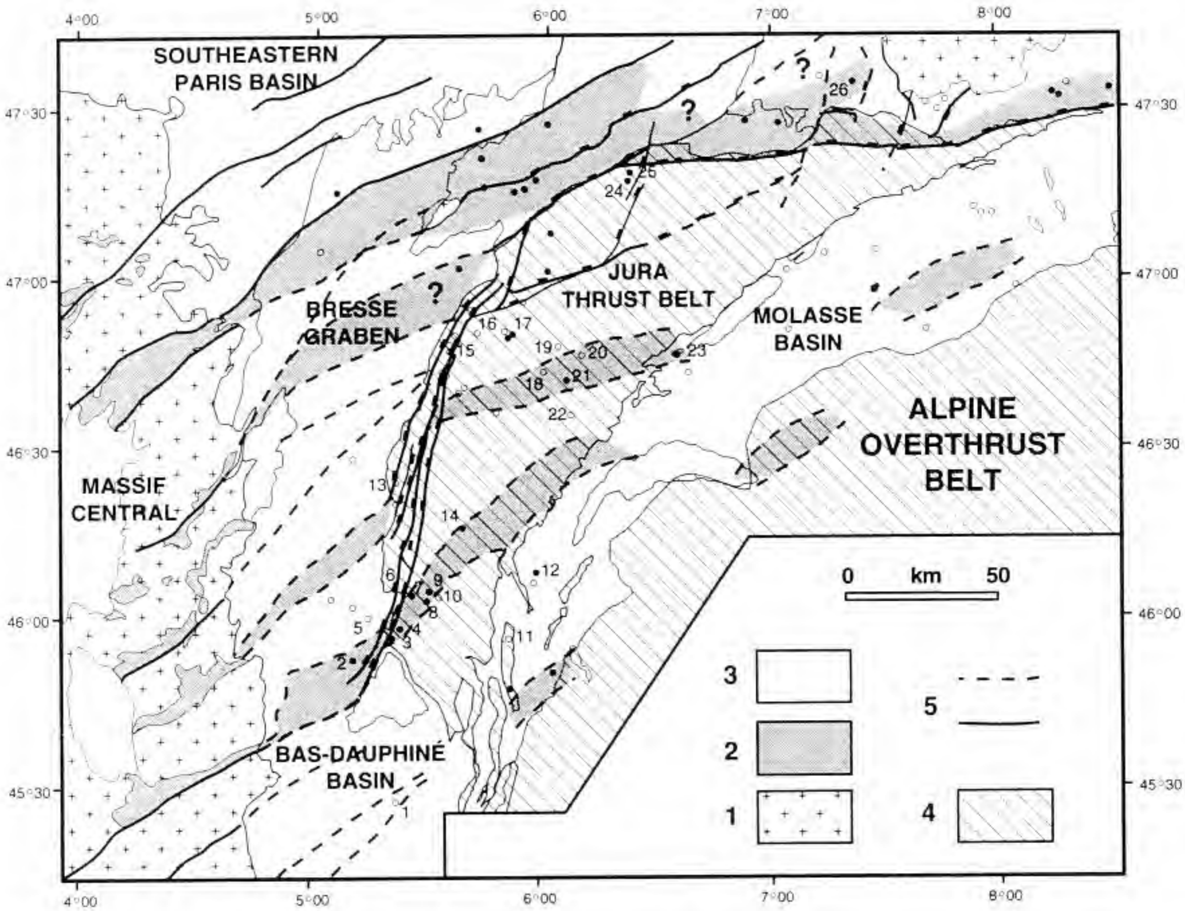
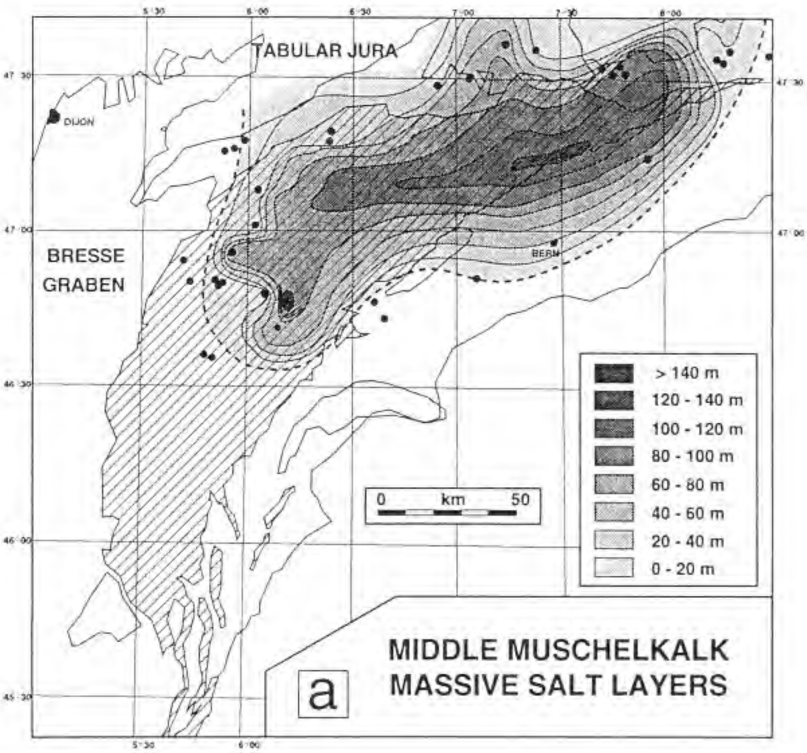


FIG. 5. Relationship between basement faults and the location of the Jura thrust front. The deep-seated basement faults in the Bresse graben are inferred from gravimetry data (BRGM, 1980; Truffert et al., 1990; modified).
1: Autochthonous Palaeozoic crystalline basement; 2 Late Variscan Permo-Carboniferous troughs (outcropping, drilled or imaged on seismic profiles); 3: Autochthonous Mesozoic cover and Tertiary sediments; 4: Jura fold-and-thrust belt and Western Alps; 5: Late Variscan deep-seated lineaments.
Boreholes: 1: Paladru; 2: Blyes; 3: Vaux-en-Bugey; 4: Torcieu; 5: Cormoz; 6: Bugey 101; 7: Bugey 102; 8: Chatillon; 9: Chaleyriat; 10: La Chandelière; 11: La Taillaz; 12: Humilly; 13: Poisoux; 14: Charmont; 15: Lons-le-Saunier; 16: Grozon; 17: Valempoulières; 18: Toillon; 19: Essavilly; 20: Laveron; 21: Chatelblanc; 22: Risoux; 23: Treycovagnes; 24: Orsans; 25: Buez; 26: Knoerringue.

accompanied by the development of the sinistral Rhine-Saône tranform zone which linked them. During the Miocene, a tectonically controlled depression developed, crossing the area of the future Jura in the prolongation of the Rhine Graben, through which communications were established between the latter and the Molasse Basin (Ziegler, 1990, 1994).

Folding of the southern Jura Mountains initiated during the Burdigalian as evident by syn-tectonic Miocene strata located in front of the forlelimb of the Gros Foug thrust-fold (Deville et al., 1994; see Encl. 1, section n°2). The main folding phase spanned Seravallian to Tortonian times (Laubscher, 1987). However, compressional deformation of this fold-and-thrust belt continued during Pliocene times and, based on geodetic data and morphologi-



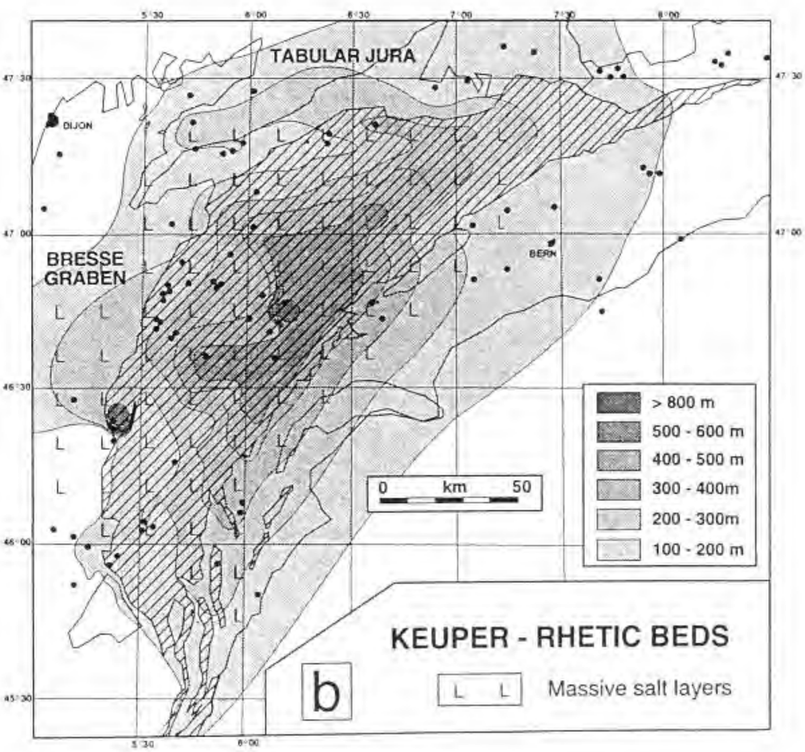


FIG. 6. Relationship between development of the Jura fold-and-thrust belt and the distribution of Triassic evaporites (thicknesses in meters).
a) Isopach map of the Muschelkalk massive salt layers.
b) Isopach map of the Upper Triassic (Keuper-Rhetic) layers (Lienhardt, 1984; modified).
Note that these maps are based only on well data, thus iso-values are tentative, especially in the eastern central part of the belt where no data are available.

cal criteria, persisted into the Recent (Fourniguet, 1978; Jouanne et al., 1995). During the development of the Jura fold-and-thrust belt, some Cenozoic extensional fault systems within the cover were compressionally reactivated, in particular the Rhinegraben faults and flexures (Laubscher, 1981), thus contributing to its present architecture. Moreover, its localization appears to have been preconditioned by the geometry of the Permo-Carboniferous half-grabens (see Encl. 1, sections n°7 and n°8; Laubscher, 1986) and the superimposed Early Mesozoic Burgundy trough in which the distribution of Triassic salts, acting as major detachment levels, played an eminent role (Fig. 6). In the latest stages of the Jura folding, post-dating thin-skin décollement, some of the Permo-Carboniferous crustal discontinuities were reactivated, causing local tectonic inversions of deep-seated Late Paleozoic troughs and related uplifts of both basement and the overlying deformed cover (see Encl. 1, section n°3; Philippe, 1994, 1995).

Geometry and Kinematics of the Jura Fold-and-Thrust Belt

As early as 1907, Buxtorf (1907, 1916) proposed a thin-skinned tectonic model for the Jura Mountains, involving detachment of the deformed Mesozoic and Cenozoic strata from the basement at the level of Triassic evaporites. Laubscher (1961, 1965) was the first to construct balanced cross-sections through this foldbelt and to quantify the bulk shortening achieved in it. Moreover, he developed the so-called "distant push" (Fernschub) hypothesis, according to which deformation of the Jura is mechanically coupled with the Alpine orogen by means of a regional sole thrust, located in Triassic evaporites. This thrust extends from the Alps through the Molasse Basin to the Jura Mountains where it splits and ramps up. As such, he considers the Molasse Basin as forming an integral part of the thin-skinned Jura allochthon. An alternate model was advanced by Ziegler (1982, 1990) who, based on reflection-seismic data from the Molasse Basin, proposed that an intra-crustal sole thrust, rooted along the northern margin of the Aar Massif extends through the Molasse Basin and ramps up into sediments along the inner margin of the Jura Mountains.

In the following we discuss the set of regional balanced structural cross-sections and their palinspastic restoration, given in Encl. 1, and focus on the relationship between the Jura allochthon and the autochthonous basement. These cross-sections are based on surface geological data and, where available, integrate reflection-seismic and well data.

Section 1: Chartreuse Massif - Bas-Dauphiné Basin

This profile, which is partly constrained by reflection-seismic and borehole data, is located to the south of the area where the Jura orocline branches off from the Alps. The thickness and composition of the deformed sedimentary sequence is similar as in the Jura Mountains. The autochthonous basement dips gently eastwards under the allochthonous Chartreuse Massif. The sedimentary fill of the Bas Dauphiné Basin has been imbricated into a consistently west verging stack of narrow thrust sheets which are detached from the autochthon at a Late Triassic or more likely at an Early Jurassic (Aalenian?) level. Triassic series are extremely reduced and devoid of evaporites and rocksalt (see Fig.6). Such a lack of basal ductile layers is responsible of the development of a typical high-tapered foldbelt achieved by imbricate foreland-verging thrusts, in good aggreement with analytic models of fold-and-thrust belts and accretionary wedges (Davis et al., 1983; Davis and Engelder, 1985). The individual thrusts ramp up through the entire Mesozoic sequence and do not employ subsidiary potential detachment levels provided by Callovian-Oxfordian and Berriasian marls.

The total amount of shortening documented in this section amounts to some 20 km.

Section 2: Southern Jura

This profile, which extends from the Savoy Molasse Basin to the autochthonous Cémieu High, crosses structural elements characterized by different transport directions; therefore, strictly speaking, it cannot be balanced.

The internal Jura is characterized by large ramp-anticlines involving the entire Jurassic sequence, such as the Gros Foug and Grand Colombier structures, which are detached from the autochthon at a Keuper salt level. Smaller folds and back-thrusts are attributed to the activation of secondary detachment horizons. Syn-tectonic Molasse series date the onset of folding as Burdigalian (Deville et al., 1994), in the eastern part of the section.

The external, thrust zone is separated from the internal zone by the broad Valromey syncline in which Neocomian strata are preserved. The change in structural style observed in the frontal imbricated zone is related to the pinch-out of Keuper salts against the Ile Crémieu High, to thinning of the Jurassic sequence due to Paleogene erosion and possibly to overprinting of extensional fault systems forming part of the Bresse Graben. In the more internal parts of this zone, upright detachment folds, evolving into pop-up structures, are cored by massif salt.

On the basis of serial cross-sections in the southern Jura (Philippe, 1995), the total amount of westwards displacement of the Jura-Molasse Basin boundary along this profile is of the order of 22 km

Section 3: ECORS profile

This cross-section is constrained by the Jura-Bresse ECORS deep reflection-seismic profile (Guellec et al., 1989, 1990a and 1990b; Damotte et al., 1990; Bergerat et al., 1990; Roure et al., 1989; Truffert et al., 1990). It crosses obliquely the sinistral Vuache-Les Bouchoux wrench zone (Blondel et al., 1988; Charollais et al., 1983; Wildi et al., 1991) which is characterized by significant Mio-Pliocene offsets of about 6 km (higher estimated value; Philippe, 1995); this impedes perfect balancing of this section.

The architecture of this sector of the Jura Mountains is characterized by a northwest verging external imbricated zone, which overrides extensional structures of the Bresse Graben, and an internal zone characterized by thrusts and back-thrusts. Folds play a very subordinate role. The internal zone of the Jura is clearly elevated with respect to its external zone and the Molasse Basin. This is quite probably the consequence of partial inversion of a Permo-Carboniferous trough during the late Jura deformation phases. The presence of such a trough is indicated by the ECORS profile and the results of the Charmont well, which bottomed in Permian red beds. As this partly inverted trough appears to be strike oblique the axes of the thin-skinned fold axes defined in the Mesozoic series, its deformation probably occurred after the main phase of the Jura deformation (Philippe, 1994, 1995).

Total shortening measured in the detached cover in this section amounts to about 32 km.

Section 4: Mont Tendre - Grozon High

This section is constrained by a number of deep wells and partly also by reflection-seismic data. It clearly illustrates the changes in structural style between the internal and external zone and the occurrence of intervening, little deformed plateaux.

There is no evidence in this section for inversion of a Permo-Carboniferous trough. The activation of Keuper evaporites as the major detachment horizon is indicated by wells Laveron-1 and Toillon-1 which penetrated more than 800 m thick salt. A second, important detachment level corresponds to Early Jurassic shales, as evident by the well Risoux-1 (see Fig. 7a). The bulk of shortening achieved in this section is accommodated by folding and major thrusting in the internal zone, characterized by the Mont Tendre and Risoux nappes, an by semi-rigid translation of the external zone which overrides the Bresse Graben margin by about 7 km near Lons-le-Saunier (Chauve et al., 1988). The allochthony of the Lons, Champagnole

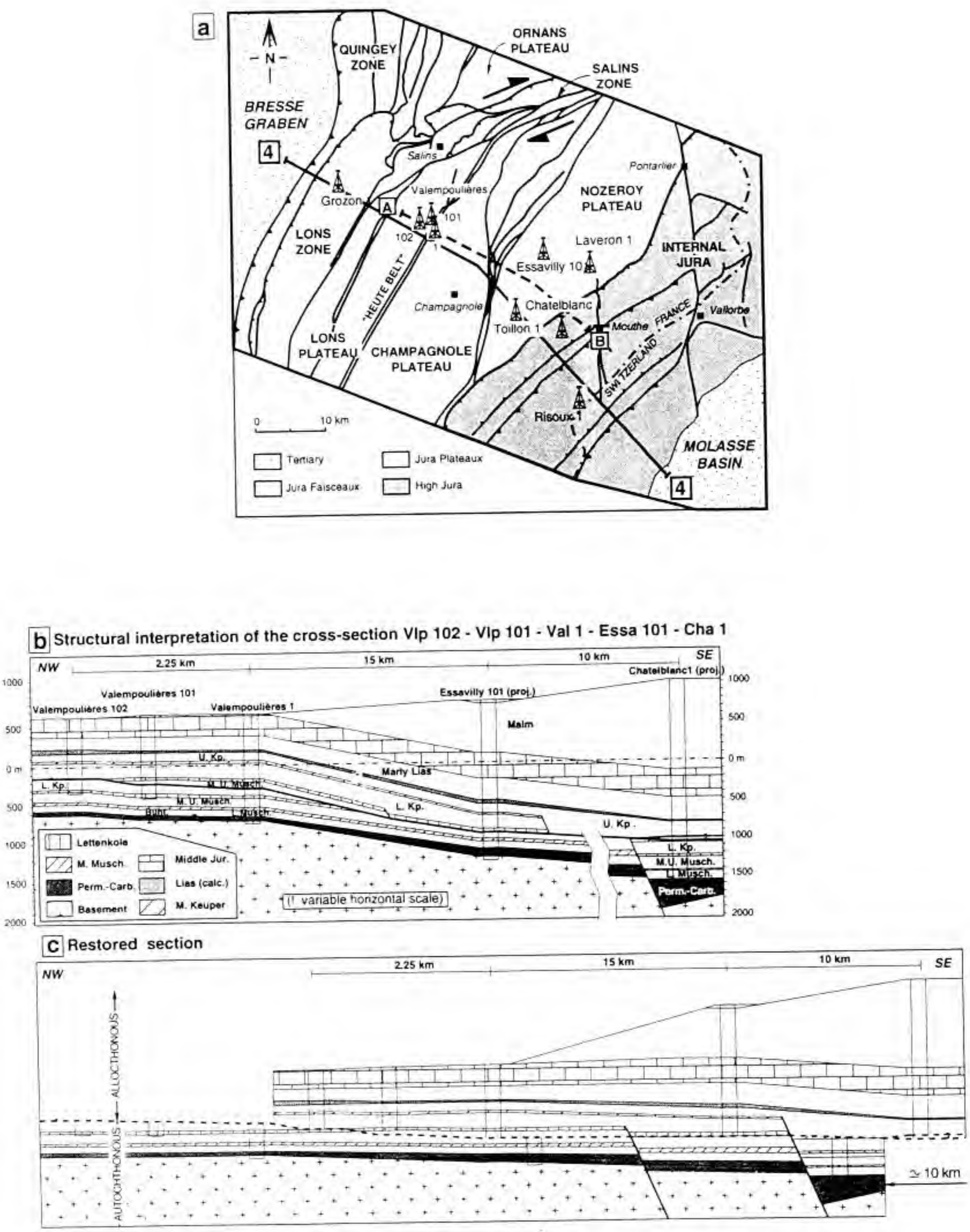


FIG. 7. **a**) Schematic structural map of the Valempoulières area.
b) Structural correlation between selected boreholes along a NW-SE transect.
c) Restoration of the previous section.

and Nozeroy plateaux, which were transported northwestwards by more than 10 km along an intra-Triassic sole thrust, without being substantially deformed, is indicated by wells drilled in the Valempoulières gas accumulation, which produce from Muschelkalk dolomites, and the wells Essavilly-101 and Chatelblanc-1 (Fig. 7b). As this flat laying thrust plane cuts along trend downwards from a middle Keuper level in the south into the Muschelkalk before rising back to a Keuper level to the north, it probably intersected a set of extensional faults which were active either during Late Triassic-Early Jurassic or possibly Eo-Oligocene times (Fig.7c).

The total amount of shortening in this section is of the order of 32 km.

Section 5: Lake Neuchâtel - Ognon fault system

Sedimentary thicknesses and the basement gradient shown in this profile are constrained by the wells Orsans-1 and Buez-1 drilled in the external Jura and the well Treycovagnes-1 located in the Molasse Basin. There are no reflection-seismic data available for this transect.

The internal zone of the Jura is characterized by a succession of major north verging faulted folds and associated back-thrusts. These are assumed to be detached from the autochthon at the level of Middle Triassic Muschelkalk salts. The geometry of these folds suggests a first phase of asymmetric detachment folding involving salt flowage in their cores, followed by a phase of fault-propagation folding. There is no evidence for major thrust sheets as seen in section 4, suggesting along-strike northeastwards gradual disappearance of the Mont Tendre and Risoux nappes. The external Jura is characterized by a large number of NNE-SSW striking sinistral wrench faults (Fig. 2); most of these faults developed during the Eocene compressional event ("Pyrenean phase") and were reactivated as normal faults during the Oligocene extension. During the deformation of the Jura, many of them were transpressionally reactivated as sinistral strike-slip faults within the detached cover as it was displaced to the northwest (Tschanz, 1990; Philippe, 1995). The Jura deformation front is marked by a set of small anticlines, striking parallel to the orocline, and a large number of NNE-SSW striking faults which form part of the Rhin-Saône transform zone. Basement faults delineating possible Permo-Carboniferous troughs, were apparently reactivated during Oligocene times and thus provided for discontinuities in the Triassic detachment levels. As such they guided the location of the frontal Besançon and Ognon zones.

Cumulative shortening in this cross-section is about 21 km.

Section 6: Grenchen Anticline - Rhine Graben

This section is essentially based on the one published by Buxtorf (1916) and an industry-type seismic profile running across the "Ferrette Jura" and the southern part of the Rhine Graben. Both Keuper an Muschelkalk are involved in the fold and thrust structures of the Jura up to its northern deformation front where Paleogene sediments of the Rhine Graben are affected by folding, as seen in the Ferrette anticline. Muschelkalk salts acted as the major decoupling horizon between the autochthonous substratum and the allochthonous Jura.

This transect is characterized by a succession of box-folds, resulting from a complex interaction of faulting and buckling induced by the presence of an efficient basal ductile layer, controlling the mechanical behaviour of the overlying brittle strata, which, in turn, are interrupted by secondary incompetent layers, such as the Oxfordian clays; locally these account for disharmonic folding. A clear distinction between an internal and an external zone is difficult. The broad Delémont syncline can be considered as an internal plateau.

The total amount of shortening in this section is of the order of 12 km.

Section 7: Aarau - Tabular Jura

This section extends from the Tabular Jura, which forms the sedimentary cover of the Black

Forest, across the narrow internal zone of the Jura which consists of two main thrust sheets; these have been pierced by a road tunnel. Reflection-seismic and well data indicate that the southern margin of the Jura Mountains closely coincides with the southern border fault of a major Permo-Carboniferous trough. Permo-Carboniferous faults were probably reactivated during the development of the Rhine Graben to the degree that they disrupted the Triassic detachment levels and thus nucleated north-verging thrusts during the Neogene deformation of the Jura (Laubscher, 1986), However, the offset on such faults was apparently insufficient to totally clamp down northwards propagation of the Jura deformation front. Northwards the basal sole thrust, which regionally is located within the Muschelkalk, rises into Keuper and Lower Dogger "opalinus-shales" (Bitterli, 1990; Jordan et al., 1990; Jordan and Noack, 1992).

The minimum estimated shortening in this section is about 6.5 km.

Section 8: Lägern anticline

The thrusted Lägern anticline is the easternmost feature of the Jura orocline. Based on reflection-seismic data (Müller et al., 1984), it corresponds in this section to a typical detachment lift-off which is slightly overturned to the North (Mühlberg, 1894). A basal heterogeneity, offsetting the detachment level, is provided by a set of deep-seated normal faults located beneath the Lägern structure, coinciding to the southern border of a Permo-Carboniferous trough.

The Lägern anticline accounts for about 2 km of shortening; as it dies out to the east this value decreases to zero.

Based on the cross-sections discussed above, which are in part constrained by wells and seismic data, we fully adhere in principle to the thin-skinned distant push model proposed by Laubscher (1961, 1965) for the development of the Jura fold and thrust belt. In order to explain its position in the Alpine foreland, it is essential to analyze the areal distribution of Triassic salts which permitted detachment of the Mesozoic and younger strata from their basement. Isopach maps of Middle Triassic Muschelkalk and the Late Triassic Keuper salts, constructed on the basis of well data, show a close coincidence with the geometry of the Jura orocline, especially in the south of the belt where the left-lateral transfer zone achieving the southern termination of the Jura thrust-belt is surimposed on the limit of Upper Triassic evaporites (Fig. 6). In the central and southwestern parts of the Jura, the main decoupling level is located at the base of the Keuper salts whereas in its northern and eastern parts middle Muschelkalk salts acts as the principal detachment horizon (Laubscher, 1961, 1986; Blondel et al., 1988; Guellec et al., 1990a and 1990b; Jordan; 1992; Philippe, 1994). Under the Molasse Basin, both salt layers decrease rapidly in thickness towards the south and disappear entirely near the Alpine deformation front.

The configuration of the external deformation front of the Jura fold- and thrust belt is highly variable. Whereas the margin of the Bresse graben is overthrusted to various degrees, the external elements of the northern and northeastern parts of the Jura bump against the undeformed foreland, corresponding to the Tabular Jura (Fig. 2). The Ognon and Lomont fault zones appears to coincide with pre-existing normal fault systems which were partly inherited from Permo-Carboniferous times and were tensionally reactivated during the Paleogene subsidence of the Rhine-Bresse system of grabens. These fault systems played an important role in the nucleation of ramping-up thrusts. The southwestern swing-back and termination of the Jura fold-and-thrust belt was apparently preconditioned by the pinch-out of Keuper salts against the Ile Crémieu High. Similarly its northeastern termination is probably related to thinning of the Muschelkalk and Keuper evaporites below a critical thickness.

Theoretical and Analogue Model Considerations

Palinspastically restored cross-sections through the Jura fold-and-thrust belt and the Molasse Basin indicate that, prior to their deformation, they had the geometry of a northwestwards tapering wedge which was underlain by a relatively smooth basement.

A possible explanation of both the interposition of the Molasse Basin between the Western Alps and the Jura foldbelt and contrasting structural styles of High Jura and Plateau zones can be provided by results of conventional models of accretionary wedges and fold-and-thrust belts.

Hubbert and Rubey (1959) showed that the maximum length of a rectangular body of rocks, undergoing rigid displacement on a horizontal plane, depends on the pore pressure at its base. From this follows that a thrust sheet, exceeding this critical length, will be internally strained according to its rheological properties. The critical taper model of Davies et al. (1983) stipulates that such internal deformation results in the development of an, in cross-section triangular, accretionary fold-and-thrust belt which continues to thicken until a critical taper is attained. Depending on the rheology and/or pore pressure at the base of the undeformed foreland sedimentary wedge, the latter can be detached from the basement and slides forward together with the internally deformed foldbelt (Fig. 8). In such a model the internal Jura corresponds to an accretionary foldbelt while its external zone represents the internally little deformed foreland wedge, the length of which depends, according to Hubbert and Rubey (1959), on the rheology of the detachment level. As at depths greater than 2 km and under geothermal gradients of 20 to 30°C/km the shear strength of salt is smaller than 1 MPa (Carter and Hansen, 1983), it is plausible that bulk detachment of the external Jura occurred once the internal Jura fold-and-thrust belt had attained a critical taper (topographic relief). Yet, as the external Jura has undergone considerable internal deformation, concentrated on pre-existing discontinuities, it cannot be regarded as a completely homogeneous rigid body, as assumed in the model of Hubbert and Rubey (1959).

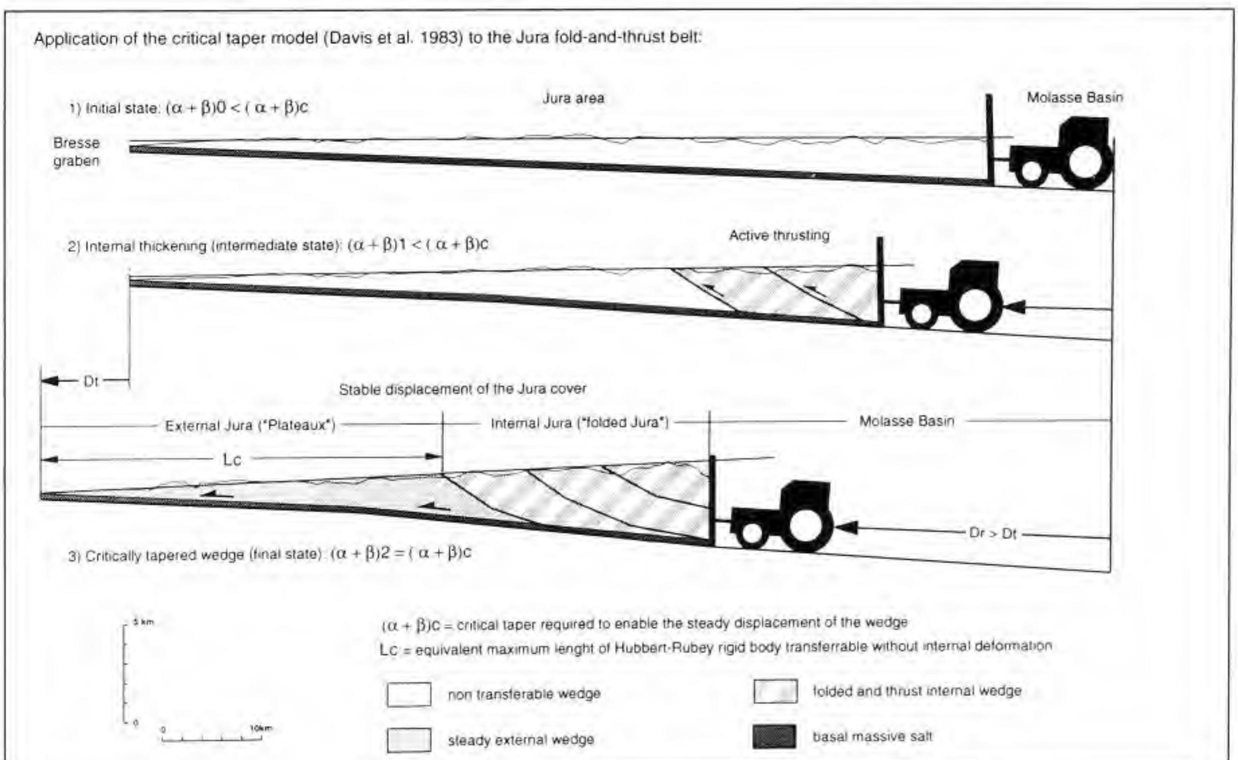


FIG. 8. Critical taper model of Davis et al. (1983) applied to the Jura fold-and-thrust belt.

However, as the Jura fold-and-thrust belt is separated from the Alps by the apparently little to undeformed Molasse Basin, the problem of stress transmission from the external Alpine massifs (Aar, Mont Blanc, Aiguilles-Rouges, Belledonne) to the Jura Mountains requires special attention (Laubscher, 1961, 1972; Mugnier, 1984; Mugnier and Vialon, 1986). To this end, a multi-layered analogue model was constructed in which limestones and sandstones are represented by sand and glass powder and rocksalt by silicon putty. The thickness of the basal silicon putty layer was varied according to the isopach of the Triassic salts. An essentially uniform thickness was assumed for the Mesozoic carbonates and shales. The Tertiary fill of the Molasse Basin was simulated by a left tapering wedge, pinching out near the middle of the model. The domain of the Jura Mtns. is represented by the left side of the model. This model was deformed at low strain rates (1-2 mm/h) by displacing a vertical back-stop from the right to the left, simulating the horizontal push of the external Alpine massifs. At several stages during the deformation of this model, transverse cross-sections were imaged by means of X-ray tomography (see Colletta et al., 1991).

Figure 9 shows the evolution in a vertical cross-section of this model and demonstrates that deformations are entirely concentrated on its left part, simulating the Jura thrust belt, whereas the right part, corresponding to the Molasse Basin, is moving to the left without undergoing internal deformation. It is also evident that deformations are initially concentrated near the thin end of the Molasse wedge but do not develop strictly in sequence. Unlike brittle models, which are characterized by in-sequence thrust propagation, models involving ductile detachment levels show a disorderly thrust sequence of thrust propagation and are characterized by the development of pop-up structures carried by fore- and back-thrusts (Ballard et al., 1987; Colletta et al., 1994). During the late stage of the model, deformation migrated towards its left boundary and stopped at the limit of the silicon layer representing Triassic salts; at the same time the more internal structures continued to grow. The final stage of the model bears considerable similarities with the structural style of the Central Jura.

From this analogue model we conclude that the combination of a tapering Molasse Basin wedge and the presence of a basal viscous layer is responsible for the stress transfer to the Jura Mountains. Whereas the initial taper of the undeformed Molasse wedge was equal to the critical taper, the initial taper of the internal parts of the Jura was below the critical taper. With progressive deformation of the internal Jura, the belt reached, in cross-section, a critically tapered prismatic shape and consequently its external parts (i.e. the so-called Plateaux) were detached from their substratum and transferred to the northwest over 10 km (see Encl. 1, section n°5).

3-Dimensional Palinspastic Restoration

3D restoration of oroclinal foldbelts has to contend with inherent space problems and requires detailed knowledge about the direction of mass transport in space and time.

Microtectonic analyses indicate a radially diverging mass transport (Fig. 10; Plessmann, 1972; Meier, 1984; Tschanz, 1990; Philippe, 1994, 1995) which is consistent with the trend of stress trajectories inferred from major fold axes (Laubscher, 1972). Palaeomagnetic data do not reveal noteworthy rotations (Johnson et al., 1984; Elderge et al., 1985; Gehring et al., 1991). Based on this, we conclude that the finite displacement map, derived from this data, reflects the centrifugal translatory motions of the different thrust elements making up the Jura fold-and-thrust belt, and that sinistral radial faults account for divergent compression in the external parts of this orocline. Lateral expulsion of Mesozoic series in the southern and northern Jura arc argues in favour of their decoupling from the autochthonous substratum. This is further supported by in situ stress measurements which demonstrate that the orientation of the principal horizontal compressional stress axes observed in the detached Mesozoic series differs from the one recorded on the underlaying basement (Becker et al., 1987; Becker, 1989). Although the present stress field of Western Europe is dominated by NW directed trajectories of maximum horizontal compression (Müller et al., 1992; Rebaï

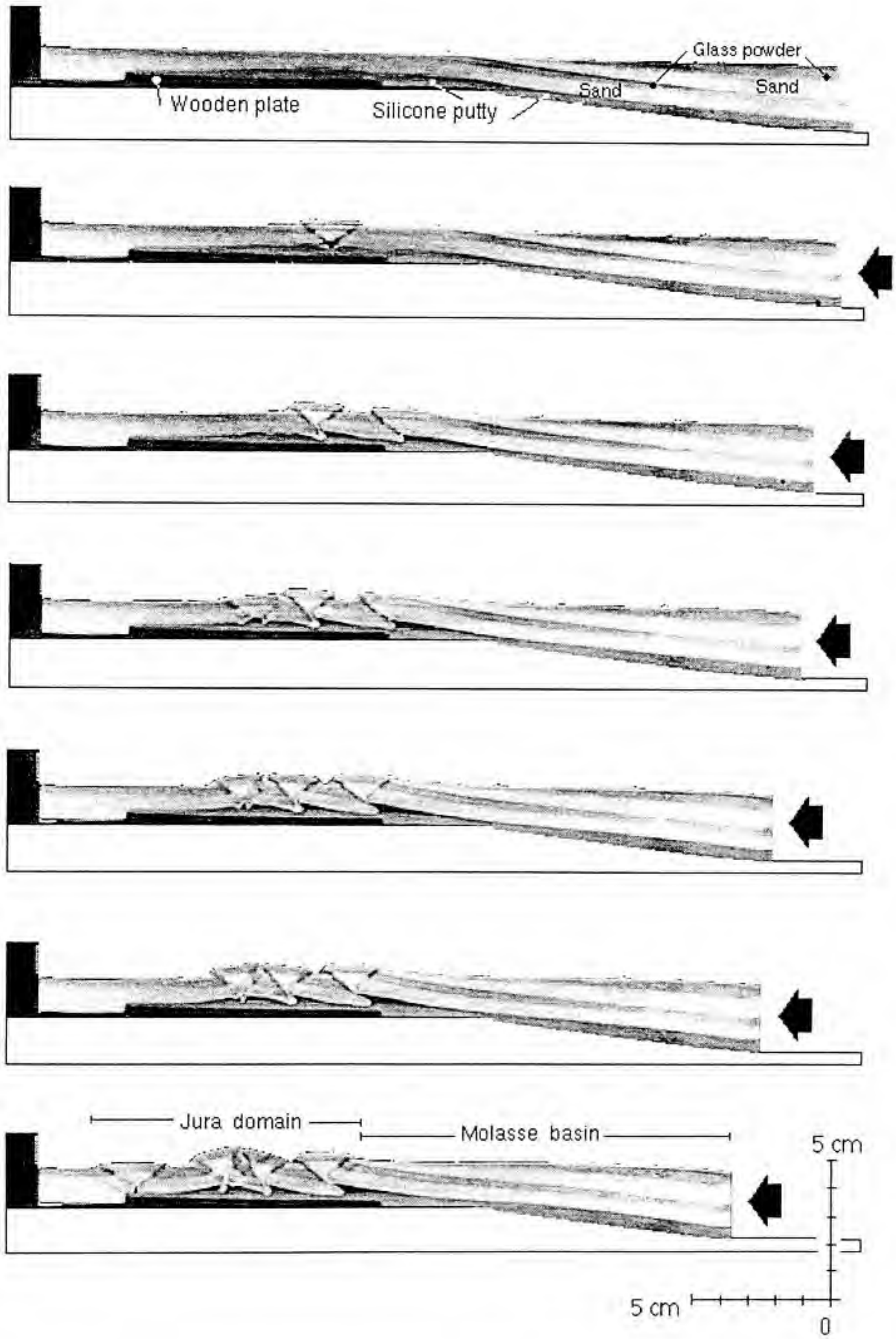


FIG. 9. Evolutionary cross-section of an analogue model simulating the horizontal push exerted by a backstop on the Alpine foreland (see text for explanation).

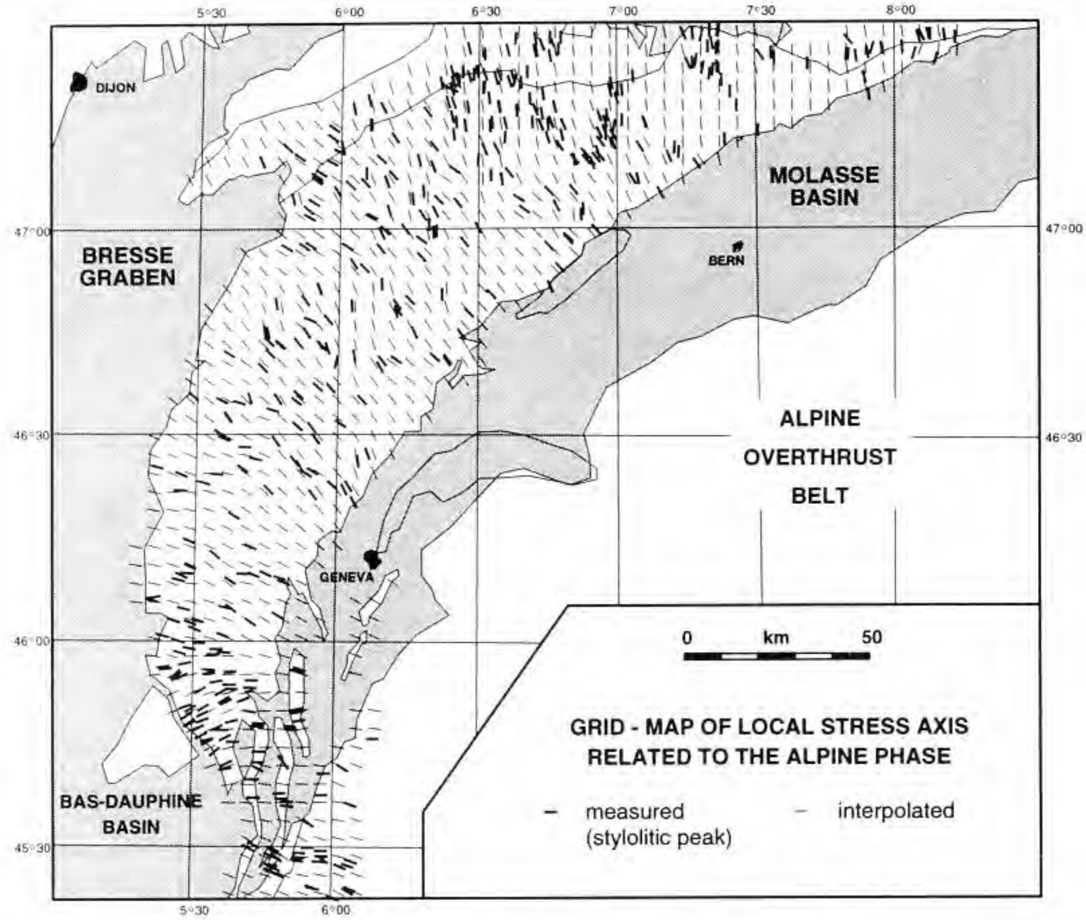


FIG. 10. Grid-map of the Mio-Piocene principal stress axes in the Jura fold-and-thrust belt based on stylolitics peaks (compiled after Plessman, 1972; Meier, 1984; Tschanz, 1990; Philippe, 1994; 1995).

et al., 1992), vertical strain-partitioning in the detached sediments of the northern Jura can deviate as much as 50° from this regional stress field.

We have applied the map balancing method developed by Laubscher (1965) and refine by Baby et al. (1993). In a first step the study area was subdivided into structural units delimited by frontal and rearward thrust faults and lateral strike slip faults (Fig. 11a). In a second step, the amount of shortening achieved within and between the different units (related to cumulative internal foldings and displacements along thrusts and back-thrusts) was derived from restored cross-sections (see Encls. 1 and 2); determined values may have an error margin of up to 25%, depending on the restoration method used. In the third step all units were retrodeformed and moved back to their pre-deformation position along transport trajectories indicated by micro- and macrotectonics

Iterative steps in retro-deformation of the Jura orocline show that, although radial forwards displacement of the different units played a dominant role, additional transpressional deformations along major wrench faults (from NW to SE: Caquerelle, Pontarlier-Valorbe, Vuache-Les Bouchoux and Culoz faults), resulting in small-scale rotations and/or shear deformation, must be invoked in order to explain the kinematic evolution of the arc. In view of the importance of these wrench faults, units bounded by them were linked into blocks as shown in Figure 11b. The transport trajectories implied by this final palinspastic model corre-

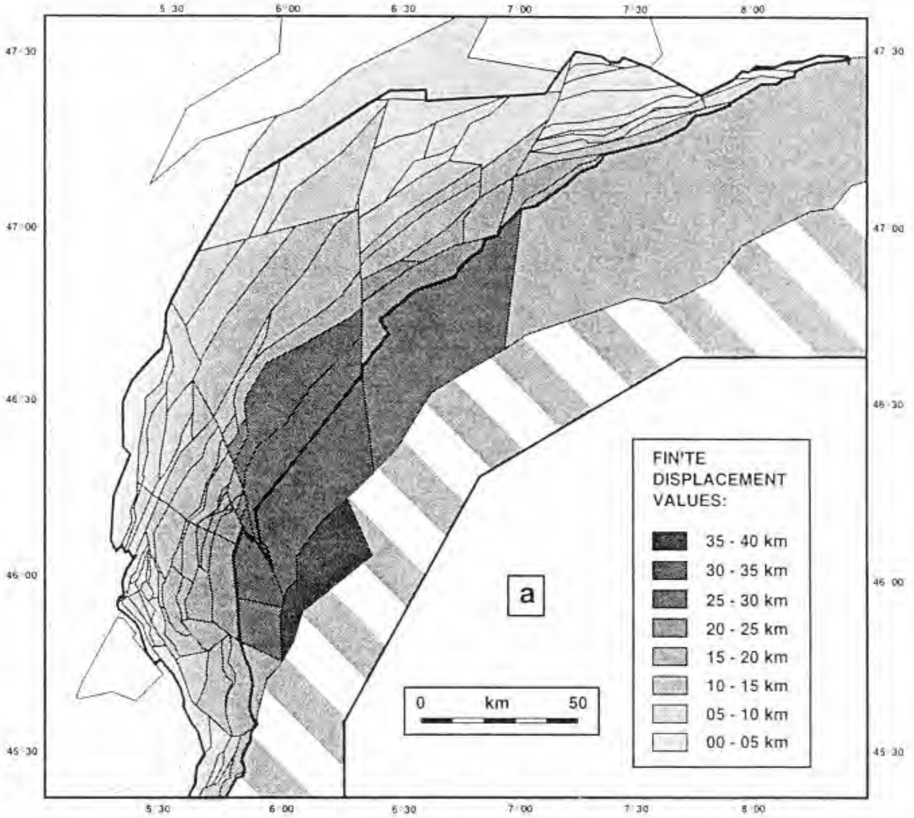


FIG. 11a. Structural map of the Jura fold-and-thrust belt divided in surface units. Finite displacement values are calculated from the Fig. 11b.

sponds closely with those derived from micro- and macrotectonic analyses (compare Fig. 11c and Fig. 12).

We conclude that radial, outwards diverging transport directions, facilitated by detachment of the Mesozoic and Cenozoic sediments from their basement and the early activation of a system of wrench faults facilitated the development of the Jura orocline. Some of these faults may correspond to pre-existing features (e.g. Vuache-Les Bouchoux fault; Charollais et al., 1983; Blondel et al., 1988; Rhine graben faults; Laubscher, 1981) whereas others (e.g. Pontarlier-Vallorbe fault; Laubscher, 1961) may have initiated during the early phases of the Jura deformation as their angular relationship to the strike of folds conforms closely to the Mohr requirement for lateral extension in a divergent system (Laubscher, 1972).

Our 3D palinspastic restoration of the Jura arc implies a 10° clockwise rotation of the Jura and Molasse sedimentary prism around a pivot located near the eastern termination of the Lägern anticline (this was advocated already by Laubscher in 1965 who estimated the amount of rotation to about 7°). However, no longitudinal extension required south of this rotation axis is far from being demonstrated. Therefore, such a rigid clockwise rotation of the detached sediments of the Plateau Molasse is certainly in fact accommodated by significant wrench faulting and/or homogeneous dextral shearing deformation, as not illustrated by the proposed final palinspastic map (for further discussions, see Burkhard, 1990). A large number of (strike-slip?) faults seems to cross-cut the Molasse Basin, according to published tectonic maps (e.g. Matter et al., 1980; Jordan, 1992); Even so, their precise signification and implication in décollement tectonics of the Jura-Molasse nappe remains under considerations.

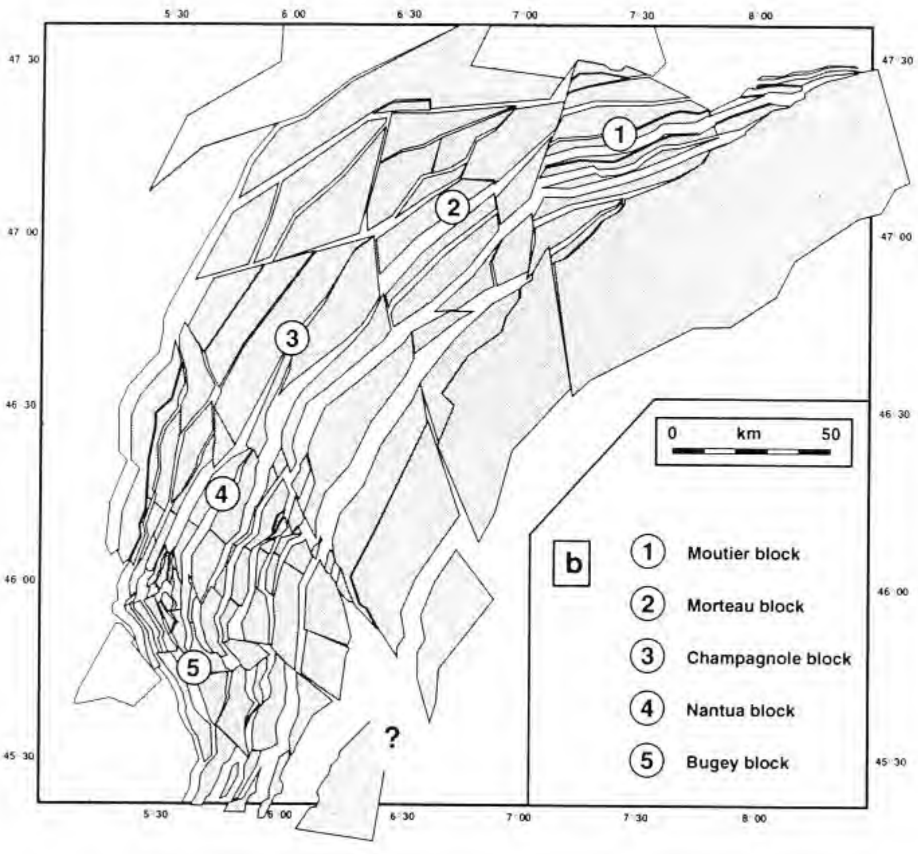


FIG. 11b. Final restored map.

Geodynamic Implications and Conclusions

Compressional stresses exerted on the Alpine foreland resulted in detachment of the Molasse Basin from its basement. Acting as a relatively rigid unit, its northwestwards displacement caused deformation of the sediments covering the Jura domain and uplift of the Molasse Basin (Laubscher, 1961; Müller and Briegel, 1980; Müller and Hsü, 1980; Trümpy, 1980). Lateral changes in bulk shortening achieved in the Jura orocline indicate a southwestwards increase in displacement of the Molasse Basin reaching a maximum of about 35 km south of Geneva where the Molasse Basin is internally deformed, as indicated by the Salève and Montagne d'Age ramp anticlines (Guellec et al., 1989, 1990a; Wildi and Huggenberger, 1993; Deville et al., 1994). In contrast, to the east of the Lägern, the structural style of the Molasse Basin is characterized by an orogen-ward dipping monocline and a well developed thrusted subalpine triangle zone (Bachmann et al., 1982; Burkhard, 1990; Stäuble and Pfiffner, 1991).

Large-scale clockwise rotational displacement of the Molasse Basin is assumed to be kinematically related to the coeval uplift of the basement-involving ramp anticlines forming the external Alpine Aar, Belledonne and Mont Blanc-Aiguilles Rouges massifs. These acted as crustal-scale mobile back-stops during the deformation of the Jura nappe (Ménard, 1977, 1988; Doudoux et al., 1982; Roure et al., 1989; Guellec et al., 1990a). Neogene differential westwards displacement of the Belledonne-Mont Blanc-Aiguilles Rouges block relative to the Aar Massif along the dextral Simplon-Rhône shear zone, located in front of the intra-Alpine Ivrea-Insubric-Periadriatic lineament (Fig. 12), accounts for the observed strain increase in the central and southwestern parts of the Jura orocline. A schematic interpretative model about

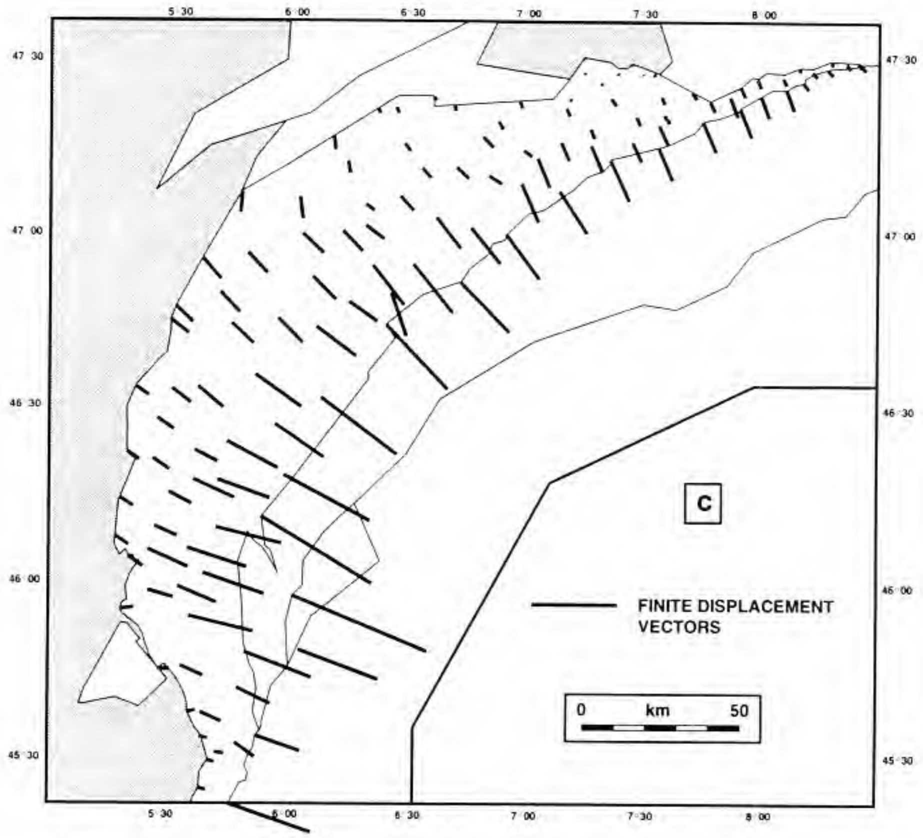


FIG. 11c. Finite displacement vectors derived from Figure 11b.

the geodynamic evolution of western Alps could be summarized as following:

(1) northwards motion and counterclockwise rotation of about 34° of the Apulian promontory,
(2) westwards thrusting anf counterclockwise rotation of about 25° of the external Alpine massifs (lateral expulsion in front of the Apulian promontory),
(3) right lateral movements along the Simplon-Rhône Line between Aar-Gothard and Mt. Blanc-Belledonne massifs. It immediately follows
 . a clockwise rotation af about 10° of the Plateau Molasse,
 . indentation, shortening and displacement of the thin Jura cover towards the northwest, and
 . centrifugual thrusting in the external parts of the Jura thrust belt accommodated by the onset of radial left-lateral strike-slip faults (strain-partitioning).

The set of surface and subsurface data at our disposal upon which we based the interpretations lead to the following outcomes:

(1) The development Jura fold-and thrust belt is governed by the distribution of Triassic evaporites, particulary rocksalt; the major transfer zone accommodating the southern termination of the belt is superimposed on the southwestern limit of the Keuper evaporites. All direct observational data show that the oldest rocks exposed in the Jura and implied by folds and thrusts are Triassic evaporites. This is also supported by distinct geometric analysis of some folds for which the calculated depth of detachment is in any case closely connect-

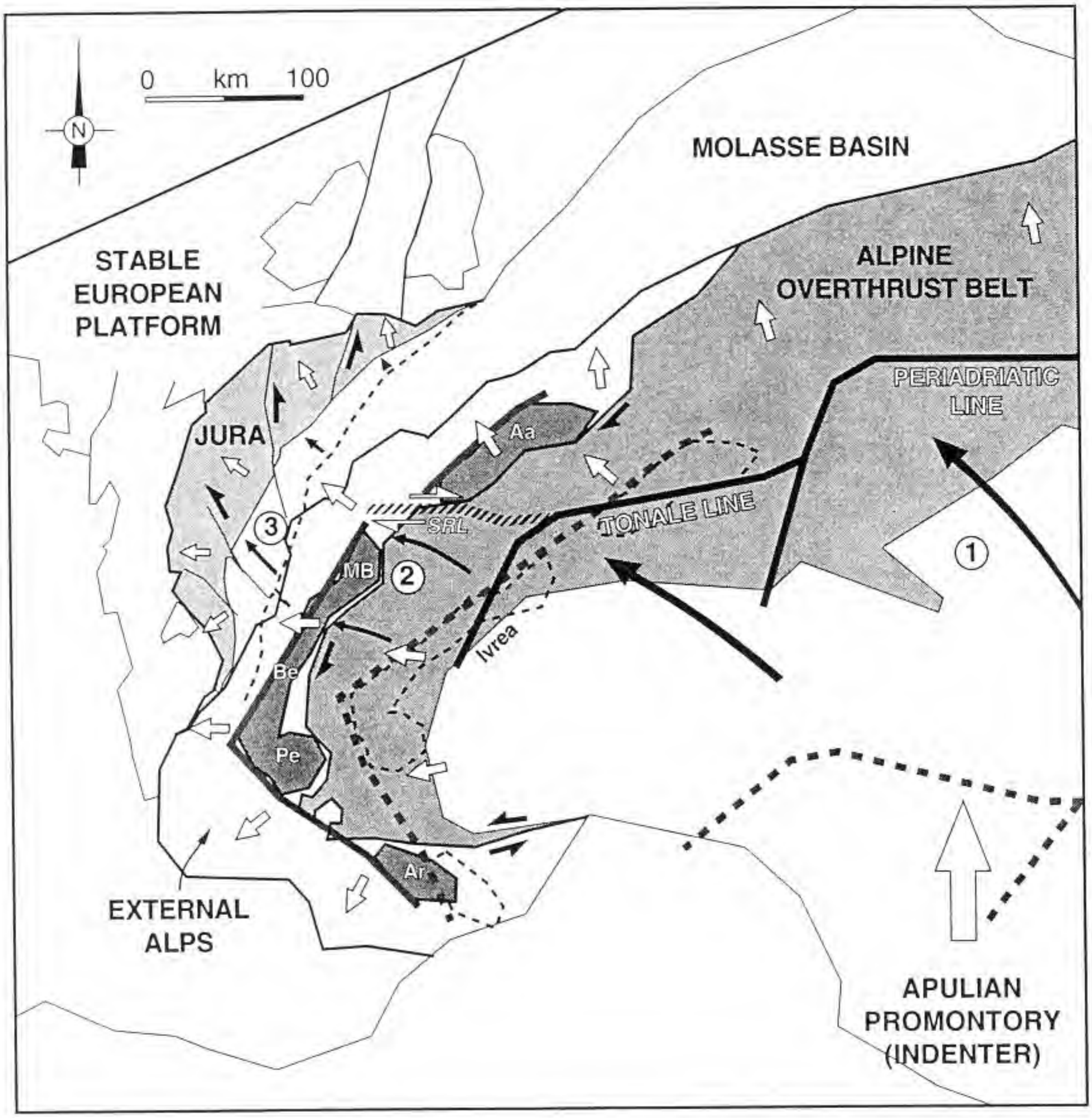


FIG. 12. Megatectonic map illustrating indenter effect of West Alpine arc and position of Jura fold-belt relative to the External crystalline massifs and Simplon-Rhône line (see text for explanation).
The palinspastic reconstitution of the Periadriatic line and External crystalline massifs is based on Vialon et al. (1989).
SRL: Simplon-Rhône Line.
External Crystalline massifs: AA: Aar-Gothard massif; BEL: Belledonne; MB: Mont Blanc - Aiguilles-Rouges massifs; Pe: Pelvoux massif.

ed to the location of Triassic evaporites (Laubscher, 1965, 1977; Mitra and Namson, 1989; Epard and Groshong, 1993). The Risoux 1 well, located on top of an abnormal structural high in the central Jura (see Encl. 1, section n°4), has revealed a duplication of the Mesozoic cover thanks to a major flat using Keuper salt and/or Early Jurassic shales. Moreover, a large number of drillholes have evidenced northwestwards displacement of about 10 km of the Mesozoic cover above a regional Triassic décollement level, both in the central Plateau zone (Fig. 7) and the western frontal part of the thrust belt (Michel et al., 1953; Chauve et al., 1988, see Encl. 1, section n°4).

(2) The arcuate shape of the Jura thrust belt and the radially diverging mass transport of the Mesozoic cover is clearly indicated by microtectonic analysis (Fig. 10) and measurements of in situ rock stresses.

(3) The set of regional cross-sections presented in Enclosures 1 and 2 demonstrates significant lateral changes in the deformation style of the Jura fold-and-thrust belt. These are largely related to the nature of the detachment level, the amount of bulk shortening and the thickness and rheological composition of the deformed Mesozoic sequence, without taking into account a significant involvement of basement. Lateral changes in bulk shortening measured in the Jura cover is only made possible if the latter is completely decoupled from its basement.

(4) As evidenced by industry-type reflection-seismic lines and the ECORS Alp-2 profile (Guellec et al., 1990a and 1990b; Deville et al., 1994), the Salève and Montagne d'Age anticlines emerging through the Savoy Molasse Basin, correspond to typical fault-propagation folds related to a sole thrust hosted in Keuper beds, thus demonstrating detachment of the Mesozoic and Cenozoic strata from the basement at the level of Triassic evaporites beneath the Molasse Basin. In the same way, reflection-seismic data in the eastern Jura (Müller et al., 1984; Laubscher, 1986) have clearly imaged decoupling between the post-Triassic cover and its basement (see Encl. 1, sections n°7 and n°8). This is fully supported by microscopic analysis of important shear zones in the basal Triassic evaporites, encountered by boreholes in the Plateau Molasse south of the eastern Jura (Jordan, 1992).

(5) The critical taper of a thrust wedge is related to the basal shear strength and in the Jura it is so low as to almost disappear. The only rocks which can provide such a low shear strength are evaporites, especially salt. Where evaporites are lacking, as below the western Chartreuse massif, where décollement follows Triassic or Liassic shales, the critical taper increases dramatically and thrust sheets are piled up on each other. As evaporites underlie Molasse Basin, they provide detachement of the latter rather than a highly tapered pile of thrust sheets at its southern margin.

(6) Simple viscous-brittle analogue models validate the dynamic arguments based on observed critical taper and are able to simulate deformation of the Jura nappe system, of which the Molasse Basin apparently forms an integral part. Moreover, they demonstrate the role of obstacles in the décollement layer for nucleation of thrusts. Such obstacles have been found to consist of faults and flexures of the Rhin-Saône transform zone (Laubscher, 1986; Noack; 1989; see Encls. 1 and 2, sections n°5, n°7 and n°8). Intrabasement thrusting would not be affected in this way, as no weak layer parallel to the top basement can reasonably be expected to have been present and displaced in a way analogous to the Triassic evaporites. To a certain extend, Late Paleozoic coalbeds could provide décollement of the Molasse Basin but such rocks are far from being uniformly distributed over the whole area; therefore they cannot be assumed to play a role comparable to that of Triassic evaporites.

On the basis of reflection-seismic data from the Molasse Basin, an alternate model is advanced by some of our Swiss colleagues: they propose an intra-crustal sole thrust, rooted along the northern margin of the Aar Massif, which extends through the Molasse Basin (Ziegler, 1982, 1990; Gorin et al., 1993; Ziegler et al., this volume) and Jura thrust belt (Pfiffner and Erard, 1995). This is resolutely opposed to our postulates developed above regarding a theory on Jura development. Although we advocate a late basement involvement in Jura deformation, there is no comparision between the amounts of horizontal shortening measured in folded and thrust sediments and in the pre-Triassic substratum where it is closed to zero. After having considered all possible angles and, keeping in mind, that all our interpretations hardly depend on data available from the Molasse Basin, we therefore see no other viable model except that of thin-skinned décollement on Triassic evaporites, i.e. the distant push model initially proposed by Laubscher (1961) for the Jura-Molasse nappe.

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Enclosures

Enclosure 1 Regional balanced cross-sections through the Western Jura and western Chartreuse subalpine chain (location on Fig. 2).

n° 1: Eastern Chartreuse massif - Bas-Dauphiné basin
n° 2: Savoy Molasse basin - Ile Crémieu High
n° 3: ECORS profile (Guellec et al., 1990a and 1990b; modified)
n° 4: Mont Tendre - Grozon high (Winnock, 1961; Laubscher, 1965; Bitterli, 1972; Martin, 1987; Guellec et al., 1990b; modified)

Enclosure 2 Regional balanced cross-sections through the Central and Eastern Jura (location on Fig. 2)

n° 5: Neuchatel lake - Ognon fault system
n° 6: Grenchen anticline - Rhine Graben (Buxtorf, 1916; modified)
n° 7: Aarau- Tafel Jura (Mühlberg, 1894; Buxtorf, 1916; modified)
n° 8: Lagern anticline (Mühlberg, 1894; Müller et al., 1984; modified)

Source : MNHN, Paris

Evolution, structure and petroleum geology of the German Molasse Basin

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ABSTRACT

The German Molasse Basin is a 300 km long segment of the of North-Alpine foredeep. In cross section, it is a composite wedge, up to 120 km wide and 0.3 to 6 km thick. Its basin fill consists of late Eocene to middle Miocene alternating marine and non-marine Alpine-derived clastics which were deposited during an estimated trans-basinal convergence of 250 to 400 km. Two successive and superposed megasequences reflect pre-extended lithosphere with a rigidity of $0.6 \cdot 10^{23}$ Nm. The older megasequence (42 to 20 Ma) coincides with the collision of the Adriatic and Penninic continental fragments with an edge-loaded European plate. The younger megasequence (20 to 8.5 Ma) shows 170 km or less of convergence and the weak and line-loaded flexure of the rising Alpine mountains. Deformation within the present Molasse Basin is dated at 12 Ma or possibly even as little as 6 to 2 Ma.

Developed oil and gas reserves amount to some $10 \cdot 10^6$ t ($80 \cdot 10^6$ bbl) of petroleum liquids and $21 \cdot 109$ m^3 of gas (735 BCF). These are contained in 59 mostly small oil and gas fields which are sourced and trapped within and below the undeformed Molasse wedge. Untested deep-gas potential exists in footwall imbrications near the Alpine front.

INTRODUCTION

The German Molasse Basin is a 300 km wide segment of the north-Alpine foredeep between the Rhine and Salzach rivers (Lemcke, 1988; Schwerd and Unger, 1981). This segment has achieved some geological coherence by a common history of mapping and petroleum exploration, manifested in more than 600 exploration and production wells, in 45 oil and gas fields, and in an estimated 5000 km of reflection seismic profiles (Lemcke, 1988). Geologically, it forms a composite wedge-shaped clastic prism which is 30 to 120 km wide and 0.3 to 6 km thick. Its age ranges between Priabonian (42 Ma) and Tortonian (8 Ma) times, and it involves a tectonic convergence of 250 to 400 km.

Our paper summarizes available geological data, describes accepted and new geodynamic

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This article includes 4 enclosures on 1 folded sheet.

interpretations, and speculates about some of the basin-forming mechanisms. Available data is consistent with the consensus of compressional tectonics within the European foreland during the Alpine collision (Ziegler, 1987, 1990; Bachmann et al., 1987). However, available data also suggests flexural response, high strain rates, and high bulk strain during the basin evolution, aspects with are in part contradictory or seemingly inconsistent.

SETTING AND GEOLOGICAL UNITS

The basin fill is a depositional body affected by tectonic deformation and erosion. It underlies the lowlands between the Alps and the Danube river (Fig. 1). To the south, the undeformed main part or Foreland Molasse is in contact with the deformed part or Folded Molasse. South of a frontal triangle zone, the Folded Molasse is an imbricate stack of up to four thrust sheets or detached folds with 1 to 5 km of estimated individual thrust transport. Typical surface features of the thrust sheets are tight north-vergent, doubly plunging synclines with steep limbs.

South of a structural contact, stratigraphically older units form three stacked North-Alpine belts, the Helveticum, the Rheno-Danubian flysch, and higher thrust systems grouped in the present paper as Austro-Alpine. Transport at the tectonic contacts between these units is polyphase and of plate-tectonic magnitude. The transport components of Tertiary age vary between 10 and 100 km. The poorly known structural style of post-stacking compression involves tight or open and duplex-type folding near the surface and presumably north-vergent imbrication at depth.

The base of the Molasse body corresponds to a regional unconformity which is onlapped by transgressive sands, shelf carbonates and dominant foredeep clastics. This unconformity truncates Mesozoic shelf sediments deposited in a passive-

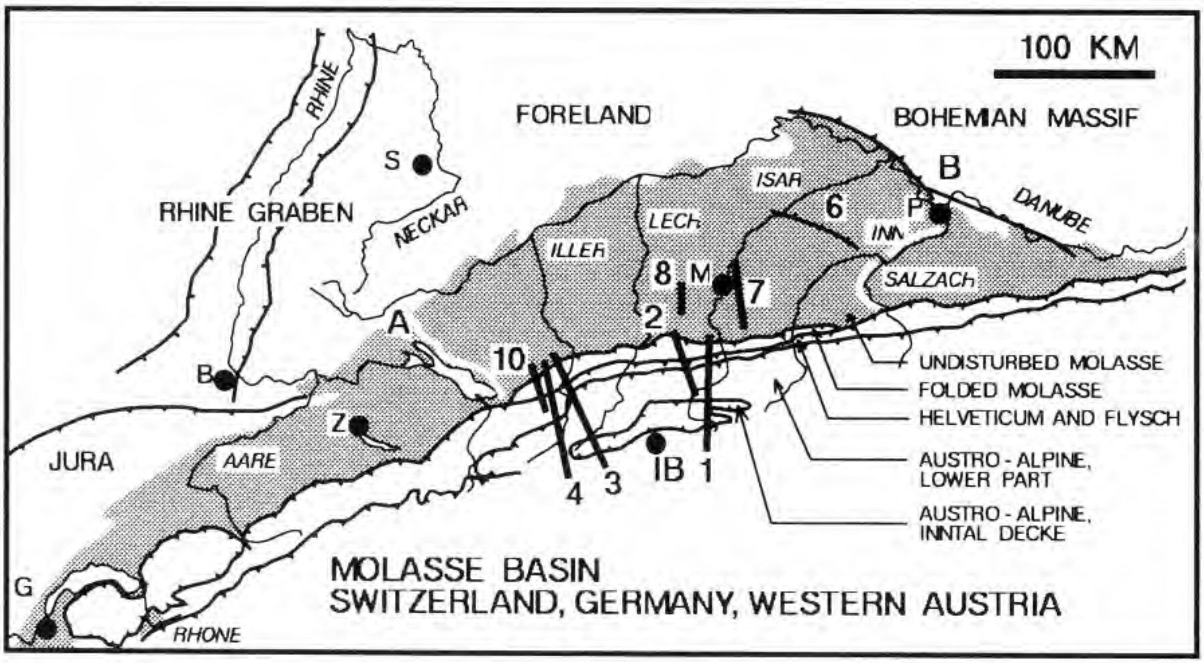

FIG. 1. Location map of north-Alpine Molasse Basin, redrawn afteBachmann and Müller (1991). Shaded: area underlain by foredeep sediments at the surface. White lines marked A and B: limits of the German segment of the Molasse Basin. Black sticks with numbers: cross-sections shown in present paper: 1, 2, 3, 4 : cross-sections on Encls. 1 and 2. 6: Landshut-Neuötting high. 7: basin cross-section given in Fig 7. 8: seismic profile Fig 8. Black dots with letters are cities: Stuttgart, Passau, Munich, Innsbruck, Zürich, Basel, Geneva.

extensional continental margin basin of Jurassic to Late Cretaceous age; their cumulative thickness increases southward from about 1 to 4 km. The structure of this former continental margin is polyphase and extensional; it shows pre-Alpine highs, rifts, and inverted basins. It is also affected by brittle extensional faulting of Tertiary age providing most of the productive hydrocarbon traps.

The crust beneath this sedimentary prism, its original southward extent, and possibly, its southward thinning, are partly defined by a gently south-dipping seismic Moho event. An abrupt southward rise of the Moho is observed near the South-Alpine Insubric line, located about 120 km south of the northern Alpine front. Perhaps this rise signals the southern edge of the foreland crust (see Ziegler et al., this volume).

STRATIGRAPHY AND BASIN EVOLUTION

Within the German Molasse Basin, Permo-Carboniferous to Tertiary sediments and subdividing unconformities record four evolutionary stages, from bottom to top, a rift-fill series, an epicontinental basin series, a passive-margin series and an Alpine foredeep series.

Late Palaeozoic continental fill of transtensional troughs overlays a Variscan basement which is coherent with the European crust at a cratonic thickness near 30 km (Ansorge et al., 1992). Several of the late Variscan troughs have been tentatively mapped, based on reflection-seismic data and on a handful of deep wells (Lemcke, 1988). Their fill can be more than 1 km thick and shows the ENE and WNW trends of the Variscan wrench-fault pattern in W-Europe and NW-Africa (Arthaud and Matte, 1977; Ziegler, 1990).

The basin floor is formed by a 1 km thick pair of shallow-marine sedimentary series. A Triassic to Middle Jurassic north-facing epicontinental basin series onlaps the basement from NW to SE and thins southward. It is overlain by a Middle Jurassic to Cretaceous passive-extensional-margin series, which faces to the south and thickens southeastwards. Isopach maps (Bachmann et al., 1987) show a regional element of southeastward thickening at rates of 10 m or less per km, and more local elements of crustal tectonics.

Molasse subcrop maps (Lemcke, 1988; Bachmann et al., 1987) show a persistence of the Mesozoic regional element in the form of a gentle bevelling toward the northwest. The thickest and most complete south parts of the basin-floor series now constitute the Alpine thrust sheets south of the present Molasse Basin.

An extensional event of Jurassic age records the shift from the epicontinental setting to the passive-margin setting. This event is important for structures and as a generator of petroleum source-rocks. It is recorded within the Alpine thrust sheets, but apparently it did not directly affect the German Molasse Basin.

Late Cretaceous to Paleogene strike-slip faulting generated a NW-trending block mosaic in the Mesozoic double sequence, in particular the Landshut-Neuötting High and the Bohemian Massif. These movements are related to the change in the Europe-Africa plate vector from southeasterly to northeasterly at 92 Ma (Dewey et al., 1989), and to compressional tectonism in the Alpine foreland (Ziegler, 1987, 1990). In the German Molasse Basin, its effects include extensive truncation of the Mesozoic sediments.

The fourth sediment sequence of the German Molasse Basin is a 0.1 to 5 km thick Tertiary clastic foredeep fill of Alpine and limited local provenance. Its relationship to the Alpine collision is clearly evident in its northwestward onlap and in its facies evolution from marine and turbiditic sandstones to non-marine clastics. It is also evident in the extensional tectonics of its substratum and, finally, in the trans-Alpine plate convergence. Since the late Eocene, subduction beneath the Adriatic plate may have begun to involve the south edge of the European continental crust. Flexural loading of the crust entailed truncation of the north parts of the Mesozoic shelf, and subsidence and marine sedimentation in the south parts. Most of the earliest Molasse sediments were probably buried beneath advancing thrust sheets. The younger Molasse formations were derived from emerging Alpine terranes and by cannibalizing the more southerly basin fill.

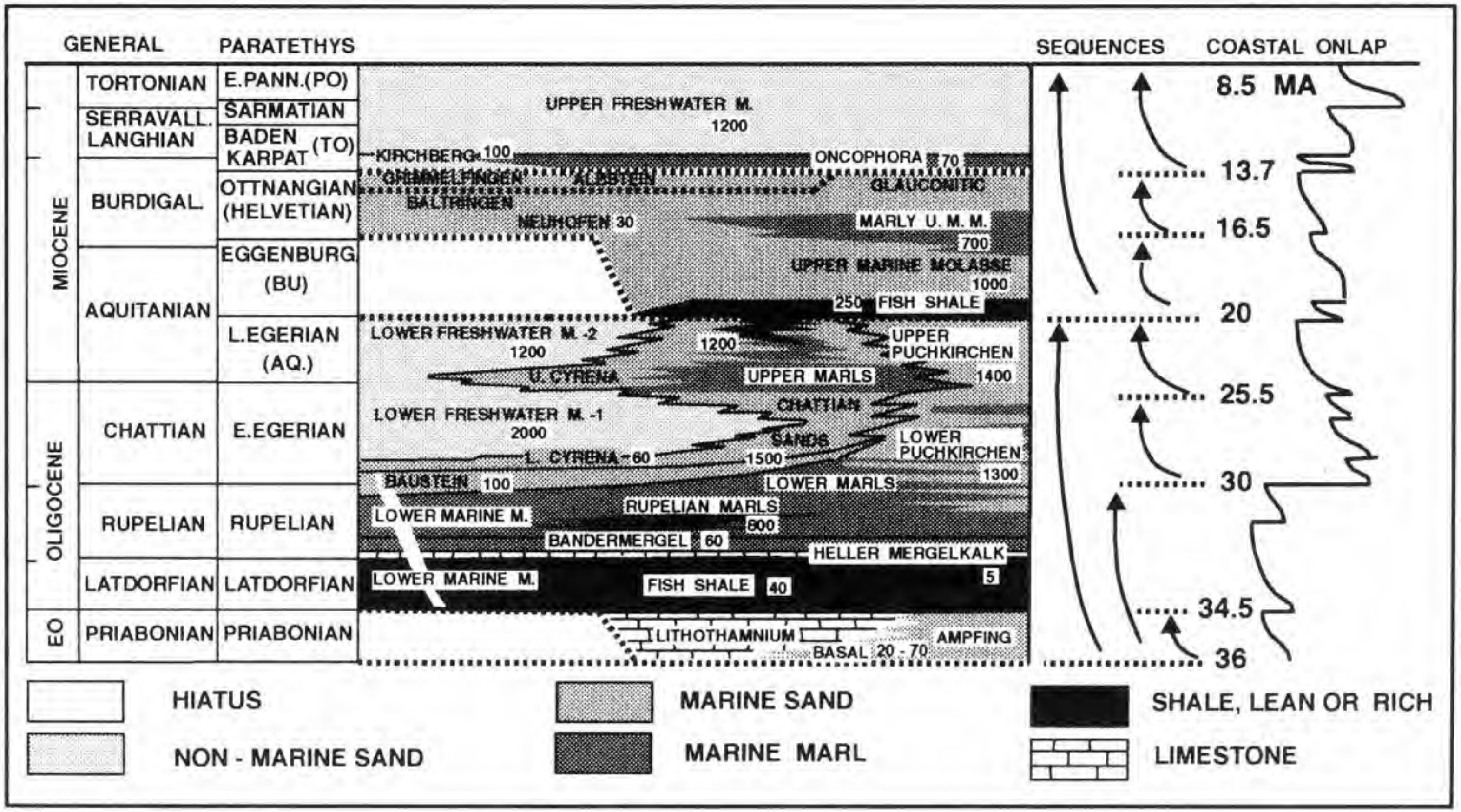

FIG. 2. Stratigraphic table of the German Molasse Basin, redrawn after Bachmann and Müller (1991), with shade patterns explained in a legend. On right, the transgressive-regressive sequences (vertical arrows) and their correlation with the coastal onlap curve (Vail et al., 1988). Chronostratigraphy, on left, after Steininger et al. (1985) and Fahlbusch (1981). Old, now obsolete stages in brackets: (AQ): Aquitanian, (BU): Burdigalian, (Helvetian); (TO) Tortonian; (PO) Pontian. Lithostratigraphy (central panel) modified from Lemcke (1988). Numbers in the centre field refer to average thicknesses in meters. Numbers to the right are geological ages.

MOLASSE DEPOSITIONAL SEQUENCES

In the German Molasse Basin, a system of four lithostratigraphic units can be recognized on the basis of biostratigraphy, mapping, and well results (Hagn and Hölzl, 1952; Ganss and Schmidt-Thomé, 1955; Hagn, 1960; Breyer, 1960). This system is still in use and includes the Lower Marine Molasse, Lower Freshwater Molasse, Upper Marine Molasse, and Upper Freshwater Molasse. Sequence stratigraphy (Haq et al., 1988), useful for dealing with high-resolution seismic data, sedimentology, and geodynamic concepts, is also used in the Molasse Basin (Lemcke, 1988; Bachmann and Müller, 1991; Jin, 1995).

Figure 2 shows a tentative correlation chart between the lithostratigraphic Molasse units and their sequence stratigraphic interpretation (Bachmann and Müller, 1991). It is similar to an east-westerly trending strike section of the German segment, but it shows neither the basal onlap geometry nor the lithostratigraphy of the Folded Molasse province. The chart shows a westward increase in non-marine facies and in non-deposition. It also shows the Molasse series divided into two major transgressive and regressive sequences.

Figure 3, keyed to the sediment bodies shown in Fig. 2, shows the northwesterly prograde pattern of onlap at the base of the Molasse. The onlap pattern shown is interpreted from well control (Bachmann et al., 1987), and it is affected not only by relative sea-level changes, but also by a regional element of elastic load flexure, as well as by local crustal tectonics.

A smoothened abstraction of the basal Molasse onlap pattern and the ages of its elements is shown in Figure 4 as dated lines or isochrons of zero deposition during the onlap. The generalized onlap pattern suggests a finite onlap rate of 0.5 cma^{-1}, which is in the same order as, and less than, the tectonic convergence rate. The geometry of the basin fill is further shown in Figures 5 and 6. The following description of the stratigraphy refers to these maps.

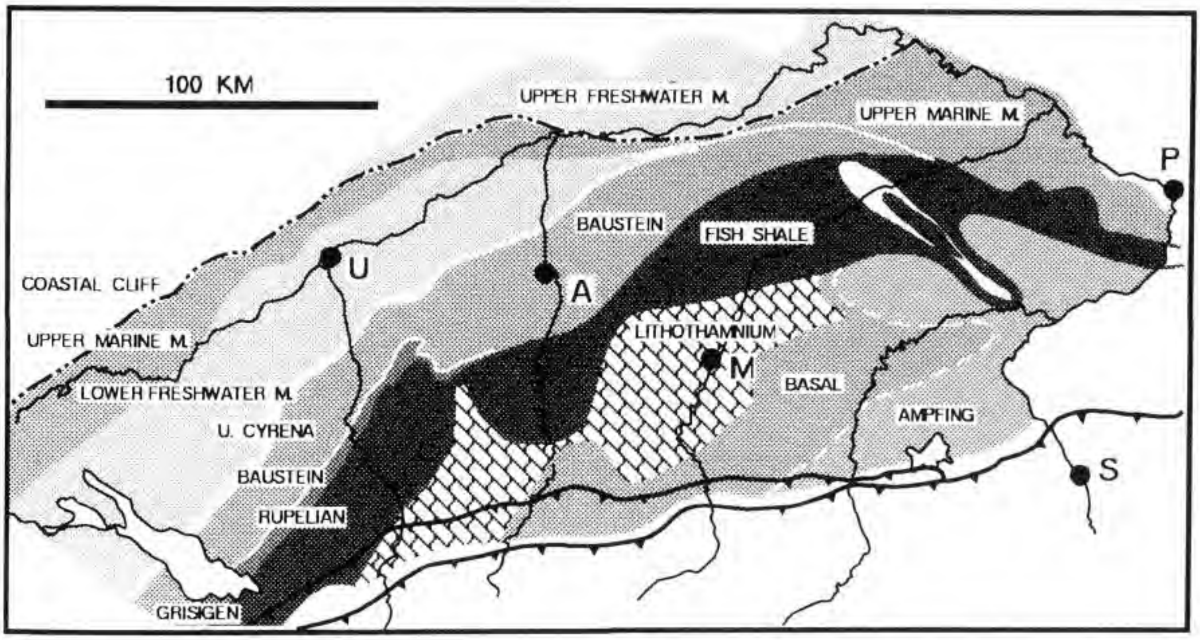

FIG. 3. Onlap map of German Molasse Basin, redrawn after Bachmann and Müller (1991). Shade patterns are geological formations, explained in legend to Fig. 2. Dash-dot-dotted line: coastal cliff of late Burdigalian age (14 Ma) in Jurassic limestones. Dots with letters: Ulm, Augsburg, München, Passau, Salzburg.

Older Molasse Megasequence

This sequence, of Priabonian (42 Ma) through lower Aquitanian (Egerian) (20 Ma) age, includes the Flysch-Molasse transition beds, the Lower Marine Molasse, and the Lower Freshwater Molasse. Late Eocene, locally derived transgressive sands, onlap the Paleocene unconformity and grade upwards into Lithothamnium limestones forming a broad platform (Figs. 2 and 3). Farther downdip, continuous sedimentation and a transition into Helvetic or Ultrahelvetic flysch units have been established stratigraphically, for example in the Hindelang-1 well (Huber and Schwerd, 1995). Everywhere in outcrop at the south edge of the Folded Molasse, evidence for this continuity is disrupted by late Miocene-age thrust tectonics and brittle strike slip (Schmidt-Thomé, 1962; Schwerd, 1983; Doben and Frank, 1983).

Rupelian and Latdorfian fully marine shales, including the organic-rich Fisch Schiefer, Mergelkalk, Bändermergel, Rupelian Marls, and Deutenhausen beds, document a generally transgressive series, in which the regressive early Chattian (30 Ma) coarse-clastic Baustein sands and conglomerates mark the transition into the Lower Freshwater Molasse.

Conglomerates, sandstones, and sandy shales of the Lower Freshwater Molasse thicken to 4 km in the Subalpine Molasse of Central Switzerland (Trümpy, 1980). An abundance of sedimentological details demonstrates changes in sea level, tectonic pulses, topographic changes and drainage slow-downs in the Alpine sediment source area (Lemcke, 1988; Müller, 1991; Jin, 1995).

Younger Molasse Megasequence

This sequence spans late Aquitanian (20 Ma) to late Tortonian (8.5 Ma) times and includes the Upper Marine Molasse, the Freshwater-Brackish Molasse, and the Upper Freshwater Molasse. The base of this sequence corresponds to an unconformity which toward the east disappears in marine shales of Chattian age (Egerian, 21 to 20 Ma) or in the marly Upper Marine Molasse of Eggenburgian age (20 to 18 Ma, Fig. 2). This unconformity can be related to strong basinal axial currents and to a narrowing of the basin as a consequence of nappe emplacement (P.A. Ziegler, personal communication). Regional structure data (Doben, 1981; Lemcke, 1988) also suggest a change in the parameters of the elastic foreland deflection at this unconformity, as induced by the load of the advancing Alpine body.

This Miocene megasequence onlaps northward through the Miocene with shallow-marine sands and silts. Beyond the present northern basin edge, a coast line with lithophagous traces of Ottnangian age (Helvetian, 18 to 15 Ma) represents an instant in transgressive onlap over Jurassic limestones. During the younger parts of this megasequence, relative sea level dropped and nonmarine sedimentation continued for another 7 or 8 Ma in the Upper Freshwater Molasse. Sedimentation ended at about 8 Ma and was followed by fluviatile erosion.

The total volume preserved of the younger megasequence (Fig. 6) is estimated at only about half of the preserved older megasequence. In part, this reflects the post-Tortonian erosion in the Folded Molasse belt, but also a change in flexural geometry of the basin shape which is as yet poorly understood.

Also the subjectively generalized onlap pattern (Fig. 4) reflects a significant but rather gradual difference between both Molasse megasequences. An older rate of 2.5 cma^{-1} is replaced in Rupelian time (33 Ma) by the much slower rates of 0.3 to 0.6 cma^{-1}. Poorly controlled onlap during the younger Molasse megasequence appears nearly stationary. Averaged over the entire basin and the total fill, the onlap rates vary between 0.3 and 0.6 cma^{-1}.

LITHOSPHERIC FLEXURE

The base of the Molasse sequence dips southward at angles increasing from 1.5° to 6° and averaging 3° (Fig. 7). Dips of the less well known Moho are steeper but roughly conformable to the

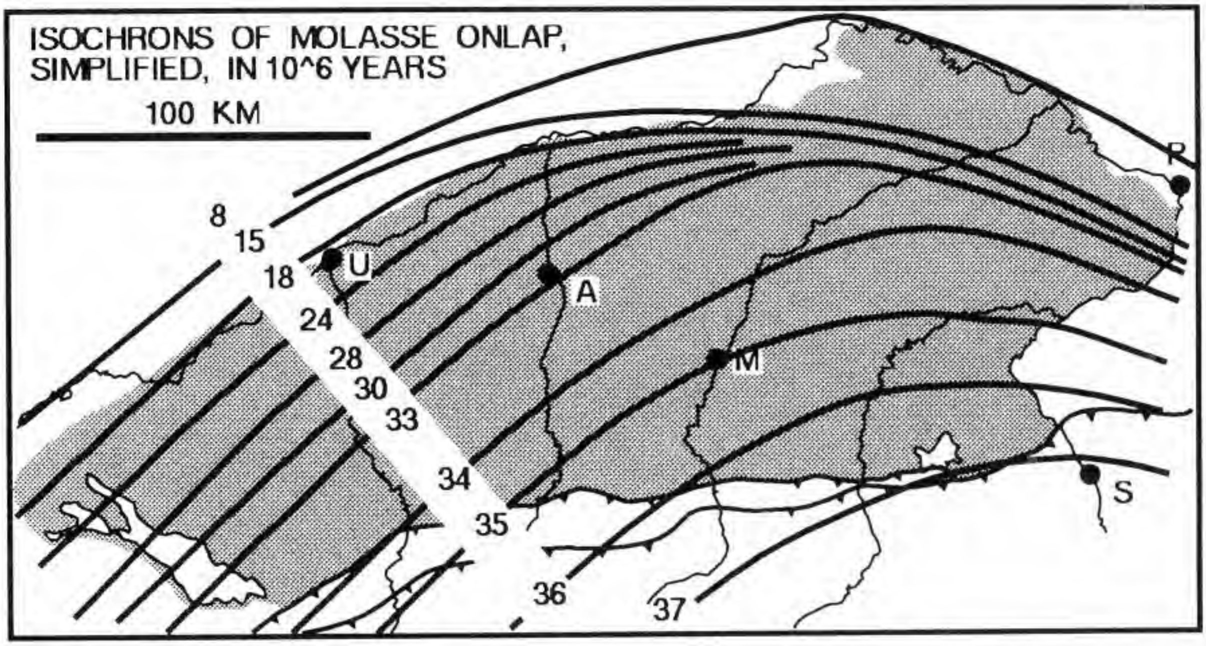

FIG. 4. Onlap map of German Molasse Basin (shaded), drawn by generalizing the onlap map of Fig. 3. Contours: lines of equal time of zero deposition. Numbers: ages in Ma.

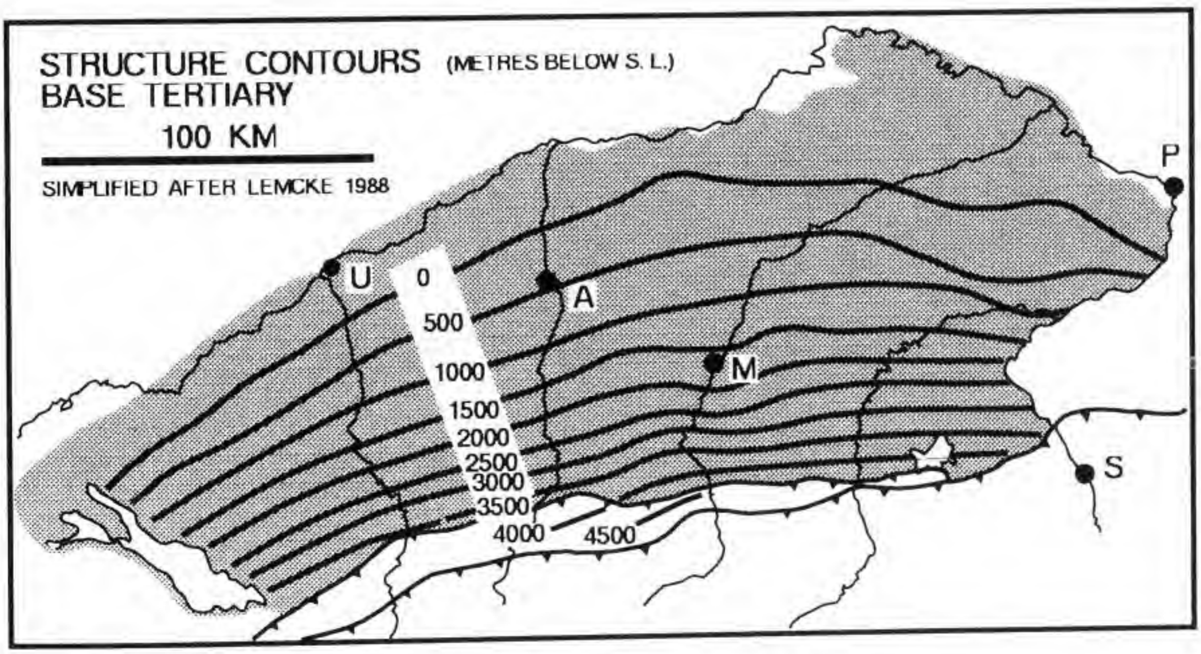

FIG. 5 German Molasse Basin (shaded), structure contour map of base Molasse in meters subsea, after Lemcke (1988).

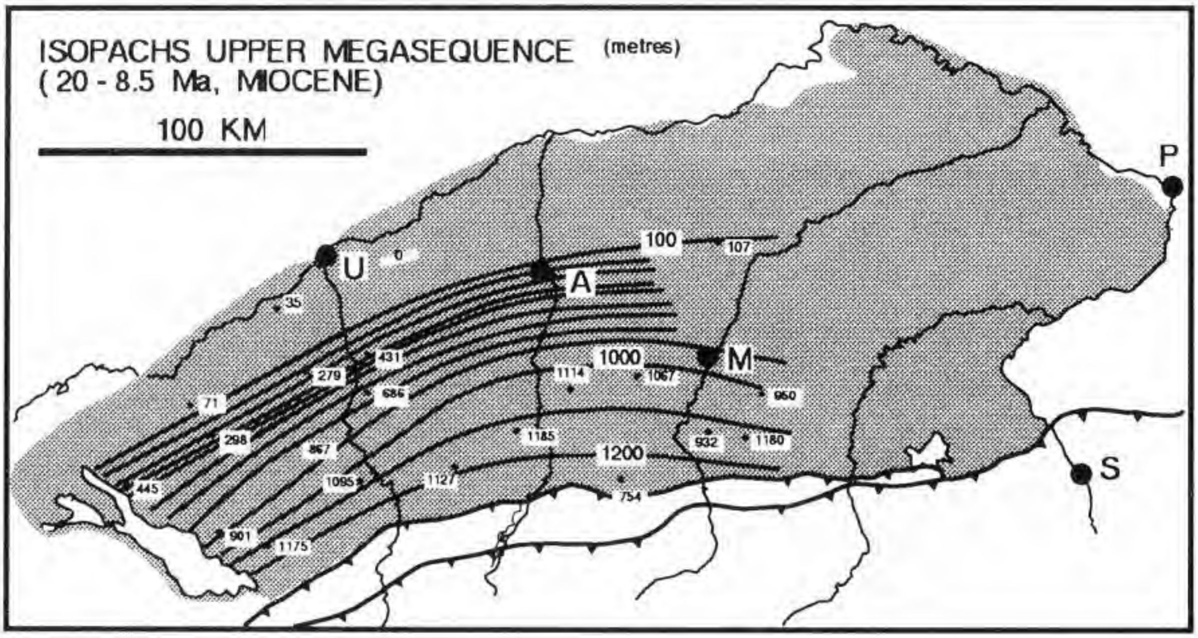

FIG. 6. German Molasse Basin (shaded), isopachs (in meters) of the Molasse Upper Megasequence as defined on Figure 2. Contoured interval is based on assorted well data shown as number fields.

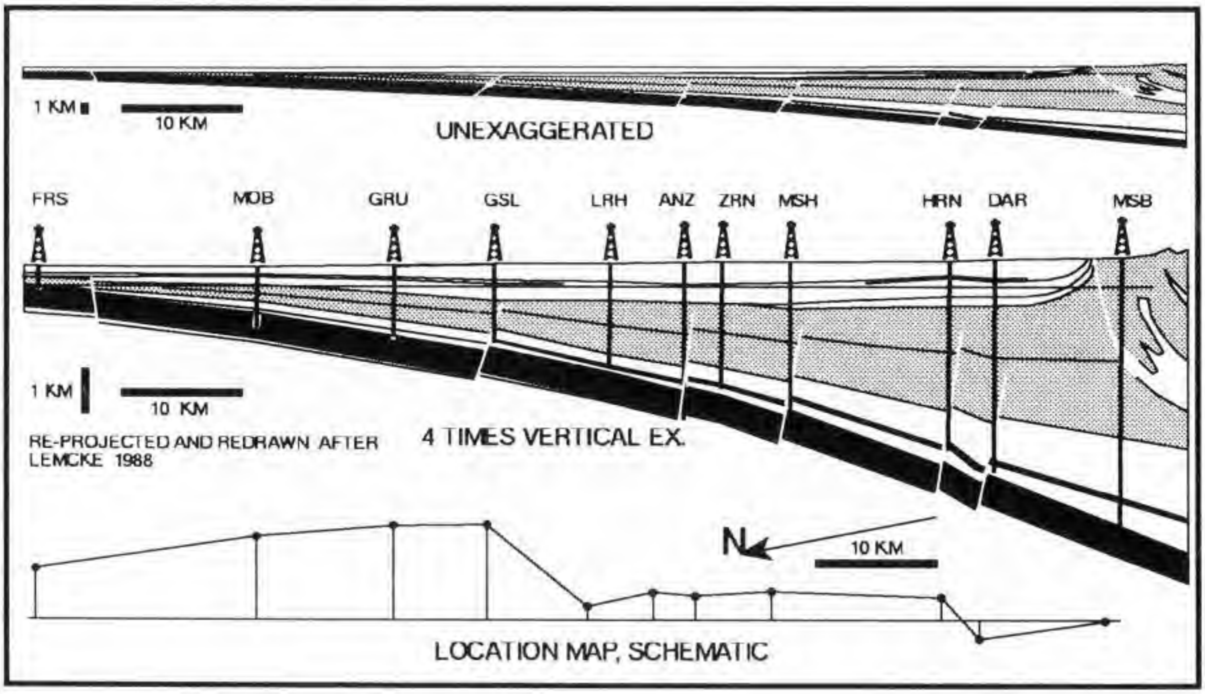

FIG. 7. Structural cross-section of German Molasse Basin based on well data, reprojected, restored to 1/1 vertical exaggeration, and redrawn after Lemcke (1988). Black: Malm carbonate of foreland series. Higher back line: Priabonian transgressive shales. Shaded: Chattian and Aquitanian interval. Groups of letters identify wells.

Molasse base (Müller et al., 1980; Giese et al., 1982; Prodehl and Aichroth, 1992). This crustal structure is interpreted as caused by elastic flexure of the lithosphere, loaded first by the Mesozoic passive-margin series, and later by subduction and by the topographic load of the Alps (Roeder, 1980; Royden and Karner, 1984).

Some of the studies of the European crust and its cumulative load have encountered unsolved mismatches and problems (Lyon-Caen and Molnar, 1989), partly ascribed to its complex origin, and partly to the mantle-induced rise of the Rhenish Massif. Therefore, we have limited present modelling efforts to an assumed flexural wavelength of 200 km, to an implied lithospheric elastic thickness of 29 km, and to an implied flexural rigidity D of $7 \cdot 10^{23}$ Nm. Although these values avoid the known mismatches, they are consistent with a European crust weakened by polyphase tectonics of Mesozoic and Tertiary age, and they can be matched by extrapolating southward the geometry of the Molasse Basin and its substratum (Figs. 5 to 7).

Normal-Faulted Molasse Basin Slope

Normal faults with throws of up to 0.2 km affect most of the German Molasse Basin and form oil and gas traps (Lemcke, 1988). These faults are of Oligocene to end-Aquitanian age, and they display a northward synsedimentary migration (Fig. 8). In the west part of the German Molasse Basin, an additional younger generation of normal faults is of lower to middle Miocene age (Lemcke, 1988).

A numerical estimate based on the geometry of circular bending (see Appendix) shows that the earlier faulting is consistent with an elastic deflection of a 20 km thick brittle upper crust with a neutral fiber at its base. In a 20 km long dip segment of the Molasse Basin floor, about 1 fault per km with a displacement of about 0.1 km would accommodate the brittle strain of elastic-load flexure.

Normal faults of the younger generation continue the outward migration of flexure, but during an interval without enough documented elastic-load flexure of Alpine origin (see Appendix). Therefore, some of these faults are perhaps related to a non-Alpine component of extensional uplift in the Rhenish Massif as defined numerically (Karner and Watts, 1983; Lemcke, 1988; Lyon-Caen and Molnar, 1989), or to the uplift of the Vosges-Black Forest Arch (Ziegler, 1994).

MEASURING SYNDEPOSITIONAL MOVEMENTS

Formation and filling of the Molasse foredeep took place during unknown amounts of crustal shortening along its south edge. To estimate the Alpine frontal strain, we must review all of its components. Following Means (1976), we define strain as the normalized change in length, and we define bulk strain as shortening or thrust overlap, measured as a distance. For the purpose of the present paper, convergence rate is bulk strain per unit time, plate convergence is Eulerian motion measured as a bulk strain along a defined small circle, and plate convergence rate is Eulerian bulk strain per unit time.

The zero edge of Molasse deposition advances (onlaps) northward relative to the basin floor. Although this onlap is largely independent of the Alpine strain, both are geologically related and therefore should be compared. Our study of Alpine-related basin progradation is illustrated by charts of the Molasse onlap (Figs. 3 and 4), by a chart of convergence rates and distances through time (Fig. 9), and by the restoration of one of our cross sections (Encl. 4).

The Northern Alpine front has advanced relative to undislocated or "pinned" (Boyer and Elliott, 1982) Molasse Basin fill. Trans-Alpine compression implies that the northern and southern Alpine fronts have approached one another. The trans-Alpine plate convergence has been traced through time by mapping the Atlantic sea floor anomalies and by Eulerian vector addition (Le Pichon et al., 1973; Dewey et al., 1973; 1989; Roeder, 1989). This data set is independent of Alpine data and quantifies both trans-Alpine plate convergence and convergence rates.

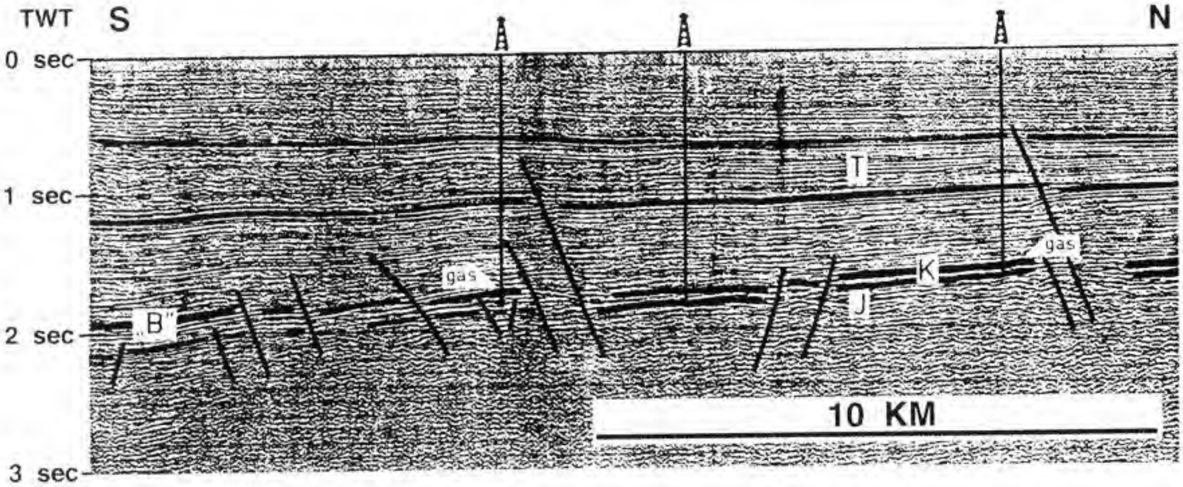

FIG. 8. Seismic dip line from the Undisturbed Molasse, east of Munich (after Bachmann et al., 1982), showing extensional normal faults, synthetic and antithetic to the southern regional dip and their successively younger synsedimentary activity from S to N.

To describe the bulk strain of the Molasse Basin, the difference must be established between plate convergence, South-Alpine convergence, and trans-Alpine convergence. Since tectonic events reach varying rates on either Alpine flank, the movements must be traced through time, by using literature data of vastly different resolution.

Convergence Data

In Figure 9 we have compiled Alpine convergence data through time since the trans-Alpine or Europe-Africa collision of latest Eocene age (38 Ma). The data are compiled from the review literature quoted in the following text segments. The vertical axis is in Ma shown as numbers at the left edge. The horizontal dimension of the five plots given is in km of convergence shown to the right of the plots. Uncertainty or ranges are expressed in shades of grey.

The output of this diagram is the plot to the right. It has been obtained by linearly subtracting from the trans-Alpine plot, the inner-Alpine convergence, the South-Alpine convergence, and the Orobic convergence. The North-Alpine convergence, therefore, is a best collective estimate.

The trans-Alpine plate convergence (38 to 0 Ma) has been determined by adding the vectors across the boundaries of the European, Adriatic, Maghrebian, and African plates (Roeder, 1989; Roeder and Scandone, 1992). The convergence shown is a vector describing the movement between Verona and Munich. Depending on the choice of Atlantic-derived data, the convergence amounts to 610 km since collision, and the rates vary between 1 cm/a and 2.2 cm/a, with an overall average of 1.4 cm/a. The rate increase since 9 Ma is an effect of the change in the Europe-Africa vector (Dewey et al., 1989).

South-Alpine bulk strain of Neogene age is shown to vary between 55 km (Schönborn, 1992) and 115 km (Roeder, 1992). Both quantities are speculative. Pulses in South-Alpine strain rates, based on stratigraphic data (Scandone et al., 1989) and radiochronologic data (Schmid et al., 1987) are shown in the chart, but are too detailed to be of much use in the Molasse Basin. An earlier South-Alpine event is identified as Orobic, and its bulk strain of 50 km is very uncertain.

Combined Intra-Alpine and North-Alpine bulk strains of Neo-Alpine age (Oligocene to post-middle Miocene, Trümpy, 1980) are widespread, polyphase, and accompanied by a crustal shortening estimated at about 110 km in the Swiss segment (Trümpy, 1980; Pfiffner, 1992; Ziegler et al.,

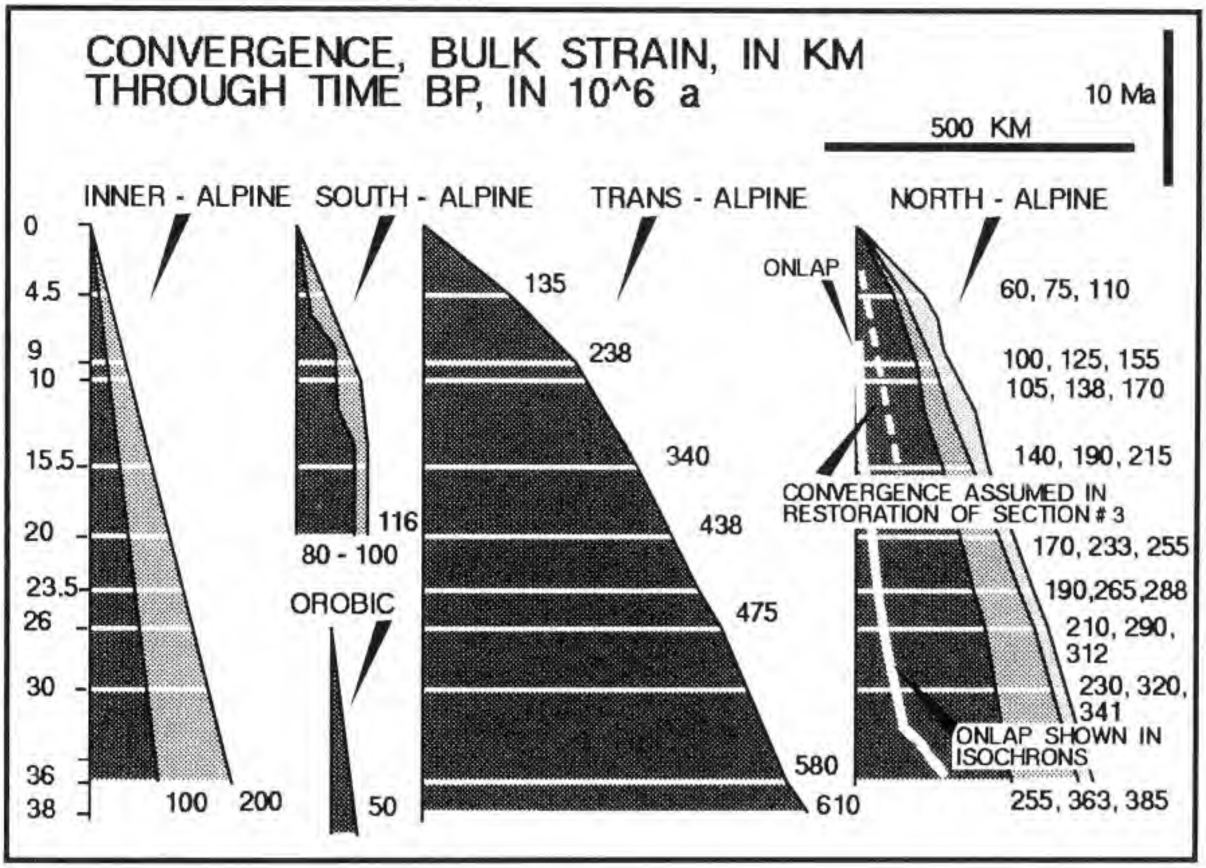

FIG. 9. Binomial Convergence-Bulk strain diagrams. Vertical: time axis in Ma. Horizontal: distance axis in kilometres.The diagrams show literature estimates of Alpine convergent bulk strain and their evolution through Oligocene and Neogene time. Dark and light shading: lower and higher amounts of convergence. Solid white line: onlap pattern shown in Figs. 2 and 3. Dotted white line: convergence used in retro-deformation on Encl.3.

this volume). For lack of more precise data, we have applied the Swiss shortening to a strain balance of the German Molasse Basin.

In the North-Alpine column of Fig. 9, we have added a strain path (white dots) of a tectonic restoration (see Encl. 4) which ignores the compilation almost completely. This reflects a lack of confidence in the compiled data, as well as in the North-Alpine subsurface structure, but not in the basic philosophy of strain compilation.

In Fig. 9 we have also added the rate of sedimentary onlap in the German Molasse Basin (white line) obtained by quantifying Fig. 4.

STRUCTURE OF THE NORTH-ALPINE FRONT

The North-Alpine front is a convergent plate boundary. Its constituent tectono-stratigraphic units have been pre-deformed, assembled, and juxtaposed at different times and places. This implies that its structure cannot be balanced, and that some of its key aspects are unknown. However, an abundance of high-quality data is available. A small part has been used to illustrate the Alpine north front in four regional cross-sections running through the key wells of Vorderriss, Staffelsee, Hindelang, and Immenstadt (Encls. 1 and 2). All cross-sections show the stacked structural units and part of their uncertainties.

Folded Molasse Unit

Müller (1970, 1978) used well control and vintage analog/single-fold seismic data to show that in many cross-sections of the Folded Molasse, the size of the synformal thrust slices is too small to fit into the space provided by the foredeep bottom. He inferred a detached and allochthonous fold belt overlying an overridden southward continuation of the Undeformed Molasse. In Fig. 10, this concept is applied to the Immenstadt area. The timing and amount of detachment at the base of the fold belt constitute significant solutions to problems of petroleum exploration along the Alpine north front.

A triangle zone along the north front of the Folded Molasse belt is documented by its north-dipping steep zone with a north-dipping backthrust implied at its base (Fig. 10). It serves as a merging termination or upper detachment to one or several blind south-dipping thrusts (Müller et al., 1988). The setting implies that the blind imbrication is coeval with the upturning of the monocline, that is, post-Tortonian (7 Ma). Further east, however, in the Perwang area of Western Austria (Janoschek, 1961; Tollmann, 1966), frontal imbrication is perhaps terminated not by an upper detachment, but by an erosional unconformity of mid- Aquitanian age (20 Ma). This interpretation would confirm the structural significance of the sequence-stratigraphic boundary at the 20 Ma-mark (P.A. Ziegler, personal communication).

In the western area, crossed by regional cross-sections 3 and 4 (Encl. 1), there is stratigraphic near-continuity between the steep zone and the adjacent syncline. However, there is a major thrust-type offset in a tight anticline near the middle of the folded Molasse zone. This vertical stratigraphic offset, typically of 3 to 4 km, separates an external Molasse unit and an internal Molasse unit, but the separating fault has not yet been traced regionally. In cross section 4, this thrust underlies four mapped Molasse folds, but its subsurface location is uncertain. Seismic data show the buried imbrication beneath the external Molasse unit (Fig. 10, Custodis and Lohr, 1974; Müller et al., 1988).

Cross-sections 3 and 4 (Encl. 1) intersect at the Immenstadt-1 well but show two contrasting interpretations of the subsurface architecture. Both versions are partly consistent with, but not clearly supported by, existing seismic data (Custodis and Lohr, 1974; Breyer, 1958). Both versions explain why the Immenstadt-1 well did not encounter the Chattian Baustein formation.

In cross-section 3 (Encl. 1), consistent with the classical imbrication model (Ganss and Schmidt-Thomé, 1955), the sole thrust of the Internal Molasse Unit passes below the bottom of Immenstadt-1. In cross-section 4, illustrating the newer concept by Müller (1970, 1978), the sole thrust of the Folded Molasse belt is cut by the well and overlays a parautochthonous Molasse unit. Both interpretations are further complicated by a low-angle backthrust merging with the upper detachment of the triangle zone.

Of the two interpretations, the version given in cross-section 4 suggests, but does not prove, a stratigraphic contact between fully allochthonous Helveticum and Molasse, later dislocated together with the Helveticum. The alternate interpretation (cross-section 3) shows only an uncritical thrust contact between Molasse and Helveticum. At present we prefer the version shown in cross-section 4, because of its regional suitability, and because of its ability to explain the juxtaposition of Molasse and Helveticum.

Helveticum Units

This group of one or more thrust sheets is located south of the Molasse unit. It is composed of an Oxfordian to Eocene shale-carbonate series, and it is interpreted as a detached part of the Mesozoic passive-margin series and pre-Oligocene Molasse Basin fill. Only a small part of this unit is present in outcrop. In our cross-sections it forms a steep-flanked, tight, and faulted antiform of thrust units. On its north flank, the antiform is overlain by Molasse and on its south flank by the Rheno-Danubian flysch unit. Basin geometry and sparse well data suggest that the Helveticum overlays

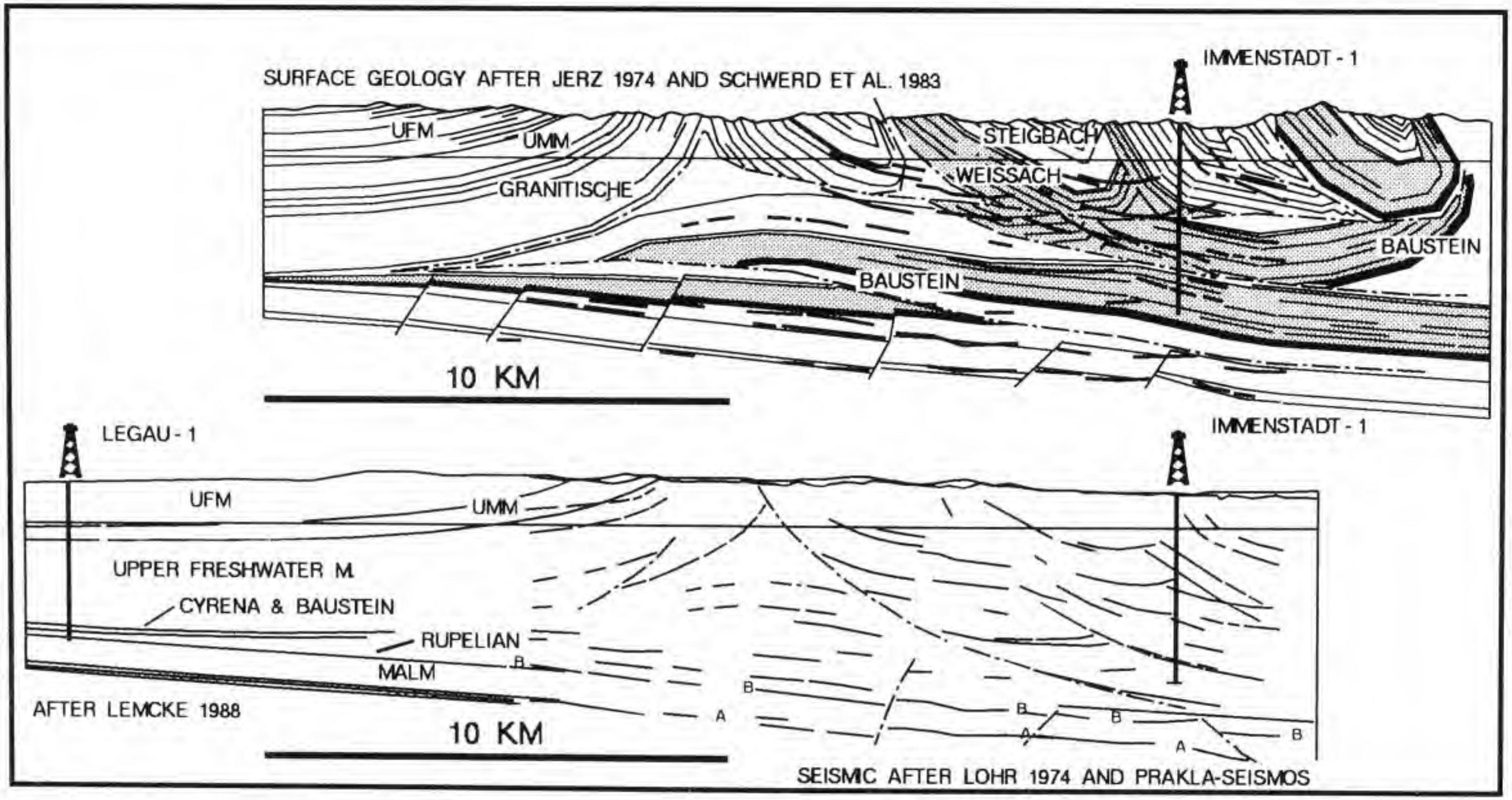

FIG. 10. Structural cross-section of Folded Molasse belt near town and well Immenstadt, juxtaposed with early analog and depth-converted seismic data. Seismic data from Breyer (1958) and Custodis and Lohr (1974). Geological data from Lemcke (1988) and obtained by construction based on surface data by Jerz (1974) and Schwerd (1983). Shaded: Weissach formation of early Chattian age (30 to 28 Ma). Black: Baustein formation of Rupelian/early Chattian age (31 to 28 Ma.) Seismic horizon A is interpreted as near top crystalline basement. Seismic horizon B is identified as near top Mesozoic carbonates.

unexplored and petroleum-prospective subthrust units of the Molasse which are within reach of the drill.

The top of the Helveticum is exposed on the steep to overturned north-flank of the antiform, where Molasse is overlying Helveticum with a minor hiatus and near-concordance, but the contact is a major and polyphase thrust fault wherever exposed. In several places (Encl. 1, cross-section 2), the top of the Helveticum appears in a tight antiformal window surrounded by flysch.

The main body of the Helveticum is known through four wells. The Kierwang-1 well (Encl. 1, cross-section 4) encountered undeformed Helveticum with an inferred thickness of 2.2 km (Müller, 1985). In the Vorderriss-1 well (Encl. 1, cross-section 1), the top 67 m of a strongly deformed Helveticum were opened (Bachmann and Müller, 1981). The Maderhalm-1 and the Hindelang-1 wells (Encl. 1, cross-sections 3 and 4) were spudded into large centres of polyphase folding and thrust imbrication (Huber and Schwerd, 1995). In outcrop, the width of the Helveticum decreases from a 10 km wide fold-thrust belt at the Rhine river to a quarry-sized sliver of indistinct shallow-water shale in eastern Bavaria.

The base of the Helveticum is a thrust fault with Molasse in the footwall, regionally mapped in Switzerland and western Austria (Trümpy, 1980) and encountered in the deep well Hindelang-1. We interpret this fault as part of the polyphase detachment separating the Folded Molasse unit from the overridden or parautochthonous Molasse (Müller, 1970, 1978). In many places, however, it is secondarily exposed by polyphase re-thrusting (Encl. 1, cross-sections 1, 3 and 4).

Flysch Unit: Subalpine, Ultra-Helvetic, Rheno-Danubian Units

Sediments attributed to this tectonostratigraphic unit are present beneath a large part of the Northern Alps and range in age from Early Cretaceous to Eocene. Several thrust sheets are distinguished (reviews by Gwinner, 1978; Doben, 1981). In the present paper, the Flysch unit is an internally featureless, 1 to 2 km thick, allochthonous body. Additional detached flysch bodies are known in the footwall of the Helveticum and have been drilled in Hindelang-1 (Müller et al., 1992). Geodynamically, the North-Alpine flysch may represent the accretionary wedge of the Austro-Alpine trench innerwall. Its base, therefore, is a major branch of the trans-Alpine collision suture.

Austro-Alpine Unit

This major tectono-stratigraphic complex is a nappe and clearly a detached part of a polyphase fold-thrust belt (Tollmann, 1976; Wessely, 1988; Roeder, 1989; Eisbacher et al., 1992). Its internal structure (Encl. 1, cross-sections 1 to 3) is not essential to the structure of the North-Alpine front. However, the depth, structure, and cutoff geometry of its base define the limits of the prospective subthrust province of the North-Alpine front.

The Austro-Alpine terrane involves a 3 to 4 km thick, extensional-margin series of Triassic to Cretaceous and, locally, to Eocene age. It underwent compressional and/or other tectonics above the Tethyan subduction during Late Cretaceous to Eocene times. Structural details show (Encls. 2 and 3, cross-sections 1-4) display a kinematic succession of detached folding, thrust faulting, and fault-bend folding. However, extensional structures of Jurassic age and facies changes of mid-Triassic age add severe complications to the thrust-fold evolution.

The base of the Austro-Alpine terrane is a thrust fault with a ramp-flat succession climbing northward through detached crystalline basement and through the sedimentary series. Over a large area, the Austro-Alpine sole fault is located within the mid-Triassic Raibl evaporite.

There are not enough data to constrain the ramp-flat sequence of this sole fault, but enough for an estimate of the finite ramp angle. The Austro-Alpine terrane has a dip width of 50 km and an estimated gross strain of 100%. The sole fault ramps up through 3 km of basement and 3 km of sediments. This suggests a finite ramp angle of 3.4°, if its origin predates the stacking of thrust sheets, and of less than 7° at the base of a stack. The second of these values is reasonable.

The depth to the Austro-Alpine sole fault is constrained by a major synform within the stack of thrust sheets near the Inn valley, by the Vorderriss-1 and several Austrian wells, by its re-emergence at the edges of the Tauern window south of the area studied, and by its up-plunge emergence at the Rhine valley. Seismological efforts to map its depth have not yet been very successful (Zimmer and Wessely, this volume).

RETRO-DEFORMATION

Ideally, the Alpine tectonic evolution should be retro-deformed through the entire time interval of foredeep deposition. This has been done conceptually and graphically (Trümpy, 1980; Ziegler, 1987; Roeder, 1989), but quantitatively and at present, this is possible only if an Alpine strain path is assumed. We have illustrated (Fig. 9) the wide range of uncertainty in the Alpine convergent strain path (see also Ziegler et al., this volume).

As an alternative project with limited scope, we have retro-deformed the Molasse and Helveticum parts of cross-section 4 of Encl. 1 in four kinematic steps to the undeformed state of the basal Molasse and its substratum of Helveticum (Encl. 4). We have used the SNIP routine (Roeder, 1991) which is a computer-aided form of hand balancing. It is not as precise as commercial balancing software, but it can be used more easily in areas with poor data control.

If we assume that the documented 60 km of bulk strain represents an arbitrary 50% of the North-Alpine convergence (Fig. 9), we obtain a finite North-Alpine convergence rate of 0.59 cma^{-1}. Applied to the 60 km of bulk strain, this rate would date the retro-deformation series as early Messinian time (10.5 Ma) to the present. This implies that the entire Folded Molasse and most of the Undeformed Molasse strata are pre-kinematic with respect to the structures within the Molasse belt. This is consistent with the opinion of most workers (Doben and Frank, 1983; Schwerd and Unger, 1981; Schmidt-Thomé, 1962; Lemcke, 1988). In the following comments, reference to geological ages of kinematic events are based on this assumed strain path.

Present State

If the convergence across the North-Alpine front has been decreasing, the present state may have been achieved up to 2 Ma before the present.

The Folded Molasse belt shows a detached Internal unit, a pinned External unit and a parautochthonous foreland with two footwall imbrications. The Helveticum terrane and the Internal Molasse unit have been jointly emplaced on a common sole fault. There is blind thrusting and complex interference between the backthrust front of the External unit and the allochthonous unit. The front of the Helveticum is a tight north-vergent antiform involving a stack of thrust sheets. The steep, north front of this antiform contains a strongly faulted depositional surface of Molasse on Helveticum.

First Restorative Step

At a steady bulk strain rate of 0.59 cma^{-1}, the state shown here would date at 3.1 Ma or early Plaisancian. As determined by SNIP graphics, the bulk strain restored in this step is 6.2 km.

The frontal footwall imbrication is restored. This results in the unfolding of the present steep zone and the adjacent syncline. It also results in a restoration of the discordant low-angle backthrust. The sole fault, common to the Helveticum and the Internal Molasse unit, ends blindly beneath the External Molasse and the undeformed foreland. The imbricate structure of the Internal Molasse unit is fortuitously located in front of the internal footwall imbrication.

Second Restorative Step

At a steady bulk strain rate of 0.59 cma^{-1}, the state shown would date at 4 Ma or early Pliocene. The SNIP-determined interval bulk strain is 5.3 km.

The Helveticum, probably overlying a carpet of involute sub-Alpine flysch, is shown as pushing an imbricate series of Oligocene and lower Miocene Molasse over a foreland series which includes beds as young as lower Chattian. At the front, this thrust is blind and ends at a northward migrating ("backpeeling") backthrust and steep zone.

Third Restorative Step

At the model bulk strain rate of 0.59 cma^{-1}, the northward advancing front of Alpine compression would have reached the south edge of the presently preserved Molasse terrane at 9.7 Ma or early Tortonian. The SNIP-determined interval bulk strain is 33.8 km.

All frontal imbrications of the Internal Molasse unit are restored and placed south of the foreland Molasse. At the present contact between Molasse and Helveticum, the tight fold is restored, but none of the fault displacement affecting the contact. The hinterland of the Molasse terrane consists of Helveticum with its tectonic cover of Flysch and Austro-Alpine terranes.

At this stage, the hinterland of the Molasse is still overlapping the foreland Mesozoic series, by transport on the Helveticum sole thrust which is older than the third restorative stage. The amount of overlap is unknown. It can be determined structurally (see Encl. 1, sections 1 and 2) or by applying a strain rate.

The Helveticum sole thrust may have started to break through the Molasse series to the surface, or it may have started the blind system with steep zone, backthrust, and "peelback" mechanism. The thickness of the southernmost, still undisturbed Molasse series is graphically estimated at 6.5 km.

Fourth Restorative Step

Displayed at a scale change (Encl. 4, cross-sections 5 and 6), this step contains a retro-deformation of the Helveticum terrane and a retro-displacement to its position downdip of the foreland Mesozoic series. Both steps are highly speculative. As modeled graphically with the SNIP routine (Roeder, 1991), this step comprises an interval bulk strain of 54.7 km. As part of this strain, the internal deformation of the Helveticum terrane has been graphically estimated at 56%, based on its documented style (Encl. 2, cross-section 4).

Tectonic Chronology

The timing of the tectonic events modeled is very uncertain; however, it is a key element in judging the availability of thrust-buried maturation of hydrocarbon source-rocks.

Retro-deformed cross-section 6 (Encl. 4) shows a concept of the basin state at the time when the Molasse fold-and-thrust belt began to form. The entire Molasse series is shown to overlap unconformably an imbrication of flysch and Austro-Alpine nappes. The basin is shown to open towards the south. At a realistic basin-bottom slope of 3°, the not yet deformed Molasse Basin extended another 60 to 80 km to the south. The stack of Flysch and Austro-Alpine nappes, we assume in this model, had been assembled and emplaced prior to the deposition of the oldest Molasse of late Eocene age. Therefore, this model implies that the North-Alpine front was tectonically relatively quiescent during Priabonian (36 Ma) to Eggenburgian times (19 Ma).

This very long quiescence is not consistent with the trans-Alpine convergence data. Therefore, it is more likely that the site of emplacement of Flysch (piggy-backing the Austro-Alpine) over its foreland of Helveticum had been located 60 to 80 km south of where it is shown in the reconstruction. The restoration shown suggests, however, that detachment of the Helveticum commenced at the beginning of the younger megacycle during the early Miocene.

The long tectonic quiescence could also be avoided by assuming much lower strain rates (Ziegler et al., this volume). However, fold-and-thrust belt tectonics (Bally et al., 1966; Boyer and Elliott, 1982; Price and Hatcher, 1983; Suppe, 1985; Roeder, 1992) tend to support the higher strain rates assumed in our model.

The structure sections, convergence rates, and retro-deformations shown in the present paper illustrate, rather than solve the problems. New work focussed on the tectonic chronology is needed to arrive at reliable, or at least internally consistent, results.

PETROLEUM RESOURCES

The German segment of the Molasse Basin is in a mature state of petroleum exploration, and production has been declining since a decade. About 600 wells were drilled in this basin between 1948 and 1985. Based on cumulative production until 1994 (Niedersächsisches Geologisches Landesamt, 1994)), the total developed reserves, from 59 mostly small oil and gas fields, is estimated at $10 \cdot 10^6$ t ($80 \cdot 10^6$ bbl) of petroleum liquids and $21 \cdot 10^9$ m^3 of gas (735 BCF) with the peak of production attained between 1966 and 1974.

Since 1992, there has been little activity to ease the steep decline of production. In 1994, the remaining proven and probable reserves in the 25 fields still on stream were $0.5 \cdot 10^6$ t ($4 \cdot 10^6$ bbl) of oil and 10^9 m^3 (35 BCF) of gas.

Figure 11 shows the basin-wide distribution of oil and gas fields. Their close relationship to Oligo-Miocene extensional fault blocks is evident. Filed names, detailed geology, and discovery histories are well summarized by Lemcke (1988).

Productive reservoir rocks (Lemcke, 1988) include Mesozoic carbonates of the Helveticum and the Molasse Basin floor, transgressive late Eocene sands, Oligocene Baustein coarse clastics, and various coarse clastic horizons within the Miocene Molasse.

Source rocks for oil and associated gas in the western basin parts, established with reasonable certainty, include the Toarcian Posidonia shales (Jurassic, 194-188 Ma) which provide the charge for Mesozoic and basal Tertiary reservoirs. The coals of the Permo-Carboniferous trough fill are near the high-temperature end of the gas window (Kettel and Herzog, 1989).

Source rocks for oil in Tertiary reservoirs of the eastern basin parts are Oligocene shales (Fisch Schiefer, 34-32 Ma). Biogenic (immature) gas in eastern Bavaria was generated in late-Oligocene and early- Miocene deeper marine shales (Schoell, 1977).

On a more speculative level, source-rock quality (Lemcke, 1988) is also ascribed to the Muschelkalk (Triassic, 240-230 Ma) and to the Helvetic Quintner Kalk (upper Malm, 150-145 Ma). In the southern region, thrust-loaded with Alpine units, any organic-rich formation of Jurassic to Oligocene age will potentially generate thermal gas, available for updip migration. Of particular interest in exploring sub-Alpine trends is the Oligocene Fisch Schiefer in subthrust position.

The western basinal region, where the dominant basin fill consists of Lower Freshwater Molasse, is barren and generally devoid of source-rocks and reservoir/seal pairs. However, in the north part of this region, there are Mesozoic reservoirs charged by Mesozoic source rocks.

Remaining Hydrocarbon Potential

Presently developed production is limited to the undeformed part of the Molasse Basin. A significant percentage of the known and possible prospects has been drilled. Undiscovered reserves from the existing geological play types are judged to be small. Any renewed exploration would therefore depend on new or untested old play types.

New play types would have to focus on the most likely hydrocarbon kitchens in the southern parts of the basin which are tectonically buried beneath Alpine nappes. Any renewed interest would also have to focus on the updip migration pathways from these kitchens.

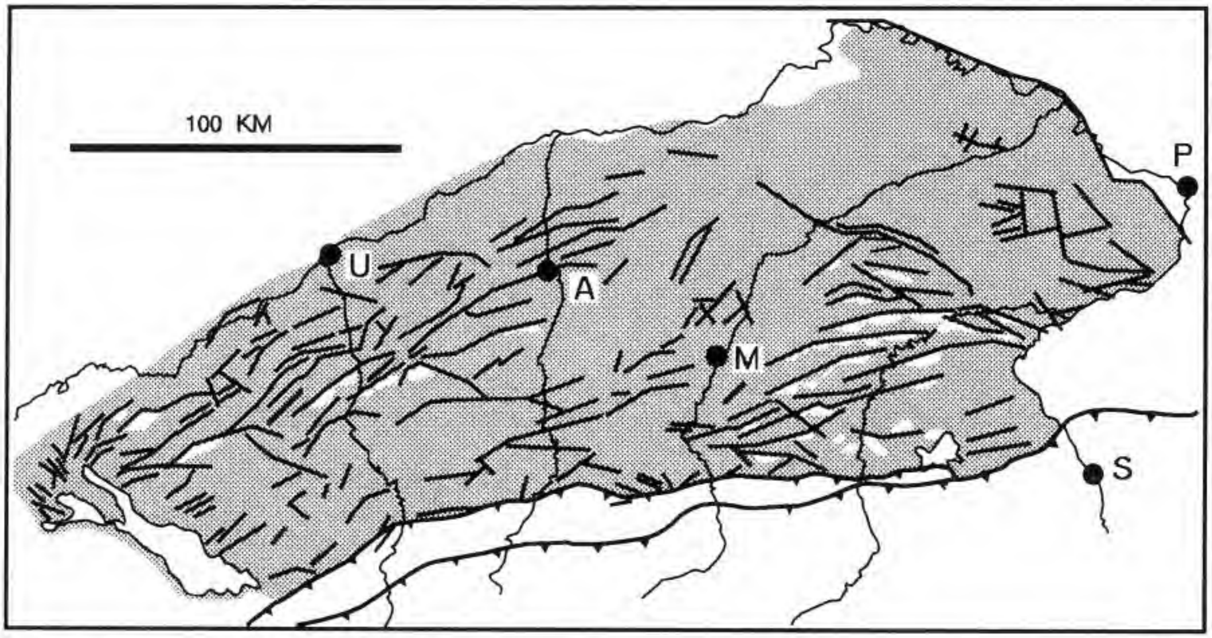

FIG. 11. German Molasse Basin (shaded) with simplified pattern of block faults and oil and gas fields (white), (after Lemcke, 1988).

The sub-Alpine hydrocarbon kitchens are poorly known. There are no publicly accessible data on source-rock volumes in the German Molasse Basin, nor in North-Alpine terranes. Applied to a basin volume of 80,000 km^3, the estimates of pooled oil and trapped gas suggest a net petroleum yield ratio of 125 tons of oil and $0.6 \cdot 10^6$ m^3 of gas per km^3 of total basin fill. These ratios are less than 10% of a range of yields for productive sedimentary basins world-wide.

This vague estimate can serve as a guide-line for exploratory decisions. An optimistic interpretation of the estimate would hold that in the German Molasse Basin, updip migration has been partially blocked by traps located downdip or south of the existing productive area. Alternatively one could assume that sizeable quantities of hydrocarbons have been lost to the surface, as suggested by Ziegler et al. (this volume) for the Swiss part of the Molasse Basin.

New Play Types

In the German segment, exploration of the Folded Molasse and Alpine fold-thrust belt took place more than a generation ago and prior to the availability of modern seismic techniques and structural theories. It generated a geological framework and a significant number of oil and gas shows from deep and needed, but perhaps poorly located wells. A modern consortial effort with federal German support has led to two world-class wildcat wells (Vorderriss-1 and Hindelang-1) and has added modern seismic data, hydrocarbon shows, and significant geological insight.

A new effort in the Folded Molasse and frontal Alpine area is presently departing from the existing data set with modern techniques, concepts, and comparative examples. The effort is focussing on mapping structural traps in sub-Alpine migration pathways. Most likely, traps in this position will be parautochthonous imbrications in front of, or just beneath, the North-Alpine front.

Such traps require good reflection-seismic definition, a convincing geological interpretation, and a reasonable thermo-structural history. Furthermore, they must be within reach of the drill, in topographically accessible locations, have a plausible charge mechanisms, and, all important, it must be demonstrate that they have an economically viable reserve potential.

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APPENDIX: NORMAL FAULTS AND CURVATURE OF FLEXURE

Extension by bending depends on the difference in arc length. The outer circumference of an arc segment of a length of (pi)(alpha) and of a thickness (delta)h is longer than the inner arc segment by an amount of (delta)l:

(delta)l = (delta)h·(pi)(alpha)/180

The end-loaded elastic-load curve has its maximum deflection angle at its free end. At a flexural length of 160 km and a maximum deflection of 36 km, the first derivative of the elastic-load curve (achieved numerically with MathCAD) suggests a near-steady decrease in deflection of about 0.8 degrees every 10 km. An arc length of 21 km taken from the middle of the foreland crust has a curvature of 2.5 degrees. Its convex upper surface has an excess length of 0.85 km, if it is 20 km thick. This excess length could be accommodated by 17 normal faults each with 0.1 km of displacement.

Enclosures

Encl. 1 Four regional structural cross-sections through North-Alpine front, compiled and constructed from data by Bachmann et al. (1982, 1987), Müller (1970, 1978, 1987), Müller et al. (1988), Doben and Frank (1983), Schwerd (1983), and Lemcke (1988). Hachure: detached crystalline foreland basement. Dark grey: Helveticum and foreland Mesozoic rocks. Mid-grey: silhouette of Austro-Alpine and Flysch units. Black: Baustein formation. Light grey: Chattian, Aquitanian, and (in sections 3 and 4): lower Chattian-age rocks.

Encl. 2 Structural details of cross-sections 1 and 2 shown in Encl. 1, based on referenced data sources and on Bachmann and Müller (1981), Huckriede and Jacobshagen (1958), Tollmann (1976), and field work by D. Roeder in 1995.

Encl. 3 Structural details of cross-sections 3 and 4 shown in Encl. 1, based on referenced data sources and on Bachmann and Müller (1981), Huckriede and Jacobshagen (1958), Tollmann (1976), and field work by D. Roeder in 1995.

Encl. 4 Retro-deformation in 5 stages of cross-section 3 of Encl. 1. Discussion in text.

Hydrocarbon exploration in the Austrian Alps

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ABSTRACT

Following discovery of several oil and gas accumulations in the allochthonous units underlying the Neogene Vienna Basin and in the autochthonous series of the Molasse foreland basin, exploration for hydrocarbons commenced in the late 1950's also in the Alpine Flysch Zone and the nappes of the Calcareous Alps.

Targets were Mesozoic and Paleogene series covering the sub-thrust authochthonous basement of the European foreland, which dips as a gentle monocline at least 65 km beneath the Alpine nappes, as well as the Mesozoic series involved in these nappes. Autochthonous reservoir rocks comprise Middle Jurassic and Cretaceous sandstones, Late Jurassic carbonates and Eocene and Oligocene sandstones. In analogy with the reservoirs of the oil and gas fields dicovered beneath the Neogene sedimentary fill of the Vienna basin, Triassic dolomites present a potential objective within the nappes of the Calcareous Alps. In the western parts of the Austrian Alps, autochthonous Triassic and Early Jurassic, as well as the allochthonous Mesozoic sediments of the Helvetic nappes present possible targets.

East of the basement spur, which projects from the Bohemian Massif under the Alpine-Carpathian nappes, autochthonous Late Jurassic shales form a major source-rock; west of this spur, basal Oligocene shales have excellent source-rock characteristics. Within the allochthonous units several potential source-rock intervals are recognized. Maturation of source-rocks was achieved during thrust-loaded subsidence caused by the emplacement of the Alpine nappes.

Exploration activity in Alpine Austria includes surface geological mapping, gravity and magnetic surveys and the acquisition of 5000 km of 2D reflection-seismic lines and two 3D surveys. Within the Flysch Zone and the Helveticum 24 and within the Calcareous Alps 8 exploration wells were drilled. The deepest well reached a total depth of 6028 m. In addition a number of wells were drilled in the imbricated sub-Alpine Molasse.

Near Vienna, the large Höflein gas field was discovered in the autochthonous Mesozoic series beneath the Flysch Zone. In Upper Austria, the well Molln-1 tested gas from Triassic carbonates of the Calcareous Alps nappes. Exploration of the Sub-Alpine Molasse yielded the light oil discovery of well Mühlreit-1. Apart from Höflein, all discoveries were subcommercial.

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This article includes 1 enclosure.

Exploration in the Alpine belt of Austria has met with only limited succes due to poor structural definition of prospects involving either autochthonous or allochthonous series. In this respect, the complexity of overburden geometries and topographic constraints on recording sufficiently dense reflection-seismic grids played an important role. Although all ingredients for a successful exploration play appear to exist within the Austrian Alpes, the risk/reward ration must be considered as lopsided under todays oil and gas price scenario.

INTRODUCTION

Austria can look back at a long and successful hydrocarbon exploration history (Brix and Schultz, 1993). The first commercial oil discovery was made 1934 with the drilling of well Gösting-2 in the Neogene Vienna Basin. In 1949 the very large Matzen oil field was found in the same basin; this field remained for a long time the largest oil accumulation found in Europe. In 1959 the first gas accumulation was found in fractured Triassic carbonates, involved in the nappes of the Calcareous Alps, forming the substratum of the Vienna Basin; this discovery was followed by a number of additional oil and gas discoveries in similar reservoirs and tectonic setting. Exploration of the Molasse Basin commenced in the mid 1950's and was rewarded in 1956 with a first oil discovery in Upper Austria and in 1960 with a gas discovery in Lower Austria. By now remaining recoverable reserves in established accumulations of Austria amount to $14.5 \cdot 10^6$ t of oil and condensate ($10^7 \cdot 106$ bbls) and $19.6 \cdot 10^9$ m^3 gas (0.73 TCF). Oil production peaked in 1955 with 3.7 t/year ($23 \cdot 10^6$ bbls/year) and gas production in 1978 with $2 \cdot 10^9$ m^3/year (73 BCF/year). In 1994 total oil and condensate production amounted to $1.2 \cdot 10^6$ t and total gas production to $1.5 \cdot 10^9$ m^3. Both oil and gas production are presently slightly declining.

Following the exploration successes in the Molasse Basin and the substratum of the Vienna Basin, interest in exploration of the Austrian Alps gradually increased during the late 1950's. Initial exploration efforts were directed at the autochthonous Mesozoic and Paleogene strata but later also at the series of the Calcareous Alps. In 1959 the well Texing-1 was the first well which spudded in the Flysch Zone and reached the autochthonous Molasse. In 1966 the first well drilled in the Calcareous Alps, Urmannsau-1, was located in a tectonic window in the vicinity of an oil seep and bottomed at 3033 m in the autochthonous Molasse. Both wells failed to discover hydrocarbons but were in so far important as they proved that the Calcareous Alps, Flysch, Klippenbelt and Helvetic Zone were thrusted over autochthonous Tertiary and Mesozoic sediments covering the gently southwards dipping basement of the foreland. By now 32 exploration wells (Fig. 1) and 10 production wells (gas-condensate field Höflein) have been drilled in the Alpine allochthon; additional wells are located in the imbricated Sub-Alpine Molasses.

GEOLOGICAL SETTING

Figure 1 provides an overview of the structural framework of Alpine Austria and its foreland. The Calcareous Alps (Kalkalpen), forming the orogenic lid of the Alpine stack of nappes, consist of the lower, middle and upper Austroalpine nappes; these were derived from the southern margin of the South Pennininc trough. During the Cretaceous gradual closure of the Alpine Tethys, these nappes were transported northwestwards (Ring et al., 1989; Eisbacher et al., 1990; Froitzheim et al., 1994; Neubauer, 1994). During the Senonian and Paleocene, increasing collisional coupling between the rising Alpine orogen and its European foreland is reflected by compressional deformation of the latter, resulting in basin inversion and upthrusting of basement blocks forming the Bohemian Massif (Ziegler, 1990). During the Paleogene, the Austroalpine nappes, together with the Pennininc and Helvetic nappes, advanced northwards and overrode the European foreland (Ratschbacher et al., 1991), causing the flexural subsidence of the Molasse Basin. By late Oligocene-early Miocene

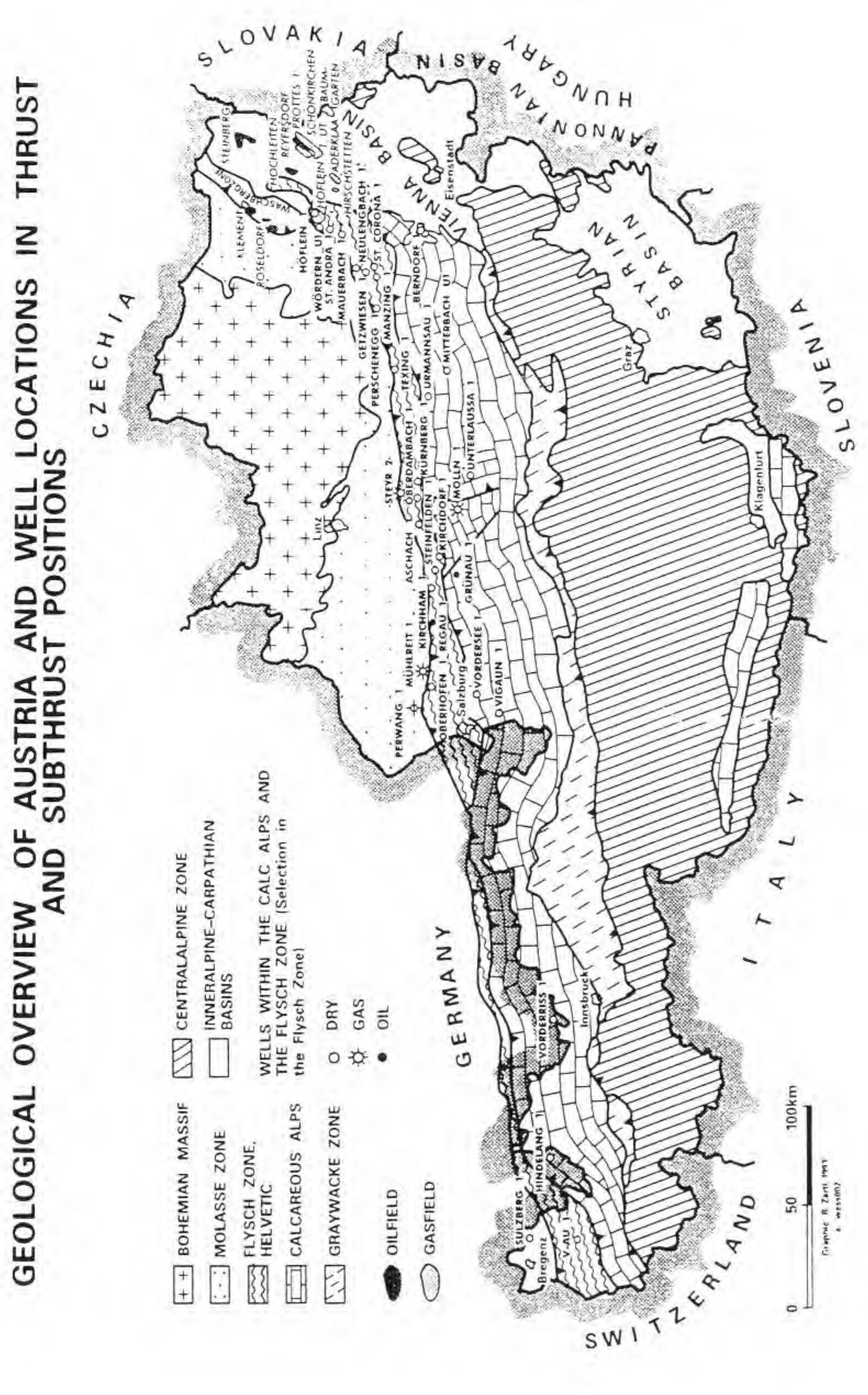

GEOLOGICAL OVERVIEW OF AUSTRIA AND WELL LOCATIONS IN THRUST AND SUBTHRUST POSITIONS
BOHEMIAN MASSIF
MOLASSE ZONE
FLYSCH ZONE, HELVETIC
CALCAREOUS ALPS
GRAYWACKE ZONE
OILFIELD
GASFIELD
CENTRALALPINE ZONE
INNERALPINE-CARPATHIAN BASINS
WELLS WITHIN THE CALC ALPS AND THE FLYSCH ZONE (Selection in the Flysch Zone)
DRY
GAS
OIL
CZECHIA
SLOVAKIA
VIENNA BASIN
PANNONIAN HUNGARY BASIN
STYRIAN BASIN
SLOVENIA
ITALY
SWITZERLAND
GERMANY
Linz
Salzburg
Innsbruck
Graz
Klagenfurt
Eisenstadt
Bregenz
0 50 100km

FIG. 1.

times, the entire nappe system was emplaced near its present position. While thrust activity persisted into late Miocene times in Lower Austria, early Miocene series seal the thrust front in Upper Austria. Results of deep exploration wells and reflection seismic data show that the European foreland crust dips gently southwards under the Alpine nappes and extends at least 65 km to the south of the present Alpine thrust front (Encl. 1).

Hydrocarbon exploration in Alpine Austria is restricted to the Calcareous Alps, the Flysch and Helvetic Zones and the sub-Alpine Molasse. Potential prospects occur in the the sub-thrust autochthonous units and within the allochthonous units.

Autochthonous Plays

The hydrocarbon habitat in autochthonous sub-thrust sedimentary series can be extrapolated from the updip, northward adjacent Molasse Basin in which a large number of oil and gas fields has been established. These produce variably from Jurassic sandstones and carbonates, Cretaceous and Eocene sands and Oligocene and early Miocene sands of the Molasse sequence (Fig. 7; Kollmann and Malzer, 1980; Kröll, 1980a).

Productive structures, involving Eocene and older reservoirs, are controlled by antithetic and to a lesser degree by synthetic normal faults; these developed during the Oligocene thrust-loaded rapid subsidence of the Alpine foreland. Such faults have generally throws of the order of 100 to 300 m. There are occasional examples of Oligocene extensional faults which were compressionally reactivated during the late phases of the Alpine orogeny. Only the well Perwang-1 encountered imbrications of Eocene and Late Cretaceous autochthonous series. Traps controlled by pre-Tertiary faults play a subordinate role. Oil and oil/gas accumulations are essentially restricted to Eocene and older reservoirs. Oligocene and Miocene sands contain in stratigraphic and differential compaction structures biogenic gas.

Along the strike of the Molasse Basin, the thickness and composition of Mesozoic and Paleogene strata varies considerably. This can be attributed mainly to the basal onlap geometry of Mesozoic strata against the Variscan basement, the regional base-Cretaceous and basal Tertiary unconformities, and the onlap geometry of the Eocene and Oligocene sediments against the palaeo-relief of the basal Tertiary erosional surface. These lateral variations have a strong bearing on the availability of reservoirs and source-rocks (Figs. 2 and 3).

West of Vienna, a spur of the Bohemian Massif projects deeply under the Alpine nappes. In this area transgressive Oligocene sands and shales rest directly on basement. The southwestern flank of this palaeo-high is controlled by a major fault (Steyer fault).

To the east of this spur, Permo-Carboniferous clastics are unconformably overlain by Middle Jurassic paralic and deltaic sands, involved in rotational, extensional fault blocks (Fig. 3). These are sealed by Late Jurassic carbonates which grade laterally into basinal shales having an excellent source-rock potential (Ladwein, 1988). The Jurassic sequence, which terminates with regressive Tithonian carbonates, attains thicknesses of up to 2000 m. It is regionally truncated by an Early Cretaceous unconformity which is related to wrench-deformations of the Bohemian Massif. Late Cretaceous sands carbonates and marls, up to 900 m thick, are truncated by a second regional unconformity which is related to compressional deformations of the Bohemian Massif during early Paleocene times. Oligocene transgressive sands overstep this erosional surface and are in turn overlain by a Miocene shaly sequence. Jurassic carbonates and sands form the reservoirs of several oil and gas accumulations. Oligocene and Miocene sands contain biogenic gas accumulations (Brix et al., 1977; Kröll, 1980a; Wessely, 1987).

To the west of the Bohemian basement spur, thin Middle Jurassic sands rest on basement and are conformably overlain by Late Jurassic carbonates attaining maximum thicknesses of some 750 m (Fig. 3). Rapid lateral thickness changes are controlled by the basal Cretaceous unconformity and associated faulting. Local porosity developments are related to fracturing and karstification of the partly reefal Jurassic carbonates. Sedimentation resumed with the transgression of Apto-Albian marine sands. Cenomanian glauconitic sands attain thicknesses of 75 m and form, together with Middle Jurassic sands, the deeper reservoir of the

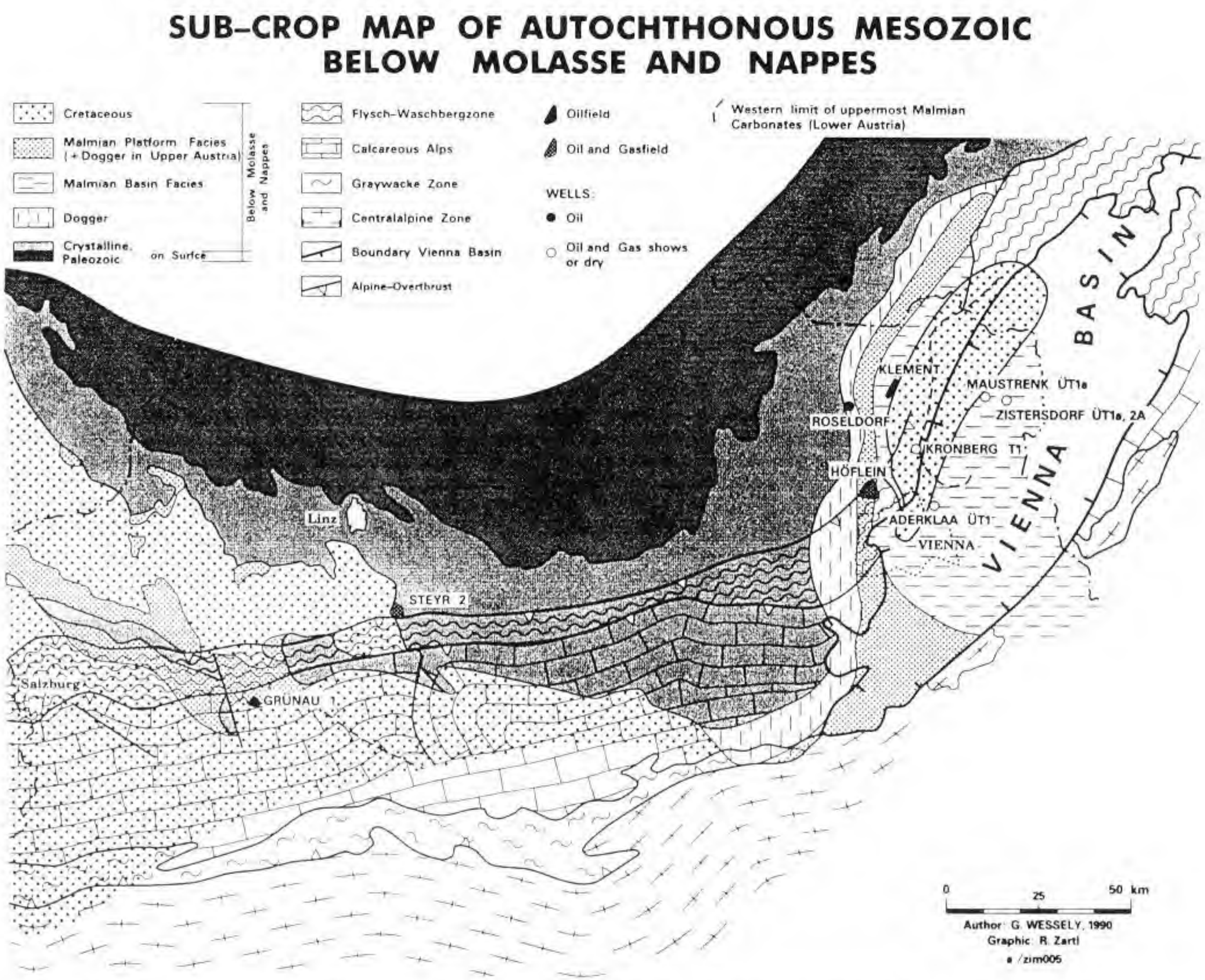


FIG. 2. The subcrop pattern of Mesozoic and older series beneath the base-Tertiary unconformity bears no relationship with the strike of the Alpine nappe systems. The map is closely constrained by wells and seimic data in the Molasse Basin and becomes progressively more hypothetical under the Calcareous Alps.

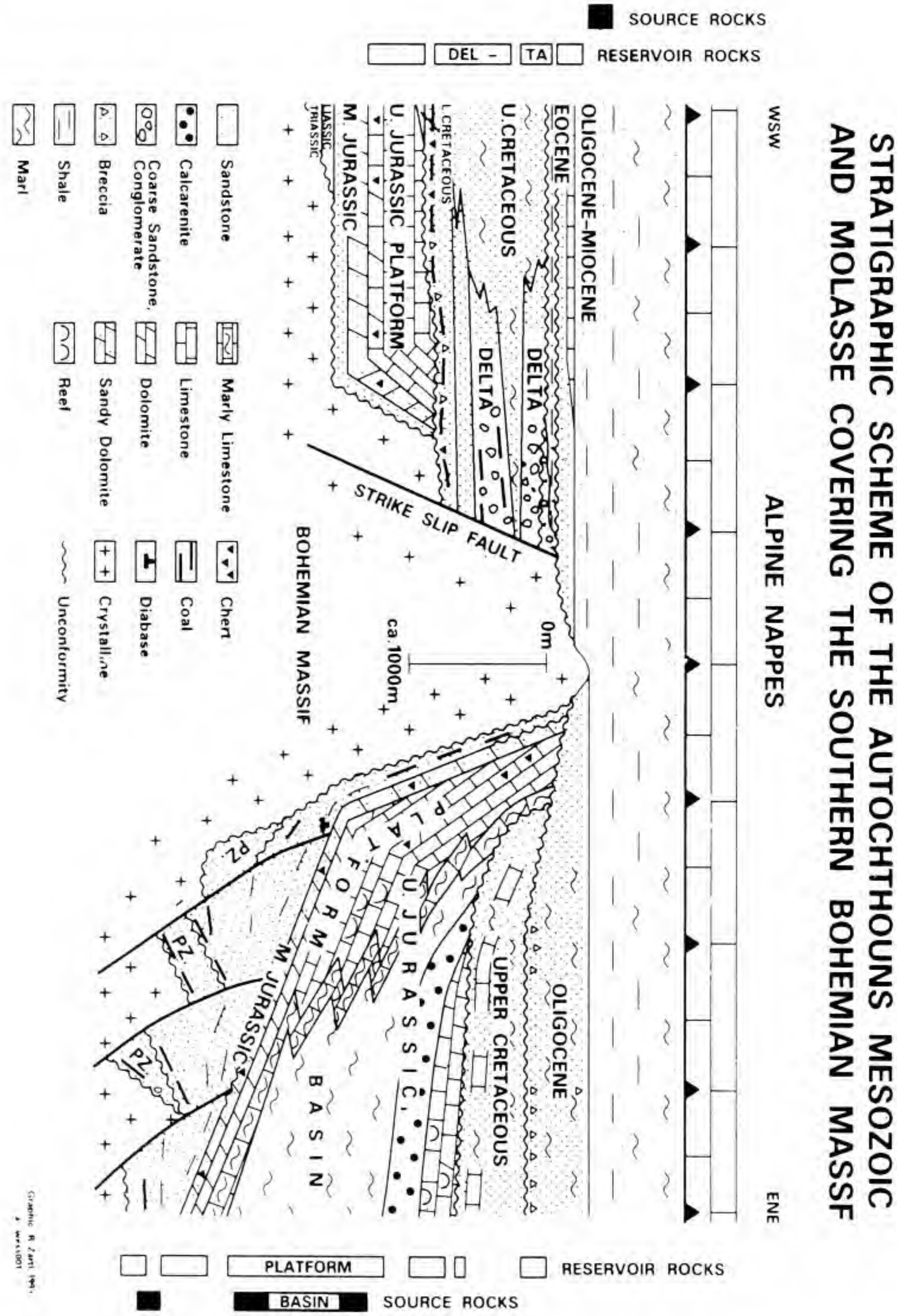

STRATIGRAPHIC SCHEME OF THE AUTOCHTHOUNS MESOZOIC
AND MOLASSE COVERING THE SOUTHERN BOHEMIAN MASSF
SOURCE ROCKS
DEL -
TA
RESERVOIR ROCKS
WSW
ENE
ALPINE NAPPES
OLIGOCENE-MIOCENE
EOCENE
U.CRETACEOUS
L.CRETACEOUS
U. JURASSIC PLATFORM
M. JURASSIC
LIASSIC
TRIASSIC
DELTA
DELTA
STRIKE SLIP FAULT
BOHEMIAN MASSIF
0m
ca.1000m
OLIGOCENE
UPPER CRETACEOUS
U. JURASSIC
PLATFORM
M. JURASSIC
BASIN
PZ.
PZ.
PZ.
PLATFORM
RESERVOIR ROCKS
BASIN
SOURCE ROCKS
Sandstone
Calcarenite
Coarse Sandstone, Conglomerate
Breccia
Shale
Marl
Marly Limestone
Limestone
Dolomite
Sandy Dolomite
Reef
Chert
Coal
Diabase
Crystalline
Unconformity

FIG. 3.

Voitsdorf field which is the largest oil accumulation of Upper Austria. These sands are capped by Turonian and Senonian clays which grade towards the Bohemian basement spur upwards into sands derived from the Bohemian Massif. Late Cretaceous sediments reach maximum thicknesses of about 1000 m. Transpressional deformations during the Paleocene resulted in a profound disruption of the Late Cretaceous shelf series and in erosion cutting locally down through Jurassic series into the basement. During the late Eocene, the Early Tertiary erosional surface was overstepped by fluvial and shallow marine sands which grade upwards into Lithothamnimum limestones. These Eocene sands form an important reservoir for oil accumulations. Rapid deepening of the area at the transition from the Eocene to the Oligocene was accompanied by the deposition of the highly organic Fish-shales, which constitute the primary source-rock for the oil accumulations of Upper Austria and Salzburg and the adjacent area of Bavaria . This rapid deepening phase was accompanied by the development on an array of essentially basin-parallel, trap-providing antithetic and synthetic normal faults. During the Oligocene, influx of coarse clastics from the advancing Alpine nappe system gave rise to the accumulation of the turbiditic Puchkirchen conglomerate and sand fans. Their southern parts were overridden during the latest Oligocene-early Miocene emplacement of the Alpine nappes, resulting in a narrowing of the Molasse basin. The middle and late Miocene Hall and Innviertel series were deposited under upwards shallowing conditions. Deep-water sands of the Puchkirchen and Hall series are charges by biogenic gas that is essentially stratigraphically trapped (Kollmann and Malzer, 1980; Polesny, 1983; Nachtmann and Wagner, 1987).

In the western parts of the German Molasse Basin, the Variscan basement is overlain by Triassic sediments, a complete sequence of Early and Middle Jurassic strata and progressively northwards truncated Late Jurassic carbonates. Cretaceous sediments have a limited distribution in the eastern part of the Bavarian Molasse basin but are thought to be more widespread under the Alpine nappes. The effects of the basal Cretaceous and Paleogene unconformities appear to be less intense than in the Austrian Molasse Basin. Although occurrence of the transgressive late Eocene sands and Lithothaminium limestones is restricted to the southern parts of the Bavarian Molasse Basin, these are likely to be present in the autochthon of the Alpine nappes. The same applies for the early Oligocene Fish-shales source-rock. Additional potential source-rocks are likely to occur in Middle Triassic carbonates and the pelagic facies to Late Jurassic carbonates (Bachmann et al., 1987; Bachmann and Müller, 1991; Bachmann and Roeder, this volume).

The subcop pattern of Mesozoic strata beneath the basal Tertiary unconformity (Fig. 2) illustrates that also in a subthrust position the distribution of both reservoir and source-rocks is highly variable. This pertains also to the distribution of Eocene reservoirs and the basal Oligocene source-rock, as illustrated by the results of the well Berndorf-1 which drilled through the nappes of the Calcareous Alps, 40 km south of the Alpine thrust front, and bottomed at 6028 m in Variscan basement after penetrating a 35 m thick autochthonous Oligocene conglomerate (Encl. 1; Wachtel and Wessely, 1981). On the other hand, rapid lateral changes can be expected across Early Cretaceous and Paleocene faults, as evident by the results of the wells Molln-1 and Grünau-1, drilled 35 km apart (Fig. 4).

Oligocene source-rocks attain maturity for oil generation beneath the frontal parts of the Alpine nappes and probably enter the gas window beneath the internal parts of the Calcareous Alps. With increasing overburden, clastic reservoirs loose their porosities. Correspondingly, the sub-thrust play has to contend with major reservoir and source rock prediction uncertainties. Potential sub-thrust traps are fault-bounded blocks having similar dimensions as in the Molasse Basin, possible compressionally reactivated Oligocene tensional structures and broad arches (Encl. 1).

Allochthonous Plays

In the Eastern Alps, the wegde-shaped Alpine system of nappes (Fig. 5) was thrusted by at least 100 km over the autochthonous floor, formed by the European foreland (Fig. 6). The complex structure and stratigraphy of these nappes resulted from multiple deformation phases, spanning Jurassic to

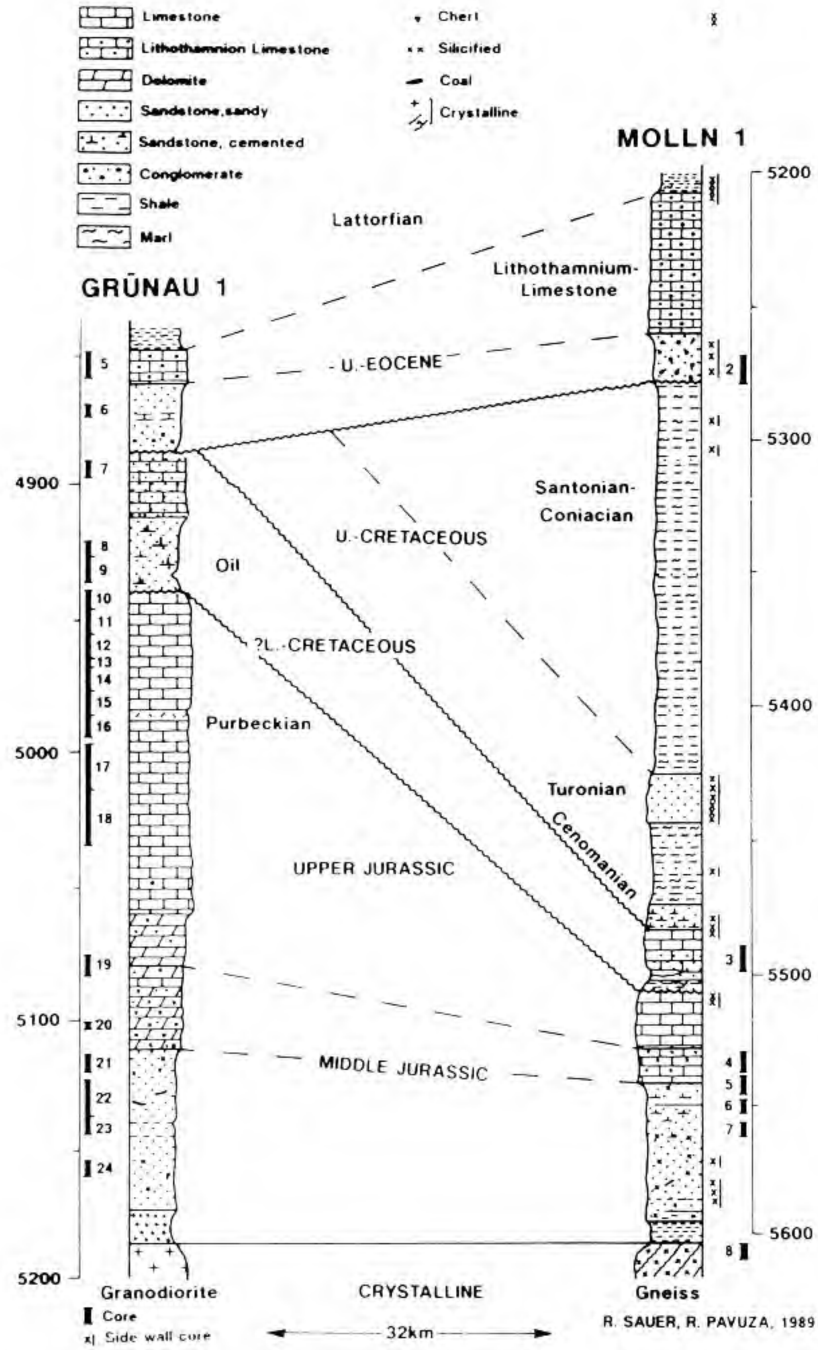


FIG. 4. Correlation of Mesozoic and basal Tertiary series penetrated by wells Grünau-1 and Molln-1, showing effects of base-Cretaceous and Paleocene foreland tectonics.

early Miocene times. These include Triassic to Early Cretaceous rifting events, culminating in the opening of oceanic basins, followed by the onset of subduction processes during the Neocomian and the Late Cretaceous collision of the Alpine orogenic wedge with the European foreland (Tollmann, 1973; Flügel and Faupl, 1987). Collision of the African and European plates controlled the emplacement of the Alpine nappe systems and contemporaneous deformation of the Alpine foreland far to the north of the Alpine thrust front (Ziegler, 1987).

Remnants of the European passive margin sedimentary prism are represented by the Helvetic and the Klippen zones (Fig. 5). The Penninic zone, corresponding to the central Tethyan region, outcrops mainly in the Western and Central Alps and is exposed in the Eastern Alps only in the Unterengadin, Tauern and Rechnitz windows. This zone comprises Palaeozoic to Mesozoic and Paleocene metasedimentary and crystalline rocks, as well as slices the oceanic crust which had formed during the Jurassic and Cretaceous opening of the Tethys (Janoschek and Matura, 1980). The Fysch Zone, which forms part of the Penninic domain, outcrops in a band paralleling the Alpine deformation front.

The Austoalpine nappes were derived from domains located to the south of the Penninic zone, corresponding to the Italo-Dinarid block (Frisch, 1979; Ziegler et al., this volume). They consist of crystalline basement and its Palaeozoic and Mesozoic sedimentary cover. The Lower and Middle Austoalpine units formed, palinspastically speaking, the southern margin of the Penninic zone and are characterized by a very low-grade metamorphic Permo-Mesozoic facies belt. Further to the south, the Upper Austoalpine units include the sedimentary sequences of the Palaeozoic Grauwacken Zone and the unmetamorphosed Permo-Mesozoic and Paleogene sediments of the Calcareous Alps.

During the Cretaceous phases of the Alpine orogeny, the Austroalpine nappes developed and during the Paleogene they moved across the Penninic zone. The Upper Austroalpine nappes of the Calcaresous Alps and the Grauwacken Zone overrode the Lower and Middle Austroalpine nappes and now rest rootless on them and the Penninic flysch. During the Oligocene and early Miocene the entire stack of nappes advance further to the north and was thrusted over the southern, proximal parts of the Molasse foreland basin.

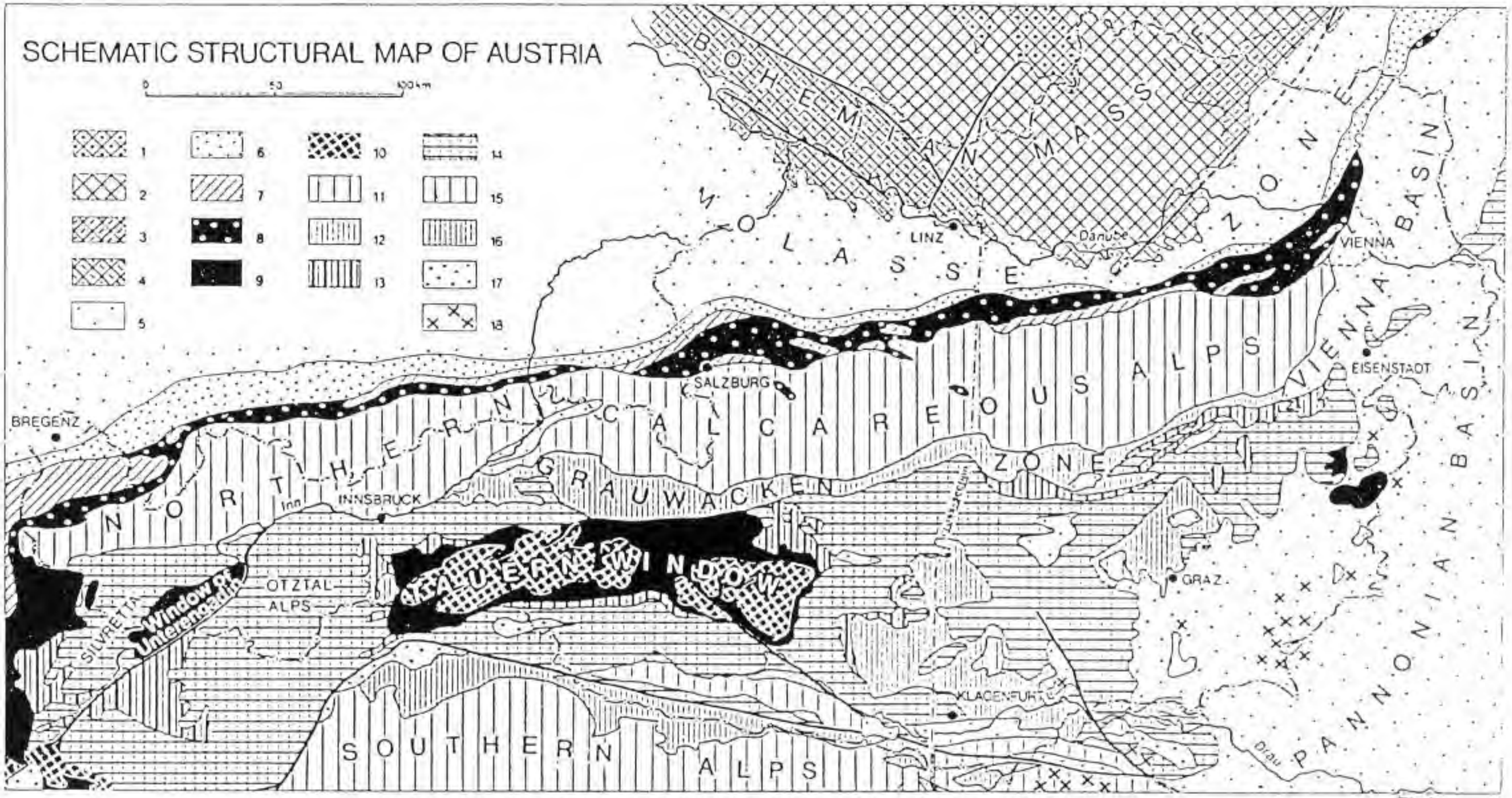
Exploration for hydrocarbons is concentrated on the deformed zone of the Molasse Basin as well as on the allochthonous Helvetic, Flysch and Calacreous Alps zones.

The stratigraphy of the Calcareous Alpine nappes, largely derived from outcrops, shows major vertical and lateral variations. Nevertheless, long-distance facies trends can be established for the main tectono-stratigraphic units of the Calcareous Alps, corresponding to the Bajuvaricum, Tirolicum and Juvavicum, and the Flysch and Helvetic zones (Fig. 7).

Nappes of the Calcareous Alps involve a sedimentary sequence ranging in age from Permo-Scythian to Paleocene. Middle and Late Triassic platform carbonates attain thicknesses in the order of several 1000 m. Late Triassic series are characterized by lateral changes from lagoonal to reefal carbonates and basinal facies partly having source-rock characteristics. Jurassic series are dominated by platform carbonates and basinal shales and carbonates. Cretaceous to Paleogene strata are dominantly developed in a clastic facies.

The fractured Middle and Late Triassic Wetterstein and Hauptdolomite, having low matrix porosities, present potential reservoirs. Limestones constitute neither reservoirs nor seals. Potential seals are provided by Permo-Scythian shales and evaporites, shales and tight sandstones of the Late Triassic Lunz formation and particularly by the Cretaceous to Paleocene Gosau group. Basinal shales and carbonates of Middle Triassic, Rhaetian and Early Jurassic age are partly characterized by elevated TOC values. Maturation of these potential source-rocks depends on their position within the nappe stack but is generally insufficient for the generation of oils.

In the Flysch Zone of the Austrian Alps, to date no hydrocarbon accumulations have been found, mainly due to a lack of porosity. In the western Helvetic Zone, reservoir development can be expected in Middle Jurassic sandstones, the Late Jurassic Quinten Limestone (Müller, 1985a and 1985b) and the Early Cretaceous Schratten Limestone, provided they were intensely enough fractured during tectonic deformation (Müller et al., 1992).



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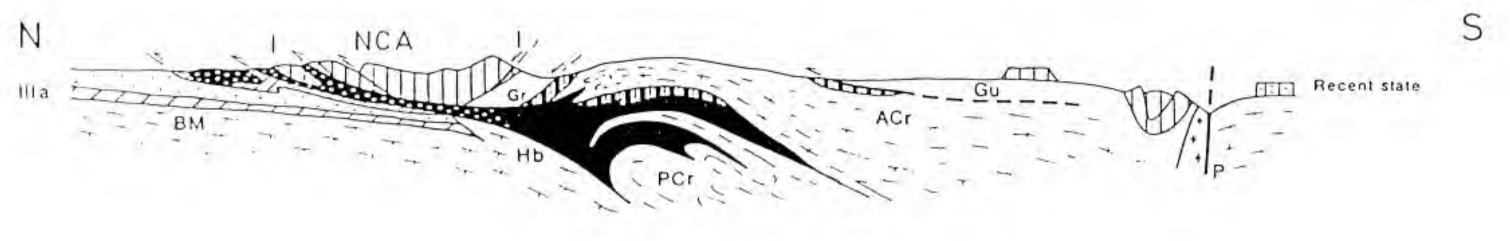
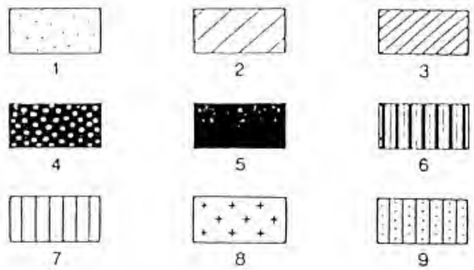
1-4 = Bohemian Massif: 1 = Post - Variscan sedimentary cover; 2 = Moldanubian Zone; 3 = Moravian Zone; 4 = Bavarian Zone; 5 = Tertiary basins; 6 = Subalpine Molasse; 7 = Helvetic and Klippen Zone; 8 = Flysch Zone; 9 = Metasedimentary rocks of the Penninic Zone; 10 = Crystalline basement of the Penninic Zone; 11-14 – Austro - Alpine Unit; 11 = Permomesozoic in Kalk - Alpine facies; 12 = Palaeozoic; 13 = Permomesozoic in Central Alpine facies; 14 = Crystalline basement ("Altkristallin"); 15 = Permomesozoic of the Southern Alps; 16 = Palaeozoic of the Southern Alps; 17 = Periadriatic intrusive masses; 18 = Neogene andesites and basalts.

FIG. 5.

SCHEMATIC CROSS - SECTION THROUGH THE EASTERN ALPS





1 = Tertiary rocks of the Molasse Zone; 2 = Extra - Alpine post - Variscan sedimentary rocks; 3 = Helvetic Zone and Klippen Zone; 4 = Flysch Zone; 5 = Permomesozoic of the Penninic Zone; 6 = Permomesozoic of the Central - Alpine facies belt; 7 = Permomesozoic of the Kal.. - Alpine facies belt; 8 = Periadriatic Intrusion; 9 = South Alpine facies; BM = Bohemian Massif; Hb = Basement of the Helvetic Zone; PCr = Crystalline Basement of the Penninic Zone; Gr = Palaeozoic rocks of the Grauwackenzone; Gu = Palaeozoic rocks of the Gurktal Sheet; ACr = Crystalline basement of the Austro - Alpine Unit; P = Periadriatic Lineament.

FIG. 6.

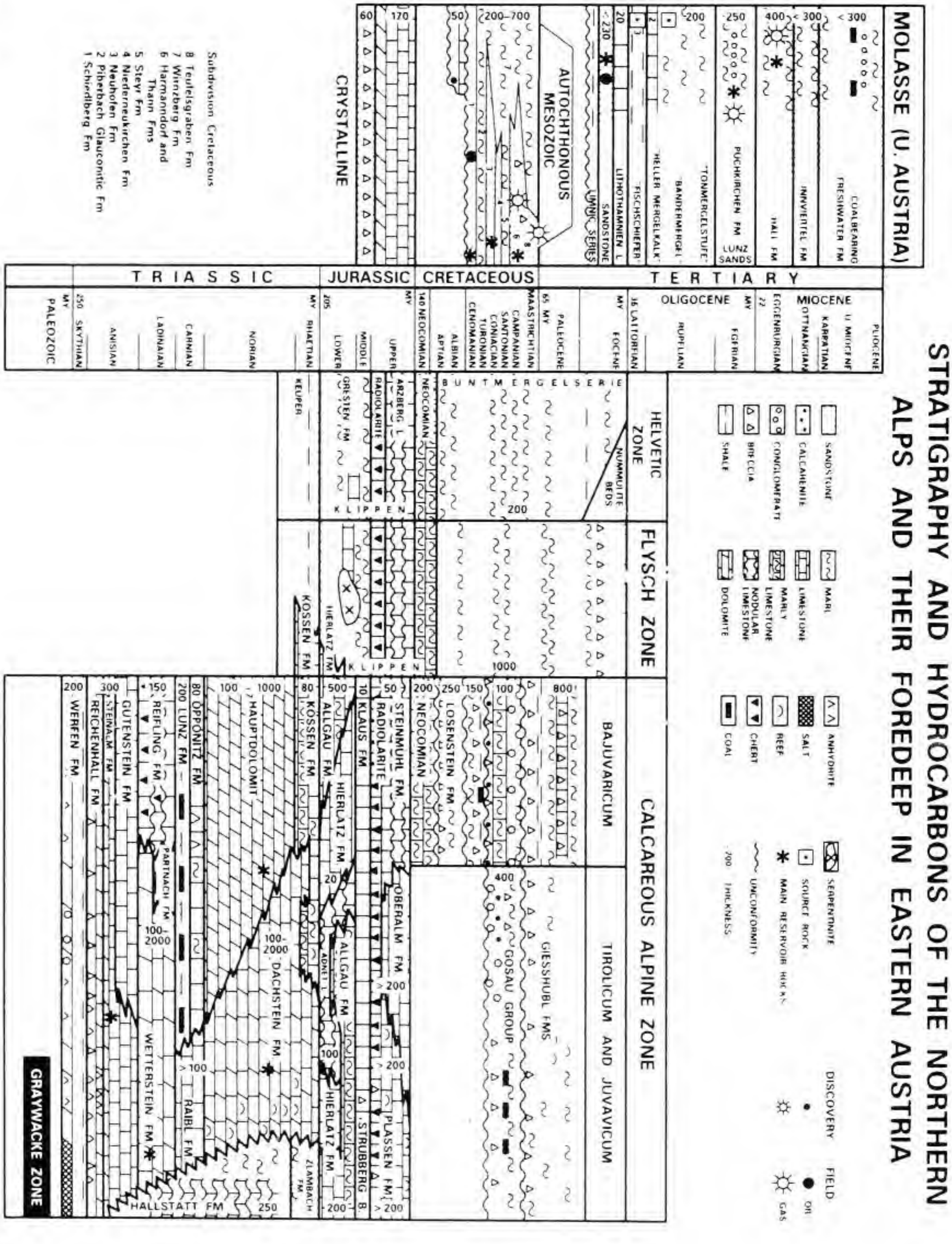

STRATIGRAPHY AND HYDROCARBONS OF THE NORTHERN ALPS AND THEIR FOREDEEP IN EASTERN AUSTRIA
MOLASSE (U. AUSTRIA)
HELVETIC ZONE
FLYSCH ZONE
CALCAREOUS ALPINE ZONE
BAJUVARICUM
TIROLICUM AND JUVAVICUM
GRAYWACKE ZONE
AUTOCHTHONOUS MESOZOIC
CRYSTALLINE
TERTIARY
CRETACEOUS
JURASSIC
TRIASSIC
Subdivision Cretaceous
8 Teufelsgraben Fm
7 Wirnzberg Fm
6 Harmannsdorf and Thann Fms
5 Steyr Fm
4 Niederneukirchen Fm
3 Neuhofen Fm
2 Piberbach Glauconitic Fm
1 Schiedlberg Fm

FIG. 7.

Within the Calc-Alpine units, potential structural traps were formed during their folding and thrusting. Unconformities, sealed by transgressive Late Cretaceous and Paleocene Gosau sediments, may provide for combination structural/stratigraphic traps (Fig. 9). Discovery of a number of oil and gas accumulations in Calc-Alpine units subcropping the Vienna basin has demonstrated that the potential of such structural and combined structural/stratigraphic traps (Kröll, 1980b). Hydrocarbon charge to such traps is provided by the autochthonous Oligocene Fish-shale west of the Bohemian basement spur and by basinal Late Jurassic shales in the Vienna Basin (Ladwein, 1988) and possibly also by source-rocks contained in the allochthonous units.

EXPLORATION RESULTS

During the exploration of the Austrian Alps for hydrocarbons large gravity and magnetic surveys were carried out. Extensive surface geological mapping and structural analyses (Tollmann, 1976a, 1976b, 1985; Oberhauser, 1980) were followd by the acquisition of 5000 km of 2D reflection-seismic lines and the drilling of 32 exploration wells. In two areas 3D seismic surveys were recorded (Geutebrück et al., 1984).

In many areas the results of reflection-seismic surveys are not conclusive. Often the autochthonous section below the Alpine nappes gives rise to better and more continuous reflections than the allochthonous units. This is particularly true for areas where the latter are characterized by steep dips. However, in areas where the allochthonous units display relatively low dips, their internal configuration can be resolved by the seismic tool, as seen, for instance, on the line given in Fig. 8 which was recorded in the area of Salzburg (Kröll et al., 1981). The 3D survey over the gas/condensate field Höflein was very successful, despite difficult terrain, surface geological and environmental conditions; it involved the recording of an irregular grid using a mixed vibroseis and dynamite source.

Drilling activity involved the drilling of 24 exploratory wells in the Flysch Zone (16 ÖMV, 8 RAG) and 8 wells (ÖMV) in the Calcareous Alps. All wells drilled in the Flysch Zone bottomed in crystalline basement with some encountering an autochthonous Mesozoic sequence, albeit with variable thickness and of different composition. Of the 8 wells drilled in the Calcareous Alps, 4 wells bottomed in crystalline basement; of these, the wells Molln-1 and Grünau-1 penetrated a thick sequence of autochthonous Eocene and Mesozoic sediments (Fig. 4). The remaining four wells drilled in the Calcareous Alps terminated within allochthonous units. In the western Helvetic Zone of Voralberg, the well Au-1 was drilled to investigate the potential of the Jurassic sequence (see Fig. 1 and Encl. 1).

Exploration wells, which in combination with geophysical data, permitted to construct the series of structural cross-sections through the northern parts of the Alps, given in Encl. 1, are the Austrian wells Höflein-1 (Grün, 1984), Berndorf-1 (Wachtel and Wessely, 1981), Urmannsau-1 (Kröll and Wessely, 1967), Mittelbach-U1, Molln-1 and Grünau-1 (Wessely, 1988; Hamilton, 1989), Vordersee-1 (Geutebrück et al., 1984) and Vorarlberg Au-1 (Colins et al., 1990) and the German wells Vorderriss-1 (Bachmann and Müller, 1981) and Hindelang-1 (Müller et al., 1992).

According to the results of the well Berndorf-1, the nappe system of the Eastern Alps were transported during late Oligocene-early Miocene times northward over a distance of at least 40 km over the the autochthonous basement and its sedimentary cover. As such the autochthonous Mesozoic series and the lower part of the Tertiary fill of the Molasse Basin were protected from erosion. At the same time source-rocks were buried to sufficient depth to generate hydrocarbons whereas diagenetic processes accounted for a significant porosity and permeability reduction in reservoir rocks. The foreland basement dips very gently under the Calcareous Alps. The basement spur of the Bohemian Massif corresponds to an axial culmination.

The Flysch and Helvetic zones form thrust wedges in front of the Austroalpine nappes and thin out beneath them or are missing completely. Towards the West, the Helvetic Zone gains in thickness and becomes more complete. Serpentinites, encountered in the well Grünau, indicate

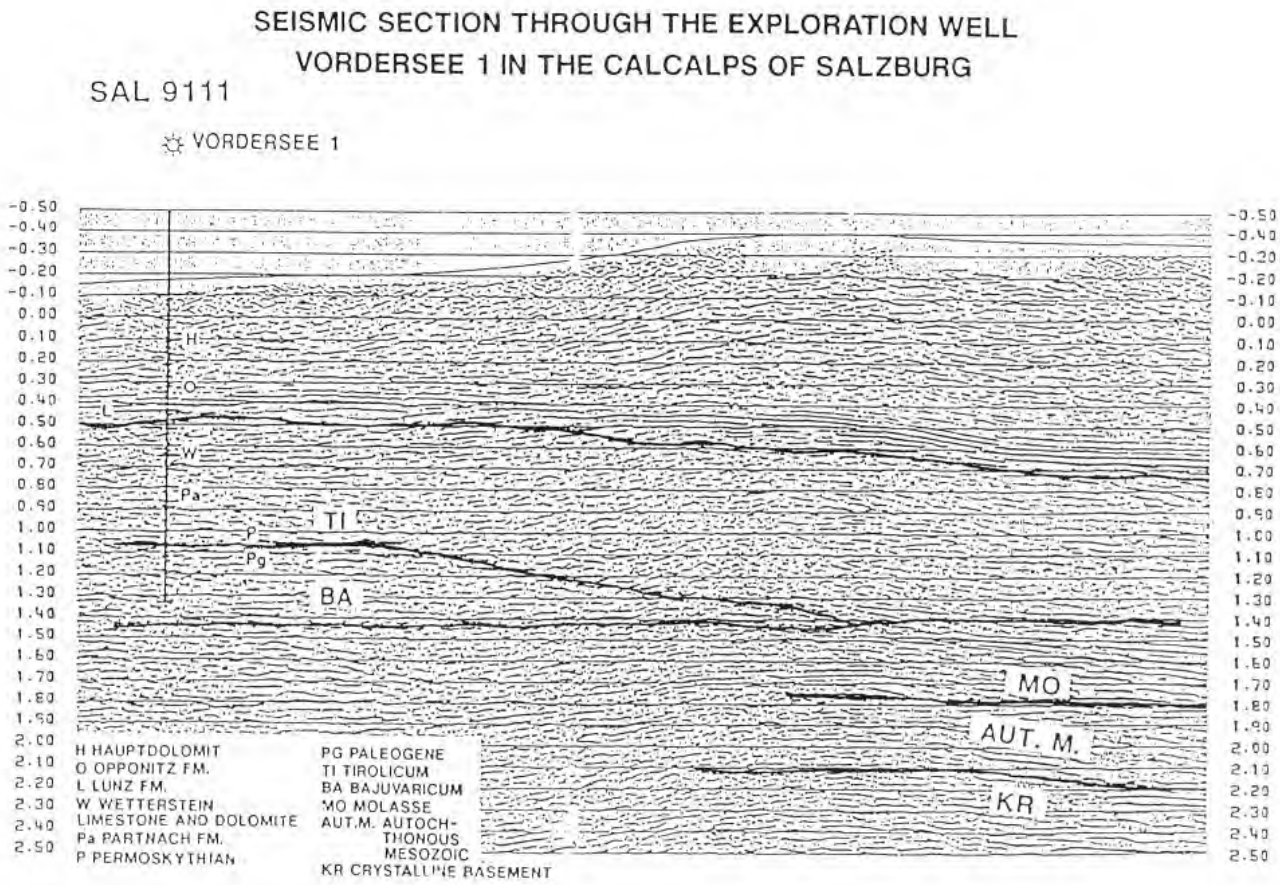


FIG. 8. For location of well Vordersee 1, see Fig 1. The well reached a total depth of 4264 m in the Bajuvaricum which thins to the south of the well.

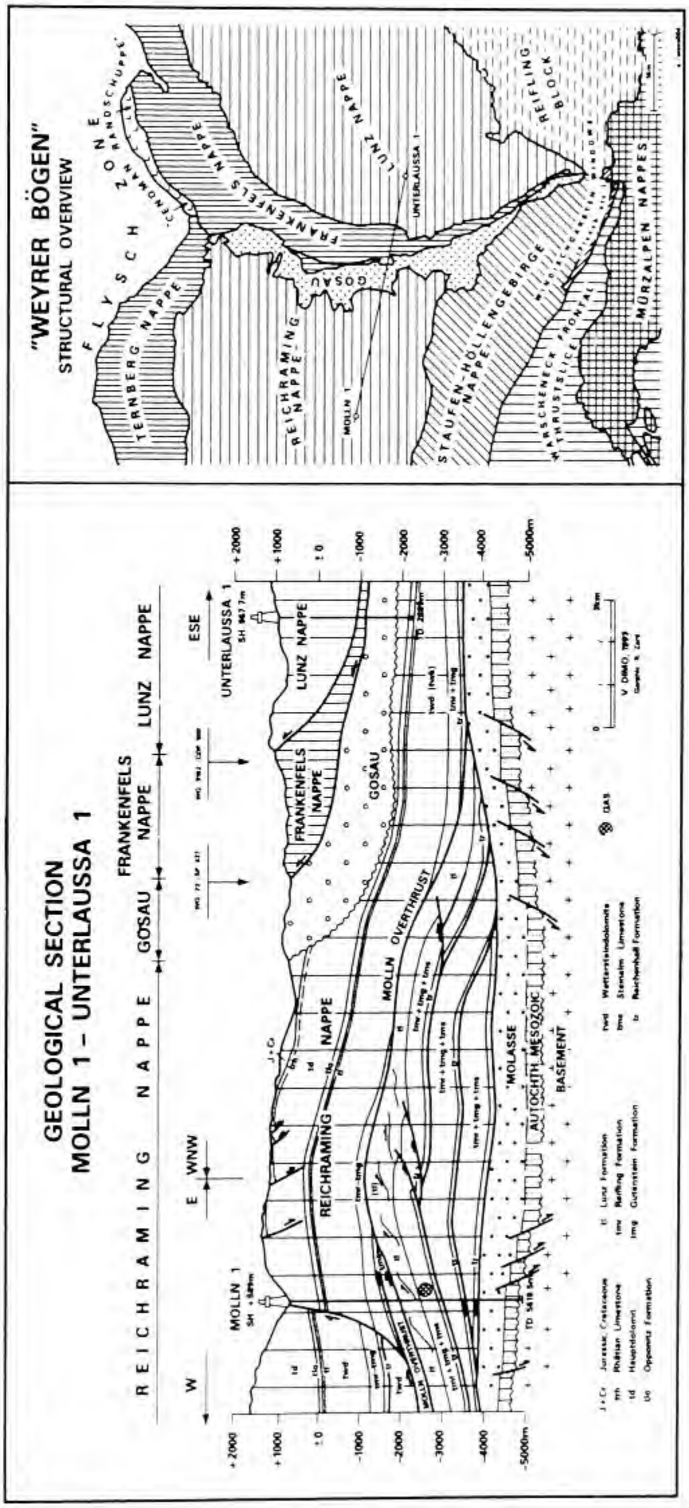


FIG. 9. for explanations see text

that the Flysch Zone is of North Penninic origin. In front of the Flysch Zone, the Molasse is folded and thrusted. This Sub-Alpine Molasse is involved in triangle zones in the West. On the Bohemian basement spur the Sub-Alpine Molasse forms a narrow belt, whereas towards the northeast, north of the Danube, the so-called Waschberg Zone consists of imbricated Molasse and Mesozoic series (Brix et al., 1977; Kröll, 1980a).

The Bajuvaricum, Tirolicum and Juvavicum units of the Calcareous Alps, which all form part of the Upper Austroalpine nappe system, display along strike dramatic changes in stratigraphic content, thickness and facies development and, consequently, in their structural style. Where Triassic platform carbonates are thin, folding plays a more important role than in areas of thick carbonates which are dominated by a relatively flat lying stacked thrust sheets.

The structural complexity of the Calcareous Alps is illustrated in Fig. 9 which crosses the "Weyrer Bögen" area, a transverse feature in the central parts of the Calcareous Alps. This structure originated by rotation, westward thrusting and duplication of two Bajuvaricum units. The lower, non rotated element is formed by the Ternberg and Reichraming nappes, corresponding to the upper, rotated element to the Frankenfels and Lunz nappes. A thick sequence of Cretaceous Gosau sediments, resting unconformably on the Reichraming nappe separates the latter from the Frankenfels and Lunz nappes. The well Unterlaussa-1 penetrated the Gosau sequence and encountered beneath it tight Triassic rocks of the Reichraming nappes. The objective of this well was to test a combined structural/stratigraphic prospect that is analogous to some of the gas accumulations occurring beneath the Vienna Basin. The neighbouring well Molln-1 encountered gas in the Reichraming nappe, thus proving the availability of hydrocarbons in the area.

HYDROCARBON DISCOVERIES IN THE ALPS

The first oil was discovered in sub-thrust autochthonous Eocene sediments by the well Kirchham-1, drilled in the Flysch Zone of Upper Austria. A much larger, and commercially exploitable accumulation is the Höflein gas/condensate field located near Vienna which contains ultimate recoverable reserves of some $7 \cdot 10^9$ m^3 (250 BCF) This field is contained in autochthonous Middle Jurassic dolomitic and cherty sandstone and deltaic sandstones (Sauer et al., 1992) involved in a horst block which was overridden by Flysch nappes (Encl. 1). Reservoir pressures are hydrostatic; the gas contains a low percentage of CO_2.

Grünau-1 was the first well spudded in the Calcareous Alpine nappes which encountered an autochthonous Mesozoic sequence and discovered at a depth of more than 4800 m light oil in overpressured Early Cretaceous sandstones (Fig. 4). Initial flow rates of more than 750 bbls/day declined, however, rapidly and the well was abandoned. Similarly, the well Kirchdorf-1, drilled north of Grünau-1, tested uncommercial quantities of oil.

Within the Sub-Alpine Molasse of Upper Austria, the well Mühlreit-1, drilled by RAG, produced considerable amounts of oil from overpressured Oligocene sandstones before being abandoned. In the Helvetic Zone of Vorarlberg, several intervals of Middle Jurassic sandstones and Late Jurassic carbonates yielded on test only gas shows and salt water. For instance, the German well Hindelang-1, which penetrated a long section within the Helvetic Zone, tested from a 300 m interval in the Early Cretaceous Schrattenkalk gas at flow rates of 3.7 MMCFF/day; after prolonged testing also this well was abandoned.

In well Urmannsau-1, located in the Calcareous Alps of Lower Austria, many oil shows were observed; however, on test fractured Middle Triassic dolomites yielded only salt water. Whether this reservoir is oil bearing at a structurally higher location is unknown.

The well Vordersee-1, drilled southeast of Salzburg in the more simply structured part of the Calcareous Alps, tested from Middle Triassic dolomites, sealed by the shales and tight sandstones of the Lunz formation, salt water only.

In contrast, the well Berndorf-1, drilled near the western border of the Vienna Basin, penetrated a very thick, complex sequence of Middle and Late Triassic carbonates from which fresh water with a maximum temperature of 45°C was tested down to a depth of 4500 m. In combination with formation water analyses from the Vienna Basin, this indicates that meteoric waters can penetrate to great depths in the Calcareous Alps, setting up a complex and very active hydrodynamic system.

CONCLUSIONS

A distinction has to be made between the sub-thrust autochthonous play and plays aimed at allochthonous prospects.

Sub-thrust autochthonous prospects are located in a depth range of 3000 to more than 6000 m. This play has to contend with a distinct reservoir risk, both in terms of the presence or absence of Mesozoic and Eocene objectives and, with increasing depth rapidly deteriorating reservoir characteristics. However, suitable seals and trapping conditions are available. Hydrocarbon charge appears to be assured, although under the deeper parts of the Calcareous Alps prospects are likely to be gas prone. High formation pressures are common and there is little chance for reservoir flushing. A further risk factor is the reflection-seismic definition of drillable structures. In this respect, a complex overburden velocity structure provides for depth conversion uncertainties and thus impedes the definition of closure of predominantly low relief extensional structures; moreover, topographic constraints on recording dense enough grids are severe in the Calcareous Alps. In view of the above the Sub-Alpine Molasse, the Flysch Zone and the frontal parts of the of the Calcareous Alps must be regarded as more prospective than the interior parts of the latter where drilling cost are very high. This is born out by the discovery of the Höflein field in the Flysch Zone and the results of the well Grünau-1 drilled near the northern margin of the Calcareous Alps. The objective of future exploration is to locate relatively high relief structures having a large trap volume which, in case of success, could justify economic field development .

Allochthonous plays have to contend in many areas with good reservoir conditions but with a poor seal and trap potential. Seal and trapping conditions are thought to improve towards the frontal and deeper parts of the Calcareous Alps. Hydrocarbon charge is provided by mature autochthonous source-rocks and possibly by source-rocks contained in the allochthonous units. Thick carbonate series can be characterized by very active hydrodynamic regimes involving meteoric waters; this could prohibit the accumulation of commercial quantities of hydrocarbons. In areas of near surface steep dips, seismic resolution is poor. However, although in area of relatively gentle dips seismic resolution is adequate to define structures at deeper levels, which may be protected from meteoric water circulation, the rugged topography often prohibits the recording of dense enough grids to establish 3-way closure of potentially prospective structures. Surface geological information and the construction of balanced cross-sections may go some way to resolve the structural complexity of the Alpine allochthon. However, in the face of objective depths in the range of 4000 to 5000 m and high drilling cost, prospects must be adequately defined before their evaluation by the drill can be justified. The discovery of significant oil and gas accumulation in the allochthonous units forming the substratum of the Vienna Basin, highlight the potential of this play. Similarly, results of the well Hindelang-1 are encouraging for exploration of the Helvetic zones.

Past exploration of the Austrian Alpine belt was rewarded with only limited success due to poor structural definition of prospects involving either autochthonous or allochthonous series. Although all ingredients for a successful exploration play appear to exist, at least in some parts of the Austrian Alps (Table 1), the risk/reward ration must be considered as lop-sided under todays oil and gas price scenario.

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Exploration Criteria in Alpine Thrust and Subthrust Areas

	Allochthonous		Autochthonous
	Calcareous Alps	Helveticum, Molasse	
Reservoir	fair	poor	poor to fair
Seal	poor	fair	fair
Trap	poor	fair	fair
Hydrocarbons	gas/condensate, oil. Toward south generation limit		
Pressure	normal	high to normal	mostly high
Seismic quality	mostly poor	fair to poor	fair to poor
Drilling frequency	very low	low	very low
Drilling depths	2000 - 5000 m	3000 - 5000 m	5000 - 7000 m
Drilling costs	moderate to high	moderate to high	very high

TABLE 1

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Enclosure

Encl. 1 Regional cross-sections through Flysch-Kalkalpen

Hydrocarbon habitat of the Paleogene Nesvacilka Trough, Carpathian foreland basin, Czech Republic

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ABSTRACT

In the Czech part of the Carpathian foreland basin, oil and gas production started in the late 1930's. The hydrocarbon potential of this area is connected with the Paleogene sedimentary fill of the Nesvacilka Trough and its Mesozoic substratum. Ultimate recoverable reserves in established accumulation in Paleogene and Jurassic reservoirs amount to 8 x 10^6 bbls of oil and 14 BCF of gas. As stratigraphic prospects in Paleogene turbiditic and Middle Jurassic transgressive sands have a considerable upside potential, the Nesvacilka Trough is regarded as the most prospective hydrocarbon province of the Czech Republic.

At the transition from the Cretaceous to the Paleocene, the southeastern flank of the Bohemian Massif was uplifted and the complex and very large Nesvacilka-Vranovice system of palaeo-valleys deeply incised into its Mesozoic and Palaeozoic sedimentary cover. During Danian to Late Eocene times, these palaeo-valleys was progressively drowned by transgressing seas and filled in with up to 1500 m thick deeper water clastics of the Damborice Group, comprising the Paleocene Tesany and the latest Paleocene-Eocene Nesvacilka formation. The Tesany formation, consisting of sand-prone, deep-water proximal distributary channels cutting into levee/overbank shales, was deposited under rapidly rising sea-level conditions. Sandy conglomerates and coarse sands, deposited in fanlobes form the reservoirs of hydrocarbon accumulations. The Nesvacilka formation, consisting predominantly of hemi-pelagic shales, was deposited under upwards shallowing conditions and filled in the remaining palaeotopography; within it possible reservoir developments are restricted to slumps and barrier bars along the margins of the palaeo-valleys. Following an Oligocene regressive cycle, the area was incorporated during the Early Miocene into the Carpathian foreland basin. During the Late Miocene terminal phases of the Carpathian orogeny, the sedimentary fill of the Nesvacilka Trough was partly scooped out and overridden by the external flysch nappes.

Oil and gas accumulations are contained in stratigraphic and in combined stratigraphic and unconformity traps involving Middle Jurassic and Paleogene sands. Hydrocarbon charge is provided by autochthonous Late Jurassic source-rocks occurring beneath the adjacent Vienna Basin and by Paleogene shales which have reached maturity

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in the deeper parts of the Nesvacilka Trough, located beneath the external Carpathian flysch nappes.

INTRODUCTION

In the Czech part of the Carpathian foreland basin, the autochthonous sedimentary cover of the southeastern slope of the Bohemian Massif, has been explored for hydrocarbons since the 1920's when close to the surface a heavy oil accumulation was discovered. In the late 1950's, the Nesvacilka and Vranovice palaeo-valleys, which are cut deeply into the Mesozoic and Palaeozoic cover of the Bohemian Massif and are filled with Paleogene sediments, were first recognized (Fig. 1). During the early 1980's, the discovery of the Urice gas accumulation, having recoverable reserves of 5.3 BCF in Paleogene sands of the valley fill, triggered an intensified exploration program. However, of 8 wells drilled within the Nesvacilka Trough, only Karlin-1, located in its deepest parts, was successful and discovered a gas accumulation at the depth of 3900 m. Despite a considerable data base, consisting of wells and 2-D reflection-seismic lines, the distribution and prediction of reservoir sands and the definition of drillable prospects remained difficult. Since 1972, 400 km of 2D reflection-seismic lines were recorded and in 1991 70 km^2 of 3D seismic coverage were acquired. The drilling of 32 wells, including 15 wildcats, has yielded one oil and two gas accumulations having combined ultimate recoverable reserves of 8x 10^6 bbls of oil and 14 BCF of gas in Paleogene reservoirs of the trough fill and its Mesozoic substratum. The Paleogene system of palaeo-valleys, which extends over an area of some 1400 km^2, constitutes the most prospective hydrocarbon province of the Czech Republic (Jiricek, 1990; Benada et al., 1990; Ciprys et al., 1995).

GEOLOGICAL SETTING

Only the northernmost parts of Nesvacilka and Vranovice system of palaeo-valleys are located in the subsurface of the undeformed Carpathian foreland whereas its greater parts have been overridden by the most external Carpathian flysch nappes (Figs. 1 and 2). The sedimentary fill of these palaeo-valleys does not outcrop and ranges, according to well data, from Early Paleocene to Early Oligocene. The topographic relief of this fluvial palaeo-valley system, which is deeply incised into Palaeozoic and Mesozoic sediments, is of the order of 1500 m. As such, it developed in response to a major uplift of the southeastern flank of the Bohemian Massif, presumably during the latest Cretaceous.

The Nesvacilka and Vranovice system of palaeo-valleys extends over a distance of some 30 km from southeast of the city of Brno under the internal Carpathian Magura nappe where its definition is no longer possible due to geophysical resolution problems (Figs. 1 and 2). The morphology of the Nesvacilka and Vranovice system of palaeo-valleys, which must have presented a spectacular sight prior to its Paleogene flooding and infilling, is defined by reflection-seismic data, calibrated by wells, and in unexplored areas by means of gravity data. From the central, southeasterly trending Nesvacilka Trough, four lateral valleys branch off to the northeast and cut through several erosional terraces (Fig. 3). These lateral valleys, which are referred to as the Otnice, Milesovice, Koberice and Zarosice valleys (Fig. 9; Brzobohaty, 1993), played an important role during the infilling stage of the valley system in terms of providing lateral clastic influx into the axial Nesvacilka Trough. Only the Koberice valley did apparently become inactive at an early stage, presumably due to river beheading in its drainage area.

The Nesvacilka system of palaeo-valleys is superimposed on a down-faulted panel of the Bohemian Massif on which little deformed Devonian and Early Carboniferous strata are preserved; this downfaulted block is referred to as the Nesvacilka graben (Fig. 2). These downfaulted Palaeozoic strata, which overlay Cadomian basement forming part of the East Silesian block, attain a thickness of

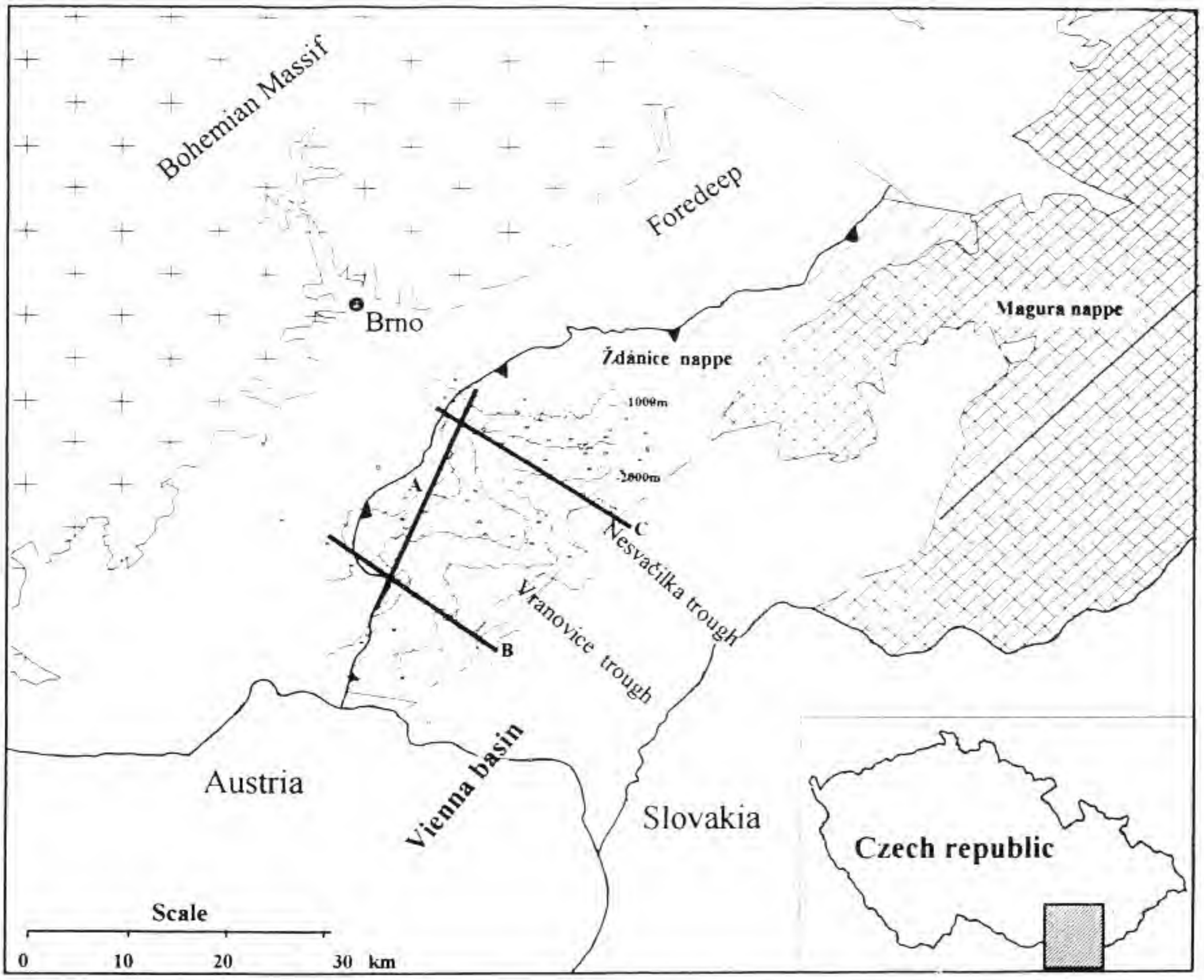

FIG.1. Schematic geological map of SE margin of Bohemian Massif showing depth contours for base of Nesvacilka and Vranovice troughs.and location of structural cross-sections given in Fig. 2.

2500 m. Middle Devonian continental sandstones rest on the crystalline basement and are overlain by Late Devonian limestones and dolomites (Fig. 3). These are followed by Early Carboniferous carbonates which pass upwards into shales and flysch-type sandstones and ultimately into a Namurian paralic sequence. A regional unconformity separates the Palaeozoic strata from a Mesozoic sequence which commences with Middle Jurassic sandstones and shales; these are unconformably covered by Callovian sandy dolomites. The entire Middle Jurassic sequence is some 300 m thick. Upper Jurassic carbonates and marls are the youngest Mesozoic strata occurring in the area and reach thicknesses of over 1000 m. At the transition from the Jurassic to the Cretaceous, the Bohemian Massif was uplifted in conjunction with major wrench deformations that must be related to rifting activity in the Arctic-North Atlantic domain and the North Sea (Ziegler, 1990). In the course of the Late Cretaceous, the flanks of the Bohemian massif were again transgressed. Based on regional palaeogeographic considerations, and as reworked Maastrichtian microfossils have been identified in the basal parts of the Nesvacilka palaeo-valley fill (Hamrsmid et al., 1990), it is assumed that also the southeastern flank of the Bohemian Massif was covered by at least a veneer of Late Cretaceous strata.

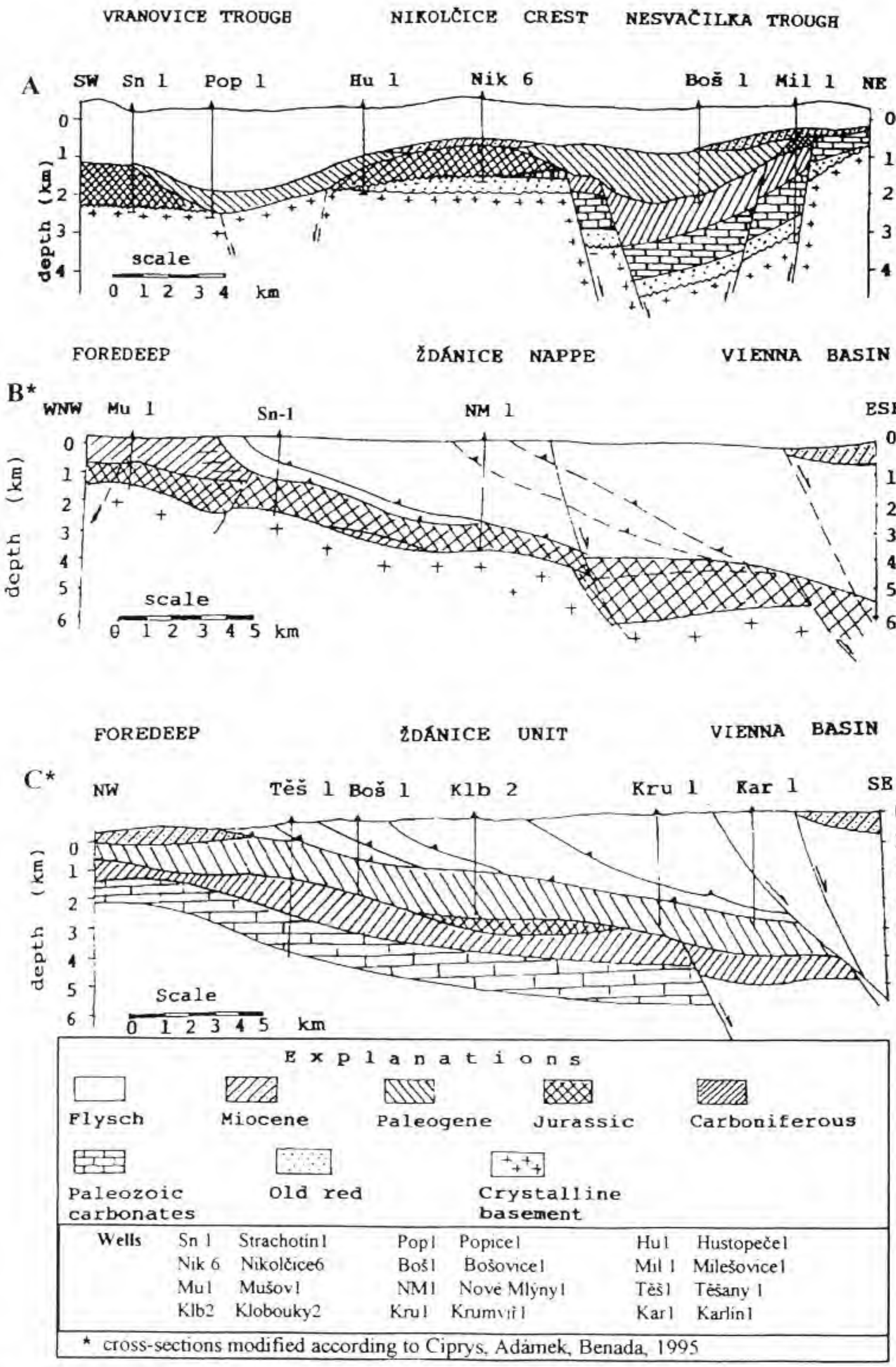

FIG. 2. Structural cross-sections through Nesvacilka and Vranovice troughs, for location see Fig. 1.

As indicated by the stratigraphic record preserved in the Central Bohemian Cretaceous Basin (Malkovsky, 1987) and in the substratum of the Austrian Molasse Basin (Nachtmann and Wagner, 1987; Wessely, 1987), uplift and internal deformation of the Bohemian Massif resumed in Late Turonian times, intensified during the Senonian and culminated during the Early Paleocene. This phase of intra-plate deformation, which resulted in the upthrusting of major basement blocks, was probably induced by compressional stresses which developed in response to collisional coupling between Eastern Alpine-Carpathian orogen and the European foreland (Ziegler, 1990).

Significant uplift of the southeastern parts of the Bohemian Massif, presumably during latest Cretaceous times, caused the development of a southeastwards directed drainage system which cut deeply into the Mesozoic and Palaeozoic strata and formed the Nesvacilka-Vranovice system of palaeo-valleys.

During the Early Paleocene, marine incursions, originating from Carpathian geosynclinal system, began to encroach on the rugged topography of the Nesvacilka-Vranovice canyons and by the Late Eocene the entire area was flooded. Marine Paleogene shales and sands, attaining thicknesses of up to 1500 m, are attributed to the Damborice Group, which, on the basis of a regional unconformity, can be subdivided into the Paleocene Tesany and the latest Paleocene to Eocene Nesvacilka formations (Fig. 4; Rehanek, 1993).

During the Early Paleocene first marine ingressions entered only the Nesvacilka Trough. However, during the Late Paleocene to Early Eocene, both the Nesvacilka and Vranovice troughs were flooded with only the high grounds of the canyon flanks still being exposed. The unconformity separating the Tesany and an the Nesvacilka formations, which cuts deeply into the Tesany formation, is attributed to a latest Paleocene temporary low stand in sea-level (Brzobohaty, 1993). By Late Eocene times, the entire area was inundat-

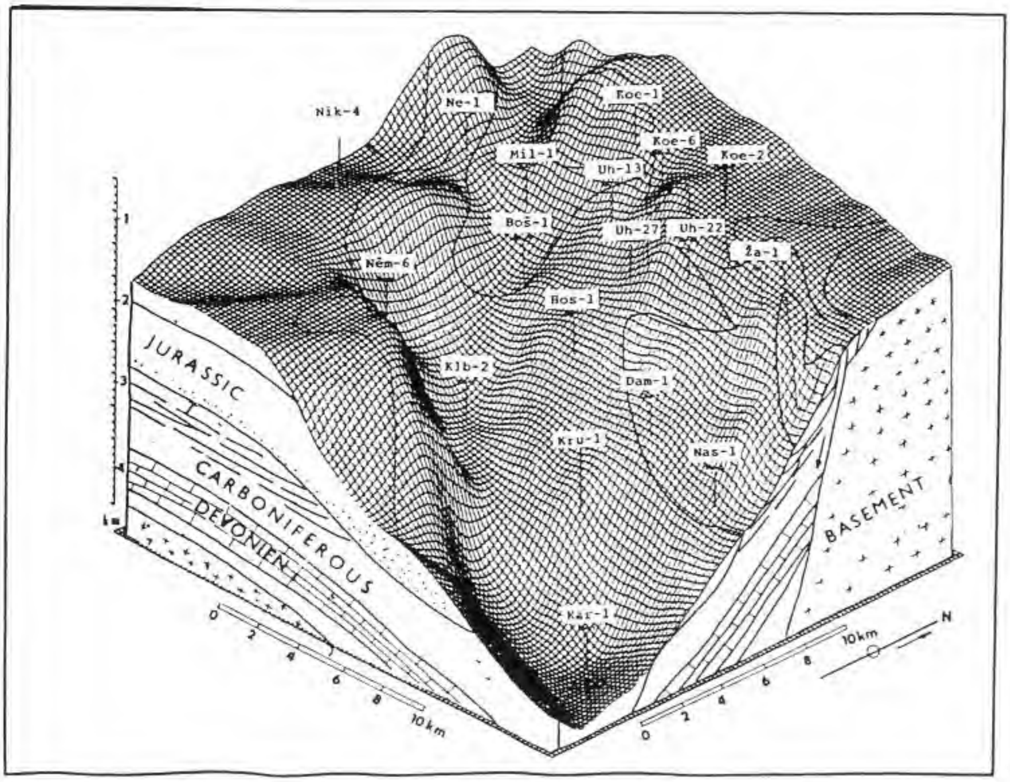

FIG. 3. Block-diagram giving base Tertiary structural relief of Nesvacilka Trough and showing subcropping Palaeozoic and Mesozoic units.

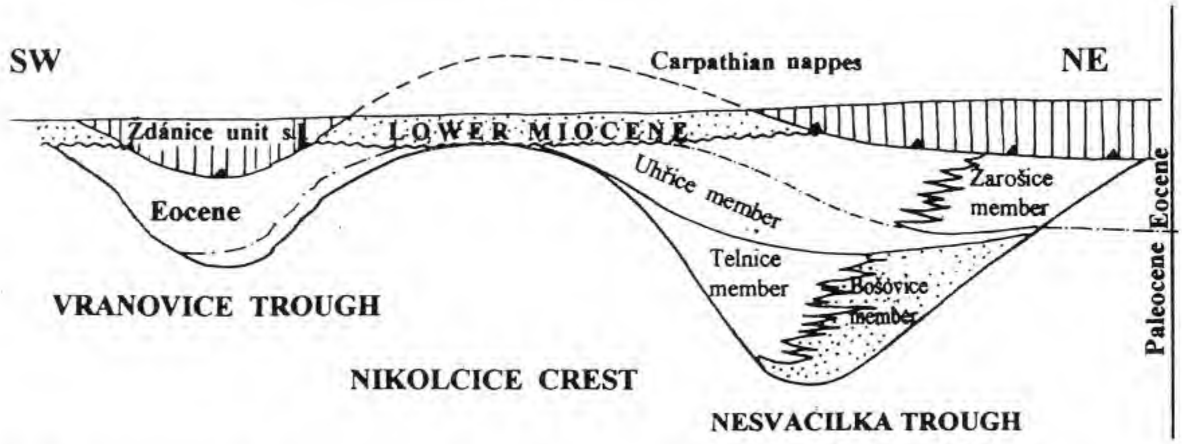

FIG. 4. Lithostratigraphy of Paleogene fill of Nesvacilka and Vranovice troughs.

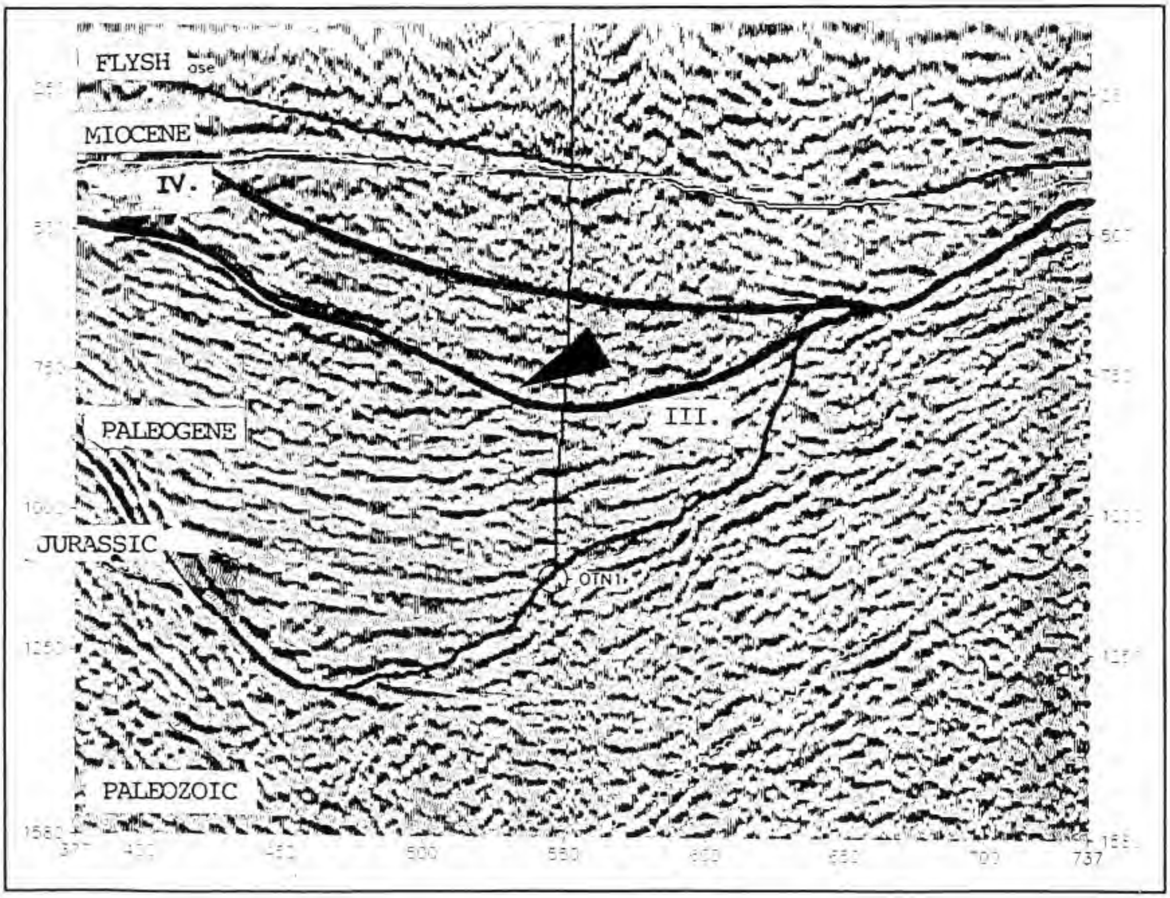

FIG. 5. Seismic Profile 250-86 showing examples of intra-Paleogene unconformities.

ed and much of the palaeo-valleys were infilled with clastics derived from the Bohemian Massif and with hemipelagic clays. At the transition from the Eocene to the Oligocene, a regional regression commenced that may be attributed to global climatic changes (Picha, 1979). However, autochthonous regressive Early Oligocene sediments occur only sporadically in some wells drilled near the northeastern margin of the Nesvacilka Trough. During the Early Miocene the area was incorporated into the Carpathian foreland basin. During the Late Miocene final phases of the North Carpathian orogeny, the upper part of the Paleogene sedimentary fill of the Nesvacilka and Vranovice trough was scooped out by thrust faults and overridden by the external flysch nappes (Fig. 2).

LITHOFACIES ANALYSIS OF PALEOGENE VALLEY FILL

On the basis of the available well data and seismo-stratigraphic criteria, the Paleogene sedimentary fill of the Nesvacilka and Vranovice troughs was subdivided into the Tesany and Nesvacilka formations and their internal lithofacies development analyzed in terms of depositional environments and the distribution of reservoir prone facies (Fig. 4; Rehanek, 1993). Based on core data, it was realized that both the Tesany and Nesvacilka formation were deposited under deeper water conditions. According to bentic foraminifera assemblages obtained from drill cores, the Tesany formation was deposited in water depth slightly greater than 200 m (Holzknecht and Krhovsky, 1987; Hamrsmid et al., 1990). The Nesvacilka formation was deposited under hemipelagic conditions. In our lithofacies analyses we followed the classification of Mutti et al. (1972).

Deposition of the **Tesany formation** was dominated by rapidly increasing water depths and high energy density current systems which came into evidence during the Early Paleocene but waned during the Late Paleocene. Waters were cold and characterized by considerable bottom currents. A turbiditic, sand-prone and a basinal shale facies, referred to as the Bosovice and Telnice members, respectively, are recognized (Fig. 4). The Bosovice member mainly consists of poorly sorted sandy conglomerates and coarse sands (lithofacies A and B); these are interpreted as fanlobe and meandering distributary channel deposits of the upper to middle parts of a submarine fan complex (Fig. 7). More locally, pebbly muds occur which are interpreted as slump deposits (lithofacies F). Channel fill deposits are characterized by polymict pebbles, brownish colour, amalgamated structures and frequent dark, plastically deformed clay chips with imprinted sand grains. The Telnice member consist of monotonous, dark silty claystones containing thin sand intercalations; these clays a interpreted as levee/overbank deposits (lithofacies D, E, F2). The presence of coaly fragments and a large amount of light micas is typical for the Telnice member.

The **Nesvacilka formation**, consisting mainly of hemipelagic clays, was deposited under gradually shallowing, warmer water conditions. A basinal and a marginal facies, referred to as the Uhrice and Zarosice members, respectively, are recognized (Fig. 4). The Uhrice member consists of monotonous, thinly bedded, variegated claystones containing silty layers (lithofacies G); coal fragments and particularly micas are conspicuously absent. Quiet bottom water conditions are indicated by traces of organic life on bedding planes. The Zarosice facies represents the shallow water, lateral equivalent of the Uhrice facies; it is only locally recognized on reflection-seismic data, such as along the mouth of the Zarosice valley, where it may include coastal barrier-bar sands. In much of the Nesvacilka and Vranovice troughs, the top of the Nesvacilka formation has been eroded prior to the transgression of the Early Miocene series.

Overall, the supply of sand to the Nesvacilka and Vranovice troughs decreased during the Late Paleocene and Eocene, probably as a consequence of progressive degradation of the palaeo-relief of the Bohemian Massif, a gradual northward advance of the shore-lines and progressive blocking of the feeder channels by increased hemipelagic clay supply.

Detailed analyses of 2D and 3D reflection-seismic data, applying seismo-stratigraphic interpretation methods, permit to unravel the internal architecture of the Paleogene fill of the Nesvacilka

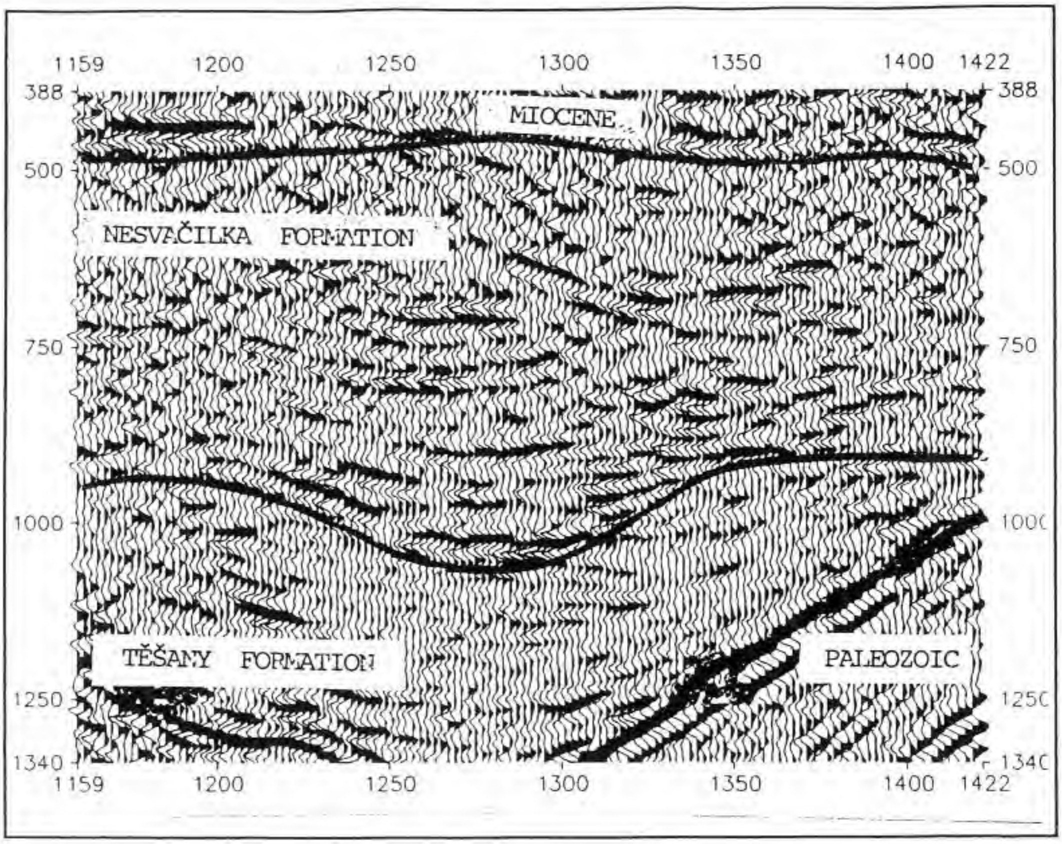

FIG. 6. Seismic Profile 308-87 showing an example of intra-Paleogene unconformity III

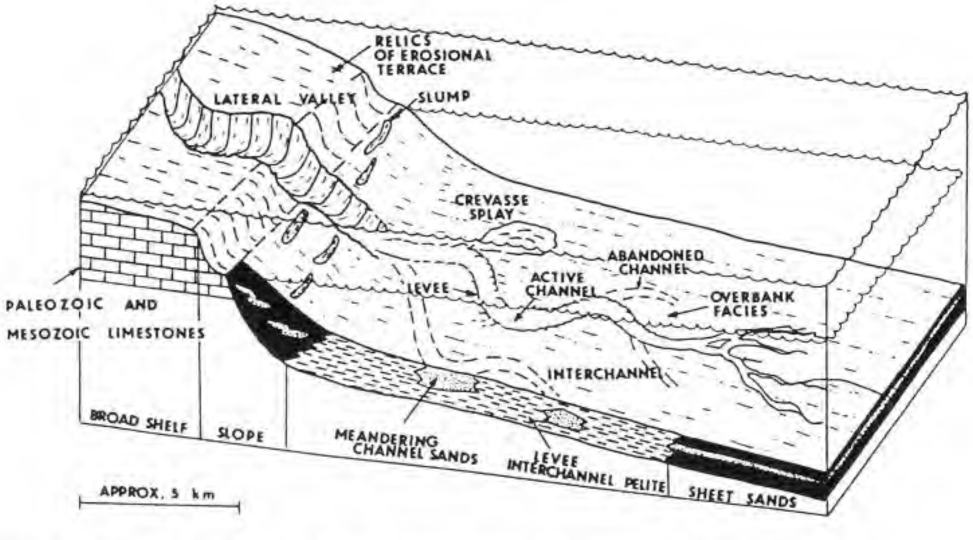

FIG. 7. Passive margin turbidite depositional model for Nesvacilka Trough (modified after Shanmugam and Moiola, 1991).

and Vranovice troughs which is characterized by a number of unconformities These are defined by reflection terminations, indicating down-cutting erosional truncation of the subcropping strata and onlap of the overlying strata; in some instances prograding and downlapping clinoforms can be observed. Examples of such unconformities are given in figs. 5 and 6. Cut-and-fill structures characterize the meandering channel deposits of the Tesany formation. Temporary low stands in sea-level and/or earthquake induced slope instabilities during Paleocene times gave rise to the development of at least two regionally correlative intra-Tesany unconformities (unconformity I and II). Unconformity I is the lowermost one within the Paleocene fill and is Danian in age; it has been recognized only in the Nesvacilka Trough. Unconformity II is of Thanetian age and is associated with a change in the Tesany depositional system and its basal onlap-relationship. The regionally recognized unconformity III marks a break in sedimentation between the Tesany and Nesvacilka formations. Unconformities IV and V occur within the Nesvacilka formation. Unconformity V coincides in the Vranovice Tough with the Middle-Late Eocene boundary but is not recognized in the Nesvacilka Through, probably due to deformation of the respective strata by thrusting during the final stage of the Carpathian orogeny.

A comparison of the age of the different unconformities recognized within the Paleogene fill of the Nesvacilka-Vranovice palaeo-valley sys-

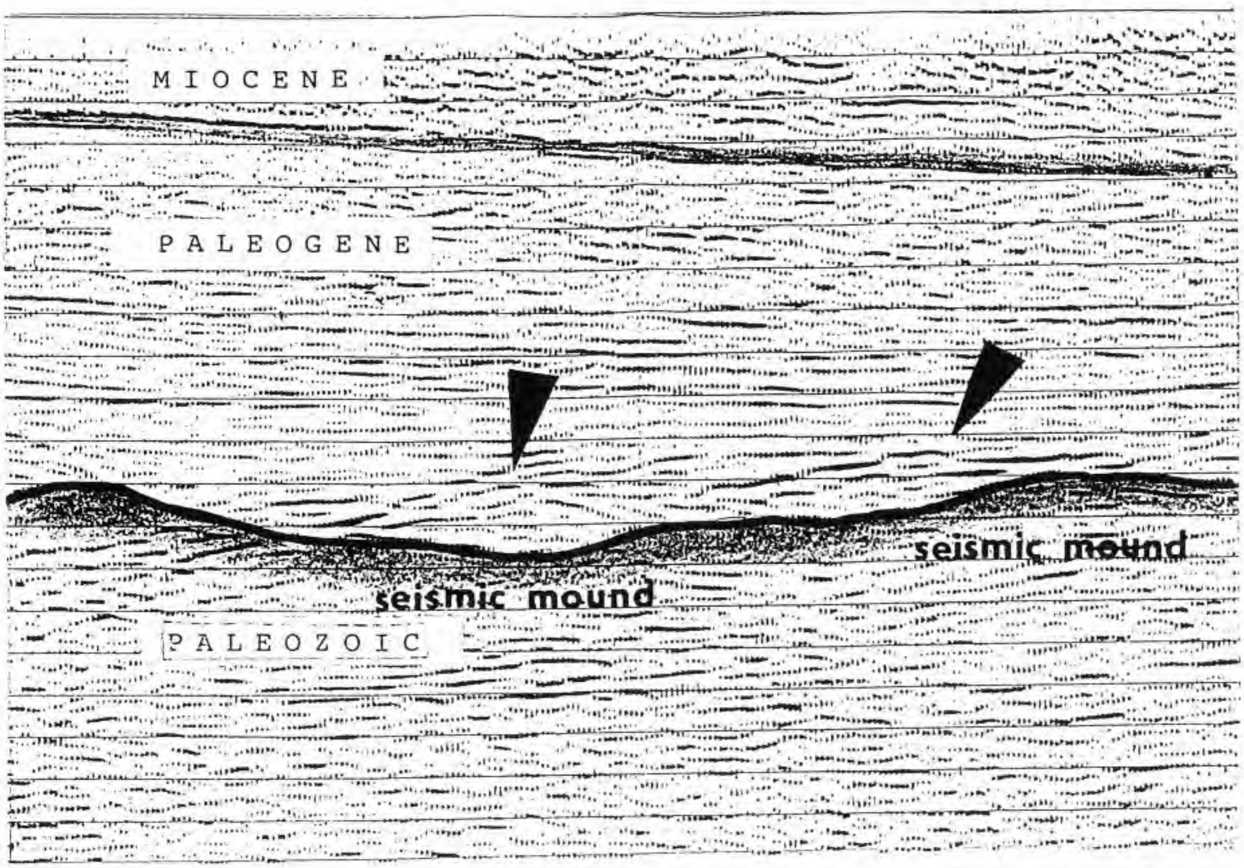

FIG. 8. Seismic Profile 377-87 showing an example of laterally shifting and accreting fan lobes in the Tesany Formation

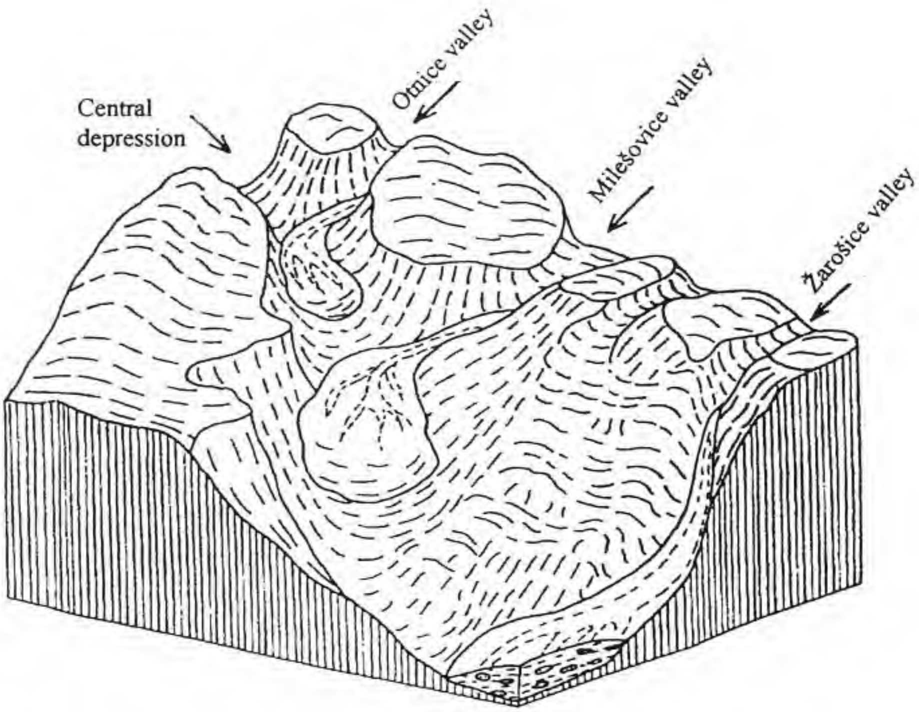

FIG. 9. Block-diagram showing palaeo-relief of central parts of Nesvacilka Trough during Paleocene times.

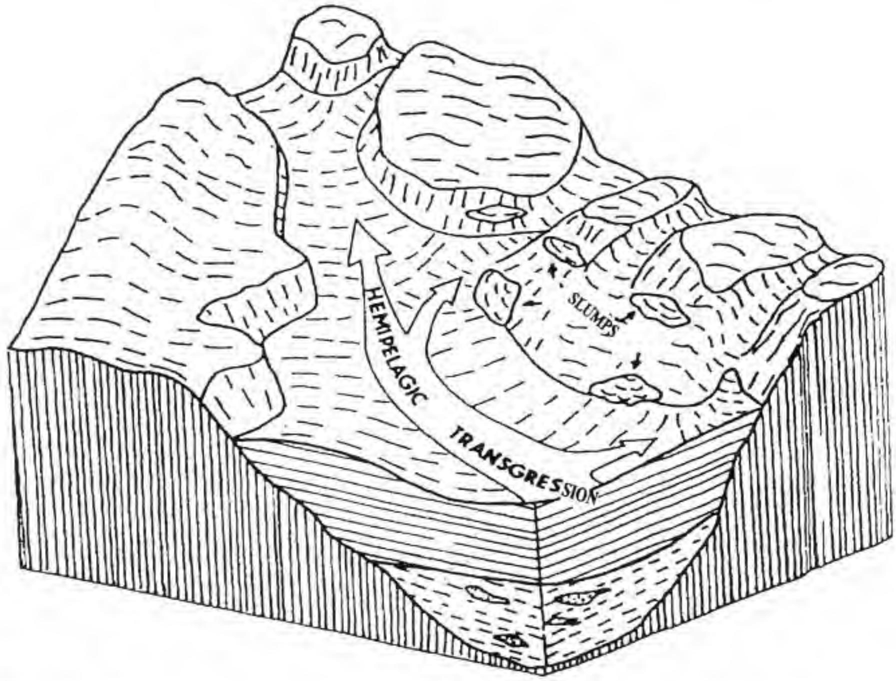

FIG. 10. Block-diagram showing palaeo-relief of central parts of Nesvacilka Trough during Eocene times

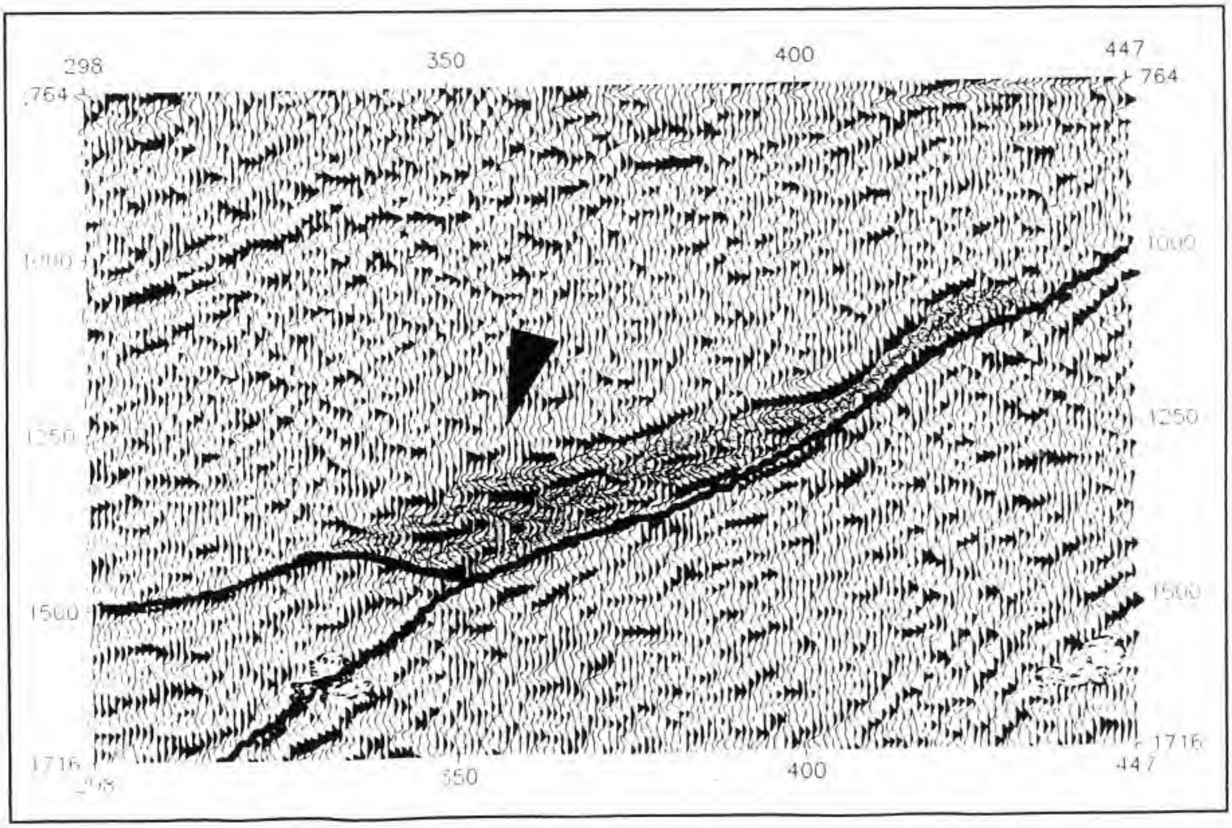

FIG. 11. Seismic Profile 243-80 showing an example of a large slump

tem and the sea-level curve of Haq et al. (1988) indicates that their development is not controlled by eustatic fluctuation in sea-level but is rather due to tectonically induced relative changes in sea-level, presumably reflecting crustal deformations caused by stresses related to the interaction of the Alpine-Carpathian orogen with the European foreland. In this respect it must be realized that the Tesany formation accumulated at a time when major compressional intra-plate deformations occurred in the Central European foreland of the Carpathians and Alps (Ziegler, 1989, 1990). However, at this stage we are unable to comment on possible dynamic processes which controlled the observed apparent sea-level changes and particularly the rapid Early Paleocene subsidence and flooding of the Nesvacilka-Vranovice system of palaeo-valleys.

DEPOSITIONAL PATTERNS AND RESERVOIR DEVELOPMENT

In our evaluation of the depositional pattern of the Paleogene fill of the Nesvacilka and Vranovice troughs, we applied the passive margin turbidite model of Shanmugam and Moiola (1991). During Paleogene times, the southeastern margin of the Bohemian Massif faced the deeper water Silesian Basin of the Carpathian orogenic system (Kovac et al., 1993). From this basin marine transgressions advanced northwards into the Nesvacilka and Vranovice troughs. However, in order to accommodate the peculiarities of the Nesvacilka-Vranovice trough depositional system, the Shanmugam and Moiola (1991) model had to be modified as shown in Fig. 7.

According to well data and detailed seismo-stratigraphic interpretations of reflection-seismic profiles, accumulation of turbiditic fan and fanlobe deposits in the Nesvacilka Trough commenced with its Early Paleocene rapid flooding and the establishment of deep water conditions. Similar deep-water fan deposits are described by Shanmugam et al (1988) from recent passive margins, such as the Gulf of Mexico. Recent fanlobes are characterized by sand-filled distributary channels and shaly-sandy levee/overbank deposits; upon compaction the sand filled channels form mounded features.

According to well data, the part of the Paleocene Tesany formation, which is bounded by unconformities I and II, is dominated by high energy sands and sandy conglomerates; these were deposited in deeper waters as fans and fanlobes. This sequence is referred to as the Bosovice member (Fig. 4). Particularly during this stage, the lateral valleys of the Nesvacilka Trough played a very important role in terms of clastic supply to the central trough (Fig. 9); the clastic load of rivers flowing through these valleys was derived from the elevated hinterland, as indicated by its polymict composition, as well as from the erosional terraces of the palaeo-valleys (Brzobohaty, 1993; Rehanek, 1993). Only along the lateral Otnice valley did turbiditic fans remained active till the latest Paleocene, whereas elsewhere coarse clastic supply to the Nesvacilka Trough had ceased earlier.

Seismic facies mapping permits to define fanlobes in the Tesany formation; Fig. 8 gives an example of a laterally shifting and accreting fanlobe which had entered the Nesvacilka Trough at the mouth of the Otnice valley. Such fanlobes appear on seismic data as mounded features and frequently display an internal two-directional downlap configuration. Similar fan-lobes had enter the Nesvacilka Trough at the mouths of the lateral Milesovice and Zarosice valleys and, after reaching the central trough, swung around and advance down its axis. According to seismic data, such laterally shifting channel deposits reach in the middle parts of fanlobes thicknesses of as much as 100-150 m where they consist of sandy conglomerates and coarse sands (lithofacies A and B according to Mutti et al., 1977). In the middle parts of the fanlobes, sandy channels alternate with fine grained levee deposits. Well logs from the middle parts of fanlobes show the characteristics of a cyclically upwards fining sequence.

Channel fills are considered as forming the most important hydrocarbon traps in the entire Paleogene fill of the Nesvacilka and Vranovice troughs. Their reservoirs are sealed by overbank/levee clays. Along the northeastern slope of the Nesvacilka Trough, the upper parts of coarse grained proximal fan deposits appear to have been cut off during temporary low stands in sea-level and are sealed by clays of the next following sequence. There are also cases of meandering channels which are cut-off up-dip by unconformities.

The upper parts of the Tesany formation, which are bounded by unconformities II and III, and the Nesvacilka formation consist mainly of monotonous hemipelagic clays. As such they reflect a significant decrease in clastic supply to the Nesvacilka Trough. As indicated by reflection-seismic data, these hemipelagic clays contain along the northeastern flank of the Nesvacilka Through in the interval between unconformity II and III a number of slumps or ponded lobes (Fig. 10; Brzobohaty, 1993). In Fig. 11 the reflection-seismic signature of such a slump feature is illustrated; its base appears to cut down into unconformity II and its internal configuration is rather chaotic, though its upper surface is marked by discontinuous, high amplitude reflectors which are indicative of a significant density-velocity contrast, and therefore also a lithological contrast. The lithological composition of such slump features is still unknown as they have not yet been tested by wells; although potentially prospective, such features carry a distinct reservoir risk (Shanmugam and Moiola, 1991).

Further potential reservoir developments may be associated with ancient shore-lines where barrier bar complexes are evident on 3D seismic data. Such shore-line sand bodies are, however, only rarely preserved as they have generally fallen victim to erosion during periods of low-stands in sea-level.

HYDROCARBON HABITAT

Within the Paleogene sedimentary fill of the Nesvacilka and Vranovice troughs two gas accumulations have been established, namely the Uhrice and Karlín fields (Fig. 12). These contain ultimately recoverable reserves 8.8 BCF gas. By now, both fields have been almost depleted. Gas production from the Uhrice field was 3.5 MMCF gas/day. This field in now being converted for underground gas storage. The reservoirs of the Uhrice and Karlín fields are formed by turbiditic sands of the Tesany formation. Traps are mainly of a stratigraphic nature and involve lateral sand pinch-outs and combined pinch-out and truncation geometries.

Apart from the Paleogene objectives, the Jurassic strata and particularly their lowermost parts, corresponding to the Middle Jurassic Gresten sandstone formation, are also prospective. These sandstones form the reservoir of the Damborice field, which contains ultimate recoverable reserves of 8 x 10^6 bbls of oil and is the largest oil field on the margin of the East-Bohemian Massif (Fig. 12). Prospects at Jurassic objective levels are formed by stratigraphic traps involving the onlap of Mesozoic strata against the regional top Palaeozoic unconformity.

According to geochemical analyses, including biomarkers, accumulations contained in Paleogene and Jurassic reservoirs were charged with hydrocarbons generated from Late Jurassic basinal marls and partly by Paleogene shales. According to rock-eval data, these source-rocks enter the oil window at a depth of about 4000 m (Ciprys et al., 1995).

Late Jurassic basinal shales, containing up to 2% of organic matter, occur in the autochthonous series beneath the Vienna Basin where they have partly entered the gas-window (Wessely, 1987; Ladwein, 1988). Hydrocarbons expelled by these shales apparently migrated laterally and updip into the Nesvacilka and Vranovice troughs. Similarly, hydrocarbons generated by Paleogene shales (up to 3.2% TOC) in the deeper, Carpathian sub-thrust parts of the Nesvacilka Trough, migrated laterally

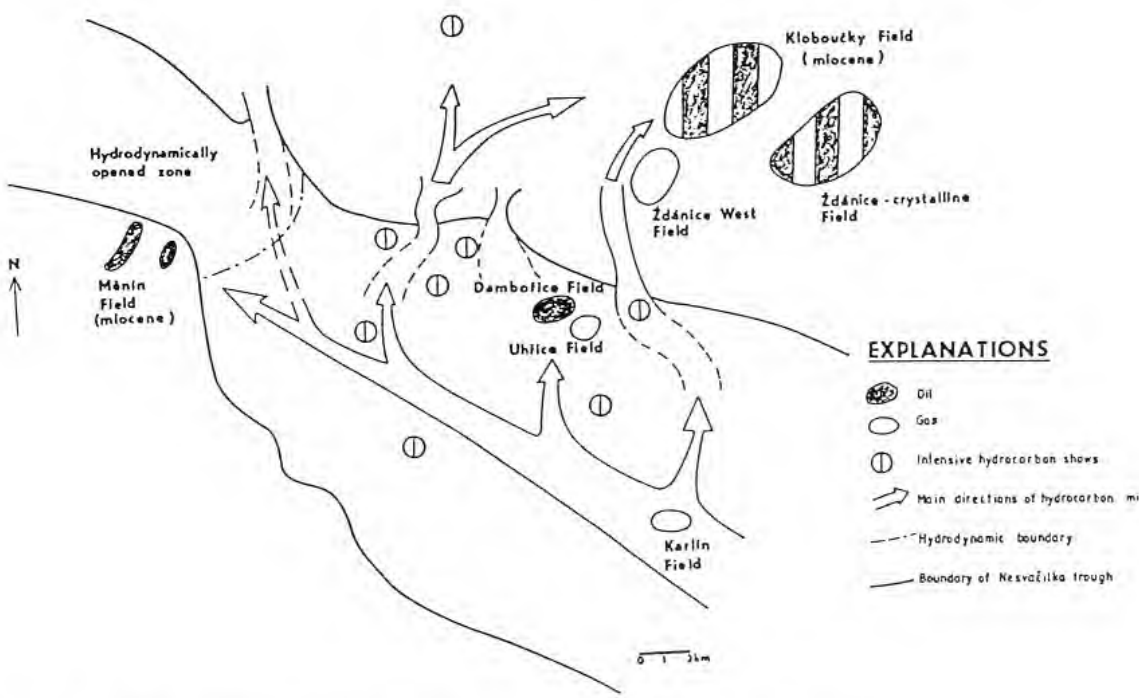

FIG. 12. Hydrocarbon accumulations of the Nesvacilka Trough and hydrocarbon migration paths.

and updip into Paleogene reservoirs. The main phase of hydrocarbon generation probably postdates the Late Miocene emplacement of the Carpathian nappes (Ciprys et al., 1995). As such, there is apparently no restriction on the hydrocarbon charge of remaining prospects recognized within the Nesvacilka and Vranovice troughs (Fig. 12).

Based on 2D and 3D seismic data, a Paleogene prospect inventory was established consisting of 12 potential stratigraphic traps involving turbiditic sands.The capacity of individual prospects was generally determined in volumes of oil and ranges from about 10 to 50 x 10^6 bbls. Correspondingly, the upside potential of as yet undrilled prospects is of the order of 275 x 10^6 bbls of oil. Nevertheless, we expect that some of these prospects are gas prone. Although an inventory of Jurassic prospects is not yet available and will be carried out after 3D seismic coverage has been enlarged, the Nesvacilka and Vranovice troughs must be regarded as the most prospective hydrocarbon province of the Czech Republic.

CONCLUSIONS

In the sub-surface of the Neogene North-Carpathian foreland basin the Paleogene Nesvacilka-Vranovice troughs correspond to a system of palaeo-valleys which were deeply incised into Palaeozoic and Mesozoic strata covering the Cadomian basement of the stable East Silesian block. These palaeo-valleys, which presumably developed in response to latest Cretaceous uplift of the southeastern flank of the Bohemian Massif, host an important hydrocarbon province which covers an area of about 1400 km^2.

Reservoirs and stratigraphic traps are provided by Early and Late Paleocene fanlobe-type turbidites, involving proximal channels filled with sandy conglomerates and coarse sandstones and more distal lenticular sand bodies related to meandering distributary channel-levee and overbank systems. During the latest Paleocene these turbidite systems became inactive in conjunction with progressive drowning out of the palaeo-relief. During the Eocene the entire valley system was infilled with hemipelagic clays containing along the tough flanks slump bodies and possible coastal barrier bar complexes. During the Early Miocene the area was incorporated into the Carpathian foreland basin. Late Miocene emplacement of the Carpathian flysch nappes was accompanied by thrust deformation of the upper parts of the Paleogene fill of the Nesvacilka and Vranovice troughs.

Within the sedimentary fill of these troughs, four regional unconformities are recognized on reflection seismic data; their development is related to tectonically induced relative changes in sea-level.

Geochemical analyses indicate that hydrocarbon charge is provided to the Nesvacilka and Vranovice troughs by autochthonous Late Jurassic marls and Paleogene shales which have reached maturity in the Vienna Basin.and beneath the Carpathian nappes, respectively Lateral and updip migration from these kitchens appears to have been very effective and is probably still going on.

Established accumulations, contained in Paleogene and Jurassic reservoirs, account for ultimately recoverable reserves of 8 x 10^6 bbls of oil and 14 BCF gas. Remaining prospects are stratigraphic traps, involving Paleocene turbiditic reservoirs, which are located at depths between 1000 and 3000 m and have an upside potential of the order of 275 x 10^6 bbls of oil. The prospectivity of Jurassic stratigraphic traps will be evaluated after additional 3D seismic coverage has been acquired. The Nesvacilka-Vranovice system of palaeo-valleys is the most prospective hydrocarbon province of the Czech Republic.

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Development and hydrocarbon potential of the Central Carpathian Paleogen Basin, West Carpathians, Slovak Republic

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ABSTRACT

The Central Carpathian Paleogene Basin (CCPB) lies within the West Carpathian Mountain chain and comprises the proximal facies of the West Carpathian Flysch Belt. This basin developed in a piggy-back position. It occupied the proximal zone of the accretionary wedge above the southwestward subducting oceanic slab, attached to the European Platform. The eastern part of the basin was affected by NE-SW compression, while its western portion was deformed by sinistral transpression. The tectonic events preceding and accompanying deformation of this basin must be related to convergent movements of the African, Apulian, and European plates.

The morphology of the basin floor was controlled by pre-Senonian nappe emplacement in the Inner Carpathians and Senonian thrusting in the Pieniny Klippen Belt. The depositional system was affected by shortening, uplift and shifting of the basin axis, which finally resulted in the termination of sedimentation during Oligocene-Egerian time. During the Paleogene-Karpatian period, the Pieniny Klippen Belt was detached from its substratum and shortened together with the CCPB, the external parts of the accretionary wedge, the Flysch Belt, and the internal parts of the Foreland Molasse Basin. Maximum shortening occurred in the Pieniny Klippen Belt and the proximal parts of the Flysch Belt. During the Badenian, shortening was replaced progressively by NE-SW extension, which spread from the hinterland and accommodated frontal shortening. As subduction ceased in the western Carpathian arc during Late Badenian, Early Sarmatian and Middle Sarmatian times, extension vectors gradually changed to ESE-WNW orientation in response to subduction roll-back in the Eastern Carpathians.

The CCPB has good quality seals, fair to poor quality reservoir units, and excellent to fair source rocks. Traps for hydrocarbons were formed before or contemporaneous with hydrocarbon maturation and expulsion. Maturation modeling in the basin is constrained by Middle Oligocene-Egerian and Eggenburgian-Karpatian thrusting, followed by uplift and erosion.

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INTRODUCTION

The Central Carpathian Paleogene Basin (CCPB) lies in the area of the West-Carpathian Mountain front (Fig. 1). The basin is bounded to the north by the Pieniny Klippen Belt and to the south by the Inner Carpathians (Fig. 2). This proximal region of Paleogene flysch deposition underwent significant shortening and uplift. These kinematics and erosion caused a vertical loss from 1 km in some areas to more than 2 km in other zones (Nemcok et al., 1977; Francu and Muller, 1983; Korab et al., 1986). Thus, the sedimentary record of the CCPB lacks the record of the terminal basin-fill succession. The present Paleogene outcrops are preserved in several structural subbasins which area separated by morphological or structural features, upheld by older rocks.

During the past decade, the structural remnants of the original CCPB were the focus of numerous studies. The Zilina, Liptov, Poprad and Hornad depressions (Fig. 3) and other localities studied for their hydrothermal potential were analyzed in detail by Salaga et al. (1976), Franko et al. (1984), Hanzel and Nemcok (1984) and Fusan et al. (1987). The hydrocarbon potential of the Sambron-Lipany region (Figs. 3 and 4) was explored and tested (Janku et al., 1987; Lesko et al., 1982, 1983; Rudinec, 1984, 1987, 1989; Rudinec and Lesko, 1984). Indications of gas and oil were encountered in the Lipany prospect (Fig. 4) by the wells Lipany Li-1 (gas), Lipany Li-2 (oil, gas), Lipany Li-3 (gas), Lipany Li-4 (oil, gas), Lipany Li-5 (oil, gas), Sambron PU-1(oil, gas), Saris S-1 (gas), Plavnica Pl-1 (oil, gas), Plavnica Pl-2 (oil) (Rudinec et al., 1988, 1989; Nemcok et al., 1977; Korab et al., 1986). Most of regions have insufficient geochemical data coverage for detailed

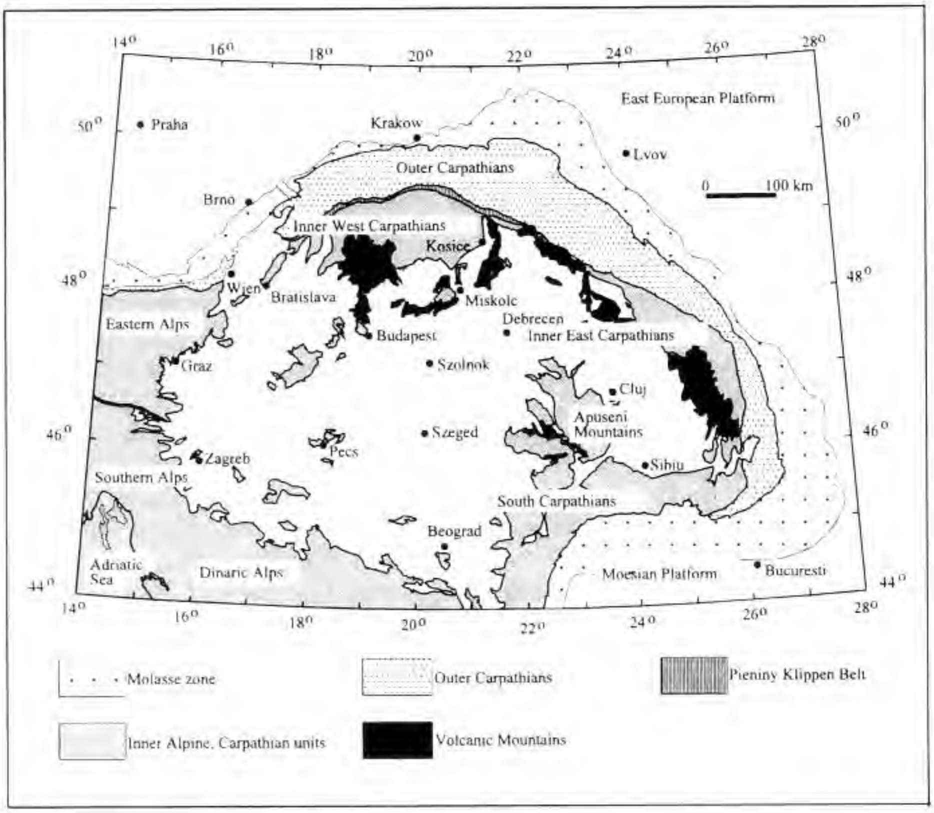


FIG. 1. Regional map of the Carpathian Arc showing major tectonic units (modified after Royden and Baldi, 1988; Sandulescu, 1988).

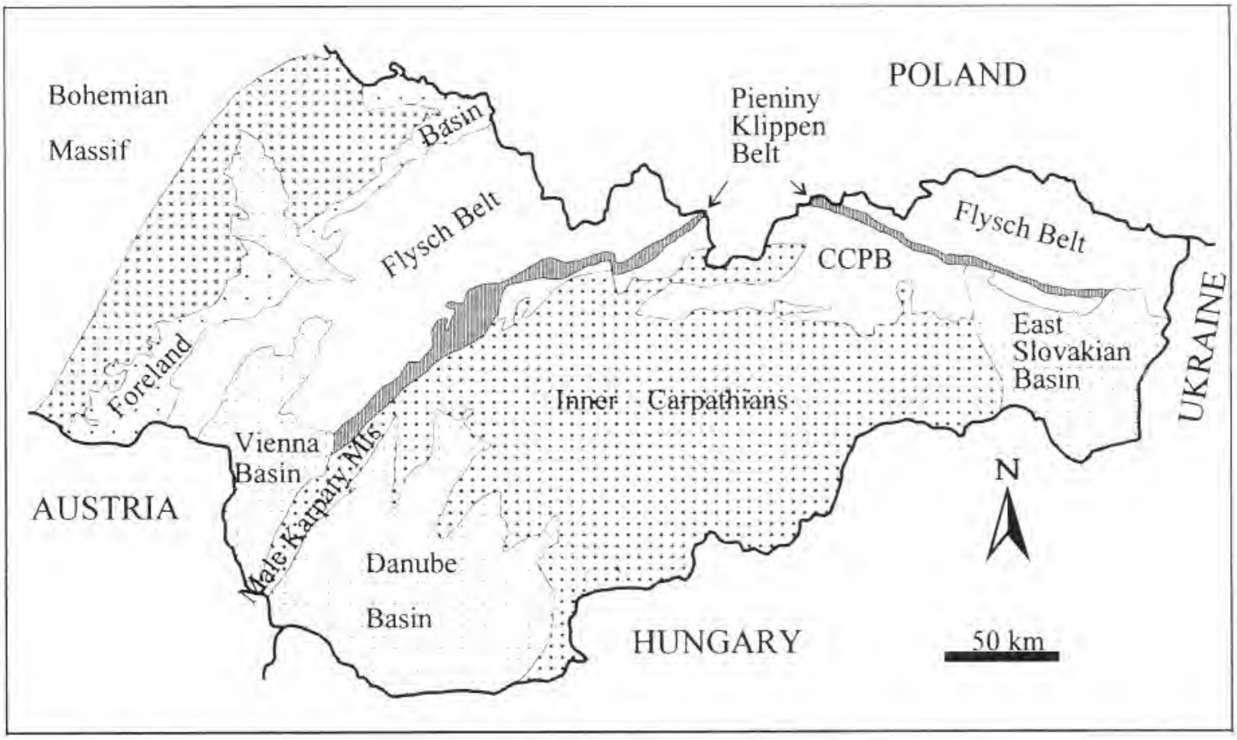


FIG. 2. Regional map of Slovakia showing the tectonic units and sedimentary basins (after Keith et al., 1991).

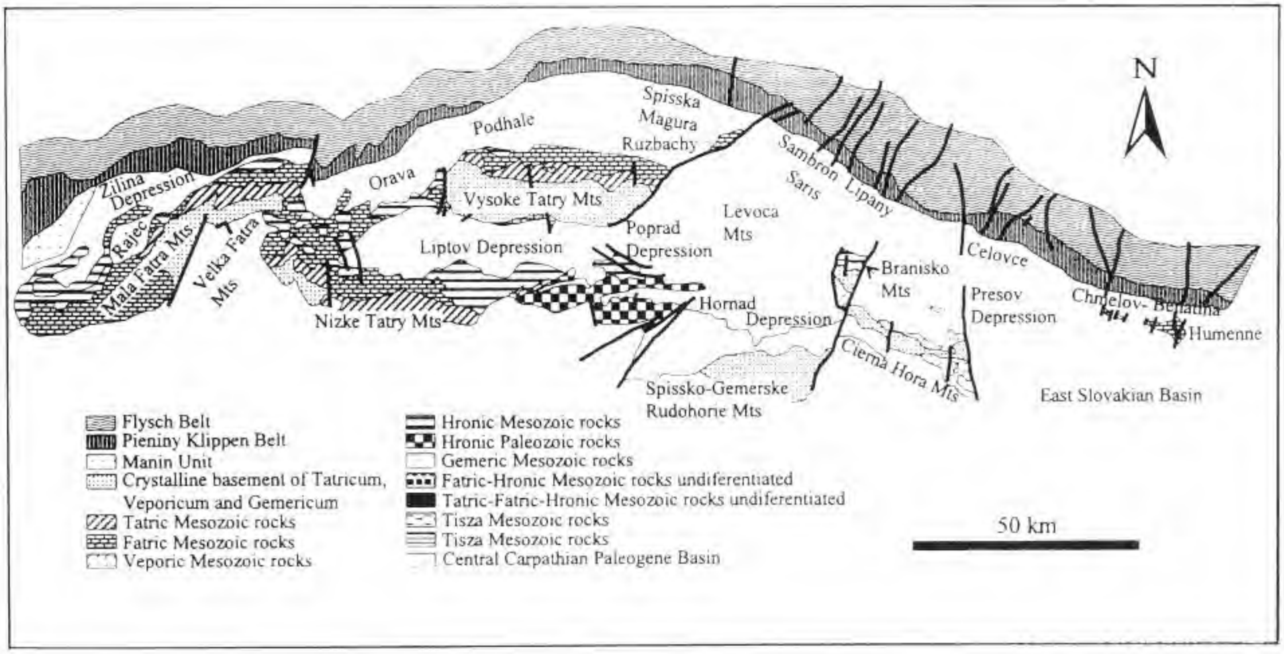


FIG. 3. Regional geological map of pre-Cenozoic surface units surrounding the Central Carpathian Paleogene Basin (after Keith et al., 1991).

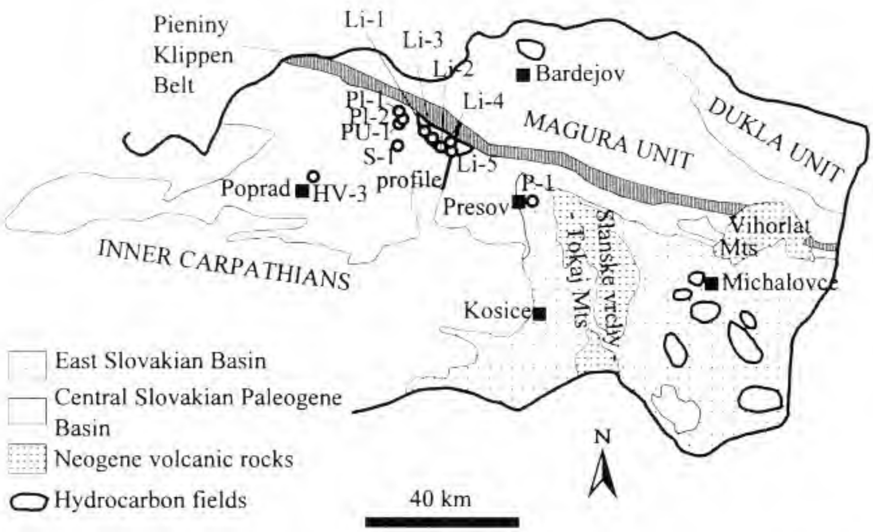


FIG. 4. Occurrences of hydrocarbon deposits, most important boreholes in eastern Slovakia (modified from Rudinec, 1989).

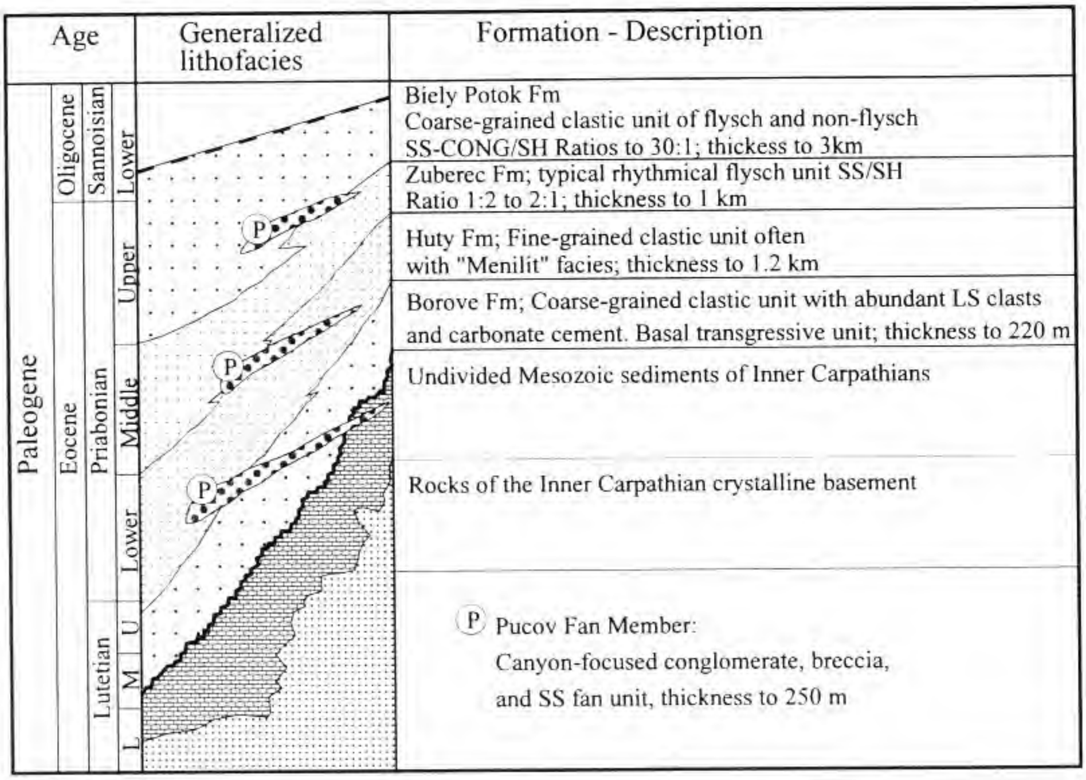


FIG. 5. Schematic lithostratigraphic column for the Central Carpathian Paleogene Basin fill (after Gross et al., 1984).

hydrocarbon evaluation. The Levoca Mountains have no reflection seismic coverage.

Each of the cited structural basins was studied separately and attempts to integrate analyses of the CCPB are rare (Marschalko, 1978, 1981, 1982; Marschalko and Misik, 1976; Marschalko and Korab, 1975; Rakus et al., 1990). The goal of this paper is to determine the model of this frontier basin and to characterize its hydrocarbon habitat by applying play concept elements which have been documented in local publications and reports. Maturation modeling was carried out by means of BasinMod™ software.

Basin models were based on available lithostratigraphical, sedimentological and structural data collected by previous workers and the authors. Structural data include measurements of faults, slickenside striations, folds, extensional veins, determination of fault displacement (e.g. Hancock, 1985; Petit, 1987; Means, 1987), measurement and determination of various fold parameters (Ramsay, 1967; Ramsay and Huber, 1983), observation of faults, vein mineralization, and cross-cutting relationships of all visible structures. Structural orientations were plotted on stereonets to analyze orientation patterns.

Fault-slip data (several thousand measurements) from more than 200 localities in and adjacent to the CCPB were used to determine palaeostress configurations. Inversion stress analysis was used to calculate principal stress orientations, magnitude ratios, and fault-slip polyphase relationships for the different events (Carey and Brunier, 1974; Angelier and Mechler, 1977; Angelier, 1990; Hardcastle and Hills, 1991). Vein and fold data, indicating the approximate orientation of principal stresses, provided a check for the afore mentioned computations.

Polyphase structural overprints were observed at most localities. A superposition of such structures permitted to observe and plot the relative movement (stress configuration) chronology at each outcrop. Timing of tectonic events was determined on the basis of the age of the deformed sediments and other geological constraints.

GEOLOGICAL DEVELOPMENT OF THE CENTRAL CARPATHIAN PALEOGENE BASIN

Data

The sedimentary fill of the CCPB is represented by the Podtatranska Group (Gross et al., 1984), summarized in the lithostratigraphic column given in Fig. 5. This sedimentary succession can be subdivided into four formations and one member.

The lowest unit of this succession is the **middle Eocene Borove Formation**, representing a basal transgressive facies, which consists of locally

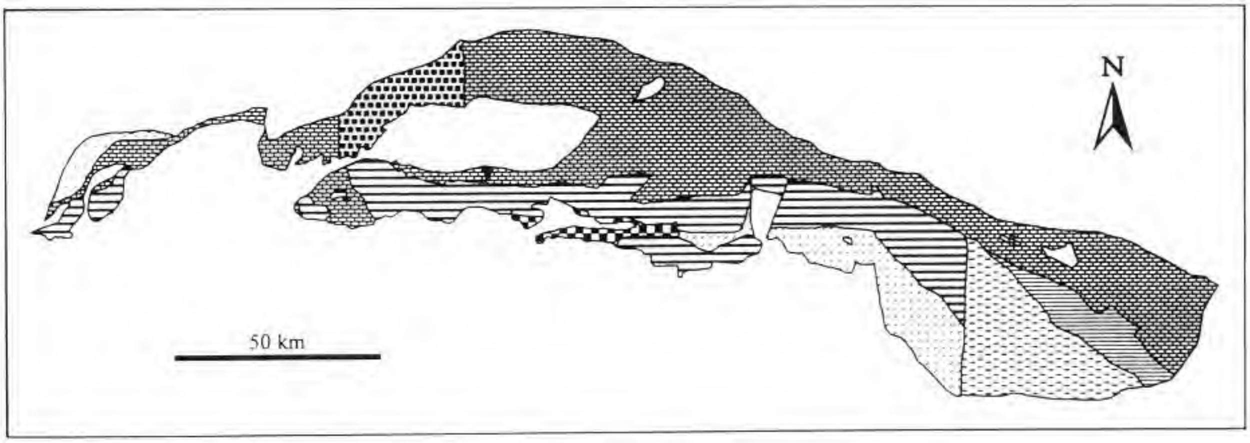


FIG. 6 Schematic geological map of pre-Cenozoic units sub-cropping in the basin floor (after Keith et al., 1991). Explanations in Fig. 3.

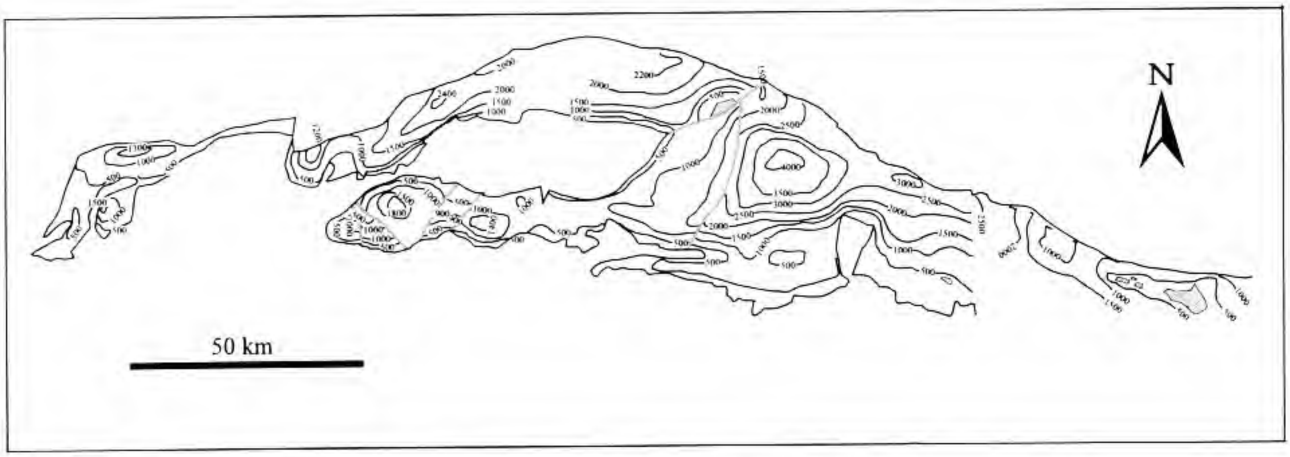


FIG. 7a. Schematic thickness map of Central Carpathian Paleogene Basin fill (after Keith et al., 1991).

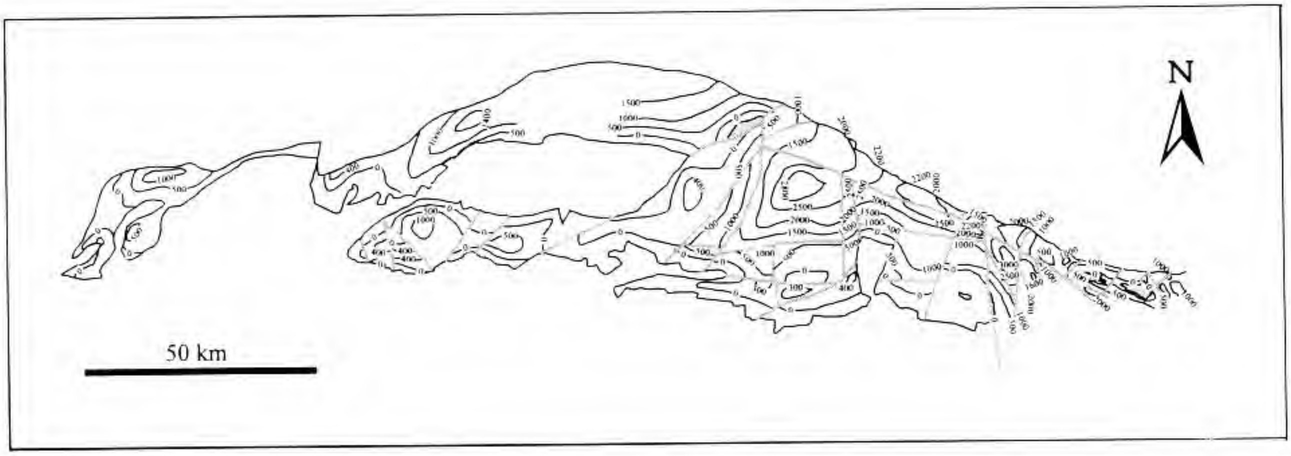


FIG. 7b. Schematic contour map of Central Carpathian Paleogene Basin floor (modified after Fusan et al., 1987). Numbers (in metres) indicate depth below the sea level.

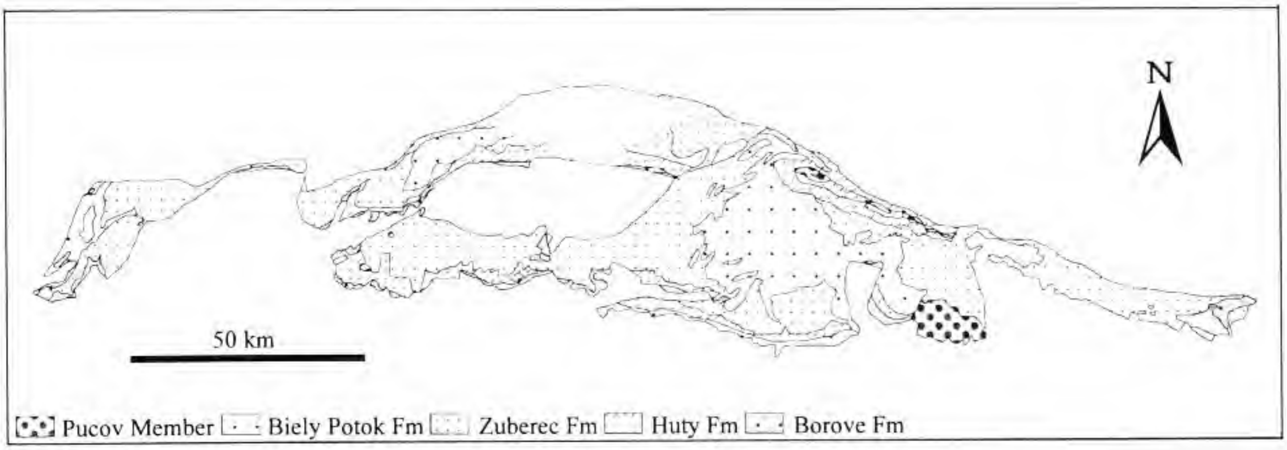


FIG. 8. Lithofacies map of Central Carpathian Paleogene Basin fill (after Keith et al., 1991).

derived breccia, conglomerate, polymict sandstone, siltstone, marl and limestone. It unconformably overlaps the pre-Senonian nappe structures of the Inner Carpathians (Andrusov et al., 1973; Biely, 1989) which consist predominantly of Mesozoic carbonates (Fusan et al., 1987) (Figs. 3 and 6).

The Borove Formation is conformably overlaid by the **middle-upper Eocene Huty Formation** which consists primarily of claystone and siltstone, containing thin interbeds of fine- to medium-grained sandstone. This unit reflects the progressive deepening of the basin. The "Menilit" facies, containing dark shale, which is best developed in the Flysch Belt, is the richest source rock and represent a member of the Huty Formation.

The u**pper Eocene Zuberec Formation** conformably overlays the fine-grained Huty deposits, indicating a change to a sandier, rhythmical flysch. The uppermost part of the basin-fill sequence is represented by the **upper Eocene-lower Oligocene Biely Potok Formation** which is a thick succession of siliciclastic flysch.

Within this succession, elongate coarse-clastic and/or brecciated carbonate turbidite fans (locally referred to as Pucov Member) can be observed at several levels. The upper portion of the regressive facies of the basin-fill sequence is missing in most of the sub-basins. The thickness of the various units is highly variable and is constrained by the morphology of the pre-Middle Eocene structures and the subsequent subsidence pattern of the basin (Figs. 7a, 7b and 8). The geometry of the basin is asymmetrical, with the deepest parts located along the Pieniny Klippen Belt (Fig. 9).

Sedimentation commenced in the CCPB contemporaneously with the development of the Carpathian arc. During the Eocene to middle Miocene period, the western part of the Carpathian orogenic belt advanced northeastwards above the southwestward subducting oceanic slab which was attached to the European Platform (Nemcok, 1993). Stress inversion studies (e.g. Nemcok, 1993) indicate that the western part of the Carpathian arc developed by sinistral transpression, while its frontal part underwent NE-SW compression. Whereas the western part of the CCPB was situated in a zone of sinistral transpression, its eastern part was located in a compressional zone (Fig. 10). The Paleogene to early Miocene stress fields computed by Nemcok and Nemcok (1994) show that a transition from NE-SW compression to N-S transpression is evident in the basin fill (Fig. 11). In this zone, as illustrated in Fig. 11, northeastward thrusting was accommodated by four large tear faults (strike-slip fault zones) which separate blocks with different uplift/erosion histories. Thrust structures related to northeastward shortening developed within a 10-15 km-wide zone along the northern margin of the basin (south of the Pieniny Klippen Belt). Shortening of up to 70 percent has been calculated for this zone. The amount of material transport by thrusting decreases southwestwards and is accommodated by folding. In the remaining parts of the basin, only strike-slip faults, N-S striking dextral and NE-SW striking sinistral

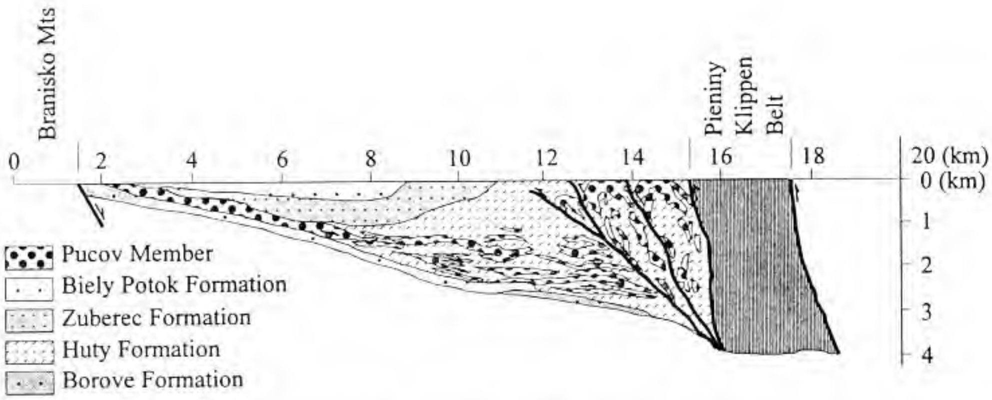


FIG. 9. Geological cross section through Central Carpathian Paleogene Basin to the North of the Branisko Mts. For location see Fig. 4.

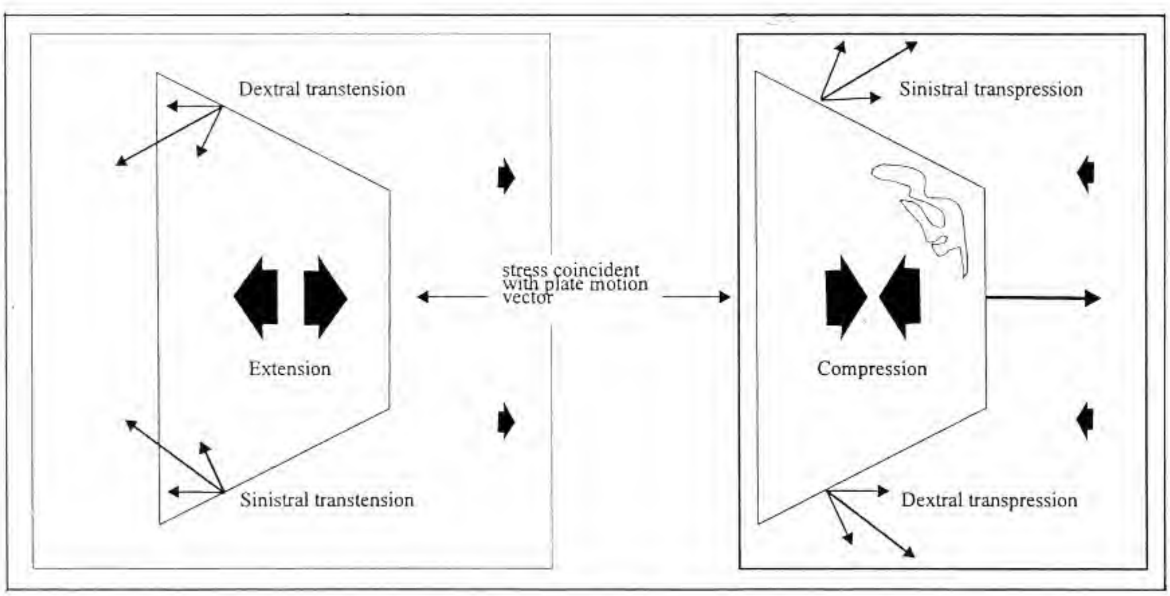


FIG. 10. Block scheme of the Inner Carpathians with Eocene - Early Miocene stress trajectories with the position and present day shape of the Central Carpathian Paleogene Basin indicated (modified after Doglioni, 1992).

faults, can be observed. Uplift, related to sinistral transpression, is indicated by apatite fission-track data (Kral, 1977, 1982; Burchart and Kral, 1982).

The youngest preserved sediments of the basin fill are early Oligocene in age. There is no evidence for continuous sedimentation during late Oligocene and Miocene times. Eggenburgian-Karpathian sediments are present in the Celovce area (Figs. 3 and 11), forming a small piggy-back basin carried by contemporaneously shortened Central Carpathian Paleogene slices. The sediments record each thrust event by the occurrence of coarse-grained sediments in the sequence. Further subsequent shortening is evidenced by Eggenburgian-Karpatian bedding becoming more highly inclined, often approaching vertical. Badenian sediments and younger rocks present in the area do not show any evidence of thrusting. Their deformational features comprise only strike-slip and normal faulting. Structural studies (Nemcok et al., 1993) indicate that the normal faulting related to NE-SW extension spread in time progressively northeastward from the Carpathian hinterland. A similar scenario is indicated by hinterland volcanism of Egerian to Sarmatian age, which becomes progressively less contaminated by crustal material with time (Poka, 1988; Salters et al., 1988; J. Lexa, 1994, personal communication).

Sarmatian arc-related calc-alkaline volcanic rocks can be found as far to the northeast as the Pieniny Klippen Belt and indicate, together with structural data, that extension progressively affected the area of the CCPB and the Pieniny Klippen Belt. Stress inversion studies (Nemcok et al., 1993) determined that during the middle-late Miocene the trajectory of extension sigma3 changed from NE-SW to WNW-ESE.

Interpretation

The CCPB development and subsequent overprint model is constrained by the structural evidence discussed above. During the Eocene to middle Miocene period, the ancestral West-Carpathians (present Inner West Carpathians) advanced northeastward over the subducting European Platform. A tapering, foreland accretionary wedge, comprising the Pieniny Klippen Belt, Flysch Belt, and allochthonous parts of the foreland molasse basin, formed as a result of progressive stacking of thrust sheets. Flysch depocenters were located immediately in front of the advancing ancestral Carpathians. The future Pieniny Klippen Belt area underwent significant uplift during early Eocene times, as indicated by apatite fission track data (Kral, 1983). This out-of-sequence thrust unit formed the northern boundary for the CCPB; this barrier was breached, however, during the early-middle Eocene by southwestward transgressions originating from the area of the Flysch Belt. Thus, the original CCPB formed since this time the proximal part of an extensive flysch depositional system (Rakus et al., 1990) which included regions of the future Pieniny Klippen Belt and the Flysch Belt. Differential thrusting in the Pieniny Klippen Belt out-of-sequence thrust unit is responsible for the different age of these transgressions in various parts of the CCPB in which the basal facies unit varies in age from Ypresian to Priabonian. For instance, the onset of basal facies deposition is Ypresian to Lutetian in the Zilina Depression (Gross et al., 1984; Samuel, 1985), Lutetian in the Orava and Sambron-Lipany area (Gross and Kohler, 1987) and Lutetian to Priabonian in the Liptov Depression, Levoca Mts., and Hornad Depression (Gross, 1985; Gross et al., 1980, 1982, 1984; Marschalko, 1965, 1966, 1981; Marschalko and Radomski, 1970; Durkovic et al., 1984). Some parts of the Pieniny Klippen Belt are characterized by the same Ypresian-aged depositional succession as the CCPB, as for the Pribradlovy Paleogen in the Zilina Depression (Gross et al., 1984). These deposits are interpreted as indicative of channels which linked the external parts of the frontal accretionary wedge with the CCPB and cut through the Pieniny Klippen Belt. A similar channel is known from the eastern part of the basin, to the East of Plavnica, Sambron-Lipany area, where it is filled by upper Eocene-lower Oligocene flysch sediments (Nemcok, 1989). Thus, the barrier provided by the Pieniny Klippen Belt probably consisted of an irregular chain of islands, which was significant enough to give rise to the development of divergent paleocurrent systems, controlling sedimentation in the Flysch Belt and in the CCPB.

The transgressive coastal onlap relationship between Paleogene sediments and the Mesozoic

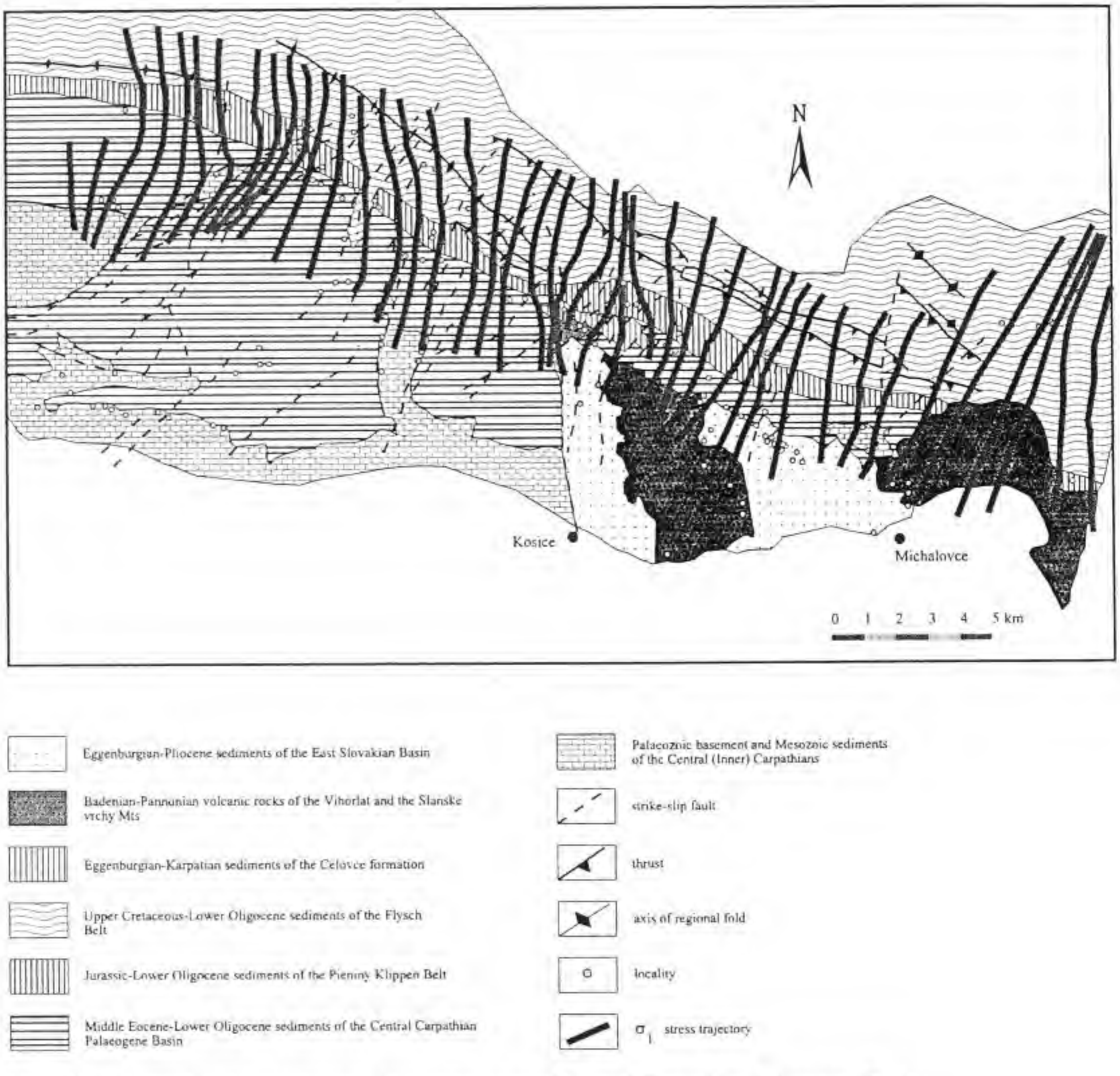


FIG. 11. Sigma1 stress trajectories in eastern parts of Central Carpathian Paleogene Basin and adjacent areas during Paleogene to early Miocene shortening (after Nemcok and Nemcok, 1994).

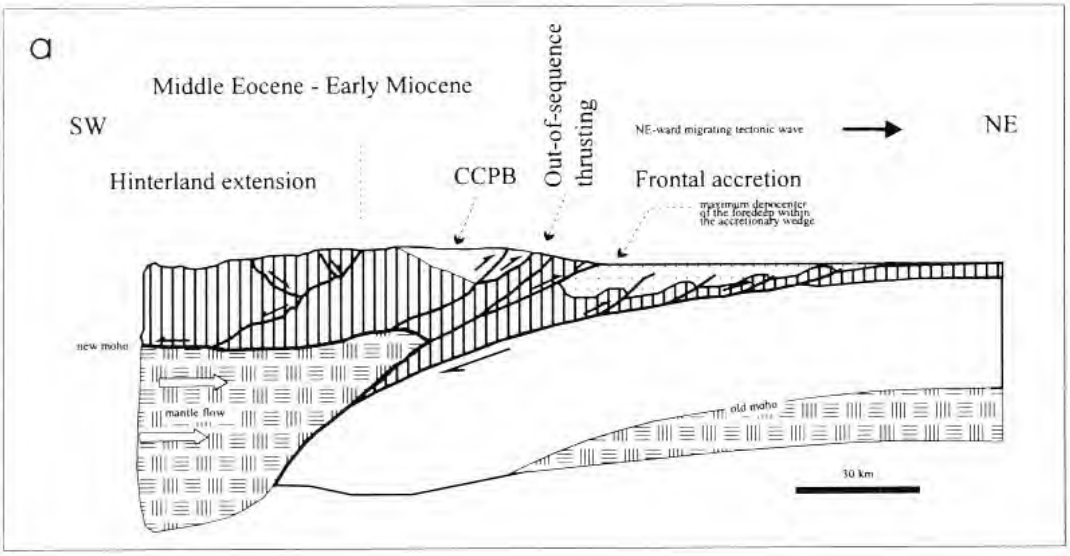
nappes forming the basin floor can be observed in outcrops along the southern margin of the CCPB and is also evident in wells. The basin floor was characterized by a considerable topographic relief which was upheld by various Mesozoic carbonate units. Progressive drowning of this relief gave rise to the development of buried-hill features. In the Pieniny Klippen Belt, there is only minor evidence of a corresponding transgressive coastal onlap of Paleogene sediments (Gross and Kohler, 1987), due to a strong tectonic overprint. The middle Eocene to Badenian structural position of the CCPB within the ancestral Carpathians is shown in Fig. 12. The basin floor was in most of areas generally inclined towards the Pieniny Klippen Belt (Figs. 9 and 12). However, a very dynamic evolution of the CCPB is indicated by changes in subsidence and uplift rates for different sub-basins (Rakus et al., 1990; Marschalko and Korab, 1975; Gross et al., 1980; Marschalko, 1978; Marschalko and Misik, 1976; Samuel, 1985) and by shifting of their axes (Marschalko, 1978). Additional evidence for differential uplift of a southern clastic source is indicated by the occurrence of north-vergent slump blocks at different stratigraphic level. For instance, in the Zilina and Orava Depressions, the slump structures have a late Eocene to early Oligocene age (Samuel, 1985), in the Liptov Depression, their age is middle Eocene (Gross et al., 1980) to late Eocene-early Oligocene (Gross et al., 1980, 1982), and in the eastern portion of the basin, syndepositional slump structures occur in the middle Eocene Borove Formation (Marschalko, 1965; Gross and Marschalko, 1981; Marschalko et al., 1966), in the middle Eocene-upper Eocene Huty Formation (Marschalko, 1965; Marschalko et al., 1966), in the upper Eocene Zuberec Formation (Gross, 1964, 1965; Marschalko, 1966; Marschalko and Radomski, 1970; Gross and Marschalko, 1981) and in the upper Eocene-lower Oligocene Biely Potok Formation (Gross et al., 1982; Marschalko, 1965, 1981). Clastic supply to the basin from southern sources was controlled by channels, as evident by lateral thickness and lithology changes of the flysch series.

After subsidence of the CCPB had ceased in early Miocene times, the Central Carpathian piggyback basin was overprinted by multiple tectonic phases which accompanied eastward movement of the Carpatho-Pannonian plate during the late phases of the Carpathian orogeny (Fig. 12). At the same time, while new portions of the remnant Outer Flysch Basin and molassic foreland basin were accreted to the frontal accretionary wedge, the tectonic setting of the CCPB changed from a region of the frontal accretion to a region of hinterland extension (Fig. 12). In the western portion of the CCPB, this change had a different character and progressed from frontal transpression to transtension.

The complexity in the later erosional history of different sub-basins is indicated by the fact that various structural remnants or sub-basins lack some units of the original basin-fill. The Zilina, Orava, and Liptov Depressions in the west have undergone the greatest Neogene uplift and erosion as indicated by Figures 7a and 8. In the Zilina Depression, portions of the sedimentary succession above the Zuberec Formation have been removed by erosion. This is also true for the southern part of the Orava Depression. The Biely Potok Formation is absent from the Liptov Depression. Moreover, the uppermost known portion of the Biely Potok Formation does not represent the final regressive phase of the basin-fill sequence (Gross and Marschalko, 1981; Gross et al., 1980). The thickness of the final regressive facies, although unknown, should be added to the stratigraphic units when constructing a subsidence and thermal model. Vitrinite reflectance data (Fig. 13) suggest that in the Sambron-Lipany area 1.5 to over 2 km of sediments have been removed by Neogene erosion (Francu and Muller, 1983; Korab et al., 1986).

After the middle Eocene-early Miocene period, the eastern part of the CCPB was affected by extension while frontal shortening continued (Fig. 12). At the end of the middle Miocene, the direction of the extension changed from a NE-SW to WNW-ESE orientation, driven by the subduction roll-back in the eastern parts of the Carpathian Arc (Royden et al., 1982, 1983a, 1983b; Nemcok, 1993). During the middle Miocene, the Vysoke Tatry Mountains (Fig. 3), located within the CCPB, were uplifted and subjected to erosion, as indicated by apatite fission track data (15 Ma; Cambel et al., 1990). Palaeocurrent patterns in the basin indicate that the Vysoke Tatry structure did not exist during the Paleogene.

Data from the western part of the CCPB indicate a different tectonic history. The middle



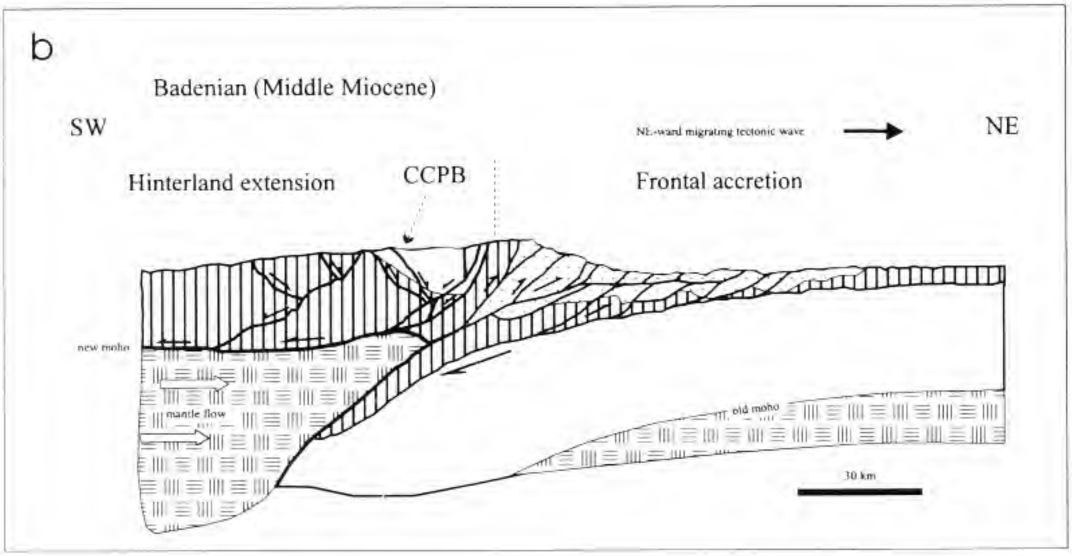


FIG. 12. Sketch profiles through eastern part of West-Carpathians showing north-eastward migration of the region of hinterland extension through time (modified after Doglioni, 1992).

Eocene to early Miocene period was characterized by sinistral transpression. Apatite fission track data from the Mala Fatra Mts. and Velka Fatra Mts. (Fig. 3) indicate that their uplift was active already during early Miocene times (Cambel and Kral, 1989).

Earthquake focal mechanisms indicate that horizontal N-S compression and W-E extension affect the area of interest (Gutdeutsch and Aric, 1976). Remeasurements of geodetic polygons indicate uplift in the western parts of the CCPB and coeval subsidence of its eastern portions (Kvitkovic and Plancar, 1979).

APPRAISAL OF THE PETROLEUM POTENTIAL OF THE CENTRAL CARPATHIAN PALEOGENE BASIN

Reservoir rocks are present within the CCPB fill and in the older, underlying sedimentary succession of the basin floor (Keith et al., 1991). Reservoirs observed within the Paleogene sequence are

(1) coarse clastic and carbonate units of the basal transgressive Borove Formation,
(2) sandstone and rare conglomerate units of the Zuberec and Biely Potok Formations, and
(3) coarse clastic turbidites of the Pucov Fans.

The quality of these reservoirs is fair to poor with an average porosity of 8 to 10% Reservoirs observed in units outside Paleogene sequence include:

(1) Eggenburgian and Karpatian sandstone and conglomerate of local extent in the Presov and Celovce areas (Fig. 3) and
(2) Mesozoic carbonate forming the basin floor which have average porosity from 1 to 14.5%. In most cases, the porosity of these carbonate reservoirs has been enhanced by fracturing. Along southern margin of the CCPB, there is evidences for their pre-transgressional karstification.

Seals within the Paleogene sequence are represented by shale of the Huty, Zuberec, and Biely Potok Formations. Other potential seals are represented by the Eggenburgian and Karpatian shale of the Presov and Celovce areas.

The quality of **source-rocks** within the CCPB sequence was determined by a limited amount of scattered data collected by various agencies in Slovakia mainly in the Sambron-Lipany area. Values of the total organic carbon content (TOC) varies between 0.1 to 1.5% for the fine-grained flysch clastics of the Zuberec Formation (14). In the Menilit shale member of the Huty Formation, TOCs of 1.1 to 10.3% were reported (Simanek et al., 1981; Hokr, 1981). Data for the Biely Potok and Borove Formations are not available. Shales of the Huty Formation, the best and thickest source-rock of the CCPB fill, has rather large areal extent within sub-basins, as indicated by Figures 8 and 9. The ratio of hydrogen and oxygen indexes of the samples from the Zuberec Formation (Fig. 15) indicates that type III kerogen (terrestrial) is prevalent with some samples indicating type II kerogen (marine-phytoplanktonic and zooplanktonic). However, it should be mentioned that interpretation of a hydrogen/oxygen index diagram in terms of type of organic matter is very hazardous, especially in the face of low TOC (matrix effect) and a high degree of maturation (Bessereau, personal com.). Various Mesozoic source-rocks have a rather low TOC (0.1-0.6%) and type II/III kerogen (Keith et al., 1991).

Potential traps can be subdivided into structural, stratigraphic, and combination structural/stratigraphic types. Structural traps are formed by:

(1) high-side thrust and anticlinal traps in a 10-15 km wide zone along the northern margin of the basin,
(2) high- and low-side normal fault traps in the remaining parts of the basin, and
(3) strike-slip fault and drag fold traps.

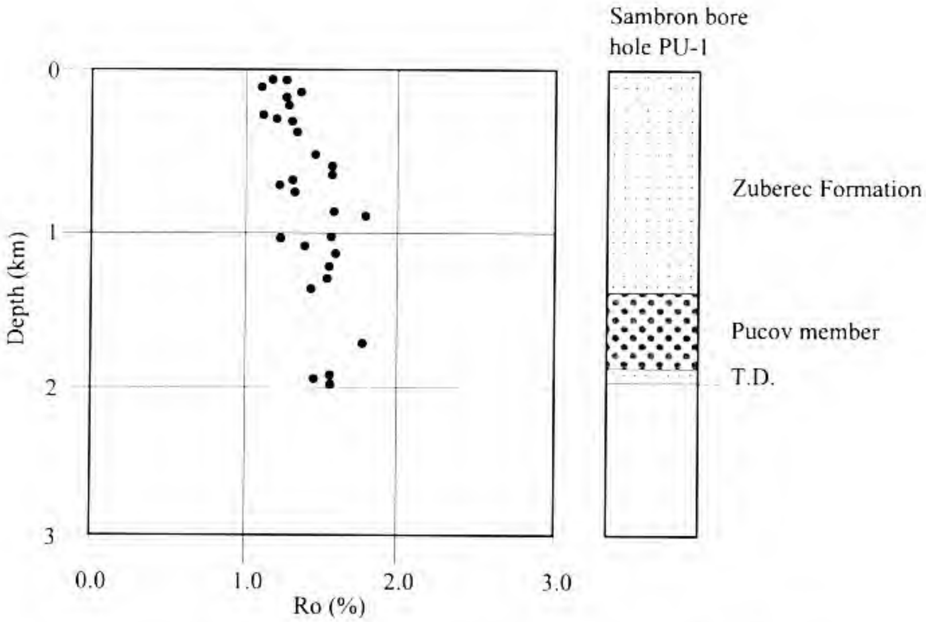


FIG. 13. Plot of vitrinite reflectance (Ro versus depth) for the Sambron PU-1 borehole (after Francu and Muller, 1983).

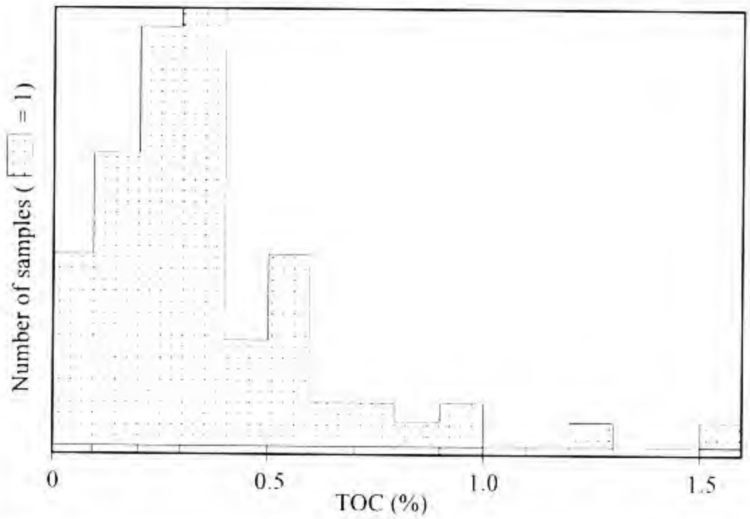


FIG. 14. Histogram of Total Organic Carbon (TOC) content of the Zuberec Formation in the Sambron-Lipany area (after Keith et al., 1991).

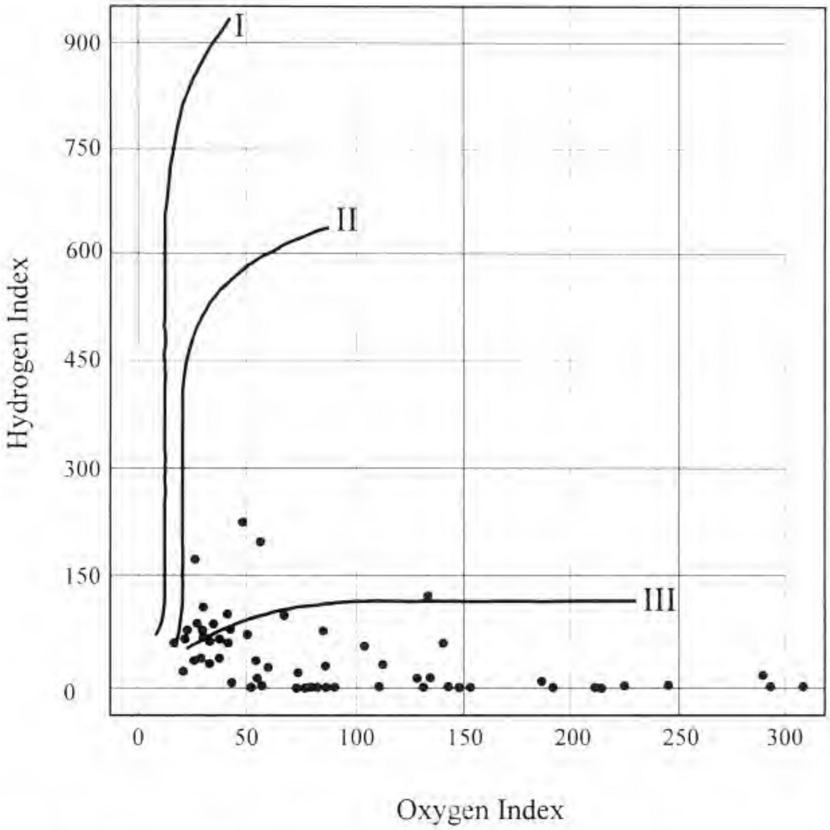


FIG. 15. Plot of Hydrogen Index versus Oxygen Index for the Zuberec formation in the Sambron-Lipany area (after Keith et al., 1991).

Stratigraphic traps are represented in the form of

(1) Pucov fans,
(2) turbiditic sandstone units,
(3) carbonate buildups on topographic highs and
(4) buried hills upheld by Mesozoic carbonates.

Combination structural/stratigraphic traps include folded Pucov fans and turbiditic sandstone units pinching-out on thrust toes.

Both horizontal and vertical migration paths can be envisaged. Lateral migration may have occurred through coarse-grained clastic units, along thrust planes, or along subhorizontal décollements. Vertical migration paths may be provided by highly permeable fractured zones, associated with strike-slip and normal fault systems.

Maturity analyses were only available from the Sambron, Lipany, Saris area (Fig. 3). As indicated by geohistory modeling, the maturity of the source-rocks in the basin sequence varies considerably due to individualized subsidence, structural and erosional histories of the different sub-basins of the CCPB. This variability is best indicated by the pyrolysis T_{max} determinations on borehole and surface samples in the structurally complex Sambron-Lipany area (Fig. 16). The location of wells discussed is shown in Fig. 4. Structures of this area comprise stacks of steeply dipping slices (Fig. 9), cut by a system of strike-slip faults (Fig. 3), accommodating inhomogeneous shortening. Burial histories of individual structures are highly variable.as indicated by T_{max} values which range from the top of the oil window (well Lipany-2), to within the oil window (surface, wells Lipany-3, 5,

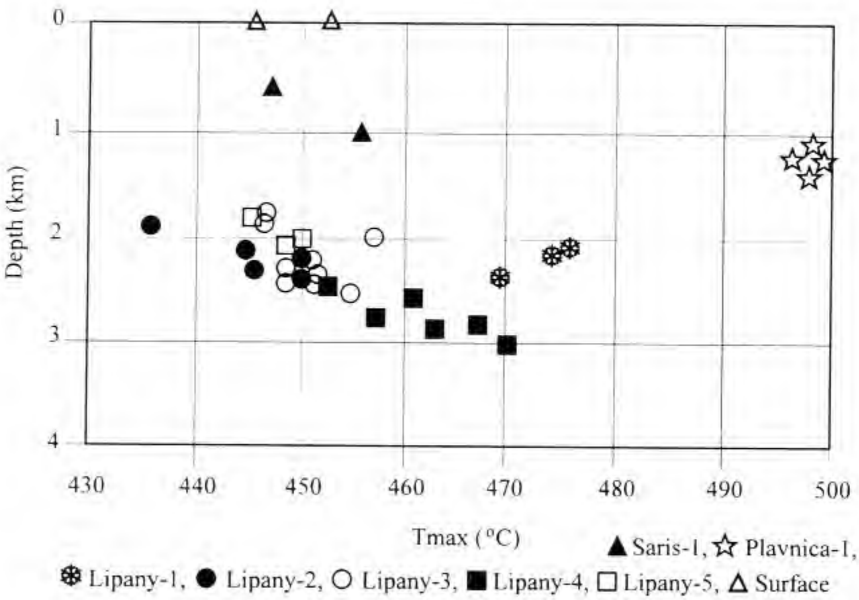


FIG. 16. Plot of Tmax versus depth for the Zuberec Formation from bore hole and surface data in the Saris-Lipany area (after Keith et al., 1991).

Saris-1), the bottom of the oil window (well Lipany-4), to the wet gas zone (well Lipany-1) or even higher maturity levels (well Plavnica-1). Analyzed samples were shales of the Huty and Zuberec Formations. Vitrinite reflectance data from the Sambron PU-1 borehole (Francu and Muller, 1983), mostly from the Zuberec Formation, indicate oil maturity to the onset of wet gas generation. Hydrocarbon generation modeling by BasinMod™ was carried out for most of areas; however, analytical data were only available for the Sambron-Lipany area. Figure 17 shows two models for the Lipany-1 well. Model A, is a simple one which assumes only sedimentary burial and erosion events (sed. mod.) and provides approximate results, whereas model B estimates sedimentary and tectonic burial and erosion (thrust mod.). Estimates of 1.5 km of missing sedimentary sequence, as indicated by vitrinite reflectance data (Francu and Muller, 1983; Korab et al., 1986), were taken into account. This is a lower limit of the suggested missing thickness range. As compared with analytical data (Fig. 16), an upper limit of missing strata of about 2-2.5 km appears to be appropriate.

Similar modeling, trying to estimate the missing parts of the sequence and taking the general asymmetry of the CCPB into account (Figs. 9 and 12), indicates that the Huty Formation and lower part of the Zuberec Formation entered the oil generation window at 30.5-27 Ma (sed. mod.) and 25-21 Ma (thrust mod.) in the Orava region, 31.5-31 Ma (sed. mod.) and 32-31 Ma (thrust mod.) in the Levoca Mts., 30 Ma (sed. model) and 23.5 Ma (thrust model) in the Sambron area, 29.5 Ma (sed. mod.; Fig. 17a) and 24.5 Ma (thrust mod.; Fig. 17b) in the Lipany area, 10 Ma (sed. mod.) and 13 Ma (thrust mod.) in the Celovce area, and 13.5 Ma (sed. mod.) and 15.5 Ma (thrust mod.) in the Presov Depression. The Huty Formation and lower part of the Zuberec Formation entered the gas generation window at about 20 Ma (sed. mod.) and 19-12 Ma (thrust mod.) in the Levoca Mts. In contrast, these source-rocks never reached the oil generation window in the Zilina, Liptov and Poprad depressions.

Fig. 17a indicates that, according to both models, the Lipany area is within the liquid hydrocarbon-generation window (sed. mod.: at depths of 2-2.8 km; thrust mod. at depths of 1.6-2.8 km). Here, the deformation occurred during middle Eocene to middle Sarmatian time (49-12.5 Ma), with the strongest shortening taking place between middle Eocene to Burdigalian times (49-17 Ma). As in other areas, hydrocarbon generation appears to have been contemporaneous with structural trap formation. Different ages of hydrocarbon generation in Celovce area and Presov Depression are caused by lower thicknesses of the Paleogene sequence and the lower Miocene burial.

CONCLUSIONS

The evolution of the CCPB was governed by the convergence of the African, the Apulian, and the European plates during the Alpine orogenic cycle. Main phases of the basin evolution can be summarized as follows:

(1) The morphology of the CCPB floor developed during the Late Cretaceous as a consequence of the emplacement of pre-Senonian Inner Carpathian nappes and by Late Cretaceous shortening of the Pieniny Klippen Belt.

(2) During the Paleogene, the West Carpathians thrust sheets advanced progressively towards the European Platform. Differential shortening and uplift of the area of the future Pieniny Klippen Belt, together with the emergent Inner Carpathians, accompanied the subsidence of the CCPB piggyback basin. The irregular island chain of the Pieniny Klippen Belt forming its northern boundary, was an effective barrier that created divergent paleocurrent systems in the Flysch Belt and the CCPB. In the latter over 4000 m of Eo-Oligocene sediments accumulated.

(3) Subsidence and sedimentation patterns in the CCPB were controlled by active thrust tectonics resulting in shifting of the basin axis, uplift of some areas and progressive basin shortening. During Oligocene-Egerian time, this basin was deformed and uplifted to the extent that sedimentation ceased and its fill was subjected to erosion. However, continued tectonic activity was accompanied by the development of strike-slip faults which accommodated the unequal north- or northeast-vergent thrust motion of different slices.

(4) During the Paleogene-Karpatian period, the Pieniny Klippen Belt was detached from its substratum and shortened. At the same time, the CCPB, the Flysch Belts, and some of the foreland molasse units were shortened. Maximum shortening occurred in the Pieniny Klippen Belt and the proximal parts of the Flysch Belt.

(5) At the end of the early Miocene, the area of the CCPB was progressively affected by extension, which spread from the hinterland and accompanied shortening of the frontal accretionary wedge. The last significant shortening in the Zilina and Orava parts of the Carpathian arc occurred during the late Badenian. Final shortening occurred in the area north of the Vysoke Tatry during the early Sarmatian and in the area east of the Vysoke Tatry during the middle Sarmatian (Nemcok et al., 1993). Later, NE-SW extension, accommodating frontal shortening, changed orientation to WNW-ESE, driven by subduction rollback in the eastern parts of the Carpathian arc (Royden et al., 1982, 1983a, 1983b; Nemcok, 1993).

The hydrocarbon habitat of the CCPB can be summarized to include the following play concept elements:

(1) Good seals are abundant and are scattered throughout the entire fine-grained portion of the sedimentary succession.

(2) Poor to fair quality reservoir units are developed which have poorly constrained shapes and are not easily predicted, particularly within the lower portion of the sedimentary succession.

(3) Fair to excellent source-rock horizons have been identified; however, they are not distributed throughout the sedimentary column. Although oil has been discovered in the basin, it appears to be more gas prone.

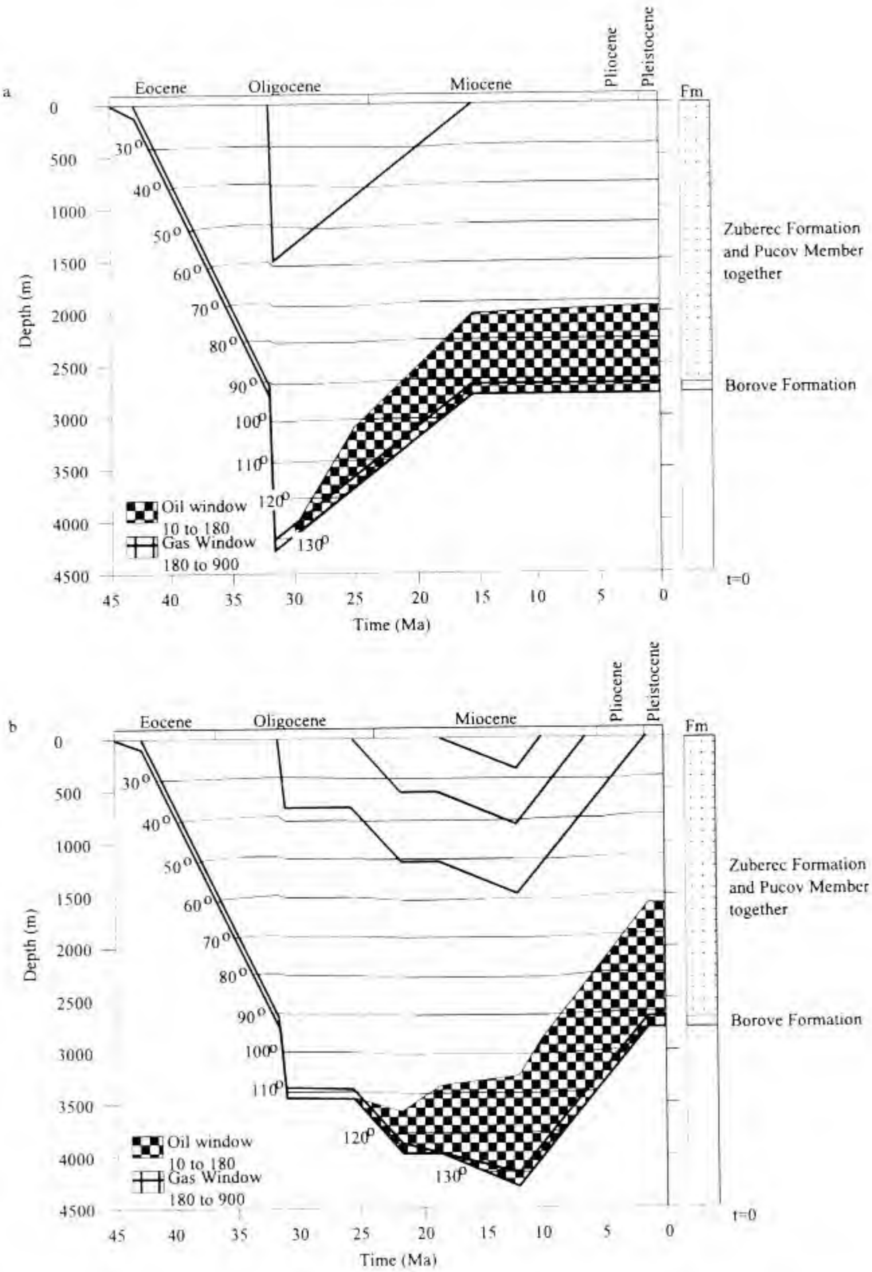


FIG. 17. BasinMod™ geohistory curve from Lipany-1 bore holedata (after Keith et al., 1991); a) sedimentary burial and erosion model, b) sedimentary and tectonic burial and erosion model.

(4) Both vertical and horizontal migration paths are provided for by faults and/or stratigraphic relationships.
(5) Traps were formed prior to or contemporaneously with the maturation of the source-rocks and the expulsion of hydrocarbons.
(6) Neogene tectonics, uplift and erosion may have caused destruction of some pre-existing accumulations.

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Structure and hydrocarbon habitat of the Polish Carpathians

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ABSTRACT

Integrated geological and geochemical studies provide an insight into the structural evolution of the Polish Carpathians and their hydrocarbon habitat and permit development of a consistent scenario for the history of their petroleum systems.

The Polish parts of the Western Carpathians are a classic fold-and-thrust belt which was largely thrusted northeastwards over the European Platform. This study centers on the Outer Carpathians and the adjacent European foreland, both of which are major petroleum provinces since a century.

The sedimentary sequence of the European Platform comprises thick Paleozoic and Mesozoic series, deposited in northwest trending grabens; these are unconformably overlain by Neogene Molasse. Although a few oil accumulations have been discovered in pre-Tertiary series, most of the gas produced in Polish Carpathians comes from the Neogene foredeep sequence.

The Outer Carpathians consist of several major tectonic units, involving Cretaceous and Paleogene flysch. These units exhibit strong lateral facies and thickness changes; these were induced by differentiation of the basin floor into cordilleras and sub-basins during successive compressional episodes, culminating in the Laramide inversion episode. During the Paleogene, flexural basin subsidence commenced. Beginning with the Late Oligocene, the basin geometry was modified by the successive emplacement of nappes, ending during Sarmatian times. These allochthonous units contain a number of accumulations which are reservoired in Cretaceous to Oligocene sands. Moreover, they contain the source-rocks for all oils discovered both in the autochthon and the allochthon. The Nosowka oil field, which may be sourced by Palaeozoic series, is an exception. The best potential source-rocks are the Early Oligocene Menilite shales which exhibit strong vertical and lateral variations in their Total Organic Content (2-15%) and Hydrocarbon Index (200-750); organic matter varies from a good type II (marine origin) in the lower part of the formation to a dominantly type III (continental origin) in its upper part. Additional potential source-rocks are the Albo-Aptian Spas and the Early Neocomian upper Cieszyn shales; however, these appear less prolific and their contribution to presently pooled oils has not yet been established.

Although present-day maturation patterns developed mainly after the nappe emplacement,

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initial source-rock maturation probably occurred during the pre-compressional stages in the southern part of the Silesian basin and the Dukla and equivalent basins, due to accumulation of thick sedimentary sequences. Integration of all data permits to postulate two episodes of migration: 1) Late Oligocene long distance migration from internal units, completed by later remigration of previously accumulated oils; this scenario may account for oil pooled in the Mesozoic platform reservoirs; 2) post-thrusting, short distance migration occurring within each basin from its deeper parts towards adjacent highs; this scenario probably accounts for oils pooled in the reservoirs of the allochthon. This two stage migration hypothesis requires testing by palinspastic basin modelling.

INTRODUCTION

Despite their complex surface geology, the Outer Carpathians were recognized as a major petroleum province already during the middle of the last century. Oil and gas fields were discovered in very distinct habitats (Encl. 1). Whereas gas fields are mainly restricted to the Neogene Carpathian foredeep, some oil fields are located in the autochthonous foreland, but most of the oil accumulations occur in the different allochthonous thrust units.

Although a common source-rock, namely the Oligocene Menilite shales, has been proposed for a long time for the oil accumulations contained both in the autochthon and allochthon (Ulmishek and Klemme, 1990), only recent geochemical analyses were able to establish a clear oil to source-rock link. This paper presents the results of integrated geological and geochemical studies and outlines the evolution of the petroleum systems of the Polish Carpathians. Particularly, the vertical and lateral distribution of the potential source-rocks, as well as oil to source-rock correlations, are compared with the present occurrence of hydrocarbons. The present and inherited maturation stages of these source-rocks are discussed in the light of present and past geometries of the fold-and-thrust belt in order to establish the timing of hydrocarbon generation episodes and to identify migration paths between kitchens and traps.

GEOLOGICAL FRAMEWORK

The area under discussion encompasses two distinct domains: 1) the Outer Carpathians, which consist of several nappes involving allochthonous Cretaceous to Paleogene terrigenous sequences, detached from their initial substratum during the Miocene thrusting phases, 2) the southern extension of the European foreland platform, which is partly overlain by Neogene synflexural sequences (essentially Miocene Molasse), deposited in the Carpathian foredeep, and which is partly overriden by the Carpathian nappes.

European Foreland

The **pre-Tertiary substratum** of the foreland is characterized by the development of a system of elongated NW-SE trending horsts and grabens, the Holy Cross Mountains, Miechow trough and Lublin basin, which were inherited from Variscan deformations and were subsequently affected by Late Permian to Jurassic rifting and Late Cretaceous to Paleocene inversion episodes (see e.g. Znosko, 1974; Ksiazkiewicz et al., 1977; Koszarski, 1985; Poprawa and Nemcok, 1988-1989; Osczypko et al., 1989). The Precambrian metamorphic and crystalline basement is overlain by Late Cambrian to Permian sedimentary formations. Among them, 1000 m of Devonian red clastics, followed by carbonates, were deposited within the Miechow trough. These are overlain by up to 400 m of Early Carboniferous carbonates and a few hundreds of metres of Namurian clastics; there is only indirect evidence for Late Carboniferous coal seams. The Precambrian to Paleozoic substratum is unconformably overlain by thick Mesozoic terrigenous and carbonate series which accumulated in grabens that were partly inverted during the Paleocene.

The **Carpathian foredeep** contains a Neogene Molasse which rests unconformably on truncated Mesozoic, Paleozoic strata or even Precambrian metamorphic rocks (Ney et al., 1974; Oszczypko and Slaczka, 1989). The Carpathian foredeep basin is narrow and relatively shallow (around 500 m) in the western part of the area, and widens and deepens (up to 2500 m) in the east (Wdowiarz, 1974). The southern parts of this basin are overriden by Outer Carpathian thrust sheets, the amplitude of which ranges from 45 km along the western cross-section (Encl. 2, section III, Roca et al., 1995) to some 70 km along the eastern cross-section (Encl. 2, section I, Roure et al., 1994). The Molasse consists mainly of fine and coarse clastic series which were deposited in a marine environment of variable salinity; its age ranges from late Burdigalian to early Tortonian (for further references see Oszczypko and Slaczka, 1989).

Outer Carpathian Units

Cretaceous and Paleogene flyschs are practically confined to the Carpathian allochthon and can be subdivided into distinct tectono-stratigraphic units, according to their structural position and lithostratigraphic composition (see e.g. Bieda et al., 1963; Kotlarczyk et al., 1985; Garlicka et al., 1989). From north to south, namely from the external to the more internal zones, the Polish Outer Carpathian flysch comprises the following units (Encls. 1 and 2, Fig. 1):

The **Stebnik unit**, which is mainly developed southeast of Przemysl in the Ukraine, is recognized to the southwest in wells Cisowa 1 and Kuzmina 1, located some 35 km south of the Carpathian thrust front (Ney et al., 1974; Ksiazkiewicz et al., 1977; Garlicka et al., 1989). This unit consists of a relatively thin, folded, mainly Miocene series, which includes lower Burdigalian evaporites and upper Badenian conglomerates; locally, it also involves uppermost Eocene and early Oligocene horizons, including the Menilite shales.

The **Borislav-Pokut unit**, which is today only exposed in the Ukraine, involves Early Cretaceous to Neogene series. In southeastern Poland, this unit is represented by small tectonic slices which outline the thrust contact between the Skole and Stebnik units south of Przemysl. The subsurface occurrence of the Borislav-Pokut unit beneath the Skole unit has been the topic of much debate (Wdowiarz and Jucha, 1981), due to its prolific hydrocarbon potential. Polish hydrocarbon exploration was initiated in this area some time before it spread to the Ukraine.

The **Skole unit**, which is restricted to the eastern part of the Polish Outer Carpathians, occupies large parts of the Ukraine. In Poland, it progressively pinches out northwestwards and disappears west of Tarnow. Where present, it consists of a thick and rather strongly deformed Early Cretaceous to early Miocene (Aquitanian or lower Burdigalian) flysch sequence. This nappe has been detached from its initial basement along Early Cretaceous black-shales.

The **Silesian unit** is present in the entire Outer Carpathians and corresponds to the major outcropping flysch nappe. It is made up of a thick sequence of latest Jurassic to Oligocene flysch series which is detached from its initial substratum along Early Cretaceous black-shales. At its northwestern front, the Sub-Silesian unit appears in tectonic slices and windows (Cieszkowski et al., 1985; Ksiazkiewicz et al., 1977). This Sub-Silesian unit is still identified southeastwards at Sanok (enl. 1) but probably disappears completely farther to the southeast. Unlike the Skole and Silesian nappes, the Sub-Silesian unit comprises only a thin sedimentary sequence, consisting of condensed Cretaceous and/or Paleogene strata which are frequently interrupted by hiatus (Ksiazkiewicz, 1960; Garlicka et al., 1989).

The **Dukla unit** represents the uppermost tectonic unit of the external Carpathian edifice of Poland. It only crops out in the eastern part of the Polish Outer Carpathians. Westwards, it is completely hiddened beneath the Magura nappe, the front of which even reaches the Silesian unit. There, its occurrence has been proved by deep boreholes and tectonic windows. Although still under discussion, the so-called Obidowa-Slopnice and Grybow sub-units could form its northwestern prolongation. The Dukla unit consists of thick,

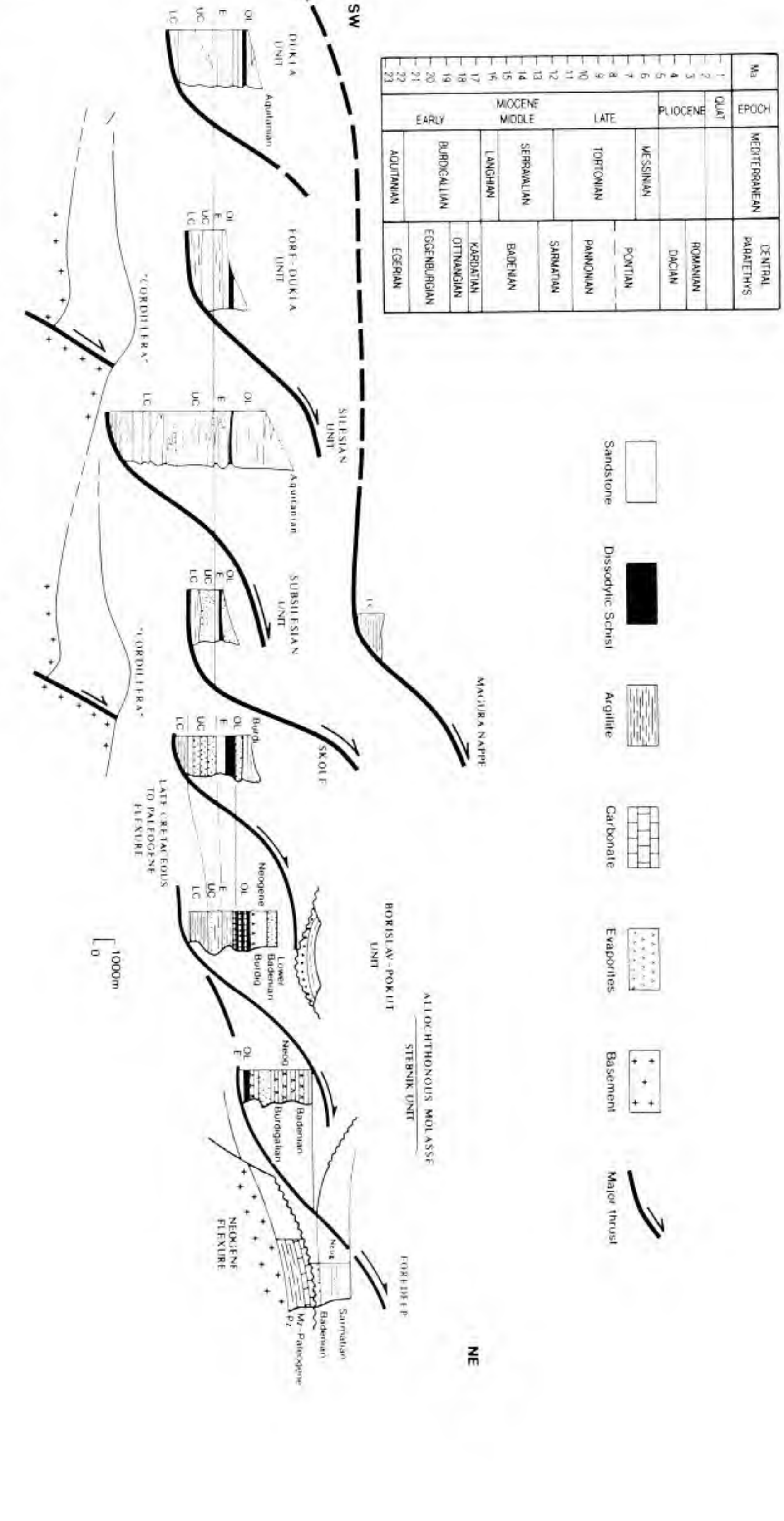

FIG. 1. Synthetic stratigraphic columns of major Outer Carpathian flysch units.

extremely deformed Cretaceous to Paleogene deep-water flysch and late Oligocene and Aquitanian Molasse sequences. In the central Polish Carpathians, the Dukla unit rests directly on top of the Silesian nappe. Elsewhere, it is separated from the latter by an additional tectonic unit, made up of Cretaceous and Paleogene condensed sequences, referred to as the "Fore-Dukla" unit in the southeast and as the "Michalczowa" unit in the west (Cieszkowski, 1992). Here, the Dukla frontal fault dips steeply beneath the south-verging backthrust structures of the Silesian unit (Ksiazkiewicz et al., 1977; Kusmierek, 1979, 1988). These units are considered as deformed paleo-highs.

The **Magura nappe**, which consists of Cretaceous to middle Miocene flysch sequences, represents the highest flysch unit of the Western Carpathians and separates the other external flysch units from the Pieniny Klippen belt (Cieszkowski, 1992; Oszczypko, 1992). It is presently restricted to the southwestern Polish Carpathians. However, isolated erosional klippen of Magura affinities are encountered further to the northeast on top of the Silesian unit, thus indicating that this nappe had previously extended further north.

The interpretation of the Sub-Silesian and Fore-Dukla units (and their equivalents) as paleo-highs or "cordilleras" (Ksiazkiewicz, 1965, 1975; Kusmierek, 1988, 1990, 1995) is based on the occurrence of repeated breaks in sedimentation and condensed sequences, but has also been inferred from paleocurrent studies, rapid thickness variations, submarine slides and turbidites on their slopes to the adjacent the basins. By applying cross-section balancing techniques and analogue modelling, the origin of these cordilleras has been related to multiple compressional phases, ranging from earliest Cretaceous up to the major Paleocene Laramide phase which induced inversion of Triassic to Jurassic extensional structures also in the Carpathian foreland (Ellouz and Roca, 1994; Roca et al., 1995).

These paleo-highs may have played a major role in the location of the décollement levels and, consequently, in the occurrence of complex and early structures (Roure et al., 1993, 1994). Early Cretaceous shales, which represent a potential décollement level, can be partly or totally eroded over these highs; this creates a discontinuity within the detachment level. In this respect, it is significant to note that major deformations, involving triangle zones or antiformal stacks of numerous duplexes, occur in areas which evolved as cordilleras during Cretaceous and Paleogene times. Examples are the Fore-Dukla (Roure et al., 1993, 1994) and the Sub-Silesian units (Roca et al., 1995). Other complex structures of the Outer Carpathians, such as the Borislav-Pokut zone in Ukraine, can also be interpreted as buried duplexes with coeval folding of the shallower units, or as triangle zones with associated backthrusts, which developed in response to the disappearance of the Early Cretaceous black-shale horizon. The Przemysl sigmoid itself is directly related to such lateral changes in décollement levels (see Encl. 1; Ellouz and Roca, 1994).

Main Stages of Mesozoic-Cenozoic Geological History

Palinspastic reconstructions of the Carpathians and adjacent areas since the Cretaceous have been attempted on the basis of balanced cross-sections (Ellouz and Roca, 1994). They show that the flysch sequences were deposited in elongated NW-SE trending basins. These basins are located in the prolongation of the NW-SE trending structures evident on the European Platform, as demonstrated at least for the external units of the Outer Carpathians. It is also evidenced that the same main tectonic events have controlled their structural and depositional history prior to emplacement of the nappes onto the foreland.

The stratigraphy (Fig. 2) and lithology of both foreland and allochthonous units have been the subject of extensive field work, reported in many papers and maps (see e.g. Ksiazkiewicz et al. 1962; Karnkowski and Oltuszyk, 1968; Sokolowski, 1976; Osika, 1980; Kotlarczyk et al., 1985; Koszarski, 1985; Poprawa and Nemcok, 1988-1989).

Following the Variscan orogeny, the Central European Platform underwent regional extension during Late Permian to Jurassic times. This led to the development of a complex pattern of intracontinental rifts and the opening of an oceanic domain in the internal Carpathians. In the European fore-

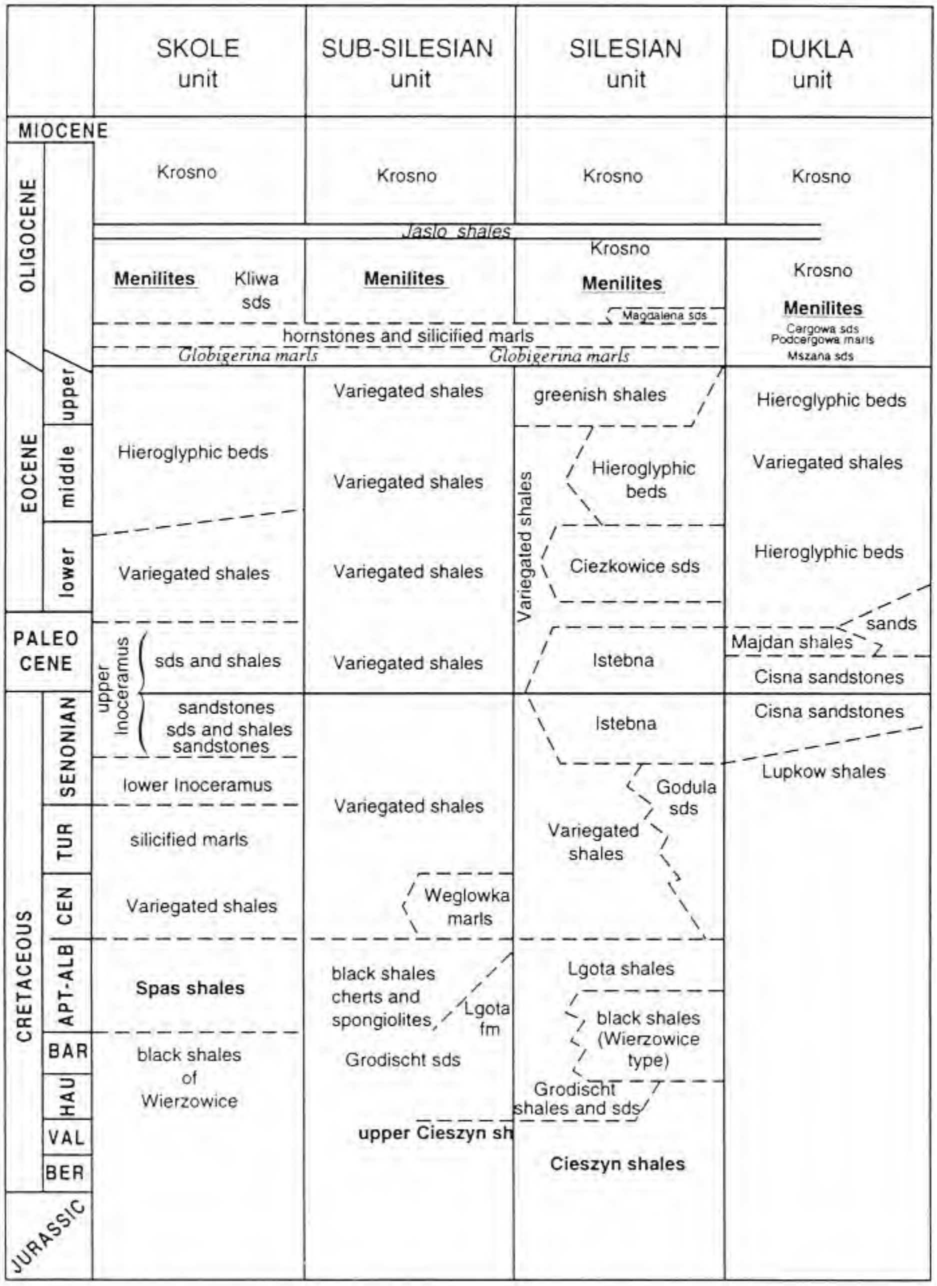

FIG. 2. Schematic stratigraphic chart of Polish Carpathians (after Gucik et al., 1962, Bieda et al., 1963).

land, major NW-SE trending structures developed during this extensional phase, including depressions and flanking highs. The now inverted Pomorze-Holy Cross Mts-Przemysl ridge corresponded to the axis of a major depression which was the site of dominantly marine sedimentation during the Jurassic and the entire Early Cretaceous. In contrast, the Miechow and Lublin basins, which flanked these troughs, were characterized by continental to shallow marine sedimentation (Koszarski, 1985; Kotlarczyk et al., 1985). At the end of Triassic, the Eo-Kimmerian phase caused uplift and erosion in the Polish Carpathian foreland. This phase has been interpreted either as compressional (Znosko, 1974; Kotlarczyk et al., 1985) or as resulting from transtensional regime (Gradinaru, 1984). During the Dogger, syn-rift tectonic subsidence led to sedimentation of clastics deposited within paleotopographical depressions in a continental and a progressively more marine environment. The overall transgression culminated during the Malm with the deposition of carbonates which reach 1500 m in thickness in the area of Tarnow-Debica. Carbonates of this age are also known from the northern margin of the Silesian and Skole basins. Southeastwards, they are unknown but are expected to be present on the basis of palinspastic considerations.

The Jurassic-Cretaceous boundary is characterized by the occurrence of tectonic movements (Late-Kimmerian phase) which can be also related to a transtensional phase. In the domain of Outer Carpathians, this phase may have initiated development of the so-called cordilleras, the presence of which is indicated by facies and current patterns since the beginning of the Early Cretaceous (Ksiazkiewicz 1960, 1965; Winkler and Slaczka, 1992). Although the entire area subsided in response to lithospheric cooling (Roca et al., 1995; Ellouz and Roca, 1994), the incipient cordilleras were repeatedly reactivated during the Cretaceous.

During the Early Cretaceous, the foreland was a dominantly continental area and underwent erosion. In the Outer Carpathians, the Early Cretaceous series is well developed in the Silesian basin (up to 800 m) and in the Skole basin (about 500 m). In the Dukla and equivalent basins, only very fragmentary data suggest that the Early Cretaceous series was deposited in a shallow to deep marine environment. It is still debated whether this basin formed part of the Silesian basin, at least during the Early Cretacous, or whether it was already well individualized. According to Ellouz and Roca (1994), the Michalczowa and the Grybow ridges, located to the north and south of the individualized Dukla basin, respectively, acted as major clastic source-arear for the Magura and Silesian basins during the Late Cretaceous.

The lowermost members of the flysch series of the Silesian and Sub-Silesian units are the Berriasian lower Cieszyn shales and limestones, and the Valanginian upper Cieszyn shales (Fig. 2). Sediments of Valanginian age, ascribed to Cieszyn formation, have also been identified in the Fore-Dukla unit (Rabe profile; Kusmierek, 1979) and could be also present in external part of the Silesian unit (Czarnorzeki profile; Kusmierek, 1985). Shaly Barremian to Albian series occur in Silesian and Skole basins (Wierzowice shales overlaid by Lgota shales in Silesian basin, Spas shales in Skole basin). Locally, these are interrupted by thick turbiditic sandstones (Grodicht in Barremian-Aptian) which were supplied from the northern margin into the Silesian and Sub-Silesian basins. At the beginning of Cenomanian, global sea-levels rose and the foreland was widely transgressed. Sands, grading upwards into marls and limestones, were deposited in the foreland while the Carpathian flysch basins were starved; in the Silesian basin, cherts and radiolarites are overlain by variegated shales, and in the Skole basin siliceous marls were deposited. However, in response to progressive uplift of the Silesian ridge, an increasing amount of sand was supplied to the southern part of the Silesian basin where it accumulated in the large deep sea-fans of the Godula formation. In the Magura and Skole basins, clastic sedimentation resumed only during the Senonian (Inoceramus formation). The clastics of the Skole basin were derived from its northern margin. In the Dukla basin, clastic deposits (Lupkow shales) are as old as early Senonian.

Whereas emplacement of the Inner Carpathian thrust-and-fold belt was completed during Turonian time, the Pieniny Klippen Belt was emplaced towards the end of the Late Cretaceous. During this Laramide phase, structures were strongly reactivated and the Polish trough in the foreland was inverted, resulting in uplift of the Pomorze-Holy Cross Mts. megaridge and concommittant isolation of Miechow and Lublin depressions. Thrust loading of the European foreland by the stacked nappes of the Inner Carpathians caused its progressive flexural subsidence and the development of the Paleogene Carpathian foreland basin. The latter was characterized by strong thickness variations, reflecting a succession of basins and cordilleras, but in a general southeastward thickening pattern. From the late Oligocene onwards, this pattern was modified by progressive emplacement of the Outer Carpathian nappes and concommittant emergence of cordilleras, supplying clastics to adjacent basins.

During the Paleocene, the Silesian cordillera continued to supply great amounts of sands to the Silesian basin (upper Istebna formation); these sands covered most of the basin and pass upwards into variegated shales. In the Skole basin, the Inoceramus sands and shales grade upwards into variegated shales, whereas in the Dukla basin, the Cisna sands are overlain by the dark Majdan shales. Clastic transport directions indicate the development of a new source-area south of the latter (Ksiazkiewicz et al., 1962). During the early Eocene, Hieroglyphic beds were deposited in Dukla basin, whereas in the Skole, Sub-Silesian basins and the outer part of the Silesian basin, variegated shales and thin-bedded flysch accumulated. At the same time, thick-bedded conglomeratic Ciezkowice sandstones were supplied to the southern part of Silesian basin. During the middle Eocene, sedimentation became more homogeneous in the Outer Carpathians due to a general decrease in tectonic activity. Regionally, the Hieroglyphic beds pass upwards into mainly pelitic, mostly shaly sediments. These are overlain by a thin horizon of Globigerina marls which was deposited over extensive areas of the Outer Carpathians. The age of this regional marker horizon is variably given as intra-late Eocene (Koszarski et al., 1974; Ksiazkiewicz et al., 1962), latest Eocene (Bieda et al., 1963; Poprawa and Nemcok, 1988-1989) or as straddling the Eocene/Oligocene boundary (Osika, 1980; Kusmierek et al., 1985).

During the Oligocene, the Menilite-Krosno formation was deposited in the Outer Carpathian domain (Fig. 2). Its basal member is made up of the sub-Menilite shales, cherts and cherty marls; in the Dukla basin, these are replaced by a dominantly sandy sediments (Mszana and Cergowa sands). The Menilite shales s.s. are dark, bituminous shales which were deposited in a pelagic or hemipelagic environment under anoxic conditions. Sands are only locally developed (Magdalena sands in the southwestern part of the Silesian basin, Kliwa sands in the Skole basin). The thickness of the Menilite shales is highly variable, ranging from 100 m in the Silesian basin to 900 m in the Dukla basin (Mruk and Kusmierek, 1991, unpubl. data). On the basis of nannofossils, where present (are often dissolved due to high organic content), an early Oligocene age is indicated for the Menilite shales (NP 21-NP 22; Muller, personal communication). These shales are overlain by the Krosno flysch, grossly dated middle-late Oligocene (NP 24-NP 25). However, the transition from Menilites to Krosno facies is gradual and diachronous, as attested by the isochronous Jaslo shale marker (Bieda et al., 1963; Jucha, 1969). Through time, the Krosno depocenter shifted progressively northeastwards from the Dukla basin (thick Krosno beds older than the Jaslo horizon) to the Skole basin (Jaslo bed at the boundary between Menilite and Krosno deposits). The Menilite-Krosno formation, as a whole, exhibits considerable thickness variations and reaches several thousands of metres in the basin centres (up to 2500 m in Dukla basin, 4000 m in Silesian basin and 1200 m in Skole basin), whereas it is absent in areas corresponding to cordilleras. During the deposition of the lower Menilite shales, these cordilleras corresponded to submerged paleohighs (no reworked material in condensed Menilite sections adjacent to these cordilleras). Later, in the middle-late Oligocene, their role as detrital sources to adjacent basins is well demonstrated (Fig. 16; Ksiazkiewicz et al., 1962; Kusmierek, 1988, 1990; Mruk and Kusmierek, 1991, unpubl. data).

In the following, we give a tentative timetable for the emplacement of the different tectonic units of the Outer Carpathians (Roure et al., 1994; Roca et al., 1995). The proposed sequence of events should not be seen as consisting of quasi-instantaneous tectonis episodes. More likely, it reflects continuous deformation during late Oligocene to mid-Sarmatian times, accounting for a large amount of supra-crustal shortening, as evident from the structural cross sections (Encl. 2). Uncertainties in this time-table arise from the erosion of the youngest deformed flysch and of the neo-autochthonous Molasse sequences (-

Ksiazkiewicz et al., 1977; Karnkowski, 1983, 1986; Oszczypko and Zytko, 1987).

The Magura nappe was mainly structured during the late Oligocene. During the early Miocene, while sedimentation of the upper Krosno formation continued in the Skole basin, the Dukla/Obidowa-Slopnice unit and the Fore-Dukla/Michalczowa unit were emplaced. The Silesian, Sub-Silesian and Skole units were mainly deformed and emplaced during the late Burdigalian and early Badenian. At this time, the present Carpathian foredeep was formed and filled with Molasse-type sediments. Its innermost parts, the Stebnik and Borislav-Pokut units, were deformed during the Sarmatian. With this, orogenic movements ended in the Outer Carpathians and the area was subjected to post-orogenic uplift and erosion.

DISTRIBUTION AND CHARACTERISTICS OF SOURCE-ROCKS

Extensive geochemical surveys were performed on the sedimentary series of the Outer Carpathians. More than 700 rock samples were collected from outcrops and cores, resulting in a high density data set for the eastern part of the Polish Outer Carpathians, whilst in their western parts sampling was more erratic (Encl. 1). In addition, few samples of the autochthonous series were collected in wells. All samples were first analyzed by Rock-Eval pyrolysis to identify potential source-rocks by their total organic carbon content (TOC), quality (HI), as well as their maturation stage (Tmax). Subsequently, additional analyses were carried out on selected samples of identified source-rocks, in order to precise their type (elemental analysis on kerogen, extract analysis, but also pyrochromatography at 550°C and preparative pyrolysis; Vandenbroucke et al., 1988). In addition, a large amount of surface samples were analyzed and dated on the basis of their nannoplankton content (Muller, personal communication).

Potential Source-Rocks In Autochthonous Series

Only very few analyses were carried out on the **pre-Cenozoic formations** of the foredeep and the nappe substratum. In the well Zagorzyce 1, a few intercalations of Early Carboniferous black-shales/silty shales were found to have a TOC around 1.5%, but a low HI around 100 mgHC/gTOC. In a well located in the Western Outer Carpathians, some Late Carboniferous coal beds show a good petroleum potential (TOC=47% and HI=450).

More than 80 samples of the **Miocene** autochthonous series were analyzed (Fig. 3; Kotarba et al., 1987). Samples from the western, central and eastern parts of Polish Carpathian foredeep are characterized by low TOC values, in the 0.5% to 0.8% range, and HI less than 120. The slight differences observed from one area to another are not significant. The Rock-Eval results and chromatograms on extracts (Kotarba et al., 1987) suggest a continental origin for this organic matter.

Potential Source-Rocks In Allochthonous Series

In the Outer Carpathian nappes, all formations were sampled and are discussed below, starting with the less organic-rich ones.

The **Late Cretaceous-Paleocene formations** (Istebna/Inoceramus formations) have very low average TOC (0.6%) and HI (<130 mgHC/gTOC). The **Eocene** Variegated shales and Hieroglyphic beds have a lower average TOC (0.3%) and a HI in the same range. In both cases, the organic matter is of probable type III (Fig. 4). It is concluded that these formations are essentially devoid of potential source-rocks. The **Late Cretaceous** Lupkow shales, which are restricted to the Dukla basin, have been sampled in only one site where they are overmature; their TOC averages 1%. One sampled site is not sufficient to make a conclusive statement about the source-rock potential of this series.

Early **Cretaceous black-shales** were sampled in different units (Fig. 7): presumably upper Cieszyn formation in Silesian unit (Stepina area)

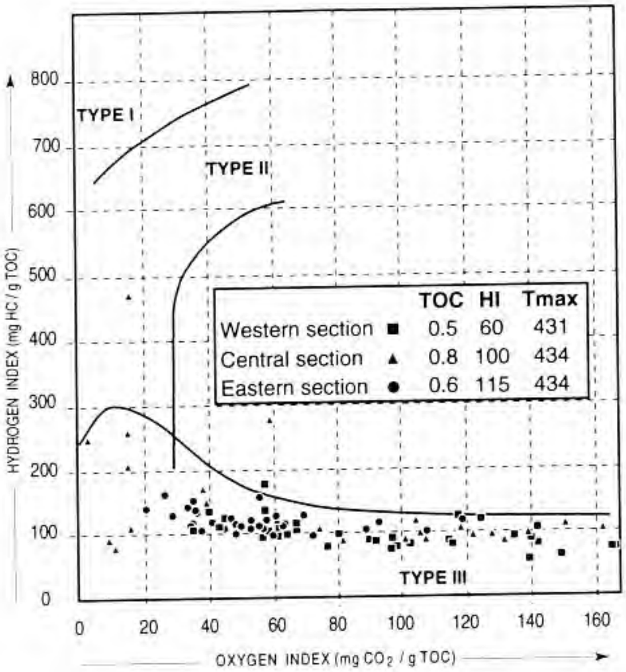

FIG. 3. Hydrogen Index-Oxygen Index diagram for Miocene formations in the Carpathian foredeep.

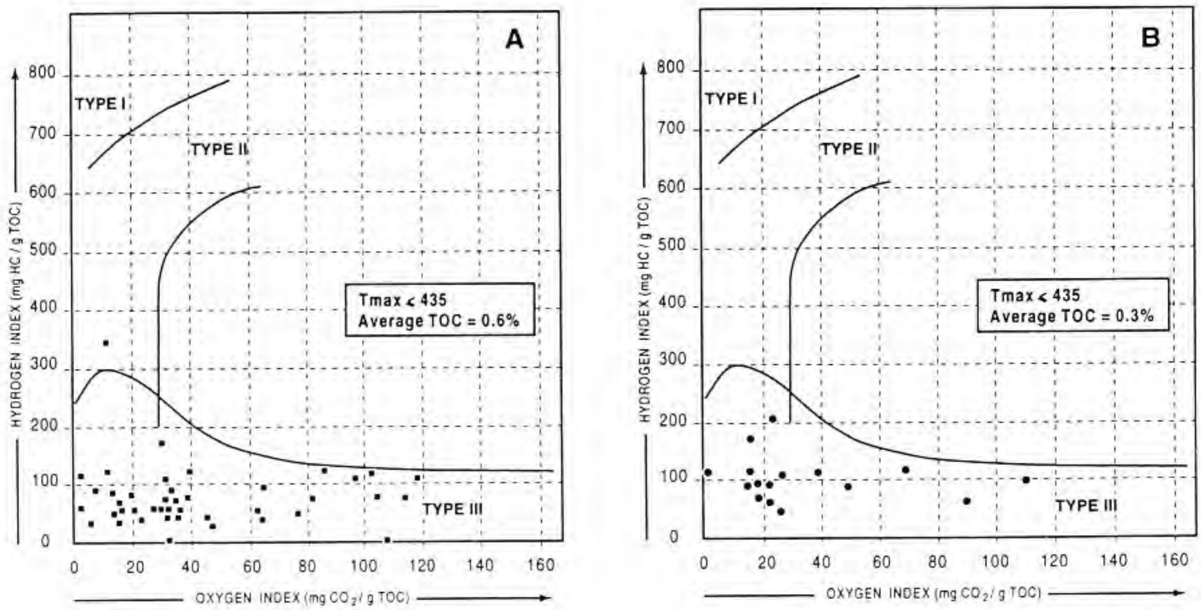

FIG. 4. Hydrogen Index-Oxygen Index diagrams: A) for Late Cretaceous-Paleocene formations (Istebna/Inoceramus formations), B) for Eocene formations (Variegated shales, Hieroglyphic beds).

and Fore-Dukla (Rabe area), Spas formation in Skole basin (Krzeczkowa area, Kuzmina 1 and 2) and the partly time-equivalent Lgota formation of the Fore-Dukla unit (Rabe area).

The **Neocomian upper Cieszyn member** can attain a thickness of 300 m and consists of alternating thin-bedded sandstones, shales and scarce siderites with locally high sand content. Berriasian to Valanginian shales sampled in the Stepina area have TOC values ranging from 1.5 to 3% (average 2%), but rather low HI of 100 mgHC/gTOC. In the Rabe area, Valanginian shales yield TOC up to 1.5 with HI of 120, despite their high maturation level (T_{max}>455) corresponding to the end of the oil-window (Espitalié et al., 1985-1986). The organic matter is of type III according to the HI-OI diagram (Fig. 6), kerogen analyses (Fig. 10) and chromatograms on extracts (Fig. 5). These shales could be considered as a potential source-rock, especially in the Rabe area. Although the facies of this formation is considered to be monotonous over wide areas (Ksiazkiewicz et al., 1962), it is questionable whether the results of limited analyses can be extrapolated to the whole area.

The **Albo-Aptian Spas formation**, is predominantly made up of black-shales which in their lower part contain sandstone intercalations. The Spas formation has a rather uniform distribution within the Skole basin. The TOC of these shales varies between 1.5% and almost 4% (average of 2%), thus expressing a vertical variability within this formation. Their average petroleum potential remains rather low : HI from 90 to 180 for a T_{max}=442°C (i.e. at the beginning of oil-window). Chromatograms on extracts (Fig. 5) and elemental analysis of kerogens (Fig. 10) clearly indicate a type III for this organic matter. Despite a rather low petroleum potential, the Spas shales can be considered as a reasonable potential source-rock. Considering the homogeneous facies development of this formation in the entire Skole basin, results of our analyses may be applicable for the whole basin.

The **Albian Lgota formation** is widely developed within the Silesian and Sub-Silesian basins and also in the Fore-Dukla zone. Dark shales are present in the middle part of this formation and alternate with thin-bedded sandstones. Their average TOC and HI are low (1% and 100 mgHC/gTOC, respectively). Accordingly, this formation has questionable source-rock characteristics. Nevertheless, the scarcity of sampling and high lateral facies variations do not allow definite conclusions on its source-rock potential.

The **early Oligocene Menilite shales** have been extensively sampled in outcrops and a few wells in the central and eastern parts of the region (Fig. 7). In contrast, in the western part, sampling was limited due to outcrop conditions and the làck of cores. The Menilite shales have been considered since a long time as a very good source-rock; our study confirms this assessment : high TOC up to 15% and high HI up to 700 mgHC/g TOC have been measured on some samples. However, our analyses point out more clearly the high vertical as well as lateral TOC (2% to 15%) variability of this formation. This is well exemplified by two stratigraphic cross sections through the Outer Carpathians (Fig. 8). A similar variability is observed in terms of quality of the organic matter with HI ranging from 200 to 750 mgHC/gTOC (Fig. 9). The wide scattering of the values observed in the OI/HI diagram is due to variations in quality of the organic matter, as al! the samples but those from the Dukla unit are immature. Accordingly, the organic matter ranges from a very good type II to a type II + III.

In contrast, samples from late Oligocene Krosno formation have systematically low TOC (<1%) and low HI (<200), regardless of which area is considered. The organic matter is of type III (Fig. 9). These geochemical results are consistent with a high input of terrigenous material.

The variations in content and quality of organic matter in the Menilite formation were studied in more details at two sites in the Silesian basin-Lukawica (Fig. 11) and Rudawka-Rymanowska (Fig. 12) -and at one site on the southern margin of the Skole basin-Frysztak 3 well (Fig. 13). At these locations, the approximate thicknesses of the Menilite formation are 120, 250 and 300 m, respectively. Although sampling densities were not equal, the following general observations can be made:

(1) below the Globigerina marls, the so-called sub-Menilite shales, when present (A and possibly F1), have a middle TOC and HI (<300) and are very likely of type III.

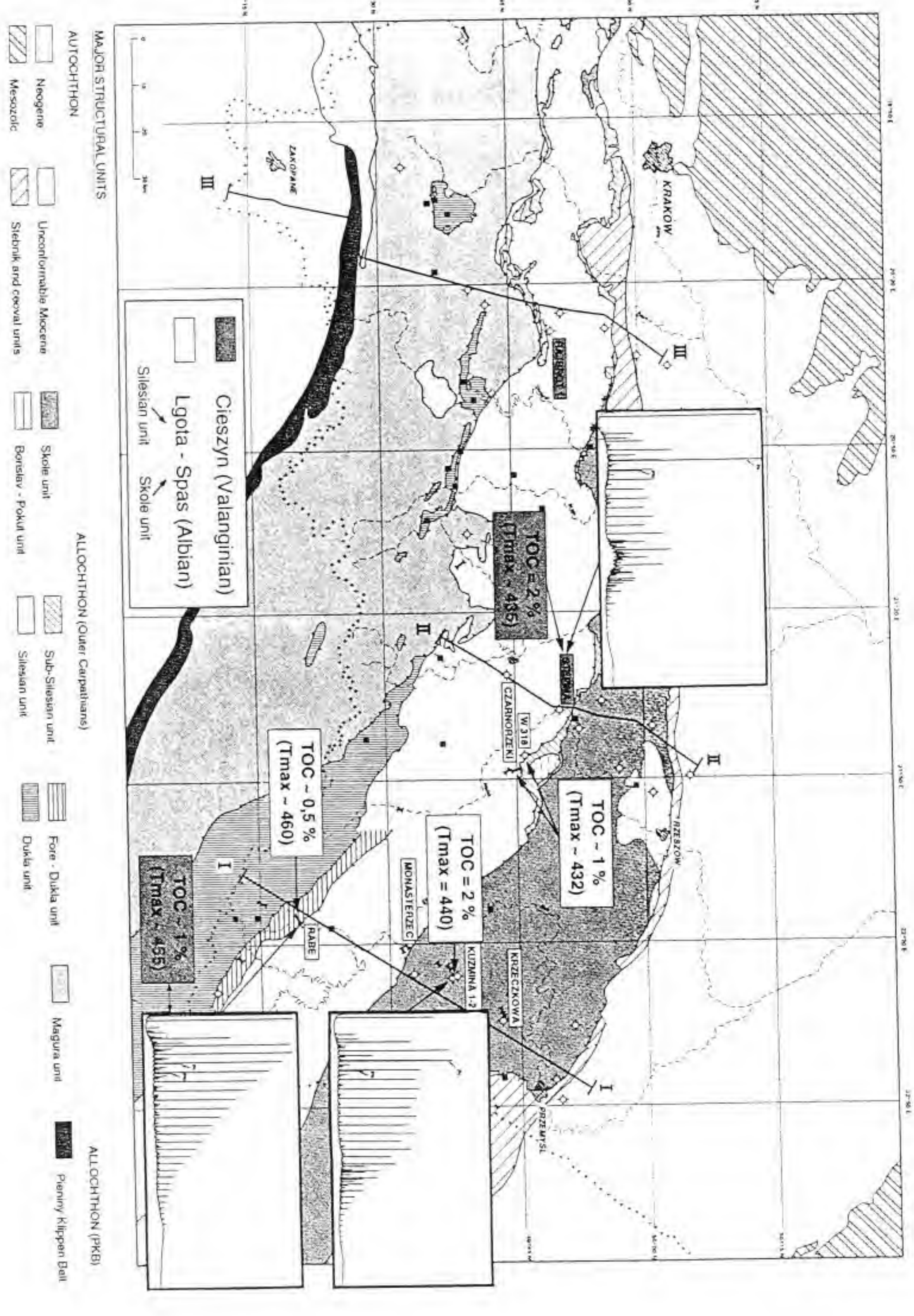

FIG. 5. Source-rock characteristics of Early Cretaceous Cieszyn and Lgota/Spas formations.

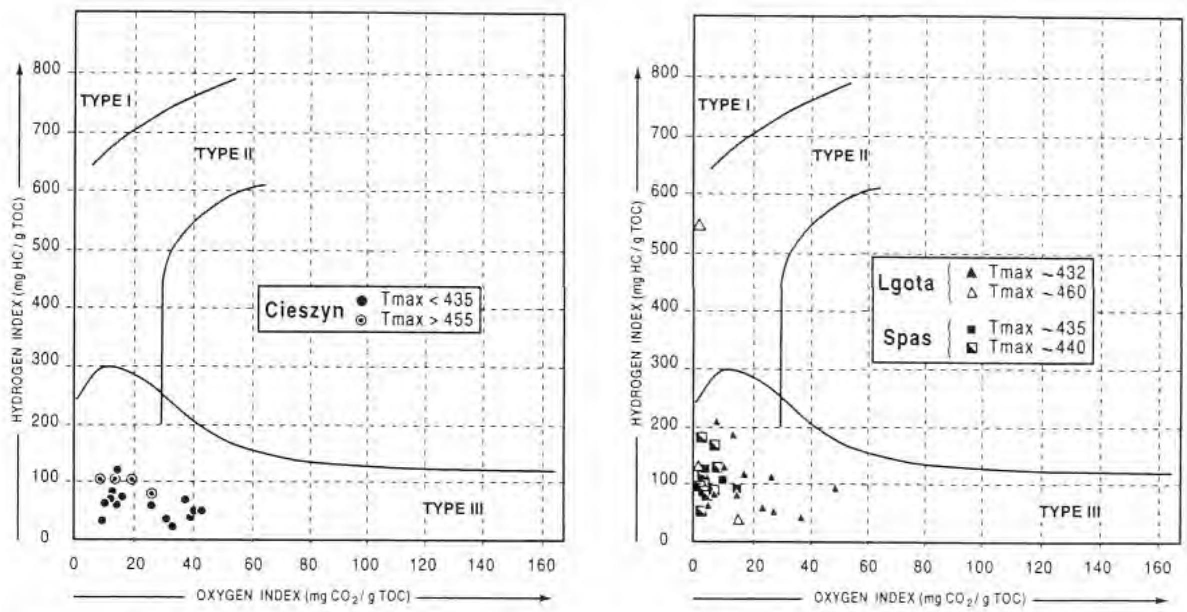

FIG. 6. Hydrogen Index-Oxygen Index diagram for Early Cretaceous formations

(2) the highest TOC and HI are consistently encountered in the lowest part of the Menilite shales, just above the Globigerina marls (samples F2, L1, L2, B). The organic matter is of very good type II as demonstrated by chromatograms and location of the kerogens in a Van Krevelen diagram (Fig. 10); its marine origin is confirmed by petrographic studies which show that it is essentially amorphous, and probably derived from marine phytoplankton. However, a very low contribution of terrestrial material (ligno-cellulosic debris) is present.

(3) higher up in the sequence, the Menilite shales exhibit a progressive decrease in TOC which corresponds to a more or less well marked decrease of the HI. This reflects primarily an increasing admixture of continental organic matter to the marine organic matter (C, D, F3, L3). Still higher in the formation (L4, F4 to F6), the terrestrial organic matter becomes predominant. This terrestrial input is marked by large amounts of saturated hydrocarbons in the C20-C30 range, a CPI>1 and a Pr/C17>>1 (the latter also confirms the low maturity level of the analysed samples).

Terrestrial input varies considerably within the Menilite formation. At the Frysztak 3 site, most of the organic matter is of continental origin. In the easternmost part of the Skole unit (Kniazyce, Gora Krepak), most of the lower Menilite series is characterized by predominantly terrestrial organic matter exhibiting low HI (<300), but alternatively high TOC (>8%) or low TOC (<3-4%) (see also Figs. 9 and 11). These results suggest at a basin scale a very heterogeneous pattern of "pure" marine versus dominantly continental organic matter within the Menilite formation and, thus, reflect a complex palaeogeographical framework during deposition of this formation.

Only few papers have been published on the depositional conditions prevailing during the accumulation of the Menilite formation in the Ukrainian Carpathians (see Koltun, 1992): development of the Menilite facies has been ascribed to the presence of an oxygen-depleted zone on the shelf and upper continental slope due to upwelling conditions, whilst the Krosno beds were deposited in deeper, more-oxygenated waters. In the Polish Carpathians, synchronous occurrence of the organic-rich (lower) Menilites, regardless of their location in the basin, speaks for a regional control on factors responsible for the accumulation of organic matter, such as a relative sea-level rise, possibly

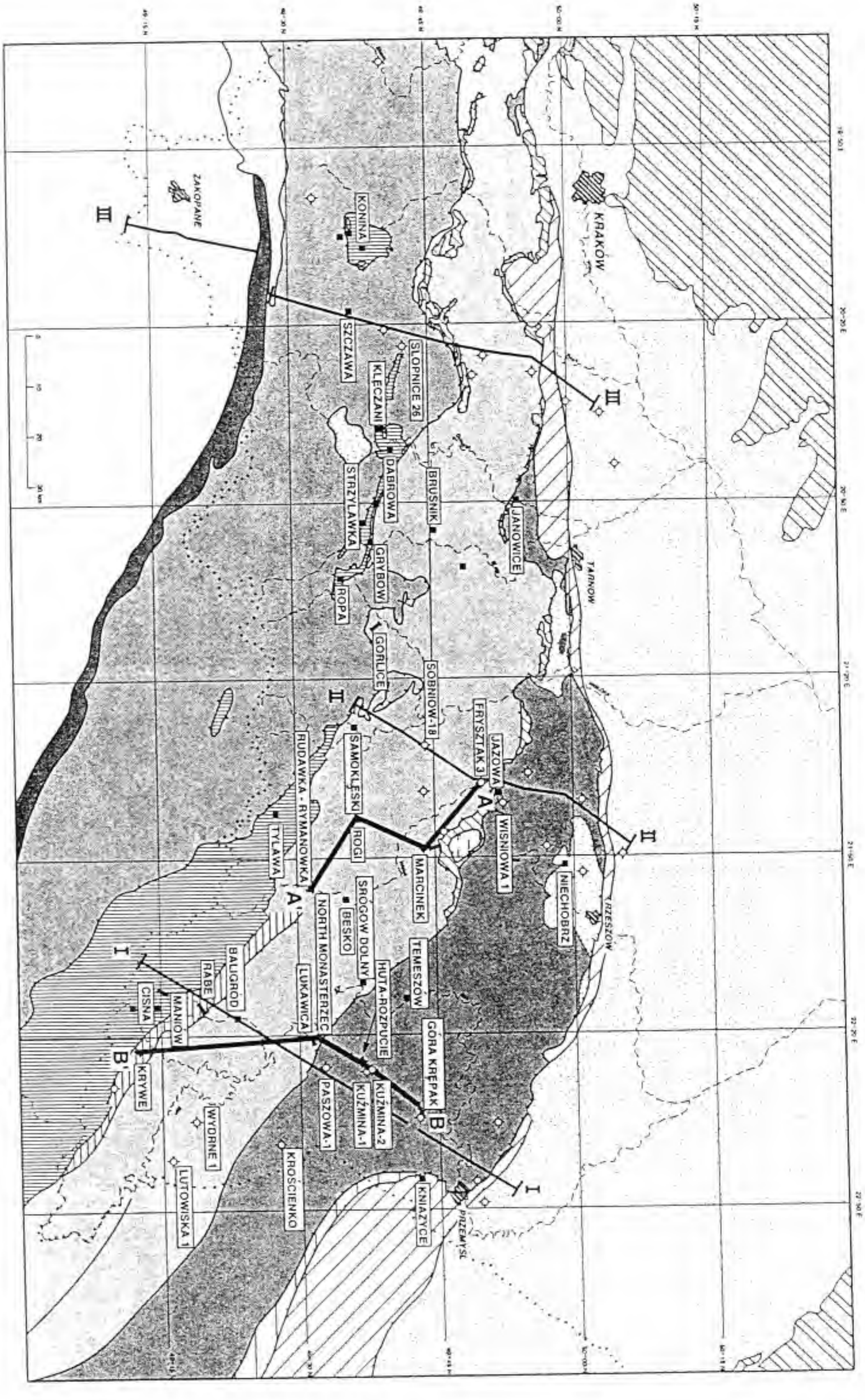

FIG. 7. Lacation map of Menilite and Krosno formations samples (wells and outcrops) and cross-sections A-A' and B-B', give in Fig. 8, and regional sections, given in Encl. 2.

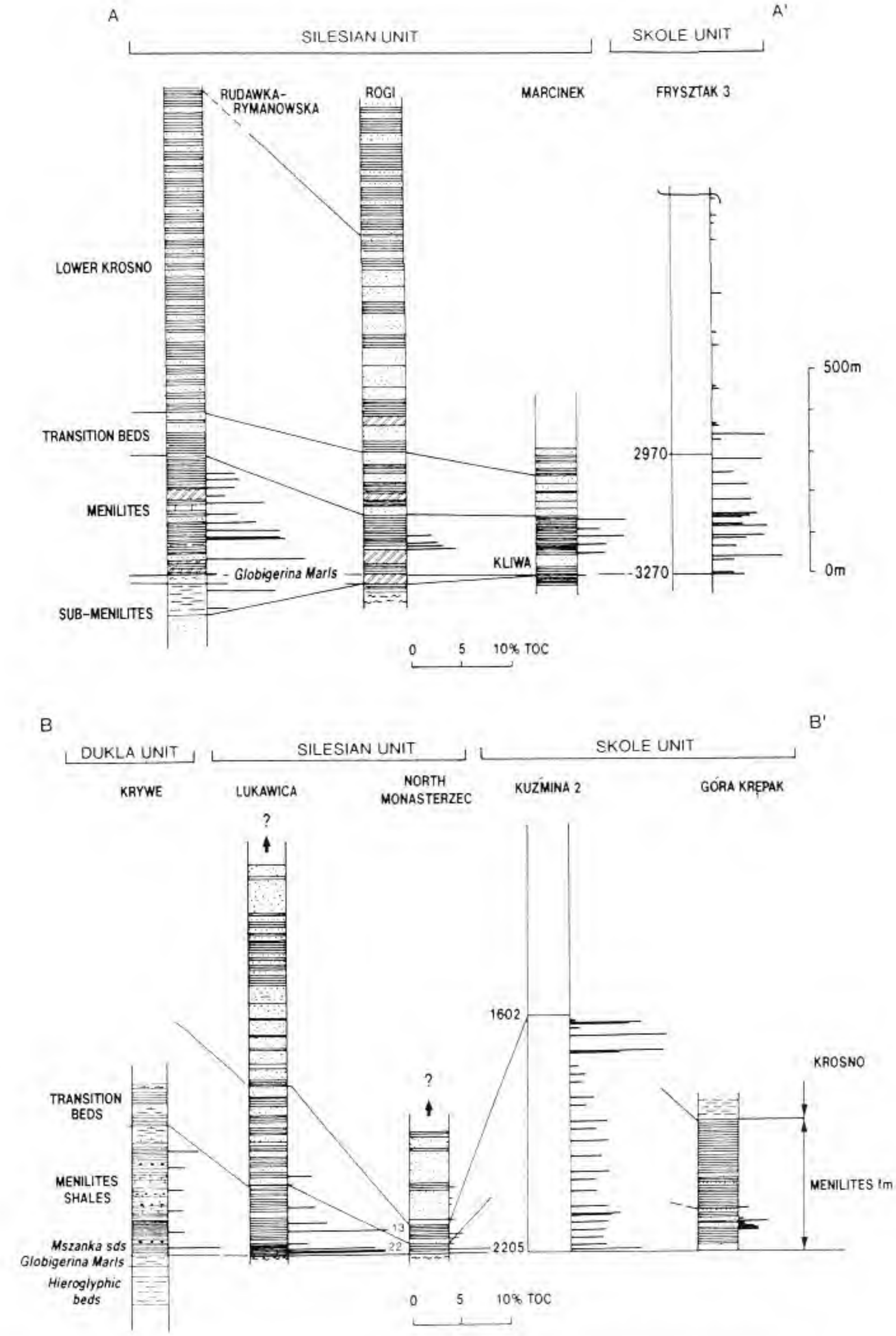

FIG. 8. Stratigraphic ross-sections A-A' and B-B', showing variations in thickness and TOC values of the Menilite shales (top of Globigerina marls has been taken as a horizontal marker).

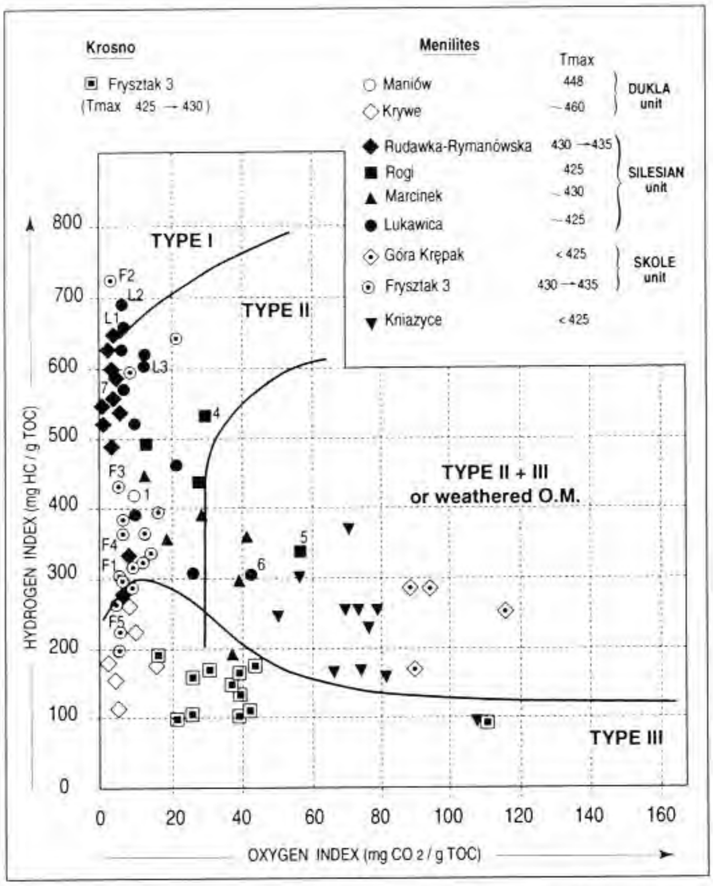

FIG. 9. Hydrogen Index-Oxygen Index diagram for Menilite and Krosno formations

enhanced by restricted bottom water circulation due to a complex sea-floor topography involving basins and cordilleras. The complex vertical as well as lateral variations in terrestrial input may be related to the interference of such factors as the emergence of cordilleras supplying terrigenous material to the adjacent sub-basins, and gravitational mass flow processes which locally can strongly modify the organic facies as seen, for example, in lake Tanganyika (Huc, 1988). In this respect, the eastern part of the Skole basin differs from the other sub-basins in so far as terrestrial influx from the platform commenced already at the beginning of the Oligocene.

In conclusion, the Menilite formation contains the best potential source-rocks. However, the observed substancial lateral and vertical variations in organic content and quality require high density data for a reliable evaluation of the petroleum potential of the Carpathians. Several Early Cretaceous shales also represent potential, though less prolific, source-rocks; these are the Spas shales in the Skole basin and the upper Cieszyn formation in the Silesian basin and Fore-Dukla unit. In the Dukla basin and equivalent units of the western Carpathians, the occurrence of Cretaceous source-rocks has not yet been demonstrated but is possible. Lastly, Carboniferous shales and coals may be considered as potential souce-rocks; however, such a statement has to be confirmed by additional investigations.

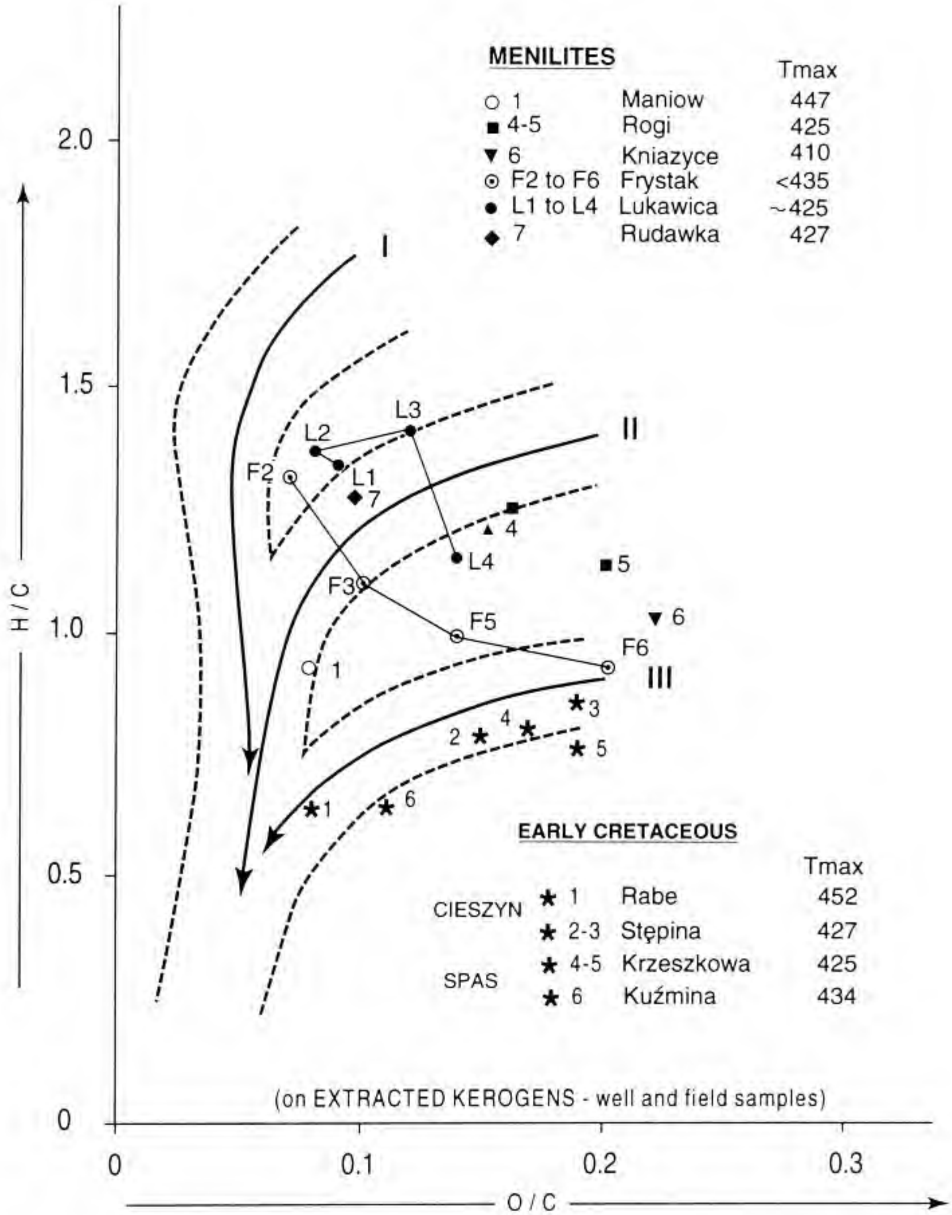

FIG. 10. Van Krevelen diagram for Early Cretaceous and Menilite samples.

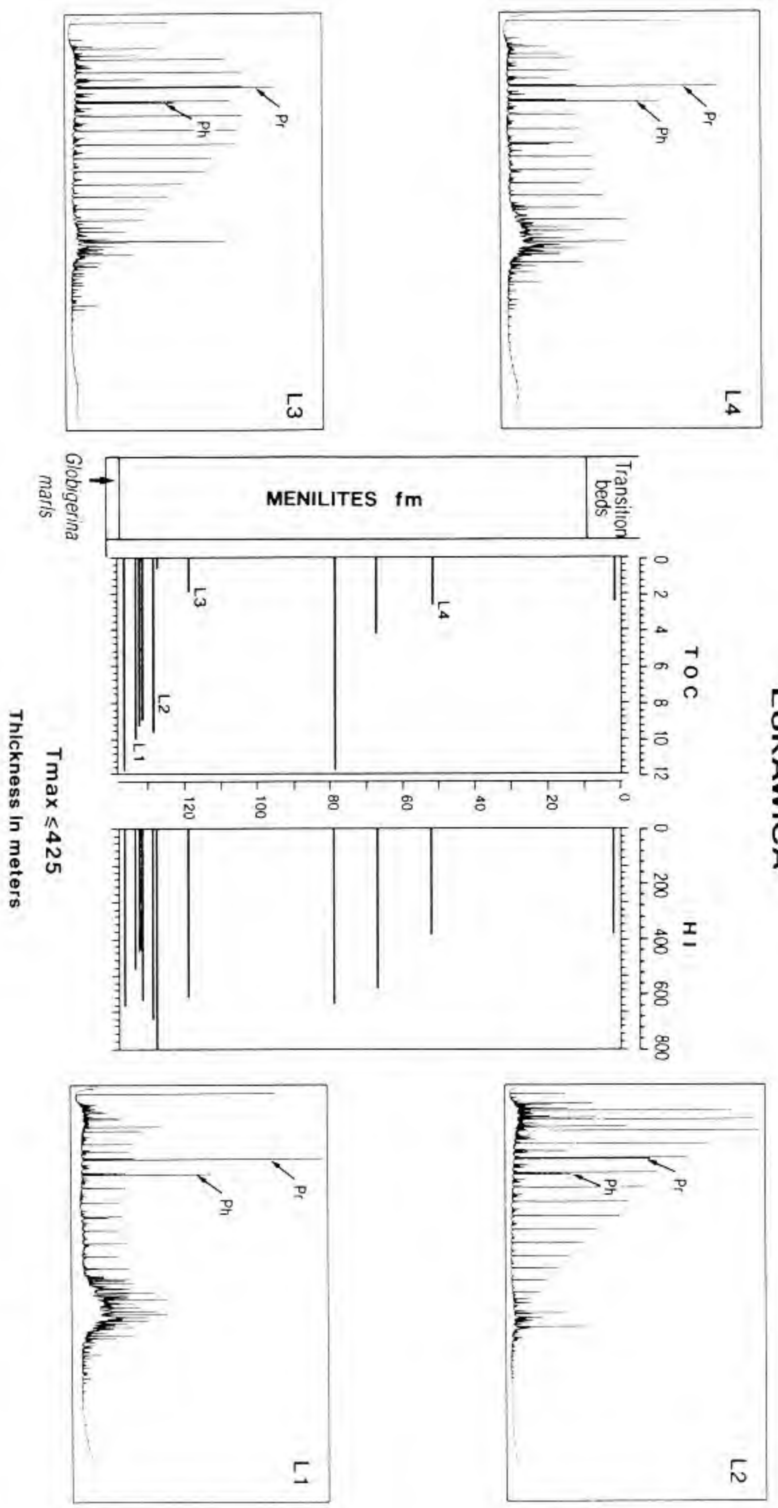

FIG. 11. Main geochemical results at Lukawica site.

PETROLEUM OCCURRENCES AND SOURCE-ROCK CORRELATIONS

Petroleum Occurrence

In the autochthon (Encl. 3), more than 70 gas fields produce from Miocene reservoirs. These fields are mostly located in the eastern, wider and deeper part of the Carpathian foredeep in the subsurface prolongation of Holy Cross Mountains, the so-called Upper San River High, where Neogene sediments directly rest on a Precambrian substratum. In the western Carpathian foredeep, a few gas fields also produce from Miocene reservoirs beneath the Outer Carpathian thrust. The gas contained in these accumulations has been demonstrated to be biogenic in origin (Kotarba, 1987, 1992). The occurrence of thermogenic gas is restricted to few Triassic, Late Jurassic and Late Cretaceous reservoirs (Kotarba and Jawor, 1993).

In addition, about 10 fields produce oil from Late Cretaceous sandstones and Late Jurassic carbonates. These are located west of the Upper San River High. Few oil accumulations have been discovered in Palaeozoic strata which are preserved in blocks adjacent to this structure on its southwestern (Early Carboniferous and Late Devonian reservoirs) and its northeastern flank (Ordovician-Silurian reservoirs).

In the allochthon (Encl. 3), more than 80 oil fields and a few gas fields, all of relatively small size, have been found in Cretaceous to Oligocene reservoirs. Most of them are located within the Silesian unit and produce mainly from (Cretaceous)-Paleocene Istebna sandstones and/or Eocene Ciezkowice sansdstones (Fig. 2). In the southeastern part of the Silesian unit, oil comes from Oligocene Krosno sandstones. In the Skole unit, only about ten fields, all but one located at its southern edge, produce oil from Oligocene Kliwa sandstones (Encl. 1). In addition, some accumulations are located in the Magura unit (Cretaceous-Paleocene Inoceramus beds), the Dukla and equivalents, and the Sub-Silesian unit (Cretaceous Weglowka marls and sands). All gas accumulations are sourced by thermogenic gas (Kotarba, 1987).

Geochemical Analyses of Oil And Source-Rock Extracts

From the 90 or so oil accumulations occurring in the different plays of the Carpathian foredeep and the Outer Carpathians, 65 representative oil samples were collected and analysed (Ten Haven et al., 1993). In addition, 12 source-rock samples were selected from the general geochemical screening and their extracts were analysed by Gas Chromatography-Mass Spectrometry (Ten Haven et al., 1993) to be compared to oil samples.

Geochemical analyses of oil samples from reservoirs of different ages permit the distinction of two oil families, the global composition of which (saturated hydrocarbons between 50% and 70% and less than 20% of NSOs) indicates an overall similarity (Ten Haven et al., 1993).

The first oil family, represented by only one sample (Nosowka; Encl. 3), has an isotopically light signature, abundant C29 steranes, and lacks, in contrast to the second family, characteristic biomarkers such as the oleanane.

The second oil family comprises all other oil samples from the foredeep as well the allochthon (Encl. 3), between which no significant differences could be observed (Ten Haven et al., 1993). This family has an isotopically heavier signature and is characterized by the presence of oleanane and 28,30-*dinor*-hopane, in some oils the highly branched C25 isoprenoïd alkane (HBI) and in one sample several additional biomarkers characteristic of terrigenous input.

The presence or absence of oleanane is of key importance for determining the origin of oils because this biomarker can be used both as a facies and as a stratigraphic marker. As it is an Angiosperm-derived marker, its occurrence implies a land-plant contribution to the respective organic matter. On the other hand, as Angiosperms appeared during the Early Cretaceous and proliferated during the Late Cretaceous and Tertiary (Moldowan et al., 1993), the presence of oleanane implies than the respective source-rocks must be Cretaceous or younger. However, its absence does not necessarily speak for a pre-Cretaceous source-rock because this can also be linked to an organic matter of purely marine origin. The origin of the two other main biomarkers is unknown: the 28,30-

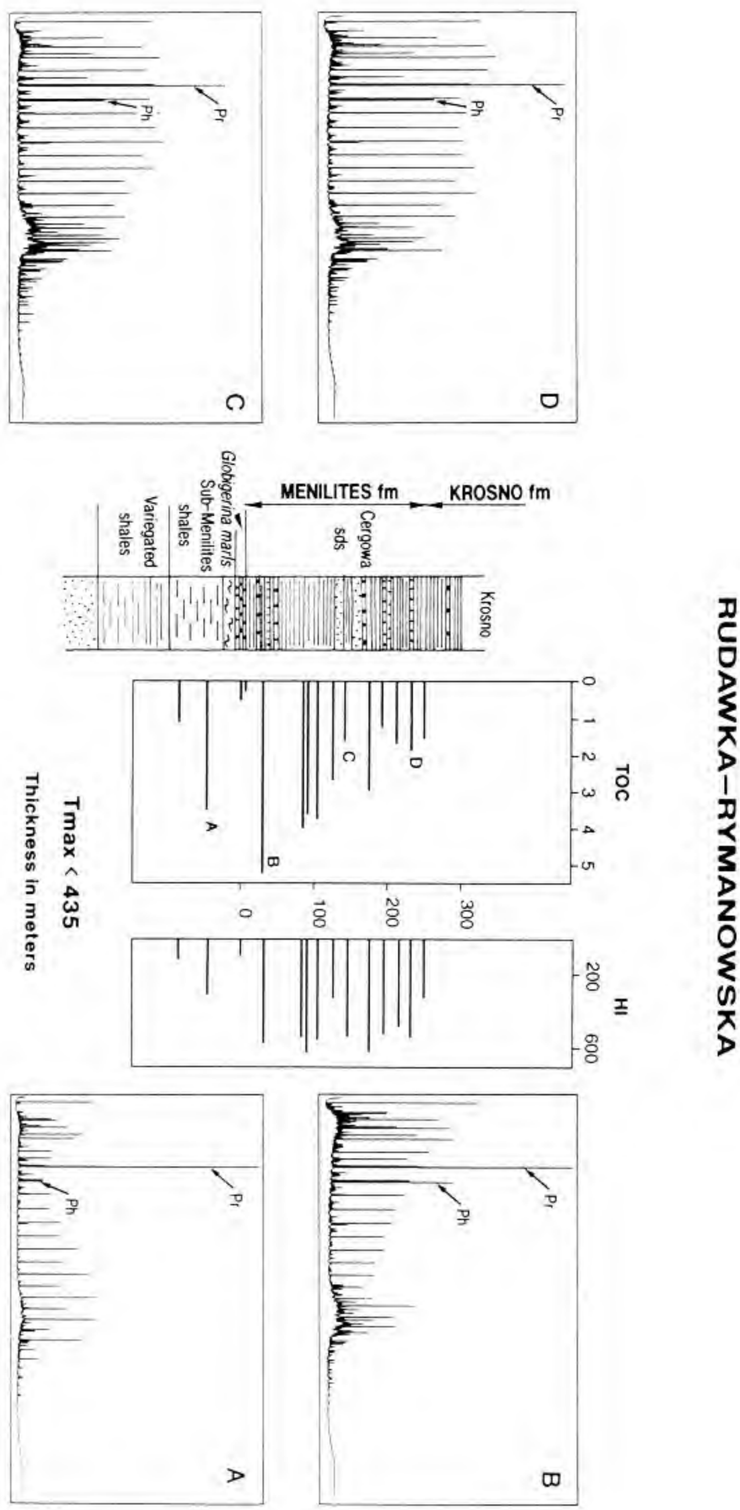

FIG. 12. Main geochemical results at Rudawka-Rymanowska site.

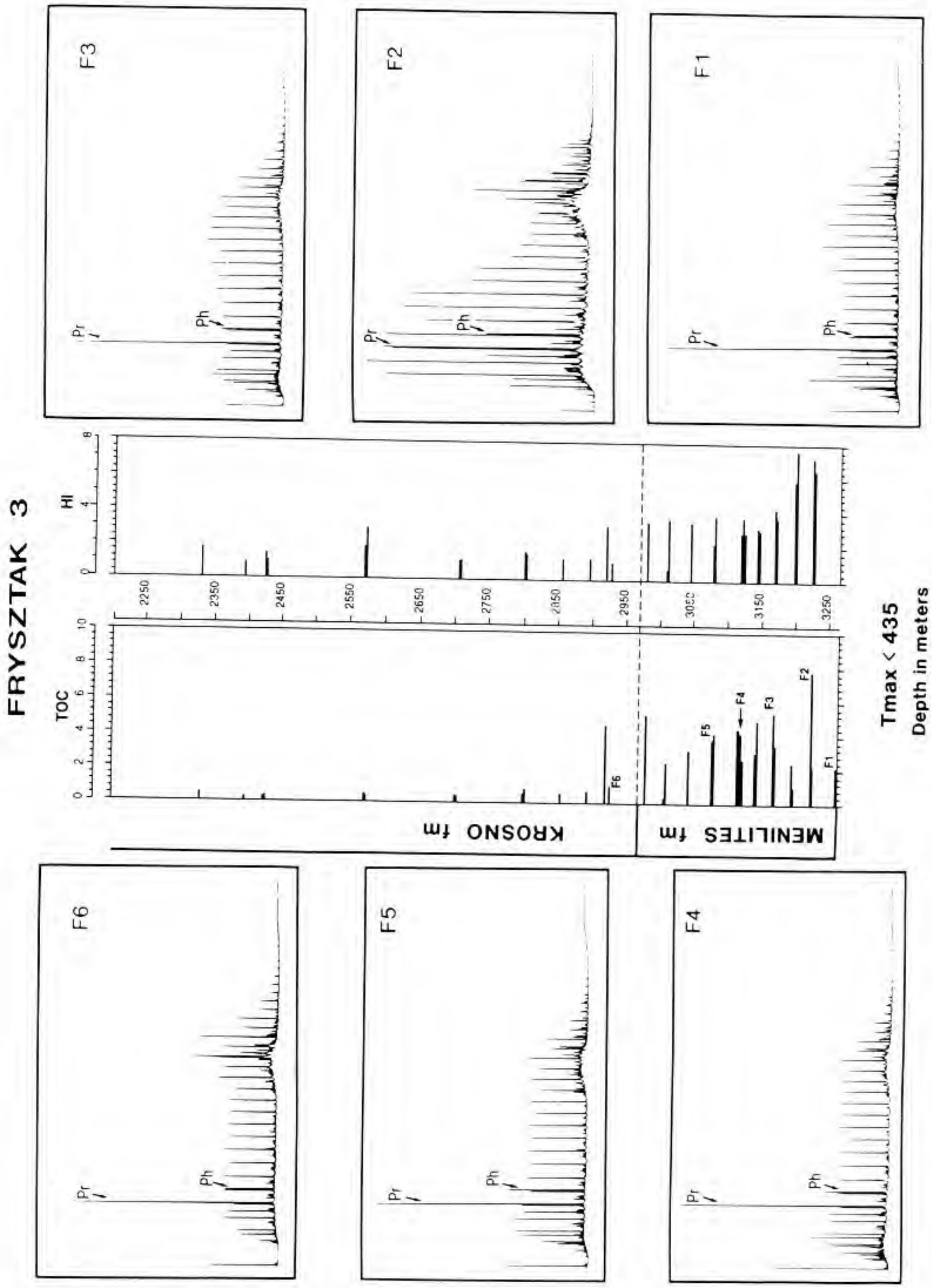

FIG. 13. Main geochemical results of Frysztak 3 well.

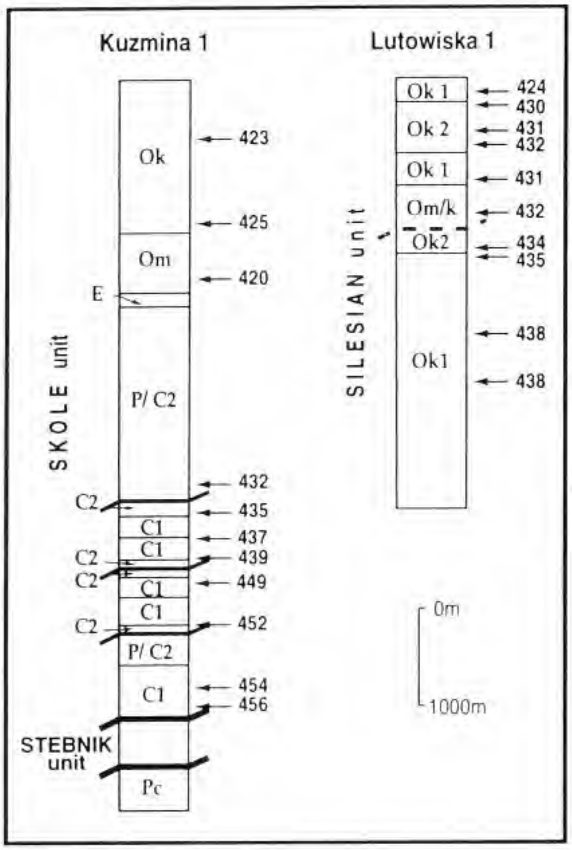

FIG. 14. Two examples of post-compressional maturation trends, based on Tmax from Rock Eval analyses.
Legend : 423 : Tmax. O : Oligocene; k : Krosno; k2 : middle Krosno; k1 : lower Krosno; m/k : intermediate beds; m : Menilites; E : Eocene; P : Paleocene; C2 : upper Cretaceous; C1 : lower Cretaceous; Pc : Precambrian.

dinor-hopane might be of bacterial origin, the HBI could be sourced by diatoms (Ten Haven et al., 1993). However, these markers are considered as useful tools for oil/oil and oil/source-rock correlations.

Based on sulphur contents and relative abundance of these biomarkers, the second family can be grouped into the following sub-sets (Encl. 3):

Group A is characterized by the dominance of higher-plant derived triterpanes and is represented by only one oil, located on the central northern margin of the Silesian unit.

Group B is characterized by abundant 28,30-*dinor*-hopane and comprises the oils reservoired in Kliwa sandstones of the Skole unit.

Group C lacks these specific characteristics and embraces all other oils found in reservoirs of the different allochthonous tectonic units.

Group D also lacks these specific characteristics but differs by a high sulfur content. It includes all samples from the foredeep, except Nosowka.

Hydrocarbon extracts from source-rocks were made on 3 Early Cretaceous samples (Cieszyn, Spas formations), 9 samples from Menilite shales covering the range of different compositions and organic contents, and 2 samples from Carboniferous series.

The analysis by GC-MS showed that all the Menilite samples contain oleanane, except in Lukawica L2; the absence of oleanane in this sample is very likely related to its pure marine origin, well documented by Rock-Eval, elemental and GC analyses (Fig. 11). The other biological markers (HBI, 28,30-*dinor*-hopane) were also found but with strong abundance variations. One sample (Kniazyce -see location on Fig. 7) contains numerous additional biomarkers of terrigenous origin. These variations within the Menilite shales confirm variations in terrestrial input ranging from a "pure" land-plant organic matter (Kniazyce) to a "pure" marine organic matter (L2) with all the intermediate compositions.

In contrast, Early Cretaceous samples contain neither oleanane nor any other of the above mentioned compounds. No other specific biomarker has been identified which could permit to unequivocally give an Early Cretaceous source for a crude-oil.

The Early Carboniferous black-shale and Late Carboniferous coal samples gave typical biological marker fingerprinting which significantly differ from each other.

Oil/Source-Rock Correlations

The characteristics of the first oil family (Nosowka oil) exclude a Late Cretaceous-Tertiary source-rock and suggest a **Paleozoic source-rock.** Moreover, they also appear atypical for oils derived from an Ordovician shale or a Devonian carbonate source-rock. The fingerprints of the Nosowka oil and of the Late Carboniferous coal sample exhibit certain similarity, suggesting a possible correlation. However, this conclusion remains speculative as it is based on a single sample only, and definitely requires additional investigations.

The presence of oleanane in all oils of the second family indicates that the **Menilite shales** are the source-rock for these oils. However, a correlation of the different groups with one particular lithofacies of the Menilites remains questionable as oils obviously have an average composition and are derived from a wide spectrum of organic facies recorded within this formation. The oil of group A might be tentatively correlated with Menilite shales having a content in terrestrial organic matter (Kniazyce and possibly Frysztak upper Menilite samples). The oils of groups B and of C, the more frequent ones, could be derived from an intermediate facies, characterized by a mixture of marine and terrestrial organic matter. Such an assumption is reasonable, considering the vertical and lateral variations in organic facies at a basin scale. The geographically restricted occurrence of group B oils is striking, but no satisfactory explanation can be given for it.

The group D oils from the foredeep have been clearly generated and expelled from Menilite shales. Although not supported by the maturity data, the high sulfur content of these oils suggests an earlier stage of generation and expulsion as compared to the other oils of this family (Ten Haven et al., 1993).

Mixing of oils derived from Menilite shales with oils generated from other source-rocks might be possible, even if it cannot be proven to-date. It is speculated that, within the Outer Carpathian overthrust, the Early Cretaceous source-rocks might have contributed to the presently pooled oils, particularly in the southeastern part of the Silesian basin (Cieszyn formation) and in the Skole basin (Spas formation); in Grobla field, a contribution from a Carboniferous source-rock cannot be excluded.

EVOLUTION OF PETROLEUM SYSTEMS

Present and Inherited States of Maturation

The present-day state of source-rock maturation has been investigated using Rock-Eval maturation index (Tmax) from outcrop and core samples. In particular, data from deep boreholes (Kuzmina 1-2, Paszowa 1) and/or wells penetrating complex tectonic structures (Frysztak 3, Lutowiska 1) provided useful insight. In this way, data of Lutowiska 1 or Kuzmina 1 wells (Fig. 14) clearly evidence that the present-day maturity trend is the result of a post-thrusting evolution: in Kuzmina 1, Tmax regularly increases from 435 to 455 within a succession of three stacked recumbent folds, involving normal and overturned early-Late Cretaceous series (Fig. 15).

Assuming this feature as representative for the entire Outer Carpathians, maturity levels have been drawn along the eastern and central cross sections (Fig. 15). Some conclusions can be reached by analysing these cross sections and by considering additional scattered maturity data in the western zone. However, these conclusions must be regarded only as a first approach of the present state of maturity in Outer Carpathian owing to the tectonic complexity of the area and the lack of sufficient maturity data.

The Menilite formation is immature in the whole Skole unit, except in Frysztak area where the Menilites are buried underneath the Silesian and the Sub-Silesian thrusts. In contrast, the Early Cretaceous Spas shales might be in the oil window in the southern part of this unit.

No maturity data is available in the Silesian unit of the western Outer Carpathians. Elsewhere, Menilite shales could be mature in the southeastern part of this unit and within some deep synclines to the northwest. Early Cretaceous source-rocks could be mature in most thrust sheets of this unit.

The Grybow, Fore-Dukla, Dukla and equivalent western units exhibit all a high level of maturity. The maturity stage in the Magura unit cannot be evaluated for lack of data.

The autochthonous Miocene source-rock is immature to slightly mature (T_{max}<440) over the sampled depth range, from 400 to 3300 m.

Problem Of Inherited States Of Maturation

The present-day maturity of the source-rocks is the result of a complex thermal history which involves a first pre-compression episode, during which sedimentary overburden provides for source-rock burial and maturation, and a second post-compression episode during which their burial is only caused by tectonic emplacement of thrust sheets. Actually, large parts of the Skole and possibly also the Silesian units did not undergo tectonic burial and their sedimentary sections were only eroded since thrust emplacement. In contrast, tectonic overburden was important in more internal units. Furthermore, this process was more important in the western parts of the Carpathians where the Magura nappe was largely thrusted onto both the Grybow and Obidowa-Slopnice nappes, than in the eastern parts where this nappe was probably thinner and its thrusting onto the Dukla unit more restricted. An another case of tectonic overburden is the Fore-Dukla triangle zone (cross section I, Fig. 16) where a high and reversed maturity trend has been observed in the outcropping section (Tmax decreasing in older formations). This feature is reliably accounted for by the Neogene emplacement of several duplexes invoked for this structure (Roure et al., 1993; Roca et al., 1995); therefore, the present-day maturity trend has been acquired lately. In these examples, the tectonic overburden severely overprinted the pre-compression source-rock maturity stage.

Actually, modelling is required to integrate all parameters controlling hydrocarbon generation in an effort to separate between pre- and post-thrusting maturation levels. However, some reasonable approximations can be made for areas which have been only eroded since thrust emplacement. Owing to erosion, source-rocks underwent decreasing temperatures, even assuming a constant heat flow. The period of time taken into consideration is short (about 15 Ma) and the impact of this factor on maturation has been demonstrated to be neglegible. Consequently, the source-rock the maturation level remained constant since thrust emplacement. Therefore, the present Tmax, as measured in outcrops, reflects the pre-thrusting maturation pattern.

This qualitative approach leads us to consider that in the Skole basin, neither the Early Cretaceous series nor the Menilite shales have entered the oil-window prior to thrust deformation. In the Silesian basin, Early Cretaceous strata were located within the oil-window and the Menilite shales might have entered it, at least in the deepest parts of the basin.

Possible Migration Trends

Hydrocarbon migration models for Outer Polish Carpathians must account for the following facts:

(1) Oils found in the autochthonous Mesozoic platform reservoirs were sourced from the Oligocene Menilite shales. This implies long distance migration from an oil kitchen located in the presently allochthonous units. Oil charge from the Skole basin, although located adjacent to the platform, can be excluded because the Menilite shales are immature in this basin. Therefore, hydrocarbons must come from more internal units.

(2) The Paleogene basin was dissected by cordilleras which were repeatedly reactivated, particularly during middle-late Oligocene times.

(3) In the autochthon, the Cretaceous and most Jurassic reservoirs are sealed by the shaly and evaporitic basal series of the Badenian Molasse. Therefore, effective oil entrapment could not occur before the middle Miocene.

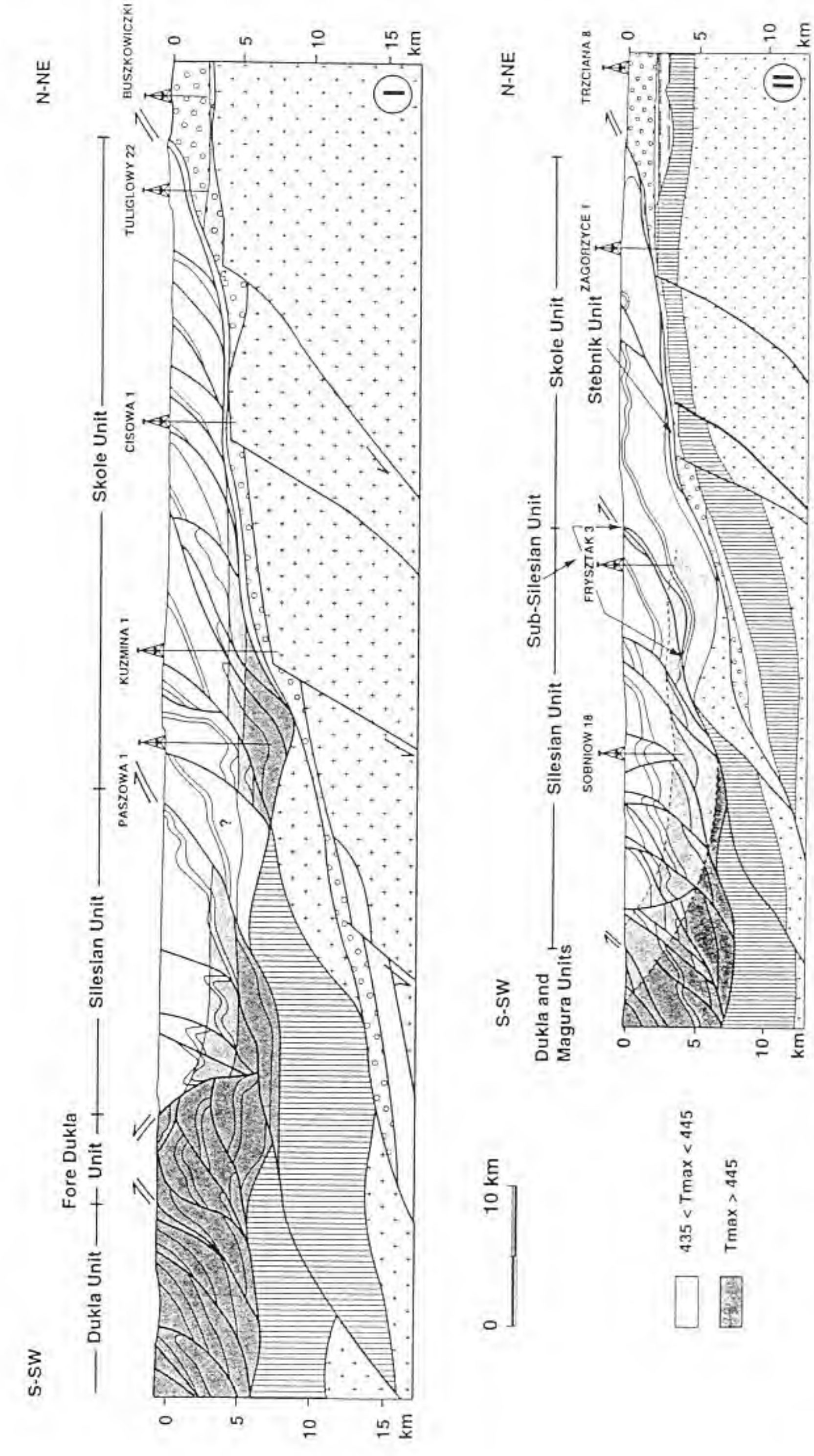

FIG. 15. Tentative reconstruction of the present-day states of maturity along the eastern (I) and central (II) cross-sections -with two maturation levels : Tmax = 435 i.e. beginning of the oil-window for type II and III, Tmax = 445 i.e. intra oil-window (Espitalié et al., 1985).

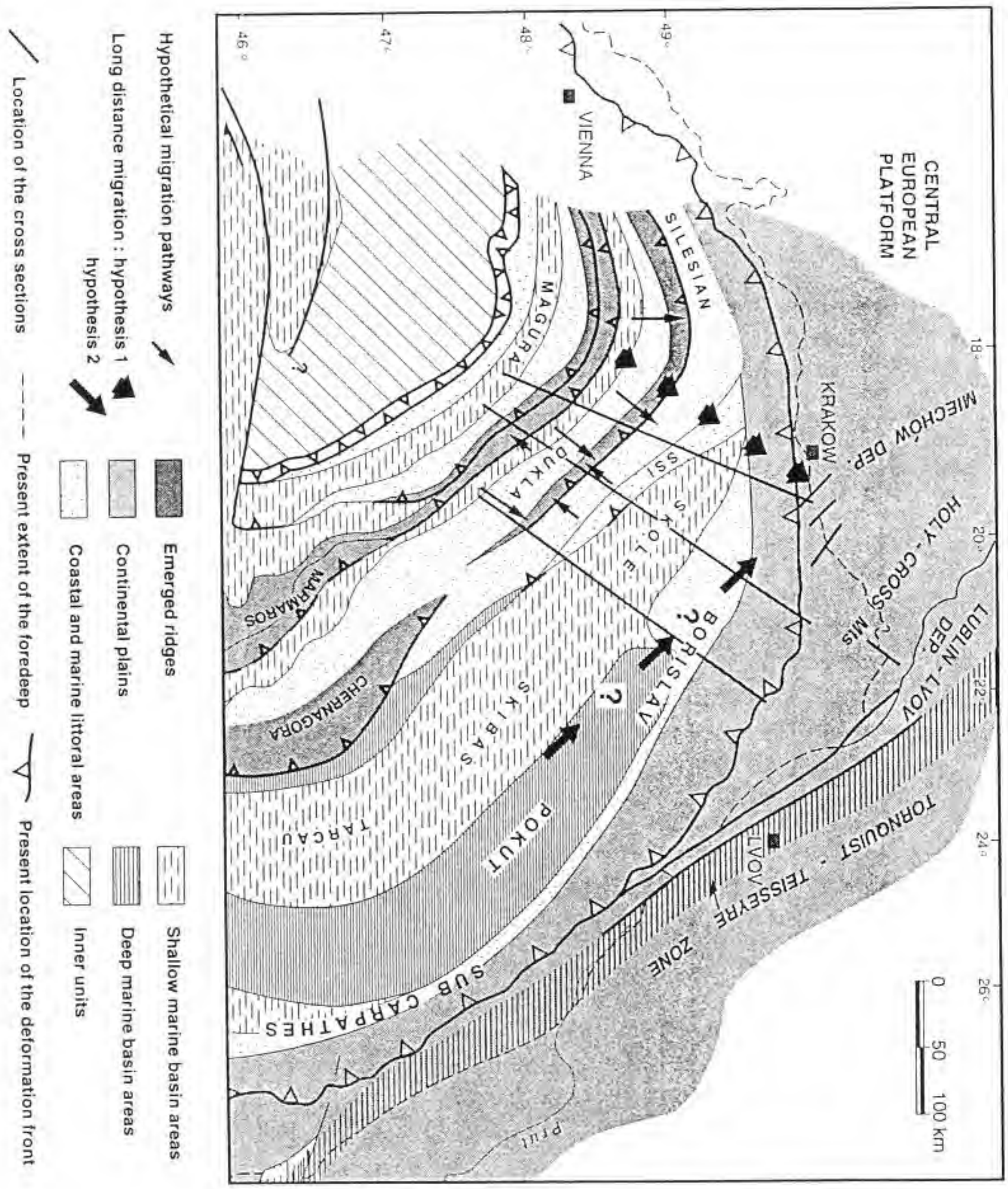

FIG. 16. Hypothetical migration trends before the main phase of thrusting (palinspastic restoration at late Oligocene - simplified from Ellouz and Roca, 1994).

Integration of these different constraints implies an accurate knowledge of the geometry and kinematics of the tectonic units versus the maturation state kinetics at the different stages of the structural history of the basin. Particularly questionable is the whether long distance migration, charging platform reservoirs, has indeed taken place. The presence of cordilleras prior to the main deformational phase and the generally complex geometry of the entire Carpathians, favour a short distance migration hypotheses. Actually, this question would require forward modelling to explore intermediate and pre-deformation geometries. Modelling would also be useful to precisely define the maturation state of the different source-rocks, to evaluate their respective contribution and to locate mature areas prior to thrust deformation. Thus, the scenario proposed below is only a preliminary attempt which will have to be confirmed or modified, once adequate modelling is available.

In this scenario, we assume an early generation of oil, prior to the Neogene deformation of the Carpathians, from Early Cretaceous source-rocks in the Silesian basin and possibly in Dukla basin, as well as from Menilite shales in the deep parts of the Dukla and Silesian basins. Oil probably started to migrate towards the palaeo-highs of internal units of the Outer Carpathians before their Neogene deformations (Fig. 16). However, volumes of oil were probably limited and restricted to the Fore-Dukla sub-unit which would have been charged from the adjacent Silesian and Dukla basins. Only Early Cretaceous oil might have migrated towards the Sub-Silesian unit. These early oil accumulations were subsequently restructured and destroyed by remigration and/or overmaturation.

At the end of the Oligocene, emplacement of the Magura nappe strongly increased maturity of the Menilite shales, particularly in the Obidowa-Slopnice unit in western areas, resulting in possibly important oil expulsion and migration. At this time, the Obidowa-Slopnice unit was not yet folded. Moreover, emplacement of the Magura nappe increased flexural subsidence of the whole area, thus favouring regional updip long distance migration towards the Mesozoic platform from the Obidowa-Slopnice unit (Fig. 16) (note that migration from the Dukla basin is not realistic as this basin was only slightly overridden by the Magura nappe and was separated from the platform by the wide Skole basin). However, the Mesozoic reservoirs in the foreland were at this time not yet sealed. A possible scenario is entrapment of oils in intermediate structures, the closure of which was reduced during the subsequent continued flexural subsidence of the basin, causing the release of oil towards the platform. However, another hypothesis could be a much younger structuration. Actually, this apparent inconsistency between an assumed late Oligocene age of deformation and oil expulsion, and a Badenian age for trapping, could result from errors in dating the onset of the deformation. As a matter of fact, reworked faunas are frequent in flysch sequences; therefore, their age can easily be overestimated. Assuming a younger age for the deformation of the entire Outer Carpathians (no older than Badenian, even for the Magura thrust emplacement) would provide a simple solution for both tectonics and petroleum migration. Whatever the hypothesis, this scenario is in agreement with the geochemical analyses which suggest a slightly earlier time of generation and expulsion for the oils of the foredeep than for the oils trapped in the allochthonous reservoirs.

During and after thrusting, short distance migration occurred in each separate tectonic unit, from its deeper parts towards the adjacent highs. This scenario appears to apply for the Dukla and equivalents and the Silesian basins. Remigration from the Silesian unit very likely accounts for accumulations in the Sub-Silesian unit. In contrast, accumulations within the Skole unit can only be explained by remigration from the Silesian unit and/or by a contribution from Cretaceous source-rocks contained in the southernmost part of the Sklole basin. This peculiar situation might account for the existence of the distinct group B oil in this basin. In contrast, no specific oil group can be ascribed to the other units; C-group oil is present in all units and no conclusion can be drawn from a single A-type sample, the location of which might be rather due to facies variation of the Menilite shales.

SUMMARY AND CONCLUSIONS

Extensive geochemical surveys carried out in the Polish Outer Carpathians confirm that the Oligocene Menilite shales are the primary source-rocks, that they are characterized by strong lateral and vertical heterogeneities and vary from a very good type II to a mixed type II+III. In addition, some Early Cretaceous shales appear to play the role of secondary potential source-rocks; however, their contribution to oils accumulated has not been clearly established.

The Menilite shales have been identified as the main source for all the oils trapped in the overthrust as well as most of the oils, except one, trapped in the foredeep. This correlation is supported mainly by the common presence of oleanane in these oils. Minor biomarker differences observed must be partly related to changes in organic facies of the Menilite shales.

There are indications for two distinct episodes of oil generation : 1) an early generation phase occurred prior to and at the very beginning of compressional deformation due to deep sedimentary burial of the Menilite shales in Dukla and Silesian basins; 2) a late generation phase occurred after thrust emplacement of the allochthounous units due to tectonic burial of the source-rocks.

Available data suggest that most of the oil accumulations in the allochthon may be charged by oils generated during the post-thrusting stage. On the other hand, the source area and the timing of the oil charge of traps occurring in the foredeep remain debatable. The preferred working-hypothesis envisages a combination of early long distance migration from the northwestern internal Outer Carpathian units, oil accumulating in intermediate structures and their subsequent destruction during the main orogenic phases, causing updip migration of oils into the foreland. Detailed basin modelling is required to determine whether this hypothesis is valid. An alternate hypothesis invokes lateral long distance migration from the most internal parts of the Borislav-Pokut basin (Fig. 16). Such migration could be facilitated by sandstone intercalations in the Menilite shales along the northeastern margin of this basin. This would imply a marked northwestward component in the flexural subsidence of the incipient Carpathian foreland basin for which we have, at this time, no evidence.

For the Ukraine, relatively short distance migration can be invoked for the oils trapped in Mesozoic reservoirs of Lopushnia field, located in the autochthon just in front of the Borislav-Pokut nappe (see Izotova and Popadyuk, Sovchik and Vul, this volume). A Menilite origin has been clearly established for this oil (Lafargue et al., 1994; Ellouz et al, in preparation). Reconstruction of pre-thrusting geometries permits to consider migration along the flexural trend from the more internal parts of the Borislav-Pokut and possibly the Skiba basins where Menilite shales were already mature. Comparison of the geological settings of the Ukrainian and Polish Carpathians at the end of the Oligocene, shows major differences between the two areas (Fig. 16). In Ukraine, oil kitchens are located rather close to the foreland and are not separated from the latter by cordilleras. Therefore, the scenario developed for the Ukraine cannot be directly applied to Poland. Thus, the petroleum system of the Polish Outer Carpathians may be unique in the entire Carpathians fold-and-thrust belt.

Several points require further clarification:

(1) Have Early Cretaceous potential source-rocks indeed and substantially contributed to the hydrocarbon habitat of the Carpathians as suggested here for the Skole unit?

(2) Do Carboniferous coals and/or shales represent an additional source for oils occurring in autochthonous reservoirs?

(3) Do Late Jurassic basinal shales and carbonates of the autochthon represent an alternate source, as evident in the Vienna basin of Austria (Ladwein et al., 1991)? In the Polish and Ukrainian Carpathian foreland, Late Jurassic neritic carbonates grade laterally into pelagic carbonates and shales (Izotova and Popadyuk, this volume). If these series has source-rocks characteristics and if it is preserved beneath the Skole unit of Poland, it may have contributed to its hydrocarbon habitat. However, the present deep burial of Jurassic sediments (Encl. 2) suggests that they are post-

mature for oil generation and thus, may yield gas only.

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Enclosures

Encl. 1 Tectonic map of the Polish Outer Carpathians (after Zytko et al., 1989; Sokolowski, 1954; Widzinski, 1950-1954), showing distribution of oil and gas fields (modified after Petroconsultant, 1994) and location of the cross-sections given in Encl. 2.

Encl. 2 Geological cross-sections through the Polish Carpathians (for location see Encl. 1) Section I and II: modified after Roure et al. (1994), section III: modified after Roca et al. (1995)

Encl. 3 Stratigraphy and structural location of oil and gas reservoirs in the autochthon and allochthon. For explanations see text. Stratigraphy of the autochthonous reservoirs: N1-Miocene, Cr2-Late Cretaceous, J3-Late Jurassic, J2-Early Jurassic, T-Early Triassic, C1-Early Carboniferous, D-Devonian, S-Silurian, O-Ordovician, Cm-Cambrian.
Stratigraphy of the allochthonous reservoirs: Ok-Oligocene Krosno beds, Okl-Oligocvene Kliwa sds, Ocr-Oligocene Cergowa sds, Omg-Oligocene Magdalena sds, Eh-Eocene Hieroglyphic beds, Ec-Eocene Ciezkowice sds, Pis-Paleocene Istebna beds, Pin-Paleocene Inoceramus sds, Cr-Pin-Paleocene-Cretaceous Istebna sds, Cwm-Late Cretaceous Weglowka marls, Cws-Early Cretaceous Weglowka sds.

Source : MNHN, Paris

Oil and gas accumulations in the Late Jurassic reefal complex of the West Ukrainian Carpathian foredeep

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ABSTRACT

The Cretaceous and Cenozoic Ukrainian Carpathian foredeep basin is underlain by an extensive Late Jurassic, reef fringed carbonate platform. The latter forms part of the extensive system of carbonate platforms which developed during Late Jurassic times on the northern shelves of the Tethys Ocean.

Late Jurassic reefal and back-reef carbonates form the principal reservoirs of 4 hydrocarbon accumulations containing ultimate recoverable reserves of some $37 \cdot 10^6$ bbls of oil and condensate and 1.3 BCF of gas. These fields are contained in two trap types, including erosional highs and roll-over structures related to a major Paleogene erosional phase and subsequent Neogene subsidence of the Carpathian foredeep. Jurassic carbonates and Cretaceous sandstones are the principal objectives in the sub-thrust play of the outer Carpathian nappes. Although the potential of this play has not yet been exhausted, reservoir prediction and reflection-seismic definition of prospects entail considerable risks. Hydrocarbon supply is not considered to be a mayor risk factor in this sub-thrust play.

INTRODUCTION

In the foreland of the Ukrainian Carpathians and beneath their external nappes two oil fields (Kokhanovka and Lopushnya), one gas/condensate field (Letnya) and one gas field (Rudky) were discovered in Late Jurassic carbonates. These fields include additional pay sections in Cretaceous and Miocene sandstones. These fields are under development and contain cumulative ultimate recoverable reserves of $37 \cdot 10^6$ bbls of oil and condensate and 1.3 BCF of gas (Fig. 1). Two blocks containing Jurassic carbonate and Cretaceous sandstone prospects in a sub-thrust position are currently under exploration.

In this paper we address the evolution of the West-Ukrainian Late Jurassic carbonate platform and the habitat of hydrocarbon accumulations asso-

Izotova, T. S. & Popadyuk, I. V., 1996. — Oil and gas accumulations in the Late Jurassic reefal complex of the West Ukrainian Carpathian foredeep. *In*: Ziegler, P. A. & Horváth, F. (eds), Peri-Tethys Memoir 2: Structure and Prospects of Alpine Basins and Forelands. *Mém. Mus. natn. Hist. nat.*, **170**: 375-390. Paris ISBN: 2-85653-507-0.

ciated with it. Although the potential of this Late Jurassic carbonate play has not yet been exhausted, reservoir and seal prediction and reflection-seismic definition of prospects entails considerable risks, particularly in the Carpathian sub-thrust play.

During the Late Jurassic the northern margin of Meso-Tethys was occupied by large reef-bearing carbonate platforms; these extended from the Jura Mountains of France and Switzerland through southern Germany and Poland into the pre-Carpathian domain of the Ukraine and Romania and eastwards via the southern slopes of the Caucasus into Central Asia.

In the western Ukraine, Late Jurassic carbonates occur in the Carpathian foredeep basin and overstep eastwards the Volyn-Podolian margin of the Precambrian East-European Platform (Fig. 1). These carbonates rest on Palaeozoic sediments except in the northwest where they are underlain by a Middle Jurassic siliciclastic series. In turn, the Late Jurassic carbonates are overlain by Cretaceous sediments which, in some areas, are deeply truncated by a Paleogene unconformity. This erosional phase, which is related to the inversion of the Polish Trough and the uplift of the Malopolska Massif of Poland (Ziegler, 1990), resulted in the complete removal of Cretaceous deposits in the northwestern part of the Carpathian foredeep where Jurassic strata are unconformably overlain by Miocene siliciclastic sediments. In the central parts of the Ukrainian foredeep, the so-called Kolomiya palaeo-valley, Mesozoic and Palaeozoic sediments were completely eroded. Elsewhere partially truncated Late Cretaceous sediments are preserved beneath the Paleogene erosional surface (see Sovchik and Vul, this volume).

Late Jurassic carbonates occupy an up to 150 km wide belt in which they attain thicknesses of the order of 500 to 1000 m; a gradual increases in thickness towards the South is evident. In the southern parts of the foredeep, the Late Jurassic carbonates form part of the autochthonous sequence which extends a considerable distance beneath the external nappes of the Ukrainian Carpathian (Fig. 2). Late Jurassic carbonates are located at a depth of about 1 km in the Volyn-Podolia area and at depths of 7-8 km beneath the Carpathian nappes.

STRATIGRAPHY AND FACIES DEVELOPMENT

The stratigraphic framework of the West-Ukrainian Late Jurassic strata was developed by A. Alth (1881), V.I. Slavin (1958), V.Ya. Dobrynina (1961), Ya.M. Sandler (1962), and V.N. Utrobin (1962) The presence of Oxfordian, Kimmeridgian and Tithonian biozones was established by V.G. Dulub (1963, 1964) and later summarized in the stratigraphic scheme of the Jurassic (Dulub et al., 1986). Fig. 3 provides a summary of the lateral facies and thickness changes of litho- and chronostratigraphic units along a selected profile through the West-Ukrainian Late Jurassic Basin.

Oxfordian strata comprise the Rudky and Sokal formations which are lateral equivalents. The Rudky formation consists of oolithic, pelitomorphic and sometimes biohermal limestones which attain thicknesses of up to 150 m; to the southwest these reefal carbonates give way to shaly fore-reef carbonates. The essentially lagoonal Sokal formation is developed in the north-eastern parts of the basin and consists of gray siltstones, shales bearing plant imprints and sandy limestones. Late Oxfordian deposits consist of 10 to 30 m thick multicoloured shaly limestone, containing towards the eastern basin margin intercalations of conglomerates, sandstones and anhydrites.

During **Kimmeridgian and Tithonian** times, progressive subsidence of this shelf was accompanied by the development of a coherent barrier reef, corresponding to the Oparia formation, which consists of gray to light coloured and mottled limestones attaining thicknesses of up to 1000 m. Reef building organisms include corals, sponges, algae, stromatoporides and bryozoans. However, limited core material does not permit detailed facies reconstructions within this reef complex which is basinward offset by thin, deeper water shales and limestones. **Kimmeridgian** back-reef strata correspond to the Rava-Russka formation which consists of a sequence of lagoonal dolomites, dolomitic limestones and anhydrites, ranging in thickness between 20 and 250 m. **Tithonian** back-reef strata are represented by the Nizhnev formation which is composed of light gray and cream

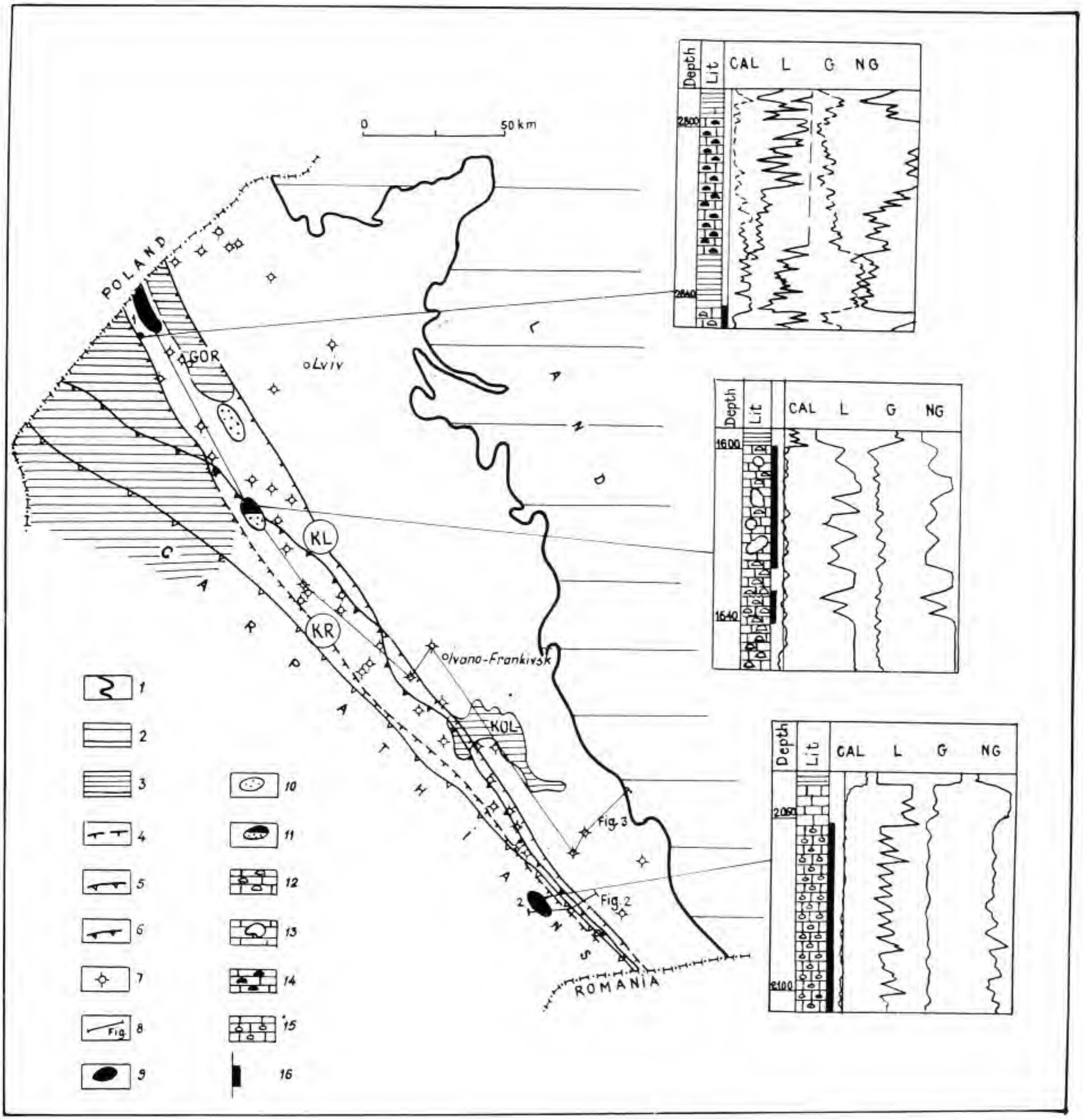

FIG. 1. Regional setting of Late Jurassic basin of Western Ukraine and representative log columns. 1-boundary of basin, 2-area lacking Late Jurassic sediments, 3-area of eroded Late Jurassic sediments (GOR: Gorodok valley, KOL: Kolomiya valley), 4-main normal faults (KL: Kalush-Gorodok, KK: Krakovets-Precarpathian), 5-front of Sambor nappe, 6-front of Boryslav-Pokutian nappe, 7-main wells, 8-trace of cross-sections, 9-oil fields (1: Kokhanovka, 2: Lopushnya), 10-Rudky gas field, 11-Letnya gas/condensate field, 12-boundstone, 13-karstified boundstone, 14-interbedded pelitomorphic limestones , bioclastic, wackestone and packstone, 15-bioclastic limestone, friable wackestone and grainstone, 16-porous intervals.

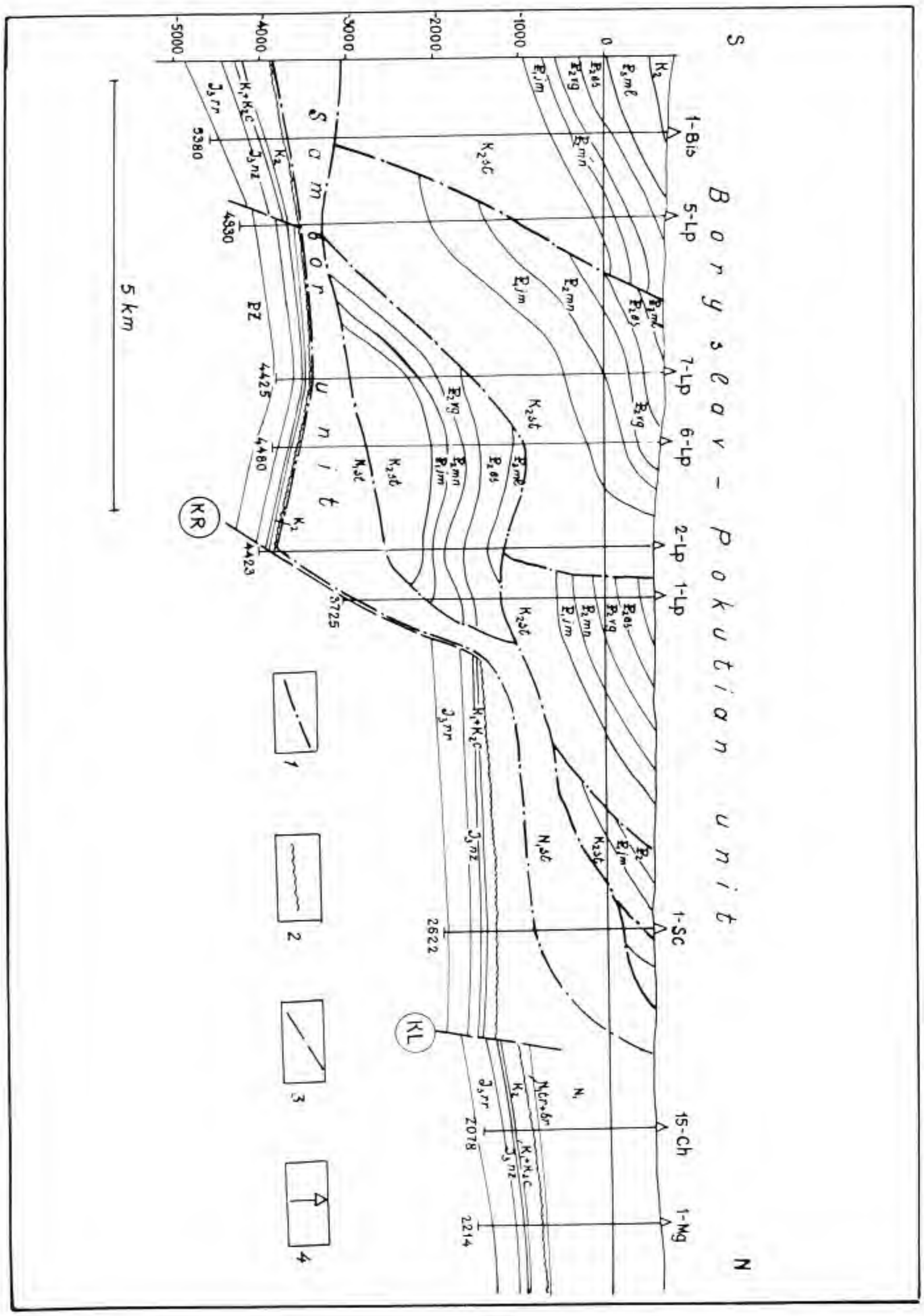

FIG. 2. Geological cross-section through southeastern Carpathian foredeep (for location see Fig. 1). 1-thrust faults, 2-top-Mesozoic erosional surface, 3-normal faults (KL: Kalush-Gorodok, KK: Krakovets-Precarpathian), 4-wells (1-Bis: Biskiv, LP: Lopushnya, 1-SC: Solonec, 15-Ch: Kovalivka-Chereshenka, 1-Mg: Migivska)

coloured detrital and algal limestones attaining a thickness of 200-250 m.

The Jurassic carbonates are conformably overlain by Early Cretaceous shales and limestones. Sedimentation continued through Late Cretaceous times, but was interrupted during the Paleocene and resumed only during the Neogene During the Palaeogene erosional phase, and in conjunction with the Neogene subsidence of the Carpathian foredeep basin, the geometry of the Late Jurassic carbonate platform was modified to such a degree that resolution of its internal configuration by reflection-seismic data meets with considerable difficulties.

METHODS OF INVESTIGATION

For the sub-surface reconstruction of the Late Jurassic carbonate platform the technique of sedimentological analysis of wire-line log data was applied (SALD). This technique relies on the fact that the log response of a rock unit reflects its mineralogical-petrographical composition and texture. As such, also the structural relationship between the different associated rock units could be determined. Based on a quantitative analysis of a suite of petrophysical logs, calibrated by limited core data, a sequence of lithofacies types was identified and their lateral and vertical relationship established. The following types of wire-line logs, which are generally available for exploration wells, were used: micro-lateral, induction, gamma-ray, neutron-gamma-ray, acoustic and caliper (Izotova and Push, 1986; Izotova et al., 1993). Applying the SALD technique, each well was analyzed in an effort to define the different lithofacies types and to establish their stratal succession and their boundaries. Subsequently log correlations between wells were used to develop facies models. The biostratigraphic framework for these sedimentological models.was erected on the basis of palaeontological data obtained from cores.

The SALD technique permits to extrapolate facies interpretations from limited core-controlled data points across the entire basin and thus allows to obtain a better impression of lateral facies and potential reservoir developments. In this respect, the quantitative analysis of wire-line logs in terms of clay content of carbonates, as well as their texture and porosity, are of particular importance. In in order to be able to readily compare the log response of the different lithofacies types, readings were plotted in so-called 8-ray diagrams (Fig. 4). These diagrams give a quantitative range of the response in the usual FSU logging units: gamma-ray (G) in gamma-ray units, laterolog (LAT) and micro-laterolog (ML) in ohmm, sonic (AL) in msec/m. Neutron-gamma-ray (NG) is calibrated in relative units which show the ratio of neutron-gramma activity of the respective strata. Caliper units (CAL) are shown as the ratio between the borehole diameter and the diameter of the drillbit. Porosity (Kp) is presented in % and characterizes the texture of carbonates. The ration of sequence anisotropy (Tk) quantitatively expresses the maximum and minimum deviations of resistivity responses from an average value; conventionally four groups are recognized, ranging from isotropic ($Tk \approx 1.0$) to anisotropic ($Tk < 0.25$). All wireline log parameters utilized for the Tk determination were average-weighted to the sequence thickness. In Fig. 4 the range of these responses are shown in 8-ray diagrams.

Based on an integration of macro- and microscopic core analyses and wire-line data, and following the fundamental work of J.L. Wilson (1980), the Late Jurassic West-Ukraine carbonate shelf was subdivided into nine genetically related lithofacies belts, as summarized in Fig. 4. For each of these facies belts an example of the standard log expression and an 8-ray diagram are given in Figs. 1 and 4. Comparing these 8-ray diagrams, it is obvious that each lithofacies is characterized by its own log response and by the degree of differentiation of the respective logs.

For instance, carbonates which were deposited below the storm wave-base in the fore-reef domain (belts 1 to 3) are characterized by high average gamma-ray readings, high velocities, comparatively low resistivity and an average differentiation of the laterolog and neutron-gamma-ray curves. This is a function of interbedding of carbonates, marls and shales. As reservoir rocks are absent in these facies belts, they are of little interest for oil and gas exploration.

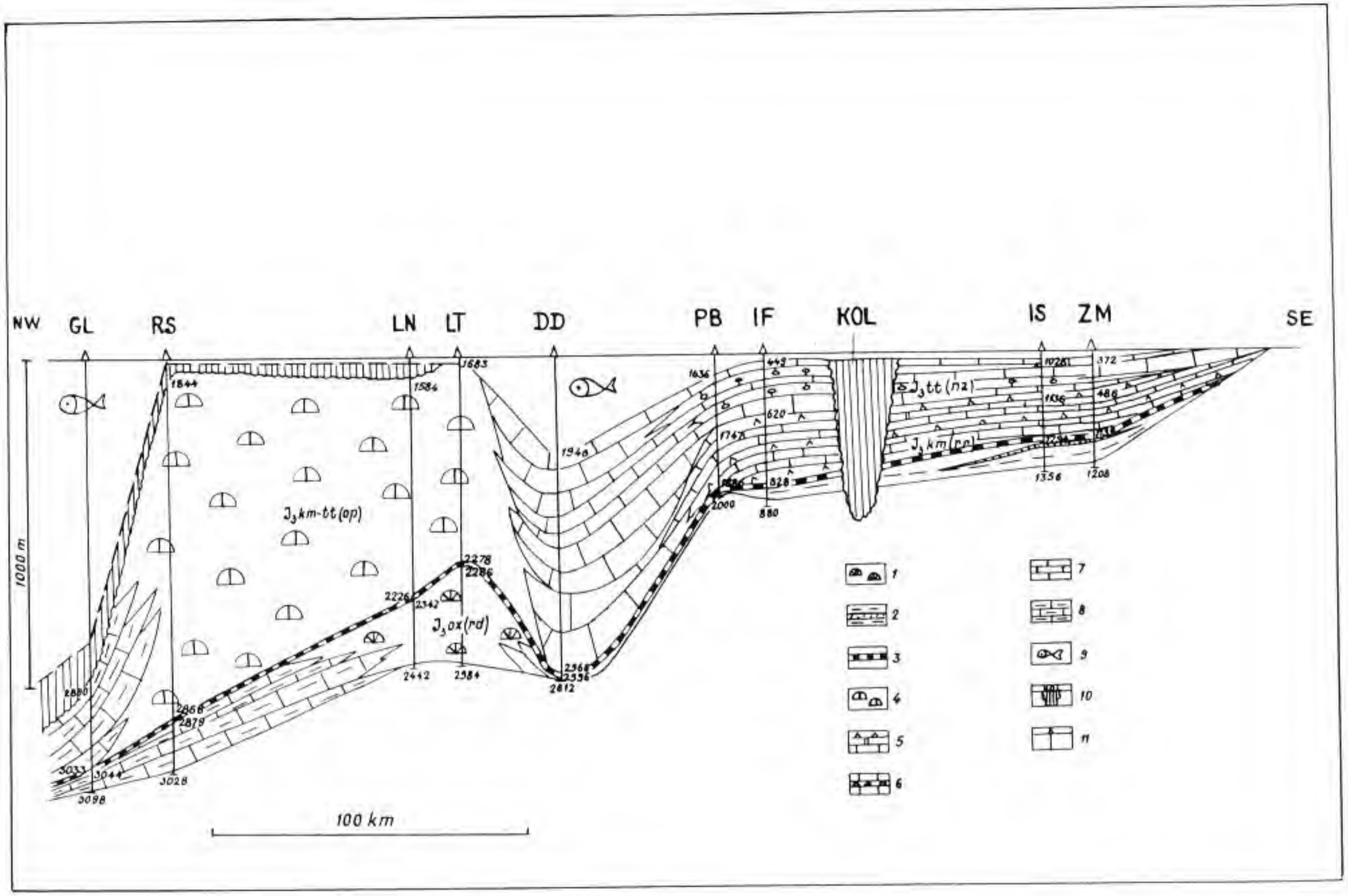

FIG. 3. Reconstruction of Late Jurassic carbonate platform. 1-Oxfordian Rudky reef, 2-Oxfordian shelf facies (Sokal fm.), 3-multicoloured horizon, 4-Kimmeridgian-Tithonian Oparian reef, 5-Kimmeridgian inner shelf and lagoonal facies (Rava-Russka fm.), 6-Tithonian shelf (Nizhnev fm.), 7-back-reef depression facies, 8-fore-reef facies, 9-palaeo-water-depth, 10-erosion during Paleogene (KOL: Kolomiya valley), 11-wells (GL: Gluvin-1, RS: Rosivska-1, LN: Lanivska-1; LT: Letnys-1, DD: Dedushychyn-1, PB: Pivnichni-Bogorodchany-1, IF: Ivano-Frankivsk-1, IS-Ispas-1, ZM: Zamosc-2).

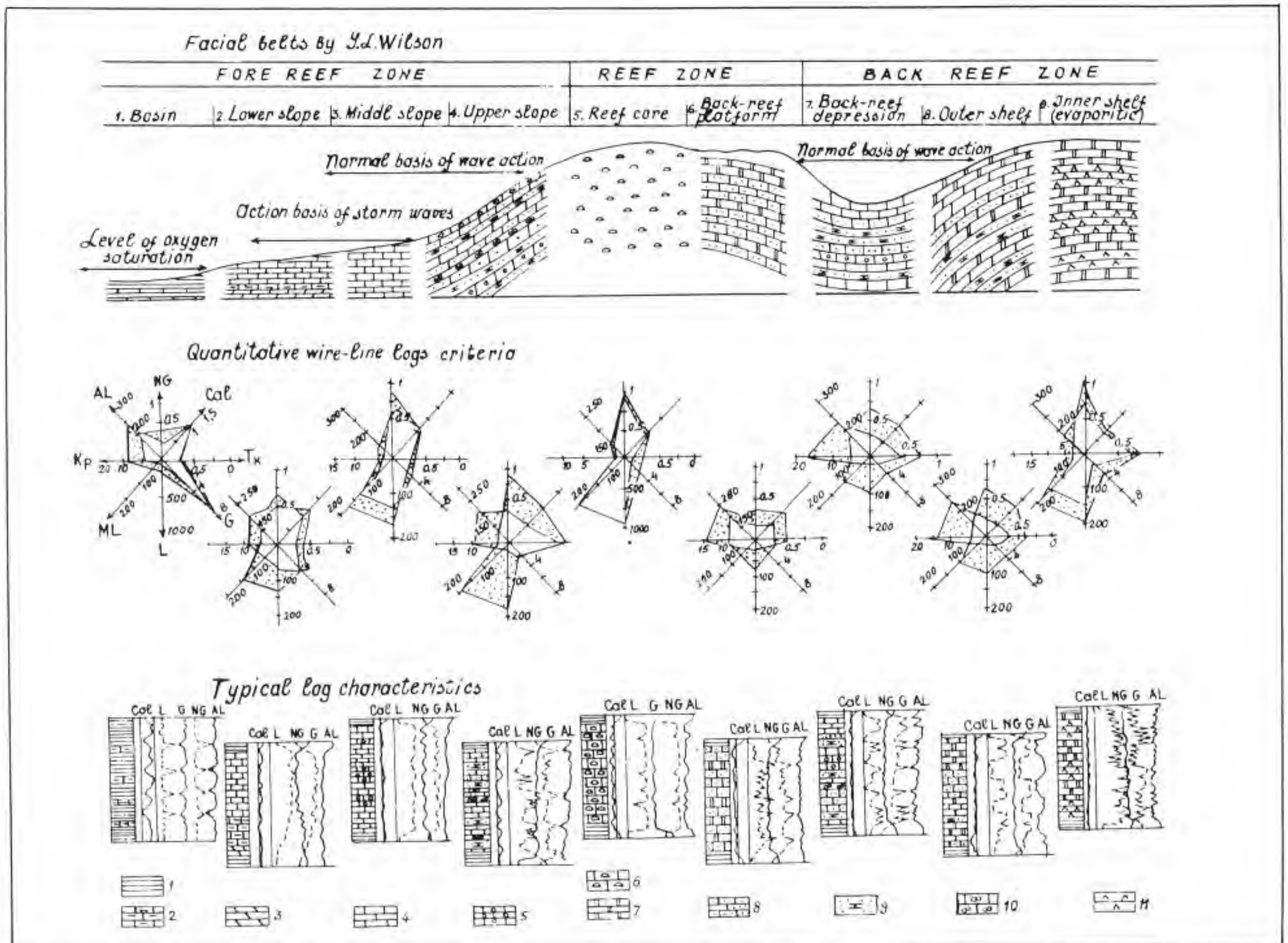

FIG. 4. Quantitative criteria for wire-line facies recognition and representative lithological columns (for explanations see text). 1-shales, 2-shaly limestone, 3-marl, 4-micritic limestone, 5-bioclastic limestone, 6-boundstone, 7-dolomite, 8-grainstone, 9-lime mudstone, 10 limestone breccia, 11-anhydrite.

Reefal carbonates (belt 5), which were deposited under warm, normal salinity, clear water conditions above normal wave base in response to a high bioproductivity, have an extremely low clay content; their gamma-ray response is generally low (not more than 2.5 gammas) and shows little variation. Although the initial texture of these reefal carbonates was partly obliterated by re-crystallization, outlines of framework builders are generally still preserved; therefore, the depositional environment of these carbonates can be determined (Reading, 1990; Wilson, 1980). Carbonates corresponding to the reef core are characterized by the most homogeneous texture and log expressions. Reefal limestones are generally characterized by a high resistivity (1000 Ohmm and greater), high secondary gamma-ray activity and low interval velocities (150 msec/m); natural radioactivity does not exceed 2.5 gammas. All wireline curves are weakly differentiated. A typical log of a reef core, that was not affected by karstification, is given in Fig. 5 for the well Mostovska-2. This section is practically isotropic. These limestones have resistivities of 800 Ohmm, velocities of 150 msec/m and a natural radioactivity of less than 2.5 gammas. The homogeneity of these rocks is interrupted by a porous interval at the depth of 1960-1966 m, possibly corresponding to a grainstone intercalation.

The reef-foreslope (belt 4) and the back-reef shelf (belts 6 to 8) consist of parasequences which are characterized by a variety of limestone facies and textures. Textural variations are reflected by strong variations of Tk. A typical example is provided by the well Lopushnya-4, which is located in facies belt 8. This well penetrated detrital limestones, deposited in normal marine waters, which are characterized by textures ranging from coarse to fine grained. All logs, except the gamma-ray curve, are highly serrated, indicating the presence of several porous intervals in carbonates having a low shale content (Fig. 6). Indeed, the back-reef zone is where the best reservoir developments have been observed with individual reservoirs having porosities in the 5 to 30% range. These reservoirs host the main oil and gas discoveries.

The evaporitic platform (belt 9), consisting of interlayered anhydrites and intertidal dolomites and limestones, has its own characteristic log response (see Fig. 4). The presence anhydrite layers and a high content of shaly limestone downgrades the reservoir potential of this facies belt.

PALAEO-RECONSTRUCTION AND REGIONAL FACIES MODEL

Based on the analysis and correlation of about 300 wells, the evolution of the Late Jurassic carbonate shelf of the West-Ukrainian foreland basin was reconstructed and its hydrocarbon potential further assessed.

In **Oxfordian** times, the Tethys Sea transgressed over the area now occupied by the Carpathian foredeep and advanced across the Volyn-Podolian margin of the Precambrian East-European Craton. During the initial development phase of the West-Ukrainian carbonate shelf, Early Oxfordian shallow marine strata overstepped in the northwestern part of the Ukraine an Early and Middle Jurassic deltaic sequence and gradually transgressed over the margin of the East-European Craton. North of the present day Krakovets fault, a hydroid-coral reef developed, attaining a thickness of 100 m; in back-reef areas, detrital and oolitic limestones, grading shore-wards into the sandy carbonates of the Sokal formation, were deposited. During the Late Oxfordian, reef growth was interrupted, probably in response to a rapid rise in relative sea-level, inducing the accumulation of widespread, 10 to 30 m thick shaly limestones which contain multicoloured horizons and cover the earlier reef complex.

During the **Kimmeridgian**, development of the Oparian barrier reef commenced, slightly basin-ward from the Oxfordian Rudky reefs. The pre-reef parasequences, corresponding to J.L. Wilson's facies belts 3 and 4 (Fig. 4), were only encountered in three wells drilled in the northwestern parts of the basin where they consist of bedded lime-mudstones and litho- and bioclastic calcarenites. The reef core is developed along the Krakovets fault up to where it is crossed by the outer Carpathian nappes. South of this point, the Oparian reef has not been reached by wells; however, its southward continuation beneath the

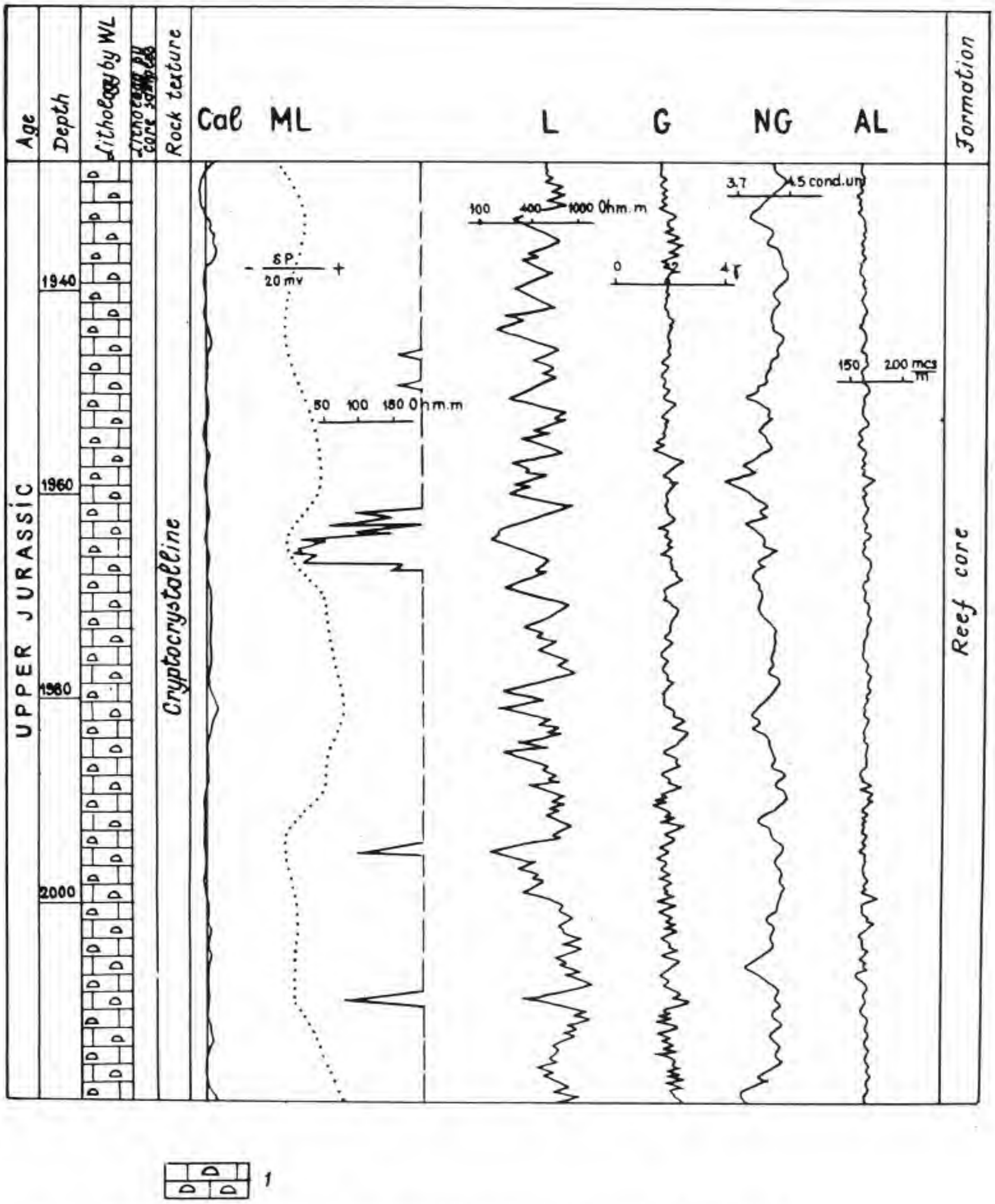

FIG. 5. Wire-line log response of reef-core facies in well Mostovska-2. (1-boundstone)

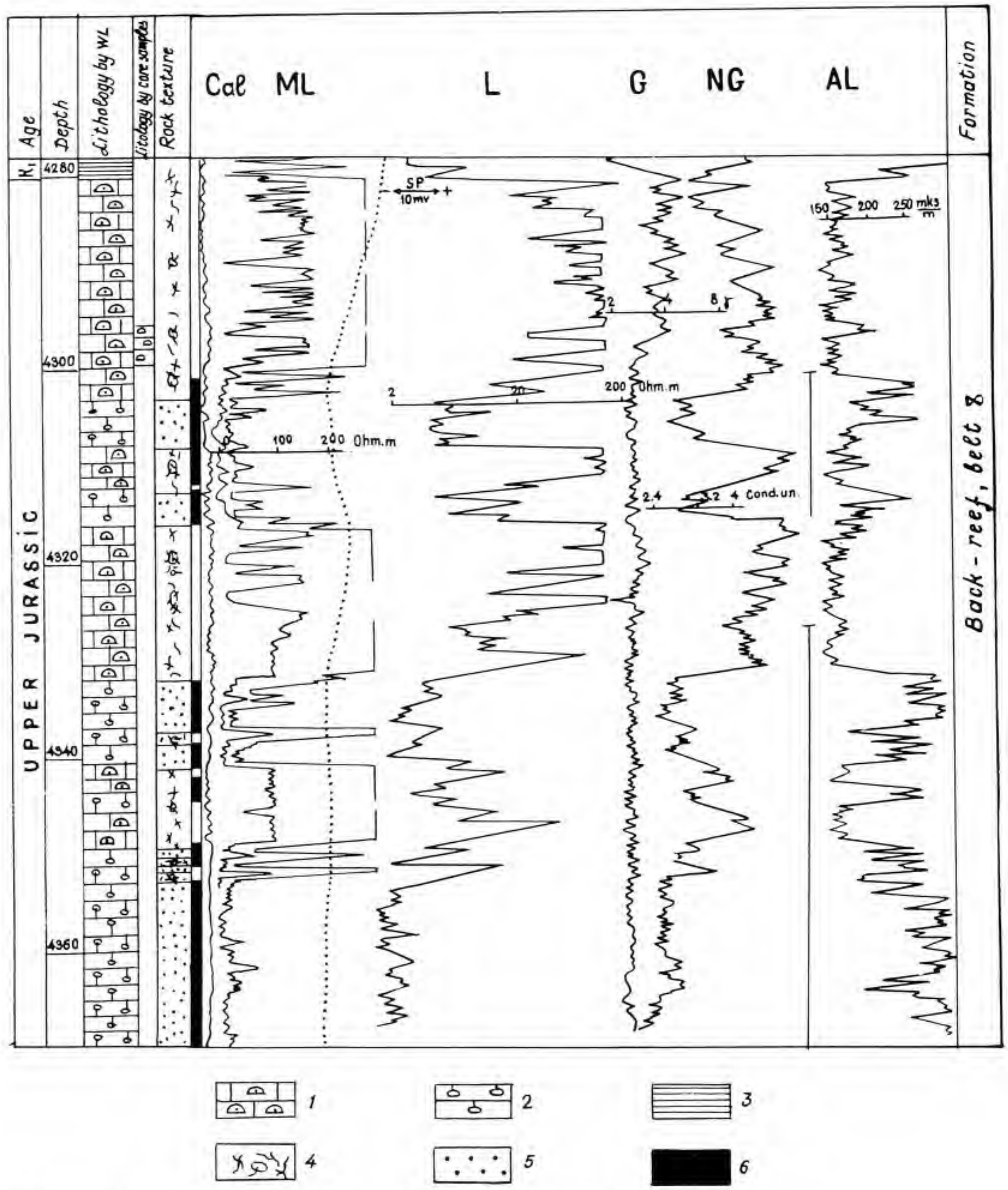

FIG. 6. Wire-line response of back-reef facies (belt 8) in well Lopushnya-4. 1-bioclastic limestone, 2-wacke-, pack- and grainstones, 3-calcareous shales, 4-microvuggy and fracture porosity intervals, 5-intergranular porosity intervals, 6-porous intervals.

Carpathian thrust sheets, at depths of 6 to 8 km, is indicated by reflection-seismic data (Figs. 1 and 7).

Reef growth was accompanied by the development of a relatively narrow, open marine back-reef shelf (facies belt 6) and a deeper water back-reef trough (facies belt 7) which was offset to the east and northeast by an open marine shelf (facies belt 8) grading laterally into a wide lagoonal shelf (facies belt 9). Sedimentation in this lagoon was characterized by a rhythmical alternation of limestones, dolomites and anhydrites, reflecting cyclical changes in water salinity during the deposition of the Rava-Russka formation. Limestones of this parasequence are composed of mud- and sand-sized particles and algal laminites. These were partly dolomitized or anhydritized. Fractured limestones, in part containing micro-vugs, are also encountered. These limestones, which have thicknesses of 10 to 25 m, are interbedded with dolomites and anhydrites. The limestone content of the Rava-Russka formation increases upwards towards its transition to the Nizhnev formation. The gradual decrease in anhydrite intercalations, and their total absence in the Nizhnev formation, indicates a progressive de-restriction of the back-reef lagoon. The limestones of the Nizhnev formation range in texture from mudstones to grainstones. Skeletal remains include bryozoans, coral, sponges and algae. Oolitic and nodular limestones can include a considerable amount of foraminifera.

During the **Late Tithonian**, environmental conditions became more uniform, probably due to a slight deepening of the basin and decreasing reef growth (decreased bioproductivity). Throughout the Ukrainian part of the Late Jurassic shelf, thick, regionally correlative bioclastic carbonates were deposited. These prograded towards the deeper waters back-reef trough. On Figs. 3 and 4 the white area shown behind the Oparian reef reflects the remnant water depth prior to deposition of the Early Cretaceous sediments. In the central parts of the Late Jurassic back-reef remnant trough, Tithonian carbonates are conformably overlain by Early Cretaceous limestones, containing some thin shale intercalations; to the East, these limestones give way to shales with clastic intercalations.

HYDROCARBON HABITAT

Source Rocks

Geochemical analyses of source-rocks are a traditional Achilles heel of Ukrainian geologists, particularly of the older generation. Therefore, special publications addressing the geochemistry of hydrocarbons contained in Late Jurassic reservoirs are lacking. However, based on regional geological considerations, we assume that possible source-rocks, which may have charged Late Jurassic reservoirs with liquid hydrocarbons, may be associated with the deltaic Middle Jurassic sequence of the northwestern parts of the foredeep whereas in its southeastern parts the Oligocene Menilites shales may be the primary source-rock. The gas contained in the Rudky and Letnya fields is probably of biogenic origin generated in Miocene strata. In view of the lack of reliable and up-to-date data we must desist from further discussions and speculations on this subject.

Reservoir Development

In view of strong lateral facies variations in the Late Jurassic carbonates, development of commercially viable reservoirs is very variable and differs in origin in the different facies belts.

Within the reef core, depositional interskeletal vugs and cavities are not preserved due their infilling with lime muds and subsequent re-crystallization. However, karstification of the reef core during the Paleogene erosional phase and as a result of sub-surface water circulation, caused the development of good reservoir porosities and permeabilities. Moreover, there is evidence for intra-Jurassic early leaching porosity developments.

In back-reef areas (facies belts 7 and 8) the best reservoirs are associated with friable limestones which are characterized by various textures, ranging from mudstones to grainstones. Of special interest are algal laminites which are very porous and intensely fractured (Markovsky et al., 1991).

Secondary reservoirs are provided by Early Cretaceous, Cenomanian and Miocene sandstones.

Seals

Early Cretaceous shales provide a sub-regional seal for Late Jurassic carbonate reservoirs. The sealing capacity of the anhydrites occurring within the Rava-Russka formation has not been established. Miocene shales and evaporites provide seals for Jurassic carbonates subcropping the Paleogene unconformity. In the southeastern part of the Carpathian foreland basin, shales of the flysch nappes, which are thrusted over Paleogene erosional surface, can provide effective seals for the autochthonous Mesozoic reservoirs (Fig. 7).

Traps

Established hydrocarbon accumulations are contained in two trap types, namely erosional highs and low-amplitude roll-over structures (Fig. 7). Both trap types are associated with the Paleogene erosional phase and the subsequent development of the Carpathian foredeep basin during which the structural configuration of the West-Ukrainian Late Jurassic carbonate shelf was profoundly modified.

During the Paleogene the entire area was raised above the erosional base level, resulting in the development of a southerly trending drainage system, as evident by the incision of palaeo-river valleys. Some of these cut through the Cretaceous and Late Jurassic strata and even into the underlaying early-Middle Jurassic and/or Palaeozoic sediments (see Sovchik and Vul, this volume). In the northwestern parts of the area, where Cretaceous sediments were completely removed during this erosional phase, Late Jurassic carbonates, both of the reefal and back-reefal type, uphold elongate palaeo-topographic highs. These were onlapped by transgressive Miocene shales and sandstones. The Badenian Baranivska shales and the Tyrassian gypsum and anhydrites provide effective seals for the Jurassic carbonates. Accumulations of this "sub-Badenian" type, which produce from Jurassic carbonates and Miocene sands, are the Rudky gas field, the Letnya gas/condensate field and the Kokhanovka oil field (Fig. 7)

Although development of the Neogene Carpathian foreland basins was accompanied by fault-controlled down-flexing of the foreland, this type of normal faulting was not as diffuse as for instance in the Austrian part of the Molasse basin, where it led to the development of a large number of mainly antithetic fault traps (fault throws of the order of 100-200 m; Kollmann and Malzer, 1980), but was concentrated on a few major faults. Amongst these, the Krakovets fault with a synthetic normal throw of 3000 m is the most important one (Fig. 2). So far no traps associated with Neogene normal faults have been established in the Ukrainian part of the Carpathian foreland basin. However, several low amplitude anticlinal roll-over structures are associated with the footwall block of the Krakovets fault; such a structures form the traps of the Lopushnya oil field (Figs. 2 and 8).

Beside the established trap types, there is some scope for additional traps. For instance, within the back-reef area between the Krakovets and Kalush faults, seismically mappable buried hill features occur which are upheld by the relief of the base Late Jurassic unconformity. These palaeotopographic anomalies, which have amplitudes of about 200 m, influenced sedimentation during the deposition of the Late Jurassic carbonates and, due to compaction drape, are also evident at Early and Late Cretaceous structural levels. To the southwest of the Krakovets fault some potential traps may be associated with the depositional configuration of the Oparian reef trend. In this area the reef envelope has a relief of 400 to 600 m towards the back-reef area and some 800 m towards the fore-reef area where sedimentation rates were considerably smaller than the bioproductivity in the reef core. Lateral variations in reef height and/or Paleogene valley incisions may provide for a wide range of possible traps, located at depths of 7-8 km beneath the Carpathian nappes. Some of these prospects are seismically mappable.

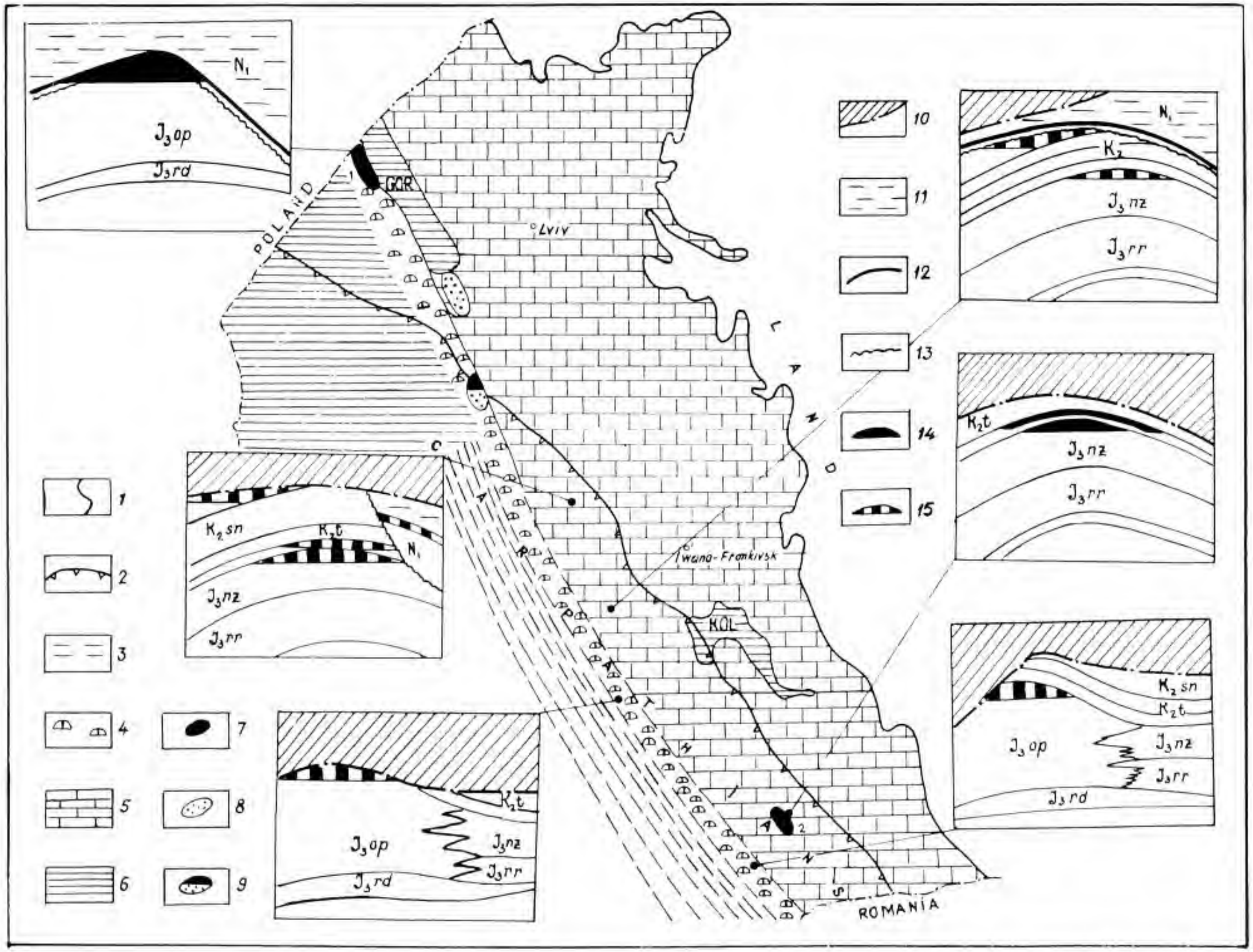

FIG. 7. Late Jurassic facies distribution and summary of established and potential Mesozoic trap types in the Carpathian foredeep. 1-boundary of Late Jurassic carbonate platform, 2-Carpathian nappe front, 3-fore-reef zone, 4-barrier reef complex, 5-back-reef zone, 6-area of eroded Late Jurassic sediments (GOR: Gorodok valley, KOL: Kolomiya valley), 7-oil fields (1: Kokhanovka, 2: Lopushnya), 8-Rudky gas field, 9-Letnya gas/condensate field, 10-nappes, 11-autochthonous Neogene, 12-Badenian seals, 13-top-Mesozoic erosional surface, 14-established fields, 15-potential traps.

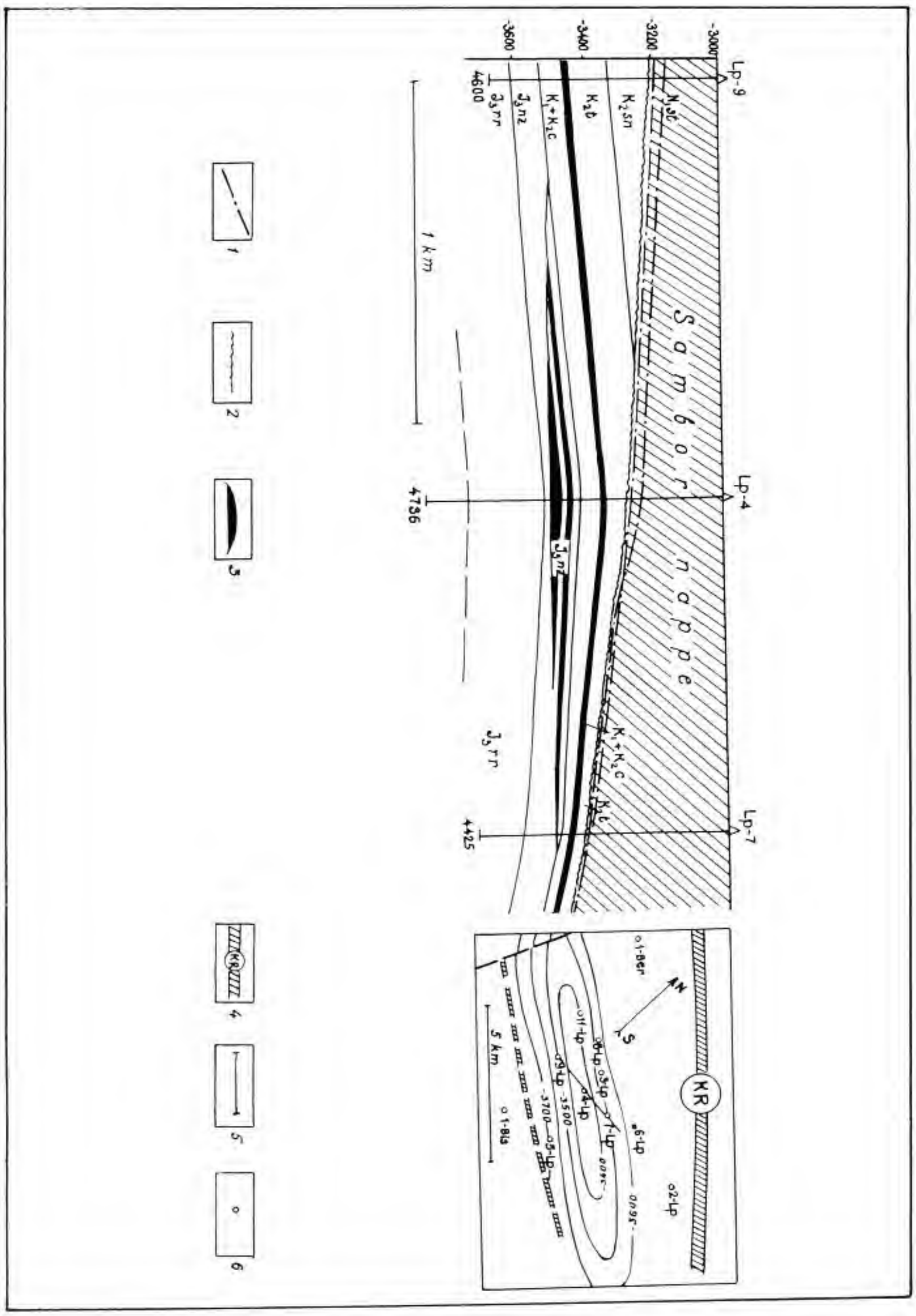

FIG. 8. Lopushnya oil field, cross-section and top-Jurassic structure map. 1-Sambor nappe sole-thrusts, 2-top-Mesozoic erosional surface, 3-pay zones, 4-Krakovets-Precarpathian regional fault, 5-cross-section line, 6-wells (LP: Lopushnya-, 1-Ber: Beregomet-1, 1-Bis: Biskiv-1).

OIL AND GAS FIELDS AND REMAINING HYDROCARBON POTENTIAL

Between 1940 and 1950 the Rudky gas field and the Kokhanovka oil field were discovered in the northwestern part of the Ukrainian Carpathian foredeep basin. Somewhat later the Letnya gas/condensate field was found in the same area. All three fields are of the sub-Badenian type and produce from Jurassic carbonates and onlapping Miocene sands. The Late Jurassic reservoirs of the Kokhanovka and Letnya fields consist of karstified reefal limestones of the Oparian formation which were eroded and leached during the Paleogene erosional phase to form erosional highs. Porosities reach 20%, permeability is fracture enhanced and pay thicknesses range up to 20 m. The Rudky field produced from karstified back-reef carbonates.

In 1984 the Lopushnya oil field was discovered in a sub-thrust position in the southeastern part of the Carpathian foredeep (Fig. 8). It established production from Jurassic carbonates and Cretaceous sands, forming part of the autochthonous Mesozoic sequence, which are involved in an anticlinal structures having an amplitude of some 200 m. This structure is clearly evident on reflection-seismic data. Production comes from Early Cretaceous and Cenomanian sandstones and Late Jurassic carbonates, sealed by Cretaceous shales.

Initial production from the Jurassic reservoir of well Lopushnya-4 amounted to 1130 bbls/day. This reservoir is formed by partly dolomitized lime-mudstones and grainstones as well as by algal limestones, characterized by abundant micro-vugs and fractures, of the Nizhnev formation which is here developed in facies type 8. Fig. 6 provides logs for the productive carbonate interval of well Lopushnya-4. Porosities of productive intervals are in the 8-20% range; the best reservoirs are formed by friable limestones which make up about 80% of the pay section.

In the entire southeastern part of the Ukrainian Late Jurassic carbonate shelf the Nizhnev formation is the prime objective horizon. Net reservoir thicknesses range between 10 and 60 m and are mainly tied to friable and micro-vuggy algal limestones.which have a regional distribution and are only lacking in Paleogene palaeo-valleys where the Nizhnev formation was partially or totally eroded.

In the vicinity of the Lopushnya field, seismic surveys permit to map a number of similar structural prospects beneath the Carpathian nappes.at depth of 5 to 7 km. To the south of these highs, reefal build-ups are expected which are encased in Early Cretaceous shales; the latter are only partially truncated by the Paleogene unconformity and are sealed by the Carpathian Sambor and Borislav-Pokutian flysch nappes. In the northern parts of the external Carpathians, where the Oparian reefs were stronger exhumed by Paleogene erosion, additional prospective structures have been mapped at the Jurassic objective level in a sub-thrust positions. However, despite of visible progress in the development of the sub-thrust play, it is still poorly evaluated, mainly due to insufficient reflection-seismic control. It is questioned whether a possible charge risk is a serious down-grading factor for this area. Yet, the integrity of sealing horizons may present a potential risk factor, as indicted by the failure of the recently drilled Tatalivke and Petrovets wells.

To the North of the Carpathian nappe front some prospects of the sub-Badenian type are recognized. Two of these were tested by the recently drilled exploration wells Vyzhomla-1 and Tyniv-2, located to the northwest of the Letnya gas/condensate field; both wells tested oil from karstified Jurassic carbonates.

It is concluded that the Ukrainian Carpathian foredeep still holds promising prospects, particularly in the sub-thrust autochthonous Mesozoic series. These warrant further evaluation by reflection-seismic detailing and drilling. The model presented for the Late Jurassic carbonate shelf and its reservoir potential requires further refinement as new core data becomes available.

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New data on the structure and hydrocarbon prospects of the Ukrainian Carpathians and their foreland

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ABSTRACT

In the West-Ukraine Carpathian and Volyn-Podolia hydrocarbon provinces 81 oil and gas fields have been discovered. These conatin ultimate recoverable reserves of some $1.2 \cdot 10^9$ bbls of oil and condensate and 15.5 TCF gas. The majority of these fields are located in the external parts of the Carpathians fold and thrust belt and in its adjacent foreland basin. Drilling of deep and super-deep wells resulted in the discovery of 7 oil accumulations in the depth range of 4-6 km.

The nappes of the Ukrainian Carpathians were thrusted during Oligocene and Miocene times over the European foreland platform over a distance of at least 35 km and possibly as much as 75 km. The prospectivity of the sub-thrust autochthonous series is highlighted by the Lopushnya oil field in the Bukovina part of the Carpathians, one field in Romania, 11 fields in Poland and 19 fields in Slovakia. These accumulations, which are partly sealed by the flysch nappes, produce from a variety of reservoirs that were charged with hydrocarbons generated from Paleogene and possibly also Mesozoic source-rocks. The Ukrainian autochthonous sub-thrust play holds the potential for further important hydrocarbon discoveries in structures associated with down-faulting of the foreland crust.

Gas accumulations occurring in the Carpathian foredeep, the Biliche-Volitsa zone, and in the frontal Carpathian structures of the Sambor unit are charged by biogenic gas. In the allochthonous Carpathian flysch, Early Cretaceous and Paleogene sands involved in the Borislav-Pokut, Scybia and Silesian nappes are the principal objectives, as indicated by the occurrence of a number of oil and oil-and-gas fields which are charged by hydrocarbons generated from Paleogene and possibly Early Cretaceous source-rocks. Pressure data and formation water salinities indicate that shales associated with the base of major nappe units act as seals. Intra-formational seals provide for stacked hydrocarbon accumulations in dip-closed anticlinal structures beneath the cover of higher nappes. Sub-thrust allochthonous and parautochthonous anticlinal roll-over structures hold a considerable potential for future discoveries, particularly if existing reflection seismic resolution problems can be solved.

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INTRODUCTION

The West-Ukrainian hydrocarbon province covers an area of some 44000 km^2 and includes the Carpathian fold belt and its Volyn-Podolia foreland. As oil extraction started in this area already around 1771, it is one of the oldest hydrocarbon provinces of the world. In 1909 the Borislav field was discovered from which subsequently 1.92 million tons ($14.2 \cdot 10^6$ bbls) of oil were produced. In 1924 the first gas field (Dashava) was discovered. After a period of little activity, modern exploration intensified in the 1950's after the discovery of the Dolina and Bitkiv oil fields. By now over 90 hydrocarbon fields have been discovered. At present 31 oil, 7 oil-and-gas, 6 gas-condensate and 37 gas fields are in production. Oil production peaked in 1967 at a level of $2.86 \cdot 10^6$ t/year ($21 \cdot 106$ bbls) whereas gas production peaked in 1969 at the level of $12.57 \cdot 10^9$ m^3/year (470 Bcf). Ultimate recoverable reserves in established accumulations amount to some $163.3 \cdot 10^6$ t ($1.2 \cdot 10^9$ bbls) of oil and condensate and $415 \cdot 10^9$ m^3 (15. 5 TCF) of free and associated gas. By spring 1994 cumulative production amounted to $104 \cdot 10^6$ t ($770 \cdot 10^6$ bbls) of oil and condensate and $277 \cdot 10^9$ m3 (10.3 TCF) of gas.

The Carpathian part of the West-Ukrainian oil and gas province is limited to to the northeast by the Bilche-Volitsa zone, corresponding to the Carpathian foredeep basin, and to the west by the Neogene Transcarpathian Depression. The Carpathian thrust- and fold-belt, involving mainly Cretaceous and Cenozoic flysch series, has been subdivided, according to nappe correlations, into an outer Sambor unit, which borders the Bilche-Volitsa zone, and the progressively more internal Borislav-Pokut, Skiba, Silesian and Dukla-Chernogora and Magura units (Fig. 1) (see also Bessereau et al., this volume). To date 38 gas accumulations have been established in the Biliche-Volitsa and the Sambor zones whereas 36 oil and oil-and-gas accumulations were found in the Borislav-Pokut zone. In the internal zones of the Carpathians so far only 2 oil accumulations were found.

The Transcarpathian Depression, which contains up to 2000 m of Neogene clastic sediments, halites and volcanic rocks, is an extensional basin which developed on top of the inner Carpathian nappes; its structure is complicated by the diapirism of Miocene salts. This basin hosts four gas accumulations in Neogene sands.

The Carpathian foreland is occupied by the Volyn-Podolia platform. Its eastern parts are underlain by Precambrian basement which is covered by a westwards expanding wedge of Riphean to Carboniferous sediment, attaining maximum thicknesses of some 7000 m, and a relatively thin veneer of Mesozoic and Cenozoic series. In contrast, the western parts of this platform are floored by folded Palaeozoic sediment which were deformed during the Caledonian and Variscan orogenies; these are covered by up to 2000 m of Mesozoic and Cenozoic sediments. The boundary between these two basement provinces corresponds to the Tornquist-Teisseyre line, a major tectonic lineament which was reactivated time and again during the Mesozoic evolution of the Carpathian geosynclinal system and its Alpine destruction.

The Volyn-Podolia platform is characterized at top-Mesozoic level by a relatively shallow, gently southwestward dipping monocline that shows a low level of structuration. Beneath the Carpathian thrust front, this surface drops down abruptly to depths of 3 to 9 km along a system of major normal faults, such as the Krakovets fault. Based on geophysical data, the foreland crust extends some 75 km beneath the Carpathian edifice of stacked nappes (Fig. 3). Devonian, Late Jurassic and Cenomanian carbonates and Early Cretaceous sands of the Volyn-Podolia platform and its extension beneath the Carpathian thrust and fold belt host a number of oil accumulations.

At present the West-Ukrainian hydrocarbon province includes 38 fields which produce from reservoirs occurring within the Cenozoic, Cretaceous, Jurassic and Devonian strata of the Carpathian foreland and the Carpathian sub-thrust autochthonous sequences, and 36 fields which produce from Cretaceous and Cenozoic reservoirs involved in the folded and thrusted structures of the Carpathians. Two oil, one oil-and-gas and three gas fields each have initial technically recoverable reserves in excess of $30 \cdot 10^6$ t ($200 \cdot 10^6$ bbls) oil and oil-equivalents. One oil and seven gas fields are in the $10\text{-}30 \cdot 10^6$ t ($75\text{-}200 \cdot 10^6$ bbls) class;

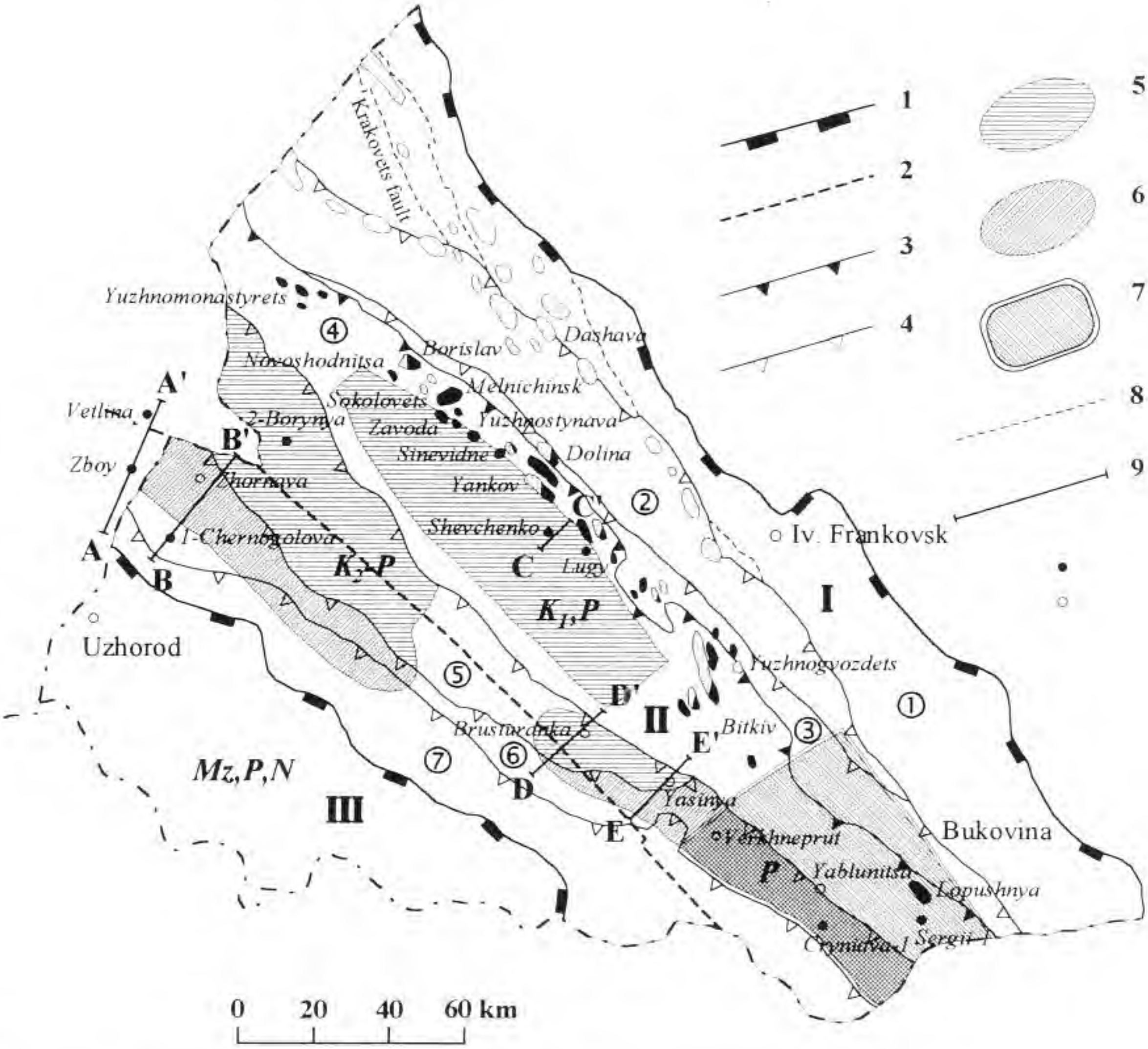

FIG. 1. Tectonic units of the West-Ukrainian Carpathians and oil and gas fields. oil fields: black, gas: fields dotted

I: Carpathian foredeep - 1. undeformed Biliche-Volitsa zone, 2. Deformed Sambor zone, 3. Borislav-Pokut nappe.

II: Stacked Carpathian nappes - 4. Scybia nappe, 5. Silesian nappe, 6. Dukla-Chernogora nappe, 7. unprospective Magura nappe.

III Transcarpathian Basin

Line symbols: 1. Basin outlines, 2. approximate western limit of autochthonous foreland crust, 3. outcropping boundaries of Carpathian nappes, 4. subsurface nappe fronts, 5. prospective area in exposed nappes with age of objective series, 6. prospective area in sub-thrust allochthonous and parautochthonous units with age of objective series, 7. prospective area in sub-thrust autochthon, 8. regional normal faults affecting foreland and autochthon, 9. Lines A-A', B-B', C-C', D-D': location of cross-sections given in Fig. 4

Well symbols: black- drilled wells; open- proposed well locations

Oil fields: black; gas fields: dotted

67 fields (34 oil, 33 gas) each have reserves below $10 \cdot 10^6$ t ($75 \cdot 10^6$ bbls) oil and oil equivalents.

In the past exploration activity was largely limited by the drilling capacity. Whereas up to 1965 no wells were drilled to depths of 4000 m, such wells made up 10% and 38% of all wells drilled during the periods of 1966 to 1970 and 1971 to 1975, respectively. In the 1970's 13 wells were drilled in the Carpathian foredeep and within the Carpathians to depths of more than 6000 m. Amongst these, the wells Sinevidne and Shevchenko reached total depths of 7001 and 7520 m, respectively. Although deep wells have yielded important new structural and stratigraphic information, only a fraction of the expected reserve potential of the deep plays has so far been proven up.

For the entire area recoverable reserves in established accumulations (production, proven and probable reserves), amounting to some $400 \cdot 10^6$ t ($3 \cdot 10^9$ bbls) of oil and oil equivalents, account for approximately 43% of its expected ultimate reserve potential. In the Carpathian foredeep some 55% of the ultimate potential reserves have so far been proven up. Although plays at the depth interval of 4-7 km are thought to hold a considerable potential (close to 30% of the total regional potential reserves), large areas are still poorly explored.

VOLYN-PODOLIA PLATFORM AND CARPATHIAN AUTOCHTHON

The eastern Volyn-Podolia platform forms part of the Precambrian East-European Craton, the western limit of which is defined by the Tornquist-Teisseyre zone that coincides with the Palaeozoic Caledonian and Variscan deformation fronts. This northwest-southeast trending line extends from Poland into the Ukrainian Carpathian foreland where it forms the western limit of the deep Lublin-Lviv Palaeozoic basin. Southeastwards the Tornquist line projects beneath the Bilche-Volitsa and Sambor zones of the central and southern Ukrainian Carpathians.

The the Precambrian basement of the Volyn-Podolia Platform dips gently westwards and reaches depths of some 9 km to the northwest of Lviv. It is covered by up to 800 m of Riphean continental clastics and a westwards expanding wedge of Cambrian, Ordovician and Silurian sands, shales and carbonates, attaining maximum thicknesses of some 4000 m near the Late Caledonian deformation front which was established by boreholes to the west of Lviv. After a short break at the transition from the Silurian to the Devonian, sedimentation resumed with the accumulation of Early Devonian red beds which are overlain by Middle and Late Devonian carbonates, Early Carboniferous shales and carbonates and a Late Carboniferous paralic, partly coal bearing sequence which is exploited in the Lublin-Lviv coal basin. Also the Late Palaeozoic sediments form a westward expanding wedge which attains maximum thicknesses of some 3000 m in the area of Lviv, to the west of which they were deeply truncated as a consequence of their Variscan deformation (Rizun and Sen'kovskiy, 1973; Wjalow and Medwedew, 1977).

During Permian and Early Mesozoic times, the Volyn-Podolia Platform and its extension beneath the Carpathian nappes was an area non-deposition and erosion. In contrast, contemporaneous rifting activity resulted in and the subsidence of the Polish Trough and the opening of the oceanic Magura basin in the internal Carpathian domain (Birkenmajer, 1986; Kovac et al., 1993). In the area of the Volyn-Podolia Platform sedimentation resumed during the Middle and Late Jurassic, initially with the deposition of a deltaic series followed by the establishment of a broad carbonate shelf (Izotova and Popadyuk, this volume). Marine sedimentation persisted until the end of the Cretaceous when the western parts of this shelf were deformed, uplifted and subjected to erosion in conjunction with the inversion of the Polish Trough (Ziegler, 1990; Bessereau et al., this volume). During the Paleogene erosional phase a system of southwesterly trending palaeo-valleys developed. These cut deeply into the Mesozoic and Palaeozoic cover of the Volyn-Podolia Platform (Fig. 2). This Paleogene palaeotopographic relief was inundated during the Eocene and Oligocene in conjunction with the development of the Carpathian foreland basin in which sedimentation persisted until

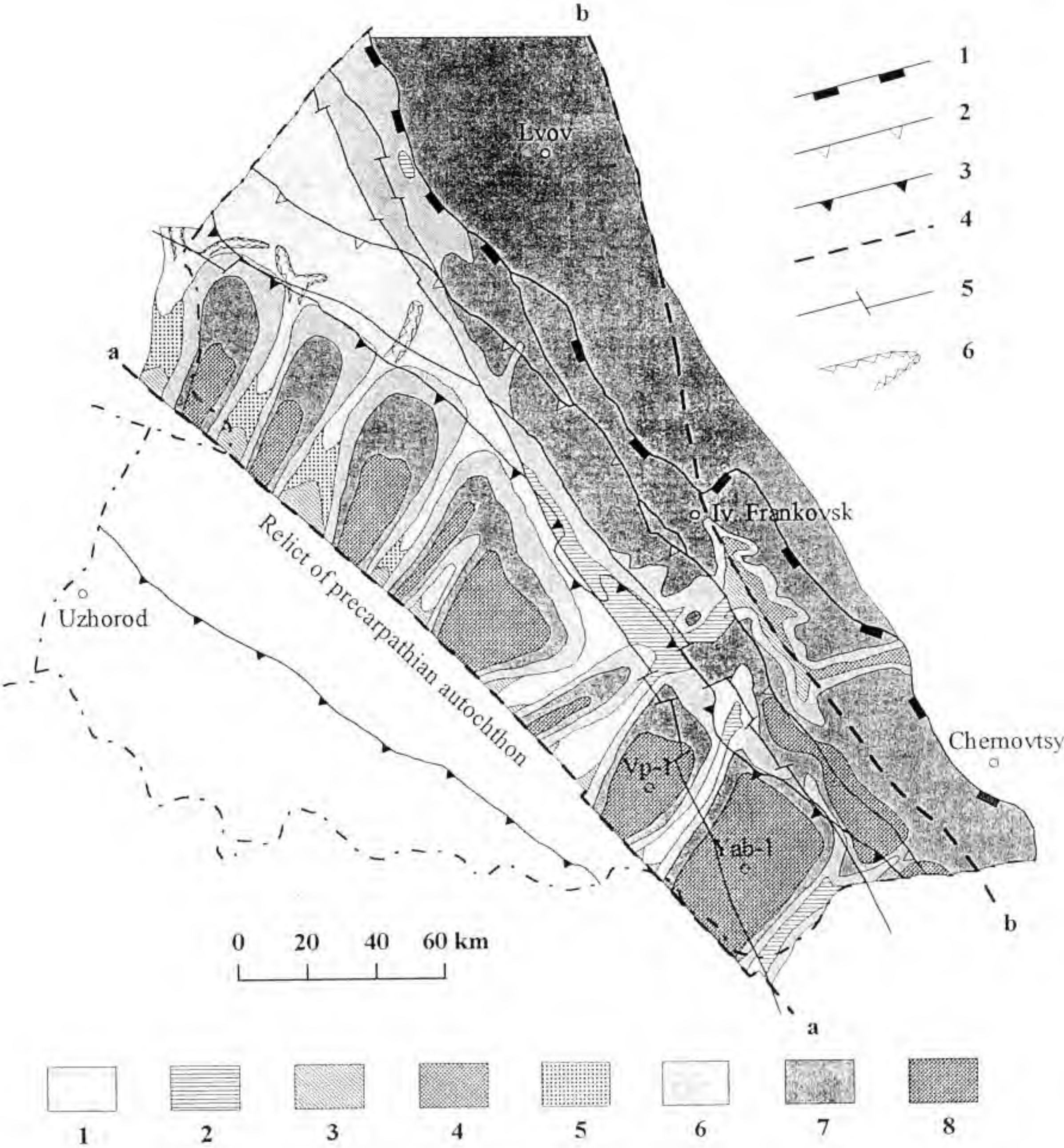

FIG. 2. Tentative subcrop map of pre-Neogene erosional surface of sub-thrust autochthon.

Line symbols: 1. Eastern margin of Biliche-Volitsa zone, 2. Carpathian thrust front (eastern margin of Sambor unit), 3. Eastern margin of Scybia nappe, 4a) Western margin of autochthonous foreland crust, 4b) Tornquist-Teisseyre Line separating Precambrian East-European Platform from Palaeozoic crust of Central Europe, 5. major normal faults affecting autochthon, 6. incised valleys of the pre-Neogene erosional surface

Area symbols (subcropping units): 1. Upper Proterozoic, 2. Cambrian, 3. Ordovician to lowermost Devonian, 4. Early Devonian red beds(Dniester series), 5. Devonian-Carboniferous, 6. Jurassic, 7. Cretaceous, 8. Paleogene

Vp - Verkhneprut, Yab - Yablunitsa proposed stratigraphic tests

Pliocene times. Subsidence of this basin was accompanied by the development of a system of synthetic normal faults amongst which the Krakovets fault, having a throw of up to 3000 m, is the most important one. Emplacement of the Carpathian nappes on the passive margin of the Volyn-Podolia Platform commenced during the late Oligocene and terminated at the end of the Miocene (Ellouz and Roca, 1994).

Major reservoirs established on the Volyn-Podolia Platform and on its southwestward extension beneath the Carpathian nappes are Devonian sands and carbonates, Late Jurassic carbonates and Early Cretaceous, Cenomanian and Paleogene sands (Izotova and Popadyuk, this volume). Potential source-rock are shales of Upper Proterozoic, Cambrian and Silurian age, Early Cretaceous shales and the Oligocene Menilites shales. Plays aimed at Palaeozoic reservoirs and source-rocks are limited to the West by the Tornquist-Teisseyre zone.

HYDROCARBON ACCUMULATIONS AND PROSPECTS IN THE CARPATHIAN AUTOCHTHON AND PARAUTOCHTHON

In view of the above, the reservoir potential of the autochthonous foreland which extends deep under the Carpathian nappes, is restricted to the Late Jurassic carbonates and Early Cretaceous, Cenomanian, Paleogene and Neogene sands. In the Ukraine, 20 oil and gas fields have been established in autochthonous sediments beneath the Carpathian nappes at depths up to 4300 m. An example is the high out-put Lopushnya oil field which produces from stacked Mesozoic and Paleogene reservoirs at depths between 4000 and 4300 m (Izotova and Popadyuk, this volume). The potential of the sub-thrust play is highlighted by the discovery of the Frasyn field in Romania, 19 fields in Slovakia and 11 fields in Poland. These accumulations, which are partly sealed by the flysch nappes and produce from a variety of reservoirs, are charged with hydrocarbons generated from Paleogene and possibly also Mesozoic source-rocks.

In the Ukrainian sub-thrust play, the distribution of Mesozoic reservoirs is controlled by the westward shale-out of the Late Jurassic carbonates (Izotova and Popadyuk, this volume) and by the Paleogene erosional unconformity which truncates all Mesozoic objective horizons. However, this unconformity can also contribute towards reservoir development by means of karstification of Jurassic carbonates. Fig. 2 presents a tentative Neogene subcrop map which is based on well and reflection seismic data. Beneath the Carpathian nappes this erosional surface is located a depths ranging from 2 to 9 km, as shown in Fig. 3.

The sub-thrust autochthonous sedimentary sequence includes, apart from several reservoir horizons, Lower Cretaceous and Neogene seals. In addition, sheared shales at the base of the flysch nappes have a sealing capacity as evident, for instance, in the stratigraphic Grinyava-1 well. Hydrocarbon supply to autochthonous sub-thrust prospects does not appear to be a problem and is apparently provided by the Oligocene Menilites shales and possibly also by Early Cretaceous shales and Late Jurassic sediments developed in an off-reef facies (see Bessereau et al., this volume). The occurrence of oil accumulations down to depths of 5300 m highlights the potential of this play and raises doubts whether over-maturity of source rocks is a limiting factor.

In the deeper parts of the Carpathian sub-thrust play, reflection-seismic definition of drillable structures at the level of the autochthonous and parautochthonous series has so far been difficult. However, a number of oil accumulations have been discovered at depths of 4300 to 5800 m (Juzhnomonastyrets, Novoskhodnitsa, Sokolovets, Zavada, Melnichinsk, Yuzhnostynava, Jankov and Juznogvozdets fields). At present a number of structural leads are recognized which require detailing and the application of the most modern reflection-seismic techniques. Zones of interest are in the southeast the North Bukovina transverse uplift (5-6 km depth) and in the northwest the area covered by the Sambor nappe (Fig. 3). Similarly prospects may be associated with the platform marginal faults, including roll-over structures in down-thrown hanging wall blocks (Izotova and Popadyuk, this volume).

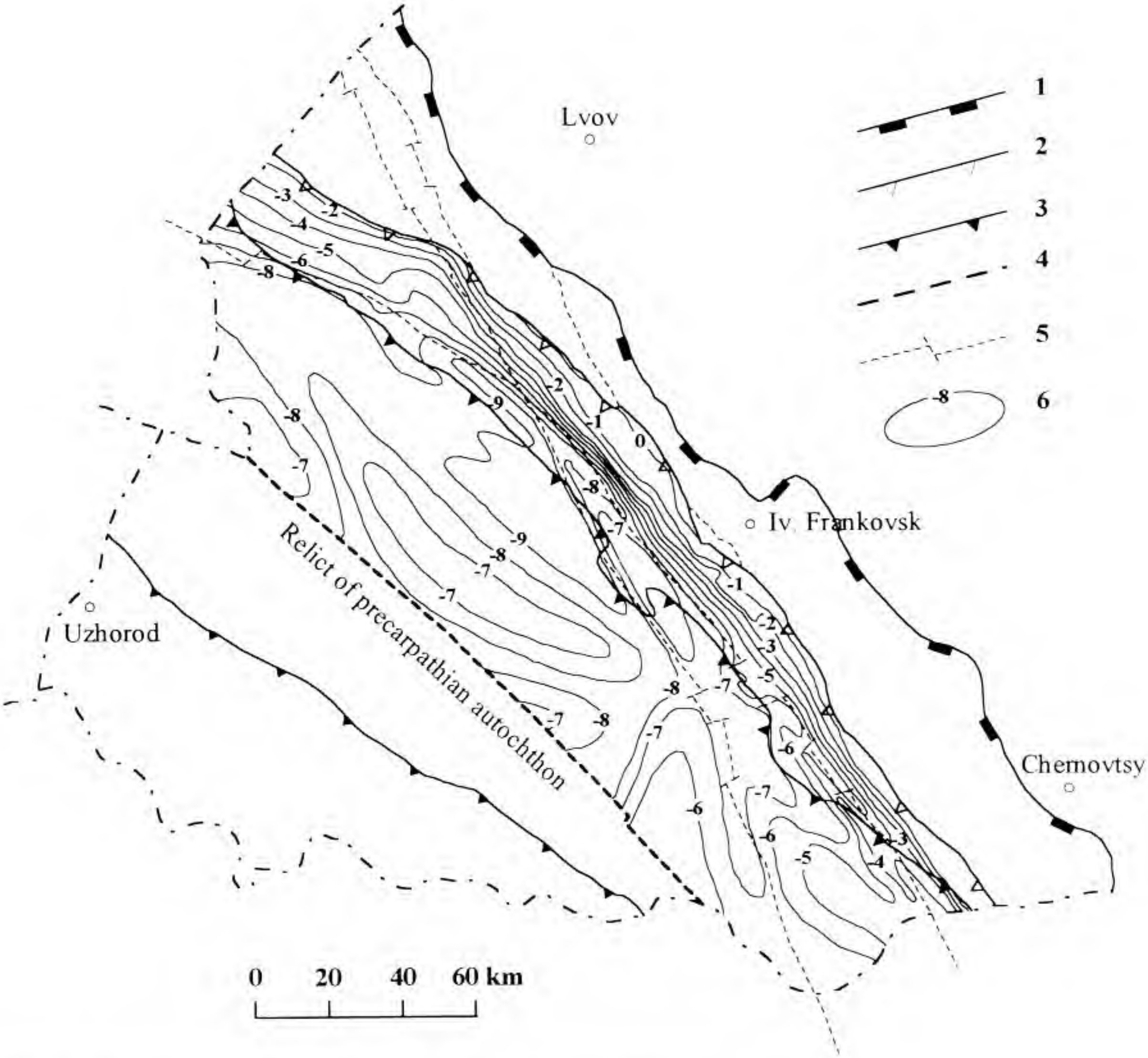

FIG. 3. Structural map of the pre-Neogene unconformity of sub-thrust autochthon and foreland.
Line symbols: 1. Eastern margin of Biliche-Volitsa zone. 2. Carpathian thrust front (eastern margin of Sambor unit). 3. Eastern margin of Scybia nappe. 4. Western margin of autochthonous foreland crust. 5. major normal faults affecting autochthon. 6. Depth contours of Paleogene unconformity in km.

The autochthonous sub-thrust play of the Carpathians still holds considerable potential and, despite certain seismic resolution problems and great objective depths, should not be unduly downgraded.

CARPATHIAN NAPPES

The Carpathian nappes involve a continuous sequence of Early Cretaceous to Miocene shales and flysch-type sandstones which attain thicknesses of 5 to 8 km. These clastics accumulated in deeper water basins that were floored by extended continental crust and possibly partly by oceanic crust. Sands were shed into these basin from the rising Carpathian orogen and to a lesser degree from the Volyn-Podolia shelf. A tentative palinspastic restoration of the Ukrainian Carpathians suggests an overall shortening of some 230 km since the Cretaceous; of this, about 180 km was achieved during the Late Oligocene to Pliocene folding and thrusting of the outer Carpathian flysch units (Ellouz and Roca, 1994).

Reflection-seismic data and results of deep wells, such as Sergii-1 (drilled 15 km to the West of the Carpathian thrust front, bottomed in autochthonous Badenian 5023 m), show that the Carpathian nappes were thrusted a minimum of 35 km over the autochthonous foreland. However, on reflection-seismic data the foreland crust can be traced westwards at least as far as the Chernogora nappe which is apparently floored by parautochthonous continental basement; this would imply a nappe transport of up to 75 km. How much of the basement of the Borislav-Pokut, Skiba and Silesian basins was subducted at the leading edge of the parautochthonous block, which underlies the Dukla-Chernogora nappe, is a matter of debate. However, Neogene calcalkaline volcanic activity in the Transcarpathian basin testifies to the subduction of a large amount of crustal material which had underlain these flysch basins (Szabo et al., 1992).

The thickness of the Carpathian nappe stack is adequately constrained by wells and reflection seismic data in the Sambor and Borislav-Pokut zones but is only partially known in the internal Carpathians where it may exceed 8000 m. To this end, plans have been formulated to drill two test wells in the Dukla-Chernogora nappe. The well Verkhneprut is scheduled for a total depth of 8000 m. The Yablunitsa test with a planned total depth of 5950 m will be located on a deep seated autochthonous structure (Fig. 1).

The Biliche-Volitsa zone, which corresponds to the little deformed Carpathian foredeep, and the adjacent frontal Sambor zone of the Carpathian thrust belt are well explored and contain 42 gas accumulation and one oil-and-gas field. Their reservoirs are formed by Late Jurassic carbonates and Upper Cretaceous and Neogene sandstones. Similar to the adjacent Polish fields, the gas is of biogenic origin. The oil is most likely related to Palaeozoic source-rocks.

On the other hand, exploration activity aimed at evaluating the hydrocarbon potential of the main body of the Carpathian allochthon was in the past at a low level and amounted to only 5% of the total exploration effort.

Drilling activity was mainly directed towards the assessment of the hydrocarbon potential of the Early Cretaceous an Paleogene series of the Scyba nappe and of the Paleogene series of the Silesian zone. Most wells were located on surface structures. Results show that anticlinal structures, which rely for closure on thrust faults, are wet whereas thrusted anticlines with a 4-way dip closure contain hydrocarbons, sometimes in stacked accumulations sealed by shales intercalated with the reservoir sands. Although regional seals are provided by the Oligocene Menilites shales, the high sand/shale ratio of the objective section probably plays an important role in the apparently limited sealing capacity of individual thrust faults (Sovchik, 1979; Sovchik and Krupsky, 1988).

On the other hand, the deep well Gryniava-1, which drilled through the base of the Chernogora nappe 7 km to the west of its erosional edge, tested from the Oligocene Krosno flysch of the underlying Silesian nappe under anomalously high formation pressures gas at commercial flow rates. Together with the results of similar exploratory wells, this suggests that sheared shales at the base of major nappes do have a considerable sealing potential. Moreover, under high pressure condi-

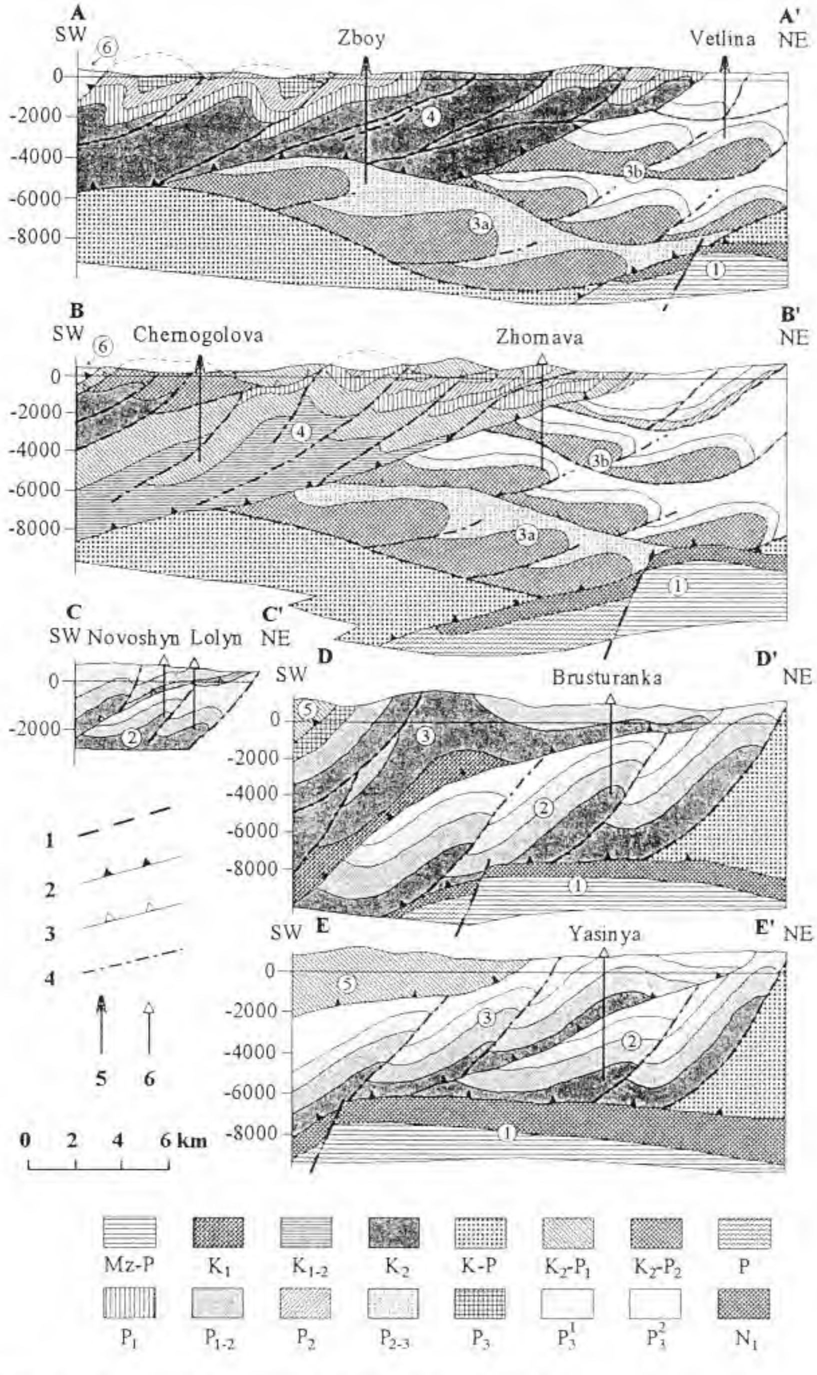

FIG. 4. Structural cross-section, for location see Fig. 1
Tectonic units (numbers in circles): 1. autochthon, 2. Scybia nappe, 3. Silesian nappe (3A Obidov sub-zone, 3B Predukla sub-zone), 4. Dukla nappe, 5. Chernogora nappe, 6. Pokulets nappe
Line symbols: 1. normal faults affecting autochthon, 2. nappe boundaries, 3. boundaries between sub-nappes, 4. thrust faults within nappes
Well symbols: black- drilled wells; open- proposed wells

tions, intra-formational shales appear to have a higher sealing potential than under hydrostatic conditions. This notion is in keeping with the results of the Lopushnya field which is partly sealed by the overlying Sambor nappe and partly by intra-formational Mesozoic shales. Supporting evidence for effective hydrodynamic separation between nappes is provided by significant changes in the salinity of formation waters, as seen, for instance, in the wells Borynya-2 and Zboy-1 (Fig. 4a, Durkovic et al., 1980). Whereas nappes exposed at the surface are deeply invaded by meteoric waters, normal salinities are encountered under sub-thrust conditions; as such, this probably contributes towards the preservation of hydrocarbon accumulations under sub-thrust conditions.

Generally it is observed that the degree of organic metamorphism and diagenetic deterioration of reservoirs is lower in the parautochthonous and external nappes than in the overlying, more internal nappes (Khain and Sokolov, 1990). That indeed viable reservoirs occur at considerable depths is illustrated by test results of the well Shevchenkovo-1, which recovered water from Cretaceous sands at depths of 6930-6990 m at the rate of 16 m^3/d, and Lugy-1 which recorded from Cretaceous sands flow rates of water of 12 m^3/d from interval 6180-6260 m and 58 m^3/d from interval 5430-5525 m. In the Borislav-Pokut zone, commercial flow rates of oil (27-1633 b/d) were obtained from Paleogene sands in the following fields: Yuzhnomonastyrets (4945-4962 m), Novoshodnitsa (4365-5050 m), Sokolovets (5704-5796 m), Zavada (4390-5050 m), Melnichinsk (4497-4790 m), Yuzhnostynava (4677-4712 m, Yankov (5183-5292 m) and Yuzhnogvozdets (4080-4386 m).

Surface geological mapping, results of deep wells and reflection-seismic data indicate that the internal nappes are characterized by a considerably greater structural complexity than the more external nappes (Fig. 4). On the other hand, frontal thrust structures are generally steep and rely on fault closure whereas more internal structures are characterized by a lower relief and large anticlinal roll-overs. As structures of this type do not exclusively rely on fault closure, they have a considerably greater potential to contain commercial volumes of hydrocarbons.

Keeping the above developed concepts in mind, future exploration should be aimed at assessing the hydrocarbon potential of those parts of the nappes which are covered by more internal nappes. This applies specifically to the inner parts of the Scybia and Silesian nappes which are covered by the Silesian and the Dukla-Chernogora nappe, respectively. The principal targets are Oligocene Krosno sands involved in role-over anticlinal structures.

This play concept is illustrated by the Brusturanka and Yasinya prospects shown in Figs. 4d and 4e; both of these planned wells are aimed at structures within the Scybia nappe which are covered by the Silesian nappe. However, as definition of the prospective structures is hampered by poor seismic resolution, these wells carry a considerable structural risk. In the southeastern parts of the Carpathians, interesting exploration targets are sub-thrust prospects beneath the Chernogora nappe in which first encouraging results were obtained in well Grinyava-1. Drilling targets are here the Krosno sands of the Silesian and Scyba nappes as well as the Mesozoic and Cenozoic series of the underlying autochthon. Similar prospects may exist beneath the Chernogora-Dukla nappe in the Silesian nappe as illustrated in Figs. 4a and 4b. An example is the Zhornava prospect, located 6.5 km to the southwest of the erosional edge of the Dukla nappe, which aims at evaluating the potential of the Krosno sands involved in a gentle anticlinal structure.

Due to the complex structuration of the internal nappes, reflection-seismic resolution of such sub-thrust prospects is generally poor. However, if seismic resolution can be improved by applying modern technology, such as 3-D surveys, a new cycle of successful exploration may be opened. It is anticipated that future exploration in the Carpathian allochthon will be rewarded with the discovery of a number of small and medium sized hydrocarbon accumulations occurring at a depth range of 4-6 km.

CONCLUSIONS

Although exploration for hydrocarbons in the Ukrainian Carpathians and their immediate foreland can look back on a long and successful history, the application of modern reflection-seismic technologies may open a new cycle of exploration activity. Experience gained during past exploration cycles shows that there is no shortage in hydrocarbon supply to properly sealed traps occurring in the sub-thrust autochthonous and foreland series as well as within the Carpathian allochthon. Both the autochthonous sub-thrust play and prospects within the Scyba, Borislav-Pokut and Silesian nappes are oil-prone.

In the southeastern parts of the Carpathians, the autochthonous substrate with its Mesozoic and Paleogene reservoirs is a of zone of prime interest. Within the Carpathian allochthon, the Paleogene and Early Cretaceous sands of the Scyba and Borislav-Pokut nappes and the Paleogene sands of the Silesian zone are likely to yield additional discoveries.

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Tectonic setting and hydrocarbon habitat of the Romanian external Carpathians

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ABSTRACT

The external part of the Romanian Carpathians hosts three more or less discrete hydrocarbon provinces, namely the Bistrita-Trotus and the Carpathian Bend provinces of the East Carpathians and the Getic Depression province of the Southern Carpathians. All three provinces appear to be intimately related to Oligocene-Early Miocene oil-prone source rocks; however, a contribution from Cretaceous source rocks cannot be excluded. The Bistrita-Trotus and the Carpathian Bend provinces are characterized by thin-skinned nappes, involving Cretaceous, Paleogene and Neogene sediments, which override a deeply subsided autochthonous foreland. The Getic foreland basin contains Eocene to Pliocene molasse-type sediments which are involved in basement-controlled compressional and transpressional structures.

In the Bistrita-Trotus province, mostly shallow, small and medium fields produce from Paleogene flysch of the Marginal Folds nappe involved in complex structures beneath the Tarcau nappe.

In the prolific Carpathian Bend province, Oligocene to Pliocene shallow marine, deltaic series, involved in the Tarcau, Marginal Folds and Subcarpathian nappes, contain multiple reservoir-seal pairs. Structural traps are associated with all nappe units. Unconformities, related to the different compressional phases, provide for additional traps. Established fields are contained in relatively shallow structures which attained their present configuration during the terminal Pliocene deformation phase.

In the Getic Depression, oil accumulations are closely related to the distribution of Paleogene series; gas-prone Mio-Pliocene source rocks charged Late Miocene and Pliocene reservoirs, involved in structural and combined stratigraphic/structural traps of the southern part of the Getic Depression.

Deep seated structures of the external Carpathians fold-and-thrust belt have probably a considerable hydrocarbon potential; however, definition of such prospects requires improved reflection-seismic resolution. Subthrust plays, aiming at the sedimentary cover of the underthrusted foreland, are restricted to the northern part of the Eastern Carpathians.

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INTRODUCTION

Statistics show that Romania is among the first oil producing countries of the world. First oil production has been recorded in 1857 at a rate of 275 tons/year. However, the extraction of crude at Mosoarele, Poieni, Doftana and Pacureti, located in the Romanian provinces of Moldavia and Valachia, has been mentioned by foreign travellers already since the first half of the 16th century.

In 1861, the first well was dug mechanically at Mosoarele, Moldavia. In 1900, Romania was the third largest oil producer of the world with an annual production of $0.3 \cdot 10^6$ tons/year. In 1953-1955, the oil output of Romania was 9-$10 \cdot 10^6$ tons/year, and in 1976 a maximum oil output of $14.6 \cdot 10^6$ tons was achieved (Fig. 1). After 1976, crude production in Romania decreased gradually and more rapidly during the last years. In 1994, oil production was at the level of $6.4 \cdot 10^6$ tons.

For a better understanding of the geology of Romanian and the evaluation of prospective areas detailed geological maps and synthesis have been drawn up over the years. All geophysical methods were applied in regional and detailed research efforts, especially seismic ones. In onshore prospective areas, the density of available reflection-seismic coverage amounts to about 1.75 km profiles per km^2 and in the Black Sea off-shore to about 2.4 km profiles per km^2.

About 400 wells deeper than 3500 meters have been drilled in an effort to explore the deep structure and hydrocarbon potential of the country. The deepest well of Romania was drilled in the Baicoi field and reached a total depth of 7025 meters.

The Carpathian fold-and-thrust belt is the least explored part of Romania. Its very roughly topographic relief presents difficulties in the acquisition of reflection-seismic lines and its very complicated internal structure is often difficult to resolve.

TECTONIC SETTING

During the Alpine evolution of Romania, two distinct depositional areas evolved in the external

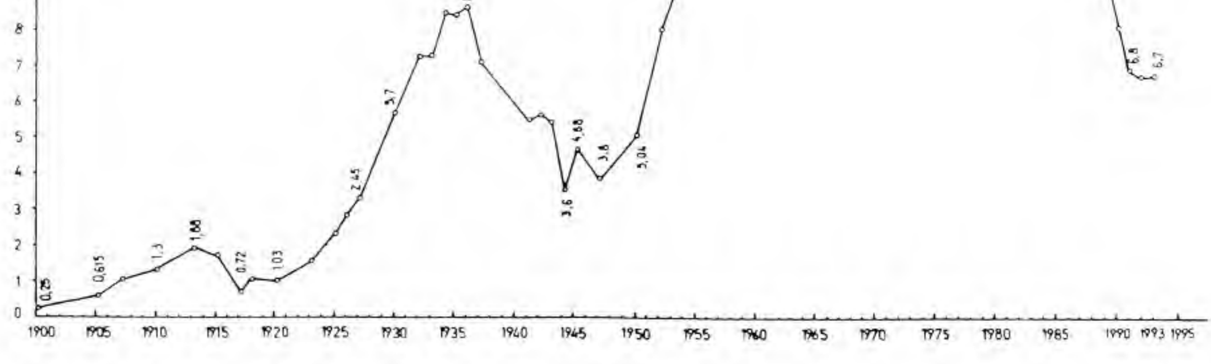

FIG. 1. Oil production of Romania during the period 1900-1993.

zone of the Carpathians, namely the Paleogene flysch and Neogene molasse basin of the Eastern Carpathians and the Paleogene and Neogene molasse basin of the Southern Carpathians (Getic Depression). Both basins were compressionally deformed during the successive Neogene Styrian (20-15.5 Ma), Moldavian (12-11 Ma) and Valachian (1.5-1 Ma) phases, giving rise to the development of a system of nappes and thrust sheets which form the external Moldavides. During these deformation phases, the flysch and molasse series were folded, faulted and thrusted in sequence over the foreland, formed by the Moldavian, Scythian and Moesian platforms (Fig. 2).

In the Eastern Carpathians, orogenic movements at the end of the Paleogene and the beginning of the Neogene (Older Styrian phase, 20-18 Ma) were accompanied by intensified uplift of the internal Moldavides and the development of a rapidly subsiding foreland basin (Sandulescu, 1988). During the Younger Styrian phase (15.5 Ma), coinciding with the beginning of the Badenian, the Tarcau and Marginal Folds nappes were emplaced; this was accompanied by the development of evaporitic conditions in the foredeep basin. During the Early Sarmatian Moldavian pulse, the entire package of Paleogene and Neogene nappes was underthrusted by the foreland platforms, which form the autochthon of the Subcarpathian, Marginal Folds and Tarcau nappes, both in the Eastern and Southern Carpathians.

Based on geophysical data and the results of deep wells, the autochthonous foreland extends a considerable distance beneath the external nappes of the Carpathians. Minimum figures are 20 km in the Moldova Valley, 30 km in the Bistrita and Trotus valleys, 15 km in the Prahova Valley and 10 km in the Olt Valley (Figs. 2 and 3). Beneath the external Carpathian nappes, the autochthonous foreland is dissected by a system of basin parallel, predominantly synthetic normal faults and transverse faults which were active during its early Sarmatian rapid subsidence (Dicea, 1967, 1995; Dicea and Tomescu, 1969). In the Eastern Carpathians, the main basin parallel, synthetic faults are the Campulung Moldovenesc, Solca and Siret faults; transverse faults generally coincide with the Bistrita, Trotus, Putna and Buzau valleys (Fig. 4).

Although these faults affected only the autochthonous foreland and its sedimentary cover, the structural relief generated by these faults influenced the architecture of the overlying nappes and folds. For instance, between Bistrita and Putna of Vrancea valleys, where the autochthonous platform is located at depths greater than 5500 meters, several subunits of the Marginal Folds nappe are defined by surface and subsurface data. Their axes can be followed over tens of kilometres, both at the surface and beneath the Tarcau nappe (Fig. 5). Based on this criterion, most of the oil and gas accumulations of the Eastern Carpathians were discovered. However, north of the Bistrita Valley, the autochthon rises to depths of 4000 to 3000 m. In this area, the high position of the platform blocked part of the Marginal Folds nappes west of the Gura Humorului-Bicaz threshold (Fig. 3). Correspondingly, the flysch formations of the Marginal Folds nappe were intensely deformed and in some areas, particularly north of the Cracau Valley, small and thin slices of the Marginal Folds nappe occur beneath and in front of the Tarcau nappe. Further north, the continuity of folds is difficult to follow from half-windows. North of the Moldova Valley, the Marginal Folds nappe was encountered only to the west of this foreland threshold as one or two slivers beneath and in front of the Tarcau nappe (Fig. 3). In the border area towards the Ukraine (Suceava Valley), the platform deepens again and several superimposed subunits are evident beneath and in front of the Tarcau nappe (Fig. 6; Gluschko and Kruglov,1971).

South of Slanic-Oituz and Vrancea half-windows (Fig. 7, section F), the Marginal Folds nappe is covered by the Tarcau nappe, the thickness of which varies between 2000 and 5000 meters, as indicated by well data. Between Slanic of Buzau and the Dambovita valleys, in the area where the Carpathian deformation front swings around into a western direction, Sarmatian-Pliocene molasse sediments cover the Paleogene flysch units up to the Cretaceous flysch nappes (Fig. 7, section I and Fig. 8). This series attains thicknesses of over 5000 m and thins eastward towards the foreland. Sarmatian-Pliocene series record the Valachian deformation phase which gave rise to the development of a series of hydrocarbon accumulations, structurally trapped in Paleogene strata of the Tarcau and Marginal Folds nappes (Fig. 7, section F).

Between the Dambovita and Danube rivers, the South-Carpathian foredeep is characterized by

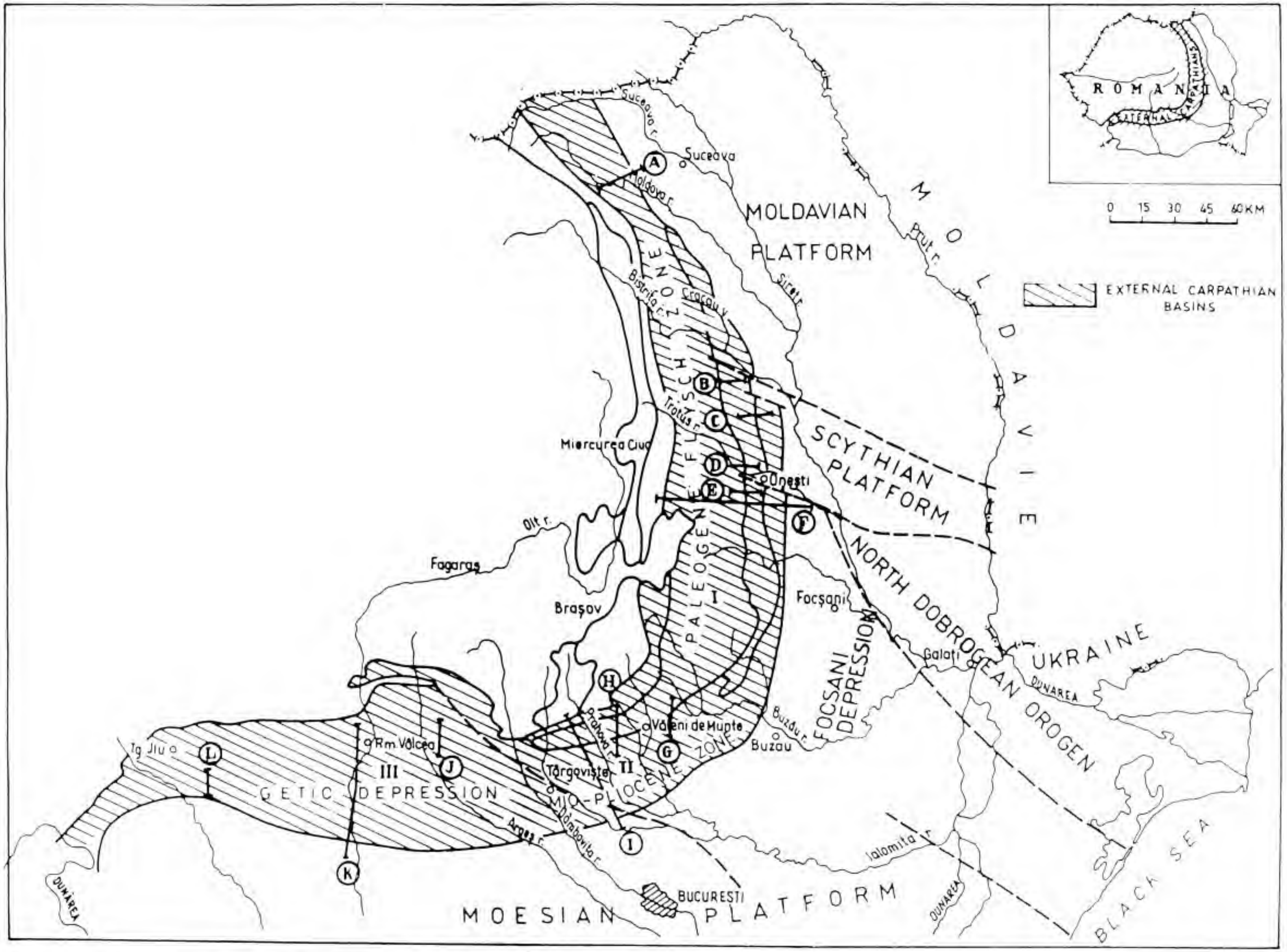

FIG. 2. Tectonic sketch map of Romanian External Carpathians (after Geological Institute of Romania), showing location of cross-sections A to L.

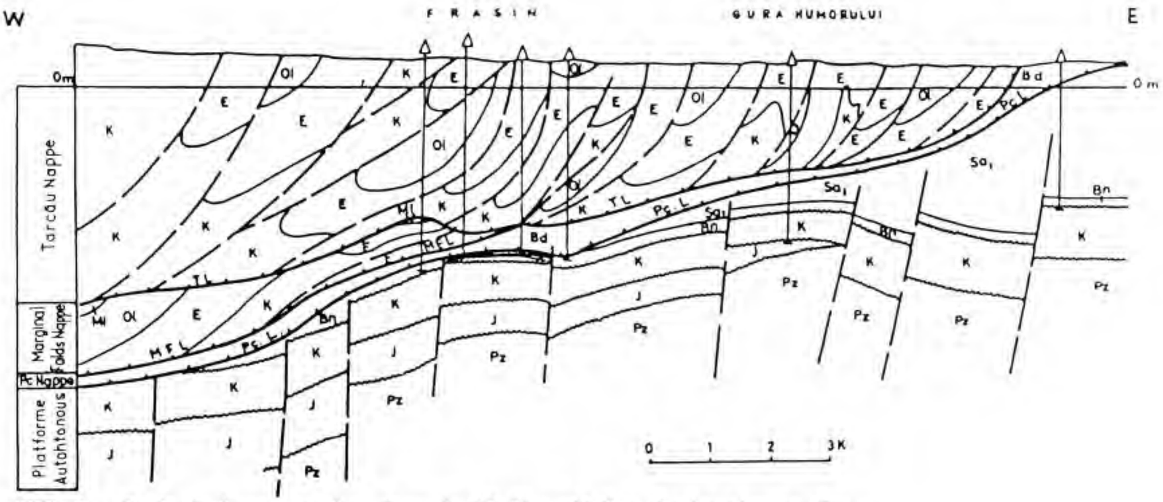

FIG. 3. Geological cross section along the Moldova Valley (for location see Fig 2, trace A). Pz - Palaeozoic, J - Jurassic, K - Cretaceous, E - Eocene, O - Oligocene, Mi - Miocene, Bd - Burdigalian, Bn - Badenian, Sa - Sarmatian, T.L. -Tarcau Line, M.F.L. - Marginal Folds Line, P.C.L. - Pericarpathian Line

the thick Paleogene and Neogene molasse deposits of the Getic Depression (Fig. 7, section K). Northward, Neogene series overstep the Cretaceous flysch nappes and the Mesozoic crystalline elements of the Southern Carpathians. Unlike in the Eastern Carpathians, the sedimentary fill of the Getic Depression is not involved in thin-skinned thrust sheets but in basement involving compressional and transpressional structures. The southern boundary of the Getic Depression is formed by the Pericarpathian fault, a major foreland verging upthrust (Fig. 7, section K and Fig. 19). The sedimentary series of the Getic Depression record the Older and Younger Styrian and the intra-Sarmatian Attic compressional phases; the Valachian phase was of minor importance in this area. On the external flank of the Getic Depression, Middle Sarmatian-Pliocene series overlap compressional structures in which Paleogene series are thrusted over Lower Sarmatian sediments (Fig. 7, section K). The area hosts a large number of hydrocarbon accumulations contained in structural traps.

PETROLEUM SYSTEMS

In the East-Carpathian Outer Moldavides, Paleogene and Neogene sediments attain thicknesses up to 5000 m. These consist predominantly of shaly and sandy series which contain multiple reservoir-seal pairs and major source rock intervals.

The most important source-rock of the East-Carpathian fold-and-thrust belt are the dyssodilic Rupelian-lower Burdigalian shales which have a total organic carbon content (TOC) ranging between 3.7 and 29.8%. In the domain of the Marginal Folds nappe, two main shale package occur within the Rupelian-lower Burdigalian interval (Fig. 9). The lower, Rupelian sequence is represented by the lower Menilite and lower Dyssodilic shales and their equivalents (shaly horizon of Pucioasa formation); these vary in thickness between 80 and 280 m. The upper, Chattian-lower Burdigalian interval consists of the upper Dyssodile shales, their equivalents (Vinetisu beds and Slon breccia) and the upper Menilites; it ranges in thickness between 50 and 100 m. These source-rock intervals are separated by the Kliwa Sandstone which was derived from the Carpathian foreland platform and presents an important reservoir. Both source-rock intervals, as well as the

Kliwa Sandstone, are also present in the Tarcau nappe. The latter contains additional reservoir-seal pairs in the Pliocene series (Fig. 9). Secondary source-rock intervals and reservoirs occur in the Eocene and Miocene sequences. In the Tarcau nappe, source-rock intervals are also present in the Cretaceous series and may have contributed to the accumulated oils (see Stefanescu and Baltes, this volume).

Reservoir parameters of the Kliwa Sandstone in the different parts of the external Carpathians and their foreland basins are summarized in Table 1. In the Marginal Folds nappe, to the north of Slanic Valley, the Kliwa Sandstones form a single, 20-170 m thick unit; a second objective horizon is formed by sandstones and conglomeratic beds occurring in a 60-120 m thick interval which straddles the Oligocene-Miocene boundary (Gura Soimului beds; Fig. 9). In the Carpathian Bend zone, the Kliwa Sandstones are only exposed in the Tarcau nappe where they are developed in two intervals. The Lower Kliwa horizon ranges in thickness between 100 and 150 m whereas the Upper Kliwa horizon attains thicknesses in the 200-300 m range (Fig. 9).

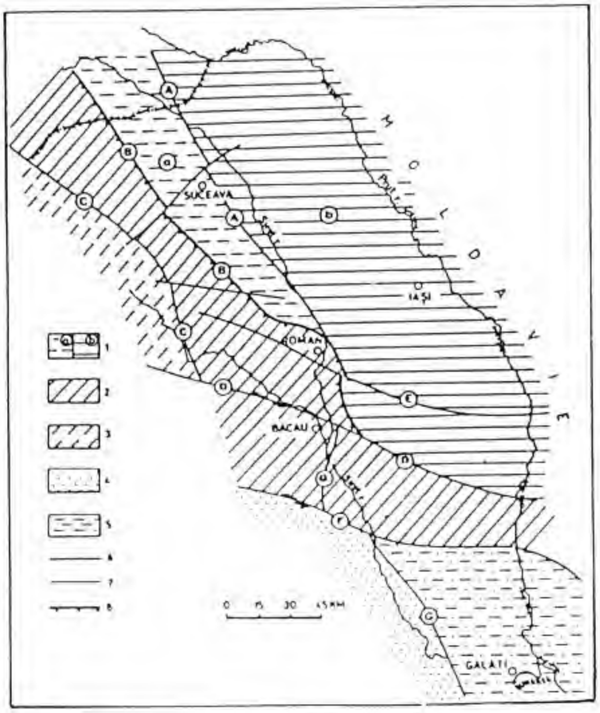

FIG. 4. Tectonic Sketch of East Carpathians Foreland (after Visarion and Sandulescu, 1981). A network of longitudinal and transverse faults having horizontal and vertical displacements, controls the architecture of flysch nappes. A-Siret Fault, B-Solca Fault, C-Campulung Moldovenesc-Bicaz Fault, D-Bistrita Fault, E-Vaslui Fault, F-Trotus Fault, G-Peceneaga-Camena Fault.

In the Getic Depression, source-rocks occur in the Late Cretaceous, Eocene, Oligocene and Sarmatian series (Fig. 9). The Oligocene series contains in its lower parts a 300-800 m thick sandstone and conglomerate sequence, referred to as "Horizon B", which forms an important reservoir; in some areas a predominantly shaly sand sequence, referred to as "Horizon A", is also developed. Facies analyses indicated that during the Oligocene, sands were shed into the Getic Depression mainly from the Southern Carpathians and, to a lesser degree, from the Moesian Platform (Fig. 10).

In the Carpathian Bend zone and in the Getic Depression, Oligocene, Miocene and Pliocene formations contain multiple reservoir-seal pairs. The facies development of individual reservoirs and sealing units was locally influenced by syndepositional tectonics. The Lower Burdigalian and Badenian salts provide the seal for many oil accumulations in the External Carpathians. Nevertheless, the Pliocene Meotian series is the most prolific objective in the Carpathians Bend zone and in the Getic Depression. Figure 11 provides a regional isopach map of the Meotian series on which sand-shale ratios are superimposed. The reservoirs properties of Meotian sands are summarized in Table 1.

In the External Carpathians, traps are mainly of the structural type and include anticlinal features, partly cut by thrust faults, and structures which are modified by the diapirism of salt. Stratigraphic pinch-out and unconformity traps play a subordinate role.

Tectonic unit	Stratigraphic Interval	Reservoir Rock	Thickness [m]	Porosity [%]	Permeability [mD]	Oil and Gas Fields
Tarcău Nappe	Eocene	Tarcău Sandstone	40 - 300	10 - 12	3 - 10	Păcurița, Văsiești, Tașbuga, Stîrmini
	Oligocene	Kliwa Sandstone	200	16	50	Dofteana - Bogata
Marginal Folds Nappe	Oligocene	Kliwa Sandstone	50 - 200	10 - 20	10 - 120	Geamăna, Tașbuga, Asău, Moinești
			150 -200	8	1 - 5	Ghelința
			350	25	400	Runcu - Buștenari
	Burdigalian	Sandstones and Conglomerates	20 - 250	7 - 18	2 - 120	Mihoc, Păcurița, Zemeș
Subcarpathian Nappe and Getic Depression	Eocene	Sandstones	45 - 80	15 - 21	8	Cosești
	Oligocene	Kliwa Sandstone	50 - 250	13	38	Tescani
		Sands and Sandstones	10 - 70	20 - 27	10 - 500	Săpunari, Vîlcele, Merișani
	Burdigalian	Microconglomerates and Sandstones	50 - 100	12 - 27	27 - 170	Tescani, Alunu, Tg. Jiu
		Sands and Sandstones	10 - 250	15 - 28	3 - 2000	Cîmpeni, Teiș, Băbeni, Țicleni
Foredeep and Getic Depression	Sarmatian	Sands and Sandstones	2 - 200	12 - 70	1 - 350	Ceptura,Boldești,Țicleni
	Meotian	Sands and Sandstones	2 - 200	20 - 40	24 - 500	Băicoi,Moreni,Boldești, Bucșani, Țicleni
	Dacian	Sands	10 - 40	20 - 37	18 - 2500	Boldești, Bucșani, Finta

TABLE 1. Reservoir Rocks in Eastern and Southern Carpathians

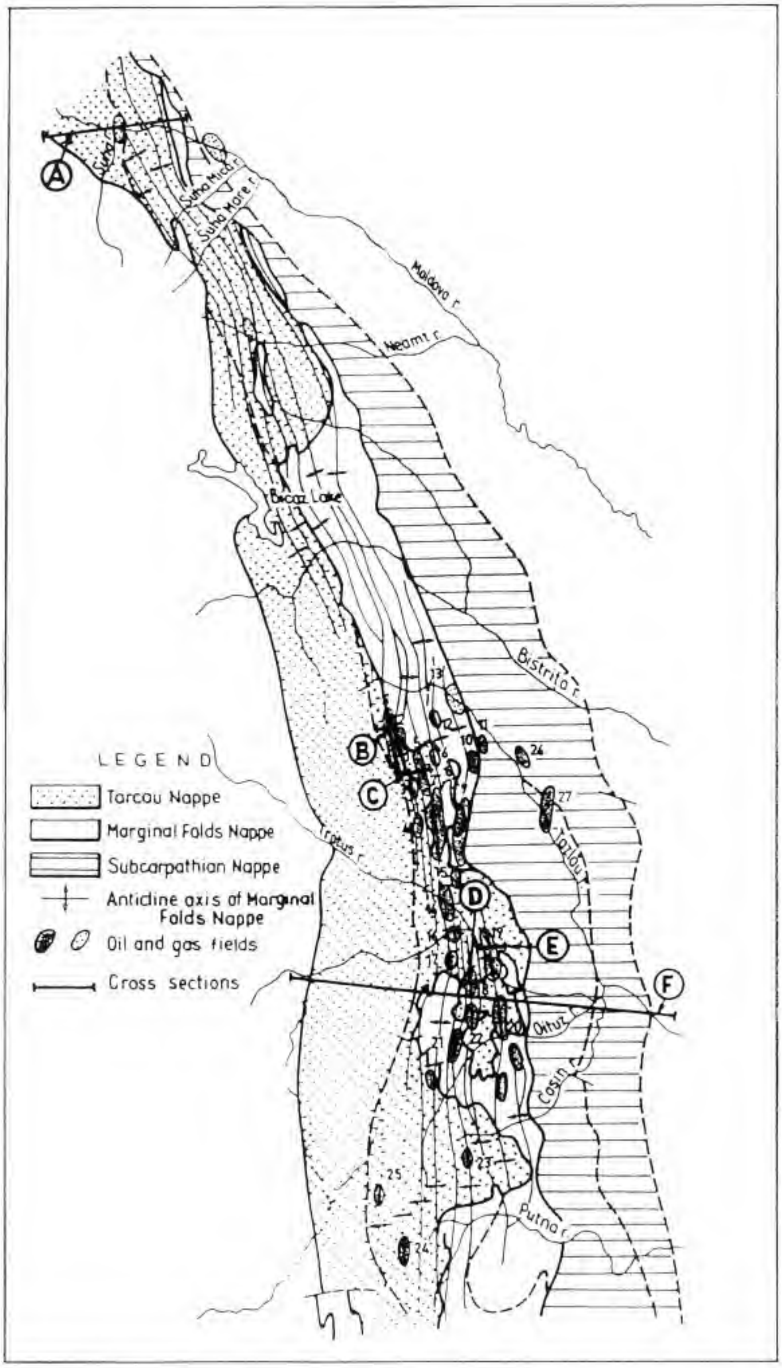

FIG. 5. Oil and gas fields from the Bistrita-Trotus Province. 1-Geamana, 2-Gropile lui Zaharache, 3-Chilii West, 4-Tasbuga, 5-Toporu-Chilii, 6-Arsita, 7-Zemes-Cilioaia, 8-Foale-Moinesti, 9-Uture-Moinesti oras, 10-Cucuieti, 11-Mihoc, 12-Frumoasa, 13-Tazlaul Mare, 14-Comanesti, 15-Vasiesti, 16-Darmanesti, 17-Doftenita, 18-Pacurita, 19-Dofteana-Bogata, 20-Slanic-Fierastrau, 21-Cerdac, 22-Slanic Bai, 23-Lepsa, 24-Ghelinta, 25-Ojdula, 26-Campeni, 27-Tescani.

MAIN PRODUCTIVE AREAS AND NEW PLAYS IN THE EXTERNAL CARPATHIANS

Main productive areas of the External Carpathians of Romania are the Paleogene Flysch Zone, the Mio-Pliocene Zone and the Getic Depression (Fig. 2).

Paleogene Flysch Zone

The Paleogene Flysch Zone comprises the Tarcau and Marginal Folds nappes of the Eastern Carpathians and the Carpathian Bend Zone (Fig. 7).

In the **Tarcau nappe**, conditions for generation, accumulation and preservation of hydrocarbons were not ideal. Only in the internal parts of this nappe, which were overridden by the Audia nappe, Oligocene source-rocks were buried to depths at which they entered the oil generation and partly even the gas generation window (Fig. 12a). Many potential anticlinal and thrusted anticline traps crop out and therefore have been destroyed by erosion. Only locally were oil accumulations discovered in the Tarcau nappe; these produce variably from the Oligocene Kliwa, the Oligocene-Lower Miocene Fusaru and the Eocene Tarcau sandstones. Generally, these accumulations are located above oil accumulations which produce from structures of the Marginal Folds nappe. Examples of such accumulations are the Zemes field in the Moldova region and the Geamana, Comanesti, Vasiesti, Pacurita and Dofteana-Bogata fields in the Tazlau-Oituz river area (Fig. 5).

Main hydrocarbon prospects of the Paleogene Flysch Zone are associated with the **Marginal Folds nappe** where it is covered by the Tarcau Nappe. In these areas, tectonic overburden provided for maturation of the Oligocene source-rocks (Fig. 12b). Most of these hydrocarbon accumulations are contained in massive or stacked reservoirs involved in thrusted folds and faulted anticlines; there are also examples of stratiform accumulations, sealed by faults or salt layers, and unconformity traps. Accumulations are concentrated along major structural axes which project northward and southward from half-windows in the Tarcau nappe.

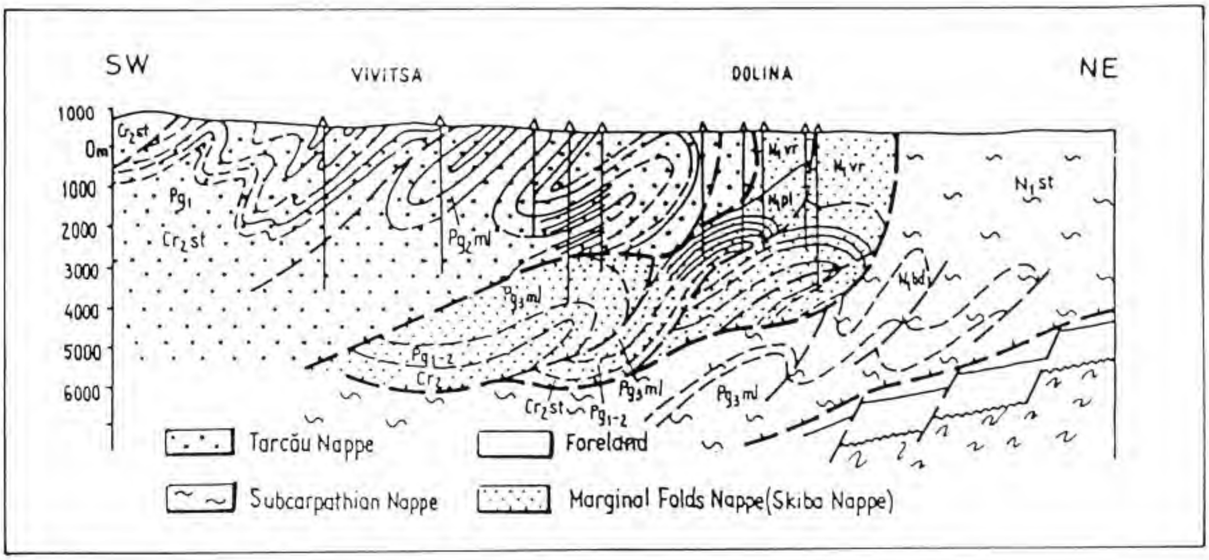

FIG. 6. Vitvitsa-Dolina geological cross-section, Ukrainian border area (after Gluschko and Kruglov, 1971). J_{2+3}-Middle-Upper Jurassic, Cr_2-Middle Cretaceous, Pg_1-early Paleogene, Pg_{1+2}-early-middle Paleogene, Pg_3ml- late Paleogene (Oligocene) Menilites, N_1pl-early Miocene Polianski Formation, N_1vr-early Neogene Vorotasce Formation, N_1db-early Neogene Dobrotov Formation, N_{st}-early Neogene Stebnik Formation. (for location see Fig. 2, trace A)

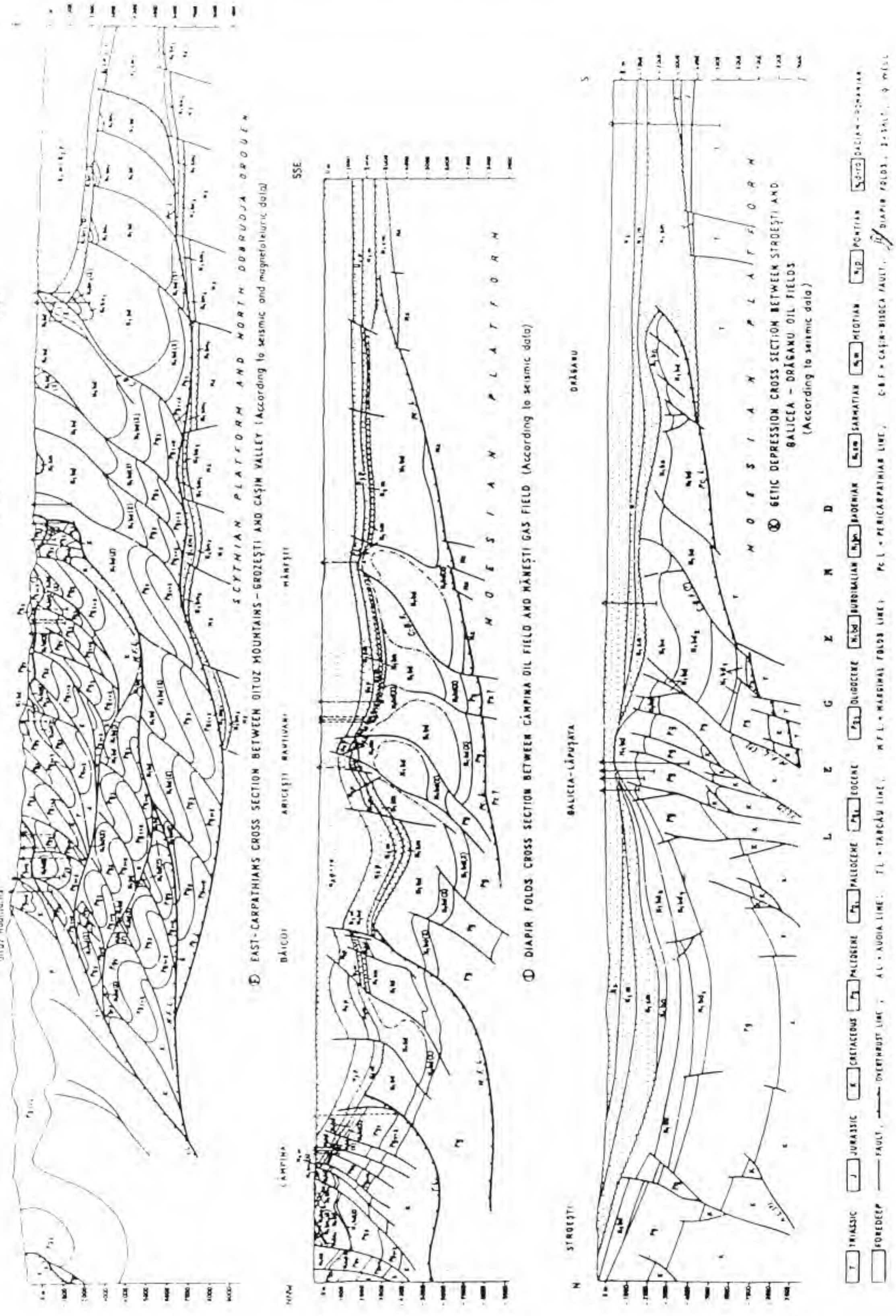

FIG. 7. Cross sections through External Carpathians (for locations see Fig. 2 traces F, I and K).

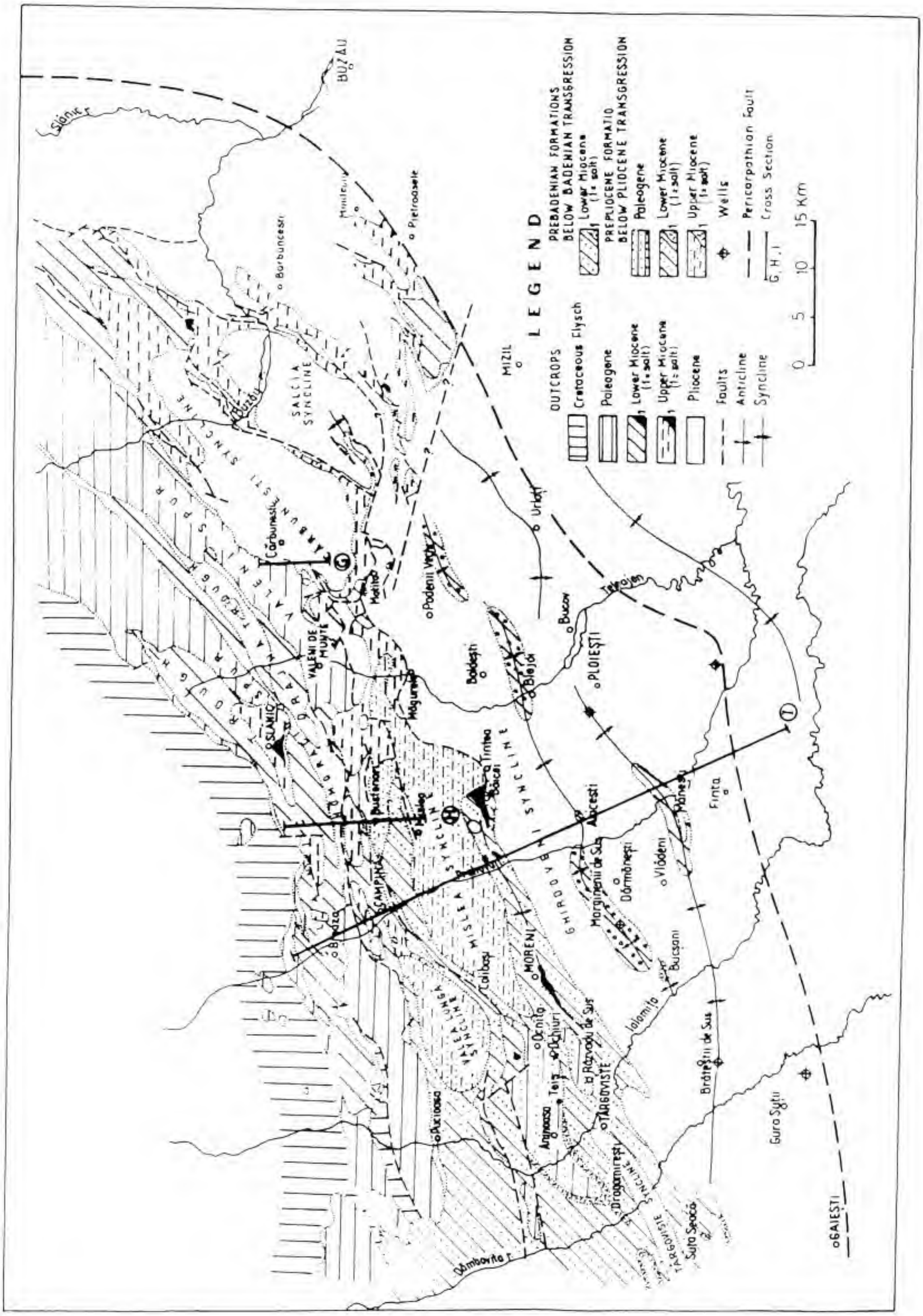

FIG. 8. Geological map of Mio-Pliocene Zone (from Romania lithostratigraphic map, after Patrut et al.,1973). G-I-Cross sections.

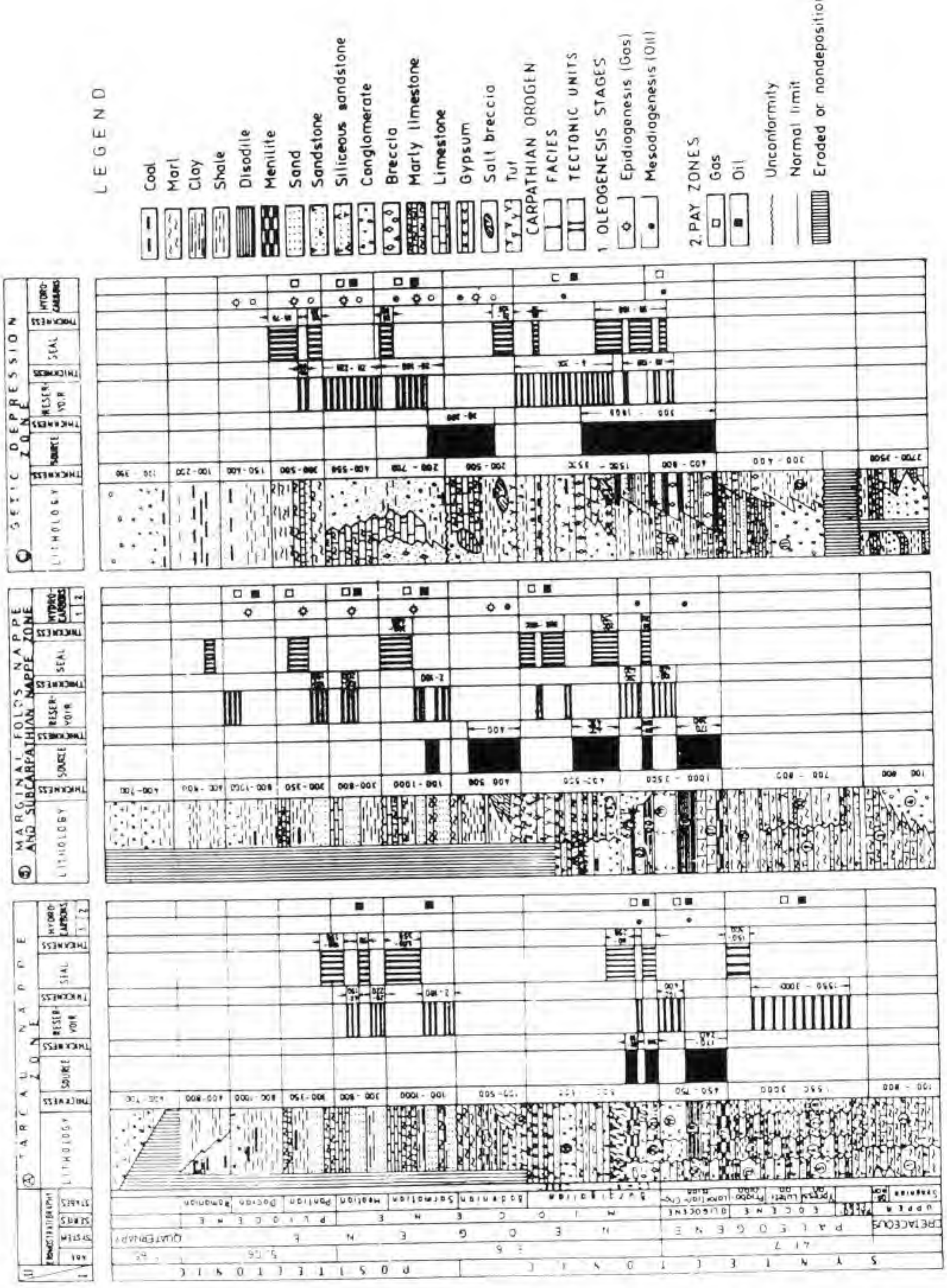

FIG. 9. Stratigraphic columns of Paleogene Flysch Zone, Mio-Pliocene Zone and Getic Depression. Cretaceous: 1-Horgazu beds, 2-Hangu beds, 3-Casin beds, 4-Piatra Streiului conglomerates; Paleogene-Miocene: 5-Tarcau sandstone, 6-Intermediate facies, 7-Colti facies, 8-Lesunt beds, 9-Gresu beds, 10-Buciasu beds, 11-Sacel conglomerates, 12-Sotrile beds, 13-Podu Secu beds, 14-Plopu beds, 15-Bisericani beds, 16-Fusaru-Krosno facies, 17-lower Menilites and white bituminous marls, 18-lower Disodiles, 19-lower Kliwa sandstone, 20-Vinetisu beds, 21-Podu Morii beds, 22-upper Kliwa sandstone, 23-Upper Disodiles and Menilites, 24-upper Disodiles, 25-Goru-Misina(Gura Soimului) beds, 26-Pucioasa-Fusaru beds; Miocene: 27-Harja beds, 28-Cornu beds, 29-Brebu conglomerates, 30-Slanic tuff.

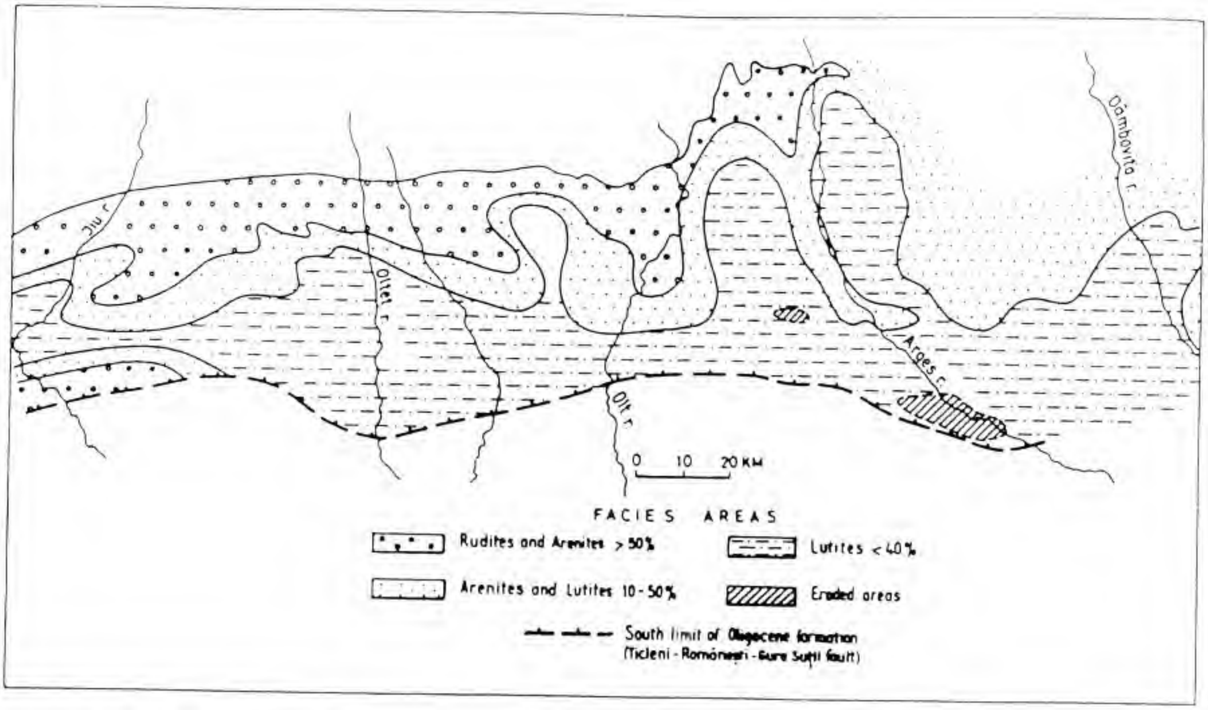

FIG. 10. Getic Depression. Lithostratigraphic map of Oligocene formations.

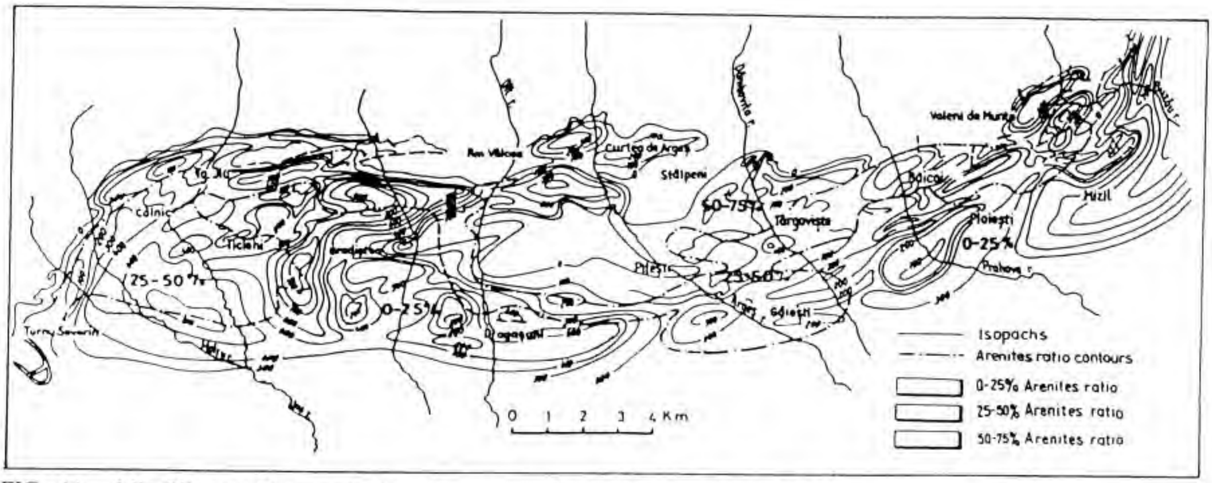

FIG. 11. Mio-Pliocene Zone and Getic Depression. Isopach map and sand/shale ratio of Meotian Formation. Sand/shale ratio in %.

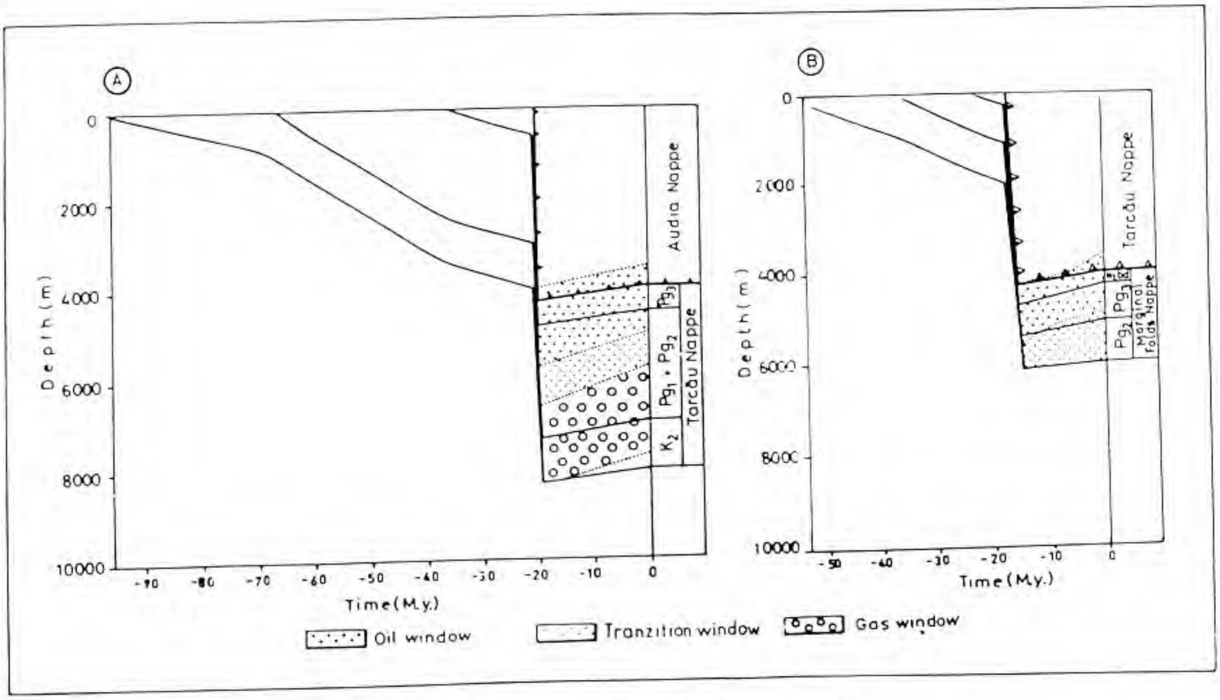

FIG. 12. Burial history and hydrocarbon kinetics diagram of Oligocene source-rocks from Tarcau (A) and Marginal Folds (B) nappes (kerogen type II-constant heat flow 50 mWm^{-2}).

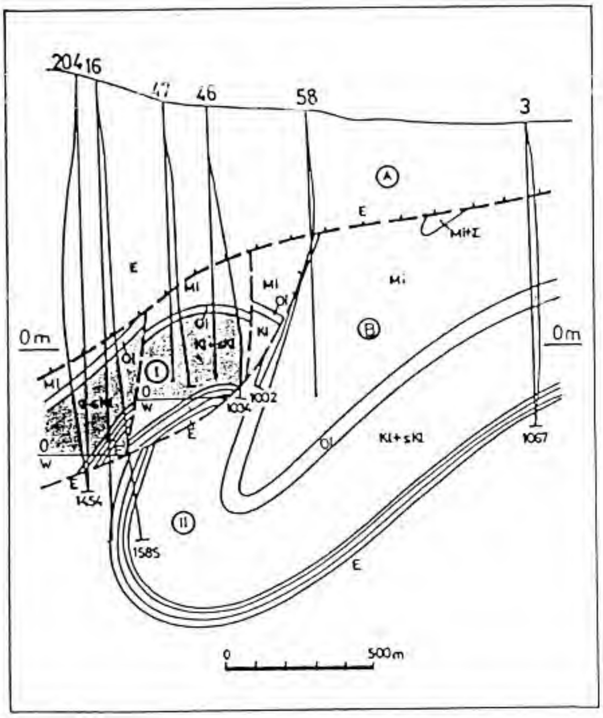

FIG. 13. Geamana oil field (for location see Figs. 2 and 5, trace B). A-Tarcau Nappe, B-Marginal Folds Nappe, I-First Scale Fold, II-Second Scale Fold, E-Eocene, O-Oligocene, Mi-Miocene, Kl-Kliwa Sandstone, SKl-Supra-Kliwa Formation (after Matei, 1973).

The sub-thrust continuation of these features was established by locating wells in the projection of established surface features and by limited reflection seismic data. The configuration of such sub-thrust Marginal Folds nappe structures varies from very steeply flanked anticlinal and thrusted features in the west to more gentle ones in the east. However, all features are characterized by a very complex internal configuration, as shown by the examples discussed below from the Tazlau-Oituz river area (Fig. 5).

The **Geamana oil field** (Fig. 13) is the westernmost productive structure of the Marginal Folds nappe occurring beneath the Tarcau nappe north of Trotus valley. In this field, the thickness of the latter ranges between 300 and 1000 m. The trap is formed by a complex faulted and thrusted fold, involving Eocene to early Miocene series. Production comes from the Oligocene Kliwa and Supra-Kliwa sandstones and the Gura Soimului beds. Additional pay sections occur in Eocene formations of the Tarcau nappe; these are in thrust contact with the Oligocene reservoirs of the underlying Marginal Folds nappe. The deepest wells were drilled to nearly 1600 m and tested two small thrust slices. Additional slices may occur at greater depths.

The **Zemes-Tazlau-Cilioaia oil field** (Fig. 14) is located at a few hundreds meters depth beneath a thin, complex zone of imbrications, attributed to the Tarcau and Marginal Folds nappes. Main accumulations are contained in two relatively gentle, though faulted and thrusted anticlinal structures. Producing intervals are Oligocene and Miocene sandstones. The sole-thrusts of tectonic slices covering this structure are only partly sealing, as indicated by the overspill of the western accumulation into the overlaying thrust slice. Pre-Oligocene objective have not yet been tested in the crestal parts of the trap-providing structures.

The **Pacurita oil field** (Fig. 15) was discovered at very shallow depths in the Tarcau, Tazlau and Marginal Folds nappes in an area where the Tarcau nappe is unconformably covered by Sarmatian sediments of the Comanesti basin. The Tazlau nappe is considered as a subunit of the Tarcau nappe. Production was obtained from Eocene reservoirs involved in the Tarcau nappe and from Oligocene reservoirs of the Tazlau and Marginal Folds nappes. In the Marginal Folds nappe, the

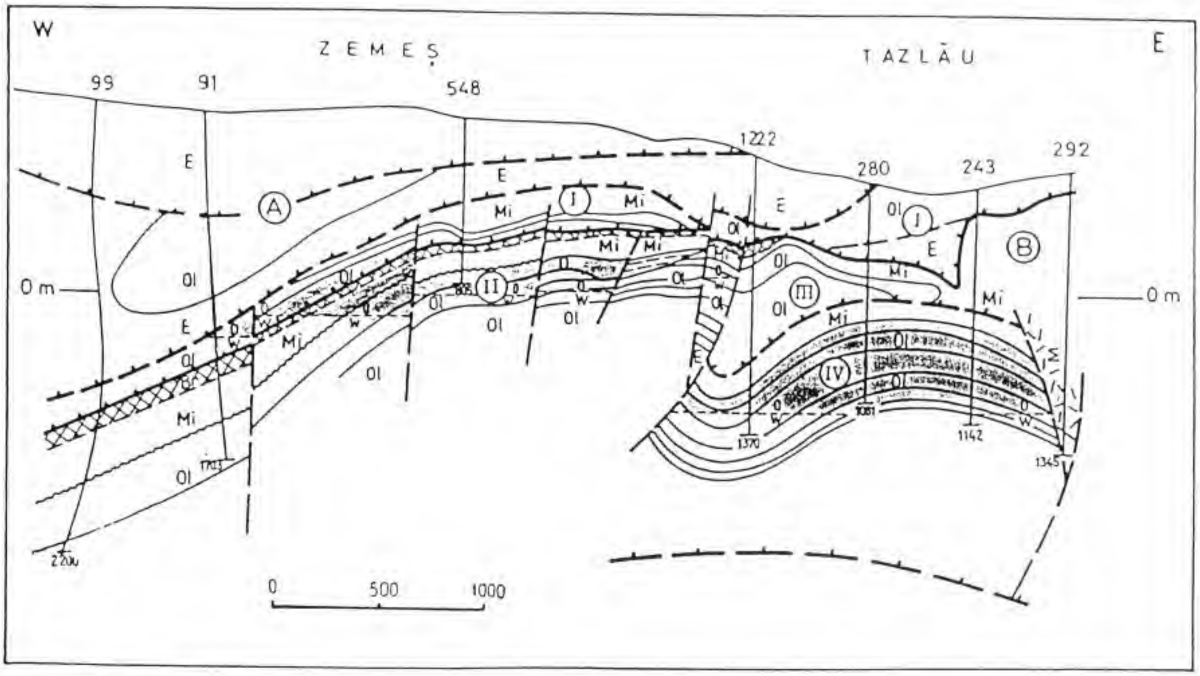

FIG. 14. Zemes-Tazlau-Cilioaia oil field, Marginal Folds Nappe (for location see Figs. 2 and 5, trace C). I-First Scale Nappe, II-Second Scale Nappe, III-Third Scale Nappe, IV-Fourth Scale Nappe (after Giurgiu et al., 1970).

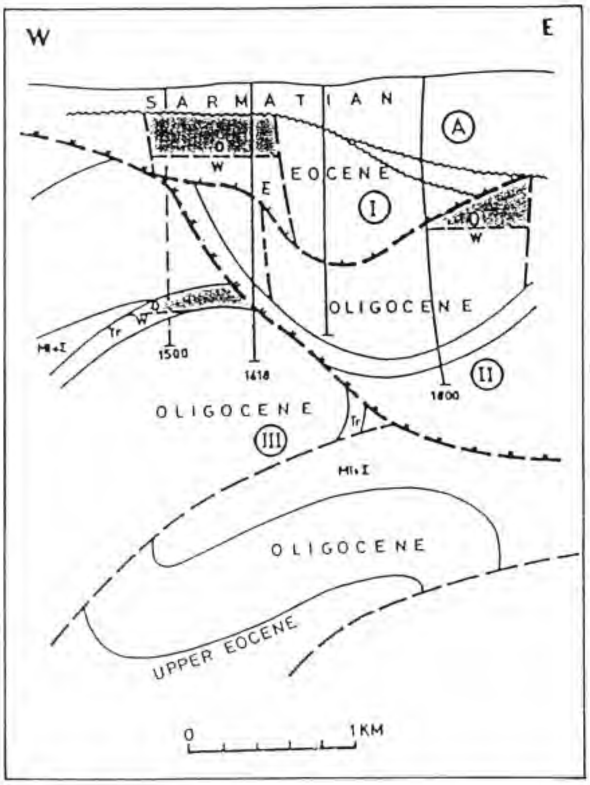

FIG. 15. Pacurita oil field (for location see Figs. 2 and 5, trace D). A-Comanesti post-tectonic Basin, I-Tarcau Nappe, II-Tazlau Subunit from Tarcau Nappe, III-Marginal Folds Nappe, O/W-oil-water contact (after Caminschi, 1973).

producing interval corresponds to the Gura Soimului beds; deeper objectives have not yet been tested. Deep seated imbrications of the Marginal Folds nappe are anticipated and may provide further prospects in this already productive, tectonically very complex area.

In the **Dofteana-Bogata oil field** (Fig. 16), wells spudded in the Sarmatian Comanesti Depression and the Tazlau unit of the Tarcau nappe, penetrated three thrust slices of the marginal units of the Tazlau nappe. The Marginal Folds nappe was not reached at depths of about 2200 m. Oil was discovered in Miocene and Oligocene reservoirs of the Tazlau nappe and its marginal thrust slices.

The southernmost discovery in the Bistrita-Trotus province of the East-Carpathian Flysch Zone is the **Ghelinta oil and gas field** (Fig. 5). It is located about 25 km to the west of the Tarcau nappe front and produces at a depth of 2100-2200 m from Kliwa sands involved in the Marginal Folds nappe. Discovery of this field proves that the thickness of the Tarcau nappe is variable and not everywhere prohibitive.

The definition of new prospects in the Marginal Folds nappe, both in a subthrust position beneath the Tarcau nappe and in tectonic half-windows of the latter, requires detailed reflection-seismic control. However, data acquisition is often hampered by a rugged relief. Nevertheless, results of previous exploration activity shows that hydrocarbon supply and reservoir risks are rather low for Marginal Folds nappe prospects. In parts of the northern East-Carpathians, the Marginal Folds nappe is poorly developed beneath the Tarcau nappe, rests behind the foreland threshold and is highly tectonized (Fig. 3); in this areas foreland structures offer the primary prospects.

Mio-Pliocene Zone

The Mio-Pliocene Zone comprises the southern and southwestern parts of the Subcarpathian, Marginal Folds and Tarcau nappes (Fig. 8). In this area, thick late syn- and in part post-orogenic molasse deposits cover the foreland, the Subcarpathian and the more internal nappes (Fig. 7, section I). The Subcarpathian nappe was emplaced during Sarmatian-Badenian times and was reactivated during the Pliocene Valachian phase (Dicea, 1995).

In the Suceava-Slanic valleys sector of the Subcarpathian nappe only few oil and gas accumulations were discovered. In this area, the Subcarpathian molasse nappe is superimposed on foreland formations and only few structurally closed traps could be established. Correspondingly, traps involving the foreland series play a more important role.

However, in the area delimited to the east by the Slanic of Buzau Valley and to the west by the Dambovita Valley, the Mio-Pliocene Zone hosts the most prolific hydrocarbon province of Romania (Fig. 7, section I and Fig. 8). Here, Oligocene and Miocene source-rocks were deposited in a continuously and strongly subsiding basin, characterized

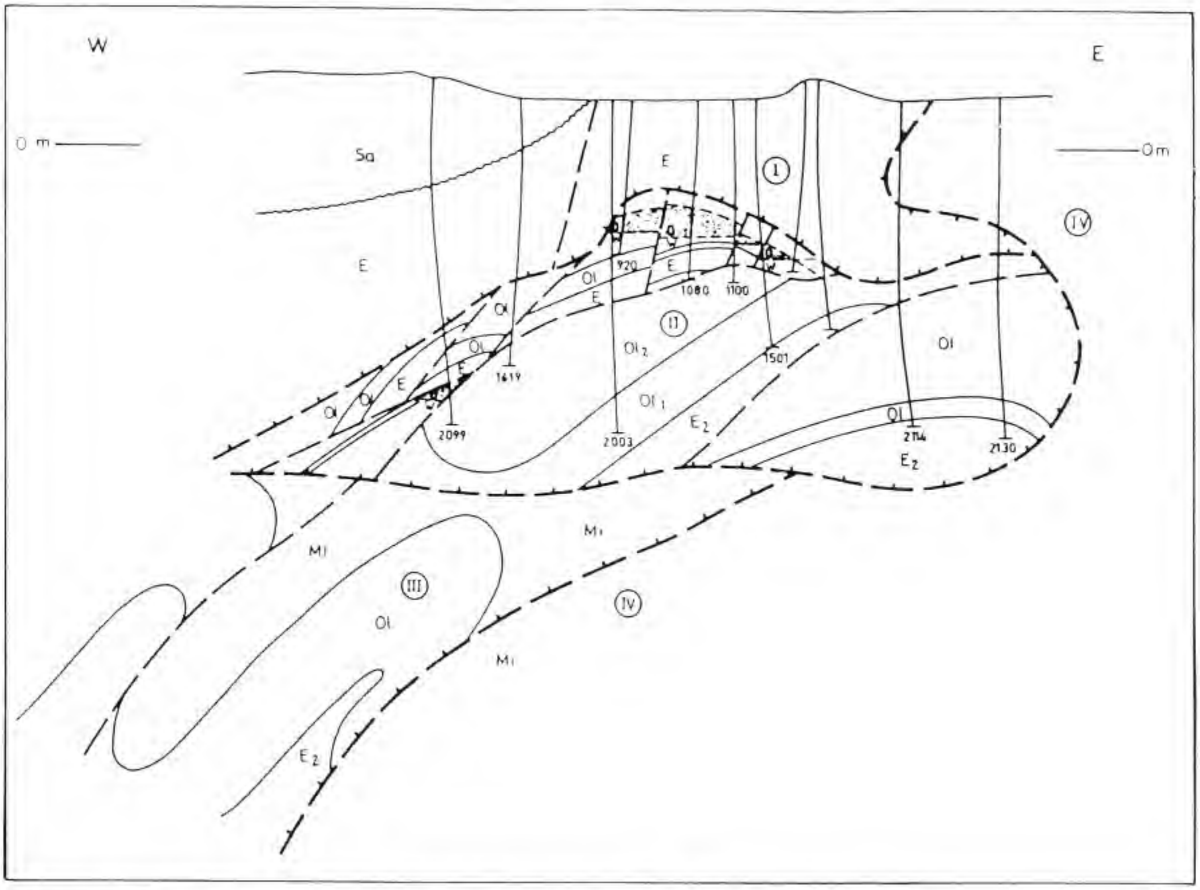

FIG. 16. Dofteana-Bogata oil field (for location see Figs. 2 and 5, trace E). I-Tazlau Subunit from Tarcau Nappe, II-Marginal Subunit from Tarcau Nappe., III-Marginal Folds Nappe, IV-Subcarpathian nappe (after Caminschi, 1973).

by a normal thermal gradient. In the structuration of this area, halokinetic mobilization of Burdigalian salt played an essential role during the accumulation of upper Burdigalian and Pliocene series. Involvement of the entire Paleogene flysch and Neogene molasse series in the intra-Pliocene Valachian compressional deformations have contributed to the development of most of the structural traps, some of which are cored by salt diapirs.

Main productive intervals occur in the Oligocene, lower and upper Miocene and Pliocene series (Meotian, Dacian and Levantin formations, Fig. 9). The Oligocene-lower Burdigalian Kliwa Sandstone is quartzous whereas the younger Miocene and Pliocene sands are calcareous. Oligocene productive horizons vary in thickness between 2 and 60 m (Bustenari field); Miocene sands range in thickness between 10 and 90 m (Teis field) and Sarmatian sands between 50 and 100 m (Boldesti field) (Table 1). Meotian sands are the main producer in the Carpathian Bend area. The Meotian formation contains productive complexes which change along strike in thickness from 125 m in the Barbuncesti field to 80 m in the Boldesti and to 34 m in Bucsani field and, in a dip direction, from 25 m in the Ocnita field to 40 m in the Moreni and to 50 m in the Finta field.

Involvement of the Mio-Pliocene molasse sequences in the Valachian folding phase is also responsible for the development of structural traps involving Oligocene and lower Miocene formations of the Subcarpathian, Marginal Folds and Tarcau nappes. The Tarcau nappe is involved in the Bustenari-Runcu anticlinal trend. In the Baicoi and Moreni structures, Oligocene series, forming probably a part of the Marginal Folds nappe, were intercepted. The front of the Paleogene Peri-

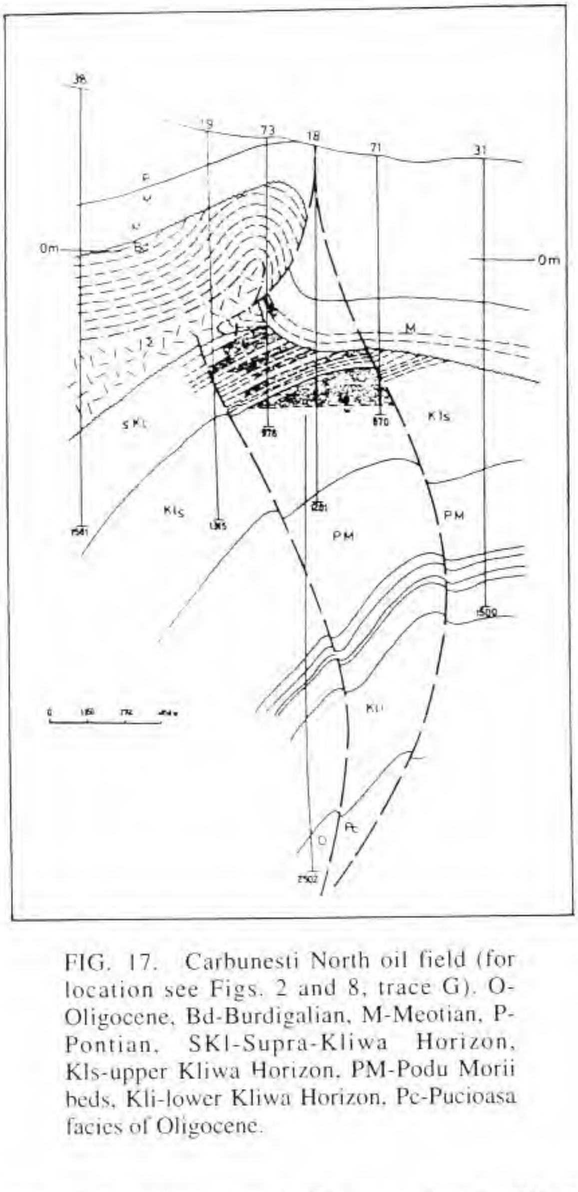

FIG. 17. Carbunesti North oil field (for location see Figs. 2 and 8, trace G). O-Oligocene, Bd-Burdigalian, M-Meotian, P-Pontian, SKl-Supra-Kliwa Horizon, Kls-upper Kliwa Horizon, PM-Podu Morii beds, Kli-lower Kliwa Horizon, Pc-Pucioasa facies of Oligocene.

carpathian nappe is probably associated with the Bucsani-Aricesti-Pietroasele-Monteoru alignment (Fig. 8; Dicea, 1995).

Oil and gas accumulations, reservoired in Oligocene and Miocene sands, are contained in structural traps, such a thrusted anticlinal features (Runcu, Gura Ocnitei fields), diapiric folds (Baicoi, Moreni, Bucsani fields) and faulted and unfaulted anticlines (e.g. Boldesti, Margineni, Podeni fields), as well as in stratigraphic traps associated with the basal transgressive surface of the Meotian formation (e.g. Carbunesti, Runcu-Bustenari, Campina, Margineni fields). This shows that these hydrocarbon accumulations have formed only after the Valachian deformation phase, that is, during the late Pliocene and Pleistocene.

Well data from the Carbunesti oil field (Fig. 17), which produces from Oligocene, Burdigalian and Meotian sands, give evidence for the two-phase development of this structure. The basal Meotian unconformity truncates Oligocene and Miocene series and was itself deformed during the Valachian compressional phase. The trap is provided by a folded and faulted unconformity surface and the presence of thick Burdigalian salt on the western flank of the structure.

The **Bustenari-Runcu oil field** (Fig. 18) is contained in complex imbrications of the Tarcau nappe, involving Oligocene and Burdigalian strata, truncated by the basal Meotian unconformity, which in turn was folded and faulted during the Valachian deformation phase. At shallow levels, production comes from Meotian, Burdigalian and Oligocene sands. A deep seated imbrication, involving Oligocene reservoirs, is also productive.

In the central part of Mio-Pliocene Zone, the most important structural trend is formed by the well-known diapiric Tintea-Baicoi-Moreni trend (Fig. 8). This structure, which is limited to the north and south by two large synclinal trends (Fig. 7, section I), contains the largest reserves of the entire Eastern Carpathians. Deep wells drilled during the last years on the Baicoi (7025 m), Moreni (5500 m) and Runcu (3600 m) structures proved the presence of Oligo-Miocene objectives and oil shows at deep and ultra-deep levels. However, poor reflection-seismic resolution of structural prospects at these depths has so far hampered exploration efforts (Dicea, 1995).

New targets in the Mio-Pliocene Zone are deep structural traps involving Oligocene and Miocene reservoirs. Reservoir development and hydrocarbon charge appear to be assured; the main risk lays in the reflection-seismic definition of drillable structures. In addition, there is a potential for stratigraphically trapped pools along the flanks of the already drilled up shallow structures.

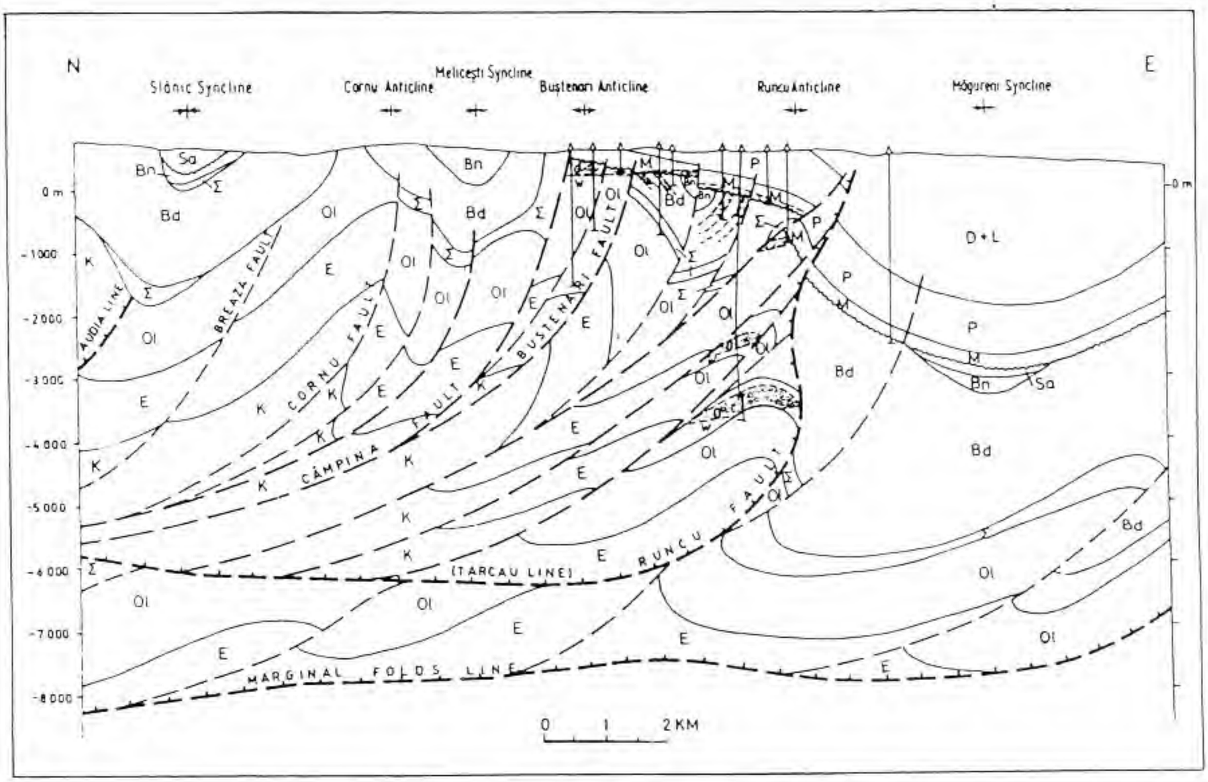

FIG. 18. Bustenari-Runcu oil field and deep prospects (for location see Figs. 2 and 8, trace H). K-Cretaceous, E-Eocene, O-Oligocene, Bd-Burdigalian, Bn-Badenian, Sa-Sarmatian, M-Meotian, P-Pontian, D+L-Dacian+Levantin, S-salt (modified after Albu et al.,1982).

Getic Depression

The Getic Depression corresponds to the South Carpathian foreland basin which is filled by Eocene to Pliocene molasse-type series, deposited on Mesozoic carbonates and Palaeozoic series of the Moesian Platform. The area was affected by the Older and Younger Styrian and the intra-Sarmatian Attic deformation phases. Along the Pericarpathian fault, which delimits the deformed area to the south, Lower and Middle Miocene strata are overthrusting Lower Sarmatian series. This fault is sealed by the onlapping and overstepping Middle Sarmatian to Pliocene molasse sequences (Fig. 7, section K).

The stratigraphic column of the Getic Depression is given in Figure 9. Eocene and Oligocene strata attain thicknesses of up to 5000 m in the northern parts of the Getic Depression and onlap southward the top-Cretaceous unconformity. Eocene calcareous sandstones and conglomerates grade upwards into a sandy marly section. Lower Oligocene sandstones ("Horizon-B") are followed by 300-800 m thick marls and shales, containing sand lenses ("Horizon-A"). The facies distribution of Oligocene series is summarized in Figure 10. Neogene strata reach thicknesses in the order of 2000 to 3000 m; they contain major lower Burdigalian and middle Badenian salt intercalations. During the Miocene the southern parts of the Getic Depression were overstepped.

The distribution of oil and gas accumulations in the Getic Depression is summarized in Figure 19. In the northern parts of the depression, where thick Paleogene sediments are present, Oligocene source rocks have entered under normal geothermal gradients in the oil window at depths of 3500-4500 m and have at present reached peak maturity (Fig. 20). The generated hydrocarbons migrated updip to the south and charged structural

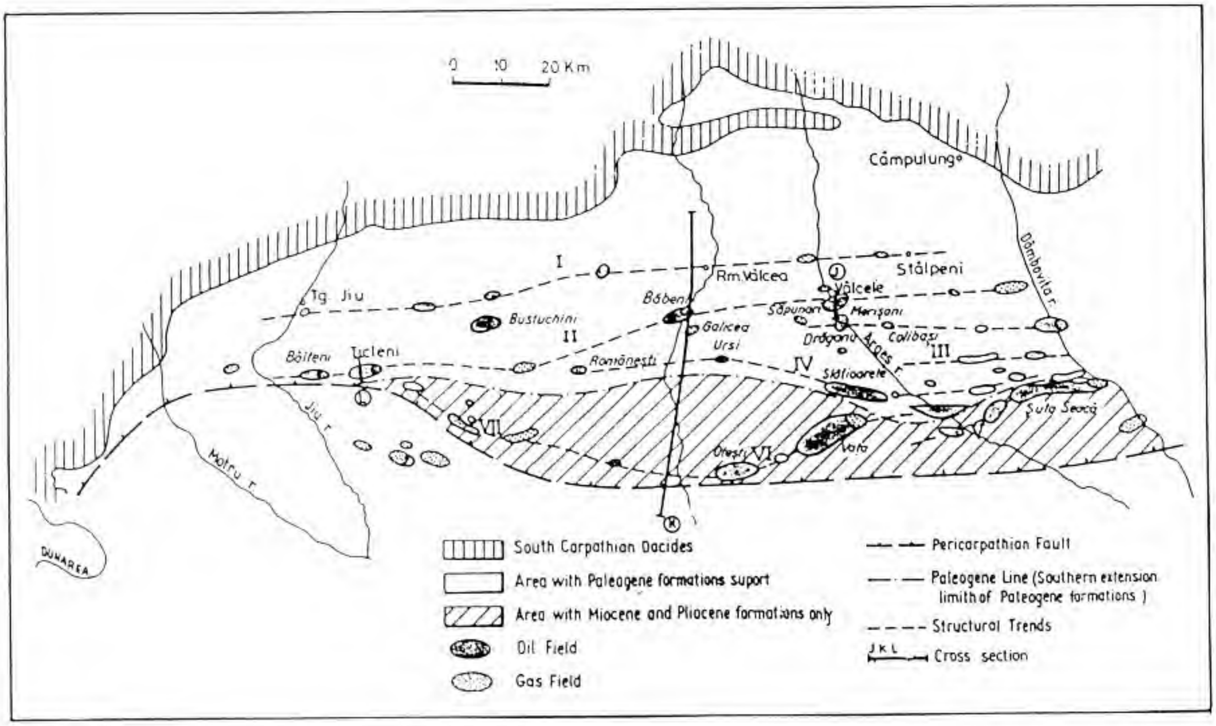

FIG. 19. Tectonic sketch and hydrocarbon pool alignments of Getic Depression.

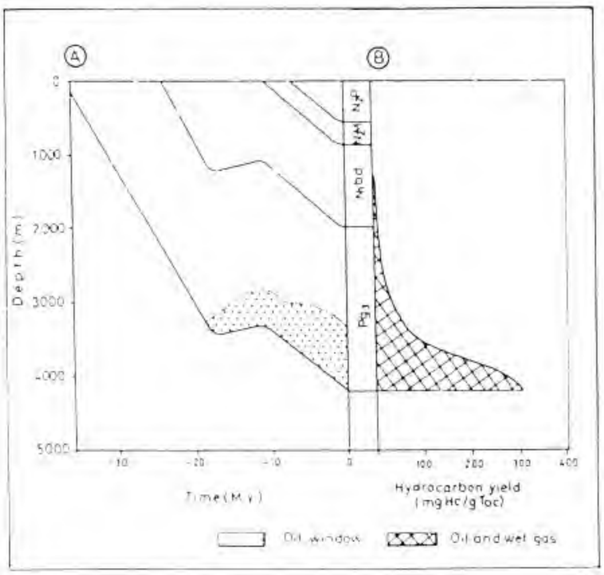

FIG. 20. Getic Depression. Burial history, hydrocarbon kinetics and yield diagram for Oligocene source rocks (kerogen type II, constant heat flow 46 mWm^{-2}).

and stratigraphic traps. Oligocene reservoirs are productive in a number of structures (e.g. Valcele, Merisani, Draganu, Sapunari). In Burdigalian to Sarmatian times, these structures developed during multiple deformation phases and were partly modified by Pliocene faulting. Fields are associated with the four distinct structural trends, designated in Figure 19 with the Roman numerals I to IV. In these fields production comes, apart from Oligocene sands, also from Burdigalian, Sarmatian and Meotian sands. Examples are the Valcele and Ticleni fields, both of which are located on trend II, well to the north of the Paleogene onlap edge.

The **Valcele oil field** (Fig. 21) produces from the Oligocene "Horizon-A" fan sands and from Burdigalian reservoirs. Structurally and stratigraphically trapped accumulations are associated with the basal Burdigalian and basal Badenian unconformities.

The **Ticleni oil and gas field** (Fig. 22) is contained in an anticlinal structure which grew during Sarmatian time and was modified by Pliocene faulting. Oil and gas pools are hosted in upper Burdigalian sands and Sarmatian and Meotian sand pinch-out traps.

In the external parts of the Getic Depression, which are situated to the south from the Paleogene onlap edge, Burdigalian series attain thicknesses of over 1000 m and contain sands with good reservoir properties; however, these reservoirs lack a hydrocarbon charge. In this area, Badenian, Sarmatian and Pliocene series attain thicknesses of over 4000 m south of the Pericarpathian fault and contain gas-prone source rocks intervals (Fig. 7, section K). These source rocks charged a number of gas fields on the structural trends V-VII (Fig. 19). Oil charge to these fields (e.g. Otesti), which are reservoired in Sarmatian, Meotian and Pontian sands, is presumably related to longer range migration from the northern parts of the Getic Depression.

CONCLUSIONS

The main hydrocarbon producing areas of the Romanian External Carpathians are the Bistrita-Trotus province and Carpathian Bend province of East-Carpathians and the Getic Depression of the South-Carpathians. All three areas are characterized by a deeply subsided autochthonous foreland which was thrusted under the Carpathian orogen.

In the Bistrita-Trotus province, the majority of fields produce from Paleogene flysch of the Marginal Folds nappe, involved in complex structures beneath the Tarcau nappe, generally in the vicinity of half-windows in the latter. Established fields are generally located at shallow depths. The potential of deeper seated structures is still poorly evaluated and requires the recording of extensive reflection-seismic surveys in a topographically difficult terrain. Oligocene-Early Miocene oil-prone source-rocks are well developed and provide abundant hydrocarbon charge to closely associated reservoirs; there are insufficient geochemical data to determine whether there is also a contribution from Cretaceous source-rocks.

In the prolific Carpathian Bend Zone, in which Mio-Pliocene molasse-type series are well developed and cover the Tarcau, Marginal Folds and Subcarpathian nappes, ample hydrocarbon charge is provided by Oligocene-Early Miocene and possibly Cretaceous source-rocks. Multiple reservoir-seal pairs are developed in Oligocene to Pliocene shallow marine deltaic series. Structural traps are associated with all nappe units. Unconformities, related to the different compressional phases, provide for additional traps. Most fields are contained in relatively shallow structures which attained their present configuration during the terminal Pliocene deformation phase. Deep wells indicate a good reservoir development at depth. Limited reflection-seismic data suggest the presence of deep-seated prospects requiring definition by extensive surveys.

The oil accumulations contained in Oligocene and Miocene reservoirs of the northern parts of the Getic Depression are closely related to the distribution of the Paleogene series containing mature source-rocks. The oil and gas accumulations of the southern parts of the Getic Depression are charged

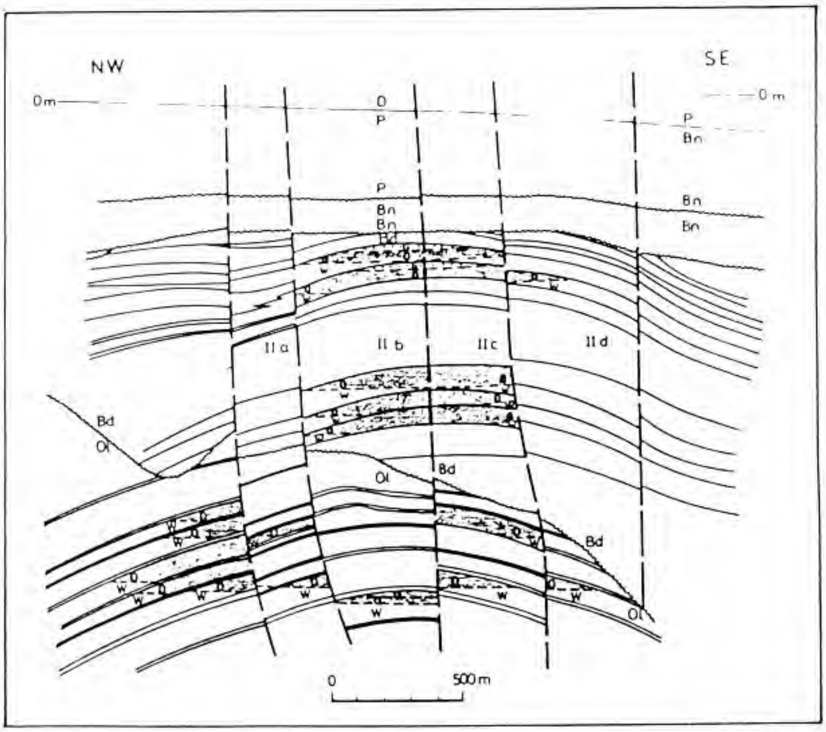

FIG. 21. Valcele oil field.(for location see Figs. 2 and 19, trace J) O-Oligocene, Bd-Burdigalian, Bn-Badenian, P-Pontian, D-Dacian, O/W-oil-water contact (after Popa, from Paraschiv, 1975).

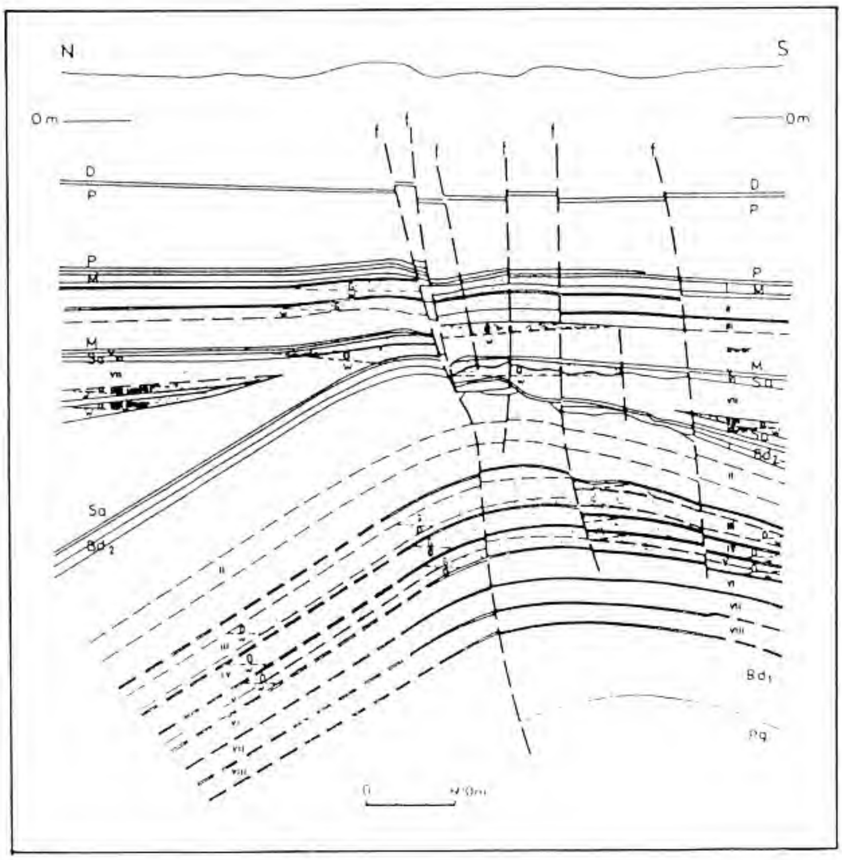

FIG. 22. Ticleni oil and gas field (for location see Figs. 2 and 19, trace L). Pg-Paleogene, Bd1-lower Burdigalian, Bd2-upper Burdigalian, Sa-Sarmatian, M-Meotian, P-Pontian, D-Dacian, I-VIII-producing horizons, F-Faults (after Ioachimciuc, 1970).

by gas-prone Mio-Pliocene source-rocks and are reservoired in late Miocene and Pliocene sands, involved in structural and combined stratigraphic/structural traps which developed during Mio-Pliocene deformation phases. In this area, the potential of deep seated prospects also requires further evaluation on the basis of extensive reflection-seismic surveys.

In the northern parts of the East-Carpathians, where the autochthonous foreland has not subsided to excessive depths, the Mesozoic sedimentary cover of the Moldavian Platform may host viable prospects. The deep Neogene Peri-Carpathian Focsani Depression hosts a number of gas accumulations. Similar accumulations may occur along the external flank of the entire Carpathian foredeep.

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Do hydrocarbon prospects still exist in the East-Carpathian Cretaceous flysch nappes ?

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ABSTRACT

Although Cretaceous strata are involved in most of the East-Carpathian nappes, only the Ceahlau, Bobu, Teleajen, Macla and Audia nappes are composed mainly of Cretaceous flysch. Seeps occurring in the Teleajen nappe and shows obtained from the Macla nappe suggest the existence of a petroleum system which is related to the Cretaceous flysch.

Effective Barremian-Aptian source-rocks are present in Teleajen nappe and Cenomanian-Turonian source-rocks occur in Macla nappe while potential, mostly Albian source-rocks are largely developed in the Bobu and Teleajen nappes.

Reservoir rocks are present in all Cretaceous flysch nappes. However, as these reservoirs generally occur in very high structural positions, they are deeply eroded, except in the Bobu and Teleajen nappes where they can be structurally trapped beneath more internal thrust sheets. Lateral facies changes may provide for the stratigraphic traps in these tectonic units.

We conclude that hydrocarbon prospects still exist in the subthrust parts of the Teleajen and Bobu nappes.

INTRODUCTION

Romania is located in area where the Carpathian Alpine structure draws a large loop and is thrust a considerable distance over the autochthonous foreland. Prior to attaining its present structural configuration (Fig. 1), the on-shore areas of Romania underwent a long and complex geological evolution during which a large number of sedimentary basins developed, both under extensional and compressional conditions. In most of these basins source- and/or reservoir-rocks were deposited. Unfortunately, only part of these sedimentary basins evolved into petroleum systems which today produce hydrocarbons.

Although none of the proven petroleum systems of Romania can be clearly related to Cretaceous source-rocks, a few fields located on the Moesian Platform are thought to be possibly charged by hydrocarbons generated by Cretaceous source-rocks (Paraschiv, 1979). Similarly, oil fields of the East-Carpathians, producing from Tertiary reservoirs (Dicea, this volume), may have been partly charged with hydrocarbons generated by Cretaceous source-rocks (Roure et al., 1993; see

Stefanescu, M. & Baltes, N., 1996. — Do hydrocarbon prospects still exist in the East-Carpathian Cretaceous flysch nappes ? *In*: Ziegler, P. A. & Horváth, F. (eds), Peri-Tethys Memoir 2: Structure and Prospects of Alpine Basins and Forelands. *Mém. Mus. natn. Hist. nat.*, **170**: 427-438 + Enclosure 1. Paris ISBN: 2-85653-507-0.
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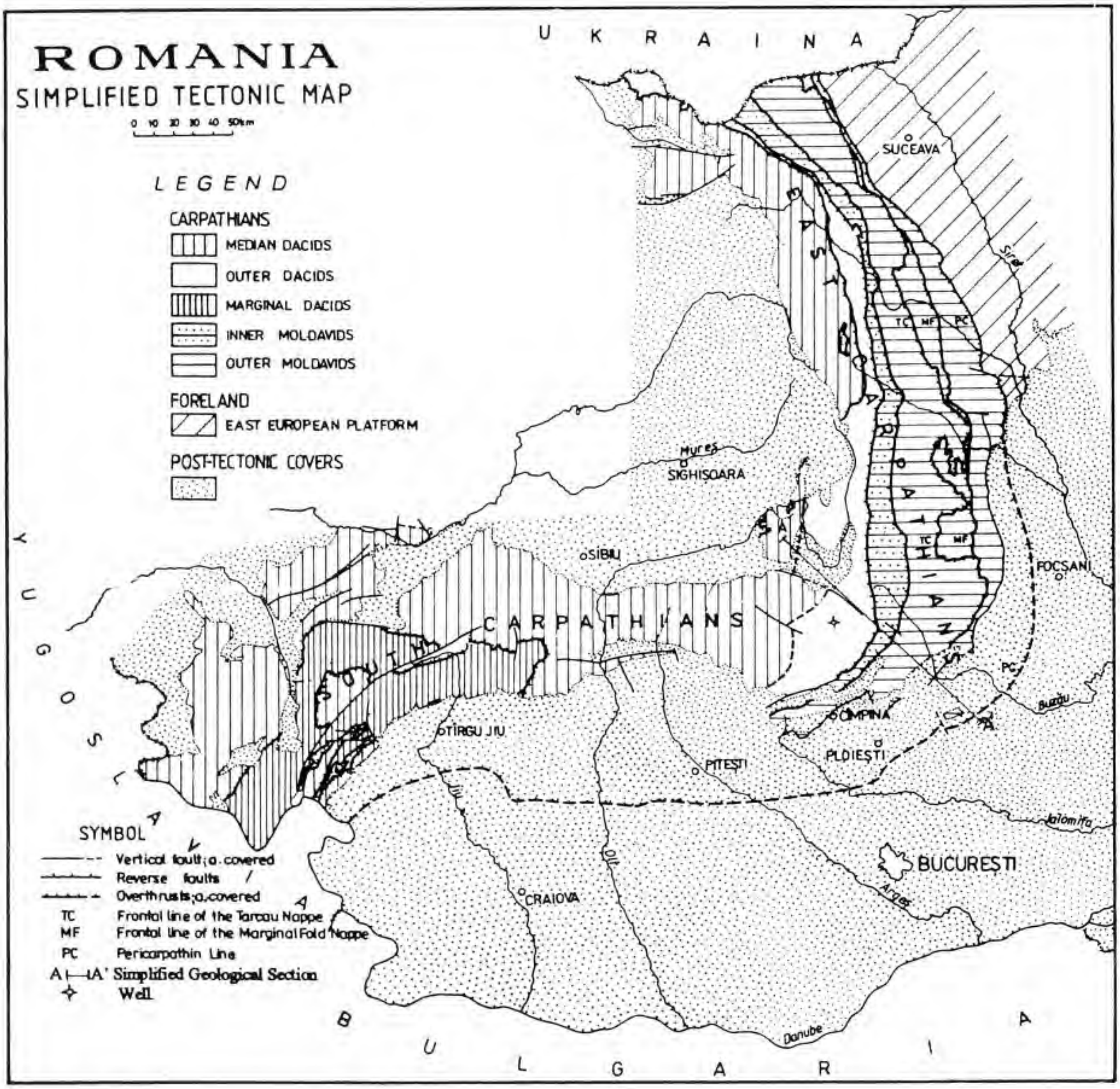

FIG. 1. Romania, simplified tectonic map

also Bessereau et al., this volume). This would not be all that surprising if we take into account that Cretaceous formations are well developed in the Romanian on-shore areas and that, on a world-wide scale, about one third of the total reserves are related to Cretaceous source-rocks and reservoirs (Klemme and Ulmishek, 1991).

On the Moesian Platform, Early Cretaceous carbonates (limestones and dolomites), clastics (clays, marls and sandstones) and evaporites (gypsum) reach total thicknesses of 300 to 1000 m. After an Aptian break, sedimentation of marls and sands resumed during the Albian and persisted till the end of the Senonian, albeit with uneven thicknesses and areal distribution. The Moesian Platform extends deeply under the East-Carpathian fold-and-thrust belt. In the latter, synorogenic Cretaceous shales and flysch-type sediments are several kilometres thick and crop out in extended areas.

It is the ungrateful task of this paper to attempt to demonstrate that there are still some hydrocarbon prospects in the Cretaceous flysch nappes of the Eastern Carpathians, despite the fact that today they are considered as being of little interest from an explorationists view.

STRATIGRAPHY AND STRUCTURE

Cretaceous sediments are involved in almost all of the East-Carpatnians nappes (Sandulescu et al., 1981a and 1981b, Sandulescu, 1984). In the Outer Dacids and Inner Moldavids they are mostly developed in flysch facies and are involved in a group of seven nappes, known in the Romanian geological literature as the Cretaceous Flysch Nappes. These are the more internal Black Flysch, Baraolt, Ceahlau and Bobu nappes of the Outer Dacids, and the more external Teleajen, Macla and Audia nappes of the Inner Moldavids (Fig. 1 and Encl. 1).

The **Black Flysch nappe** has a very complex internal structure and is characterized by four distinct imbrications. Each of them displays more or less distinct sequences which accumulated on a common basal mafic complex consisting of intraplate-type basalts, tholeiitic and calc-alkaline rocks (Sandulescu et al., 1981b). In the three external imbrications, flysch-type series were deposited during both Tithonian-Neocomian and Barremian-Aptian times. The slaty shales of the Tithonian-Neocomian flysch are very rich in graphite.

As all imbrications were affected by high-pressure/low-temperature metamorphic processes (Sandulescu et al., 1981b), the hydrocarbon potential of the Black Flysch nappe must be considered as negligible.

The **Baraolt nappe** is mostly made up of Early Cretaceous sandy-calcareous flysch which does not contain well developed source-rocks. Moreover, reservoirs involved in this nappe take in a very high structural position and, consequently, are deeply eroded. For these two reasons, the Baraolt nappe is also considered as non-prospective.

Despite the lacking prospectivity of the Black Flysch and Baraolt nappes, these two tectonic units have been mentioned here only because they form part of the Cretaceous Flysch Nappe system; they will not be further discussed in this paper.

The **Ceahlau nappe** is characterized by a very complicated internal structure, involving large imbrications. In the entire nappe, Tithonian-Neocomian sandy-calcareous flysch is well developed and is conformably overlain by a Barremian-Aptian sandy-shaly or shaly-sandy flysch sequence. In the internal parts of this nappe, Barremian-Aptian flysch is unconformably overlain by a thick pile of Albian conglomerates which, in turn, are unconformably covered by a hemipelagic to pelagic late Vraconian-Turonian sequence. Locally the latter rests directly on Aptian flysch. In contrast, in the more external imbrications of the Ceahlau Nappe, the Barremian-Aptian flysch grades upwards into a thick Albian sandy-shaly or sandy flysch sequence.

The **Bobu nappe** has a relatively simple internal structure which is characterized by large folds, involving Aptian-Turonian deposits developed in different facies. Aptian-early Albian strata are developed in a shaly-sandy flysch facies; middle Albian series consist of massive sandy flysch, locally containing conglomerate lenses. The lower parts of the late Albian (early Vraconian) are represented by shaly flysch which grades upwards into hemipelagic to pelagic series of the uppermost Albian (late Vraconian) and early Senonian.

The **Teleajen nappe**, which involves Hauterivian to Turonian series, has also a relatively simple internal structure, characterized by large, partly faulted, vertical or recumbent folds. Its sedimentary sequence begins with Hauterivian black-shales; these are overlain by Barremian-Aptian shaly-sandy flysch, containing black-shale intercalations. This flysch grades upwards into an up to 3 km thick sequence of Albian-Turonian shaly-sandy flysch, containing thick intercalations of massive sandy flysch.

The **Macla nappe** is characterized by a highly imbricated structure, involving only parts of its entire Albian-Turonian sequence. This unit consists of shaly flysch, containing red to purplish and black shale intercalations. Locally, in the highest part of this sequence, a thin level of sandy flysch is recognized.

The Ceahlau, Bobu, Teleajen and Macla nappes are overlain by a common late Senonian-early Miocene post-tectonic cover, attaining thicknesses of up to 1300 m.

The **Audia nappe** is characterized by a complex internal structure which is controlled by the lithological composition of the involved sedimentary sequences. Tightly imbricated structures occur in areas where mostly shaly Hauterivian-Aptian and Cenomanian-Turonian deposits dominate. In contrast, in areas where the upper part of the Audia Nappe sequence, consisting of Senonian to Eocene massive sandy flysch, crops out, the structural style is dominated by large synclines, separated by narrow, faulted anticlines.

The present overthrusted relationship between the above mentioned nappes was established during the following successive stages of the Carpathian orogeny:

(1) During Mid-Cretaceous times the Black Flysch and Baraolt nappes were emplaced; at the same time the internal parts of the Ceahlau nappe were folded and subsequently its leading edge covered by Albian-Turonian post-thrusting series.
(2) During intra-Senonian times the Ceahlau, Bobu, Teleajen and Macla nappes developed
(3) During the early Miocene (intra-Burdigalian), the latter were thrusted over Audia nappe and together with it over the Paleogene flysch zone.

SOURCE ROCKS

Fig. 2 summarizes the distribution of source- and reservoir-rocks in the Ceahlau, Bobu, Teleajen, Macla and Audia nappes. Although all five nappes contain source-rock intercalations, these occur at different stratigraphic levels. The oldest source-rocks accumulated during the Tithonian-Neocomian whilst the youngest were deposited during the Turonian. The longest period of source-rock accumulation spans the Tithonian-Aptian interval whereas the shortest one occurred during the middle Turonian.

In the **Ceahlau nappe**, the entire Tithonian-Neocomian interval contains around 2% TOC (for lithological details see Stefanescu and Micu, 1987); the kerogene is of Type I-II and is over-mature, as indicated by a R_0 greater than 2. The thickness of Tithonian-Neocomian source-rock shales varies between 350 m in the internal parts of the Ceahlau nappe and 500-600 m in its external parts. The Aptian and Albian flysch has a TOC of 1.1%; the kerogene is only of Type II and is also over-mature. Potential source-rock intercalations in the Aptian-Albian sequence have cumulative thicknesses ranging between 250 and 600 m.

Analysis on outcrop and well samples from the shaly-sandy Albian-Cenomanian flysch of the **Teleajen nappe** showed a TOC content varying between 0.8 and 1.35% and a kerogene of sapropelic type. R_0 values indicate that this flysch is at present located in the upper part of the oil window. The effectiveness of the Albian-Cenomanian flysch source-rocks is proven by the occurrence of oil and condensate seeps within the Teleajen nappe. The cumulative thickness of source-rock intercalations is of the order of 550 to 700 m.

The black Early Cretaceous series of the **Audia nappe**, which are associated with a few seeps, were more systematically analyzed (Baltes

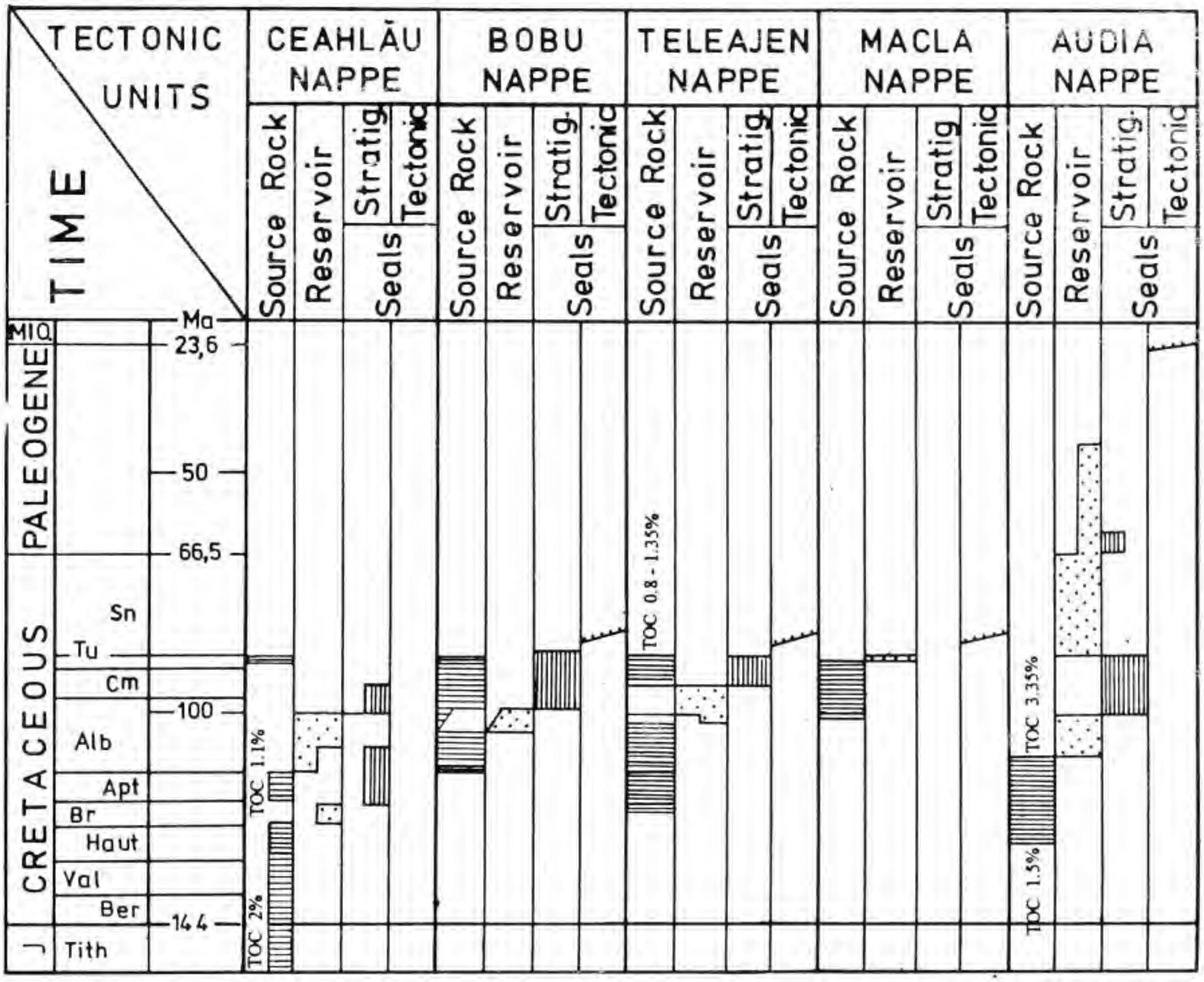

FIG. 2. Stratigraphic distribution of source- and reservoir-rocks in the Cretaceous Flysch Nappes.

et al., 1984) than the other Cretaceous formations of the Eastern Carpathians. Outcrop and core samples indicate a TOC range of 1.05-3.35%. Optical studies of the kerogen identified at least five main types of structured and one type of amorphous-colloidal organic matter. As a result of different and successive thermal flows, and also as a consequence of local burial histories and present tectonic positions, the organic matter shows extremely variable values of organic metamorphism ranging between R_0 0.55 and 2.02.

The Neocomian-Barremian black-shales, which contain siderite lenses, have an average TOC content of less than 1.5% and therefore a very weak petroleum generation potential. In contrast, Aptian-lowermost Albian shales, which attain a thickness of up to 200 m, yield TOC values reaching up to 3.35% and therefore have a much higher hydrocarbon generation potential; in places this sequence is still located near the top of the oil window whereas elsewhere it has passed through it an entered the gas generation window ($R_{0\ max}$ 1.9%).

It should be kept in mind, that the TOC values quoted above represent the residual potential of the respective source-rocks, the original organic content of which was presumably considerably larger.

Apart from the above discussed and identified source-rocks, the East-Carpathian Cretaceous flysch contains several additional potential source-rocks, the lithology of which compares favourably with proved source-rock intervals. For instance, the Albian flysch of the Bobu nappe is developed in a similar facies as the Albian flysch of the Teleajen nappe; the pelitic background of the Aptian flysch of the Teleajen nappe is similar to that of the Aptian flysch of the Audia nappe; the Cenomanian-Turonian shaly flysch of the Macla nappe can be compared with the Aptian black-shales of the Audia nappe. In this respect, is is noteworthy that wells drilled in the Macla nappe encountered important gas shows.

In conclusion, results of geochemical analyses and/or the occurrence of seeps indicate that, with the possible exception of the Bobu nappe, all of the above discussed Cretaceous flysch nappes contain effective source-rocks.

POTENTIAL RESERVOIRS

As shown in Fig. 2, all nappes under discussion contain reservoirs; however, their facies and stratigraphic position is different in each nappe.

The oldest potential reservoir section occurs in the **Ceahlau nappe** and consist of strongly tectonized, thick and massively bedded Barremian.flysch sandstones. The same nappe contains over 1500 m of Albian polymict conglomerates and sandstones, presenting a second objective section.

In the **Bobu nappe** the potential reservoir section is again Albian in age, but is developed in a thick bedded, sandy flysch facies which, macroscopically, shows good porosities. This section attains thicknesses of up to 1000 m and tends to wedge out towards western margin of this nappe where it is developed in a shaly flysch facies.

In the **Teleajen nappe** the potential reservoir section is of Late Albian-Cenomanian age; it is developed in a thick bedded, sandy flysch facies which in places attains thicknesses of the order of 1000 m and displays apparently good porosities (Fig. 3).

The **Macla nappe** is almost devoid of potential reservoir rocks. Only in places a Turonian feldspathic, thick bedded sandstone occurs, which never is thicker than 50 m.

The **Audia nappe** contains two potential reservoir sections. The older one is Albian in age and consists of 50-200 m thick, siliceous, hard, very low porosity sandstones. The younger section is Senonian-Eocene in age and and consists of 800-1000 m thick, massive sandy flysch that is characterized by good porosities.

SEALS AND TRAPS

Most of the above discussed potential reservoir sections are encased in thick shales, partly developed in source-rock facies; these provide viable top and seat seals (Fig. 2). Shale intercalations in the reservoir sections can provide for local intra-formational seals. Only the Turonian sand in the top part of the Macla nappe and the Senonian-Eocene sands of the Audia nappe apparently lacked an initial normal stratigraphic seal. However, both the Macla and Audia nappes are now tectonically sealed by shales contained in the basal parts of overriding nappes. Thrust faults can also provide potential seals for subjacent tectonic units. This does, however, not apply for areas where Albian conglomerates form the base of the Ceahlau nappe. Surface and subsurface geological data indicate that horizontal displacement on the individual Cretaceous flysch nappes is on average in excess of 10 km.

Lateral shale-out of potential reservoir sections may provide for stratigraphic (lithologic) traps. Such a shale-out is indicated, for instance, for the Albian-Cenomanian sandstones of the Teleajen nappe. Similar lateral facies changes may also occur in the other nappes.

In terms of potential structural traps, it must be kept in mind that the internal structuration of each nappe differs. The Ceahlau, Macla and Audia nappes, particularly in areas where only shaly deposits outcrop, display a very complex, generally tightly imbricated structure which is not favourable for the development of structural traps and the preservation of hydrocarbon accumulation. In contrast, the internal structure of the Bobu and Teleajen nappes, and in those parts of the Audia nappe

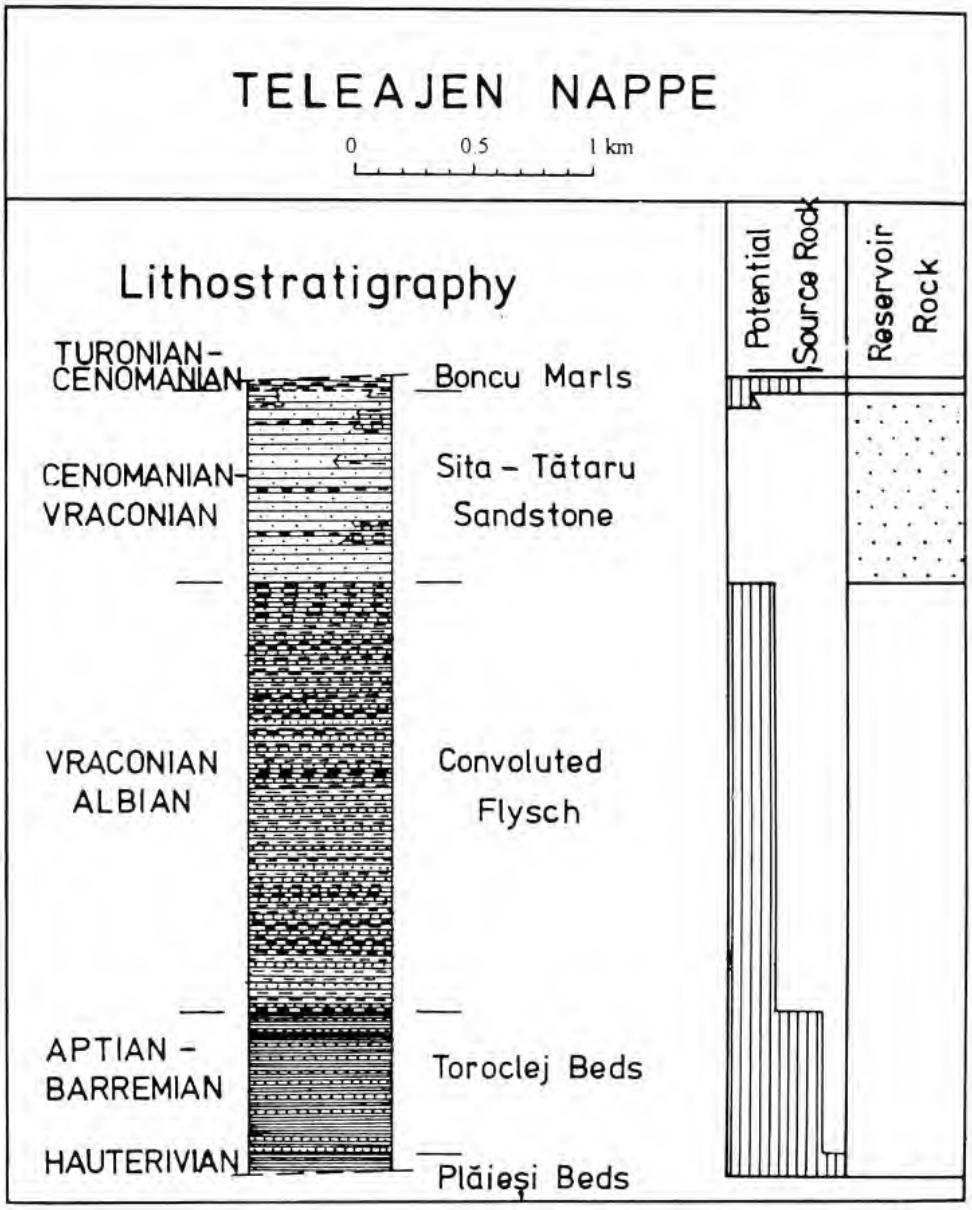

FIG. 3. Lithostratigraphy of the Teleajen nappe

where thick Senonian-Eocene sandstones are present, is characterized by large synclines and anticlines which are partly affected by steeply dipping thrust faults (Fig. 6). In such faulted structures, most of the reservoir sections occur in a high structural position and are consequently deeply eroded. Therefore, potential structural traps are restricted to features which are sealed by overriding nappes.

HYDROCARBON PROSPECTS

Taking the above into account, conditions for generation, accumulation and preservation of hydrocarbons is probably most favourable in the Teleajen nappe. Therefore, the following discussion will focus on this tectonic unit.

The stratigraphic sequence involved in the Teleajen nappe consists, in ascending order, of the following members (Fig. 3): locally the sequence begins with 50 m of Hauterivian black clays and siltstones. These grade upward into up to 500 m thick Barremian-Aptian shaly-sandy flysch which has the same black-shale pelitic background as the Hauterivian. The next following unit consists of 2500 m of Albian shaly-sandy and sandy-shaly flysch, containing thin black-shale intercalations. This member is capped by 750 m of late Albian-early Cenomanian massive sandy flysch, locally containing lenses of polymict conglomerates. The topmost member consists of 20-400 m of late Cenomanian-Turonian shaly-sandy flysch. Older than Hauterivian deposits are not known from the Teleajen Nappe which, like all the tectonic units of the Cretaceous Flysch nappes, is completely detached of its initial basement. The Hauterivian-Albian and late Cenomanian-Turonian series contain effective source-rocks whereas the Late Albian-Early Cenomanian sands have excellent reservoir characteristics (Fig. 4).

During the Coniacian the area of the future Teleajen nappe was overridden by the Ceahlau and Bobu nappe and prior to the Santonian, the Teleajen nappe was thrusted over the next external unit, the area of the Macla nappe, which, in turn, was also deformed. At the same time the entire sedimentary sequence making up the Teleajen nappe was folded for the first time. During the late Senonian to Oligocene, an over 1000 m thick sequence of neo-autochthonous sediments was deposited on the Teleajen and the other Cretaceous Flysch nappes. During this time, the area of the Audia nappe had not yet been deformed and formed part of the Carpathian foreland basin. During the early Miocene, the Ceahlau, Bobu, Teleajen and Macla nappes were thrusted over the Audia nappe and, together with it, over the Paleogene flysch zone. During this late phase of the Carpathian orogeny the Teleajen nappe was deformed a second time.

Fig. 5 summarizes the burial and maturation history of the Teleajen nappe. In parts of the future Teleajen nappe, Hauterivian-Early Aptian series had already entered the oil window during the Coniacian, that is prior to its involvement into the Carpathian orogen (Fig. 5a). However, hydrocarbons generated and expelled from these source-rocks accumulated either in stratigraphic traps in Late Albian-Early Cenomanian reservoirs or were lost as at that time structural traps had not yet been formed.

At the end of the Coniacian, subsidence of the Teleajen basin ceased and, prior to the late Campanian, the Ceahlau and Bobu nappes were thrusted over the Teleajen nappe, which in its turn was thrusted over the Macla Nappe. At the same time the internal parts of the Teleajen Cretaceous flysch basin were strongly deformed, resulting in the development of structural traps.

During Campanian to end-Oligocene times, the Teleajen nappe subsided under the load of its neo-autochthonous sedimentary cover. During this time, the Late Aptian-Turonian source-rocks entered and partly passed through the oil window (Fig. 5b). Hydrocarbons generated presumably accumulated in earlier formed structural traps. During the early Miocene phase of the Carpathian orogeny, the configuration of these traps was modified, causing loss of hydrocarbons. However, Albian-Turonian rocks, including reservoir- and source-rocks, remained till the present in the oil window. This is in keeping with the occurrence of light oil seeps in the Teleajen nappe, immediately in front of the Bobu nappe.

The late Albian-Cenomanian reservoirs of the Teleajen Nappe are involved in imbricated, folded structures. In areas where this nappe outcrops, these structures are deeply eroded and their original stratigraphic seals have been removed (Fig. 6). Therefore, the outcropping parts of this nappe must be considered as essentially non-prospective. However, further to the West, where the Teleajen nappe is covered by the Bobu nappe, subthrust structures may still be effectively sealed, either by Cenomanian-Turonian marls or by the sole-thrust of the Bobu nappe. Structuration of the Teleajen nappe beneath the Bobu nappe is presumably characterized by a similar style as in its outcropping parts. Consequently, if beneath the Bobu nappe appropriate structures, involving thick Late Albian-Cenomanian reservoirs, can be identified, these may present viable exploration targets. However, in view of the complexity of the structuration of the Teleajen nappe, it remains to be seen whether such targets can be defined by the reflection-seismic tool.

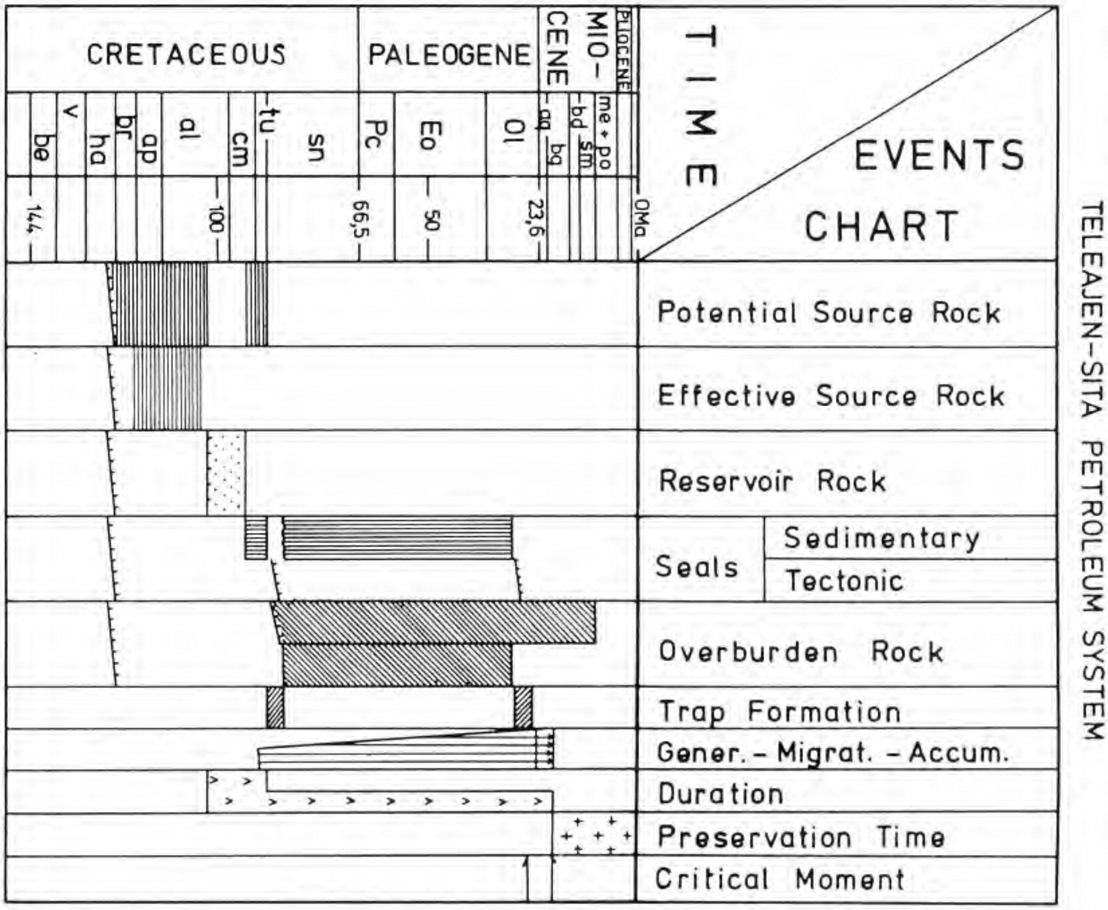

FIG. 4. Teleajen nappe petroleum system

CONCLUSIONS

The Cretaceous flysch of the Eastern Carpathians contains several well developed reservoir- and source-rock intervals. Geochemical analyses and the occurrence of seeps show that source-rocks have variably entered the oil and even the gas window.

During the evolution of the Cretaceous flysch basin towards a petroleum system, early generated hydrocarbons accumulated either in stratigraphic traps or were lost due to unfavorable timing between peak generation and the formation of structural traps. Hydrocarbons accumulated in stratigraphic traps, were presumably destroyed during the orogenic phases.

Part of the source rocks are at present still located within the oil window. Hydrocarbons generated during the syn-deformational stages of the Carpathians were presumably structurally trapped. Polyphase deformations and late uplift and erosion of the Carpathians resulted in destruction of such accumulations, mainly by removal of their stratigraphic seals. However, in sub-thrust positions, such accumulations, which are either stratigraphically or tectonically sealed, may still be preserved, for instance in the area where the Teleajen nappe is covered by the Bobu nappe, or the latter is covered by the Ceahlau nappe.

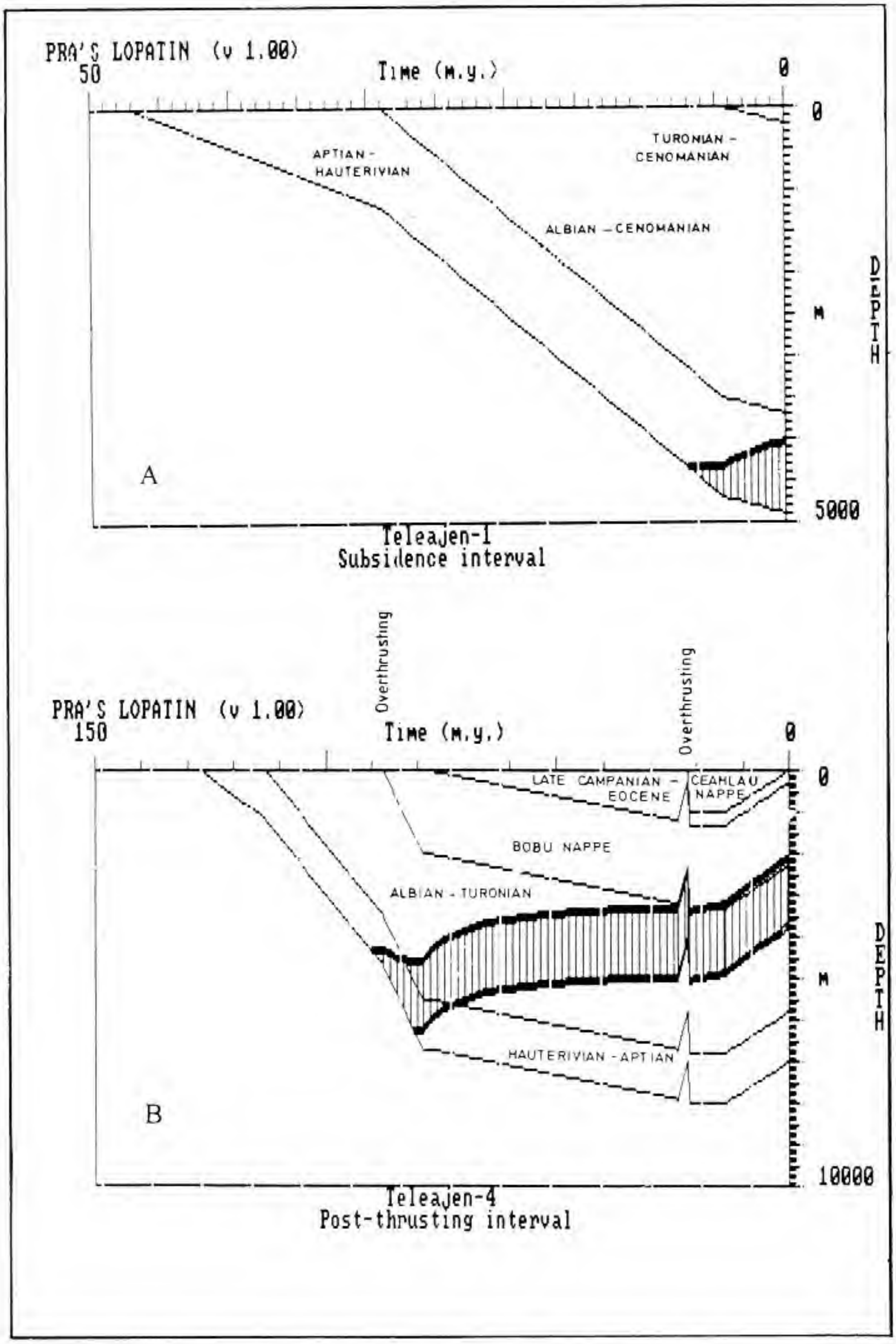

FIG. 5. Teleajen nappe burial and maturation history. Vertically hachured area corresponds to oil window.

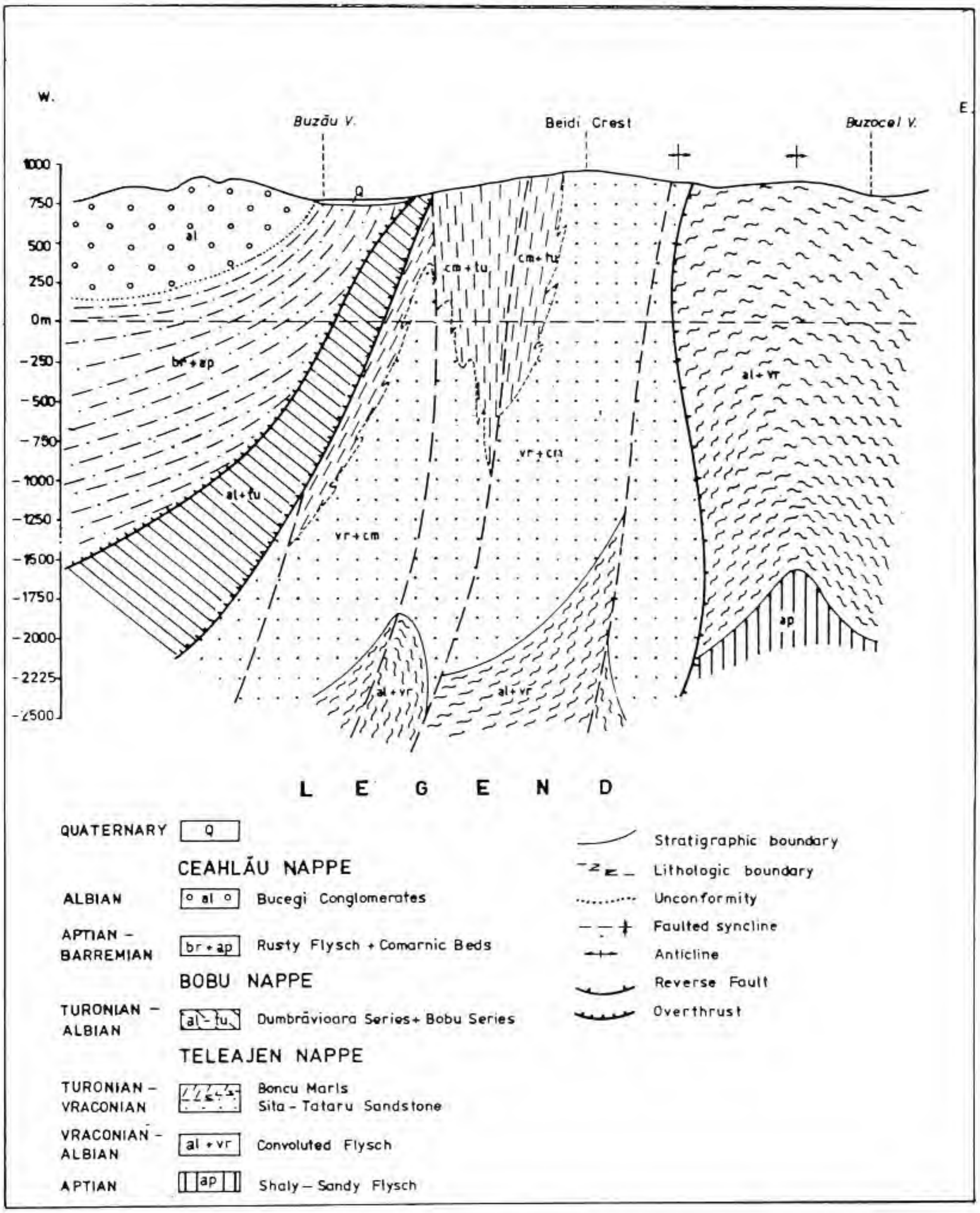

FIG. 6. Geological cross-section at the boundary between the Bobu/Teleajen nappes.

The prospectivity of the East-Carpathian Cretaceous flysch nappes cannot be ruled out, particularly in terms of subthrust prospects sealed by the sole-thrusts of higher nappes. The main risk of such a play is the reflection-seismic definition of drillable structures. Hydrocarbon charge and reservoir risks play a secondary role.

We conclude, that albeit speculative hydrocarbon prospects still exist in the East-Carpathian flysch nappes of Romanian.

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Enclosure

Encl. 1 Persani-Ciucas-Pietroasa simplified geological cross-section. For location see Fig. 1.

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Neoalpine tectonics of the Danube Basin (NW Pannonian Basin, Hungary)

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ABSTRACT

The structure of the pre-Tertiary substratum of the NW Pannonian Basin is traditionally interpreted in terms of subvertical Tertiary strike-slip faults controlling the subsidence of major pull-apart basins. However, based on a recent reevaluation of reflection-seismic data the middle Miocene structure of the basin is dominated by a number of low-angle normal faults.

The gently dipping basement of the European foreland can be traced some 200 km to the SE beneath the extensionally collapsed transition zone between the Eastern Alps and the Carpathians. This suggests a large-scale allochthoneity of the Alpine edifice underneath the NW Pannonian Basin.

The compressional pre-conditioning of the substratum of the Neogene NW Pannonian Basin was always assumed to be a key factor in the formation of extensional structures by reactivation of pre-existing weakness zones. Based on reflection-seismic data, such an interaction between Cretaceous compressional décollement levels and Miocene low-angle normal fault planes indeed occurred, although in more complex manner than previously assumed.

INTRODUCTION

This paper discusses the Neo-Alpine (sensu Trümpy, 1980) evolution of the area which straddles the junction between the Eastern Alps and the Western Carpathians and is occupied by the NW Pannonian Basin, more specifically by the Hungarian part of the Danube Basin. The following Neo-Alpine structural stages are recognized:

(1) Early Miocene "escape" tectonics which follow on the heel of the Paleogene Mesoalpine compressional phase,
(2) Middle Miocene syn-rift tectonics,
(3) Late Miocene-Pliocene post-rift tectonics and
(4) Quaternary-Recent neotectonics.

Tari, G., 1996. — Neoalpine tectonics of the Danube Basin (NW Pannonian Basin, Hungary). *In*: Ziegler, P. A. & Horváth, F. (eds), Peri-Tethys Memoir 2: Structure and Prospects of Alpine Basins and Forelands. *Mém. Mus. natn. Hist. nat.*, **170**: 439-454 + Enclosures 1-3. Paris ISBN: 2-85653-507-0.

This article includes 3 enclosures on 2 folded sheets.

This papers focuses on problems related to the Mid-Miocene syn-rift tectonics which underlay the formation of the Danube and related basins (Fig. 1).

The middle Miocene Danube Basin has been interpreted by many authors (e.g. Bergerat, 1989; Vass et al., 1990) as a large pull-apart basin. In this paper I document the presence of a system of major low-angle normal faults in this basin which are evident on reflection-seismic data, calibrated by wells. The presence of these detachment faults and the lack of major throughgoing middle Miocene strike-slip structures contradicts the traditional pull-apart basin interpretation. Note that in this paper the terms low-angle and detachment fault are used interchangeably. The new observations and interpretations are summarized in a regional structural transect across the NW Pannonian Basin, more specifically, across the Danube Basin.

The Neoalpine Danube Basin, which forms part of the larger Pannonian basin complex is superimposed on an earlier Cretaceous and Paleogene compressional realm, as inferred from the Alpine structure of the surrounding thrust-fold belts, such as the Alps, Carpathians and Dinarides (Fig. 1). Based on well and reflection-seismic data, these structures can be traced with considerable confidence at depth through the Danube Basin (Figs. 2 to 4). Moreover, these data show, that the compressionally pre-conditioned "memory" of the substratum of the Neogene Danube Basin played a significant role in localizing Miocene extensional faults, partly involving the tensional reactivation of pre-existing compressional decollement levels (e.g. Grow et al., 1989; Tari et al., 1992). The seismic line drawings given in this paper (see also Tari and Horváth, 1995; Tari, 1995a) show that reactivation of abandoned Eoalpine thrust fault planes occurred frequently, however, in a more complex manner than anticipated by many authors.

GENERAL NEOALPINE TECTONO-STRATIGRAPHY OF THE NW PANNONIAN BASIN

The locally very thick Neogene sedimentary fill of the NW Pannonian Basin, which exceeds in the centre of the Danube Basin 8 km, can be subdivided into two major units (for an overview see Royden and Horváth, 1988). The upper unit is late Miocene to Pliocene in age (Sarmatian/Pannonian; 13.8-0 Ma) and forms the post-rift sedimentary succession which accumulated in response to regional thermal subsidence of the area. The thickness variation and spatial distribution of the underlying middle Miocene (Karpatian/Badenian; 17.5-13.8 Ma) succession is largely controlled by syn-rift structural features. Deposition took place in fault-bounded half-grabens. Some of the deeper subbasins of the Pannonian Basin were clearly formed by extensional detachment faulting (Tari et al., 1992).

Note that I placed the syn-rift/post-rift boundary stratigraphically earlier than Royden et al. (1983). Commonly this boundary is placed at the Pannonian/Sarmatian boundary (i.e. ~10.5 Ma); however, based on a review of the available well and seismic data this boundary must be placed between the upper and middle Badenian, some 3.3 Ma earlier (for a detailed discussion, see Tari and Horváth, 1995).

STRUCTURE OF THE DANUBE BASIN BASED ON REFLECTION SEISMIC DATA

The following discussion on the structure and evolution of the Danube Basin is based on a systematic structural and seismostratigraphic interpretation of a reflection-seismic grid, including some 200 lines covering the Hungarian part of the NW Pannonian Basin (Tari, 1994).

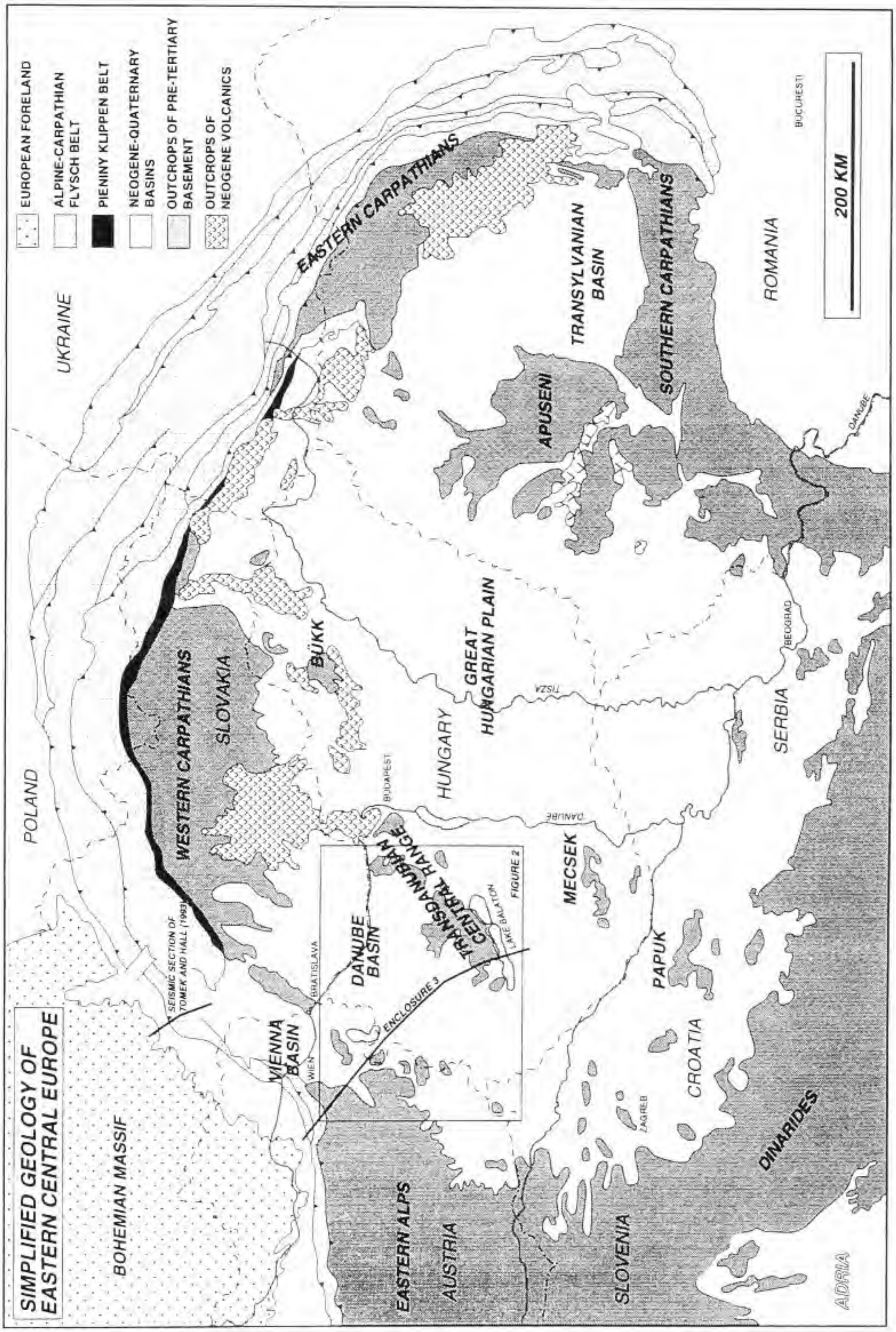

FIG. 1. Simplified geologic map of the Carpathian/Pannonian system showing the location of this study.

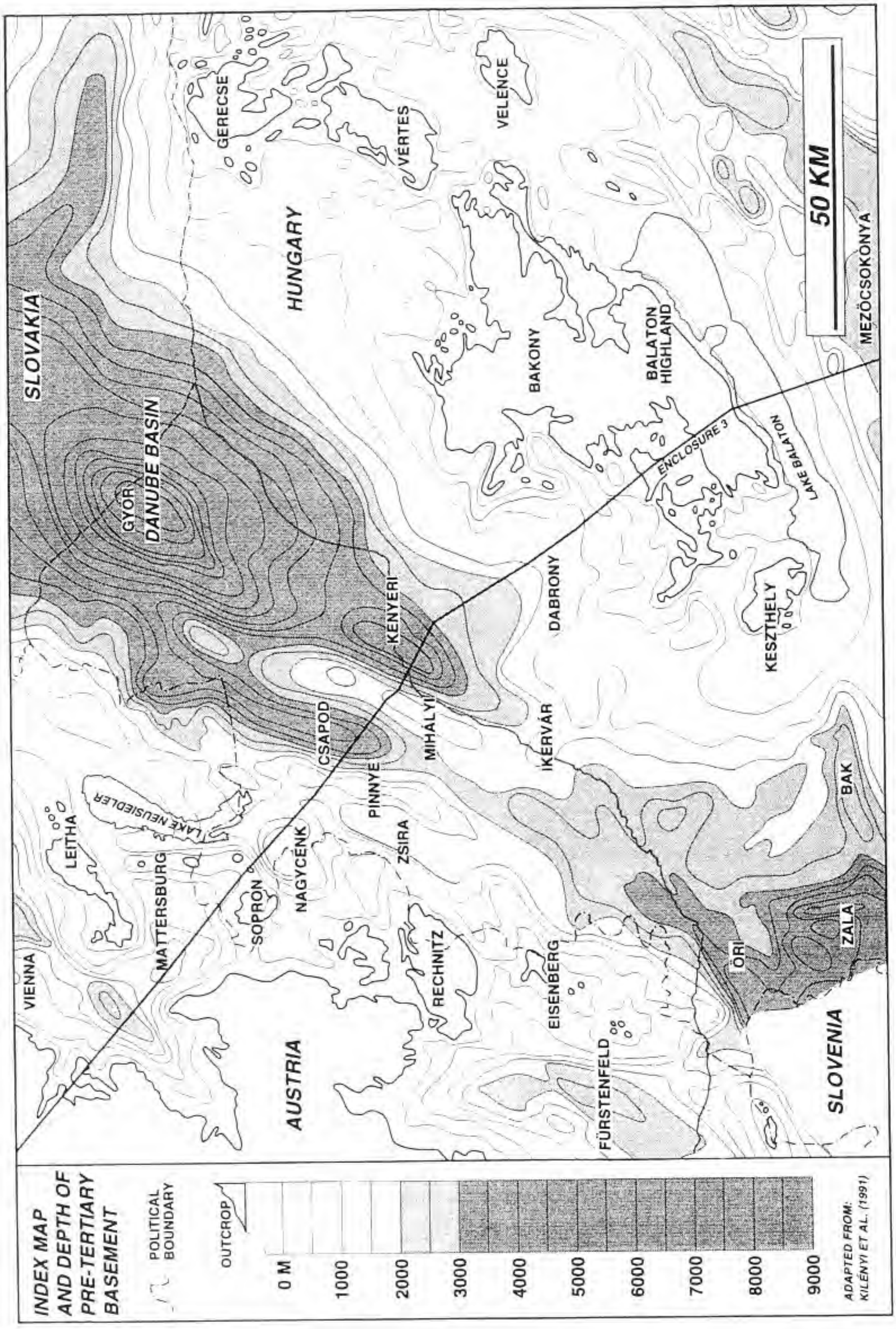

FIG. 2. Index map and depth of pre-Tertiary basement in the NW Pannonian Basin. For location see Fig. 1.

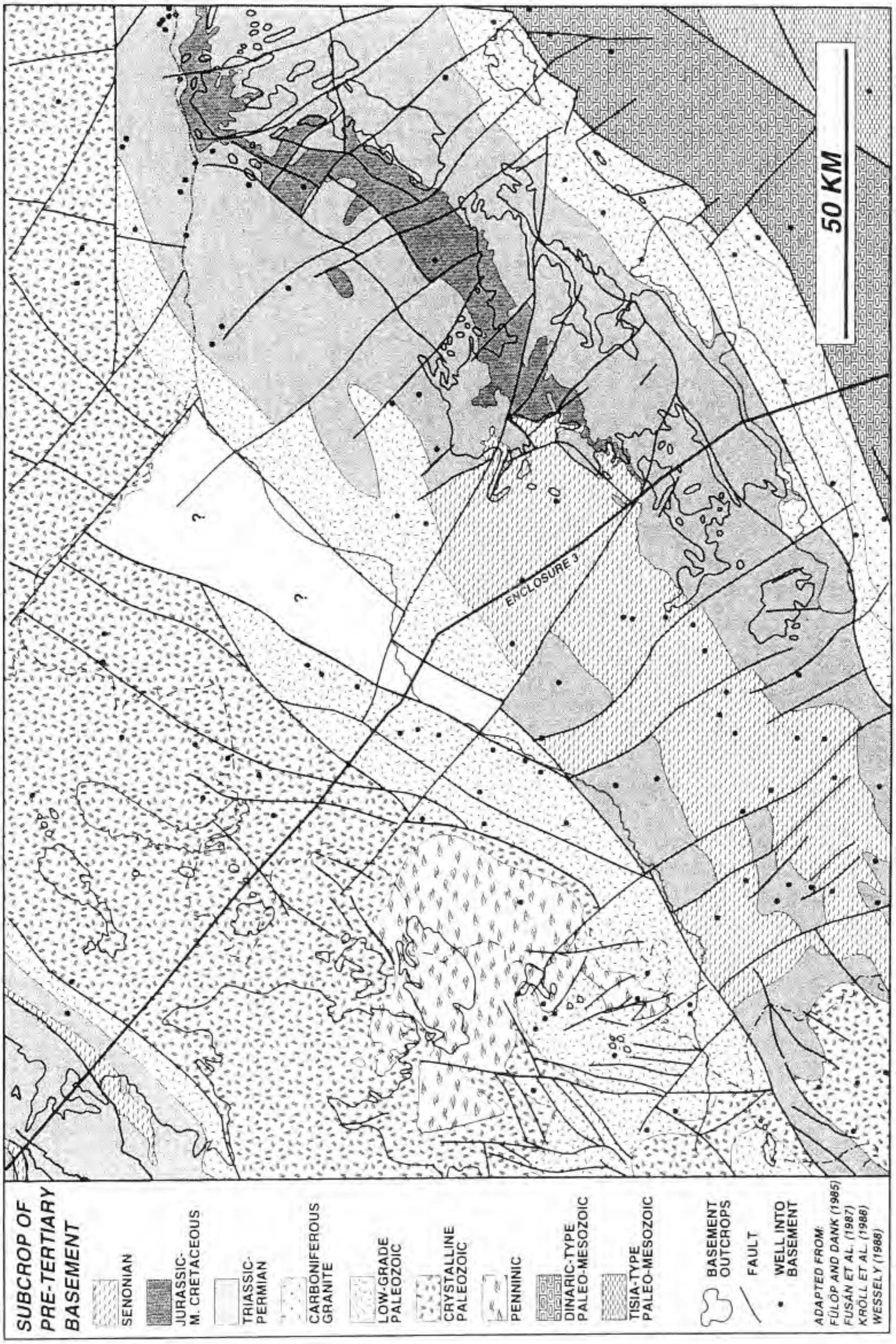

FIG. 3. Subcrop of pre-Tertiary basement in the NW Pannonian Basin. For location see Fig. 1.

Seismic Data Set

From this extensive data set five characteristic seismic profiles were selected; these are given in Enclosures 1 and 2. Three additional profiles from the same area were presented by Tari and Horváth (1995). Whereas the deep reflection profile of Posgay et al. (1986) is shown in Encl. 1 with and without interpretation, four industry-type reflection-seismic lines are reproduced as line drawings on Encl. 2. All the sections are migrated and the industry profiles are displayed at a 1:1 scale for a velocity of 5,000 m/s (16,400 ft/s); datum is 100 m (328 ft) above sea level.

Alpine Stratigraphy in Terms of Seismic Signatures

Fig. 5 gives a simplified summary of the stratigraphy of the Hungarian part of the NW Pannonian Basin; it is based on a detailed Phanerozoic lithostratigraphy described by Tari (1995b). While the thickness data are well known for the upper 10 km of this composite section, thickness relations are poorly constrained for the Palaeozoic of the Austroalpine units and for the Mesozoic of the Penninic unit. Interval velocities of the major units were compiled based on velocity surveys in selected wells and reported interval velocities in several seismic surveys. Fig. 5 also shows the interval velocities which were adopted for the depth-conversion of selected seismic sections. Moreover, its righthand column identifies the seismic mapping horizons as shown in Enclosures 1 and 2.

Characteristic Reflection Seismic Examples

Enclosure 1 shows the MK-1 deep reflection profile (for location see Fig. 4) of Posgay et al. (1986) as well as its line drawing interpretation by Tari (1994).

At the NW end of this line, Palaeozoic crystalline rock, attributed to the Lower Austroalpine nappes, crop out in the area of the Sopron Mtns. (Figs. 2 and 3); well data near the trace of this line indicate that Palaeozoic basement holds up also the Pinnye High (Körössy, 1987), seen at line km 18. This high is flanked by two Neogene half-grabens, the Nagycenk Basin to the NW and the Csapod Basin to the SE (Ádám et al., 1984). The Mihályi High, evident near line km 40, is upheld by low-grade metamorphic Palaeozoic rocks (Balázs, 1971, 1975).

Both the Pinnye and Mihályi highs are bounded on their SE flank by major low-angle normal faults on their SE side (see below the industry seismic profiles). The fault which bounds the Mihályi High has been referred to by several authors as the Rába fault (for a detailed discussion see Szafián and Tari, 1995). To the S of this fault, the basement is covered by a Late Cretaceous sedimentary succession. At line km 50 the middle Miocene syn-rift sequence shows a clear thickening in the Kenyeri subbasin of the Danube Basin. Further to the SE, pronounced reflector packages within the pre-Senonian basement suggest the presence of a number of NW vergent thrust faults (Tari, 1995a).

The four industry seismic lines, for which line drawings are given in Encl. 2 come from the northwestern part of the Hungarian Danube Basin (Fig. 4). In this area, the Neogene basin fill displays a general monoclinal dip to the E. While the post-rift Pannonian succession covers all the pre-Tertiary basement structures, the syn-rift middle Miocene (Karpatian-Badenian) can be found only in local subbasins, delimited by faulted basement highs. The two prominent Bük-Pinnye and Mihályi-Mosonszentjános basement highs strike to the NE-N and delimit the Csapod subbasin (Fig. 2).

Starting from the SW, the consecutive dip-oriented (i.e. NW-SE) seismic sections C1, C3 and C5 reveal the gradual deepening and widening of the Csapod subbasin that is related to an increase in offset along a major detachment fault on its northwestern flank. This clearly low-angle fault flattens at depth and therefore can be regarded as a listric normal fault sensu Bally et al. (1981). This fault corresponds to the Alpokalja or Répce Line of Fülöp (1989, 1990), separating low-grade metamorphosed Palaeozoic rocks from crystalline rocks. The seismic data clearly show that this fault is not a strike-slip fault (cf. criteria given by Harding, 1990), as previously suggested by a number of

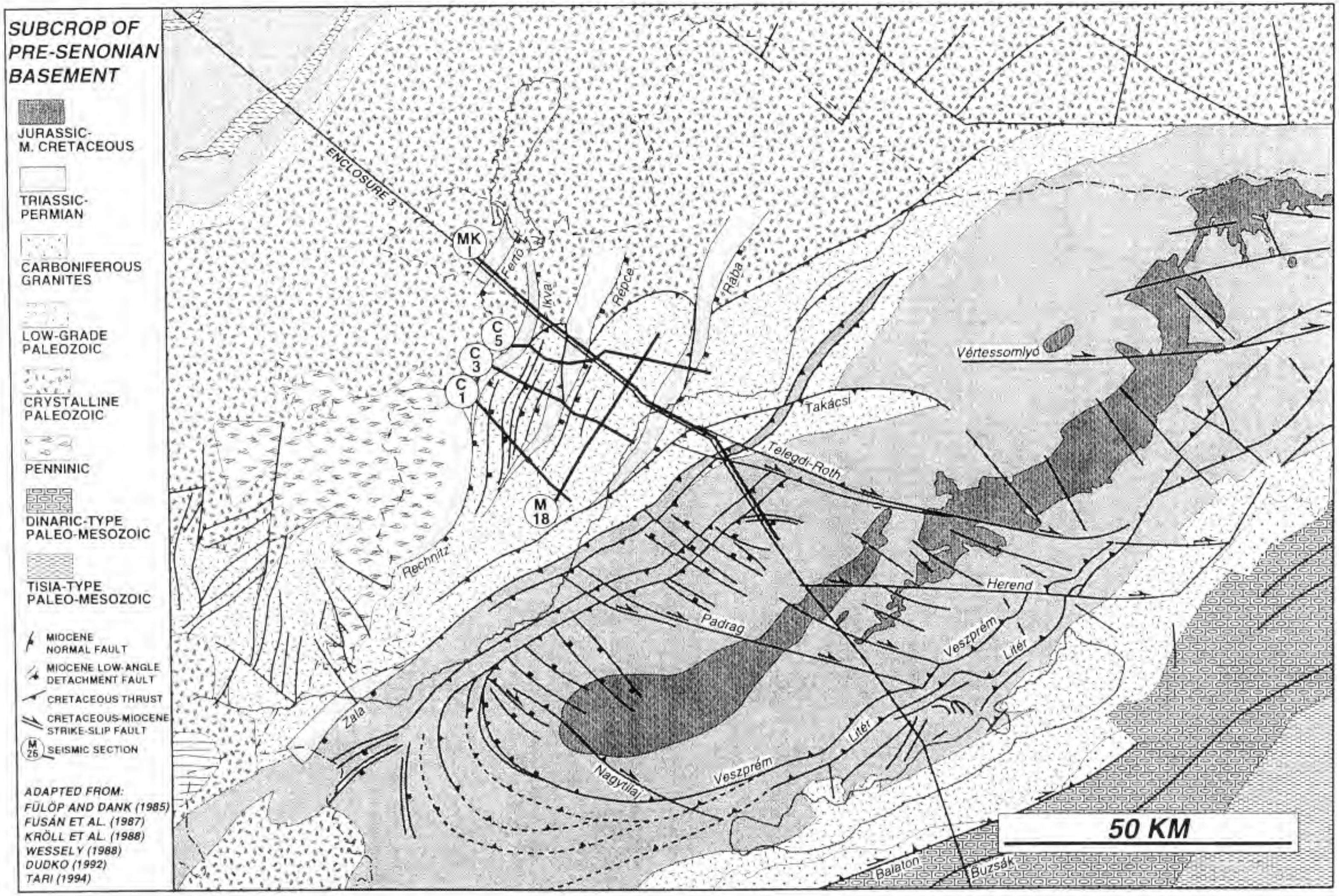

FIG. 4. Subcrop of pre-Senonian basement and Alpine structural elements revealed by reflection-seismic data in the Hungarian part of the NW Pannonian Basin (Tari, 1994), compare Fig. 3. For location see Fig. 1.

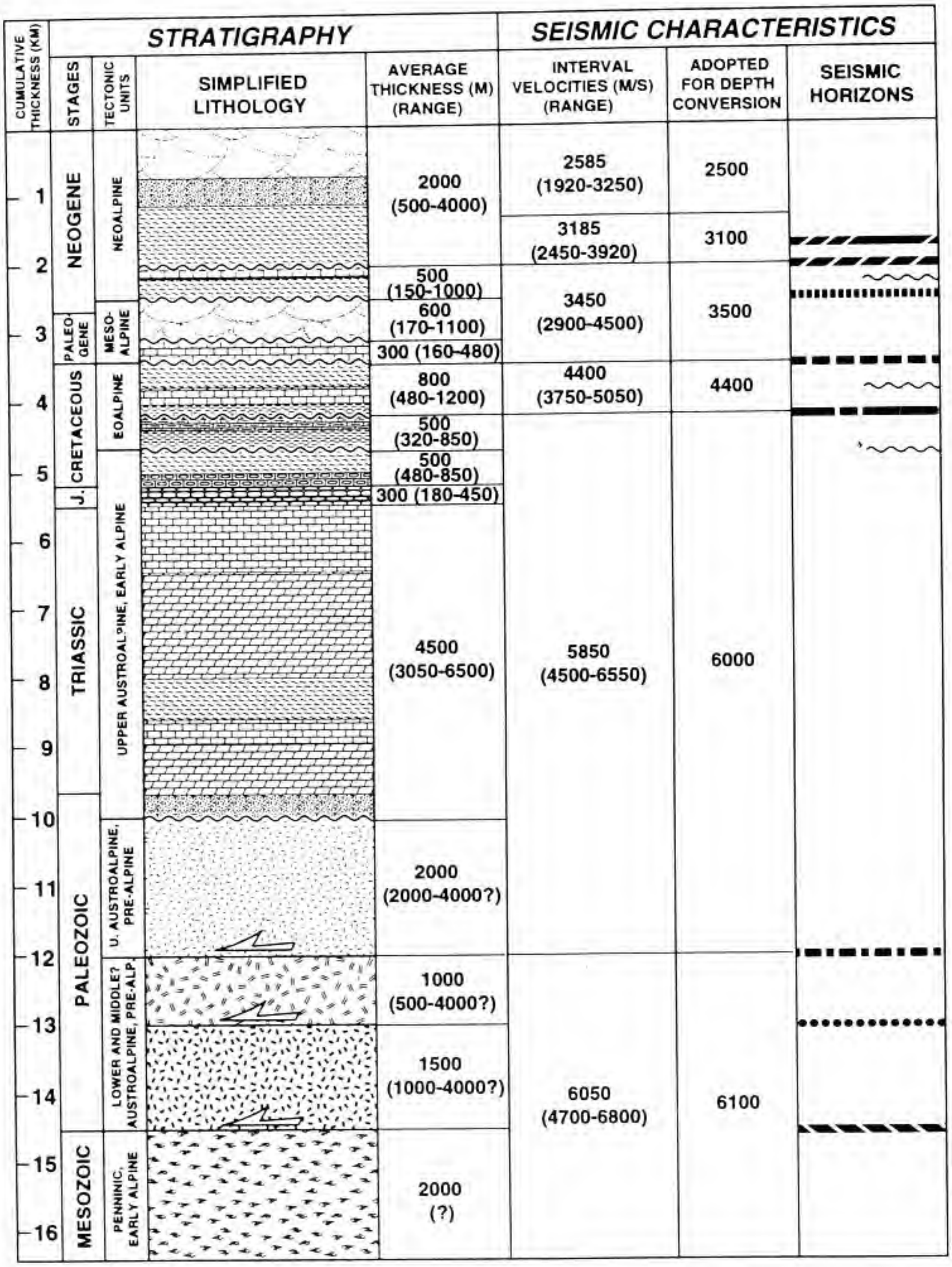

FIG. 5. Lithology and seismic characteristics in the NW Pannonian Basin (Tari, 1994).

REGIONAL STRUCTURE TRANSECT BASED ON DEPTH-CONVERTED REFLECTION SEISMIC SECTIONS

The location of the regional transect given in Encl. 3 is shown on Figs. 1 to 4. This section starts in the N in the European foreland, crosses the Eastern Alps, the Vienna and Danube basins, and terminates in the S at Lake Balaton, at the Mid-Hungarian shear zone.

The northern, Austrian part of this transect is based on a section published by Wessely (1987). In the Eastern Alps three tectonostratigraphic levels are recognized (Wessely, 1988). The lowermost level corresponds to the crystalline basement of the European foreland and its autochthonous Mesozoic cover which dips gently to the S. Beneath the Vienna Basin, the autochthonous Mesozoic cover is preserved. This cover is, however, missing under those parts of the Eastern Alps which are located in the projection of the Bohemian Massif; over this basement spur, which was intercepted by the well Berndorf 1, Mesozoic strata were eroded in conjunction with latest Cretaceous and Palaeocene compressional foreland deformations (Wessely, 1987). To the NW of this well, the European foreland crust is covered by deeply eroded autochthonous Mesozoic series, thin Late Eocene sands and carbonates, Oligocene-early Miocene flysch and middle Miocene molasse (Zimmer and Wessely, this volume) .

The next level is represented by the allochthonous Alpine nappes, which outcrop to the W of Vienna (Fig. 1). The tectonically highest unit corresponds to the Upper Austroalpine nappes; in the Northern Calcareous Alps these can be subdivided from top to bottom into the Upper Limestone Alps and Graywacke zone (Juvavicum), Ötscher nappes (Tirolicum, or Göller nappe system) and the Frankenfels-Lunz nappe system (Bajuvaricum, Hamilton et al., 1990). All these nappes are "cover nappes" (Tollmann, 1989), in so far as they were detached from their basement. These nappes exclusively consist of unmetamorphosed Mesozoic sediments, except for the uppermost Juvavic nappe which has a low-grade metamorphosed Palaeozoic substratum (Graywacke zone). These units are underlain by the Middle and Lower Austroalpine thrust sheets which also outcrop along strike. The presence of a Penninic unit at depth is problematic in the area of the Vienna Basin due to lack of well control (Wessely, 1988; Zimmer and Wessely, this volume).

The uppermost level is represented by the Neogene succession of the Vienna Basin. The structure section crosses the southwestern corner of this basin (Fig. 2), where normal faults bound the 2-3 km deep Neogene basin. These normal faults are shown to sole out and merge with the base of the underlying Alpine nappe complex (Encl. 3). Note that this is the only modification I made to the original sections of Wessely (1988), who thought that the normal faults also affected the autochthonous European foreland crust. These normal faults were thought by many authors to accommodate sinistral strike-slip movements required for the opening of the Vienna pull-apart basin (Royden et al., 1982; Fodor, 1991, 1995; Fodor et al., 1990). The inferred left-lateral offsets along these major faults, however, could not be documented (Wessely, 1988).

The Vienna Basin is separated from the Danube Basin, which underlays the Little Hungarian Plain, by a composite basement high which trends perpendicular to this transect. This high consists of the Leitha and Sopron Palaeozoic basement blocks (Fig. 2) which are attributed to the Lower Austroalpine unit. These blocks are bounded to the southeast by major normal faults and are separated by the small Mattersburg Neogene basin.

The section crosses the Austrian/Hungarian border just to the N of the Sopron Mts. and from there follows the trace of the deep reflection-seismic section MK-1 given in Encl. 1 (Ádám et al., 1984; Posgay et al., 1986). Further to the S, the section follows the continuation of the MK-1 line through the Bakony Mts., which was processed only to 4 s TWT time (Ádám et al., 1985). This part of the section, however, is constrained by surface geology (e.g. Császár et al., 1978).

In the northwestern part of the Danube Basin the pre-Tertiary basement exhibits a characteristic basin-and-range morphology. Individual subbasins (Mattersburg, Nagycenk, Csapod, Kenyeri) are separated by basement highs (Leitha, Sopron, Pinnye, Mihályi). All of these subbasins are controlled by major SE-dipping middle Miocene normal faults. The crustal section clearly shows that at authors. In the following this low-angle normal fault is referred to as the **Répce fault** (Tari, 1994).

The Répce fault plane itself can be traced between the terminations of more or less coherent SE-dipping basement reflectors of the Pinnye high and the overlying chaotic seismic facies which corresponds to coarse-grained clastics shown stippled on profiles C3 and C5. This facies unit represents alluvial talus which was deposited synchronously with the initial activity of the Répce fault. This facies unit was penetrated by the nearby Csapod-1 well which encountered an about 500 m thick Karpatian succession of conglomerates and breccias (Körössy, 1987).

The Badenian syn-rift fill of the Csapod subbasin is just slightly asymmetric and documents only little or negligible fault growth. This indicates that much of the normal faulting had occurred right at the beginning of rifting, i.e. during the Karpatian. Strikingly similar seismic examples of analogue basins were published from the Basin and Range province by Effimov and Pinezich (1981) and from the Newark Basin by Costain and Coruh (1989).

Interestingly enough, coherent basement reflector packages below the Mihályi High described a roll-over anticline (section C5) which apparently is associated with the large normal offset on the Répce detachment fault. The normal offset on this fault can be estimated by restoring displaced prominent basement reflectors in the Bük-Pinnye and Mihályi-Mosonszentjános highs. Such reconstructions suggests that the magnitude of offset along the Répce fault varies along strike between 4-10 km (horizontal component), with error bars being on the order of 0.5 km.

Note that, within the basement, the Répce detachment fault shows up as prominent fault-plane reflectors (section C1). Comparable fault plane reflectors, originated from similar detachment fault planes, were reported from Utah (e.g. von Tish et al., 1985) and Arizona (e.g. Frost and Okaya, 1986).

Since many intra-basement reflecting horizons could be correlated with considerable confidence in this area, I mapped certain Eoalpine basement units based on their seismic character (Tari, 1994). The best geometric constraint is provided by the Penninic succession which has a very distinct, highly reflective seismic expression and a well-defined top (section C1).

Since the Upper Austroalpine unit is lithologically markedly different from the Middle Austroalpine (very low-grade to low-grade versus medium-grade metamorphics) their contact is interpreted to correspond to a pronounced change in reflectivity (section C1). The Upper Austroalpine unit is characterized by short, but strong reflectors in contrast to the underlying Middle Austroalpine which has a mostly transparent character.

In the S, the Upper Austroalpine unit can be found right on top and in fault contact with the Penninic(section C1). This relationship was indeed observed along strike in outcrop at the Eisenberg Mountains (Figs. 2 and 4) where the Upper Austroalpine Hannersdorf series have a poorly understood tectonic contact with the Penninic succession (e.g. Pahr, 1980; Schmidt et al., 1984; Tollmann, 1989).

On sections C1 and C3 another detachment fault can be interpreted to the NW of the Bük-Pinnye high which controlled subsidence of a smaller syn-rift graben. Tari (1994) referred to this detachment fault as the **Ikva fault** and to the associated basin as the Zsíra subbasin (Fig. 2). In the S the Ikva fault is detached on top of the Penninic unit. Farther to the NE, however, the fault shows gradually decreasing normal offset and flattens out close to base of the inferred Middle Austroalpine unit (section C5).

Whereas in the S (section C1) the Répce fault seems to flatten close to or into the boundary between the Upper/Middle Austroalpine units (see strike section M18), it apparently ramps down to deeper structural levels along strike, i.e. to the NE. As can be shown on a number of strike sections, the Répce fault plane describes a synform, the axis of which plunges to the SE. Note that this pronounced synclinal feature is remarkably displayed on section M18. Looking at several dip lines, the Répce fault has a pronounced "spoon" shape in the basement with maximum displacement along the long axis of the spoon. Interestingly enough, the Répce fault plane climbs up in terms of physical depth farther to the NE, but it ramps down in a tectonostratigraphic sense into the Lower Austroalpine unit (section M18).

least the Fertö and Répce faults maintain their low-angle dip (~30-40°) to mid-crustal depth. Moreover, the Fertö fault (also called Balf fault by Fodor, 1991) appears to merge at depth into a prominent surface which I interpret as the base of the Austroalpine nappe complex. The mid-crustal geometry of the Rába fault is not clear because the data deteriorate in the southeastern part of the crustal profile.

The average dip (~5°) of the European foreland in the northwestern part of the transect is well constrained by the wells Raipoltenbach-1 and Berndorf-1. Extrapolating this dip to the SE, the top of the European foreland can be tied into the northwestern end of the crustal profile MK-1 given in Encl. 1. A prominent reflection doublet at 4.2 s TWT time (~11 km depth) is interpreted as originating from the boundary between the autochthonous European foreland crust and the overriding Alpine nappes. This reflection event can be correlated further to the SE with a slightly steeper dip (~10-15°), to about 20 km depth, beneath the Mihályi high. Beyond this point, poor seismic data quality does not permit to follow this surface farther to the SE (Encl. 1). Subhorizontal to slightly SE-dipping strong reflector packages below this interface are thought to emanate from crystalline rocks of the European foreland crust. Beneath the Pinnye high, at about 6 s TWT time (~14-18 km), some strong NW-dipping reflectors are tentatively interpreted as being related to a 10-15 km wide Mesozoic (Jurassic?) half-graben located in the distal parts of the European passive margin. The boundary fault of this half-graben dips toward the Jurassic Penninic ocean, located some distance to the S.

Regarding the depth of the Moho discontinuity along the transect, the map of Posgay et al. (1991) shows this surface in an elevated position at 26 km beneath the Mihályi high. In the central part of the crustal seismic section (Encl. 1) very low frequency reflectors between 9-10 s TWT may correspond to the Moho. Below the Eastern Alps the depth of the Moho descends to 35 km and more (Meissner et al., 1987).

In the southeastern part of the Danube Basin, the pre-Tertiary basement displays a general monoclinal dip to the NW (Fig. 2). Here the Eoalpine structures are reasonably well-known near the top of the pre-Senonian basement, based on the interpretation of the industry seismic profiles (Tari, 1995a). There are a number of NW-verging nappe structures (Fig. 4) that display an Eoalpine deformational style similar to that of the Upper Austroalpine nappes of the Eastern Alps.

The Bakony Mts. consist of the Devecser and Halimba synclines which are superimposed on a much larger synclinal feature trending to the NE. These synclines appear to float above two regional thrust surfaces which can be correlated from outcrops in the Balaton Highland to their subcrop in the Danube Basin (Tari, 1994). Whereas the structurally higher Veszprém thrust is associated with a Carnian detachment level, the deeper Litér thrust generally follows a Middle Triassic detachment level. As these thrusts do not have a clear seismic expression at depth, the interpretation given in Encl. 3 must be considered as conceptual.

To constrain the deep structure of the transect across the Bakony Mts., another crustal line was projected into the regional structure section from some 60 km to the NE (Tari , 1994). The overall synclinal geometry of the Bakony Mts. might be caused by two NW-verging deep thrusts, involving the crystalline basement of the Middle and Lower Austroalpine units. Probably these deep thrusts are responsible for the along-strike appearance of Palaeozoic rocks in the the Velence area, as shown in Figs. 2 and 3.

In the transect, the correlative anticlinal structure under Lake Balaton is distorted because of early Miocene activity along the Balaton "Line". Unfortunately this fault does not have a clear seismic expression. Its reverse fault character at shallow depth was documented by drilling (e.g. Balla et al., 1987; Körössy, 1990). In the transect, the Balaton fault was placed immediately to the N of the Karád wells which bottomed in Dinaric-type Palaeozoic carbonates (Bérczi-Makk, 1988). It is postulated here that the Balaton fault flattens at depth into the base of the Austroalpine nappe system. As an indirect argument for the flattening of the Balaton Line at lower crustal levels an analogy with the the Insubric Line of the Alps is invoked. Heitzmann (1987) and Schmid et al. (1987) demonstrated that the Insubric Line flattens northward at a mid-crustal level and appears to be the westernmost extension of the Periadriatic-Balaton Line system.

Summing up, the most striking result of our transect is that the European foreland crust dips gently (5-10°) beneath the Eastern Alps and extends over a distance of at least 150 and possibly as much as 200 km into the area of the Danube Basin. This supports the earlier held view of Wessely (1987). Although this finding is in keeping with the Central Alpine transect (see Ziegler et al., this volume), it is at odds with the results of a deep crustal profile across the northern Carpathians (Tomek and Hall, 1993) which images an abrupt steepening of the European foreland crust to 70-80° beneath the external Carpathians in Slovakia, about 150 km to the East of the Vienna Basin (see Fig. 1).

DISCUSSION

It appears that during the last decade the role of middle Miocene syn-rift strike-slip faulting in the opening of the Pannonian Basin was overemphasized (e.g. Tari, 1988). Further confusion arose from the fact that throughout the Pannonian Basin numerous flower structures were observed within its post-rift sequence. These structures, however, are related to very recent, and in some cases even currently active fault zones; therefore they cannot be easily assigned to the middle Miocene syn-rift stage of the Pannonian Basin. Instead, they suggest later transpressional deformation of this basin, resulting in the development of local inversion structures (Tari, 1994).

The same holds true for the NW Pannonian Basin as well: seismic evidence clearly shows that low-angle normal faults play an eminent role in the distribution of the different Eoalpine units in subcrop of the Neogene Danube Basin (Fig. 4). Although some oblique-slip movements may have occurred along these detachment fault planes, it must be emphasized that these features are **dominantly** extensional.

The alternate structural model which is proposed here invokes dominantly low-angle normal faulting and opposes the earlier model of dominant strike-slip movement along subvertical faults (e.g. Balla, 1994). The extensional model was tested along the structural transect given in Encl. 3 by gravity modeling (Szafián and Tari, 1995). Preliminary results show that the model based on middle Miocene extensional detachment faulting is indeed viable as a good match was found between the observed and calculated gravity anomalies.

Earlier I addressed the problem of the widely held **assumption** that Cretaceous overthrust planes were reactivated during the Miocene as low-angle normal faults in the Pannonian Basin (e.g. Grow et al., 1989; Tari et al., 1992; Horváth, 1993). Based on the seismic illustrations given in this paper (see also Tari and Horváth, 1995; Tari, 1995a) it is now clear that Neoalpine low-angle normal faults did indeed interact with abandoned Eoalpine thrust fault planes. Such interaction, however, seems to have been more complex than had been anticipated.

The following three ways of interaction between extensional faults and pre-existing thrust faults can be visualized (Fig. 6):

(1) an earlier thrust plane is extensionally reactivated over its entire length (e.g. Ratcliffe et al., 1986),
(2) a newly formed, relatively steeply dipping normal fault soles out at depth into an originally compressional, tensionally reactivated detachment level (e.g. Bally et al., 1966) and
(3) newly formed, extremely low-angle normal faults cut through pre-existing thrust faults and ramp-anticlines (e.g. Wernicke et al., 1985).

Ivins et al. (1990) reviewed the factors, such as geometry, intact/pre-existing fault strength and fluid pressures, which determine whether extensional reactivation will or will not occur, although they did not specifically study the above described geometries.

In the NW Pannonian Basin, reactivation of Cretaceous thrust planes by middle Miocene low-angle normal faults occurred dominantly in the second manner and less typically in the first manner. The geometry where shallow thrusts are cut by

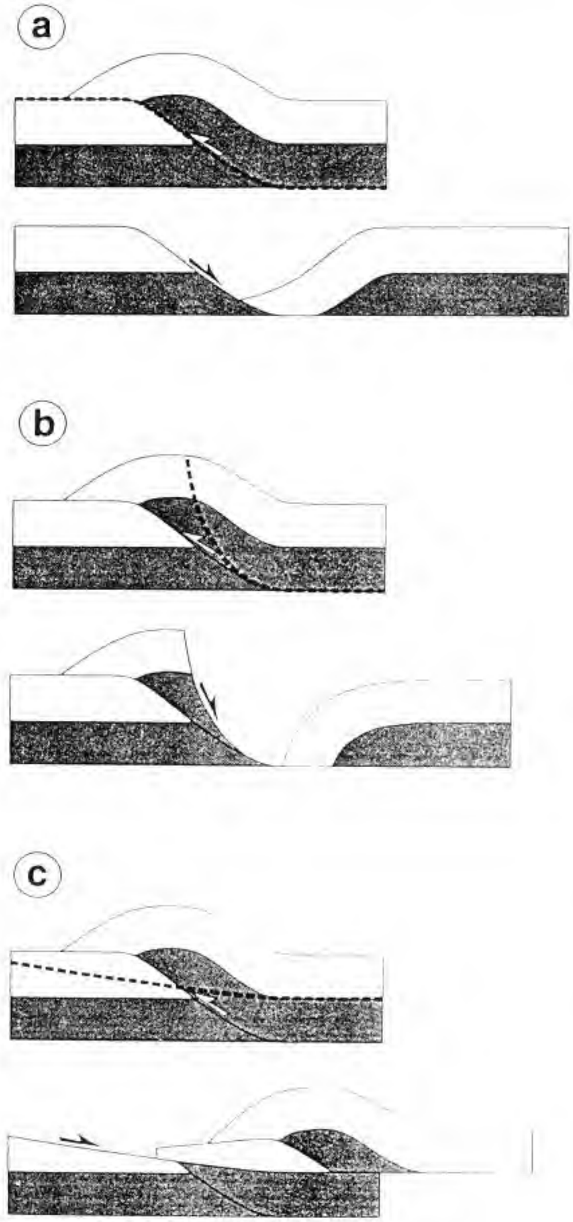

FIG. 6. Different modes of reactivation of pre-existing thrust faults during low-angle normal faulting. In the NW Pannonian Basin mode "b" is the predominant, whereas locally mode "a" can also be observed.

extremely low-angle normal faults was not observed.

The conclusion is that even though reactivation occasionally occurred, Cretaceous overthrust and Miocene extensional detachment fault planes rarely coincide at the top of the pre-Neogene basement or at shallow depth (i.e. <1 km) beneath it. That is why in map view the Répce and Rába faults appear to "ignore" the Austroalpine nappe contacts (Fig. 4). Towards deeper intra-basement levels, however, pre-existing thrust and subsequent low-angle normal faults frequently merge into each other (see also seismic illustrations by Tari, 1995a; Tari and Horváth, 1995).

CONCLUSIONS

The apparent lack of any syn-rift strike-slip structures and the presence of several major low-angle normal faults in the NW Pannonian Basin suggest a primarily extensional origin for the entire Danube Basin as opposed to the traditionally held pull-apart basin interpretation.

Based on the evaluation of reflection-seismic data, the European foreland crust dips at an angle of 5-10° beneath the Alpine/Pannonian junction and extends at least 150 km and possibly as much as 200 km from the Alpine deformation front under the Danube Basin.

The compressionally pre-conditioned "memory" of the basement of the Neogene NW Pannonian Basin influenced the geometry of the subsequent continental extension; partial reactivation of pre-existing regional compressional décollement levels guided the geometry of the newly forming extensional faults.

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Enclosures

Encl. 1. MK-1 crustal reflection profile from Posgay et al. (1986) and its recent reinterpretation by Tari (1994). For location see Fig. 4. See also Encl.3 for the depth-converted geologic section along this seismic profile.

Encl. 2. Line drawing interpretations of the C1, C3, C5 and M18 reflection-seismic sections, adapted from Tari (1994). For location see Fig. 4.

Encl. 3. Regional structure transect across the NW Pannonian Basin by Tari (1994). For location see Figs. 1 to 4. The Austrian part of the transect is based on a geologic section published by Wessely (1988). The middle part of this section is based largely on the deep reflection-seismic profile shown in Encl. 1.

Structural-stratigraphic evolution of Italy and its petroleum systems

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ABSTRACT

The geological evolution of Italy was controlled by Triassic-Early Jurassic rifting, culminating in the separation of the European, African and Adriatic plates, by their interaction during the Mid-Jurassic-Early Cretaceous opening of Tethys and the Mid-Cretaceous and Cenozoic closure of Tethys. During the break-up stage, a complex system of carbonate platforms and intervening troughs developed on the Adria plate. These contain laterally discontinuous Middle Triassic to Early Jurassic oil source-rocks. Cretaceous to Eocene subduction of oceanic basins and Oligocene to Recent continental collision governed the development of the Alpine and Apennine orogens. In foredeep basins associated with these evolving fold- and thrustbelts, thick Neogene flysch successions were deposited. These contain some oil source-rocks as well as sizable amounts of biogenic gas.

Italy's URR amount to $131 \cdot 10^6$ t of oil and condensate and $743 \cdot 10^9$ m^3 gas. The oil and gas fields discovered in Italy can be grouped, according to charge-providing source-rocks and processes controlling their maturity and trap development, into several petroleum systems. Definition of some of these systems remains, however, tentative due to insufficient data.

The main oil fields, accounting for 13% of Italy's URR, were charged by Triassic-Jurassic source-rocks. These attained maturity during the Late Neogene in rapidly subsiding foreland basins. Hydrocarbons generated charged by prevailingly vertical migration block-faulted foreland structures and anticlinal features of the external thrustbelts. The charge factor of Triassic source-rocks is low. The retention capacity of seals is generally limited.

Miocene flysch is the source of thermal gas and light oil in the Apennine thrustbelt. Moreover, Miocene flysch series contain minor and marginal accumulations of bacterial gas in the eastern parts of the southern Alps, the southern Adriatic Sea and in western Sicily. The bulk of Italy's hydrocarbon reserves consists of bacterial gas contained in Pliocene and Pleistocene flysch. Although bacterial gas accumulations occur in many basins and in different traps, optimal conditions prevail in the northern Apennine foredeep which is characterized by high sedimentation rates, multiple reservoir/seal pairs provided by highly efficient turbidites and synsedimentary trap forming conditions.

List of abbreviations: HI=hydrogen index, HC=hydrocarbons, SPI=source potential index,

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This article includes 3 enclosures on 1 folded sheet.

TOC=total organic carbon, URR=ultimate recoverable reserves

INTRODUCTION

Italy's on- and off-shore basins and external fold- and thrustbelts host a large number of hydrocarbon plays. Many of the already discovered hydrocarbon accumulations can be related to specific source-rocks. The interdependence of factors and processes controlling the formation of hydrocarbon accumulations has been discussed in previous reviews of Italy's petroleum geology (Pieri and Mattavelli, 1986; Riva et al., 1986; Mattavelli and Novelli, 1988, 1990; Mattavelli et al., 1993; Zappaterra, 1990, 1994)

Since 1944, about 2500 exploration wells were drilled in Italy's on- and off-shore plays. These resulted in the discovery of numerous oil and gas fields (Fig. 1) having cumulative URR amounting to $131 \cdot 10^6$ t (950×10^6 bbl) of oil and condensate and $743 \cdot 10^9$ m^3 (27.6 TCF) gas (cut off date 31.12.1994). Italian fields produced in 1994 a total of $4.9 \cdot 10^6$ t of oil and $20.6 \cdot 10^9$ m^3 gas. This approximately corresponds to 14.1% of Italy's energy requirements.

This paper is mainly, but not exclusively, based on the data and conclusions previously published by the authors and their colleagues of Agip S.p.A. The stratigraphic and structural evolution of Italy is reviewed in a geodynamic framework with special emphasis on the source-rock habitat. A preliminary classification of Italy's petroleum systems, referring to examples of commercial accumulations, is presented. However, in some cases, the paucity of information on source-rocks does not permit an adequate definition of the respective charge system.

GEODYNAMIC EVOLUTION OF ITALY

The Mesozoic and Cenozoic stratigraphic and structural record of the Italian sedimentary basins and fold- and thrustbelts has greatly contributed to the understanding of the geodynamic evolution of the Central Mediterranean area, main stages of which were the Triassic-Jurassic break-up of Late Palaeozoic Pangea, resulting in the opening of the Tethys system of oceanic basins, the separation of the Eurasian, African and Adriatic (also referred to as Apulia or Italo-Dinarid) plates, and the development of passive margins, followed by the Cretaceous-Cenozoic convergence and collision of Africa and Europe, causing the deformation of the intervening Adria plate and the development of the Alpine and Apennine orogens (Biju-Duval et al., 1976; Laubscher and Bernoulli, 1977; Tapponier, 1977; Bernoulli et al., 1979a, 1979b; Dercourt et al., 1985, 1986; Ziegler, 1988, 1990; Dewey et al., 1989; Boccaletti et al., 1990).

In the following we retrace the Triassic to Neogene evolution of Italy in a plate-tectonic framework. The following four stages can be distinguished (Fig. 2):

Permo-Triassic Rifting Stage (Fig. 3a)

At the end of the Hercynian (Variscan) orogeny, during which the Pangea super-continent was consolidated, a tensional tectonic regime prevailed in the Central Mediterranean area. Initially continental clastics were deposited in incipient rifted basins. As these were gradually invaded by the transgressing Tethys Sea, evaporitic series were deposited. This was followed by the establishment of a complex system of carbonate platforms and intervening deeper water troughs. By Late Triassic times, the Apulia Platform was flanked to the East by the Olenus-Pindos and Sub-Pelagonian and to the West by the Lagonegro deeper water troughs, and to the North by the South-Alpine, Austroalpine and Piedmont rift systems.

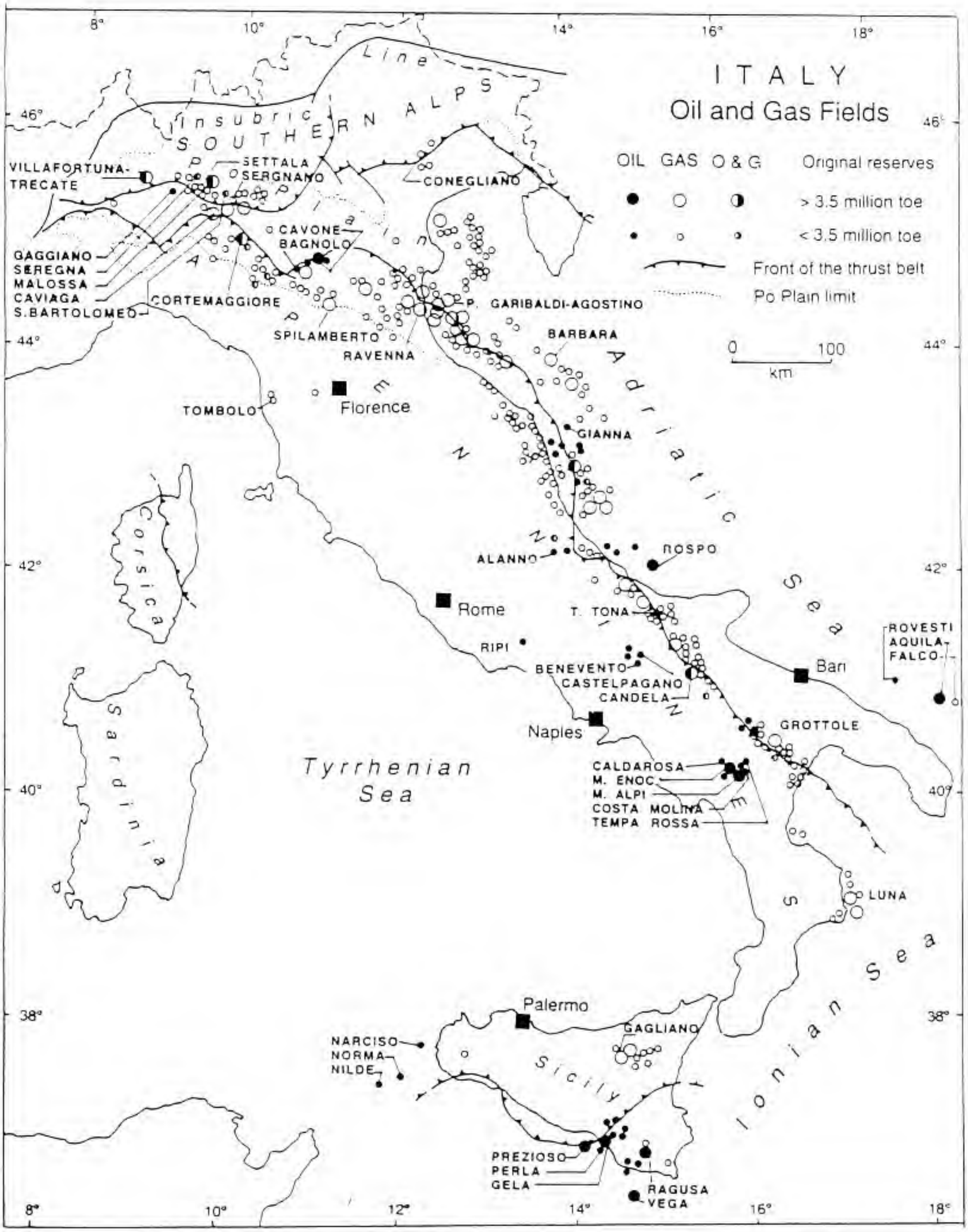


FIG. 1. Italy: oil and gas fields. Fields mentioned in the text are named.

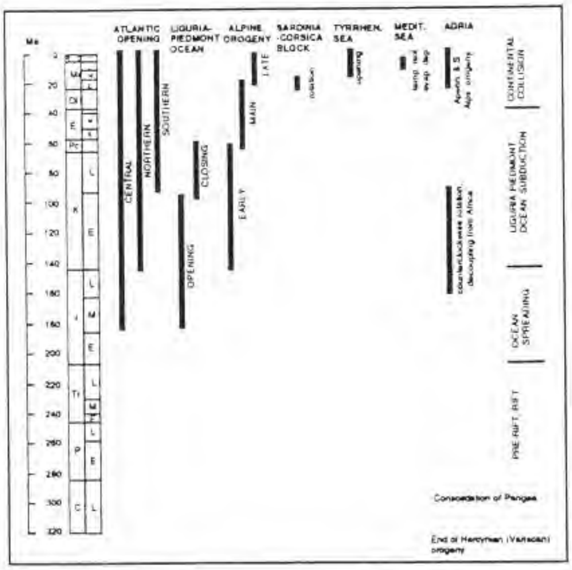


FIG. 2. Chronologic chart of the main geodynamic events in the Mediterranean and Atlantic areas.

Jurassic Sea-Floor Spreading Stage

Rifting activity continued during the Early Jurassic and culminated in early Mid-Jurassic crustal separation between the Adriatic and European plates and the transtensional opening of the Alboran-Ligurian-Piedmont-Penninic ocean. During the Late Jurassic, the oceanic Vardar Basin began to close in response to sinistral translation of Africa relative to Europe, induced by the progressive opening of the Central Atlantic. At the same time, the relatively small Adriatic plate was decoupled from Africa along a transform shear zone and began to rotate counter-clockwise. On platforms carbonate deposition continued throughout Jurassic times. However, rift-induced further break-up of platforms and their post-rift subsidence led to the establishment of additional deep basins characterized by carbonate, shaly and cherty sediments.

Cretaceous-Eocene Subduction Stage (Figs. 3b and 3c)

Subduction processes commenced in the Central Mediterranean domain already during the Late Jurassic onset of closure of the Vardar Ocean. During the Cretaceous, rapid opening of the North Atlantic caused rotation of the Adriatic block and the initiation of a transform subduction zone along its northern margin (Ziegler et al., this volume). With the Senonian onset of counter-clockwise convergence of Africa-Arabia with Europe, suduction zones rapidly propagated into the Western Mediterranean. Gradual closure of the Ligurian-Piedmont-Penninic Ocean went hand in hand with progressive uplift of the Alpine and Apennine ranges and the shedding of flysch into remnant oceanic basins and gradually evolving foreland basins.

Oligocene to Recent Continental Collision Stage (Figs. 3c and 3d)

Following subduction of the Ligurian-Piedmont-Penninic Ocean, the Adriatic plate was framed by the continent-to-continent collisional Dinarid, Alpine and Apennine orogens. Its passive margin sedimentary prisms were overridden by nappes, consisting of oceanic crustal slices and their deep-water sedimentary cover, and were themselves detached along basal, mostly evaporitic, levels. Thrust-loaded subsidence of foredeep basins, paralleling the evolving Alps and Apennines, provided accommodation space for thick syn-orogenic flysch series. Progressive migration of the thrustbelt-foredeep systems towards the foreland was accompanied by drowning out of the latter and the deposition of thick clastic wedges, consisting of flysch series grading upwards into shallow water and partly alluvial deposits. Kinematic considerations suggest, that post-Tortonian compressional features developing along the external front of the Apennine, are the result of active thrusting, combined with passive subsidence of the foreland lithosphere (Patacca and Scandone, 1989; Scandone et al., 1992).

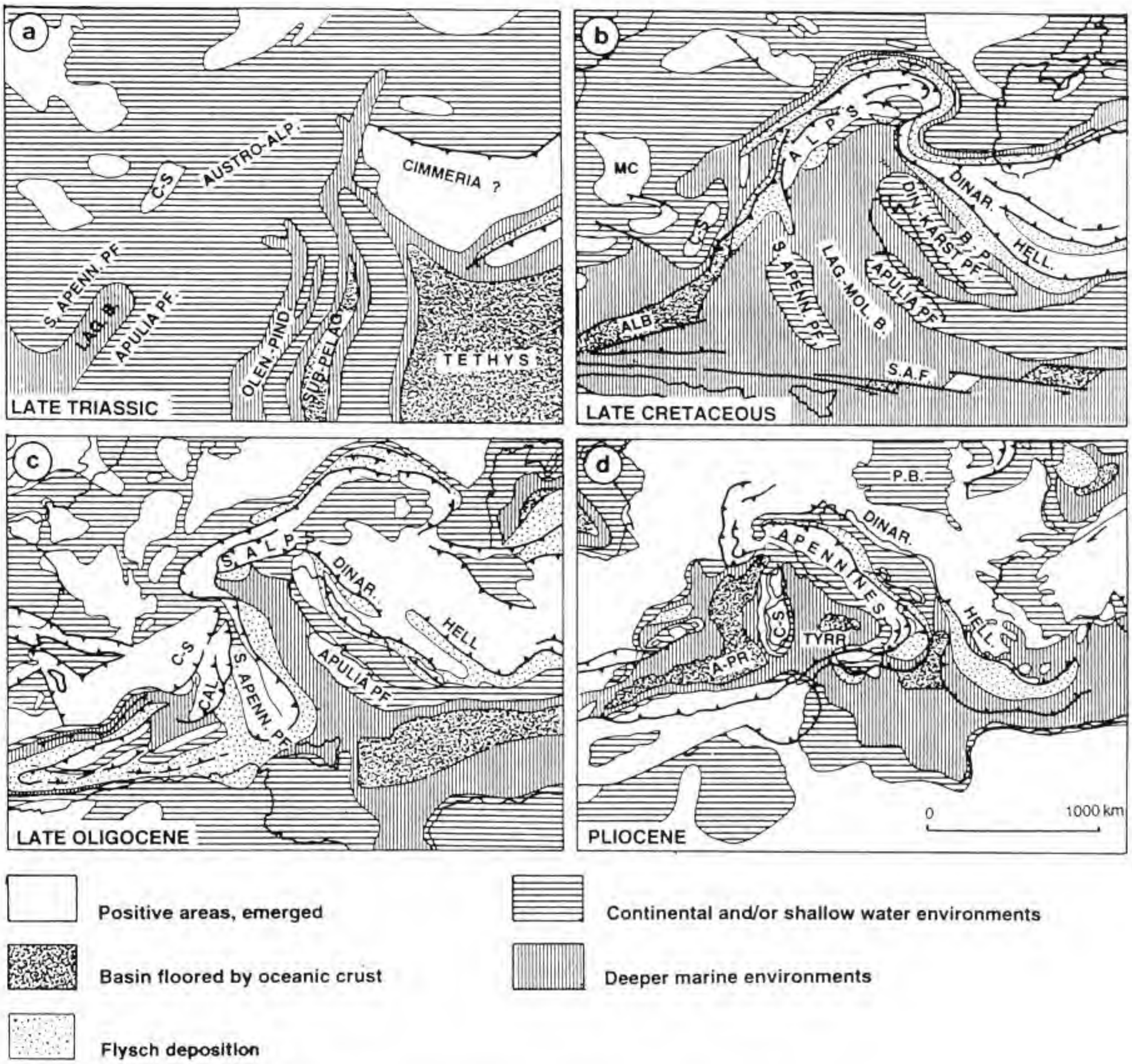


FIG. 3. Paleogeographic-geodynamic maps (general framework simplified after Ziegler (1988) and modified for the central Mediterranean area). LAG. B: Lagonegro Basin; C-S: Corsica-Sardinia block; OLEN.-PIND: Olenos-Pindus Basin; SUB PELAG: Sub Pelagonian basin; MC: Central Massif; S.A.F: South Anatolian Fault Zone; ALB: Alboran (Liguria-Piedmont) ocean; DINAR: Dinarides; HELL: Hellenides; DIN.-KARST PF: Dinarides-Karst Platform; APULIA PF: Apulia Platform; LAG.-MOL: Lagonegro-Molise Basin; S. APENN. PF: S Apennine Platform; P.B: Pannonian Basin; TYRR: Tyrrhenian Basin; A-PR: Algero-Provençal Basin.

From Late Miocene times onward, the Tyrrhenian and Peri-Tyrrhenian areas were subjected to extension, causing their collapse and the subsidence of the episutural Algero-Provençal and Tyrrhenian system of basins.

During Late Miocene times, temporary isolation of the Mediterranean Sea (Hsü et al., 1977) resulted in an evaporation-induced lowering of the sea-level and the widespread deposition of Messinian evaporites. As a consequence of earliest Pliocene re-opening of communications with the Atlantic Ocean, normal sea-levels and salinities were established again.

According to prevailing subsidence mechanisms, the sedimentary successions of the foreland, as well as those involved in the Apennine and South-Alpine thrustbelts, can be subdivided into the following tectono-stratigraphic sequences (Fig. 4):

(1) the **pre-rift and syn-rift sequence** commences with Permo-Triassic continental clastics, resting on Hercynian basement, which include Triassic evaporites and/or shales and carbonates, deposited prior to the regional Tethys transgression. It contiues with shallow marine carbonate banks which, during Early to Middle Jurassic times, may evolve into deeper water basins where shales and cherts are associated with carbonates,

(2) the **passive margin sequence** ranges in age from Middle Jurassic to Late Cretaceous and consists of shallow water carbonate banks separated by wide, deeper water basins. Local pelagic carbonate platforms, characterized by frequently condensed sequences, hardgrounds, episodes of submarine erosion and stratigraphic gaps, are present (Santantonio, 1994),

(3) the **foredeep sequence** generally commences with shales to shaly carbonate deposits, which reflect rapid flexural subsidence of the foreland to considerable water-depths. This initial transgressive unit grades upwards into flysch, supplied by clastics derived from the rising orogens. Depending on clastic supply and subsidence rates, flysch series can grade upwards into shallow marine and continental sediments (e.g. Po Plain Pliocene-Quaternary foredeep). As the axes of foredeeps migrated in time towards the foreland, facies boundaries are time transgressive. Moreover, synsedimentary compressional deformation of the proximal parts of foredeep basins, causing the development of sea-floor topographic anomalies, influenced the distribution of turbiditic sands; these accumulated preferentially in synclinal areas whereas anticlinal ridges wre characterized by hemipelagic sediments (Pieri and Mattavelli, 1986).

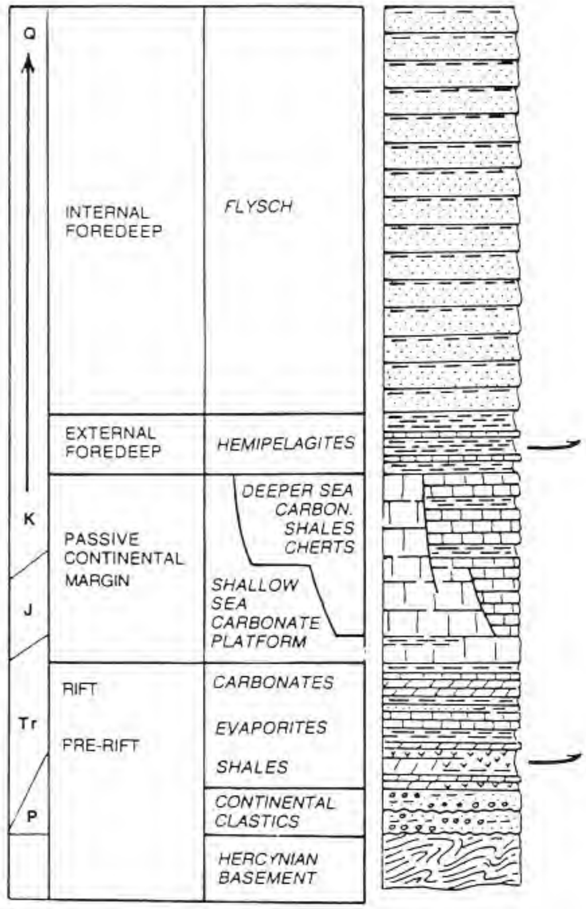


FIG. 4. General sedimentary succession of the Adria continental margin. Flysch deposition is strongly heterochronous. Arrows correspond to main detachment levels.

In areas corresponding to peripheral bulges (e.g. large parts of Puglia region, Fig. 5), the foreland is not incorporated into the foredeep basin and the entire Cenozoic sequence consists of discontinuous shallow-water carbonates.

STRUCTURAL AND STRATIGRAPHIC FRAMEWORK

The main structural units of Italy are the Alpine and Apennine fold- and thrustbelts, the foreland of which corresponds to the stable continental block of the Po Plain and Adriatic Sea. To the South, the Apennine finds its continuation in the Calabrian-Sicily arc. Its foreland is formed by the oceanic Ionian Sea and the continental Pelagian Shelf (Fig. 5). The structural cross-sections given in Enclosures 1-3 show that the autochthonous foreland crust extends up to 100 km under the Apennine thrustbelt and as much as 70 km beneath the Southern Alps.

The structural style and configuration of the Alpine and Apennine fold- and thrustbelts are controlled by the thickness and rheological composition of the sedimentary sequences involved in them, as well as by the geometry of the Mesozoic basins out of which they evolved. During the evolution of the Alps and Apennines, many of the main extensional faults, controlling the distribution of Mesozoic platforms and basins, were compressionally reactivated and often developed into main thrust faults (e.g. boundary between Northern and Southern Apennine). Correspondingly, the different tectono-stratigraphic units of these orogenic belts conform, with a few exceptions, to Mesozoic palaeogeographic zones.

Southern Alps

In the Southern Alps, the following four structural zones are distinguished (Doglioni and Bosellini, 1987; Consiglio Nazionale delle Ricerche, 1989):

(1) the South-Alpine **Lombardy fold and thrust arc** developed during the Late Cretaceous and Late Miocene out of a Triassic rifted basin (see Ziegler et al., this volume). Its external units emerged during the Messinian low-stand in sea-level and are sealed by undeformed Plio-Pleistocene sediments, attaining thicknesses of up to 2.5 km (Encl. 1, sect. 1)

(2) the **Verona (Lessini Mts.)** area was not affected by Alpine deformations and represents the outcropping part of the Po Plain foreland which is bounded to the Northwest by the frontal elements of Lombardy thrust belt and to the Northeast by the Schio transcurrent fault

(3) the WSW-ENE trending **Veneto folds and thrusts** were activated during the Miocene and are tectonically still active. Their southern front corresponds to the morphological boundary between the pre-Alps and the Veneto Plain (Encl. 1, sect.2)

(4) the NW-SE striking **Dinaric folds and thrusts** find their on-strike prolongation in the Dinarides and the Hellenides. Whereas crustal shortening in the northern Dinarides terminated in Mid-Miocene times, their southern parts and the Hellenides are still active to day.

The stratigraphic succession of the Southern Alps reflects the development of the northern Adria continental shelf. Late Permian to Early Triassic continental clastics and local evaporites are covered by Middle to Late Triassic carbonate platforms and intervening anoxic basins. During Jurassic times, deep water basins developed in the Lombardy and western Veneto areas whereas the Verona zone was occupied by a pelagic carbonate platform. In the East-Veneto, and the Dinaric zone, shallow-water carbonate platforms persisted into Late Cretaceous times. In the Lombardy Basin, Late Cretaceous and Oligo-Miocene flysch, records the gradual uplift of the Southern Alps

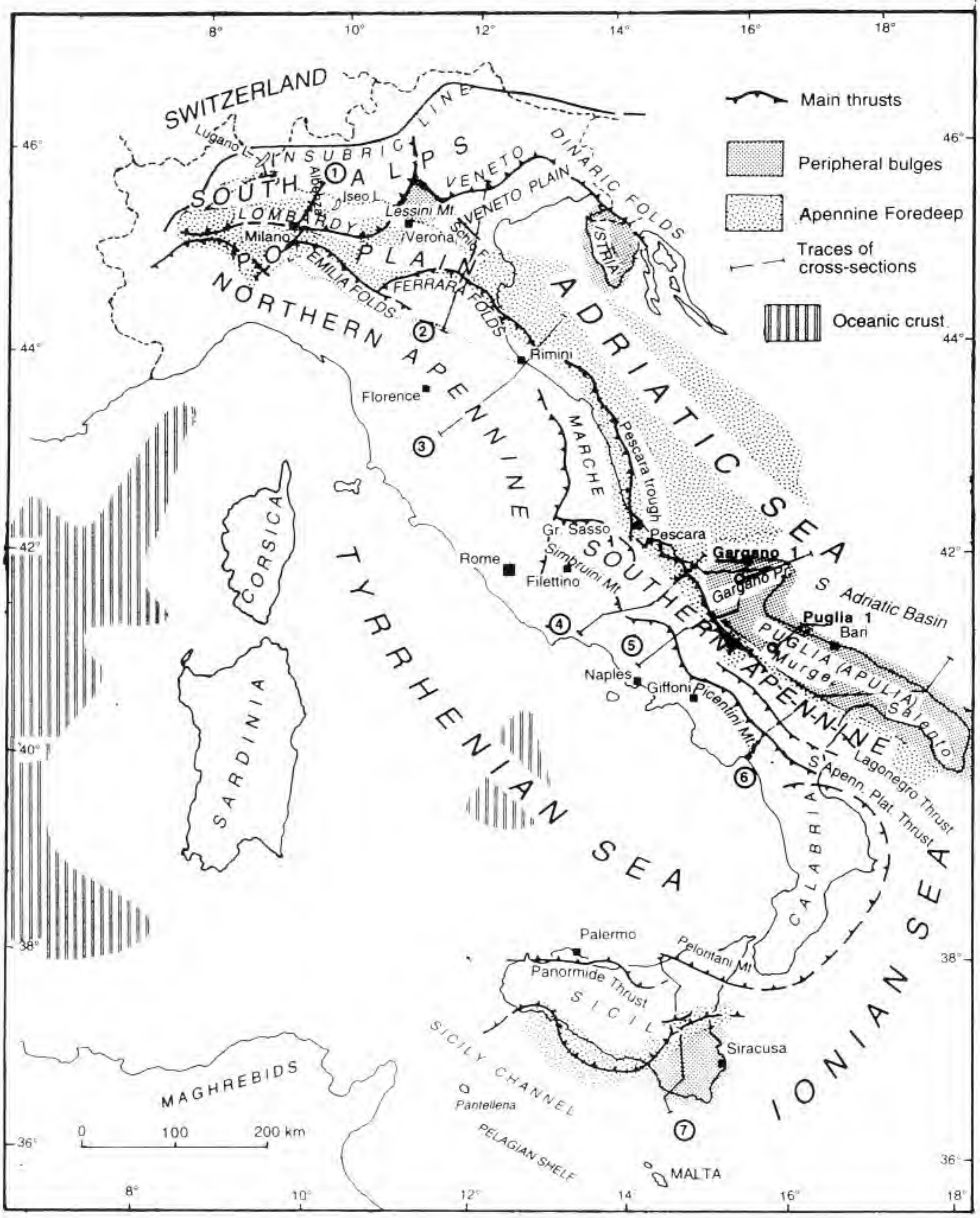


FIG. 5. Italy: main structural units. Traces of sections iven in Encl. 1, 2 and 3.

(Ziegler et al., this volume), whereas in the Veneto and Dinaric zones flysch sedimentation commenced only in the Eocene. In contrast, Paleogene-Miocene strata are represented on the Verona High by shallow-water carbonates.

Apennine

The Apennine, which evolved on the western passive margin of the Adriatic plate, can be subdivided into three main sectors which are bounded by major thrust fronts. The internal parts of the Apennine were affected by Neogene extensional faults related to the opening of the Tyrrhenian Sea.

In the area of the **Northern Apennine**, the Mesozoic continental shelf was characterized by a relatively uniform stratigraphic succession, consisting of a Late Triassic continental clastic and evaporitic-dolomitic series, earliest Jurassic shallow-water carbonates and mid-Early Jurassic to Paleogene deep-water carbonates. The transition to flysch accumulation occurred on the distal shelf, corresponding to the internal Apenninic unit, during the Late Oligocene and progressed during the Miocene and Pliocene into the domain, now represented by the northeastern and eastern external Apenninic units and ultimately into the area of the present foredeep basins.

During the Alpine orogenic cycle, these sedimentary sequences were detached from their basement at the level of the Late Triassic Burano evaporites (Encl. 1, sect. 3). In the Emilia and Tuscany regions, the internal parts of the Adria passive margin successions are widely covered by the Ligurides nappes which consist of obducted ophiolitic fragments and deep-water sediments, deposited in the Liguria-Piedmont oceanic basin.

The **Southern Apennine** consists of four tectono-stratigraphic units which differ in the composition of their Mesozoic series. Each unit corresponds to a major nappe. During the stacking of these nappes, extensive detachment of flysch units from their carbonate substratum occurred. In comparison with the Northern Apennine, the southern one is characterized by a more complex architecture and a greater amount of shortening (Encls. 2 and 3, sect. 4, 5, 6).

The tectonically upper-most, and therefore the most internal unit, is composed of flysch sequences attributed to the possibly oceanic Ligurides domain. The next lower unit, corresponding to the South-Apennine platform, consists of Mesozoic-Paleocene platform carbonates, Early Miocene carbonates and Middle Miocene flysch. It overlays a unit which is derived from the Lagonegro trough and consists of Mesozoic-Paleogene-Lower Miocene calcareous, cherty and shaly sequences, followed by Middle Miocene quartzarenitic flysch. The lower-most unit, and therefore the most external one, is analogous to that of the Puglia foreland and is characterized by thick platform carbonates ranging in age from Jurassic to Cretaceous, which are unconformably covered by thin and discontinuous Paleogene and Miocene shallow water carbonates. The Cretaceous-Paleogene sequence may however change westward to deeper water carbonate sediments.

The frontal thrust sheets of the Southern Apennine involve allochthonous Miocene flysch and fill the foredeep which is limited to the N-E by the Puglia foreland.

The southern-most sector of the Apennine, corresponding to the Calabrian arc, is characterized by the internal Calabria-Peloritani nappes which involves Hercynian basement; this nappe is possibly derived from the northwestern margin of the Alboran-Ligurian Ocean.

Current palaeogeographic reconstructions place the deep-water Lagonegro Basin between the Apulian and South-Apennine carbonate platforms (Pieri, 1966; D'Argenio et al., 1973; Mostardini and Merlini, 1988). An alternate interpretation proposes that the Lagonegro Basin was located to the west of the South-Apennine platform, was thrusted during the Langhian-Tortonian over this platform and was enveloped during the Messinian-Pliocene by the South-Apennine nappe (Marsella et al., 1992).

The **Sicily-Apennine** occupies the northern and central parts of the island and links up through the the Sicily Channel with the Maghrebides of North Africa (Encl. 3, section 7). The stratigraphy of the external units of the Sicily Apennine and their tectonic relationship with the autochthonous foreland are still poorly known. Such units may occur beneath the more internal Imerese and Panormide nappes and may be characterized by a similar sedimentary succession as seen in the foreland of southeastern Sicily. The Imerese nappe consists of sedimentary sequences which are similar to the Lagonegro unit of the Southern Apennine. The Panormide units consist of Late Triassic to Early Cretaceous shallow-water carbonates. The most internal units, representing a continuation of the Ligurides and the Calabria-Peloritani nappes, occur in northeastern Sicily.

In general, the tectono-stratigraphic units of the Southern Apennine can be correlated with those of the Sicily-Apennine. The most external units of the Sicily-Apennine consist of Miocene flysch and post-flysch clastic successions which are detached from their substratum and fill a Pliocene-Quaternary foredeep.

Foreland of the Southern Alps and Apennine

The Mesozoic-Paleogene-Miocene series of the foreland in the subsurface of the Po and Veneto plains correlate with those of the external units of the Southern Alps (Pieri, 1984). In the northeastern Adriatic Sea, the Dinaric succession is recognized, whilst from Rimini to Pescara and in the South-Adriatic the composition of the foreland sequences does not significantly differ from those of the external Northern Apeninne. In this area, flysch sedimentation commenced in a very broad foreland basin during the Pliocene and grades upwards into Pleistocene deltaic series.

South of Pescara, the Apennine foredeep basin is separated from the Adriatic Basin by the Puglia (Apulian) Platform which can be regarded as a peripheral bulge that was affected by Plio-Pleistocene normal faults. According to the results of the deep wells Puglia-1 (TD 7070 m) and Gargano-1 (TD 4853 m), this platform consists of thick Cretaceous and Jurassic shallow-water carbonates, Late Triassic Burano anhydrites and dolomites and Middle to Early Triassic and Permian carbonates and clastics.

The Sicily foreland succession consists of Late Triassic-Early Jurassic shallow-water carbonates and interspersed deeper-water troughs in which the organic-rich carbonates and shales of Noto and Streppenosa Formations were deposited, representing important source-rocks. Middle Jurassic to Eocene series consist of deep-water cherty limestones. These are overlain by calcareous-marly Oligocene to Miocene strata. Late Cretaceous to Late Miocene shallow-water deposits are only known from the Siracusa area. During the Neogene the Pelagian Shelf was transsected by the northwest striking Pantelleria rift system.

SOURCE-ROCK HABITAT

Middle and Late Triassic rift-induced subsidence of often limited inter- and intra-platform deeper-water troughs, characterized by poorly oxygenized, stagnant bottom waters, was favourable for the deposition and preservation of organic-rich shales and carbonates and thus the accumulation of oil-prone source-rocks. Similar conditions developed also in intra-platform subtidal ponds and lagoons. About 90% of the oil tapped in fields so far discovered has been generated by Middle and Late Triassic source-rocks (Mattavelli and Novelli, 1990). A similar setting is indicated for the Late Triassic and Early Jurassic Noto and Streppenosa source-rocks of Sicily.

During Middle and Late Jurassic and Cretaceous sea-floor spreading and ocean subduction stage, several anoxic events occurred, such as the one related to the regionally recognized basal Turonian Bonarelli Bed (Farrimond et al., 1990). Due to limited thickness, these organic-rich intervals cannot be regarded as effective source-rocks. However, in the Southern Apennine an effective Cretaceous source-rock (unknown in outcrops), may have generated the oils trapped in the Costa Molina and related fields (Fig. 1).

During the Cenozoic collisional stage, no euxinic enviromments developed. Nevertheless, preservation of mainly terrestrial organic matter in the distal parts of Neogene turbiditic fans favoured the generation of light oils in the Apennine thrust belt and bacterial gas in the foredeep (Mattavelli and Novelli, 1987; Mattavelli et al., 1992a).

Sedimentological and geochemical characteristics of Italian source-rocks have been extensively analyzed and discussed (Pieri and Mattavelli, 1986; Mattavelli and Novelli, 1987, 1988, 1990; Brosse et al., 1988; Stefani and Burchell, 1990, 1993; Zappaterra, 1994). Here we discuss only the geochemical characteristics of source-rocks which play an important role in the hydrocarbon habitat of Italy.

Mesozoic Source-Rocks of the Southern Alps and Po Plain

The most important source-rocks are the Middle Triassic Besano (Grenzbitumen zone) and the Meride formations which crop out in the western part of the Southern Alps near the border between Italy and Switzerland (Fig. 6).

The up to 16 m thick Besano Formation is characterized by alternating laminated dolomites and black shales. It is an excellent source-rock with an average TOC content of 12% and maximum values of 40% (Fig. 7a). The kerogen, mainly amorphous organic matter (75%, Fig. 7b), has a high generation potential averaging about 60 kg HC/t and ranging up to 200 kg HC/t.

The overlying 300 m thick Meride Formation consists to 83% of limestones and to 17% of argillaceous limestones, marls and back shales and has an average TOC content of only 0.65%. Limestones may be considered as lean source-rocks, whereas argillaceous intercalations yield average TOC values of 2.4%. The organic matter consists to 75% of land-plant material.

The Besano and Meride Formations are the source for oils reservoired in the deep Mesozoic carbonates (>6000 m) of the major Villafortuna-Trecate field. In the past the role played by Middle Triassic source-rocks in the hydrocarbon habitats of the Po Plain and the Southern Alps was underestimated. Stefani and Burchell (1993) considered the clay-rich Rhaetian series as the main oil contributors. However, it is now realized that the Besano and Meride Formations are the most important source-rocks of Northern Italy.

These formations were deposited in the Lombardy intra-carbonate platform basin, which covers about 1200 km^2 and subsided in response to Late Anisian-Ladinian crustal extension, accompanied by widespread volcanic activity (Bernasconi and Riva, 1993; Ziegler et al., this volume). At depositional sequence scales, rapid increases in water-depths, either due to regional transgressions or accelerated tectonic subsidence rates, are often associated with enrichment in organic matter (Creaney and Passey, 1993; Stefani and Burchell, 1990; Katz and Pratt, 1993). Ample supply in land-derived nutrients inducing phytoplankton blooms, led to the development of eutrophic conditions and a reduced level in carbonate production. Accumulation of the highly organic Besano shales reflects severe anoxic bottom waters and a reduced influx of carbonates from adjacent platforms. In contrast, the Meride Formation, forming part of the same depositional sequence (Gaetani et al., 1991), reflects increased carbonate production on platforms from which frequent turbidity currents transported lime-muds into the Lombardy basins, thus causing dilution of the organic matter.

Late Triassic increased crustal extension caused disintegration of the widespread Norian carbonate platform into a system of highs and intervening troughs in which micritic limestones were deposited under anoxic conditions (Aralalta Group). These are covered by transgressive, argillaceous Rhaetian series, consisting of the basal Riva di Solto Shale and the upper Zu Limestone sequence (Fig. 7c Stefani and Burchell, 1990). The Riva di Solto Shales vary in thickness from 2 km in the Lake Iseo depocentre to less than 100 m on local palaeo-highs; lateral facies and thickness changes are related to syndepositional tectonics (Pieri and Mattavelli, 1986).

In outcrops, Late Triassic sediments are generally over-mature, except on long-lived palaeo-highs (see Table 1). Basal Rhaetian argillaceous deposits have average residual TOC content of 2-3% ranging up to maxima of 5%. A less important increase in TOC is observed in the upper parts of the Zu Limestone; samples of immature black shales on palaeo-highs yielded TOC values of 0.7-1.5% and have a generation potential of 1-3 kg HC/t. Both Rhaetian anoxic events correlate with transgressive episodes and are characterized by predominantly land-plant derived organic matter. As such they are gas- and gas-condensate prone source-rocks (Fig. 7b), as indicated by the Malossa gas-condensate field (GOR 1000; Mattavelli and Margarucci, 1992).

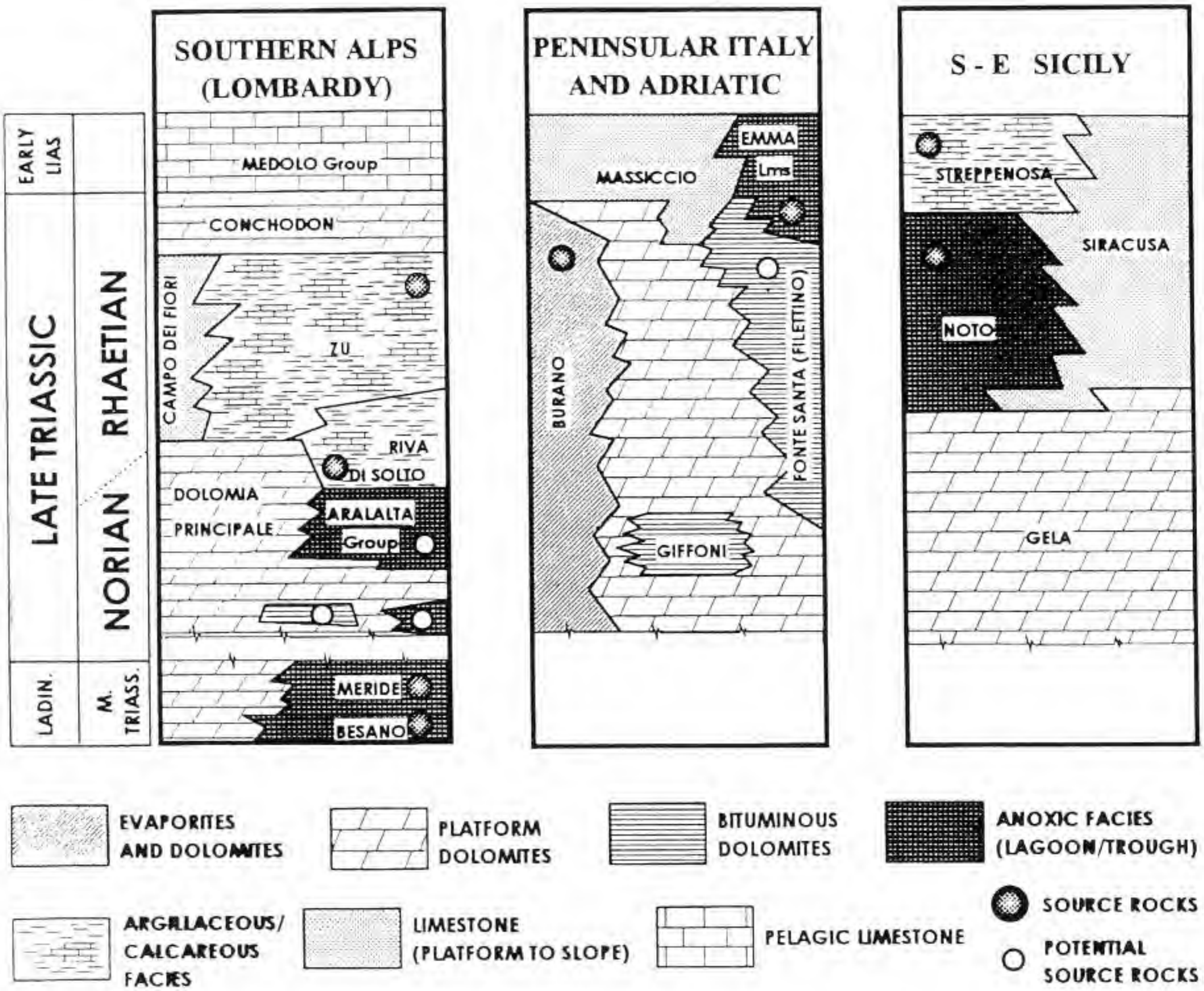


FIG. 6. Middle-Late Triassic, Early Lias stratigraphic charts of Southern Alps, Peninsular Italy - Adriatic and Southeastern Sicily. Relationships between the main litostratigraphic units and occurrence of Triassic effective and potential source rocks.

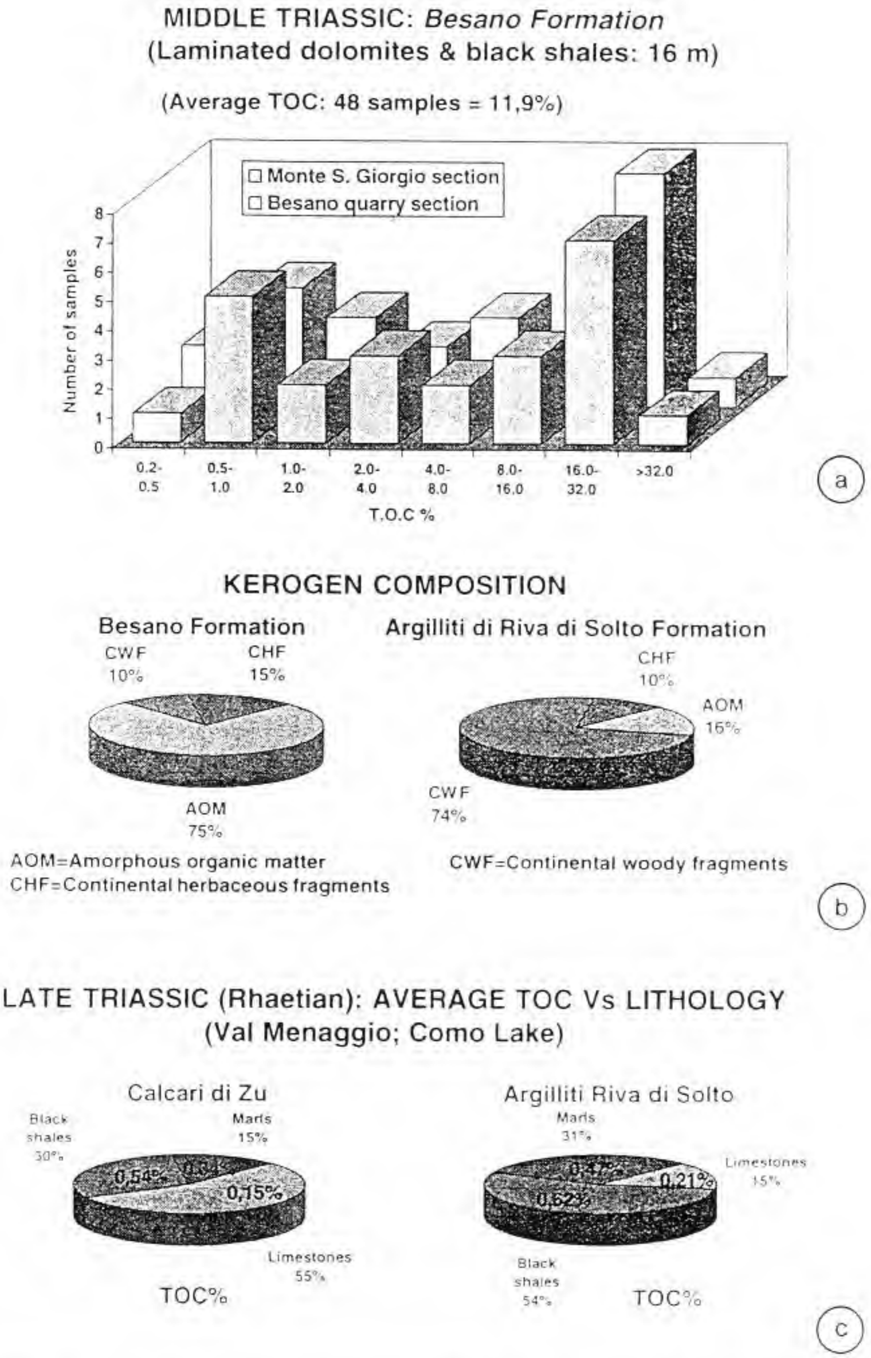


FIG. 7. a: Middle Triassic Besano Fm, S Alps. TOC vs number of samples of two significant outcrops in NW Lombardy.
b: Triassic source rocks in Southern Alps, Lombardy, kerogen composition of Besano (Middle Triassic), and Riva di Solto (Late Triassic) Fms. Amorphous organic matter prevails in Besano Fm, while continental woody fragments dominate in the Riva di Solto Fm.
c: Late Triassic Rhaetian sources in Southern Alps. Average TOC vs lithology in Zu and Riva di Solto Fms. Shale and marls are effective source rocks, while the limestones are mostly lean source rocks.

In a broader context, frequent occurrences of anoxic carbonates and shales have been reported from the Hauptdolomite platform of the Northern Alps (Müller-Jungbluth, 1968; Koster et al., 1988) and the time-equivalent Dolomia Principale of the Southern Alps (Jadoul, 1985).

Mesozoic Source-Rocks of the Southern Apennines and Adriatic Foreland

Reconstruction of the Late Triassic source-rock habitat in the Southern Apennine is hampered by the paucity of outcrops, an apparently complex palaeogeographic setting and the Neogene nappe structures (Ciarapica et al., 1987).

In the Simbruini Mts. (Filettino, 60 km E of Rome), Late Triassic platform carbonates change laterally to alternating dolomites, laminated dolomites containing very thin layers of marls and anoxic black shales. The distribution of organic matter is extremely heterogeneous. TOC values of gray dolomites are <0.1%, range in laminated dolomites between 0.4 and 3.2% and are >45% in centimetre thick shale layers (Fig. 8a). The kerogen is immature (R_0=0.4%), is predominantly of marine origin (Fig. 8b) and has a high HI (600-800 mg HC/g TOC). The average generation potential is 2 kg HC/t and exceeds 200 kg HC/t in shales.

In the Picentini Mts (Giffoni, 60 km SE of Naples), Late Triassic organic shales and laminated dolomites yielded a rich ichtyofauna (Boni et al., 1990). Organic-rich layers have an average TOC content of 4.5% and a generation potential of up to 572 mg HC/g TOC; the kerogen is mainly algal in origin. Sedimentological criteria indicate that this succession was deposited in a relatively shallow, subtidal lagoonal trough, surrounded by extensive carbonate platforms.

Similar anoxic Late Triassic successions are known from the Gran Sasso Range. This basin may extend to the northeast into the off-shore where the Late Triassic Emma limestone sourced the Gianna oil accumulation (Adamoli et al., 1990).

Evaporitic euxinic environments, favourable to preservation of organic matter, developed also during the deposition of the Burano Formation, as indicated by the correlation of Adriatic oils with organic shales intercalated with the Burano evaporites (Paulucci et al., 1988; Mattavelli and Novelli, 1990). However, as only few wells penetrated the Burano Formation, the geometry and areal extent of this hydrocarbon generating basin is largely unknown.

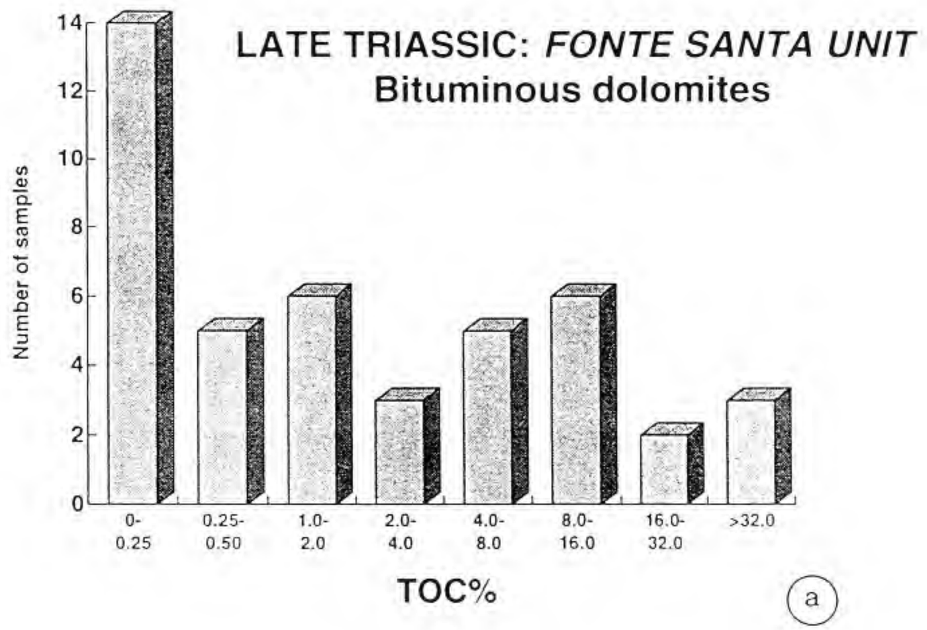
In conclusion, the area of the Southern Apennine and the Adriatic foreland was occupied during Late Triassic times by an extensive tidal carbonate-evaporite platform in which discontinuous euxinic sub-basins, extremely variable in size and shape, developed. These sub-basins can be subdivided into lagoonal troughs, over which platform carbonates prograded (e.g. Giffoni Basin), and rifted lagoonal troughs which later evolved into deeper water basins (e.g. Emma Basin) (Fig. 6; Zappaterra, 1994).

Mesozoic Source-Rocks of the Southeast Sicily Foreland

Geochemical data indicate that the heavy oil accumulations of southeastern Sicily (e.g. Gela, Ragusa, Perla, Prezioso, Vega fields) were charged with hydrocarbons generated by the carbonate-dominated Rhaetian Noto and the shaly Hettangian

Location	Average TOC		Ro %
	Riva di Solto Sh	Zu Lst	
Val Menaggio (W of Como L.)	0.52	0.45	1.0 - 2.14
Val Seriana (N of Bergamo)	0.50	0.25	1.38 - 3.28

TABLE 1



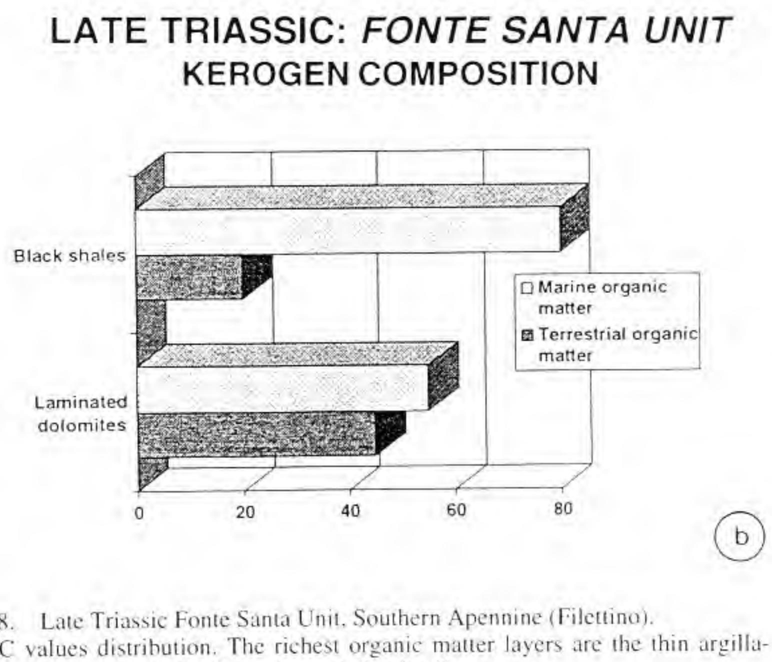


FIG. 8. Late Triassic Fonte Santa Unit, Southern Apennine (Filettino).
a: TOC values distribution. The richest organic matter layers are the thin argillaceous intercalations in the "Bituminous Dolomites".
b: Kerogen composition.

Streppenosa formations; the latter attains in depocentres thicknesses of up to 3000 m (Fig. 6; Pieri and Mattavelli, 1986; Mattavelli and Novelli, 1990). Sedimentological evidence suggests that these source-rock successions, only known from well data, were deposited in limited tensional/transtensional basins which subsided in a carbonate platform (Brosse et al., 1988; Catalano and D'Argenio, 1983).

The Noto Formation exhibits an average TOC content of around 1%; values in the 3-10% range were obtained from marl and black-shale intercalations (Brosse et al., 1988). Its generation potential ranges from 3 to 5 kg HC/t. The kerogen is mainly type II (Novelli et al., 1988) and has an HI of up to 900 mg/g TOC (Fig. 9). On the other hand, the Streppenosa Formation is characterized by a lower TOC content (average 0.35%) and generation potential (0.5 kg HC/t) and by type III kerogen consisting mainly of woody material.

As observed in the Southern Alps and in Northwest Europe, Rhaetian and Hettangian source-rocks were deposited during cycles of rising sea-levels (Creaney and Passey, 1993; Ziegler, 1990).

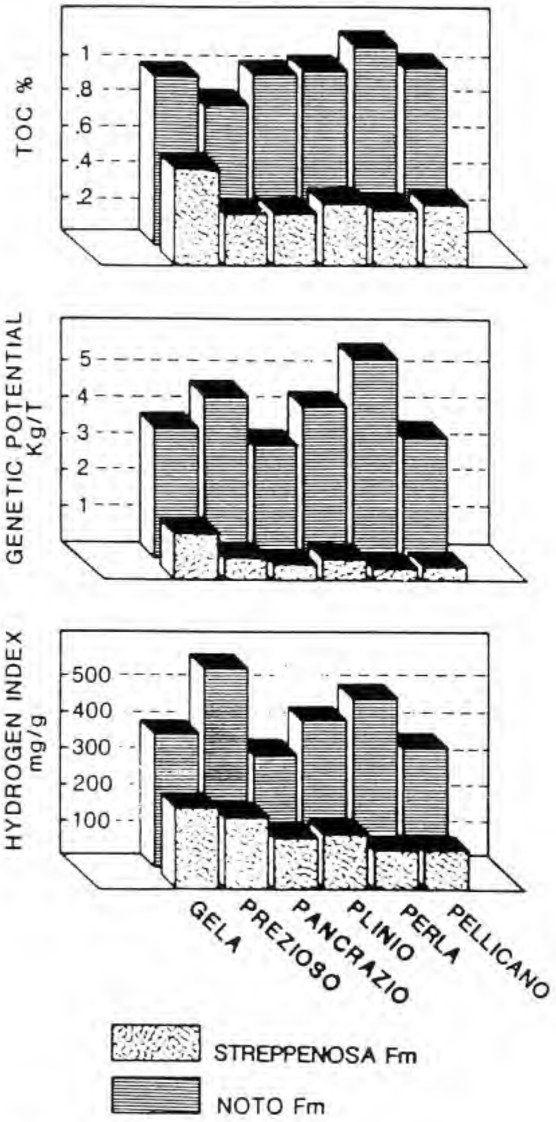


FIG. 9. Late Triassic-Early Lias Noto-Streppenosa Fms, Sicily. Average TOC and Rock-Eval values relevant to some key wells of South Sicily offshore. The Noto Fm is the effective source rock, while Streppenosa plays a minor role as co-source.

Cenozoic Source Rocks

In the evolving Cenozoic foreland basins of the Southern Alps and Apennines, siliciclastic turbiditic series attain thicknesses of several kilometres. In these rapidly subsiding basins, the floor of which was affected by synsedimentary compressional deformations (Pieri and Mattavelli, 1986), conditions for development of anoxic conditions were not favourable. However, predominantly land-plant derived organic matter was preserved, mainly in in the distal parts of siliciclastic turbidites, due to relatively high sedimentation rates, preventing its oxydation. In contrast, very high sedimentation rates, characterizing the proximal parts of submarine fans, apparently account for dilution of organic matter (Mosca and Dalla, 1993).

Geochemical analyses demonstrate that the turbiditic Langhian-Tortonian Marnoso-arenacea Formation has an average TOC content of 0.68% and that the kerogen consists to 80% of land-plant derived matter (Riva et al., 1986). This formation generated the light oils occurring in the external parts of the Northern Apennine (e.g. Cortemaggiore field) (Fig. 10; Riva et al., 1986). Similarly, isotopic data and molecular parameters suggest that the Tertiary flysch of the Southern Apennine generated the light oils contained in e.g. the Castel-

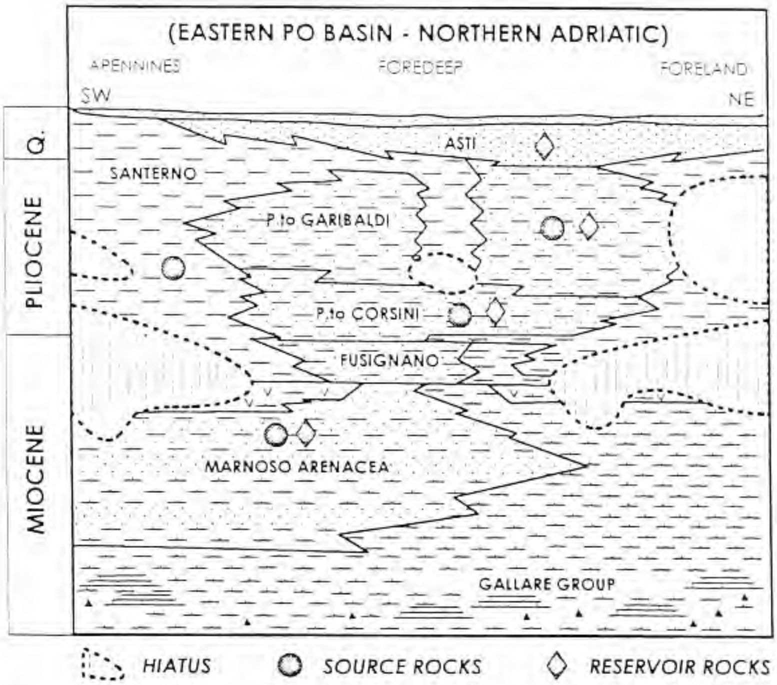


FIG. 10. Late Neogene stratigraphic chart of Eastern Po Plain and North Adriatic. Relationships between the Neogene lithostratigraphic units and main reservoirs and gas prone source rocks.

pagano and Benevento fields (Mattavelli and Novelli, 1990). Tortonian sand-shale successions of the Veneto, the Southern Adriatic and Western Sicily contain only biogenic gas.

In the thick Plio-Pleistocene turbiditic series of the Po Plain and the Northern Adriatic foredeep, enrichment in organic matter occurs mainly in the shaly distal basin plain deposits of highly efficient turbidites (Mutti, 1985). Organic-rich layers can have a TOC content of about 0.7% of which 85% is land-plant derived. Detailed analyses of core material show that the TOC content of re-sedimented clays is 2 to 5 times greater than that of hemipelagic ones (Mattavelli et al., 1992b). The huge volume of Plio-Pleistocene turbiditic series (several thousands of km^3), containing mainly allochthonous land-plant derived organic matter, represents Italy's main source-rock. 80% of Italy's ultimate recoverable gas reserves, corresponding to 2/3 of its entire hydrocarbon resources, consist of biogenic gas trapped in the Northern Apennine foredeep.

PETROLEUM SYSTEMS

In the following we summarize the different petroleum systems of Italy. These are outlined in Fig. 11.

Mesozoic Besano-Meride System, Western Po Plain

In the subsurface of the Po Plain, the South-Alpine and North-Apennine thrust fronts delimit their common foreland (Fig. 5). As Mesozoic carbonate series were virtually unaffected by Tertiary compression, Triassic-Jurassic extensional structures are preserved. Gravity data indicate the presence of a basement high southwest of Milano (Cassano et al., 1986) over which Mesozoic series decrease in thickness to a few hundred meters.

FIG. 11. Approximate geographic distribution of Italian Petroleum Systems.

This high is buried beneath 4000 to 6000 m of Cenozoic strata.

On this palaeo-high, the Gaggiano (1982) and Villafortuna-Trecate (1984) light oil fields were discovered; the latter is one of the largest on-shore oil fields of Europe and has URR of $20 \cdot 10^6$ t ($150 \cdot 10^6$ bbl). Both fields are contained in block-faulted structures which were charged by vertical migration from Middle Triassic source-rocks. The reservoir of the Gaggiano field is located at a depth of 4600 m and consists of two dolomitic layers in the top part of the Meride Formation, providing for a limited trap volume. The reservoir of the Villafortuna-Trecate field is formed by Middle Upper Triassic dolomites, located at depths from 5500 to 6300 m. Both accumulations are sealed by Mesozoic pelagic limestones which are capped by thick Tertiary shaly flysch (Bongiorni, 1987; Novelli et al., 1987; Schlumberger, 1987).

These light oil fields (Gaggiano 36° API, Villafortuna-Trecate 43° API; 0.2% sulfur content) are significantly overpressured (about twice as high as hydrostatic). Overpressure developed in the impervious strata of Oligocene-Miocene sediments during the Neogene rapid subsidence of the Po foreland basin (Novelli et al., 1987). These pressures were subsequently transmitted to the underlying Mesozoic carbonates, the hydraulic continuity of which was disrupted by Tertiary compressive tectonics.

Subsidence analyses indicate that Triassic source-rocks attained maturity for oil generation only during Pliocene-Pleistocene times. The late generation of petroleum may be partly related to the build-up overpressures that retarded hydrocarbon expulsion (Chiaramonte and Novelli, 1986; Mattavelli and Novelli, 1988; Hao Fang et al., 1995).

The genetic potential of source rocks in Gaggiano field exhibits a SPI value (SPI=Source Potential Index, Demaison and Huizinga, 1991) of 2 t HC/m^2 while in the outcrops of the Southern Alps their value is around 4 t HC/m^2. Consequently, a low charge factor may have to be assigned to Besano-Meride source rocks, following the genetic classification of Demaison and Huizinga. Furthermore, the lack of significant oil occurrences in Cenozoic series indicates that the pressures at the top of these light oil accumulations did not exceed the entry pressure of their seal (Hunt, 1990).

In conclusion, the Besano-Meride petroleum system is characterized by low a charge factor (i.e. hydrocarbon yeld), vertical migration paths and good a integrity of the seal

Mesozoic Riva Di Solto System of Southern Alps

In 1973 the Malossa gas-condensate field was discovered in the external parts of the Lombardy thrustbelt which are buried under up to 2 km thick Plio-Pleistocene series of the Po Plain (Encl. 1, section 1). Subsequently additional marginal gas-condensate accumulations, such as Seregna and San Bartolomeo, were discovered in the vicinity of Malossa (Errico et al., 1980; Mattavelli and Margarucci, 1992).

The Malossa field is contained in a northwest striking thrust-anticline which evolved during the Miocene deformation of the Lombardian thrustbelt by reactivation of a Late Triassic extensional fault block. Such northerly trending extensional structures were separated by intervening depocentres in which the Rhaetian Riva di Solto and Zu source-rocks were deposited (Fig. 4-1; Pieri and Mattavelli, 1986; Bertotti et al., 1993). These have been identified as the sources of the hydrocarbons reservoired in the over-pressured, fractured low-porosity (3%) Norian and Early Jurassic carbonates of the Malossa field; from these gas migrated into fractured younger Jurassic and Cretaceous carbonates (Mattavelli and Margarucci, 1992). Lateral and top seals are provided by Cretaceous marls and argillaceous limestones and thick Oligo-Miocene flysch. In areas of more intense faulting, substantial amounts of gas escaped into Pliocene reservoirs and mixed with indigenous biogenic gas (e.g. Caviaga field; Mattavelli et al., 1983).

As the Rhaetian source-rocks are not present in the compressionally deformed Malossa palaeohorst, its charge was probably provided by a lateral graben-shaped basin. These source-rocks have a generation potential of 1-3 kg HC/t and are several hundreds of meters thick (SPI value between 1 and 2 t HC/m^2); as type III kerogen predominates, they have a limited oil potential and are gas prone. This accords with the low gravity of the Malossa condensate (53° API) and a GOR of 1000. Overpressures were responsible for the presence of a monophase hydrocarbon fluid in the reservoir with a dew point of 398 kg/cm^2 (39 MPa=5661 psi). The burial history of such a basin, derived from wells and seismic data, suggests that hydrocarbon generation and expulsion started already during the Early Jurassic (Mattavelli and Margarucci, 1992), possibly charging extensional traps. This is supported by the presence of pyrobitumens with different maturity levels. Much of these earlier entrapped hydrocarbons were probably lost to surface during the Middle-Late Miocene deformation of the Lombardy thrustbelt. The Malossa gas-condensate was probably charged during the Plio-Pleistocene subsidence of the Po Basin.

We conclude that the Riva di Solto petroleum system is probably undercharged, depends on lateral and vertical migration and is characterized by a poor seal integrity.

Mesozoic Emma System of Central Adriatic

In the Pescara Trough of the central Adriatic, Late Pliocene frontal elements of the Apennines involve foreland inversion structures. In these, the Late Cretaceous-Late Eocene Scaglia Formation, which consists mainly of pelagic marly carbonate, contains in turbiditic calcarenite intercalations. These form the reservoir of several accumulations of sulphur-rich (4-10%) heavy oils (7-20° API) (e.g. Gianna field). Geochemical data show that these oils are early expulsion products of the Late Triassic Emma Limestone, encountered in boreholes. The high sulphur content of these oils suggests a low activation energy for the kerogens (Mattavelli and Novelli, 1990; Mattavelli et al., 1991, 1992a).

The Pescara Trough is characterized by a low geothermal gradient (av. 22°C/km) and rapid Plio-Pleistocene subsidence/sedimentation rates (up to 1000 m/Ma). Subsidence analyses show that the Emma Limestone entered the oil window only during the Late Neogene at a depth of more than 5000 m. At the same time, Cretaceous and Paleogene carbonates were compressionally deformed. Expelled oils migrated vertically into the growing structures; their accumulation in calcarenitic layers presumably impeded diagenetic processes and preserved their original porosity. As the fractured marly limestones of the Scaglia Formation have a limited seal potential, some oil escaped and accumulated in Miocene carbonates (Alanno and other minor onshore fields).

The Emma petroleum system is probably undercharged, depends on vertical migration and has a poor seal integrity.

Mesozoic Burano System of Southern Adriatic

The Rospo oil field was discovered in 1975 (André and Doulcet, 1991; Héritier et al., 1991). Its heavy (11° API) and sulphurous (6%), immature oil was generated by anoxic Late Triassic Burano carbonates, which underwent a similar thermal history as the Emma Limestone (Mattavelli et al., 1991). The reservoir is formed by karstified

Albian-Cenomanian limestones, sealed by Messinian anhydrites and marly limestones. The trap is of a paleotopographic-diagenetic type and is mainly hydrodynamically controlled.

The southern parts of the Adriatic Sea are separated from the Apennines by the Puglia Platform and as such are closer associated with the Dinarides-Hellenic foreland basin. The Rovesti and Aquila oil accumulations are located to the East of the Puglia Platform in small Miocene horst blocks, down-faulted with respect to the Puglia Platform (Encl. 3, sect 6). Of the two accumulations, the Aquila field is scheduled for development by means of horizontal wells (Oil and Gas Journal, 1993). Its reservoir is formed by high-porosity (15%) turbiditic calcarenites of the Scaglia Formation which is unconformably overlain and sealed by Oligocene marls. The turbidites were derived from the Puglia Platform.

According to biomarkers, the oils of both accumulations belong to the same family (Paulucci et al., 1988) and show similarities with oil-extracts from the Late Triassic Burano Formation. The Aquila oil is under-saturated and vertically density stratified (36-22°API; Schlumberger, 1987). Numerical simulations indicate that the source-rock entered the oil window oil during the Late Cretaceous (±70 Ma) and that gas-condensate generation commenced during the Late Oligocene (±25 Ma). The main oil expulsion phase occurred during the Late Eocene (45-40 Ma; Mattavelli et al., 1991).

Despite an analogue source-rock and reservoir model, the evolution of the Aquila area differs considerably from the Emma area. Low maturity heavy oils may not have been expelled from the source-rock due to lack of fracturing and may later have been cracked to lighter oils (Palacas, 1983).

Available data do not allow to assess the charge factor of the Burano Petroleum System; migration is probably vertical and seal integrity is good.

Mesozoic Noto-Streppenosa System of Southeast Sicily

The on- and off-shore fields of southeastern Sicily account for most of Italy's URR of oil. Oils are characterized by low gravity (5-20° API) and a high sulphur content (2.6-9.8%); well preserved n-alkanes indicate that they are not biodegraded but are early expulsion products of the Noto and Streppenosa source-rocks (Fig. 6; Pieri and Mattavelli, 1986; Mattavelli and Novelli, 1990). The genetic potential of these source-rocks is rather low, with SPI value estimated at about 3 t HC/m^2. Expelled oils migrated into Late Triassic platform dolomites, sealed by the Noto-Streppenosa series (e.g. Gela and Ragusa fields), and into Early Jurassic limestones deposited on a platform, prograding over the source-rocks, which is sealed by Early Jurassic argillaceous limestones and marls (e.g. Vega and Perla fields). The integrity of these seals was reduced by Tertiary tectonics, as evident by asphalt seeps and frequent shows in post-Jurassic series. Traps are provided by faulted anticlines (e.g. Ragusa and Gela fields) which developed in response to Late Neogene compressional reactivation of Jurassic extensional structures. In contrast, the off-shore Vega field is contained in a Late Cretaceous combination trap.

Numerical modelling indicates that in the area northwest of Gela, the main phase of oil generation occurred during the last 5 Ma when the Sicily Apennine foredeep basin developed (Novelli et al., 1988). Strong subsidence of the foredeep initiated the onset of oil generation and expulsion at low maturity levels, particularly from the Noto Formation. High heat-flow, related to rifting activity on the Pelagian Shelf and Tyrrhenian Sea, may have contributed to maturation. Conversely, the burial history of the Noto Formation in the drainage area of the Vega field indicates that oil generation commenced during the Paleocene and peaked during the Late Miocene-Pliocene and, as such, clearly post-dates trap formations (Schlumberger, 1987).

In conclusion, the petroleum system of southeastern Sicily is undercharged, depends on lateral migration and is characterized by a relatively poor seal integrity.

Miocene Flysch System of Po Plain and Emila Folds

The Cortemaggiore gas-oil field and additional minor oil accumulations are located in the Emilia Arc, a major subsurface unit of the external Northern Apennine (Pieri and Groppi, 1981; Pieri, 1992). The Cortemaggiore accumulation is contained in a thrust-anticline; clean, well sorted Messinian sands contain wet gas and Tortonian sands light oil (35-40° API). Minor amounts of thermal gas leaked into Pliocene reservoirs where they mixed with biogenic gas (e.g. Spilamberto field; Mattavelli et al., 1983). Trap formation commenced during the Late Miocene and persisted into Plio-Pleistocene times (Pieri, 1992). Thrust structures, detached from Mesozoic carbonates, are cored by Miocene flysch and are unconformably overlain by Messinian marls and sands, Pliocene shales and Pleistocene sands.

Bio-markers, carbon isotope values and pristane/phytane ratios permit to differentiate between the Cortemaggiore oil group from Mesozoic oils (Riva et al., 1986). The presence of oleanane, an Angiosperm-derived bio-marker (Moldowan et al., 1993), is indicative of land-plant derived kerogen, and suggests that hydrocarbons were sourced by the 800-1000 m thick Langhian-Tortonian Marnoso-arenacea flysch succession which has a SPI of 1.6-2.0 t HC/m^2 (Mosca and Dalla, 1993). Mathematical modelling shows that oil generation occurred during the last 3 Ma at depths of 5500 to 7000 m (Chiaramonte and Novelli, 1986). Rapid Neogene thrust-loaded subsidence and deep burial favoured the generation of thermal gas, amounting in energy equivalents to 12 times the oil reserves of this province (Mattavelli and Novelli, 1990).

The Cortemaggiore petroleum system must be regarded as undercharged, depends on vertical migration and is characterized by a poor to adequate seal integrity.

Tortonian Biogenic Gas Systems

Tortonian biogenic gas plays a subordinate role in the hydrocarbon habitat of Italy. In the Veneto, the small Conegliano biogenic gas field is reservoired in Tortonian carbonate sands involved in a Pliocene anticlinal structure. In Western Sicily turbiditic sands of the Terravecchia Formation, involved in a Pliocene fold, host the small Lippone-Mazara biogenic gas accumulation. In the southern Adriatic, the Falco-1 well, drilled in the vicinity of the Aquila oil field, established a biogenic gas accumulation in Pliocene and Messinian sands and Tortonian limestones (Paulucci et al., 1988).

Plio-Pleistocene Biogenic Gas Systems

Bacterial gas, associated with the immature Plio-Pleistocene turbiditic series of the Apennine foredeep, is by far the most important hydrocarbon resource of Italy (Mattavelli et al., 1983; Mattavelli and Novelli, 1988). 70% of the biogenic gas reserves are located in the Northern Apennine foredeep and in the Northern Adriatic Sea where high subsidence rates and a low geothermal gradient were conducive to the generation and preservation of large volumes of bacterial gas; moreover, basin-plain highly efficient turbidites (Mutti, 1985) provide for an excellent sand-shale ratios and laterally continuous sand sheets. In contrast, the Southern Apennine foredeep is characterized by laterally discontinuous, channelized turbidite sands. Generation of biogenic gas was syn-sedimentary and is probably still going on.

Main traps are provided by synsedimentary thrust-anticlines (e.g. Ravenna field) and by gentle anticlines adjacent to the thrustfront (e.g. Porto Garibaldi-Agostino field). In the northwestern Po Plain, structural, stratigraphic and combination traps are associated with the Messinian unconformity (e.g. Sergnano and Caviaga fields). In the same area, up-dip shale-outs of sands provides for stratigraphic traps (e.g. Settala field). The large Barbara field in the Northern Adriatic Sea is contained in a gentle drape fold. Gas often occurs in stacked accumulations, separated by less than 1 m thick clays. Gas saturations give rise to reflection-seismically detectable direct hydrocarbon indicators (amplitude/frequency anomalies) (Pieri and Mattavelli, 1986; Schlumberger, 1987; Mattavelli et al., 1988; Mattavelli and Novelli, 1988). In the Southern Apennine foredeep, gas can be trapped in palaeotopographic and /or structural highs upheld by Mesozoic carbonates, unconformably sealed by Pliocene shales (e.g. Grottole field), and in stratigraphic and combination traps involving Plio-Pleistocene sands (Sella et al., 1990, 1992).

Small extensional basins associated with the opening of the Tyrrhenian Sea contain minor biogenic gas accumulations in Plio-Pleistocene shallow marine sands (e.g. Tombolo field).

Petroleum Systems Related to Uncertain Source-Rocks

Quite a number Italian oil and gas accumulations were charged from source-rocks which, so far, have not yet been identified. However, in some cases, their stratigraphic position can be inferred from geochemical data.

The **Cavone and Bagnolo oil fields**, located in the external part of the Ferrara foldbelt of the Northern Apennine, are probably related to Triassic source-rocks. Both oils are heavy (20-23° and 16° API, resp.) and sulphur rich (3-4 and 5%, resp.) but differ in their carbon isotope compositions ($d^{13}C$: Cavone -28.9 to -30.8%°, Bagnolo -22.5%°). Molecular parameters are similar to the Besano-Meride system. In the area of these fields, inferred Triassic source-rocks are located at depths of 5000 to 7000 m and attained maturity during the last 6.5 Ma (Wygrala, 1988).

The Cavone field is contained in a dissected, thrust anticlinal structure, involving Early Jurassic and Cretaceous carbonate reservoirs, sealed by Middle Jurassic marly limestones and Early Cretaceous marls (Nardon et al., 1991); Neogene compressional deformation of a pre-existing Jurassic extensional fault block reduced the seal capacity and allowed gas to escape to the surface. The reservoirs of the Bagnolo field are formed by Cretaceous platform carbonates, sealed by Late Miocene shales.

The marginal **Ripi field**, located to the Southeast of Rome, was probably also charged by Triassic source-rocks. The area is heavily tectonized, permitting hydrocarbons to migrate vertically through fractured Mesozoic and Miocene carbonates into an irregularly structured Tortonian sand-shale sequence. Discovered during the past century on the basis of oil seeps, this field produced during the past years on average 1000 t/y (7300 bbl/y) of 21° API oil with a sulphur content of 3.7%.

The **Castelpagano group of small oil accumulations**, located in the Southern Apennine northeast of Naples, produce 30-43° API, low sulphur oil from Early Miocene and Cretaceous limestones, involved in parautochthonous thrust structures covered by the Lagonegro nappe (Encl. 4, sect. 5). The presence of oleanane in the Castelpagano oil suggests a Tertiary origin. Possible source-rocks are Messinian shales which have a TOC content of 0.95-1.25 and a generation potential of 3-5.2 kg HC/t. These oils are similar to those associated with biogenic gas, reservoired in Pliocene sands, in the eastwards adjacent Candela and Torrente Tona fields (Mattavelli and Novelli, 1990; Casero et al., 1991).

The Costa Molina group of oil fields of the Southern Apennine are also trapped in parautochthonous compressional structure beneath the Lagonegro nappe (Encl. 3, sect. 6 Mostardini and Merlini, 1988). Production comes from low porosity Cretaceous and Miocene carbonates, sealed by tight Miocene limestones and/or marls of the Lagonegro nappe. The limited capacity of these seals is indicated by oil seeps. Oil gravities range between 12.6 and 20.6° API and sulphur content is around 3%. Molecular parameters seem to suggest a Late Triassic-Early Jurassic source for these oils (Mattavelli and Novelli, 1990). However, recently an organic-rich Cretaceous facies has been prposed as a potential source candidate for similar oil in the Tempa Rossa field (Roure and Sassi, 1995). Numerical modelling indicates that oil generation and expulsion set in during the Late Neogene (Casero et al., 1991).

In the Ionian Sea, off-shore Calabria, the **Luna gas field** produces from Serravallian-Tortonian conglomerates and sands (porosity 9-22%), sealed by Tortonian and Pliocene clays and marls; these are involved in the external parts of the upper allochthonous units (Schlumberger, 1987; Roveri et al., 1992). This clearly thermal gas contains minor amounts of ethane and propane and is devoid of non-hydrocarbon gases. Heavy isotope values of methane suggest that it was generated at depths >6000 m from an unknown source rock, presumably during the Late Neogene.

In northeastern Sicily, the **Gagliano group of gas-condensate fields** produce from low porosity quartzose Oligo-Miocene turbiditic sands of the allochthonous Numidian Flysch. The gas has clearly a thermal origin; associated condensate (55° API) contains small amounts of oleanane, indicating at least a contribution from Tertiary sources. The trap-providing imbricated thrust-anticlines developed only in Early Pliocene times and have a limited retention potential, as indicated by frequent seeps (Schlumberger, 1987; Mattavelli and Novelli, 1990).

In the Sicily Channel, the **Narciso group of small oil fields** are reservoired in low-porosity (5%) Oligocene fossiliferous limestones involved in thrust structures which are partly overridden by the Flysch nappes. The 21-39° API oil, containing 1-2% of sulphur, was probably generated by Mesozoic source-rocks of unknown age. The presence of a CO_2 gas cap in the Nilde field must be related to volcanic activity associated with the development of the Pantelleria rift system (Schlumberger, 1987).

CONCLUSIONS

The hydrocarbon accumulations of Italy are concentrated in the external parts of the Alpine and Appennine fold- and thrustbelts and their forelands. A schematic sketch of the main petroleum plays of Italy is shown in Fig. 12. Minor accumulations occur, however in the more internal parts of the Apennines. The bulk of Italy's ultimate recoverable hydrocarbon reserves consists of biogenic gas contained in Plio-Pleistocene turbiditic sands of the Northern Apennine foredeep basin. Italy's oil and gas accumulations can be attributed to a number of more or less well defined petroleum systems.

The Middle Triassic-Early Jurassic petroleum systems are related to the development of isolated larger and smaller anoxic depressions within expansive carbonate platforms, resulting from early rifting phases, preceding the opening of the Alboran-Liguria-Piedmont ocean and the isolation of the Adria plate during the separation of Africa from Europe. Shaly and carbonate dominated source-rocks, partly associated with evaporites, which were deposited in these depressions, can attain thicknesses of 2 km and more. Their TOC content varies vertically and laterally and reaches maxima during transgressive periods, giving rise to a reduction of carbonate influx from flanking platforms.

Compared with world-wide examples of source-rocks, the SPI of Italian Middle Triassic to Early Jurassic source-rocks is low (Fig. 13). Correspondingly, its Triassic-Early Jurassic petroleum systems are generally undercharged. Moreover, it must be realized that the distribution of potential source-basins beneath the thick sedimentary fill of the foreland basins and particularly under the Apennine nappes is largely unknown. It is interesting to note, that on a global scale, Triassic source-rocks represent only 1.2% of all known source-rocks (Klemme and Ulmishek, 1991) whereas the bulk of Italian oils were generated by Triassic source-rocks. In many parts of Italy these reached maturity only during the Neogene empacement of the Alpine and Apennine nappes and the associated rapid flexural subsidence of the respective foreland basins; the Mesozoic Aquila and Vega kitchens are exceptions.

Unlike in other Tethys realm basins, Middle Jurassic to Cretaceous source-rocks play a very subordinate role in the hydrocarbon habitat of Italy. A probable exception is the Costa Molina petroleum system.

During the Alpine orogenic cycle, the rather lean Miocene flysch petroleum system developed, which plays a role in the Po Plain, the Emila foldbelt and in the Castelpagano area of the Southern Apennine. Bacterial gas was also generated in the shaly Miocene series of the eastern Southern Alps, the Southern Adriatic and in Western Sicily.

During the Neogene emplacement of the Apennine nappes, huge thicknesses of Plio-Pleistocene flysch series accumulated in the Po Plain and the Northern Adriatic foreland basin. High sedimentation rates, low temperature gradients, the availability of multiple reservoir seal pairs involving laterally continuous turbiditic sands and hemipelagic shales, and syndepositional compressional deformations provided ideal conditions for

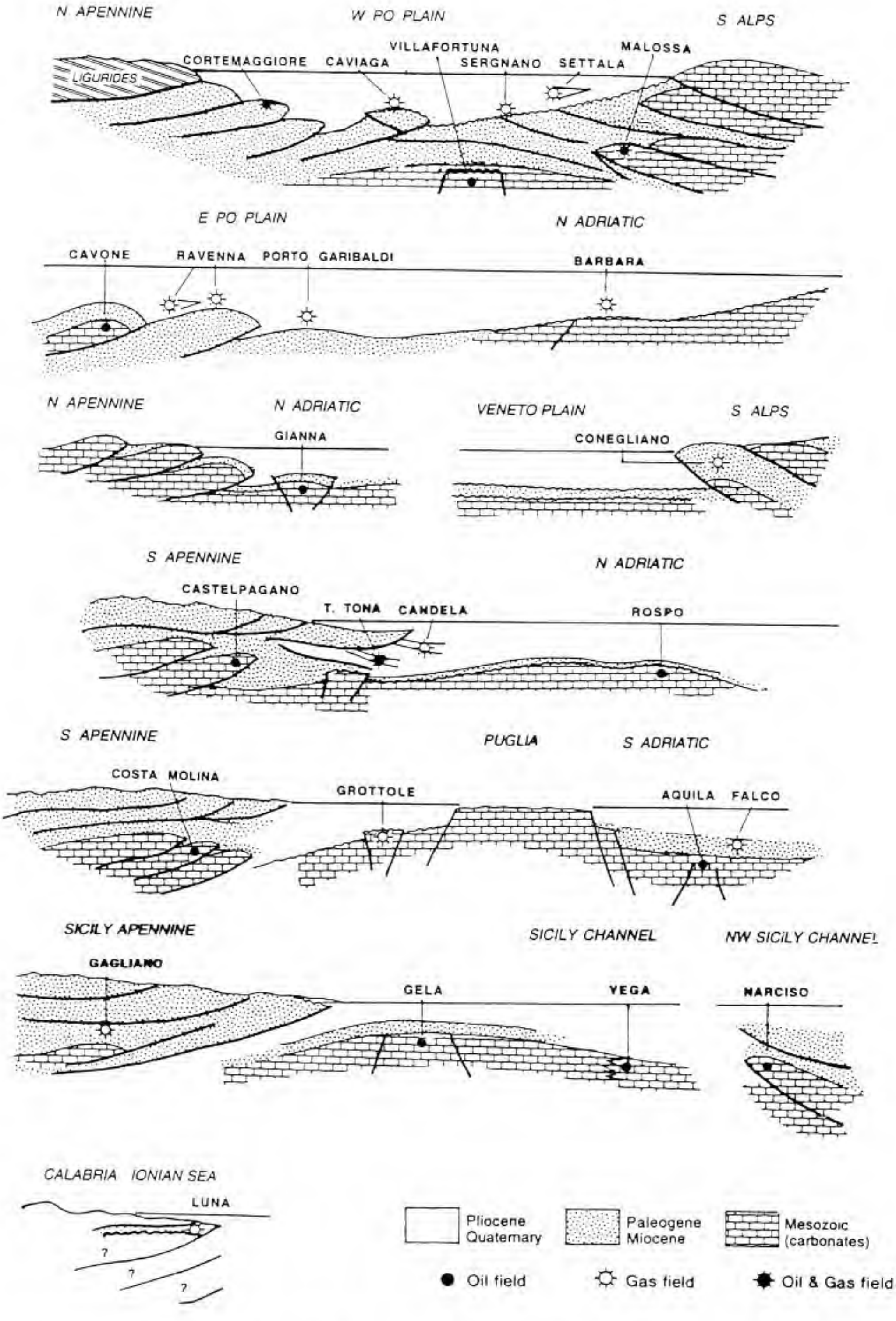


FIG. 12. Schematic sketches of the main Petroleum Plays in Italy.

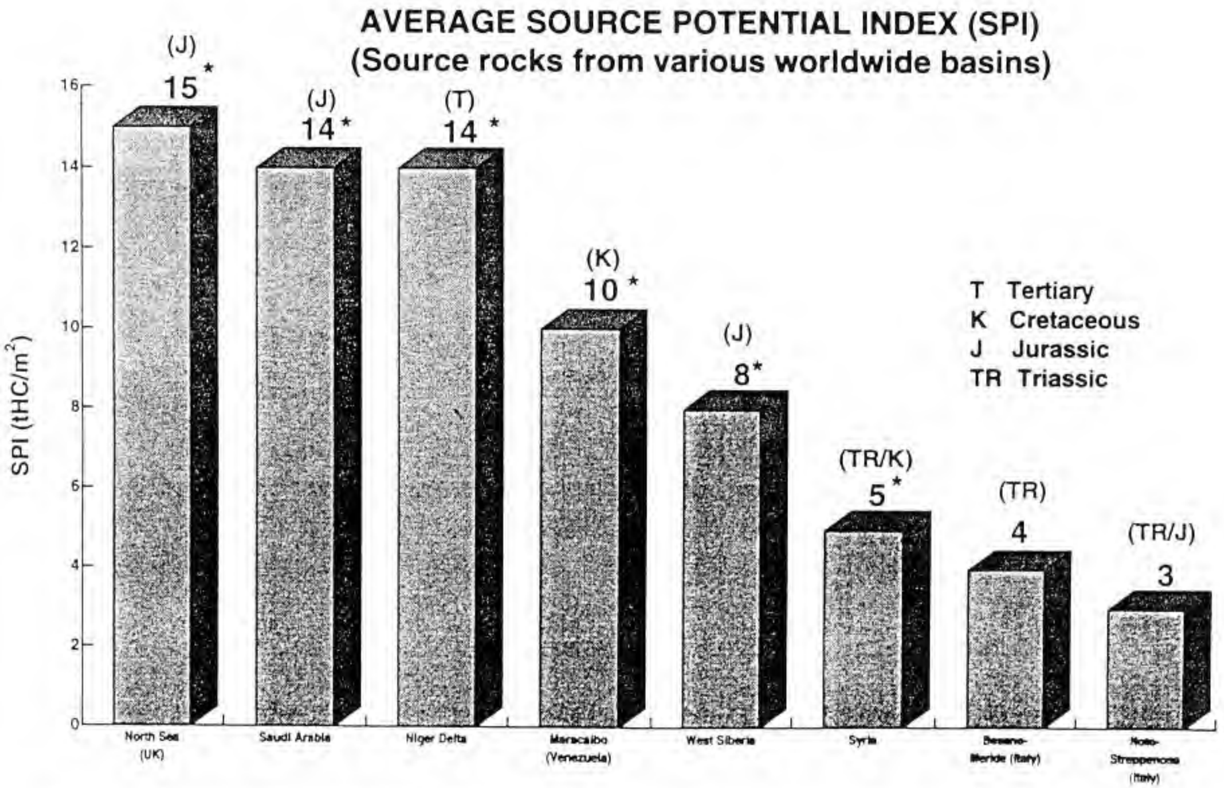


FIG. 13. Average Source Potential Index (SPI) of some petroleum systems (data after Demaison and Huizinga, 1991).

the generation and entrapment and retention of bacterial gas, accounting for 70% of Italy's hydrocarbon resources. In the Southern Apennine foredeep and in the Peri-Tyrrhenian Neogene basins such ideal conditions were not realized.

Sub-thrust plays, aiming at Mesozoic and Cenozoic objectives in compressional foreland and parautochthonous structures, covered by the Apennine nappes, have met with success in the Castelpagano and Costa Molina areas of the Southern Apennine. Such plays have to contend with difficulties in reflection-seismic prospect definition, and above all, with hydrocarbon charge uncertainties.

At present, Italy's recoverable hydrocarbon reserves consist to 87% of biogenic gas and to 13% of Triassic oil. Future discoveries in frontier areas, such as the external thrustbelts and the deep Mesozoic objectives of the foreland basins could change this situation.

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Enclosures

Enclosure 1 Geological cross-sections through the Northern and Central Italy.

Enclosure 2 Geological cross-sections through the Southern Italy.

Enclosure 3 Geological cross-sections through the Southern Italy and Sicily.

Relationship between tectonic zones of the Albanides, based on results of geophysical studies

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ABSTRACT

The Albanides link the Dinarides and the Hellenides, with which they form the southern branch of the Mediterranean Alpine Belt. Our analysis of the Albanides and their extension into the Adriatic Sea integrates surface geological observations, well data and results of seismological, refraction- and reflection-seismic, gravity, magnetic and geoelectric surveys.

The evolution of the Albanides began with the Triassic subsidence of their Hercynian substratum under a tensional regime, culminating in crustal separation and opening of the Subpelagonian and Hellenic-Dinarid oceanic basins. The Alpine orogenic history of the Albanides spans Late Jurassic to Quaternary times and can be subdivided into a Late Jurassic-Early Cretaceous early-tectonic, a Mid-Cretaceous to Eocene main-tectonic, an Oligocene-Miocene late-tectonic and a Plio-Pleistocene neo-tectonic cycle.

The Albanides consist of two major palaeogeographic domains. The Internal Albanides formed part of the oceanic Subpelagonian Trough, whereas the External Albanides developed out of the western passive margin and continental shelf of the Adriatic plate. During the early-tectonic phase, the ophiolitic Mirdita nappe was obducted onto the margin of the Adriatic plate. This was accompanied by the development of a flexural foreland basin. During the main-, late- and neo-tectonic phases, progressive westward advance of the orogenic front was coupled with a westward shift of the foredeep basin axis to its present location at the margin of the Adriatic Sea. The External Albanides evolved out of the Ionian Mesozoic shelf sedimentary prism and the superimposed foredeep wedge. The Albanides are underlain by autochthonous continental basement which was little deformed during their evolution.

The ophiolites of the Mirdita nappe give rise to major gravity and magnetic anomalies, indicating that its thickness ranges between 2 and 14 km. Reflection-seismic and gravity surveys carried out in the External Albanides and the Adriatic Sea define distinct structural belts which are related to different tectono-stratigraphic units.

The most important oil and gas accumulation are found in the Ionian and Sazani zones and in the Periadriatic Depression which extends into the Adriatic off-shore. Structuration of the Ionian and Sazani zones occurred during the late- and neo-tectonic phases. The carbonate-dominated Late Trias-

Frasheri, A., Nishani, P., Bushati, S. & Hyseni, A., 1996. — Relationship between tectonic zones of the Albanides, based on results of geophysical studies. *In*: Ziegler, P. A. & Horváth, F. (eds), Peri-Tethys Memoir 2: Structure and Prospects of Alpine Basins and Forelands. *Mém. Mus. natn. Hist. nat.*, **170**: 485-511. Paris ISBN: 2-85653-507-0.

sic to Late Cretaceous series of the Ionian, Kruja and Krasta-Cukali zones contains several rich to very rich source-rock intervals. In the Ionian zone, Late Cretaceous, Paleocene and Eocene carbonates and Oligocene flysch-type and Miocene molasse-type sandstones form the reservoirs of the main oil and gas accumulations. In the Periadriatic Depression, Tortonian-Pliocene Molasse-type clastics form the primary reservoirs.

INTRODUCTION

The Albanides form an integral part of the southern branch of the Mediterranean Alpine Orogen which extends from the Southern Alps, through the Julian Alps, the Dinarides and the Albanides into the Hellenic arc. The Albanides have since long been the subject of intense surface geological studies which are summarized in a voluminous literature (Aubouin, 1973; Aliaj, 1987; Dalipi, 1985; Kodra and Gjata, 1989; Melo, 1986; Ndoja, 1988; Papa and Kondo, 1968; Papa, 1981; Shallo et al., 1989; Valbona and Misha, 1987).

The Albanides are subdivided into an internal and an external zone (Fig. 1). The Albanian Internides are formed by the Mirdita ophiolite nappe which was derived from the oceanic Subpelagonian Trough. This very large ophiolite nappe forms the orogenic lid of the Albanides and overrides the essentially sedimentary Korabi and Gashi nappes. The Externides comprise the Krasta-Cukali and Kruja nappes, derived from the Pindic and Gavrovo zones, respectively, and the Ionian and Sazani zones. The Krasta-Cukali and Kruja nappes are thrusted over the Ionian system of thin skinned thrust sheets and folds which are detached from their basement at the level of Permo-Triassic salts. The latter borders the little deformed Adriatic foreland, corresponding to the Adriatic Platform (Figs. 2 and 3).

Sedimentary series involved in the Albanides range in age from Ordovician to Quaternary and record their evolution. Following the Variscan orogeny, the area of the future Albanides began to subside during Permo-Triassic times under a tensional regime which culminated in the Late Triassic opening of the oceanic Subpelagonian Trough and the Hellenid-Dinarid Ocean, also referred to as the Vardar Ocean. Sea-floor spreading terminated during the Late Jurassic when closure of the Vardar Ocean and the Subpelagonian Trough commenced at the onset of the Alpine orogenic cycle. In Albania, the Alpine orogeny spanned Late Jurassic to recent times and can be subdivided into four cycles. During the Late Jurassic-Early Cretaceous palaeotectonic cycle, the Subpelagonia Trough was closed and the ophiolitic Mirdita and the sedimentary Korabi nappes were emplaced on the passive margin of the Adriatic plate. This was accompanied by the development of a first flexural foreland basin. During the Mid-Cretaceous to Eocene maintectonic cycle, the Pindos zone was incorporated into the orogen, forming the westward advancing Krasta-Cukali and Kruja nappes; during this time the Ionian zone was incorporated in the foreland basin. During the Oligocene to Miocene late-tectonic phase, the Ionian zone was deformed and during the Plio-Pleistocene neo-tectonic cycle, the most external peri-Adriatic foreland Basin was compressionally deformed; at the same time tensional Neogene basins subsided on the Mirdita nappe (Fig. 2).

In recent years, geophysical data have been increasingly and successfully applied in an effort to unravel the structural configuration of the Albanides and to establishing the relationship between their different tectono-stratigraphic units. Our analysis of the Albanides and their extension into the Adriatic Sea integrates surface geological observations, well data and the results of seismological, refraction- and reflection-seismic, gravity, magnetic and geoelectric surveys. (Aliaj, 1998; Arapi, 1982; Langora et al., 1983; Lubonja et al., 1977; Nishani, 1985; Sulstarova, 1987).

CRUSTAL CONFIGURATION

Refraction, gravity and magnetic data indicate that the thickness of the crust increases from about 30 km in the central parts of the Adriatic Sea and

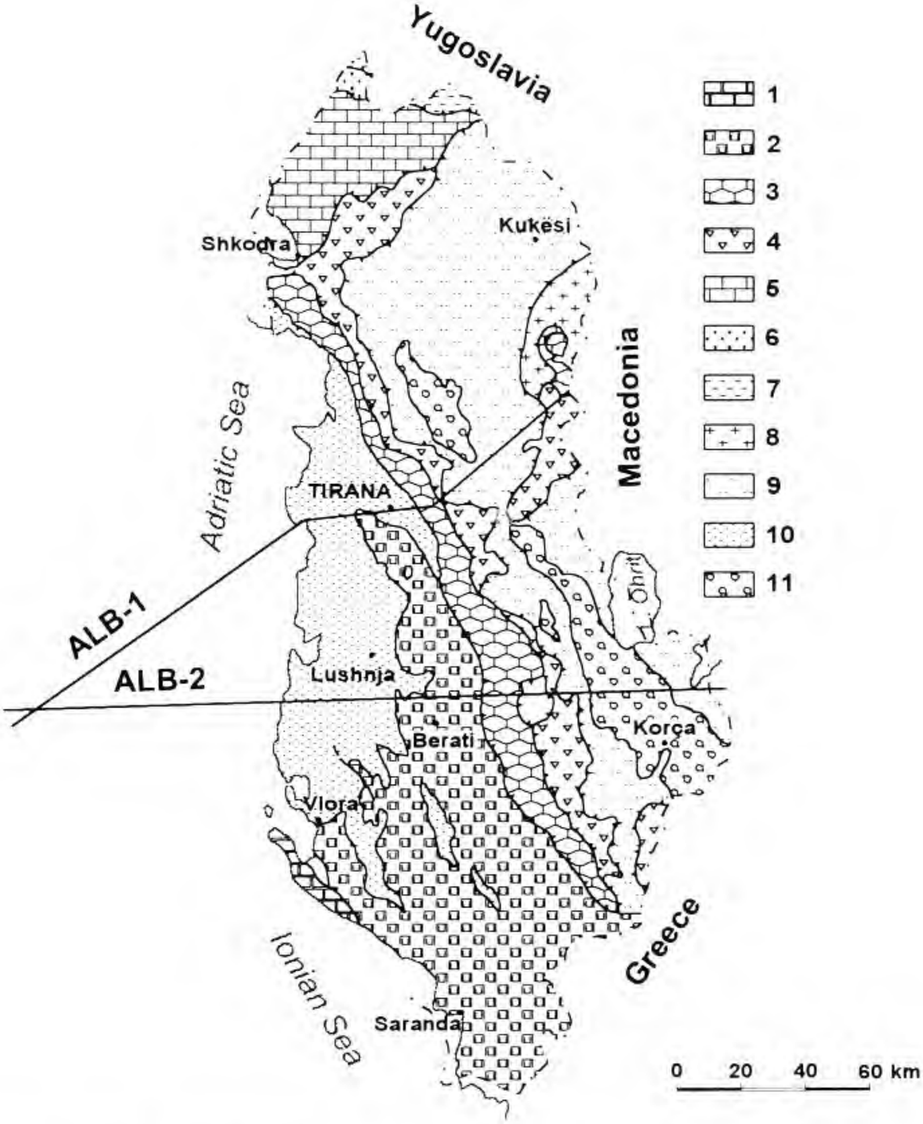

FIG. 1. Schematic tectonic map of Albania. Tectonic Zones: 1) Sazani, 2) Ionian, 3) Kruja, 4) Krasta-Cukali, 5) Alps, 6) Vermoshi, 7) Gashi, 8) Korabi, 9) Mirdita, 10) Periadriatic Depression, 11) Neogene depressions on Mirdita nappe. ALB-1 and ALB-2 locations of transects given in Figs. 2 and 3.

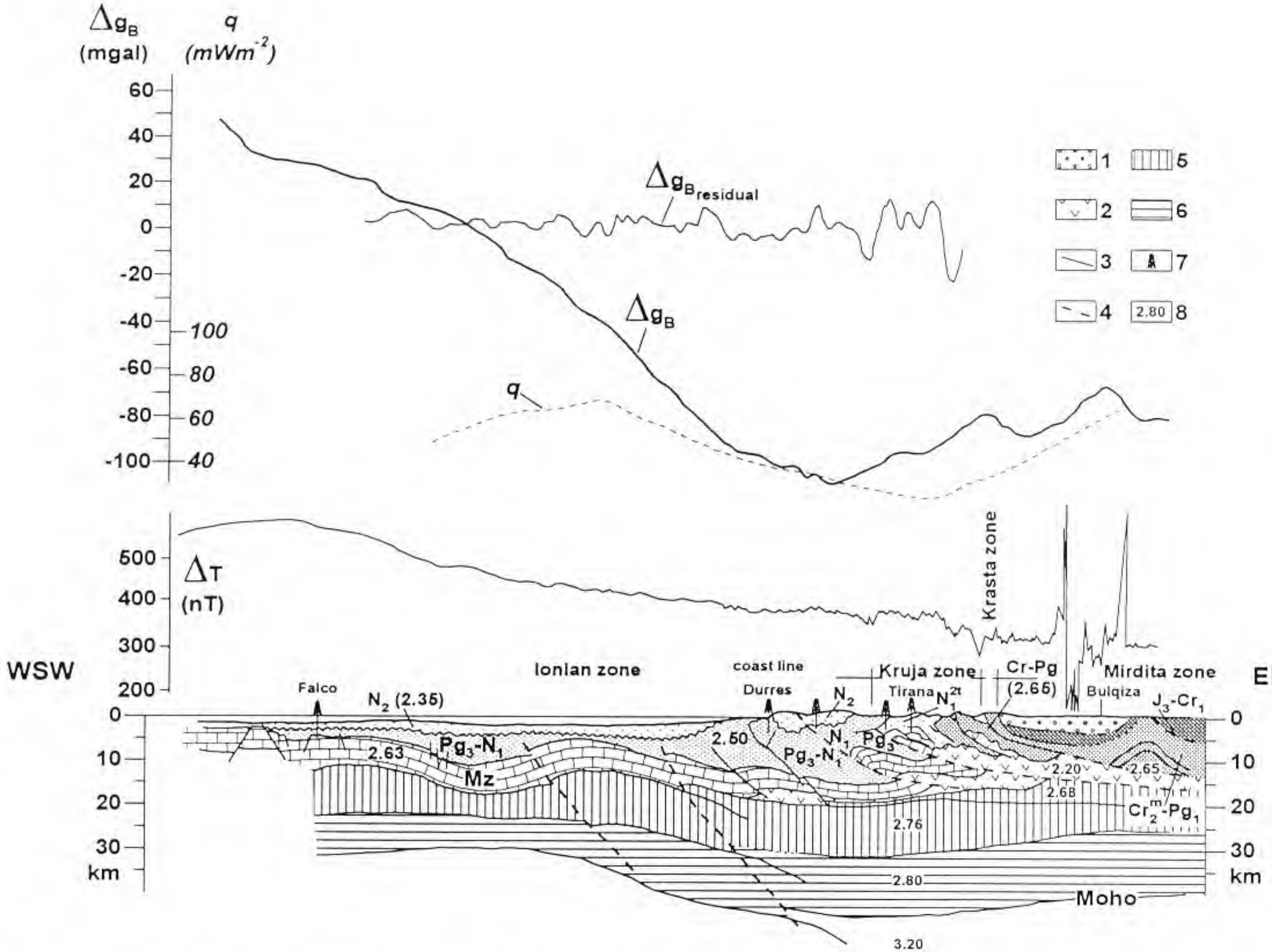

FIG. 2. Geological-geophysical transect ALB-1 through north-central Albania. For location see Fig. 1.
N-Neogene, Pg-Paleogene, Cr-Cretaceous, J-Jurassic, MZ-Mesozoic.
1) ultrabasic rocks, 2) salt, 3) faults, 4) crustal fractures, 5) upper crust, 6) lower crust, 7) upper mantle, 8) deep wells, 9) density g/cm^3.

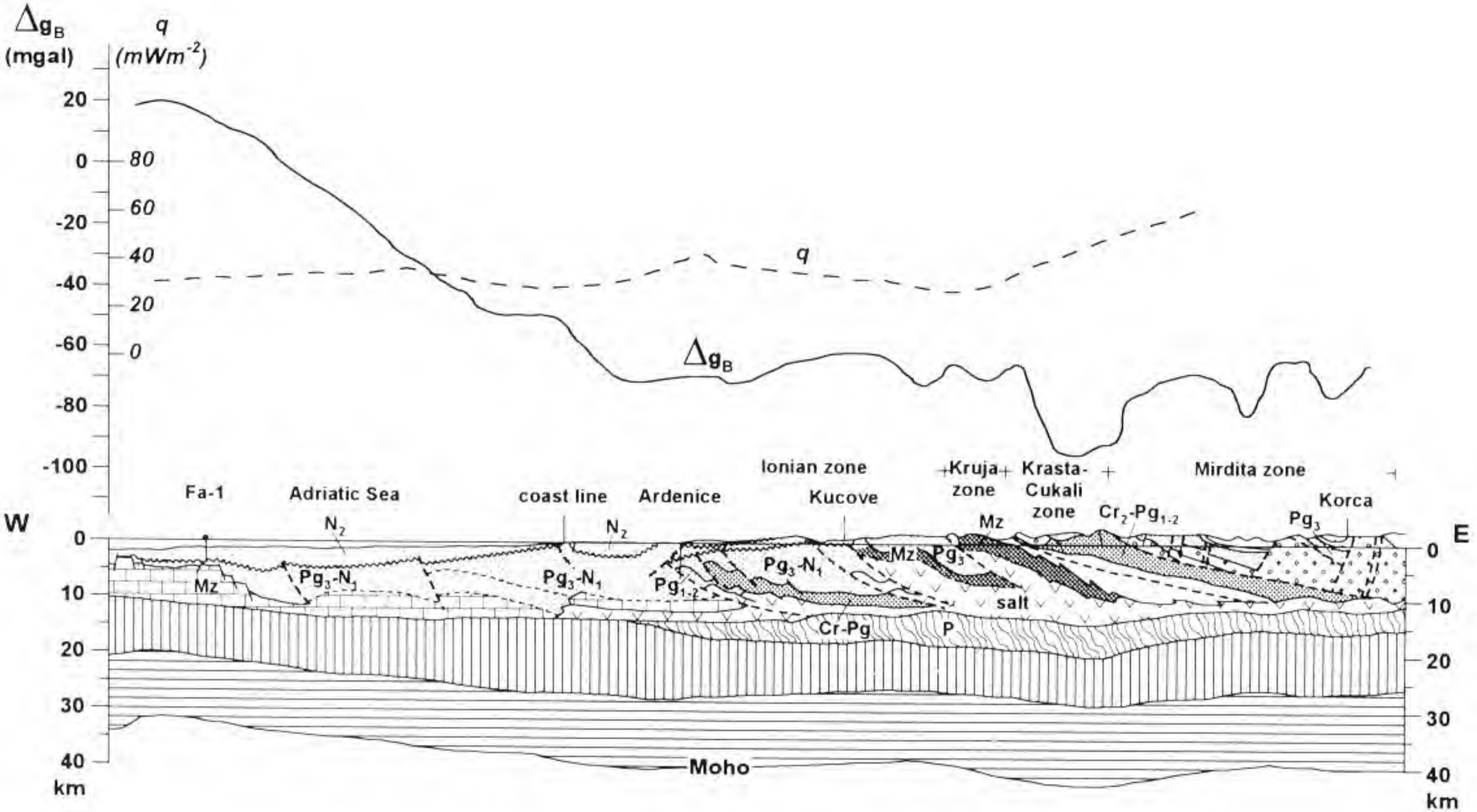

FIG. 3. Geological-geophysical transect ALB-2 through central Albania. For location see fig 1. Same legend as Fig. 2.

the Ionian zone to 43 to 52 km under the Albanian Internides (Figs. 2 to 4). Maximum crustal thicknesses are observed in the northern parts of Albania. Sediments attain thicknesses of minimum 12 km in coastal areas, increase to 13-14 km in northwestern Albania and up to 15-17 km in the central parts of Albania. In the southern coastal area the top of the basement is located at a depth of about 12 km and descends to 17 km in northern Greece.

The third order Bouguer gravity map (Fig. 5) shows the presence of a broad positive anomaly in the southern coastal area which is separated by a discontinuous negative trend from the sharp positive anomaly that is associated with the front of the ophiolitic Mirdita nappes. Refraction data confirm that the coastal positive Vlora anomaly coincides with a relatively shallow Moho whereas gravity minima correlate with greater crustal thicknesses. The configuration of these gravity anomalies suggests that the crust is subdivided into several blocks, varying in crustal thickness, which are separated by a system of deep reaching fracture zones (Fig. 5).

The northwestern positive gravity anomaly of the Shkodra area, which trends towards the Mirdita zone, suggests that the crust thickens to the North under the Albanian Alps as well as to the South in the Durresi area. The negative Durresi-Tirana-Elbasani and Permety anomalies, which are indicative of greater crustal thicknesses, are partly separated by the Vlora-Berati negative anomaly, corresponding to thinner crust. This subdivision of the crust into blocks is partly also reflected by the pattern of magnetic anomalies (Fig. 6). In the southern Adriatic Sea, the crust is characterized by a more uniform thickness of about 30 km (Fig. 2; Morreli et al., 1969; Montanari, 1989; Rigo and Caprarelli, 1980; Cadet et al., 1980).

Refraction- and reflection-seismic data, as well as gravity models and geoelectric soundings, suggest that the entire Albanides are underlain by an autochthonous continental basement complex which dips gently to the East (figs. 2 and 3). However, granitic basement is apparently involved in the Gashi zone of northernmost Albania (Fig. 1).

Although there is only limited evidence for involvement of the continental crust in this major thrust belt, seismological data indicate that deep transverse and longitudinal fractures transect the crust (figs 5 and 6; Aliaj, 1988; Grazhdani, 1987; Sulstarova, 1987; Bakiaj and Bega, 1986). Such crustal fractures, which are associated with earthquake epicentres, appear to coincide with the boundaries between the Krasta-Cukali and Kruja, the Kruja and Ionian and the Ionian and Sazani zones and occur also within the Ionian zone (figs. 1, 5 and 6). These fractures delimit blocks characterized by different crustal thickness, depth and gravity response. Although these fractures appear to influence the architecture of the Albanides, the age of their development is uncertain. However, earthquakes associated with them show that they are at present tectonically active. During the different orogenic cycles of the Albanides they may have controlled the subsidence pattern of the evolving foreland basin and guided the deformation of the sedimentary cover of the basement.

INTERNAL ALBANIDES

The Internides of the Albanides are dominated by the Mirdita nappe which occupies an up to 70 km wide belt in the western part of Albania. It rests in thrust contact on the Krasta-Cukali nappe, which reappears in erosional windows along the eastern border of Albania in the Peshkopia, Okshtuni and Gramozi areas. In the northern part of the Peshkopia window, the Korabi nappe appears beneath the Mirdita nappe. The Gashi nappe is restricted to northernmost Albania and appears to plunge southwards under the Mirdita nappe (Fig. 1). The Mirdita nappe is thought to be derived from the Subpelagonian Trough whereas the Korabi, Gashi and Krasta-Cukali nappes correspond to the Pindic margin of the Subpelagonian Trough.

Mirdita Zone

The Mirdita nappe consists of a basal Middle-Late Triassic ophiolite sequence which varies con-

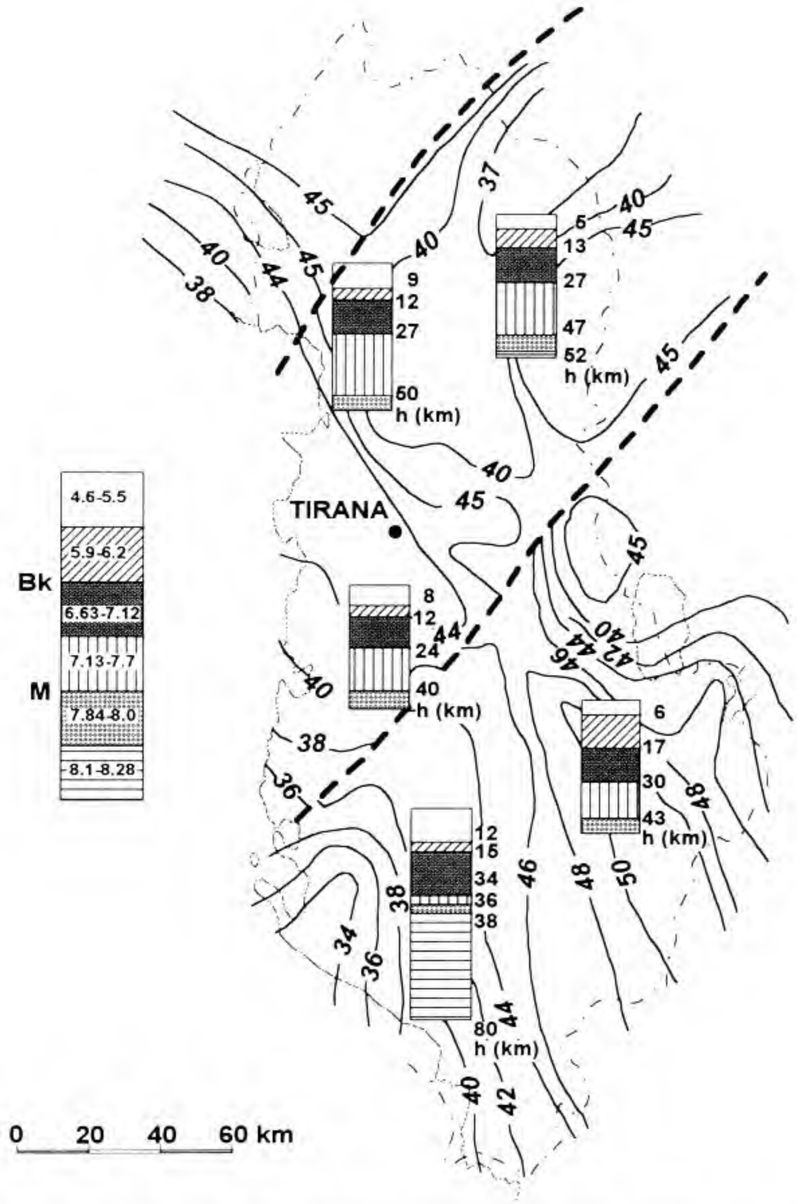

FIG. 4. Crustal thickness of Albania with selected velocity profiles. Contours depth to Moho in km. Bk: top crystalline basement, M: Moho.

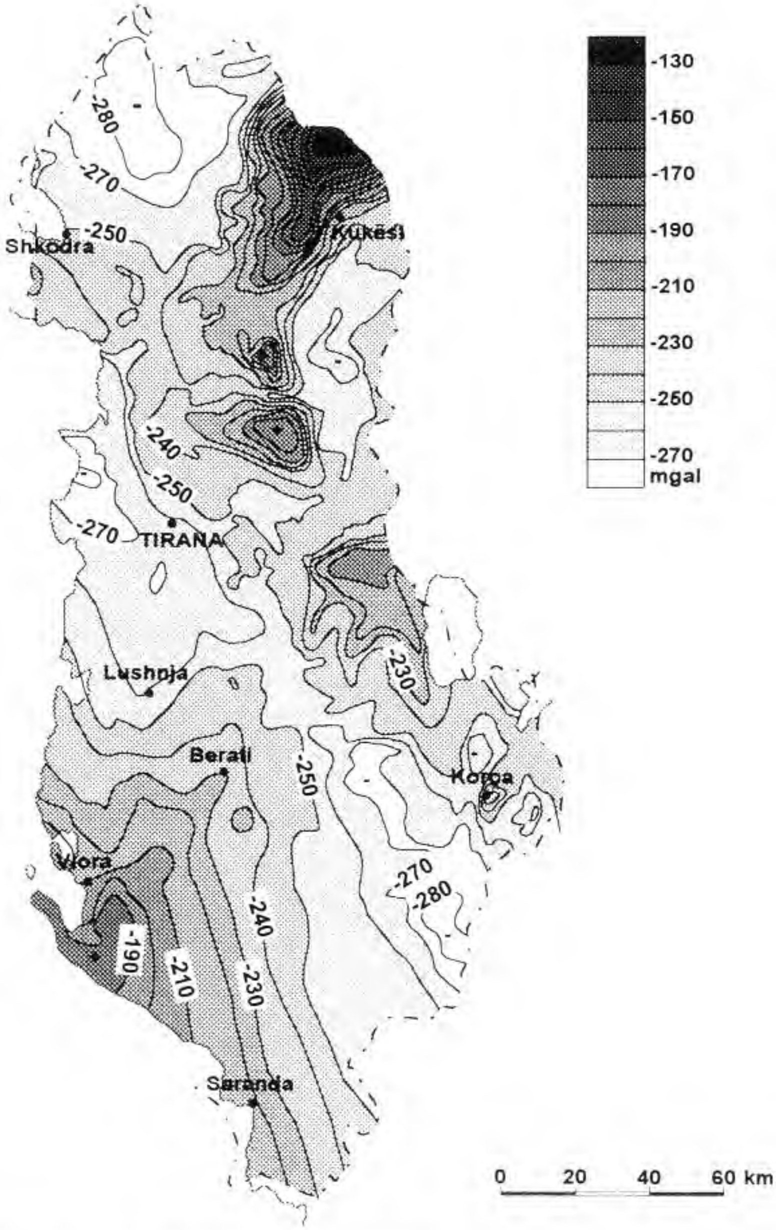

FIG. 5. Bouguer anomaly map of Albania.

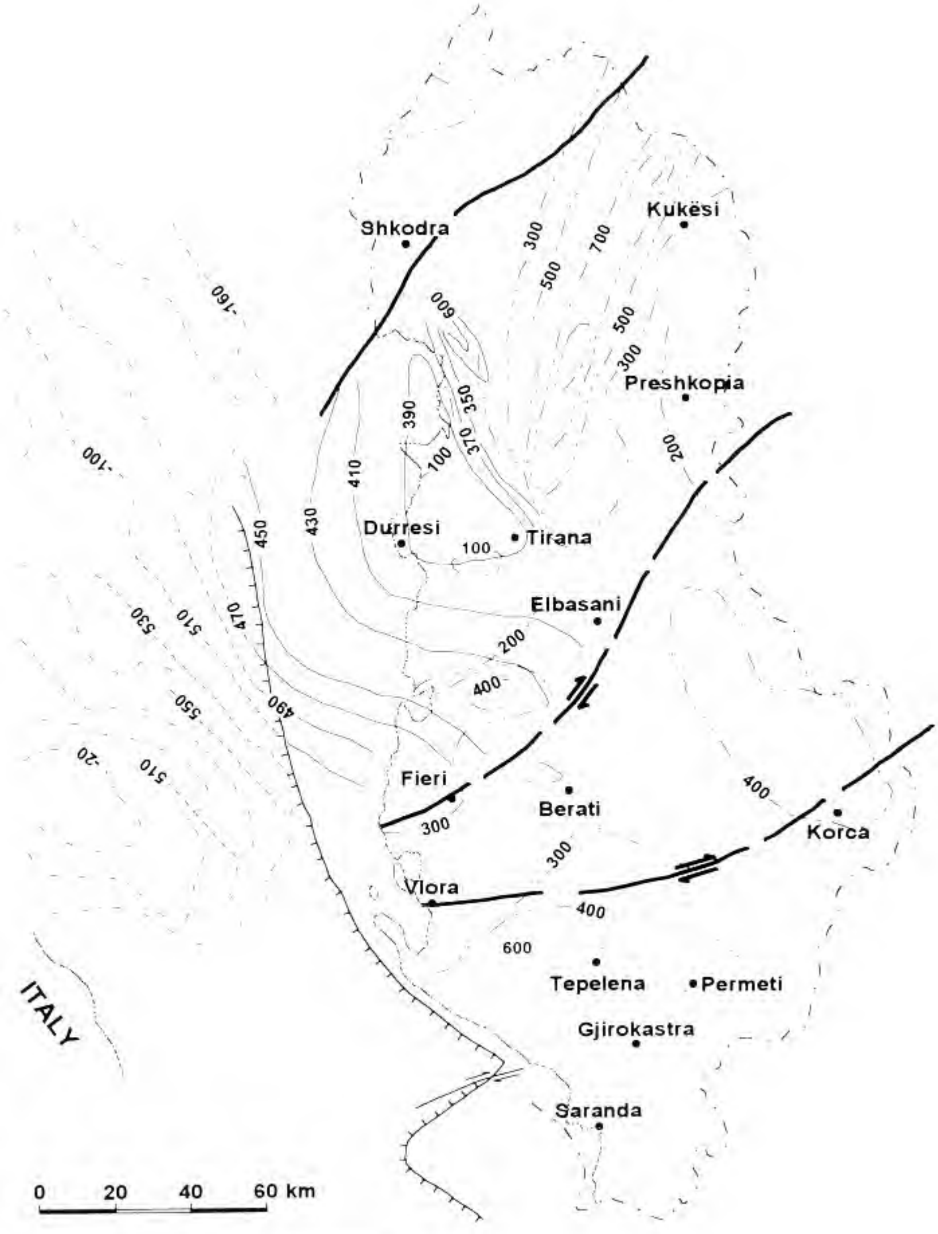

FIG. 6. Aeromagnetic map of Albania.

siderably in thickness and is overlain by Triassic and Jurassic carbonates (Fig. 7). In some areas the sedimentary cover of the ophiolites ranges upward into Late Jurassic (Tithonian) to Early Cretaceous (Berriasian) flysch. Eocene and younger continental clastics were deposited on top of the Mirdita nappe after its emplacement on the Adriatic foreland.

The Mirdita nappe was obducted on the western margin of the Adriatic plate during latest Jurassic. During the main- and late-tectonic phases, the Mirdita nappe was transported westwards on the back of progressively deforming more external sedimentary nappes as a passive orogenic lid which was internally only little deformed. The erosional Peshkopia, Okshtuni and Gramozi windows in the Mirdita nappe correspond to antiform features of the underlying Krasta-Cukali and Korabi nappes (Figs. 1 and 2). During the late-orogenic and neo-tectonic cycles, the extensional Burreli and Korca basins subsided on top of the Mirdita nappe; these form part of the Aegean-Pannonian collapse system and contain up to 5-6 km of Eocene to Late Miocene continental conglomerates and sands (Figs. 1 and 7).

The Mirdita ophiolites are interpreted as remnants of oceanic crustal material of the Subpelagonia Trough. Their basal parts consist of ultrabasic and gabbroic rocks. Ultrabasic rocks range from harzburgite to dunite and display different degrees of serpentinization. Basic rocks include troctolites, gabbro-olivinite, gabbro and gabbro-norite. Higher up in the sequence plagiogranites and quartzdiorites occur. The top part of the unit is formed by 5 km thick volcanic rocks. Gabbro-pegmatite and pyroxenite dykes occur in ultrabasic rocks, gabbro-pegmatites in gabbros and micro-tonalite, plagiogranite and porphyry dykes in plagiogranites and quartzdiorites. These dykes form phyllonitic complexes.

These ophiolitic rocks have average densities of 2610 kg/m^3 and an elevated magnetic susceptibility. Correspondingly, the Mirdita nappe gives rise to strong gravity as well as sharp magnetic anomalies tracking its outlines (Figs. 5 and 6). Figure 5 shows that the Mirdita nappe is characterized by several major gravity anomalies which are separated from each other. These anomalies, which partly also coincide with major magnetic anomalies, correlate with ultramafic massifs (Boltz, 1963; Lubonja et al., 1967).

The cross-sections given in Figs. 2 and 8 give examples of such anomalies. Deep boreholes drilled in the Bulqiza Massif, which is associated with a 48 mGal positive anomaly, have encountered at depths of 1000 to 1320 m ultramafic rocks with densities ranging between 2740 and 3310 kg/m^3 and a predominance of 3180 kg/m^3. Gravity modelling, applying this information, permitted to determined that the Mirdita allochthon attains a maximum thickness of about 6 km in the area of the Bulqiza Massif and overlays the sedimentary Krasta-Cukali nappe which surfaces to the west in front of the Mirdita nappe and to the east in the erosional Peshkopia and Okshtuni windows (Lubonja et al., 1967).

Surface geological data show that the northern parts of the Mirdita nappes are almost separated from their southern parts to the East of Tirana along the southwesterly trending erosional Peshkopia and Okshtuni windows in which the Krasta-Cukali zone outcrops in an antiform structure, involving Paleogene and Cretaceous flysch. These windows are readily recognized on the Bouguer map where they correspond to an elongate negative anomaly (Figs. 5 and 9).

The Mirdita nappe attains a maximum thickness of about 14 km in the ultramafic Kukesi massif in northeaster Albania from where it thins towards the West and Southeast to about 2 km. Away from ultramafic massifs, the magnitude of the Bouguer anomalies decreases rapidly (Fig. 10). However, a direct correlation between the Bouguer and magnetic anomalies cannot be established as ultramafic rocks have a highly variable magnetic susceptibility.which is generally greater than that of the encasing serpentinized ophiolitic rocks. In the southeastern part of Albania, the Mirdita nappe attains thicknesses of up to 10 km near the city of Korca (Fig. 3).

A reflection-seismic profile through the Neogene Burreli Basin, located to the northeast of Tirana, shows that it rests on non-reflective Mirdita ophiolites (Fig. 11). Seismic horizon 1 corresponds to the top of the Mirdita ophiolites and the base of the Neogene sedimentary fill of the Burreli basin. The base of the Mirdita ophiolite nappe corresponds to the strong horizon 2 reflector which gently rises from about 3.0 sec TWT at the eastern end

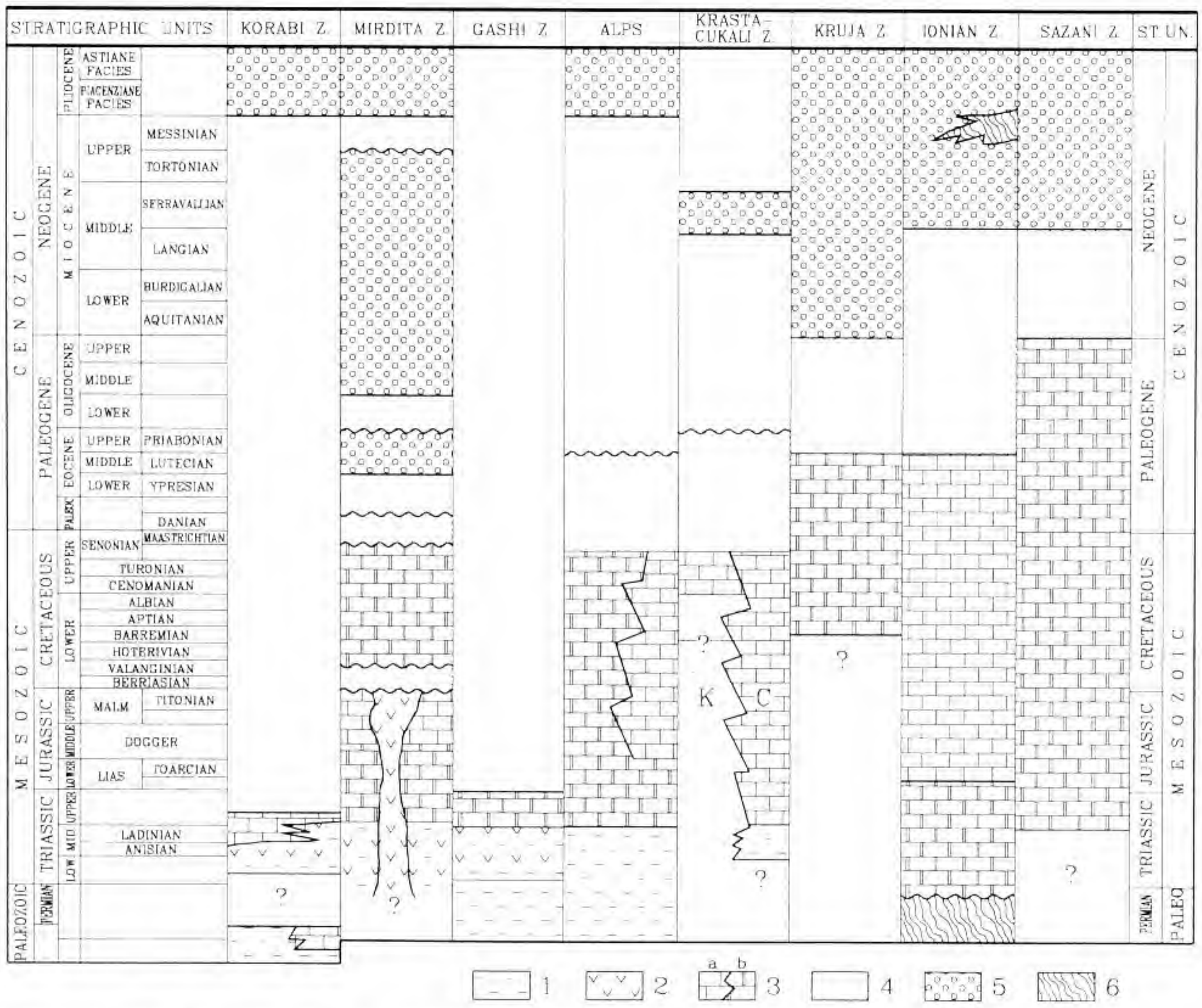

FIG. 7. Chrono- and lithostratigraphic chart of tectonic zones of Albanides.
1) lower terrigenous series, 2) diabase and spillite-keratophyr, 3) carbonates, a: pelagic, b: neritic, 4) flysch and pre-molasse, 5) molasse, 6) evaporites.

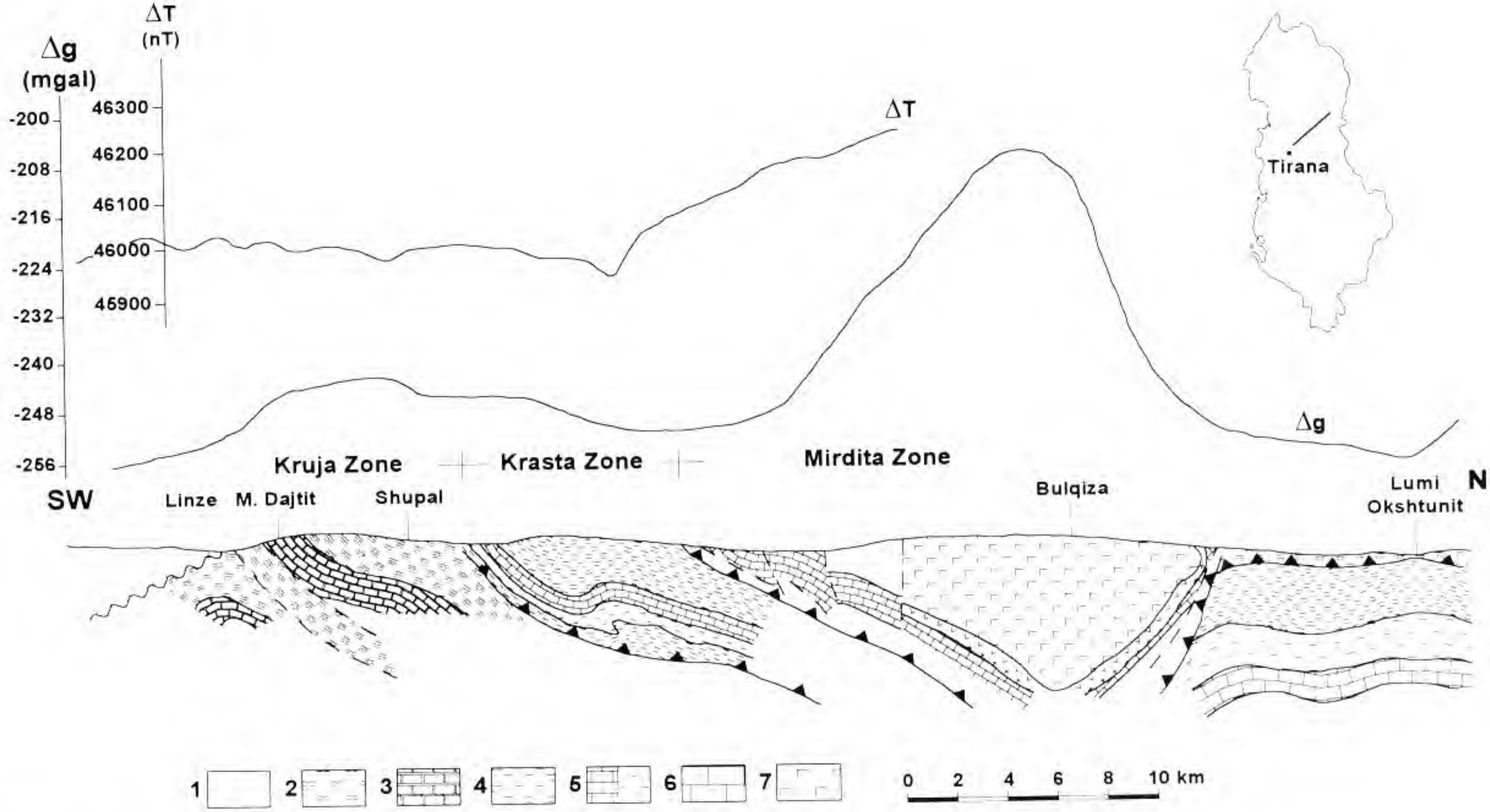

FIG. 8. Geological-geophysical cross-section through Kruja-Krasta and Mirdita zone, northeast of Tirana. Legend: 1) Tortonian, 2) Paleogene flysch, 3) Late Cretaceous-Eocene carbonates, 4) Tithonian-Early Cretaceous flysch, 5) Late Triassic-Early Jurassic carbonates and radiolarites, 6) Late Triassic carbonates, 7) Mirdita ophiolites.

of the profile to about 2.1 sec TWT at its western end. Beneath this sole-thrust reflector, subhorizontal events are attributed to the thrusted sedimentary sequences of the Krasta-Cukali and Kruja zone.

Based on reflection- and refraction-seismic data and gravity modelling, we postulate that autochthonous basement underlies the entire Internides of the Albanides.(Figs. 2 and 3). In view of this, the root of the Mirdita nappe is apparently not located within the territory of Albania and has to be sought further to the East. However, on-trend continuation of the Mirdita nappe into Greece leaves little doubt about its derivation from the Subpelagonian Trough.

Korabi Zone

The Korabi zone involves Silurian and Devonian schistose sandstones, conglomerates and metamorphic limestones and a Carboniferous flysch-type series which were deformed during the

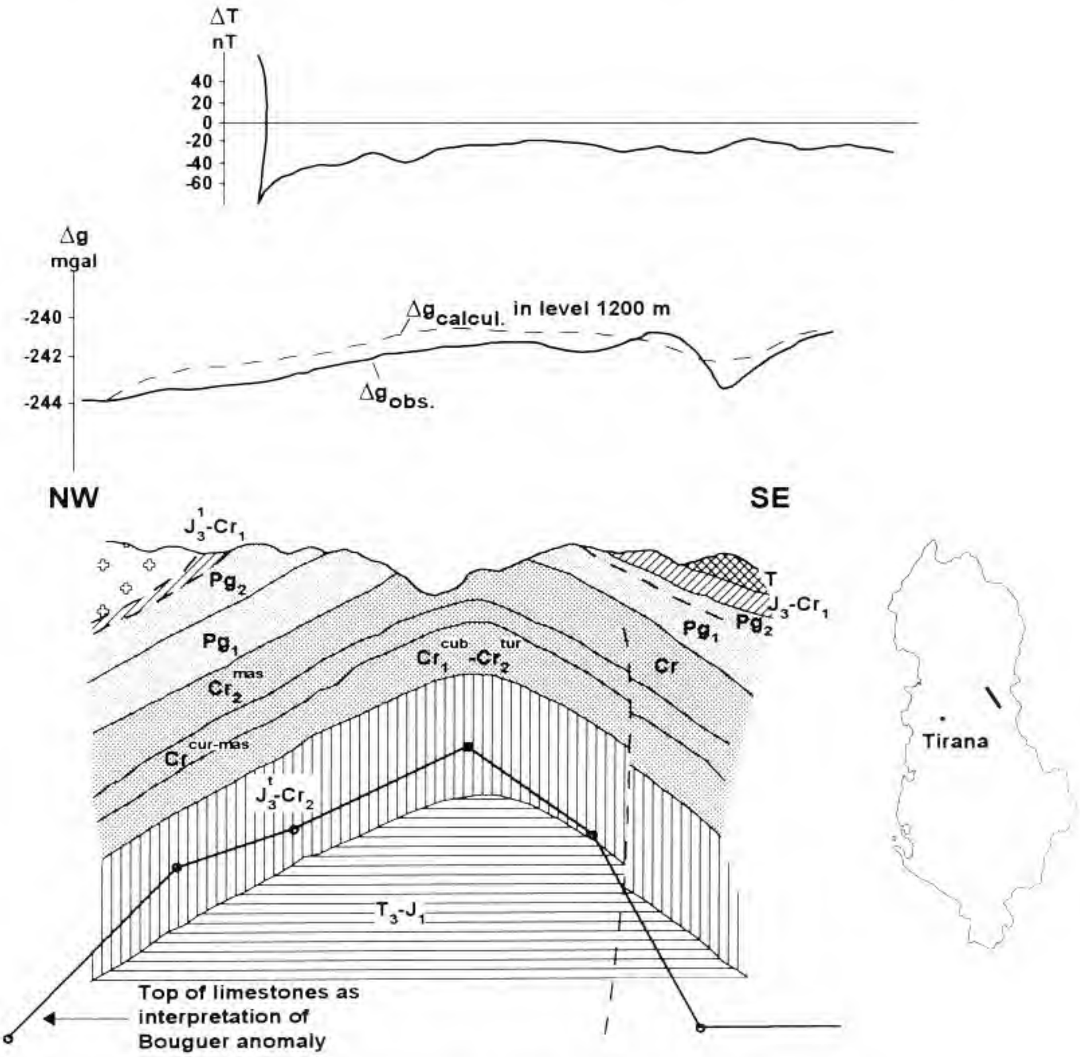

FIG. 9. Geological-geophysical cross-section through Paleogene and Cretaceous flysch exposures of Okshtun window. Abbreviations: T-Triassic, J-Jurassic, Cr-Cretaceous, Pg-Paleogene.

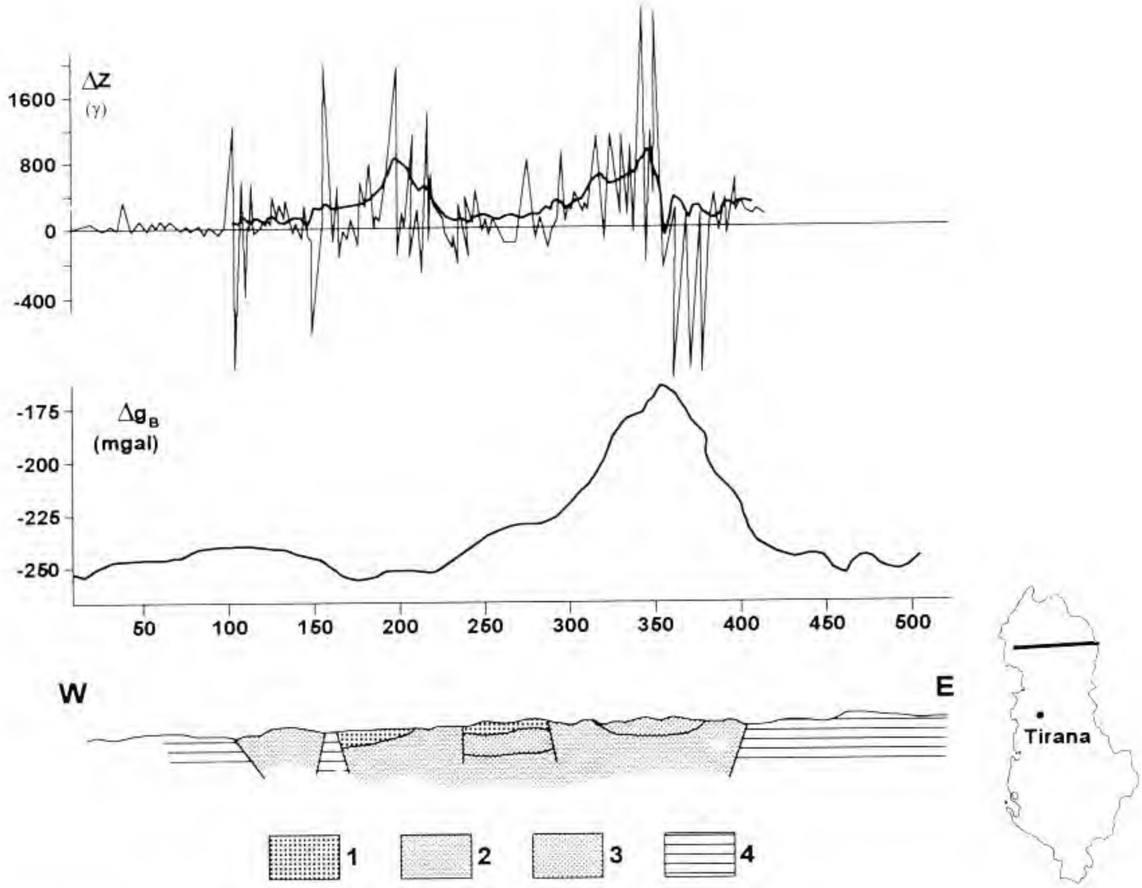

FIG. 10. Geological-geophysical cross-section through Mirdita nappe, Shkoder-Kukes area. 1) effusive rocks, 2) ultrabasic rocks, 3) gabbros, 4) sediments.

Variscan orogeny. These are overlain by Permian conglomerates, gypsum and anhydrites, Early and Middle Triassic clastics and Late Triassic carbonates, containing basic and alkaline volcanics and dykes. After a major hiatus, carbonate sedimentation resumed during the Senonian. Eocene series are developed in molasse facies (Fig. 7). Alpine deformation of the Korabi zone gave rise to the development of open folds and thrusted anticlines, partly cored by evaporites

The tectonic map of Albania, given in Fig. 1, shows that the Korabi nappe is confined to the northern parts of the Peshkopia window whereas in its southern parts the underlying Kruja and Krasta-Cukali nappes rise to the surface.

Gashi Zone

The Gashi zone correlates with the Durmitori zone of the Dinarides and occurs only in the northern-most parts of Albania. It consists of slightly metamorphosed clastic rocks and carbonates ranging in age from Palaeozoic to Mesozoic. Early and Middle Triassic clastics are overlain by Late Triassic carbonates.(Fig. 7) This series is overlain by a basement involving thrust sheet which, in turn, is covered by a sequence of metamorphosed intermediate and acidic and basic volcanics. The latter are attributed a Palaeozoic and Middle Triassic age. Field relationships indicate that the Gashi zone was overridden by the Mirdita nappe.

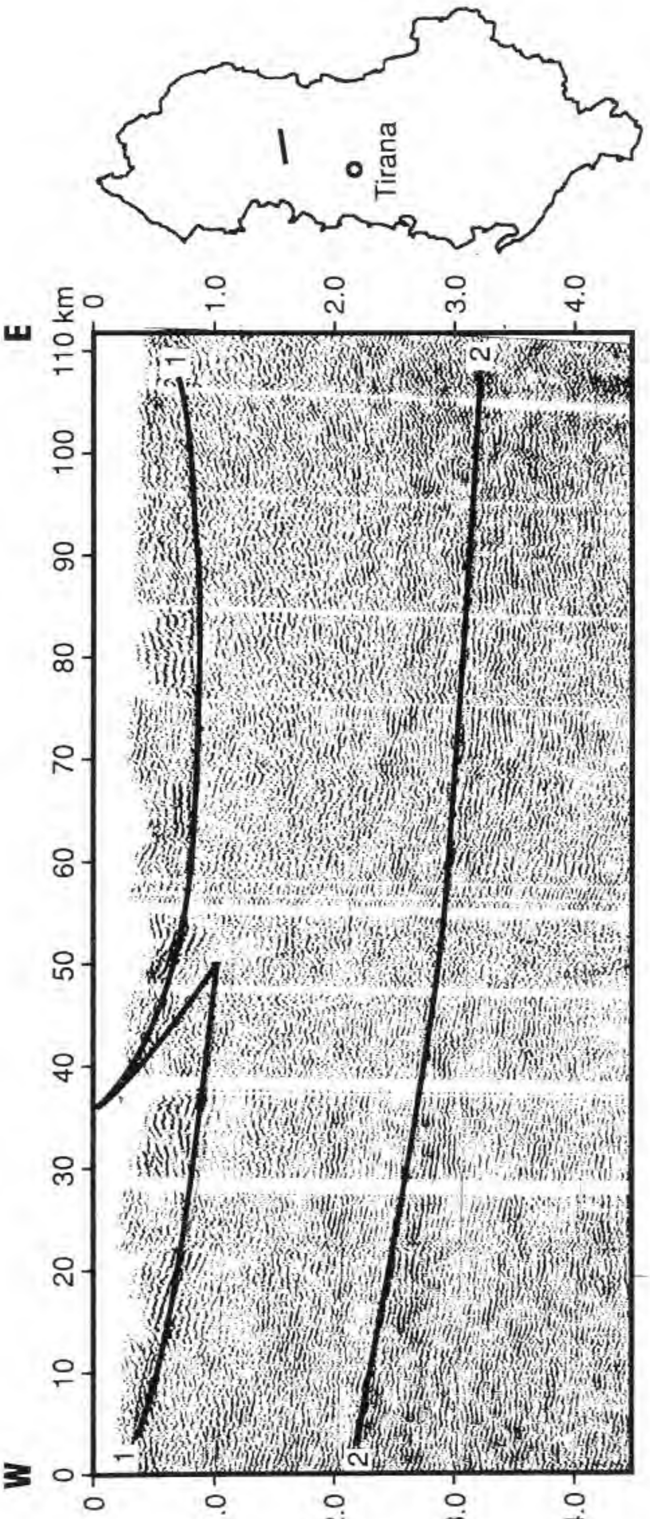

FIG. 11. Reflection-seismic profile through Neogene Burreli Depression (Nishani, 1985). Reflectors: 1) base Neogene sediments, top Mirdita ophiolites, 2) base Mirdita ophiolites.

EXTERNAL ALBANIDES

From East to West, the external elements of the Albanides consist of the Krasta-Cukali zone, which is assigned to the Pindic domain, the Kruja zone which corresponds to the Gavrovo-Dalmatian domain, and the Ionian zone and the Sazani zone which border the Adriatic foreland. In northern Albania, the Albanian Alps zone, which takes in an intermediate position between the Gashi and the Krasta-Cukali zones, is correlated with the Higher Karst zone of former Yugoslavia.

Geophysical data show that the External Albanides are underlain by a little deformed basement complex which deepens from 9-10 km under the Adriatic Sea to about 20 km near the Mirdita thrust front (Figs. 2 and 3). The top of the basement reaches a maximum depth of 12 km in the Durres-Tirana-Shengjini region which is characterized by gravity and magnetic minima (Figs. 5 and 6). Deep crustal fractures, corresponding to seismogenic zones, occur in the Ionian zone to the west of Durres (Fig. 2). The 15th April 1979 earthquake, which had a magnitude of 7.2 on the Richter scale, was associated with the Durres fracture zone. Focal mechanisms show that it was of a compressional nature, suggesting that crustal-scale imbrications, as shown in Fig. 8, are at present tectonically active.

Albanian Alps Zone

In the Albanian Alps, Permian sandstones and conglomerates are the oldest rocks exposed. These are covered by Early and Middle Triassic clastics, interbedded with tuffs. Late Triassic series consist of limestones and dolomites (Fig. 7). Neritic, biogenic limestones containing chert intercalations were deposited during Jurassic and earliest Cretaceous times. Maastrichtian-Danian onset of flysch sedimentation documents the incorporation of the Alps zone into the Albanides foreland basin. Deposition of flysch persisted into Eocene times. The Permian to earliest Cretaceous shelf series attain a thickness of 2300-3100 m whereas the flysch series ranges in thickness between 700 and 1200 m.

The Albanian Alps are characterized by a system of imbricate ramp anticlines which developed during the late-tectonic phase when the Alps zone was thrusted over the Krasta-Cukali zone (Fig. 12)

Krasta-Cukali Zone

The Krasta-Cukali zone correlates with the Pindic zone of Greece. It takes in an intermediate position between the Internides and the Externides of the Albanides and can be subdivided into the northern Cukali and the more internal Krasta zones.

The **Cukali sub-zone** involves Middle Triassic clastics and volcanic flows, Early Triassic to Cretaceous carbonates and includes latest Jurassic radiolarites. Flysch sedimentation commenced during the Maastrichtian and persisted into the Eocene (Fig. 7).

The Cukali zone is characterized by large thrusted anticlines which are overridden by the Mirdita and the Alps nappes. The contact between the Cukali and the underlying Kruja zone is marked by a major thrust fault. The schematic cross-section given in Fig. 12, crosses the southern parts of the Alps, the Cukali zone and its thrust contact with the Mirdita nappe. The gravity anomalies associated with the Alps and the Mirdita nappe are presumably the effect of the great thickness and density of Mesozoic carbonates and the ophiolites, respectively. On the other hand, the systematic gravity gradient across the Cukali zone could be either due to neo-tectonic enveloping of Mirdita ophiolites by the Cukali thrust sheets, or the involvement of a considerable thickness of high density Palaeozoic rocks in the Cukali thrust sheet.

The **Krasta sub-zone** extends as narrow belt from Shkodra to Leskovik (Fig. 1). In this zone Late Jurassic to Albian flysch, Maastrichtian carbonates and Late Maastrichtian to Eocene flysch are exposed (Figs. 7 and 8). On the other hand, flysch prevails in the Cretaceous to Paleogene series of the Okshtuni-Peshkopia window which transects the Mirdita nappe. Geoelectric surveys

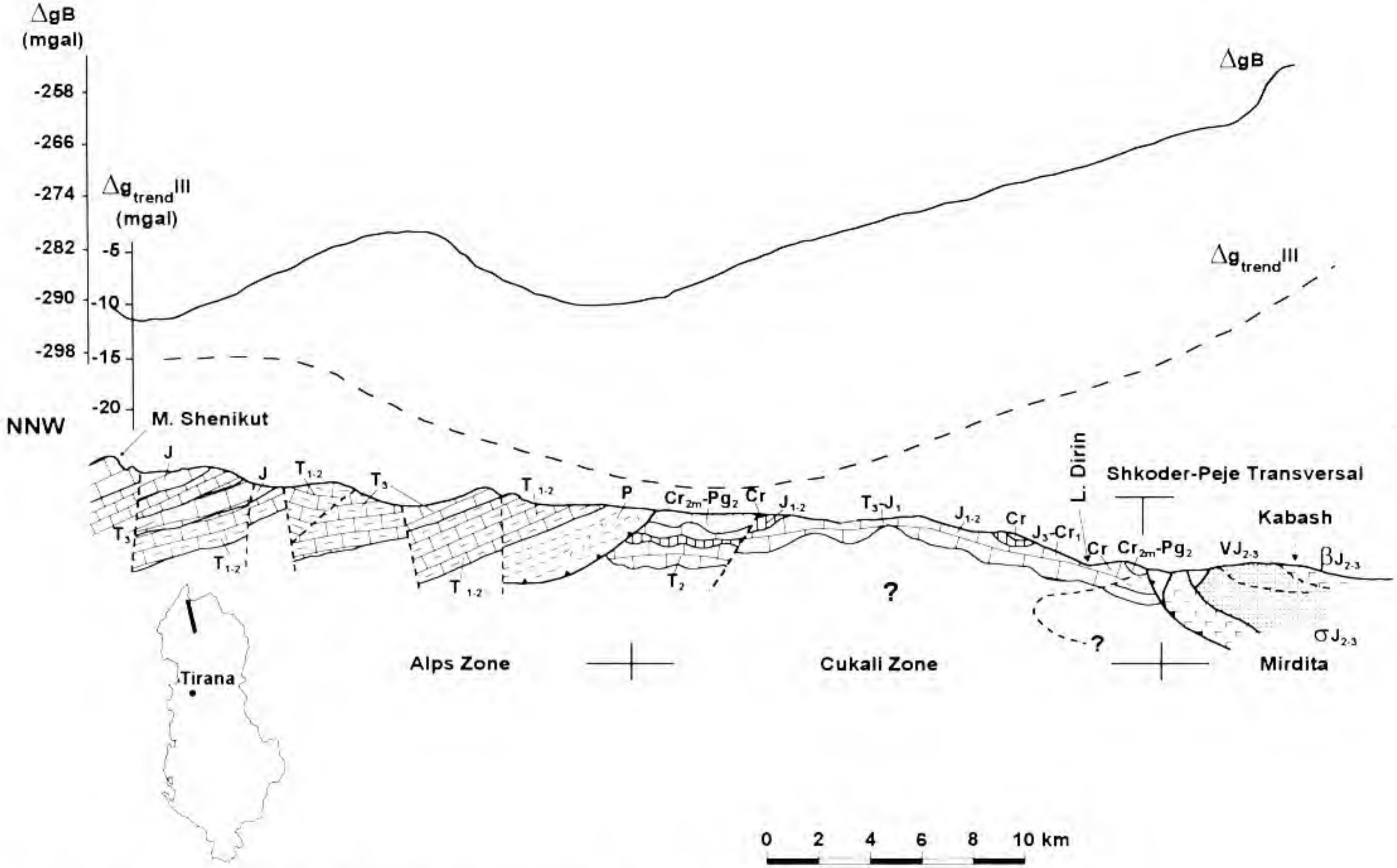

FIG. 12. Schematic geological-geophysical cross-section through Alps, Cukali and Mirdita zone, Northern Albania. Abbreviations: P-Permian, T-Triassic, J-Jurassic, Cr-Cretaceous, Pg-Paleogene.

indicate that the total flysch package has a thickness of 2000-2500 m. Although not exposed, gravity data suggest that the Mid-Cretaceous flysch is underlain by Triassic to Late Jurassic carbonates (Fig. 9). A deep well, drilled on the Okshtun structure, confirming this concept, could open new areas for oil and gas exploration.

Kruja Zone

The Kruja zone forms the prolongation of the Dalmatian zone of former Yugoslavia and links up to the South with the Gavrovo zone of Greece.

This zone is characterized by 1500 m thick Cretaceous to Middle Eocene neritic carbonates and 5 km of Late Eocene to Oligocene flysch (Fig. 7). Main deformations occurred during Mid-Oligocene to early Mid-Miocene times. Locally a Tortonian series, developed in a continental sand facies, rests unconformably on a variety of older strata (Fig. 8).

According to a reflection-seismic profile recorded across the Kruja zone in the area of Tirana (Fig. 13), two distinct, sub-parallel reflectors are observed at depths between 1.8-2.2 sec. TWT (horizon 2) and 2.9-3.3 sec TWT (horizon 3). Horizon 2 was identified as the top of the carbonates. The nature of horizon 3 is still uncertain but could correspond to the base of the Mesozoic carbonates and the top Permo-Triassic salts. Alternatively these reflectors could be related to the sole thrust of the Kruja zone and the top of a deeper, more external tectonic unit, involving a thick flysch package and basal carbonates. This could open up yet an other hydrocarbon play, requiring, however, the acquisition of new reflection-seismic profiles with better resolution than hitherto available.

Ionian Zone

The Ionian zone of Albania is the equivalent of the Ionian zone of Greece. It occupies a large part of the Albanian Externides and corresponds to a thin-skinned fold and thrust belt which is detached from the basement at the level of Permo-Triassic evaporites (Figs. 3 and 14). Late Triassic to Early Jurassic neritic limestones and dolomites are covered by Middle Jurassic to Eocene pelagic limestones containing cherts (Fig. 7). The thickness of these carbonates range between 2.5 and 4 km. Late Eocene to Aquitanian series are developed in flysch and flyschoid facies and attain thicknesses of 4 to 6.5 km. An angular unconformity at the base of the Burdigalian to Serravallian clay and marl series, truncates anticlinal structures and testifies to a first folding phase. Serravallian to Messinian sediments are developed in a molasse-type facies. Burdigalian to Messinian strata attain a thicknesses of up to 5 km.

The Ionian zone hosts 11 producing oil and gas fields (Fig. 17). These are reservoired in Late Cretaceous and Paleogene carbonates and in sands of the Eo-Oligocene flysch series. These accumulations are contained in structural, stratigraphic and combination traps.

Hydrocarbons show geochemical variations which are related to different source-rocks. Main source-rock intervals occur in the Late Triassic and Late Cretaceous carbonate series (Diamanti, 1992).

Sazani Zone

The Sazani zone corresponds to the most external unit of the Albanides and marks the transition to the Adriatic foreland platform. It is characterized by thick Cretaceous to Eocene limestones and dolomites (Fig. 7). Burdigalian to Tortonian marls transgress over an erosional surface and attain thicknesses of up to 5000 m (Fig. 15). Flexural subsidence of this part of the foreland basin was accompanied by Neogene normal faulting. Tortonian and older strata are involved in frontal thrusted structures.

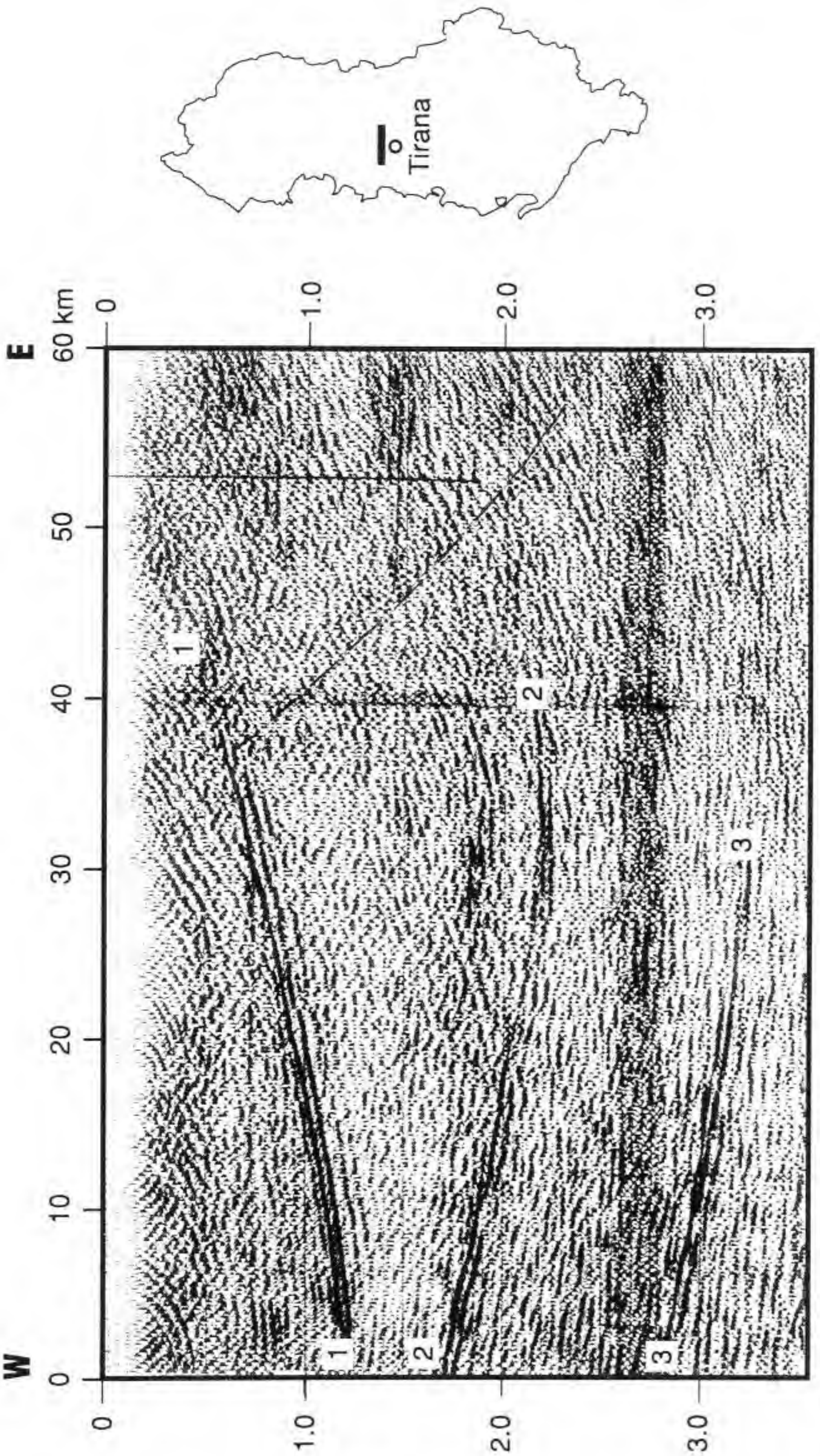

FIG. 13. Reflection-seismic line through Neogene depression near Tirana (Nishani, 1985). Reflectors: 1) base Tortonian, 2) top carbonates, 3) uncertain deep level.

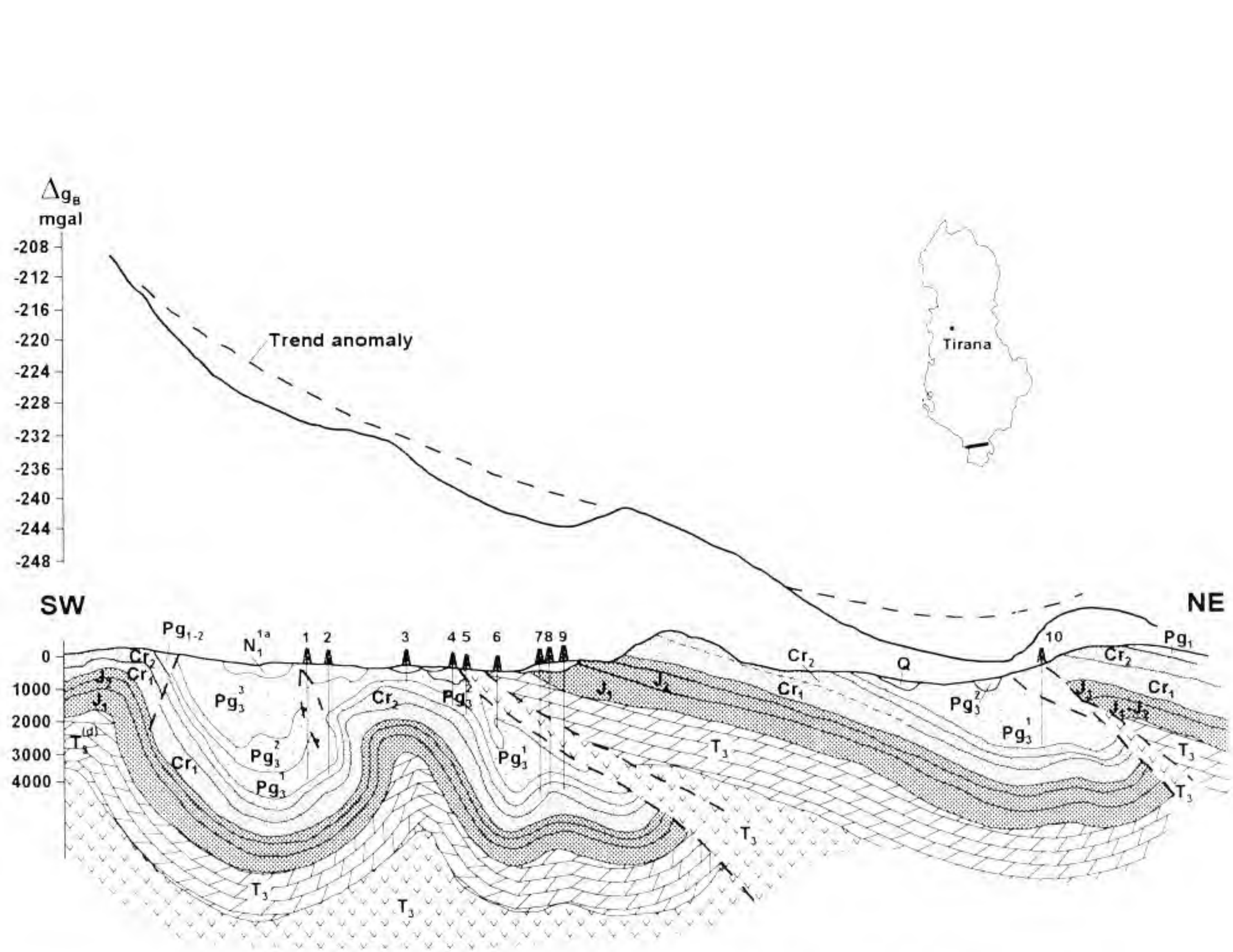

FIG. 14. Geological-geophysical cross-section through Ionian zone, Saranda-Gjirokastra area, Southern Albania. Abbreviations: T_3-Late Triassic evaporites and dolomites, J-Jurassic, Cr-Cretaceous, Pg-Paleogene, N-Neogene, Q-Quaternary.

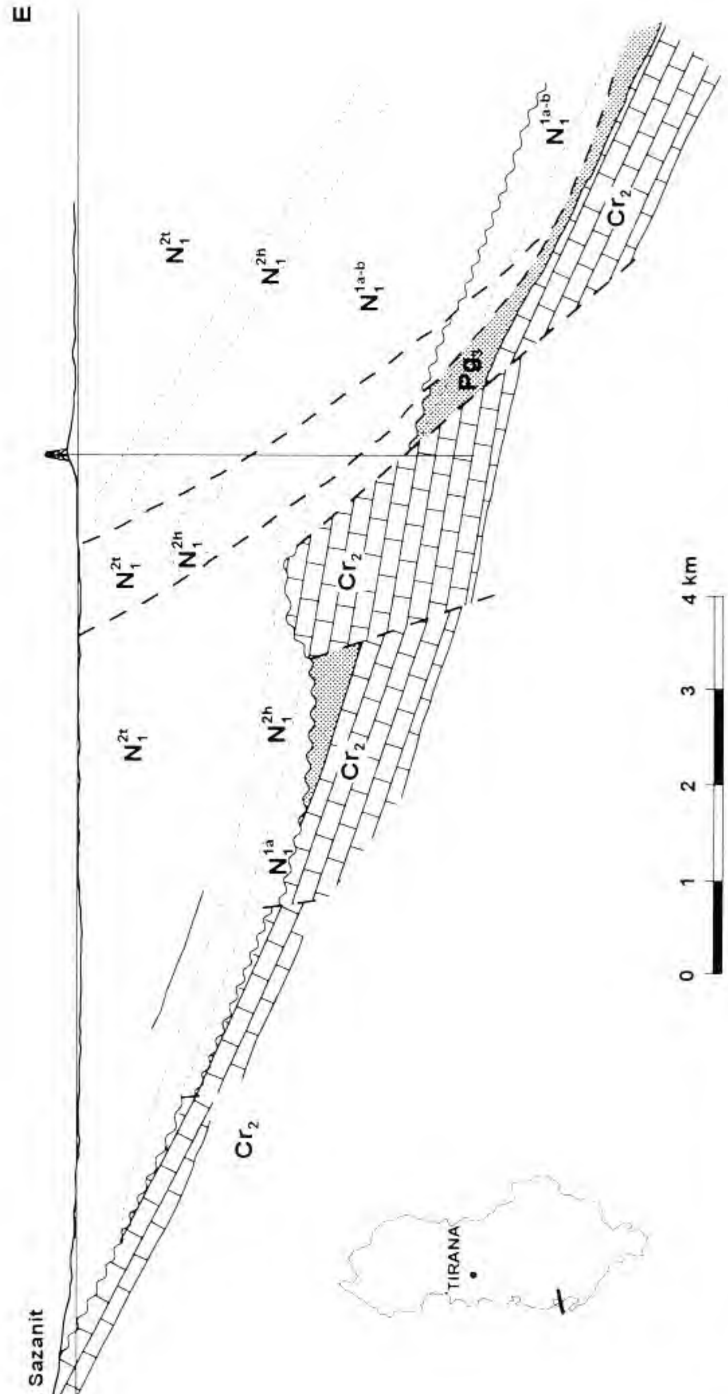

FIG. 15. Structural cross-section through Sazani Zone in Vlora Region. Abbreviations: Cr_2-Late Cretaceous, Pg_3-Oligocene, $N_1{}^{2a-b}$-Aquitanian-Burdigalian, $N_1{}^{2h+t}$-Helvetian and Tortonian.

PERIADRIATIC DEPRESSION

The Periadriatic depression is filled with continental and deltaic Miocene and Pliocene series which prograde into the Adriatic Sea and are covered by Quaternary deposits.

On-shore, the proximal parts of the Periadriatic Depression cover the northeastern parts of the Sazani and Ionian zone and partly also the Kruja zone (Fig. 1). Tortonian sandstones and clays rest unconformably on deformed older strata and are themselves involved in compressional structures (Figs. 8 and 16). Pliocene clays, sandstones and conglomerates rest unconformably on Tortonian and older series and document a two-phase deformation history of this area.

Off-shore, the Neogene sedimentary wedge expands rapidly to some 5-7 km and rests in the central Adriatic on Oligocene flysch, attaining a thickness of 2-3 km, and on Cretaceous and Eocene carbonates; these series are involved in extensional fault blocks (Figs. 2 and 3). The latter give rise to local gravity and magnetic anomalies (Frasheri et al., 1969; Richeti, 1980; Rigo and Caprarelli, 1980). Geophysical data indicate that in the Adriatic the Neogene series are underlain by up to 6 km of Mesozoic strata, which, in analogy with drilling results from the Italian Apulia platform, consist of basal Triassic clastics and evaporites and Late Triassic to Eocene carbonates. Late Triassic series, deposited in depressions can have source-rock characteristics and can provide a hydrocarbon charge to such fault blocks (see Anelli et al., this volume). Away from the central parts of the Adriatic, compressional structures play an increasingly important role as the coast is approached. The occurrence of possibly crustal scale thrusted anticlinal features, as shown on Fig. 2, is indicated by the available geophysical data. Such features are characterized by positive gravity and negative magnetic anomalies, suggesting uplift of thick carbonate units. Such features are located on trend with the Sazani zone. A further uplift of the Mesozoic carbonates may occur in the Durres area; this feature is interpreted as a deep-seated ramp anticline which is detached from the basement at the level of Permo-Triassic evaporites.

In the on-shore parts of the Periadriatic Depression, so far eight commercial oil and gas fields have been discovered (Fig. 17) These are reservoired in the Neogene molasse series, which provide for stacked accumulations in structural and stratigraphic traps. At shallow depths, tar deposits and accumulations of biodegraded oils occur. At greater depths, undegraded oil accumulations with gas caps and gas/condensate accumulations with oil legs are present. Hydrocarbon generation and migration had commenced already during the deposition of the Tortonian series, that is prior to the Early Pliocene final structuration of the area. Multiple source-rock intervals, having a regional extent, are recognized (Fig. 18; Diamanti, 1992).

CONCLUSIONS

The Albanides are a typical, west-verging Alpine thrust belt, the internal parts of which are characterized by a stack of ophiolitic and sedimentary nappes whereas its external parts are formed by thin-skinned thrust sheets.

Geophysical data have greatly assisted in understanding the architecture of the Albanides. Gravity modelling indicates that the Mirdita ophiolite nappes attains thicknesses ranging between 2 to 14 km. Reflection seismic data confirm the nappe structure of the Albanian Internides and define the thrust front of the Ionian and Sazani zones towards the undeformed Adriatic foreland. Moreover, geophysical data indicate that the foreland crust extends essentially unbroken for at least 100 km from the thrust front of the Albanides under their internal nappes where its top is located at depths of about 20 km. Correspondingly, the Albanian sedimentary basin extends from the Adriatic Sea deep under the Externides and Internides of the Albanian orogen. Earthquakes indicate that neo-tectonic deformations include the compressional reactivation of crustal-scale fractures.

The onset of flysch deposition in the different tectono-stratigraphic zones of the Albanides and their foreland indicates that latest Jurassic-earliest Cretaceous obduction of the ophiolitic Mirdita

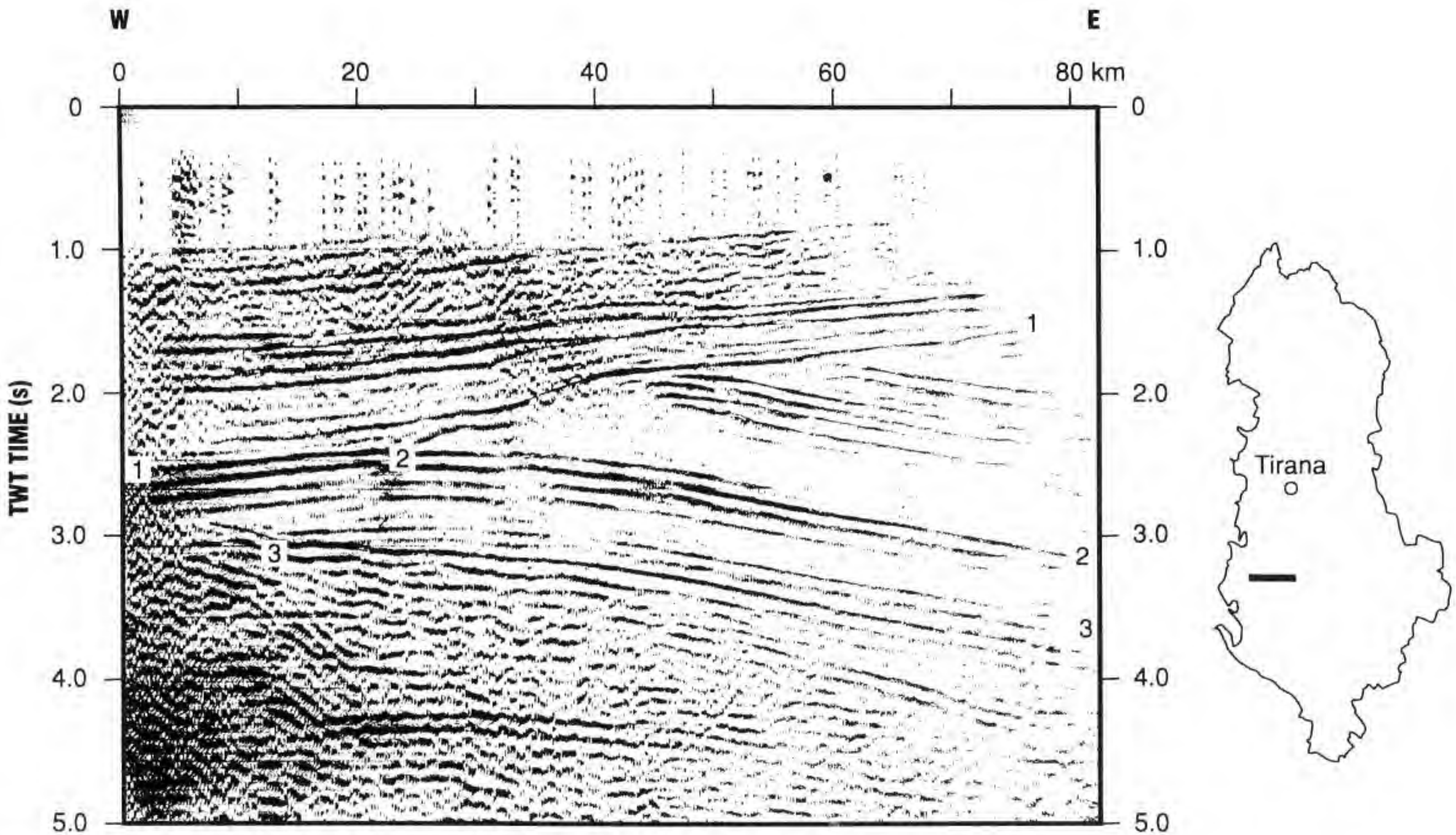

FIG. 16. Reflection-seismic profile through Periadriatic Depression, northwest of Berati. Reflectors: 1) base Tortonian unconformity, 2) top carbonates, 3) deep event.

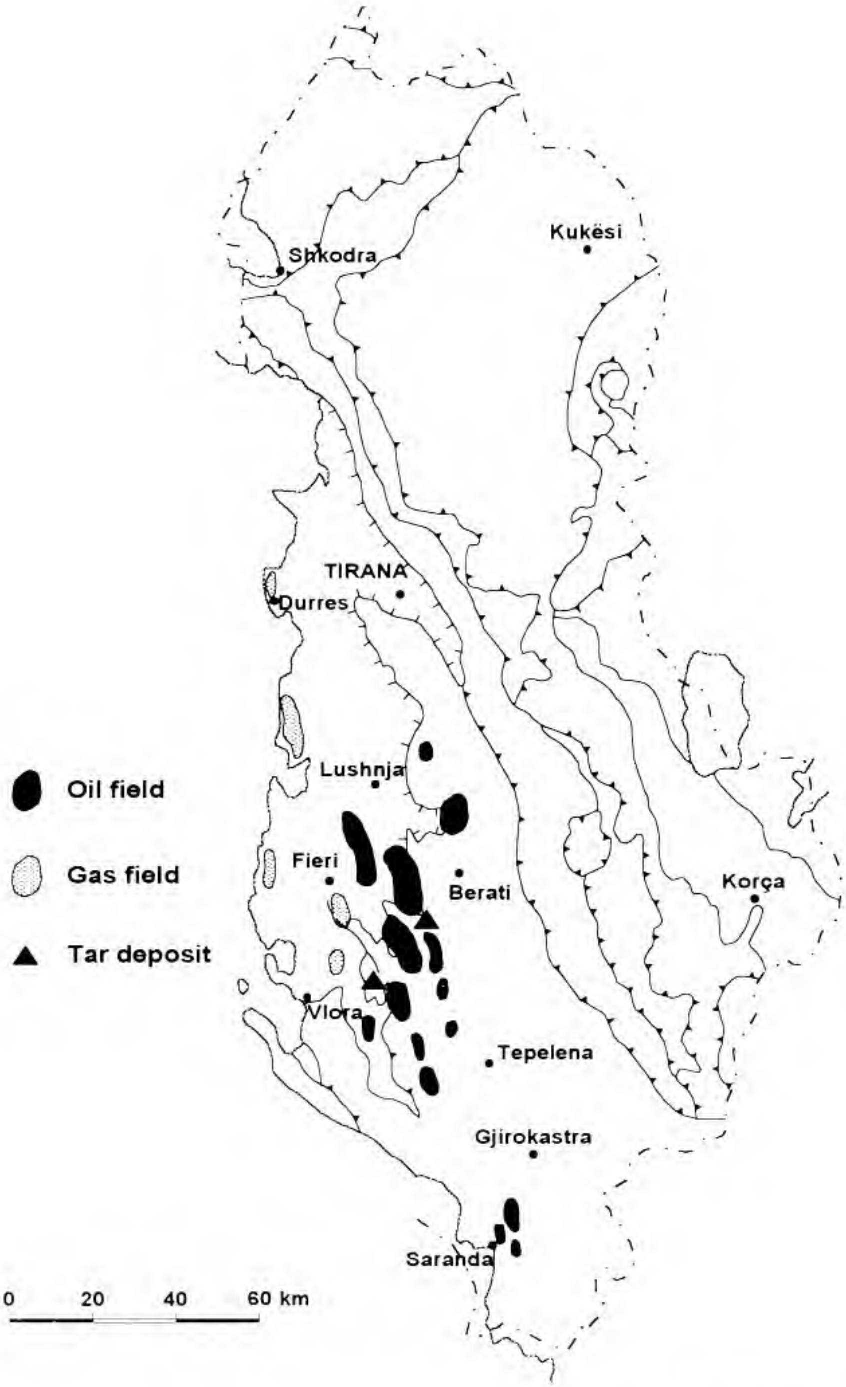

FIG. 17. Oil and gas fields of Albania. Tectonic boundaries are the same as in Fig. 1.

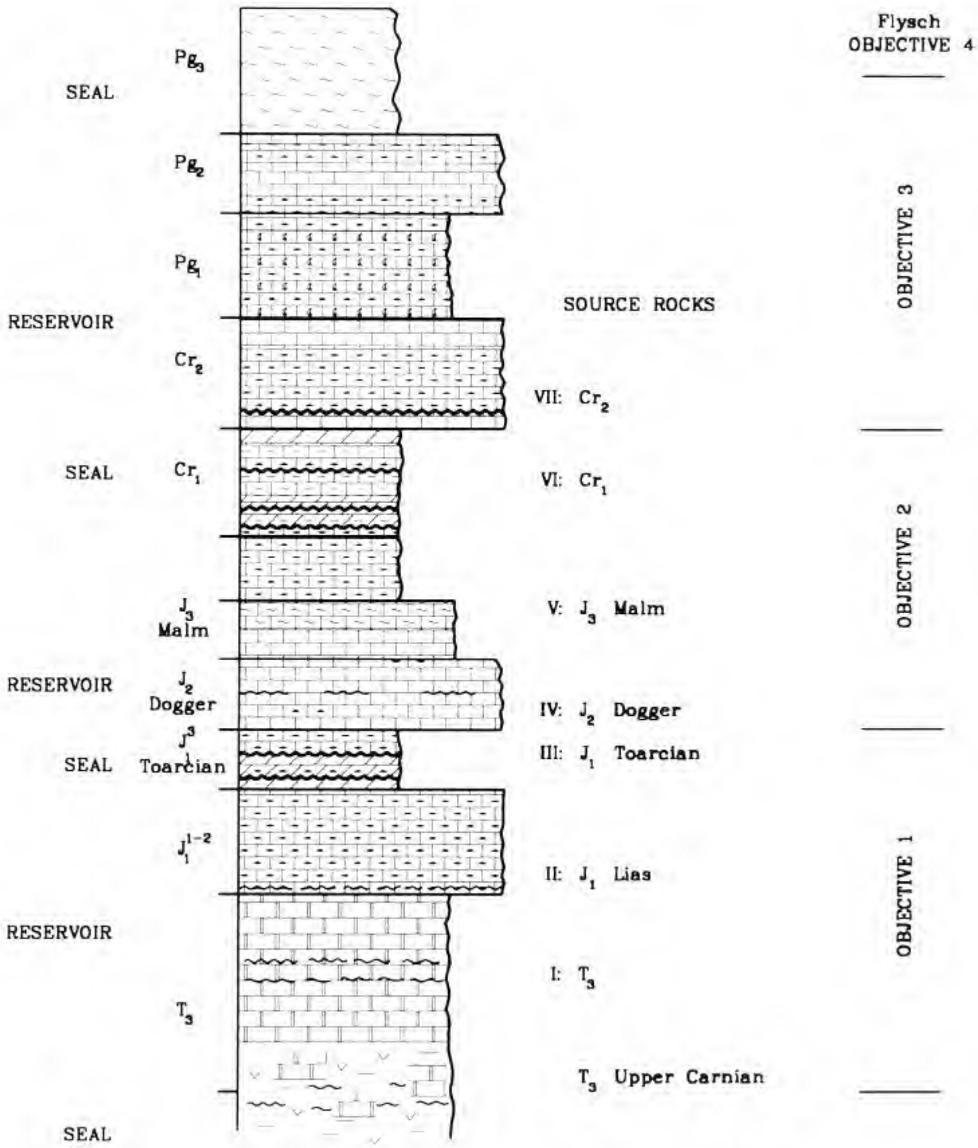

FIG. 18. Stratigraphic column showing position of main source-rocks in Albania.

nappe was accompanied by the development of a flexural foreland basin. During the Cretaceous to Neogene orogenic phases of the Albanides, progressively more external zones were incorporated into this foreland basin whereas more internal zones became involved in an orogenic stack of nappes. The main phase of nappe displacement occurred during the Paleogene when the Krasta-Cukali and Kruja nappes were emplaced. Neogene two-phase deformations characterize the Ionian zone and the Periadriatic Basin.

The thin-skinned thrusted and folded structures of the Ionian zone and the Periadriatic Basin host a number of oil and gas accumulations. Several source-rock intervals, having a high oil and gas generation potential, are recognized within the Mesozoic carbonate series. Multiple reservoir/seal pairs are associated with the Neogene molasse series which is the main producers in the Albanian Basin. Additional reservoirs are developed in the Mesozoic to Eocene carbonates. These are involved in folds, ramp anticlines and imbricate thrust sheets. Prospects in the Adriatic off-shore have a high priority since they are located in water depths of often less than 100 m.

The most important oil and gas accumulation are found in the Ionian and the Periadriatic Depression which extends into the Adriatic off-shore. Structuration of the Ionian and Sazani zones occurred during the late- and neo-tectonic phases. The carbonate-dominated Late Triassic to Late Cretaceous series of the Ionian, Kruja and Krasta-Cukali zones contains several rich to very rich source-rock intervals. In the Ionian zone, Late Cretaceous, Paleocene and Eocene carbonates and Oligocene flysch-type sandstones form the reservoirs of the main oil and gas accumulations. The Tortonian-Pliocene Molasse-type clastics of the Periadriatic Depression contain source-rocks and mainly stratigraphically trapped gas accumulations.

Established hydrocarbon accumulations are areally and depth-wise restricted. The great thickness of the sedimentary series, the occurrence of good quality source-rocks at several stratigraphic levels, the suitable relationship between source-rocks and reservoir/seal pairs and the structuration of the external zone of the Albanides provide excellent conditions for entrappment and preservation of major hydrocarbon accumulations, not only at shallow but also at greater depths. In view of the above, we conclude that the remaining hydrocarbon potential of the Albanian on-shore and off-shore is significant.

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Maps:

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Tectonic map of Albanian Republic, 1: 200.000. Inst. Geol. Studies, Faculty of Geology and Mining, Polytechnic University, Tirana, 1985

Source : MNHN, Paris

Crimean orogen: a nappe interpretation

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ABSTRACT

Detailed surface geological analyses and a review of subsurface and published palaeontological data refute the long held notion that the area of the Crimean Highlands was affected by a major Late Triassic-Early Jurassic orogeny, referred to as the Early Cimmerian orogenic pulse. The Taurian flysch, which contains abundant reworked Carboniferous to Jurassic rock fragments and which was previously thought to be of Late Triassic to Early Jurassic age, has been redated on the basis of Ammonites as Hauterivian to Aptian. Correspondingly, an older-over-younger relationship, typical for nappe tectonics, has been established for the Late Jurassic carbonates, the Taurian flysch and the Albian-Aptian autochthonous series which crop out in the Crimean Highlands.

The topographic relief of the Crimean Highlands is upheld by 800-1000 m thick Late Jurassic reefal and platform carbonates forming the Yayla nappe. In the southern parts of the area this nappe rests on the Taurian nappe and further north on autochthonous series. The Taurian nappe is composed of Hauterivian to Aptian flysch. The Yayla and Taurian nappes each account for horizontal northward transport of supra-crustal rocks over a distance of at least 30-40 km; their root zones, which may be located off-shore, have not yet been identified. The youngest autochthonous strata which are overridden by these nappes yield a middle Late Albian age. The thrust contact between the Taurian nappe and the autochthon is sealed by Cenomanian limestones. Therefore, the main deformation phase of the Crimea, during which the Yayla and Taurian nappes were emplaced, is dated as Late Albian and thus correlates with the Austrian phase of the Alpine orogenic cycle.

In the northeastern part of the Crimea, subsurface data give evidence for Eocene and younger compressional reactivation of the frontal parts of the Yayla nappe, corresponding to the Vladislavovka nappe which involves Late Jurassic to Paleogene strata. The effect of these Late Alpine deformations on the Crimea Highlands is difficult to evaluate for want of a Late Cretaceous and younger stratigraphic record.

Popadyuk, I. V. & Smirnov, S. E., 1996. — Crimean orogen: a nappe interpretation. *In*: Ziegler, P. A. & Horváth, F. (eds), Peri-Tethys Memoir 2: Structure and Prospects of Alpine Basins and Forelands. *Mém. Mus. natn. Hist. nat.*, **170**: 513-524. Paris ISBN: 2-85653-507-0.

INTRODUCTION

The southern most parts of the Crimean Peninsula are occupied by highlands which are upheld by folded and thrusted, partly reefal Late Jurassic carbonates and Early Cretaceous flysch series. These northeasterly trending highlands, which rise to an elevation of nearly 1500 (highest peak Roman-Kosh Mtn. 1543 m elevation), have a length of some 170 km and are up to 50 km wide. Geophysical data indicate that the Crimean orogen extends off-shore in the direction of the Dobrogea and the Greater Caucasus as well as to the south into the deeper waters of the Black Sea (Khain, 1994).

The Crimean Highlands are one of the best studied parts of the Alpine chains of Eastern Europe. Surface geological mapping had already commenced during the second half of the 19th century. By the late 1960's more than 1300 papers had been published on this area. This voluminous literature was synthesized by M.V. Muratov in volume VIII of the "Geology of the USSR" (1969). In this fundamental work the classical model for the evolution of the Crimean orogen was developed. According to this model the geosynclinal stage of the area terminated with the compressional deformation of the Taurian flysch group during Late Triassic-Early Jurassic times. This Early Cimmerian orogenic pulse was supposedly followed by a prolonged period of tectonic quiescence during which Late Jurassic carbonates were deposited on the eroded Early Cimmerian fold belt. Compressional deformation of the area resumed only during the Neogene.

In this model the Crimean fold belt was regarded as an inverted basin in the foreland of the Alpine chains that lacked the classical nappe tectonics which characterize the true Alpine orogens. Although, with the gradual acceptance of plate tectonic concepts, this interpretation began to be seriously questioned, Muratov and Tseisler (1982) continued to adhere to their model. On the other hand, after the publication of the paper by S.L. Byzova (1980), Khain (1994) visualizes a more complex evolution of the Crimean Highlands, involving Early Cimmerian deformation of the Taurian flysch trough, followed by Late Cimmerian (pre-Tithonian) southward thrusting of the Late Jurassic carbonates and Late Alpine deformation of the eastern parts of the Crimean orogen.

Yu.V. Kazantsev (1982) were the first to revise the classical model of Muratov (1969) in the sense of nappe tectonics, however, without considering a revision of the age of the Taurian flysch complex. Although these early efforts were strongly criticized (Archipov et al., 1983a, 1983b; Byzova et al., 1983), additional impulses were given to the more mobilistic nappe model by the recognition of nappes in the Kerch Peninsula located to the east of the Crimean Highlands (Kazantsev and Beher, 1987; Kruglov and Tsypko, 1988). This discovery encouraged the author of this paper to carry out further surface geological studies in the Crimean Highlands. In this respect the excellent and readily accessible outcrops in the Salgir, Tonas and Sukhoy-Indol valleys were closely studied (Fig. 1). Based on the results of these studies, as well as on a review of publications and available subsurface data, a nappe model was developed for the Crimean Orogen (Popadyuk and Smirnov, 1991). In this paper we discuss the rationale which underlies this model and support our concepts with a series of structural cross-sections through the Crimean Highlands (Fig. 3).

TECTONO-STRATIGRAPHIC UNITS OF THE CRIMEAN HIGHLANDS

The distribution of the major tectono-stratigraphic units making up the Crimean Highlands, as well as the location of the transects discussed below, are given in Fig. 2. In essence five tectono-stratigraphic units are recognized. These are, in ascending order, the autochthonous unit which includes Early Cretaceous and Middle Jurassic strata, the allochthonous Taurian flysch group, the age of which was for a long time a matter of dispute, the allochthonous Late Jurassic Yayla unit and the largely post-tectonic Late Cretaceous to Cenozoic unit.which records evidence of Late Alpine deformation.

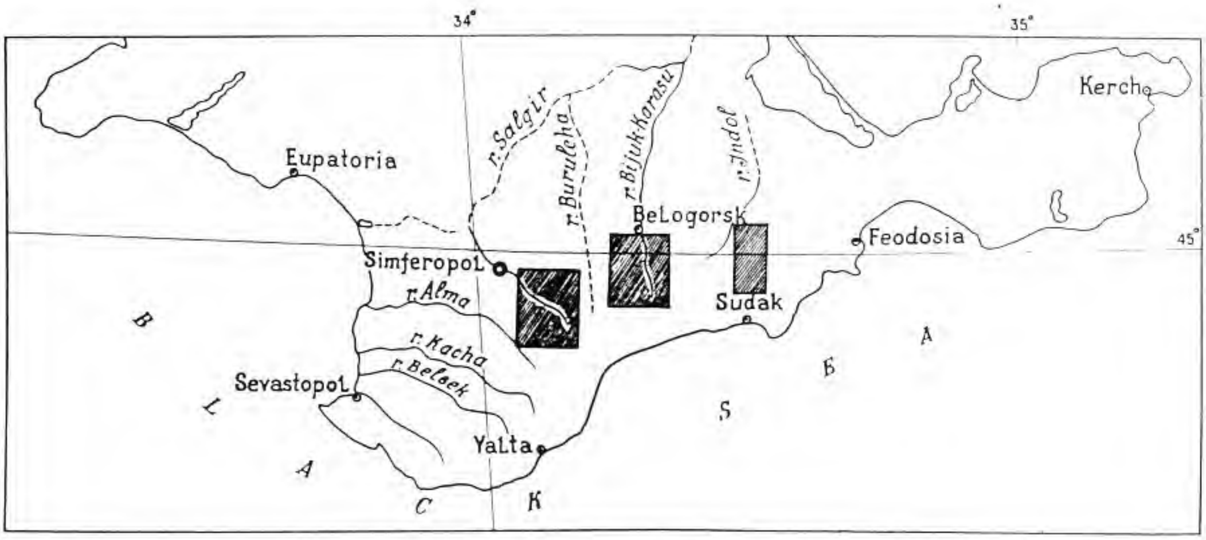


FIG. 1. Location map showing areas of detailed field work.

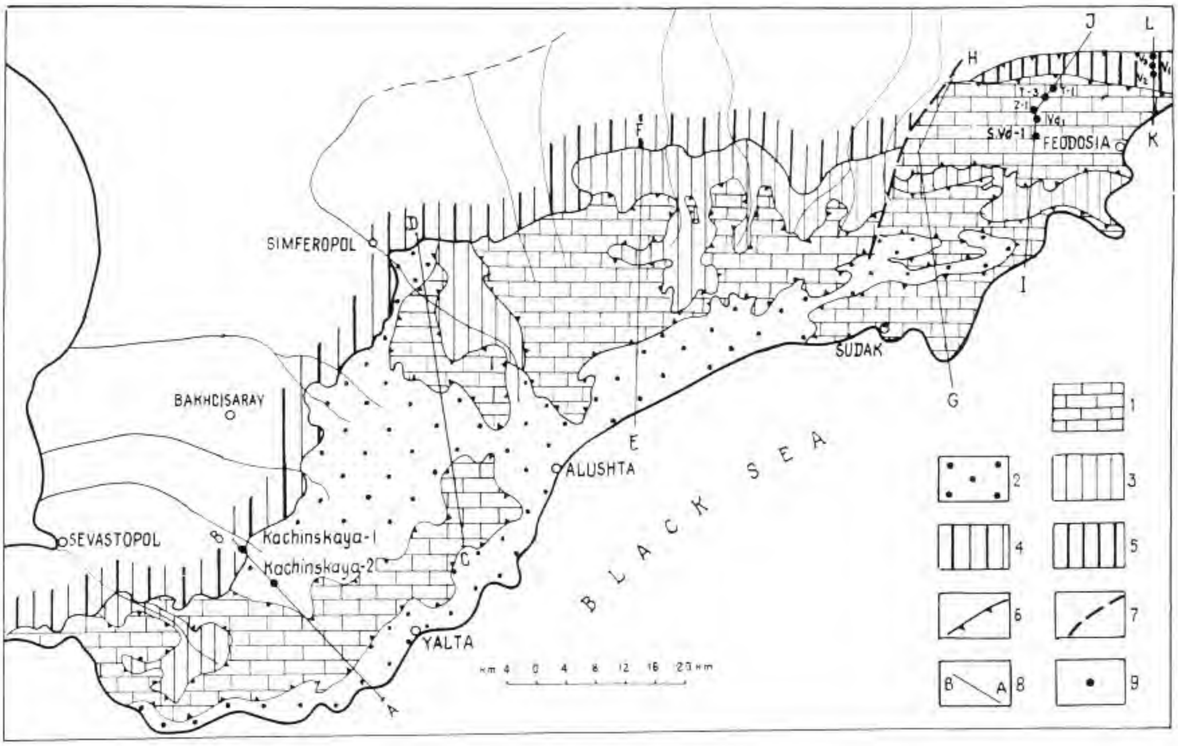


FIG. 2. Tectonic sketch-map of Crimean Orogen.
1: Yayla nappe; 2: Taurian nappe; 3: Early Cretaceous autochthonous series; 4: Late Cretaceous-Miocene neo-autochthonous series; 5: Vladislavovka nappe; 6: main thrust-faults; 7: postulated normal fault; 8: location of cross-sections given in Fig. 3; 9: well locations

Autochthonous Unit

The structurally lowest unit, which outcrops in the northern valleys of the Crimean Highlands, is referred to in our model as the autochthonous unit. In outcrops it consists of generally weakly deformed dark-gray, plastic Albian and Aptian clays, Barremian thin bedded flyschoïd sands, silts and shales, Hauterivian to Valanginian rhythmically bedded calcareous shales, silts and limestones and Berriasian calcareous shales and limestones. According to surface geological mapping and the results of boreholes, this sequence attains thicknesses of the order of 3.5 to 4.5 km and rests disconformably on a Middle Jurassic continental, coal-bearing series which exceeds 1 km in thickness. There is no information on the age, lithology and thickness of the pre-Mid-Jurassic autochthonous series and the basement on which it rest. Where control is available, the is no evidence for Late Jurassic carbonates in the autochthonous series, either due to their non-deposition or due to pre-Cretaceous erosion.

Taurian Flysch Unit

The Taurian flysch group, which was already recognized by Vogdt in 1902 as a special unit (Menner et al., 1947), overlays in outcrops Albian and older autochthonous strata. The Taurian flysch consists of intensely deformed dark-coloured, often black, rhythmically and thinly bedded shales, siliciclastic sands and silts. Within this sequence a coarse-clastic facies, referred to as the Eski-Odra formation, is locally recognized; it consists of sandstones, fine to coarse conglomerates and in some cases even giant limestone blocks which yield fossils of Carboniferous, Permian, Triassic and Lower Jurassic age. In some areas the Taurian flysch contains diabase bodies and tuffs.

The Taurian group has yielded numerous fossils such as Ammonites, Brachiopodes, Pelecipods, Gastropods, Beleminites, Crinoides, Foraminifera and plant imprints. As the majority of these fossils stem from carbonate olistoliths contained in the coarse-clastic Eski-Odra facies, they must be regarded as reworked. Their age ranges from Carboniferous to Permian, Middle and Late Triassic to Early and Middle Jurassic. The shales of the Taurian group contain only rare fossils. Conventionally the Taurian flysch group has been assigned a Late Triassic age. The Eski-Odra formation, regarded as forming part of the Taurian Group, was, however, thought to have an Early Jurassic age. By some the Taurian flysch was thought to range downwards into the Middle Triassic (Dagis and Shvanov, 1965) and the Eski-Odra formation to be as young as Middle Jurassic (Shalimov, 1960).

In some areas of the western Crimean Highlands, the shales of the Taurian group contain limestone blocks and olistostroms which yield early and middle Early Jurassic fossils Menner et al., 1947, see pp. 69-70). The occurrence of Early Jurassic faunistic remnants was also noted by Muratov (1960), Kazakova (1962) and Koronovsky and Mileyev (1974). A.S. Moiseev, one of the best experts in Crimean geology, remarked explicitly that the Late Triassic limestone blocks "represent themselves only blocks within the Taurian shales, similar to blocks of Permian rocks" (see Menner et al., 1947, pp. 58). A similar view was held by O.G. Tumanskaya who observed between the Bodrak and Salgir rivers that Permian, together with Liassic and Triassic limestones, formed a cliff consisting of blocks of different size which had settled into Triassic shales containing *Pseudomonotis caucasica* (see Menner et al., 1947, pp. 55). On the southern shore of the Crimea, near the village of Rybachiye, located to the west of Sudak, Taurian flysch equivalent strata yielded Bathonian fossils (Lychagin et al., 1956). From the Eski-Odra sequence, considered as having an Early Jurassic age, Middle Jurassic as well as Permian and Triassic fossils were retrieved (Muratov, 1960, 1969; Koronovsky and Mileyev, 1974). According to Shalimov (1960), the lower levels of the Eski-Odra formation (the so-called Salgir unit) contain Early Jurassic fossils whereas higher levels yield a mixture of Middle Jurassic, Late Triassic, Permian and Carboniferous fossils. However, the most interesting finds were made by Dekhtyareva et al. (1978) who describe from the Eski-Odra formation near Simferopol, apart from Triassic faunas, the occurrence of Ammonites straddling the Hauterivian-Barremian and the Middle-Late Aptian boundaries.

From the above it is concluded, that all Carboniferous to at least Jurassic fossils, which were retrieved from limestone or sandstone blocks as well as from sandstones and shales of the Taurian group, are reworked and therefore cannot be interpreted as giving the depositional age of the Taurian flysch. In this respect, even such delicate pelagic shells as *Pseudomonotis caucasica Witt* were reworked and are now found in a secondary position. For example, V. Bodylevsky described intensly deformed black shales containing blocks of calcareous sandstone and quartzite with purplish-red nodular siderite from which he retrieved, apart from *Pseudomionotis caucasica*, also *Halobia* and *Proarcestes* (Late Triassic). In some cases broken pieces of *Pseudomonotis* are found in friable shale intercalations, indicating, according to V. Bodylesky, that they were reworked in submarine flows or by wave action. More rarely *Pseudomonotis* occurs in nodular siderite where mainly *Halobia* is present (see Menner et al., 1947, pp. 65). In our opinion this does, however, not necessarely mean that these shells were reworked, as first assumed by V. Bodylevsky.

On the other hand, it is significant, that Late Jurassic carbonate fragments have never been reported from the Taurian flysch. Morover, it is interesting to note that the Hauterivian-Aptian Ammonites were also found in carbonate fragments, suggesting that they were reworked from the Yayla allochthon as carbonates of this age are not known from the Crimean autochthon.

At present no reliable in-situ fossils have been identified which could date the Taurian group. However, considering the youngest Ammonites, the Taurian flysch is provisionally assigned a Hauterivian to Aptian age.

In view of its intense deformation, the depositional thickness of the Taurian flysch cannot be determined. Moreover, no data are available on the direction of clastic transport. On the other hand, the occurrence of coarse mass flow deposits and of olistoliths and olistostroms within the Taurian flysch, composed of Carboniferous to Early Jurassic carbonates, indicates that it was deposited under tectonically increasingly unstable conditions in a deeper water basin of as yet unknown origin.

Yayla Unit

In outcrops, the Yayla unit overlays variably the Taurian flysch or different parts of the autochthonous unit. The Yayla unit is composed of Late Jurassic, partly reefal carbonates, which range in age from Kimmeridgian to Tithonian, and attain maximum thicknesses of 800-1000 m. Locally a basal conglomeratic unit is evident which usually is assigned to the Oxfordian-Early Kimmeridgian. In a few places the Yayla unit appears to include Neocomian carbonate flysch. The Yayla carbonates, which uphold the relief of the Crimean Highlands, are deformed to various degrees, though distinctly less intense that the Taurian flysch.

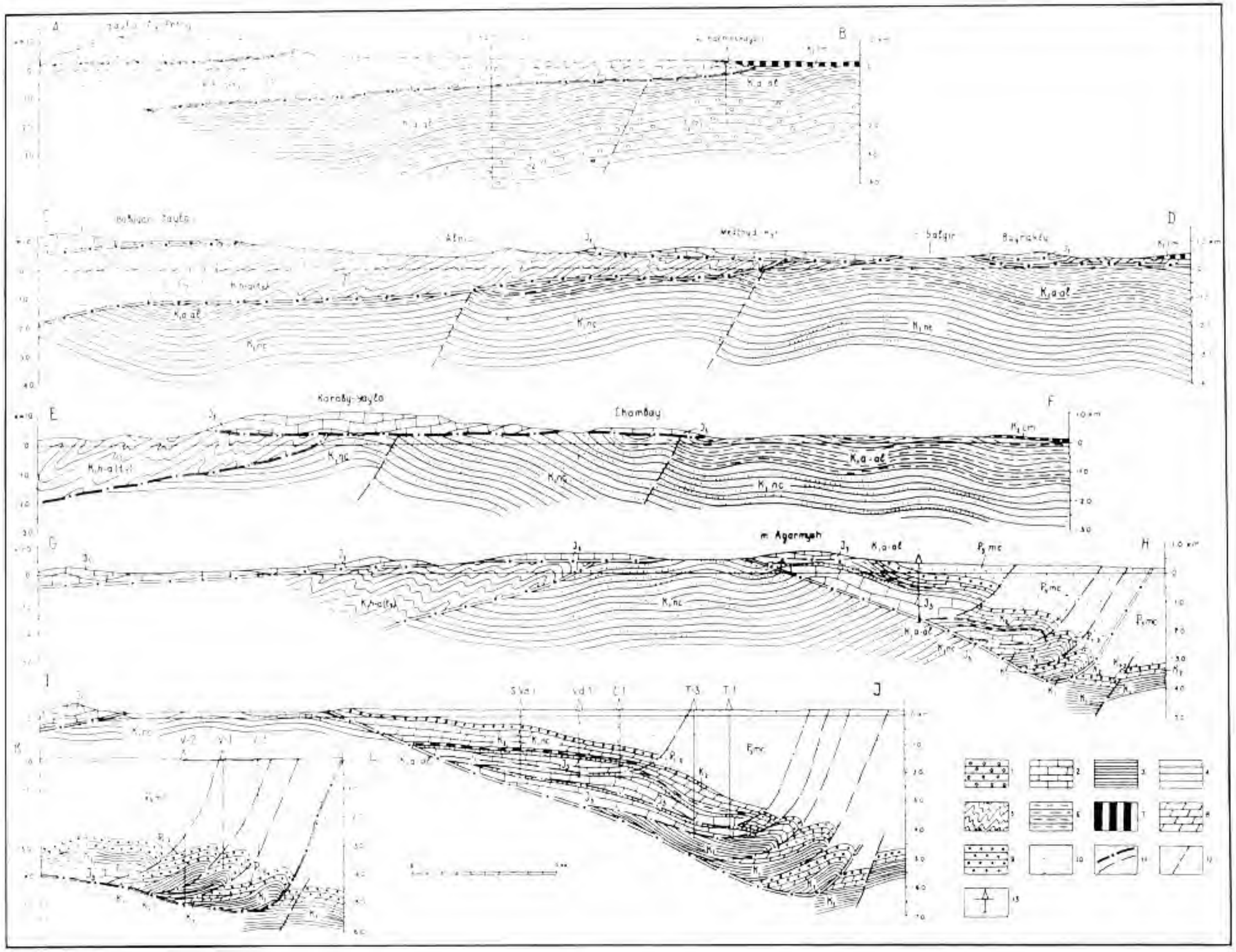
Late Cretaceous and Cenozoic Neo-autochthonous Unit

Along the northern margin of the Crimean Highlands, Late Cretaceous and Cenozoic strata unconformably overlay the mildly deformed autochthonous unit and the more intensely deformed Taurian and Yayla units. In the northeastern part of the Crimea, Cretaceous and Paleogne strata, together with Late Jurassic carbonates, are involved in the so-called Vladislavovka nappe, a structure that obviously records Late Alpine reactivation of the Yayla nappe (Fig. 2).

REGIONAL STRUCTURAL CROSS–SECTIONS

In Fig. 3 we present a set of regional structural cross-sections through the Crimean Highlands. These transects are based on detailed surface geological mapping and on an integration all available subsurface data. In the following results of our studies along the different transects will be discussed.

FIG. 3. Regional structural cross-sections through Crimean Orogen (for location see Fig. 2)
1: Middle Jurassic shales and silts containing rare coal intercalations; 2: Late Jurassic limestones and conglomerates of Yayla nappe; 3: Early Cretaceous shales, sandstones (Vladislavovka nappe); 4: Neocomian flysch-type shales and limestones (autochthonous unit); 5: Hauterivian to Aptian Taurian flysch series; 6: Albian-Aptian shales (autochthonous unit); 7: Cenomanian neo-autochthonous clastics and carbonates; 8: Late Cretaceous limestones, marls, shales and silts; 9: Paleocene shales, silts, marls and limestones; 10: Oligocene (Maïkop group); 11: thrust faults; 12: normal faults; 13: wells



Belbek Valley

The cross-section through the Belbek Valley area is built on surface geological data and the results of wells Kachinskaya-1 and -2 (Fig. 3, profile A-B). Kachinskaya-1 was spudded in gently north dipping Cenomanian limestones resting unconformably on the tightly folded Taurian flysch. Upon penetrating the base of the latter, the well entered into Aptian-Albian shales forming part of the autochthonous sequence. After drilling some 750 m of Early Cretaceous shales, a continental, coal-bearing Middle Jurassic sequence was encountered. The well was terminated before reaching the base of this Middle Jurassic series. Kachinskaya-2 was spudded in Taurian flysch and entered Albo-Aptian clays at a depth of 920 m; after penetrating 1854 m of little deformed Early Cretaceous and 1169 m of Middle Jurassic sediments, the well was abandoned.

Based on surface geological data and extrapolation of the well information it appears that the thickness of Taurian flysch increases southwards to about 1.5 km. The Yayla-Ay-Petry mountain is capped by some 1000 m thick Late Jurassic carbonates which rest directly on the Taurian flysch. To the southwest of the Belbek Valley, the Yayla carbonates cover the entire Taurian flysch and are in turn overstepped by Cenomanian deposits (Fig. 2).

The base of the Yayla and as well as of Taurian allochthon correspond to gently southward dipping thrust planes which are sealed by Cenomanian sediments along the northern margin of the Crimean Highlands. This indicates that these allochthons were emplaced during the Late Albian on a little deformed autochthonous sequence, consisting of Aptian-Albian shales which rest disconformably on a thick Middle Jurassic sequence.

Salgir Valley

In the topographically lowest part of the Salgir Valley, dark-grey to black, plastic and weakly deformed Albian clays outcrop in several tributaries of the Salgir river (Fig. 3, profile C-D). On the eastern slopes of this valley, massive Late Jurassic limestones, forming the Yayla plateau, rest directly on Albian clays. The contact between the Jurassic carbonates and the Albian clays is concealed by scree slopes and giant blocks which have tumbled down from the edge of Yayla plateau. However, as this contact is clearly marked by a system of springs, it can be readily mapped. The contact between the Albian clays and the Late Jurassic limestones is interpreted as a sub-horizontal thrust fault.

Along the southern and western slopes of the Salgir Valley, the Early Cretaceous sequence extends stratigraphically downwards into the Aptian and consists of dark-grey clays containing silty and fine-grained sandstone intercalations. In one of the tributaries of the Salgir river the Aptian sequence is directly overlain by the Taurian flysch; the latter consists of strongly deformed, rhythmically bedded shales, silts and quartzitic sandstones. However, the contact between the Aptian clays and the Taurian flysch is poorly exposed. Similarly, the contact between the Taurian flysch and the overlaying Jurassic carbonates is essentially concealed by scree. In this area the Jurassic sequence consists of massively bedded Kimmeridgian conglomerates with a carbonate matrix; these grade upwards into the massif Tithonian carbonates. The components of the Kimmeridgian conglomerate are made up of quartz, sandstones, metamorphic shales and limestones.

Topographically speaking, the Late Jurassic conglomerates and limestones form the uppermost unit. Further south, this unit rests on highly deformed Taurian flysch and further to the north directly on weakly deformed Aptian and Albian clays. Where present, the Taurian flysch overlays Early Cretaceous clays and thus forms the middle unit. The Albian-Aptian clays form part of the basal autochthonous unit. Both the Taurian flysch and the Late Jurassic Yayla unit are allochthonous and were apparently thrusted northwards over the autochthonous Aptian-Albian clays. Across the Salgir Valley, erosion has cut through the Yayla and Taurian thrust sheets into the underlying autochthonous unit; as such the Salgir Valley represents a tectonic half-window. Surface geological evidence indicates the the autochthonous series extend a minimum of 20 km under the combined Yayla and Taurian thrust sheets. The Bayrakly and

Medshyd-Kyr hills are upheld by klippen of the Yayla thrust sheet which are partly underlain by the Taurian thrust sheet. Although the southern part of the cross-section C-D is based on limited subsurface data and is therefore schematic, it suggests that the Yayla thrust-sheet was transported northwards over a distance of at least 35 km along a sub-horizontal thrust-fault, separating it from the Taurian thrust-sheet, which in turn traveled a similar distance over the autochthonous foreland. Both the Yayla and the Taurian thrust-sheets can be regarded as major north-verging nappes.

Tonas Valley

The Tonas Valley is located in the head-waters of the Biyuk-Karasu river (Fig. 1). South of the city of Belogorsk, weakly deformed Aptian-Albian clays occupy the valley floor (Fig. 3, profile E-F). 5.5 km south of Belogorsk, outcrops of thinly bedded flyschoid dark-gray shales, silts and sandstones of Barremian age are observed. Further upstream, progressively older strata are exposed; a more shaly section is followed by monoclinally north dipping rhythmically bedded calcareous shales, siltstones and limestones of Valanginian-Hauterivian age and Berriasian calacreous shales and limestones. The lower parts of this sequence are characterized by 10-15 m thick cycles, each of which commences with a limestone conglomerate which grades upwards into limestones and calcareous shale. The basal parts of the Cretaceous are characterized by 1-2 m thick limestone beds. Along the road leading up to the Alikot-Bogas pass, more strongly deformed calcareous shaly flysch, containing sand intercalations, is observed; these sediments are interpreted as equivalents of the Barremian.strata seen south of Belogorsk.

In the Tonas Valley a 3.5-4.5 km thick sequence of Albian to Berriasian flyschoid autochthonous series can be observed. However, its base has not been seen and it is unknown what underlies it.

Four kilometres to the south of Belogorsk, the hills flanking the Tonas Valley are capped by massively bedded Late Jurassic limestones and alternating limestones and shales of Tithonian-Berriasian age. Since these carbonates rest on Albian shales, they are attributed to the Yayla nappe. Apparently this nappe rests in the Tonas Valley directly on various levels of the autochthonous sequence. The Taurian nappe is only recognized under the southern parts of the Karaby-Yayla hills from where it continues to the Black Sea coast. The Karaby-Yayla hills form part of a major Yayla nappe klippe.(Fig. 2). Similar to the Salgir Valley, the Tonas Valley is interpreted as a tectonic half window which developed in response to erosion of the allochthonous units. Surface geological criteria indicate for the Tonas Valley area a minimum of 15 km northward transport of the Yayla nappe over the autochthonous foreland.

Sukhov-Indol Valley

On the eastern slopes of the Idol Valley, Mount Agarmysh represents the northernmost remnant of the Yayla nappe (Fig. 3, profile G-H; Fig. 4). Late Jurassic limestones, which uphold this mountain, rest directly and in thrust-contact on Albian-Aptian autochthonous clays. As nowhere else, it is here possible to closely observe the internal structure of the Yayla nappe. Mount Agarmysh provides in a cliff-face of over 400 m and in a quarry on its southern flank excellent exposures. These show that the upper parts of Mount Agarmysh are composed of gently folded, massif Tithonian limestones (Fig. 4b). Along the southern flank of Mount Agarmysh, fine-grained Neocomian siliciclastic flysch and conglomeratic limestones, intercalated with calcareous shales and reminiscent of the Berriasian series of Tonas Valley, are exposed. In the Agarmysh quarry, an olistostrom is recognized which is composed of large Late Jurassic limestone and Neocomian siliciclastic flysch blocks; these are encased in a shaly matrix of unknown age (Fig. 4a). The thrust contact between the Tithonian to Neocomian sediments of the Yayla nappe and the underlaying Apto-Albian shales is sub-horizontal and appears to be paralleled by a subsidiary thrust fault located some 110 m above the sole-thrust of the Yayla nappe.

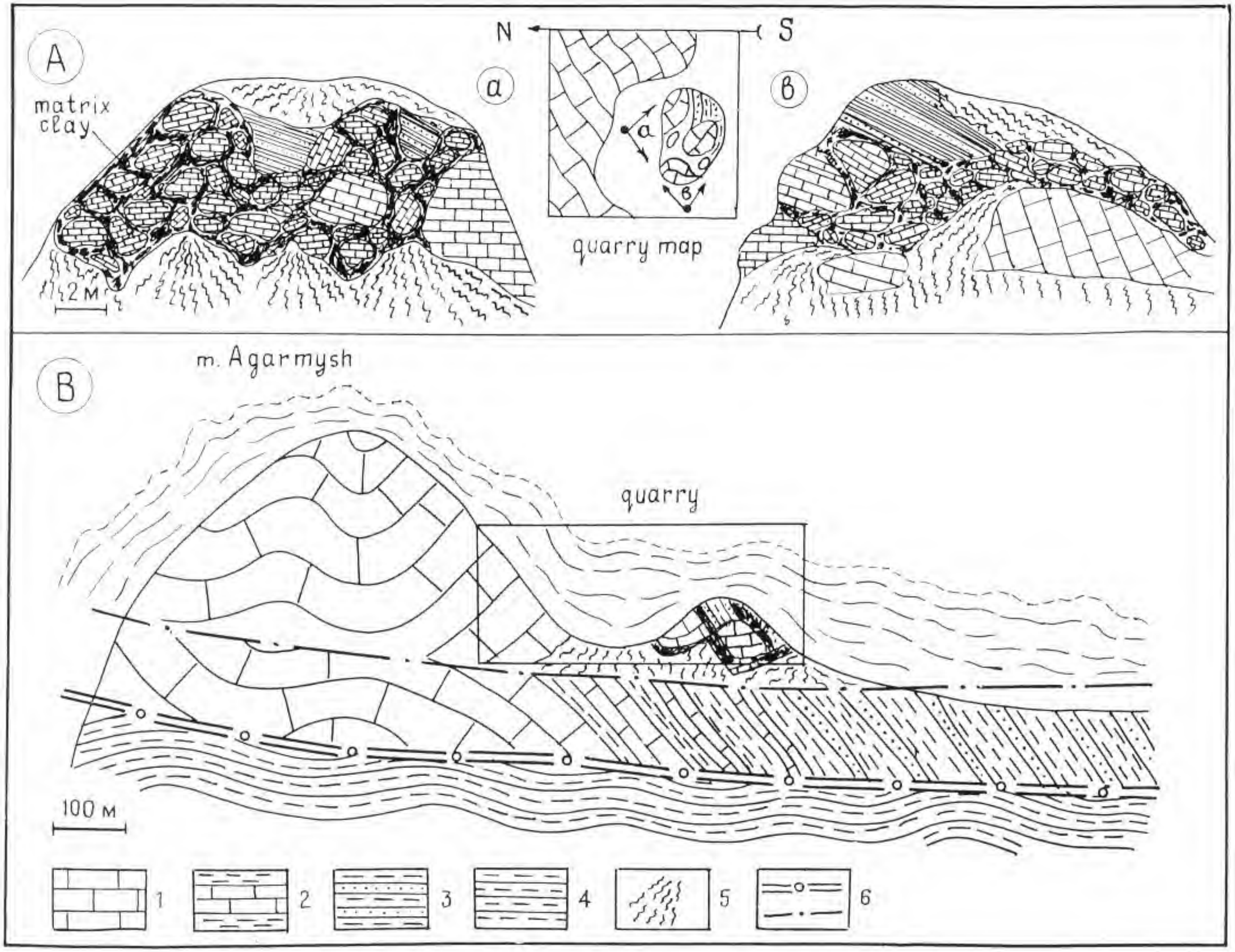


FIG. 4. Geological cross-section through Mount Agarmysh
A: north and west wall of Agarmysh quarry; B: Agarmysh Mountain showing location of quarry; 1: Massif Tithonian limestone; 2: Tithonian-Berriasian flysch-type shales and limestones; 3: Neocomian sandy flysch; 4: Aptian-Albian dark-gray, non-calcareous shales; 5: talus; 6: sole-thrust of Yayla nappe and intra-Yayla thrust fault

The northern part of section G-H is based on sub-surface data which indicated that the sole-thrust of the Yayla nappe plunges northwards and that it was apparently active during Cenozoic times as evident be the imbrication of Late Cretaceous and Paleogene sediments. A similar pattern is shown in profiles I-J and K-L which cross the Vladislavovka nappe and are closely controlled by boreholes. Several wells drilled in this area penetrated the basal sole-thrust and bottomed below Late Jurassic carbonates in little deformed Neocomian sediments (Voloshyna, 1977).

The southern parts of profile G-H illustrate that the Taurian nappe appears beneath the Yayla nappe to the south of an erosional window in which Neocomian flyschoid series of the autochthon are exposed. On the basis of this profile, a horizontal northward transport of the Yayla nappe, amounting to some 35 km, can be postulated.

DISCUSSION

Based on the results of our research, we propose that the configuration of the Crimean orogen is dominated by the Yayla nappe, which involves mainly Late Jurassic carbonates, and by the Taurian nappe, which is composed of Early Cretaceous flysch. Palaeontological, surface geological and sub-surface data clearly demonstrate that emplacement of both nappes resulted in the superposition of older over younger strata.

The Yayla and the subjacent Taurian nappe were thrusted northward over an autochthonous series which includes Middle Albian.and older strata. At their northern margin, both the Yayla and the Taurian nappes are overstepped by undeformed Cenomanian and younger sediments which also seal their basal thrust-faults. Within the basal parts of these post-tectonic transgressive sediments, reworked Hauterivian, Barremian, Aptian and Albian faunas were recognized, together with limestone conglomerates and fragments of Taurian flysch; these basal beds grade upwards into undisputable Cenomanian limestones as seen, for instance, in the Alma and Bodrak valleys. At some stage the question of the lower age limit of the neo-autochthonous series gave rise to considerable controversy. However, in view of a definitely Mid-Albian upper age limit of the autochthonous series, we conclude that the final emplacement of the Yayla and Taurian nappes occurred during the Late Albian.

Therefore, the main deformation phase of the Crimean Highlands can be safely dated as Late Albian. As such it coincides with the Austrian phase of the Alpine orogenic cycle and not, as previously postulated, with the Early Cimmerian phase (Muratov, 1960, 1969; Muratov and Tseisler, 1982). Similarly, a Mid-Cretaceous compressional phase governed the development of the Dobrogean orogen for which a previously postulated Early-Cimmerian compressional deformation must be rejected on the basis of newer litho- and biostratigraphic data (Gradinaru, 1984).

For the northeastern part of the Crimean Highlands, subsurface data clearly demonstrate an Eocene and younger reactivation of the Yayla nappe; this reactivated zone is referred to as the Vladislavovka nappe. The extent to which the remainder of the Crimean Highlands were affected by these Cenozoic deformations is uncertain. However, the erosional edge of the Cenomanian and younger post-Austrian neo-autochthonous series suggest at best a regional upwarping of the entire area and minor faulting. Such upwarping of the entire Crimean Highland probably entailed a modification of the geometry of the Yayla and Taurian nappe sole-thrusts.

The Yayla nappe, which at present consists of a system of klippes, is presumably rooted off-shore in the Black Sea. Remnants of this major nappe, which attains thickness of some 800 to 1000 m, indicate that is was transported in a northwesterly direction over a distance of at least 40 km; its lateral extent is about 140 km. Although the internal structuration of the Yayla nappe is very complex and is far from being resolved, its sole-thrust is sub-horizontal and plunges only northwards in the domain of the Vladislavovka nappe.

In a strike direction the Taurian nappe extends over a distance of 140 to 150 km. In a dip direction its distribution is very variable; in the Yalta-Bakhcisaray sector it has a length of at least 40 km whereas in the Alushta-Sudak sector and to the southwest of Yalta it does not exceed 10-15 km (Figs. 2 and 3). The thickness of this complexly deformed nappe is difficult to estimate. Although a minimum thickness of 1 km is ascertained by well data, it is likely that the thickness of this nappe increases to 2 or more km to the south in non-eroded areas.

CONCLUSIONS

The Crimean Orogen consists of the Taurian and Yayla nappes which were thrusted northwards across the autochthon over a distance of at least 40 km during the Mid-Cretaceous Austrian phase of the Alpine orogeny. There is no evidence for an Early-Cimmerian compressional deformation of the Crimean Highlands. Paleogene reactivation of the Crimean orogen appears to be restricted to its northeastern parts, though contemporaneous broad arching of the entire Crimean Highlands cannot be excluded.

Although our understanding of the architecture and the timing of deformation of the Crimean orogen has greatly advanced, much remains to be done in order to understand the evolution of the basin out of which this thrust belt evolved. Some of the outstanding questions are: Did emplacement of the Yayla and Taurian nappes involve in-sequence or out-of-sequence thrust propagation? From where did the Yayla carbonate platform originate and what processes governed the subsidence an closure of the Taurian flysch trough? What is the derivation of the Carboniferous, Permian, Triassic and Early Jurassic olistoliths and olistostroms contained in the Taurian flysch? What is the geotectonic setting of volcanic activity associated with the Taurian flysch? Can indeed a relationship be established between the evolution of the Dobrogea and the Crimean basin?

At this stage we are still far away from attempting a palinspastic and historical restoration of the Crimean Orogen, which in all probability would contribute materially to the understanding of the evolution of Tethys and its flanking platforms.

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3D geometry and kinematics of the N. V. Turkse Shell thrustbelt oil fields, Southeast Turkey

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ABSTRACT

In 1984 NV Turkse Shell (NVTS) acquired the first 3D survey in Turkey, over the Beykan field. Between 1989 and 1992 an additional seven 3D surveys have been acquired. The 3D seismic data has enabled substantial advances in the understanding of detailed subsurface structure, deformational history and the relationships between oilfields. As a consequence, 3D results have led to fundamental revisions of structural maps in both heavily drilled (Beykan) and lightly explored (foreland, Kayakoy West Deep) areas. This work has provided a detailed picture of the lateral extent, internal geometry and hydrocarbon distribution within the major structures of the NVTS Lease Areas. The relationship between stacked imbricate structures, and the geometry of the underlying foreland setting has also been mapped in detail.

The classical view of the genesis of the SE Turkey Foothills Belt is that the terranes presently incorporated in the imbricated zone formed a positive (forebulge) area at the Arabian Platform edge in the Cretaceous. During the Late Cretaceous, coeval with Upper Mardin and Kastel Formation deposition, the platform edge was significantly deformed ultimately leading to imbrication. The resulting structures are typical examples of imbricates formed in a foreland setting. Thrust sheets are relatively thin, comprising a layered sedimentary sequence, and have a length which is several times greater than their thickness. The basal decollement plane formed in argillaceous units within the Silurian Dadas Formation. In the west, thrust imbricates are relatively thin and gentle folding accompanied thrusting. In the east the imbricates are much thicker and larger amplitude folding led to steeper dips in the deformed strata. In the area occupied by the Kurkan, Kayakoy and Kayakoy West structures, deformation resulted in a stack of smaller imbricates. These differences are caused by 1) the presence, in the east, of a clastic sequence overlying the argillaceous units of the Dadas Formation and 2) the resulting difference in frictional behaviour along the basal slip plane.

GILMOUR, N. & MÄKEL, G., 1996. — 3D geometry and kinematics of the N. V. Turkse Shell thrustbelt oil fields, Southeast Turkey. *In*: ZIEGLER, P. A. & HORVÀTH, F. (eds), Peri-Tethys Memoir 2: Structure and Prospects of Alpine Basins and Forelands. *Mém. Mus. natn. Hist. nat.*, **170**: 524-547. Paris ISBN: 2-85653-507-0.

INTRODUCTION

Since the discovery of relatively small and complex oil fields in the Foothills Belt of the Taurus Mountains in SE Turkey (Fig. 1), this area has been the subject of extensive geological and geophysical surveys. Using a variety of tools, a considerable number of additional structures have been identified and successfully drilled. A complete understanding of the relationship between individual structures has, however, remained elusive. The acquisition of 3D seismic data within the NV Turkse Shell Lease Areas has enabled substantial advances in the understanding of detailed subsurface structure, deformational history and relationships between fields. The results have had a significant impact upon the interpretation of the hydrocarbon habitat in the Foothills Belt, from migration path to fluid contacts, and will provide a basic framework for future reservoir characterisation work.

This paper concentrates upon the mechanics and kinematics of deformation and resultant subsurface geometry together with the implications for exploration and development activity in this densely drilled but still only partially understood area.

NVTS THRUSTBELT EXPLORATION AND 3D SEISMIC HISTORY

NV Turkse Shell (NVTS) has been actively exploring in Turkey since 1953. In 1960 the Kayakoy-2 well discovered the first commercial oilfield in the Foothills Belt (Fig. 2). Subsequently 300 wells have been drilled by NVTS in this structural setting resulting in the discovery of 26 fields. Exploration and development has employed a variety of tools over this period, including gravity surveying, 2D reflection seismic, refraction seismic, 2D swath acquisition, borehole gravimetry and recently 3D seismic. Progress in improving subsurface structural definition has been irregular, largely in step with incremental advances in geophysical technology.

The delineation of the major structures within the NVTS Lease Areas was largely established by combining drilling results with gravity and seismic data and surface geology studies. Some 10 years after the first discovery, the major structural elements (Beykan, Kurkan; Fig. 2) had been located, the variation in structural style across the Lease Areas described in general terms, the decollement level identified and the timing of the main phases of deformation established.

The various historical exploration methods employed, while adequate to map out the overall shape of most of the producing structures (Fig. 3), largely failed to delineate both the steeper southern flanks and the internal complexities of the imbricate sheets. Considerable uncertainty in the position of the southern boundary (thrust) fault in several fields hampered the optimal development of the crestal and leading edge areas of a number of accumulations.

In 1984 NVTS acquired the first 3D sesmic survey in Turkey, the 90 km^2 Beykan survey over the western and central parts of the field. Between 1989 and 1992 an additional seven 3D surveys have been acquired, all but two within the Foothills Belt setting (Fig. 4).

This paper summarises the results of initial seismic interpretation work involving four 3D surveys, 14 fields and some 220 wells. Work concentrated primarily upon the Cretaceous Mardin Group resrvoir sequence. Structural interpretation made extensive use of horizon attribute analysis (dip/azimuth) in combination with fault data from well penetrations.

STRUCTURAL SETTING AND REGIONAL STRATIGRAPHY

The Taurus Mountain Belt represents a zone of major Alpine deformation which formed in response to the collision of the Arabian Platform and Eurasia (Fig. 5; Sengör and Yilmaz, 1981; Yilmaz, 1993). The zone can be subdivided: (Fig. 6):

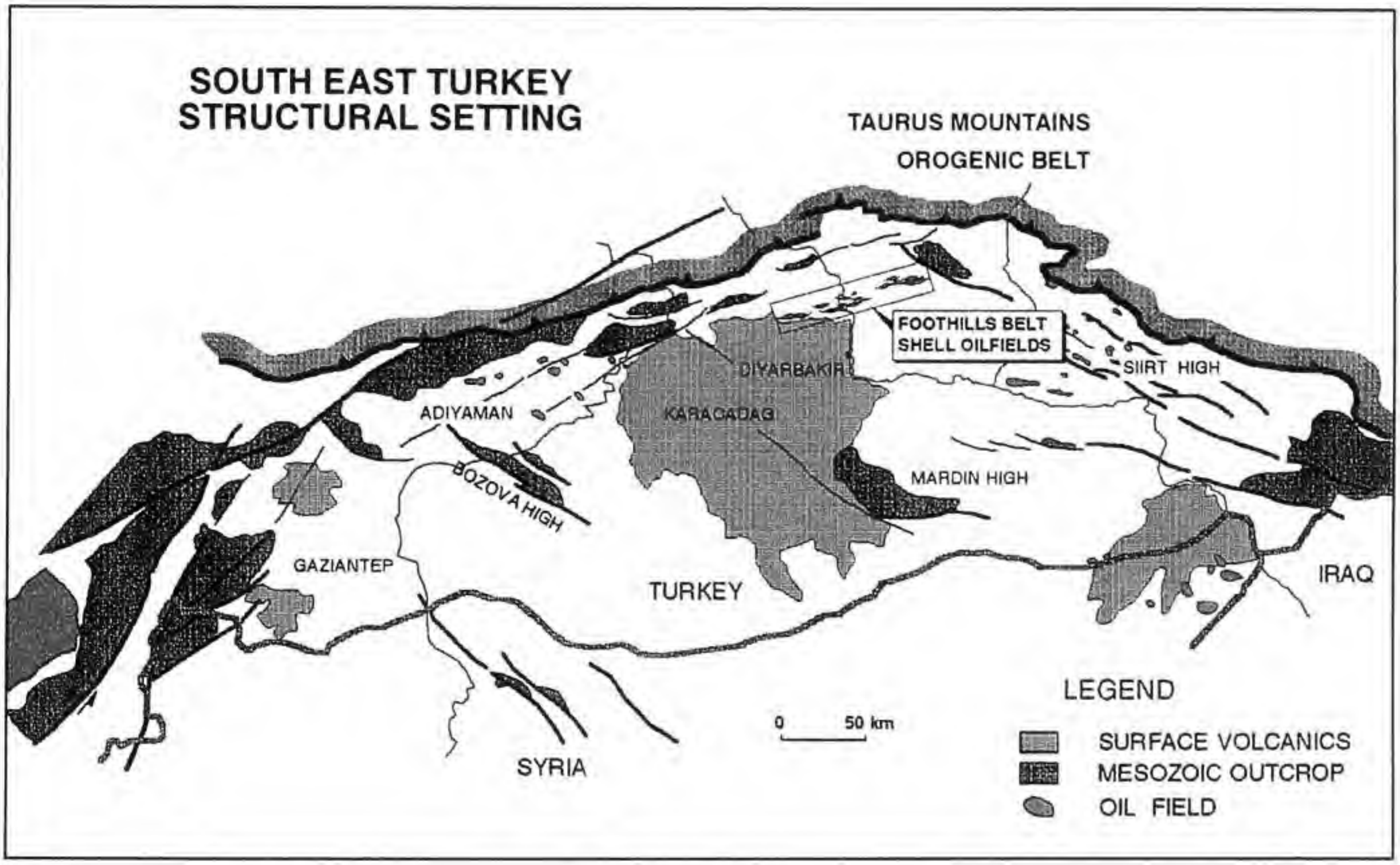

FIG. 1.

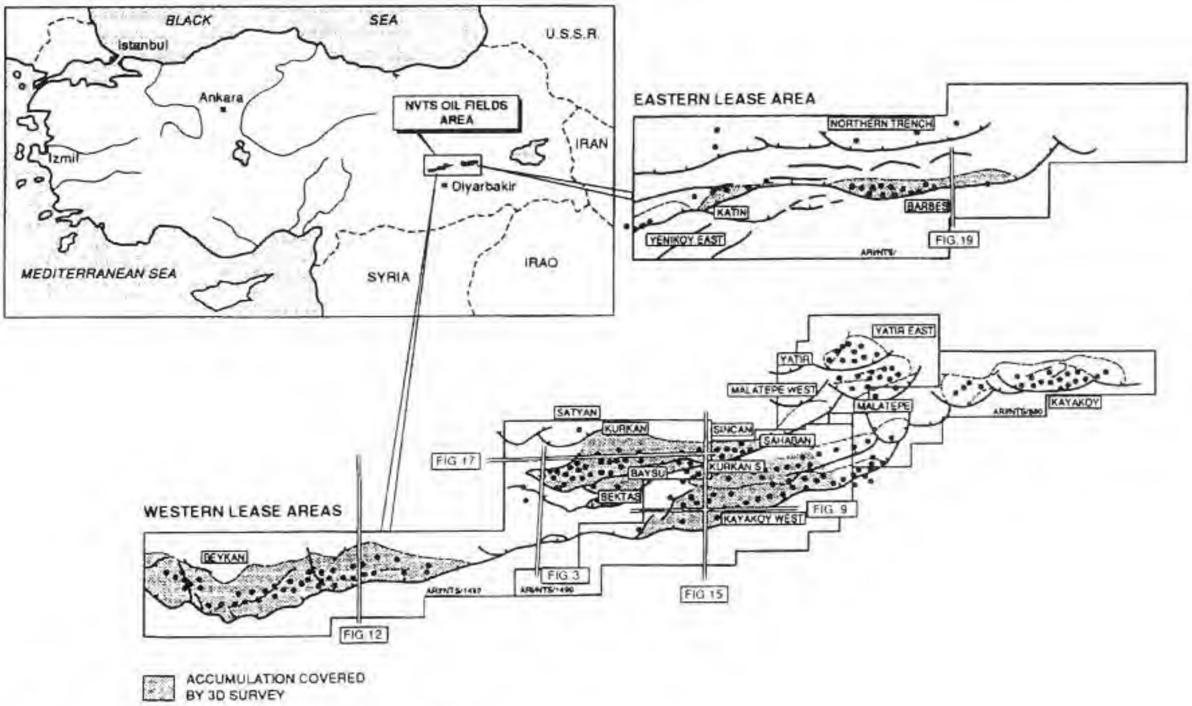

FIG. 2 Location map of NVTS lease areas

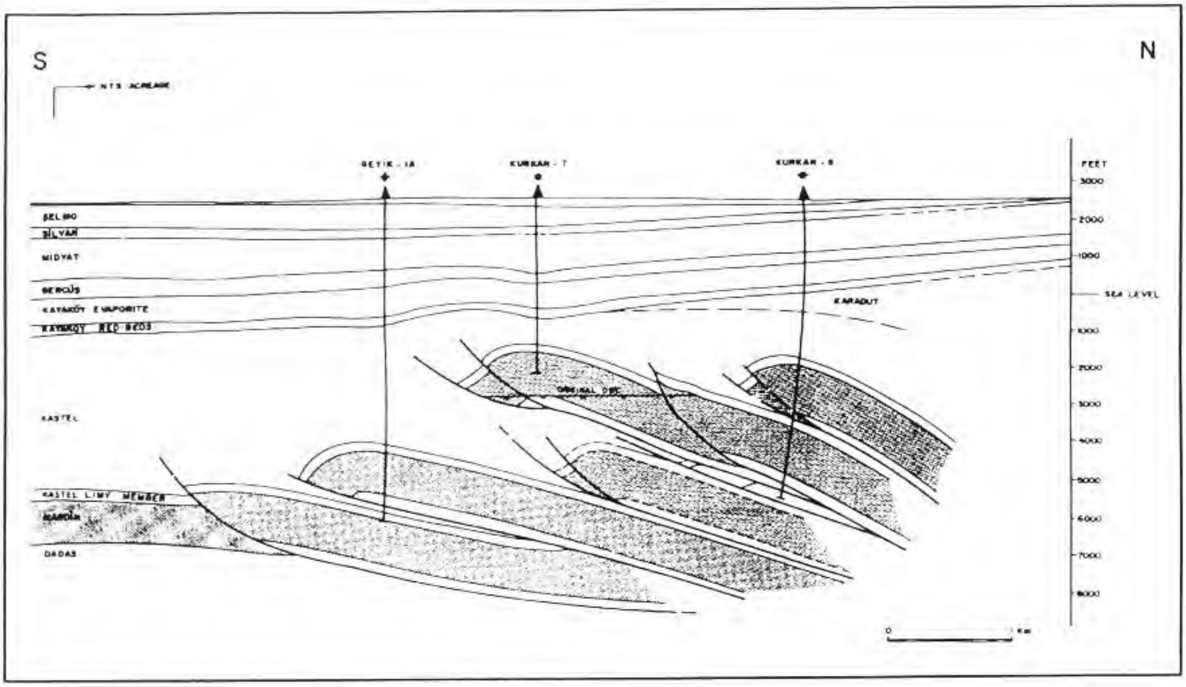

FIG. 3. Geological cross-section, Western Lease Areas

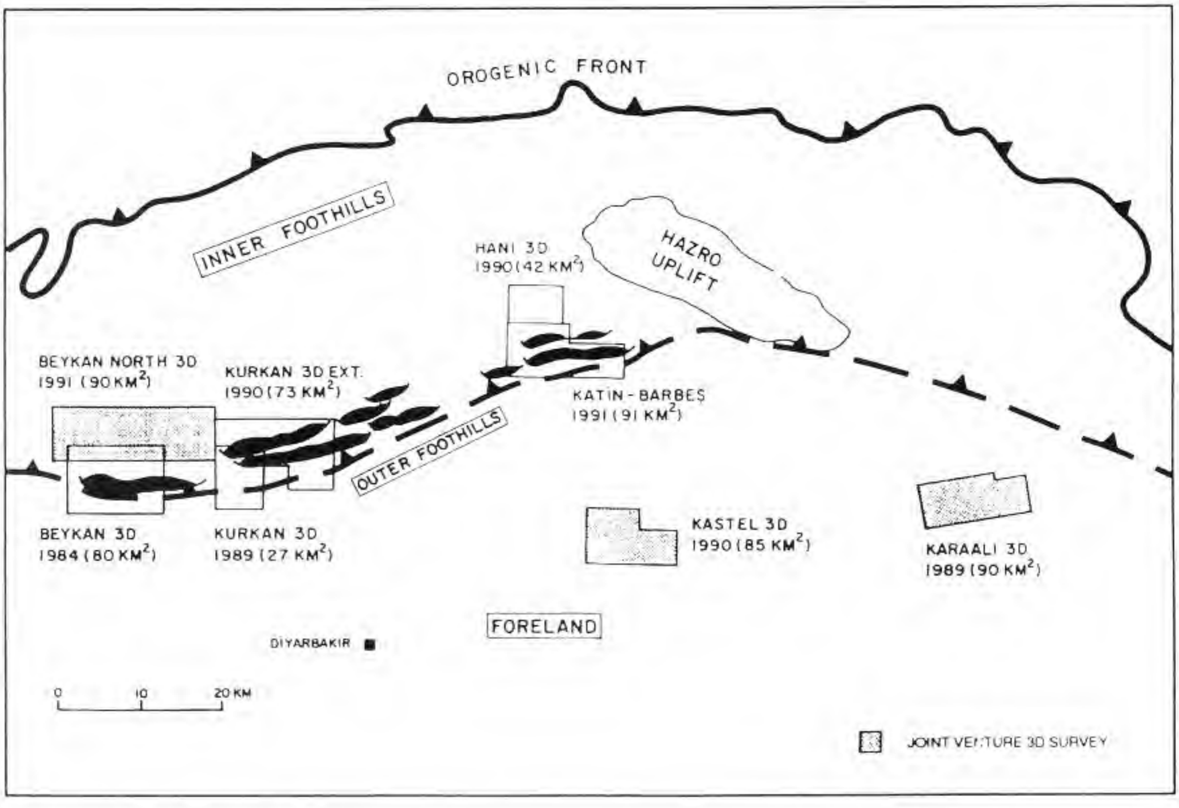

FIG. 4. NV Turkse Shell 3D activity

(1) Orogenic Belt characterised by large scale uplift and nappe tectonics,
(2) Foothills Belt characterised by imbricate thrusting and disharmonic folding,
(3) Foreland characterised by gentle folding and normal and reverse faulting.

Pre-deformation, the areas presently incorporated within the Foothills Belt and the Foreland were situated on the northern margin of the Arabian Platform, where during much of the Cretaceous (Upper Aptian to Lower Campanian), platform carbonates were deposited. The collision of the Arabian Platform with Eurasia, initiated during the Late Cretaceous, led to the formation of the elongated Kastel trough coupled with a forebulge ahead of advancing ophiolitic nappes (Fig. 7a; Horstink, 1971). The foredeep rapidly filled with marls and shales of the Kastel Formation (Fig. 7b) which also contains abundant clasts of both ophiolitic and platform carbonate associations. The latter represent erosional products derived from the forebulge which, under the influence of the advancing nappes, was subsequently rapidly submerged. Consequently, shallow marine Kastel deposits now overlie the upper Mardin erosional surface (Fig. 7c).

Continued compression transmitted by the ophiolitic nappes advancing over the platform sediments, resulted in the imbrication and overthrusting to the south of the foredeep sediments and their basement. Basal thrusts are primarily developed in relatively incompetent Palaeozoic sediments and cut up-section through the more competent Mardin Group rocks. The southern edge of the thrustbelt coincides with the original forebulge area and marks the boundary between the Foothills Belt and the Foreland.

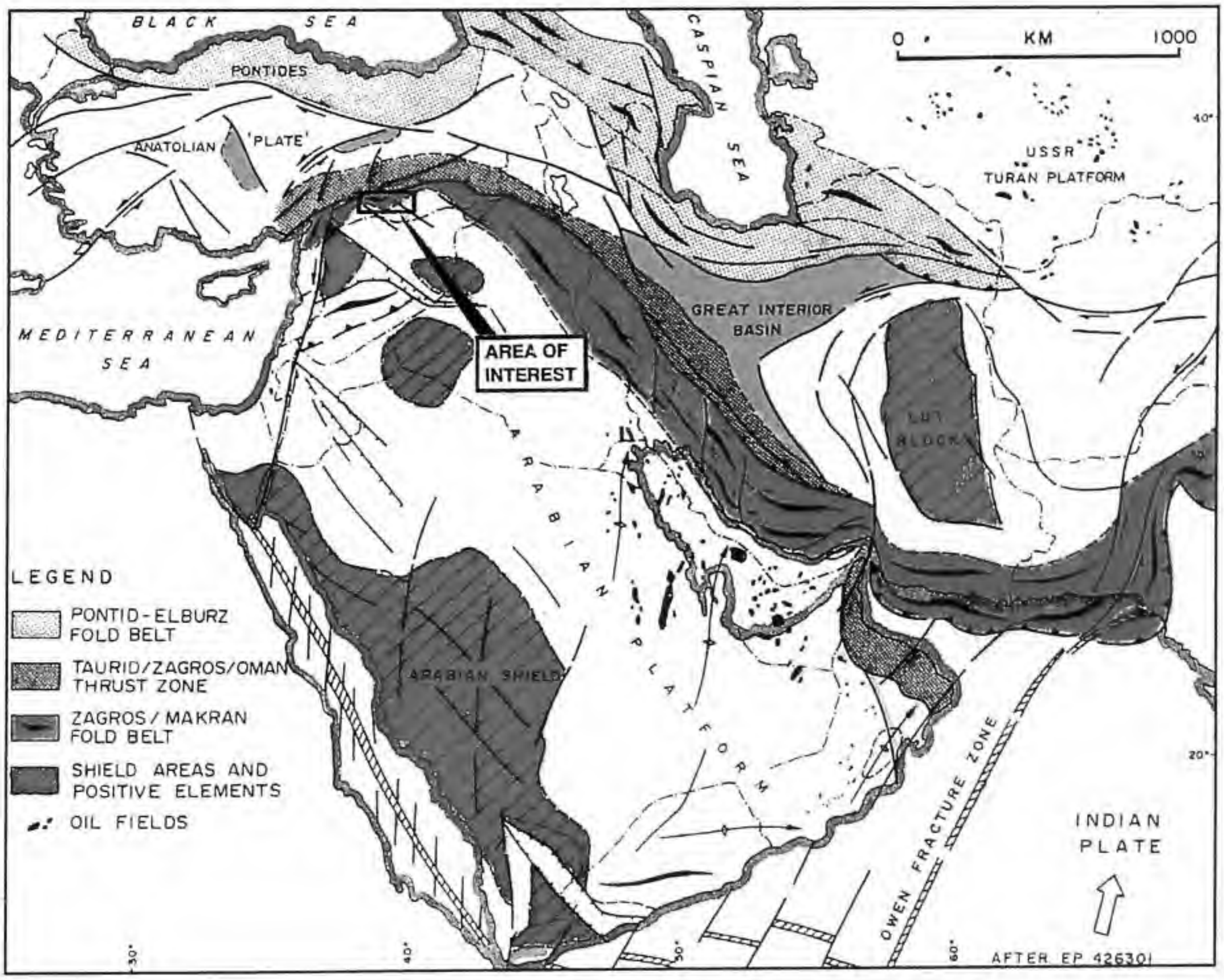

FIG. 5. Middle East mega tectonic setting

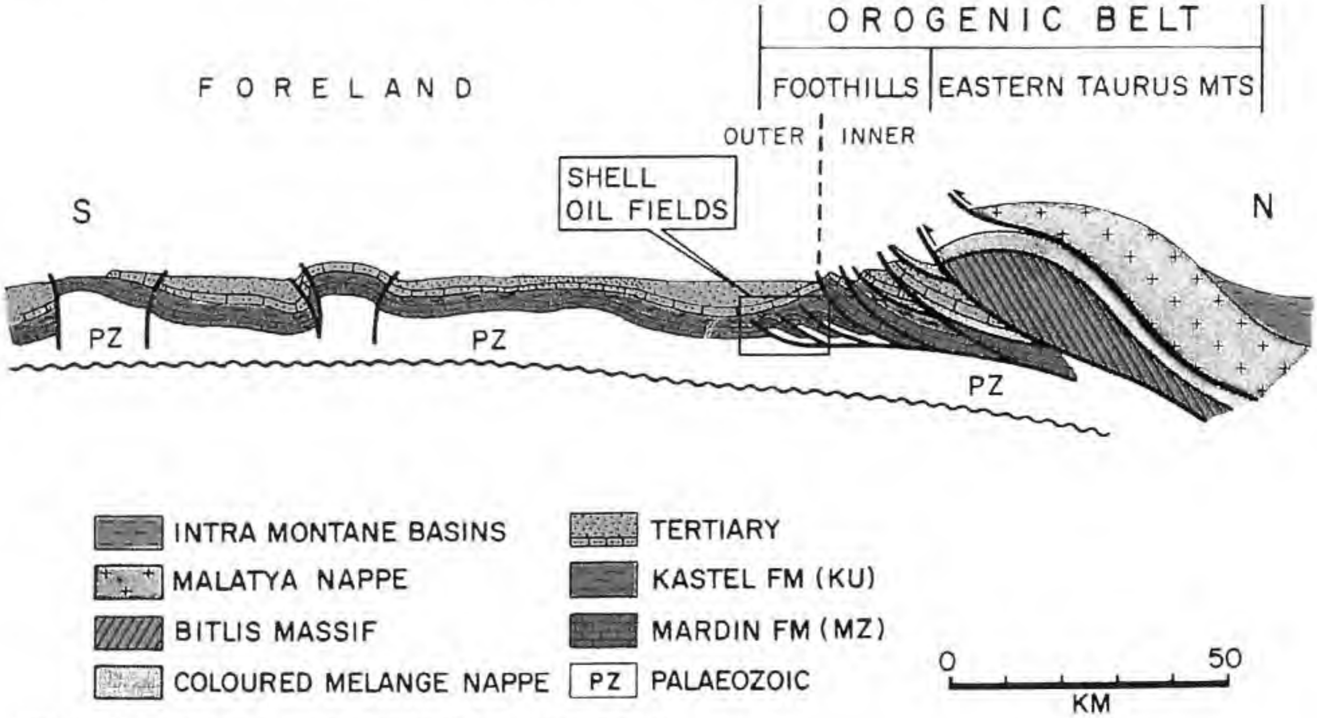

FIG. 6. Schematic cross-section, Southeast Turkey orogenic belt

Following the emplacement of the ophiolitic nappes in the Late Cretaceous, the compressional tectonic setting gave way to a predominantly extensional regime. The Foothills Belt thrust imbricates and the Foreland were buried by shallow marine sediments during the Lower Tertiary (Fig. 7d). Rapid facies variations indicate a series of changes in sealevel, reflecting continuous interaction between the Arabian Platform and Eurasia.

During the Middle Miocene the marine environment was gradually replaced by continental conditions (Fig. 7e). Yilmaz (1993) argues that this was caused by the thrusting of a metamorphic nappe complex over the northern edge of the Arabian Platform following the consumption of oceanic crust. The northern part of the Foothills Belt was the scene of imbricate thrusting (Fig. 7e) deforming the Late Cretaceous structures and their Tertiary overburden. In the southern parts of the Foothills Belt and in the Foreland, the advance of the metamorphic nappe complex led to gentle folding and reverse faulting.

The majority of the SE Turkey oil fields are found in the southern part of the Foothills Belt, within a relatively narrow zone along the Foreland margin. The stratigraphy of the area within which the NVTS oil fields are located may be summarised as follows (Fig. 8):

(1) a Lower Palaeozoic sequence overlying a Cambrian and older basement and containing incompetent Silurian shales. The latter formed the main detachment level for the overlying thrust units and represent the regional source rock interval,

(2) an Upper Palaeozoic to Upper Cretaceous sequence of competent rocks of variable thickness. The lower sequence of Devonian clastics and carbonates is unconformably overlain by Permian and Triassic clastics and carbonates which are in turn progressively eroded out to the west. The Lower Cretaceous Mardin Group succession, including dolomitised shelf carbonates, thins to the north and west. The uppermost members of the Mardin Group are absent in the structures within the NVTS Lease Areas (cf. Cater and Gillchrist, 1994),

(3) an Upper Cretaceous to Palaeocene sequence of shales and marls with intercalated flysch-type deposits (Kastel Karadut

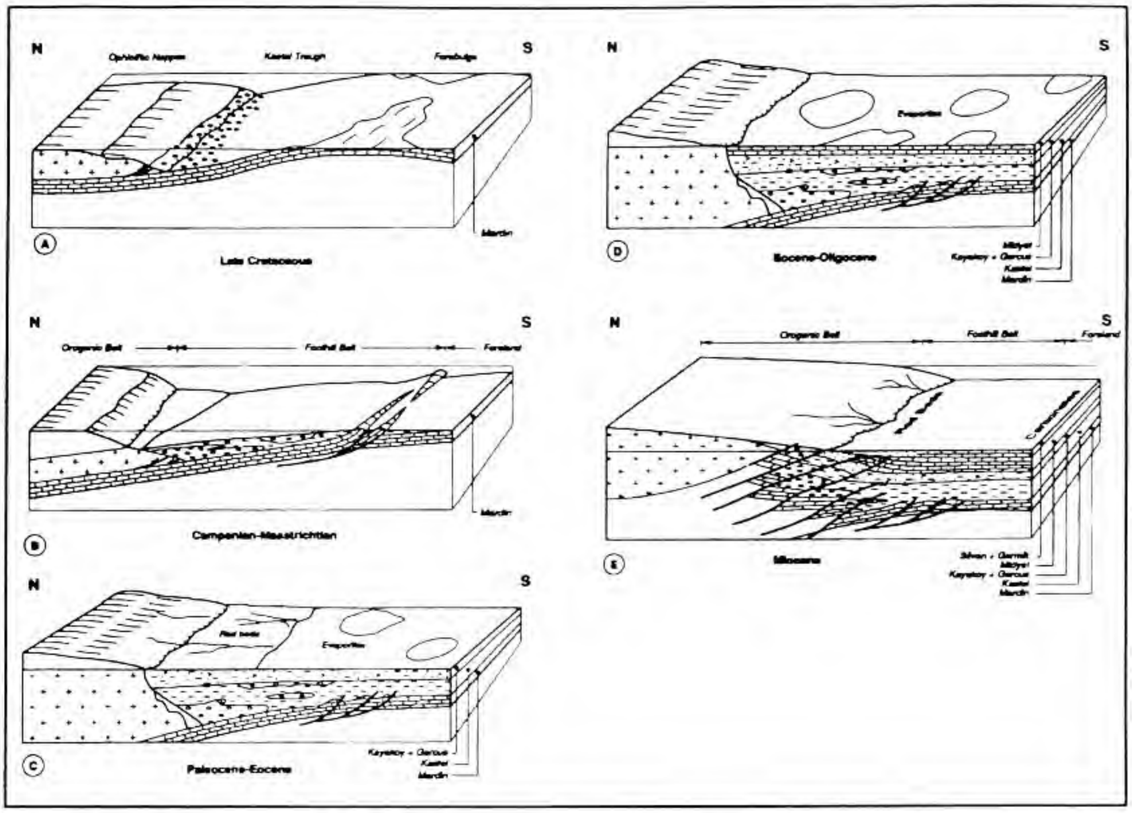

FIG. 7. Sequential development of the Southeast Turkey Foothills Belt

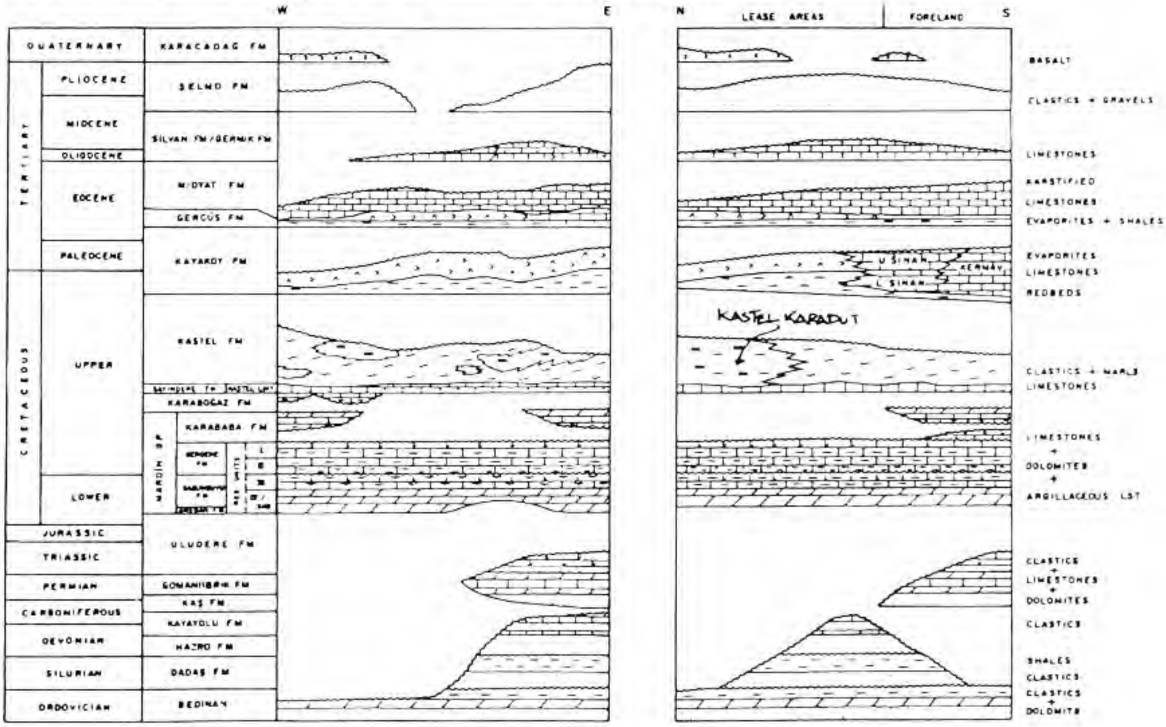

FIG. 8. Generalised stratigraphy of the NVTS Production Licences and adjacent areas

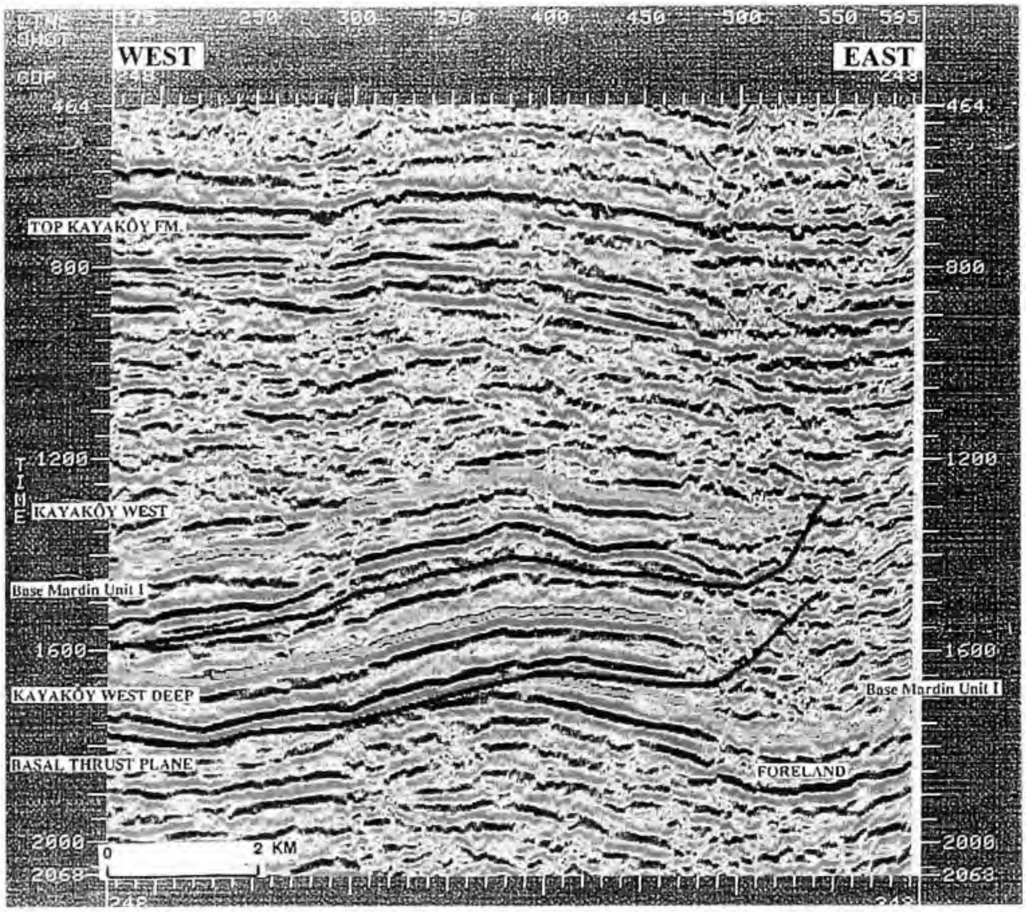

FIG. 9. Interpreted crossline 248

Facies). This thick and incompetent sequence disconformably overlies the Lower Cretaceous and thins rapidly to the west and south,

(4) a thick Eocene to Recent sequence of evaporites and shallow shelf limestones capped by basalts, unconformably overlying the Palaeocene.

DETAILED STRUCTURAL GEOLOGY OF THE NVTS LEASE AREAS

Improved 3D seismic resolution has enabled previously unseen structural detail to be recognised and relationships between the major structural units to be established. New 3D maps of most fields resemble their predecessors in gross geometry but differ fundamentally in internal detail. As discussed below, the new data has led to an overall simplification and reduction in the number of thrust sheets, combined with a radical redefinition of internal block geometries, fault patterns and inter-field relationships.

1 Western Lease Areas

Within the Western Lease Areas (Fig. 2) there are four major thrust sheets of which three have been mapped in detail. From bottom to top these sheets are: Kayakoy West Deep, Beykan -Kayakoy West -Kurkan South and Kurkan -Sincan. The latter is overlain by the Bozalan -Satyan structure, as yet not remapped.

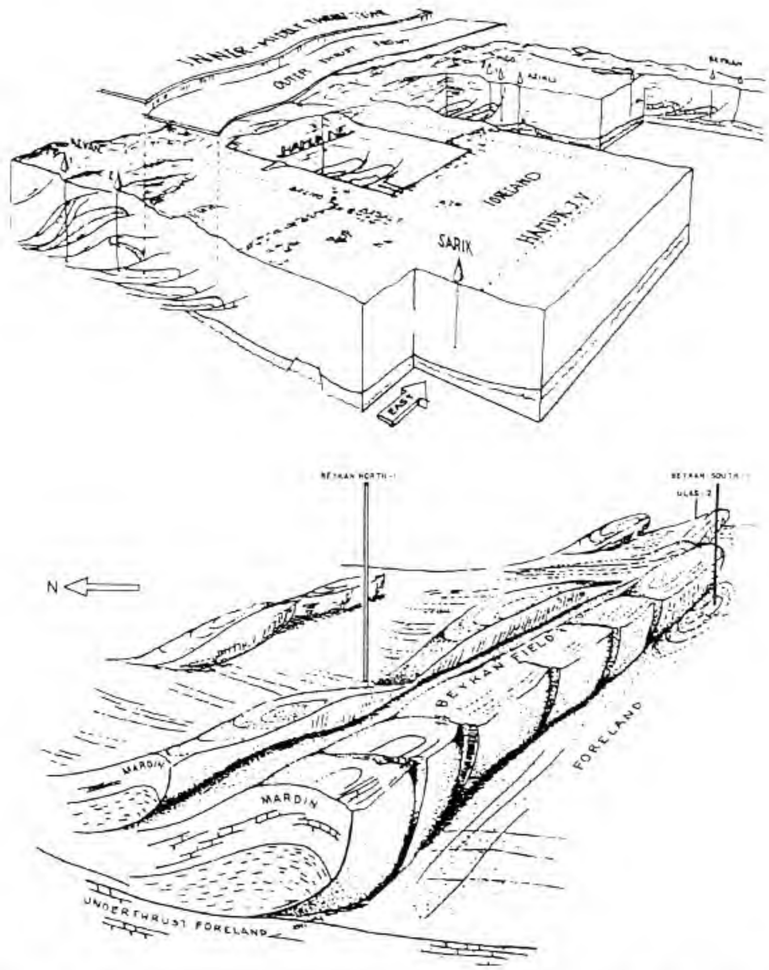

FIG. 10. Historical 3D models of Southeast Turkey, Foothills Belt structures

1.1 Kayakoy West Deep Structure

The Kayakoy West Deep field is the deepest producing structure within the Western Lease Area. Due to the poor 2D quality and the heavily dip-biased nature of the 2D grid, the structure remained poorly understood.

Following 3D interpretation the structure is now interpreted as an elongated E-W trending structure. In its central and eastern portions the structure represents a typical thrust-bound anticlinal structure with the leading edge overhanging the foreland sequence by several hundred metres. It is bounded in the north by the basal thrust of the overlying Kayakoy West structure by which it is directly overlain (in the east the structure is underlain by the foreland) (Fig. 15).

On the western flank the frontal thrust dies out and the structure passes laterally into a low-relief monoclinal structure (Fig. 9), merging with the autochtonous foreland sequences. The throw of the frontal thrust is at a maximum in the central area of the field where the structure is broadest and has maximum relief. The foreland shows most intense deformation beneath the area of maximum thrusting.

1.2 Beykan - Kayakoy West - Kurkan South Complex

The accumulations in the Beykan -Kayakoy West -Kurkan South complex lie within a single, heavily internally faulted thrust sheet (Figs. 2 and 10). The Beykan field was discovered in 1965 and by 1992 a total of 48 wells had penetrated the structure. 2D interpretation (Fig. 11) delineated an elongate thrust-bound, E-W trending anticlinal structure, internally cut by a number of E-W trending normal faults densest within the structural culmination (Fig. 14).

Interpretation of the Beykan 3D survey has shown that the Beykan field consists of an elongate thrust structure which is densely faulted, particularly on its steeper northern flank (Figs. 12 and 14). In addition to the classical compression related fault and fold features identified throughout the Foothills Belt, a complex series of cross cutting elements have been identified. The leading edge of the field is no longer mapped as smoothly cuspate, but instead is interpreted to be sinuous, complex and intersected by numerous discrete crestal faults.

Internal deformation within the Beykan structure comprises both normal and reverse faults, with numerous splays branching off the field-bounding thrust. Well-expressed backthrusting affects the flank of the field. Some of the crestal faults are oblique to the main axis of folding and appear to constitute conjugate sets. In addition to the crestal features the field is cut by two sets of steep, near N-S trending reverse faults which effectively compartmentalise the structure (Fig. 14).

The Baysu, Bektas, Kurkan South, Sahaban, Yesildere and Kayakoy West oil fields are located within the Kayakoy West -Kurkan South structure (Figs. 2 and 15). These accumulations share a common oil-water contact (OWC) at 1020 m subsea

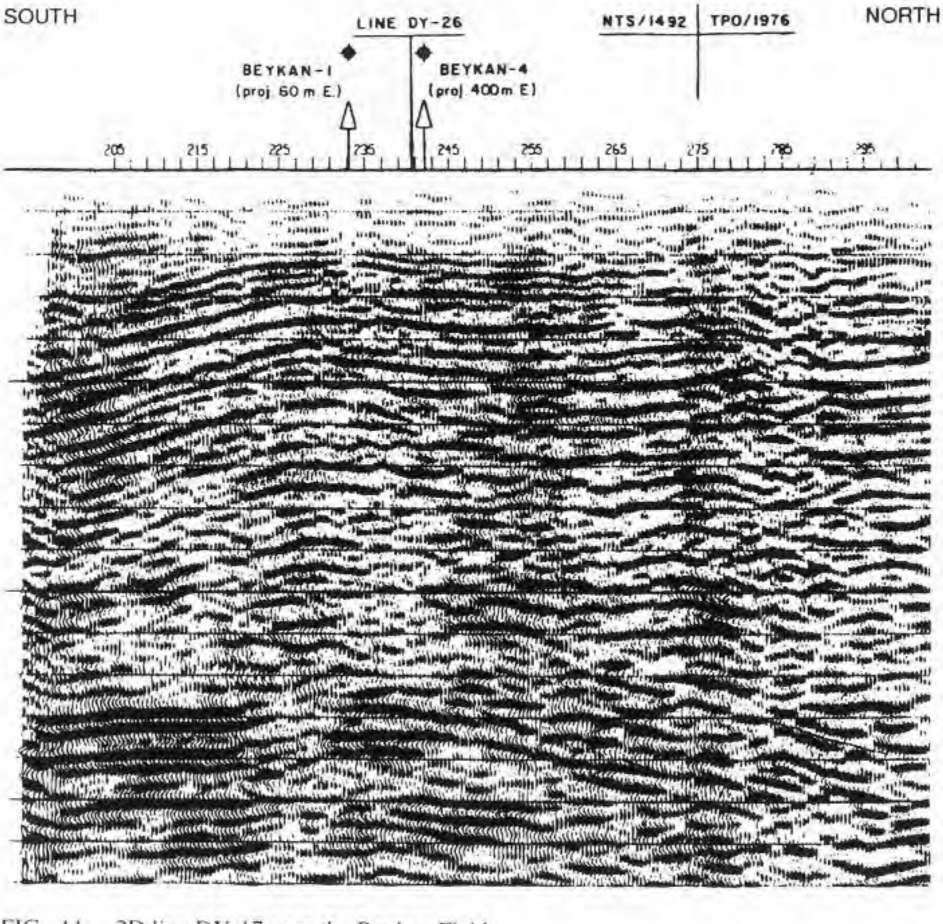

FIG. 11. 2D line DY-17 over the Beykan Field

although the controlling mechanism was not identified prior to 3D seismic acquisition. Following 3D interpretation, delineation of several of these fields and the distribution of oil within the Kayakoy West -Kurkan South complex was significantly altered. Instead of a series of relatively small cuspate structures separated by thrust faults, the sheet is now seen to comprise a single larger accumulation, combining the Kayakoy West, Kurkan South, Yesildere and Sahaban fields. It is connected to two satellite blocks forming the Bektas and Baysu fields (Fig. 16).

This large, single thrust sheet is bounded to the south by a frontal thrust and internally crosscut by several fault sets. These consist of numerous shorter, oblique reverse faults, predominantly trending WNW-ESE, with a secondary conjugate set trending ENE-WSW. In the southwestern area of the thrust sheet, well developed backthrusting is oriented sub-parallel to the frontal thrust.

1.3 Kurkan - Sincan Complex

The Kurkan accumulation is the second largest producing field in the NVTS Lease Areas and consists of an elongate thrust-bound anticlinal structure which has been subdivided into the Kurkan and Sincan oil fields (Figs. 2 and 15). The structure has been penetrated by 48 wells and was historically interpreted as a relatively unfaulted imbricate structure with the culmination located in the southwest of the field.

The gross geometry of the fields following 3D interpretation is broadly similar to previous interpretations. However, the structure is interpreted to be considerably more heavily faulted than previously realised. Faulting is predominantly reverse with fault planes often sub-vertical. The dominant fault trend is ENE-WSW with a less prominent conjugate set trending ESE-WNW. A smaller num-

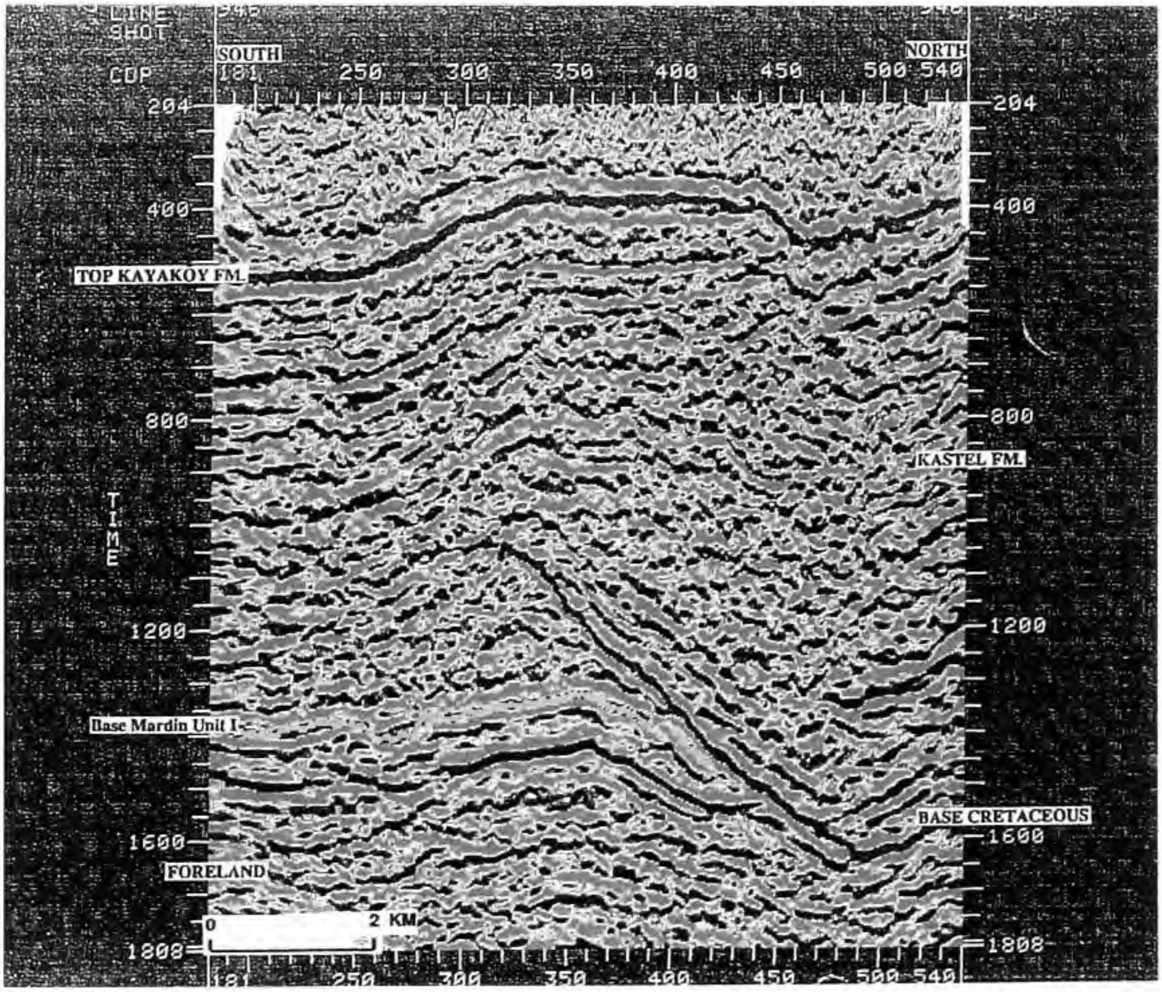

FIG. 12. Inline 946 over Beykan structure and underlying foreland

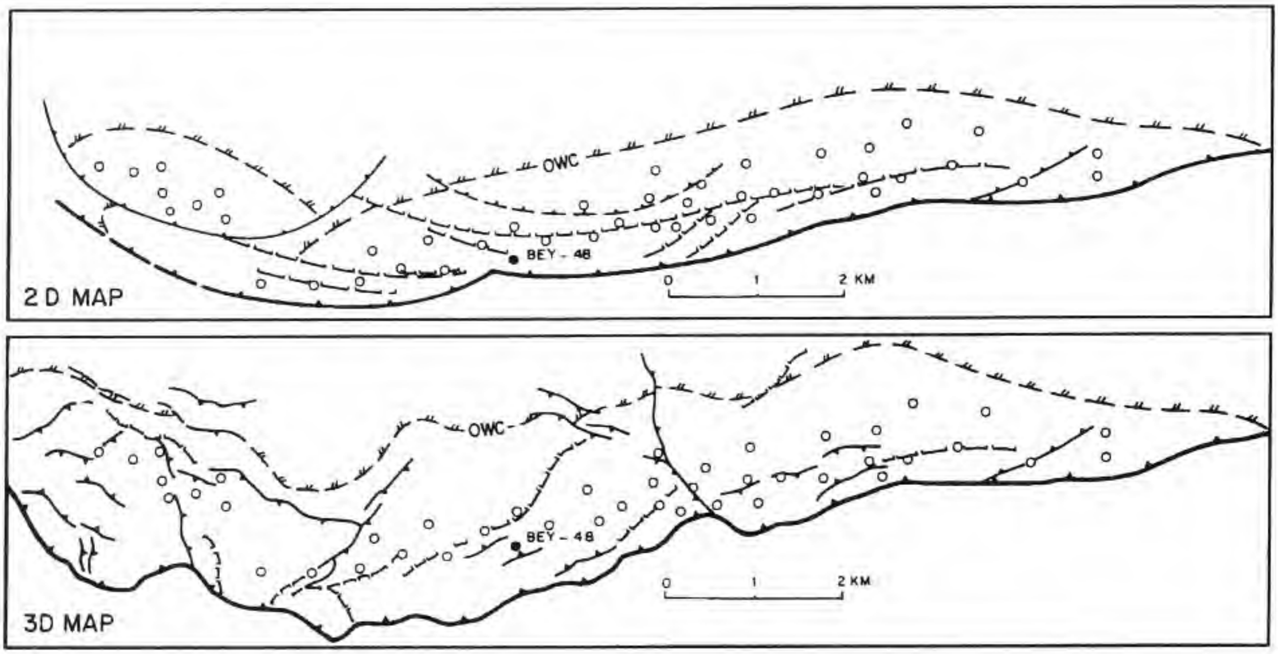

FIG. 13. Beykan Field summary maps. Pre-, and post-3D interpretation.

ber of faults trend either N-S (i.e. near perpendicular to the main thrust), or E-W forming backthrusts. The leading edge of the structure is a sinuous surface, locally offset by intersecting faults.

The culmination on the southwestern edge of the structure is adjacent to a major sidewall ramp (Fig. 17). The Top Mardin surface plunges some 350 m deeper to the west of this ramp. It is likely that more intense buckling related to thrusting over this sidewall ramp was responsible for the presence of larger faults in the western part of the Kurkan field.

2 Eastern Lease Areas

In the Eastern Lease Area (Fig. 2) two major thrust sheets are present (Katin -Barbes, and the overlying sheet making up the fields of the Northern Trend). The poorly developed Caytepe feature to the south may be an incipient thrust structure but seems to have more affinities with the foreland setting.

2.1 Katin - Barbes Structure

The Katin-Barbes imbricate is structurally rather simple, consisting of two thrust-bound anticlinal structures, separated by a saddle (Fig. 2.). The crestal area is relatively tightly folded and the northern flank dips at 25-40°, progressively steeper towards the west (Fig. 18). A deeper Palaeozoic gas-condensate accumulation is located within the same thrust sheet, almost directly above the decollement level. Within the Mardin resevoir in the Katin-Barbes oil field, faulting is predominantly parallel to the thrust front, consisting of a combination of concave upwards cuspate, south hading backthrusts and lower angle normal faults on the northern flank (Fig. 19). On the eastern margin of the Barbes structure the frontal thrust passes laterally into a sidewall ramp, the thrusted Mardin sequence merging with the foreland (Fig. 20).

To the north and overlying the Barbes field, the Kastel sequence with abundant Karadut facies sediments, has also been involved in thrusting and a complex series of intra-Kastel Formation unconformities, slumps and onlapping relationships may be identified.

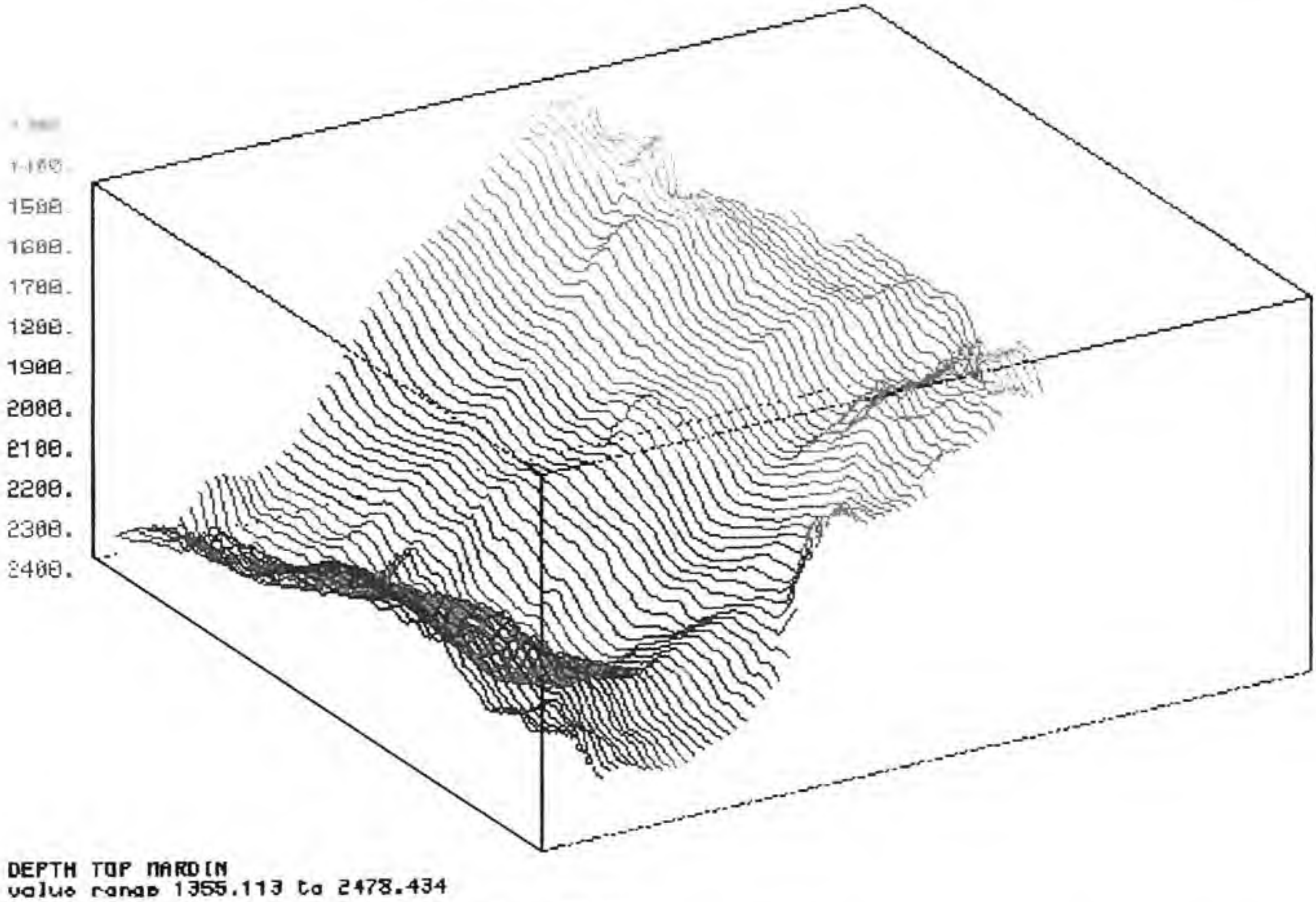

FIG. 14. Isonmetric display Beykan Field viewed from WNW.

3 Foreland Areas

In contrast with the large number of wells within the thrust structures situated directly to the north, the number of wells penetrating the autochtonous Mardin sequence of the foreland within the Lease Areas is small (Figs. 6, 10 and 12). Although hydrocarbons shows were encountered in some of these wells, producible quantities of oil have not been found to date.

The foreland area south of the Beykan Field has previously been mapped as being largely undeformed. The foreland Mardin Group was penetrated in two deep, dry Beykan wells. Oil was, however, produced on test from the exploration well Beykan South-1. Footwall deformation leading to the formation of potential traps, was not previously observed on 2D data largely as a consequence of the poor data quality beneath the overhanging Beykan structure (Fig. 11). 3D data however, shows that foreland deformation is more complex and intense. The foreland underlying the Beykan Field can be seen to be deformed and a low relief footwall ramp anticline is interpreted to have formed immediately beneath the Beykan structure (Fig. 12).

The 3D azimuth map of the foreland area south of Kayakoy West and Kayakoy West Deep shows the Top Mardin Group surface to be cut by two main fault sets (Fig. 18). Close to the intersection with the frontal thrust (effectively the northern margin of the foreland), a series of reverse faults are interpreted. These faults (possibly incipient thrusts) are parallel to the thrust front, their spacing decreasing northwards into the footwall. A second set of predominantly reverse faults is aligned near-orthogonal to the first set. The orientation of these faults seems to rotate from N-S in the west to NNW-SSE in the east of the survey area, (i.e.

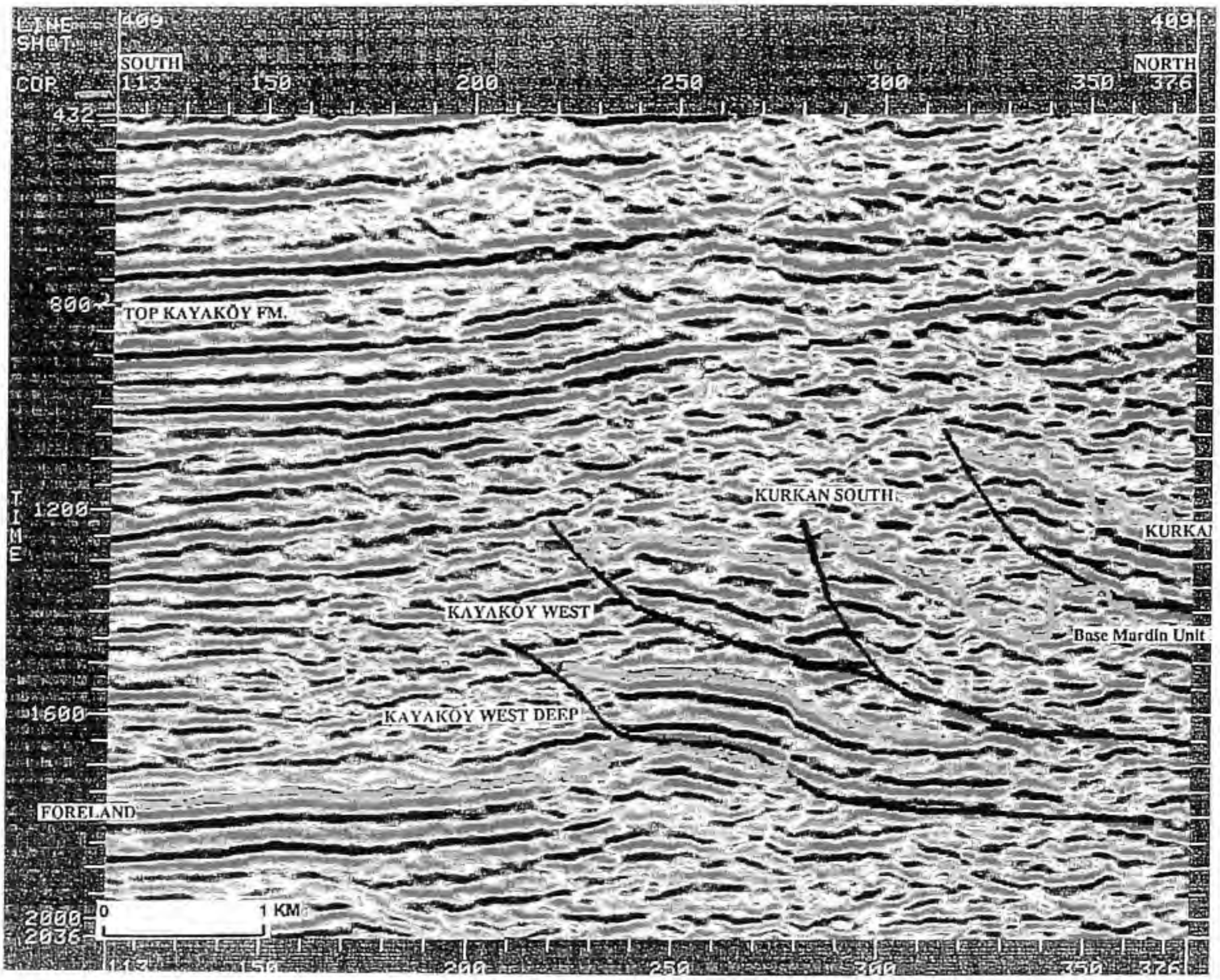

FIG. 15. Interpreted inline 409.

remaining nearly perpendicular to the curved thrust front). These faults are linear with a sub-vertical dip.

Foreland deformation is significantly more intense than previously realised, particularly immediately beneath the imbricated sequences. It appears that the deformational effects in the foreland of the thrusted overburden are restricted to an area within 1-2 km of the fault plane/foreland intersection. Beyond this zone both the intensity of faulting and folding decreases rapidly until the dominant faults appear to be near N-S lineations interpreted as incipient or aborted sidewall ramps.

RESERVOIR DISTRIBUTION AND PLATFORM IMBRICATION

Typically, oil fields occupy crestal positions within imbricate structures. By far the most important accumulations are found in the Lower and Upper Cretaceous carbonates of the Mardin Group. The best reservoir rocks are dolomites with secondary intercrystalline porosity and grain supported limestones (Cordey and Demirmen, 1971; Cater and Gillchrist, 1994). Upper Palaeozoic sandstones also act as reservoir rocks in the imbricates in the east. The carbonates of the Kastel Formation form the top seal.

The better Mardin reservoirs were formed by diagenetic processes which caused the formation of dolomites. Dolomitisation cross-cuts stratification

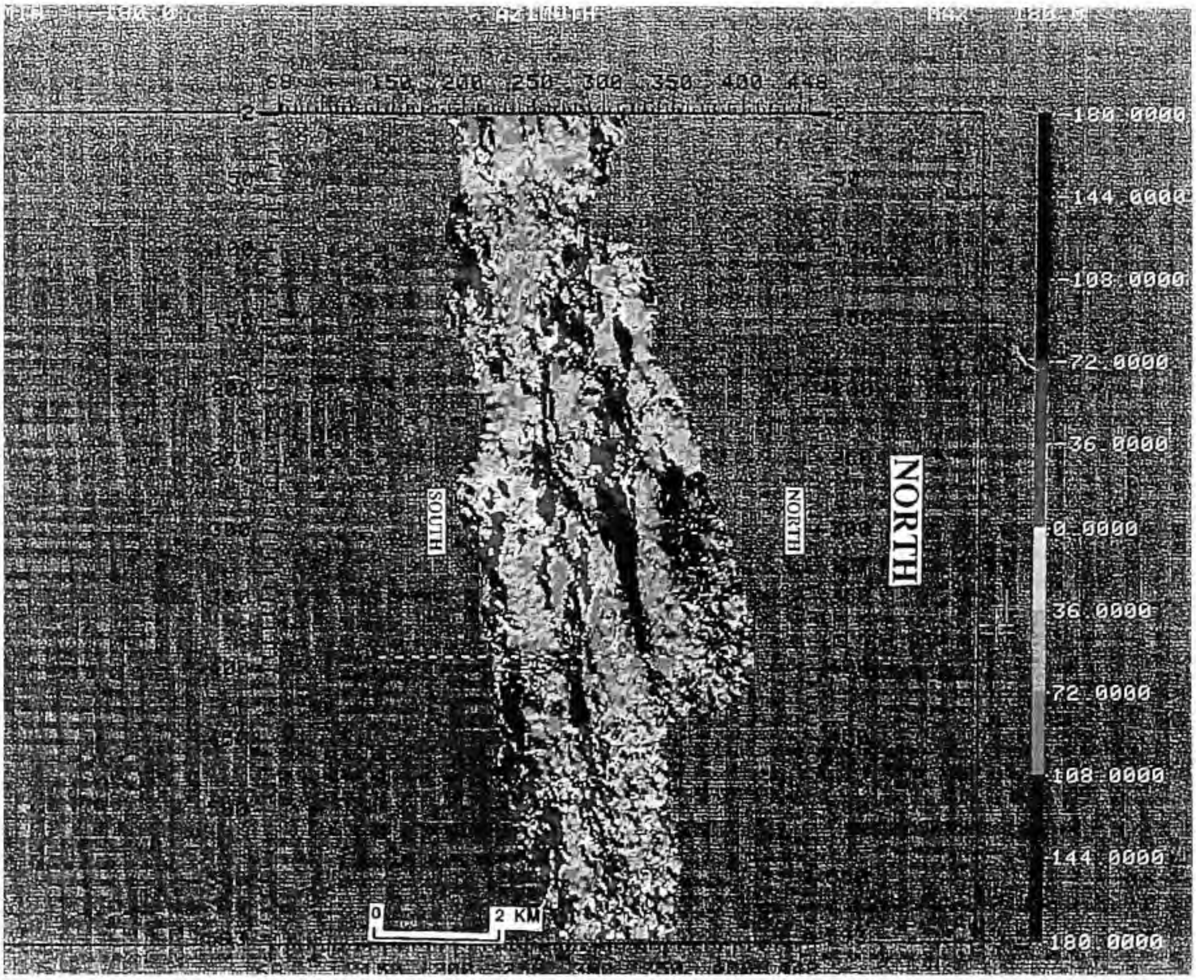

FIG. 16. Azimuth map, base Mardin unit I, Kurkan south-Kayakov west.

and is variable in thickness (Cordey and Demirmen, 1971; Cater and Gillchrist, 1994). The presence of thick dolomites within the oil fields area contrasts with both the foreland setting to the south and the thrusted sequence of the Foothills Belt to the north. The results of some 25 exploration wells in these areas demonstrate that dolomite development in the Upper Mardin sequence is very limited or entirely absent.

Within the Beykan field dolomites in the Upper Mardin sequence seem to be primarily developed in an area parallel, but slightly offset to the north of the field culmination. There is evidence that this relationship exists in analogous imbricate structures. This spatial distribution of dolomites, combined with the observation of evidence for subaerial exposure led to the conclusion that the terranes presently incorporated in the imbricated zone formed a positive area during the deposition of the uppermost Mardin Group in the Late Cretaceous (Cordey and Demirmen, 1971).

Imbrication of the platform margin occurred during the deposition of the complex sequences of the Kastel Formation (Fig. 21). The preferentially dolomitised areas, related to crestal areas in individual structures, seems to imply that the location of the eventual breakthrough of the basal thrust plane coincided with the Mardin Forebulge Area. The deformational history is complex, some structures forming as folds prior to imbrication while others imbricated without appreciable folding.

As a result of the continuous growth of the Mardin structures, reworking of the Kastel Karadut sediments (Fig. 8) occurred. Kastel Formation sediments on the crests of the imbricating structures were eroded and redeposited as clastic wedges at the leading edges of the imbricates. Eventually, Karadut sedimentation was restricted to the deep-

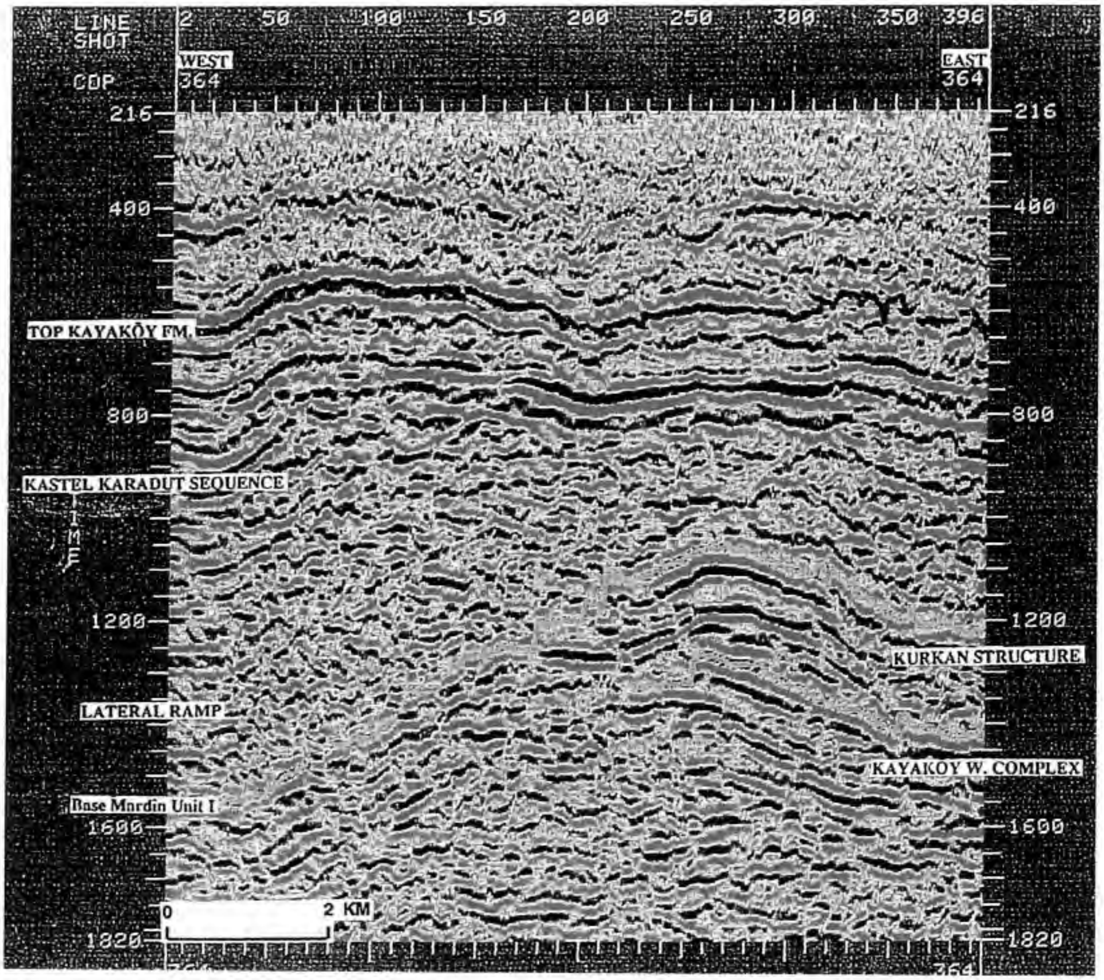

FIG. 17. Crossline 364 showing lateral ramp on Kurkan flank.

ening trough between the imbricated platform edge and the orogenic belt in the north. The Karadut complex reached an ultimate thickness of over 2000 m (including tectonic repetitions) to the north of the production licences. To the south the imbricated zone at the platform edge must have formed an effective barrier to Karadut sedimentation. This is supported by the virtual absence of Karadut sediments in the Foreland.

OIL MIGRATION PATHS AND FIELD FLUID CONTACTS

The oil produced from the Foothills Belt oil fields has been generated from thick marine Silurian shales of the Dadas Formation. The kitchen is interpreted to be located to the north of the producing fields, in the trough which separates the Orogenic Belt from the Foothills Belt. Oil migration probably post-dates formation of the Mardin forebulge, with the main phase of expulsion taking place during deposition of the Kastel Formation. Deposition of the upper part of this formation, ie. sediments of the Kastel Karadut facies, buried the source rocks to a depth sufficient for maturity to be reached. Expulsion probably continues to the present day.

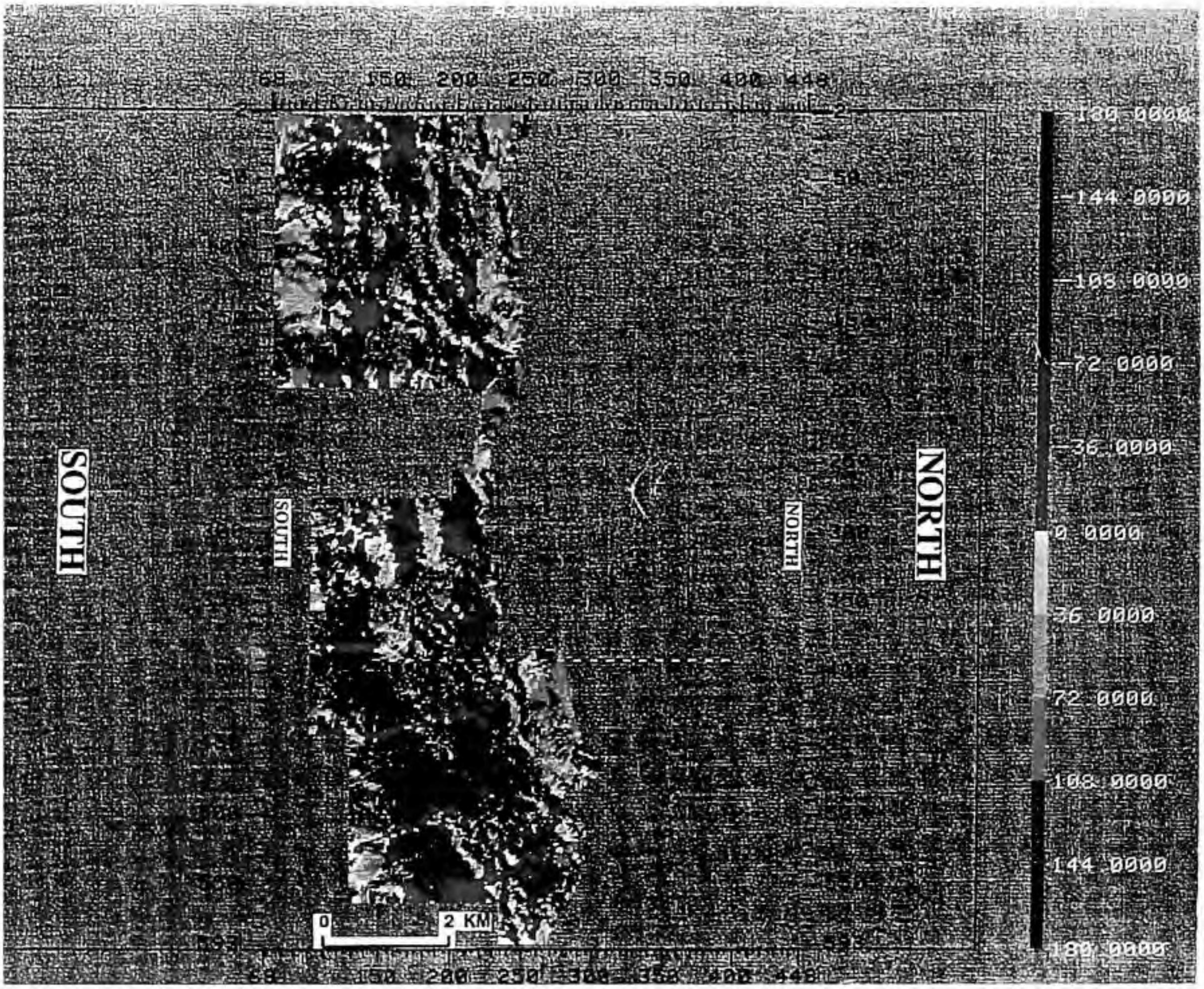

FIG. 18. Azimuth display, base Mardin unit I forland, Kurkan extension 3D.

Despite the complexity of the thrustbelt structures, the original fluid contacts exhibit a remarkable consistency between fields.The Beykan field (Fig. 13) for example, has an OWC at 1080±20 m subsea, while the fields in the Kayakoy West - Kurkan South structure share an OWC at 1020±50 m subsea.

Following 3D mapping there is substantial evidence that fluid contacts are structurally controlled. The accumulations in the Kayakoy West - Kurkan South complex, interpreted to share a (near-)common OWC, can be demonstrated to be structurally interconnected (Fig. 23). The position of the original fluid contact is controlled by a structural spillpoint mapped between 1040 and 1100 m subsea. Spill probably follows the structural saddle between the Kayakoy West -Kurkan South complex and the Beykan field, oil eventually escaping to the north via the Kurkan western flank or an unidentified structure to the north of Beykan. The shallow Kurkan -Sincan accumulation (Fig. 3) may have been partially directly charged by hydrocarbons from the kitchen to the north, with an additional contribution from the Kayakoy West -Kurkan South complex beneath, oil migrating vertically, probably via faults, through the ramp beneath the Kurkan structure.

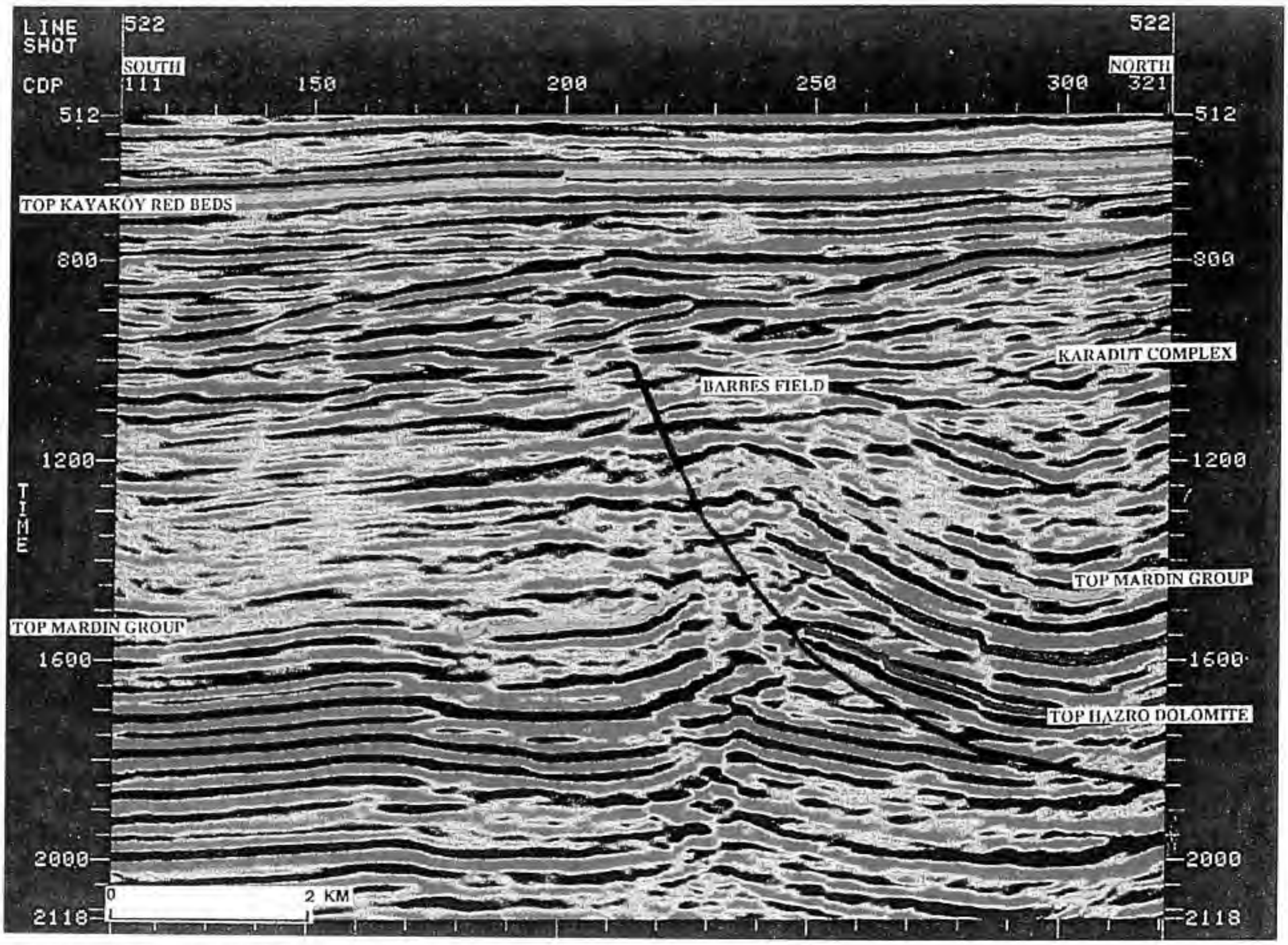

FIG. 19. N-S line 522, Katin-Barbes 3D

KINEMATICS AND MECHANICS OF THE IMBRICATION OF THE MARDIN SEQUENCES

The producing structures on the southern edge of the SE Turkey Foothill Belt are typical examples of imbricate structures in a Foreland setting. Thrust sheets are relatively thin, comprising a layered sedimentary sequence and have a length which is several times greater than their thickness. Imbricate structures in this setting are thought to have been initiated as if they were "pushed from behind" by some tectonic force over a decollement horizon (basal slip or detachment plane).

In the Eastern Lease Areas (Fig. 2) a clastic sequence overlies the argillaceous units of the Dadas Formation. This clastic sequence is progressively cut out to the west such that in the Beykan area the Cretaceous carbonate sequence lies almost directly on top of the basal shales (Fig. 12). The observed effect is that, in the east, imbricates are significantly thicker and contain in their basal sections a sedimentary sequence with fundamentally different mechanical properties compared to the overlying carbonates. In a mechanical sense the thrust sheets are thus composed of a relatively stiff, brittle beam of carbonates where the brittleness is enhanced by dolomitisation, underlain by an eastward thickening sequence of less brittle clastics on a weak, argillaceous substratum. The difference in thickness has traditionally been taken as the primary explanation for the difference in tectonic style. The difference in mechanical properties is probably an equally significant factor.

In the west (e.g. Beykan Area) folding took place on a smaller scale as a consequence of the

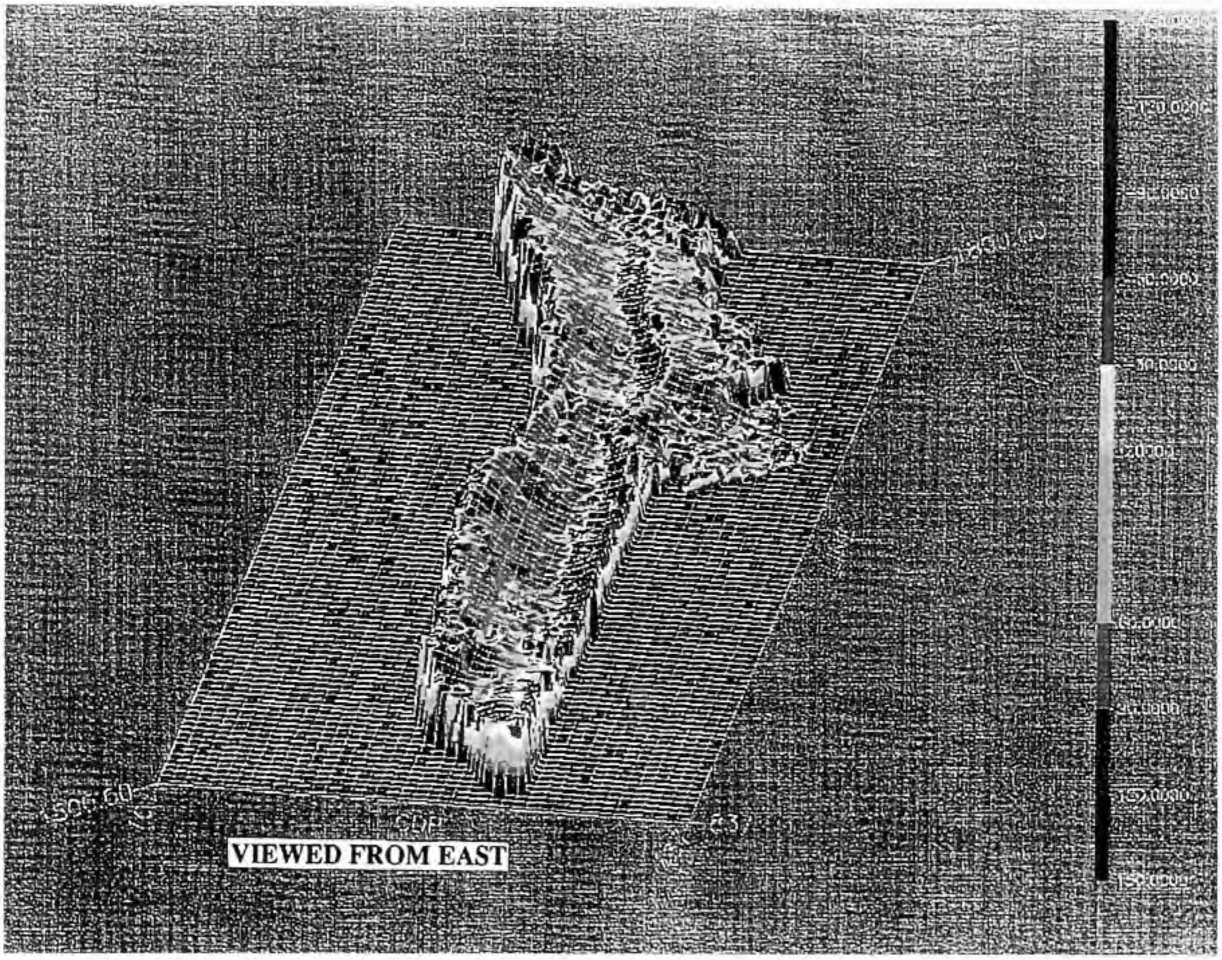

FIG. 20. Isometric display, base Mardin group, Katin-Barbes structure, with azimuth overlay.

reduced thickness of the stratigraphic section involved (Fig. 23). Initial shortening was accompanied by gentle folding, as evidenced by the low structural dips encountered in wells and on seismic. Continued shortening caused the basal thrust/slip plane within the Dadas Formation to ramp upwards through the brittle carbonate sequence of the Mardin Group and thrust the allochthonous block over the foreland (Fig. 23). In the east the increased thickness of the thrusted sequence led to larger amplitude folding and higher structural dips. Increased shortening was therefore accomodated prior to the breakthrough of the basal thrust. Folding prior to thrusting was further amplified by considerable ductile deformation of the clastic Dadas sequences within the cores of the anticlines.

The thickness of the eastern thrusted sequence must have caused vertical loading on the basal slip plane to be higher compared to that of the thinner western sequence. The likely effect was that, with a higher load, the frictional characteristics along the basal slip plane changed and hindered the movement of the thrust sheet (Goff and Wiltschko, 1992). Finite element analysis of tapered thrust models (Mäkel and Walters, 1993) has clearly demonstrated that variation in basal friction does influence the resulting geometry of imbricated structures. Reduction of friction, either by changing characteristics of the rock involved or by increasing pore pressure, tends to result in larger thrust sheets (Mandl and Shippam, 1981).

In a mechanical sense the thrust sheets of the Kurkan, Kayakoy and Kayakoy West structures (Fig. 17) resemble that of the Beykan structure, i.e. thin Dadas sequence and relatively gentle folding. However, in contrast with Beykan where the deformation resulted in one large imbricate structure,

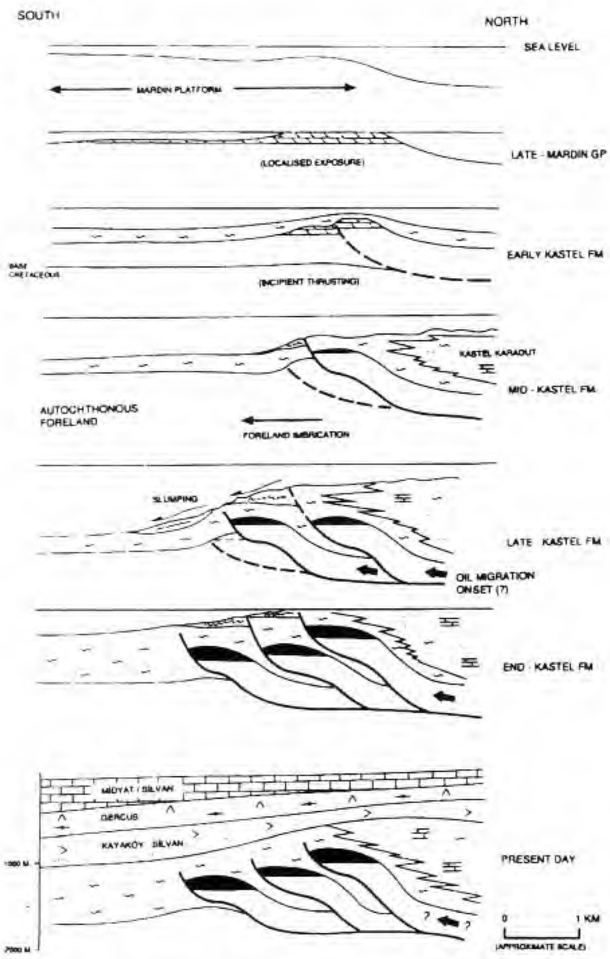

FIG. 21. Schematic evolution of NVTS Lease Areas oil-fields.

the Kayakoy West area consists of a stack of smaller imbricate structures (Fig. 24). Since differences in slab thickness are not apparent, the changes are interpreted to relate to greater friction along the basal slip plane beneath the Kayakoy West area compared to the Beykan structure. Little data is available on the composition of the rock within the slip plane. Relatively minor differences in the composition of the slip plane rocks, e.g. an increased presence of siltier material, could be responsible for increased friction (Goff and Wiltschko, 1992).

The last parameter which may have influenced the geometry of the structures is the shape of the footwall ramp, which in turn affected friction on the basal thrust plane. As a result of propagation of the basal slip plane ahead of the leading imbricate (Mäkel and Walters, 1993), subsequent foreland deformation may have been superimposed upon structuration which predated the main compressional phase. Pre-existing faults in the foreland sequences may have caused the formation of side-wall ramps within the thrust sheets. Indeed in the Beykan and Kayakoy West areas (Fig. 18) there is evidence to support this. Thus, the geometry of the footwall, in combination with frictional behaviour along the slip plane, was most likely responsible for the observed differences in the deformational styles of the individual thrust sheets.

The geometries of the Kurkan and Kayakoy West structures (Fig. 3) and the influence of basal friction during their creation conform with the classical interpretation of foreland propagation of

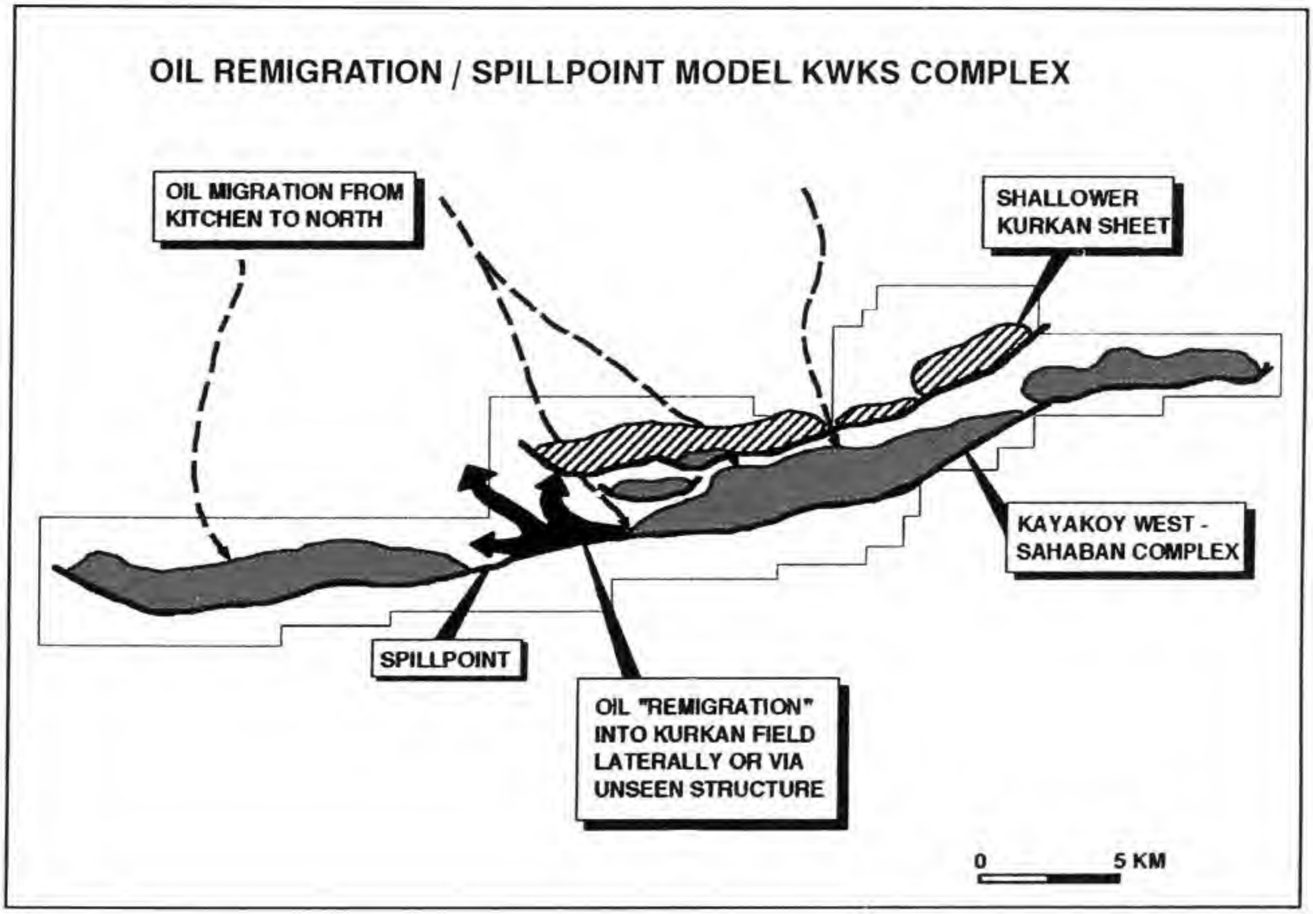

FIG. 22.

imbrication. This implies that the imbricate closest to the hinterland forms first where a steep thrust intersects the basal slip plane (Mäkel and Walters, 1993). When the imbricate overrides the foreland, the stress regime and hence the deformation in the footwall adjusts itself to the compressional force and the load exerted by the imbricate. The location of the next imbricate then depends primarily upon the mass of the overriding thrust sheet and the frictional characteristics along the basal slip plane. In principle, a higher load and higher friction will lead to shorter imbricates (Mandl and Shippam, 1981; Goff and Wiltschko, 1992).

The formation of the second imbricate will tend to steepen the first (Li Huiqi et al., 1992). Indeed the Kurkan structure seems to have a much steeper northern flank than the underlying Kayakoy West structure (Fig. 15). The Kayakoy West Deep structure (Figs. 9 and 15), overrides the foreland in the east but in the west is still connected to the foreland through a transfer zone expressed as a monocline. Thus it seems that Kayakoy West Deep is partly an incipient thrust structure. It was most likely formed after the Kayakoy West structure was pushed over a footwall ramp, but the next imbricate (Kayakoy West Deep) failed to complete its development. Indeed the evidence from other areas (e.g. Beykan and the Caytepe structure in the Eastern Lease Area) seems to indicate that foreland imbrication was the dominant process.

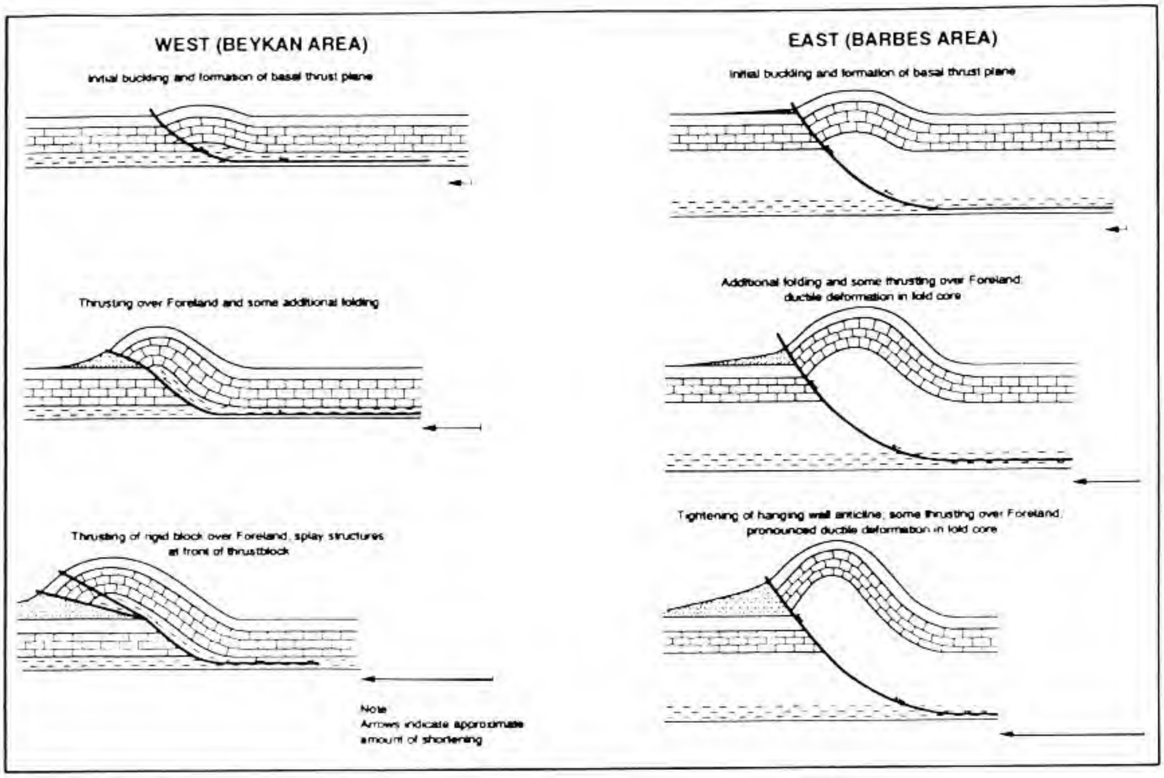

FIG. 23. Influence of stratigraphic thickness and rock properties on imbricate geometry.

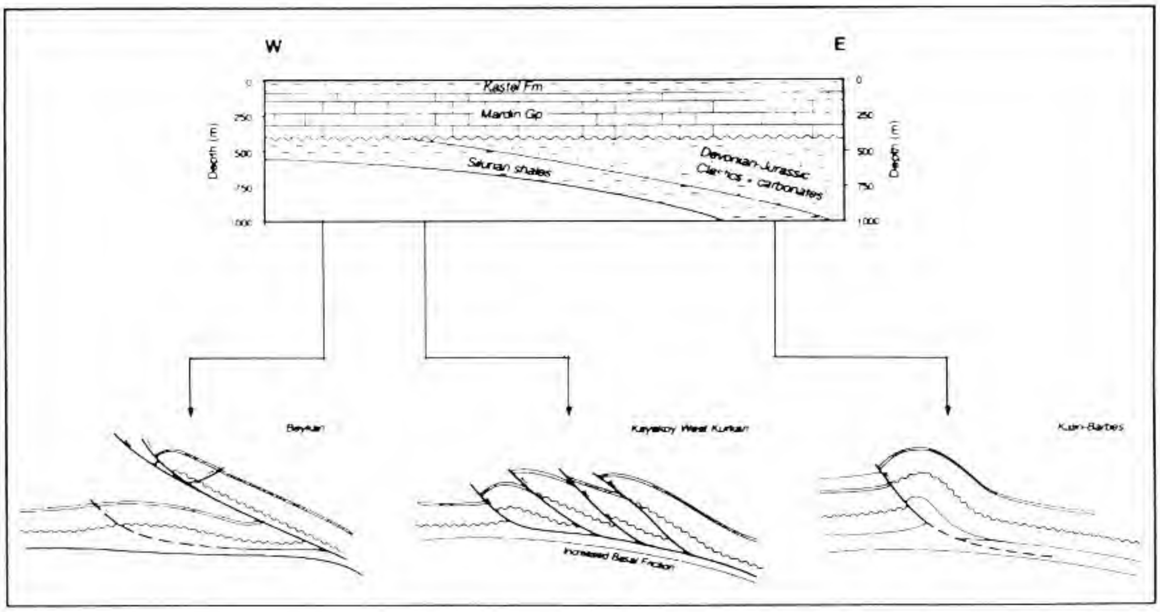

FIG. 24. Difference in imbricate geometry as a result of stratigraphic thickness and frictional properties.

POTENTIAL IMPACT OF 3D RESULTS ON APPRAISAL AND DEVELOPMENT IN SE TURKEY

The results to date of the acquisition of 3D seismic in the SE Turkey thrustbelt can be summarised as follows.

(1) 3D seismic data has been demonstrated to substantially improve the delineation of subsurface structures compared with previous 2D results. Definition of internal faulting and fault block boundaries, identification of the interconnectivity of producing fields and correlation of major faults have been radically improved. This has also positively impacted upon the general understanding of how these structures formed and the parameters which influenced their ultimate geometries,

(2) 3D results have led to fundamental revisions of structural maps in both heavily drilled (Beykan) and lightly explored (foreland, Kayakoy West Deep) areas. Interpretation of previously unseen attic areas have been demonstrated in several mature fields and the new structural maps are likely to provide a basic framework for future reservoir characterisation work.

Acknowledgements - The authors wish to thank *NV Turkse Shell* for permission to publish these results. The interpretation of the seismic data represents our personal views and subsequent work may have resulted in refinements to the subsurface model. Many thanks to Rifat Güler for draughting the enclosures, to our former colleagues in production geology and exploration in *NVTS* for providing additional material and constructive criticism, and to Alf Garnett who provided the Katin -Barbes 3D material.

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LIST OF ENCLOSURES

Enclosure 5 : a. 2D seismic line through the Arzacq Basin, time migration ; *b.* 2D seismic section through the Rousse and Meillon fields.
Enclosure 6 : a. South Aquitaine, 3D seismic surveys ; *b.* 3D Meillon survey, Rousse and Meillon gas fields ; *c.* Dip structural cross-section through the Rousse and Meillon gas fields ; *d.* Dip cross-section through the Upper Lacq oil field and the giant Deep Lacq gas field.

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Enclosure 2 : Regional balanced cross-sections through the Central and Eastern Jura. n° 5 : Neuchatel Lake — Ognon fault system. n° 6 : Grenchen anticline — Rhine Graben. n° 7 : Aarau — Tafel Jura. n° 8 : Lagern anticline.

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Enclosure 2 : Structural details of cross-sections 1 and 2 shown in Encl. 1.
Enclosure 3 : Structural details of cross-sections 3 and 4 shown in Encl. 1.
Enclosure 4 : Retro-deformation in 5 stages of cross-section 3 of Encl. 1.

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Enclosure 1 : Tectonic map of the Polish Outer Carpathians and location of the cross-sections given in Encl. 2.
Enclosure 2 : Geological cross-sections through the Polish Carpathians (for location, see Encl. 1).
Enclosure 3 : Stratigraphy and structural location of oil and gas reservoirs in the autochton and allochthon.

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Enclosure 1 : Geological cross-sections through the Northern and Central Italy.
Enclosure 2 : Geological cross-sections through the Southern Italy.
Enclosure 3 : Geological cross-sections through the Southern Italy and Sicily.

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Peter A. ZIEGLER (Geological-Paleontological Institute, University of Basel, Switzerland, retired petroleum geologist) and Frank HORVÀTH (Geophysical Institute, Loránd Eötvös University, Budapest, Hungary) convened the American Association of Petroleum Geologists Symposium held in The Hague out of which this memoir developed.

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