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Erosion and Sedimentation

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Erosion and Sedimentation

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Preface

Geology today demands the knowledge of a large number of fundamental ideas which are provided by neighboring sciences. The student, or the geologist at the beginning of his career, runs the risk of losing himself in a multitude of details and complications, which will seldom be useful to him. To guide him, we have endeavored to prepare a work which is easy to read, not overburdened with tables and graphs, and which avoids swamping the reader with all the details of many specialized disciplines. Our aim is to stress the general facts and to bring out the conclusions which will enable the reader to reconstruct the great events that have taken place in the past on the surface of the earth.

This volume frequently uses results obtained by geomorphologists and biologists, but it will not overlap with the treatises and manuals written by them, since its point of view is quite different. In attaining this objective, morphology is but one of the many ways which can serve to trace the history of our globe.

The very important problem of uniformitarianism arises here. Like most geologists at the present day, we believe that the history of the surface of the earth can be explained almost entirely on the basis of observation of present-day phenomena. It is important, however, to note some limitations. Even among the processes which occur today, there are some which it is very difficult to observe (for example, turbidity currents). Furthermore, many mechanical, physical and chemical phenomena resulting from internal geodynamic processes, such as folds, metamorphism, and granitization, are only known by their effects.

It is necessary, also, to consider the time factor: most of the events which modify the surface of the earth have a duration which greatly exceeds that of a human life. The conditions which have existed on the earth have changed considerably since the beginning of what we may call "Fossiliferous Times", that is to say, the last 600 million years. In fact, Organic Evolution cannot be doubted by anyone, and there must have been periods when the continents were entirely without a plant cover. Life, which seems to have appeared in the sea near to coasts, may have reached the ocean depths at about the same time as the surface of the continents emerged.

We can also deduce the existence of a Geochemical Evolution, which develops in a way parallel to the evolution of Life. It is thus important to be familiar with current work over a wide range of subjects.

Finally, uniformitarianism relies, by definition, on the characteristics of the period in which we live, but it is necessary to note that this has been profoundly affected by the events occurring during the Pleistocene. These ("Ice Age") times have had few equivalents in the earth's history and it is necessary to go back to the Carboniferous or to the Precambrian to find glaciations comparable to those of the last million years. The present period is still under the influence of the recent glacial phases and of the regression accompanying the last extension of the ice. This is apparent from the morphology and from the impoverishment of the flora and fauna. It is, therefore, essential to take account of these facts when one attempts to reconstruct the history of the earth.

We have, therefore, found it necessary to begin this book with a rapid survey of the climates, past and present, paying special attention to the conditions of the Quaternary glaciations.

Secondly, we have stressed the close relationship between erosion and sedimentation on the land surfaces and in the seas: detrital sedimentation results directly from erosion, the distribution of saline deposits is associated with the solubility of each of the elements of the hydrosphere, the existence of an association between erosion and vegetation is shown by the formation of soils, while limestone sedimentation is essentially due to the activity of marine organisms. These factors have grouped themselves in various ways in the course of geological time, and these combinations have allowed, in large measure, the interpretation of the succession of formations studied in stratigraphy.

Note for the English-language edition.—For the purposes and needs of our English-speaking readers, we have recast the chapter arrangements in large measure, adding some material, and introducing locality indications which might not otherwise have been obvious. A glossary of selected technical terms has been added.

Introductory Note

One of the most general factors touching upon morphology and also governing geology is *the elevation of the continents*. A more or less continuous emergence has, in fact, been proved by the occurrence of erosion surfaces on the old continents, and also by the existence of more limited, Quaternary terraces. The African, Scandinavian, North and South American and Australian shields, for example, have undergone repeated planation, the stages of which it is possible to enumerate and document in detail. Simultaneously, the *oceans have tended to become deeper*. The *epeirogenic* movements of the continents are thus matched by those movements which we have called *bathygenic*[1] affecting the oceans.

We have also to take into account the *regional movements of continents and oceans*, which can, superficially and temporarily, obscure this general scheme. While certainly there is a general tendency for the oceans to become deeper while the continents are being uplifted, the existence of periodic transgressions, marked by the overflowing of the sea on to the land areas, indicates an opposing tendency. The causes of such transgressions are varied: *eustatic transgressions* (p. 22), in particular, appear to be well established. Their effects are superimposed on those due to orogenesis, or to foundering, to which the crust of the earth is constantly subjected. The movements, of which we observe the traces, are thus always relative.

The explanations which have been given of continental erosion are thus varied; as more problems are envisaged, the greater number of unknowns suggest more solutions. There is no lack of general theories, but we cannot adopt, in its entirety, any one of them. In fact, many theories are founded on groups of interesting and well-documented facts, but they may only be applicable in limited circumstances. The common failing of these ideas is an excessive generalization, that is to say, application to instances where they can no longer be verified, thus demonstrating that they are opinion rather than fact.

[1] From βαθύς, deep, and γενιχός, to bring about.

Contents

I

Zonation and the Geographical Cycle

Two main causes have been advanced to explain the geomorphic evolution of the earth's crust.[1] Both processes give rise to the relief of the earth's surface: these are zonation (geographic—vegetational and climatic) and the geographical cycle. It will be seen that far from opposing each other, these two processes are often complementary.

The study of *climatic distribution* has been particularly advanced in the vast territory of the Soviet Union (K. K. Markov, 1956). Zonation depends essentially on global characteristics: the shape of the earth, its revolution around its axis and its thermal regime. There must always have been varied climates which influenced paleogeography, geomorphology and sedimentation. However, the experience of geologists leads to the conclusion that these zones can only be demonstrated in landscapes produced since the time when Life became widespread over the land. For example, in the Precambrian the character of the detrital sediments is monotonous throughout. The late Precambrian glaciation demonstrates an ancient climatic period of the earth's history, but the climatic zonation which would have accompanied it is not apparent in the sedimentary succession, owing to the absence of fossils.

On the other hand, geologists have been able to establish a definite climatic zonation for the Carboniferous and especially for the Permian. It seems likely that it will also be possible to do the same for all periods back to the Devonian, that is to say from the time when the continental flora were sufficiently developed. The boundaries of zones are constantly changed by the whim of cosmic and geological fate, and the landscapes become modified according to the state of plant evolution. This only becomes comparable to the present day with the appearance of the Angiosperms.

Thirteen major geographical zones[2] may be distinguished today. These are from the poles to the equator: 1. the glacial zone; 2. the tundra; 3. the temperate forests; 4. the wooded steppes and forests with clearings; 5. the

[1] The authors use the term "Glyptogenesis", first introduced by de Martonne. Glyptogenesis is the process of sculpturing of the lithosphere through the agency of the atmosphere, hydrosphere, biosphere and pyrosphere. The term is rarely used in English and is here given as "geomorphic evolution", "sculpturing", or "morphogenesis". Translator and Editor.

[2] This scheme is modified from that of L. S. Berg for Eurasia. Similar zones occur on most continents. Only Antarctica is limited to one zone.

steppes; 6. the mediterranean zone; 7. the semidesert zone; 8. the high latitude deserts; 9. the subtropical forests; 10. the low latitude deserts; 11. the tropical wooded steppes or savannas; 12. the humid tropical forests; 13. the equatorial rain forest zone. The first two zones are the result of the Pleistocene glaciation.

In general, periods of orogenic calm and of the great "Tethys" transgressions (see pp. 29 and 67) display the fewest geographical zones, whilst periods of orogenesis have a zonation which is more marked and diverse. We speak of the "thalassocratic" (ocean-dominated), as opposed to "epeirocratic" or land-dominated periods.

W. M. Davis (1905) developed the concept of *cycles of erosion*. These are applicable to some, but not all, of the climatic types given and in general practice govern the development of land forms: a landscape passes through a stage of youth, a stage of maturity and a stage of old age. The succession of these three stages constitutes the *geographical cycle*, comparable with the "geomorphic cycle" of A. C. Lawson (1894). This theory permits the clear and precise explanation of the various phenomena related to the evolution of a certain type of landscape. Like all hypotheses, the Davisian cycle is only an ideal representation, and the true picture diverges from it to some extent. None of it astonishes the geologist, since he is accustomed to changes of climate and hence to varied conditions of erosion. He is also accustomed to see orogenic and epeirogenic movements taking place, which result in a change of base level. Indeed, the perfect geographical cycle is a limited idea which is practically never realized. However, this conception remains valuable and useful if it is applied with a critical approach.

The example which approaches it most closely is, curiously enough, the cycle of arid erosion, although Davis had primarily the humid-temperate environment in mind. There are two reasons for this: firstly, deserts which are typical of this cycle are most often situated (by chance!) in more stable areas less subject to orogenesis; and secondly, the conditions of arid erosion are such that these deserts attain rapid maturity. The latter is not the case in temporate regions. The Davisian cycle can be applied nicely to the Central Asian deserts, for example the Tarim Desert.

Cycles exist, however, which are much less influenced by climate than by the character of the rocks, as for example in the evolution of karst (limestone) scenery. There are also complex cycles where a number of factors occurring simultaneously, or in succession, combine to give the landscape its character.

It will be seen that the morphological evolution of ancient continental platforms is slow, and usually is undisturbed by movements of any great magnitude. The persistence of ancient shields makes them a suitable choice for research into morphogenesis: the differences in their individual histories and their diversity of appearance are caused partly by climate. It has thus been possible to criticize with justice the *Davisian cycle*, and more or less to substitute for it geographic zonation as the principal explanation of geomorphology. However, it would appear that in the case of ancient platforms, the two approaches are quite compatible.

The relief of these continental platforms, which have been generally reduced to a peneplain, lends itself well to the study of the function of zonation, while more recent mountain chains which border these platforms are still undergoing a complete cycle. During their long and slow history, the erosion agents have been streams, whose courses have varied little and have only been rejuvenated from time to time by the gentle periodic uplift of the shields. As a result, the areas of deposition which correspond with their mouths, have remained constant for a very long period of time. These regions, at the junction of land and sea, are those which have been most easily invaded by marine transgressions. There, in the alternation of marine and continental beds, our stratigraphic history is most clearly written.

CLIMATES AND GEOMORPHIC EVOLUTION

The influence of climates in present-day morphology is so obvious that it seems desirable to refer back to ancient climates whose analogous action has left a delicate imprint on geological formations. This has been shown by both stratigraphers and pedologists.

Everywhere the climates of the Quaternary have left unmistakeable traces. It may be said that the greater part of the earth's surface has hardly been modified since the Pleistocene. Thanks to C-14 dating, this is nowadays taken to have ended about 10,000 years ago. The period since then is called the Holocene or Recent. We are actually living in a mild interglacial or postglacial type of climate. The Pleistocene climates will be discussed below outlining the chief characteristics of the glacio-pluvial periods and the interglacial periods. Following that, the nature of arid phases will be discussed, comparing them with the present epoch.

Finally a short section will be devoted to the role of ancient climates. Little detail is known of these, but they hold an important key to the explanation of deposits and of organic evolution in the course of geological time.

A few general points follow.

Ideas Concerning Paleoclimates

It is increasingly apparent that morphogenesis and sedimentation are strictly dependent upon climate. It is necessary to assemble as many as possible of the recent findings on ancient climates in order to reconstruct precisely the different ways in which the earth's crust has been superficially molded.

The *fossil flora and fauna*, if compared with present populations, furnish useful information on this subject. But it can be reasonably objected that the habitat of an animal or plant-group can expand or contract and that the expectation of life of one form can be very different from that of its near ancestors. The temperature threshold related to the solubility of certain salts such as calcium carbonate and magnesium carbonate (Stehli, 1958) correlates the distribution of certain faunas to physico-chemical factors and makes it possible to build up a definite though incomplete climatology. This is the measurement of *paleotemperature* (see further below) which gives precise information about the climates of geological periods during the last 100 million years.

Glacial remnants, *eolian deposits* and *evaporites* (to which several sections will be devoted) are also excellent guides to the reconstruction of ancient climates.

Finally, the soils and particularly traces of laterization complete the story which the geologist unfolds to mark the zonal boundaries for each period of ancient geography.

Measurement of Paleotemperatures.—Since 1946, at the suggestion of H. C. Urey, a method of calculating the temperature of carbonate precipitation in the sea has been developed. The isotopes O^{16}, O^{17} and O^{18} of oxygen are in different proportions in atmospheric and sea water, the latter being rich in heavy isotopes as a result of differential evaporation. Furthermore, among carbonates precipitated in the sea the proportion of oxygen isotopes depends upon the temperature of formation: a difference of 1° C. produces a modification of 0·02% in the ratio O^{16}/O^{18}. Thus from the seasonal growth rings of a belemnite guard it is possible to reconstruct its birth in summer and, four years later, its death in spring. However, a source of error is introduced by isotopic variations within the oceans themselves: thus it is known at the present day, that in tropical latitudes the surface waters are 0·02% richer in O^{18} than those of the Arctic Ocean; furthermore, other variations must be taken into account near coasts, at the mouths of rivers, or in seas with a high evaporation. Such sources of error can be reduced by using pelagic organisms from the open sea. And not least, there remains the problem relating to variations of isotopic ratio during geological time.

It is emphasized that a knowledge of the temperature of the seas at a given epoch, plus all the accessory information that can be furnished by geology, combine to permit an approximate reconstruction of the whole climatic picture.

Quite important results have already been obtained in this sphere. Qualitatively, there is evidence of a temperature rise at the beginning of the Upper Cretaceous to a maximum about 85 million years ago. This was followed by a temperature fall at the end of the Cretaceous. The temperatures were as follows: for the Senonian Chalk with flints of England, from 14·2° to 23·8° C.; for the Chalk of the Upper Maestrichtian of Denmark, from 12·8° to 14·3° C., these two regions being situated between 52° and 56° north of the equator. From *Belemnitella americana* of the Maestrichtian of the southeastern United States, situated between 33° and 41° north, the

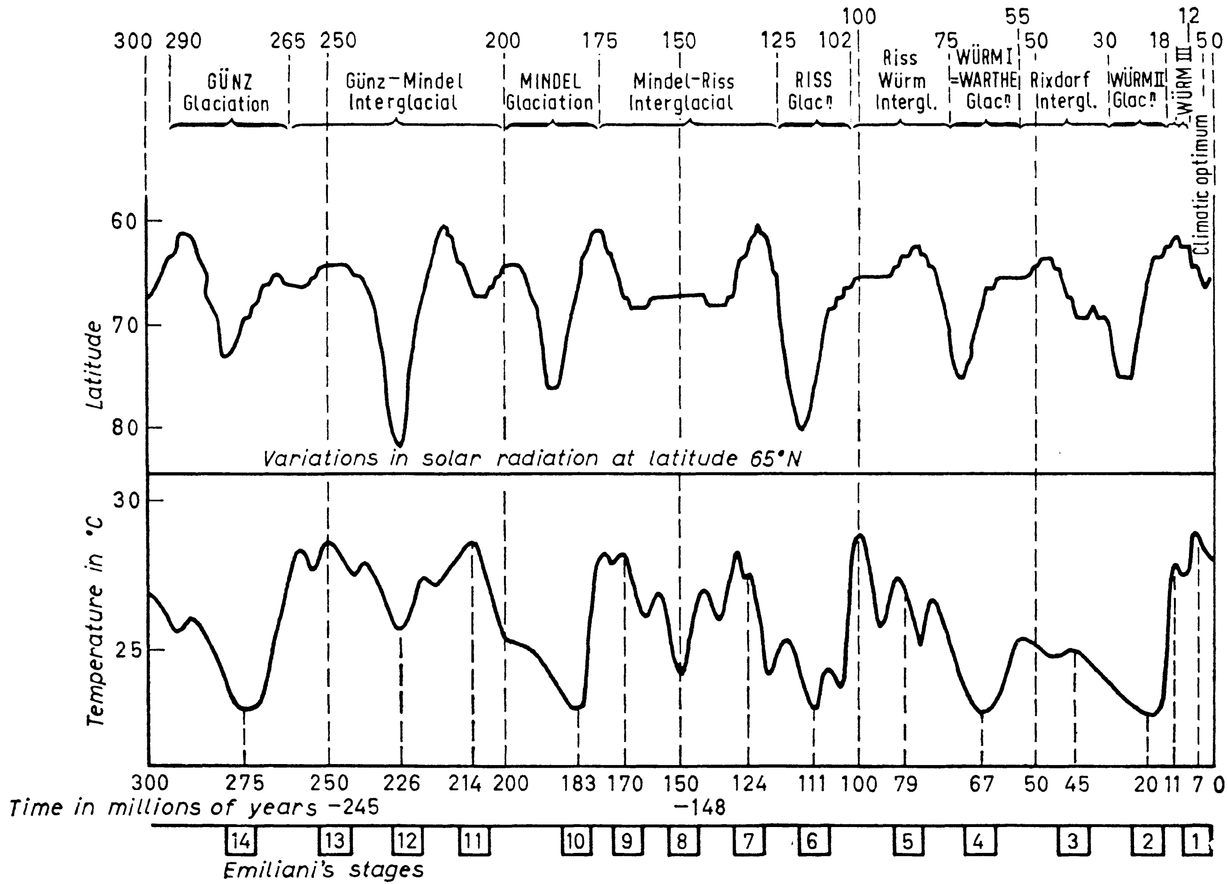


FIG. 1. PALEOTEMPERATURES DURING THE QUATERNARY (after C. Emiliani)

calculated temperature varied between 11·9° and 18° C. The accuracy of the analytic method is believed to be about ±1° C.

With regard to the Tertiary and Quaternary era, the work of Emiliani (1955 and 1958) on benthonic and pelagic foraminifers, has resulted in a useful series of temperatures back to 65 million years B.P. It may be noted that an enormous effort must be made to extract collections of *single* species from among the foraminifers of all genera in an average core of benthonic species required for the preparation of each sample of 5 milligrammes from which the isotopic ratio is measurable. The benthonic foraminifers obtained from central Pacific deep sea cores, indicate that the temperature of the bottom water was nearly 12° C. in Oligocene time, 35 million years ago. Subsequently it gradually became cooler, being 6·5° C. in the Miocene, 22 million years ago, and falling to 2·5° C. in the last 500,000 years. From the deep ocean sediment coming from the Arctic Ocean it can be said that the temperature of the polar seas in Oligocene times was approximately the same as off the Atlantic coast of Morocco today at a depth of 1,300 feet. This is confirmed by the presence of Oligocene pine and fir forests in Greenland. The ice cap there was very small if not completely absent.

By the same method, the *pelagic* foraminifers from the same beds show the fluctuations in *surface* water temperature. The calculated changes can be recorded throughout the oceans. These measurements, applied to the Quaternary, have been produced at the same time as the dating of corresponding beds by radiocarbon. These go back to about 70,000 years. The dating of older cycles has been made by the extrapolation and correlation of the sedimentary changes with phases of retreat and advance of the glaciers, a method which may be effective as far back as 1,000,000 years. However, no direct correlation has yet been possible between the older glacial phases and specific marine oscillations, and further isotopic dating (e.g. by thorium-protactinium ratio) is being developed.

Climate and Modification of the Earth's Magnetic Field.—The magnetic axis of the earth can be likened to a powerful dipole (bar magnet) included in the core of the earth. As the result of the gyroscopic effect of rotation, this axis today coincides to within about 25° of the geographical poles. There is no doubt that during geological time, the magnetic field of the earth has fluctuated.

To discover these variations and in particular the position of the poles at each period, *thermoremanent magnetism* (T.R.M.) of the grains of iron and titanium oxides in volcanic rocks has been used. This magnetism was acquired during the cooling from a higher temperature through a characteristic critical point for each mineral (Curie point).[1]

At ordinary temperatures the same oxides can also obtain an *isothermal remanent magnetism* (I.R.M.) when they are subjected to prolonged exposure

[1] The Curie point is 575° C. for pure magnetite and 675° C. for hematite; it alters in the solid solutions of magnetite (Fe_3O_4) and ulvo-spinel (Fe_2TiO_4).

to a magnetic field such as the earth's. This type of remanent magnetism is particularly interesting in sedimentary rocks, but in lavas, it can falsify the measurement by superposing itself on the previous one.

Some other types of remanent magnetism can be envisaged elsewhere: by slow crystallization, by viscous remanent magnetism (V.R.M.), by remanent magnetic hysteresis (R.M.H.), by total (T.T.R.M.) or partial thermoremanent magnetism (P.T.R.M.) which are theoretically possible, but not actually demonstrated.

It has been observed that variations of the magnetic declination (of many degrees) have occurred continuously during the past 400 years and more. They are thought to be due to movements of currents produced in the earth's core. Even the magnetic center of the earth is eccentric, lying about 340 km. toward the west Pacific. It has been suggested that the core position fluctuates with a 40-year period.

Any series of superposed lava flows or sedimentary successions rich in iron oxides (red sandstones) have shown that the direction of the axial bipole has reversed itself with an irregular periodicity varying from a few hundreds of thousands of years to several million. For example, at the beginning of the Quaternary a compass needle would have pointed South.

Even if the axis of the dipole has not varied much since the Miocene, it seems to have shown a diversity of direction in earlier geological periods. By plotting a large number of samples of magnetic declinations (as well as *inclinations*), an approximate geographic pole is obtained. In this field the studies of British scientists (Blackett, Runcorn *et al.*) in particular, extend over rocks of all ages back to the Torridon Sandstone (Precambrian).

This leads us to consider the migration of the magnetic poles, the position of the poles, the relative equator on the earth's surface, and the arrangement of climatic patterns on the land.

According to Opdyke and Runcorn (1959), the North Pole would have been situated near the Hawaiian Islands in the Algonkian, and southwest of that position toward the center of the Pacific in the Cambrian. It then approached the Mariannas in the Silurian, returned toward the northeast in the Devonian, and became situated in Manchuria in the Pennsylvanian. The pole would have been to the west of Kamtchatka in the Permian; near the Sikhota Alin in the Trias; to the south of Timor in the Cretaceous[1]; in the Arctic Ocean to the west of the Isles of Long in the Eocene; and near its present-day position in the Miocene.

Such variations should have led to considerable climatic modifications in the geological formations. On the whole, a certain constancy of zonation can be established throughout the paleogeographic history of the globe. On the other hand, the earth presents a form which does not easily lend

[1] Note however that measurements on the Upper Cretaceous basalts of Madagascar indicate a magnetic axis coincident with that of the present day (A. Roche, L. Cattala and J. Boulanger, 1958).

itself to the displacement of the polar axis. It follows that the paleomagnetists envisage a revival of the theories of the continental drifting, without always accepting the displacements envisaged by Wegener. In their eyes, the magnetic axis would remain practically fixed. It is the crust and mantle of the earth which would have been displaced, and certain continents would have moved more than others. Australia, for instance, would have shifted considerably.

However, the continental positions thus envisaged do not always correspond with the known distribution of fossils, or with glacial evidence from that continent.

Within the last decade an alternative theory has developed from two sources: geophysical (Egyed), and oceanographic (Heezen). This is the theory of global expansion, which would have the continents spread apart as the earth gradually expanded, under geochemical differentiation. The mid-ocean ridges would thus represent "hot zones", marked by volcanoes where the cracks were constantly being healed by molten extrusions.

THE CLIMATE DURING THE QUATERNARY GLACIATIONS

1. Glaciated Regions

Observation of present-day mountain glaciers does not adequately explain the origin of all forms of terrain which can be seen in northern Europe, nor does the exploration of ice sheets in their present form permit observation of the effects which they have on the underlying rocks.

The Quaternary glaciations coincided with a special set of climatic conditions which do not occur on a comparable scale today.

Many writers imagine that the basic requirement for glaciation is a heavy rainfall which is higher than that of the preceding period: that is, *a glacial phase corresponds to a pluvial phase*. Temperature permitting, precipitation falls in the form of snow, or rain. But provided the snow does not remain powdery and stuck fast in névé and in ice, it seems important that the temperature should not be too low. (It may be noted that snowfall in the Antarctic takes place chiefly in summer.) Also, a lowering of wind speed has been found to favor the accumulation of ice. Without these conditions, no ice accumulates. This is observable in some of the coldest, blizzard-swept regions of the world, as at Verkhoyansk which has recorded a temperature of $-71{\cdot}3°$ C.

The general increase in atmospheric precipitation has been attributed to a number of terrestrial or cosmic causes. This is the so-called *Pluvial Theory* that is still widely quoted. As will appear below it is probably based upon a circular argument; ice advances lead to displacement of climatic belts, snow lines fall and heavy rains occur in previously drier latitudes—but *not* universally. Mean temperature lowering is the most significant factor in glacial advance. In Equatorial Zones it is now suspected that the pluvial phases were hot times and interpluvial (dry) phases were equivalent to the glacial epochs.

A slight cooling was sufficient to induce a glaciation; certainly a fall of 6° C. mean temperature corresponds to approximately 800 miles latitudinal displacement of the limit of the glaciers.

For the last glacial period (the Würm), L. Trévisan (1940) attributes a change of 4,000 feet in the level of the snow line (compared with the present day) in the Apennines, to a fall of about 6° C. in the mean summer temperature accompanied by an increase of about 55 inches in the annual precipitation. Similarly, L. C. W. Bonacina (1947) suggested that a fall of 1·7° C. in the mean temperature of Scotland would give rise to small glaciers on Ben Nevis, 4,440 feet high.

Such a change decreases the evaporation over land and sea, increases the percentage of snow in the precipitation, and lowers the snow line. The glacier itself, once formed, accentuates these results by cooling the air around it. In summer, the sun's heat is lost in the melting of the ice. Thus, a minor initial cause would seem to be responsible for the great effects produced by the glaciations.

The lowering of sea level is of primary importance. It leads to the breakup of oceans into segments, particularly in the north. There is also a reduced exchange of waters between the arctic and temperate seas over the Wyville-Thomson ridge, a submarine barrier which links the Shetlands with the Faroes (Ewing and Donn, 1958). Such a lowering of sea level also contributes to a general cooling.

In the present day, cold and heavy air accumulates above the ice sheets, forming *glacial anticyclones*, such as occur over Greenland and Antarctica. There they produce centrifugal winds, in fact strong blizzards, which sweep snow as far as 200 miles beyond the limits of the ice. These anticyclones also cause pack ice to drift over the sea. Yet the center of the ice-sheet is marked by calms or moderate winds. The form of the anticyclones is little known; it changes seasonally while at the same time the anticyclones are moving about. Their origin would be controlled by latitude, the low thermal capacity, and the conductivity and great reflecting power of snow. The combined action of these varied properties gives rise to extraordinarily low temperatures (−86° C.) in the center of Greenland and the Antarctic while the temperature is near 0° C. on their coastal fringes.

During the Quaternary, an anticyclonic zone dominating the Scandinavian ice sheet was responsible for the periglacial eolian features of this region, including the deposition of loess. The moist winds, coming from the Atlantic, covered Europe with snow. In eastern Europe the snow line swung northward; further south, the same Atlantic winds brought rain at all seasons, supplying the glaciers of the Apennines, the Balkans, the Caucasus, and Asia Minor. Near the edge of the ice sheet, the anticyclone would have become weaker so that the dry northeasterly winds could blow out beneath

TABLE I.—CHRONOLOGY AND CLIMATES OF THE QUATERNARY IN EUROPE

N.B.—From —290,000 the second column follows the short chronology, after C. EMILIANI (1955).

GLACIAL AND INTERGLACIAL PHASES IN NORTH AMERICA	AGES IN YEARS	INDICES OF PALEOTEMPERATURE	GLACIAL AND INTERGLACIAL PHASES, EUROPE: NORTH GERMANY AND EUROPE	GLACIAL AND INTERGLACIAL PHASES, EUROPE: ALPINE CHAIN (after PENCK and BRUCKNER)	INDUSTRIES	MOVEMENTS OF SEAS (MEDITERRANEAN)	ALTITUDES OF MARINE TERRACES	PHASES IN THE PARATETHYS: BLACK SEA	PHASES IN THE PARATETHYS: ARALO-CASPIAN BASIN
	The Present 0 +600				Aurignacian etc.	**Flandrian** transgression = Versilian	0 +6 ft to 10 ft	modern expansion	regression of Novo-Caspian
WISCONSIN: **CARY** **TAZEWELL**	—8,000 —10,000 —12,000 —15,000	see Table II	**POMERANIAN MASURIAN**	**WURM III**					
WISCONSIN: **FARMDALE** **IOWAN**	—18,000 —20,000 —28,000	minimum insolation	**WEICHSELIAN**	**WURM II**	Mousterian	pre-Flandrian transgression = Grimaldian	—330 ft to —295 ft	regression of New Pont-Euxin	Khvalyaskian = 2nd pluvial (expansion)
	—35,000 —45,000 —53,000	maximum insolation cold to temperate climate	**RIXDORF**			**Neotyrrhenian** Tyrrhenian transgression III = Normanian	+15 ft	Karangat transgression	1st Interpluvial = Atel regression
	—60,000 —67,000 —73,000	minimum insolation glacial climate	**WARTHE**	**WURM I (= RISS 2)**		Nomentian II slow "chellean" transgression (Tyrrhenian transgression II = "Monastirian")	+65 ft		
SANGAMON	—79,000 —85,000 —101,000	maximum insolation	**EEMIAN**	**Riss-Wurm Interglacial**		Tyrrhenian transgression I **Eutyrrhenian**	+100 ft	Uzunlar transgression	KHAZARS: Upper Khasarian (expansion)
ILLINOIAN	—110,000 —111,000	minimum insolation	**SAALE**	**RISS**	Acheulian	Post-Milazzian regression (Nomentian of Italy)		regression	KHAZARS: Lower Khasarian (expansion)
YARMOUTH	—125,000 —135,000 —138,000 —148,000 —160,000 —170,000 —175,000	maximum insolation minimum insolation maximum insolation		**Mindel-Riss Interglacial**		"Milazzian" transgression = Sicilian II **Paleotyrrhenian**	+210 ft to +180 ft	Early Pont-Euxin transgression (brackish)	Urundzhik regression = Upper Backinian [=Likhvin]
KANSAN	—183,000 —185,000 —200,000	minimum insolation	**ELSTER**	**MINDEL**	Clacto-Abbevillian	regression (Flaminian of Italy)			Lower Bakinian expansion = Apscheronian transgression Pre-Apscheronian
AFTONIAN	—210,000 —214,000 —226,000 —235,000 —245,000	maximum insolation minimum insolation maximum insolation	**CROMERIAN**	**Gunz-Mindel Interglacial**		Sicilian transgression I = Calabrian II = Upper Villafranchian Post-Calabrian regression Emillian transgression	+330 ft to +260 ft —330 ft —65 ft to 100 ft		Aktchaghylian transgressions
NEBRASKAN	—265,000 —275,000	minimum insolation		**GUNZ**		Calabrian regression I = Lower Villafranchian	—130 ft	Tschaudian transgres-sion (brackish)	
	—290,000							Kujalnik	
SIERRIAN	—1,100,000					**Acquatraversian**			Dacian
	—1,200,000					**ASTIAN = PLAISANCIAN**			

TABLE II.—SUCCESSION OF CLIMATIC AND HUMAN EVENTS OVER THE LAST 20,000 YEARS

AGE IN YEARS B.P.*	PALEO-TEMPERATURES (DEPARTURES)	EUROPE	NORTH AMERICA
0	0	Present Time	
—600	—1° C	Mediaeval Cold Phase	
—1,000........	*+0·5°*	**DUNKERQUIAN III**	
—1,600........	*+1°*	**DUNKERQUIAN II** (*Mya*) Roman Cold Phase	FLORIDA Emergence
—2,000 A.D. ... B.C. ...	—0·5°		
—2,300........	*+1°* "CLIMATIC OPTIMUM"	**DUNKERQUIAN I** (*Limnaea*)	*Late Silver Bluff*
—3,700........	*+2°*	**CALAISIAN II** (*Late Littorina*) **FLANDRIAN**	*Middle Silver Bluff*
—5,500........	*+2.5°*	**CALAISIAN I** **MAXIMUM** (*Middle Littorina*)	*Early Silver Bluff*
—7,000........	*+2°*	**OSTENDIAN II** ***Neolithic*** (*Early Littorina*)	*Late Champlain*
—8,300	—0.5°	Bothnian Glacial (Ancylus—Baltic—Lake)	COCHRANE Glacial
—9,000........	*+1°*	**OSTENDIAN I Mesolithic** (*Yoldia; de Geer Zero Varve,* 6839 *B.C.*)	*Early Champlain*
—10,500........	—3°	FENNOSCANDIAN GLACIAL (Salpausselkä)	VALDERS Glacial
—11,000........	—2°	**ALLEROD**	*Two Creeks*
—12,500........	—4°	VELGAST GLACIAL	PORT HURON Glacial
—13,000........	—3°	**BOLLING**	*Lake Arkona*
—15,000........	—7°	GOTIGLACIAL	CARY Glacial
—16,000........	—6°	**MASURIAN**	*Brady*
—18,000........	—8°	DANIGLACIAL	TAZEWELL Glacial

* Before Present. Warmer phrases are shown in italic type; cooler phrases in roman type.

the wet southwesterly winds which were drawn in. The former were carrying the loess which was deposited in northern Germany.

There is a parallel history for the northern part of America.

The increase in the thickness of the Greenland ice cap is equivalent to 14 inches of water per year. It is particularly important on the edge of the ice sheets where precipitation is more abundant. The origin of this snow would be from cyclones, marginal to the anticyclones, especially during the seasonal displacement of the latter. The hypotheses vary considerably among writers, and it remains for meteorological conditions to become better known (Charlesworth, 1957, pp. 640–680.)

It would seem that a large accumulation of ice can only occur on terra firma. This is so in the Antarctic and, in the northern hemisphere, in Greenland, which has provided a base sufficiently solid for glaciers. Nearer the pole, the Arctic ocean is covered by a layer of ice only a few feet thick (*Nautilus* expedition, 1958.)

Ice sheets terminate at ocean margins and at plains with a semiarid climate. Also in anticyclonic regions the ice tends to be worn away. The topography underlying the glaciers also modifies their extent and their ideal form, giving rise to side branches and eccentric glaciers.

HYPOTHESES.—A large number of hypotheses have been developed to explain the fundamental causes of the Quaternary glaciations. Some hypotheses are based upon a suggested alternation of the glaciations between the northern and the southern hemispheres. It seems likely, on the contrary, that the glacial phases of the two hemispheres were essentially contemporaneous. Certain people assume that the present is a "postglacial", i.e. nonglacial, epoch; this is quite incorrect, for there are about 30 million cubic km. of ice still locked up in Antarctica and Greenland. In "normal" geological epochs there were probably no great ice caps. It is very likely that we are in a mild interglacial period, with the prospect of a new glaciation occurring within the next few tens of thousands of years. However, the Würm glaciation is sometimes considered to be weaker than its predecessors, so it is possible that the Quaternary glaciations are in their decline at this stage. However, Emiliani's curve (fig. 1) suggests that each glacial maximum reached an approximately equal amplitude. Charlesworth (1957) classifies hypotheses for the causes of glaciation in five groups:

1. The *topographical* hypotheses, according to which there has been a general uplift of the continents, and raising of their height in relation to the geoid. In fact, a universal epeirogeny occurred during the late Precambrian and the Carboniferous (p. 16). An uplift of the Pleistocene seems also to have been a planetary phenomenon. It was accompanied by a world-wide lowering of sea level ("tectono-eustatic").

2. The *geophysical* hypotheses, related to the displacement of the Poles. According to an old hypothesis very rapid polar changes occurred so that the glaciations of North America would correspond with the interglacials

of Europe and vice versa. This view cannot be upheld, for there is in fact a correlation between the glaciations as well as between the interglacials from one side of the Atlantic Ocean to the other. The rapid displacement of continents has also been invoked (Wegener). On the other hand, in recent years, paleomagnetic surveys suggest a *slow* polar change that took the South Pole

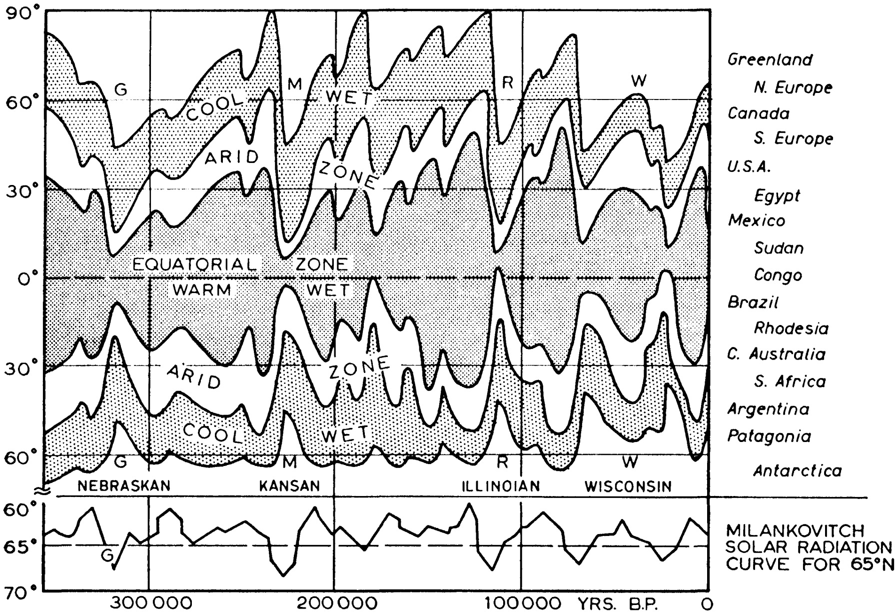


FIG. 2. RECONSTRUCTION OF WORLD CLIMATIC BELTS FOR THE QUATERNARY, BASED UPON THE ASSUMPTION THAT HIGH SOLAR RADIATION WILL INCREASE THE EVAPORATION RATE, BROADEN THE EQUATORIAL AND ARID ZONES, AND SHRINK THE ARCTIC (GLACIAL) REGIONS. GREATLY REDUCED REACTIONS ARE VISUALIZED FOR ANTARCTICA, WHICH IS BELIEVED TO HAVE REMAINED ICE-COVERED, EVEN DURING THE INTERGLACIAL PHASES

A hypothetical correlation is suggested using the Emiliani interpretation of the Milankovitch solar radiation curve (the "short Pleistocene"); the same principle would be involved if the Soergel ("long Pleistocene") correlation is used. (Graph prepared by R. W. Fairbridge.)

into the Antarctic continent in the late Tertiary. This played a major role in developing our greatest glaciated continent.

3. The *atmospheric* hypotheses. It has been thought that the glaciations are related to the amount of carbon dioxide in the atmosphere (from the burning of industrial hydrocarbons), or to volcanic dust. Both have some influence on the world climate (instrumentally recorded), but at the present juncture it is thought to be minor.

4. The *planetary* hypotheses, related to variations in the position of the earth's axis with respect to the ecliptic or to that of solar radiation received by the earth. Such hypotheses are developed with the knowledge that the

four cardinal points show a gradual migration during 20,700 years, and that in about 91,800 years the eccentricity of the earth's orbit varies from 0·00331 to 0·0778. Most important, perhaps, is the cycle of the obliquity of the earth's ecliptic (41,000 years). The Yugoslav astronomer-mathematician Milankovitch (1920, 1940), in a series of elegant calculations, worked out the effective radiation received ("insolation") for any latitude on the globe. The curve for 65° N. is probably most important because this is the mean latitude of maximum ice development during the Pleistocene (excluding Antarctica, which does not vary greatly from "glacial" to "interglacial", receiving as it does very little insolation at such high latitudes). The Milankovitch curve indicated about seventeen peaks over some

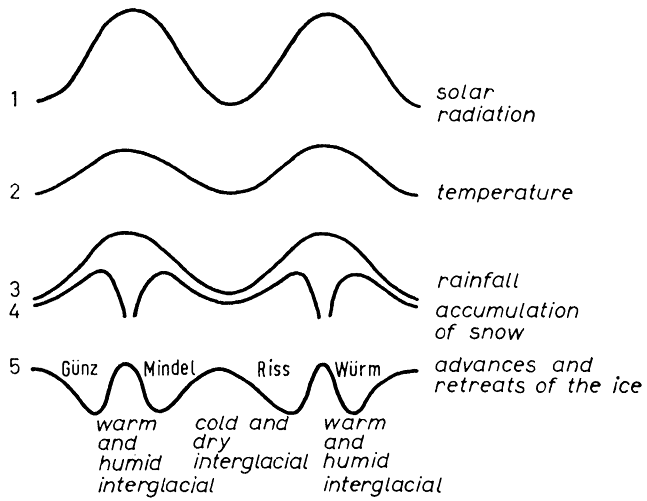


Fig. 3. The Causes of the Quaternary Glaciations according to the Hypothesis of G. C. Simpson

600,000 years; however, ice masses once developed melt very slowly, and the Hungarian mathematician, Bacsák (1955), calculated that most glacial stages would embrace at least two or even three of these oscillations. The geological record certainly indicates two or three peaks for the great glacial phases. However, the suggested correlations that have been made with Alpine and continental ice advances are still hypothetical, and based upon numerical coincidences.

5. The *cosmic* hypotheses. Among these, the hypothesis of G. C. Simpson (1938) is the most interesting because it groups together on five corresponding curves (see fig. 3) two supposed cycles of solar radiation, the temperature, the rainfall, the accumulation of snow, and finally, the real fluctuations of the glacial periods. The last of these curves, we now know, was grossly oversimplified by Simpson, but the principle suggested may well apply to multiple radiation peaks. Recent observations from rockets and satellites

suggest very little variation in the "Solar Constant", but they do show up to 200% variability in the ultra-violet emissions, and these control the production of ozone in the outer atmosphere. Ozone contributes to the earth's "thermal blanket", which controls our effective heat budget (Fairbridge, 1961).

By applying the O^{18} isotope method to tropical pelagic foraminifers from deep sea cores, Emiliani distinguished forty points of paleotemperature over a period estimated to be about 300,000 years. These stages are com-

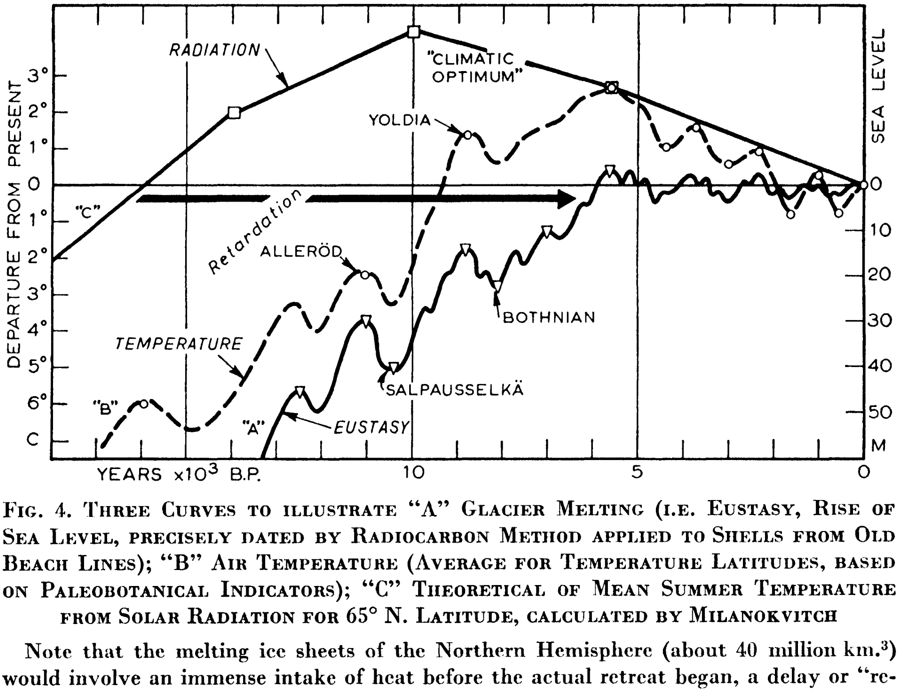


FIG. 4. THREE CURVES TO ILLUSTRATE "A" GLACIER MELTING (I.E. EUSTASY, RISE OF SEA LEVEL, PRECISELY DATED BY RADIOCARBON METHOD APPLIED TO SHELLS FROM OLD BEACH LINES); "B" AIR TEMPERATURE (AVERAGE FOR TEMPERATURE LATITUDES, BASED ON PALEOBOTANICAL INDICATORS); "C" THEORETICAL OF MEAN SUMMER TEMPERATURE FROM SOLAR RADIATION FOR 65° N. LATITUDE, CALCULATED BY MILANOKVITCH

Note that the melting ice sheets of the Northern Hemisphere (about 40 million km.³) would involve an immense intake of heat before the actual retreat began, a delay or "retardation" calculated to be about 10,000 years. (Graphs prepared by R. W. Fairbridge.)

pared with the Milankovitch curves of maximum and minimum insolation. A delay of about 5,000 years must be allowed between maximum insolation and the rise in temperature which it initiates. An even longer delay (c. 10,000 years retardation) must be allowed for melting, because of the latent heat factor and other delay controls (Fairbridge, 1962) (fig .4.)

The geologist is aware that the Quaternary glaciations are not the only glaciations in the earth's history. Two other apparently universal glaciations are well known: these are the late *Precambrian glaciation* and the *Carboniferous glaciation*. The late Precambrian or "Eocambrian" glaciation was a little more than 600 million years ago, the Carboniferous glaciation about 240 to 215 million years ago, whereas the Plio-Pleistocene glaciation began more than 1 million years ago (being initiated in Antarctica and Greenland).

Minor glacial stages are noted in South America and South Africa during the middle Paleozoic. If the three great glaciations alone are considered, it may be noted that they are not separated by regular intervals, although common geological traits can be observed among them. They have all followed great continental periods, marked by the abundance of detrital terrestrial sediments, which are often red in color and suggest an arid or semiarid climate. Moreover, they added to this continental sedimentary record, and were then followed by marine deposition, beginning with recognizable transgressions. If the help of geology is invoked in the search for the causes of glaciations, there is room for inquiry into the causes of such "continentalization", which follows important orogenies (the Precambrian, the Variscan and the Alpine) but is separated from them by heavy erosional phases represented by conglomeratic deposits of "molasse" type. These phases seem to indicate positive epeirogenic tendencies. Thus, the "topographic hypothesis" seems worth considering. Although it is rarely favored by European writers, it postulates a relative rise and fall of the effective continental land areas. Perhaps such rise and fall is related to phases of expansion of our planet (the Egyed-Heezen theory, p. 8).

TABLE III.—THE SEQUENCE OF EVENTS IN THE THREE GREAT GLACIAL PERIODS OF GEOLOGICAL TIME

SUCCESSION OF EVENTS	EOCAMBRIAN GLACIATION	SAKMARIAN GLACIATION	PLEISTOCENE GLACIATION
VI *Eustatic transgression*	**LOWER CAMBRIAN**	**WORDIAN**	**FLANDRIAN**
V *Playas and Salt Deposits*	Ex. Salt Range	Zechstein of Europe, Russia, Greenland. Salt of U.S.A.	Ex. Lake Eyre, Great Salt Lake
IV *Icesheet*	Scandinavia, Greenland, Ghana, Togoland, Congo	Australia, India (Talchir), Africa (Dwyka) to **SAKMARIAN**	Scandinavia, Canada, Part of Southern Hemisphere
III *Continentalization* II *Morphogenesis, Molasse*	**UPPER INFRACAMBRIAN** (*s.l.*)	**STEPHANIAN NEW RED SANDSTONE**	**VILLAFRANCHIAN and PLIOCENE**
I *Orogenesis*	**Assyntic** of Europe **Penokean** of America **Ketilides** of Scandinavia	**Variscan** of Europe **Moghrabine** of North Africa **Altaides** of Asia, etc.	**Alpine**

2. *Pluvial Regions*

During each glaciation, the areas peripheral to the ice fronts received heavy precipitation, not in the form of snow but as rain. Boundaries between glacial zones and pluvial zones are in some places very clear. This is so in Utah, where ancient Lake Bonneville is of pluvial origin. Large numbers of extensive old pluvial areas have been traced all through the arid southwest of the United States.

The pluvial period is characterized by swollen rivers and lakes. This becomes an important phase of continental erosion and of alluvial deposition in the oceans and intermontane basins.

The climates of those zones which today are arid have undergone the greatest changes. Indeed it is likely that there was no time free from deserts during the Pleistocene, but there were certainly times of amelioration of the climates of the present arid regions. This is supported by the enormous extension of endorheic basins during certain stages of the Pleistocene, compared with what they have become today.

There is a meteorological problem, that has not yet been solved, about the climate of nonglacial regions during ice ages. According to Flohn (1952) the cooler world climate would cause a 20% reduction in evaporation from the critical regions of the oceans, and moreover, the eustatic withdrawal from the continental shelves would reduce the area of effective evaporation by about 5%. We might expect a generally more arid period to coincide with the coldest phases.

But there is another factor. The glacial reduction in the width of climatic belts would increase the temperature gradient from equator to ice margin, thus greatly accelerating mean wind velocities. The stronger winds would increase the evaporation rate, and might make the glacial stages more rainy. It is a paradox.

Although the matter is not finally resolved it seems that both factors have some application. In the climatic zone of strong westerly winds just on the equatorial side of the ice front, there is geological evidence of heavy pluviation at some stages of each glacial phase. There is evidence of strong winds (large dunes) as well as increased rainfall. In Europe the ice advance to Central Europe pushed the zone of westerlies down to the Mediterranean, and the north shore of Africa received increased rains. In North America, the strong westerlies reached across California to Arizona, New Mexico and Texas.

Thus along the northern borders of dry or desert lands in the northern hemisphere the climate became much wetter at times during the ice phases (e.g. Spain, Italy, North Africa), but the deserts themselves were not eliminated. The dry zones may even have expanded at certain stages; they certainly advanced equatorward and in Africa ice-age dunes are to be traced from the Sahara and the Sudan, pushing right down to the northern

TABLE IV.—APPROXIMATE CORRELATION OF THE CLIMATIC STAGES OF THE QUATERNARY IN THE ARID AND SEMIARID ZONES OF AFRICA

N.B.—In East Africa it is customary to compare the Pluvials with Glacials, but recent C^{14} dating suggests that the tropical pluvials are high radiation periods, and the dry periods in low latitudes are glacial equivalents. Because Moroccan chronology has been more complete for the beginning of the Pleistocene, it has been necessary to allow for an older pluvial than the one which had been accepted as the first.

<table>
<tr><th>EUROPEAN PHASES</th><th>PLUVIAL PHASES OF MOROCCO</th><th>PLUVIAL PHASES IN E. AFRICA</th><th>PLUVIAL AND ARID PHASES IN SAHARA</th></tr>
<tr><td rowspan="2">FLANDRIAN</td><td rowspan="2">RHARBIAN</td><td rowspan="2">"Mid-Recent pluvial" (c. 5,000 yr. B.P.)</td><td>Present arid phase</td></tr>
<tr><td>GUIRIAN (Neolithic)</td></tr>
<tr><td>WURM III</td><td rowspan="2">SOLTANIAN = 4th pluvial</td><td rowspan="2"></td><td rowspan="2">post-Saourian arid phase</td></tr>
<tr><td>WURM II</td></tr>
<tr><td>WURM I = RISS II</td><td rowspan="2">OULJIAN = 3rd Interpluvial</td><td rowspan="2">GAMBLIAN = 4th pluvial</td><td rowspan="2">3rd pluvial = SAOURIAN (Arterian)</td></tr>
<tr><td>Riss-Wurm Interglacial</td></tr>
<tr><td>RISS</td><td>TENSIFTIAN = 3rd pluvial</td><td>3rd Interpluvial</td><td>post-Ougartian arid phase (Acheulean and Aterian)</td></tr>
<tr><td>Mindel-Riss Interglacial</td><td></td><td>KANJERAN = 3rd pluvial</td><td>2nd pluvial = OUGARTIAN Acheulean (End of Pebble Culture)</td></tr>
<tr><td>MINDEL</td><td>AMIRIAN = 2nd pluvial</td><td>2nd Interpluvial</td><td>post-Mazzerian arid phase</td></tr>
<tr><td>Gunz-Mindel Interglacial</td><td>2nd Interpluvial</td><td>KAMASIAN = Great pluvial</td><td>1st pluvial = MAZZERIAN (Pebble Culture)</td></tr>
<tr><td>GUNZ II</td><td>SALETIAN = 1st pluvial</td><td rowspan="2">Early pluvial</td><td rowspan="3">post-"Villafranchian" arid phase</td></tr>
<tr><td></td><td>1st Interpluvial</td></tr>
<tr><td>GUNZ I</td><td>MOULOUYAN = post-Pliocene regression = early pluvial</td><td rowspan="2">KAGERIAN 1st pluvial Oldowayan</td></tr>
<tr><td>CALABRIAN</td><td>MOGHREBIAN transgression</td><td>"VILLAFRANCHIAN" Beginning of Pebble Culture</td></tr>
</table>

borders of the Congo. From the other side, the sands originally coming from the Kalahari, pushed all the way up across Angola pressing back the jungle far into the Congo. The same thing is seen in Australia and parts of South America.

During interglacial periods, the opposite probably happened. Increased radiation increased the tropical rainfall, the tropical jungles spread out north and south, revegetating the sands and "fixing" the dunes. The desert, for its part, moved poleward and the north African coastal lands became drier. The westerlies moved poleward too, and the mild temperate rains returned to northern Europe and the northern region of the United States and Canada.

This is the interpretation first enunciated by the great geographer Albrecht Penck. It has, however, been commonly displaced by another theory, viz. that increased rainfall occurred the world over during glacials and that interglacials were universally dry. Accurate dating of "pluvial" formations in Africa and elsewhere (by radiocarbon) is beginning to disprove the "glacial = pluvial theory", but there is still much to be learned in this field.

The question of whether pluvial periods were contemporary with the Eocambrian and Carboniferous glaciations should be considered. The existence of such pluvial periods is hypothetical and difficult to support on paleontological grounds. However, lakes and marshes are known in the northern hemisphere which were contemporaneous with the Stephanian and Sakmarian glaciers of the southern hemisphere. These are the limnic basins where the coal of the Upper Pennsylvanian was deposited during a general regression of the sea. These limnic basins maintained a flora with a Carboniferous character in a climate which remained *locally* comparable, yet within the framework of a changing zonation (H. and G. Termier, 1958).

3. *Humid Tropics and Equatorial Zones*

N.B.—Based on theory of "glacials = pluvials"; note possible alternatives.

The alternating climates of the Pleistocene also left their mark on the hot and humid regions. It will be seen that the *tropical humid climate* is characterized by an alternation of one dry and one wet season. This rhythm, sometimes accentuated by the monsoon, favors the formation of laterite, a thick red soil (see Chap. 6). In contrast, the *equatorial climate* which is constantly hot and humid tends to remove the soils by gullying.

During pluvial times the zone of alternating seasons favored laterization, while during the interpluvials there was an even greater tendency toward dryness than there is today. During the interpluvial periods, the equatorial zone experienced only a slightly lower temperature, approaching that of the tropical humid climate.

Table V shows a comparison between the Quaternary history of Guinea and Guiana. These two regions of similar latitude, but somewhat different climates, illustrate the different tendencies described in the preceding paragraph. However, they are both based upon the assumption that "glacials = pluvials". It may be alternatively given as "glacials = inter-pluvials".

Using the second interpretation, it may be seen that an identical process (laterization) may occur at quite different times. *Guinea* (West Africa) has in fact a tropical climate, still humid in places where the forest vegetation is progressively giving way to savanna. These conditions may suggest the end of the Holocene pluvial period, and similar cool dessication in the past

TABLE V.—COMPARISON BETWEEN THE QUATERNARY HISTORY OF GUINEA AND GUIANA

WORLD-WIDE CHRONOLOGY	GUINEA (after Maignien, 1958)		GUIANA (after B. Choubert, 1957)		
Age in years	CLIMATE	SOIL	CLIMATE	SOIL	ALTITUDE OF COASTS
0 **Present epoch** —2,000	tropical (savanna)	hardpan	equatorial	erosion	0
Climatic optimum —5,000	Sudan-Guinean humid (forest) tropical (savanna) Sudan-Guinean humid Sudan—dry	erosion hardpan slow ferrallitization hardpan	equatorial	erosion	7 feet
WURM II **Ouljian** **WURM I**	tropical alternately dry and humid	ferrallitization	tropical equatorial tropical	ferrallitization erosion ferrallitization	16 feet
Tyrrhenian II **Riss-Wurm Interglacial**	Sahelian—dry	hardpan	equatorial	erosion	32 feet
			Period of faulting		
Tyrrhenian I			tropical equatorial	ferrallitization erosion	65 feet
RISS	equatorial (forest)	erosion and laterization	tropical	ferrallitization	
Mindel-Riss Interglacial	semiarid	hardpan	equatorial	erosion	115 feet
MINDEL	warm and humid (forest)	ferrallitization	tropical	ferrallitization	120 feet to 180 feet
Gunz-Mindel Interglacial	warm and dry	pediplanation	equatorial	erosion	236 feet
GUNZ			tropical	ferrallitization	256 feet to 318 feet
LOWER VILLAFRANCHIAN			equatorial	erosion	395 feet

resulted in the formation of "hard pan". During the pluvial periods, its tropical climate was suitable for laterization. Guiana (South America) is today covered with equatorial forest and subject to a very hot and humid climate. Its soil suffers heavy erosion, despite the presence of vegetation. During the interpluvial periods, the cooler climate led to intense laterization as in the tropical humid climate.

4. Marine Terraces, Eustasy and "Glacial Control"

The glacial phases, as previously stated, had profound effects upon the oceans. Terraces and beaches progressively emerged during the Quaternary, and indicate that the altitude of mean sea level has varied considerably. This relative level was changed not only by local deformations of the earth's crust (orogenic movements and epeirogenic warping), but also by other causes. E. Suess (II, p. 841; III, pp. 1669 and 1673) was one of the first to recognize that modifications could also be brought about by changes of the level of the sea itself. These changes had a general effect upon the whole globe and he called them *eustatic movements*.

Sea level plays a part of major importance in sculpturing the land, because it is the chief base level, below which, atmospheric erosion only makes itself felt in very local areas (for example to the north of the Caspian Sea). These oscillations can have very important consequences. Precise measurements which have been made recently, show that mean sea level (formerly considered as a stable datum) undergoes measurable variations at the present time. According to the average of tide-gauge records throughout the world, sea level rose at a mean rate of 1·2 mm. per year from 1890 to 1950.

The study of eustasy or "eustatism" as noted by R. W. Fairbridge (1948) must take place on continents with a stable coast line, not those affected by recent orogenic movements. This author thinks that it would be desirable to select as a datum for this information the *level of low tides which is considered to be the actual base level of subaerial erosion*, and which coincides in Australia, and on coral islands of the tropical oceans, with the littoral terrace cut by the sea in the reef or eolian limestones.

Several types of eustasy can be visualized:

(*a*) One would be the result of deformation of the ocean floor due to "bathygenic" movements (tectono-eustasy). This phenomenon was visualized by Charles Darwin a century and a half ago to explain the "drowning" of central Pacific atolls.

(*b*) It might also be thought that eustasy is dependent upon the equilibrium of erosion and sedimentation. This concept has two complementary aspects: 1. For C. Arambourg (1952) and A. Cailleux (1952), erosion, so important on coasts as well as inland, has resulted in isostatic uplift of continents as they are progressively lightened. This causes the shore lines to retreat more and more. Consequently, the highest resulting terraces are the oldest. 2. For Fairbridge (1948), increasing sedimentation on the ocean floor causes sinking. This is a phenomenon which is often encountered: the elevation of areas of sialic crust and the subsidence of areas of simatic crust. The filling of the ocean by sediments has also actually raised the level of the sea, regardless of tectonic movements (the theory of Suess).

(*c*) Glacial eustasy is particularly well known. In 1837, Agassiz noted that the ice which accumulated during a glacial period locked up the water so that it became a "rock" on terra firma, and in 1842, McLaren concluded that the melting of this ice must have raised the level of the seas.

In 1912–1913 this hypothesis was put forward by F. B. Taylor as a glacial-eustatic theory. According to this concept, the last negative glacial-eustatic movement of the Quaternary would have lowered the level of the seas by 330 feet (E. Antevs, 1928). This was contemporaneous with the Würm glaciation as Baulig (1925) has shown with regard to the basin of the Durance. Such a eustatic lowering of sea level resulted in a world-wide marine regression. During the transgression following the release of the glacial melt waters, it has been suggested that the sea would have "drowned" the area it invaded, with erosion of salient features and with the filling of hollows.

R. A. Daly (1934) introduced the term "glacial control" for the geological consequences arising from variations of glacial-eustatism. It has been used to explain the history of coral reefs (particularly in the South Pacific and Indian Oceans) and the drowning of the greater part of the world's coast lines. This submergence followed a notable period of surface sculpturing which has become hidden beneath the sea (see pp. 48–55).

It now seems well established that each glacial phase of the Quaternary corresponds with a regression of the sea, and that each interglacial phase corresponds with a transgression. This correlation seems sufficient to prove glacio-eustasy, as the regressions correspond with the ice advances. But it is not as simple as this, because the effect of epeirogenic and orogenic movements is superposed upon the eustatic effects. It is difficult to separate these factors and this difficulty has led to numerous discussions by specialist workers.

"Glacial control" can be demonstrated in two ways:

1. By the calculation of the quantity of ice locked up in the ice sheets and in the great mountain glaciers, since their approximate area can be closely estimated. But this procedure should not play too large a part in any hypothesis since the exact thickness of the ice is not known. Today's total ice volume is at least 30 million km.^{3}, and the Würm maximum at least 70 million km.^{3}

2. It might appear easier to measure the variations in sea level observed in the Quaternary formations. Melting some 40 million km.^{3} of ice would raise sea level 330 feet. Or expressed another way 1 mm. change of sea level equals 400 km.^{3} of ice added or removed. Thus at the two extremes: the

pre-Flandrian regression indicates a level of at least —330 feet, while the Sicilian I transgression indicates a level of at least +330 feet. The sum of these absolute values would be 660 feet. This would correspond with the difference between a maximum storage of water in the form of ice and the minimum storage, the extreme case with practically no ice at the poles. This would suggest that the present epoch lies midway between these two extreme cases. It is difficult to concede that it is a question of eustatism alone, for it is known that there were epeirogenic movements during the Quaternary. It might be that the height of 300 and possibly 600 feet for the Sicilian I partly represents the mean uplift experienced by the continents during the Quaternary Era. A considerable time elapsed from the Sicilian "high" to the Würm "low", during which slow tectono-eustasy could operate.

When they talk of glacial-eustatism, geomorphologists and geologists always consider those events which have accompanied and followed the Pleistocene glaciations. In fact if such a phenomenon occurred during the Quaternary, it must be recognized that similar causes would have produced similar effects in more ancient times. Accordingly, the Precambrian and Paleozoic glaciations of geological history must have had glacial-eustatic effects.

The continental shelf, which constitutes a well-marked geomorphological entity, is the next question to be considered (see pp. 48–55).

If a corresponding body of water were removed from the present oceans, all the epicontinental seas less than 660 feet deep would be dried up, so that, to cite only a few examples, the English Channel, and the Baltic and the Sunda Seas, would disappear. The continental shelf would be almost entirely above sea level and the oceans would begin at the junction of this shelf with the continental slope. The euphotic zone would then descend to a part of the present continental slope and would modify the marine population there. However, the colonization of such steep slopes is more precarious than on gentle inclines and the surface available for neritic life would be much reduced.

It will be shown that the level attained by the Sicilian transgression (see Table I, p. 11) is found today at over 300 feet above sea level. If, as most writers agree, the whole surface of the continental shelf was dissected by subaerial erosion, the real changes between the lowest and highest points of mean sea level during the Quaternary is then 1,000 feet, over 300 feet more than the effect attributed to "glacial control". If it is postulated that the continents are rising at a more or less steady rate, this difference can be explained as the amount of epeirogenic uplift.

Eustatism apparently justifies the interpretation that Quaternary *marine terraces* represent different positions of mean sea level (C. Dépéret, 1913–1924). More recent works (A. C. Blanc, M. Gignoux, etc.) show that the history of ancient shore lines is very complicated and that much geomorphic evidence has been destroyed. It can be said rather that the terraces represent marine incursions to different levels, the elevations of which are controlled partly by eustatism and partly by epeirogenesis.

Marine Terraces of the Pleistocene.—1. The emerged surfaces at +260 to +330 feet.

These levels, which are among the oldest, represent an episode slightly younger than the Tertiary-Quaternary boundary. In Europe and in North Africa they can be correlated with the *Sicilian*. [Some workers also include the Calabrian terraces, at c. 500–600 feet in the earliest Pleistocene.]

On the Atlantic coast of North America, the Brandywine deposits, situated at 270 feet, are composed of gravels which have been interpreted as Lower Pleistocene alluvial deltas derived from the crystalline province of the Piedmont (C. W. Cooke). The terrace forming the upper part of it could correspond with the Aftonian (Günz-Mindel). The same surface has received the name *Hazlehurst level.*

On the Pacific coast, at Santa Cruz, 56 miles south of San Francisco, a terrace is recognized at +260 feet.

2. A number of levels immediately beneath, at 210 to +180 feet, which occur in Europe and in North Africa are known either as the *Milazzian*, or the *Sicilian II.* (At the old type area, Milazzo, this terrace is deformed, but the name persists by priority.) On the eastern coast of North America there are two levels, one at +210 feet, the *Coharie*, in North Carolina, the other at +170 feet, the *Sunderland*, in Maryland. These levels may correspond with the Mindel-Riss (=Yarmouth) interglacial.

3. A number of levels follow which are called the *Tyrrhenian* terraces. These are mostly contemporaneous with the Riss-Würm (=Sangamon) interglacial. Generally they occur at altitudes between +130 and +15 feet. In North Africa and in Italy the oldest terrace commonly occurs at +100 feet and is called the Tyrrhenian I; and a second terrace at +65 feet is sometimes known as the Tyrrhenian II (Main Monastirian). On the southeastern coast of the United States, in central Florida, the *Okefenokee* level (Veatch and Stephenson, 1911) is at +145 feet; the *Wicomico*, in South Carolina is at +100 feet; the *Penholoway* in Georgia and South Carolina is at +70 feet; and the *Talbot* is at +40 feet. On the Pacific coast of the United States, south of San Francisco, another terrace occurs at +95 feet. In Australia, an equivalent terrace is widely recognized between +100 and +120 feet, and another at +40 feet.

4. At about 25 feet there is one of the most widespread and well-preserved terraces in the world, which goes by the name "Late Monastirian" and by various other names. Regrettably at Monastir (Tunis) it is deformed, but the name is generally retained. This pre-Würm platform is represented in the eastern United States by the *Pamlico* surface at 25 feet and also in Australia.

5. Perhaps still relating paleontologically also to the Tyrrhenian, there is in North Africa the "Tyrrhenian III" (=Ouljian) or "Epi-Monastirian".

This may be the equivalent of the French Normanian, which is at about +15 feet. It is often difficult to distinguish such terraces from those of the post-Würm, that is to say, of the highest Flandrian. The confusion which occurs here comes from the fact that it is difficult to distinguish, in the absence of characteristic fossils, any level between 15 and 25 feet which *preceded* the retreat of the Würm, from another level of the same height which *followed* this retreat and which corresponds with a maximum for the Flandrian transgression.

6. Holocene terraces at low elevations, formed much nearer the present day are very widespread. One first described at 10 feet in France is named the Calaisian (Dubois, 1924), and a younger one at about 5 feet is called the Dunkerquian. Because such terraces are rapidly destroyed on coasts exposed to heavy wave action they are often difficult to follow from one district to another. In the Mediterranean they are collectively known as the Nizza terrace. Similar terraces are met on the eastern coast of the United States, such as the *Silver Bluff* surface at Miami (at an altitude of 5–10 feet), in the Hawaiian Islands at *Kapapa Beach*; and in Australia three levels are carved in a dune formation between +10 and +12 feet, between +4 and +6 feet, and between 1½ and 3 feet.

In Scandinavia most of these Holocene beaches are emerged, sometimes stranded far inland by the postglacial isostatic uplift. On the south side of the Baltic and westward to the Rhine delta, on the other hand, they are often tilted below sea level, and only known from coastal borings. The record, however, is better preserved than elsewhere, and multiple little oscillations have been identified, each corresponding to a slight climatic oscillation. Corresponding to the Calaisian are the *Littorina* beaches, to the Dunkerquian I the *Limnaea* beaches, and to the Dunkerquian II the *Mya* beaches.

It must also be remembered that there was a slight rise in sea level during the first half of this century, which coincided with a climatic amelioration in the northern hemisphere.

The chronological evaluation of terraces must be considered with some caution. Altimetric data alone may be misleading. They have, on occasion, suffered warping due to tectonic deformation. Moreover, some terraces have overlapped others, so that terraces of different ages can occur at almost the same altitude. This happens in the case of the Ouljian, the Calaisian and the Dunkerquian.

Comparison between Marine and Pluvial Lake Terraces.—It has been stated that the Pluvial periods in temperate latitudes are contemporary with the glaciations. The Interpluvial periods are therefore contemporaneous with the Interglacial periods and with marine transgressions, which deposited sediment upon the terraces. The enlargement of lakes during the Pluvial period was similar to the marine transgressions, but alternate with them. There is no question of establishing correlations between terraces of these two origins. There is a striking comparison between the Black Sea, which behaved as a branch of the Mediterranean, and the Aral-Caspian basin, which experienced a pluvial history. An early regression of the Black Sea corresponded with the expansion of the *Khosarian* Pluvial period of the Caspian. The Black Sea then rose to the *Uzunlar* and *Karangat* phases. The latter corresponded with the *Atel* regression in the Aral-Caspian basin. The regression of the *New Pontian-Euxinian* (contemporaneous with the Würm) corresponded with the *Khvalynskian* expansion of the Caspian. Finally, while the Black Sea has been lately experiencing an expansion, the Aral-Caspian basin is experiencing a retreat which started in the *New Caspian* stage and continued into the present epoch to produce the geographic form which we know today.

CLIMATES OF THE PRESENT DAY

Present climates are described in detail in specialized works. It is not intended here to repeat these descriptions but merely to place the present climates within the framework of geological history.

The zones which affect present-day geography have developed from the late Pleistocene climate but are modified by two important phenomena: the *changes in the centers of precipitation* accompanying an increase in temperature, and a great *marine transgression* arising from the melting of the Quaternary ice sheets. In the temperate belts the decrease of precipitation and the rise in temperature together cause evaporation of bodies of water stored on the continents. The relatively rapid and spectacular melting of the ice sheets, due to the postglacial rise in temperature, must not be regarded as the only source of water capable of raising sea level. Water from the evaporation of pluvial lakes must also be considered.

It must be emphasized that a large number of the lakes of our present epoch are clearly in regression. Among the largest areas of nonmarine water in the world, the Caspian and the Aral Seas, the Shotts, Lake Eyre and the Great Salt Lake occupy very small areas compared to the great areas that they covered during the Pleistocene. Thus, some of them only temporarily contain water or are even true playas. In fact several of these are the remains of ancient seas, but their great extension dates from the Pleistocene and they were contemporary with the glacial phases. Although they are often found in regions far from where ice sheets were formed, they must be considered as part of the effects of the glaciations. They indicate the general importance of atmospheric precipitation in displaced zones, whether as snow or rain. The extension of temperate latitude lakes in the Quaternary can thus be attributed to the "pluvial periods" and correlated with the glaciations. Tropical lakes, however, may well have coincided with times of stronger evaporation from the oceans during interglacial phases.

Regression of the temperate lakes occurred at the same time as regression of the glaciers, but at a slower rate. As the temperature rises, the ice melts and disappears. Some of the resulting water drains out to sea or is retained in basins and gives rise to lakes. A simple rise of temperature is not the only cause of evaporation from lakes formed during the pluvial period. The dryness of the atmosphere must also be considered. Although the drying up of the lakes took place slowly, it did not do so at a constant rate, so that while evaporation affects a vast sheet of water, the latter may be so large that the atmosphere in the neighborhood becomes humid and the water may be partly reprecipitated. The smaller the area of water becomes, the less are the chances for the atmosphere to become saturated. The water vapor is then carried up into the atmosphere and condenses to form rain which falls outside the original area.

Evaporation has a further consequence; that is the tendency toward the drying up of large parts of the land, formerly pluvial. The period about 5,000 years ago was called the "Climatic Optimum" in northern Europe; since then there has been a cooling and in the subtropics a desiccation. The way in which aridity has spread during the recent history of the globe is very striking. Man has left his mark and can be held responsible for much deforestation and even for some of the expansion of deserts. It is true that the conservation of forests and an intelligent irrigation can delay this process, but comparisons with similar periods which occurred during geological time, before the appearance of human beings, clearly indicate that this is a natural crisis.

Such crises have marked effects on the continental area of the biosphere. Numerous plants and animals whose extension was checked by the preceding glaciation have finally been destroyed by drought. Once more, man, in exterminating the last representatives of mammals and birds, is not the real agent of nature's destruction. It is deplorable that man has not employed his collective intelligence soon enough to check these extinctions. The Mediterranean region, the cradle of civilization, was undoubtedly less arid during classical antiquity than at the present time. It is not thought that the great invasions alone were responsible for the increasing dryness. It has even been suggested that drought, bringing famine, was the cause of large population displacements (Elsworth Huntington, 1907).

The existence of periods during remote geological time when playas were present have been recorded mainly by deposits of evaporites. The three periods during which salinity occurred most frequently followed the three principal periods of glaciation (see Table III, p. 17). The evaporites of other epochs are due to local lagoons or to saline coasts. The drying of the continental atmosphere, observable at the present time, can thus be expected at times during every glacial period.

However, despite its tendency towards aridity, the present period offers much diversity of climate. On the site of ancient ice sheets, now melted, there are lakes and stream systems which have none of the characters of playas.

In the temperate zones, the mean temperature and the distance from the sea determine the climatic regime. It is also necessary to add the factor of altitude, which in such a region often leads to vertical climatic divisions, easily detected by vegetation zones. This is characteristic of all high mountains such as the Alps, and is very marked on Kilimanjaro in equatorial Africa.

The monsoon climate, exceptionally well developed today, also must receive attention. Regions largely surrounded by sea such as the East Indies and southeast Asia, not only have a constant temperature, but also a seasonal rainfall. These hot and humid countries are covered with a luxuriant vegetation, which plays an active role in their geomorphic development and the sedimentation which accompanies it.

In the same way, the temperature of 15° C. constitutes a threshold of solubility for a large number of substances. At present, the winter isotherm of 15° C. in the oceans marks an important climatic limit to which a number of organisms are sensitive (Eckman, 1952).

Thus, although the climate of the present epoch appears to be clearly defined, one can recognize the presence of some characteristics of other climatic types, which are either residuals or have recently appeared. It is this climatic diversity which has allowed the successful geological use of the methods of uniformitarianism. However, it must be remembered that although the present has corollaries with the past, uniformitarianism is unable to account for all the events of the geological periods. Some examples are examined in the following section.

TETHYS TRANSGRESSIONS, MONSOON CLIMATE AND EXTENSIVE LATERIZATION

At certain times in the earth's history, the seas extended over continents of low relief, particularly in the Tethys (the ancient Mediterranean-Himalayan) region. The term *Tethys transgressions* has been given by the authors (1952) to these marine expansions. One of their characteristics is the uniformity of marine faunas and associated continental flora. The uniformity of marine faunas indicates, without doubt, that a temperature not less than 15° C. extended as far as the polar circles. The flora reveal two factors; that the plants are analogous to those of present-day monsoon countries, and that their wood is often devoid of annual rings. Thus, it can be said that during the Tethys transgressions there was a widespread occurrence of monsoon climates. This was a result of the penetration by the fronts into the heart of the major land masses, thus increasing the number of paralic basins, spreading moisture, and maintaining fairly constant temperatures.

From the time of the expansion of the continental flora during the Devonian, rich forests spread over lands experiencing this monsoon climate. The action of humus, associated with the growth of the forests, led to an acid alteration of the underlying rocks. This rock breakdown gave rise to a particular sedimentary cycle which will be dealt with at length. The final state of this alteration is laterization (pp. 142–146).

The homogeneity of marine faunas, which allows the recognition of a Tethys transgression, seems to be demonstrated by the widespread occurrence of *Archeocyatha*, as long ago as the Early Cambrian. The record of another transgression is found in the Middle Ordovician, where reefs spread as far as the Arctic regions; there is another in the Gothlandian and one in the Middle Devonian. But the first time that the extension of a warm fauna, a monsoon flora devoid of seasonal rings, and a laterization clearly coincide, is in the Carboniferous, from the Visean to the Westphalian. The extension of Tethys seas is first indicated by the uniformity of marine faunas in the Visean. The development of vegetation on the land took place after the seas had begun to retreat in the Late Namurian. It was at that time and in the Late Westphalian, that the luxuriant paralic flora developed, giving rise in certain regions to lateritic soils (H. Termier and G. Termier, 1958).

The general occurrence of a monsoon climate did not appear again until the Jurassic (Lias and Dogger) wherein there was an extension of warm marine faunas, a large distribution of paralic flora and the development of lateritic soils. Formation of the latter continued after the Tethys transgression and into the Late Jurassic.

In the Cretaceous, from the Barremian-Aptian to the Lower Senonian, the widespread transgression which led to a uniform marine fauna also favored the development of forests and the extension of lateritic soils, which gave rise, among other things, to the French bauxite deposits.

The Eocene, a new warm period of Tethys transgression, was a phase of expansion of the nummulitic facies, and of laterization.

Finally the Miocene, the last warm period of the Tertiary, following the Tethys transgression of the Eocene-Oligocene, is the period with the greatest known laterization in the world (possibly because these formations, being more recent are the best preserved).

INTERMEDIATE CLIMATES

The extreme climatic phases through which the earth's surface has passed have just been described. The glaciations associated with epochs of playas, correspond to periods of high relief and of maximum positive epeirogenesis. The Tethys transgressions with their monsoon climate correspond with periods of low relief and of epeirogenetic minima.

The passage from one to the other of these two contrasting phases implies a series of intermediate climates. This can be seen particularly well between the Permian and the Lias. During the time of Permian playas, the eustatic oscillations seem to have passed progressively into the Tethys transgression. The dry cool climate of the early Permian changed gradually to the hot arid climate of the Trias, before the seas penetrated sufficiently far into the continents to influence them with the humid monsoon climate.

Counter currents, a sort of reflex to the Tethys transgressions also can be seen. These have scarcely modified the general temperature and climate. The authors (1952) have called them *arctic transgressions* and *circumpacific transgressions*. The latter were accompanied by a certain local return to aridity in regions far from the Pacific.

Thus, periods with a mixed climate can be recognized. It has already been pointed out that laterization must have extended from the time of the Tethys transgressions of the Westphalian (Carboniferous), the Upper Jurassic and the Miocene. Another interesting and equally well-known time is the Late Devonian, which, without giving rise to a major glaciation, nevertheless possessed glaciers in South America and South Africa. In addition, wood from trees of that time displays seasonal rings.

These climatic reconstructions may help to bring some understanding of the history of the earth's geomorphic development, the agents of which have changed on a number of occasions during the course of geological time.

2

Earth Movement and Geomorphology

The molding of landscapes can be classified in two distinct categories according to the predominant earth movements. These are the old platforms and the orogenic zones. Slow movements of broad amplitude are characteristic of the former, and rapid folding, localized and spectacular are typical in the latter. The results of these movements vary according to their magnitude. In the two categories considered, it may be noted that the dynamic evolution of a region has an immediate effect on the river system; certain stream valleys are found to be antecedent, or controlled by orogenesis, examples of which are to be seen in many countries; others are dislocated by capture or by reversal of their direction of flow. The detailed study of river systems yields valuable information concerning the history of movements experienced by orogenic zones and old platforms. Unfortunately the constant upheaval of orogenic zones rarely allows investigation into their history.

EPEIROGENIC MOVEMENTS

The movements of continental areas which chiefly involve sinking and uplift, with or without fractures, are generally opposed to those of mountain zones. If these movements give rise to large-scale structures, such structures are scarcely contorted, in contrast with folded zones. On the other hand, it often happens that much larger areas are affected by epeirogenesis than by orogenesis. It appears well established (Termier, 1956) that epeirogenic movements of the foreland are closely associated with orogenic movements in the folded zones.

To a certain extent epeirogenic movements have effects similar to eustatic movements. They constitute the chief motive power for the processes of geomoıphology by changing the relationship between continent and base level.

It is important to recall that epeirogenic movements do not affect all parts of a continent equally. They separately affect each of the independent blocks which, collected together as in a mosaic, form a continent. This must not be visualized too diagrammatically for the blocks often behave as vast swells and basins of the substratum, delimited by flexures or fractures and controlled by "plis du fond" (Argand's term for deep-seated crustal warping).

These swells and basins have been called "*anteclises*" and "*syneclises*" in the U.S.S.R., terms which help to stress their mosaic or ellipsoid form.

In order to visualize the deformations experienced by a continent, it is particularly important to compare the age and amplitude of the epeirogenic movements which have affected each one of these elemental blocks respectively.

Among the traces left by epeirogenic movements are the transgressions and regressions localized on continental areas, the various pediplanations and peneplanations and their rejuvenations which followed the uplift of mountain chains, the antecedence of river courses, and finally, the fractures and foundering affecting the old continents.

This is particularly true of transgressions onto old Precambrian platforms: the Scandinavian Shield affected during the Middle Cambrian, the Canadian Shield (*sensu lato*) during the Late Cambrian in the west, and in the south the upper valley of the Mississippi. Another example from more recent chains is the westward Carboniferous transgression in the Appalachian region (P. B. King, 1955). It is noticeable that these large epeirogenic movements are generally contemporaneous with important movements in nearby orogenic zones. "Inversions of relief" have a tendency to become stronger in the foreland, the intermediate massifs and the geanticlines of continental structure during the uplift of mountain chains. However, it is remarkable that the amount of material affected by epeirogenesis is undoubtedly greater than that affected by orogenesis. It may be thought that orogenesis is subordinate, so that epeirogenesis prevails over orogenesis and even causes it.

Where pediplanation and peneplanation has occurred, the most spectacular example is, without question, that of an orogenic zone which has been leveled to a pediplain, in contrast with the folding which preceded it. But a region rarely experiences just one planation. After the planation immediately following a tectonic paroxism, others follow, and plane down the relief produced by rejuvenation and the purely epeirogenic uplift. These planations are not confined to orogenics and can affect vast territories. There is nothing surprising in this, since the phenomenon of planation depends essentially on the position of a continent in relation to base level (see p. 103) and to the duration of its stability in this position.

PLANATION SURFACES

According to W. M. Davis, the state of *youth* in a region which is experiencing a cycle of erosion, is characterized by the persistence of remnants related to the "initial" topography. This may be the topography following a marine transgression, or a rejuvenation, or even resulting from folding. The state of *maturity* is characterized by the total disappearance of the "initial" topography. Finally the state of *old age* evolves slowly toward his ideal concept which is the *peneplain*.

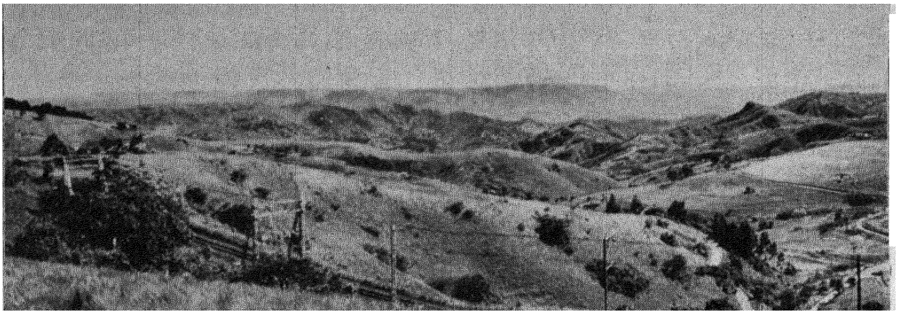


FIG. 5. VIEW OF NYUSWA IN NATAL, SOUTH AFRICA (30 miles from the coast)

The tabular surface (altitude 2,800–2,000, feet) visible in the background is Lower Tertiary. In the foreground, the deep valleys have been excavated in the Quaternary. This is "the Valley of a Thousand Hills". In the middle distance on the extreme right there is an Upper Tertiary surface (altitude 1,570 feet). Between the two surfaces, there are sandstone residuals. (Photograph: L. C. King.)

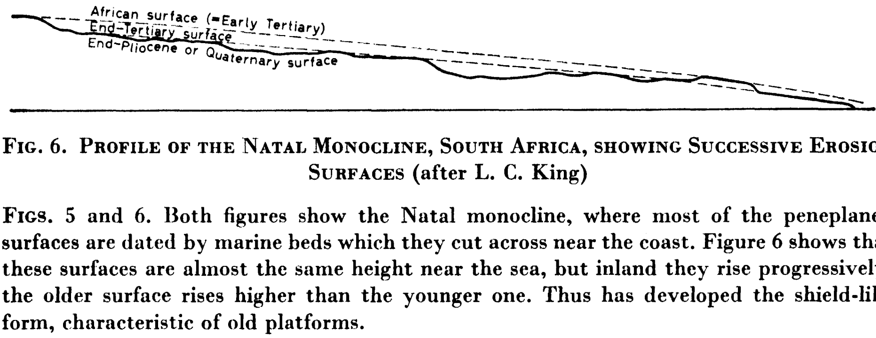


FIG. 6. PROFILE OF THE NATAL MONOCLINE, SOUTH AFRICA, SHOWING SUCCESSIVE EROSION SURFACES (after L. C. King)

FIGS. 5 and 6. Both figures show the Natal monocline, where most of the peneplaned surfaces are dated by marine beds which they cut across near the coast. Figure 6 shows that these surfaces are almost the same height near the sea, but inland they rise progressively; the older surface rises higher than the younger one. Thus has developed the shield-like form, characteristic of old platforms.

The "Gondwana" landscape of the Jurassic was peneplaned during the Early and Middle Cretaceous. Uplift led to renewed erosion terminated by a peneplanation at the beginning of the Tertiary (see Fig. 3). Other phases of denudation occurred at the beginning of the Miocene and in the Pliocene. The latter surface (Late Tertiary) (see Fig. 3) has been uplifted (nearly 6,000 feet in the Drakensberg). This was followed by the downcutting of the Quaternary rivers which are still active. These flow from east to west and have inaugurated a new cycle of erosion.

For many geographers, the state of youth in a cycle of erosion can be recognized by steep slopes on which there is sliding of detrital material. But this feature is only encountered in regions with great available relief, that is, in mountain chains and in high plateaus dissected by deeply entrenched rivers. Consequently a state of youth can never occur in broad depressions such as the Paris Basin and the Gulf Basin.

To the authors, youth, which is by definition the first stage of a cycle, is of importance only in a region newly subjected to erosion. But it is evident that the *appearance* of the three stages envisaged differs in a mountainous region subjected to folding and vertical movement, from that in a subsiding alluvial plain, the base level of which is suddenly lowered.

The age of exhumed surfaces varies from continent to continent. In southern Norway the Precambrian surface remains partly intact; in the Belgian Congo portions of Carboniferous topography are known. L. C. King (1957), believes that from the end of the Mesozoic era to the present day all South African surfaces which have formed successive steps in the landscape are synchronous with, and correspond to, those of Brazil (figs. 5 and 6).

Thus arises the question whether the uplift of continents, that is, epeirogenesis, is a widespread phenomenon simultaneously affecting large land masses, and in fact all the shields.

Certain developments stand out in continental evolution, and the synchronous evolution of Africa and Brazil, according to King, may be taken as an example. In the two regions, the formation of the present "rift valleys" has resulted from a powerful upwarping of the plateau in the Late Tertiary and in the Quaternary. From such upwarpings, lines of weakness were produced from time to time, which were predetermined by structural lineaments of the Precambrian platform. These must have been active at least since the end of the Paleozoic and coincide with the maxima of uplift. The "rift valleys" correspond, in fact, with the highest point of uplift of the planed shield (3,000–10,000 feet) in East Africa.

The deformations experienced by shields are clearly shown by the patterns of river systems and by the coastal sculpture. Such deformations, comprising syneclises and anteclises (see p. 41), affect the vast region of the Kalahari. The rise and fall of structural axes also affects the Brazilian shield. The drainage system there has undergone considerable alteration, including inversion of direction of flow and river capture (fig. 7). The coast line of the Serra do Mar is composed of eroded cliffs, resulting from the *submergence* of the region, while toward Cabo Frio in the east, the shore is characterized by lagoons and saltpans indicating *stability* of the platform.

Pediments and Pediplains

The most perfect planations imply a succession of complex phenomena, which include transportation and deposition of sediments as well as erosion. In a mountain zone, each hillside presents a form of torrential, homogeneous rainwash, giving rise to talus cones at the foot of the principal slopes. Such cones tend to coalesce by extending their bases, and thus form *alluvial slopes*. **In an arid or semiarid climate,** a *piedmont slope*, called a *pediment* (McGee; Bryan), or *rock floor* (Davis) or *rock plane* (Johnson) is often formed by erosion at the foot of a mountain. This slope may continue upward and may reach between the mountain summits, thus breaking up the chain by the formation of "embayments".

Thus pediments are surfaces cut into hard rocks at the foot of mountains

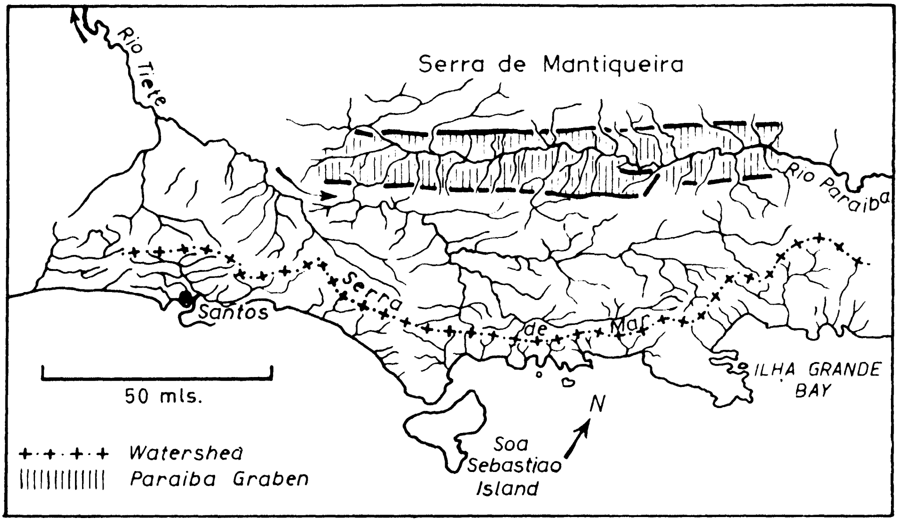


Fig. 7. Sketch Map of the Serra do Mar and the Paraiba Graben, Brazil (after L. C. King)

Following a planation in the Middle Tertiary and the cutting of valleys in the Younger Tertiary, this area was affected by considerable earth movements. The drainage underwent marked modification; only those streams whose sources lay on the eastern slopes of the Serra do Mar flow directly to the coast. By contrast, the Rio Tiete now flows 2,000 miles before reaching the sea.

which they continue to erode. At their highest point, they join the mountain side at a break of slope called a "knick" (=crease or sharp bend, in German) and at their base they pass imperceptibly into the plain, which is gradually covered with material wrested from the mountain tops. This material builds up a special alluvial slope called a *bajada*. It is an alluvial fan which may be over 800 feet thick. A particular feature of a pediment is that a profile drawn parallel to the break of slope (marking the junction between pediment and mountain) shows that thalwegs or interfluves rarely occur. The process which has led to the development of pediments still remains obscure. According to L. C. King, they are formed by the action of sheets of water which, during the tropical rainy season, are not directed into channels but operate over the whole surface.

There is such a continuity between pediment and bajada that for certain writers, such as Cotton, the term "piedmont slope" applies to both the pediment and the bajada. At the final stage, when the pediments surrounding an ancient mountain massif coalesce and spread over it, they form a *pediplain* broken by erosional remnants (e.g. monadnocks, inselbergs, or kopjes).

The formation of pediments is the result of a continuous mixing of material brought down by erosion. This mixing is the ultimate cause of the perfection of the pediplain's planation. The view held by Baulig, which differs somewhat from that of King, is as follows: erosion brings down hard rocks from the mountain core (crystalline rocks, quartzites, etc.). When they reach the plain, the streams, carrying their maximum load, encounter softer rocks, such as weathered granite. The hard rocks then act as an abrasive and wear away the river banks: this is *lateral corrasion* (Gilbert, 1877). Furthermore, streams easily shift their beds in loose ground, such as the distributaries of a delta. The abraded rocks vary in resistance, the hardest remaining in relief. Plant cover also gives unequal protection from one rock to another and a weathered layer of alteration, which may be a calcareous or siliceous crust, is formed. The nature of the rocks which is the chief factor in geomorphological differentiation, is here of material consequence. Granite which produces a homogenous arenaceous zone of alteration exhibits the smoothest profile. It can be said that *pediments are the product of arid climates.*

In a region of great stability, pediplanation leads to a balance between aggradation and degradation, the *panfan* of Lawson. If the region has no outlet to the sea, the material deposited accumulates in an endorheic basin (see p. 104) and the base level becomes higher. The existing relief is progressively smothered by these deposits. But there are few regions of the earth's surface which remain stable for a long time and it can be said that, **geologically speaking, an indefinite stability is inconceivable.** Connecting a basin with the sea leads inevitably to degradation, while its isolation induces accretion. Sometimes a change of climate or an epeirogenic movement postpones the time of burial of a pediment beneath alluvial cones. Consequently *a pediplain is always the indicator of crustal stability for a relatively long period of time.*

During geological time the rejuvenation of mountains and the subsidence of alluvial zones were important factors in the formation of pediments. By the progressive lowering of base level or by increasing the slope of rainwash, there is the almost indefinite continuation of a phenomenon which would certainly prevent congestion by sediments. Under these conditions an equilibrium between the action of erosion and sedimentation is produced. Undoubtedly, the periods without continental vegetation, that is to say, the periods before the Devonian and the arid periods since that time, were ideal for the formation of pediments.

Tangential Erosion and "Sheet-flood"

Erosion by rainwash and small gullies must be included in the consideration of the formation of planed surfaces. This erosion is called "sheet-flood" (MacGee, 1897) and appears to be the process which G. Choubert (1945, p. 729) has called *desert or tangential erosion.* It is a form of erosion which "proceeds by the successive cutting of whole horizontal slices, and thus gives rise to perfect plains". This is the case of the regs (stony deserts) which are ranged in terraces. The formation of pediplains is a similar phenomenon. Several agencies seem to underlie this erosion. These certainly include the river system and especially pluvial erosion by "sheet-flood"; also eolian corrasion and deflation. In the case of a plain of erosion, the principal factor is the stream system.

Cryoplanation

The term cryoplanation was proposed by Kirk Bryan for a type of erosion operating under periglacial conditions on gentle slopes. J. Alexandre (1957) has suggested that planation in the Central Ardennes could result from *cryergy*. This latter is defined by Y. Guillien (1949) as the combination between congelifraction, solifluction and cryoturbation and its results may be added to those of rainwash. Cryergy operates best on slopes of less than 10° and on the less resistant rocks. In this way, partial planations are produced, giving rise to pediments which correspond with the *cryoplanation* of Bryan. In a large valley these surfaces merge with the top of an alluvial plain, which may be represented by an ancient terrace.

Given that a more rapid thawing occurs on the adret[1] than on the ubac[1] solifluction is most strongly developed on the less sunny slope. Thus, the valleys develop an asymmetrical form.

Hamadas, Regs and Sai

In the Sahara, the name *hamadas* is given to uplifted plains, especially denuded plateaus, exposed to the wind in a desert climate.

Regs and *serirs* are planed areas with a covering of boulders, which tumble from the surface of the hamadas or from the plains below them. The term reg is generally reserved for the low plains used by caravans. Moreover, this term is applied commonly to all bouldery ground which has been subjected to deflation. The wind acts as a fan, and carries away the grains of sand, leaving in place the stones which are too heavy to move. These boulders finally become polished and varnished. This "desert varnish" or "patina" is a ferruginous and siliceous coating gradually applied by the alternation of dew during cold nights and evaporation by the hot sun.

The regs can be: (1) Valley fill brought by the large oueds into closed basins (allochthonous regs). Such are the Oued Trahart and the Oued Tekouiat in the Tanezrouft, also those of the Kemmé valley (Enneri Kemmé) which formerly, carrying rhyolitic boulders, led to the Tibesti Serir in the direction of the Syrtes, or (2) Eluvium (residual soil) resulting from disintegration of the substratum in places (autochthonous regs), for example in the Libyan desert (Lelubre, 1952).

A similar concept is that of the *saï* of the Tarim Desert, which corresponds to a piedmont plain. It is barren and formed from a kind of reg with stones polished by the wind. It has acquired a desert varnish and is laid out like

[1] *Adret:* slope exposed to the south, facing the sun and protected from the north winds. *Ubac:* slope exposed to the north and thus in the shade.

FIG. 8. NORTHWEST PART OF FAHAL ALORRAT IN THE SAHARA
(Photograph: S. E. T. P. Jean Guglielmi)
Note the inselbergs and the pattern of a temporary rainwash network on the peneplaned surface.

"macadam". These stones are the product of the disintegration of much larger blocks, broken up by the alternation of strong insolation and freezing nights. The saï are thus confined to the autochthonous reg.

Peneplains

A peneplain (W. M. Davis, 1909) resembles a plain but preserves undulations and traces of relief. In principle, such a structure is essentially due (particularly in temperate and humid climates) to erosion which lowers the crests and slopes of hills.[1]

[1] To relief features standing above the planed landscape, the various names "inselberg" or "monadnock" are given in arid and humid countries, and "nunatak" in glaciated regions (fig. 8).

Many writers (L. C. King, among others) today tend to think that the term pediplain ought to be substituted for that of peneplain, used in geological literature. Certainly, many planed surfaces are formed by pediplanation. It may be thought, however, that more complex phenomena have taken part in the construction of large and almost universally recognized peneplains.

The Precambrian peneplain of Africa (called the sub-Cambrian in Scandinavia) and the pre-Permian[1] of Europe and North Africa are of this type. Their surfaces can be seen to cut across folds. Fluvial erosion alone cannot be invoked to explain the formation of these peneplains, and marine erosion is even less likely. There is no doubt that a number of erosive activities, such as preliminary pediplanations, tangential erosion during arid periods and perhaps also pedogenesis, must be considered. In fact, certain geological phenomena are known, which occur only in areas that have almost attained planation. This is the case where laterites are formed. These rocks appear in the zone where there is a fluctuation of water level. It is possible that this is a process associated with the cutting of peneplains in areas covered with vegetation (see p. 142). It must be mentioned that planed erosion surfaces, which have been formed by the lateral corrasion of a river, also exist.

Roles of Erosion Surfaces in Geology and in Paleogeography

It would seem that erosion of the earth's relief by rainwash producing a minutely etched planing would finally reduce the continents almost to sea level. This planation then, making a "clean sweep", prepares the ground for new geological phenomena. It often marks a "continentalization" over a long period of time. The ancient shields, which are also called "old platforms", are none other than peneplains which have almost attained a final form. Situated in the center of continents, they show a persistent tendency toward epeirogenic uplift. It is on plateaus, sometimes uplifted and peneplained (often those which have undergone initial pediplanation) that desert dunes (Sahara) and ice caps (Scandinavia) have developed. These have perfected the planation of the continental surface. The same plains of subaerial erosion, reduced almost to a coastal platform, offer little resistance to marine transgressions. The smallest eustatic movement raising the mean sea level permits invasion of the continental area by the sea. The African coast of the Gulf of Guinea illustrates well this type of low coast. The coast of Senegal is similar (R. Laffitte, 1949, p. 247).

In fact, most of the transgressions which can be reconstructed from geological history are the products of surfaces of planation, covered or not, as the case may be, with their mantle of eluvial or alluvial detritus.

Surfaces of pedimentation, of tangential erosion, of cryoplanation, of ferralitic alteration, of planing by ice sheets and even (although reduced to a narrow strip) the coastal platform, amount to *a superficial leveling of the*

[1] That is to say post-Autunian or pre-Saxonian. It is also called post-Variscan.

land surface during periods of stability of the earth's crust. During certain geological periods this leveling seems to have attained a universal extent. The *Subcambrian surface* seen on every single continent, and the *post-Variscan surface* may again be noted. It is significant that these two surfaces which followed periods of intense orogenesis, occurred at times when the earth's climates displayed strong contrasts, and also when continentalization and epeirogenesis were most active. In the intervening periods, the surfaces formed were less perfect. These imperfect surfaces are the peneplains where universal correlations remain less certain.

OLD PLATFORMS

The old platforms gradually built up by the Precambrian orogenesis and planed several times, occupy a unique position in the geomorphic evolution of our planet (E. Suess). Their distinction lies in their unique tectonic behaviour. Broadly speaking, the upwarpings of shields are swellings of so large a radius of curvature that they appear to be simple vertical displacements. These are the *epeirogenic movements*. In addition the continents have undergone local warpings, archings or depressions. The terms *anteclises* and *syneclises*, introduced into Soviet literature (see A. Bogdanoff, 1958), may be applied to these results. The typical example is furnished by the Moscow syneclise, which downwarps the Russian platform.

Slow Evolution of Deserts

Taken as a whole, the shields are proof of a general tendency toward uplift. This tendency appears to be permanent, while at the same time, geomorphic action is in process of dismantling them. In the Congo Basin and in South Africa (pp. 33–35; figs. 5 and 6), the later erosion surfaces succeed one another like the giant steps of a staircase, so that the most ancient surfaces are also the highest.

A large number of the old platforms have evolved into **deserts**. For most readers, deserts are inseparable from heat, but they ought to be associated with a state of little surface water. Indeed, the old platforms are abraded as far as the granite where large outcrops of crystalline rock occur. On the one hand, the peneplained rocks are suitable for the formation of stagnant pools with muddy bottoms (where the water does not flow); on the other hand, their erosion produces sand which in vast flat regions is carried by the wind. In this way deserts develop, regardless of the prevailing climate. Consequently the moors of Dartmoor in southwest England and the heath of Lüneburg in northern Germany, should be called deserts, because, although situated in the midst of the temperate zone, they are without much running water. Among famous deserts, the Sahara must obviously be cited as a hot tropical desert, but the Gobi is a cold desert. These two examples are related by morphologically similar forms.

Correlation and Dating of Erosion Surfaces

The importance of the epeirogenic uplift of continents in relation to the Quaternary platforms has already been noted. It had an analogous effect on peneplaned areas and should be considered in the relative dating of peneplains produced on the continents, that is, on the old platforms.

In South Africa, Cahen and Leperson (1952) and L. C. King (1954) distinguish eight stages of erosion since the Carboniferous. These correspond to recognizable erosion levels, which rise above each other, like the steps of a staircase (see p. 34). L. C. King (1957) has also observed that the same erosive phases affected the Brazilian shield, and likewise the Australian shield.

In southern Norway, traces of the sub-Cambrian surface can also be seen to dominate the more recent "Paleic" (probably early Tertiary) surface. Considering negative movements which occurred between the phases of planation, there is no doubt that *the old platforms were subjected to a regular epeirogenic elevation the results of which can still be seen at the present time.*

OROGENIC ZONES

Earth movements in orogenic zones do not differ fundamentally from those affecting the old platforms. They are more intense but they are localized, and might be regarded as repercussions of the latter.

Young Mountains and Old Mountains

One of the ideas, poorly defined from a geologist's viewpoint, is the difference between *young mountains* and *old mountains.* This notion, a purely morphological one, must not be confused with the age of their initial geosyncline, nor with that of rejuvenation. Thus the Central Massif was subjected to Variscan folding and was rejuvenated, but not folded again, during the Alpine orogeny. Thus the Central Massif is an "old" mountain mass. The Pyrenees, whose substratum is composed of material of the same age and composition as the Central Massif, were refolded at the beginning of the Tertiary. Thus they are a "young" chain like the Alps. It must therefore be concluded that the ideas concerning young mountains and old mountains are bound up with the differences between orogenic movements, the generators of localized folds, and epeirogenic movements that affect the continental block as a whole. Although the latter have the most widespread ultimate effect, the former are responsible for the most spectacular relief. They thus furnish material ready for sculpture by erosion, and allow the accomplishment of more complete geomorphological cycles. Again it becomes necessary to distinguish between:

(1) deep-seated warping ("plis du fond" of Argand), like those which have given rise to the Pyrenees,

FIG. 9. THE KANCHENJUNGA MASSIF (HIMALAYAS). VIEW TOWARD THE SOUTHWEST OVER THE VALLEYS OF NEPAL FROM CAMP 2, JANNU (18,370 feet).

Note the parallel, knife-edged ridges, and in the distance the snow-covered crests of the high summits (Gipfelflur). (Photograph: French Himalayan Expedition, 1959.)

(2) superficial folds or "plis de couverture"[1] like the visible part of the Jura,
(3) gravity slide nappes like those of the Alps and Apennines, and
(4) zones raised by faults as in the Andes.

"Gipfelflur" and Steps

The rise of a mountain chain occurs in several phases, which can be subdivided into a series of discontinuous movements. These phases are always accompanied by vigorous erosion, which is continuous although variable in intensity. Furthermore, most chains exhibit mountain spurs or lateral steps, forming shoulders delimited by flexures and often even by faults (as in the style of the Bergamask Alps).

Almost all the summits of the axial zone of a mountain chain have peaks rising to similar heights. This concordance of summit levels constitutes the "Gipfelflur". Various theories to explain its origin have been advanced, in particular by Davis (1911) who saw in the concordance of summit levels evidence of an ancient planation or peneplain. In reality this "Gipfelflur" which characterizes young mountains is far from being truly planed. As shown by A. Penck (1919) it results from erosion by approximately equidistant streams, barely entrenched on a vast arch which represents the maximum uplift of the chain. The ideal points of departure of the streams and the general conditions of erosion, including the nature of the material attacked, are practically homogeneous in this mountainous zone.

ROLE PLAYED BY FAULTS

Large faults are known, which have continued to move throughout geological time and which, like the fault in southern Norway, have been active since the Precambrian. On the other hand, the "living fault" (San Andreas Fault) in California (fig. 10) developed much later. Traces of such movements can occasionally be found in the stratigraphic column. Erosion near faults which bound horsts produces coarse material which rapidly passes to finer sediments. Accumulations of breccias are known along faults. These indicate a prolonged tectonic activity or repeated renewals of movement at intervals of time which may be in the order of 50 million years.

The role of faults in static geomorphology must not be considered purely on the basis of their *last movement.* Thus faults are known which may be called "young" when a little of the fault plane still forms an escarpment, they are still more "recent" when the fault plane is polished and striated. There are "mature" faults whose relief is partly abolished, and "old" faults where the relief has become completely obliterated. But these terms only concern the *present form* of faults and can reveal nothing of the date when movement commenced, for it is obvious that the speed of alteration

[1] Translator's note.—"Plis de couverture" (Argand) are a superficial type of folding resulting from the sliding of strata over a lubricating layer and compression into folds.

of the fault plane or of the relief can vary infinitely according to circumstances. In a Carboniferous basin, a fault of Variscan age, thus an old fault, can have moved in recent times, producing a fault plane which can be found intact. On the other hand a fault of Late Tertiary age, thus of recent formation, which has not moved again, may have had time to largely disappear from the landscape. The shattered zones accompanying faults also

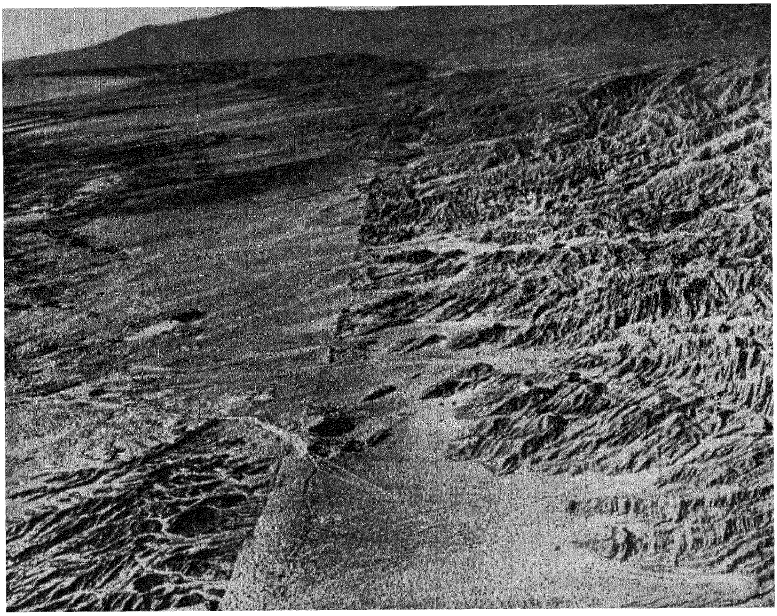


FIG. 10. THE "LIVING" SAN ANDREAS FAULT, CALIFORNIA
(Photograph: Laboratory of Physical Geography, Paris)
This fault is clearly visible along the whole of its length (about 620 miles) from north of San Francisco to the Salton Sea. Its fresh appearance is due to repeated movements, at least since Tertiary times.

serve to draw attention to them. This happens, for example, in granite masses, where fault breccias which are chloritized stand out green against the virgin rock.

If landscapes are viewed on a large scale, faults can often be clearly picked out by their straightness. The east coast of Madagascar coincides with a long fault for at least 700 miles. Inland, the point where faults intersect the surface topography may sometimes be shown by the presence of water. The Great Glen Fault of Scotland, marked by a string of lakes along its length, is a particularly good example.

The geomorphological role of faults is illustrated in Belgium by the "Eifel Fault" described by C. Ancion and J. Van Ham (1955). To the south of Seraing "the fault acts as an impervious barrier between the permeable sands of the Devonian outcropping to the south, and the impermeable shales of the Westphalian, lying to the north". The waters thus ponded up, only escape by transverse valleys where they form important springs marking the position of the fault.

In Central Morocco, in the region of Dechra Ait Abdallah and of Jebel Aouam, the faults and outlines of overfolds often cause springs which give rise to a greener vegetation in an almost arid landscape. In general, the paths of hypabyssal rocks and of mineralization follow those faults which are sufficiently deep to reach the zone of instability of the earth's crust.

3

Submarine Morphology and its Relationship to Continental Evolution

THE OCEAN FLOOR

The Continental Margin (fig. 11)

The continents, composed of light sial,[1] extend under part of the sea. This projection, which may be called the *continental margin*, exhibits a characteristic profile all round the oceans.

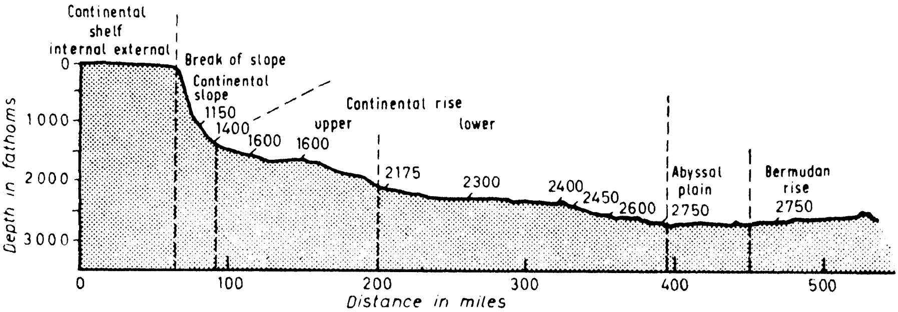


FIG. 11. TYPICAL PROFILE OF THE SEA FLOOR OF THE NORTHEAST OF THE UNITED STATES, ANYWHERE BETWEEN GEORGES BANK AND CAPE HATTERAS (after Heezen, Tharp and Ewing)

Proceeding from the land above sea level the following divisions may be distinguished:

1. *The continental shelf* possesses a very gentle slope (average 0·07°) and extends to depths of 300 to 1,000 feet.

2. The *continental slope* is steeper than the shelf, its slope being from 3° to 5°, and it extends to depths near 8,000 to 9,000 feet.

3. The *continental rise* although it descends to 16,000 feet has a gentler inclination than the continental slope.

The continental margin corresponds partially with the *precontinent* of Bourcart (1958), although this term includes only the shelf and the slope.

[1] Sial: Term formulated by Suess, designating the *si*licate and *al*uminum components which form the greater part of granitic material (density about 2·7).

The Continental Shelf

This is a platform inclined gently seaward from the shore to a depth of 300 to 1,000 feet below sea level. From these depths, the slope increases abruptly to form the continental slope. This margin, which on submarine maps appears flat and smooth, is very extensive to the southeast of New Brunswick and Newfoundland and also to the west of the English Channel. The continental shelf displays geomorphic features which most geologists and oceanographers agree were cut by subaerial erosion. Submarine rivers and canyons which form a continuation of rias generally to a depth of 150 feet are cut into the shelf. J. Bourcart insists that meanders and elbows of capture are present, and that the canyon floors recall those of streams flowing through unconsolidated sediments. The slight irregularities of the continental shelf occasioned by islands and banks might be considered as monadnocks scattered over a peneplain. Deeper down, at about 300 feet, there are narrow channels (the Toulon roads) and deposits, seemingly brought by muddy flows from the continent. The larger valleys and the principal submarine canyons form a continuation of the subaerial valleys as far as the continental slope, but some of them only begin at a depth of 300 feet or so. It might be thought, according to J. Bourcart, that their upper segments have been choked by sandy sediment. However, the hypothesis of Daly, Keunen and Ewing has attributed their excavation almost exclusively to turbidity currents and their formation to the time of the Flandrian transgression. In the Hudson Canyon, 12,000 feet deep, fossils of Quaternary (Wisconsin) age have been dredged and shallow water facies found which have been transported by current action (Richards and Ruhle, 1955).

According to Shepard (1948) and Bourcart (1958) the excavation of valleys, now submarine, took place in several stages: at the end of the Miocene, the end of the Pliocene, and in the middle of the Quaternary.

There is thus no difference of structure between the continental shelf and the adjacent continent. Their junction, which corresponds with base level, is often fringed by maritime coastal plains, which, during geological time, were built up by river alluvium on the substratum of sial, just as sediments have accumulated in the sea. They correspond to a number of different positions of base level. It may be concluded that *the continental shelf, like low-lying continental areas, is the site of transgressions and regressions.*

Where it is adjacent to folded regions, the continental shelf may possess an analogous relief. This is the case in the region of the Maritime Alps (Bourcart, 1958).

The *shore*, which is the dividing line between land above sea level and the continental shelf, is very variable in position, dependent upon the level of the sea in relation to the continent. The many reasons for the variations in the positions of the shore line will be considered later (pp. 65 and following).

During major regressions an important part of the continental shelf appeared above the sea and evolved into a newly uplifted maritime plain furrowed by river systems which deposited alluvial sediments.

To support this statement a few typical examples of submerged continental shelves will be described.

The Sunda Shelf (fig. 12) is a continental shelf occupying 1,150,000 square miles (about three times the area of France) composed of igneous and metamorphic rocks and covered with recent sediments. Its surface is marked by the traces of a large river-drainage network, Molengraaff's *River Sunda* (1922), which flowed northward; there is also another following a south-southeasterly direction toward the Strait of Macassar. These streams, subaerial a short while ago, are partly buried, in the same way that their tributaries have been covered by recent sediments carried by rivers. But their connection is borne out by the distribution of fish in the streams of Borneo and Sumatra. The fish of Kapoewas (western Borneo) are very different from those of Mahakam (eastern Borneo) but very similar to those of the Sumatra streams. Because of this, Weber considered that they were all descended from the ichthyological population of the Pleistocene Sunda basin. To these stream beds of subaerial origin must be added the channels and cul-de-sacs excavated by certain tidal currents as they were forced into narrow passages (Straits of Banka and Sunda, and the region of a Thousand Islands off Jakarta). Finally, rocky islands and submarine ridges which are a continuation of the mountain chains from the surrounding land represent the ancient topography (R. van Bemmelen, 1949).

The Arafura Shelf (figs. 13 and 14) extends between Cape York and the Sahul Shelf and is now submerged. Its northern edge is bounded by the Merauke ridge, south of New Guinea, and it is bounded on the west by the Aroe Islands. These islands, whose altitude rarely rises above 50 feet, are separated from each other by narrow marine channels, the "Soengeis". Tidal currents occur in this region to depths of 300 feet. The terrestrial fauna of the Aroe Islands has an affinity with that of New Guinea (Birds of Paradise) and with that of Australia (Kangaroos). This proves that the Arafura shelf was emergent during the Pleistocene and permitted communication between the two regions. Since 1857 (Wallace), the "Soengeis" have been considered, like the rest of the channels, to be part of a Quaternary river system. Fairbridge (1951) attempted to reconstruct this as follows: a number of major streams flowed across the shelf, one to the south of New Guinea in the present-day Snow Mountains Trough, and another in the Arafura depression. The "Soengeis" of the Aroe islands would have been coastal streams of secondary importance. The last submergence of the shelf would date from the time of the Flandrian transgression. Deep dissection of the streams occurred during the late Pleistocene, but was amplified due to local warping. This is not surprising, since the East Indies nearby were experiencing maximum orogenesis. It has since led to a superficial emergence of the Aroe Islands.

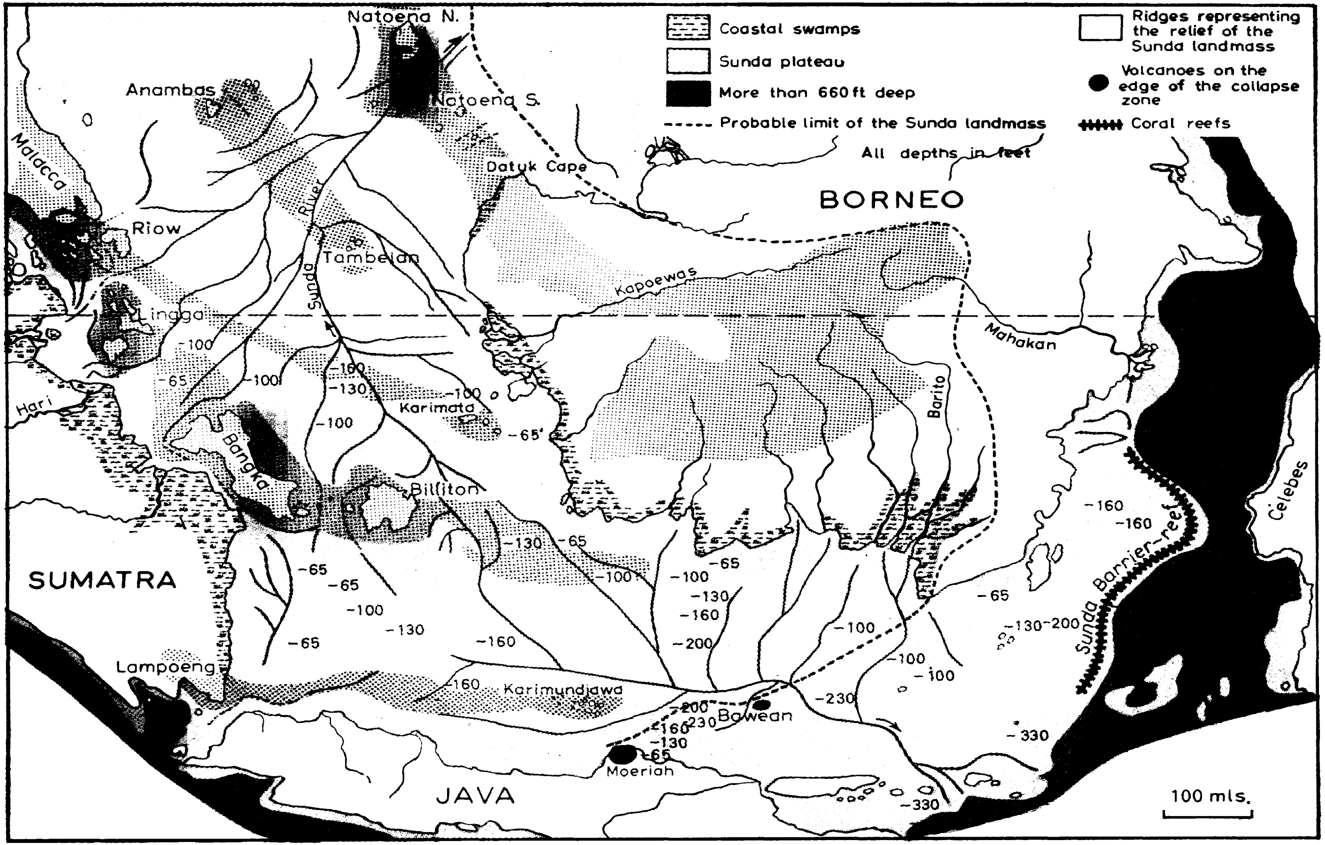


FIG. 12. THE SUNDA SHELF, AUSTRALIA, SHOWING THE ANCIENT STREAM SYSTEM AND ITS RELATIONSHIP TO THE RIVERS OF THE SUNDA ISLANDS. THUS, THE RIDGES, WHICH ARE A CONTINUATION OF THOSE OF THE LAND ABOVE SEA LEVEL REPRESENT THE ANCIENT RELIEF OF SUNDA (after Molengraaff)

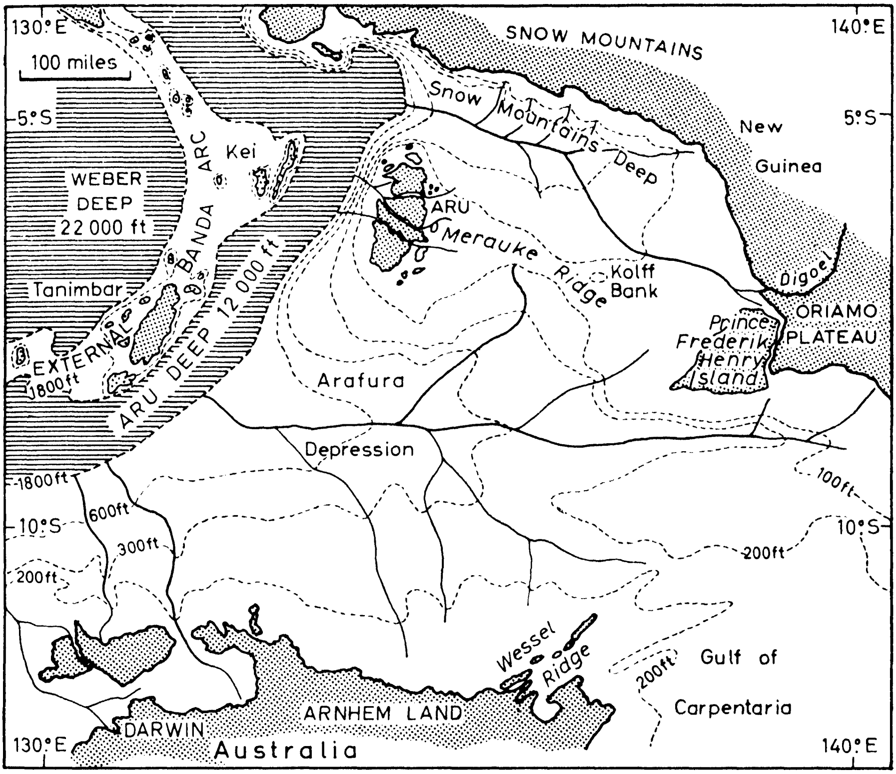


Fig. 13. The Continental Shelf North of Australia showing the Submerged Fluvial Network and the Islands of Aru (after R. W. Fairbridge, 1951)

The dotted lines represent submarine contours.

The *Tyrrhenide* (Forsyth Major, 1883) is an ancient land which sank progressively beneath the sea during the Middle Miocene when, as a result of repercussions from the Alpine movements, the Mediterranean sea came into existence. Two examples of the continental shelf which formed part of it, and which today border the Tyrrhenian sea, are described here. The Tuscany coast to the west of Pisa has been studied by A. Segre (fig. 15). In that region, he recognized an area which was partly above sea level during the Würmian. A common extension of the Arno and Serchio rivers can be discerned, and also an extension of two other rivers and their tributaries (the Meloria and a stream bed cut between the Rosignano Sovay and the island of Gorgona). As the upwarp of Tuscany resulted from the buckling of the Apennines (in the Neogene), it is likely that this river system crossing the present coast line is no older than the Quaternary. The French Mediterranean coast borders a more ancient land in which Variscan massifs, such as the Maures of Esterel, and the Lower Tertiary mountain chains, such as the

FIG. 14. THE ISLANDS OF ARU. A VERTICAL AERIAL PHOTOGRAPH OF SOENGEI MAIKOOR, SHOWING OLD FLUVIATILE SEDIMENTS ON WHICH A RIVER CHANNEL WAS SUPERIMPOSED, DISSECTED DURING A PHASE OF LOW SEA LEVEL DURING THE PLEISTOCENE, AND THEN INVADED BY THE FLANDRIAN TRANSGRESSION

Note the small structures (torrent bedding) etched out by the Soengei, i.e. waterway. (Photograph: Fairbridge.)

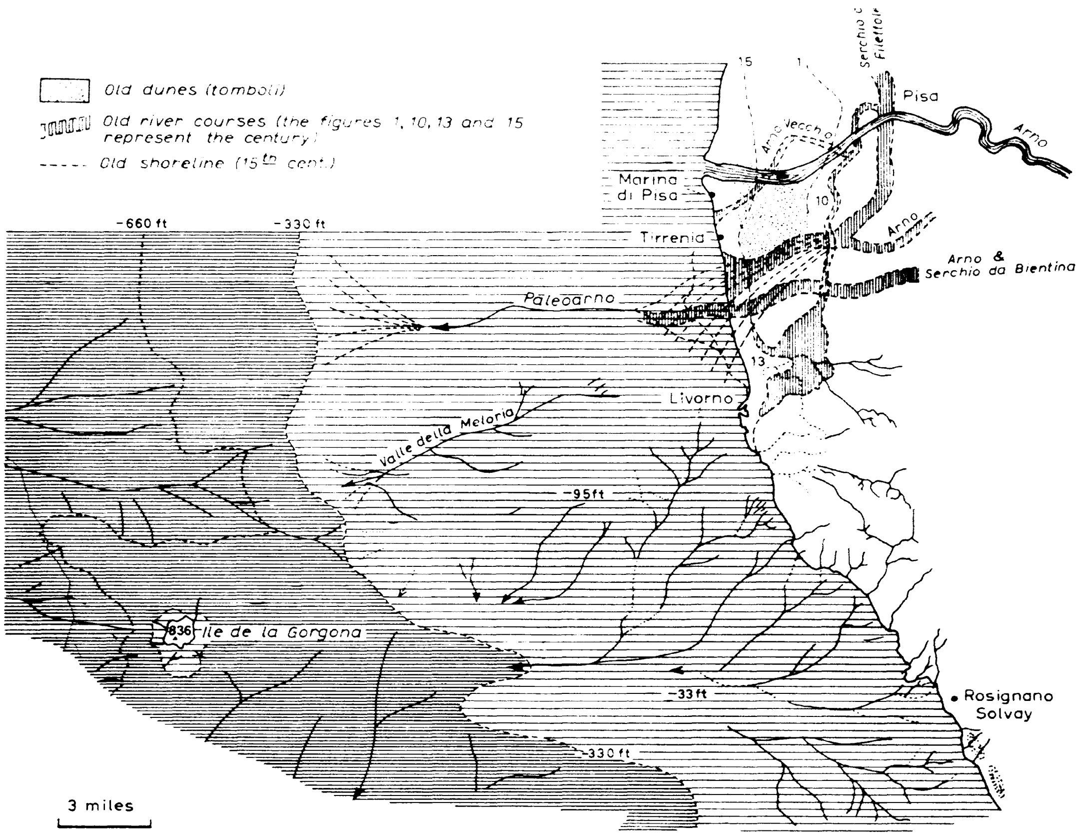


FIG. 15. THE COAST OF TUSCANY (ITALY) AND ITS CONTINENTAL SHELF (after A. Segre)

In wide horizontal shading—the sea floor from 0 to 300 feet; in narrow horizontal shading—the sea floor below 300 feet.

Pyrenees and the mountains of Provence, can be included. The subaerial river system of that region dates from the Pontian. The Pliocene sea invaded these valleys and formed rias, including those of the Rhone. Since that time there have been periods when they have been above sea level, as in the Villafranchian. J. Bourcart (1958) has shown evidence of the

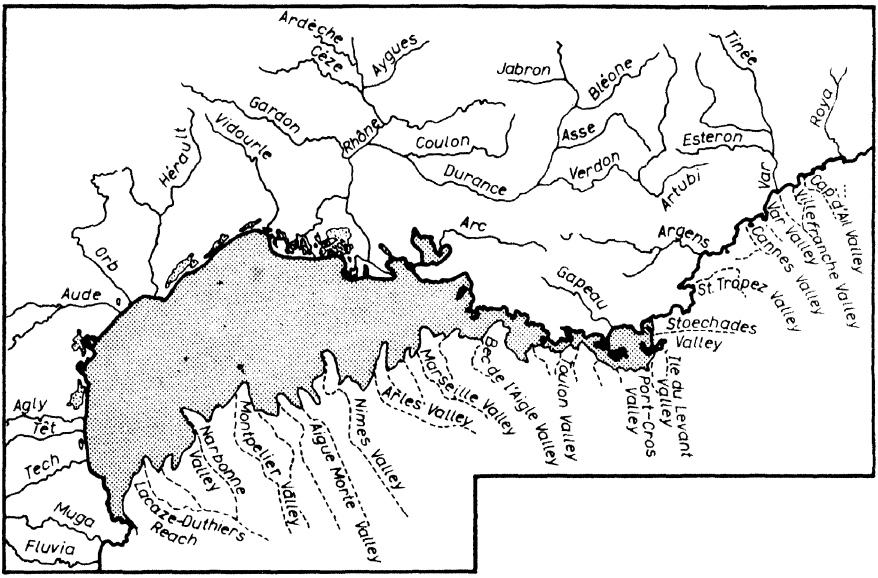
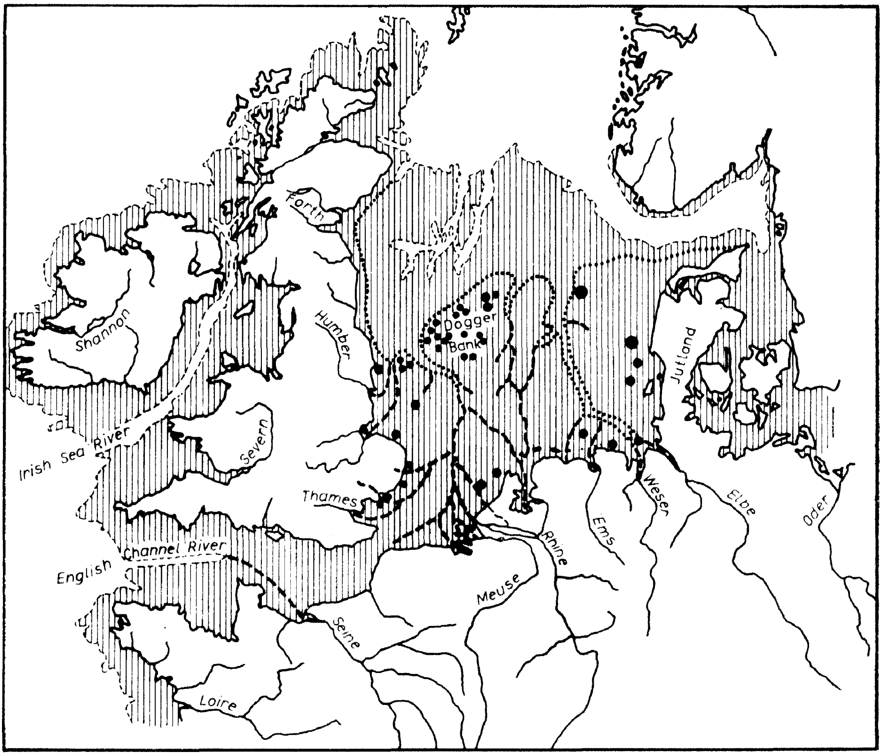


FIG. 16. THE MEDITERRANEAN COAST OF FRANCE AND ITS CONTINENTAL SHELF (Submarine valleys after J. Bourcart)

In gray—the detrital terrigenous sediments which have spread out on the continental shelf.

existence just off the coast of submarine canyons continuous with the present-day coastal streams (fig. 16) except in a part of the Gulf of Lions where recent sedimentation has obliterated the connection.

Finally, a well-known example, that of the *English Channel* and the *North Sea*. From the end of the Miocene until the Pliocene, a peneplain furrowed by rivers covered the region between France and England. The continual alternations of transgression and partial retreat of the sea during the Quaternary, are recorded here. The last regression was again that of the Würm. The river system of that period, which certainly must have had much in common with that of the Pliocene, has been reconstructed (fig. 17). The Seine was extended as far as the north of Brittany by the Channel stream, the Rhine received the Meuse and the Thames, and the Weser and the Ems were tributaries of the Elbe.



● Peat–bogs (submerged)

||||||| Areas exposed during the Wurm, today less than 160 feet deep (•••••)

Submerged river courses

FIG. 17. THE CONTINENTAL SHELF AND ITS "RIVER SYSTEM" IN THE ENGLISH CHANNEL AND NORTH SEA

Old peat deposits (radiocarbon-dated to the Early Holocene) dot the area "drowned" by the Flandrian transgression.

The Abyssal Plains

At about 16,000 feet, no more traces of continental structures belonging to the sial remain. The ocean floors are entirely composed of sima.[1] Normally they are covered by a thin layer of pelagic sediment; globigerina ooze in the shallower regions, and red clay in the deep and extensive ocean bottom. However, Ewing and his colleagues have shown the existence of networks of canyons on certain abyssal plains, accompanied by continental detrital sedimentation, composed chiefly of sands formed in shallow water (fig. 18).

[1] Sima: a term introduced by Suess, designating basic and ultra basic rocks containing *si*lica and *ma*gnesia of basaltic and peridotic type (density about 3).

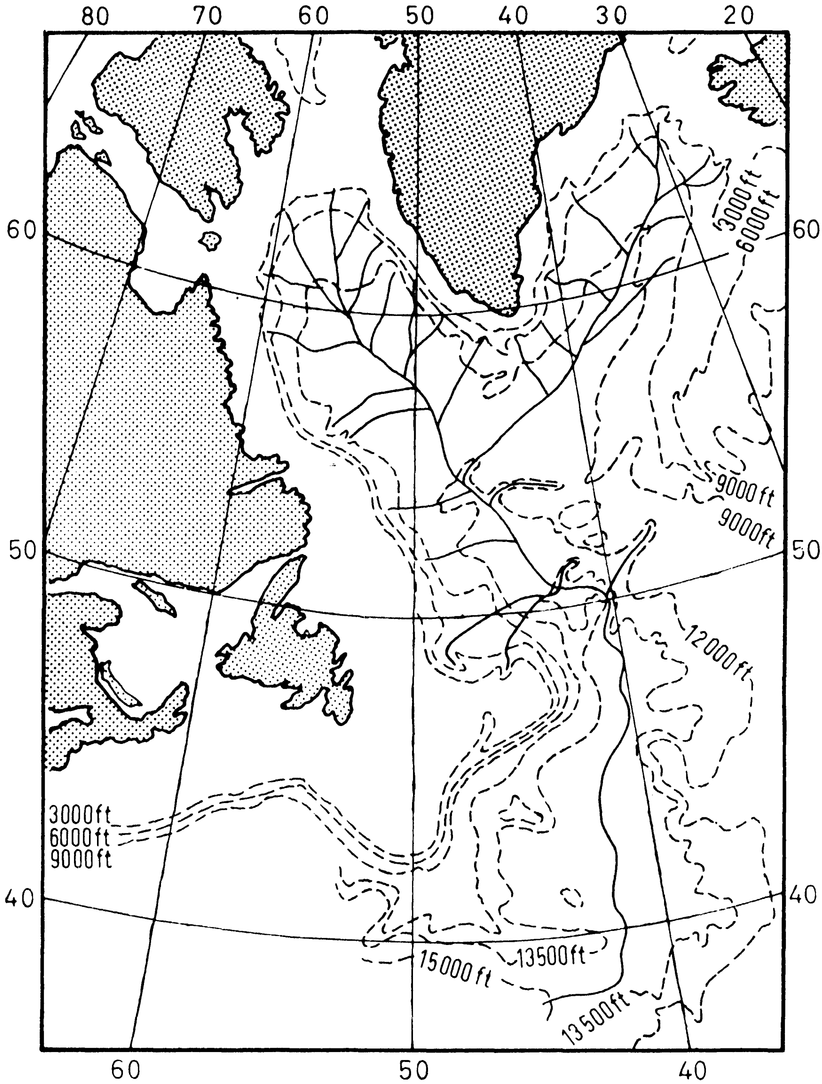


FIG. 18. THE MID-OCEANIC CANYON OF THE NORTHWESTERN ATLANTIC (after Heezen, Tharp and Ewing)

The main trunks of the Atlantic abyssal canyons commence in the Arctic region and then enter the abyssal plain of the Atlantic through narrow gaps. They appear to receive the canyons from the continental margin as tributaries. Considering the depth at which they are found, and the gentle slope of the abyssal plains, the only hypothesis which permits an explanation of these canyons and the detrital sedimentation which accompanies them, is that they are due to the work of turbidity currents. The abyssal plains display a relief (abyssal hills), much of which is due to volcanic and fissure eruptions, chains of seamounts and volcanoes such as those of Hawaii, and the mid-Atlantic ridge.

HISTORY OF THE OCEAN DEEPS

Much emphasis has been placed on the tendency of ancient shields to rise. On the other hand, the ancient sea floors had a tendency to sink. There is thus the contrast between the light sialic material which comprises the continents and the heavy sima material, which forms the bottom of the oceans. There is also the idea of continental flexure in the area between them. Since the Cretaceous some of the shields have risen up to 10,000 feet at their center, despite the counteraction of erosion. What, then, is the corresponding figure for the sinking of the ocean deeps? The only precise information on this question is furnished by seamounts (guyots). These are submarine mountains, generally volcanic, whose flat summits are suggestive of subaerial erosion. They sometimes also support coral formations which must certainly have lived in shallow water. But the flat summits of the seamounts, whose erosion goes back to the Early Cretaceous, are often found sunk today to a depth of more than 3,000 feet. The oceans cover more than seven-tenths of the earth's surface and it is likely that they were still more extensive before the advent of the major orogenies which have built up the continents. It is thus logical to say that their sinking (abyssal sedimentation not counterbalancing continental erosion) has been less than the elevation of the shields. It is also probable that these two opposing movements did not occur at a constant rate, nor did they regularly overlap. On the continent phases of stillstand can be reconstructed which have resulted in extensive planations; and conversely, even slight movements can be detected which have allowed wide transgression by the sea. Despite the fact that great progress has been made in recent years, our knowledge of the ocean floor is much less detailed than our knowledge of continental geology. When the epeirogenic history of the continents can be reconstructed and paralleled with the bathygenic history of the oceans, an enormous step will doubtless be made in the research into the fundamentals of geology (orogenesis, contraction and dilatation).

Some aspects of the fauna of the deep sea at the present time will be considered next. A. Brunn (1957) noted that most of the abyssal fauna are of very recent geological age. Menzies and Imbrie (1958) note that the abysses—the true abysses commence at 16,000 feet—contain no more archaic species than the epicontinental seas. Alone among them the Monoplacophore *Neopilina* (at more than 11,700 feet) is a deep-sea species. It seems, according to statistics drawn up by the authors, that each genus or family of benthonic formaminifers stay within a constant range of depth (with a slight descent for the Paleozoic genera). On the other hand the sponges, echinoderms, brachiopods and bryozoans, show the tendency of more recent genera and families to populate the deep sea (for example the deep sea corals). Thus, 86% of Tertiary genera, 14% of Mesozoic genera and no Paleozoic genera are found in the abysses. In all cases, the depth at which most relict fauna live, as well as the benthonic foraminifers, the bryozoans and the crinoids, rarely exceeds 6,500 feet.

From these facts it is clear that most forms which have survived since the Paleozoic era have lived in shallow water. In the tidal zone, the distribution of genera is as follows: 66% for the Tertiary, 30% for the Mesozoic and 4% for the Paleozoic.

In fact most of the Paleozoic forms have disappeared and are not taking refuge in the abysses.

The sea bottom shows considerable relief. If the topography which is obviously continental and unmodified is taken into account, there are a number of structures remaining, whose origins are the object of discussion. These structures are: the submarine mountains and canyons, the seamounts, and the mid-Atlantic ridge.

SUBMARINE MOUNTAINS AND SEAMOUNTS

(fig. 19)

Volcanoes form irregularities on the sea floor which, because there is little active submarine erosion, are well preserved, and can be easily seen. They appear as submarine constructions which are more or less conical in shape. Volcanoes are known with their lower slopes submerged but with their summits above water, forming islands. For example, the islands

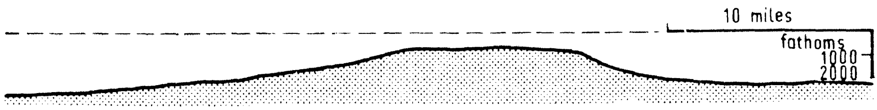


FIG. 19. SEAMOUNT TO THE NORTHEAST OF BIKINI (CENTRAL PACIFIC) (section after Emery, Tracy and Ladd, 1954)

stretching out from the Cameroons into the midst of the Gulf of Guinea have their bases at 6,000 or 10,000 feet below the level of the Atlantic. Such volcanoes experience pluvial and marine erosion on their subaerial summits. Should this process produce a peneplain, and be followed by a marine transgression, the submarine mountain would take the form of a truncated cone and would produce the form which is today familiar as a seamount (guyot). Guyots and submarine mountains extend along major lines of fractures in the Gulf of Alaska (Menard and Dietz, 1951), in the mid-Pacific (the mid-Pacific chain; Marshall Islands) as well as on the mid-Atlantic ridge (Heezen, Ewing, Ericson and Bentley, 1954).

Extensive landslides often occur on isolated volcanic cones and seamounts, leaving deep gashes in the sides and building up piles of debris on the lower slopes. This results in the circular pattern of the truncated cone being converted sometimes into a star shape (Fairbridge, 1951).

A diagrammatic map of the North Atlantic (Bruce C. Heezen and M. Tharp, 1957) also shows large seamounts: the Great Meteor at −1,469 feet and the Cruiser at −961 feet. Water-worn pebbles have been dredged from seamounts on the mid-Atlantic ridge which are today at a depth of over 1,000 feet.

Among other specimens collected from the mid-Atlantic ridge is a fragment of basalt coming from a depth of 14,000 feet which has a probable age (calculated by the helium method, Carr and Kulp, 1953) of 30 million years. This would indicate that the Tertiary basalt eruptions whose date should be verified by other methods, were followed by a period of emergence, during which erosion led to the formation of pebbles. Traces of lava flows have been located at a depth of over 3,000 feet.

The most famous seamounts are those of the Pacific. The mid-Pacific chain comparable with that of the Hawaiian islands, but today entirely submerged, is 1,724 miles long, and from 28 to 709 miles wide. It leaves Necker Island (Hawaii) and heads westward as far as 170° E longitude. Intersected by passages, 13,100 to 14,700 feet deep, it possesses sharp peaks, ridges and seamounts. The planed summits of the latter are at various altitudes. Work of the U.S. Navy, published by Hamilton (1956), has revealed that at least some of the seamounts are still covered by deposits often of reefy nature, and of Aptian or Cenomanian age. From this evidence they conclude that the "truncation" of these mounts, preceded the reef deposits. As a result of their discovery, it is possible to date the subaerial morphology as Cretaceous. A later subsidence caused the death of the Cretaceous reefs.

The seamounts of the Marshall Islands are not far from the end of the mid-Pacific chain, but they are not part of an alignment. They form isolated volcanic cones. Bikini atoll, a special example, is situated on a seamount linked by a submarine ridge to the Sylvania seamount. This was probably always deeper and was not covered by corals (Dobrin and Beauregard Perkins, 1954; Raitt, 1954) (fig. 154). The pattern of Bikini atoll, with its coral crest, shows landslide-type gashes and disturbances on the lower slopes suggesting slides as in the case of other oceanic volcanic cones (Fairbridge, 1951).

THE FRONTIER BETWEEN LAND AND SEA

Probability of a "Continental Flexure"

The tendency of continents to rise and oceans to sink suggests that at their junction an intermediary zone exists which takes the form of a flexure.

However, the "continental flexure" (Bourcart, 1926) is a structure which has been much discussed and which yet remains hypothetical for the most part. To it may be attributed spasmodic movements along the length of the continental border, and it would seem that submersion (transgression) of the continental margin could be the result of it. In fact, the flexure would give rise to epeirogenic movements which vary in amplitude according to the position of the axis. This may in fact be zero at the neutral line and it thus has effects distinct from those of eustatic changes of sea level, which are universal.

There is little evidence that a well-defined "continental flexure" is present, even potentially, at all junctions of continents and oceans. Where it is observed it represents a zone of weakness, which may coincide with mobile zones, some of which give rise to orogenies.

However, in an important article, Cotton (1955), searching for typical examples, notes that occasionally the linking up of river terraces resulting from uplift with the deposits of an estuary resulting from subsidence (example furnished by the work of Oestreich, 1938, on the Rhine) could be attributed, as by J. Bourcart, to a continental flexure with discontinuous movement. Alternatively, it might be that the coasts of such a region are unequally submerged because they are affected by an oblique flexure. However, it is also possible to attribute many such changes to orogenic movements. Nonetheless, it may also be said that where continental flexure is apparent and becomes increasingly accentuated, it can pass into a local geosynclinal orogeny.

Cotton (1955) distinguished three types of coast line: the stable coast line; the unstable coast line, which would be subject to a marginal flexure; and a tectonic coast line, terminated by a monoclinal flexure or even by a fault. On the stable coast line, it is difficult to show evidence of these deformations, but on the tectonic coast line they are strongly marked. This is the case in the region of Durban, South Africa, studied by L. C. King (see p. 34, figs 5 and 6).

The Coast Line

The coast line is not the true junction between the continents and the oceans. This is really represented by the edge of the continental slope. The coast line is an extremely variable line, which is apparent as the margin of the land above the sea at any given instant. Its position varies according to the state of the tide. In the geological history of a region it is rare to find any trace of it.

Rocky coasts often show cliffs which may be undercut at the level of the intertidal zone. Many fine examples are known in limestone regions from all parts of the world. Such overhangs are, like river gorges, susceptible to collapse. The cliff rises above a more gentle slope which sinks under the sea. This has been called the abrasion platform. The highest point of this surface, which is marked by a break of slope, is the limit of high water at neap tide. It often corresponds with the upper part of the beach, which is composed of gravels and sands, either fluvial or marine, or loams of continental origin. The lower parts of this material are carried into deeper

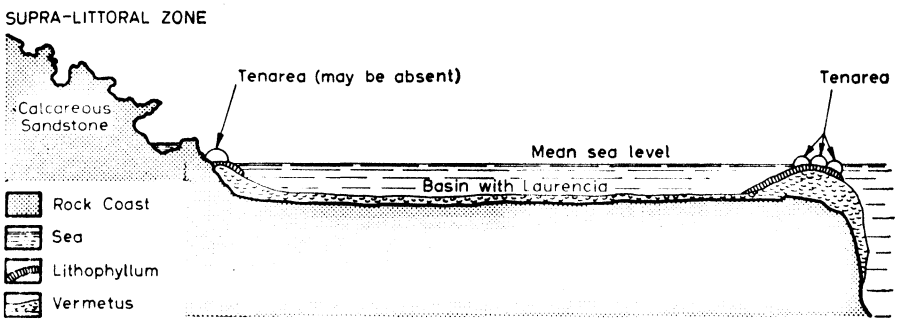


FIG. 20. SECTION ACROSS THE LITTORAL VERMETUS PLATFORM AT TORRE DEL ISOLA, SICILY (after Molinier and Picard)

FIGS. 21 to 23. THE LITTORAL PLATFORM NEAR TIPASA, ALGERIA (Photographs. G. Termier)

FIG. 21. MESOLITTORAL ZONE CORRESPONDING APPROXIMATELY TO MEAN SEA LEVEL, WITH THE SUPRALITTORAL ZONE ABOVE

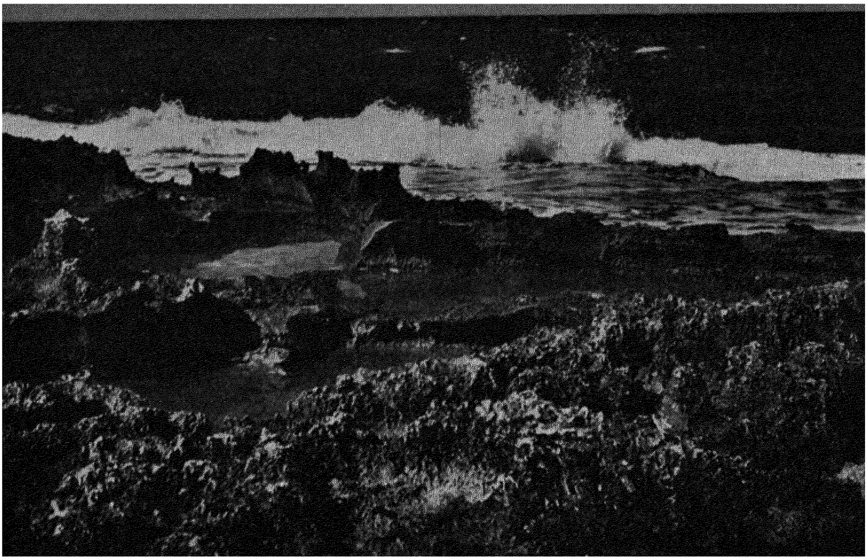


FIG. 22. GENERAL VIEW OF THE SHORE

In the foreground, the jagged supralittoral stage, and in the background, the smooth mesolittoral stage on which the waves are breaking.

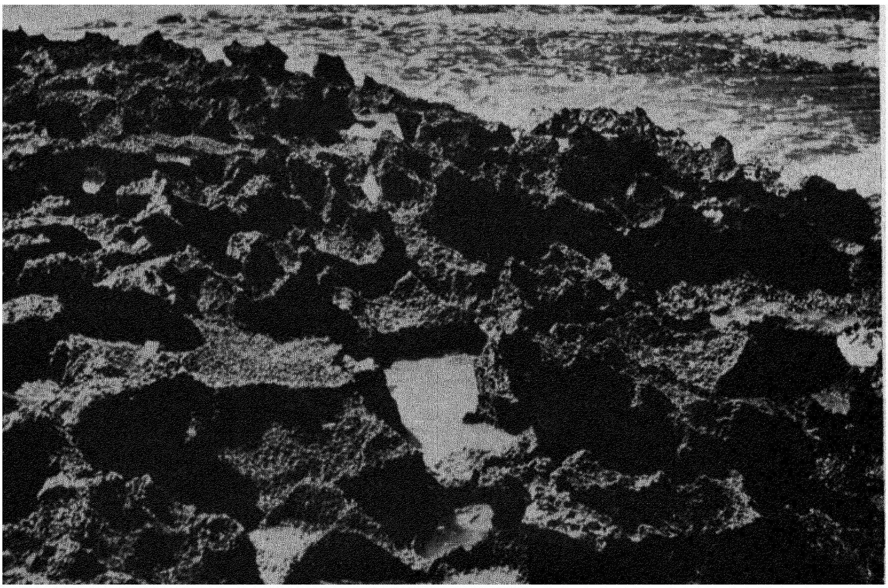


FIG. 23. DETAIL OF THE SUPRALITTORAL ZONE SHOWING ALVEOLAR EROSION (HONEYCOMB WEATHERING)

water by marine currents in a manner analogous to deltas advancing into the sea. J. Bourcart (1955) calls these beaches *continental beaches*, whereas *marine beaches*, which are further seaward, are generally composed of elements which differ from those of the adjacent continent, because they are transported by longshore currents and often contain marine animals or vegetable debris.

Marine Abrasion Platforms

The abrasion platform, which borders certain coasts is essentially the result of erosion by the sea. It is composed of a *supralittoral zone* pitted by honeycomb weathering, where the principal agents are spray, certain boring organisms, and the pH of marine pools. There follows a *mesolittoral zone*, forming a narrow bench which corresponds approximately to mean sea level. The Mediterranean and Atlantic coasts of Morocco (between Rabat and Casablanca) provide good examples of this type of platform. But these surfaces are generally destroyed by erosion and are rarely found intercalated in a geological sequence. The calcareous beaches and the consolidated dunes (eolian calcarenite) of which they are formed at least in part, disintegrate easily and are unlikely to be preserved for indefinite periods.

That part of the coast which has been worked by agents of the tidal zone is called the "strand". Its dimensions are very variable. Its width in the bay of Mont Saint-Michel is 12 miles. In Normandy, it has received the local name of *vey*. It attains the exceptional depth of 63 feet in the Bay of Fundy (Canada) but it is more often between 3 and 10 feet. Biologically the strand corresponds to the *mesolittoral zone*.

Above this zone, the supralittoral zone can be compared, in certain respects, with semiarid regions (Guilcher, 1954). Because it is battered by spray, it has practically no plant cover. Thus, the work of rainwash is unhindered. This zone is, moreover, drenched by heavy rainfall which must be added to the "heavy sea". The run-off of water received can total more than 500 inches per year. It results in conditions particularly favorable for erosion. The cliffs are not so affected, for their altitude preserves them from most of the spray, but they can experience a karst-like type of evolution. To this may be added the action of ice, but, more generally, wind action forms dunes analogous to those of deserts.

In these two zones at the margins of warm seas already saturated in calcium carbonate, solution of limestones occurs. This solution is due to variations of pH in the spray pools of these zones. The variations themselves are mainly referable to the diurnal and nocturnal activity of algae (Davy de Virville, 1934; Emery, 1946) which absorb carbon dioxide during the day and expel it during the night, and partly due to the temperature differences which act in the same way (Revelle and Fairbridge, 1957). Thus at night, the carbonate cement of calcareous sandstones is dissolved, the grains of quartz are loosened and are then detached by molluscs browsing on the tufts of algae. Emery cites the Littorinas of California, which, numbering about 2,600 in an individual count, have each been able to dislodge 0·3 gm. of grains in twenty-four hours, while providing, at the same time, dissolved CO_2 (fig. 40). Afterwards the waves clear away the material thus dissociated.

FIG. 24. MARINE ABRASION PLATFORM ON THE ATLANTIC COAST AT RABAT, MOROCCO, NEAR THE HOSPITAL OF MARIE-FEUILLET (Photograph: H. Termier)

In the foreground, the supralittoral stage can be recognized by its very rough surface, with hollows containing deposits of salt. In the background, the almost completely flat platform (with low algal rims) corresponds to the wave-cut bench (mesolittoral stage).

Under rather similar conditions, as the spray pools dry out, and often on nearby parts of the same littoral zone, there is, in contrast, precipitation of $CaCO_3$ which becomes the cement of calcareous sandstones. Examples of this occur on all tropical coasts, and detailed descriptions have been supplied from Australia, Florida and from the Pacific islands. The process is also particularly well developed on "beachrock" (calcarenite).

The edges of pools are specially corroded, according to the composition of the rocks which form them. A break-down of the minerals in igneous rocks is also observed under the action of spray. This disintegration is assisted also by the mechanical action of wetting and drying, and also, as suggested by J. Bourcart, by the crystallization pressures of marine salt incrustation. The least sensitive to this alteration are quartzites, as well as the joints of all rocks hardened by limonite. These two materials are thus likely to remain in relief. It may be noted in passing that the phenomenon of honeycomb weathering (p. 99) which is encountered at the sea margin outside the littoral stages proper, is also well known in arid climates. Very fine examples can be seen in the Ahaggar, where igneous rocks are hollowed out to form many taffoni (honeycombs). The role of water, and especially salt water, appears less important here than that of sand and deflation. Thus the question arises whether the same agent may be sometimes responsible for the erosion of the littoral zone as for the erosion of the shore. The latter is also the place of dune formation comparable with that of the desert, and is subject to very strong winds.

The littoral abrasion platform is fairly common. Examples are known all along the rocky coasts of California, Florida, the western Mediterranean, in the Red Sea, on the Atlantic coast of Morocco and South Africa, and in Australia. Its presence chiefly depends upon the nature of the rock attacked. It affects limestones almost as easily as the calcareous sandstones or calcarenites, resulting from consolidated dunes or "beachrocks". It is less clearly marked on coasts of granite, basalt, and of shales with or without alternating sandstone.

An abrasion platform is found on most coral reefs, because their calcareous material often has little secondary cement and renders them particularly vulnerable to this type of erosion. This platform is sometimes partly covered with a calcareous sand (forming small islets or sand cays); often it supports remnants of older eroded reefs as mushroom-shaped rocks or jetsam ("negroheads"). These are indicative of the vigorous erosion to which the reef surface has been subjected.

In Australia, the calcareous littoral platforms are particularly widespread between one and two feet above low tide level and are raised at their outer margin. This is often encrusted with calcareous algae. The platform itself is often almost perfectly smooth. Its planation is mainly attributed to biochemical and physiochemical action, which is linked with the amount of carbon dioxide in the sea water.

Displacement of Shore Lines

The shore line in the strict sense, is only observable at the present time or at a time very near to its "functional" state. It is formed by detrital sediments, more or less coarse-grained and easily transportable, so that when it ceases to be functional it is rapidly "obliterated". Moreover, if a rapid burial and the fossilization of certain shore lines occur, their remains are often difficult to interpret. Cobbles in a geological sequence may indicate a beach or a beach barrier, or they may indicate other geological phenomena, such as the uplift of a submarine fold.

The displacement of shore lines is of considerable importance in the history of the planet, but the wide transgressions and regressions established in stratigraphy and drawn on paleographic maps are only the culmination of a large number of small advances and retreats, where the detail is not known.

One of the chief arguments enabling the position of old shore lines to be determined is furnished by old cliffs, but it is sometimes difficult to distinguish these cliffs from nonmarine cuestas. One criterion would be "constant altitude along the foot of an escarpment for a considerable distance" (Guilcher, 1954, p. 27), but warped shore lines also exist, to which this criterion could not be applied. In the geological column, the problem of cliffs occurs in rather a different fashion. Unless the cliff is a truly modern one, it is not rare to observe in a well-dated marine sequence, whole sections tilted or otherwise, belonging to an older series. The fragments of the earlier formations must be the remains of a cliff or a small island attacked by the sea.

Transgressions

A transgression is the invasion by the sea of a land previously above sea level.

Many writers have thought that marine incursions, in the stratigraphic record, were the result of eustatic oscillations. In this way, R. Ciry (1954) explains major transgressions across continental areas, such as the one which occurred in the Cenomanian.

It has been shown (p. 23) that "glacial control" alone may be taken as an example of a true change in the *volume* of oceanic waters. Nevertheless, many displacements of shore lines which have taken place during geological periods are of a vertical magnitude often exceeding that which would have been caused by glacio-eustasy during the Quaternary. Explanations other than glacial eustatism must be invoked when large-scale glaciation did not occur. If most changes of sea level are considered, even as recently as the Miocene and Pliocene, the almost total absence of glaciation renders the notion of glacial eustatism inapplicable. Other factors must therefore be sought which are capable of producing the same effects. The terms tectono- and sedimento-eustatism have also been mentioned. Even in the Quaternary, a sinking of the continental shelf of at least 330 feet is superposed on glacial eustatism. This sinking could in part correspond to a movement of the continental flexure (Bourcart, 1926).

Local transgressions also occur in geosynclinal belts and local basins. The history of the Carpathians (Termier, 1957, p. 706) and the Mediterranean (Termier, 1957, p. 773) has furnished good examples.

The condition of a subaerial surface at the moment of transgression has a certain geomorphological importance. When encroaching upon an irregular land surface, the transgression, like an impounded lake, fills the low areas first. It penetrates river mouths and produces estuaries, rias, and fjords, as the case may be. It is an *ingression.* The Flandrian transgression is a good example of this process. When a transgression invades a surface which is worn down to a peneplain and is practically without relief, it advances as a sheet with a continuous front, so that even a thin layer of water may cover vast areas. Such a transgression explains the recent history of gulfs in northern Europe such as the Wadden Sea (of northern Holland and Germany) or even the whole Baltic.

This last statement leads more or less implicitly to the concept of continental areas. E. Haug (1900) opposed the idea of transgressions due to earth movements in continental areas, in favor of transgressions generated in the more localized realm of geosynclines.

In fact, modern knowledge no longer favors the idea of a contemporaneity between the transgressions of different geosynclines. Each one evolved according to its own history and the transgressions only occurred during certain phases of this drama (H. and G. Termier, 1956, pp. 215–222).

If the surface is visualized as covered by seas, it is the transgressions upon continental areas which are the most important and have lasted the longest time. As has already been noted, most of them were accompanied by hot and humid climates, suitable for the development of life.

The authors have shown (H. and G. Termier, 1952) that the migrations of marine invertebrates were stimulated at the time of these transgressions on to the continental areas. Such migrations usually followed the displacement of bodies of warm or cool water.

At the time of the *Tethys transgressions* the paleogeography shows that the waters encroached northward and southward from Tethys, followed by a huge spread of organisms. During the *arctic transgressions* which can be distinguished in the northern hemisphere, and which immediately followed the Tethys transgressions, the displacement of oceanic water was toward the south, that is, toward Tethys, bringing with it a retinue of invertebrates known only up to then in boreal regions. The nordic fauna did not consciously migrate to the southern regions, but their distribution was only made possible by the presence of water sufficiently warm to allow free-swimming larvae to live longer before becoming fixed (H. and G. Termier, 1952, p. 34).

Regressions from the principal continental areas, which had less sedimentary impact than the transgressions, seem to correspond with marine invasions of the *circumpacific* regions. These invasions had an orogenic origin. Moreover, on the emerging continents, the climate was often harsh, arid, and sometimes followed by glaciations.

The extent of the great transgressions of geological history is much wider than the glacio-eustatic transgressions of the Quaternary. They may affect the major part of the old platforms as well as the continental shelf. Certainly, the opposing tendencies of the rising of continents and the sinking of ocean floors must have played a great part in the halting of transgressions and the commencement of regressions. In order to account for glaciations, a general uplift of shields has been invoked and in the same way, to explain the major transgressions, a general uplift of ocean floors must have occurred concurrently with the erosion of continents. In this case, as noted by J. Bourcart (1955) deformations of the earth's crust must have occurred simultaneously. In the authors' opinion, epeirogenic movements are mainly responsible.

Geodetic transgressions and regressions are also visualized by Fairbridge (1961). These could result from the immediate adjustment of the hydrosphere to the geophysical requirements of a rotating spheroid, whenever there is a shift of the poles relative to the lithosphere, such as is envisaged by the paleomagnetic investigators. It is calculated that if the pole of rotation migrated by 1° of arc along a certain meridian, at the intersection of that meridian and the new equator there will be a rise of sea level (maximum) on opposite sides of the globe; on the equator at 90° from these maximal nodes there will be no change. At each of the new poles and their related quadrants there will be maximal *regressions*. It is possible, therefore, that some of the great transgressions and regressions of geological history, such as the Tethys oscillations, and other cyclic tendencies, may be only associated with certain quadrants of the globe, and thus not strictly eustatic or epeirogenic but geodetic. Thus a transgression in two quadrants is matched by a regression in the two opposite ones.

As mentioned earlier, nearly all of the continents today are surrounded by a submerged continental shelf, on which traces of subaerial erosion, river channels and courses, and mountain ridges can be recognized.

Two periods comparable with the end of the Quaternary epoch are known. They are the Early Cambrian and the Permian. These were two periods of general transgression, but they were limited in extent. They followed important glaciations, which were contemporaneous with general regressions. During the marine retreats, the continental shelf was almost completely above sea level and the shore lines must have passed abruptly to great depths, via the continental slope.

The reduction or disappearance of practically the whole neritic zone established on the continental shelf, implied by such a fall of sea level, has not been sufficiently emphasized. *During the course of the regression contemporaneous with the glacial maxima, the neritic fauna and flora were practically ousted from their living space.* The succeeding transgression created a new continental shelf and thus a new neritic zone for which the potential fertility was very much diminished. Of the old neritic flora and fauna only impoverished "reserves" remained and repopulation was mainly due to an influx of pelagic larvae from the open sea.

Since organic productivity on the continental shelf is the principal control of limestone sedimentation (provided that the temperature is adequate), the alternation between "high" and "low" sea levels corresponds also with times of high and low carbonate sedimentation (Fairbridge, 1955).

Thus it can be seen that there is little in common between the glacial eustatic transgressions, which are generally of minor extent, as in the case of the Flandrian transgression, and the major transgressions which occurred throughout the great geological eras. These took place according to a well-defined rhythm. They commenced with an overflow from Tethys and not by an expansion of polar influences, which would have been the case, had there been melting of the Arctic and Antarctic ice caps. The latter transgressions favored certain regions and certain biotopes, in turn, but there was always a great potentiality present for refertilizing the sea bed. They covered vast surfaces, but they were not universal. They were only partial movements of ingression and regression over the shelf and continental areas and did not cause the complete exposure or immersion of these two zones everywhere in the world.

Thus, during general transgressions due to glacial eustatism, the installation of marine biotopes must have occurred progressively on practically virgin ground. This took place by successive associations as seen in the present-day nonalgal marine plants and reefs.

Present Coast Lines

In concluding this chapter devoted to the frontier between the continents and the oceans, it may be said that due to glacial eustatism, the consequences of which have probably been exaggerated, the present is a period of almost general transgression, the *Flandrian transgression*. As Guilcher (1954) has said, "coast lines all over the world are thus coasts of submergence, except where very recent tectonic uplift of a magnitude exceeding that of the eustatic transgression has occurred, and an exception is made for shore lines affected by postglacial isostasy".

Turning back in geological time, there probably were periods comparable with the present in the Early Cambrian and the Permian. But most eras have been characterized on the one hand by regressions, and on the other hand by nonglacial eustatic transgressions. The concept of shore lines with quite another form from our own thus arises. The present forms of our shores show a few well-defined types which aid in the reconstruction of ancient landscapes.

The geographer Gulliver (1899) proposed a subdivision of coasts based on their *initial* forms, where the relief and the particular form would have had a continental origin, while the *sequential* forms would result from profound modifications caused by the action of the sea. Most present-day coasts seem to fit well into the second category. On the other hand the former are exceptions today because they are barely conceivable. These exceptions could have developed during almost instantaneous major modifications of the paleogeography, for example when the sea drowns a continental landscape in the manner of a dammed lake invading a valley. This is a purely theoretical idea which does not take into account the sucession of geographical changes which have passed almost imperceptibly from one state to another during the immense span of geological time.

An Enigma: the Carolina Bays (figs. 25 and 26).—The coastal plain between northern Florida and New Jersey contains a number of regular elliptical depressions, each about 5,000 feet long, orientated N. 45° W., filled with marshes, and fringed by low sandy ridges. These are "Neptune's racetracks". A number of interpretations have been offered to explain these amazing structures. The most interesting, suggested by W. Cooke (1954), is connected with the earth's rotation, because of the particular direction of the long axes of the ellipses. This elliptical form might arise from the gyroscopic effect of the earth's rotation on a lake, where all the water would have taken part in a gyratory movement, and created a whirlpool.[1] The force maintaining the action of these whirlpools would have been furnished by the tides, while a connection existed between the lagoons and the sea, that is, while the position of the "bays" remained on the coastal plain near to sea level. It may also be held that the "bays" were produced at different times, since they occur at seven different levels, the youngest still being at sea level. The most typical were formed during the time of the Talbot sea level (see p. 25).

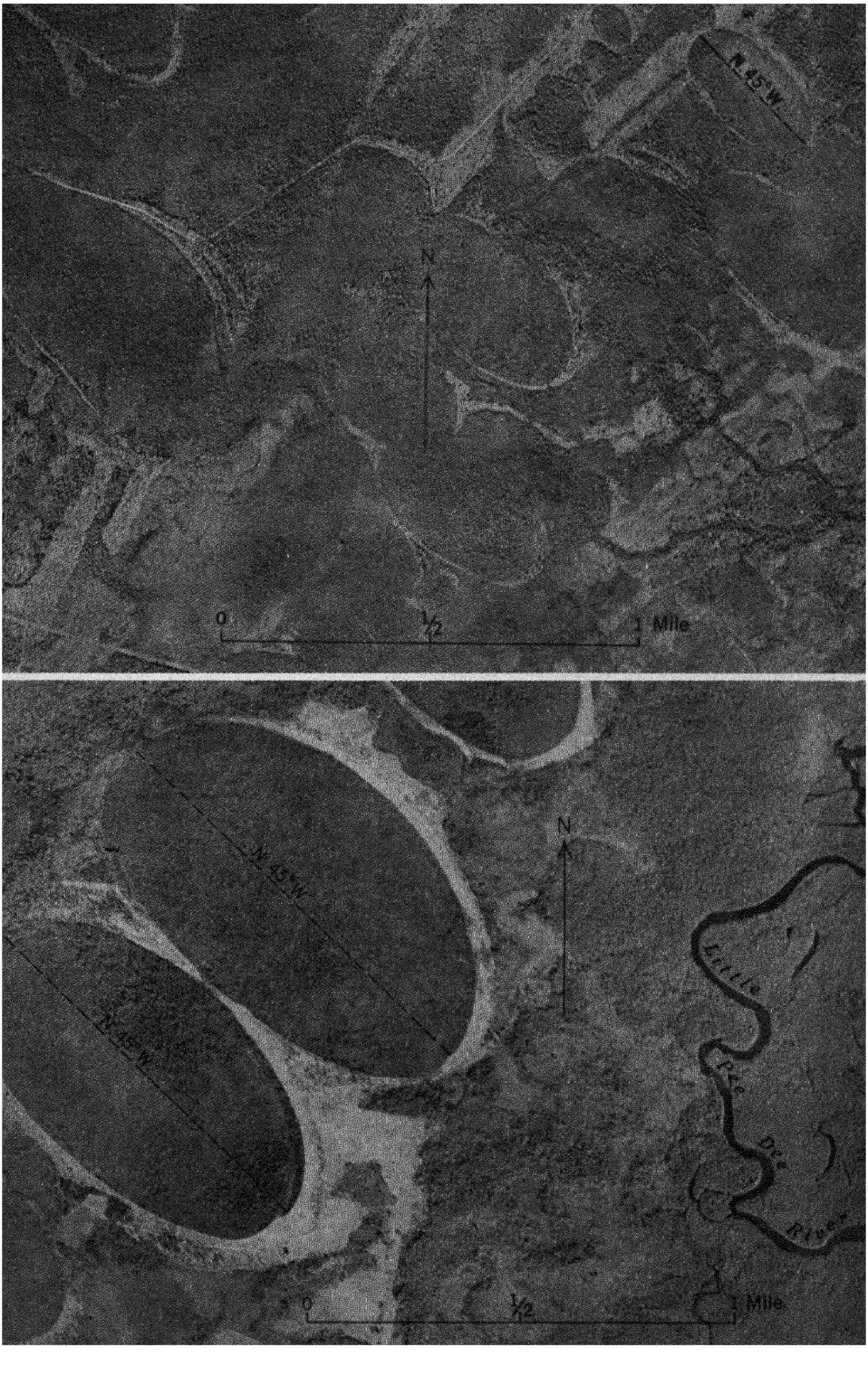
The whirlpools could have originated either in two marine currents with different directions, or between a current and a calm region, or at the place where a current passed an obstacle. The tidal currents, in effect, produced whirlpools when they reached stretches of calm water. In the same way, there are reversals of currents in estuaries.

The growth of terrestrial plants and algae has gradually transformed the lagoon into a salt marsh enclosed by an elliptical sandy ring. If the level of the sea falls, the marine marsh passes into a freshwater marsh, which will be described later (pp. 181–183).

[1] The dominant orientation of the long axes of these ellipses is, in fact, comparable with that of the trade winds which are universally held to be affected by the rotation of the earth.

FIGS. 25 and 26 (*opposite*). TWO EXAMPLES OF "CAROLINA BAYS" ATTRIBUTED TO WHIRLPOOLS ON A TIDAL FLAT

Above: view of part of Horry County region, South Carolina. Below: the Pee Dee Islands, Marion County, South Carolina. (*Geol. Surv. Prof. Pap.* 254–I, kindly provided by C. Wythe Cooke.) The sea is situated to the southeast of this region at the present time.


N 45°W
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4

Erosion

A. EROSION IN RELATION TO SEDIMENTATION

Erosion and sedimentation are the agents which, under the influence of climate, epeirogenesis and orogenesis, have molded the surface of the continents.

The former is essentially the alteration of rocks by meteoric action; the latter is the transportation and deposition of the material freed by erosion. Gravitational forces play a very important role in sedimentation.

Moreover, subsidence, a common phenomenon of the earth's surface, can be initiated by the weight of a heavy load of material such as lava, ice, or an accumulation of sediment. Sedimentary loading takes place particularly beneath deltas on the continental shelf.

Perhaps insufficient notice has been taken of the fact that subsidence has occurred throughout geological time, and that there is subsidence, not only of the ocean bed and continental shelf, but also of the continents themselves, especially in basins and certain alluvial plains which receive an abundance of sediment. Subsidence generally has effects on sedimentation and causes localized thickening. Fine examples are provided by the paralic coal basins centered on the continents, and by coral reefs such as those in the Devonian of Belgium (p. 288), the mid-west of America and in N.W. Australia.

In the case of a paralic basin, the bottom tends to remain more or less constant with respect to sea level, or oscillatory, throughout its evolution. Subsidence of the sea bed is generally contemporaneous with the elevation of an adjacent continent.

B. AGENTS OF EROSION

Erosion is the attrition of the superficial part of the earth's crust.

E. Haug (*Traité*, I, p. 406) wished "to reserve this term for the accessory phenomenon which accompanies transportation by water, and which consists of attrition of the bed and banks by the material thus transported". But his example has not been followed.

Erosion is a matter of common observation and is essentially characteristic of the superficial evolution of the earth's crust by the mechanical and chemical actions of the hydrosphere (rain and rainwash, glaciers, the sea, humidity and evaporation); by the action of the biosphere (plants, animals, man); by the action of the atmosphere (wind); and by variation of temperature. It is one of the principal external dynamic agents of geology. Its consequences are everywhere considerable. In the first place it is responsible for all the land forms visible on the surface of the continents. These *forms are all of a transitory nature.* The study of the geological history of any region of this planet shows that topographic relief is in the process of continual modification. In the second place, since the material cannot be lost from the globe, the immediate result of erosion is the production and deposition of the material which it has dislodged. *Erosion begets sedimentation.* This pair assume prime importance in the mechanism of the geological drama, previously defined (Termier, 1956, pp. 215–222).

The principal agents of erosion will be mentioned in passing, not so much to describe once again phenomena which are well known to all, but to inquire into their limits, variations and processes. During this review relationships between various phenomena will become apparent.

1. *Agents of the Hydrosphere*

The hydrosphere in its many forms (the sea, rivers and rain) is largely responsible for erosion.

(*a*) **Rain.**—Rainwater includes traces of various ionized elements: Na, K, Ca, Mg, HCO_3, Cl, Br, I, SO_4, PO_4, O_2, O_3, of which a few, particularly the halogens (Cl, Br, I) and the alkalies (Na, K) appear to come directly from sea water. According to Tamm (1953), the acidity (pH) of rain can vary between 5·9 and 4·3. It is increased in industrial regions by the acquisition of SO_3, iodine and fluorine from smoke (Gorham, 1955). The NO_3 ions which are abundant in rain, particularly in tropical regions, have been attributed to the electrical fixation of atmospheric nitrogen (Hutchinson, 1954) or to the oxidation of ammonia (Hoering, 1957).

The role of atmospheric ozone in the composition of rain water is certainly considerable. It appears to oxidize the halogens and thus fix them, then to liberate them in the gaseous state (Caeur, 1938). Behne (in Correns, 1956) notes an increase of iodine and bromine relative to chlorine (the ratio Cl/Br, which is 292 in sea water, is 87 in rain water: sampled at Göttingen). In this way, according to Conway, rain deposits annually $0{\cdot}69 \times 10^{14}$ g. of Cl over the continents. Another effect of the ozone is the polymerization of free radicals, emitted either by industrial smoke or by plants such as the conifers, which would result in a hydrocarbon fog (*smog*, studied by Went). These hydrocarbons would be carried to the soil by rain.

(*b*) **Rainwash** (figs. 27 to 31).—Rainwash is the principal eroding agent of the hydrosphere. No rock, however hard, can resist it, for it acts in a number of ways: by mechanical force, by solution and by abrasion with transported material.

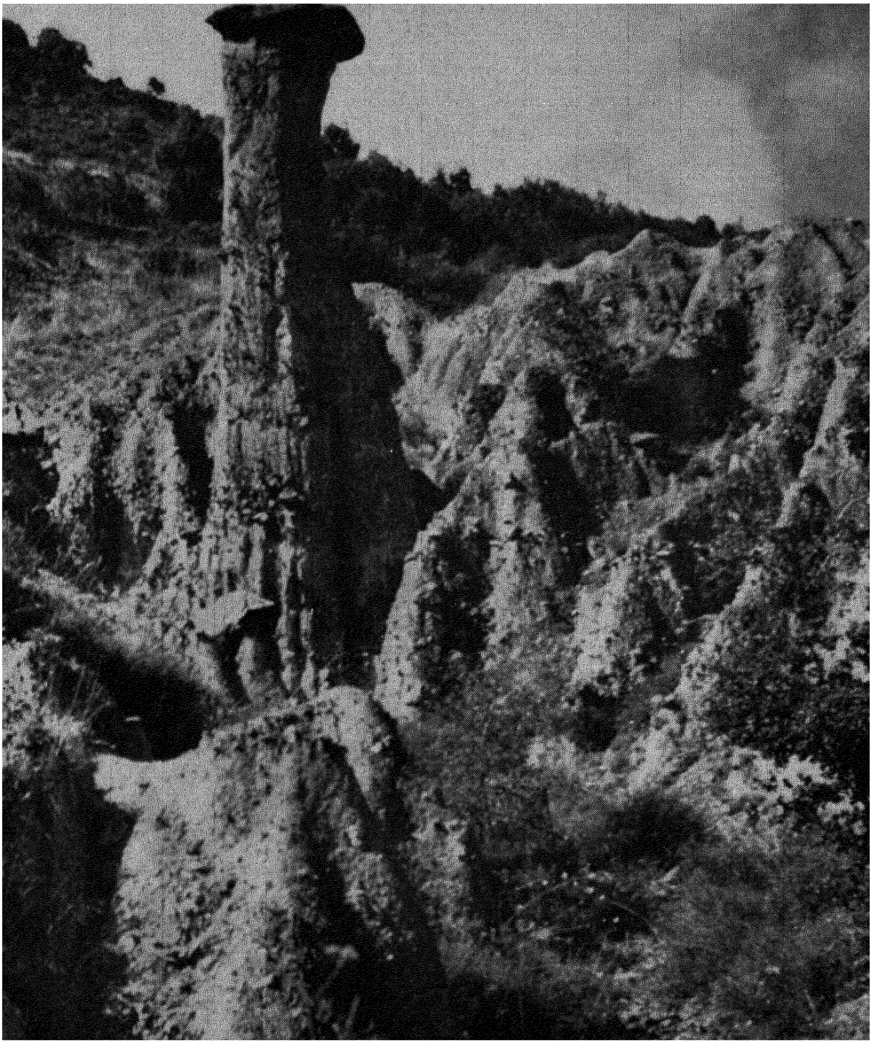


FIG. 27. TYPICAL LANDSCAPE OF "BADLANDS" WITH EARTH PILLARS ("CHEMINÉES DE FÉES"), TÊT VALLEY (EASTERN PYRENEES, FRANCE) (Photograph: G. Termier)
Note the stratification of the pebbles.

The term rainwash is not intended to imply stream networks, that is, river systems with beds which are provisionally fixed, or thalwegs, where the water passes gradually from numerous small trickles to larger ones, and is eventually concentrated in rivers and streams. Such networks represent an organized streaming impelled by gravity and adapted to the geological and climatic type of the region. Rainwash, even in its primitive form, the

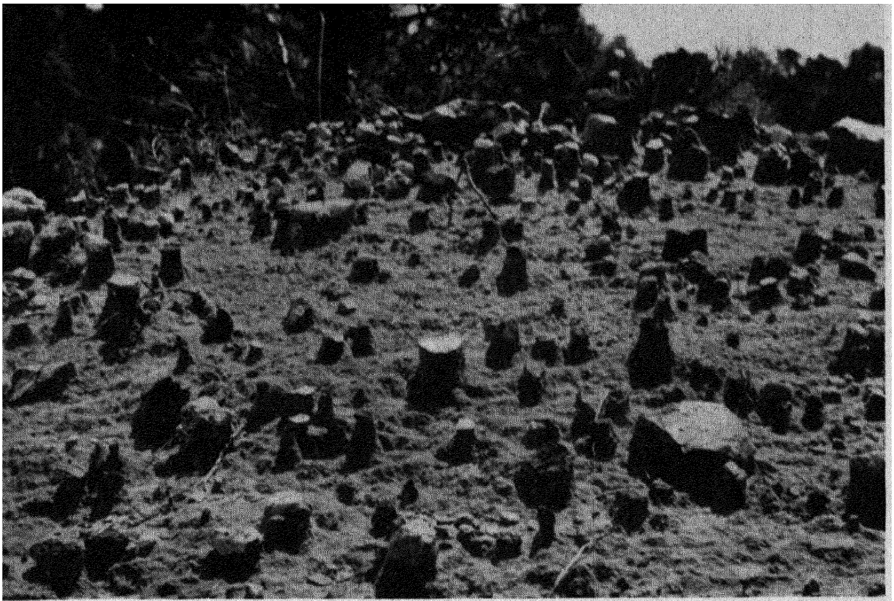


FIG. 28. SMALL-SCALE EARTH PILLARS IN A SANDPIT AT BOUCHES-DES-RHONE, FRANCE, TO THE WEST OF BEDOIN, NEAR CARPENTRAS (Photograph: G. Termier)

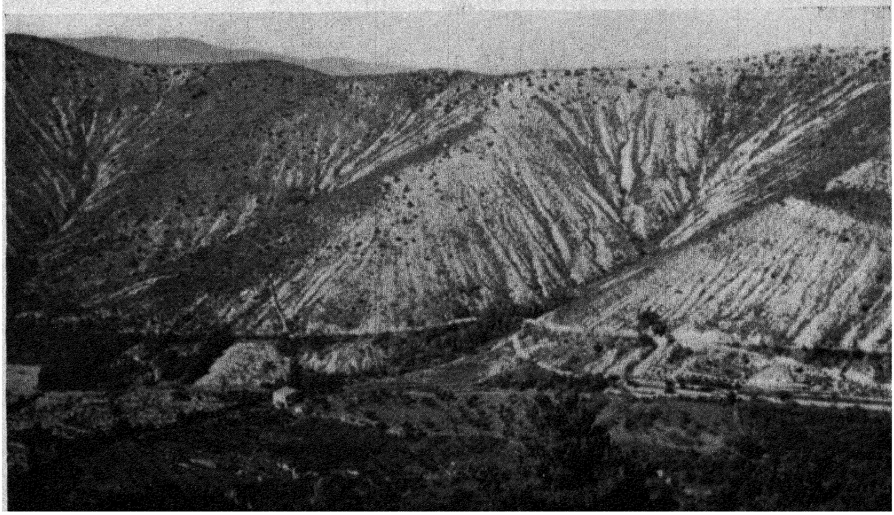


FIG. 29. GULLIES IN VERY SOFT SEDIMENTARY ROCKS IN THE TECH VALLEY AT THE FOOT OF CANIGOU (EASTERN PYRENEES) (Photograph: G. Termier)

Rainwash almost always follows the lines of steepest slopes.

FIG. 30. GULLYING. NOTE THE COMMENCEMENT OF THE CANYON

FIG. 31. NOTE THE LATERAL RAVINES LEADING TO THE CANYON

FIGS. 30 and 31. THE FORMATION OF GULLIES OR RAVINES IN THE QUATERNARY LOAMS OF THE EL MERS REGION (THE NORTHERN CENTRAL ATLAS), ALGERIA

(Photograph: H. Termier)

pluvial state, is a very important agent of erosion. For example, in loose ground the *unrestrained waters*, which have just fallen and are not channeled into pre-existing thalwegs, cause considerable degradation, particularly in semiarid regions, where they can isolate small pillars and ridges, or "badlands" and earth pillars (figs. 27 and 28) on slopes composed of soft rocks. On extensive nearly flat surfaces, rainwash selects its direction of flow more or less haphazardly in temporary courses which become connected in an amazing network which dries up when the downpour is over. This happens

FIG. 32. THE SUPERFICIAL NETWORK OF RAINWASH CHANNELS ON THE SURFACE AT TADEMAÏT (SAHARA), SEEN FROM THE AIR IN 1952 (Photograph: G. Termier)

on the surface of the rigorous desert region of Tademaït (fig. 32). Finally, very heavy rainstorms give rise to temporary lakes in hollows, or to sheets of water in which sedimentary material previously accumulated, is reworked and redeposited. To these water sheets, the name sheet-flood has been given (see p. 37).

The mechanical role of falling rain is particularly important on soft rocks on slopes. It has a considerable effect in arid regions where there is very heavy rainfall after long periods of drought. Branching ravines edged with smalls cliffs, channels and, rarely, tunnels appear. Together these constitute "badland" topography. Sometimes they cause earth flow or "creep" (less than 3 cm. a year). The salt mountains of North Africa are an extreme case. These are sculptured by the rain into delicate shapes which are quickly washed away.

"Badland" topography is common in arid regions of soft sandstones and pebble beds. Their sculpture is essentially due to rainwash by the rain falling on fairly steep slopes, sometimes producing ravines and ridges with extraordinary ramifications. Fine examples occur in the United States (Arizona, Dakota, Utah and Wyoming) and even in dry and sparsely wooded regions of France, such as certain mountain spurs of the eastern Pyrenees (the Têt valley) and several areas in Provence (Roussillon, near Apt). In Morocco, there are "badlands" which have been formed in the saline clays of the Permo-Triassic.

(*c*) **Rivers.**—Rivers form after the rainwash stage (the latter being important only on bare ground). The grandeur of rivers, as they sweep between embankments in large towns, must not hide the fact that they are the natural outlets for water which has fallen upon continents. They follow the line of greatest slope, because they obey the laws of gravity. The stream pattern may be completely altered if a new obstacle suddenly appears, if an epeirogenic or orogenic movement changes the position of the line of greatest slope or if erosion finally removes certain barriers. Thus, since the end of the Tertiary, in less than a million years, the courses of the Mississippi and of the Zambezi, have undergone directional changes as extreme as a reversal of flow. During historical time, the wandering of river courses is a common occurrence in geography. It happens in the rivers of closed basins, such as the Tarim, and in great rivers such as the Mississippi and the Hwang Ho. The mouth of the latter has shifted over 300 miles during the last 2,000 years. The flow of streams depends on elements which are called "constant" but from the geologist's point of view their temporary and precarious nature must be emphasized (figs. 33 and 34).

(*d*) **The Sea.**—The role of the sea in erosion seems unimportant beside the work of rainwash. Wave action is the chief erosive agent on coast lines. It attacks the cliffs, and above all, removes material, most of which has been transported by streams. On the sea bottom, the abrasive work of currents is considerable, and those marine zones free from contemporaneous deposits ("hard-grounds") may be attributed to them. However, their sedimentary action is more important because they often carry material along the shore. Turbidity currents (see p. 202) heavily loaded with sediment may form deep gashes in the sea bed. It has been pointed out that they are responsible not only for the hollowing out of channels but also for the cutting of submarine canyons.

The volume of material removed by marine abrasion from a small peninsula on the Atlantic coast of France (Cornard) has been estimated at about 16 million cubic yards in eight centuries. This material seems to have been completely utilized by sedimentation to build up bay mouth bars and "debris", which, by filling up the bays, has smoothed out the coast line. In the present creeks, between the islands of Oléron and Ré, and the whole of the Straits of Antioch, the amount of detritus in suspension (essentially argillaceous, because it is derived from the reworking of the Kimmeridgian marls (Mathieu, 1954)) is estimated at approximately 2·5 cm^3 for 1000.

(*e*) **Phreatic Waters.**—The phreatic zone is of considerable importance in the weakening and solution of superficial elements of the earth's crust. This action by underground water has been studied by Chebotarev (1955) in the great artesian basin of Australia. He has compared it to that of other subterranean water elsewhere in the world, in oilfields, mud-volcanoes and geysers.

Most of this water comes from rainwash. It sinks into fissures of well-defined catchment areas, but some of the water is held within the sediments.

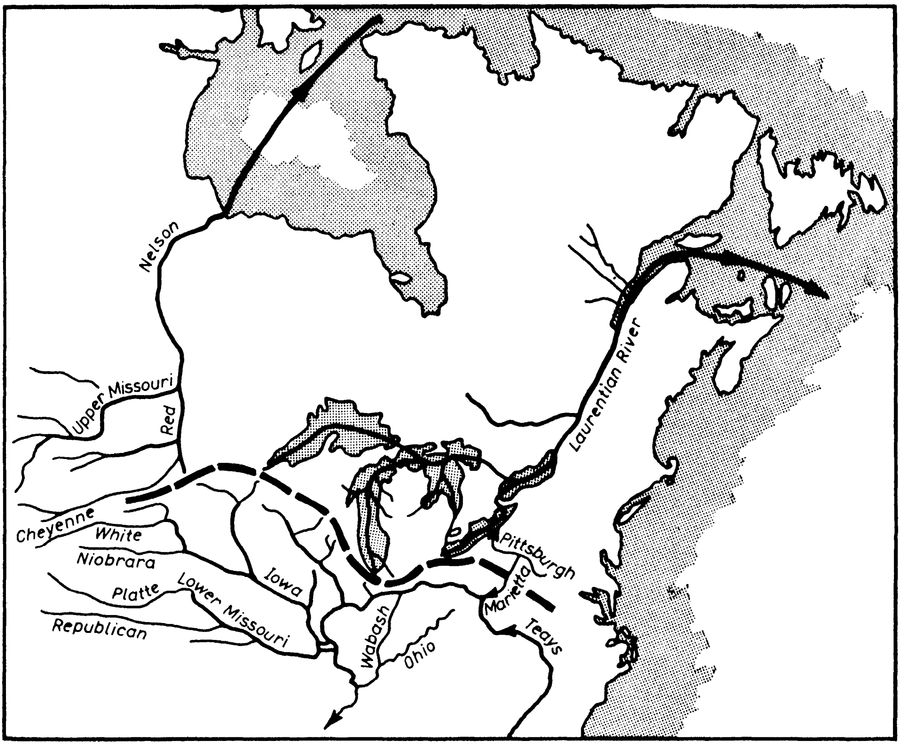


FIG. 33. PRINCIPAL PREGLACIAL RIVERS OF THE CANADIAN SHIELD

It will be seen that (p. 193) the aquatic muds contain a great deal of water, part of which succeeds in escaping by fissures during compaction, the rest remaining within the sediment. The names "connate waters" and "fossilized brines" have been given to these waters, which undergo what might be called a veritable diagenesis (see p. 335). They probably evaporate into gases included in the rock, particularly those of the hydrocarbon facies. A minute fraction penetrates by capillary action into those rocks in contact with the water table (the *capillary fringe*). These waters, are generally very rich in salts. One of these types of sedimentary waters is represented by the chloride solutions which nearly always accompany petroleum and natural gas.

Waters penetrating the surface of the lithosphere from the hydrosphere, particularly during the cycle of meteoric alteration, must also be taken into account. This cycle shows a series of states from the disintegration of transported detrital material to the ionization of the molecules during chemical solution. The elements of erosion can be recognized there. The soluble salts are all in the ionized state: cations (Na^+, K^+, Ca^{++}, Mg^{++}, H^+, Al^{+++}, Fe^{++}, Fe^{+++}) and anions (HCO_3^-, CO_3^{--}, Cl^-, SO_4^{--}, and subordinate NO_2^-, NO_3^-, SO_3^{--}, OH^-, F^-, SiO_3^{--}) and finally NH_4^+.

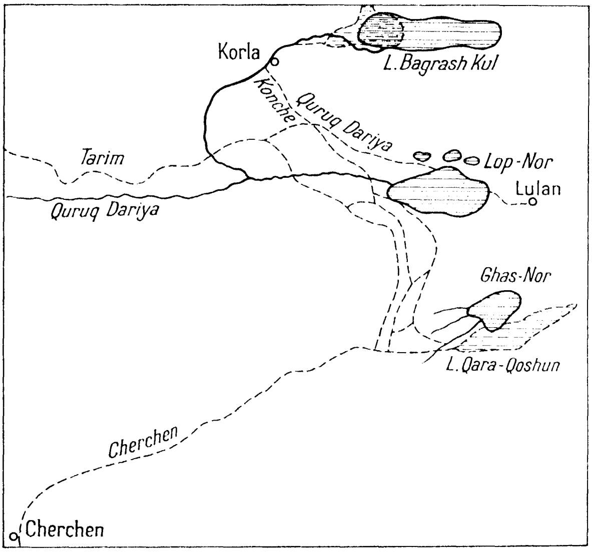


FIG. 34. THE DISPLACEMENT OF THE TARIM AND THE LOP-NOR DURING HISTORICAL TIMES (after an ancient Chinese map *in* Wagner and Himly)

The dotted lines show the present lakes and rivers.

The natural waters contain CO_2 or carbonic acid, H_2CO_3, which, being unstable, decomposes to give either H_2O and CO_2 or H^+ and HCO_3^-, from which bicarbonates can be formed. H_2CO_3 is a solvent which readily dissolves carbonate rocks.

However, the water table (differing from the surface waters) impregnates the rocks where it occurs, thus creating a permanent zone of absorption, whose surface varies only with the height of the hydrostatic level. Nevertheless, this flow is only appreciable in the upper part of the water table. Its lower part, in which there is less water, moves slowly, and justifies the term *static zone.* The upper part of the phreatic zone, which has a low concentration of salts and where movement is relatively rapid, constitutes the *zone of deposition, or discharge*. The surface of this zone fluctuates. The term *zone of cementation* as used by ore geologists corresponds to this zone of deposition. In the zone of cementation there is decalcification, and impoverishment in magnesium, while the ore minerals (gold, silver, copper, cobalt and manganese) are often concentrated in sulfides of secondary origin.

Deeper down, in the static zone, the phreatic water has a higher concentration of chloride, sodium and sulfate ions, while most of the carbonates are precipitated.

Naturally the concentration of different elements in the phreatic zone fluctuates with the varying composition and decomposition of the rocks bathed by these zonal waters.

Chebotarev (1955) distinguishes three principal geochemical categories of subterranean waters; *bicarbonated* waters, *sulfated* waters, and *chlorinated* waters. The *acid* waters including sulfuric or hydrochloric acid must also be mentioned. Bicarbonated waters are especially related to the residual detrital phase of the ortho- and para-eluviums (p. 137); sulfated and chlorinated waters are associated with the accumulation of calcium carbonate and chlorosulfates.

The phreatic waters will be referred to again in the section on soils (p. 136).

(*f*) **Artesian Basins.**—The most important phreatic waters, especially in arid countries, are artesian basins. Among these may be distinguished: *open basins*, emptying into rivers and lakes situated in the low areas; *half closed basins*, where drainage affects only half the basin; and finally *closed basins*, where the drainage behaves as in endorheic basins.

(*g*) **Chemical Composition of River and Ocean Waters.**—Polynov (1937) compared the percentages of various elements carried by river and ocean waters: SiO_2, Al_2O_3, and Fe_2O_3 may be considered as limited to rivers and to amount to 12·8, 0·9 and 0·4% respectively. It must be noted that these oxides are not soluble, but are generally found in the form of very stable salts. All other substances are ionized; the river water contains[1] 14·7% Ca, 4·9% Mg, 9·5% Na, 4·4% K, 6·75% Cl, 11·6% SO_4, 36·5% CO_3; sea water contains 1·19% Ca, 3·72% Mg, 30·59% Na, 1·1% K, 55·29% Cl, 7·69% SO_4, and 0·2% CO_3. Recognizing that CO_3 is gaseous it can be noted that ionized chlorine, which is much more abundant in the sea than in rivers, is also much less stable, and its concentration may be attributed to this instability. It should be noted that the solubilities of the cations are also very varied; magnesium is five times more soluble than calcium, and sodium is more soluble still.

This being so, waters may be classified according to their concentration of salts. The Russian authors Vernadsky (1933) and Chebotarev (1955), distinguish *fresh* waters whose concentration does not exceed 1%, *brackish*

[1] Percentage of total dissolved material.

or *rather saline* waters between 1 and 3·5%, and *brine* solutions whose concentration exceeds 3·5%. These waters are normally stratified according to their specific gravity, and flow in layers of definite density.

(*h*) **Sea Ice** (figs. 35–38).—The present-day distribution of circum-

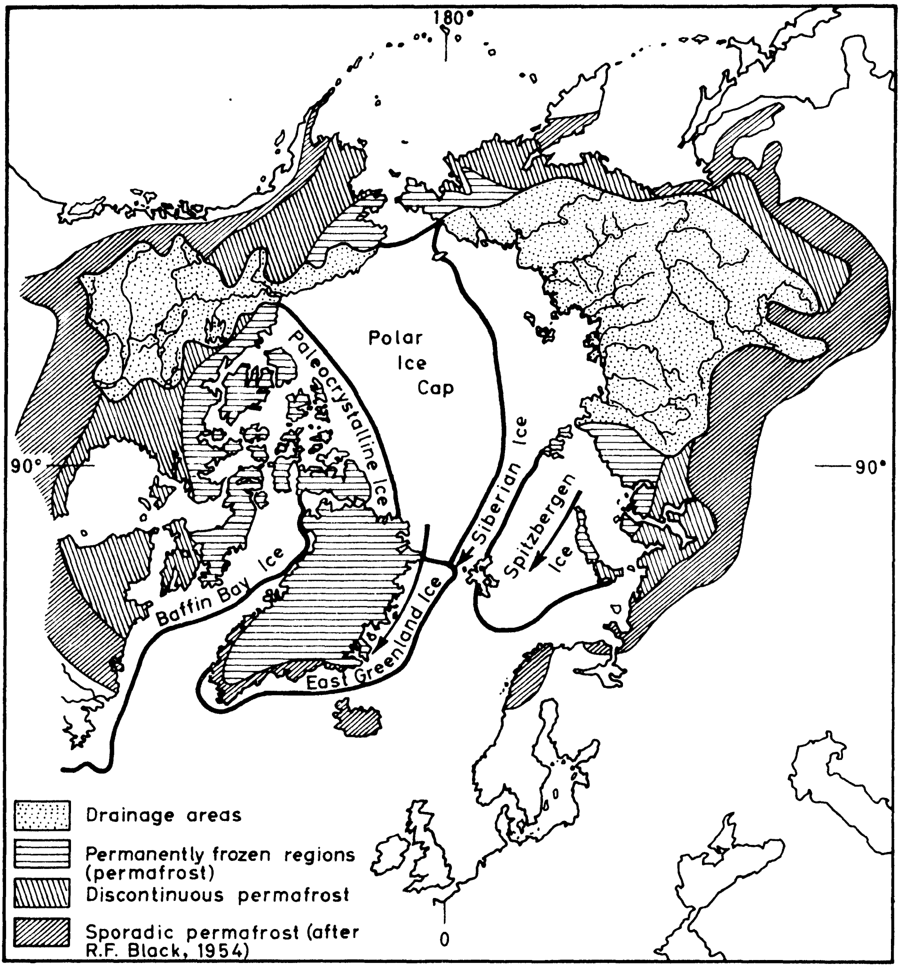


FIG. 35. THE DISTRIBUTION OF GLACIERS IN THE ARCTIC REGION

polar ice of the northern hemisphere does not throw much light on their past history. According to L. Koch (1945) the Arctic ice and regions tributary to it, like the Greenland Sea, come from Siberia and the Canadian Shield, the two principal areas of fresh-water drainage. Siberia provides the *Siberian ice*, which skirts the borders of nothern Asia and normally moves between Severnaya Zemlya, Franz-Josef Land and Spitzbergen. The

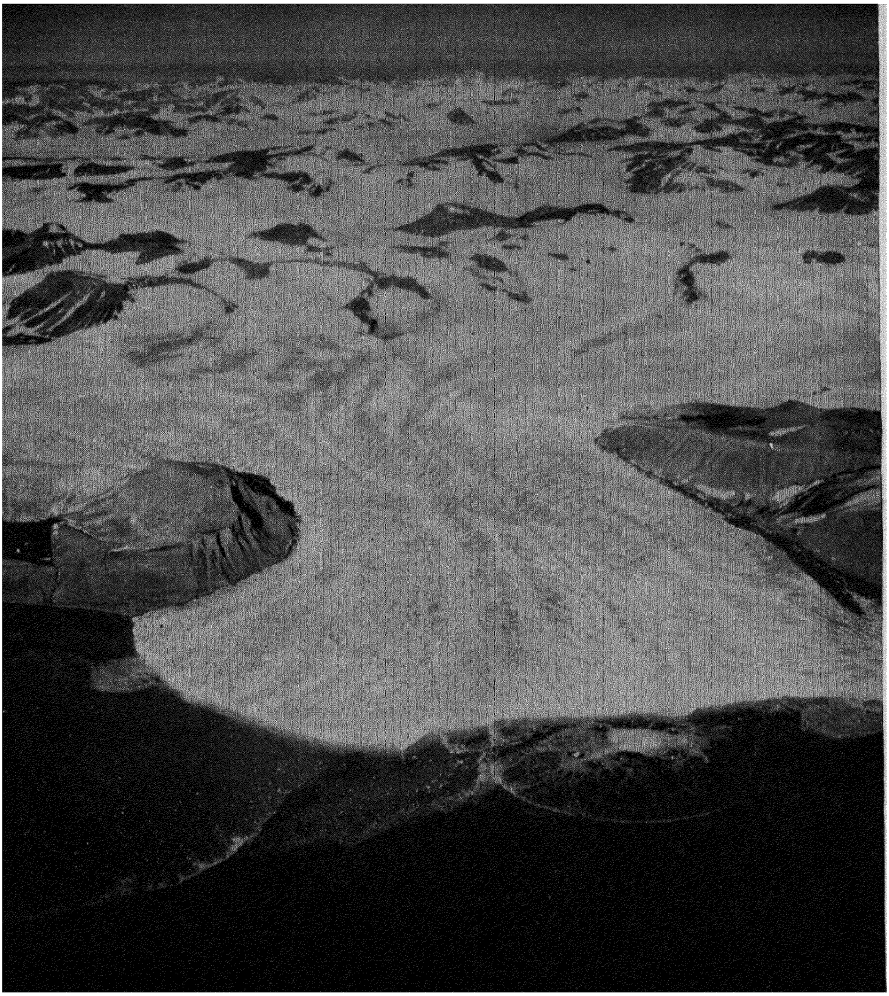


FIG. 36. CROLLBREEN, ON THE WEST COAST OF STORFJORDEN, SPITZBERGEN
(Photograph: B. Lüncke, 4th August, 1936, Enerett—Copyright Norsk Polarinstitutt)
Note the advance of the glacier into the sea and its breakup into icebergs. It must be remembered that the glacier carries a large amount of detrital glacial sediment into the sea.

Canadian Shield feeds the *North Pole ice* between Alaska and northeast Greenland. In turn, these principal areas of ice development feed the "packs".

The Spitzbergen ice is situated between Severnaya Zemlya, Franz-Josef Land, the south of Spitzbergen, Novaya Zemlya and the mouth of the Ob. The drift ice derived from it carries sediment (sand, stones and plant debris)

Fig. 37. Penckbreen, South of Van Keulenfjord, Spitzbergen
(Photograph: B. Lüncke, 10th August, 1936, Enerett—Copyright Norsk Polarinstitutt)
A glacier descending to the sea. Note the medial moraine, the lateral moraine clearly visible on the left, and the terminal moraine in the foreground. The subglacial streams give rise to a "sandur", a layer of fine-grained sandy material.

from the east and the south of Spitzbergen, sometimes as far as eastern Greenland and Iceland.

The paleocrystalline ice (fossil ice) (Nares 1878) occurs to the north of the archipelago of Arctic America, of Grant Land and of northern Greenland. It flows towards the *Baffin Bay ice.* Finally, the *east Greenland ice* comes from two sources—the North polar ice and the Siberian ice.

The distribution of ice fringing eastern Greenland is not uniform. According to L. Lock (1945) three principal types of ice may be distinguished on the basis of their origin:

(i) ICE DETACHED FROM GLACIERS; composed chiefly of *Icebergs* and of smaller blocks (*Calf ice*).

(ii) RIGID ICE; formed along the coasts, and comprising the *ice foot* fringing the shore. Ice from fjords is often only temporary. If it remains more than ten years, it is given the name of *Sikussak ice.*

(iii) DERIVED ICE as follows:

The *land ice* formed at the mouth of fjords and bays which, after one to three years, drifts toward the open sea before sinking; and *pack* ice formed

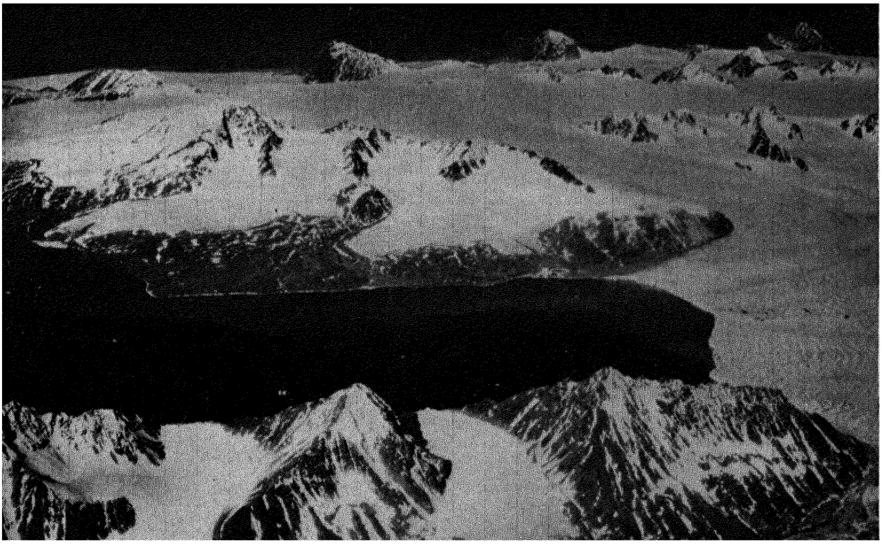


FIG. 38. MÖLLERHAMMA, KING HAAKON PENINSULA AND LILLICHÖÖKFJORDEN, SPITZBERGEN (Photograph: B. Lüncke, 14 September, 1936, Enerett—Copyright Norsk Polarinstitutt)

Note the concave relief close to the glaciers and the nunataks (peaks or crests not covered by ice).

in the sea from fresh water, which floats on top of salt water. The irregular surface of pack ice forms "hummocks". When it is between one and three years old, it is simply called *pack* (the Siberian ice, the Spitzbergen ice and the ice of Baffin Bay), but when it is more than five years old it is called *arctic pack* (North Polar ice, paleocrystalline ice). The floating parts of the derived ice have been classified according to their size and form:

field, more than 3 miles in diameter; *floe*, more than ½ mile; *small floe*, more than 650 feet; "*glacons*", smaller than 650 feet; *cake*, much smaller than 650 feet and flat; *bits*, less than 18 inches, grouped into *brash ice.*

"Hummocks" are fragments of ice broken by pressure. They rise up on the surface of the sea, forming "*growlers*" when they exceed 5 feet above the surface. They are often found in groups.

The ice which forms glaciers consists of various layers which continually change: the top 160 feet are rigid, composed of blocks which move parallel with one another, perpendicularly or obliquely to the ice shed. These slide over the deeper layers, which are continually changed by viscous or plastic flow. The differences in movement between top and bottom layers are due to their different loads.

2. *Agents of the Atmosphere: Eolian Erosion*

The wind is the principal atmospheric agent of erosion with the exception of the humidity of the air, and the capillary water whose influence in rock disintegration is quite important. But, wind removes very little material by itself, even at high velocity. Its abrasive power is principally due to its content of dust and sand. The effects of eolian erosion are characteristically developed on a small scale, and they are not so apparent on a large scale. In piedmont sediments and terraces, the abrasion of sand grains (rounded and frosted), pebbles (*dreikanters*) and sculptured ridges (*yardangs*) may be noted. In hard rocks, abrasion by sand thrown up by the wind undermines pillars, which assume characteristic forms (fig. 39).

Erosion of fertile soils is the most important effect of the wind. It has roused anxiety in governments of a number of states, particularly in deforested zones. The wind lifts up the dry soil in the form of fine dust. This soil erosion has become a serious problem in the Middle East, which formerly was a cultivated or well-timbered region. It is also a problem in the great plains of the United States, which have been under heavy cereal cultivation during the last 50 or 100 years. Imprudent exploitation and short sightedness in the United States, risks results analogous to systematic destruction by invading Mongols. The arable land is carried up into the air and deposited elsewhere. In this case, cultivation seems to initiate desert conditions.

The influence of wind is not restricted to hot arid regions. Its effect is seen in all subdesert regions, whatever the climate. Thus the coastal supralittoral zone, which was noted earlier, has many characteristics in common with deserts. For example, it supports coastal dunes comparable with those in sandy deserts. The periglacial zones are equally rich in detrital material and display phenomena completely comparable with those of hot deserts; loess, reg, soil polygons, "dreikanters" and dunes.

The wind transports material which is not fixed, such as soils, dust, and sands, leaving great barren areas of rock and pebbles. The latter make up the *reg* of the surface of the hamadas (see p. 38). J. Walther (1891) calls this ablation by wind, *deflation*. The load of fine material in movement in the air is a powerful abrasive, capable of wearing away the hardest rocks. It gives rise to honeycomb weathering, comparable with taffoni, and is even

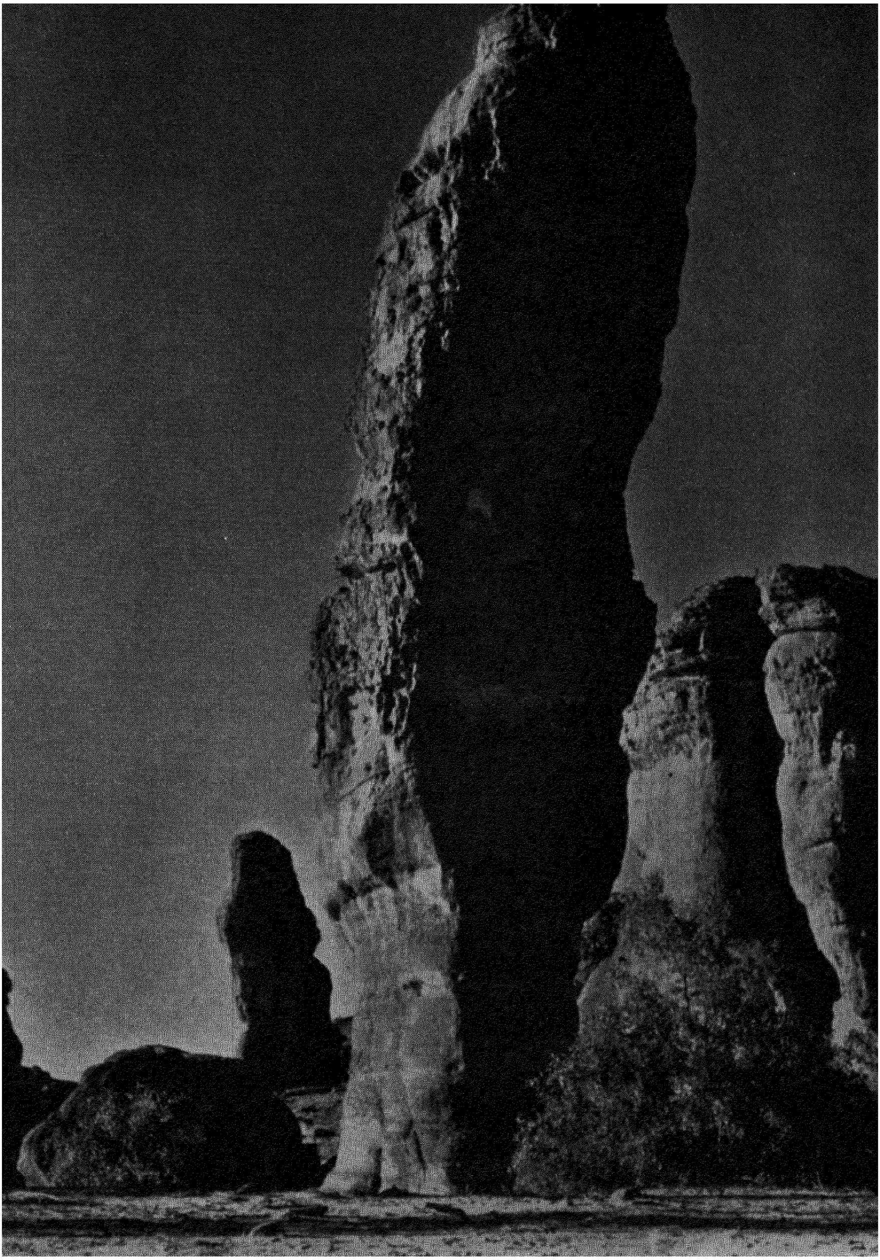


FIG. 39. TYPICAL EOLIAN EROSION IN CAMBRO-ORDOVICIAN SANDSTONE (INNER TASSILI), NEAR ZOUAR (TIBESTI), CENTRAL SAHARA (Photograph: Freulon)

able to reduce local topographical levels. From blocks of small size, wind action carves three-sided pyramids or "dreikanter". Even the surface of these eolian pebbles is pitted with many coalescing hollows. They also become rounded and frosted like grains of sand.

In the indurated clays of the Tarim, especially those of Qum-darya, there is deflation orientated in a northeast or north-northeast direction by prevailing winds. These have sculptured the "yardangs" which consist of ridges and trenches, with sand piled up in their hollows.

The "pans" of South Africa are also in part formed by deflation. They are depressions indicating the beds of oueds. They are generally dry and only contain water after rains, but their floor is usually covered with mud and salt. They are thus a variety of playa (p. 117). According to South African writers, the "pans" initially formed by water action have been hollowed out and completed by the action of whirlwinds during the dry season. In course of time, this wind erosion has extended and deepened the pans, so that some have become joined. Thus, not only deflation (ablation) occurs, but also corrosion (sculpturing by the wind).

Alternation of cold and heat adds, in all countries, to the disintegration of rocks whose components possess different coefficients of expansion.

3. *Agents of the Biosphere*

The biosphere plays a considerable role in erosion. All organisms depend for their sustenance more or less directly on the soil, and thus on the subsoil. Bacteria and plants derive immediate benefit from the subsoil, and assimilate inorganic salts from it directly. They live on the substratum and thus attack it mechanically and chemically. Thus there is *an association between erosion and vegetation* because the plants which attack the bare rock create a soil which, at the same time, protects the rock from other agents of erosion. Reforestation is the only known method of stabilizing mountainous slopes, and of counteracting their degradation by mountain torrents. Plantations of rushes and pines stabilize dunes by hindering their movement by wind. This *association between erosion and vegetation* will be discussed in detail later (p. 133).

In the sea, particularly in shallow water and near coasts, burrowing organisms play a role comparable with the plants on land and help form a "submarine soil" (Termier, 1952, pp. 113–118). The plant and animal organisms of the fixed benthos stabilize the mud or sand on which they live. This point is discussed later (p. 222).

If, in fact, forest and bush are agents of conservation, then animals (and man, the most terrible animal of all) are agents of destruction. They are incapable of directly utilizing the chemical substances of the soil and the energy of the sun, and are led by their nature to destroy the plants. The pursuits of forestry and agriculture are the human activities which seriously disrupt the natural equilibrium between erosion and vegetation. Over-grazing by cattle has disastrous consequences. Those regions which are fertile nurtured the more advanced civilizations and were developed by sedentary peoples and cultivators who respected, at least empirically, the laws of hydrology and agronomy. By contrast it is significant that the semi-arid and arid zones, which can be seen to increase at the expense of former fertile regions, were in Prehistoric times occupied chiefly by stock-breeding people and often by nomads. The establishment of these peoples in the semi-fertile regions of Eurasia, which coincided largely with the fall of the Roman Empire and the beginning of the great invasions, proved to be the end of the large forests whose bushes were burnt, saplings uprooted and grass trampled upon. Certain Greek cities of the past were so aware of this peril that they prohibited the breeding of goats (L. Robert, *Hellenica*, VII, pp. 161–170).

Most modern writers claim that man is the dominant factor in aridity. In some countries the desert has become established, and there has been practically no return to the former state following the removal of vegetation. These countries may be contrasted with those regions where the equatorial forest has again spread over an abandoned town, as it has done over the ruins of Angkor in Cambodia. In the arid zone where evolution towards a desert state is undeniable, it might be thought that man could have fought this tendency better than he has. But his action has merely precipitated a phenomenon from which there is hardly any escape. However, there have been other periods of disastrous aridity in the geological past. Man cannot be held responsible for the disappearance of animals and plants brought about by the Permo-Triassic climate. So today, quite apart from the role of man, the planet is almost certainly involved in a long-term climatic deterioration.

The Role of Boring and Burrowing Organisms in Marine Erosion.— Most agents of marine erosion differ little from those which affect the continents: water, ice, wind, and chemical substances are most important. But to these the work of organisms must be added. If organisms which attack wood are excepted, "boring organisms" are present in negligible quantities on the continental surfaces. On the other hand, they abound in the seas, especially in the littoral zone. They attack soft rocks such as limestones, particularly coralline limestones, and dune sandstones situated in the shallow littoral zones. Most phyla are represented in this ecological category, namely, "burrowers and borers".

The boring algae, which generally belong to the blue algae (Schizophytes), probably penetrate the limestones by solution. Thus, shells and even pebbles disappear. Some of the sponges, the Clionidae, behave almost in the same manner.

Some of the worms, chiefly the annelids, and some of the molluscs (Pholads, Patellas and Chitons) excavate holes in rocks. To these must be added littorinas and other gastropods, whose method of attack has been specially studied in Indochina, by P. Fischer (1953) (fig. 40). A velvety mat of green algae in the intertidal zone becomes "browsed", as in a grass field by goats or cattle; the characteristic scraping by gastropod radulae is seen on all rocky coasts.

In almost all latitudes, the regular echinoids live in depressions which they have hollowed out. In tropical regions their action is on the increase

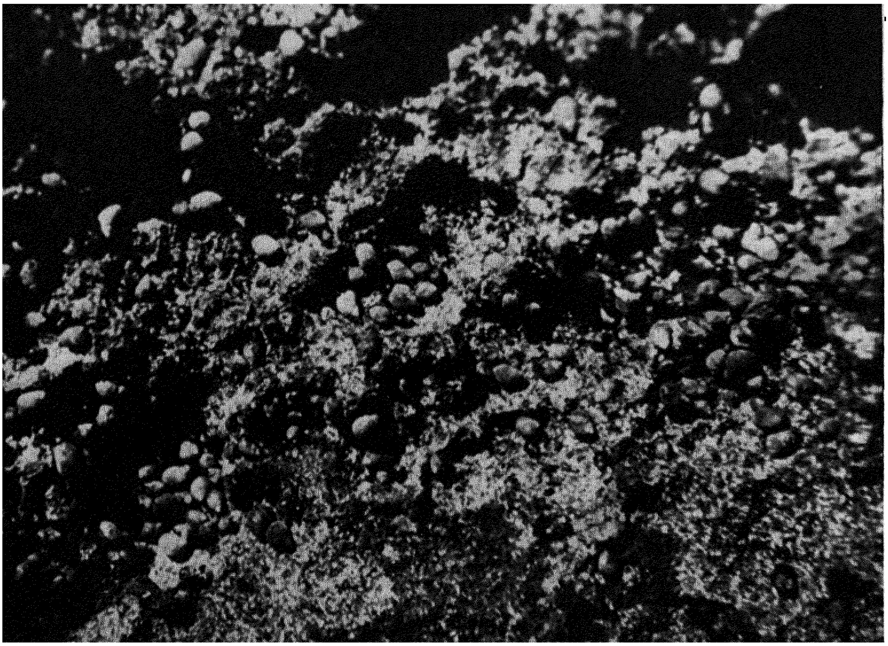


FIG. 40. THE SUPRALITTORAL STAGE WITH LITTORINA GASTROPODS WHICH CONTRIBUTE TO ITS EROSION

Near Tipasa, Algeria. (Photograph: G. Termier)

and the role of *Echinometra mathaei* (Blainville) in the undermining and abrasion of reef limestones in the East Indies is quite considerable. (Umbgrove, 1947.)

Crustaceans, and certain fish such as the blennies, which live in holes and hollows, although they are not true boring organisms, enlarge their hollows and accelerate rock disintegration. Other fish such as the scara (parrot fish) browse on living corals and, at the same time, on part of their polypary. These fish are thus erosion agents acting contemporaneously with the construction of reef rock.

At the other extreme, there are *sediment-forming organisms* which contrubute to the formation of rocks. This is true of many burrowing organisms, such as sea urchins and sea slugs, which ingest coral sand and debris from shells or coral colonies and which reject a fine limy mud. This also is true of non-burrowing organisms such as fish (mentioned earlier) and oysters. Members of the Calypso expedition observed fish browsing on living coral thus: "They attack the sound part of the colonies and tear the living part and some of the limestone skeleton at each bite. After passing through the digestive tract, calcareous debris of the dimension of sand or silt is rejected. These animals browse continually and the result of their activity, although small in detail, becomes, in the end, important, because of the continuous nature of their nutritive functions. This is probably the origin of the calcite sands and silts of the Red Sea." (V. Nesteroff, 1955.)

Practically all echinoids, worms, crustaceans and gastropods, thus expel small pellets of fine sediment, which result in the formation of muds and later of fine-grained rocks (Moore, 1939). It will be shown that the argillaceous part of these muds, acquires particular characteristics (p. 222).

There is thus an equilibrium between *organisms and rock*. This equilibrium can be seen in almost diagrammatic form on the Florida coast, where R. N. Ginsburg (1953) has studied the "beachrock". This is a friable calcareous sandstone formed from the debris of algae (*Halimeda*, Lithothamnions), shells, skeletons, and fragments of rock which normally form a beach. They are only found in the intertidal zone of coral regions and their thickness rarely exceeds 8 feet. The consolidation of "beachrock" is attributed to the precipitation of calcium carbonate in solution in water, perhaps under the influence of bacteria living on decomposing organic matter (very abundant on this coastal fringe). Furthermore, due to warm temperature, precipitation occurs in the upper layers of the water. Each grain of sand provides a center of crystallization for aragonite. The surface of this friable rock often displays alveolar erosion or honeycomb weathering. The speed of erosion suggests that such an accumulation might not be preserved, as it stands, in the fossil state. Erosion takes place almost as quickly as the rock is formed. This process also occurs in Bermuda (Prat, 1935). It would seen desirable to study the various limestones found near reefs, from the Visean to the Jurassic. They show microscopic similarities to these Recent sandstones, since they possess rolled organic limestone elements cemented by calcium carbonate. Some of the sand grains are oolitic or sometimes pseudo-oolitic.

C. SOME PECULIARITIES OF EROSION

1. Jointing in Rocks as a Guide for their Erosion

The detailed etching by erosion often follows lines and pre-existing surfaces which are part of the rock structure. In general, jointing (and "parting"), the natural division into more or less regular blocks, aids erosion in the same way as it aids quarrymen by facilitating access to surface layers or even deeper rocks.

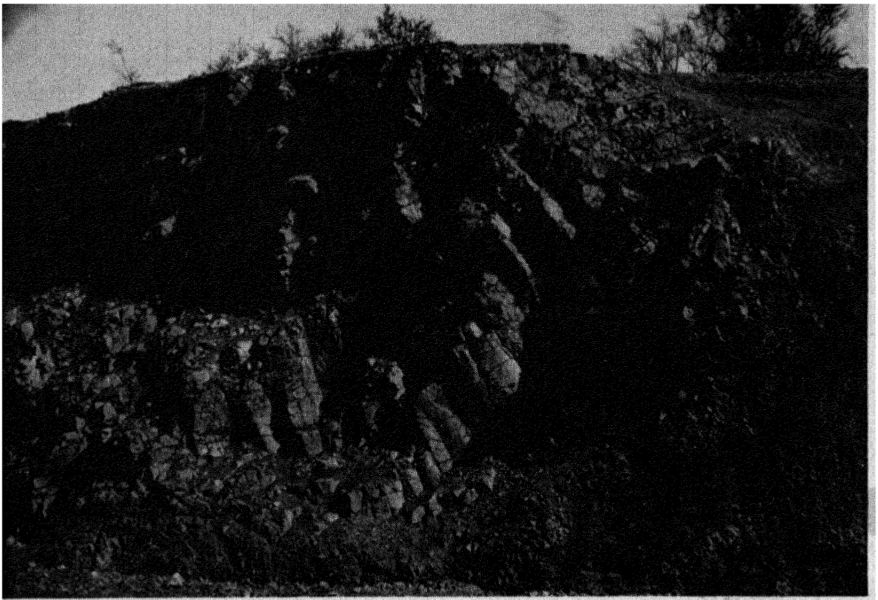


FIG. 41. COLUMNAR BASALT IN THE PERMO-TRIASSIC OF MOROCCO
(Photograph: H. Termier)

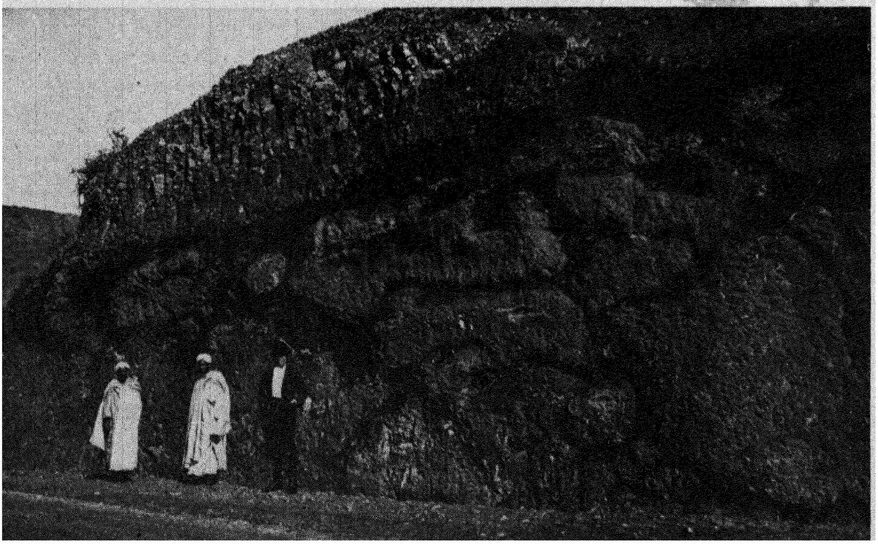


FIG. 42. BASALT IN THE PERMO-TRIASSIC, NEAR MERZAGA (CENTRAL MOROCCO)

The lower part of the section is *a pillow lava.* The lava flowed into a lagoon, the rapid chilling producing "pillows". The lagoon is suggested by red gypsiferous clays in the neighborhood. The rock, which is often dark-green, contains much red chalcedony.

The upper part of the section shows *well-preserved columnar structure* (prismatic jointing developed during cooling) which is rare in such old lavas. The rock is a doleritic basalt containing labradorite and pigeonite.

Fig. 43. Spheroidal Weathering of a Permo-Triassic Basalt in Morocco

Right bank of the Oued el Mouater (between Oujda and Berguent). (Photograph: G. ermier)

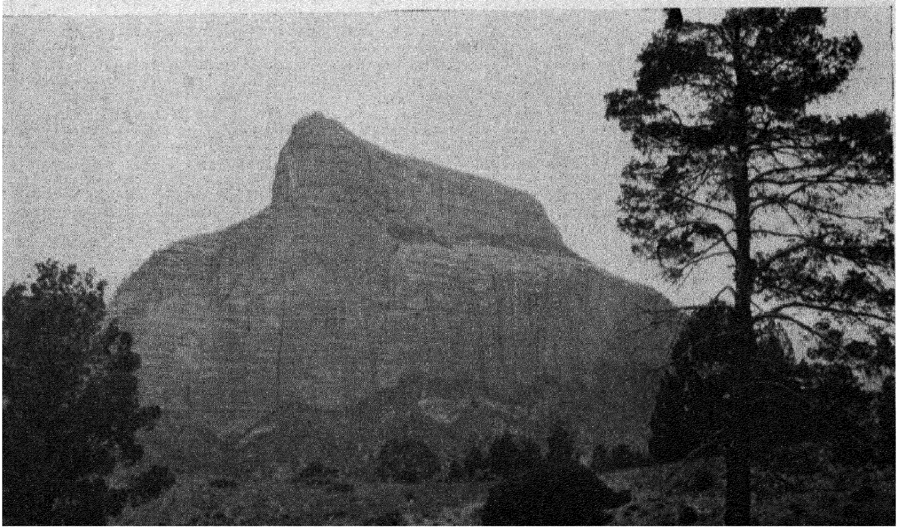


Fig. 44. The "Cathedral" of Tilouguit. An Example of Erosion attacking Conglomerates

"Pontian" (Pliocene) of southern Morocco. (Photograph: H. Termier)

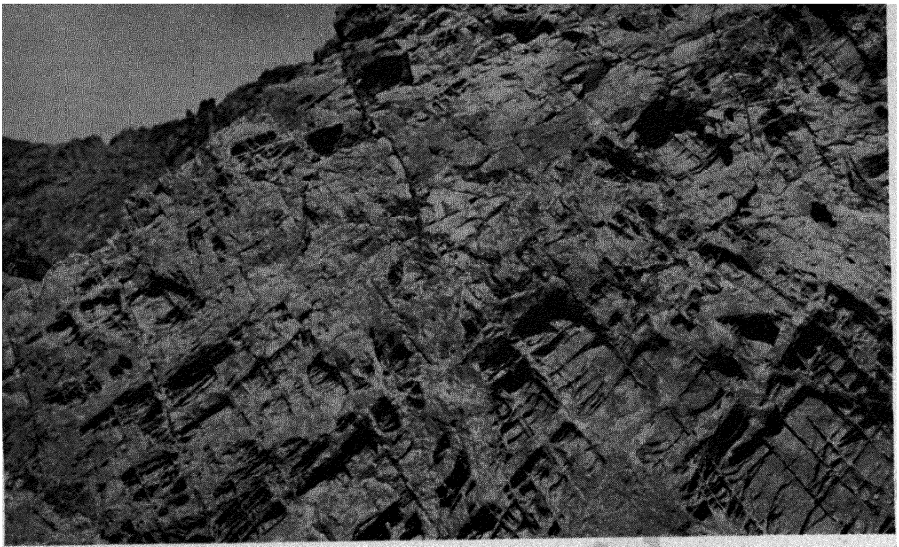


FIG. 45. THE ERODED SURFACE OF A MICRODIORITE DIKE (WHICH CUTS A BAND OF MARBLE)

The hardened joint-planes of the rock form a network of ribs which enclose hollows several inches deep. Taferout n'Tizi Ilrès: on the ridge on the northeast border of the granodiorite massif of Tichka, Upper Atlas, Morocco. (Photograph: H. Termier)

FIG. 46. THE EROSION OF A GRANITE DOME IN A TEMPERATE CLIMATE: A "BALLOON" IN THE VOSGES SEEN FROM HOHNECK (Photograph: G. Termier)

Erosion of a homogeneous rock.

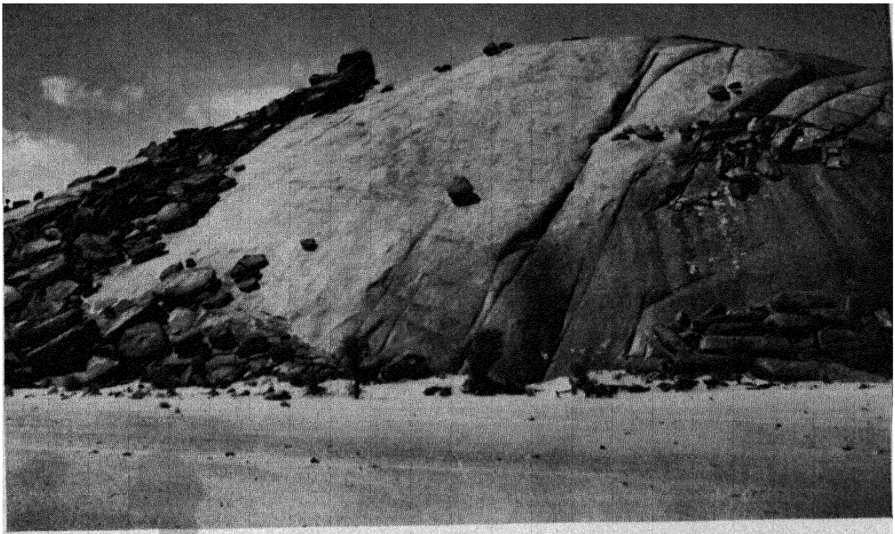


FIG. 47. THE EROSION OF A GRANITE DOME UNDER AN ARID CLIMATE: A GRANITE MASS IN THE NORTH OF THE HOGGAR MOUNTAINS, NORTH AFRICA (Photograph: G. Termier)

Concentric exfoliation.

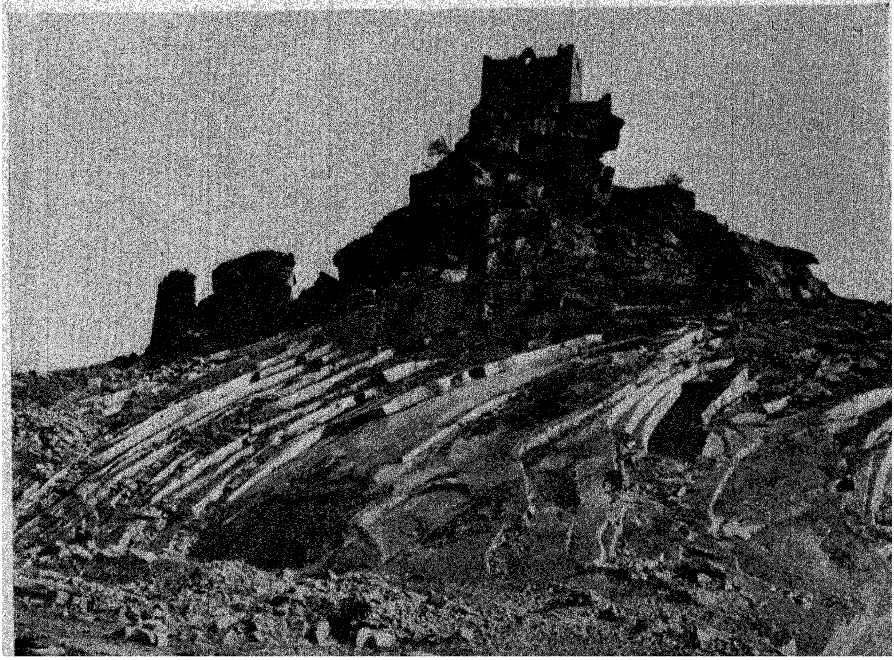


FIG. 48. PARALLEL EXFOLIATION STRUCTURES IN PALEOZOIC SCHISTOSE HORNFELS, WELL EXPOSED IN A QUARRY, west of the ruin of Flossenberg, near Weiden, Oberfalz, Germany

The tower stands on a granite which cuts across the bedding of the hornfels. (Photograph: Akermann)

FIG. 49. AN EXAMPLE OF JOINTING IN GRANITE

The gneissic granite has biotites oriented in the plane of the horizontal joints. The vertical joints cut the foliation squarely, and there is thus a tendency toward spheroidal weathering. Les Bouillouses, Eastern Pyrenees, France. (Photograph: G. Termier)

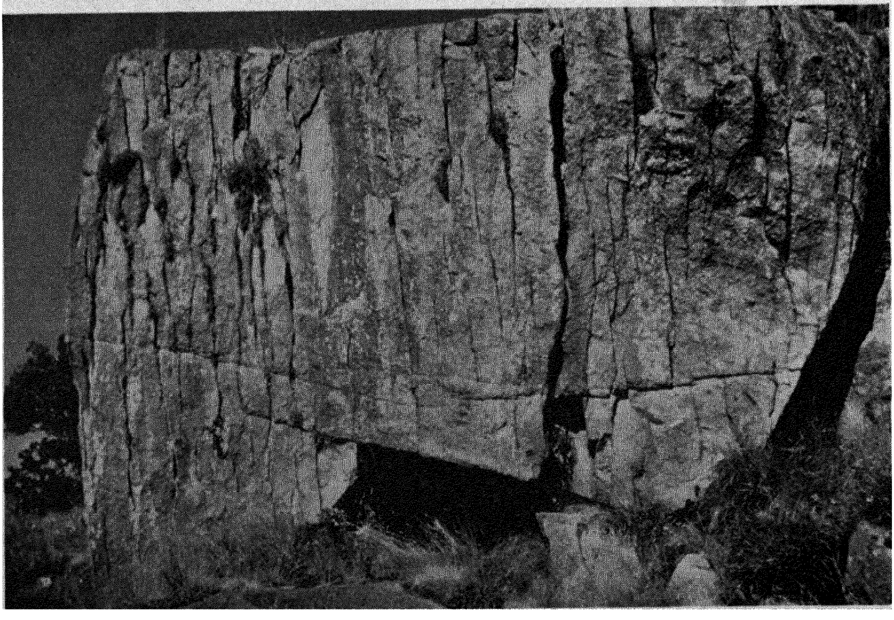


FIG. 50. ANOTHER EXAMPLE OF JOINTING IN GRANITE

The foliation, due to inclusions and micas, are oriented at an angle of about 30° to the horizontal. Some of the joints follow this direction, but the principal joints are subvertical. La Llagonne, Eastern Pyrenees, France. (Photograph: G. Termier)

In sedimentary rocks, bedding planes are due to changes in the conditions of deposition. Joints are fissures perpendicular or oblique to the stratification and cleavage results from orogenic pressure. Erosion exploits each of these. As was observed in the Paleozoic of Central Morocco, even shales can be distinguished by their type of jointing (H. Termier, 1936).

FIG. 51. FLAGGY SANDSTONES SHOWING CONCENTRIC WEATHERING (UPPER VISEAN, MISSISSIPPIAN, OF CENTRAL MOROCCO)

This fairly coarse sandstone (grain size about 1 mm.) has been regularly jointed, and circulating water has cemented the walls adjacent to the fractures more thoroughly than the interiors of each prism. The latter have been hollowed out by erosion. The hammer handle (16 inches long) gives the scale. (Photograph: Henri Termier)

Volcanic rocks occur in thick bands, in slabs or platy layers (phonolites and dacites). To these must be added forms due to shrinking during cooling, including basalt prisms. This type of jointing may be seen in the peridotites of New Caledonia, where forms due to cooling and various fissures have given rise to cubes, parallelepipeds and rhombohedrons. These will eventually be detached from each other and rounded into spheroids (Chételat, 1947). In fact, when a rock is, on the one hand, homogeneous and, on the other, divided by three systems of fissures, it breaks up into rhomboid blocks. Their edges become rounded and spheres are finally produced. This phenomenon, which is frequent in petrographic types ranging from granites to basalts, may be followed step by step in most igneous rocks, especially under humid climates. In granites, it results in a chaos of rounded rocks as seen in Brittany, the Central Massif and the Pyrenees (for example at Targassonne, between Font-Romeü and Bourg-Madame).

In an earlier work (Termier, 1957, pl. XXXIV and XXXV), evidence was adduced to show that the breakup of orientated crystalline rocks into balls takes into account the dominant direction of the grain, if such exists. The cleavage finally occurs along the joints.

To conclude, when the climate is known, it is usually fairly easy to identify rocks in the field by their jointing.

2. *Some Details of Erosion Methods*

To understand the processes of geomorphic sculpture, it seems appropriate to review the behavior of erosion agents according to the materials provided and to the environmental and climatic conditions.

Among hard rocks, it is possible to distinguish:

- erosion which leads to the formation of taffoni (in granites, greenstones, metamorphic schists, limestones and sandstones).
- alveolar erosion (in sandstones).
- eolian corrosion (in hard or soft rocks).
- karst erosion (peculiar to soluble rocks).

Taffoni.—The taffoni are alveoles or honeycomb structures of disputed origin. They only affect *vertical* walls or overhangs of denuded rock (granites and rhyolitic tuffs). In Corsica, where they were first described, their dominant orientation is toward the south, the direction of maximum insolation. They are elongated parallel with the stratification, the cleavage and the joints. They generally widen toward the bottom, and are subdivided into compartments. They may also be seen in Greenland and Iceland. B. Popoff (1937) thought that they had been hollowed out by the mechanical action of columns of warm air, rising up and displacing the cold air lying in the bottom of these cavities. This interpretation can be applied to other regions, notably to the Sahara, where taffoni have been observed by the authors in the volcanic lavas of Atakor (Ahaggar). No chemical alteration has taken place (fig. 52).

According to A. Cailleux, this mechanism would have to be preceded (as in alveolar erosion) by the alternation of freezing and thawing, although it rarely freezes in the Sahara.

The study of taffoni in Corsica by Ottmann (1956), has shown that, even if some taffoni are still developing, a large number of others are now "dead". The former may be recognized where they enclose sand or thin flakes of the hollowed rock. The "dead" ones are covered with lichens. Some of those at Scandola point (Gulf of Oporto) are found today under the sea and date from a period of regression. Near Calvi, a dune, thought to be of "Tyrrhenian age", rests on a rock which is riddled with taffoni.

FIG. 52. "TAFFONI" IN THE CRATER OF MZARAF AROURI NOYED, A VOLCANO IN THE HOGGAR MOUNTAINS (CENTRAL SAHARA)

The rock is a trachyphonolite. (Photograph: G. Termier)

It would seem that the Corsican examples may be explained by large diurnal variations in temperature, and the daily repetition of soaking and drying. Toward morning, a considerable humidity occurs in these deep cavities.

Alveolar Erosion.—1. INLAND. Alveolar or honeycomb erosion occurs on vertical walls where it emphasizes the stratification and jointing of sandstones, limestones, and crystalline schists. It is due neither to wind action nor to chemical alteration. According to E. Haug (*Treatise*, I, p. 378) this type of erosion occurs on rocks previously protected by a ferruginous coating ("patina" or desert varnish); the irregularity of distribution of the alveoles would be due to the chance removal of this protective varnish. Alveolar erosion is often found in periglacial regions. In this instance, according to A. Cailleux (1953), a film of frozen water would settle permanently in the bottom of the alveoles. This film would constantly attract warm air and atmospheric vapor, which would seep into the interstices and refreeze during the night or during the winter. This is the mechanism of corrosion, which tends to be continuous.

2. FRINGING THE SEA. With regard to the platform of marine abrasion, it has been seen (pp. 63–65) that a hollowing-out of cupolas occurs in the supralittoral zone (fig. 53). Moreover, the authors noted, with P. Muraour, fine examples of alveolar erosion in the Oligocene sandstones of Dellys

FIG. 53. ALVEOLAR EROSION ON THE SUPRALITTORAL SURFACE OF THE GRANITE OF PLOUMANACH, BRITTANY, FRANCE (Photograph: G. Termier)

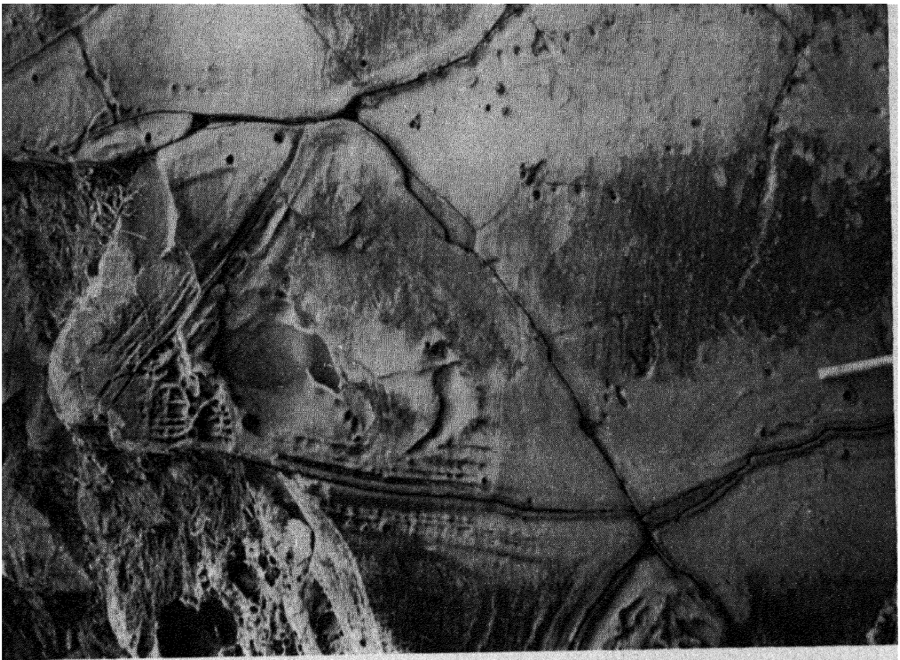
(Algeria) often about a hundred feet above the water (figs. 54 and 55). These are clearly phenomena of corrosion which cannot be explained by any of the preceding hypotheses. Just as in deserts with their ferruginous varnish, there is often a "case-hardening" at the coast associated with the precipitation of carbonates and sulfates ("pelagosite"), which also favors honeycomb weathering (Revelle and Fairbridge, 1957, p. 258).

D. CORROSION

Rain (p. 73) is an important factor in the corrosion of surface rocks.

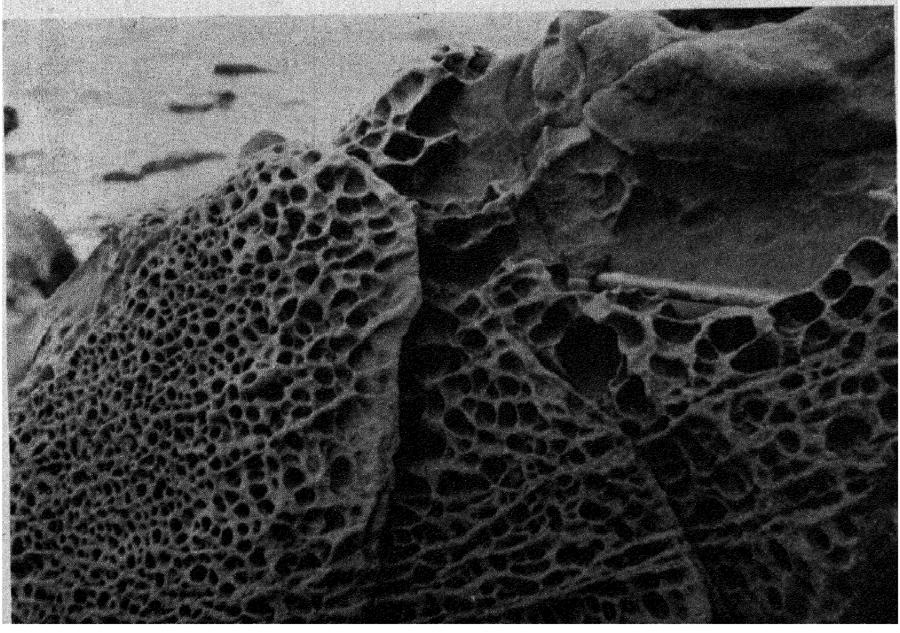
The Role of Salt.

The role played by marine salt is certainly very considerable. It is one of the elements of spray and, consequently, one of the agents of the supralittoral stage. Moreover, it is often carried by the wind. Near the coast, in Ceylon, more than 90 lb. per acre per year are deposited.



FIGS. 54 and 55. ALVEOLAR EROSION IN THE DELLYS SANDSTONE, ALGERIA, ON THE COAST OF KABYLIE

Note in the upper figure, the development of hollows along the fissures. (Photograph: G. Termier)



The Role of Humic Acids.

The humic products of plant decomposition on land covered by vegetation, yield acids capable of attacking the underlying rocks. Their action is important in the formation of soils (see p. 133).

The Role of pH.

Variations of pH between day and night, which can occur in plant zones in marine, as well as land environments, can result in considerable degradation of rocks.

E. STRUCTURAL SURFACES

This is an important term (B. de la Noë and E. de Margerie, 1888), which applies to the whole of a hard rock surface where erosion has removed the overlying softer rocks which masked it. Erosion could be due solely to the activity of running water and rainwash, or even to landslipping by solifluction or "creep" (*décoiffement*, Lugeon, 1949) of poorly consolidated ground. The surface thus exposed is similar in appearance to ground which has been produced by dissection. It may be horizontal beds, anticlinal or synclinal folds of stratified rocks, or the outer edge of granitic batholiths. The structural surface is an *exceptional* and temporary form since *its being exposed implies that it immediately becomes prey to erosion.* Furthermore, at the time of its formation, a fold rarely exhibits the diagrammatic appearance of a structural surface. But the idea is useful, both in hydrology and geomorphology (fig. 56).

FIG. 56. THE STRUCTURAL SURFACE OF A SMALL ANTICLINE AT BJØRKO, NEAR OSLO, NORWAY (Photograph: G. Termier)

5

Morphology

LARGE-SCALE EROSION FORMS

Definition of Base Level

Geomorphic sculpturing is chiefly concerned with erosion and transport of material. Thus, it partly depends on the effect of gravity. The action of rainwash on continental surfaces may be related to one or a number of fixed surfaces, below which fluvial erosion no longer occurs. The name of *base level* has been given to these theoretical surfaces (Powell, 1873).

The level of the sea has long been considered stable, and under the name of *mean sea level* it is referred to as the *main base level*. But this surface changes vertically in its relation to the continent, depending on eustatic, epeirogenic and orogenic movements. These include deformation of the sea bed as well as changes in the volume of ocean.

"Glacial control" is among the factors which have modified sea level. At the time of maximum cold, each glaciation resulted in the locking up of enormous masses of water in the form of ice, and led to a low sea level. On the other hand, the interglacial periods corresponded with the more or less complete melting of the glaciers, and with the return to circulation of the water thus liberated. Only Antarctica seems to have been very little affected (probably due to its relative brevity of the interglacial intervals). Today (between 1885 and 1951) a rise of mean sea level of about one-sixteenth of an inch per year has been measured. This rise corresponds with a period of increasing mean temperature, and the melting of glaciers.

The change of altitude of the main base level automatically sets in motion the mechanism of transgression and regression of the sea. A modification of fluvial erosion also occurs. Lowering of this level allows a stream, which has already attained a profile of equilibrium, to recommence erosion and to incise or intrench itself to a new profile of equilibrium.

When rivers do not terminate at the sea, but join a major stream, it is the bed of the major stream at their confluence that becomes the base level of erosion. In rare cases, a lake, like Lake Geneva, the Great Salt Lake of Utah, or the Great Lakes system, forms a base level and retards the morphogenetic evolution of the region, since its altitude is generally much higher than mean sea level.

At other times, *local and temporary base levels* are formed by outcrops of hard rock underlying a soft rock. The river does not show a regular profile but is broken up by these outcrops into a series of segments or *quiet reaches*, separated by "Knick points".

In karst regions (p. 302) the water table acts as a base level. The same thing probably happens in ground which has a thick bed of clay, such as in lateritic regions (as in Guiana, see p. 154).

There are two useful terms relative to base level (introduced by Surell), namely the idea of the *fixed point* to describe the place where the stream reaches base level (see further below), and the *inflection point* to denote the place which separates the zone of erosion from the zone of alluvial deposit formation.

Exorheic and Endorheic Basins

In an *exorheic basin*, the rivers flow to the sea and become adjusted to the main base level.

An *endorheic basin* (fig. 57) has no outlet to the sea. The base level can

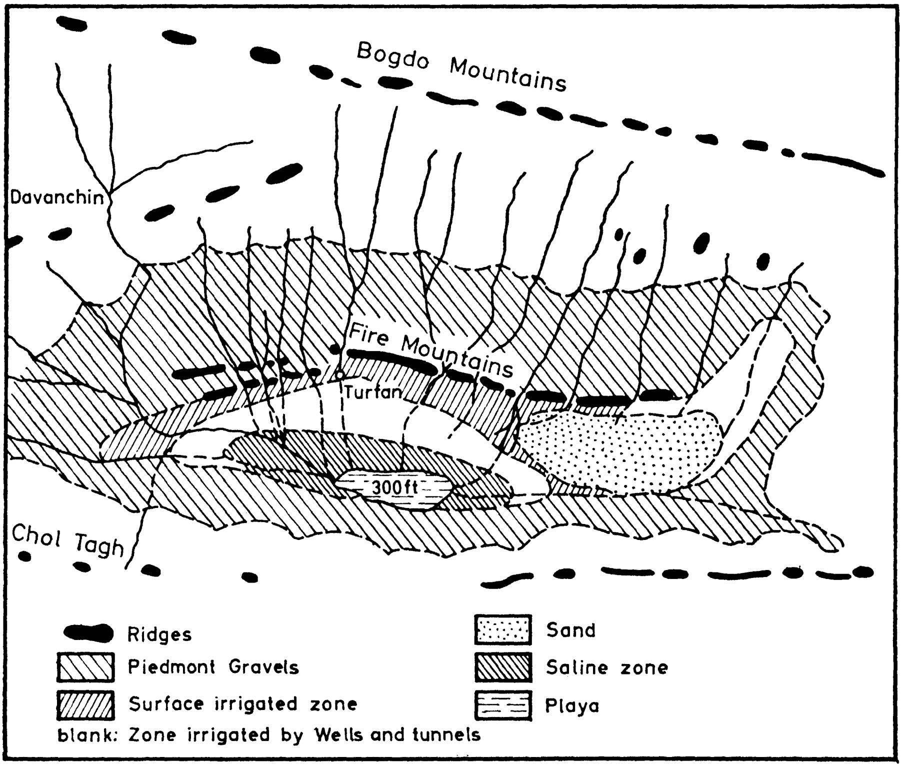


FIG. 57. THE TURFAN BASIN, CENTRAL ASIA (after Ellsworth Huntingdon)

only be formed by a lake, which is generally temporary, and often saline (playas), and may change its position. An example is Lop Nor in Central Asia. Streams are not usually permanent there and are lost in alluvial deposits and sands. Such closed basins in dry regions may be assumed to have a *variable and temporary local base level.*

In fact, in an arid or semiarid country, the question of base level becomes complex. The oueds[1] behave in fact as local base levels on which variations in the altitude of sea level have no effect. Some rivers are lost in the sands, far from their origin. Their base level is thus the lowest altitude which they are able to reach.

G. Choubert (1946) noted that, as a result, the phases of cutting and filling of the North African oueds during the Quaternary have been independent of marine regressions and transgressions except in the immediate proximity of the coast (for a few miles only). Along the shores of the Mediterranean and elsewhere in similar latitudes (but NOT in the tropics), it is probably true that the *pluvial periods* are coincident with the *glaciations* and correspond exactly with the marine *regressions*, while the *interpluvial periods* are coincident with the *interglacials* and correspond with the *transgressions.* The special cycle of oueds in endorheic regions occurs in the following manner: "during a *pluvial period,* all the oueds deepen their beds simultaneously along their whole length. At the end of the *pluvial period* there is a commencement of infilling everywhere, with a phase of coarse pebbly deposits; and during the *interpluvial period* fine loamy sedimentation becomes the rule in all basins. These phenomena are synchronous, as far as can be judged, along the whole length of a oued, as well as from one oued to another." In this case, the terracing of the oueds may be correlated with the climate.

Geographers like to distinguish also, *arheic* regions, where there is no trace of regular stream flow. This concept seems to be chiefly theoretical. It should apply to deserts, but there are in fact, often traces of stream flow over the desert regions (see fig. 32), except in totally sand-covered areas.

Water Courses

The natural drainage of rainfall and rainwash gives rise to watercourses as varied as the land which they cross, and the climates to which they are subjected.

From time immemorial, man has been impressed by the various geomorphic developments which result from the flow of rivers, from the general regularity of their water supply, and from the nature of the rocks which they cross. A series of phenomena allow the complete picture to be visualized. This is the concept of the *profile of equilibrium* (Dausse, 1872), which implies the law of *regressive erosion,* to which are added variations of discharge, of

[1] Translator's note.—Oued = wadi, a ravine in the desert, occupied by water only during the scarce rain showers.

speed, and of the stream's load of solid material. These are indeed factors which depend upon the climate, and imply the stability of the earth's surface (which is far from true once one ceases to think in terms of the human life span).

Surell's laws, which are not found in his basic treatise (1841), are expressed by Haug (Treatise I, pp. 408–412) as follows:

"1. The deepening process of a river by the flow of water takes place from its mouth upstream, leaving a fixed point at the base of its slope, which is base level. Its movement is thus regressive.

2. The longitudinal profile leaves base level in a regular curve. This curve is concave toward the top and tangential to the horizontal in its lower section; and upstream it trends sharply upward so that it becomes tangential to the vertical."

O. T. Jones defines the ideal profile of equilibrium as a curve whose limit is a parabola; he gives the formula for it. But this is theoretical because the profile is actually a composite curve, showing breaks at each important confluence and irregularities caused by the nature of the rocks which it crosses. It is a uniform slope varying regularly, a "grade" (Davis; J. H. Mackin, 1948). This is because stream erosion progresses by cutting its bed from its mouth toward its source. It is evident that a smooth form acquired by both the stream bed and its cross section, occurs only when there is an observable equilibrium between the effectiveness of erosion agents and the resistance of rocks. As pointed out by Baulig, this equilibrium can be identified by a mantle of moving debris which is subject to continual modification. In a region unaffected by orogenic and epeirogenic movements the profile of equilibrium, once established, is modified progressively more and more slowly, while still moving towards a goal which is rarely realized.

The stages passed through by a stream as it approaches its profile of equilibrium have served to define cycles of erosion. The nearer a stream is to this limit, the nearer it is to the state of old age.

In mountain regions with a young relief (p. 42), profiles are never smooth. In the torrential state, watercourses draining the many channels of the headwater region surge over the rocks, produce severe abrasion, and carry blocks to the bottom of its steep slopes. These accumulate and form a talus cone along with better sorted sediments or *colluvium.* This is material which rolls or slides down slopes.

The Preservation of River Traces and the Vicissitudes of River Courses.—Water flowing over the surface of the continents follows the line of the greatest slope, and thus strictly obeys the laws of gravity. On continental areas, where the morphology has, in general, varied little for millions of years, the traces of rivers have often been preserved.

Tectonic depressions provide compulsory routes for running water. As long as these areas are stable, the water courses are permanent. In recently folded regions—in the western Alps, for example—superficial upheavals of the earth's crust have led to considerable variations in river courses. In the ancient shields, less spectacular uplifts of a larger radius of curvature (syneclises and anteclises) have brought about important changes. Indeed, on almost plane surfaces, the smallest lowering of base level can, without considerably altering the *trace* of the flow of the water, suffice to *invert its direction of flow*. The geomorphic history of various continents provides many examples of this: the Upper Paraguay, the Zambezi, the Nile and the Missouri (p. 79, fig. 33).

A Few Definitions.—Features of present-day geomorphology show that the influence of geological structure is twofold. Firstly, it operates statically: the river system makes use of geological irregularities such as synclines and faults (fig. 58). Further, it follows the edge of strata (a number of examples of this will be shown). The heterogeneity of the rocks crossed exerts considerable influence on the details of the river's course. Its second mode of operation is more dynamic, for it is conditioned by orogenesis and epeirogenesis, that is, by movements of the earth's crust.[1]

The hypotheses are difficult to substantiate if confined to a geomorphological viewpoint. On the other hand they become more certain when the *geological history of the area studied is reconstructed*. From this viewpoint, two ideas seem to be important; a valley is said to be *superimposed* on a deep structure, when the stream which gave rise to it was established on a surface, or in terrain, above it. It then cuts down and progressively penetrates into the deep structure or the terrain. A river is *antecedent* when it has begun to flow over a terrain before the terrain is folded. The river then cuts into it in proportion to the speed of folding, and is thus established before the tectonic deformation occurs.

[1] To describe these facts, geographers make use of two nomenclatures, one of which is purely descriptive, and shows the relationship between the form of the valleys and their geological structure. The other genetic nomenclature is more difficult to apply and is, in part, more or less hypothetical.

In the descriptive nomenclature, valleys are classified by their relationship to the dip of the beds. These are: *monoclinal* or *homoclinal* rivers, that is parallel with the direction of dips on the flanks of folds; *cataclinal* (Powell, 1875) where the direction of flow is opposite to the direction of dip; *anticlinal* or *synclinal*, that is to say, along the axes of anticlines and synclines; *diaclinal* (Powell, 1875) which cross folds. This list of names is not of great interest to the geologist.

Valleys are also classified in relation to the general slope of the land surface, which differs often from the dip of the beds. This classification has been adopted by Davis. However, his actual usage is sometimes genetic, and sometimes descriptive. The terms utilized are: *consequent* valleys (Davis, 1889–1890) which follow the slope; *obsequent* valleys falling in a reverse direction to the general surface slope; *subsequent* valleys (Jukes, 1862) developed "subsequently" to the consequent streams into which they flow. These are generally monoclinal water courses, which flow along the foot of steep slopes, because they excavate their paths across the less resistant rocks. The term "resequent" is not used, since this term lacks an unequivocal definition.

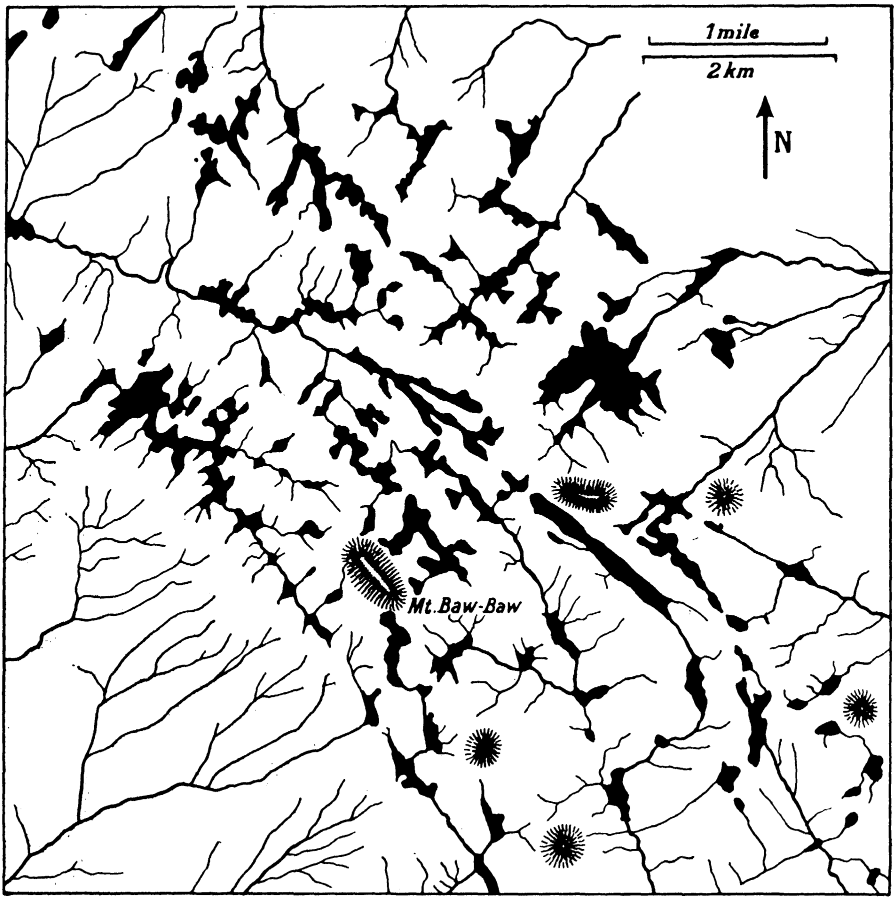


FIG. 58. THE DISTRIBUTION OF RIVERS AND LAKES ALONG THE FRACTURES OF THE GRANITE PLATEAU OF MOUNT BAW-BAW, VICTORIA, AUSTRALIA (after W. Baragwanath)

The relationship between streams and geological structure having been thus described, the consequences of modification in relief will follow. First there is the rejuvenation of this relief by vertical movements: a depression of the sea bottom, a fall in sea level for any other reason, and the uplift of continents, increase erosion until a new profile of equilibrium is established. This relationship between streams and vertical movements can be seen, both in the surface hydrography and in the "karstic" stıeam system. Fine examples of such rejuvenation are found in Norway and along the Dalmatian coast.

As mentioned above, local downwarping or uplift can lead rivers to reverse their courses either completely or in part. Thus modifications have occurred in the relationship of the Rhine and the Doubs. Captures occur and change the area served by certain rivers. This was the case of the Zambezi in South Africa at the end of the Tertiary. Certain stream patterns are completely dismembered. One example of this type, in the Sahara, dates at least from Cretaceous times. Some regions become swampy, while the drainage of others improves (Renguist, 1945). The great lakes of Finland, like those of North America, once emptied northwards. The reversal of their outlets is recent and resulted from the Quaternary glaciations and the uplift of the land, which followed the melting of the ice.

Thus it can be said that hydrography is extremely sensitive to the variations of orogenesis and epeirogenesis and that the *history* of a stream network is a very important factor in reconstituting the recent evolution of the lithosphere. On the geological scale of time, exorheic and endorheic basins are not permanent. At best it may be noted that the hydrographic evolution of old platforms is slower and relatively stabilized, so that several of them possess streams that are tens of millions of years old.

Captures.—The term *capture* is used to describe the annexation by a basin of a stream belonging to another basin.[1] Captures provide some of the most spectacular changes in river systems.

Captures are extremely frequent features in morphogenesis. Captures by overflow, that is, passive captures by oscillation on talus fans, are worthy of special note. There are famous examples, such as the Rhine, which changed its course during the Quaternary: it once flowed westward to join the Doubs, but then turned northward. In equatorial Africa, the Logone was partially captured in this way by the Benue (J. Dresch). Such reversals are commonly caused by tectonic movements, some of which are of quite small amplitude. G. Choubert (1945) explained the capture of the Moroccan oueds in the Pliocene in this way.

Alluvial Plains and Terraces (figs. 59 and 60).—On reaching a broad plain, a stream flows gently across it carrying only fine material, which it deposits on its banks and at its mouth. It thus builds up an alluvial plain which spreads gradually at the expense of the continental shelf. Often, but not always, as it passes through this alluvial plain it digs a winding, constantly changing course. This is the case of the Mississippi, the Danube and the Hwang Ho. These rivers are choked with sediment and are characterized by meanders, of which a large number are dead, abandoned, or even filled up.

These large rivers, and many smaller ones, construct alluvial plains over which they flood and on which they can shift their courses. The floodings are irregular, but generally not more frequent than one per year. During these floods, which are unsuited to the transport of sediment, the deposits are not great: 2·4 mm. for the Ohio flood in January–February,

[1] Two classic types of capture are distinguished:

1. *Active captures* due to the work of regressive erosion, where a ridge forming a watershed which formed the limit of the deeper basin disappears.

2. *Passive captures* which are stream diversions, due partly to disastrophic forces, and partly to aggradation, so that the stream spills out of its valley into a lower basin.

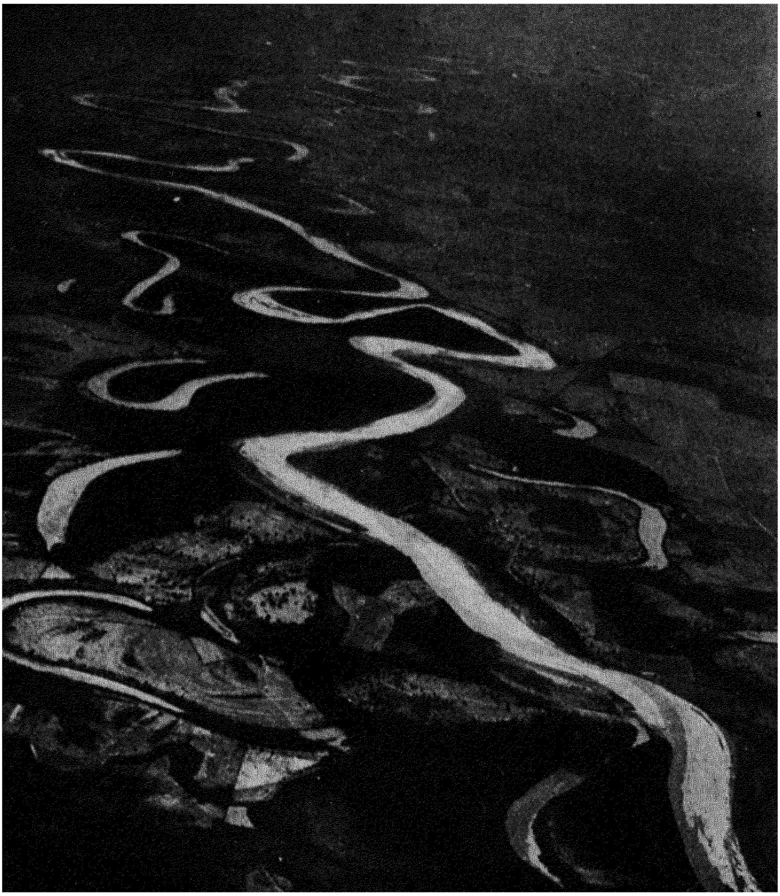


FIG. 59. THE COURSE OF A RIVER FLOWING OVER FLAT COUNTRY: THE MISSISSIPPI SEEN FROM THE AIR

Note the number of meanders: of these, some form part of the actual river, whereas others ("ox-bow lakes") are normally isolated from it. (Photograph: G. Termier)

1937; 3 cm. for the Connecticut flood in March, 1936. On the concave bank, a natural levee composed of the larger elements is sometimes formed.

Alluvial plains appear to be composed chiefly (80–90%) of "point bars". These are accretionary deposits formed on the inside of the convex bank of a river loop. They are lateral growths caused by circulatory movement or helicoidal flow, associated with the curvature of the channel. The material deposited has been torn from concave banks. There follows a lateral migration which can be from 0 to 2,500 feet per year. An example is the Kosi river, in north Bihar, India, according to Ghosh (1942). Lateral migration occurs at 30 feet per year in the Yukon and at between 150 and 250 feet per year on the Mississippi. This migration of the river takes place from one side of the valley to the other, and the valley, which is the sum total of all these meanderings, is in its lower course largely filled by beds of these accretion deposits (Wolman and Leopold, 1957).

Wolman and Leopold have demonstrated that, under normal conditions, the height of the surface of an alluvial plain remains stable in its relation to the level of the stream bed. The flood deposits are in effect controlled by

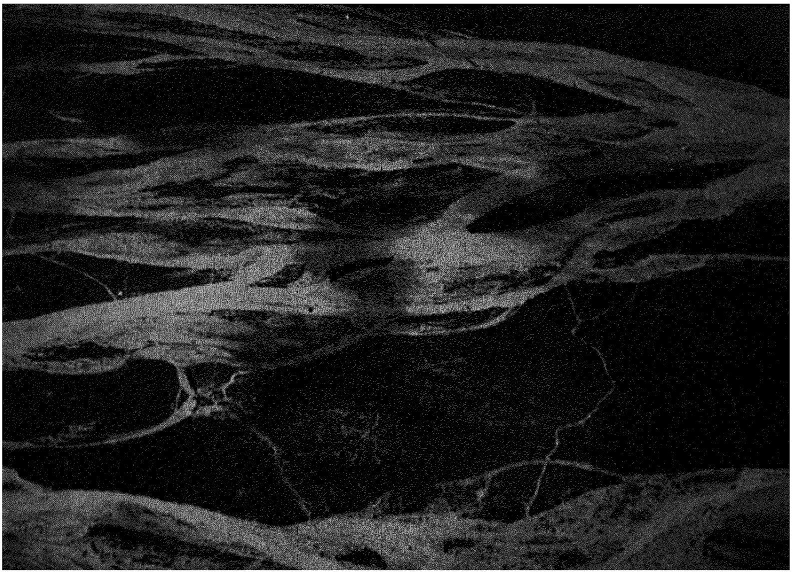


FIG. 60. THE FLOODING OF THE PIAVE, A TORRENTIAL STREAM IN NORTHERN ITALY. Aerial view taken during World War I

the lateral migration. Consequently, when terraces occur they must be the result of changes in the general conditions determining base level (e.g. orogenic or epeirogenic movements, and eustasy or climatic modifications). These causes allow infinite combinations.

Certainly, aggradation in the lower courses at present day has its origin in the rise of base level (eustatic), following the end of the Pleistocene glaciation. In Mediterranean latitudes, aggradation is also related to the drop in rainfall.

In the case of exorheic basins, it may be noted that the bed of a sluggish river is choked with sediment, which is deposited on the banks, and the river cuts its channel anew during flood time. But, in endorheic basins, the downcutting has always been proportional to the flow. Thus, the Mississippi, which receives one of the greatest volumes of water in the world, fills up its bed and vacillates over its alluvial plain. There is no doubt that periods of downcutting of a surface coincide with the uplift of the surface, while periods of aggradation correspond with a decreasing flow or even with slight downwarping. It may be noted that, during the Quaternary, the relative uplift of the land in respect to sea level coincided with the glaciations, that is, with the periods of greatest rainfall. It follows that all conditions were favorable for downcutting (in temperate and Mediterranean latitudes).

Normally, when the continental shelf is wide, streams terminate in an *alluvial delta* which advances into the sea (deltas often become established in basins which were originally subsiding). However, many of them, particularly those on the eastern coast of the Atlantic, have had their lower valleys invaded by the sea. These are maintained free from alluvium, at least in part, by tidal currents whose force is often much more than that of the river (except perhaps at flood time). These are *estuaries* like that of the Gironde.

The nature of the terrain over which the river passes, modifies the form of the river bed considerably. Through hard plateaus composed of limestones and sandstones the stream cuts its bed deeply, whatever the climate. Gorges and canyons are just as likely to occur in a temperate limestone region, such as the Causses (the Tarn gorges, France), as in the sandstones and shales of an arid region such as Arizona. In these rocks, river erosion takes place, not only by the action of the current with its load, that is, by the action of gravity, but also by the action of eddying currents. These develop especially at breaks in the profile of equilibrium, where rapids and waterfalls are produced. A type of alveolar erosion of the river bed then occurs (potholes which may join up and form channels). On valley sides, such rivers often undermine their banks, irregularly forming overhangs and grottoes which are ranged according to the state of downcutting. Fine examples are seen in the Doubs valley and the Tarn gorges of France, as well as in several streams in Morocco. The absence of abrasive material in the limestones, explains the irregularity of the profiles of karstic rivers.

Water Courses in Arid Countries (fig. 61).—Intermittent stream flow is characteristic of some climates. Examples of intermittent streams may be seen in the *oueds* (*wadi*) of North Africa, *arroyos* of Latin America and *omirimbi* (sing. *omuramba*) of the Kalahari, in South Africa. These rivers, which do not always terminate at the sea, are dry for a large part of the year. They only flow when there are very heavy rains, but then they flow in torrents. This gives them considerable power of degradation especially because they pass through regions poorly protected by vegetation (see p. 134). Rivers of this type are encountered in all desert and subdesert regions. Those of the Gobi desert are comparable with the African oueds:

generally they are almost dry and lost in the sand, but they are capable of sudden flooding, which may be at times devastating.

The *dallols* of the left bank of the Niger are large dry valleys representing a river system fed from high Saharan land. This falls imperceptibly towards the Niger and is still almost intact. Such channels are invaded by the Niger waters in spate, which create *gullies* (secondary branches for the intermittent flooding). The drying up of these dallols seems due, at least in part,

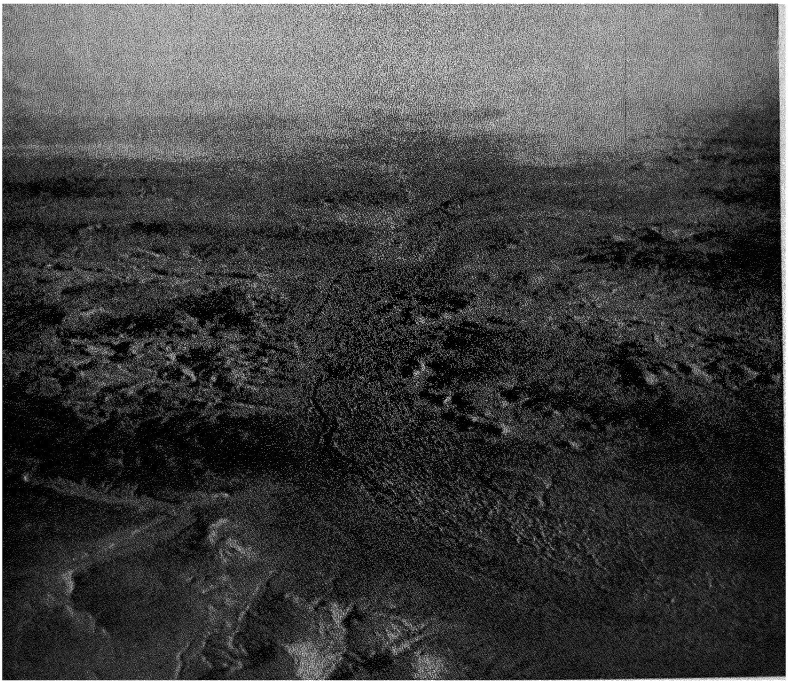


FIG. 61. ONE OF THE GREAT WADIS ("OUEDS") DESCENDING FROM THE ATLAS MOUNTAINS AND DISAPPEARING IN THE GRAND ERG OCCIDENTAL (THE GREAT WESTERN SAND SEA). THE OUED-EN-NAMOUS CROSSING THE PLIOCENE HAMMADA OF SOUTHERN ORAN. (Photographic atlas of Algeria)

to the infiltration of rainwater into sandy flat ground. If the slope is sufficient, water streams over the surface and has no time to become entrenched. Some dallols, such as the pool of Keöta (Adar Doutchi), are dammed by sand and form lakes.

Further inside the desert region, there are dry valleys with sandy floors. In Aïr, these are called *koris* and their floors are close to the water table.

Torrential mountain streams in the Barbary Atlas, have been given appropriate names by the local inhabitants. These are *asif* for an important valley, *irhzer* for narrow ravines, *talet* or *talat* for dry ravines and *agouni* for laiger ravines which are sometimes cultivated. Although they are almost always dry on the surface, they do at times possess underground water. During violent rainstorms which fall on the high ground and feed their headwaters, there are sudden floods, generally of short duration, in the course of which intense erosion takes place. Changes of the oued beds are frequent and contribute to the acceleration of erosion and to the formation of pediments. In mountain regions, downcutting is considerable and gives rise to deep gorges or defiles (kheneg) whose discharge on to the plain has received the name of *foum* in Arabic or *imi* in Berber.

The Saharan oueds certainly established their valleys during more humid periods than the present. It is sometimes difficult to determine their present base level, which is often considered to be the region where the thalweg ceases to be seen, because the water from the stream is lost before it arrives at base level. Also it is often difficult to attribute certain oueds to a particular basin, for it happens that the tributaries of one basin may be diverted toward another. In fact, these water courses must be considered to result from degradation of the Saharan river network under the action of aridity. Dubief (1953) retraced the stages of this degradation in the following manner:

(1) The oued only reaches its base level with the help of water from its lesser tributaries. (2) The oued no longer reaches its base level, even with water from its tributaries. (3) The flow of water shows breaks in the continuity of its main trunk; consequently it is difficult to determine the effective basin of the oued. (4) The tributaries become autonomous and function as main trunk streams. (5) The old main trunk disappears, or it may in a very degraded form develop a secondary artery in the present day, in one of its old tributaries.

Thus, large streams such as the oued Igharghar and the oued Tafassasset, have become of secondary importance. The Igharghar, which once had a course 775 miles long (longer than the Rhine), is at present reduced to 160 miles. The Tafassasset, until fairly recently 870 miles long (more than three times the length of the Potomac) is, today, no more than 93 miles long.

The Saharan oueds still "living" are those which rise in the Atlas, and those which flow from the southern side of the Ahaggar and the northern side of the Tassili mountains. Those in the first category include the Saoura, which flows for 500 miles, and a few oueds, which flow either to the Dra or to the Djedi. Many terminate in endorheic basins, sebkhas and maaders.

The floods feeding the Saharan oueds are in proportion to the rainfall. Although they are an annual occurrence in the Atlas, they may occur only once in five years or more in the more arid regions.

In the north of the Saharan zone, for example in Tafilelt, the floods, which are fairly regular, carry a considerable load, and the oueds become powerful agents of lateral erosion. They then establish a large plain (covered by regs) which grows continually larger, and across which they wander as they cut their channels.

Most of the oueds of arid regions, possess an *underflow*. This is a sheet of phreatic water beneath the alluvium fed in part by loss of water from the surface oued, and capable of flowing in its direction. The underflow, whose path does not always coincide with the axis of the oued, can have a heavy flow. Thus, in the Sahara, the underflows are frequently more important than the surface streams, and it may be noted that in France, in the Crau, the underflow entirely supplies the surface stream.

Allogenic Rivers Irrigating Arid Regions.—A few streams which cross arid regions rely on tributaries in their upper reaches to maintain sufficient water to reach the sea. This is the case of the Indus, the Nile, the Euphrates, the Senegal, and the Rio Grande.

The Indus owes its flow to tributaries in the Punjab, whose sources lie in the Himalaya. These tributaries are the Jhelum, the Chenab and the Ravi on the one hand and the Sutlej on the other.

Mesopotamia is a loamy desert which can be made fertile by good irrigation. This has been realized in part by means of the natural channels linking the lower courses of the Tigris with the Euphrates, and partly by means of irrigation works created by civilization of the distant past (at least 4,000 years B.C.).

The Nile, which has none of the characteristics of the Saharan oueds, also flows across the middle of the desert. Its waters come from the Sudan and from Ethiopia. Egypt owes its fertility to the Nile. The course of the river was considerably modified in several sectors during its long geological history, even with reversed flow and capture. Its middle and lower sectors are marked by oueds (here: wadis) which today are totally dry (Lawson, 1927).

This natural irrigation of normally arid country suggested to man very early that he might make the land fertile. This marked a considerable advance of the cultivators, who developed the use of the "Archimedes Screw", the water-wheel, and the Saggia (bucket-wheel). Egypt, Mesopotamia and the Lower Indus (the civilization of Mohenjo-Daro and Harappa, 1700–1500 B.C.) are very good examples.

Like the Nile, the Senegal, today 1,055 miles long, results from the junction of several water courses whose sources are in humid regions: the Bafing (the Black River which flows from the Futa Jalon), the Bakoy (the White River), swollen by the Baouli (Red River), are all situated on the northern slope of the Guinea Highlands. They cross a country which is rather arid (the northern part of the Sahelien zone) but which is not such a complete desert as the Sahara. To the north is Mauritania, the Ferlo and the Djoloff. After receiving the Faleme, which has already been in the Sahelien zone for two-thirds of its course, it receives no more permanent tributaries. Tricart (1955) noted that each year a flood of variable height alternates with a season of low water. This phenomenon is due to the seasonal character of the rainfall in the whole basin. The gradient of the Senegal is so slight that flood waters take four months to reach its mouth (from July to November). The valley is about 12 miles wide upstream from the delta, but at flood time, vast areas are slowly inundated. Adventitious or "accidental" lakes (Guiers and r'Kiz lake) then fill and act as regulators by flooding into the main river bed during periods of low water.

In humid tropical climates the rivers rarely attain their profile of equilibrium when they cross rocky outcrops. According to Baulig, this permanent irregularity, which can be seen in the persistence of waterfalls and rapids separated by quiet reaches, is due to the type of weathering. It does not, in this climate, erode resistant hard bands because it does not produce abrasive material such as pebbles or sands. Only fine silts are formed, and these are practically unable to erode.

Endorheic Basins.—Endorheic depressions are also called *bolsones* (*bolson* in the singular). This is a term designating closed depressions in the zone of Mesas in the Mexican Plateau (Chihuahua and San Luis Potosi basins, the Llano of Gigantes, and the bolson of Mapimi), as well as in much of the Basin-and-Range Province of the American Southwest (particularly Utah, Nevada and Arizona).

An endorheic basin may be established on a high plateau or at the bottom of a relict sea or a tectonic trench. In the latter cases, base level can descend a long way below the general level of the seas. In general, any isolated sheet of water, if it is permanent, forms a gathering ground for neighboring rainwash, and thus represents the center of an endorheic basin *whatever the climate in which it is found*. The lakes which are situated on old planed platforms, either in glaciated areas or in desert regions, serve as base level for the rivers which surround them.

Often endorheic basins emphasize an important paleogeographic trait and they then occur in more or less regular alignments. This is the case of the large lakes and relict seas of Paratethys: the Caspian today is −83 feet and the Aral Sea +59 feet. The African basins, the Congo and Chad-Bodelé, are similar; the relative altitude of separate sheets of water results in the drainage from one to another, often following warping of the basement. There is, then, the formation of overflow channels, which are reminiscent of fluvial valleys, but which are in fact entirely of lacustrine origin.

In the Chad region, channels and arms of the lake itself, penetrating into the shore line or separating islands aligned regularly, are called *bahrs*. They are orientated north to south, northeast to northwest, are 500–1,000 feet wide and where they leave the lake, they split into three or four branches. With respect to those *bahrs* which penetrate the shore and which could be confused with tributaries, a very slight fall in the level of the lake water is sufficient to dry them up over a large area.

The Chad basin slopes gently toward the Bodele depression. During the

Quaternary, the Chad waters flowed into this basin, by way of the bahr of Gazal, a trench 250 miles long, 65–100 feet wide, whose sandy banks are extremely irregular. This trench is connected by channels to outer basins.

The Usboi is a channel which resembles the bahr of Gazal, and which links the Aral to the Caspian Sea. This sinuous channel, about 3,000 feet wide and 70 feet deep, terminates in the Gulf of Krasnovodsk.

Among the endorheic depressions corresponding to tectonic trenches, the most spectacular is the Dead Sea (1,286 feet deep) which is situated in a graben forming a continuation of the African rift systems.

In the vast depressed region of Eritrea which reaches 380 feet below sea level, there are two large stretches of water: Lake Assale (Alel Bad) and Lake Egogibad. The prolongation of this region toward the coast of French Somaliland forms a small depression, situated to the west of Djibuti, which is occupied by a lake also called Assale. This string of lakes is parallel to the Red Sea. In Egypt, the depression of Birket Qarum (147 feet below mean sea level) near Faiyum and the large depression of Qattara (443 feet below mean sea level) are extended westward by the basin of the Siwa oasis. These depressions are bounded by sheer cliffs, suggesting that their formation was due to collapse; however, other observers consider that they correspond to deflation depressions, the desert winds having selected poorly consolidated sandy facies (possibly old delta deposits). Sheer cliffs are commonly produced in deserts, without tectonic aid.

Examples of endorheic basins also occur in Tien-Shan. Each basin is surrounded by high mountains and has a playa lake at its center. This receives all the torrential water from the mountains either directly on the surface or, indirectly, underground. In this region as in other arid regions, the rivers become choked with sediment because mechanical erosion is predominant. During this period of aggradation the base level rises. Moreover, the Tarim (Sinkiang) which is one of the largest basins surrounded by orogenic belts southeast of Tien-Shan, has, as a result of the observations of Davis and later Sven Hedin, become the classic example of an endorheic basin of which the spillway has changed in position in the course of time. The outlet 1,500 years ago was in the extreme east. At that time a playa lake, the Lop-Nor was becoming filled with alluvium chiefly derived from the Qum-Dariya. Later, a separate playa lake, the Qara-Koshun became established and the Lop-Nor was transformed into a vast plain covered with a salt crust which has taken on the form of solidified waves. The Qara-Koshun has itself become filled with detritus, so that today it is losing its importance while the Lop-Nor has been replenished by the Qum-Dariya and has become a playa lake once again (fig. 34).

Playas.—The general term *playa* is used to designate sheets of water with a high salt concentration which became established in endorheic basins.

Sebkhas.—As a result of evaporation, some closed basins are floored by a saline crust which is called *sebkha* in North Africa (the sebkha of Oran, fig. 62), *solonchak* in the region of the Caspian, and *kevirs* in Transcaspia. They are temporarily invaded by rain water, by floods, by rising ground water, or even by water drawn to the surface by capillary action. The floor of these basins is impermeable. The role of sebkhas as "evaporating pans" is unquestionable. E. F. Gautier (1908) noted their whitening from salt efflorescence during the flooding of oueds which feed them by underground drainage.

The Rann of Cutch.—An earth tremor in 1819 is said to have formed the Rann of Cutch, which is situated east-southeast of the mouth of the River Indus (Platt, 1962). This area is a depressed basin and at the present

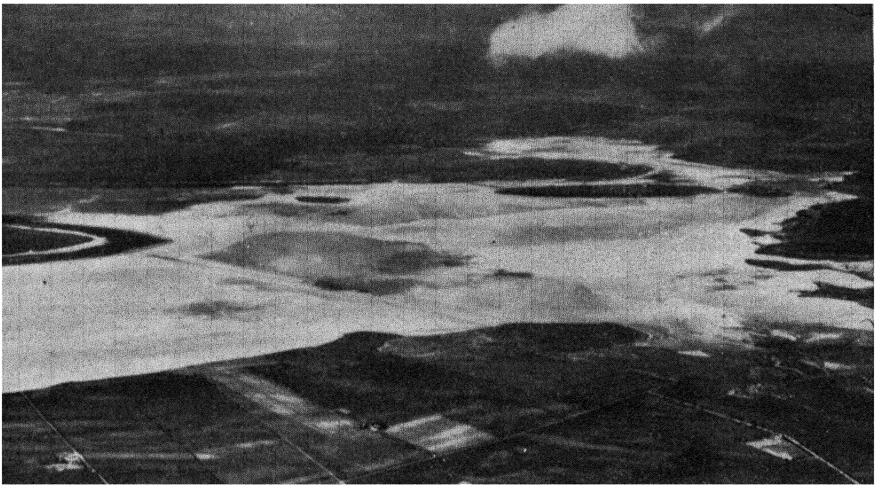


FIG. 62. THE SEBKHA OF ORAN (ALGERIA): AERIAL VIEW OF THE WESTERN PART OF THE "SEBKHA", I.E. SALT FLAT (Photograph: G. Termier)

day is an arid plain. Always humid, it contains temporary shallow basins of saline water. Inland, the Rann of Cutch passes into the arid plain of Put which in turn passes imperceptibly into the Thar desert. This is dotted with dunes reaching a height of 400 feet. Between these there are occasional lakes. The Rann and the Put are entirely desert and without vegetation or fresh water. Each year, during the southwest monsoon, the Rann is inundated by a sheet of water three feet deep. Part of this water flows to the sea, and that which remains evaporates, leaving a thick saline crust which may be between 4 inches and 3 feet thick. The Rann may thus be considered as a vast sebkha of tectonic origin.

Shotts.—Shotts are shallow, saline basins, often very extensive, which rarely contain available water, although they are always damp. They occupy the lower part of the endorheic systems of North Africa and of the Sahara (Shott of Berguent, Shott of Massiline) where they cover more than 770,000 square miles. According to M. Gautier (1953) they are zones of eolian deflation (see p. 86).

Typically the floor of a shott is formed by a breccia derived from the underlying rocks. The breccia is overlain by alluvial deposits 150–1,000 feet thick, composed by clayey sands or red clays with bands of gypsum. This is covered with a layer of fine sand (grain-size less than 1 mm.) made up of gypsum particles, with a little clay where the surface is strongly saline.

The shotts are thus the lowest impermeable areas where waters from rainwash gather after heavy rains. To this water is added that absorbed by the permeable rocks round the basin. The water table in these rocks may rise as high as the *dayas*, which are circular depressions cut in the limestones forming the upper beds of the high plains. These dayas, which are a variety of dolina, are shallow (3–6 feet) and may be a few feet to several hundred feet in diameter (as in Morocco in the Forest of Mamora, to the east of Rabat). They may drain an area of several square miles and are floored by fine sand resting on a thin, fissured calcareous crust. They support a herbaceous vegetation which makes a splash of green in the arid landscape. The oueds crossing the dayas gradually lose their water to the dayas and only rarely reach a base level which is formed by the lowest points of the basin. All this water is evaporated by the shott (Gautier, 1953). In the shott of Chergui, where the Colonization and Hydrology Service of Algeria have undertaken large-scale operations, the "sheet of water evaporated annually would have a thickness of 28 inches", equivalent to more than 1,000 million cubic yards.

The supply of water to the shotts is largely by infiltration from artesian sources, the outlets of which are scattered over the shott surfaces and are marked by mounds of saline deposits. Furthermore, the shott floors are perforated by small pipes, visible to the naked eye, through which the water rises.

The action of wind on the shotts is also important: it produces deflation of eolian origin, as for example in the Chergui shott where there is evidence of a morphological surface dominating the shott at about 165 feet, *el Gara*. The sedimentary counterpart is the existence of small dunes composed of crystals of gypsum, which are scattered over the surface of the shott.

Dayas, takyrs and vleis.—Depressions of various origin, karstic or otherwise, are temporarily transformed into pluvial lakes during the rainy season. This happens to the dayas of North Africa, the playas of South America (in the restricted sense) and the takyrs of the Aral-Caspian region (fig. 63).

Frequently the rivers of old platforms are associated with poor drainage which can be seen either at their confluence with their tributaries or near to their mouths. This poor drainage gives rise to marshes, often of a temporary nature. The *vleis* of the Zambezi Basin in South Africa are typical.

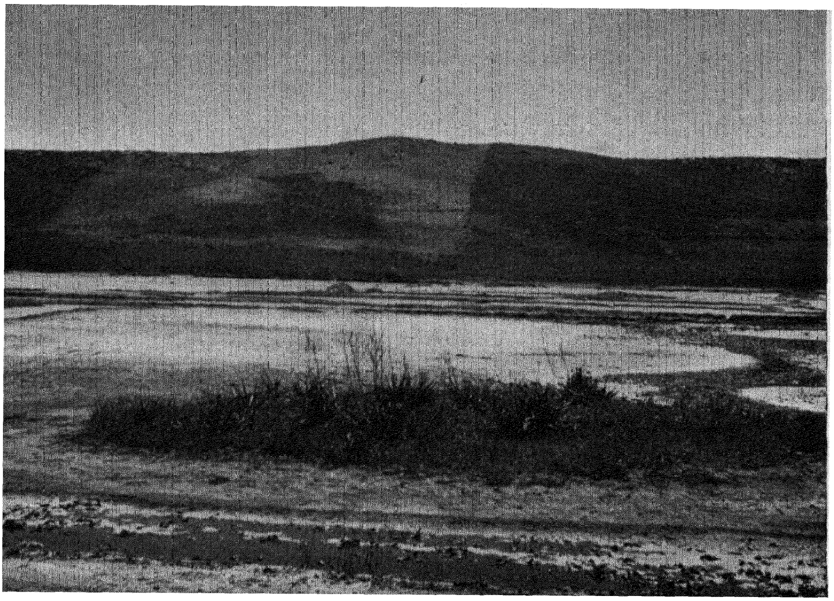


Fig. 63. A Saline "Daya" in Morocco: Agourai, at the Northern Foot of the Plateau of the Central Atlas Mountains (Photograph: G. Termier)

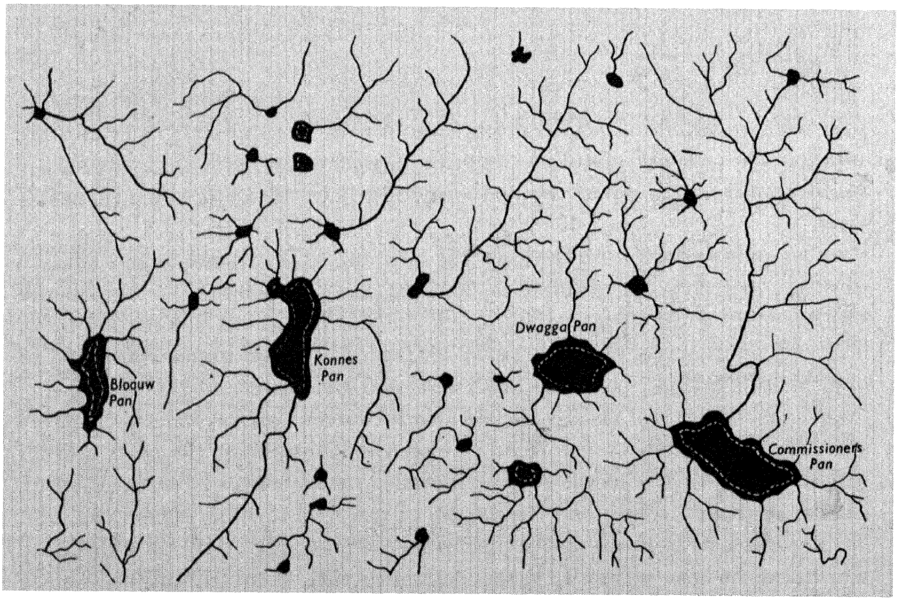


Fig. 64. Centripetal Drainage System (Endorheic Basins) of the Zone of "Salt Pans" in the Union of South Africa (after L. C. King)

They are shallow, and the water in them is evaporated by the heat of the sun. In many respects, these marshes differ little from playas. The depressions become filled up and the evaporation of water from them may reduce the flow of the stream by more than a half. Salt pans (fig. 64) which are nearly always dry are formed. These have poorly defined boundaries and may pass into prairie. Also they may coalesce to form large stretches such as Makarikari (5,000 square miles) from which emerge grassy islands. Sources of fresh water are abundant, and small lakes form.

However, the *vleis* of the great plateau of western Australia, which are greatly elongated and obviously represent ancient rivers blocked by sand and gravel, have remarkably smooth floors covered by a layer of fine sand. It may be noted that the vleis of Australia, which are not part of a functional stream, differ from those of South Africa and call to mind the sebhkas and the shotts. In fact, the authors interpret them as the last stage of degradation where the preceding stage is represented by the Saharan oueds (see p. 114).

The Consequences of Glacial Erosion

The role of glaciers in erosion is qualitatively comparable to that of streams, but quantitatively it is much more limited, because, except during glacial periods, glaciers are confined to the circumpolar regions and to high mountain chains. In dealing with glaciers, it is not possible to speak of a true profile of equilibrium. In ice sheets the speed of movement is very slow, while in the mountains it varies with the slope and can reach 78 feet per year. In all cases, however, the fashioning of the terrain by a glacier produces the same characteristics, namely, polishing and striations. The polish is produced by the ice itself and the fine particles which it contains (sand and clay). The striae are produced by blocks of rock which are carried in the bottom of the ice and which scratch the bedrock. This is true abrasion. The striae are parallel to the direction of flow of the glacier (figs. 65–68).

On a large scale, glacial erosion gives rise to forms which may often be compared to stream valleys. However, special characteristics such as cirques, U-shaped valleys in cross-section, over-deepening, and rock bars can always be recognized. On plateaus, the topography is gently undulating, or *moutonnée*.

It seems that frequently, if not always, glacial forms are derived from the valleys of pre-existing rivers, the glaciers having merely modified the effects of water erosion.

Thus in southeastern Norway (fig. 69), the *Gudbrandsdal* leaves Lake Lesja and separates the two enormous monadnocks of Dovre and Jotunheim. It then falls into Lake Mjøs having crossed the Paleic surface[1] in the region of the Sparagmite. This is an ancient valley displaying in its lower

[1] Surface probably of early Tertiary age.

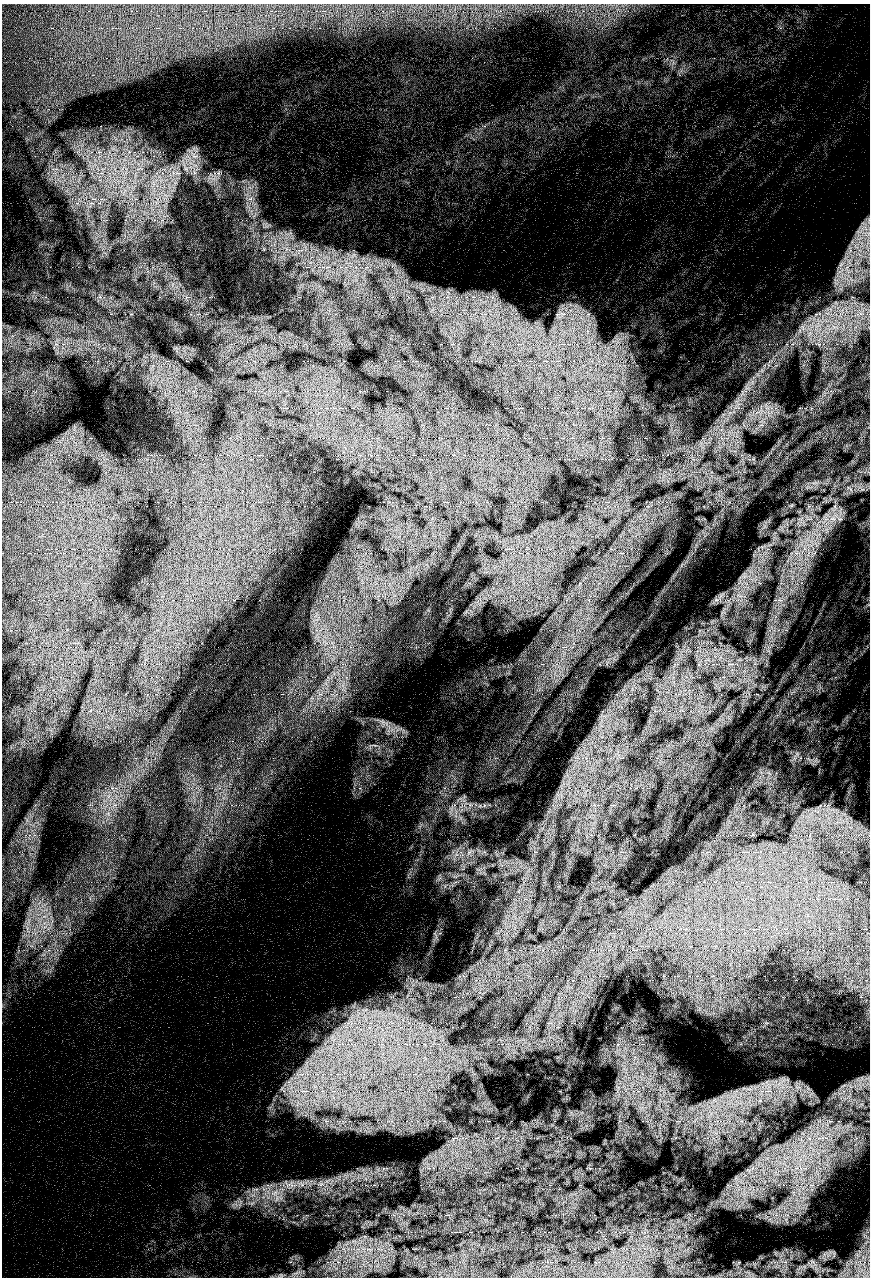


FIG. 65. THE EDGE OF A GLACIER AND ITS BEDROCK: MER DE GLACE, NEAR CHAMONIX, FRENCH ALPS

Note the morainic blocks which are pinched between the ice and the rock. As the ice advances, these produce the characteristic striations. (Photograph: G. Termier)

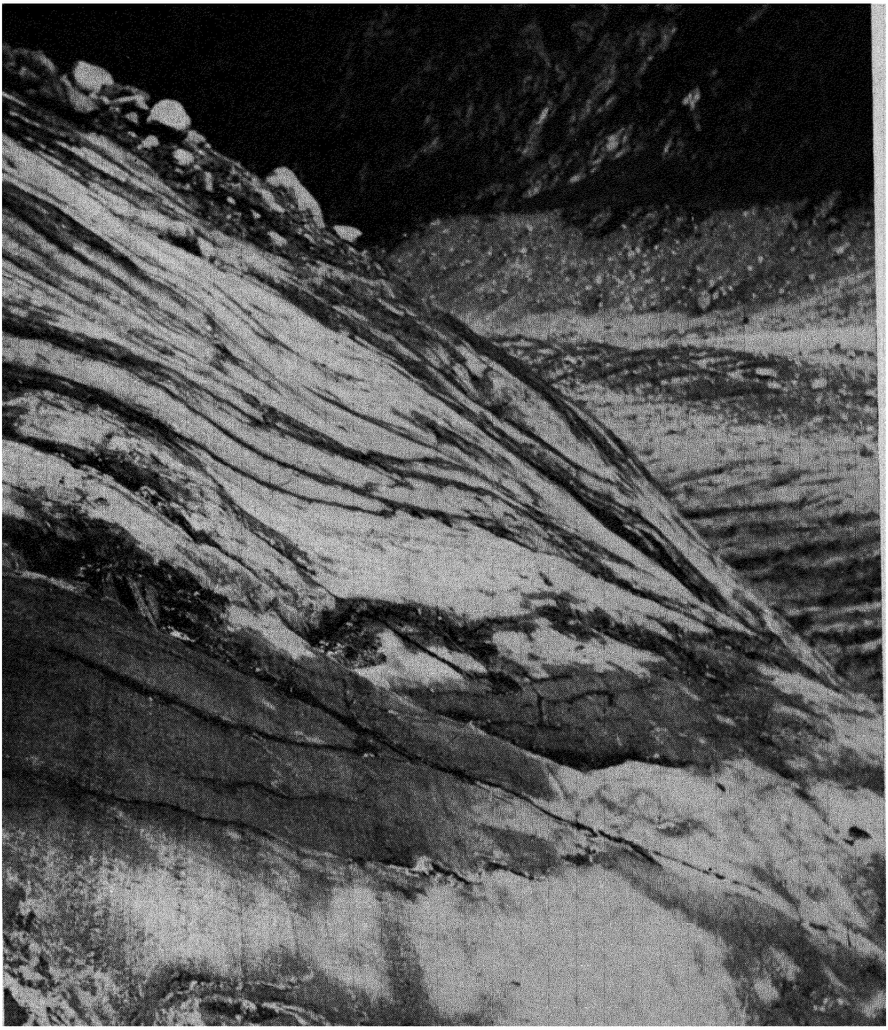
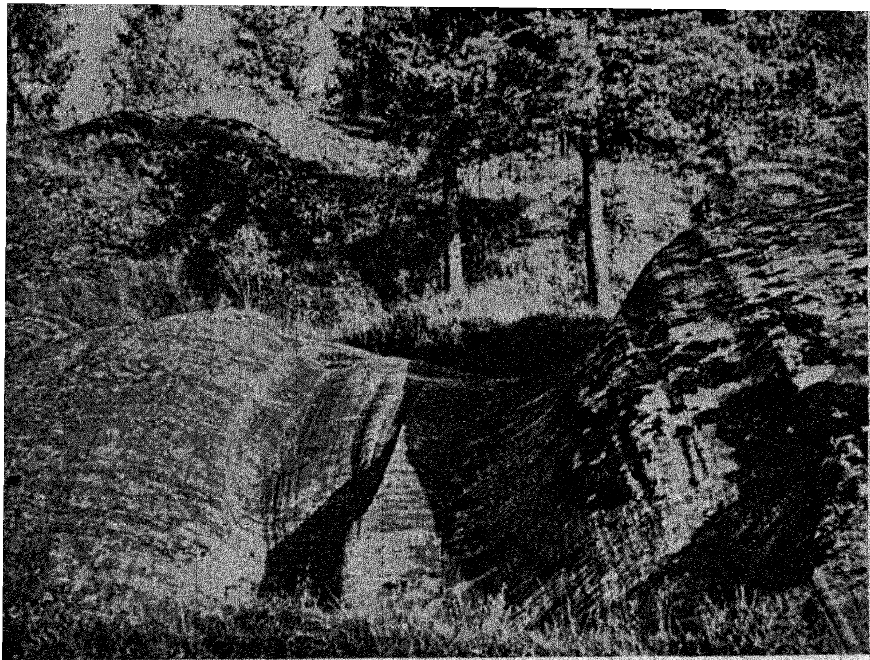


FIG. 66. STRIATED ROCKS AT THE MARGIN OF THE MER DE GLACE, LEFT BANK BELOW MONTANVERS (Photograph: G. Termier)

reaches numerous traces of glacial erosion, and in particular, numerous lateral hanging valleys. It seems that before the glacial epoch, the Gudbrandsdal was a winding valley and that the Quaternary glaciers have given it its present quasi-rectilinear form by planing away the rocky spurs and meanders. Following the retreat of the glaciers, intense erosion has led to the formation of alluvial deltas at the mouths of large adjacent valleys. Lake Mjøs, which is a continuation of Gudbrandsdal, is a lake with a natural



IG. 67. PRECAMBRIAN GNEISS SMOOTHED, POLISHED AND STRIATED BY THE QUATERNARY CE SHEET OF SCANDINAVIA (ROGNSTRAND, SOUTHERN NORWAY) (Photograph: G. Termier)

FIG. 68. STRIATION LEFT BY THE QUATERNARY ICE SHEET ON A ROCK IN FAKSE QUARRY (DENMARK) (Photograph: G. Termier)

Note the lunate "chatter marks"; the ice moved from left to right.

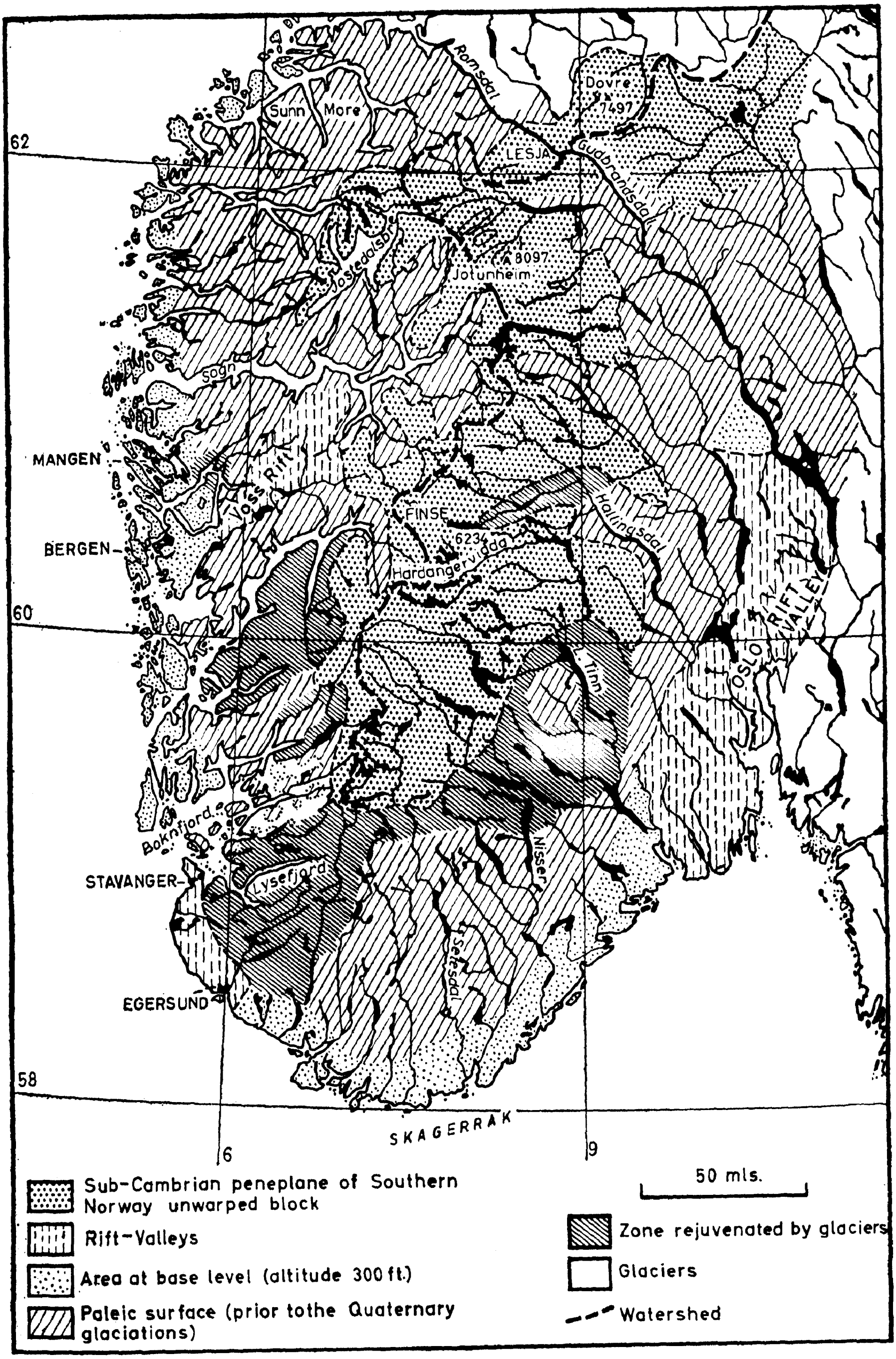


FIG. 69. SKETCH MAP OF THE STRUCTURE OF SOUTHERN NORWAY

In black, the present-day lakes. The area 300 feet above base level is an old sea shore which has been uplifted.

dam. It is very deep (1,480 feet) and was excavated by glacial erosion which followed the diıection of fault lines.

The *Hallingdal* is another valley which commences in the region of Finse, where the sub-Cambrian peneplain of Hardanger Vidden is preserved. It is marked by a string of shallow lakes which indicate the action not only of a river system but also of glacial erosion on the Paleic surface. Like the Gudbrandsdal, the Hallingdal appears to have reached maturity before the glaciation and to have been rejuvenated before the ice.

The *Lake of Tinn* is also of glacial origin (1,300 feet deep) and is at the confluence of three rivers. It is elongated like a fjord and is connected to Hitterdal Lake by a fluvial gorge which passes into Lake Nordsjø (575 feet). From the latter a river flows past Skien into the Skagerrak at the level of the faulted depression of Langesund. The system obviously owes its origin to glaciation. According to Ahlmann (1919) the deepening of the Lake of Tinn occurred during a long interglacial period. The formation of the gorge which links it with Hitterdal Lake was due to fluvial erosion. To the south of the Lake of Tinn, Gausta (6,160 feet) appears to have been a "nunatak" because it has escaped glacial erosion and must have stood above the ice sheet (fig. 70).

Lake Nisser flows into the Niddal which itself flows into the Skagerrak to the south of Arendal. In the north where it resembles a fjord, this is a shallow lake displaying roches moutonnées and youthful characteristics. Downstream it passes into a hilly region with rocky ridges typical of the Scandinavian Shield, and reminiscent of those areas where partial submergence has given rise to the Stockholm archipelago and Lake Malar. The Niddal crosses the coastal zone and continues underwater where it forms a "fjord" indicative of a former higher land level.

The *Setesdal* opens out into Bylandsfjord and ends at Kristiansand. It is a mature valley with paleic topography throughout and passes through the coastal zone at an altitude of 330 feet.

Fjords.—A fjord is a prolongation of the sea into the land. It has a U-shaped valley and is normally enclosed by sheer walls. It may have tributaries, but these are often "hanging valleys" with waterfalls. Fjords are thus genetically associated with glacial periods and generally form a number of basins in their lengthwise profile, due to the ice undulation. This "over-deepening" may exceed 3,000 feet. Superficially similar phenomena, sometimes classed as fjords (but simply "drowned valleys"), occur in regions which have never been glaciated, such as the Dalmation coasts (the mouths of the Kotor and the Cattaro). There does not seem to be any fundamental difference between the typical fjords, scooped out below sea level, and some rivers and even strings of lakes like those of southern Norway (Lakes Tinn and Hitterdal, noted above) and those of northern Sweden (Torne Träsk). Moreover, the fjords are generally continued inland by river valleys.

Fjords may be seen in a number of countries. The most famous, and which

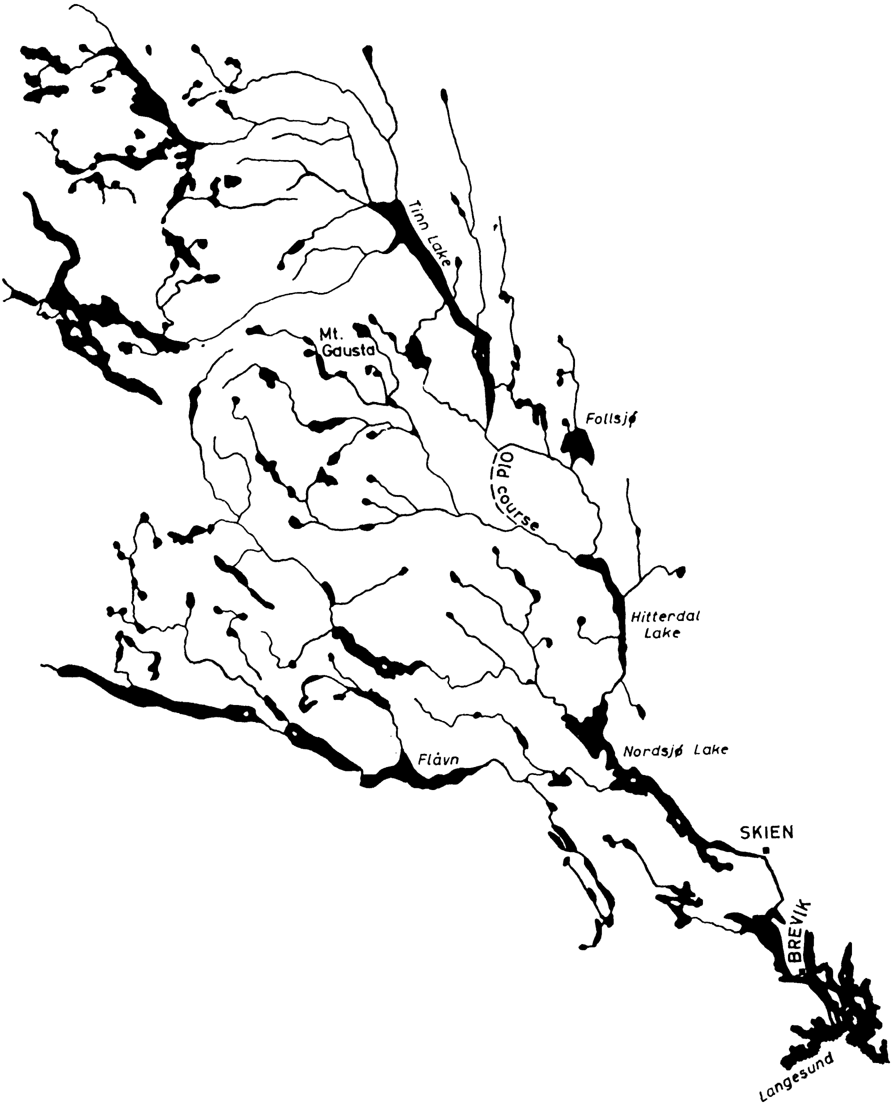


FIG. 70. THE TINN, HITTERDAL AND NORDSJØ LAKE SYSTEM (SOUTHERN NORWAY) (after Ahlmann)

Illustrating an alternation between fluvial and glacial erosion during the Quaternary.

have given them their name, are those of Norway (see fig. 69). In Scotland, there are also the "firths". In the southern hemisphere fjords are typical of the southwestern coast of New Zealand. The deepest fjord is Baker fjord in Patagonia.

A few examples from southern Norway are described here:

Lysefjord (1,427 feet deep) is a typical fjord which seems to owe its original downcutting to fluvial erosion. At a later stage it was broken up into basins by rock steps which were the result of glacial action, aided by the jointing of the schists which form the bedrock. It has been calculated that this erosion reached a depth of 1,300 feet below sea level. The fjord walls are dissected by deep ravines which descend from flat-topped interfluves. These are postglacial features cut by melt water from the ice, and the vigor of the downcutting is largely due to the absence of vegetation.

The *Bay of Bokn* is in the region of the Caledonian folds where the deformed Precambrian surface has influenced the positions of certain fjords, as for example the Sande, where the valley coincides with a syncline. Off the Bay of Bokn, the sea floor is flat and covered with Quaternary glacial deposits.

The vast system of the *Hardangerfjord*, which is also in the folded Caledonian zone, exhibits widespread planation of the shield (4,000–5,000 feet). The fjord commences near the Hardanger Vidden, while at its mouth, flat spurs still show a discontinuous surface at 65 to 130 feet. This surface represents an old base level related to an early valley system. In the Hardangerfjord there are many hanging valleys which give rise to spectacular waterfalls. These hanging valleys are sometimes produced by glacial overdeepening and sometimes, according to Ahlmann, they result from small lakes on the Paleic surface being cut across by the steep sides of the fjords. Glacial erosion is entirely responsible for the morphology of the submerged part of the Hardangerfjord, which is between 3,000 and 3,300 feet deep. One of its branches, the *Eidfjord*, is fed by a lake, the Eidfjord-Vand, from which it is separated by a terminal glacial terrace. It is a basin of glacial confluence fed by valleys originally cut by fluvial torrents and guided by fractures, which were remnants of the Paleic surface. These valleys are still V-shaped, but they have been smoothed by ice. The heads of the fjords formed in this way are called *botn, dalbotn* or *saekkedaler* (Helland, 1875) and differ from the true glacial cirque or *fjeldbotn*. They are almost the equivalent of the *trogschluss* or *kar* of the Alps.

The *Bergen Arc* is an ancient massif, occupying a special place in the Caledonian chain. It contains a system of fjords which preserve the imprint of the geological structure. One group, e.g. the Osterfjord follows more or less the northeasterly direction of the Caledonian chain, while others, particularly the secondary valleys, are perpendicular, and follow the northwesterly direction of the Precambrian folds, as for example the Fensfjord. This stream system implies a long history of erosion, accompanied by the revival of ancient structures. The Osterfjord is a mature valley with low sides along the length of its course above sea level, its submarine course alone being entrenched. In this submarine part it is a typical fjord and must have been cut when sea level was very much lower. It is a collection of deep basins, often 2,000 feet deep, separated by rock steps, one of which forms islands barring the mouth of the fjord. The glacial topography of the Bergen region has disrupted the pre-existing valley system and has created cirques at low altitudes. The ice-sheet rejuvenated the pre-existing valleys and erased the details of the levels on the coastal plain, but it has left untouched the high levels of the initial plain and thus has accentuated the available relief.

Inland, the complex tectonic zone of Voss constitutes a depression occupied by a collection of lakes and valleys which fall into the Bergen system of fjords and which, toward the east, merge into the Paleic topography of the Hardanger Vidden.

The largest fjord in Norway is the *Sogne*; it is also the deepest. Its source is in the Jotunheim, whence it crosses the Precambrian and Caledonian crystalline rocks and then, near its mouth, the Paleozoic rocks. At the mouth it is barred by a rock sill, and there is a large basin excavated by glacial action which is more than 3,000 feet deep. All its tributaries descend from altitudes which may be as much as 3,500 feet above sea level. Inland the topography remains young. The Sogne still winds like a river valley, and erosion has been guided by jointing. The *Laerdal*, which enters the Sogne, has its source on the watershed between the Atlantic slopes and the Skagerrak. This is formed by the unwarped block of southern Norway which is a flattened area covered by bogs. This ancient tableland separates the Laerdal from the Valdres which descends toward the Ringerike at the northern end of the Oslo trench. The upper reaches of the Laerdal, situated on the plateau, appear to result from the recent capture of a stream which previously flowed toward the east.

Everywhere the excavation of fjords and of hanging valleys is a very distinct phenomenon.

The *Naeröfjord* is a left bank tributary of the Sogne. Near its exit there is the famous landscape of Stalheim (fig. 71). This is an ancient valley situated about 1,300 feet above sea level, and sloping toward the west. *Naerödal* is a more recent valley which flows toward the north and has cut or captured the ancient valley and has divided it into two branches: the Sivledal and the Ophiemdal (two gorges separated by the Stalheimklev). The character of the Naerödal is due to its nearness to base level, which results in its steeper profile of equilibrium.

The Sogne exhibits many characteristic features of glacial topography. These include the "rasskars", a sort of hanging cirque which acted as a stone chute. They formed at the end of the glacial epoch, when the ice was confined to the fjord channels. Their origin was essentially due to a change in climatic conditions. Landslips of solid rock have sometimes occurred (such as that of 26th May, 1908, in the narrow Norangerdal) and these form enormous cones. The Sogne also displays "glacial slabs" (or "boiler-plates") particularly around Josterdalsbre, a broad plateau still occupied by glaciers.

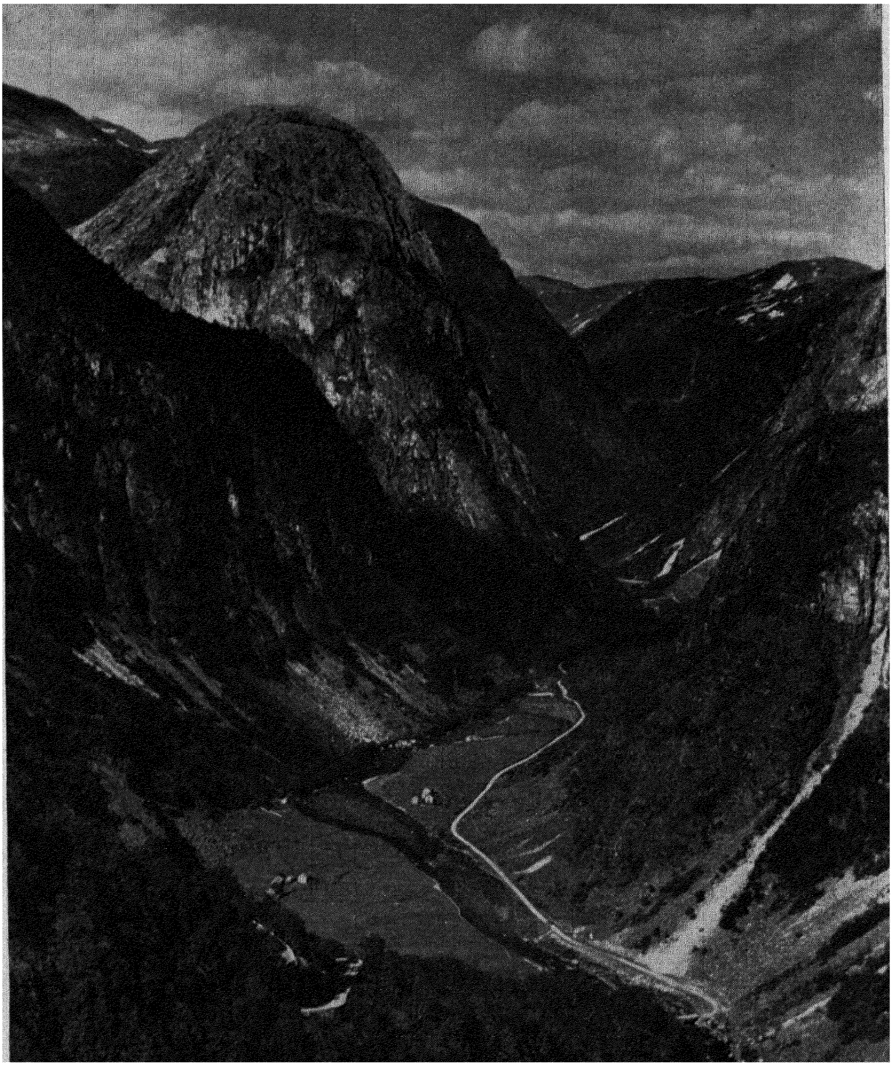


FIG. 71. THE COUNTRY NEAR STALHEIM, SOGNE DISTRICT, WESTERN NORWAY

Note the U-shaped glacial valley of Naerofjord. The dome (Jordasnuten) is formed of anorthosite. (Photograph: K. Harstad, Kunstforlag, Oslo.) Compare with Yosemite Valley, California.

FIGS. 72 and 73 (opposite). PINGOS ON AN ALLUVIAL PLAIN IN GREENLAND (permafrost phenomena)

Fig. 72. Group of pingos on Tobias Dal.

Fig. 73. The first stage. A pingo 100 feet high in Randböldal, Cape Franklin, East Greenland.

(Aerial photographs: E. Hofer, Expedition of Dr. Lauge-Koch.)


These are roches moutonnées formed by tongues of ice between which there are narrow rocky ridges. This is the result of lateral erosion by the glaciers.

Sculpturing by Ice Sheets

Regional glaciers like those of Greenland and the Antarctic today are likely to modify considerably the general features of the land which they cover. There are strong reasons to believe that the weight of Quaternary ice sheets situated on the Canadian and Fenno-Scandinavian Shields caused these continental areas to be depressed just below sea level. It would seem that the load of ice on the Greenland shield has the same effect today.

Ice sheets and regional glaciers formed in basins by the confluence of mountain glaciers (e.g. in Alaska and Tierra del Fuego) flow much more slowly than the mountain glaciers, and terminate in the sea. Their erosional role is chiefly that of grinding the bedrock to give it the appearance of roches moutonnées and to striate it in the direction of movement of the ice. Their deposits are moraines situated chiefly at the edge of the ice sheet.

Glacial sediments which can be recognized in the geological succession are dealt with on pp. 159–164.

Pingos (figs. 72 and 73) are seen in glaciated regions (eastern Greenland, northern Siberia and on the Mackenzie delta in Canada) where the subsoil remains permanently frozen ("permafrost") (see fig. 35). They are mounds rising up to 150 feet, sometimes forced up and out in the form of a crater, so that they have the appearance of small volcanoes.

F. Müller (1959) has shown that pingos are formed where phreatic water penetrates into a frozen layer of soil and forms what might be called a "hydrolaccolith". When this freezes, it increases in volume and the pressure of crystallization becomes sufficient to give the structure an eruptive character. Pingos can occur in groups or rise inside older pingos. Their mounds of circular or oval form are conical or elliptical, and the uplift by the ice leads to the wrinkling and finally the bursting of the top.

Numerous small round ponds 50–250 feet across and 10–20 feet deep, situated across northern Germany to the Paris Basin, are interpreted as late Pleistocene pingo basins. Their age is proven by pollen analysis.

6

Soils—The Relation between Erosion and Vegetation

Plato was probably one of the first authors to discuss the relationship between deforestation and the erosion of the soil. For in "The Critias" he says, in translation: "Thus, in the midst of numerous and terrible floods which had taken place during nine thousand years . . ., the soil, which these upheavals had caused to slide from the heights, did not accumulate on the land as in other countries, but rolled over the shore to be lost in the depths of the sea. So that, as happens in the small islands, our country, what now remains of it compared with what then existed, resembles a body emaciated by illness: all that was once rich and fertile land was carried away from all parts, and there remains no more than a skeleton. But previously Athens, whose soil had still not suffered any disturbance, had in place of mountains, high arable hills; the plains which we now call the fields of Phelleus[1] were filled with an abundant and fertile soil; the mountains were shaded with thick forests of which visible traces still remain" (from the French translation of Dacier and Grou, 1869, p. 312).

The close interrelationship which exists between erosion of the soil and vegetation cover is now well known, having been brought into prominence by the disasters which have occurred in the prairie regions of the western United States. There, cereals are the main crop, and during the dry season the wind carries away the fine dusty soil which is not bound by the deep roots of plants. This is comparable to the effects of wind over noncoherent rocks such as sand, where dunes are moved. On mountains, the action of rainwash on soils is similar in that it removes them from slopes where they are not held by plants.

Thus, the mantle of vegetation protects the surface of rocks, slows down erosion, and tends to stabilize the relief. Certain events in human history can without doubt be explained by the modifications which man has made in the density of the vegetation. This is particularly true of the margins of the Mediterranean. Foı example, Morocco in the Middle Ages was, according to the Arab historians (notably Ibn Khaldoun and Léon l'Africain) covered by forests. These were destroyed, and today many areas have become desert

[1] φελλεύς; a field of stony soil, or a porous rock, like pumice (Translator's footnote).

(Ben Guerir, Guercif, etc.). It is probable that this is a process analogous to that which caused prosperous areas to become more and more arid during the early periods of history, such as Palmyra in eastern Syria, or even southern Italy, where the Paestum roses are reputed to have flowered. However, in addition to the destructive role of man the sharp climatic oscillations of the last two millenia should also be considered.

In considering geomorphic sculpture throughout geological time there arises the problem of the effects of organic evolution on erosion. Plants comparable to those of present-day forests are unknown before the Upper Devonian. The Precambrian, Cambrian, Ordovician and the greater part of the Silurian periods have yielded little evidence of plants other than bacteria and algae, that is to say, aquatic plants. A protective mantle of vegetation, which may have consisted of Schizophytes on the land, seems possible only if the climate was particularly humid, or conditions were swampy. It is possible that lichens played some part.

Before the expansion of the vascular Cryptogams, rocks could have been preserved from destruction, as they are today in arid climates, by elements deposited in them forming calcareous or siliceous crusts, and by "soils" in the geomorphological sense of the term. (For example, granitic sand which may reach thicknesses up to 50 feet.)

The study of the relationships between erosion and vegetation leads to a consideration of sedimentation. Sedimentation arises from erosion, and it is weathering which leads to soil formation and the possibility of vegetation. Indeed vegetation produces organic substances and is capable of altering the chemical composition of rock minerals, either by the products of plant metabolism (e.g. humic acids) or by the purely physical action of fixation which results from the presence of the plant cover.

As has been noted by J. van Baren (1928), a soil not only shows characteristics derived from the parent rock, but also acquires new properties. Four variables influence the formation of soils: climate, parent rock, topography, and time.

Between a rock and the soil derived from it there are a series of intermediate stages or "horizons". In general, there is a lower horizon (the "C Horizon") containing minerals that differ little from those of the parent rock, and an upper part which is argillaceous. This latter part is considerably altered by superficial weathering and may have material derived from elsewhere added to it. It is divided in two: the "B Horizon" often enriched by downward chemical transport; and the "A Horizon" at the surface, greatly modified by plant action and humus. Thus the parent rock and the climate are the dominant factors controlling the formation and composition of a soil. It must be remembered, however, that the minerals of rocks can be altered without exposure at the surface by pneumatolytic or hydrothermal action.

According to D. Carroll (1934) the most stable minerals in soils are the "stress minerals", i.e. those resistant to compression. Thus zircons from granites, magnetite and ilmenite from basic rocks, and ferro-magnesian minerals from all sorts of crystalline rocks are found in soils. The latter group are ultimately destroyed by prolonged weathering, such as that which produces laterites. A soil is *mature*, when all the unstable grains have been transformed into argillaceous or other minerals more in equilibrium with the conditions present in the zone of alteration (see p. 138).

The nature of the alteration varies according to the mineral affected. *Kaolinization*, which is sometimes due to hydrothermal action, can also take place in the soil in an acid environment. More often, however, the clay formed is beidellite $(Al,Mg)_4(SiAl)_8O_{12}(OH)_{20}$ which is closely related to montmorillonite. Kaolinite is frequently one of the major components of laterites.

The pH value influences the mineralogy of soils:

at pH 4 aluminum silicate is precipitated
at pH 5 $Fe(OH)_2$ is precipitated
at pH 5·5 to pH 8 colloidal silica is formed.

Finally, in conditions of extreme acidity, aluminum silicates are decomposed and alumina (Al_2O_3) is liberated, as in lateritic soils.

Titanium oxide shows a tendency to separate from iron oxide. The alteration of titanium minerals and ilmenite to leucoxene (a variety of sphene) ($CaTiSiO_5$) is thus one of the important changes taking place in the weathering of rocks. Carroll believes that leucoxenization takes place most readily in soils rich in calcium, because leucoxene itself contains calcium. Under acid conditions anatase and brookite are formed instead of leucoxene. Thus, during pedogenesis (soil formation), the nature of the alteration of the titanium minerals depends more upon chemical diagenetic environmental conditions than on the nature of the parent rock.

Possibly the titanium oxide behaves in the same way as silica (Carroll, 1934), although colloidal titanium oxide forms between slightly different pH limits (pH 3 to pH 4 according to Brammall and Harwood, 1923). When calcium is present in the soil as a result of leaching of limestone, leucoxene is produced as a gel.

Iron solutions behave differently according to the climate. The stable iron oxide in soils is ferric hydroxide. It goes into solution at pH 3, while ferrous hydroxide does so at pH 5·5, and the peptization of colloidal ferric hydroxide occurs at pH 6·6, under the influence of humus and colloidal silica which is present in the soil. These changes take place especially in the podsols (see later).

Finally, alum is present deep down in certain saline soils (see p. 317). Australian geologists believe that the salt and the gypsum which impregnate soils of arid regions are brought there mainly by wind and rain, rather than by local drainage. The potash is derived from the bedrock.

The structure and density of soils depend upon the speed of development: if it is slow, the grains of the sandy part are fine, the ferro-magnesian minerals are almost completely destroyed, and the feldspars show the maximum degree of kaolinization.

THE RELATIONSHIP BETWEEN VEGETATION AND THE WATER TABLE

In humid, temperate regions, plants obtain enough water from the soil and from the air for their roots to be placed above the water table in the most favorable conditions. In arid countries, the humidity of the soil is slight and that of the atmosphere is variable; most plants are unable to survive in these conditions. One class, the *xerophytes*, are able to tolerate these conditions and build up reserves of water; generally they have roots which are situated above the subterranean water and its capillary fringe (p. 79). This is true of the American cacti and certain Euphorbias. Most other plants in dry regions need roots long enough to reach down to the water table, or at least to its capillary fringe; these are the *phreatophytes* (O. E. Meinzer, 1923). In the Sahara, the date palms belong to this class. Poplars, willows, some tamarisks and *Acacias* are also phreatophytes (T. W. Robinson, 1958).

Thus the position of the water table is of great importance in determining the composition and fertility of the soil. This is especially true of laterites (tropical forest soils) and of those saline soils where the water table intersects the surface. It should be noted that the *hydrophytes*, which live in water, have a distinctive ecology.

The geological role of the phreatophytes is important. In the first place, the extraction of water by them modifies the equilibrium of the hydrosphere, particularly in dry countries. Secondly, the chemical composition of the water has a marked biological effect on the plants which use it. Thus willows and poplars which cannot tolerate a high concentration of salt, live either on the upper reaches of rivers, or on alluvial plains. On the other hand, the tamarisks, *Sarcobutus vermiculatus* and *Allenrolfia occidentalis*, which can withstand very variable amounts of salt, live on the edge of playas in the desert areas of the Great Basin of America. The concentration of certain phreatophytes, such as the tamarisks along the rivers of arid countries finally results in the blocking of the rivers, causing floods and the deposition of thick layers of silt (4 to 5 feet in the alluvial plain of the Gila in the Great Basin in 1954). Finally, the plants absorb traces of rare chemical elements dissolved in the underground water. These elements can be utilized as indicators of mineralizations. For example, *Cowania stansburiana*, a common plant in Arizona, Idaho and Utah, absorbs large quantities of uranium and vanadium.

ELUVIUM

The weathering and leaching of the upper part of the earth's crust has produced *in situ* disintegration products or *eluvium*. Their formation shows the characteristics of a cycle, the *weathering cycle* (Polynov, 1937), which differs according to the rock attacked. This cycle has two end products: *residual soils* and *soluble substances* (Chebotarev, 1955).

The cycle corresponding to the weathering of *crystalline* rocks or *ortho-eluvium* (Polynov, 1937; Chebotarev, 1955) contains the following residual material: insolubles, lime nodules (if protected by a plant cover), kaolinite and illite, and may form a crust of lateritic type. The material removed includes chlorides and sulfates, some calcium carbonate, and aluminum silicates, which are deposited elsewhere as alluvium.

The cycle of alteration of *sedimentary rocks* or *para-eluvium* is more simple and leaves a residue of detrital material, aluminum silicates, and possibly some illite, while the transported materials always contain chlorides, sulfates, carbonates and aluminum silicates.

Finally the weathering of recent soils or *neo-eluvium* gives a residue of solonetz (see p. 324), gypsum and aluminum silicate, while the material removed contains chlorides, sulfates, calcium carbonate, and alumino- and ferri-siliceous substances.

The processes of weathering of rocks of the lithosphere thus lead to the formation of quartz grains, carbonates, kaolin, limonite and chlorides, and to the alteration of ferro-magnesian minerals into other, more stable minerals. Consequently there exists a mineralogical equilibrium at the surface toward which minerals under the influence of weathering tend to move.

THE ROLE OF BURROWING ORGANISMS IN THE FORMATION OF SOILS

The upper part of soil is rich in organic matter, especially decomposing plant remains. In forests, for example, fallen leaves which are turning into humus are often linked by fungal hyphae. This humus is a valuable nutrient, not only for plants and bacteria, but also for numerous burrowing organisms, especially earthworms. These feed directly on the humus of the soils, which they mix up, aerate, improve, and fertilize before replacing it. The name *mull* (Müller, 1887) is given to those reworked soils which contain traces of animals (worms, insects, myriapods) in the form of their excrements.

In the tropical zone, a very important part is played by termites. Nazaroff (1931) was the first to notice their role in the formation of laterites. Erhart (1951) has shown the association of their fossil nests with the tropical soils of Africa. The present-day habitat of tropical termites is widespread (from the forest to the desert). They have acted both as agents of diagenesis and as transporters (for example, by incorporating grains of quartz in ironstone layers). Chemically, they assist the passage upwards of lime and phosphoric acid (Bouyer, 1949). Mechanically, they loosen the soil and render it more suitable for the formation of concretionary ironstone layers.

DEFINITION AND CLASSIFICATION OF SOILS

Definition

To the geomorphologist, the *regolith* is a layer of debris formed from the underlying, more or less fragmented and disintegrated rock. This layer is, in fact, composed of eluvium and in its upper part, of alluvium. A *soil* forms on rock or on regolith as a result of the work of roots, burrowing organisms, the infiltration of water, freezing and thawing, variations in temperature and humidity, and chemical alteration: that is, geochemical and biochemical evolution. A soil is comprised of an impoverished zone, the *eluvial horizon*, and also a zone of enrichment or accumulation, the *illuvial horizon*. Among the eluvial horizons may be cited, for example, the *reg* of desert regions where the wind has removed the finer particles, leaving a concentration of the pebbles. The development of soils is barely possible on steep slopes. A true soil can only fully develop on flat surfaces or where the slope is very gentle, since the components can then be transformed without being removed. Bacteria occur there in great number and play an important part in the development of the soil.

Soils depend directly on the conditions of weathering and on the vegetation. Thus, according to the climate, soils develop in different ways. Under the influence of gravity, soils may move very slowly (*soil creep*, Davison, 1889) even though they are covered by vegetation as in New Zealand (temperate, humid climate). In periglacial zones soils move by *solifluction*, and in tropical, humid climates by mass movement. In dry climates soils are discontinuous.

Classification

The development of a soil tends to be counterbalanced by erosion. When equilibrium between formation and erosion is reached, the soil is said to be *mature*, when erosion is greater than the rate of formation, the soil is *immature*. If the soil is formed more rapidly than it is eroded, it is a *planosol*.

Immature soils may retain some of the characteristics of the underlying rock. Thus, grains of quartz, clay, and iron minerals may occur in the soils of granites; clay dominates in those of schists, while calcium carbonate is present in soils on limestones.

The soils of glaciated regions, *frozen soils*, *permafrost*, *tjäle*, where biological activity is limited, are rarely mature. These are the typical soils of tundras. They belong to the pedocals group (see below).

Three principal types of *mature soils* can be recognized (Marbut, 1935; Robinson, 1951). In temperate climates where the humus is less rapidly destroyed by bacteria than in the tropics, two types, the pedalfers and the pedocals, can be recognized. These are developed on practically all types of rock and show a number of variants according to the plant cover (fig. 74). The third type is the chernozem (*see below*, under Pedocals).

The *pedalfers* are soils from which the soluble salts are completely leached and generally occur in the forests of humid regions. They show, close to the solid rock, a banded zone preserving the original structure, but in which the iron minerals are oxidized and hydrated. The lowest part is a zone of enrichment to which the leached-out soluble salts from the upper part of

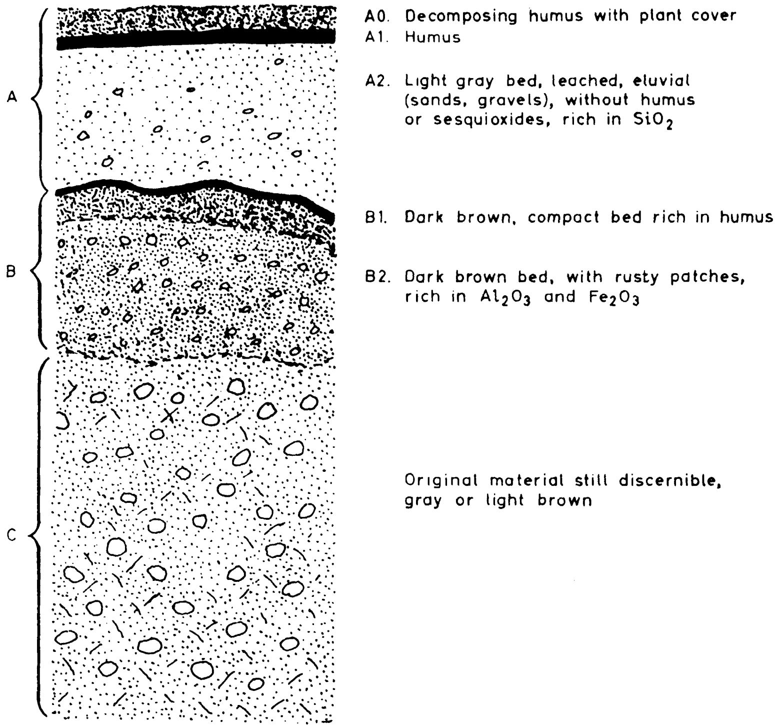


FIG. 74. SECTION THROUGH A PODSOLIZED SOIL: A MATURE SOIL WHICH HAS BEEN THOROUGHLY LEACHED AND HAS UNDERGONE CONSIDERABLE CHEMICAL ALTERATION (see p. 140)

the soil are carried. The pedalfers are enriched in *iron* and *aluminum* and are generally argillaceous in character (kaolinite and limonite). The soluble sodium, calcium and magnesium salts are carried away in solution. Grains of quartz persist unaltered. The top few inches of a pedalfer are leached of iron and most of the alkali and argillaceous material. These substances are found either in the zone of enrichment below, or are washed away by surface waters.

One of the varieties of pedalfers is *laterite*, which is formed under humid tropical climates (p. 142). A special type, known as the *gore*, is a kaolinized zone found in temperate climates at the level of the water table: it is similar to laterite in its mode of formation. The parent rock consists of andesitic tuffs. A typical example is that of Uzerche, France.

When the pedalfers are traced across the latitudinal zones their general characteristics remain constant. According to the climate, altitude, and parent rock they show some variation in detail. In the temperate zone, for example in Central Europe, they form the *brown forest soils* (Ramann, 1905) or *burozems*; in the Mediterranean zones, which are less humid and are

TABLE VI.—DISTRIBUTION OF SOILS ACCORDING TO VEGETATION AND CLIMATE

GLACIATED REGIONS TUNDRAS	*Frozen soils* (permafrost, tjale)		
	SATURATED SOILS		UNSATURATED SOILS
	STEPPES Pedocals and saline soils	PRAIRIES Hydromorphic soils and pedocals	FORESTS Pedalfers (perhaps podzolized)
TEMPERATE ZONES	Chestnut and brown steppe soils	peaty soils gley soils Chernozem	brown forest soils (= Burozem) Gore
SUBTROPICAL MEDITERRANEAN ZONE	Gray soils (= Cerozem) Saline soils soil-crusts	Chernozem —smolnitz Tirs	Chestnut forest soils ash soils of Japan yellow soils (= Jetlozem) red soils (= Crasnozem)
ARID TROPICAL ZONE	Brown and red soil-crusts hardpans		tropical ferruginous soils
HUMID TROPICAL ZONE	hardpans	Black soils (= allaphanites): regur, terra roja	ferrallites (= laterites)

warmer, the *marron soils* occur which are highly argillaceous at depth; and finally, in the hot forest zones, the amount of clay minerals increases and the decomposition of the original minerals becomes more complete. These are all *laterites in a broad sense*. There are also *yellow soils* (*jetlozem*) and *red soils* (*crasnozem*), e.g. from south of Batum on the east coast of the Black Sea; *tropical ferruginous soils*, rich in organic matter and kaolinite showing a concretionary horizon where iron oxide has segregated; and finally the *ferrallites* or *laterites* (*in the strict sense*) which are often very thick.

All the pedalfers are liable to undergo *podsolization*, which results in the migration of colloidal materials and sesquioxides through the soil under a forest cover. A *podsol* has a light-colored upper zone enriched in SiO_2, while the lower part, dark brown with rusty patches, is enriched in Al_2O_3

and Fe_2O_3. Podsolization is a chemical change which can be seen in numerous places such as northern Europe, for example.

The soils of prairies and steppes are nearly always *saturated* with soluble alkali and alkaline-earth salts due to poor drainage.

The *pedocals* are soils in which leaching is incomplete. They occur in dry regions such as those of prairies and bush. In the soils of hot countries the calcium and magnesium carbonates are leached from the upper part of the soils and carried toward the base. The intense evaporation of these regions, however, causes the deposition of the carbonates thus transported before they have reached the water table. In Mediterranean areas it appears that capillary action draws the water carrying the calcium salts to the surface, thus forming a calcareous *crust* (*caliche* = "carapace" of Pomel). Because of the low rainfall, little alteration of the argillaceous materials of these soils takes place, unless montmorillonite is present, as in the soils of the savanna (Erhart, 1956). Pedocals are of common occurrence in the western half of the United States. Saline soils fall into the same category (see pp. 322–325). In cold and subarctic countries some soils (peaty, gley soils) and the soils of tundras are also pedocals. There is a gradation between pedalfers and pedocals.

The pedocals are divided, in practice, into the prairie soils and the steppe soils. The *prairie soils*, saturated with water, are called *hydromorphs*. In cold and temperate climates, three types can be recognized: (1) *peaty* soils containing 15 to 50% of organic matter; (2) *gley* soils having, beneath the water table, a reducing, anaerobic horizon impoverished in organic matter, where bluish FeS and $Fe_3(PO_4)_2$ are formed, with prismatic jointing; (3) *chernozem*, black and rich in humus. Under a Mediterranean climate the latter are often developed on basic rocks and contain less humus (*smolnitz* of Bulgaria, *tirs* of Morocco). In humid tropical climates the hydromorphic soils are very rich in organic matter ($C/N > 17$). Among these can be included the *allophanites* (Guerasimov), black soils such as the *regurs* (black cotton soils) rich in silica and Fe_2O_3 which are developed on the Deccan basalts of India and the red soils (terra roja) of Brazil and China which are less rich in silica and Fe_2O_3 and which in some ways are similar to laterites. The *dry steppe soils* are *chestnut* or *brown* in temperate zones. In Mediterranean climates they are *gray* (*cerozem*) and form several varieties of *saline soils* and *soil crusts*. In the tropical arid zone *red* and *brown* soil crusts are dominant, and under a Sahelian (western Saharan) climate *ironpans* occur which are also found in humid tropical climates.

The *chernozem* or black soils are particularly rich in organic matter and have been classified either with the pedalfers or with the pedocals. They are produced in moist zones which are not forest-covered, and are found in Russia, in the United States and in Morocco (*tirs*). The prairie soils are rather similar.

In every respect, a soil is the product of a very long process. Many modern soils began their formation during the Tertiary. In view of the sensitivity of soils to climate, and the variation in climate during the Quaternary, it is to be expected that each soil is extremely complex, especially when the modification of the biotopes which have assisted in its formation are taken into account. Thus the black soils of Limagne (France) are peaty soils formed during the Allerød interglacial period and do not correspond to their present-day plant cover. Innumerable examples of this sort of inherited or "fossil" soil may be found, a factor which has enormously complicated the work of soil scientists.

ELUVIAL SOILS WITHOUT MECHANICAL EROSION

Ferrallitization = Laterization[1]

One of the most important phenomena in the weathering of rocks under tropical and equatorial climates is that of laterization, and probably, also, that of bauxitization of rocks. Laterization occurs in hot, humid climates under the influence of abundant rainfall. On high, cold plateaus of the inter-tropical zones, above 6,500 feet, it does not take place and weathering is reduced to disintegration. There is no active laterization today at the edge of the sea, nor in zones constantly saturated with water. In a tropical country where the forest has been destroyed, the red lateritic clay formed beneath plant cover is subjected to intense insolation, and hardens rapidly as a limonitic layer forms.

Laterite (from *later* = brick; Buchanan, 1807) is a mixture of *hydrated oxides* of iron, aluminum, manganese and titanium, the proportions of which vary with the nature of the rock undergoing alteration. In hot, humid climates these hydrated oxides are derived solely from crystalline rocks. The bright red tint of lateritic soils is due to the oxidation of iron during their development.

The hydrated oxides of aluminum in modern laterites are hydrargillite or gibbsite and other alumina gels. The hydrated iron oxides present are crystalline goethite and colloidal stilpno-siderite; by desiccation they are

[1] Among English-speaking geologists the terms ferrallitization and ferrallite are rarely used, although the terms are familiar to pedologists. *Ferrallitization* is the process leading to the formation of a *ferrallite* which is defined as a *soil material* with a high ratio of sesquioxides of iron and aluminum to silica (2.0). The term *laterite* (formed by *laterization*) has been widely, and somewhat loosely, applied to *tropical red soils* of widely differing characteristics and origins, in which the sesquioxide-silica ratio may be greater or less than 2. Pendleton (1936), quoted by Robinson (1949), would restrict the use of the term to *soil profiles* characterized by the presence of *concretionary material* or *crusts*.

The complex and controversial problem of the usage of the terms ferrallite and laterite has been discussed at length by Robinson (1949).

Although the authors have here equated the terms ferrallitization and laterization, they later (see Hardpans, p. 146, and Fossil Bauxites and Laterites, p. 147) use the terms ferrallite and laterite in the narrow and the broad senses, respectively, noted here. In translation, the authors' terminology has been strictly adhered to.—Translator.

converted into hematite. The hydrated oxides of manganese are manganite or psilomelane. The hydrated oxides of titanium, which are abundant in the laterites of gabbros and of basalt, are colloidal, but possibly take the form of doelterite (Lacroix, 1934) which has the composition, $TiO_2.2H_2O$ or $TiO_2.H_2O$. The silicates of aluminum are kaolinite and halloysite.

All intermediate stages between the unaltered silicate rock and the laterite can be recognized, but the boundary between the rock and the laterite is very abrupt. A. Lacroix (1934) has called that part of the laterite which is in contact with the rock the "zone of departure" and that part which is at the surface the "zone of concretion".

The zone of departure is characterized by the migration of part of the components of the original rock, the alkalies in the form of bicarbonate of sodium, potassium, calcium and manganese, and the silica as a hydrated colloid; this is the *migratory phase* of H. Erhart (1955).

In the formation of laterites, the energy necessary for the liberation of the alumina can be provided directly or indirectly by biochemical reactions which take place in hot climates (T. Holland). It is known that the ferro-bacteria can precipitate hydrated iron oxide, accompanied by alumina (Ehrenberg, Vinogradsky) and that diatoms in the presence of bacteria can decompose kaolin and liberate hydrated alumina (Vernadsky). However, according to Erhart (1926) the part played by living organisms in laterization is not biochemical; it is the dense mantle of vegetation, together with a thick humic soil formed under tropical forests, which provide the conditions necessary for laterization. The soil of these zones, which is fixed and preserved from erosion, remains permeable, and allows the circulation of water containing humic or other substances produced by the decomposition of the vegetation. In this way, some kaolins (Madagascar) are produced and since the humus is acid, soils rich in silica and hydrated alumina are formed.

According to A. Lacroix (1934) the variation in level of the water table due to the alternation of very wet and drier periods is of great importance. The first rains carry the soluble salts and the humic substances deep into the soil, where they gradually attack the unaltered rock beneath. After complete saturation of the soil, the excess water carries away with it the migratory products. During the dry season, the solutions take up ferrous bicarbonate, and rise to the surface by capillary action. There ferrous iron is oxidized, and finally hydrated ferric oxide is deposited. A similar process permits the deposition of alumina in the zone of concretion. Laterization is thus a process of "phreatic corrosion" (J. Bourcart, 1947).

The formation of an ironpan (p. 146) impedes the growth of vegetation and finally causes the cessation of laterization. The thickness of a laterite is generally about 10 feet but sometimes exceeds 100 feet.

In the upper part of the peridotites of New Caledonia (Chételat, 1947) laterization is also responsible for the concentration of nickel minerals in the form of hydrated silicates, together with hydrated silicates of magnesium.

It is of particular interest to note the metasomatic hydrothermal nature of the serpentines which are formed in New Caledonia by the alteration *per ascensum* of graywackes and peridotites, and to compare this with the process of lateritic alteration. When attacked solely by hydrolysis, without oxidation, these rocks are not readily laterized. Since the formation of the

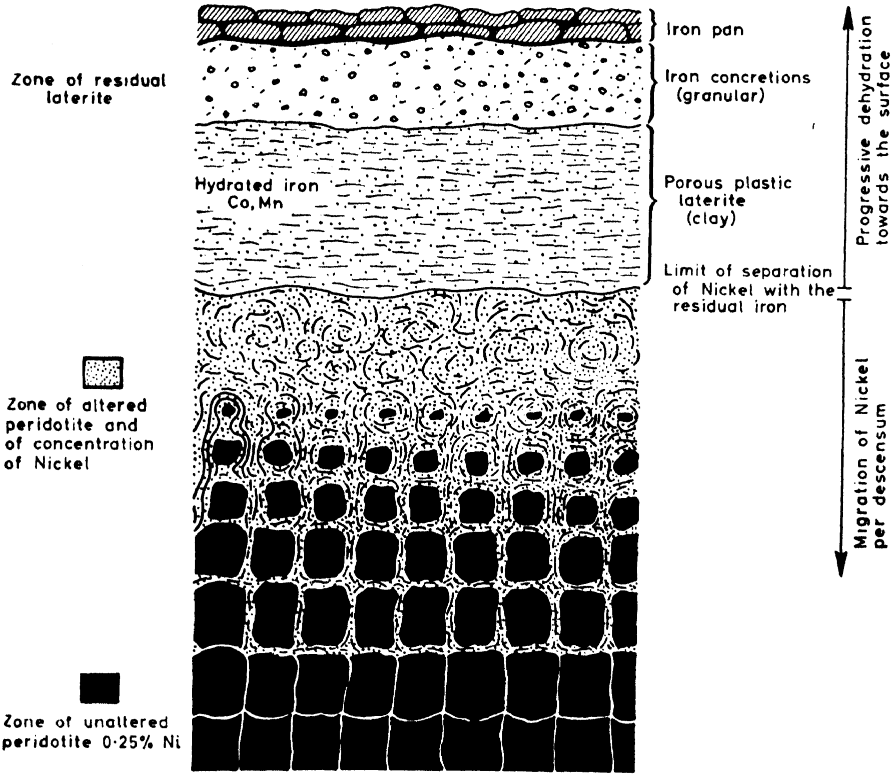


FIG. 75. SECTION THROUGH A COMPLEX LATERITE IN NEW CALEDONIA (after E. De Chételat, 1947)

This is a normal mature lateritic profile, modified by the presence of nickel in the ultrabasic country rock.

serpentine takes place with an increase in volume, the rock is not permeable, unlike the peridotite which is readily altered by the solutions circulating through joint systems. Nickel minerals are practically absent from serpentine, which does not therefore lend itself to the concentration of nickel. As has been emphasized by E. de Chételat (1947) the superficial alteration of rocks if favored by plateaus or gentle slopes, that is, where there is comparatively little erosion. Where slopes are steep, as in deeply dissected areas, the run-off of water brings about mechanical erosion and the alteration products have little chance to accumulate. Thus laterites are found more often on old platforms than in orogenic zones.

In the zone of concretion, where decomposition is virtually complete, there occurs the *residual phase* (Erhart, 1955) consisting of iron and aluminum oxides and kaolinite. The concentration of iron is made apparent by the red color of the laterite, which may be capped by an ironpan layer. There is also enrichment (particularly in bauxites) in alumina, while some chromium and titanium may be present.

The texture of the parent rock can be preserved intact in the course of alteration (G. Millot and M. Boniface, 1955). The conversion of a dunite to a laterite, and of a nepheline syenite to a bauxite occurs at constant volume. The final process is the compaction of the lateritic zone (Chételat, 1947).

Several varieties of laterite are known. The most important is *bauxite* (Dufrénoy, 1837; Sainte-Claire-Deville, 1861), which was formed from alumina gels in place of crystalline gibbsite. The composition of the original rock has considerable influence on the composition of the final laterite.

The formation of soils on basic volcanic rocks has been studied by P. Ségalen (1957) in the intertropical zone of Madagascar. It was found that vegetation became established on a recent lava flow in the following order: bacteria, fungi and algae, grasses, bracken, and then xerophytes which assisted in the breakup of the rocks. The transport of the plants to the lava flow was effected by the wind. They were kept moist by rainwater. Plant roots attacked the rocks and liberated bases (lime and magnesia, which formed alkaline solutions), silica, alumina and iron. The first clay formed is a montmorillonite. The initial soils, which are dark and thin, are rich in organic matter, exchangeable bases, and phosphoric acid. They contain many traces of unaltered rocks. They are, in short, regoliths.

Following this stage of youth, the soil undergoes a prolonged period of elaboration and deepening.

1. With a rainfall of 28 to 36 inches, the bases are not completely eliminated and nodules and filaments of calcite are formed. Montmorillonite also remains.

2. With a rainfall of 31 to 59 inches and a temperature of about 80° F., deciduous forests develop. The soil contains kaolinite, anauxite,[1] goethite, lime, magnesia and potash.

3. Above 59 inches of rain and a temperature between 74° and 77° F. the vegetation is modified and forms a true tropical forest. The bases are leached and carried away and the soil becomes acid, leading to progressive destruction of montmorillonite. The stable clays in these conditions are kaolinite and gibbsite. A small amount of iron is eliminated from the lower part of the soil and deposited in the upper part, to which it gives a red or yellow color. The lime forms stable complexes with the humus, which are

[1] Variety of kaolinite rich in silica—$Al_2Si_3O_7(OH)$.

not readily carried away by water. Phosphoric acid forms insoluble salts with the iron and aluminum.

4. When the rainfall exceeds 80 inches and the temperature is 61–68° F. the soil is similar to that above (in 3), but organic matter accumulates.

Hardpan Formation[1]

The *hardpans*, in the limited sense which is assigned to them at the present day (Maignien, 1958), are hard layers, formed at depth from the same materials as laterites (mainly ferro-magnesian minerals) and even from ferrallites[2] already present. They occur, however, in climates and under biological conditions less restricted than those necessary for the formation of laterites. In Africa, at the present time, they are found from the western Sahara (arid) down to Guinea (humid tropical). In the latter country, the hardpan occurs chiefly in the deforested areas which it tends to render sterile (this is the *bovalization* of Aubréville, 1947). Thus the development of hardpans *increases from the forests to the savannas* and towards the areas cultivated by man. The hardpans have a mineral composition similar to that of laterites: *gibbsite*, boehmite, diaspore, *goethite*, magnetite, the spinel maghemite, *hematite*, residual ilmenite, manganese dioxide, *quartz* (up to 50%), *kaolinite*, halloysite and illite (on young hardpans or those formed on alteration products). They often contain rock debris or fragments of older hardpans.

Certain hardpans, composed of sesquioxides of aluminum and kaolinite, are formed perhaps, by processes similar to those leading to the occurrence of laterites. It must be emphasized however: (1) that the alumina only occurs in the hardpans formed from ferrallites; (2) that they also carry soil crusts (see pp. 156–157) and a zone of phreatic cementation (p. 81).

The water of seasonal rains infiltrates into them, as well as running off the surface. It penetrates from the top to the bottom (*per descensum*) as in all soils, but also moves obliquely and laterally and trickles to the bottom of slopes. During the dry season, capillary action and evaporation brings about the formation of crusts at various levels in the hardpan. These thicken downwards by accumulation of the sesquioxides at the top of the water table, or at some other level (sea level, for example). It follows that the level of the hardpan can reach several hundred feet below the surface of the soil.

Theoretically, the levels of mineral accumulation in a hardpan occur in the following order from top to bottom: (1) aluminous, (2) argillaceous,

[1] "Hardpan formation" is used to convey the sense of the French noun "cuirassement", which literally means armouring (of ship, etc.) or armour-plating.

"Hardpan" is used to translate "cuirasse", literally a cuirass or breastplate. A "cuirass ferrugineuse" is ferruginous hardpan, or ironpan = the ferruginous "duricrust" of Australia (Woolnough, 1927).—Translator.

[2] See footnote, p. 142.—Translator.

(3) ferruginous, (4) manganiferous. This succession is, however, modified by the circulation of water which may carry material from neighboring areas.

The role of living organisms in the formation of hardpans is negligible. Nevertheless, the decomposition of organic matter acidifies the soil and decreases the solubility of silica and the hydrolysis of aluminum, iron and manganese. Furthermore, microbic activity, particularly in the wet season, can cause a diminution in the amount of oxygen, and, locally, can bring about anaerobic, reducing conditions. Bacteria then take the oxygen which they require from iron oxides, which they reduce. They also attack silica and calcium carbonate. When the amount of organic matter diminishes and finally disappears, microbic activity is also reduced. This favors an increase in oxidation and the formation of sesquioxides. It is apparent, therefore, that deforestation (by reduction of organic matter) will either cause or complete the formation of hardpans.

Hardpans on subarid soils have a nodular or alveolar structure, with ferruginous sandstones, and sometimes with ferruginous (limonitic) pisolites, e.g. in western Australia.

The exposure of hardpans by erosion and the removal of loose overlying material brings about a secondary enrichment in iron, and also, possibly, manganese. This forms a shiny patina known as "desert varnish". Fossil hardpans which outcrop at the surface may become broken up and incorporated into new hardpans.

Among the hardpans, the *lateritoids* (Fermor) are superficial concentrations of limonite and perhaps manganese, associated with quartzitic rocks (quartzites, quartz conglomerates, micaceous shales). Of this type are the ferruginous conglomerates of central Morocco. (H. Termier, 1936, p. 945.)

Fossil Bauxites and Laterites

In the fossil state it is difficult to distinguish ferrallites[1] from hardpans.

In view of the geographical extent and thickness of these deposits it is not surprising that they are found in the fossil state. The best known are the *bauxites* which are white, yellow, brown or red.

Their composition is close to that of the laterites, but contains less iron. They consist chiefly of colloidal hydrates of alumina (alumina gels) associated with crystalline substances including kaolinite, gibbsite and diaspore. To these may be added hydrates of iron, silica in the combined state, titanium and phosphoric acid.

There are two types of occurrence of bauxite. The first is that in which all stages between the bauxite and its substratum can be observed. It can be seen, for example, when the substratum is a nepheline syenite in which the proportion of alumina reaches 20%. Good examples are known in the central region of Arkansas (nepheline syenite underlies the Wilcox group, which is equivalent to the Ypresian), in Brazil in the district of Poços de Caldas

[1] See footnote, p 142.—Translator.

(fayaite and tinguaite are cut by viens of caldasite) and in Guinea in the islands of Los (nepheline syenites).

The second case is that where the intermediate products between the bauxite and the parent rock are poorly developed or absent. This happens very often when the substratum is a karst limestone: for example, in southern France, in Istria, in Yugoslavia and in central Kazakstan. There are two theories to explain this type of occurrence. The classical interpretation is that the solution of the limestone has produced a residual clay similar to a *terra rossa*, which has later been altered to a bauxite by laterization. Such a process does not seem impossible, since such concentration is well known in ore deposits. The limestone contains 0·15 to 0·85% Al_2O_3 and this increases to 55 to 56% Al_2O_3 in red bauxite. This interpretation has been questioned by H. Erhart (1956) who envisages the transport of laterites during a period of "rhexistasy" (see p. 153). The karstic form of the limestone has thus acted as a trap for the transported sediment. According to E. Roch (1956) the bauxites of the Durance isthmus and of Bas-Languedoc (southern France) results from the accumulation of dust carried by the wind, similar to loess. The ochres of the region of Apt and of Roussillon seem to have the same origin as the bauxites, but there the dust fell in the sea. This hypothesis takes into account the fine grain of bauxites, the absence of stratification and their lack of flints and rolled pebbles. These characters exclude the possibility of transport by running water, although they are not opposed to the formation of the bauxite *in situ* without transportation. There is also extreme variation in the thickness of the bauxite layers. In the case of the bauxites of the south of France, it is possible that the laterized crystalline rocks which were the source material were the Massif Central, the Maures, and the Esterel. Elsewhere, as in Yugoslavia, the crystalline massifs are remote from the areas of deposition and it must be assumed that transport over a very great distance occurred. In the present state of knowledge it is difficult to choose between these hypotheses, since bauxites occur very frequently on karsts and do not extend beyond them.

The principal deposits of bauxites are described below in chronological order.

PRE-CARBONIFEROUS.—Forests are unknown before the upper Devonian and it seems probable that the major part of the continents during Early Paleozoic times were devoid of a true plant cover, if lower forms of life are ignored. Bacteria, lichens and mosses do not cause laterization of rocks. In the absence of a plant cover it seems likely that hardpan formation would have been common. This would have caused the concentration of the sesquioxides of iron and manganese, and to a lesser degree, of alumina, and accounts for the red color of pre-Carboniferous continental sediments, such as the Old Red Sandstone of the Lower and Middle Devonian. It is probably also responsible for the formation of certain iron minerals. Devonian bauxites have been recorded in the Urals (Bouchinsky, 1958).

CARBONIFEROUS.—Carboniferous bauxites are known in Germany, Bohemia, Great Britain, Ireland, Russia, China and North America. N. Strakhov, E. Zalmanson and M. Glagoleva (1959) have listed a large number of deposits associated with, or peripheral to, carbonaceous terrains in the Lower Visean of the Russian platform and to the south of Timan. In the Middle and Upper Visean, a deposit also exists in Tadjikistan. In the northern part of Ayrshire (Scotland) a lacustrine bauxitic clay is believed to be a reworked laterite originally developed on basaltic lavas. These lavas are associated with the Millstone Grit and the clays are penecontemporaneous; they are of Early Pennsylvanian age (Wilson, 1922).

In the United States, the "fire-clays" of the Mercer Group are included in the Pottsville Series on the Allegheny plateau (Bolger and Weitz, 1952). These are believed to be the products of lateritic alteration subsequently transported from their region of formation in low marshy plains. The diaspore clays of Missouri, which are of Desmoinesian age (and thus slightly younger than those of the Mercer Group) are poor in alkalies, alkaline-earths and iron. Gibbsite is absent, but kaolinite, diaspore, boehmite and a chloritic mineral are present. They appear to have been formed on an Ordovician dolomite which was undergoing karstic evolution.

JURASSIC.—Laterite formation of a "biostasic" type (see pp. 153–155) seems to have extended over part of the land surface of the Lias and the Dogger. One of the best examples has been observed in the Gobi Desert in Mongolia, where the Efremov expedition (1945–1949, publ. 1954) measured a thickness of nearly 200 feet of Jurassic laterites. This was formed on a forest-covered plateau, dotted with lakes and traversed by rivers.

On the eroded Triassic limestones of the Vanoise (Savoy Alps), Ellenberger (1955) described an horizon above the water table which he has attributed to laterization. A bed of Dogger age consists of several feet of carbonaceous shale and contains *Mytilus*. It rests on 3 to 10 feet of a green chloritoid rock, of density 3·2, which is probably a hyperaluminous clay derived from a ferruginous kaolinitic clay in a reducing environment and then metamorphosed. The laterization continued for a long time on low-lying lands which emerged during the Dogger. In western Vanoise, this process apparently continued into the Malm, the base of which is a red bauxitic clay containing chloritoid, muscovite or sericite, hematite and diaspore. Possibly the silica of the Argovian radiolarites (Upper Oxfordian, according to Arkell) represents the migratory phase of these laterites.

In northeast Spain, in Catalonia, thick limestones and dolomites were deposited during the Keuper, Lias and Dogger and even during the Malm. A stratigraphic break corresponding to emergence occurred at the end of the Jurassic and the beginning of the Cretaceous, and the limestones were subjected to karstic weathering. Bauxites were deposited in pockets in the karst (Closas).

At about the same time, between the Kimmeridgian and the Urgonian, bauxites were formed at Ariège on the other side of the Pyrenees.

CRETACEOUS.—During this period a broad zone of bauxite was formed parallel to the Mediterranean in southern Europe—southern France, northern Italy, Dalmatia, Yugoslavia, Hungary, Greece.

The bauxites in the south of France (between the Var and the Herault rivers) occur on karsts formed on limestone and dolomite the ages of which vary from Early Lias to Urgonian (figs. 76–77). These rocks formed part of the emergent Durance isthmus which linked the Massif Central with those of Maures and Esterel. According to the hypothesis of E. Roch, which has already been mentioned, the bauxites were derived by wind transport from the crystalline massifs. The bauxites are overlain either by transgressive marine rocks of Cenomanian age or by lacustrine beds of the Campanian (Upper Senonian). Taking into account the time of emergence and the time required for the formation of the karst on the Urgonian limestone, the transport of the laterites would seem to have occurred between the Aptian and the middle Senonian, and the preliminary biostasy on the crystalline massifs could have been Aptian.

In Istria, in Montenegro (Nikšic), at the mouth of the Cattaro river, and on the other coast of the Adriatic at Monte Gorgano, Jurassic limestones are cut by Neocomian karsts. The depth of this karst is not more than 150 feet and locally it contains bauxite deposits, which are older than the middle Cenomanian transgression.

Istria, the Dalmatian islands, Dalmatia, Bosnia, Greece and in Italy, Apulia and Monte Gorgano were all emergent in the Senonian and a new karst was cut to about 300 feet: bauxites were deposited before the Late Senonian transgression (d'Ambrosi, 1954).

In central Kazakstan, Cretaceous bauxites occur on Devonian limestones.

In North America bauxite deposits seem to have been formed at the expense of a Lower Cretaceous (Tuscaloosa) clay, underlying marine Eocene rocks (Claiborne). The belt extends from Alabama to Georgia.

Bauxites are associated with the erosion surface which cuts the nephelinic rocks of Poços de Caldas in Brazil, and which is of Early Cretaceous age (J. J. R. Branco, 1956).

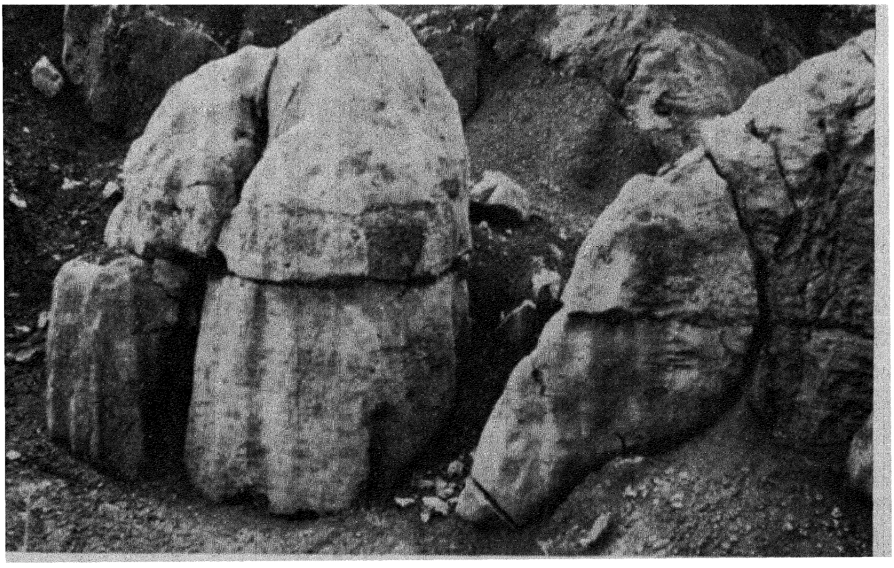
TERTIARY.—Rocks of Tertiary age allow the observation of laterites still in place and permits study of the succession resulting from "biorhexistasy" (pp. 153–155).

EOCENE.—Following the Pyrennean movements, the Durancian isthmus was peneplaned. The surface of the peneplain, still visible in parts of the Rhone valley, is covered by mottled refractory clays, containing alumina, silica and iron and, near their top, ferruginous concretions: this is a reworked laterite. The forest of Mount Ventoux occurs on fossil lateritic soils which nullify the effects of the karstic limestone beneath (P. George, 1935).

In North America traces of a similar climate are found. In Arkansas, bauxites are formed on nepheline syenites (Pulaski, Saline) underlying the Wilcox Group (equivalent to the Ypresian).

ᶦIGS. 76 and 77. AN OLD KARST, EXPOSED IN BAUXITE WORKINGS IN LOWER PROVENCE, FRANCE (Photograph: M. Casalis)

The bauxite was probably derived from the weathering of the limestone during the formaion of the karst, or it may have been brought by wind or water and trapped by the uneven urface of the karst (p. 148).



In Georgia and in Alabama (south of the Appalachians), the Knox Dolomite (Cambrian) which is cut by a karst has, at the level of a presumed Eocene peneplain, pockets of bauxite accompanied by clays and limonite in the middle of a thick residual clay.

Laterites are widespread across Africa and the old massifs of Europe (Massif Central, Brittany).

MIOCENE.—During this epoch laterites were more widely distributed than at any other time.

The whole region of the Sudan, which today is subdesert, is covered by a hardpan of lateritic origin, indicating that it was covered by forests at a time, not long ago, when the area was irrigated by a network of now dry rivers (see p. 112).

At about the same time, in the western Congo and in Uganda, the Upper Kalahari Formation was deposited on a mid-Tertiary, pre-Miocene surface. This deposit consists of eolian sands, and in places, laterites (Cahen and Lepersonne, 1948).

In the coastal region of Senegal there is, at the base of the Quaternary, an horizon which consists of 1% of siliceous pebbles derived from the Precambrian and Cambrian, and 99% of concretionary ironstone (Tricart, 1955). This represents the destruction during an arid period of lateritic soils formed earlier. The glacio-eustatic retreat of the seas at the time of the first glaciation, by lowering the water table and the base level, could have been the cause of this change in regime.

The siliceous crusts of Angola may have been formed by the migratory phase of the African laterization (Gilluly, 1951).

At the end of the Miocene in Australia, almost the whole of the country had been peneplaned and then covered by a tropical forest, beneath which laterites were formed. Definite remains of laterites are known from Queensland (Alice Plateau) to the gold-bearing region of Yilgarn. It is possible that the siliceous crust of the Lake Eyre basin, of pre-Pliocene age, was formed during the migratory phase of this laterization. In the Northern Territory laterites are localized on outcrops of argillaceous sediments and after erosion have developed a siliceous cement, known as "billy". The low silica content of Australian sediments since Cambrian times seems to be responsible for the special character of these laterites (Opik, 1956). During the Pleistocene, and at the present time, only podsols have formed in the coastal regions where the humidity is high. The basic rocks in semiarid areas have given rise to chernozems and pedocals, and in the higher regions, red loams, quite different to the Miocene laterites. Bauxites of Miocene age occur in Tasmania.

Fossil lateritic soils are formed in many parts of the East Indies and in Malaya. In the latter, granites and the volcanic rocks of the Pahane Series are covered with clay and a fossil lateritic soil. The islands of Riouw, which extend beyond Singapore, are also covered by a thick lateritic crust, which is exploited as a bauxite on Bintan (the neighboring island to Singapore). Laterites are also present in the Sunda Isles (southwest of Borneo, for example [van Bemmelen, 1940]). However, on the recent volcanic rocks of the northwest coast of Sumatra, H. Erhart (1954) has observed illitic soils between sea level and a height of 4,000 feet.

QUATERNARY.—The ferrallites of Guiana and Surinam were mentioned earlier (p. 22). These laterites, some 65 feet thick, rest on a thick clay overlying crystalline Precambrian rocks.

In Cuba the iron minerals of Mayeri are attributed to laterites. In Haiti the bauxites are derived from red ferrallitic soils associated with Eocene and Oligocene limestones which are often detrital. These bauxites, which are worked commercially, never seem to be associated with chalky limestones containing flint. Butterlin (1958) believes that these soils are formed by the pedogenetic alteration of the limestones.

In Indochina and in India the laterites which today cover vast desert areas, can only be relatively recent in origin. They are consolidated rocks, which, for a long time, have been used as building stone.

It is certain that in New Caledonia the extent of laterization is less at the present time (if it is still taking place) than it was at the end of the Tertiary.

In Africa, ferrallitization continued during the Quaternary (see p. 21, Table V). Examples can also be quoted in Guinea, Dakar and in Madagascar (see p. 145).

Laterites also occur on basaltic lavas in Hawaii.

The Theory of Biorhexistasy

In order to explain the formation of laterites and to account for their distribution in the sea and on the land throughout geological time, H. Erhart (1955) has put forward a theory based on the variations in the plant cover of the continents.

At the present time soils are formed (under equatorial forests) which are composed of "residual minerals", such as quartz (if the parent rock contains it), kaolinite, hydroxides of iron and aluminum. They are, however, poor in alkalies (Na, K), alkaline-earths (Ca), magnesium and silica from original silicates, which constitute a "soluble migratory phase". When laterization occurs, that is, when an area is covered by a thick plant cover, the minerals of the "migratory phase" are eliminated and carried towards the sea. Thus during geological time, those continents with low relief which were covered by thick forests have been able to provide the elements of limestones, dolomites and hydrated silica rocks. These were deposited in the sea as long as climatic and vegetation conditions remained constant. According to Erhart "these sediments are thus indices of great stability of the earth's crust".

It is of particular interest to apply the theory of biorhexistasy to geological problems and to determine the significance of the periods of biostasy and of rhexistasy. There is no great difficulty concerning the period of biostasy, since it coincides with phases of planation and stability of the crust. Mountains, even if covered by forests, do not give rise to laterites.

It is necessary to point out that *these phases cannot be universal*, but follow the vicissitudes of geomorphic sculpturing, which is itself the result of orogenesis and epeirogenesis. They occur on the great shield areas and on eroded mountain chains. Temporally, they are produced in periods of tectonic calm. Moreover they raise the question of climate. In warm, moist periods (or monsoon climates) the vegetation responsible for the formation of biostasic soils is widespread. However, during dry periods, whether they are warm or cold (*playa periods*, see p. 28), forest vegetation is restricted. Biostasic phases are therefore limited in time and space. For example, Fair and King (1954) have distinguished six cycles of planation on the African Shield between the Congo and the Cape of Good Hope which date from the beginning of the Mesozoic to the present day. The climate, however, often tended to be dry and prevented the occurrence of biostasic phases in each of the cycles. In actual fact, only two such phases are known (Eocene, Miocene).

The problem of rhexistasic phases is more difficult. In the first place, rhexistasy implies the disruption of a state of equilibrium, that is, the biostasic equilibrium. The phases envisaged are thus only possible on surfaces which have already undergone the effects of biostasy, and these are, in general, not very widespread.

The results of rhexistasy are identical with those of mechanical erosion which affects all rocks subject to the action of rain, running water, ice, glaciers and the wind. There is a mechanical breakup of the rocks which reduces them to fragments varying in size from blocks to colloidal particles, and even soluble materials are produced.

Mechanical erosion is most intense in regions of high relief, but is comparatively limited on planed shield areas protected by a mantle of sand (such as that present in the Sahara), or by a "carapace" of ice during glacial periods (as on the Antarctic, Canadian and Baltic shields). These shields, which represent two types of desert show neither the characteristics of biostasy, nor those of mechanical erosion. They are stable areas as are most of the old platforms which form the greater part of the continents.

If the zones of high relief and the hot and cold deserts which are unsuited to the development of biostasic conditions are excluded, there remain only the humid, planed regions supporting a protective plant cover to which the theory of biorhexistasy can be applied.

However, apart from the above restrictions, two other uncertainties upset the applications of this theory.

1. BIOSTASY.—Guiana is a plateau covered with tropico-equatorial forest. The soil of this forest is very often laterized. The climate is hot and

humid, probably too hot (78 to 81° F.) and too humid for the formation of ferrallites at the present day (p. 21). This forest seems, by its presence, to have stabilized the soil on which it lives. B. Choubert (1957) has shown that abundant rain in equatorial climates (100 to 150 inches per year) attacks laterites or bauxites which are porous and fissured to a depth of 60 feet. The penetration of the rain is only stopped by an underlying kaolonitic clay which may be 150 feet thick. The water circulates in the lateritic layer and causes a karstic type of erosion (p. 302) with dolines and caves, by the solution of oxides of iron and hydroxides of alumina. The laterite becomes porous, breaks up and is transported by the water. *This destruction of the laterite* PRECEDES *the destruction of the forest cover.*

2. RHEXISTASY.—It has already been shown that hardpan formation, for example in Guinea, follows deforestation. This hardpan is totally impermeable and protects the soil beneath it. It is of similar composition to that of ferrallites and it is more *stable* than that formed under arid or semiarid conditions, or the soils of Guiana formed under a forest cover.

It does not therefore seem possible to use the theory of biorhexistasy as a general explanation of sedimentological phenomena, although it may have local applications. In the next section, it will be shown that pedological processes can, in fact, furnish valuable information concerning such events.

Sedimentation in the Interior and at the Edges of Shield Areas

It should be noted that the observations on which Erhart based his theory were limited to well-defined basins, such as that of Chad. Basins of this type are, at the present day, surrounded by crystalline plateaus which are rarely covered by virgin forest under which laterization can occur. The successive sedimentary formations deposited in such basins are as follows: diatomites, then coarse sands and fine sands, silts and opaline kaolinitic clays (with hydroxides of aluminum and iron), sodium carbonate (natron), clays with alkali carbonates, and, finally, alluvial or reworked laterites. There is thus evidence of the transportation and deposition of the migratory phase, followed, after the dispersal of the plant cover, by the residual phase.

A succession such as this is well established in basins of similar type which have undergone an analogous geological history. Thus, the basin of Lake Eyre in Australia was surrounded during the Miocene by a peneplain on which laterization was taking place. The basin itself is covered by a siliceous crust of pre-Pliocene age, which appears to have represented the migratory phase of the pedogenesis of the laterite.

The theory of biorhexistasy can be extended to certain basins of argillaceous sedimentation (Millot, Radier and Bonifas, 1957). As has already been shown, the residual phase consists of kaolinite and hydroxides of aluminum associated with oxides of iron; other clays are deposited under different conditions. The association of attapulgite, montmorillonite, limestone and siliceous concretions (that is, an assemblage rich in silica, lime and magnesia, but poor in alumina and iron), suggests the migratory phase of laterite development. This association is found not only in lacustrine basins, but it also occurred in Eocene marine basins on the edge of the African shield when times favored the formation of laterites. These basins commonly contain phosphate horizons, as for example, at Gao. These formations are followed by iron ores containing ferruginous oolites (Tessier, 1954) together with kaolin clays and sands, which correspond well to a rhexistasic stage.

This application of the theory demonstrates clearly the continental character of sedimentation in marine coastal basins.

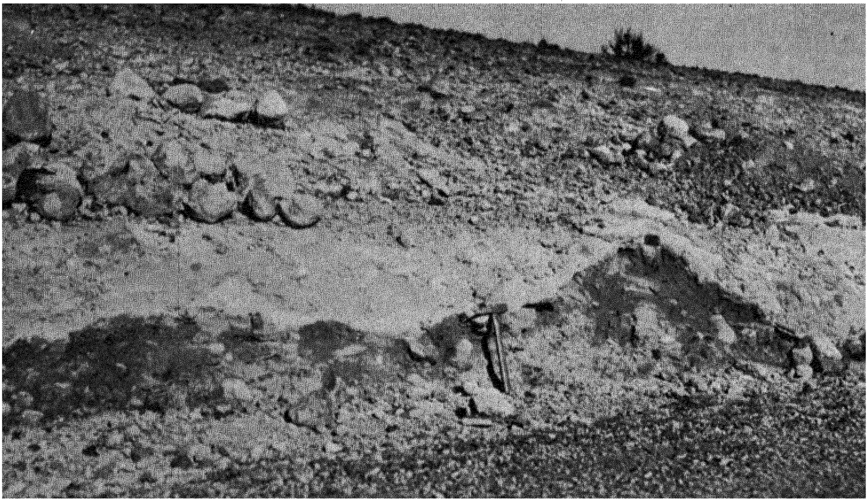


Fig. 78. The "Carapace" of Pomel. Between Oujda and Berguent, Eastern Morocco (Photograph: H. Termier)

Soil Crusts (fig. 78).—It has been shown that soil crusts are a variety of pedocals (p. 141). J. H. Durand (1956) has applied the theory of biorhexistasy to soils other than laterites. According to him, the banded crusts of the North African soils were formed by precipitation within the zone of partial saturation by water. The climate was tropical and the alternation of periods of abundant rain and short dry periods encouraged a forest cover and prevented eolian erosion. The calcium carbonate of the crusts was derived from the underlying limestones. However, the author does not explain clearly the mechanism of the solution and deposition of the calcium carbonate. He seems to invoke solution at depth by biostasy and precipitation at the beginning of rhexistasy (erosion) by rainwash in a dry climate. This hypothesis, if it is correct, supposes a rapid, even seasonal, succession of biostasic and rhexistasic phases. Deposition of these banded crusts is believed to have occurred at the end of the Villafranchian. According to M. Gigout and G. Choubert, however, formation of the main calcareous crusts of Morocco began in the Tensiftian (pluvial period corresponding to the Riss) and continued to the Ouljian (Late Tyrrhenian) (see Table IV, p. 19).

But crusts were formed in nearly all epochs of the Quaternary. Saline and gypsum crusts are related to the level of the water table and to evaporation, whereas calcareous crusts require the action of humic acids under a plant cover (forests, prairies). Powdery limestones are formed in humid and hot climates in lakes draining dense forests where the soil is a "terra rossa". In North Africa, these are characteristic of the uppermost Pliocene and the lower Villafranchian. The red loams with powdery calcareous nodules correspond to the beginning of the rhexistasic phase. Powdery gypsum crusts are formed in a similar way, but in evaporation basins of the shott type (p. 118).

Chilean nitrates are associated with soil crusts in Chile at latitudes between 19° and 26° S. (Tarapacas and Antofagasta). They are localized in dry depressions of the plateau which forms the Cordillera near the coast and inland from it. The crusts (=caliches) are 7 to 10 feet thick and are often brecciated. The cement is formed of very soluble salts which are preserved by the very dry climate of the region. The salts include $NaNO_3$, KNO_3, $NaCl$, $KClO_4$, K_2SO_4, $CaSO_4$, borates and iodates. It is of interest to note the presence of elements generally fixed by organic matter, such as the nitrogen of the nitrates. However, Lindgren (1933) believes that the salts are derived from tuffs and lavas of Jurassic and Cretaceous age which occur nearby. Other authors have suggested that they result either from the leaching of guano, or come from plant substances occurring on the slopes of the Cordillera. Finally, Goldschmidt (1954, pp. 449–453) has invoked the atmospheric oxidation of bituminous sediments of marine origin as an explanation.

The soils associated with soluble rocks are discussed later (see pp. 322–325).

7

Continental Sedimentation

Continental sediments are essentially the products of erosion and consist of *detrital* materials characterized by their size, which ranges from large blocks to fine particles of colloidal dimensions. Associated with these are dissolved substances (p. 81). Most of these sediments result from erosion by the agents of the hydrosphere (water, glaciers) and are more or less reworked by the wind. However, it is necessary to note the important category of materials formed by *chemical alteration* (e.g. soils).

It is probable that most marine sediments are formed more or less directly from components supplied by detrital erosion and chemical alteration of the land surfaces. These will be discussed later.

Torrential and Fluviatile Sediments

Most continental sediments are formed by erosion, and the most important of them are formed by streams. The action of running water is essentially mechanical and the sediments produced show a wide range of particle size. The fragments are more or less rounded according to the nature of the rock and the distance they have been transported. Decreasing size and prolonged transport lead to a series of decreasingly large materials: cobbles, pebbles, sands, silts. The size and grading of the load is rapidly established and depends upon the *competence* of the stream. That is, the ability to carry rock fragments varies with the vigor of the stream. The soluble components go into solution very rapidly, thus increasing the salinity of the water, which then has a chemical role in addition to its mechanical one.

Two other types of deposit are grouped with torrential and fluviatile sediments. One is mainly coarse-grained and is formed by the erosion of slopes and by deposition at their foot: this is the *colluvium.* The other, generally finer grained and better sorted, is carried far by streams to form *alluvium.* There is always a transition between the two types.

In connection with "embryonic" rainwash noted earlier (p. 73) there are several other types of deposits, which are mostly silts or loams (*dess* of Morocco).

Volcanic Sediments (fig. 79)

Of materials thrown out by volcanoes, it is obvious that the "stratified" lavas cannot be called sedimentary rocks. However, some of the products of eruption are delivered to the agents of sedimentation, that is to the wind and to running water. Thus, volcanic ash is subject to the same effects as eolian and fluviatile sediments which are deposited in the same place. They may be intercalated between detrital beds, and become receptacles for the preservation of fossils. Volcanic ashes also form excellent soils, whose

FIG. 79. VOLCANIC ASH FILLING A SMALL RAVINE BETWEEN BOU CHARDANE AND THE OUED ZOBZIT (CENTRAL ATLAS; REGION OF BERKINE, MOROCCO)

The ash was erupted during the Quaternary from the volcano of Yerfoud, which pierced the flank of the Jebel Ichouadda. The ash is well bedded and more or less consolidated. These sediments are preserved in a great number of isolated patches, each corresponding to a small ravine or basin formed by the erosion of the Jurassic marls. The size of the particles decreases progressively as the distance from the volcano increases. Note the strong discordance between the ash and the Jurassic rocks (lower left-hand corner).

fertility has at all times attracted farmers to regions made perilous by the proximity of eruptive centers.

The recent eruption of Paricutin (Mexico) has clearly demonstrated the role of the wind which carried the ashes in whirlwinds, and also that of rain-wash, which likewise transported them. The town of Herculaneum was buried under volcanic muds resulting from torrential rains which occurred simultaneously with the eruption of Vesuvius.

Glacial Sediments (figs. 80–84)

Glacial sediments are similar to fluviatile ones, but they are neither rounded nor sorted. They form surface moraines (terminal, lateral, median) and particularly *ground moraines* where the abrasive materials of glaciers are concentrated (see p. 121). Ground moraines or "tills" are composed of

FIG. 80. MORAINES, NEAR MONT-LOUIS, EASTERN PYRENEES, FRANCE

The blocks are rounded but are not sorted, and are enclosed in a sandy clay.

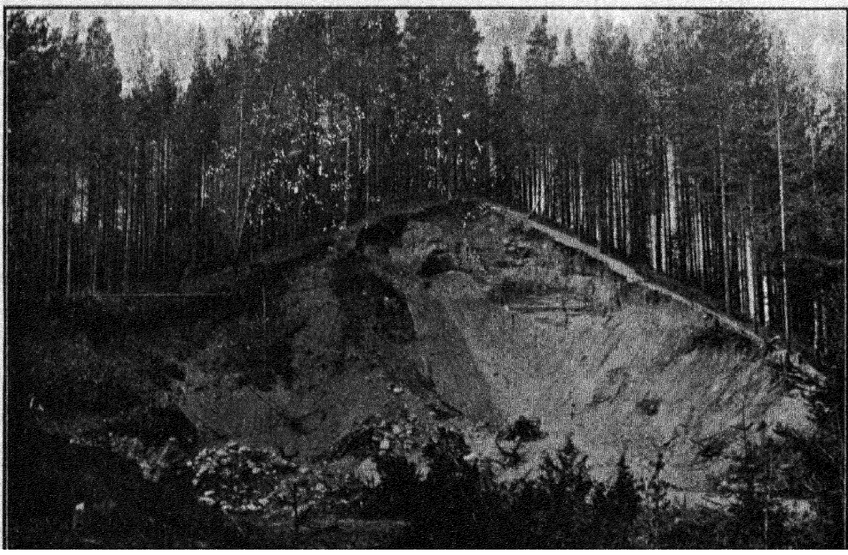


FIG 81. THE HORIZONTAL STRATIFICATION OF AN "ESKER" IN FINLAND

(Photograph kindly provided by the Finnish Geological Commission.)

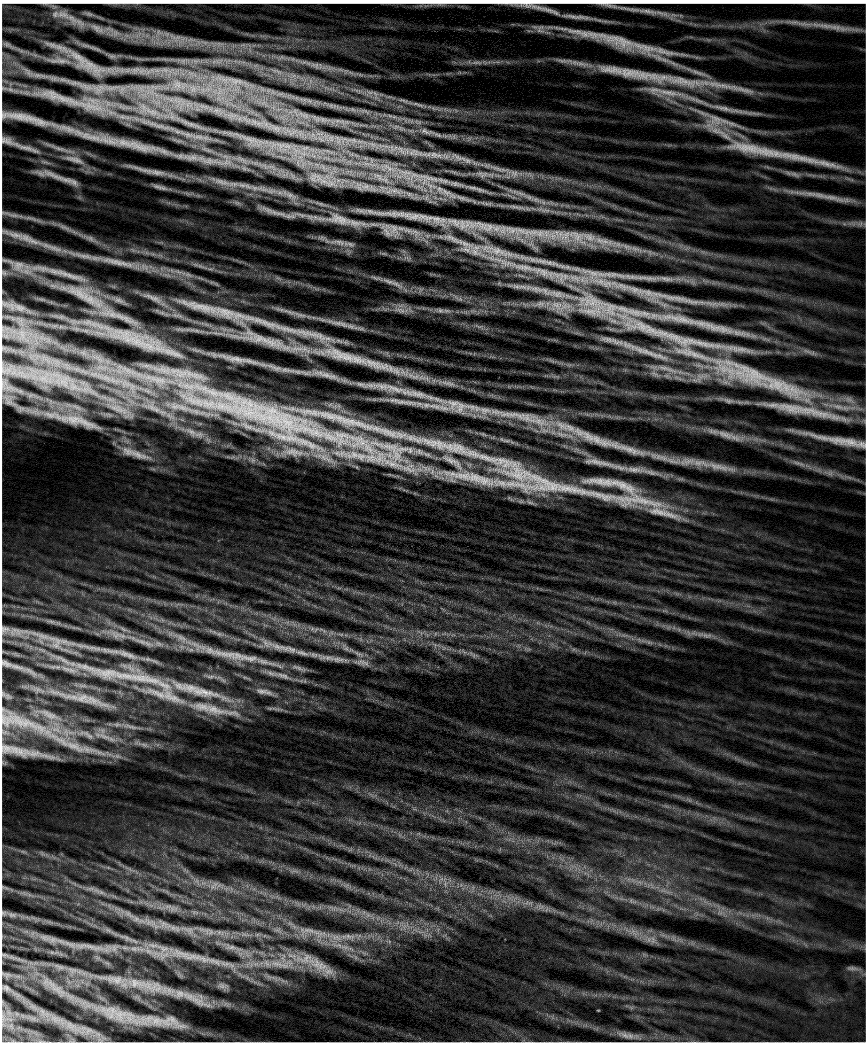


Fig. 82. Fluvio-glacial Stratification in an Esker, Ruskeahiekka, Helsinki
Showing detail of a section about 20 inches high. (Photograph: K. Virkkala)

fragments which may be very large (boulders) and fine clays. Surface moraines consist mainly of coarse materials. These sediments are found commonly in *boulder clays*. When consolidated, these are called *tillites* and represent ground moraines. In polar regions where glaciers reach the sea, icebergs may calve off and carry with them part of their moraines. These are gradually released, thus forming deposits which are both glacial and submarine.

Erratic blocks, which are often enormous, may be carried very great distances. For example, crystalline rocks of the Scandinavian shield have been carried on to the North German plain. Some erratic "nappes" exceed one mile in length.

The finest part of the glacier's sediment is carried by running water, particularly by the melt water running from the glacier itself. This sediment may be carried considerable distances before being deposited, possibly in a lake. These lacustrine deposits are called *varves* (from Swedish *varv*=bed). They are banded clays or fine sands, the characteristics and thicknesses of which are closely related to seasonal or other variations affecting the glacier and the run-off. Thus the summer layers are thicker and lighter in color, whereas the winter ones are thin and dark (since they are richer in organic matter).[1]

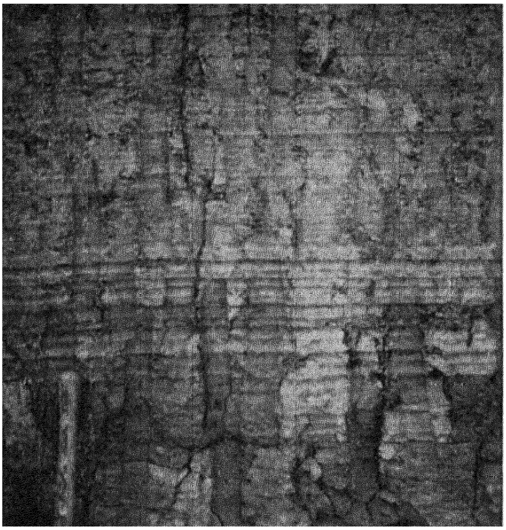


FIG. 83. STRATIFIED GLACIAL CLAY. HIPPOS-AITOLAHTI BRICKWORKS, TAMPERE, FINLAND

Note the fine stratification due to *varves*, i.e. seasonal banding in glacial lakes. (Photograph: Virkkala.)

From their observation of varves, De Geer and his school have developed a valuable method for the determination of the chronology of the retreat phase of the last Quaternary glaciation. Another aspect of the characteristic bedding of varves has been the attribution of ancient deposits of similar type to the same cause, thus identifying them as glacio-lacustrine in origin. However, it is possible to confuse the true varves with well-bedded sediments of different origin. Thus it is necessary to confirm the glacial origin of these deposits by the nearby presence of other glacial sediments, such as tillites.

During the recession of the Quaternary ice sheets, detrital accumulations

[1] The term *varve* is generally limited to those deposits resulting from glacial sedimentation. However, the word is also used to denote seasonal deposition. It is thus common to find that certain authors emphasize the seasonal character of varves. It is in this sense that M. Gignoux (1953) has applied the term to the Pliocene coastal marine clays of Gard. There, the seasonal variation in the transport of sediment from the continent corresponds with the annual alternation of dry and wet periods, or to the changes in coastal currents. The thickness of varves varies between a fraction of an inch and several inches.

were left behind, either in the form of outwash fans deposited at the snouts of glaciers by subglacial streams (os), or as fluvio-glacial ridges often filling in lakes (eskers) (figs. 81 and 82).

Glacial material, such as cobbles and gravel, also may be transported by rivers coming from the glaciers. The sediments resulting from these succes-

FIG. 84. QUATERNARY FLUVIO-GLACIAL STRATIFICATION IN THE JURA MOUNTAINS, NEAR SAINT CLAUDE, EASTERN FRANCE (Photograph: G. Termier)

sive actions have mixed characteristics clearly demonstrating their double origin, and justify the use of the term *fluvio-glacial* for these deposits. The Eocambrian sparagmite of Norway is a good example of this kind of deposit.

Frost-shattered Rocks (fig. 85)

The alternation of freezing and thawing, which occurs in high mountains or in periglacial zones is undoubtedly an agent of degradation which produces detrital materials. The absorbed water increases in volume on freezing, and thus tends to exert pressure on the walls between which it occurs. If the spaces are large enough to allow the expansion to take place, the rock will suffer little damage (e.g. vesicular basalt or very porous sandstone), but if the voids are small, their walls are pushed apart (e.g. argillaceous limestones). Such rocks are exfoliated, often fragmented and ultimately, disintegrated.

On the other hand, some uncemented rocks may become consolidated by interstitial ice and can then slide or be transported (Gripp, 1954). "Boulders" of uncemented (but frozen) sediments may thus be incorporated in conglomerates.

Sediments of High Mountains (fig. 86)

In high mountains, the steepness of the slopes augments the effects of gravity. Erosion, however, can hardly be due to rainwash, since atmospheric precipitation occurs as snow and not as rain. The principal agent of disaggregation is thus ice, aided by névé and by glaciers. Sediments resulting from the destruction of rocks in the high mountains are, therefore, partly screes and partly moraines. The screes occur either as flattened cones or fans, or in long ridges of stones when they are channeled. These "moraines" have characteristics similar to those formed by ice sheets.

Eolian Sediments

In connection with eolian erosion (pp. 86–88), the agents of erosion and the resultant deposits have been noted.

The sediments which are the agents of wind erosion are classified mainly with the sands and silts, but they are not, in their entirety, the products of this form of erosion. Even in arid countries where the wind has maximum effect, they result mainly from erosion by rainwash. This explains the thinness of eolian deposits in some parts of the Sahara where rainwash is particularly limited.

The dust in suspension in the atmosphere is derived from loams of fluviatile or pluvial origin which dry out in an arid climate. They can then be easily transported by the wind. Included in this type are the *loess* of semidesert (often periglacial) areas and the *adobe* of Central America, both of which grade into reworked loams. These loams retain a large amount of clay which gives to these eolian deposits their characteristic properties. They are exploited for pottery and brickmaking, and are potentially fertile soils (e.g. the loess of China and the adobe of Mexico, especially in the province of Sonora). The dust deserts owe their aridity solely to the dryness of the climate, because if they are irrigated they can become fertile and can be cultivated. In their natural state, they support a fairly abundant xerophilous plant cover. When the eolian dusts have little clay, their consolidation gives rise to *pelites*.

Eolian sands are generally quartzitic, although occasionally they are saline. The quartzitic sands are the principal agents of eolian erosion. Unlike the dusts, they form *dunes* which can move at a measurable rate (see p. 168). The purity of the silica sand limits their fertility and only halophilous plants can live on them. Such plants are used to fix these dunes in place. Eolian sand is used for glassmaking, as at Saint-Gobain where the Fontainebleau sand is utilized. The same sand is also used in the Murano factory in Venice. However, a small amount of iron oxide will color sands of

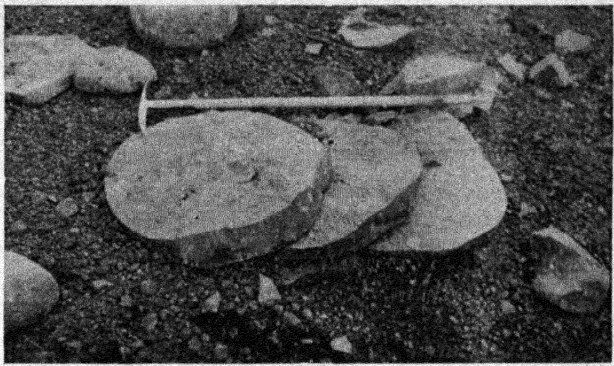


FIG. 85. FRAGMENTATION OF A BLOCK BY ICE TO FORM DISCS WHICH HAVE BEEN DISPLACED LATERALLY. CENTRAL ICELAND
(Photograph reproduced from *Précis de Géomorphologie* by Derruau, Masson, Paris.)

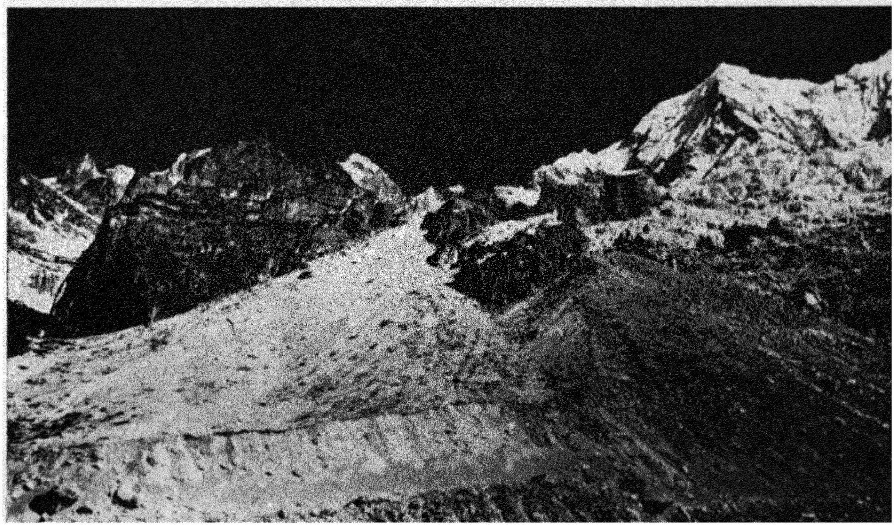


FIG. 86. EROSION AND SEDIMENTATION ON VERY HIGH MOUNTAINS

Part of the Jannu Massif (the principal range of the Himalayas), composed of migmatites and granite, surrounded by talus and old moraines. (Photograph: French Himalayan Expedition, 1959.)

hot deserts yellow or red. The origin of this iron is difficult to determine, although iron is one of the most mobile elements in the upper part of the earth's crust. It may be derived from lateritic erosion, or from soils developed on limestones.

Among sedimentary rocks, fossil eolian deposits show certain characteristics which enable them to be distinguished from fluviatile sediments, although both types show cross-bedding (Opdyke and Runcorn, 1959). The bedded layers, often several feet thick, are steeply inclined up to angles of 33° to the horizontal, which is the angle of rest of dry sand. Eolian sands contain neither marine fossils nor mica (which is either carried away by the wind or destroyed by percussion). The sand grains are well sorted, rounded and frosted by repeated impacts.

In this category can be placed the New Red Sandstone (Trias) of Britain, and certain sandstones of the western United States. These include the Weber sandstone of Utah, and the Tensleep sandstone of Wyoming (both are Pennsylvanian), the Wingate sandstone (Triassic) of Utah, and the Navajo sandstone (Jurassic) of Utah. There are, however, some differing opinions about some of these examples. Water-laid phenomena (slumps, low-angle cross bedding) are observed, and suggest that eolian sands have been blown into a shallow sea, as may be observed today along the Persian Gulf.

Coastal Sand Dunes.—In Europe, a large part of the low-lying coasts are bordered by dunes of quartz sand. The southern coasts of the North Sea and the Baltic, the coasts of the Bay of Biscay (150 miles long and 2 to 5 miles wide) and the coast of the Gulf of Lions can be quoted as examples. These dunes correspond to several periods of formation: for example, during the Pleistocene two series of dunes, which are now fixed by plants, were formed in the Bay of Biscay.

The discontinuous chain of dunes, along the southern part of the North Sea and the Baltic extend for more than 1,800 miles. Sand is also deposited in the sea in the form of banks, such as those which extend from the northern end of the Frisian Islands to the west coast of Schleswig-Holstein. These sediments often bear a particular form of bioherm: Hermella (*Sabellaria*) reefs, formed by polychaete worms which build tubes of sand grains held together by agglutin. The reefs of Heligoland in the North Sea are well known.

Dunes are generally arranged in rows parallel to the coast. In Denmark they cover about 270 square miles. The width of the dune belt is about 6 miles between Blaavandshuk and Skagen. Dunes are present not only on the islands of the North Sea, but also on those of the Kattegat (Laes, Anholt), on the southern coast of the island of Sjaelland and on the coasts south and southeast of Bornholm. Some chains are separated by belts of sand moving, for example, over old peat beds (V. Nordmann, 1928). The bare, living dunes (white dunes) move in the direction of the dominant wind. Very occasionally they occur in the form of parabolic barchans open on the leeward side. A large number (gray dunes) have been fixed by vegetation (beach grass, mosses and lichens). The destruction of this vegetation allows movement of the dune by the wind which erodes the dune and forms a hillock, parabolic in shape but open on the windward side.

In hot regions, the quartz sand of coastal dunes is often replaced by the debris of calcareous algae, foraminifers, molluscs and bryozoans. This occurs in Bermuda, the Bahamas, Madeira, the Cape Verde islands, Morocco, Egypt (Alexandria), Palestine, Arabia, South Africa, the west coast of India, Australia, Hawaii and in Ecuador. The dunes of calcareous sand show the same characteristics as those of quartz sands. The sand is, in fact, a detrital sediment and behaves as such.

R. W. Sayles (1931) has called the rocks resulting from dunes *eolianites*. Their structure sometimes betrays their origin: cross-bedding with steeply dipping foresets and ripple-marks. These ripple-marks, which record undulatory displacement of particles of all kinds, but chiefly of sand grains, are oriented perpendicular to the wind, and may be ramifying if there are two dominant wind directions. Calcareous dune rocks are probably better described as *eolian calcarenites*.

The dune sands of hot regions therefore contain a large proportion of calcium carbonate. "Beachrocks", formed in part from this, are often cemented by secondary calcite which converts them into a form of travertine (see p. 243). At high sun temperatures the first cement is aragonite, which later inverts to calcite.

The sand of coastal dunes results from erosion and transportation by waves and coastal currents, whereas the sand of deserts is mainly produced by fluviatile action.

It appears that the formation of dunes at the edge of the sea under various climatic conditions is directly related to the sterility of the coastal zone, in all latitudes, due to the high concentration of salt. As in deserts, vegetation has difficulty in establishing itself, and consists mainly of halophilous plants. The existence of vegetation capable of fixing the dunes is thus precarious and depends largely on the rainfall. Often, "dead" dunes are fixed, and between them and the sea new accumulations of sand are formed.

At the edge of the sea, the dunes are most easily consolidated, due to the circulation of water and dissolved substances which bind them together. The cement is usually calcite. This is occurring at the present day in many of the Quaternary dunes of North Africa, on the Atlantic shore of Morocco, especially at Rabat, and on the Mediterranean coast of Algeria, near Tipasa. The most extensive examples are in Western Australia.

Eolian sands of coastal dunes have polished grains. But it must be remembered that the eolian characteristics are always of secondary origin and that mechanical erosion is the work of water, while the wind acts only as the transporting agent.

Sandy Deserts (Ergs, Edeyen).—It is rare to find isolated barchans in the

Sahara, although they do exist between El Goleah and In Salah (Algeria) (figs. 87 and 88), and in Mauritania. The eolian deposits of the Sahara are dominantly areas of complex dunes, the *ergs* or sand seas, which occupy three regions oriented more or less southwest to northeast:

1. From the coast of Senegal to southern Algeria, the Mauritanian erg, the Raoui erg, the Great Western sand sea (Grand Erg Occidental) and the Great Eastern sand sea (Grand Erg Oriental).

2. On both sides of the Hoggar mountains, the Sudan erg, and those of Issaouane, Oubari, and Mourzouk.

3. On both sides of the Tibesti mountains, those of Tenere and Rebiana, and the Libyan desert.

These sand seas are thus grouped around the great crystalline massifs which provide them with detritus. The depressions which are free from sand (Qattara depression, the shotts of southern Tunisia, Touat, Bodelé) are undoubtedly due to the concentration in them of water. The sands tend to accumulate on very dry, gentle slopes. The plateaus are traversed only by lines of very low dunes. The bases of high cliffs are not covered by sand, whereas the escarpments and buttes are readily overwhelmed (Dubief, 1943).

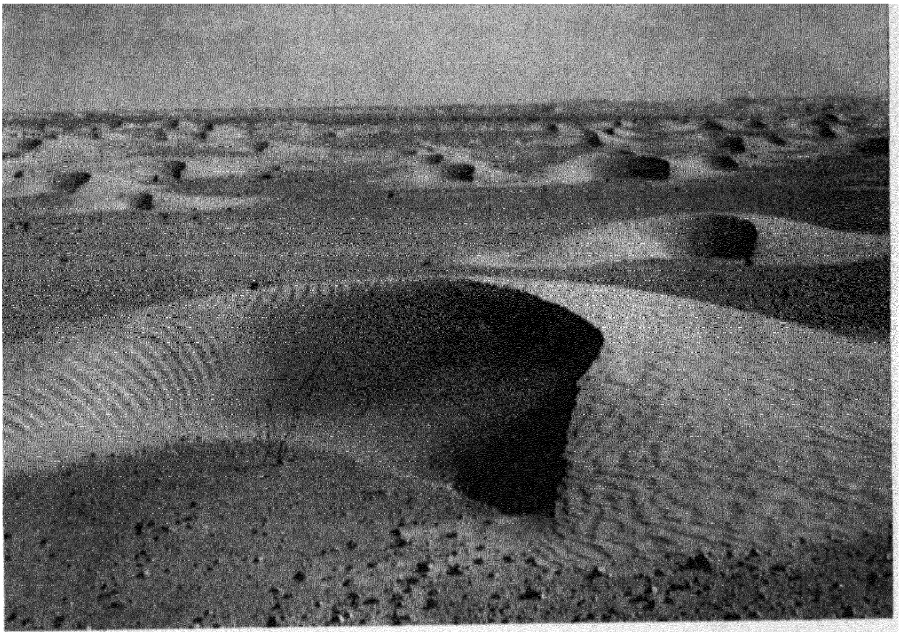
The origin of the Sahara sands was investigated by J. Dubief (1953). This author distinguishes "dust drifts" or "dry haze", which are intercontinental movements of particles less than 0·02 mm. in diameter[1] (those of the Sahara reach Algiers and even Europe), and "sand drifts" which transport material greater than 0·02 mm. over relatively small regions. The winds responsible for the latter movements alone, lead to the formation of dunes. They are linked to semiarid regions, that is, the zones where rainwash selects the material to be transported. They are related to a turbulence of thermal origin in the northern Sahara where intertropical tornadoes also play an important role. This is, in part, a seasonal phenomenon, with a maximum in spring (February in Mauritania, April–May in the southern Sahara) and a minimum in winter (November–December). The direction of the sand drift in the southern Sahara is from southwest to north in winter, and from northeast to south in summer. On the basis of much data, Dubief was led to the idea of an *annual resultant of sand drift* which seems accountable for the actual displacement of the dunes. Furthermore, he believes that the older dunes were built by a wind regime different to that of the present day. Finally, one can distinguish:

1. Ergs still in the course of development (southwest of Mauritania) where the barchans are moved on average, 330 feet per year; 2·43 inches per day for a height of 102 feet; 13·72 inches for a height of 13 feet; 52 inches for a height of 29·5 inches.

2. Old ergs which are relatively stable, although not fixed by vegetation (Grand Erg Occidental).

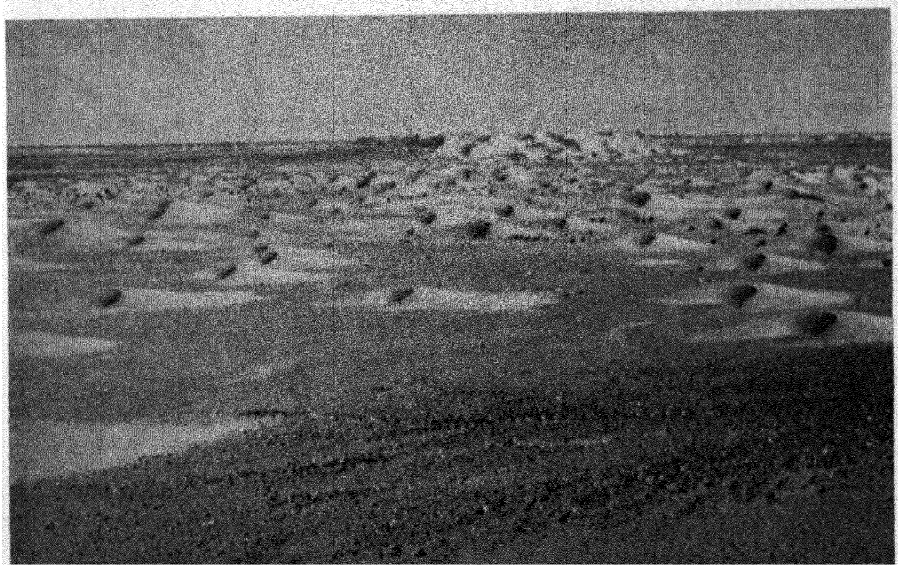
3. Ergs in process of disappearing, due to the wind transporting the

[1] Their consolidation leads probably to the rocks defined as pelites.



FIGS. 87 and 88. BARCHANES SOUTH OF EL GOLEAH, ALGERIA

In the upper photograph, note the occurrence of ripple-marks on the barchan. Between the dunes the reg is covered with polished pebbles. In the foreground of the lower photograph, the pebbles indicate the probable direction of rainwash, and in the background, the accumulation of sand leads to the formation of an erg. (Photographs: G. Termier.)



material to other regions (Erg Chech, western part of the Grand Erg Oriental), resulting in the formation of an undulating sandy plain, often with a wavelength of a mile or more (Mereie, and south of the Erg of Mourzouk).

4. Ergs, the development of which has been arrested and the form stabilized by vegetation in less arid climates (dead ergs of the northern Sudan).

Old dune formations can be preserved in a stabilized region. They then show characteristic relief which has been termed *bult* or hummock topography in the Kalahari and in Senegal.

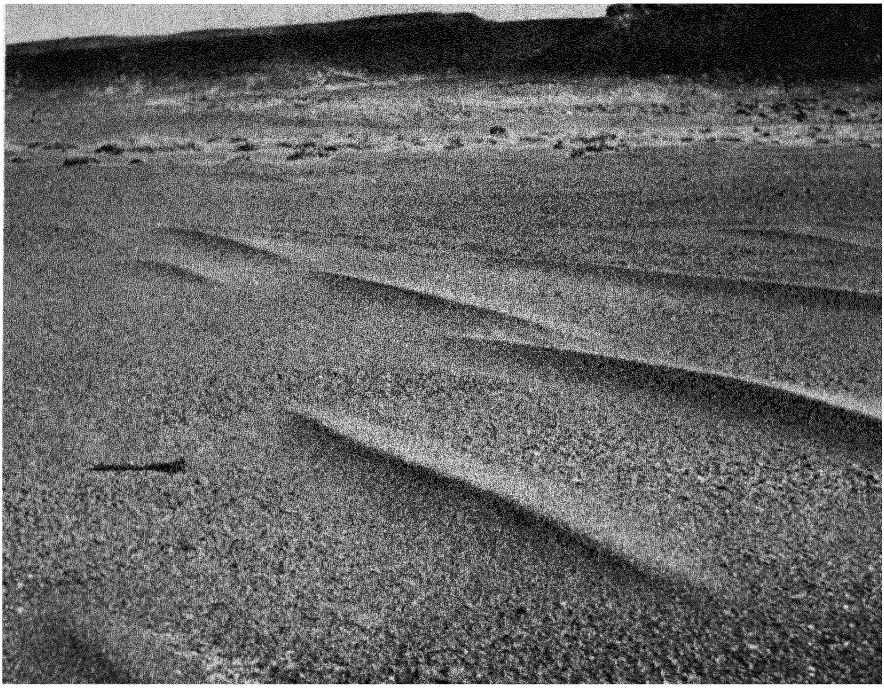


FIG. 89. THE INITIATION OF DUNES IN VERY COARSE SAND TO THE SOUTH OF IN SALAH, ALGERIA (Photograph: G. Termier)

Dunes of Salt and Gypsum.—The action of wind on the dry surfaces of playas produces considerable deflation. Thus the drained lagoons on the high delta of Senegal are covered with evaporites, particularly common salt (which hinders the growth of vegetation). The salt is picked up by whirlwinds, which are frequent at the end of the dry season. In the playas (sebkhas) thus formed, the crystallization of the salt results in the breakup of the bottom muds and the formation of particles of 0·1 to 2 mm. size, which are themselves picked up by the wind and deposited in ridges and small dunes. As much as 6 inches may accumulate in 24 hours (Tricart, 1954).

The action of the wind on the North African shotts is important. Deflation of the Chergui shott has revealed a wind-abraded rock (gara) which is evidence of a morphological surface dominating the bottom at a depth of about 160 feet. The sedimentary counterpart is the existence of dunes of gypsum crystals, which are scattered about the floor of the shott.

FIG. 90. CONSOLIDATED CALCAREOUS DUNES ("EOLIANITE") NEAR THE SHORE OF THE MEDITERRANEAN, IN THE REGION OF TIPASA, ALGERIA (Photograph: G. Termier)

The *vlei* (pp. 119–121) of Western Australia are often bordered by dunes of gypsum sand. Lake Frome, which is a great playa in the depression of Lake Eyre and receives about 12 inches of water a year, has dunes of both sand and gypsum.

The action of the wind has been invoked by T. Monod to explain the saline deposits of Ijil in the Adrar of Mauritania which rest directly on Precambrian rocks.

8

Lacustrine Sedimentation

Stratification of the Water: Dimictic and Monomictic Lakes

At the present day, freshwater lakes occur in the temperate zones, the humid tropics, and in subarctic regions.

When a lake is very deep, the surface waters show seasonal variations in temperature, but toward the bottom there is a layer of water where the temperature remains constant at about 4° C. (39° F.). At this temperature, the density of fresh water is at its maximum. Normally, due to the effect of the wind, a circulation is established at the surface.

Dimictic lakes (Hutchinson, 1957) are those deep freshwater lakes in temperate climates in which the water is, normally, overturned twice a year. The first mixing of the water begins in spring under the influence of the wind when the water has a uniform temperature of 4° C. ("spring overturn"): the bottom thus becomes aerated. In summer the surface is warmed and the circulation occurs only in the surface waters. Thus the lake becomes stratified: the upper mobile part is the *epilimnion*, while the lower stagnant part is the *hypolimnion* which becomes a poorly aerated environment. In autumn, the surface of the lake gradually cools and when the epilimnion reaches a temperature of about 4° C., a second complete circulation begins ("fall overturn") which again aerates the bottom of the lake. But in winter, when the lake is completely frozen over the circulation ceases and the upper layers of the water have a lower density than those toward the bottom. Thus the waters again become stratified. The mixing of the water takes place twice a year: hence the name of dimictic lake.

Subtropical lakes are called *monomictic* (Hutchinson, 1957) since the stratification of summer alternates with a single complete overturn during the winter. This is due to the fact that the water is never below 4° C.

Intermediates between these two types of lake exist. Thus, in Kentucky, Tom Wallace Lake is dimictic in cold years but monomictic in warm ones. The Great Lakes of North America are of such large dimensions that although situated in the temperate zone, they are rarely stratified. The effect of the wind causes instead an almost continuous circulation (Hough, 1958).

Biology and Sedimentation: Eutrophic, Oligotrophic and Apatotrophic Lakes

Living organisms in lakes and the seasonal sedimentation vary with the degree of oxygenation of the water.

Photosynthesis by plants causes localized production of oxygen at varying depths (the limit of the euphotic zone) dependent on the form and population of the lake. This upper zone corresponds approximately to the epilimnion: since it favors the existence of living organisms it is called the *trophogenic zone*. The plants present take in carbon dioxide which is carried to the lake by rain and by tributary rivers. These rivers generally contain calcium leached from rocks, which is present in the water as the unstable bicarbonate: aquatic plants take CO_2 and HCO_3^- from it. It follows that the pH reaches 8 to 9 and, exceptionally, may rise to 11. If the calcium is sufficiently abundant, it may be deposited as limestone on the floor of the lake. Much lower, in the hypolimnion, there is the *tropholitic zone* where photosynthesis cannot take place. The organisms which live in this zone produce only carbon dioxide. There is thus a drop in pH, and in summer it may produce acid conditions in this environment. But if enough calcium carbonate is precipitated in the epilimnion, it may neutralize the acidity of the epilimnion as it settles.

The *eutrophic* lakes are those in which living organisms are abundant. Oxygen is used up on the one hand by plants and animals in the epilimnion, and on the other hand by the decomposition of dead organisms in the hypolimnion. Here the conditions thus rapidly become reducing. An example of a eutrophic lake is that of Lake Minnesota, a small lake to the south of Mankato, Minn. The sediments are particularly rich in organic products (sapropels, copropels, humic acids) and the fineness of the grain size allows the retention of a large amount of water (more than 85% according to Swain, 1956). It is in such lakes in Europe and Asia that the *Botryococcus* waterbloom develops. A special case is that of the small lakes of northeast Greenland, for example those of the Isle of Ella, which are fed by melt water from snow. The arctic climate is so rigorous that plankton die under the winter ice. In Langsø Lake, which is very deep, the incomplete oxidation of this planktonic material leads to the formation of humus which darkens the water in winter. The bottom of the lake thus becomes covered with gyttja (p. 226) which decomposes further and liberates H_2S. It follows that the fauna decreases toward the bottom, the larvae of Chironomidae alone living on the bottom (Andersen, 1946).

In the *oligotrophic* lakes, which contain fewer living organisms, the consumption of oxygen is less and the conditions in the hypolimnion remain oxidizing. The concentration of mineral substances in the water of the lake is generally the cause of the reduction in numbers of the living organisms. Thus, when the water contains more than 100 mg. of calcium per liter (alkalitrophy, Hutchinson, 1957) the existence of plankton is impossible even in the presence of nutrients (see p. 234). This happens in Lake Träske in Sweden (Naumann, 1929–1932).

It has been shown (p. 27) that in arid regions lakes at the present time tend to form playas. There are intermediate types between these saline lakes and freshwater ones. According to the cations in solution, the existence of living organisms may or may not be possible. Thus the presence of magnesium favors animal life, as, for example, in Lake Boulak near the Caspian Sea where the saline concentration is 28·5% and the lake is well stocked.

Swain and Meader (1958) have given the term *apatotrophic* to those lakes in which the water is brackish and contains living organisms. The type example is Lake Pyramid (Nevada), the largest of the residuals of Lake Lahontan which existed in the pluvial periods of the Pleistocene. The salts dissolved in the water total 4,700 p.p.m. (including 7·5 p.p.m. Ca, 111 p.p.m. Mg, 1,570 p.p.m. Na, 128 p.p.m. K) and in this organisms do occur, although they are very much restricted. These include a planktonic flora of diatoms, myxophytic algae, *Chara* and *Potamogeton* and a planktonic fauna of *Daphnia*, while ostracodes which are represented chiefly by their shells, are gradually disappearing, and gastropods are extinct, being represented only by their shells. The adjacent country is volcanic. The principal river to the south is the Truckee River which carries silts. The deep waters of the lake are pushed along by this river from south to north, while the surface waters are agitated in the opposite direction by dominant winds from the north. In spite of this the water is stratified, with a boundary 60 to 90 feet below the surface.

The pH of the water is essentially alkaline, due to the salts of sodium and potassium dissolved in it: pH 9·2 at the surface, pH 9·1 at the bottom, and 8·9 in the sediments. The slow, but continuous, decrease in pH with depth can be attributed to the release of H_2S by decomposing organic matter, and to the acidity of the terriginous sediments. The waters are slightly oxidizing, whereas the sediment is often slightly reducing. The sediment on the bottom consists either of diatoms (diatomite) or diatoms mixed with a mass of detritus of volcanic origin. These sediments are coarse-grained, though the particles are usually not more than 2 mm. in diameter. The organic matter present is in the form of bitumen which always contains sulfur. The quantity is comparable to that of eutrophic lakes. Of the hydrocarbons present, 72% are saturated and the remaining 28% are aromatic. The quantity of amino-acids and humic acid present is intermediate between that of deposits of eutrophic and oligotrophic lakes.

Organic Matter in Fossil Lacustrine Sediments

Modern methods of analysis of small quantities of organic matter (in particular, chromatography) have given information on the character of the fossilized organic matter in sediments, and has allowed the reconstruction of the conditions of deposition.

The Precambrian bitumens of Minnesota have been examined in this way (Swain, Blumentals and Prokopovitch, 1958). The *Thomson Slate* of Knifian Age (c. 1,500 million years B.P.) is a thick series deposited in an elongated basin and contains 200 p.p.m. of bitumen which consists mainly of hydrocarbons, together with humic acid. From the relatively high value of the carbon-nitrogen ratio ($C/N = 4$) and the presence of 15% of carbohydrates in this organic matter, its origin can be attributed essentially to Dinoflagellates (which contain cellulose). The small amount of saturated hydrocarbons indicate that the environment was not very saline (low alkalinity) and it seems likely that sedimentation took place in an oligotrophic freshwater lake. Concordant with the Gunflint Formation (see p. 177), the *Rove* (Animikie age, c. 1,000 million years B.P.) consists of graywackes (volcanic tuffs consolidated under water) passing into argillites. The graywackes contain 450 p.p.m. of bitumen which is mainly saturated hydrocarbons with some asphalts and tar. Material of this type is formed in oligotrophic lakes which are poor in alkalies and in which relatively little protein is deposited. Relatively abundant aromatic hydrocarbons indicate a certain degree of salinity (or alkalinity) of the water. In the Rove argillites, the bitumens form only 180 p.p.m. and contain only a small proportion of hydrocarbons. The ratio $C/N = 3{\cdot}3$ indicates an environment where the living organisms were rich in protein. The graywackes thus seem to have been deposited in an oligotrophic saline lake and the argillites in a coastal sea nearby, in an environment comparable to the Gulf of Mexico today. There was thus an alternation of these two conditions.

It is not impossible that the Timiskaming deposits which are perhaps contemporaneous with the Thomson Slate, were also formed in a lake (MacLaughlin, 1955).

Very recently, the Eocene lacustrine beds of the Green River have yielded information which has permitted a reconstruction of the evolution of the lake in which they were deposited. At first, the lake was hyposaline (i.e. of low salinity) in which paraffinic hydrocarbons were deposited. The appearance of cyclic aromatic hydrocarbons marks an increase in salinity. In the final stage of hypersalinity (i.e. very high salinity) when mineral salts are precipitated, the organic matter is present as asphalts and nitrogenous substances (Hunt, Stewart and Dickey, 1954).

The Deposition of Iron and Silica

According to Ruttner (1953) and Hutchinson (1957) the deposition of iron in lakes is controlled by Eh (oxidation-reduction potential) and the pH (acidity or alkalinity). The iron is leached during rock weathering and carried into the lakes by streams. Under reducing conditions and at a pH of about 6, *ferrous* iron is 100,000 times more soluble than *ferric* iron under oxidizing conditions when the pH is 8·5 (Cooper, 1937).

In a dimictic lake, reducing conditions occur near the bottom in summer, and in winter lead to the formation of ferrous compounds from iron in suspension. These compounds include the bicarbonate, sulfate, and organic complexes which remain *in solution* in the hypolimnion. When the spring and fall overturns bring oxygen to the bottom of the lake the iron is oxidized and *precipitated* in the form of ferric compounds which are relatively insoluble. When the bottom waters contain H_2S, direct precipitation of iron sulfide (soluble only at low pH) can occur. But, more often, the iron sulfide of bottom sediments of lakes can be attributed to the reduction of ferric iron within the sediment.

In contrast, the deposition of silica in lakes is closely linked with living organisms. This has been clearly shown by Ruttner (1953) in connection with the pullulation of diatoms and Silicoflagellates in the epilimnion. Water also contains silica in colloidal form and in true solution as H_4SiO_4 (Alexander, Heston and Iler, 1954). The solubility of colloidal silica decreases with the pH, slowly between 11 and 6·5 and more rapidly between 6·5 and 4·5 (Correns, 1949). The hypolimnion of lakes poor in lime, where the pH varies between 5 and 7, is thus a favorable environment for the precipitation of silica. There is still, however, considerable uncertainty concerning the behavior of silicic acid.

On the bottoms of Lakes Ontario, Michigan and Huron an argillaceous clay rich in diatoms is at present being deposited, in which there is an alternation of layers (each less than 1 inch thick). The layers are either dark in color due to the presence of iron sulfide, or gray, in which the iron is present as hematite. The amount of iron and silica present is small, but this alternation does demonstrate the functioning of a seasonal iron cycle.

The Banded Iron Ores of the Precambrian

The principal iron deposits of the world, which are of Precambrian age, represent a particular type of sedimentation which has not recurred in later times. These are the *banded iron ores* where the iron is associated with silica and perhaps with phosphates. Among others may be noted the Shamvaian ores of Rhodesia, dated between 2,850 and 2,650 million years B.P.; the Minnesota ores 1,800 to 1,100 millions years old; the deposits of Lapland (Kiruna, Gällivare) (1,150 millions years old); those of Manchuria (Anshan) about 1,000 million years old, and perhaps also the deposits of Itabira, in Brazil.

Van Hise and Leith (1911) and later Launay (1913) believed these to be sedimentary deposits. Other authors have since, however, attributed an eruptive origin to them. In 1950, Sakamoto put forward the hypothesis that they had been formed in basins of lacustrine type. Backlund (1952) has compared those of Lapland to the well-bedded deposits which are formed in the present day in the deep stagnant lakes of Fenno-Scandinavia. These modern sediments contain silica, clay, calcareous mud, bituminous organic matter, phosphorous compounds and oxides of manganese intimately mixed with limonite.

According to J. L. Hough (1958) the banded deposits of Minnesota, where silica and iron alternate, were deposited in a region of low relief (mature physiography) since coarse detritus and the more mobile ions in solution (Ca^{++}, Mg^{++}, Na^{+}, K^{+}) are absent. The climate was warm and temperate (or subtropical) with moderate rain which leached the surrounding regions. The basin of deposition was a very deep freshwater lake which was stratified, showed a monomictic cycle, and was without communication with the sea. The environment was oligotrophic, but poor in calcium. During the summer the Eh was only slightly modified since there was little deficit of oxygen, and the pH of the hypolimnion which was enriched in CO_2 fell slightly to 5–7. The transported elements were carried in the form of colloidal suspensions. The silica was deposited mainly in the summer and the iron oxide during the winter "overturn". Thus the banding can be explained as a seasonal phenomenon similar to varve formation. The Gunflint Formation, which forms part of the banded formations of Minnesota, has siliceous bands containing organisms (blue algae, fungi, siphomycetes, and possibly a form of flagellate) which give some indication of the phytoplankton of this environment (Tyler and Barghoorn, 1954).

It is possible that the conditions of deposition of the banded ores result from the interaction of a climate of unusual type and high atmospheric carbon dioxide content during these early times (Siever, 1957). It must be added that Macgregor (1951) sees in the oxidation of the iron of the Rhodesian ores the proof of the presence of free oxygen in this atmosphere.

9

Transitional Coastal Zones—Lagoons, Estuaries and Deltas

Estuaries and Deltas

It will be recalled that rivers, the principal carriers of detrital sediment, form either an estuary or a delta where they meet the sea. The river does not, however, entirely lose its identity at this point, but tends to flow over the sea for some distance, carrying with it some part of its sedimentary load, and in some cases following the course of a submarine canyon (p. 48).

In estuaries, the transgressive character of the present period is clearly demonstrated, since the sea penetrates more or less deeply into the subaerial course of the river. This phenomenon is widespread along the coasts of the world. The extreme cases are the rias (p. 66) and the fjords (p. 126). The sediments often form a bar in front of the estuary (fig. 91). They also cover the floors of fjords and rias and form muddy banks (fig. 92).

Deltas (pp. 72, 112) are formed by the deposition of sediment in the distributaries of a river (often of estuarine type) on an open continental shelf. They are generally of considerable age and are built up by a mosaic of simple deltas representing the successive stages of the evolving delta. For example, the thickness of the deposits of the great Mississippi delta is 20,000 feet for the Paleocene and Eocene of the Gulf Coast, and 17,000 feet for the Miocene and Pliocene of Louisiana.

Offshore Bars (fig. 91)

The interplay of fluvial and marine currents on a gently inclined continental shelf may build up, particularly near an estuary, a bank of more or less coarse debris brought by the river. This is known as the *offshore bar* or *barrier-island* (*cordon littoral*, Fr., *Nehrung*, Ger., *lido*, It., *flèche*, *épi*, *pulier*, Artois). Such a barrier, which takes a long time to build, plays an important morphological role:

1. Behind it, the sea is more or less isolated in a lagoon in which fluviomarine sediments accumulate and tend to fill it. This zone may pass through a series of stages: lagoon (sometimes with salt marshes), then muddy tidal flats (slikke), salt marshes (schorre), then polder or salt meadows, and finally firm ground no longer covered by the sea at any time, although this may be

marshy in places. This evolution differs from that of a delta (although the conditions of establishment are almost identical) in that the delta develops on a subsiding platform.

2. As long as the lagoon is present the initial stream will attempt to find an outlet over or through the bar and will be deflected from its original course.

3. The bar is subject to the local climate of the littoral zone where wind action is predominant It often serves as a focal point for the formation of dunes and beach sands.

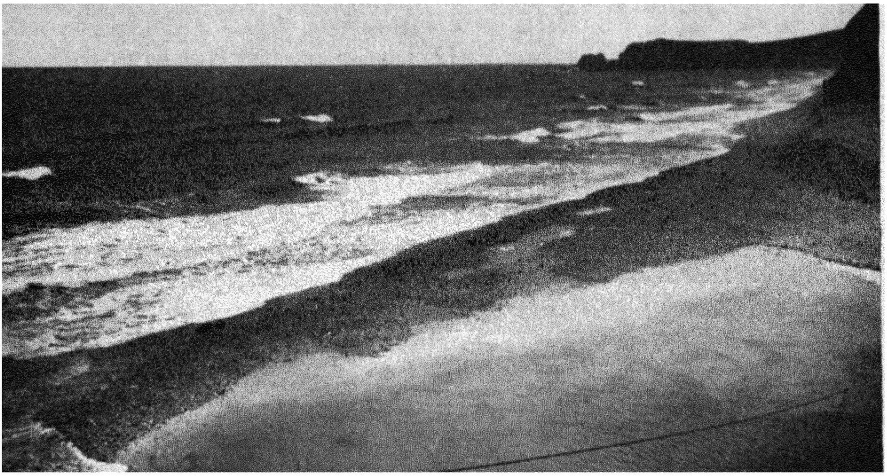


FIG. 91. SAND BAR TEMPORARILY CLOSING THE MOUTH OF A RIVER, NORTH BASQUE COAST OF SPAIN (Photograph: G. Termier)

If the coast is bordered by islands the bar may form a link between them or may link the islands to the coast. The formation of such a bar, or *tombolo*, is due almost entirely to the action of waves and marine currents. A typical example is that of Monte Argentario, northwest of Civitavecchia (near Rome), where the island is joined to the Italian mainland by three tombolos. In the south of France the peninsula of Giens, to the south of Hyères, is formed by a rocky island linked to the coast by two tombolos.

Frequently, tombolos appear to have been developed on top of banks of Posidonias (e.g. at Porquerolles and in Sicily, at the Punta d'Alga near Marsala (J. J. Blanc)).

Fluvio-marine Muds

The material in suspension in a river forms a "muddy bung" (the *bouchon vaseux* of L. Glangeaud) oscillating upstream and downstream with the tide. It is chiefly these suspensions which give rise to the fluvio-marine muds

(C. Francis-Boeuf, 1947). They are very fine-grained and rich in iron and organic matter. The precipitation of these muds takes place in those zones subjected to violent current action where the penetration of salt water from the sea into the estuary probably brings about changes in the electrostatic equilibrium of the colloidal matter in suspension. Warmth, as well as salinity, accelerates deposition of the mud (Berthois, Chatelin and Marcou, 1953).

Sediments forming on coasts at the present time derive much of their material from rivers, although some part of the sediment is provided by the sea itself. Deposition takes place largely at the mouths of rivers and the character of the sediment justifies the use of the term "fluvio-marine sedimentation".

As has been emphasized by C. Francis-Boeuf (1947) "the geographical importance of the estuary is determined by a number of factors: (1) the river discharge; (2) the gradient of the river bed; (3) regional tidal action."

The mud of estuaries is inhabited mainly by annelids and pelecypods (*Scrobicularia*, mussels, oysters, *Cardium*, *Mya*), seaweeds (*Fucus lutrarius*, *Enteromorpha*), Naiadales (*Zostera*, *Spartina*) and when the climate permits, mangroves. Bacteria are abundant, especially the aerobic iron bacteria (*Leptothrix*, *Crenothrix*) which are able to concentrate the iron of their surroundings (Harder, 1919, Andrée, 1920).

The mud is of mixed origin, the mineral part being derived mainly from the land, whereas the bulk of the organic matter is provided by the sea and organisms living *in situ*. The marine contribution contains a large proportion of plankton, chiefly diatoms. Fluvio-marine muds are rich in water (80 to 100%) and the water of the estuary may contain 200 g./liter of suspended material (mouth of the Kapachez in Guinea, cited by Francis-Boeuf, 1947).

Fluvio-marine muds are characterized by a form of "physiological activity" (Francis-Boeuf, 1947) due to the presence of abundant living bacteria. Below the surface, the mud is anaerobic and certain bacteria produce hydrogen sulfide (from alkaline sulfates) which reacts with the iron oxides concentrated by the iron bacteria to give iron sulfide (see p. 232). This contributes to the blue-black color of the mud.

The size of the grains of a mud may be less than 0·001 mm.; that is, it may be of colloidal dimensions.

The "Tangue" or Calcareous Mud of Brittany.—The *tangue* is a complex clay formed in the shelter of the marshes of northeast Brittany (France), particularly to the south of Mont-Saint-Michel (fig. 101). The deposition of this mud is associated with the transgression of the sea across the area, although the role of the sea seems to be secondary in its formation. According to Y. Millon, it is a fluvio-lacustrine silt which has been reworked by the sea. This clay is very stiff, but is strongly thixotropic (with the consequent danger of forming quicksands). It is characterized by 25–61% of calcium carbonate and is, in fact, a marl. The activity of the sea is limited to the effects of burrowing organisms and the supply of marine shells by currents. These shells are rapidly reduced to powder by Cyanophyceae and then incorporated in the mud. Organic matter and iron are present in only small amounts, in contrast to the estuarine muds. Moreover, the size of the grains is never less than 0·002 mm.

The tangue is a bedded deposit, each layer corresponding to one or more spring tides. The surface is often ripple-marked, especially in the sandier parts, and when it has dried for a long time, mudcracks are formed.

Small solution hollows develop where the calcium carbonate is dissolved by water draining from the salt marshes further inland (Phlipponeau, 1956).

The muds of estuaries and the tangue form mud flats, while the salt marshes occur at a higher level.

The Soils of Salt Marshes.—In the salt lagoons isolated behind barrier beaches a deposit of clay and mud forms, which is colonized by algae and vascular plants. As a result, soils begin to form on the mud. In tropical climates, marshes of this type become covered with mangroves. In western Europe the muds develop into salt marshes and then salt meadows populated with *Salicornia* (Marsh Samphire or Glasswort), *Puccinellia* (Meadow grass or Spear grass), *Suaeda* (Seablite) on the sandy clays, and *Zostera* (Eelgrass) on the muds. As on beaches, the mixing of the soil by burrowing worms is of great importance. Once plants have begun to grow, they act as a trap for further sediment. The marshes remain broken up by salt water channels, or creeks.

Measurements of the rate of sedimentation on the salt marshes of Scolt Head Island, Norfolk, England, have been made by J. A. Steers (1959). Between 1937 and 1957 the depth of sediment had increased from $\frac{5}{16}$ to nearly 9 inches at one site, from $\frac{3}{4}$ to 9 inches at another and from $\frac{1}{2}$ to 8 inches at a third. In the marshes of the Dovey estuary, Wales, F. J. Richards has noted an increase of $2\frac{1}{2}$ inches in 100 months in a plant colony of *Glyceria*, and of 4 inches in 54 months in an association of *Glyceria* and *Armeria*.

Sedimentation is still more rapid in the tropical zones populated by mangroves (and accompanied by *Cymodocea* and *Spartium*). These colonize muddy, sheltered coasts near the mouths of rivers and assist in the spread of the mud banks toward the sea, by fixing them and then bringing about the accumulation of as much as 30 feet of additional material during recent times. Three zones can be recognized in mangrove swamps; in the lowest zone, nearest the sea there occurs *Rhizophora mangle*, above this comes *Avicennia* and still higher up, *Conocarpus*.

Tidal Mud Flats (figs. 92, 93 and 94).—In the higher part of the zone reached by tides, muds may be deposited during the ebbing of the tide. These form tidal mud flats situated between the levels of mean high and mean low tides. The soft mud is composed of putrid, colloidal organic matter mixed with sand. The diameter of the grains of this mud is less than

16μ. These mud flats are without vegetation and are furrowed by tidal channels, so that it is not surprising that the distribution of this mud suffers frequent modification.

However, colonization by *Spartina* and *Salicornia* transforms the mud flats into "high mud flats" and prepares the ground for other vegetation.

The development of the mud flat also occurs in tropical climates. The mud (also known as "poto-poto" in West Africa) is there colonized by mangroves, the roots of which assist in the fixation of more mud particles.

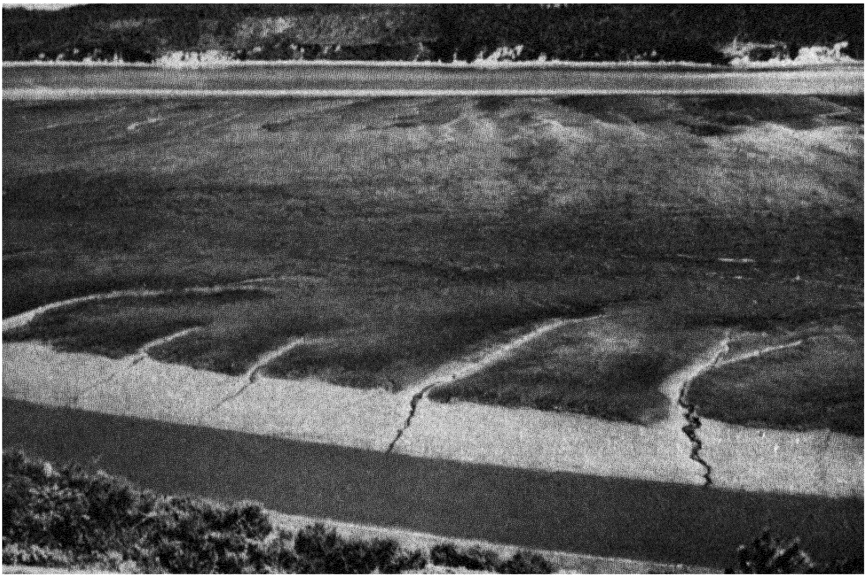


FIG. 92. SALT MARSH AND MUD FLAT IN THE MIDDLE OF THE RIA OF THE RIGADEO, GALICIA, SPAIN (Photograph: G. Termier)

Salt Marsh (figs. 92, 93, 94 and 95).—Fixation by vegetation modifies the character of the mud flats which then become *salt marshes* (covered by the sea only at spring tides) and sometimes used as light grazing land for cattle. They have horizontal surfaces formed of very firm colloidal clay, which may be accompanied by sand and gravel brought by the wind and by rivers. Situated in sheltered estuaries and in rias, the salt marshes are covered by a network of tide channels which are more or less interconnected. Between the channels the vegetation grows profusely; in western Europe it is halophilous and is dominated by *Juncus maritimus*, which also traps mud particles. Due to the period of submergence of the salt marshes by the tide the fauna is amphibious.

The evolution of the salt marshes of the Atlantic coast tends toward the progressive desalting of the ground, leading to the formation of salt meadows without channels. On the Mediterranean coasts, where salt marshes rarely form, the rushes which first colonize the sand are replaced by *Salicornia*. This is followed by an efflorescence of salt and finally by an invasion of Cyanophyceae.

FIG. 93. THE SOUTHERN SHORE OF THE RIA OF FAOU, FINISTERE, BRITTANY

The development of marine marshes with mud flats behind a bar. To the north (top of the photograph) is the shiny surface of the mud flat, free from vegetation. To the south, the marine marsh is dark-colored and covered with vegetation. It is drained by a network of sinuous channels. (Photograph: French *Aéronautique navale*.)

The slow formation of the salt marshes show that these are established on stable coasts where the mean sea level is almost constant.

The Salt Marshes of the Rias.—The banks of certain rias, such as that of the River Tagus at Lisbon and that of the Bou Regreg at Rabat, where the sea flows far into the estuary, are occupied by a variety of mud flats in process of developing into salt marshes (fig. 96).

"Sansouires"—Wet Sand Tracts in the Rhone Delta.—Lagoons and salt pans may be left isolated on alluvial plains at considerable distances from the sea. These receive water only from rainfall or irregular seasonal flooding. The salt in them consequently remains fairly constant in quantity but is from time to time diluted, according to the season. Silt derived from flood waters gradually fills these basins and is fixed by the plant population which is little different from that of the salt marshes.

The Mississippi Delta (fig. 97).—This sedimentological unit, which has received much attention from American geologists, has been an area of deposition and subsidence since Jurassic times. The modern delta of the Mississippi is, in fact, built up of a mosaic of old deltas which have been

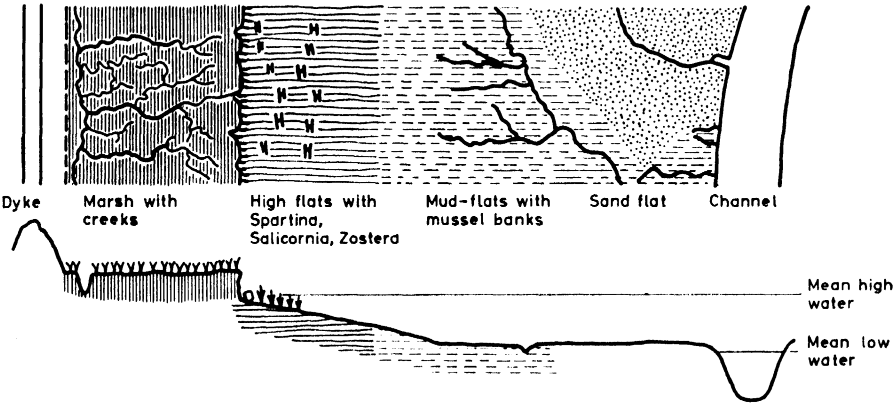


FIG. 94. THE WADDEN SEA, FRIESLAND, HOLLAND

Diagrammatic plan and section showing the position of the mud flat and the marine marshes (after Van Straaten).

partially destroyed and partially preserved. The maximum thickness of the sediments is found toward the open sea. The coast of Louisiana, comprising a series of lagoons closed by barrier islands, lies over the ancient delta of the Mississippi. At the present time the river is building a series of more or less concentric zones at the mouths of the distributaries which are extending the delta seawards in the following manner (Scruton, 1955):

1. *Swamp* deposits.

2. Silts and sands of the *delta front*—up to 65 feet thick in the river mouths. These deposits, which are composed of more than 50% of silt, are constantly reworked by waves and currents and are well sorted. They are deposited in horizontal beds, sometimes ripple-marked, or they may be current-bedded. Some lenses are rich in lignitic constituents.

3. *Prodelta.*—This is an area of deposition in front of the true delta, forming a continuous band 1 to 10 miles wide and reaching a depth of 180 feet. These are soft clays commonly containing only 1·6% of sand and consisting mainly of terrigenous material (clays and silts mixed with fragments of shells and marine organisms). This material is thinly bedded

($\frac{1}{10}$ inch) with some thin lenticular intercalations. This is deposited from suspension where the fresh water of the rivers meets the salt water of the sea.

4. *Clays of the open sea.*—These are clays and silts of fluviatile origin which are deposited on the ocean floor and contain shells and organic debris of marine origin. These thin deposits extend 20 miles out from the coast in waters 115 to 150 feet deep.

5. *Marginal deposits.*—These are the thin heterogeneous mixtures of sand and clay deposited on the continental shelf round the margins of the delta. They are worked over by burrowing organisms, and toward the open sea become coarser grained as they are winnowed by current action.

6. The *sands* ringing the delta result from the destruction of old delta deposits. They appear to have been derived originally from the southern end of the Appalachians.

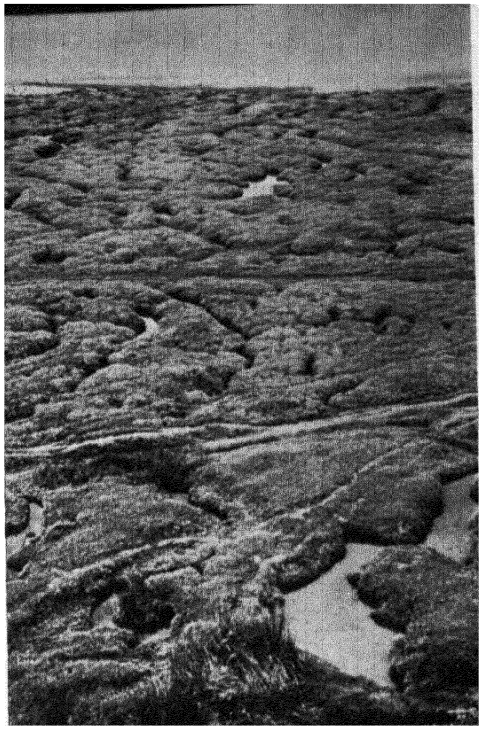


FIG. 95. THE VEGETATED SALT MARSH ("SCHORRE") AT THE EDGE OF AN ESTUARY IN BRITTANY (SAINT-POL-DE-LYON) (Photograph: G. Termier)

It should be noted that the delta of the Mississippi contains almost exclusively fine-grained sediments, to the exclusion of pebbles. Coarse sediments are, on the contrary, abundant in the deltas of swift flowing rivers, for example in the Crau, the old delta of the Durance.

Lagoons on the Borders of Enclosed Seas

These lagoons are not very different in their sedimentary evolution from the playas associated with endorheic basins. Like these, they are fed rarely with either salt or fresh water, and undergo evaporation under the effects of strong winds.

They are thus natural salt pans, like those of the arid part of the Gulf of Mexico (Laguna Madre) and at several places in the Mediterranean. There is the example of the pools of Languedoc containing water of salinity comparable to that of the sea, but bordered by small dunes of salt.

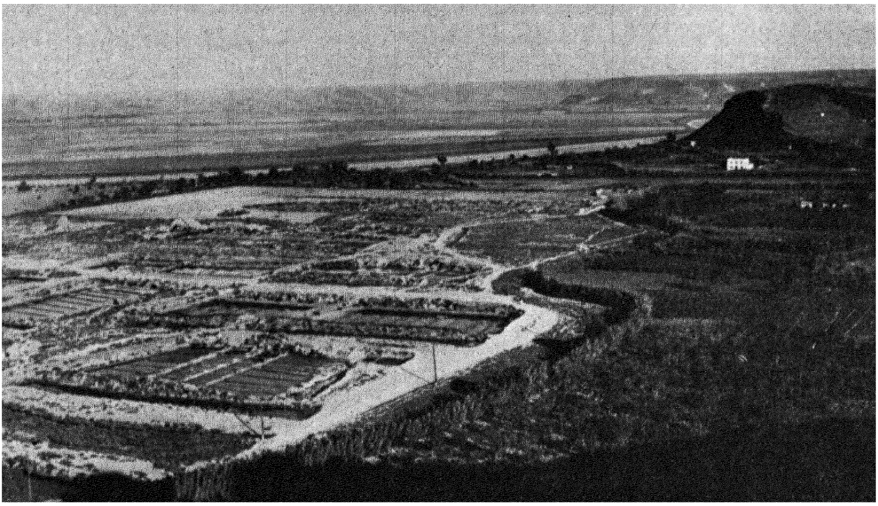


FIG. 96. THE SALT MARSHES OF BOU REGREG, UPSTREAM FROM RABAT, MOROCCO

These marshes are on the left bank of the river about 1¼ miles from the sea. An amphibious fauna, consisting mainly of numerous crabs, still flourishes. (Photograph: H. Termier.)

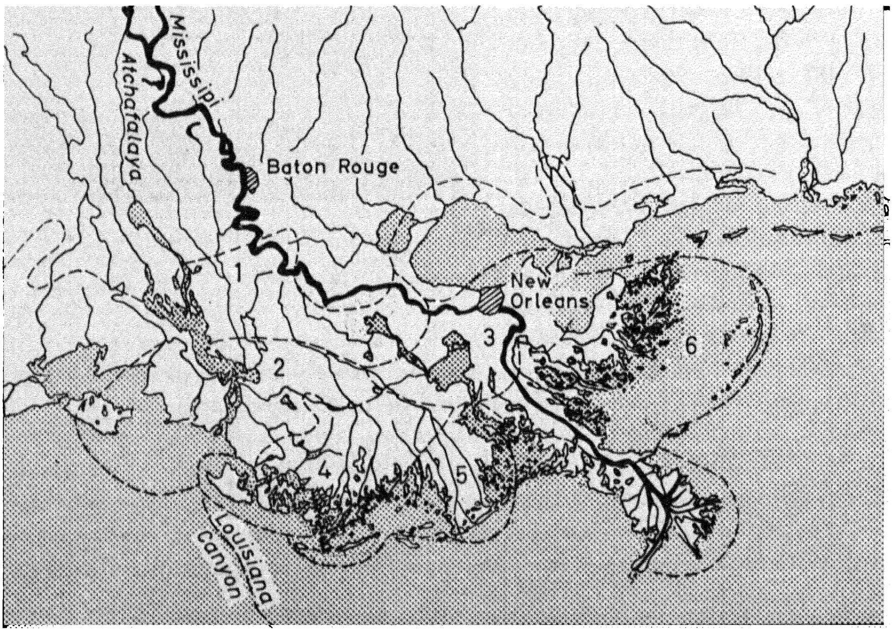
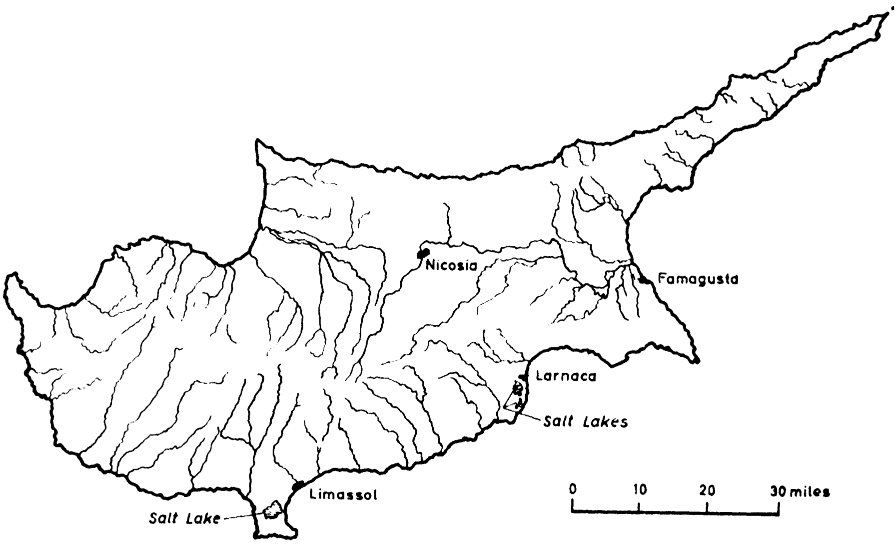
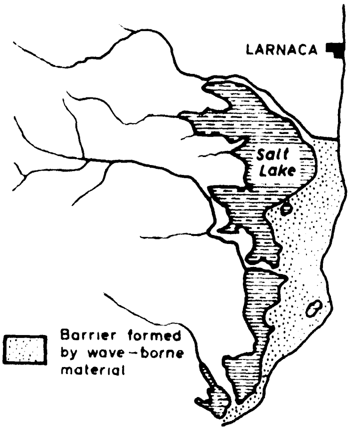


FIG. 97. THE DELTA OF THE MISSISSIPPI AND THE SUCCESSIVE STAGES OF ITS CONSTRUCTION

The inner broken line represents the position of the coast line about 5,000 years ago (after Scruton, 1955).

Similarly, on the coast of Cyprus, to the south of Larnaca (figs. 98 and 99) there is a salt lake that is separated from the sea by a loose barrier thrown up by the waves. This lake is not more than 3 feet deep, and its





FIGS. 98 and 99. THE SALT LAGOON OF LARNACA, CYPRUS

surface is 6 feet below that of the Mediterranean. It follows, therefore, that the sea water with a lower concentration of salt filters through the barrier. The rivers which flow into the lagoon of Larnaca are insufficient to compensate for the loss of water by evaporation and hence salt is deposited.

In Egypt, on the border of the desert, such lagoons are abundant. The salt lake of Mex, near Alexandria, is an example. There a coarse sand of sodium chloride grains with ripple-marks is deposited on a black clay at the rate of 3 to 5½ inches per year.

The shores of the Red Sea also have salt lagoons where salt is deposited and blown into dunes by the wind.

The Role of Beaches in Sedimentation throughout Geologic Time (figs. 100, 101 and 102)

For a long time it has been supposed that ripple-marks of marine origin could only be formed on the foreshore, that is, on that part of the coastal region between high and low tide limits. It is now known that they can occur at all depths. There is good evidence, however, that the largest surfaces covered by ripple-marks may be situated in shallow waters or on beaches.

Moreover, geological terrains with ripple-marks often show characteristics which indicate that they were formed in very shallow water, close to the land. In a number of cases they are associated with reddened detrital rocks. Also, there are commonly thick formations where the ripple-marks are localized at the surface of most of the beds. The Belt Series show the most famous examples. Typical examples have also been observed in the Carboniferous rocks of central Morocco (fig. 124, p. 213).

The sandy shores of the present-day coasts of Brittany are often covered by ripple-marks over great distances. Their pattern is accompanied by the traces of a wide variety of marine invertebrates; tracks, faecal pellets of burrowing organisms, funnel-shaped hollows dug by crabs (fig. 118, p. 210), small holes due to amphipods (*Talitrus*), etc., etc. It should be noted that this wide strand is only possible in such areas as Brittany, the Bay of Fundy and N.W. Australia, where, with 40–50-foot tides, the beach may extend several miles seaward at low tide.

The logical interpretation of great stratigraphic series characterized by ripple-marks thus seems to be, in more than one case, the existence of an ancient strand (or shore), that is, a coastal region affected by tides.

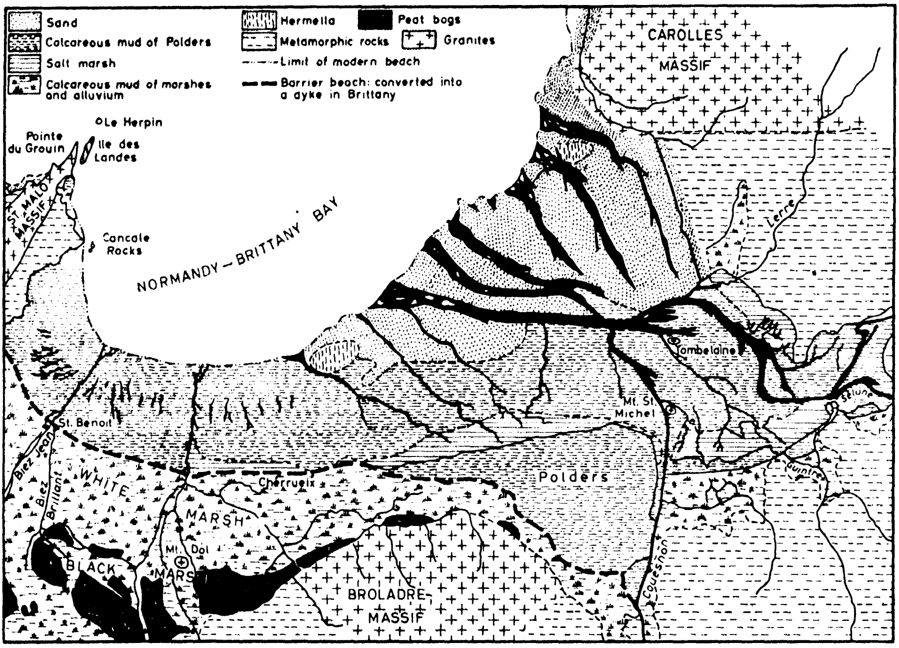
Formation, Sedimentation and Colonization of Basins on the Borders of Ancient Continental Shelves

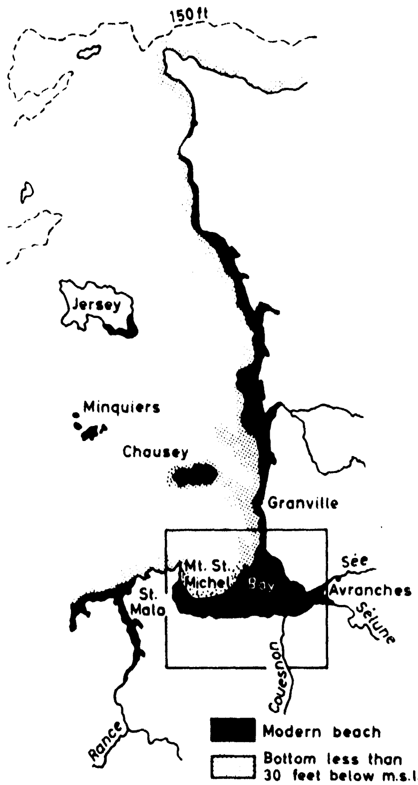
There are undoubted geomorphic similarities between the coasts of most continents as can be proved by a comparative examination of the Atlantic coast of Europe, the coast of Languedoc, the Gulf of Guinea, the Gulf of Mexico and the Gulf of California. The mouths of rivers invaded by the sea (C. Francis-Boeuf, 1947) are generally transformed into rias, often even into barred lagoons. This is especially apparent where the land is deeply dissected.

Sedimentation along such coasts is primarily due to the deposition of continental materials, although there is also a marine contribution in the from of reworked detrital sediments, facies constructed by organisms, and

FIG. 100. THE BAY OF MONT-SAINT-MICHEL, NORTHERN FRANCE, SEEN FROM THE MOUNT

(Photograph: G. Termier)





FIGS. 101 and 102. THE FLOOR OF THE BAY OF MONT-SAINT-MICHEL (after Phlipponeau, 1956) evaporites deposited in the lagoons. Moreover, the coastal vegetation, which includes seaweeds, marine grasses, and the vegetation of the forests and prairies, contributes organic material to the sediments being deposited. The bottoms of river mouths and lagoons are often stagnant, so that this

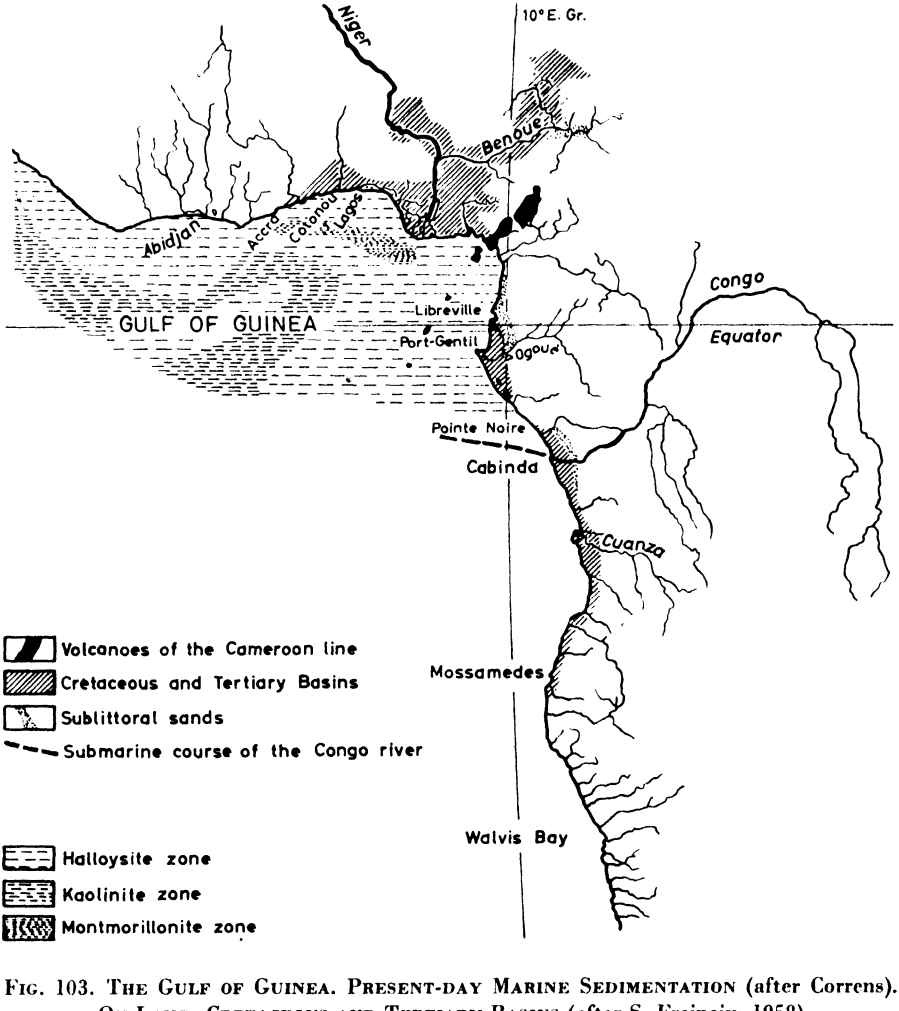


FIG. 103. THE GULF OF GUINEA. PRESENT-DAY MARINE SEDIMENTATION (after Correns). ON LAND: CRETACEOUS AND TERTIARY BASINS (after S. Freineix, 1958)

organic matter may retain part of its reducing characteristics, and give rise directly to hydrocarbons and carbon, and indirectly to sulfides.

In the geological column there are many examples which are comparable to this present-day environment: (1) almost all Permian outcrops, excepting those of the Tethys (in Transcaucasia); (2) all the paralic coal basins; (3) a number of Devonian deposits bordering the Old Red Sandstone continent, which are in part griotte limestones and in part alum shales; (4) a number of horizons of the Lower Cambrian where marine beds of shallow water type alternate with sandstones and redbeds.

Sedimentation in these basins therefore depends to a large extent on the pedologic evolution of the continents. In fact, in the paralic basins, the continental influence is at its maximum. These basins are generally fed by streams which have drained from the continents and which have transported soil particles eroded during their passage. It is in such basins that the theory of biorhexistasy (p. 153) can most legitimately be applied.

Finally, two paleobiological aspects of this subject should be noted. In the first place, it is in such basins that synchronous marine and continental fossils exist side by side, thus permitting the establishment of stratigraphical correlations. This is exemplified by the Westphalian in the Franco-Belgian basin and by the Rhaetic of Scania, southern Sweden. In the second instance, the faunas of contemporary basins, although they may be very close together, enjoy a relative independence since they are linked to rhythms which differ in detail for each basin. This may be due, in part, to variations in microclimate. As an example, the marine and lagoonal faunas of the Cretaceous in the Gulf of Guinea (Freineix) (fig. 103) may be noted. The ultimate closing, by a sand bar, of certain lagoons favored the geographical isolation of the plants and animals living there and allowed the phenomenon of speciation to operate.

10

Marine Detrital Sediments

PROCESSES AND STRUCTURES

There is little erosion on the sea floor outside the areas of the abrasion platforms and cliffs. This action is virtually linear since it only attacks the littoral zone and thus differs quantitatively from erosion by rainwash which acts on the whole of continental surfaces. But waves and currents roll pebbles and sand along with them and give rise to a characteristic facies. Moreover, the greater part of the detrital sediments derived from the land find their way into oceanic basins, where they become saturated with sea water, and thus chemical solutions, and are mixed with the organisms living therein. The modifications which they undergo are rapid since their generally fine grain size greatly enhances the surface area available for reaction with the marine agents.

Marine or freshwater muds may occur, in which the solid component is less than half (20–50%) of the total, the remainder being water. In the lacustrine molasse of Öhningen (Germany), some marly deposits still contain 41·3% of water, which results in numerous landslips in this plastic, semi-liquid material (Wegmann, 1955), which is nevertheless of Miocene age.

These characteristics make it clear that submarine sediments can behave sometimes as solids, sometimes as liquids, or more commonly as a mixture of the two.

Littoral detritus is thoroughly reworked by the sea. On the coast of Provence, for example, on the basis of their granulometric composition, J. J. Blanc (1958) distinguished the following representative types:

1. *Unsorted* sediments of varying size comprising the talus and screes at the foot of cliffs which have not been attacked by the surf and the waves.
2. Sediments *moderately sorted* forming beach deposits which have been subjected to the swash and backwash of wave action, and coastal currents.
3. *Well-sorted* sediments of creeks and bays in which occur the debris of marine organisms (e.g. pieces of the red alga *Jania*). It is on these deposits that Posidonias and Cymodaceae become established.

STRATIFICATION

Sedimentary rocks are deposited mainly in horizontal beds. For this, they behave *as liquids*. Exceptions to this rule become more frequent as the components become coarser. However, since mass is the principal factor in deposition, the ideal tendency is toward horizontality.

Among the exceptions may be quoted *cross-bedding* of continental and marine deposits, which principally affects sands, gravels and conglomerates. In particular, it occurs in *lenticular deposits.* These take the form of the topography at the place of sedimentation. For example, the talus cones of streams become conglomerate lenses after burial under later formations, and a bioherm forms a lenticular or conical mass. *False unconformities* and *intraformational unconformities* can often be explained by the effect of "creep" (see p. 77).

The horizontality of beds is favored by that of the floor, but it is not always thus and the surface of the floor may slope or be deeply gullied. Furthermore, it may be in an orogenic zone or on a stable continental area. Deltaic areas are an example of deposition on an inclined, moving floor.

Stratification joints (bedding planes) result from the presence of clay or marl beds limiting banks of sandstone or limestone. They seem to result from those "imponderable" particles (Lombard) which remain in suspension just above the bottom and which are kept moving by the currents which have deposited them. Stratification joints (bedding planes) correspond to periods of complete calm.

On the borders of basins, horizontality of beds does not occur, and they terminate in a "feather edge".

The thickness of strata varies with the series: V. C. Kelley (1956) has defined the *stratification index* as the number of beds multiplied by 100, divided by the thickness.

Features of Stratification

Cross-bedding.—This is the most common feature of sediments and results from deposition by currents, of either wind or water. Fluviatile or deltaic sediments, as well as sand dunes show this characteristic type of stratification in which the inclination of the beds may reach 33°, under eolian conditions, but rarely exceeds 25° when water-laid. The slope of the beds descends in the direction of the current.

Several varieties of cross-bedding can be distinguished. The *tabular type* is that in which the beds are horizontal and the stratification is oblique. It is characteristic of deposits formed by water currents both in rivers and in the sea (figs. 104–106). A variety of this type is "herringbone" stratification which is due to marine currents of inconsistent direction. Thus, in certain beds the slope is reversed. The structure of the Pliocene beds on the route from Algiers to Birmandreis is believed to be of this type (figs. 107 and 108).

There also occurs a form of stratification known as *lenticular* cross-bedding, which is also associated with aquatic deposits. This type takes the form of a meniscus, without a trace of horizontality and is often found in alluvial deltas.

In the case of dune deposits the structure is sometimes *tabular*, but more often the cross-bedding takes the form of wedge-shaped segments (*wedge-bedding*).

Submarine Sliding (Slumping).—When soft sediment is deposited on an inclined sea floor, particularly in regions liable to seismic shocks, irregularities of stratification may be caused by slumping (Fairbridge, 1946). The

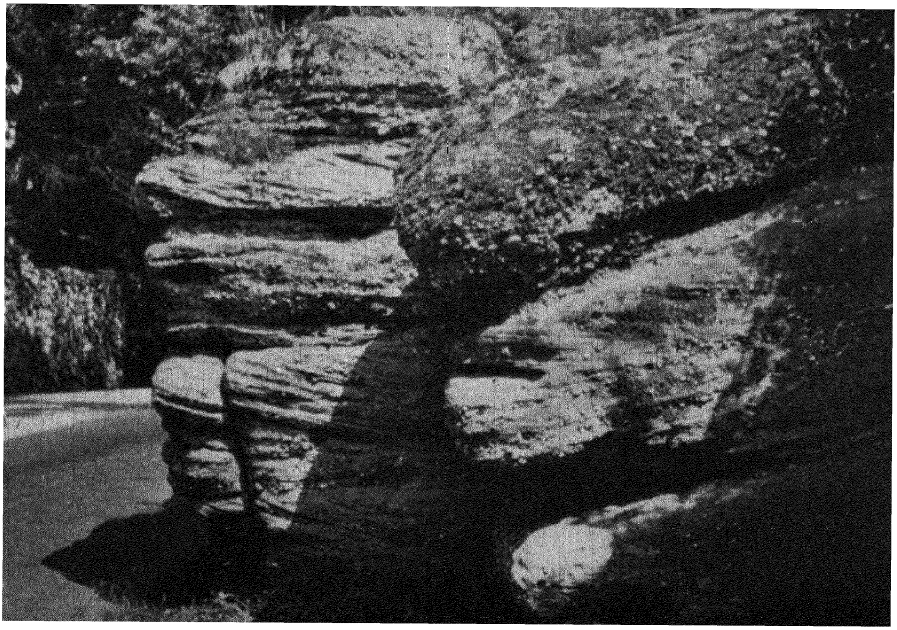


FIG. 104. COARSE SANDSTONES OF THE TRIASSIC (BUNTERSANDSTEIN) OF MT. ODILE, VOSGES, FRANCE

A typical example of tabular cross-bedding. (Photograph: G. Termier.)

viscosity of the sediment allows it to retain, in part, the structure due to its movement, and thus, small folds are formed (fig. 109).

Penecontemporaneous Deformation of Deposits.—This occurs in areas where abundant sediment is laid down and earth movements are frequent, as in depositional basins in orogenic zones (in the foredeeps of geosynclines and in intermontane basins). It also occurs in subsiding areas in mountainous regions.

A well-known example is in the region of the Isthmus of Corinth, where a great thickness of Neogene sediments has accumulated and is subject to frequent earthquakes which are clearly associated with the intermittent

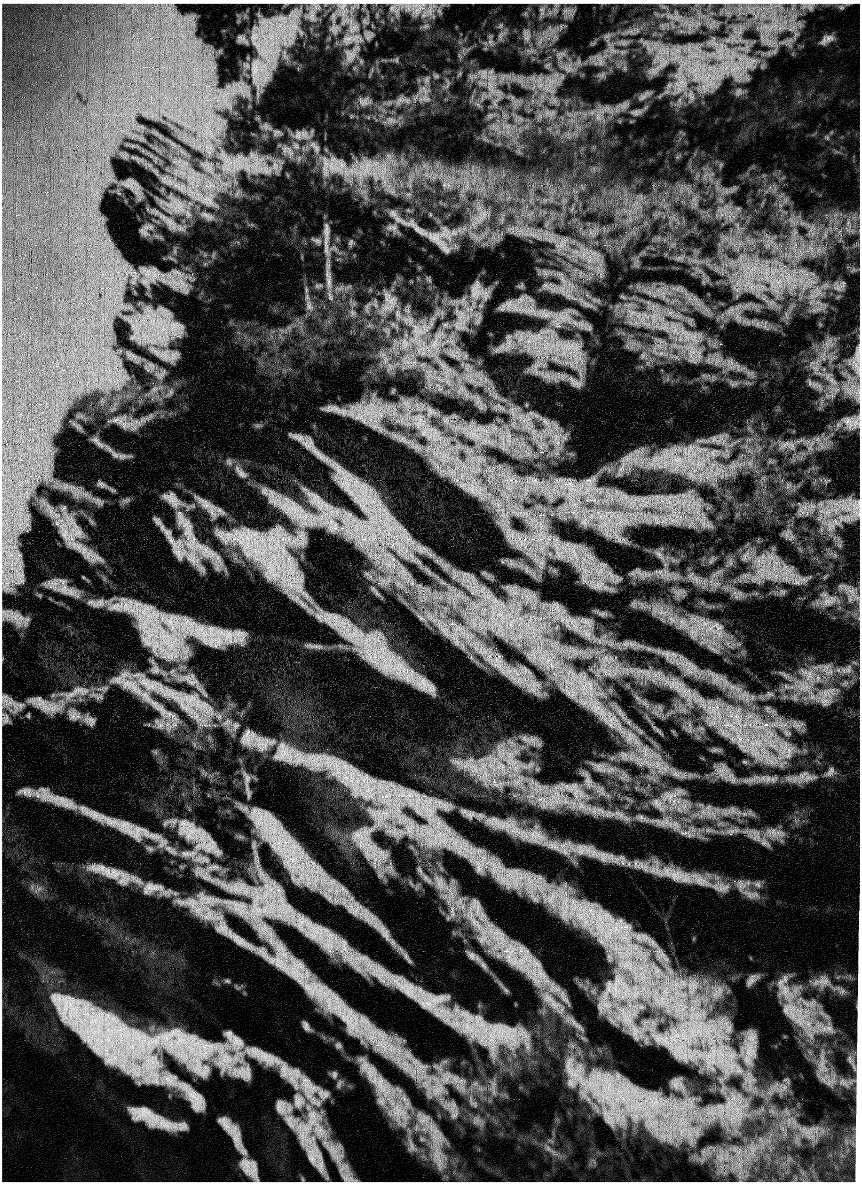


FIG. 105. TABULAR CROSS-BEDDING IN A CALCAREOUS SANDSTONE
Valley of Saint-Martin du Vercors, Southern France. (Photograph: G. Termier.)

sinking of the Aegean Sea. The horizontal beds are consequently broken by a great number of faults which are well exposed in the walls of the Corinth Canal (figs. 110 and 111). This phenomenon is very common in subsiding sediments.

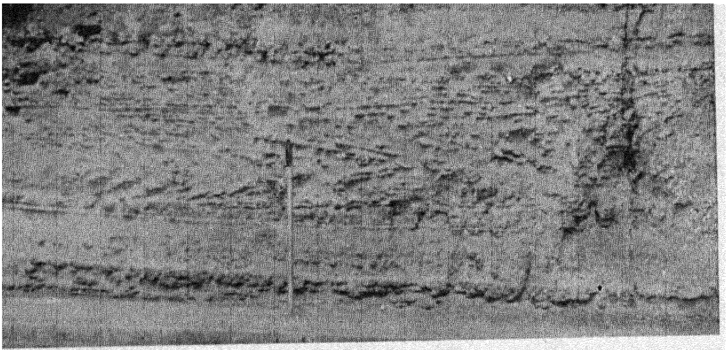
Penecontemporaneous deformation can also cause fissures which are often nearly vertical and are accompanied by more or less prominent off-

FIG. 106. TABULAR CROSS-BEDDING IN A MARINE LIMESTONE HORIZON OF URGONIAN AGE, JURA MOUNTAINS, FRANCE

The current came from the left. (Photograph: G. Termier.)

setting of the strata. These fissures are filled by sediment coming from beds higher in the succession (autocicatrization, "clastic dikes", Pruvost, p. 143, 1954) (fig. 112).

Tepee Structures.—In the Guadalupian (Middle Permian) of the Guadalupe Mountains, symmetrical folds which may reach 30 feet in height have been described (Newell, Rigby, Fischer, Whiteman, Hickox and Bradley, 1953). These folds rest on horizontal beds and their crests are buried in horizontal sediments. These structures have been named *tepees* because their shapes resemble Indian wigwams. The materials forming these *tepees* are bedded dolomites formed on the continental shelf behind the reef (back-reef facies) (pp. 245–247). The explanation proposed by the authors cited above, is a kind of diapirism of thin gypsum beds which have subsequently been leached out along joints which follow the axes of these diapirs. The growth of a crust of gypsum or caliche as a result of recent weathering has given rise to comparable structures in the sandstones of the Delaware Basin.



Figs. 107 and 108. Cross-bedding of "Herringbone" Pattern, in Pliocene Sediments. Between Algiers and Birmandreis, North Africa

In fig. 107, the upper horizontal beds cut across the inclined beds, whereas these beds pass asymptotically into the lower horizon. The beds have not been markedly affected by folding. (Photograph: P. Muraour.)

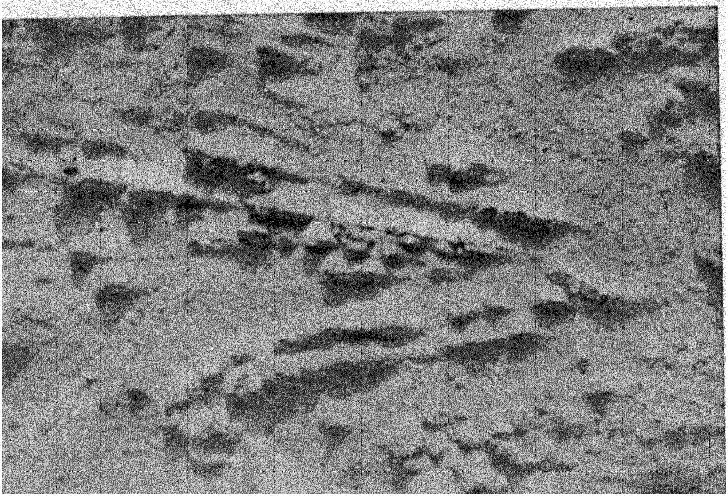


Fig. 108. Close Up View of the Same

Since the beds above and below are horizontal, it appears that the folding has been produced during diagenesis (fig. 113).

Diagenetic Deformation of Evaporite Beds.—The stratification of beds of soluble salts such as NaCl or $CaSO_4$ always raises problems related to the

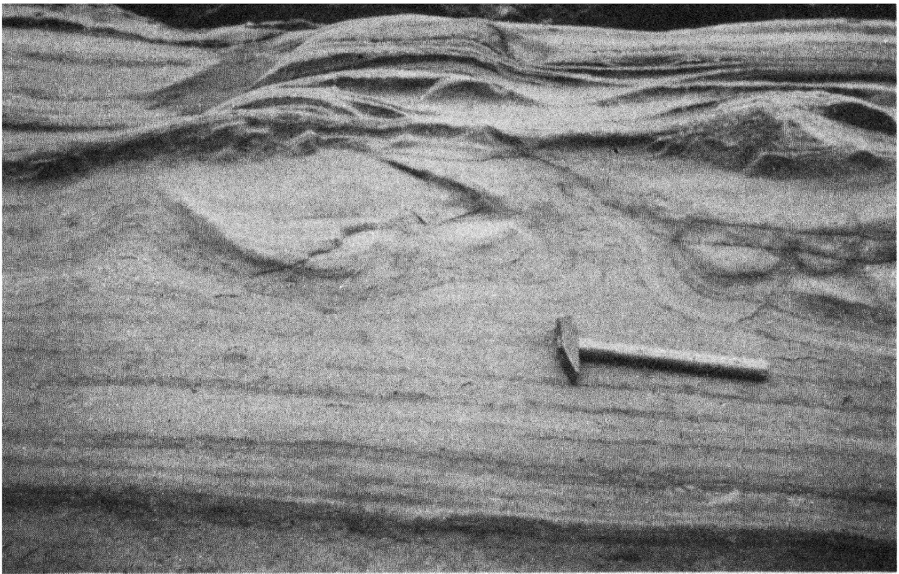


FIG. 109. BEDDING DISTURBED BY SUBMARINE SLUMPING (SLIDING) DURING DEPOSITION
Oligocene sandstone of Dellys, Algeria. (Photograph: G. Termier.)

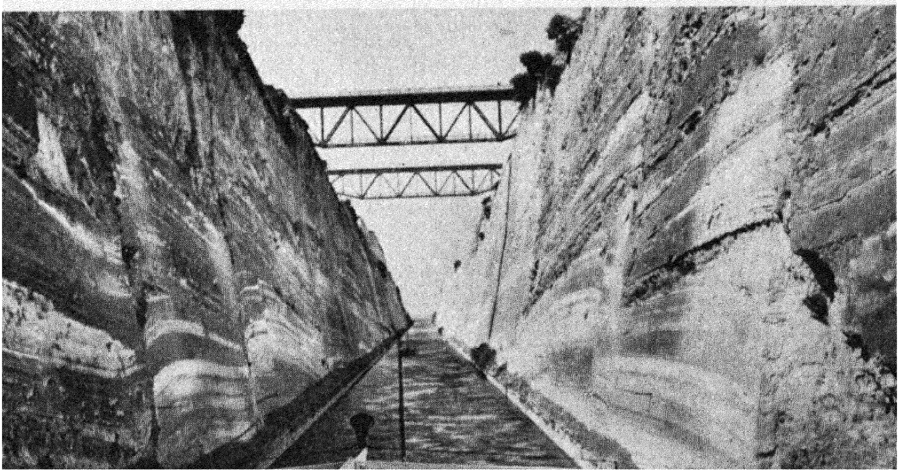


FIG. 110. THE ARTIFICIAL CUT OF THE CORINTH CANAL, GREECE
Note the numerous, late Quaternary, faults exposed in the walls

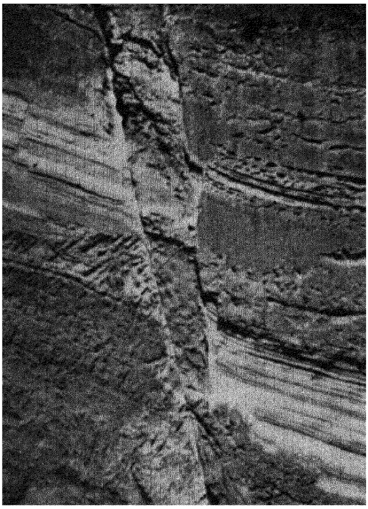
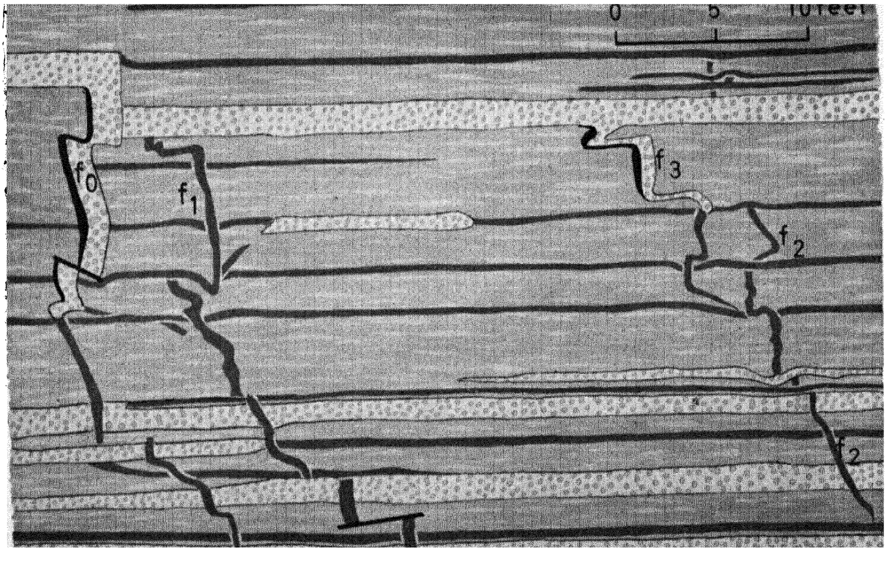


Fig. 111. Corinth Canal
Detail of a fault.



Sandstone Argillaceous sandstone Clay

Fig. 112. Intraformational "Clastic Dikes"

Section about 1¼ miles from Babouch, between there and Tabarka, Tunisia (after C. Gottis).

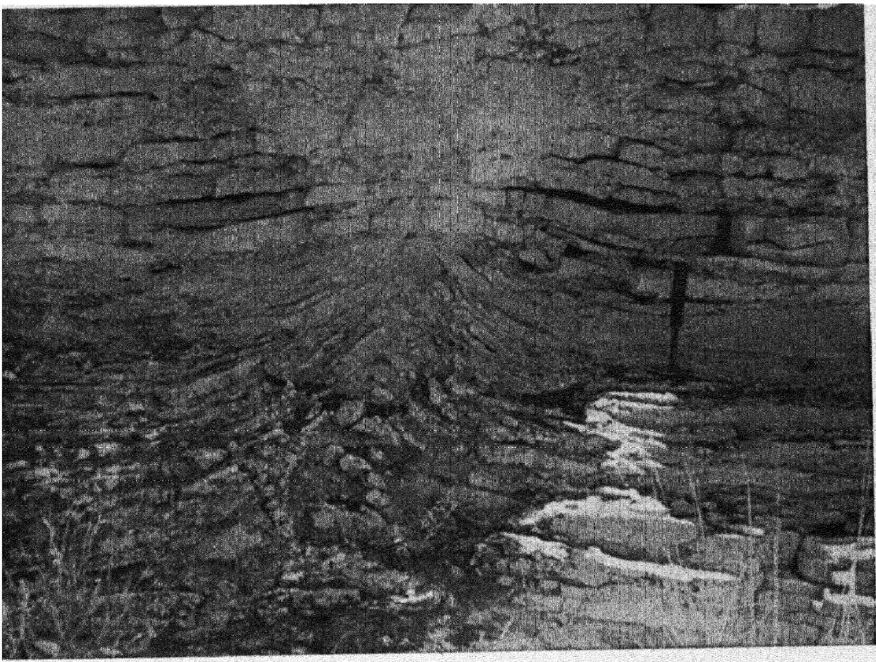


Fig. 113. A Typical Example of "Tepee" Structure. Permian Dolomitic Limestone (Grayburg-Queen Sequence). South Tank Canyon, in the Center of the Guadalupe Mountains, New Mexico (Photograph: Donald W. Boyd)

Fig. 114. Puckering of a Bed of Kieserite ($MgSO_4.H_2O$) in a Bed of Carnallite between Two Layers of Rock Salt. Potash Mine in the Triassic Zechstein of Heringen, Germany (Photograph: Wintershall)

maleability and plasticity of these rocks. This plasticity is particularly marked in the crumpling of certain beds which expand during recrystallization (fig. 114). The transformation of anhydrite ($CaSO_4$) into gypsum ($CaSO_4.2H_2O$) takes place with an increase in volume. It follows therefore, that deformation will occur simultaneously with diagenesis, giving rise to domes or folds (fig. 115). It has been called "endogenic folding" by Grabau.

It is possible that such beds have also undergone some deformation during their deposition and have retained structures attributable to submarine sliding (see above). Von Gaertner (1932) attributes the formation of alabaster nodules in beds of anhydrite or slightly dolomitized gypsum in the southern Harz mountains to this mechanism (fig. 116).

Density Currents

A *density current* is a mixture of liquids or gases, having a uniform density different from that of the normal environment. This occurs where rivers enter the sea and the fresh water floats on the salt water. The density may be intermediate between that of the surface water and that of the bottom water, in which case the current occurs between the two and constitutes an "interflow". Finally the density of the current may be greater than that of the environment; the current then flows along the bottom as an "underflow" or "bottom" flow.

Turbidity currents (Forrel, 1885) are a particular case in which the high density is due to material in suspension. Two examples of density currents in the atmosphere are nuée ardente and dust drifts. One of the best known aqueous examples is the turbidity current of Lake Meade (the reservoir lake behind the Hoover Dam) fed by the Colorado and Virgin Rivers (H. R. Gould, 1951).

In a river, a turbidity current is formed when only suspended material is present. Flocculated sediment and also coarser material on the bottom constitute a density current. The subaquatic slopes of the deltas of the Rhine and the Rhone, where these rivers cross Alpine lakes, show trenches cut by density currents and subaqueous levees similar to those of subaerial deltas.

In the oceans, the existence of turbidity currents has to be presumed. Evidence of such a current on the Grand Banks of Newfoundland has been given by Ewing and others (fig. 117). The rapid transport into the sea of large volumes of sediment from rivers, from torrential rain storms or from volcanic eruptions, or the agitation of the bottom by storms, marine currents or tsunamis, or even mass movements of the sea bed, are liable to cause slumping and the flow of muds down submarine canyons. The presence of coarse detritus on the floors of the oceans, moreover, can only be explained by means of such currents.

These turbidity currents obey the same laws as streams, so far as the

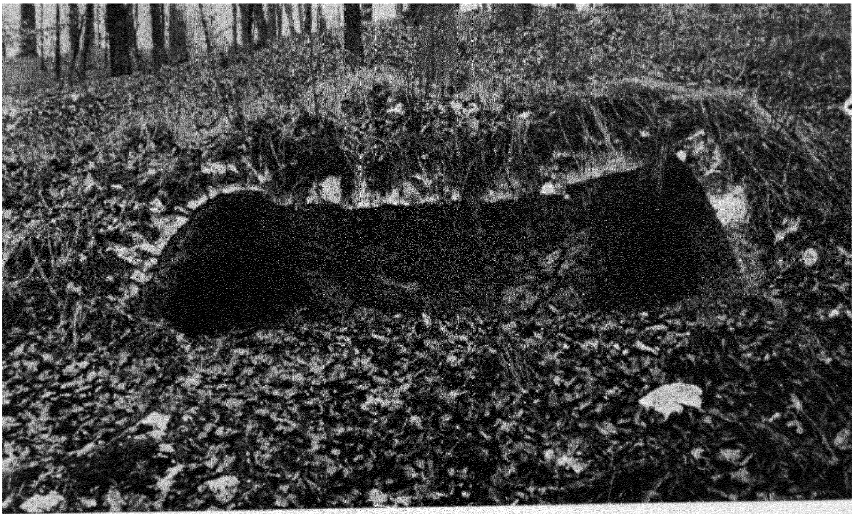


Fig. 115. A Dome of Gypsum formed by Enterolithic Folding (Grabau's Term) during the Conversion of Anhydrite to Gypsum. Triassic Zechstein (Werra Anhydrite). Southern Harz Mountains, Germany (Photograph: von Gaertner)

Fig. 116. Nodules of Alabaster in a Bed of Dark Anhydrite, or Slightly Dolomitized Gypsum. Triassic Zechstein (Werra Series). Southern Harz Mountains, Germany (Photograph: von Gaertner)

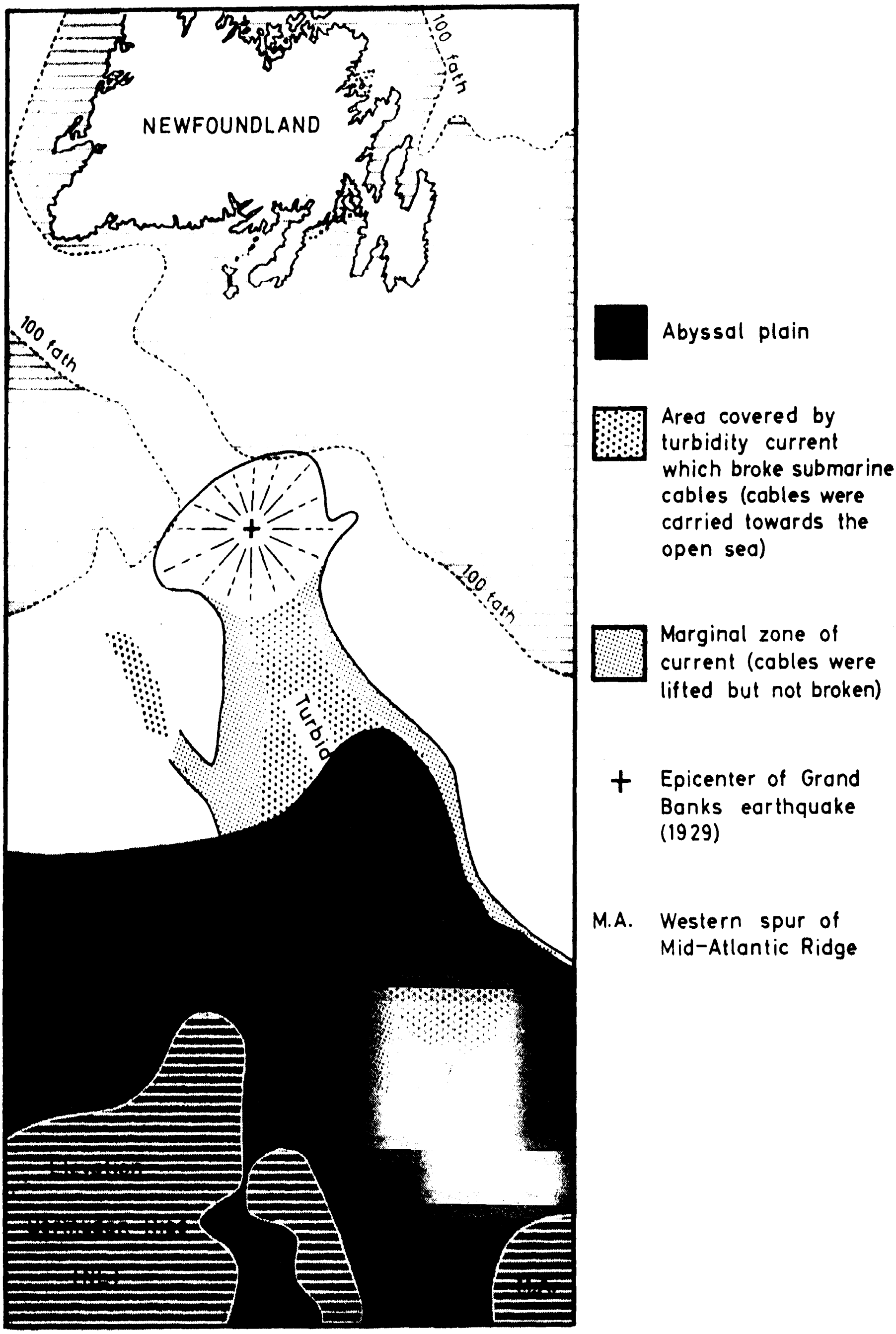


FIG. 117. THE TURBIDITY CURRENT OF 1929, FROM THE GRAND BANKS, NEWFOUNDLAND (after Heezen, Ericson and Ewing, 1954)

deposition of sediment and the channeling of flow are concerned. It is difficult to define their base level since it may be the bottom of the deep sea or a local level represented by a layer of water of the same density as the turbidity current. Thus it may be supposed that turbidity currents run toward the foot of the continental slope and erode it because they are situated high above their base level represented by the abyssal plains (see p. 56).

According to Shepard, the average slope of the continental shelf is 0° 07′, while that of the continental slope is 4° 17′ and that of submarine canyons is 2° 30′. It is necessary to bear these figures in mind when considering the continental shelf and slope because many diagrams grossly exaggerate the vertical scale and hence the angle of slopes.

Dredging in submarine canyons shows that their sides may be cut in bare rock whereas the bottom is covered with sand and mud. In the submarine canyons of Provence, J. J. Blanc (1958) described very fine-grained fluid muds quite distinct from surface muds. If, as Daly and Kuenen suppose, the canyons are due to abrasion by turbidity currents and are cut more slowly than a river would erode, the absence of coarse sediments seems to indicate that they have now reached their profile of equilibrium and are no longer being eroded.

The speed of a turbidity current is proportional to the square root of its effective density. The lower part, more heavily loaded, thus travels faster than the upper part. It follows, therefore, that the grain size of the sediment decreases from bottom to top: this is one mode of origin of "graded-bedding". The study of such sediments has been undertaken by Kuenen and Migliorini (1950) and later by Kuenen and Menard (1952). The superposition of several layers, each showing graded bedding, is one of the most typical examples of a sedimentary rhythm (see p. 354). Each sequence is normally made up as follows: (1) coarse-grained detrital sediment, (2) medium-grained detrital sediment, (3) fine-grained detrital sediment. It may be continued by colloidal and then calcareous layers. Sometimes the sequence may be incomplete, lacking either the coarse material at the base or the fine-grained upper layer.

The graywackes (in the German sense—see Glossary), such as the conglomerates and sandstones of the "Culm" in the Harz mountains, contain typical examples of graded beds. From the Archean to the Tertiary, and particularly in geosynclines, such sequences are, however, comparatively rare. The graywackes and similar formations constitute only 15–20% of the sandstones in the geosynclinal zones of, for example, the Apennines (Migliorini). These graywackes, *which must be deposited in deep water* (Bailey, 1930), consist of grains of angular sand set in an argillaceous groundmass, sometimes with boulders or fragments (pseudo-tillites) not unlike the material of glacial moraines. They cannot be confused with fluvial formations because they only show minor cross-bedding, and often alternate with Radiolarian cherts. Moreover, they have commonly suffered considerable deformation which occurred contemporaneously with their deposition: intraformational disturbances in general, distortion of bedding, slumping and various other structures ("pull-apart", "crinkling", "dented-bedding", "clay pebbles").

The variations and complications in the effects of turbidity currents and their deposits are due to several causes: the stratification of ocean waters, and the presence of "interflows" and "underflows". The transport of sand in a turbidity current formed from an argillaceous suspension does not increase the density, but causes a reduction in mobility and suppresses turbulence. The high salinity of sea water causes flocculation of the clay which explains how the viscosity and the thixotropy (physical properties of argillaceous suspensions) can be modified by salinity changes. There is, however, little difference between turbidity currents in sea water and in fresh water. In a lake the suspensions tend simply to fill in depressions, whereas in the sea they tend to form a deposit of uniform slope whose inclination is so small that the turbulence no longer hinders the fall of the clastic and flocculated material to the bottom.

It is thus possible to recognize the existence of mudflows which at the present time and in the past have occurred on the broad gentle slopes of the continental shelf, and in particular, mudflows which have been accentuated by their association with orogenic zones. Such flows explain the existence, often observed in deep waters, of coarse, well-sorted sediments, and the remnants of terrestrial vegetation and organisms, such as foraminifers, normally found near the coast or in shallow water (Phleger, 1951; Heezen, Ewing, Menzies, 1955). A good example of this occurs off the north coast of California where La Jolla canyon opens on to the San Diego deep by way of a delta fan at a depth of 2,000 to 3,000 feet. Sediments are carried to the San Diego deep by mass movement in the form of a turbidity current moving like a river and flowing across the delta. The displacement, en masse, by sliding, is thought by Shepard (1951) to have been favored by the accumulation of marine grasses which gave some cohesion to the sediments. It is probable that earthquake shocks play an important part in initiating submarine mudflows and it is likely that such processes materially assisted in the filling of oceanic basins (Heezen, 1954).

The behavior of turbidity currents raises the problem of geomorphic sculpture of continental shelves. At the margin of a geosyncline having a *narrow* continental shelf which is terminated by a steep slope and which is subject only to weak wave action and negligible subsidence, it seems likely that submarine slumping will form steeply inclined terraces comparable to those of lacustrine deltas where slumping takes place over a broad front without the formation of canyons. When, however, the continental shelf is *broad*, there is a tendency for subsidence to occur under the weight of accumulating sediment and there is a similarity to great deltas with gentle slopes where very little slumping normally takes place.

It seems possible that, in the second case, turbidity currents were initiated at the edge of the shelf at the end of glacial phases. It may be that the canyons observable today may have been largely produced in this way during the Pleistocene. This explanation, put forward by Kuenen (1950), is closely linked to the hypothesis of glacial-eustatism. It has been shown (p. 48) that this hypothesis of subaerial erosion is highly probable since it appears that a large part of the continental shelf was above sea level during the glacial phases of the Pleistocene.

GRAVITY SLIDES (OLISTOSTROMES AND OLISTOLITHS)

Olistostromes (from the Greek ολισταινω = to slide, στρωμα = accumulation) are sediments formed by accumulations resulting from sliding. Large exotic or erratic masses may be present within the accumulation and are termed *olistoliths*. Large exotic accumulations of this sort in the southwestern United States are known as "chaos".

Examples from Sicily and Italy.—Examples of olistostromes are fairly numerous and the "argille scagliose" (P. Bianconi, 1840) of the Apennines and of Sicily (Flores *in* Beneo, 1956*a*; Beneo, 1956*b*) may be cited as an example.

In Sicily (Beneo 1956*a*) the search for petroleum by geophysical prospecting has shown the existence, in the center and south of the island, in prolongation of the plain of Catania, of a trench filled with argillaceous materials more than 25,000 feet thick. The structure of these plastic masses is very heterogeneous. They contain "flakes" of limestone, sandstone and eruptive rocks, and show definite evidence of submarine sliding and turbidity flow. These are argillaceous breccias and "argille scagliose". From the stratigraphic point of view, the "allochthonous" character of this formation explains why it is that this rock, formed mainly during the Tertiary and Quaternary, contains rock fragments of all ages. Among these fragments are the famous boulders of Sosio, which contain one of the best marine Permian deposits in the world. They consist of several great masses of limestone resting on a Triassic argillaceous limestone, which in turn rests on Miocene clays (D. Napoli Alliata, 1953). Normal sediments are intercalated with this disordered series.

In the central Apennines (Beneo, 1956) the chaotic "argille scagliose" and the Flysch of *Pontian* (Upper Miocene) age, are similar in type to those of Sicily. This is a polygenetic assemblage, whose argillaceous matrix contains a microfauna indicating ages of Upper Cretaceous to Miocene. The rocks contain blocks of manganese shales or sandstones from the Oligocene, granular limestones and breccias with large reworked foraminifers, "scaglia rossa" with *Globotruncana stuarti* from the Campanian-Maestrichtian, and also serpentinized rocks.

Some Other Examples of Gravity Slides

Gravity slide deposits are found in almost all parts of the world, but only in recent years have been recognized as such. Some of the earliest to be reported are in the Cretaceous of the Carpathians, in the Jurassic of Scotland, and in the Ordovician of Quebec. Some examples, familiar to the authors, are noted below.

In the region of *Dechra Aït Abdallah* (central Morocco), the coral limestone of the Middle Devonian is incorporated in sediments of Strunian age. The Givetian reefs are preserved only in this manner (J. Agard, P. Morin, H. Termier and G. Termier, 1955). In *Chaouia Sud*, the unsorted conglomerates of *Biar Setla* form a band continuing for more than a mile and attaining a thickness of 650 feet. This then splits up into a train of coral limestone blocks, and thereafter thins and disappears. These blocks could easily be mistaken for lenses contemporaneous with the enclosing shale (H. Termier, 1936, p. 397).

At *Ben-Zireg* near Colomb-Béchar (Southern Algeria) banks of Devonian rocks are enclosed in the Visean (Pareyn, 1955). They form a *Wildflysch*, resting on Fammenian limestones. The succession begins with shales and a ferruginous layer which appears to have resulted from pedogenesis (soil formation). Above this is the Flysch containing mudflows full of large blocks, in which the stratigraphic succession is inverted. The series ends with a well-bedded sandy shale, without boulders.

In the massif of *Djurdjura* (Algeria) to the northwest of the Haizer on the south flank of Djemaa bou Sero, the Oligocene conglomerates contain "enormous masses of Liassic limestone which simulate true outcrops and which correspond to cliff faces falling into the sea" (Flandrin, 1948).

Olistostromes are similar to the "sedimentary klippes" described by P. Lamare (1947, 1948 and 1950) in the *Mendibelza massif*. There, gigantic exotic blocks are included in the polygenetic "puddingstone" of the Albian (Lamare, 1947) or of the Cenomanian (Casteras, 1952).

In *northern Syria* large blocks of Carboniferous and other Paleozoic sediments, up to several hundred feet in length were first reported by Dubertret. They were mapped by Fairbridge and Badoux, who discovered that they were totally enveloped by Upper Cretaceous chalks. Apparently there was an important E.–W. fault along the northern border of Syria, which was active during much of Upper Cretaceous time. Progressively more and more blocks became detached from the submarine escarpment and then slid down a steep slope into deep water. They are also associated with slump structures.

In *western Georgia* (U.S.S.R.), M. F. Dzvelaia (1954) described argillaceous breccias of the Sarmatian, enveloping at many places blocks of Upper Cretaceous and Eocene limestone, which may be more than 30 feet long and which were derived from a collapsing cliff.

In the *Crimea* the position of the fossiliferous Permian limestones is similar to the Permian of Sesio.

Finally, in *western Venezuela*, Renz, Lakeman and van der Meulen (1955) have also described a chaotic sedimentary series in the Lower Tertiary (Eocene and Oligocene) succession.

According to Beneo (1956*b*) olistostromes are formed by submarine slumping and turbidity currents. The coarsest material slides from its initial position and the finest particles are redeposited in stratified form. In the case of sediments such as the "argille scagliose" which are associated with the uplift of the Apennines, there is no doubt that the olistostromes are a direct consequence of the elevation of the folds. They thus correspond to orogenic sedimentation of the "flysch" type, a phenomenon which progressively affects larger areas as uplift continues. This explanation seems adaptable to most of the examples quoted. Among the numerous types of sediment to which the name "flysch" has been given, it is essentially the "Culm" and "graded-bedded" graywackes of Germany (Harz) and the "wildflysch" of the Alps, which appear to correspond with the olistostromes.

The olistoliths are blocks of indurated rock, chiefly limestone or reef limestone, or greenstones of eruptive origin. Their great resistance allows erosion to demarcate and isolate them from softer rocks which are carried away more rapidly by waves and currents. Once separated, these hard rocks become vulnerable and can be broken down. Bioherms (p. 257) which are often of comparatively small dimensions may form giant blocks and become incorporated in olistostromes.

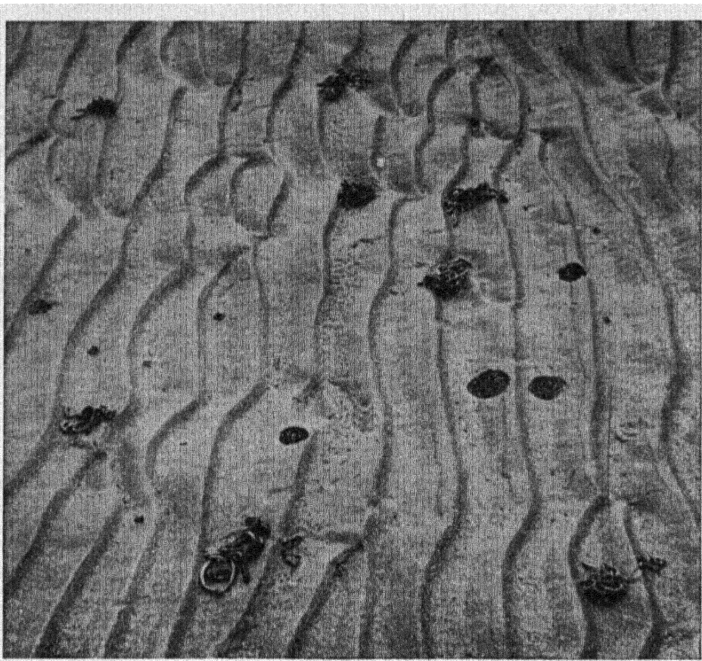
SOME TYPICAL STRUCTURES

Detrital sediments are the result of the transport and deposition of rock fragments which have been detached from the surface of the earth by erosion. The mechanism of sedimentation and the agents of erosion and transport have already been discussed. It must be stressed here that features resulting from transport and deposition can be preserved in sediments and so give the rock a "label" of origin.

Ripple-marks (figs. 118 to 124)

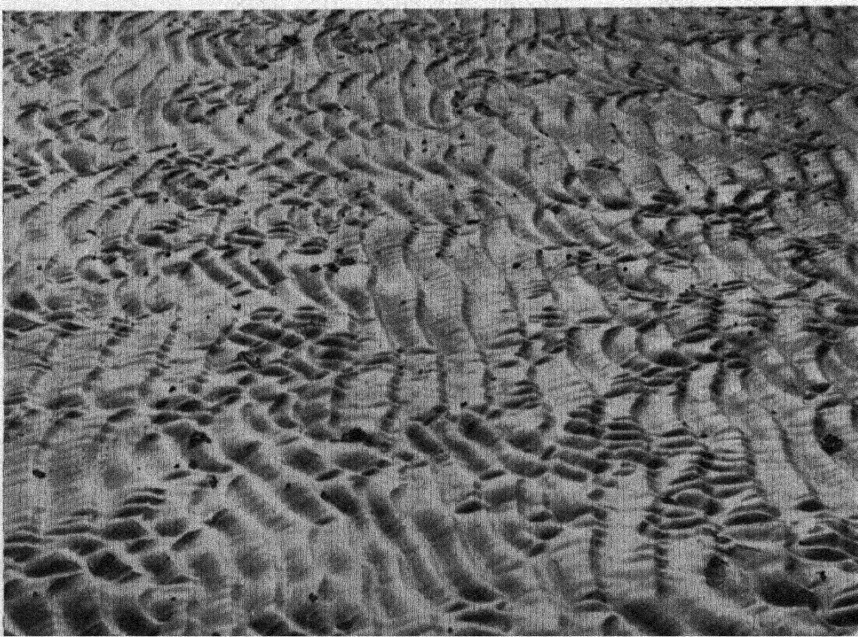
Ripple-marks are periodic structures of undulatory type formed by parallel crests, regularly spaced on the upper face of beds which were originally sandy. They are formed on the surface of desert sands, on fluviatile deposits, marine strands, and also on marine silty muds at all depths. Their dimensions vary considerably. These features indicate that the environment of deposition has been subjected to a current of air or water and they probably represent a regular succession of "swells" and "nodes". There are complex ripple-marks formed where the original currents came from varying directions (figs. 119 and 122).

Ripple-marks formed by currents of water or air are asymmetrical, and



Figs. 118 and 119. Marine Ripple-marks on the Beach at Saint-Efflam, Brittany

Note the complex type of ripple-marks in fig. 119, where two systems interfere. In fig. 118 casts formed by *Arenicola* worms, the funnel-shaped apertures made by crabs and the holes made by a small amphipod can be seen. (Photographs: G. Termier.)



can be distinguished from those patterns which are symmetrical (Kindle, 1917), and are related to simple wave oscillation pressure, generally in waters of intermediate depth on the continental shelf or in lakes.

In the fossil state, eolian ripples are relatively rare since they tend to be transient and have little chance of being buried under a subsequent layer. On the other hand, it is common to find traces of fluviatile and, especially, marine ripple-marks. The latter, which seem to be the most

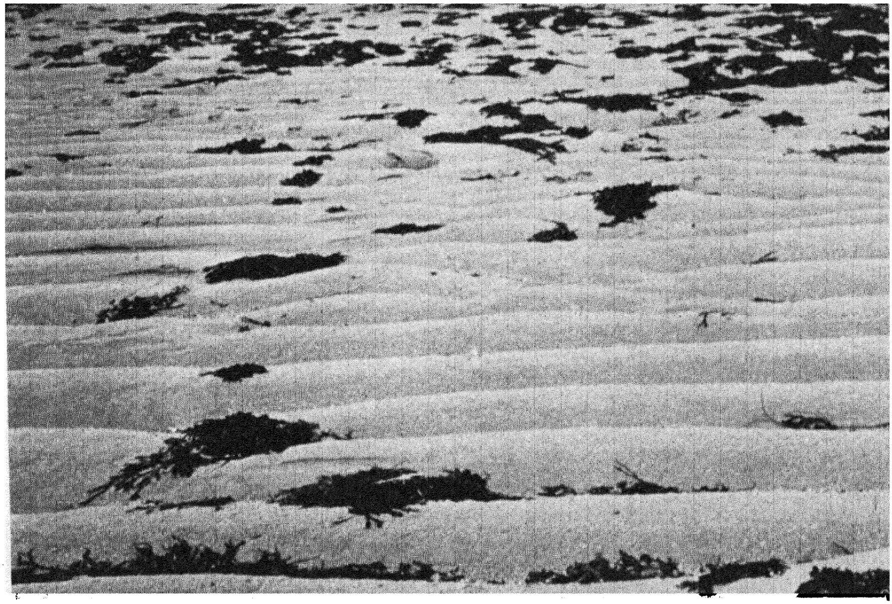


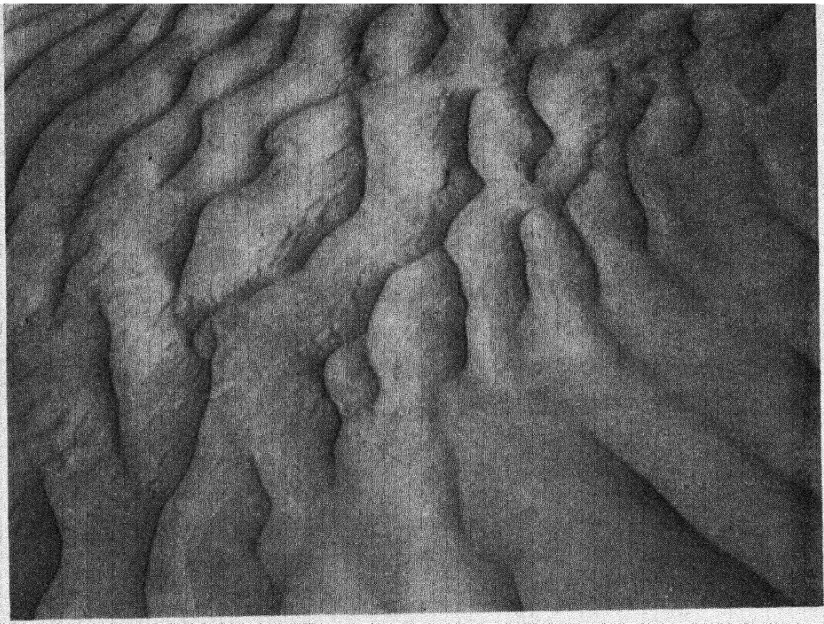
FIG. 120. LARGE RIPPLE-MARKS ON THE EAST COAST OF THE ISLAND OF OLERON, ATLANTIC COAST OF FRANCE (Photograph: G. Termier)

The *Fucus* sea-weed indicates the scale.

common, often show impressions due to crawling or burrowing organisms (burrows, wisps of excrement, etc.). Some sedimentary series are ripple-marked on the upper surface of all the beds. This suggests a quasi-seasonal periodicity such as the alternation of deposition and reworking on the tops of ridges. This may also happen in the case of very large beaches which are uncovered for part of the year.

Rill-marks (figs. 125-126)

Tidal currents which occur during the retreat of the sea from a beach form a pattern of fine channels, particularly where the water is retarded by obstacles, pebbles or shells. These channels or *rill-marks* formed on the surface of moist, soft sand can be preserved by fossilization. These markings



Figs. 121 and 122. Complex Ripple-marks on the Surface of Dunes in the Region of El Goleah, Algeria (Photographs: G. Termier)

FIG. 123. NEAR ABLEDID, MOROCCO. SURFACE OF A SLAB OF MIDDLE JURASSIC SANDSTONE WITH RIPPLE-MARKS (Photograph: H. Termier)

FIG. 124. FOSSIL RIPPLE-MARKS ON THE SURFACE OF A MISSISSIPPIAN (VISEAN) SANDY SHALE AT KOUDIA SIDI YEDDINE, CENTRAL MOROCCO, 6 MILES WEST OF AZROU (Photograph: G. Termier)

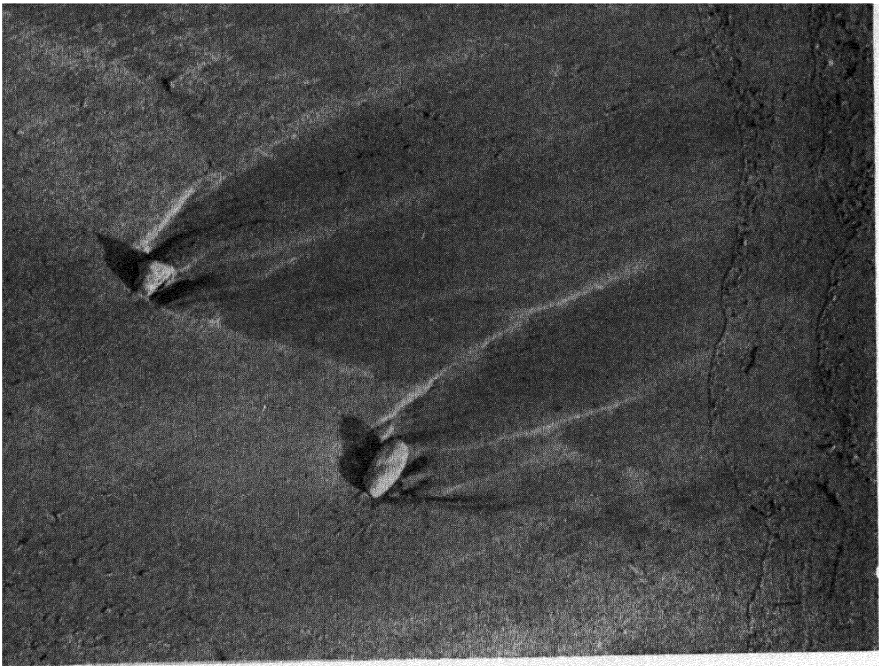


FIG. 125. TYPICAL RILL-MARK ON THE BEACH AT CASTIGLIONE, ALGERIA

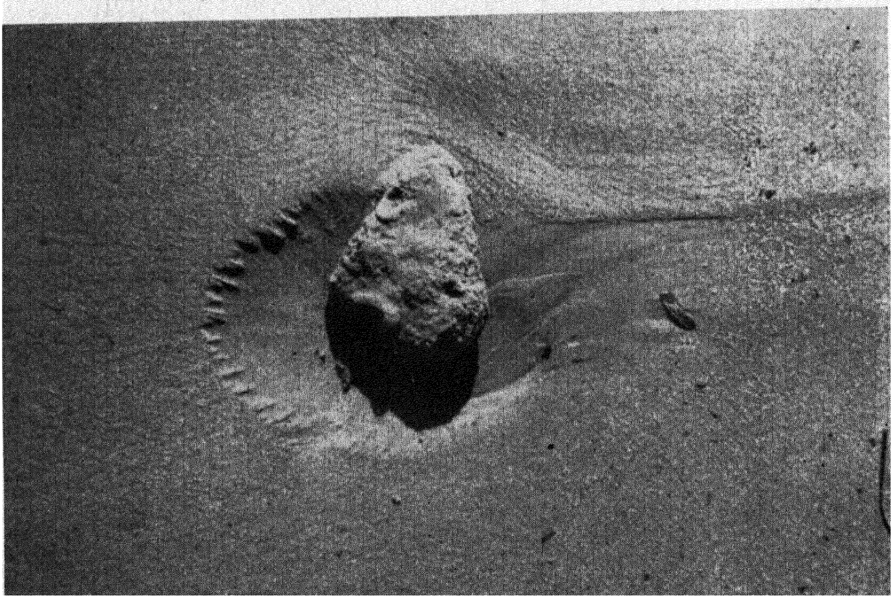


FIG. 126. RILL-MARKS FORMING A "PSEUDO-FOSSIL" AROUND A PEBBLE. ALSO AT CASTIGLIONE, ALGERIA (Photographs: G. Termier)

sometimes resemble the imprint of shells (fig. 126) and might then be called pseudo-fossils.

Tidal Fringes (fig. 127)

The advance and retreat of waves over a beach leads to the transportation of a great variety of objects. As a result of storms, the size of these objects may be considerable; the largest are thrown to the top of the beach and the smaller ones collect at the foot. During calm periods the waves

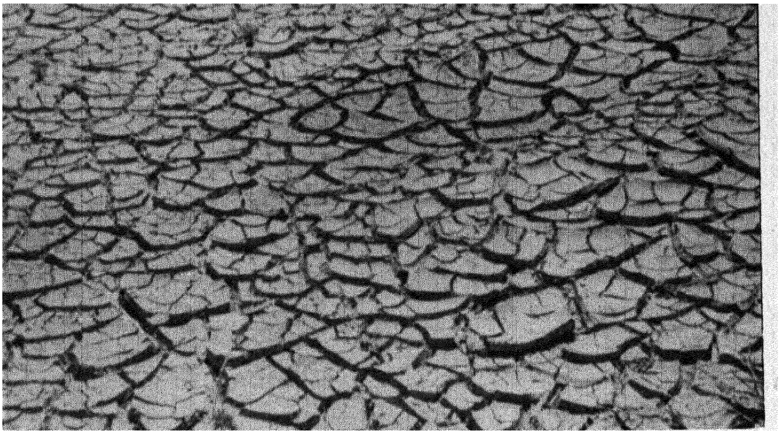
FIG. 127. PART OF A "TIDAL SCUM" ("FRANGE DE MARÉE") COMPOSED OF SAND AND FINE ORGANIC PARTICLES DERIVED FROM SEA-WEEDS ON THE BEACH AT SIDI FERUCH, ALGERIA (Photograph: G. Termier)

The waves come from the direction of the top of the picture.

deposit a fringe of fine sediment which accompanies scum, and forms complicated festoons. It does not seem impossible that these traces could become fossilized and thereby furnish proof of the existence of a coast line.

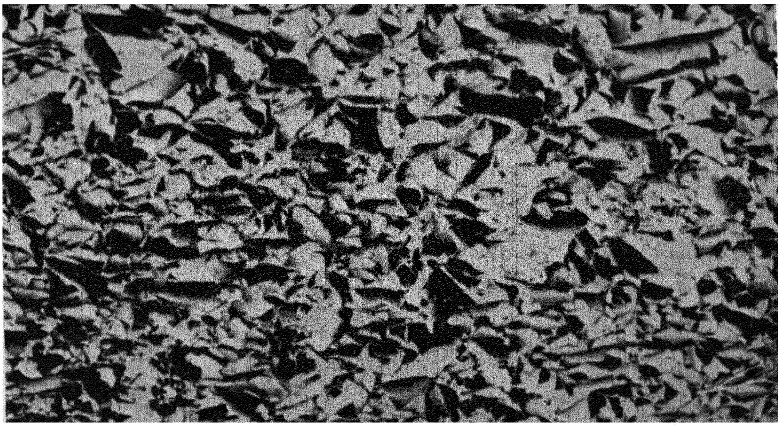
Mud Cracks (figs. 128 to 132)

When the argillaceous or calcareous muds on the bottom of a river, sea or playa are exposed to the sun, their surfaces dry out and shrink, forming patterns of polygonal cracks. Sometimes this surface forms a dry, thin film which breaks away from its substratum and curls up (fig. 128). These films when detached by the wind, may be swept together to form, after burial, a particular type of conglomerate (edgewise conglomerate). When clays are lithified by compaction and cementation, they become mudstones or limestones with fossil mud cracks on their surfaces (figs. 130–131).



FIGS. 128 and 129. MUD CRACKS IN THE DRIED-OUT BED OF THE MAGDALENA RIVER, SONORA STATE, MEXICO

Note the curling up of the mud lamellae in the lower picture. (Photographs: G. Termier.)



Soil Polygons

The *soil polygons* which have been observed in periglacial and tropical regions appear to be varieties of mud cracks with which are associated pebbles comparable to those of "regs" (fig. 132).

The "elephant-skin" structure which covers the surface of certain blocks or bands of silicified sandstone (quartzite), as in the case of sandstone of the forest of Fontainebleau (near Paris) (figs. 133–135) may be related to

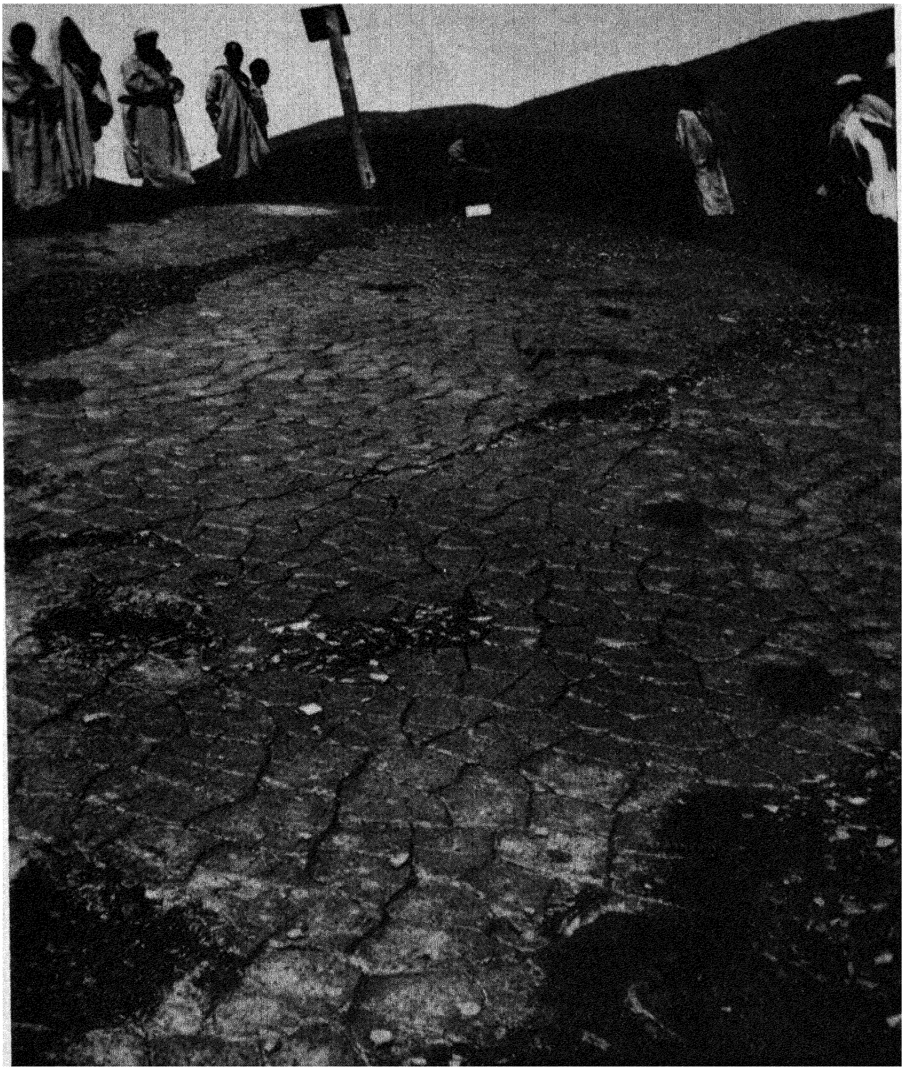
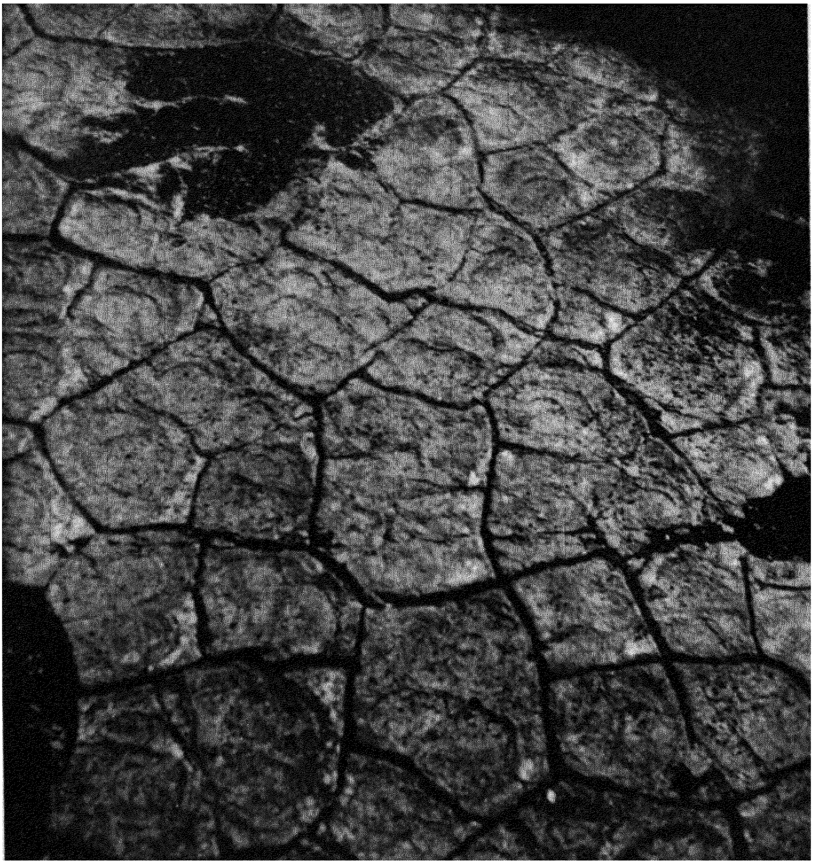


Fig. 130



FIGS. 130 and 131. CALCAREOUS MUDSTONE WITH MUD CRACKS AND THE FOOTPRINT OF A DINOSAUR. LOWER JURASSIC OF THE REGION EAST OF DEMNAT, AIT OUARIDÈNE (ATLAS OF MOROCCO (Photographs: H. Termier)

mud cracking. This structure indicates a colloidal state of the surface during diagenesis.

Alveolar and Cavernous Sands

A variety of sand has been observed on certain beaches (Charentes) in which deposition is closely connected with the presence of scum on the water. This variety is characterized by the presence of bubbles, alveoles and pockets due to the occurrence of substances with high surface tension (R. Baudoin, 1949). This structure often affects ripple-marks and gives shelter to some organisms. Moreover, it may become fossilized.

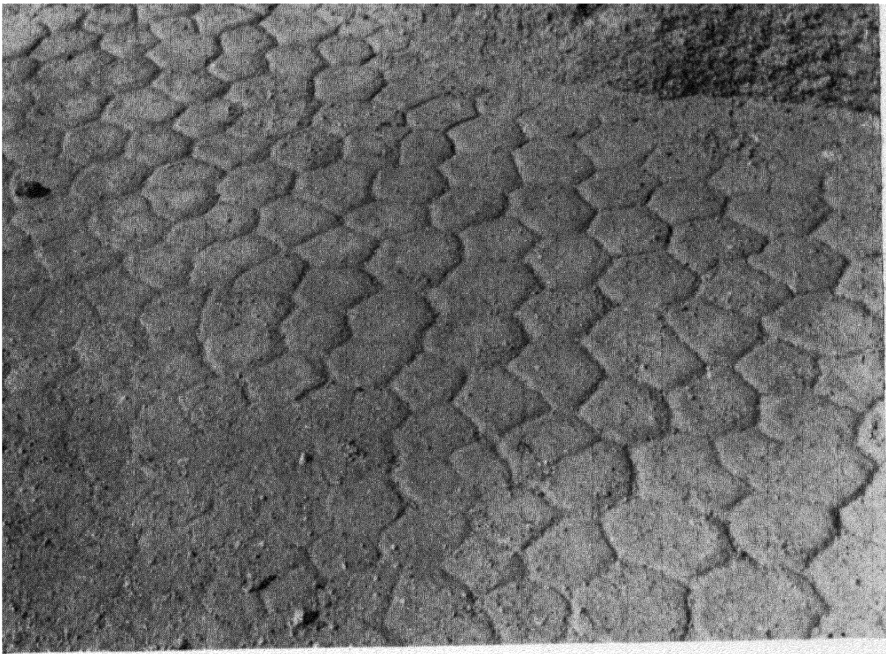


FIG. 132. REGULAR POLYGONAL STRUCTURE OF MUD CRACKS ON THE DRY FLOOR OF A SMALL LAKE. NORTH OF THE HOGGAR MOUNTAINS, NORTH AFRICA (Photograph: G. Termier)

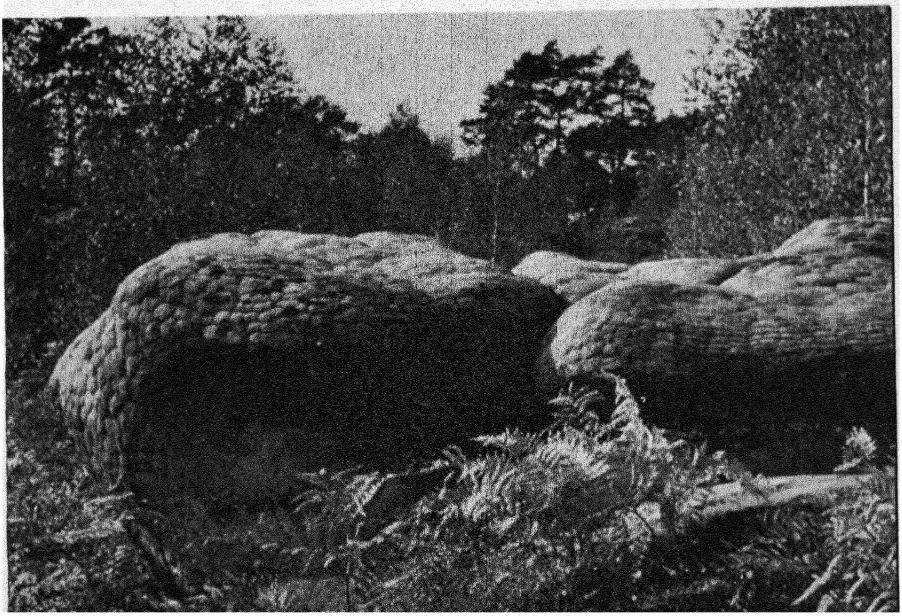
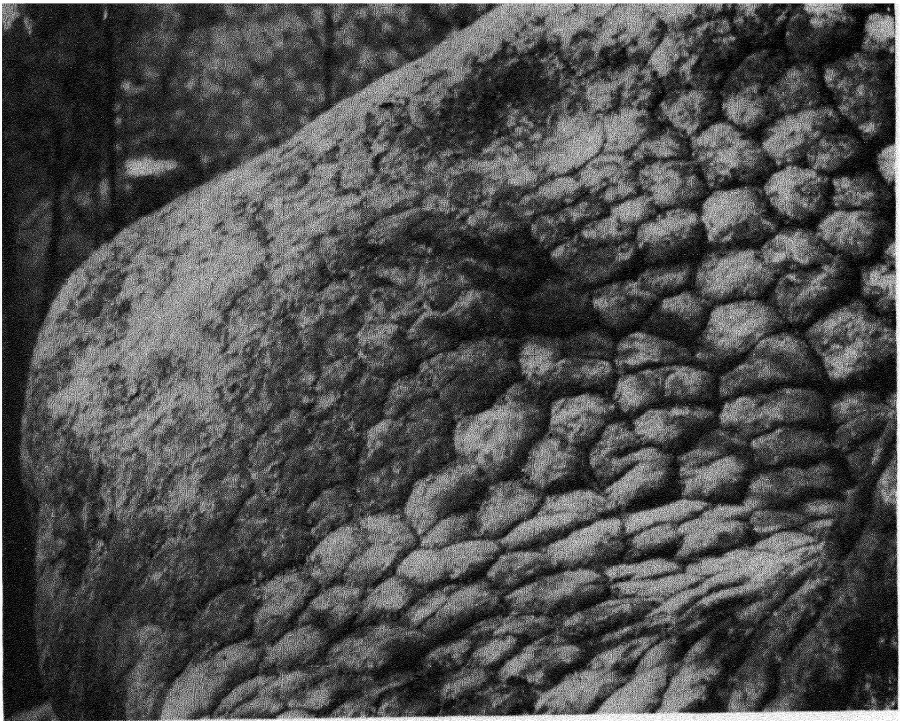


FIG. 133. "ELEPHANT-SKIN" STRUCTURE (*see also* figs. 134 *and* 135)



FIGS. 133, 134 and 135. POLYGONAL JOINTS OR "ELEPHANT-SKIN" STRUCTURE IN THE FONTAINEBLEU SANDSTONE, NEAR BARBIZON, SOUTH OF PARIS, FRANCE (Photographs: G. Termier)

II

Marine Sedimentation

There is a great contrast between purely marine sediments and continental sediments. The latter as previously shown, are generally, detrital, while the sediments formed in the sea are made up principally of skeletons or organic secretions and of the products of evaporation. The oceans are weak agents of erosion. But since the sea forms the base level for the rivers of continents, it receives the major part of the detrital sediment eroded from continents. The sediments of the coasts are, therefore, mixtures.

COASTAL SEDIMENTATION
(fig. 136)

In various well-defined regions of the world (Friesland, North Germany, the Gulf of Lions, Italy, Mesopotamia, the edge of the platform of the Sunda Islands, the Gulf of Mexico, to quote only a few well-known examples) the land is encroaching upon the sea. However, this is not due to the uplift of the continents or to changes in depth of the ocean floor. This gain affects only the continental shelf. In fact, the accumulation of sediment which is tending to increase the size of the emergent areas is later than the very recent Flandrian transgression (pp. 12 and 69), and locally tends to cancel out its effects. This sedimentation is a "filling-up" by subaerially eroded material and particular care should be taken not to confuse it with a regression. It seems that it may be to those continental deposits which spread over the marine domain that J. Bourcart (1955) gave the name of *invasion*, the meaning of which is, to say the least, ambiguous.

There appear to be four principal factors governing marine sedimentation (M. Dreyfuss, 1954): the deposits themselves, water depth, water agitation, and deformation of the sea bed. Together, these factors assume particular importance because of the *combined role of the depth and the movements of the bottom and because of the idea of a surface of equilibrium between sedimentation and erosion.* "When this surface is situated above the bottom, sedimentation occurs. There is a transfer of material or some erosion when the surface of equilibrium is situated at the level of the bottom or below it" (J. Barrell, 1917).

The most reliable indication of the depth of marine sedimentation is the evidence of temporary emergence in the tidal zone (mud cracks, ripple-

FIG. 136. AERIAL VIEW OF THE WEST COAST OF MEXICO, BETWEEN THE MOUTH OF THE RIO DE SANTIAGO AND MAZATLAN

Note the old hilly islands, covered with forest and dark in color, which have been joined to each other and to the mainland by an alluvial coastal plain built up by fluvio-marine sedimentaion. (Photograph: G. Termier.)

marks, rain pits). Nevertheless, a considerable number of marine organisms are, with some reservations, good indicators of depth (Termier, 1952) (see p. 57).

The idea of "bottom control" over sedimentation and erosion has also been suggested by A. Lombard (p. 355).

MARINE DEPOSITS

The Clays

The substance of marine clays (or muds) is made up of detrital fragments eroded from the land surface and transported by currents. The organisms which they contain derive their nutrients from the water rather than from the substratum.

On the surface of continents there are deserts and fertile soils, and similarly, at the bottom of the sea there are sediments of many types, some of which are suitable for the establishment of biotopes rich in living organisms. These "fertile soils" are the *clays* or *muds*, the fine particles of which are capable of absorbing many substances, especially organic matter. The homogeneity of these deposits is due to the presence of finely divided, decomposing organic debris, often algal, which forms what J. Bourcart (1939) has called *algon*. This tends to change into humus.

The "clays" are deposited in all bodies of still water whether they are lagoonal, fluvio-marine (p. 179) or deep sea. They are composed essentially of very fine-grained detrital material (including some *silt*), siliceous or ferruginous gels and flakes of clay minerals. The clay particles absorb preferentially certain chemical elements, for example boron (S. Landergren, 1945) and sulfur, which may occur in organisms as trace elements. They form a medium where bacteria can live in abundance, the number being a direct function of the fineness of the grain. Bacteria are found at all depths but are not uniformly distributed, as has been shown by ZoBell (1938, 1946). In the clays they produce humic acids and extract mineral substances from the water, instead of taking them from the substratum.

The muds are saturated with water and contain in the case of those from the Baltic, up to 50% and even 80% by weight, according to Debyser (1957). This water is rich in various salts derived from the mineral matter.

Fixation of the mud is often the work of organisms; the plants of the submarine "prairies", and the algae, arrest fine particles mechanically in the same way as mangroves (p. 182). Oysters, *Cardium*, crinoids, gastropods, holothurians and certain fish, such as eels, also accumulate mud particles. In Charente-Maritime, the bottom of the Marenne shelters 500 million (500,000,000) oysters which precipitate about 200,000 tons of mud each year.

The acids of the digestive tracts of marine organisms can alter the composition of the clay minerals ingested by the animals. A. E. Anderson (1958) has been able to show experimentally that, in a closed aquarium, a bentonitic montmorillonite exposed for five days to sea water was partially altered to brucite $Mg(OH)_2$. The same clay taken in by oysters, other pelecypods and fish was returned in the faecal pellets, having been altered by the acids of the alimentary canal; the brucite had disappeared and the crystalline structure included potassium in the clay layering structure.

There is undoubtedly a relationship between all the chemical processes of sedimentation in the sea and in lakes. The link between them is represented by the Schizophytes (bacteria and blue algae), which react in various ways according to the local conditions. In this category are the "waterblooms" which, in a highly nutritious environment, can form the source of hydrocarbon-rich, siliceous and sulfurous sediments. The Cyanophyceae can give rise to carbonaceous substances (bogheads) in the form of "waterblooms" and can also precipitate calcareous "biscuits", stromatoliths and calcareous muds. Finally, the sulfate-reducing bacteria can (by the liberation of CO_2) cause the precipitation of carbonates (Cl. Lalou, 1957) as well as bring about the formation of iron sulfide by their liberation of H_2S.

Glauconitic Sediments

The formation of glauconite in the presence of organic matter is characteristic of the marine environment. Many places are known where glauconitic sediments are being formed today. Glauconite occurs, for example, off Japan (both east and west coasts), in Monterey Bay, California, and in the Mississippi Delta. According to Galliher (1933, 1936) glauconite begins to form at a depth of 50 to 65 feet. It is also formed much deeper, but generally at depths less than 1,000 fathoms (6,000 feet). This mineral is a silicate of potassium and iron: $2(K_{1\cdot5})(Fe^{+++},Mg,Al,Fe^{++})_{4-6}(Si,Al)_8O_{20}(OH)_4$, with Fe^{+++} about 2, which occurs most frequently as detrital grains in green clay, sands and limestone. It is also known as a mineral formed *in situ*, commonly filling the tests of foraminifers and the canals of sponge spicules, as thin coatings, in small accumulations and in botryoidal aggregates. It may be formed by the alteration of a number of minerals: biotite is the most important (Galliher, 1939; Carozzi, 1951), but pyroxenes, felspars, clay minerals or colloidal silica (Takahashi, 1939) can be the original material. The alteration of biotite appears to begin by hydration with bloating of the mineral. In the black muds of the Bay of Aomori (Japan), Takahashi and Yagi have observed the alteration of faecal pellets of invertebrates to glauconite. The evidence indicates generally that a loss of silica and alumina occurs, and there is an enrichment in potassium and ferric iron. According to J. Bourcart (1958) and his colleagues, the internal glauconitic casts of foraminifers result from the alteration of an original cast of colloidal iron sulfide, at first to ferric oxide then by the adsorption of silica, to glauconite in an oxidizing environment.

Glauconitization is thus a chemical transformation which is not unlike laterization and chloritization. Like these two phenomena, it only takes place between precise limits, that is, in an open sea of normal salinity where the alkalinity can reach a pH of 8 to 9. In a brackish environment, such as that of the lagoon of Kasumiga-uri (Japan) the alteration is incomplete. In this case, there is also a loss of silica and enrichment by iron, but it is always ferrous iron, and the mineral formed resembles chlorite (chlorite, it should be noted, forms in shallow, sublittoral waters) (Hadding, 1932). The observations of Japanese workers prove that the temperature of the water is important and must not be less than 60° F. Moreover, glauconite is only formed in an anaerobic reducing environment, where organic matter is present. For this reason it is often accompanied by iron pyrite (Takahashi, 1939; Cloud, 1955). It seems that the action of bacteria, and also of humic acids and unstable organic salts of iron, play an important role in glauconitization.

Since glauconite is formed in sediments characterized by their richness in organic matter, it is necessary to add that this is usually a shallow-water, littoral facies which is readily reworked. This explains, perhaps, the apparent opposition between the reducing environment of formation and the richness of the mineral in ferric iron. Because glauconite is a stable and light mineral, it is easily transported and redeposited in the sediments of agitated water such as detrital quartz sands.

It should be noted that glauconite is a common mineral in transgressive horizons which often rest on a crystalline basement. For example, in northern Europe, the first beds of the Lower Cambrian, the Tremadoc beds of the Upper Cambrian and the Skiddaw beds at the base of the Lower Ordovician; in France, the Latharingian (Lower Lias) of Mont d'Or, Lyons, the Pliensbachian (Lower Lias) of Charollais, the Gault facies of the Aptian and Albian (Cretaceous); in North America, the Lower Cambrian and the Middle and Upper Cambrian of the mid-continent all contain glauconite. This appears to be due to the encroachment of a shallow sea over broad crystalline areas. Eckel (1914, p. 56) has calculated that the glauconite of all the Cretaceous series contain 25×10^{10} tons of iron oxide.

It is interesting to note that glauconite can cause *landslips*. The town of Algiers is built on the northern side of a high hill whose slopes are often very steep and pass locally into cliffs. These are composed of Pliocene rocks which are glauconitic marls overlain by calcareous sands. Rain water which percolates through this horizon dissolves calcium carbonate, but when it arrives at the contact with the glauconitic marls, the calcium is fixed and the potassium is liberated. The water therefore becomes distinctly alkaline and reaches a pH of 9. The marl is deflocculated and hydrolysis of the alumino-silicates occurs. The marl becomes liquid and the over-lying beds slide down the slopes and collapse (M. Proix-Noé, 1946; G. Drouhin, M. Gautier and F. Dervieux, 1949).

Finally, it may be noted that recently a method has been developed for the determination of the *absolute age* of rocks using glauconite.

Zones of Acidity in Marine Muds (see also p. 234)

In general the upper part of the mud is oxygenated to a depth of several inches, largely because certain benthonic organisms there lead a burrowing existence: the bacteria are *aerobic* and multicellular organisms (metazoans and metaphytes) can live there. Lower down in the unconsolidated mud (which can reach a thickness of at least 300 feet) the bacteria are *anaerobic* and other organisms are unable to survive. The anaerobic bacteria have the ability to enter into the geochemical cycles of iron, sulfur, carbon and nitrogen (see Chapter 18).

The important role of muds in relation to the organisms of sea floors has an important practical repercussion relating to researches into the origin of petroleum. This is a good example of the support that biology can give to geology. In some muds it is possible for organisms to live some depth below the surface of the sediment: these sediments are the *gyttjas* (Swedish term—adopted in sedimentology by E. Wasmund, 1930, and later by Krejci-Graf 1955) where the benthonic fauna and the plankton abound.

Dy, Sapropels and Gyttja

These bottoms are rich in nutrient substances (the eutrophic environment). In the *dy* (Swedish, mud, mire, used by E. Naumann, 1930), the organic matter is in colloidal form: this is found in rias and silted estuaries. The *sapropels* (Potonié, 1906) are the entirely anaerobic muds, often called foetid muds ("faulschlamm") or putrid muds from which hydrogen sulfide diffuses into the water. The origin of hydrocarbons can, in part, be attributed to chemical changes taking place in sapropels. In fact, the sapropels result from the accumulation of microscopic plant material (diatoms, pollen grains), other planktonic organisms rich in fats, and small crustaceans. Several chemists (Laurent, 1863; Engler, Maihle, 1922, etc.) have obtained synthetic petroleum by distilling or hydrogenating plant pulp comparable to sapropels.

Well-oxygenated sediments contain less organic matter (oligotrophic environment) than the "gyttja", but calcareous muds and limestones that give off a foetid odor when they are broken open (see p. 229) are well known.

The Amount of Organic Matter

L. Fage (1950) has emphasized that the amount of organic matter in sediments has considerable effect on the abundance of benthonic organisms, and that this is true for shallow, as well as deep, water. The maximum amounts of organic matter occur in calm waters, either in estuaries (5–15%) or near the edge of the belt of mud which surrounds the land mass, the "mud line", where the amount is 6%. Consequently, when these zones are well oxygenated the faunas are most abundant. Organic matter present in abyssal and bathyal sediments is largely due to plankton, where the amount may fall to 0·02% (Trask, 1939).

Among the deep sea sediments, the Globigerina ooze may reach 3·45% of organic matter (Correns, 1939), an amount intermediate between that of the blue muds and the red clays.

Upwelling currents (see p. 234) bring to the surface nitrogen and phosphates which can be assimilated by phytoplankton, and thus enrich the marine environment in nutrient substances. Californian sediments formed in such an environment contain nearly 10% of organic matter.

When the topography of the sea floor provides basins where sedimentation can occur under calm conditions and decreasing oxidation, the accumulation of organic matter is encouraged. This is typical of fjords and lagoons.

The organic matter of sediments is present in three forms: *pectins* which are decomposed rapidly; *chitins*, attacked less rapidly by bacteria; and *celluloses* (of plants and tunicates) which are decomposed very slowly in an anaerobic marine environment.

It seems likely that the organic matter on the sea bottom may be assimilated by micro-organisms, including bacteria (Waksman, 1933; Baier, 1935). These may live symbiotically with mud eaters or may be absorbed by them as food (MacGinitie, 1934; Portier, 1938; ZoBell, 1946).

Organic Matter in Fossil Marine Sediments

Traces of carbonaceous matter are known to occur in the slates of the Lake Rice Series of Manitoba (2,500 million years old). The more recent argillites of Cuyuna (1,000 million years old) included in the Biwabik Formation of Minnesota and contemporary with those of the Gunflint, have yielded 380 p.p.m. of bitumen, poor in hydrocarbons. It is sufficiently rich in asphalt, and sometimes contains sulfur, to indicate a shallow coastal marine environment receiving abundant clastic and tuffaceous sediment.

An interesting section in the Middle Devonian marine sediments of the Mount Union region (Pennsylvania) has been described by Swain (1958) who determined qualitatively and quantitatively the character of the organic matter present. This method allows the variations in the paleogeography to be followed very closely.

1. The *Ridgeley sandstone* (part of the Oriskany, at the top of the Lower Devonian) contained 230 p.p.m. of bituminous extracts with 16% of free sulfur.

2. The *base of the Newton Hamilton Formation* (equivalent to the Onondaga) is discordant and transgressive over the preceding sandstone which it has reworked. This is an argillaceous sand, thinly bedded, which contains lenses of shale. It contains 179 to 210 p.p.m. of bituminous extracts, with 0 to 8·6% of free sulfur. The amount of hydrocarbons in the bitumen is considerable, and there is about 0·15% of humic acid. This deposit was formed in a very shallow epicontinental sea.

In the rest of the calcareous shales of the Newton Hamilton Formation, the amount of bitumen varies between 74 and 605 p.p.m. with an average of 100 to 300 p.p.m. In calcareous shales about 16 feet below the top, the amount of organic carbon reaches 3·49%. Free sulfur is generally absent, but may reach 26% of the extract (=45 p.p.m.). Where it is absent, however, it is replaced by gypsum crystals which may be partially altered to sulfur. There is apparently a genetic relationship between the gypsum and the sulfur which is probably associated with the iron sulfide formed as the organic matter is deposited (see p. 232).

3. At 44 feet below the top of the *Newton Hamilton Formation*, a thin-bedded, argillaceous limestone with fossil debris, large brachiopods, and phosphatic scales (? fish), contains grains of quartz and crystals of pyrite. Bitumens comprise 171 p.p.m. of the rock, there being very little asphalt. Sulfur is also present (19 p.p.m.). This is a deposit of an epicontinental sea, near to a coast line, or on a shoal.

4. At the top of the *Newton Hamilton* beds greenish gray, thin-bedded shales with crystals of calcite contain bitumens (400 p.p.m.) very rich in asphalt. This bed is, perhaps, comparable to that formed on a prodelta (p. 184).

5. The Newton Hamilton Formation also contains brachiopod-bearing shales deposited on the margin of the continental shelf yielding asphalt-rich bitumens.

6. In the same formation a greenish *pelagic shale* contains *Tentaculites*, *Styliolina*, and ostracodes (*Pholidops*, *Bollia*), together with rare brachiopods. This horizon shows undulating bedding and is almost devoid of asphalt.

7. The black *Marcellus Shale* is particularly rich in organic matter (2·57% of organic carbon); the bitumens reach 2,248 p.p.m. and are largely saturated hydrocarbons. The rock contains 1,458 p.p.m. of sulfur and 1% of humic acid. The sulfur is concentrated at certain levels. Siliceous concretions and finely divided pyrites are also present. The fauna, consisting of small brachiopods and pelecypods, is typical of the pyritic facies. The sediment is often sapropelic.

8. The base of the *Mahantango* beds is a sandstone, but the rest of the series is mainly shale with sandy lenses. The amount of bitumen varies between 20 and 1,000 p.p.m., with an average of 100 p.p.m.

A little higher, a shale representing a prodelta facies, contains carbonaceous fragments. It contains 1,012 to 1,014 p.p.m. bitumen, which is mainly saturated hydrocarbons, and also 0·08% of sulfur. The same facies in the rest of the succession contains 0·04 to 0·4% of humic acid.

9. The principal facies of the *Middle Mahantango* is a shelly mud deposited on the continental shelf. This also is thinly bedded and contains relatively little organic carbon (0·21%). The bitumens vary from 183 to 1,402 p.p.m. and contain 20% of sulfur. Polar organic compounds are poorly represented.

10. A *shelly, argillaceous sandstone of the Mahantango* which was deposited in an epicontinental sea contains 198 p.p.m. of bitumen (mostly saturated hydrocarbons). This sandstone consists of well-bedded layers, each about 3 inches thick.

11. The base of the Portage Group (Upper Devonian) is formed by the Tully (shaly) Limestone. This is a shelly rock deposited toward the open sea and contains 186 p.p.m. of bitumen and 0·5% of humic acid.

12. The succeeding member of the succession, the *Burket* (black) *Shale* is a thick-bedded, micaceous, carbonaceous, sapropelic deposit containing brachiopods, pelecypods and fragments of trilobites. It contains 443 p.p.m. of bitumen, poor in polar compounds (resins and pigments), and 2% of humic acid. Toward the top it becomes more silty and then more calcareous. Deposited further from the coast, the amount of bitumen is smaller (153 p.p.m.) as is the humic acid (0·1%).

The Portage Series is a regressive series in Pennsylvania. According to Clarke, it was contemporaneous with coastal glaciers which have striated the surface of the sandstones immediately beneath, but proof of glaciers is lacking and the striations are more likely due to gravity slides. The sea in which horizons 11 and 12 (above) were deposited was thus fresher water than that of the Middle Devonian.

Foetid Limestones, Bituminous Limestones ("stinkstones", anthraconites, lacullans)

The clays discussed above are not the only sediments formed in a reducing environment. There are many examples of foetid limestones containing pyrite and bitumen. Among present-day sediments there are calcareous muds (of Red Sea reefs) which Nesteroff (1955) attributed to the excreta of browsing fish and to corals. Although white in color, they give off a slight odor of H_2S, become black if they are placed in a closed bottle (probably due to the action of Rhodothiobacteria), and form small crystals of pyrite 1 mm. long. A similar occurrence is known in the Great Barrier Reef of Australia (Marshall and Orr, 1931). In many atoll lagoons the white carbonate mud becomes black and saturated with H_2S only an inch or so down.

The southern Jura was occupied during Kimmeridgian times by an atoll with a lagoon extending 28 miles from north to south and 15 miles wide (Cerin, Vaux Saint-Sulpice, Orbagnoux, Saint-Champs). The sediments deposited in the lagoon were thinly bedded bituminous limestones devoid of clays. They are 13–16 feet thick at the edges and 200 feet thick in the center of the basin. To the west and northwest the reef sediments of the margins are dolomites, oolitic limestones and calcarenites, while to the east and the south, coral reefs are associated with the dolomites. These separate the basin from the pelagic limestones of the Tithonian facies.

In bituminous limestones, the abundant Dinoflagellates and Hystrichospheres are believed by G. Deflandre to be the source of the bitumens. Radiolaria and diatoms are also present in this limestone.

The rocks are banded by the alternation of thin beds, one to several millimetres thick. Light-colored bands of calcite debris from *Miliola* and *Radiolaria* tests alternate with dark bands composed of lignitic debris, pollen, diatoms and hystrichospheres in which the organic material is concentrated. This is a deposit from still water in a closed basin. Patches of chalcedony and recrystallized calcite resulting from the diagenesis of the soft sediment are common. Iron is absent. The organic material has not altered to hydrocarbons, but is still as a *proto-petroleum*. The abundant CO_2 which saturates the environment appears to have stopped the reducing action and impeded the formation of hydrocarbons. Enrichment in sulfur has taken place, and this has been derived partly from micro-organisms and partly from organic matter (Y. Gubler and M. Louis, 1956) (fig. 137).

Bituminous limestones are often found associated with bituminous shales where they occur as small nodules, as in the alum shales. From samples taken from the Lower Cambrian up to the Middle Ordovician, Landergren (1954) has been able to show the existence of two phases of carbon, organic and inorganic, distinguished by their $^{12}C/^{13}C$ ratios.

The ratio of the stable isotopes of carbon $^{12}C/^{13}C$ in organic matter is higher than that in carbonates. In the *Megalaspis limbata* limestones of the Lower Ordovician of Scandinavia, the ratio varies between 88·43 and 89·18, being much lower in the red limestones formed under oxidizing conditions than in the gray limestones formed under reducing conditions. In stagnant water, where fermentation is taking place, the carbon is enriched in the lighter isotope which comes from the carbon dioxide released from the organic matter. If the organic carbon dioxide is in excess of that in the sea water (and derived from the atmosphere), then the precipitated calcite will be enriched in the lighter isotope of carbon. As a general rule, the value of the $^{12}C/^{13}C$ ratio of a marine limestone can be used to indicate whether it was deposited in calm or agitated water. This is confirmed by additional evidence: the presence of glauconite accompanying the enrichment of the light isotope in the *Megalaspis planilimbata* limestone (Lower Ordovician) points also to the existence of calm conditions.

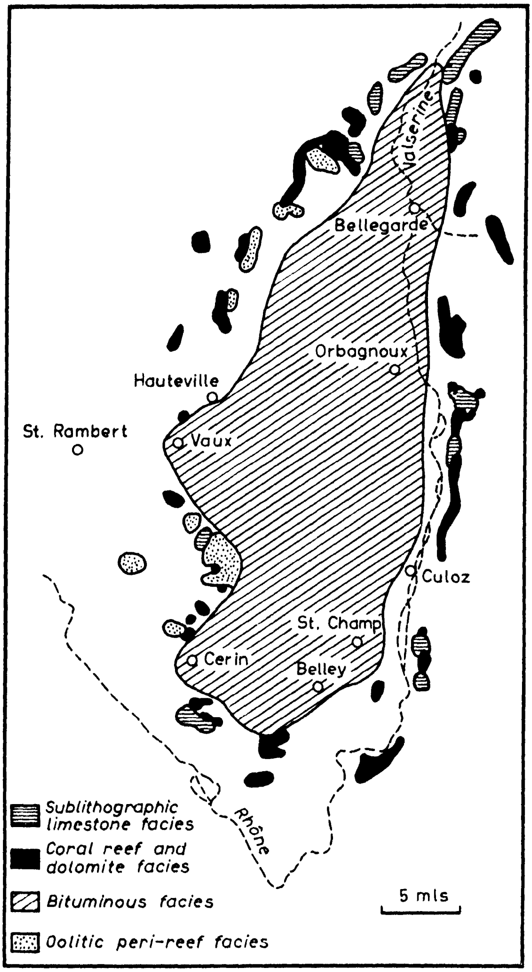


FIG. 137. THE KIMMERIDGIAN ATOLL OF CERIN, SOUTHERN JURA, FRANCE (after Gubler and Louis)

Petroleum Source Rocks

For a sediment normally rich in organic matter to give rise to a petroleum bearing series, it seems necessary that there must have been continuity of sedimentation. Thus, in the petroleum-bearing Recent clays of Grande Isle (Louisiana), the deposition of which was quite normal, a core, 115 feet long, has been obtained which represents uninterrupted deposition for at least 12,000 years (Smith, 1954). Similar sediments have been recorded in California by Trask (1956). Hydrocarbons may impregnate lagoonal or lacustrine sediments providing deposition was continuous over a sufficiently long period of time.

The evolution of petroleum-bearing sediments is brought to its final stages by diagenetic processes. The role of bacteria is undoubtedly very important since they can cause modifications of both organic and mineral matter in the sediment, particularly in the superficial zones. Thus, the sulfate-reducing bacteria produce C_{10}–C_{27} hydrocarbons (Jankowski and ZoBell, 1944). Possibly the synthesizing activity of the bacteria is assisted by high pressures within the sediment (Oppenheimer, 1950). Finally, it must be mentioned that higher organisms (corals, gorgons, sharks) containing hydrocarbons may be included in these deposits (Bergman, 1949; see also p. 237).

The littoral and neritic zones occupying the continental shelf are regions where terrestrially derived sediments are thickest, most rapidly deposited, and richest in organic matter. They are, therefore, particularly suited to the formation of hydrocarbons. These sediments can also be carried into deep water. The sandy zones and those which are swept by violent currents are not favorable to the accumulation of hydrocarbons because they are continually subjected to the oxidation effects. The most favorable environments, though not the only ones, are those where reducing conditions are established, and where hydrogen sulfide is accumulating, that is, in the *euxinic environments* and in the *sapropels* (see above). Hydrocarbons also form in the middle of coral reefs and, in general, all over the continental shelf.

HYDROCARBON FACIES

Prokopovitch (1952) distinguished three marine environments favorable to the development of *waterbloom* (see p. 223): (1) certain deltas such as those situated on the coast of the Gulf of Mexico; (2) certain shallow seas (such as that in which the Silurian deposits of Bessarabia were formed); (3) shelf seas (such as that in which the kukersite deposits of Estonia occurred). It will be shown later (pp. 234–238) that upwellings of water in the sea favor the development of waterbloom. Freshwater lakes are also known where conditions are comparable to those resulting from such currents.

E. Wasmund (1930) distinguished *sapropels*, where the O_2–H_2S boundary is in the water or very near the surface of the clay, from the *gyttjas* (see p. 226), where the O_2–H_2S boundary is situated far below the surface of the sediment and aerobic benthonic life is possible. The gyttjas contain planktonic and benthonic organisms in abundance. They are rich in nutrient substances and so correspond to the *eutrophic environment*. They are thus distinguished from the sapropels which correspond to the *hypertrophic* (see p. 234) or *dystrophic environment.*

E. Naumann (1930) used a slightly different terminology: the *sapropels* include, in part, the *gyttjas* in which it is possible to determine the origin of the organic matter, and the *dys* where the organic matter is colloidal and precipitated. It seems that this last situation may often occur in certain estuaries, for example in the silted estuaries of southern Russia (Dnieper, Bug), in the Black Sea, and in the rias of the northern and western coasts of the Spanish peninsula.

The preservation of organic matter in these environments tends to produce hydrocarbons. Such environments are known as "euxinic" and have been known for a long time in the Black Sea. In contrast to this is the "oligotrophic" environment, rich in oxygen, such as occurs in the bottom waters and in which the quantity of organic matter is small. The sediments formed in this instance are essentially mineral in character; that is, detrital sediments and limestones.

In the facies corresponding to the hydrocarbon sediments, there is a distinct contrast between the sapropels and the gyttjas. The former, which often correspond to the conditions associated with rising currents (upwellings, see p. 234), are argillo-siliceous sediments with nektonic and planktonic fossils (at Holzmaden in Germany, benthonic organisms are associated with driftwood). The gyttjas, on the other hand, have a quite different fauna composed mainly of juvenile forms belonging to those species which frequent the open sea in the adult stage, and other, benthonic, forms. These are the equivalent of the spawning grounds of the modern seas which are restricted to the upper part of the continental shelf and are colonized by seaweeds and marine grasses.

The formation of hydrocarbons in the submarine "fields" of seaweeds is accentuated by the presence of certain organisms rich in them, particularly the wracks (*Fucus*).

The kukersite of Estonia, where the benthonic fauna contains an abundance of bryozoans, has features intermediate between the two types.

The large amount of hydrogen sulfide in fossil sediments leads in both cases to active pyritization. But the gyttjas, from which a part escapes to the "sulphuretum"[1] contain a large proportion of limestone, and are light in color. These are represented most often by pyritized marls.

A description of hydrocarbon sedimentation would be incomplete without mentioning the underground waters associated with petroleum. Hydrocarbon deposits where there has been no displacement of the oil are generally associated with water very rich in chlorides.

Research into petroleum sedimentation has produced many publications which it would be inappropriate to review here.

Sedimentary Sulfides

One of the consequences of the anaerobic, reducing environment in which some clays are deposited, is the formation of sulfides of which the most important is iron pyrite.

[1] The surface layer of unconsolidated clay and water impregnated with H_2S.

The iron is linked to the organic material to form true complexes (Baas Becking and Moore, 1959). It seems that the proportion of organic matter and (ferrous) iron is generally 3 to 1. These complexes form chiefly in the black foetid muds, for example, in estuaries, during the reduction of marine sulfates. According to Rochford (1952) the muds of estuaries contain 8·5 to 19% of organic matter, and consequently the amount of material affected is considerable. The density of the sediment decreases with the increase in the amount of iron, a fact which appears paradoxical. This is, however, largely compensated by the amount of organic matter.

Alum Shales.—The alum shale facies has been studied in detail in Scandinavia where it occurs in the Middle and Upper Cambrian. These are black shales containing alum, which results from the decomposition of finely disseminated sedimentary iron pyrite. The alum shales contain nodules and sometimes bands of foetid, black limestone (anthraconite and lucullan), which often exhibit cone-in-cone structure. The whole of the sediment is more or less bituminous, and is a true sapropel. It was formed under anaerobic conditions in a shallow sea. It contains algae of the type *Morania* (Henningsmoen, 1956) and sometimes pyritized fossils. The presence of iodine is evidence of the algal origin. The amount of boron (0·014% B = 0·045% B_2O_3) present indicates, according to Landergren (1945) that the deposit was definitely laid down in a marine environment. Thermal alteration, in places, due to diabase intrusions, reduces the amount of boron to 0·009% (=0·029% B_2O_3). The rise in temperature due to metamorphism has driven off some of the hydrocarbons in the rock but has not affected the $^{12}C/^{13}C$ ratio in the organic carbon (91·11 $\pm$ 0·12) or that of the carbonate carbon (89·52 $\pm$ 0·16) (Landergren, 1955). The alum shales of Sweden also contain vanadium (2·6 g. per 1,000 kg.) which is combined with the porphyrins derived from chlorophyll (see p. 237) (Bader, 1937). The same sediments also contain molybdenum and tungsten.

The *Kolm* (Swed. Kol, coal) is a variety of cannel-coal, with a high ash content, which occurs in the alum shales of the south of Sweden. It is rich in uranium, which reaches nearly 1% (Eklund, 1946) whereas the normal alum shales contain only 50 to 100 gm. per 1,000 kg. The shales also contain thorium (0·6–2 p.p.m.), according to Koczy (1949). The method of dating by the lead-uranium ratio has indicated that the age of the kolm is about 440 million years (Nier, 1939), but more recent determinations suggest that this figure should be about 15% higher. Sphalerite (zinc sulfide) is sometimes found in the alum shales of Sweden (Lundergardh, 1948).

The Kupferschiefer.—In the lower Zechstein (Permian) of Germany, there occurs the Mansfeld Kupferschiefer, which is rich in fish remains and plant matter and is comparable to the alum shales of Sweden. They have, however, undergone mineralization which has altered their chemical composition, and in particular, the amount of uranium, copper sulfide and molybdenum.

Goldschmidt (1937) thought that these shales were due, entirely or in part, to the erosion and redeposition of humic soils enriched in copper salts. Furthermore, mineralizing solutions played a part in the subsequent diagenesis, bringing in molybdenum and uranium. Certain beds also show an enrichment in zinc and cadmium.

From an examination of long core samples obtained from the Baltic off Stockholm between Bornholm and Landsortdjupet, Debyser (1957) compared the chemical and mineral constitution of organic muds containing pyrite and siderite, with their Eh and pH. He came to the following conclusions: "(1) The favorable environment for the formation of pyrite is found at depth within the sediment, which implied in this particular case that it is not syngenetic, but is post-depositional. (2) There is agreement between the theoretical limits of stability of the pyrite and the percentage of sulfur *in situ*, these being considered proportional to the amount of sulfide. (3) There exists a relationship between the pH and the Eh, on the one hand and the chemical constitution of the sediment, on the other. (4) The alternations in the nature of the deposit favor its diagenesis because they bring into close contact heterogeneous components, which must move into equilibrium with each other."

Ascending Currents (Upwellings)

The *ascending currents* ("upwelling", Sverdrup, Johnson, Fleming, 1942, pp. 140, 501, etc.) are those currents of cold, dense water which rise from the bottom toward the surface of the sea. They are generally due to the effect of offshore winds, with counter-currents at depth, that drive the cold bottom water up the continental slope to the surface. They occur especially at the points where surface currents diverge, particularly in the tropical and subtropical zones (coasts of California, Peru, Somalia and West Africa) and around the Antarctic. They are found also in closed seas such as the Black Sea, the Sea of Azov and the Caspian. These are discontinuous phenomena dependent on the constancy of the wind, and hence are seasonal. Upwellings give rise to zones of low temperature water in seas which are generally warm. Locally they may modify the climate of neighboring landmasses. The rainfall on these landmasses is usually small, and little detrital sediment is carried by the streams.

These ascending currents have important sedimentological consequences. Due to their low temperature they drive out the calcareous plankton, especially the Globigerina, and these are replaced by arctic or antarctic plankton, diatoms and flagellates. Since the terrigenous contributions are small, conditions are suitable for the formation of pure diatomaceous oozes. Furthermore, the upwellings bring to the surface and hence to the phytoplankton, an abundant supply of mineral salts (phosphates, nitrates, silica, etc.) (Revelle and Shepard, 1939). This hypertrophic environment (the nutrient environment of Sverdrup, Johnson and Fleming, 1942, p. 782) contains material derived from the continents and stored in the waters of the aphotic zone at depths of 600 to 1,000 feet, where no autotrophic vegetation exists. It thus permits a superabundant growth of unicellular plants (Prostista) in the euphotic zone; the formation of "waterbloom" often causes a change in the color of the sea. The relationship between upwellings and the abundance of phytoplankton has been observed by Trask (1939)

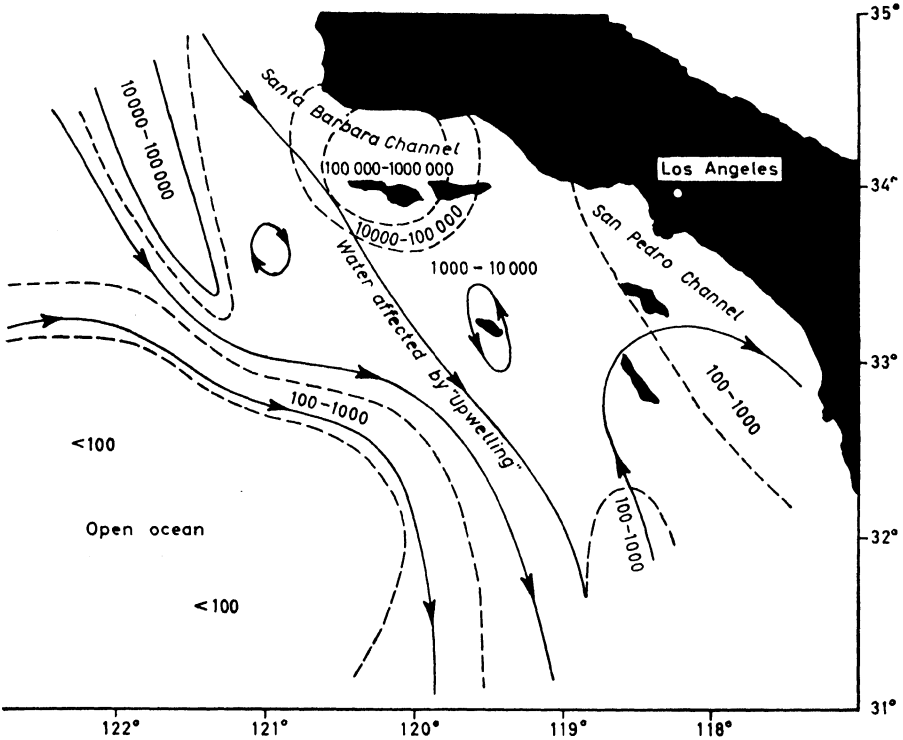


FIG. 138. THE UPWELLING (ASCENDING CURRENT) OFF THE CALIFORNIAN COAST (after Sverdrup, Fleming and Johnson)

The figures show the number of diatoms per liter of sea water. The arrows indicate the direction of the currents.

in California, by Correns (1940) near Cape Verde, by Wisemann and Bennett in the Arabian sea, and by Böhnecke (1943) off the Canary Islands and the Azores.

It has been known for a long time that the formation of sapropels (see above) is often associated with the formation of "waterbloom". The Peridineans and certain Chloromonadins secrete a mucilage which binds the individuals together. The whole of this jelly-like mass finally sinks and covers the sediment of the bottom which then develops anaerobic conditions.

The sapropels are the anaerobic sediments rich in hydrogen sulfide and deficient in oxygen. The boundary between the hydrogen sulfide layer and the oxygenated zone may be either very near the bottom or well up into the water. According to Copenhagen (1934) diatoms selectively absorb hydrogen sulfide and are then no longer a nutrient, but a poison for the benthonic organisms. The presence of the H_2S explains why the benthonic fauna of these zones is impoverished.

Furthermore, when the "waterbloom" develops, certain dinoflagellates, such as the Noctiluca, appear to secrete a paralyzing toxin (M. Brongersma-Sanders, 1948). Animals coming to the surface of the sea to feed on the plankton, or other animals which have already eaten, are poisoned and die in great numbers. These include fishes, sea birds, seals and crustaceans (shrimps and crabs). The pelecypods of the sea bed (*Donax*, *Pholas*, *Mactra*) and alcyonarias also die if they are contaminated by this environment. The numerous bodies increase the amount of organic matter available for anaerobic decomposition.

Most of the flagellates giving rise to "waterbloom" are Peridineans. In 1946–1947 they caused the death of very many fish, turtles, oysters, crabs, shrimps and barnacles on the coast of Florida by their toxins, the effect of which is comparable to curare. Mussels, which live principally on the phytoplankton are immune to these poisons and consequently are extremely dangerous to eat.

The toxicity of the flagellates seems to be linked to the change of color of their plastids, which on turning brown, seems to favor the synthesis of the toxin. This is the case, not only with the posidonians, such as *Goniaulax* and *Pyrodinium*, but also with the chrysomonadines, like *Chattonella.* This genus, by the periodic pollution of the west coast of India near Kozhikode (Malabar), has caused the death of many fish, molluscs and crustaceans (Subrahmanyan, 1954). In the port of Algiers where the water is permanently foul, the same genus produced "waterbloom" in July and August 1956, due to a rise in the temperature of the sea water (to 75° F.) and weak agitation of the water (Hollande and Enjumet, 1956).

Walvis Bay (Southwest Africa).—An example of upwelling accompanied by waterbloom and the slaughter of fish is found in Walvis Bay (M. Brongersma-Sanders, 1943, 1944, 1947, 1948, 1951, 1957). The bay is situated on the southwest coast of Africa in latitude 21° 30′ S. The sediment there is a dark green fine-grained mud with a strong odor of H_2S. In that mud there are 44 c.c. of hydrogen sulfide, for a total gas content of 79 c.c. This amount decreases in the sandier zones. The bottom is so poor in benthonic life that it can be called azoic. This zone is *situated in the open sea* below an ascending current and falls away to a depth of 450 to 500 feet. Farther away, the sediment becomes very hard and loses its foul smell. Sometimes a narrow sandy band, rich in a benthonic flora and fauna, occurs between the coast and the azoic zone.

In this latter zone, 60% (by dry weight) of the sediment is formed from the frustules of siliceous algae. It is a diatomaceous ooze enclosing the dead bodies of fish. Toward the open sea, it contains many tests of foraminifers and empty shells of *Crassatella*. In general, it is totally without carbonates, but instead contains much organic matter (the loss on ignition reaching 19·65%).

Walvis Bay is well known for the slaughter of fish which occurs nearly every year toward December (midsummer in the southern hemisphere). These destructions coincide with the "blossoming" of the Noctiluca, which redden the sea and release a most unpleasant odor. The fish of the sea bottom are the first to be affected by them, the pelagic forms follow later. The piling up of the carcases forms a wall which may reach 7 feet in height.

The bodies which thus become mixed with the diatoms also contribute to the formation of hydrocarbons which are enclosed in the sapropel. The decomposition of the diatoms furnishes oil in abundance, and it has been suggested by Pirson (1946) that the role of animals in the formation of petroleum is comparatively small. It has been pointed out, however, that the livers of certain sharks contain oil in which there occurs an unsaturated hydrocarbon, *squalene* $C_{30}H_{50}$ in the unsaponifiable fraction (Tsujimoto, Ormandy, Craven, Heilbron and Channon, 1927). This hydrocarbon on hydrogenation yields an oil closely related to liquid paraffin (Boutan, 1926, 1928). The destruction of fish accompanying the planktonic waterbloom must further increase the amount of hydrocarbons in sediment.

In the course of the diagenesis of such deposits (pp. 230–231), nitrogen is expelled more rapidly than carbon; the ratio C/N changes from about 8·4 in recent sediments to 14 in fossil sediments. The amount of organic matter decreases at the same time by about 20% (Trask, 1932). There is also a loss of oxygen, sulfur and phosphorus, but the relative proportion of hydrocarbons tends to increase. (According to ZoBell and Smith, 1952, a fresh sediment contains 10 to 20 mg. of liquid hydrocarbons per 100 gm.) Thus the organic matter tends to advance toward petroleum. It may be added that the anaerobic conditions explain the conservation of porphyrins derived from chlorophyll in marine sediments containing hydrocarbons (M. Brongersma-Sanders, 1951).

Fossil sediments similar to those formed by upwellings are frequent throughout the petroliferous series. Many of these are also rich in diatoms (Oligocene menelite shales of the Carpathians, Miocene shales of Monterey, California). Furthermore, the remains of fish and the rarity of benthonic invertebrates is also characteristic.

M. Brongersma-Sanders (1948) has suggested the following examples: the Lower Tertiary shales of Padang (Sumatra) and the Green River shales (see p. 175) which are freshwater deposits; the Kupferschiefer of the Zechstein of Germany (an interpretation differing from that of Goldschmidt, see p. 234) and the Oligocene of the northern Caucasus, which are both marine. To these may be added the bituminous shales of Toarcian (Upper Lias) age from Holzmaden (Württemberg), which show all the essential faunal and sedimentological characteristics, and are also marine. The tests of all the fossils of this horizon are pyritized, and fossilization has thus occurred in a "sulphuretum".

The Black Sea.—It seems that the case, often quoted, of the Black Sea permits the reconciliation of several theories concerning sapropels and their pyritized faunas. Eelgrass is abundant on the northern part of the continental shelf on the borders of the limans[1] and in the sea of Azov. The rivers carry a large amount of nutrients which give rise to an abundant waterbloom. In the absence of rivers, it would be necessary to invoke upwellings as an explanation. The hydrogen sulfide is finally trapped in the deep waters of the sea.

The most important characteristic of the Black Sea is the concentration of this hydrogen sulfide in the deep waters. This is similar to the conditions which give rise to the *euxinic* environment of other large seas. A concentration of hydrogen sulfide inimical to all living organisms is found at depths of 400 to 650 feet near the coasts and between 300 and 400 feet in the central region. It seems that the development of waterbloom in the limans and in the Sea of Azov is one of the chief sources of H_2S. The stillness of the deep waters then plays a part in its retention. As a result of the large quantities of hydrogen sulfide carried down, the floor of the Black Sea is very sparsely populated.

In the Sea of Azov and nearby, the shallow fresh water is enriched in nutrients brought by the rivers from the Russian platform, and abounds in algae and Protista. Waterbloom develops at certain periods (in summer) when the surface temperature rises to 75° F. The bloom of dinoflagellates, including *Exuviella cordata*, causes a reddening of the water and a great slaughter of fish, as in the bay of Temrjuk. By the beginning of autumn, the Sea of Azov, near the straits of Kerch, is covered by blue algae, and the diatoms multiply rapidly (Knipowitsch, 1926–1927). Thus the bottom of the Sea of Azov is poor in benthonic animals (three-fifths of the bottom has no living organisms). The limans of the Dnieper and the Bug are invaded by waterbloom composed of *Microcystis flos aquae* (a blue alga) at the end of July. Within a month it settles to form a sapropel, and the bottom becomes devoid of fauna, except for a few larvae of *Chironomus plumosus* (Knipowitsch, 1927; Issatchenko, 1934).

The Norwegian Fjords.—It has been shown (pp. 126–132) that one of the characteristics of the Norwegian fjords is an overdeepening by Quaternary glaciers. The fjords are therefore partially isolated from the open sea by a rock sill. In these basins, as in all estuaries into which the sea penetrates,

[1] These are the branching estuaries, closed by bars deposited by coastal currents in a tideless sea. The lagoons thus formed are slowly filled by fluvial debris filtered from the waters which flow through the bar into the sea.

there are two layers of water, resting on each other, which do not mix. At the bottom there is a layer of very dense salt water coming from the sea, and above it, fresh water flowing toward the sea. The salt water of the sea is thus trapped in the overdeepened basin. K. M. Strøm (1936–1937) has shown that in such conditions the salt water loses its oxygen progressively, the environment becomes anaerobic and hydrogen sulfide is formed by the decomposition of organic matter. To this is added a high concentration of phosphate (at least 700 mg. of P_2O_5 per cubic meter). The concentration of H_2S increases annually by 1 to 1·5 c.c. per liter. It has reached 40 c.c. per liter in the two fjords where it has been measured. Thus the conditions are favorable for the preservation of organic matter, which is deposited as a black mud containing 23·4% of organic carbon and 0·23% P_2O_5.

The presence of a sapropelic bottom inhibits aerobic life over the whole of the lower part of the basin. Sometimes movement of the waters may bring hydrogen sulfide to the surface and kill off the surface animals.

Certain littoral marine basins, such as Warnbro Sound (Carrigy, 1956), to the south of Fremantle (Western Australia) behave in a similar way to fjords. The entrance is closed by a line of reefs. The surroundings, due to wave attack, are covered with a layer of calcite sand on which colonies of Posidonias live. The center of the basin, about 30 to 60 feet deep, is calm, and its stratified waters are not disturbed by winter winds. Organic muds are thus deposited, and concentration of hydrogen sulfide occurs.

Bogheads.—Waterbloom also developed in the freshwater swamps of the coal forests. These were due to Botryococcus colonies of which the genus *Pila* flourished. They contributed to the formation of the bituminous shales, and particularly to the formation of *bogheads* or algal coals. Deflandre believes that these planktonic algae were derived from the Xanthomonadinae, some of which could secrete large quantities of oil, which could produce hydrocarbons. Bogheads are still being formed at the present time.

For example, in Australia, *coorongite*, which forms along the Coorong and on the floodplains to the north of the Murray River, is composed of myriads of *Botryococcus brauni* and other algae. This waterbloom becomes a solid, elastic substance as a result of desiccation (see p. 251, fig. 140).

Kukersite.—This is a very uncommon rock which occurs in the Baltic state of Estonia at Kukers, near to Jewe to the north of Lake Peipous. It is of Llandeilan (Ordovician) age.

Petrographically, it is a "Boghead" composed of the colonial unicellular algae, Protophyceae, or blue algae, being almost exclusively *Glaeocapsomorpha prisca* (Zalessky, 1917, 1920). This rock was used as a low-grade fuel during the last war (1939–1945). Asphalt can be extracted from it by distillation. It contains 65% C and 8% H, although the richest lenticles of asphaltite contain 84% C, 9% H, 7% O, N and S (Kogermann, 1933). The bed which is exploited is about 15 feet thick. The conditions of deposition of kukersite indicate an anaerobic environment unsuitable for living organisms.

The numerous fossils were apparently transported before being buried in the sediment. Nevertheless, the fauna is principally benthonic and very fragile. It seems, therefore, that they were coated with *Glaeocapsomorpha* at the point where periodic upwellings occurred.

"Marine Prairies"

The marine herbs (i.e. non-algal marine plants), are the most widespread group in the world and are found localized on the upper part of the continental shelf, at the infralittoral level. For more than ten years the authors (Termier and Termier, 1948) have directed attention to the fauna living among these herbs, since many juvenile forms and small organisms shelter there. It has been shown that the same features recur in a large number of pyritized fossil faunas. The authors have suggested an interpretation of the biology of these faunas and of the conditions of their fossilization *in situ*. The preservation of these faunas seems to have been the consequence of an anaerobic environment where pyrite can occur in association with hydrocarbons.

Present-day marine herbs, which were the object of only scattered observations, have, since 1953, been methodically studied in the Mediterranean by Molinier and Picard. The sediments which accompany them have been investigated by J. J. Blanc (1958), who has shown that they are coarse littoral deposits, generally poorly graded (sands, sometimes muddy). They have received substantial marine additions in the form of broken shells carried by currents. These sediments are fixed in place by algae and herbs, and in some places the accumulation amounts to 3 feet per century. The dead leaves coming from the autumn fall of Posidonias or Cymodoceae may pile up in banks which sometimes may lead to the formation of beach barriers (Punta d'Alga near Marsala in Sicily, and the Giens peninsula in Provence). Finally, the carpet of vegetation, or *matte* produced chiefly by Posidonias and their rhizomes (Molinier and Picard, 1953), retains a sand rich in organic debris which may be 8 feet thick. In spaces between the mattes there occur patches of decaying vegetation (*intermattes*). The formation of the mattes is closely related to the movements of currents. Mattes are interrupted by the channels at the mouths of small rivers (for example, the oueds of the Algerian coasts) or by currents as in the creeks of Provence. The sediments of the intermattes are also rich in organic matter. To the north of Marsala, these sediments are either shelly argillaceous sands containing small marine gastropods, or fine clays forming a bed at least 18 inches thick on which lives another biotope consisting of Cymodoceae and Caulerpa. It seems that the decaying intermattes which enclose transported dead shells (which constitute a thanatocoenosis) are formed in an environment where pyritization of shells can readily occur. Behind the mattes, which form "barrier reefs", are lagoons (paralagoons) where other non-algal plants live (Cymodoceae and Zostera). There the sediment is fine-grained and contains organic matter filtered out by the leaves of Posidonias of the "barrier reef"; the thickness of this sediment is only a few inches.

There remains one very important point. The argillaceous sands of the matte and also the clays of the intermatte and of the paralagoons are marine, but they are comparable to the soils of subaerial vegetation. Posidonia, Zostera and Cymodoceae are phanerogams held fast by their rhizomes, and colonies of these plants are comparable to a prairie. At the level of the rhizomes they probably produce humic acids, among other substances, capable of altering the composition of the sediments. These can also liberate a soluble migrant phase and leave in place a phase which is essentially argilo-siliceous.

12

Carbonate Sedimentation (General)

Most marine carbonates are secreted by animal and plant organisms (H. and G. Termier, 1952). A. Heim had thought that much was precipitated chemically in the form of drewite (an impalpable calcareous mud), because in Florida and the Bahamas, for example, such calcareous muds exist. However, in the laboratory, the purely chemical precipitation of aragonite is difficult and requires precise conditions of saturation, of pH, and of seeding.

The precipitation of carbonates by sulfate-reducing bacteria begins by a fall, followed by a rise in pH (C. Lalou, 1954). The precipitation of calcium depends mainly on the intensity of the process of respiration of, for example, *Microspira aestuarii* var. *tambi*, and occurs then in the form of small spheres covered with iron sulfide, and as concretions (B. Issatchenko, 1924).

Laboratory experiments have since shown that with small additions of basic reagents (alkali carbonates, ammonium carbonate plus ammonia, amine carbonates, etc.) the pH of a solution decreases, for example from 8·2 to 7·75, while large amounts cause the pH to rise. At the same time hydrated carbonates are precipitated which ultimately give rise to calcite. This explains why the biological processes capable of making an environment alkaline and then precipitating carbonates, at first cause a reduction in pH (G. Lucas, 1955).

Nesteroff (1956) systematically studied the organic residues remaining after the solution of carbonates. The fraction composed of bacteria was named by him *trame*, and the *substratum* was that material derived from the macro-organisms. Sometimes limestone existed only in the form of inclusions in the tissues. Organic traces are found in the ooliths of the Great Salt Lake, for example, and are due to the action of Cyanophyceae. The cement of beachrocks which are actively forming at the present day in the tidal zone (see p. 90), may also be due to the activity of bacteria. Each grain of sediment is, at first, coated with an amorphous film of lime, probably by bacteria which are active in the zone. This film subsequently crystallizes in the form of aragonite needles by physicochemical processes, and repetition of the process results in the filling of the pore spaces between the grains. The final deposit is always a calcareous film of bacterial origin. Finally the cement thus formed recrystallizes as calcite rhombehedra and the rock is then composed of a mosaic of crystals in which the original sandy texture of the beachrock is destroyed. The wave-cut bench composed of *Tenerea* of the Mediterranean coast shows an analogous cement between the algal fronds.

The Calcium Content of Sea Water

Sea water normally contains calcium ions derived from the continents. It is this which enables organisms to deposit or secrete the marine limestones. Along with all other soluble salts, calcium contributions are continuously made to the oceans by the rivers flowing into them. This material is derived by the solution of pre-existing limestones during river erosion, and also by the action of humic acids formed during the decomposition of organic matter. Calcium salts are liberated in this way by the erosion of podsols and laterites. It may be asked whether the amount of calcium carried by rivers has any influence on the deposition of $CaCO_3$. The theory of biorhexistasy, for example, attributes to it great importance. In fact, the continental shelf is the region where rivers discharge, where the waters are most actively agitated, and where the largest amount of carbonate is deposited. In closed seas the part played by rivers can be clearly seen; their role is particularly important in the Gulf of Kara-Bogaz (Caspian Sea). Marine currents, however, distribute mineral substances in the oceans; and consequently, the deposits of calcium salts are not restricted to the regions of river mouths, but will occur wherever the pH favors their precipitation. It appears possible that the great quantities of calcium carbonate required to produce the accumulation of the Chalk could be provided by the development of equatorial forest soils of modern type, during Cretaceous times. These soils could give rise to a migratory phase rich in soluble calcium salts.

The Calcium Carbonate Content of Marine Sands

Sands, sometimes entirely siliceous, which are carried by rivers to the sea are often reworked in the tidal zone. Many sandstones are, however, rich in calcium carbonate in the form of tests of calcareous organisms, and a carbonate cement. The proportion of carbonate may exceed 50% and increases progressively as the distance from the shore increases. This happens, for example, on the sandy west coast of Florida at the present time, and can be compared with the Permian sandstones of the Delaware Basin, U.S.A. (p. 245).

Certain Paleozoic stratigraphers have given to these marine calcareous sandstones the name graywacke, but this term is used in five or six different senses (p. 205), and such usage is deplorable.

Limestones of Organic Origin

Most carbonate rocks are formed in warm waters and are often associated with bioherms (reefs).

In considering the limestone muds associated with modern coral reefs, several possibilities can be envisaged. First of all, there is the frequently quoted example of the Bahamas and of Australia where the lime may be formed by chemical precipitation. In point of fact, it is possible that the breakup of green calcareous algae after death could provide the elements of the limestone muds in the form of aragonite needles (see p. 249). Furthermore, organisms, particularly fish, browsing on the coral may grind the calcareous part to a fine powder, which is excreted in the form of a limestone mud and deposited in lagoons and in the hollows of the reef. This is known to occur in the Red Sea (Nesteroff, 1955). On Bikini Island, however, there is no trace of lime mud (Marshall Islands) (Emery, Tracey and Ladd (1954)).

Chalk is formed by a different process. This friable nonrecrystallized rock is composed mainly of the shells of planktonic foraminifers and coccoliths and has not been converted by diagenesis into a more coherent rock. The small size of the organisms forming the chalk does not account for its powdery nature, since these differ little from those forming the compact limestones. It seems possible that the friable nature of the chalk was in some way linked with the almost constant occurrence of flint, which might produce an inequilibrium of pH and thus hinder the formation of a calcareous cement. To prove this hypothesis it would be necessary to compare chalk containing flint on the one hand with limestone and dolomites containing chert on the other. These latter are often very compact and are frequently silicified. It is suggested, therefore, that the absence of cement, and hence the activity of bacteria of the type described in connection with beachrocks, is probably due to the depth of deposition and to the diagenesis of the chalk. The chalk is believed to have formed in the lower part of the neritic zone, and thus was below the level at which photosynthesis could occur. During the Cretaceous, the enormous production of calcareous planktonic muds seems to have modified the calcium carbonate economy of the surface of the globe (Kuenen, 1950). During this period the limestones were deposited on the lower parts of the continental shelf and could not be kept in circulation by erosion. They thus accumulated and were removed from the calcium carbonate cycle. The only bathymetric limit imposed on such sediments is that due to the solubility of the limestone. It should be noted that it was in the Late Cretaceous that the ahermatypic corals (which make use of calcium carbonate) of deep seas appeared (see pp. 264–267).

From this, it seems that, during the periods when the sea level was low and consequently the area of the continental shelf was much reduced, carbonate rocks were deposited in only small amounts. In particular the bioherms (p. 267) are much reduced. The oscillations of sea level in the course of geological time can, moreover, be demonstrated by the variation in the amount of limestone in the sedimentary deposits (R. W. Fairbridge, 1955).

The genetic classification of organic limestones proposed by Fairbridge (1955) may, thus, be adopted:

(1) On the *continental shelf*, between depths of 0 and 650 feet or exceptionally, nearly 3,000 feet, conditions are favorable for the deposition of limestone, especially in warm water between 50° F. and 85° F. There are zones of clear water, rich in nutrients, and particularly favorable to living organisms, in which occur *bioherms*, *biostromes* (p. 262) and clastic carbonate deposits (oolites, beachrock, windborne limestone debris, and *pelagosites* deposited from sea water supersaturated with lime).

(2) In the *bathyal* and *abyssal* (continental slope and deep sea) regions between 650 and 18,000 feet in warm climates, and in open oxygenated basins, calcareous deposits are also formed. These include Globigerina and Pteropod oozes, with coccoliths and rhabdoliths. At great depths, the high concentration of carbon dioxide prevents the accumulation of carbonates. In closed and poorly ventilated basins anaerobic fermentation occurs, and a "sulphuretum" is produced. In cold regions, microplankton give rise to siliceous oozes.

Chalk belongs to an intermediate category consisting of pelagic sediments similar to the Globigerina and coccolith oozes, but occurring on the deeper part of the continental shelf. According to G. Millot, chalk forms part of the sedimentary cycle proposed by Erhart (see p. 356).

Some Examples of Detrital Limestone Sedimentation

Florida.—On the west coasts of Florida (Gould and Stewart, 1955) the deposits which are forming at the present day are unconsolidated sediments showing six zones parallel to the coast. They do not, however, form a continuous cover over the bottom.

The quartzose sands of the innermost zone contain grains of phosphorite and represent a reworked sandstone deposited during the first phase of the Pleistocene. This zone is enlarged opposite Tampa Bay and Charlotte Harbor due to the influence of marine currents.

Toward the open sea, the quartz gradually becomes less abundant and finally disappears, while the proportion of broken shell increases. The successive zones are as follows: a zone of shell debris, a zone of algal sand, a zone of oolitic sand, and, finally a zone of sand and silt composed of foraminifers. The limestone is most abundant in the deepest water. In the bays, the terrigenous sediments are very fine-grained and less rich in shells than in the inlets and the outer faces of the barrier islands.

The Delaware Permian Basin.—In this basin, which subsided in the area of western Texas and New Mexico and shows a complete Permian succession, three regions can be distinguished. From south to north, they are (1) the basin proper, (2) an intermediate zone, (3) a continental shelf.

In the intermediate zone, first biostromes and then bioherms developed on the slope, forming reefs of which the best known is the Capitan reef. These reefs are flanked by detrital limestones, *calcarenites* resembling the "beachrocks" in part.

Finally, to the north, on the continental shelf, in the region now occupied by the Guadalupe Mountains, the sedimentation was, it seems, comparable with that occurring at the present day on the west coast of Florida. The following types of sediment can be seen:

(*a*) *Calcarenites and dolomitized coquinas*, formed from reef and subreef material. This facies, which represents the immediate surroundings of the reef, encloses the remains of green algae belonging to the genera *Gymnocodium*, *Macroporella*, and *Mizzia*, the latter being the most abundant. Fusulinids, equally frequent, are accompanied by gastropods, while the small foraminifers which occur are often broken. Nearly all the tests are recrystallized. Small spherulites of calcite and dolomite (0·003 to 0·1 mm.) are also found.

(*b*) *Dolomitic and calcitic pisolites* also were formed, particularly during the period corresponding to the Carlsbad stage. The pisolites are 1·5 to 31 mm. in diameter, generally in the form of flattened ellipsoids; their concentric layers enclose a grain of detrital quartz. Sometimes, they show protrusions and hollows. They are enclosed in a groundmass of calcite and dolomite. Many of the fossils occur as nuclei of these pisolites. This facies is associated with the preceding one and is found at some distance from the reef. Most of the pisolites are believed by Ruedemann (1929) and Johnson (1942) to be of algal origin. According to Johnson, they were Cyanophyceae. Moreover, the conditions of deposition of these concretions correspond with the salinity, temperature, clarity and depth of water of living algae.

(*c*) *Fine-grained dolomites.*—About one thousand yards from the reef, a hard porcellaneous dolomite, mahogany to white in color, thin-bedded and without fossils, passes, on the one hand, into a calcarenite, and on the other into a gypsiferous bed. The indefinite outlines of fusulinids and other fossils, the structures of which have been destroyed by dolomitization, can be recognized.

(*d*) *Evaporites.*—Anhydrite is deposited several miles from the reef. The type of sedimentation giving rise to this rock seems analogous to that of the Gulf of Kara-Bogaz (Caspian Sea) which is a large basin of evaporation (see p. 243). In the case of the Permian Delaware Basin, however, the anhydrite is closely connected with the development of a barrier reef. This reef, incidentally, is not rich in corals.

(*e*) *Quartzose sandstone.*—These sandstones are associated with red shales. They have little cement and are poorly stratified. They contain 27–80% of quartz (57% on average) in the form of grains in a dolomite matrix. Most of the grains are small, angular, sometimes elongated and slightly frosted. They also contain a few large, rounded, frosted grains which are believed to have come from the shores of a lagoon. It appears that the periods of extension of the sandy sediments coincided with a diminution in the growth of the reef. The diminution may be due to a deepening of the basin.

When an alternation of quartzitic sandstones and dolomite occurs, the vertical passage from one bed to another is as follows: the base of the sandstone is always sharp without any trace of irregularity, while at its top it passes imperceptibly into a sandy dolomite.

THE ROLE OF ORGANISMS IN CARBONATE SEDIMENTATION

As was mentioned at the beginning of this chapter, an important aspect of marine sedimentation is the precipitation which produces a number of nondetrital deposits. A large part of these deposits is due to biochemical activity. This is true of carbonates generally, and is particularly true of limestones. The sea does not have the monopoly on organic limestones: they also form in fresh waters.

The Precipitation of Limestone in the Presence of Chlorophyll

The precipitation of calcium carbonate from bicarbonate solutions is a result of photosynthesis. The chemical reaction is, in effect:

$$Ca(HCO_3)_2 \rightleftharpoons CaCO_3 \downarrow + H_2O + CO_2 \nearrow.$$

The carbon dioxide which is liberated is immediately utilized by plants, and the calcium carbonate is precipitated. This precipitation is stimulated by a rise in pH and in temperature. It has already been shown that this occurs in eutrophic lakes (p. 173) principally due to the action of phytoplankton (Coccolithophoridae and even siliceous diatoms). It follows that certain plants are encrusted with calcium carbonate, some accidentally (such as the stems of reeds and the leaves of certain Bryophytes, for example *Myriophyllum*, *Elodea*, and *Potomageton*), while others (such as *Chara* and a great number of algae) always receive a coating. "Among the encrusted Cyanophyceae, the crystals generally remain fixed to the surface, the whole thallus thus becoming hard and compact" (Symoens, Duvigeaud and Van der Berghen, 1951). The mucous coating of the filaments of Schizophytes undoubtedly assists in the fixation of the calcium carbonate. In *Oscillatoria* and the Bacillariophyaceae, the crystals surround the filaments to give moniliform appearance. The Dasycladaceae, the Codiaceae and the Rhodophyceae are also encrusted in characteristic fashion. The phenomenon is widespread, and can be observed in freshwater lakes, in brackish lagoons, in alkaline peat bogs, on the beds of streams, in petrifying springs and in shallow seas.

Most *travertines* are limestones precipitated in the presence of chlorophyll in fresh waters. They often form important structures built up of successive layers. On hillsides below springs, under a forest cover in Belgium and Lorraine, travertines (*crons*) are formed by moss (*Cratoneurum*) and by Cyanophyceae (*Lyngbya*). They also dam up streams with the aid of other plants (particularly grasses) and give rise to series of waterfalls. As the limestone accumulates it attacks the bases of trees and ultimately destroys the forest.

In the semiarid Mediterranean climate of the south of Spain, the Balearic islands and in Morocco (H. Termier, 1936, p. 935) travertines are

FIG. 139. TRAVERTINE FROM A SPRING, AIN OUM ER-RBIA, MOROCCO, FORMING A MANTLE WHICH HAS ENCRUSTED THE SOIL AND IS KILLING OFF THE VEGETATION (Photograph: H. Termier)

often formed by water emerging from Jurassic and Cretaceous limestones (fig. 139). They occur especially, at the exits of underground streams. Large pisolites occur around Fès and to the east of Azrou (Ras el Ma, Morocco).

Limestones precipitated in the sea by algae are generally magnesian (see p. 343).

Limestones Ascribed to Chemical Precipitation

Some of the present consolidated limestones seem to lack recognizable organic debris. It has been suggested that they originate in an entirely inorganic, chemical process of sedimentation.

Among modern sediments, there are calcite muds (calcilutites) which could have formed in this way. They occur, for example, in the Bahamas (in particular, off Andros Island), in Florida (Florida Key between the external reefs and the Keys), and in the Society Islands (at Maiao).

Black (1933) following Kindle (1923) gave the name *drewite*, to these muds. This name seems to be appropriate for impalpable lime muds. The precipitation, in the form of aragonite, occurs because calcium carbonate is supersaturated in warm tropical seas fed by cool lime-rich currents. Nevertheless, as early as 1913, Drew supposed that the precipitation was of bacterial origin, and in 1932, Bavendamm, using bacteria from soils and from estuarine muds, showed that this was possible. His experimental work paralleled the conditions which occur in the Bahamas, in lagoons, mangrove swamps and in marshes. This confirms the comments made at the beginning of this chapter (p. 243).

In all the natural cases quoted, aragonite is formed as very fine needles (0·003 to 0·005 mm.) behaving in mass like a clay. It occurs at the indistinct boundary between land and sea in marshes, swamps and lagoons. It is possible that accumulation of these needles is assisted by algae living in the same areas and acting as sediment traps. Black (1933) and Young (1935) believe that the same process occurs in the formation of certain Stromatoliths (p. 250).

On the other hand, Lowenstam (1955) has shown that in the same deposits algae themselves produce needles of aragonite. The algae are mostly Codiaceae, but Dasycladaceae, Nemalionaceae and Chaetangiaceae also occur. Closely interlocked with the living algae, the needles separate on death either by the chemical action of bacteria or by holothurians feeding on the algae and then rejecting the inorganic residue. This appears to explain the origin of the modern limestones supposed to result from chemical precipitation.

Among the older rocks, the lithographic limestones may come into this category. This is the interpretation which Cloud and Barnes (1948) place upon the Ellenberger Group of central Texas, which is of Early Ordovician age. The rocks are fine-grained (0·002 to 0·005 mm.) and of the same order of size as the aragonite needles. They contain Stromatoliths, suggesting dolomitic episodes (probably secondary), and enclose cherts and rolled sponges.

The Role of Blue-Green Algae (Cyanophyceae)

Blue algae often form nodules in modern lakes and are known as far north as the Shetland Isles. They often enter into the composition of waterbloom. According to the climate and the composition of the environment in which they live, their sedimentological role is directed towards the deposition of calcium carbonate, of silica, or simply organic matter. Non-detrital, aquatic sedimentation on the borders of continents appears, therefore, to be principally controlled by the Schizophytes. Certain blue-green Myxophytes live in salt marshes; for example, in England, *Microcoleus*. In the playas of western Australia, allied forms fix and mold the bottom sediment, as in the case of the red mud of Lake Cowan (Clarke and Teichert, 1946). In fresh waters, travertines are often the work of Cyanophyceae, such as *Rivularia hematites*.

The section of a modern travertine crust in the Hautmont woods in Belgium was studied by Symoens (1949). It showed a surface layer with filaments of Cyanophyceae (*Lyngbya*, *Phormidium*, *Gongrosira*) and an internal layer stratified by discontinuities probably due to temporary emergence.

Algal Biscuits and Stromatoliths (figs. 140–145).—In some regions calcareous concretions associated with algae, and in particular blue algae, are formed which are given the name of "biscuits" because of their appearance and dimensions.

The best-known examples are those found southeast of Adelaide, South Australia (Mawson, 1929). They occur in a depression among the coastal dunes, situated above the water table in summer, but submerged in winter. This area, with very sparse vegetation, receives 3 inches of water in the eight or nine months of the humid season, but dries out completely in summer. The "biscuits" resemble those formed by Lithothamnion found also in Lake Karatta where they form reefs (figs. 140–141).

These "biscuits", formed in slightly saline water, show no cellular structure and are composed of aragonite. It may be noted here, that calcareous algae such as *Melobesia* form calcite. Chemical analysis of the "biscuits" shows: 43·54–51·75% CaO, 7·3–1·8% MgO, 38·43–42·73% CO_2, a trace of P_2O_5, 0·13–0·28% Fe_2O_3 + FeO, 0·17–0·02% Al_2O_3. They also contain an organic residue composed of the blue algae, *Glaeocapsa* and *cf. Schizothrix fusciculata*.

Immediately to the northwest of the region of the "biscuits", the lagoon of Coorong contains dolomite marls which seem to have been produced by precipitation by abundant micro-organisms. This sediment is known as *coorongite* (see p. 239).

The introduction of artificial drainage in the region of the "biscuits" is slowly causing the disappearance in Australia of the conditions necessary for the growth of the algae which produce them. There are, however, other regions where they seem to be forming; for example, on the coasts of Yucatan and in Bolivia.

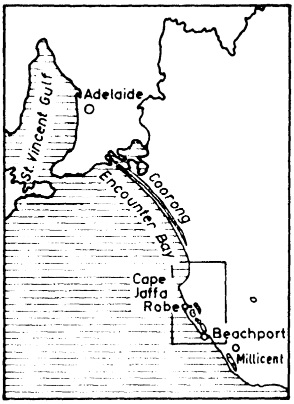
The bottoms of certain water courses in temperate regions show similar nodules. The River Bourne (Cambridgeshire, England), which dried out in 1943, contained concretions 1 to 8 inches in diameter. These enclosed a shell, a fragment of wood or a pebble and were formed by Cyanophyceae (*Phormidium* and a few *Gongrosira*). The filaments of these algae are coated with mucilage and trapped particles of loam and vegetable debris (Fritsch and Pantin, 1946).

"Biscuits" of similar origin are also formed in abundance in arms of the sea which have temporarily dried out on the borders of the St. Lawrence river (Canada).

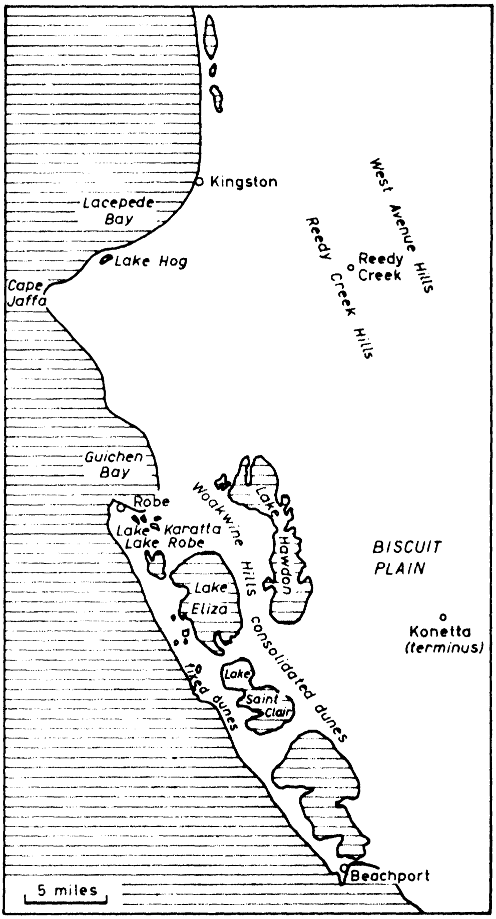
In the Permian (Kansas and Texas) the Porostromatal alga *Ottonosia* forms similar concretions.

In the Dogger (of Morocco and Normandy) marine bryozoans are associated with the "biscuits".

In all cases, the conditions are well defined and those which are important for the formation of stromatoliths should be noted. They are similar



FIGS. 140 and 141. THE LAKES AND LAGOONS WHERE ALGAL "BISCUITS" OCCUR IN SOUTH AUSTRALIA



to those which occur in playas: pH of the water, warmth, and temporary drying-out. It is possible that it was in this environment that life on the continents may have begun (figs. 142–145).

Marine Algal Limestones and Oolitic Limestones

The major part of the nondetrital limestones are those secreted either by Schizophytes (bacteria or Cyanophyceae) or by the red or green algae. It is of interest to review the respective characteristics of limestones formed by these organisms (H. Johnson).

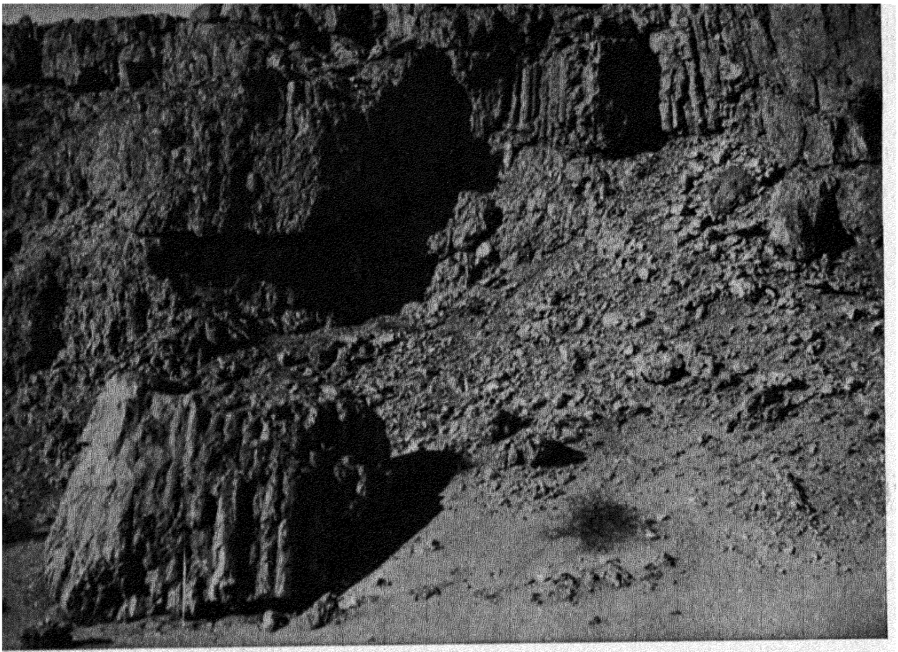


Fig. 142a. Stromatoliths (Conophyton) of El Hank, Mauritania: Eocambrian (Photograph: Menchikoff)

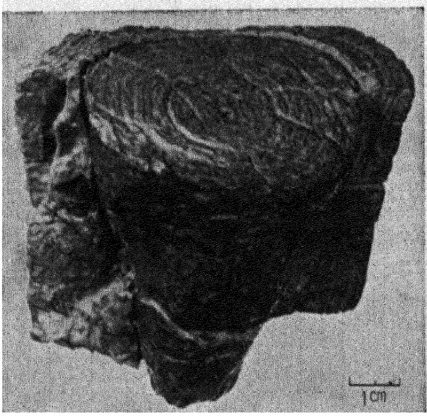


Fig. 142b. Stromatolith: Side View showing the Cone Structure (Photograph: J. Leriche)

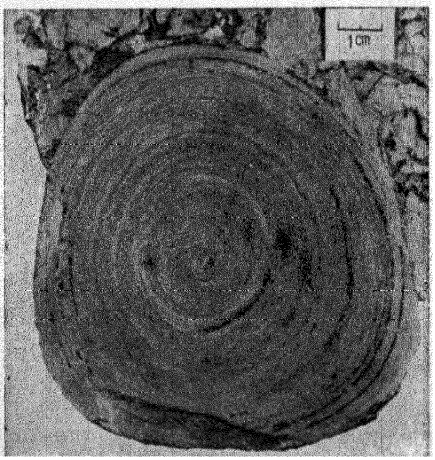


Fig. 142c. Stromatolith: Section Perpendicular to the Axis showing the Concentric Layers of Calcite (Photograph: J. Leriche)

The red algae, Melobesiae, are the main builders of the Lithothamnion limestones which have the following composition: 63–89·72% $CaCO_3$, 3·76–25·17% $MgCO_3$, 0·07–3·5% SiO_2, 0·01–1.62% $(Al,Fe)_2O_3$, 0–0·42% $Ca_3P_2O_8$, 0·03–1·39% $CaSO_4$, 0·35–11·44% of organic matter. The growth of these algae amounts to 2–7 mm. per year, with a maximum of 0·5–1 mm. per month in summer, in France, while in Samoa they grow nearly 24 mm.

FIG. 143. STROMATOLITH AND ARCHAEOCYATHA BIOSTROME. LOWER CAMBRIAN OF THE SONORA DESERT, MEXICO (Photograph: G. Termier)

(nearly 1 inch) and in the Maldive Islands about 26·5 mm. per year. Corals grow even more rapidly; in France 3–4 cms. (1¼–1½ inches) per year and at a much greater rate in the tropics.

The calcareous green algae are the Codiaceae (among which *Halimeda* represents an important fraction of the material of modern coral reefs), the Dasycladaceae (for example, *Acetabularia*, fig. 145) and the Characeae. The first two families are well known as far back as the Ordovician and have been responsible for the building of massive limestone structures, particularly in the Triassic. The Characeae, which are lacustrine, are also limestone builders.

The most productive builders are the algae which not only produce an encrustation on the thallus, but also precipitate limestone around themselves. In this category are the Porostromata (*Girvanella*, *Ottonosia* and *Somphospongia*) which are found as fossils and were probably green algae. The best examples, however, are the *Cyanophyceae* or blue algae (see p. 249). They play an important role in the formation of marine limestones, as well as in the framework of algal "biscuits". It will be recalled that the Cyanophyceae and the bacteria are very resistant to salinity and temperature

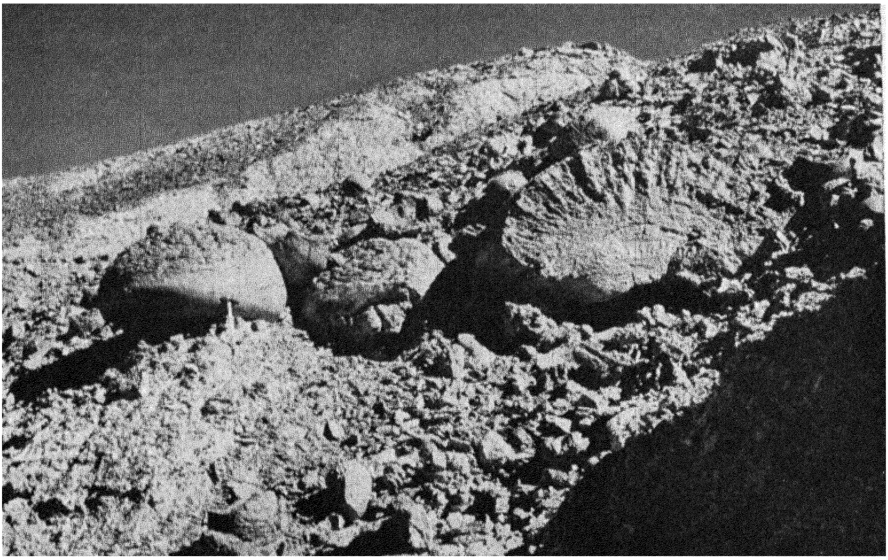


FIG. 144. COLLENIA OF THE UPPER VISEAN (MISSISSIPPIAN) OF THE FEZZAN, NORTH AFRICA (Photograph: Freulon)

change, and can survive long periods of aridity. The travertines deposited by springs may, in fact, be formed by them. It is certain that most of the calcareous ooliths owe their existence to them. The limestones secreted by the Schizophytes have all been formed in very shallow water, often in the tidal zone.

The calcareous oolites generally have a nucleus consisting of a grain of detrital quartz, a small foraminifer or a fragment of shell. This nucleus is coated with concentric layers of aragonite needles arranged parallel to the surface of the oolite, the layers being clearly separated from each other. This has been observed in the Bahamas by Illing, and at Djerba in Tunisia by G. Lucas (1954). It has been shown (p. 242) that similar pellicles occur on an organic substratum in the oolites of the Great Salt Lake. The calcareous oolites are often perforated or enveloped by algae of the Girvanella type

(Dangeard, 1952–1953; Lucas, 1954). At Djerba these oolites are cemented by calcite and have themselves recrystallized as calcite. At the present time only a few places are known where oolites are being formed:

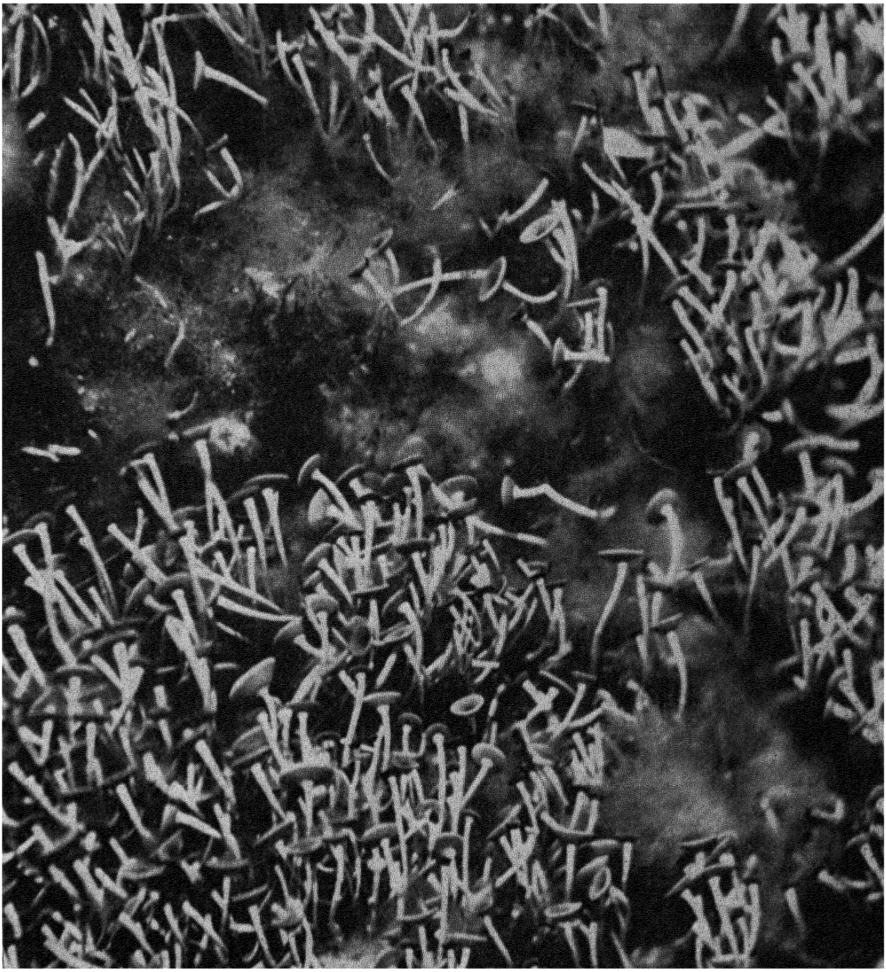


FIG. 145. FIELD OF ACETABULARIA ON THE CONTEMPORARY LITTORAL PLATFORM, NEAR TIPASA, ALGERIA (Photograph: G. Termier)

1. In a marine environment, in the Bahamas, on the Florida coasts, in the Gulf of Suez, on the Egyptian coast near Alexandria (Hilmy, 1951), at Djerba, Gulf of Gabes.

2. In salt lakes; the Great Salt Lake and in the Aral Sea (Brodskala, 1939).

The composition of the oolites of the Great Salt Lake is as follows: 84% $CaCO_3$, 5·5% $CaCO_3.MgCO_3$ and 5·6% fine mud (Eardley, 1938).

The synthesis of calcareous oolites has been achieved by Monaghan and Lytle (1956). They found it necessary, however, to increase the concentration of the carbonate ion in the water to 0·002 mol (gramme-molecule) per liter, which is very much greater than that of normal sea water. The environment where sulfate-reducing bacteria appear to be important in the origin of oolites is characterized by a deficiency of oxygen, a high pH, and the presence of H_2S and sulfides. Such an environment is likely to be very much localized. It occurs where carbonates are plentiful; for example, at the mouths of rivers, although no oolites can be seen there. The presence of magnesium in the sea causes the calcium carbonate to be precipitated as aragonite, rather than as calcite.

In calcareous oolites, diagenesis has transformed the aragonite into calcite and a second, radiating, structure is superimposed on the concentric structure. Prior to cementation, the oolites can be reworked by marine currents, or by the wind, and deposited in places other than those in which they were formed.

Oolitic limestones are common in the geological record. They occur in the Precambrian associated with *Collenia* and they have been observed in the Lower Cambrian of Morocco at Ait Anzal, associated with *Archaeocyatha* limestones (H. and G. Termier, 1947). They are also found in the Visean of Europe and North Africa, but it is in the Jurassic that this facies is best developed. For this reason William Smith, and then E. Haug, united all the stages from the Bajocian to the Portlandian into a single Oolitic subsystem.

OTHER LIMESTONE-ACCUMULATING PLANTS.—Mosses seem to be able to extract calcium carbonate from fresh water. This may account for the calcareous tufas of the valley of Hoyoux (J. J. Symoens, 1950) (see p. 247).

Globigerina and Pteropod Oozes

Globigerina oozes are frequently deposited in warm seas at depths less than about 13,000 feet and are often accompanied by Pteropods. These deposits are of pelagic origin, and although in no way connected with reef conditions, they have much in common with them, in particular the temperature and the clarity of the surface waters. Furthermore, the areas colonized by coral reefs coincide approximately with the areas of deposition of the Globigerina ooze.

Sediment of this type is found, for example, in the region of the Marshall Islands. The floor of the Red Sea is covered by oozes composed of foraminifers and pteropods (Nesteroff, 1955). Close to the reefs, the oozes contain detrital material resulting from erosion of the reef itself.

13

Reefs, Biostromés and Bioherms

Only the more important organic structures built of calcium carbonate are noted here (E. R. Cumings, 1932). *Biostromes* are composed of stratified beds containing whole organisms which lived *in situ*. If these have been transported they form a *thanatocoenosis*.

The *bioherms*[1] are generally without bedding and form mounds or domes rising above the contemporaneous beds surrounding them. They are formed of the remains of various organisms according to the temperature and the environment, and correspond closely to the ecological idea of a "reef". The rate of construction of bioherms is much greater than the rate of sedimentation in the areas between the reefs. This difference explains the dip of the beds on the margins of bioherms.

Nevertheless, stratification can be seen in reef lenses (which are, by definition, bioherms) in the Frasnian of Belgium. In this case, subsidence will account for the form of this mass of organic limestone (see p. 288).

The difference between the biostrome and the bioherm appears to lie in the relative movement of the sea floor, and in the case of the bioherm, it is as important as the rate of growth of the organisms.

The theory of Darwin, according to which subsidence is the fundamental factor in the increase in thickness of coral reefs is, moreover, invoked frequently by modern authors (Kuenen, 1933; Lecompte, 1954, 1956; Rutten, 1956; Dubourdieu, 1957).

I. NONCORALLINE ACCUMULATIONS: ROCK LEDGES AND WAVE-CUT BENCHES[2] (figs. 146–149)

In 1854, de Quatrefages described "wave-cut benches" which he considered, wrongly, to have been constructed by gastropods, whereas they are only covered by them. The structure was first observed in Sicily near to Torre del Isola, near Palermo.

The existence of wave-cut benches has been recognized in a number of places all over the world: Banyuls (on the edge of the Pyrenees), the

[1] From ἕρμα, ἕρματος (= reef, rock).

[2] The authors use the term "trottoir" = wave-cut bench, or narrow platform cut by the waves.

island of Pont-Cros (sandstones of Tuf point), Bandol (where they occur on the soft sandstone of Renecros), Cape Couronne to the west of Marseilles (Miocene Molasse), on the Algerian coast in the consolidated dunes near Algiers (at Douaouda, Fouka, Chenoua bay, Tipasa), on the northern coast of Tunisia, in Sicily (the sandstone of Palerma and Formentara and the Capucini (Syracuse limestone). They also occur on the Atlantic coast of Morocco (Rabat, Temara).

As L. G. Seurat remarked in 1935: "A character common to these occurrences is their division into an emergent portion, rising more or less just above sea level, battered by waves and corresponding to the upper part of

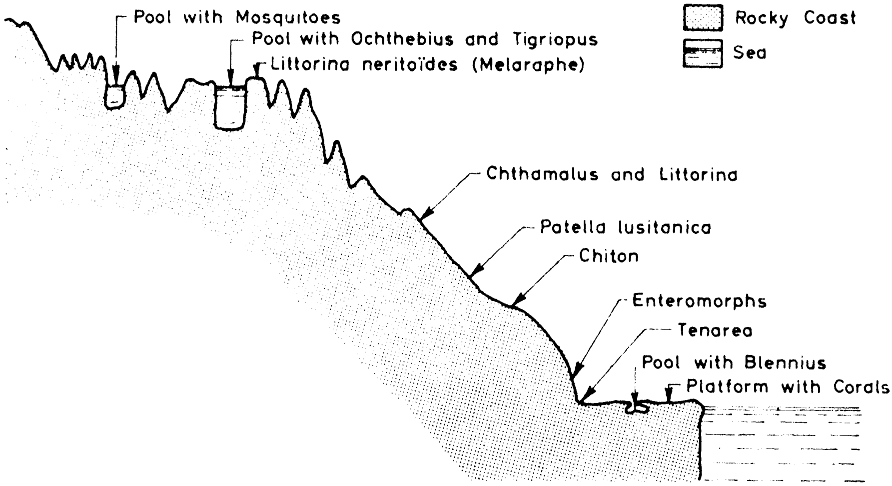


FIG. 146. THE LITTORAL ZONE OF ALGERIA, SHOWING THE WAVE-CUT BENCH-PLATFORM WITH CORALS (after Seurat)

the intertidal zone, and a part almost always submerged. The two parts are often joined by a horizontal platform, swept by the waves, which is uncovered at low tide. In some cases the emergent part of the cliff rises almost vertically, in others it forms a gentle slope with supralittoral basins swept by waves at high tide."

The supralittoral part is here the site of intense alveolar erosion (see p. 99) which might at first sight be attributed to the impact of waves. J. M. Pérès and J. Picard (1952) think that this erosion is not purely mechanical, but also physico-chemical, since "the small pools show marked variation in salinity, temperature and pH". This zone, bristling with pinnacles, normally receives only spray, except in times of storms. This is the habitat of *Littorinas*, certain Patellas, the cirrepede *Chthamalus*, Chitons, Mytilids and polychaete annelids, and is accompanied at a slightly lower level, by algae (*Lithophyllum*, *Verrucaria*, *Tenerea*, *Ulves*, *Gelidium*). In the pools crabs are abundant. In this zone erosion is rapid and the emergent rocks recede rapidly.

The part corresponding more or less to the mean sea level, the *wave-cut bench*, is a narrow platform (3 to 30 feet wide) where the waves break. This

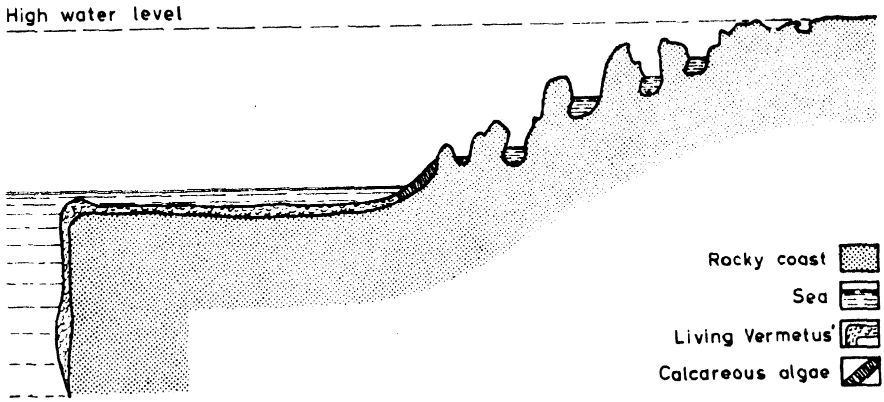


FIG. 147. LITTORAL ZONE OF THE CALCARENITE COAST OF TIPASA, ALGERIA (after Pérès and Picard, 1952)

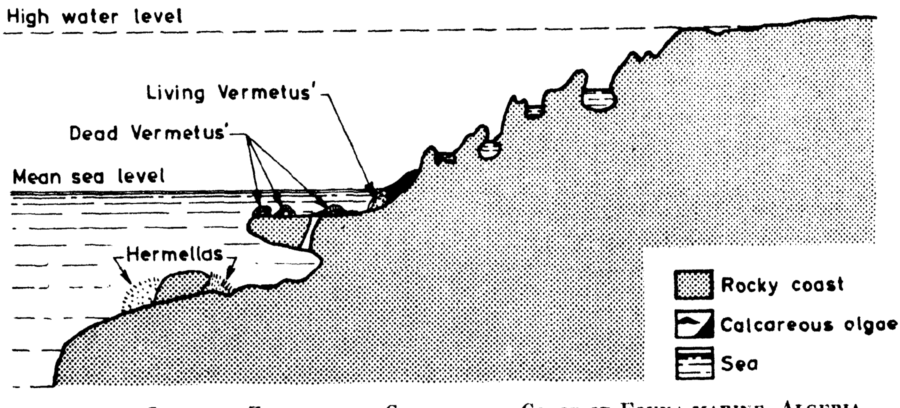


FIG. 148. LITTORAL ZONE OF THE CALCARENITE COAST OF FOUKA-MARINE, ALGERIA (after Pérès and Picard, 1952)

is covered by a thick carpet of algae which are mostly calcareous (*Jania*, *Falkenbergia*, *Ceramium*, *Cladophora*, *Cystoseira*, *Laurencia*) and among which live echinoids (*Paracentrotus*, *Arbacia*), actinids, crabs (*Carcinus*), hermit crabs (*Calcinus*), gastropods (*Cerithium*, *Oncidiella*, *Fossarus*, *Gadinia*) and fish (Blennies). The edge of this platform, washed by the sea, is often encrusted with *Vermetus*, with which occur the sponges, *Clione* and *Pomatoceras*, algae (*Jania*, *Cystoseira*, *Cladophora*) and polychaetes, Melobesiae (*Tenerea*, *Lithothamnium*) which fill up the tubes of boring pelecypods

(Lithodomes, *Cardita*), Chama, Patellas and sipunculids (*Physcosoma*), hydroids (*Plumularia*, *Sertularella*) and bryozoans. At Tipasa in Algeria and at Torre del Isola in Sicily, Vermetus hinders erosion, for this gastropod forms a dense mass several inches thick. Where the growth of Vermetus is rendered impossible, often by impurities in the water, marine erosion can attack the crust by undermining it. Such is the case between Castiglione and Fouka (Algeria) where the water is polluted by dissolved nitrates (J. M. Pérès and J. Picard, 1952). Elsewhere, Vermetus develops directly on the

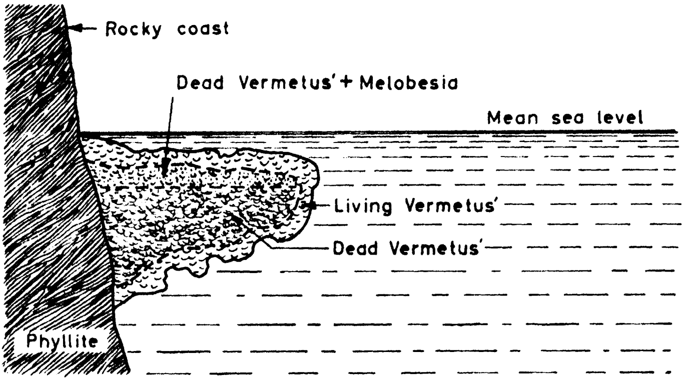


FIG. 149. THE "CORNICHE", OR ORGANIC LEDGE AT CENTUREI, CORSICA (after Molinier, 1955)

rocks instead of on a previously formed erosion platform. This has happened at Centuri (Corsica) where, according to R. Molinier (1955), they have built a horizontal ledge jutting out 12 inches from a crystalline base.

This type of coastal cornice, or rock ledge, is thus a phenomenon of marine abrasion: the sea at this point gains on the land and the biotopes which develop there contain those species which are best adapted to an amphibious existence (see p. 60–65).

In the case of platforms the encrustation of Vermetus is thickest on the outer edge, which is more exposed to the waves. It follows that they form a rim which retains the water even when the tide is abnormally low (Molinier and Picard, 1953). These authors have observed that Vermetus builds ridges over the joints in the littoral rocks, thus cutting the platform into a series of small basins. These basins also retain a shallow layer of water.

In Sicily, the edge of the zone of Vermetus reaches mean sea level. The organisms, therefore, no longer live in the most favorable environment, and are quickly covered by an encrusting layer of Lithothamnium which serves to support cushions of *Tenerea* (characteristic of the algal-covered benches) accompanied by *Patella*, *Fossarus*, *Oncidiella*, *Lasaea* and *Brachydontes*.

The Lithothamnium bench is liable to be eroded by the action of Chitons (*Middendorfia*) which burrows into it. This occurs frequently in the Adriatic and at Marsala (Sicily).

The Mediterranean bench is colonized chiefly by Melobesiae, which live at about the same mean sea level, just above the Vermetus. Outside the Mediterranean algal benches have been observed in the Marianas (Guam, Tinian, Saipan) and in many other places.

In warm seas the littoral cornice is better adapted to the Lithothamnium facies, and forms the "zone of Nullipores". The necessary condition for the growth of Melobesia is the frequent renewal of the water. Therefore, the calcareous algae are favored by the strong surf round islands, and at the points of headlands. In calm waters, such as those found frequently in the East Indies and in the Red Sea, these algae do not flourish. The reefs subjected to the surf have indented borders, as on Funafuti and in the East Indies. They are cut by narrow ramifying channels which terminate toward the land in blowholes. The Nullipores occupy the shallow basins on the terraces which are comparable to those of the Mediterranean benches. The best examples in the East Indies are on the west coast of Sumatra, the fossil dunes of the south coast of Java, the Nenoesa Islands, the east coast of Karekelong (Talaud Islands) and on the north coast of the Schouten Islands.

The Lithothamnium bench is known, in the fossil state, in the upper Tertiary of Rembang, Java (Martin, 1911).

The Rhodophyceae are relatively susceptible to strong sunlight, which dries them out. Thus they live under the cover of brown algae which are more resistant to dry conditions, or under barnacles. They can occur as much as 10 feet above the low tide level provided they are protected from desiccation as they are in the English Channel and in the Bay of Fundy between New Brunswick and Nova Scotia (Canada). In the tidal zone they are mostly encrusting forms. Sometimes they are of the ramifying type which is found on the coasts of Normandy and Brittany. Others are of nodular type and occur at Bikini in the Marshall Islands and at Haingsisi, southwest Timor, where the sea is rough.

The ramifying type, when subjected to rough water conditions, is transformed by the waves into rounded nodules. By alternate exposure of all sides of the nodule to light, these algae become uniformly colored. The salinity may vary but a considerable amount of freshening can be tolerated by the Corallinaceae.

The low tide level controls the upper limit of growth of these algae. Toward the bottom they form an uninterrupted cornice (or overhanging ledge) round the reefs at a depth of about 80 to 115 feet. The optimum zone of growth for the red calcareous algae is the sublittoral zone, which descends to 30–65 feet. However, they are sometimes found as deep as 800 feet, such as around the Funafuti atoll. The temperature of the water never seems to limit the depth, though warm seas are the most favorable. They are also found in banks along the Norwegian coast in Trondheim Fjord, around Iceland, Spitzbergen, in the White Sea, and off Greenland. As the depth increases, the algae decrease in size and the calcareous crusts are much thinner.

Oyster Bioherms (fig. 150)

In many seas, oysters build banks which are comparable to biostromes, or even to bioherms. Since oysters can tolerate variations in salinity these banks are often found in estuaries and in lagoons. The banks are mobile and may completely disperse. For example, the oyster banks (*Ostrea edulis*) near Heligoland (North Sea) described by Möbius (1883, 1893) have completely disappeared (Caspers, 1950) and have been replaced by a *Nucula nucleus* biotope living on the sediments accumulated by the oysters.

The conditions for the establishment and growth of oyster bioherms are known to be a temperature of 50° to 77° F., a salinity of 1·0 to 3·0%, a floor of mud or sandy mud, and shallow water. Such conditions are found along the east coast of North America, from the St. Lawrence River nearly to Mexico. The oyster banks tend to be abrupt and ovoid, with the central part occupied by dead shells and the flanks by the living oysters. They form barriers, the long axes of which are normally perpendicular to the dominant currents. These reefs commonly occur, particularly in waters of normal salinity, on a hard subtratum of other organisms: ophiurians (*Ophiothrix*), actinians, holothurians, mussels (*Brachidontes*, particularly in brackish water), crabs, various pelecypods (*Chione*) and gastropods (*Neritina*, *Haminoea*, *Bulla*). A number of parasites also occur (the fungus *Dermocystidium*, the Gregarinida *Nematopsis* and the trematode *Bucephalus*), predators (the polyclade *Styloches*, the crabs *Callinectes* and *Menippe*), the gastropods *Thais* and *Odostomia*, and finally the shell borers (the sponge *Cliona*, the lamellibranch *Martesia*, the polychaete *Polydora*). In winter the bryozoans and the tunicates constitute an epifauna of the oysters, but they disappear in the summer. *Crepidula* are moderately abundant (Hedgpeth, 1953).

In Atchafalaya Bay (fig. 150) the oyster banks filter the water only allowing the deposition of fine silt and clay (which is mainly chloritic) with less than 1% of sand. The deficiency of sand in front of the reef causes sand to be taken from a barrier isle and reduces the abrasive role of coastal currents.

The destruction of an oyster bank is the beginning of a normal sediment, though rather a peculiar one. For example, in Copano Bay (Texas), shell debris becomes mixed with foraminifers on a variable clay and silt floor.

In France, similar reefs are known in the pools of Languedoc where they are known locally as *cadoules* and *platières*.

Bryozoan Biostromes

Bryozoans do not, strictly speaking, form bioherms. Instead, the calcareous skeleton of these organisms form biostromes, which have been of importance in the geological column since the Ordovician.

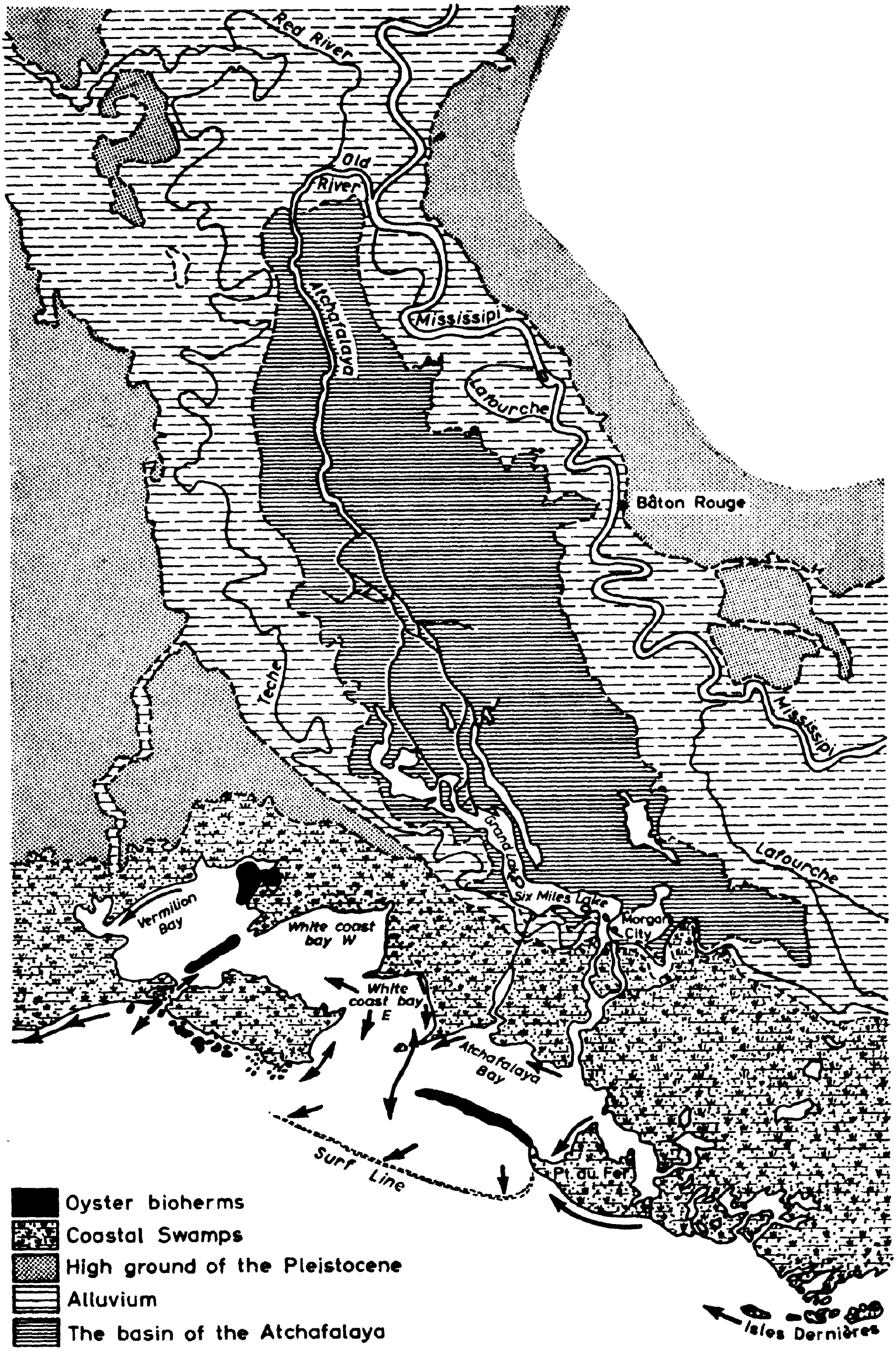


Fig. 150. The Atchafalaya River and its Mouth (after the map of the Mississippi River Commission, and Cary, 1906, for the oyster bioherms)

Note the position of the oyster bioherms.

In the Carboniferous, especially from the Late Tournaisian until the Late Visean, bryozoans almost formed bioherms. The fenestellids, then dominant, accumulated, and because of their reticulate structure acted as sediment traps. This is the "Waulsortian" reef facies of the Carboniferous limestone, well known in northern France and Belgium, in the Sahara (see p. 290) and the United States.

Modern bryozoa form extensive reefs (biostromes), for example, on the continental shelf of northwestern Australia; their presence is noted on hydrographic charts as "coral" (Carrigy and Fairbridge, 1954).

2. THE CORAL FACIES

The Madreporaria are subdivided on the basis of the biological conditions which prevailed during their development:

1. Those which form reefs in the tropical zone at less than 300 feet deep and at a temperature above 65° F. are associated with *zooxanthellae*[1]: these are the *hermatypic* corals. Optimum growth is observed in water less than 20 feet deep, and with a mean annual temperature about 80° F.
2. Those of world-wide distribution which do not form reefs, but occur as individuals or isolated colonies, perhaps in banks at depths between 200 and 20,000 feet (with an optimum between 600 and 700 feet), and at temperatures between 54° and 58° F., although exceptionally this may be as low as 30° F. They are never associated with the zooxanthellae and thus they do not require light. These are the *ahermatypic* corals.

Banks of Ahermatypic Corals (fig. 151)

These corals prefer salinities of 3·3–3·5%. The floor on which they live is composed of hard rock or very coarse deposits (for example on moraines without fine material). Marine currents carry food to them.

The Madreporaria occur outside the limits of the continental shelf, on its borders, and on the continental slope. These are not true reefs since the corals which form them are either simple (*Flabellum*, *Caryophyllia*, *Fungiacyathus*) or branching (*Astroides*, *Astrangia*, *Lophelia*, *Madrepora*). All these are impoverished in coenenchyme. They are present in the Mediterranean, for example, off Provence at between 800 and 1,600 feet, and near Ajaccio. All seem to occur on siliceous or crystalline rocks (Le Danois, 1948; J. J. Blanc, 1958). They are also found in the Atlantic ("coral patch") between 2,600 and 5,000 feet. More than one hundred banks have been counted off the Norwegian coast as far north as the Lofoten Islands, for example, at the entrances to the great fjords such as that of Trondheim. There are none in the English Channel or the North Sea since the water is very shallow and muddy.

It is necessary to emphasize the peculiar character of these corals since

[1] Symbiotic algae living in association with modern corals.—Translator.

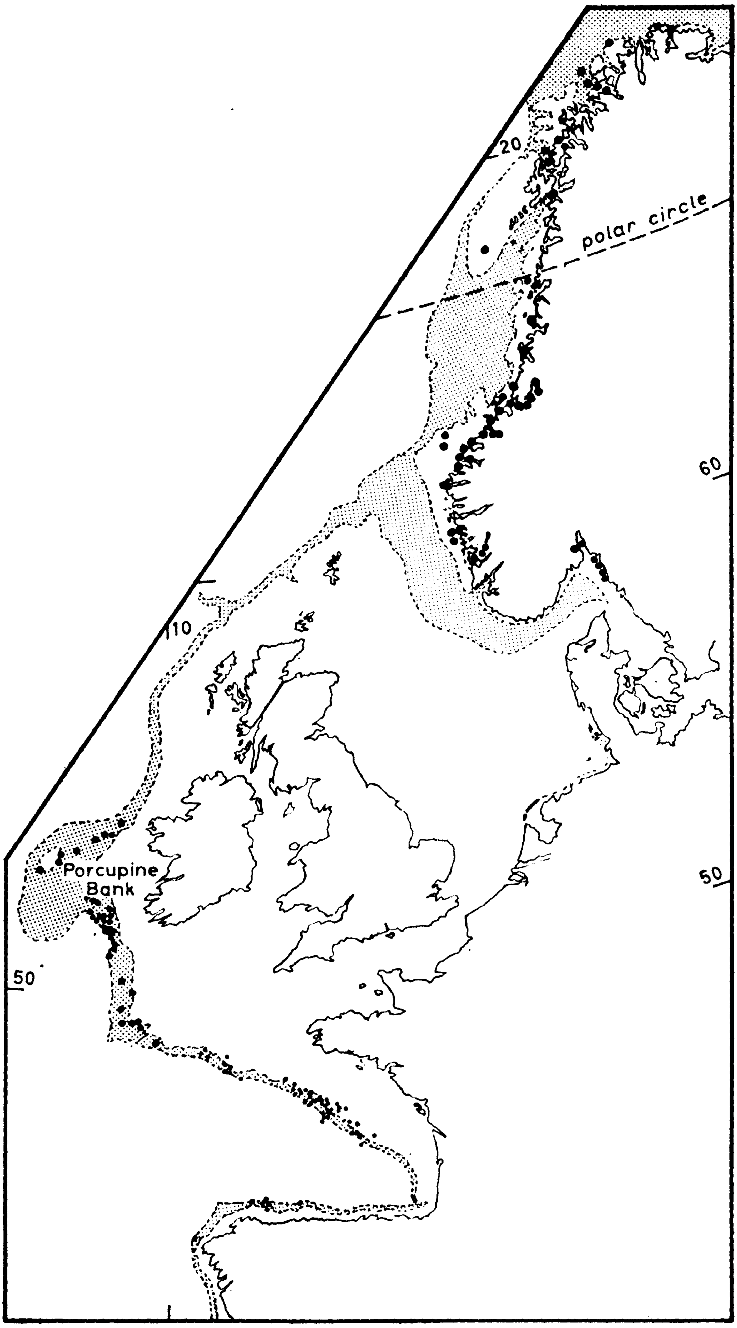


FIG. 151. DISTRIBUTION OF AHERMATYPIC CORALS ALONG THE NORTHWEST COAST OF EUROPE

In gray, the edge of the continental shelf.
(In part, after C. Teichert.)

they are as well defined systematically and bathymetrically, as those which form true reefs at depths of less than 300 feet in intertropical zones! The adaptation of forms to deep water is relatively old, judging by the cosmopolitan distribution of *Caryophyllia smithi*, which occurs without modification in arctic, boreal and antarctic waters and in the intervening deep waters. Most of these forms appeared in Jurassic or Cretaceous times, but their first association with ocean deeps only dates from the Late Cretaceous.

In the Norwegian coral banks *Lophelia prolifera* and *Madrepora ramea* are associated with the hydrozoans *Stylaster gemmascens* and *Allopora norvegica*, which also form dendroid colonies. On the banks there is a large benthonic fauna, comprising 190 species (Dons, 1944; Teichert, 1958) which includes sponges (*Ceratospongia*), coelenterates (dendroid and isolated corals, sea fans, hydras), polychaete worms (destroying skeletons), bryozoans, brachiopods (Terebratulids), echinoderms (asteroids, ophiuroids, crinoids, echinoids, holothurians), crustaceans (crabs and cirrepedes), molluscs (*Lima*, *Pecten*, gastropods and cuttlefish), tunicates and fish.

The formation and the destruction of the coral thus produces calcareous muds which cement the banks, as in the case of true reefs.

There is no doubt that the Norwegian coral banks were widespread during the Pleistocene, since some are found raised by the post-Pleistocene uplift of the Scandinavian Shield. This has brought them above the critical depth of 200 feet and they have died. Some are found as much as 75 feet above sea level in the fjords of Oslo and Trondheim.

The presence of the banks has some bearing on the origin of limestone. During Pleistocene times it is unlikely that Scandinavia was covered by forests and thus the development of the limestone of these banks was independent of all biostasic phenomena.

Another more important consequence is the occurrence of limestone formed by corals outside the true reef environment (Teichert, 1958). These banks, once they are buried, give rise to lenses, which, in the fossil state, have the appearance of reefs, although they may pass into moraines or glacio-marine sediments, or into euxinic deposits of fjords. Certain characteristics distinguish them: the number of species of coral is small and coenenchyme is absent from them. Furthermore, algae are absent since the banks are formed at depth and in darkness.

Among fossil examples there are the banks of corals which lived in deep, cold water in the Late Danian of Sweden and Denmark. These banks occur in the chalk and thus give some indication of the temperature of its formation and also indicate that the water was relatively deep. At Faxe (Denmark) limestones, now occurring 230 feet above sea level, contain branching corals (*Dendrophyllia candelabrum*, *Calamophyllium faxense*, *Lobopsammia faxensis*), while a sea fan (*Molkita isis*) is rather rare. The individuals are isolated, separated by a hard calcitic mudstone. The fauna is rich: crustaceans (the crab *Dromiopsis rugosa*), nautiloids (including

Hercoglossa danica), gastropods (*Pleurotomaria niloticiformis*, *Siliquaria*, *Cerithium*, *Cyprea*, *Tritonium*), pelecypods (*Modiola*, *Arca*, *Crassatella*, *Isocardia*), brachiopods (*Rhynchonella*), echinoderms (*Cyathidium*, *Holopus* and *Temnocidaris*) and the teeth of sharks.

In places this facies passes into a bryozoan limestone containing some of the corals, or they may be completely absent. In the latter, brachiopods (*Terebratula*, *Crania*), echinoderms (*Tylocidaris*, *Brissopneustes*, *Metopaster*), worms (*Ditrupa*, *Serpula*) and cirrepedes (*Scalpellum*) occur. A single type of rudistid, a small *Radiolites*, is also known here. The assemblage is a dwarf fauna, very rich in species and individuals, but it lacks the characteristics of the faunas of a warm sea. There are no Orbitoloideae and the corals are without coenenchyme. On the contrary, the rock contains pelagic coccoliths and foraminifers and was probably developed at a depth greater than 300 feet.

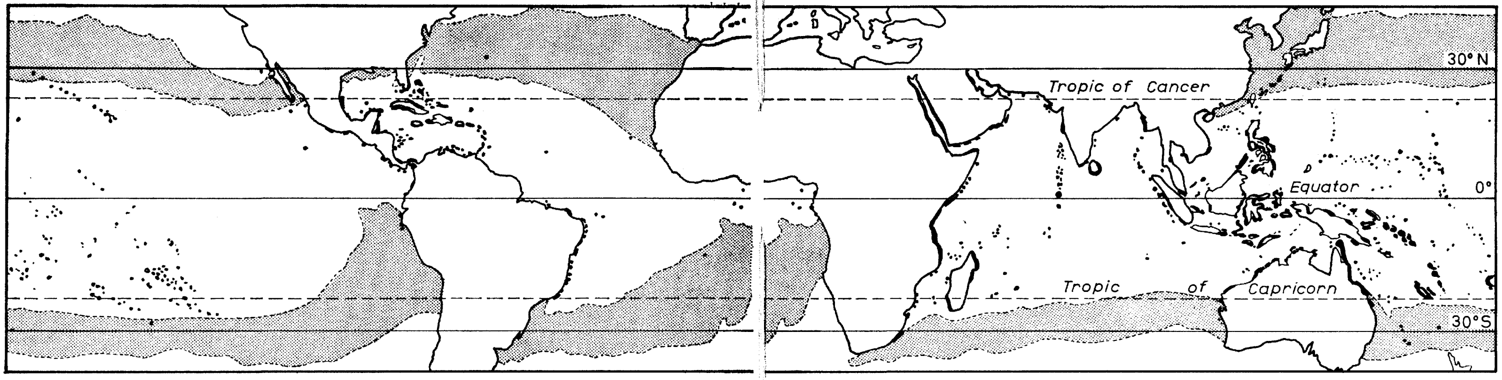
Going further back in time, it becomes more difficult to distinguish between the corals which lived at depth and those which were restricted to the reef environment. It is nevertheless clear that certain assemblages of corals were in no way associated with warm.sea conditions. This is true of the *Cyathaxonia* (D. Hill) fauna which is known from the Silurian up to the Permian. The small isolated corals which form this facies are surprisingly robust and do not seem to have lived in deep water, since they are often associated with algae. There is, in fact, some evidence to suggest that they lived in relatively fresh water and on muddy floors.

Coral Bioherms (figs. 152–155)

Much has been written about "coral reefs" since Darwin's important contributions (1837, 1889). The principal types of reef are, briefly as follows:

1. *Atolls* (Darwin, 1837; Cloud, 1958) are annular (ring-shaped) reefs surrounding a lagoon devoid of all pre-reef material, and in which detrital limestone accumulates. Their name comes from the term "atollan", borrowed from the Malay language by the natives of the Maldive Islands. Most of the atolls occur in the Indian and Pacific Oceans between the Tropics. About 330 have been counted in the Tuamotu Islands, eastern Indonesia, the Caroline Islands, the Marshall Islands, Fiji, the Maldive Islands, the Laccadive Islands, the Gilbert and Ellice Islands, the Coral Sea, the South China Sea, the Bismarck and Solomon Islands, the Great Barrier Reef of Australia and the Red Sea. In the Atlantic only two definite atolls are known (Tortugas to the southwest of Florida Keys and Rocas off Brazil).

These structures are often aligned. The Maldive Islands extend from north to south for 470 miles in a belt 65 miles wide formed by a double row of small individual atolls. Those of the Caroline Islands are northeast to southwest, but the other atolls in the Pacific are all aligned northwest to southeast. The Australian atolls are parallel to the coast.



FIGS. 152 and 153. DISTRIBUTION OF HERMATYPIC CORAL REEFS

In gray, the zone traversed by the 70° F. (21° C.) isotherm during the course of the year (simplified after Joubin.)

An atoll arises from a platform formed previously by erosion, which then subsided and on which reefs developed.

The *tabular reefs* can be considered as being derived from uplifted atolls, or from the complete infilling of coral lagoons (Fairbridge, 1950).

2. *Fringing reefs* grow directly against the coast. They may be 10 to 1,000 yards in width, but are discontinuous, with gaps opposite every fresh-water stream or river. A shallow water way, or "boat channel", often separates it from the beach.

3. *Barrier reefs* are parallel to the coast, from which they are separated by a sheltered sea. An example of this is the Great Barrier Reef of north-east Australia. Barrier reefs, like fringing ones, are found in sections or "ribbon reefs", divided by gaps and channels, sometimes exceeding 300 feet in depth.

The movements of land and sea are liable to upset the precise application of this classification to some reefs.

The Conditions for Reef Development.—The biological characteristics of the evolution of reefs have been discussed elsewhere (Termier, 1952, pp. 216–229). The equilibrium between growth, sedimentation and erosion which is associated with them, places the establishment of this facies among the principal phenomena dealt with in this book.

The marine organisms forming the reefs are particularly sensitive to climate; their constructional role is restricted to the tropical and equatorial zones. Moreover, reefs which occur on the continental shelf show evidence of the fluctuation of the sea in these areas.

The principal physiological condition for the prosperity of modern coral reefs is intense oxygenation. Verwey, in 1931, calculated that *Acropora hebes* consumed 20 c.c. of oxygen per kilogram per hour. It is largely because of this that the limit of light penetration has a considerable effect on the existence of the reef, since it fixes precisely the limit of growth of algae which are the principal generators of oxygen. Furthermore, clear water free from impurities is almost indispensable. As a general rule, suspended sediments are prejudicial to corals, which can be killed by asphyxiation (Mayor, 1924). When the suspended matter is not excessive however, the corals are capable of rejecting the mud particles from their surfaces.

Wind and waves erode reefs, which are first broken into very coarse debris (coral shingle) and then into sand. In this way "sand cays" (keys) can be formed. These are actually dunes, which begin as bars and grow under the influence of trade or monsoon winds. They are gradually covered by vegetation and form rounded or oval islands with relatively small ridges of bare sand which change direction according to the prevailing wind (see p. 280). The shingle stands up as ramparts on the windward side, and behind these there is a calm zone where mangroves can become established. Island reefs, eroded by the waves (and perhaps aided by man who uses the limestone for road metal) are gradually undermined and finally disappear.

The shingle which fringes coral islands is flanked on its inner margin by a ditch which seems to be due to the flow of water through the rampart (Mayor, 1924; Umbgrove, 1947). The sand situated in the tidal zone is cemented into a "beachrock".

Each reef includes a number of biotopes (within and below the reef). The role of the wind and the waves in the deposition of sand and coral debris is such that the most favorable conditions for the growth of reefs occur on the windward side: that is, where sedimentation is practically absent. In some cases, the swell of the waves makes the living coral take on peculiar shapes, such as the comb-like structures described by Guilcher (1954) in the northwest of Madagascar (see pp. 284–286).

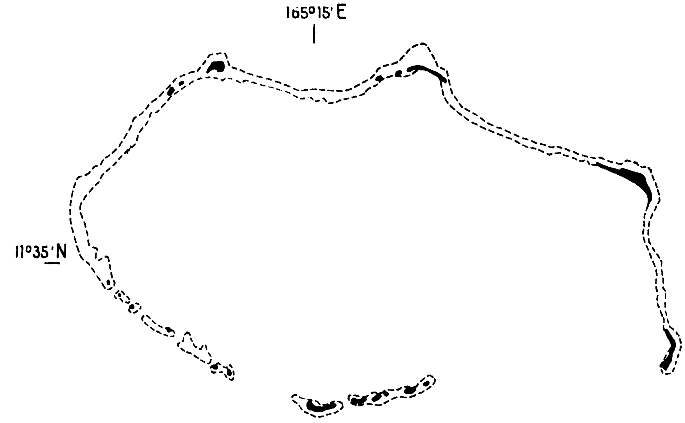


FIG. 154. BIKINI ATOLL, MARSHALL ISLANDS, CENTRAL PACIFIC (simplified after Teichert)

The coasts of volcanic islands where there is much eroded material, do not favor reefs (W. M. Davis, 1928), and the steep offshore slopes sharply limit the width of reefs.

The external zone of coral structures is often protected by a Lithothamnion "rim" which is particularly important in most reefs of the Pacific region. These algae act as a cement for the "shingle rampart". The calm water of the lagoon allows the building of secondary reefs (pinnacles or knolls) either of corals, or of pelecypods which are more tolerant of the quiet conditions than the Madrepora.

The Origin of Modern Reefs.—Of the many theories put forward to explain reef development, that of Charles Darwin is fundamental, stating that progressive submergence is necessary to build thick reef accumulations. During the Pleistocene glacial stages there was progressive lowering of sea level and cutting down of old reefs and reef foundations to form platforms down to about −300 feet. During the post-glacial transgression reefs have built up again, under eustatic influence, often without tectonic interference.

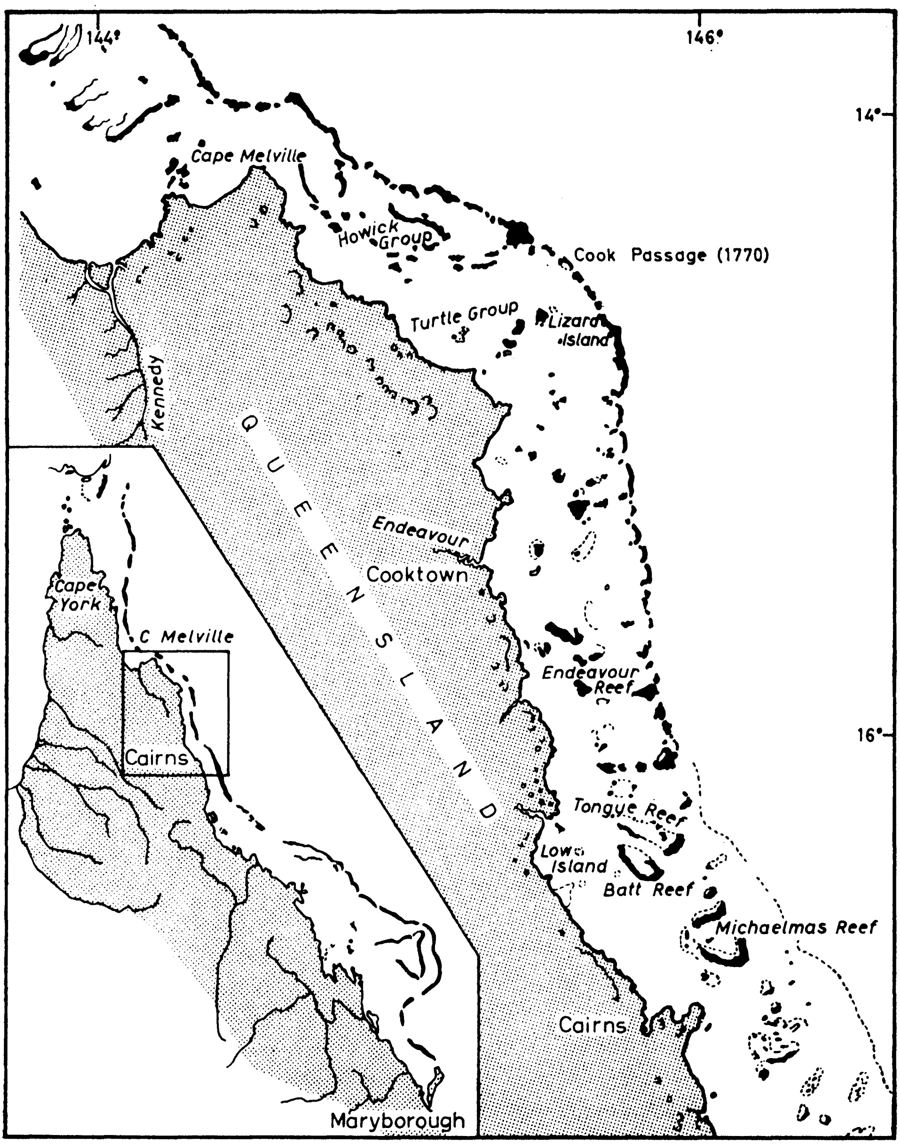


FIG. 155. THE GREAT BARRIER REEF OF AUSTRALIA (very simplified, after Yonge)

This is Daly's "theory of glacial control" (1934), which is the most important single factor in continental shelf regions.

In the mid-Pacific atolls, and tectonic regions like the East Indies, the foundations are often far below (or above!) the Pleistocene limits, and the eustatic factor may be obscured.

Thus it is necessary to fall back on Darwin's theory (Umbgrove, 1947) according to which an atoll begins as a reef fringing a (possibly volcanic) island. The island may be planed off to become a guyot or seamount (Hess, 1946). If the reef continues to grow, and maintains itself at the optimum depth, it becomes successively a barrier reef, then (when the island is submerged) an atoll. As has been observed in Queensland, reefs often occur on broad platforms. Detailed study has shown that progressive sinking of the sea bed is a common phenomenon. Subsidence explains the increase in thickness of the reefs in relatively stable bathymetric conditions, and it is now invoked to explain the considerable thickness of certain fossil reefs (p. 287).

The borings and geophysical observations on reefs in the Pacific (for example on Funafuti (Hinde, 1904), Bikini, Eniwetok, in Queensland and in Bermuda) have shown that the coral limestones reach a depth of 650–1,200 feet below the present sea level. At Eniwetok the calcareous facies (4,000 feet thick) goes back as far as the Eocene (Ladd *et al.*, 1953). These examples favor the hypothesis of subsidence.

On the other hand, many ("shelf") atolls, rise from a sea bed less than 300 feet deep (Cloud, 1958). This depth is generally believed to mark the maximum fall in sea level at the close of the Quaternary glaciation. It seems likely, therefore, that these reefs are of recent formation and that their development has been dependent on "glacial control". The last regression, progressively dropping the sea level in small oscillations over the last 4,000 years to 10 feet below its maximum in the mid-Holocene, played an important role in raising part of these structures above sea level.

Examples of Modern Reefs

The Coral Reefs of Australia.—The problems posed by the coral reefs of Australia are linked to the stability of the Precambrian shield.

West Coast (fig. 143).—All the west coast of Australia is dotted with coral reefs (Carrigy and Fairbridge, 1954). In this region, the area of the continental shelf is about 414,763 sq. miles.

The reefs are modified by tides. The range of spring tides is 35 inches at Fremantle, while the annual variation in mean sea level is about 27 inches. From Geraldton to Wyndham the height of spring tides increases; it is 22 feet at Port Hedland and 36 feet at Derby. This increase seems to be linked to the increase in size of the continental shelf and its decrease in depth. The prevailing winds south of the Tropic of Capricorn are: (1) the southeast trade winds (dry, because they have come over the continent) which blow in summer, and remain north of latitude 30° S. in winter; (2) the west winds (humid, from the Indian Ocean) which bring the winter rains. The northwest is dominated by the monsoon which comes from Java and the Indian Ocean. The monsoon winds are moist from December to March and carry rain to the Kimberley plateau in summer. Moreover, there are hurricanes ("willy-willies") in the Timor Sea which follow the coast southward and

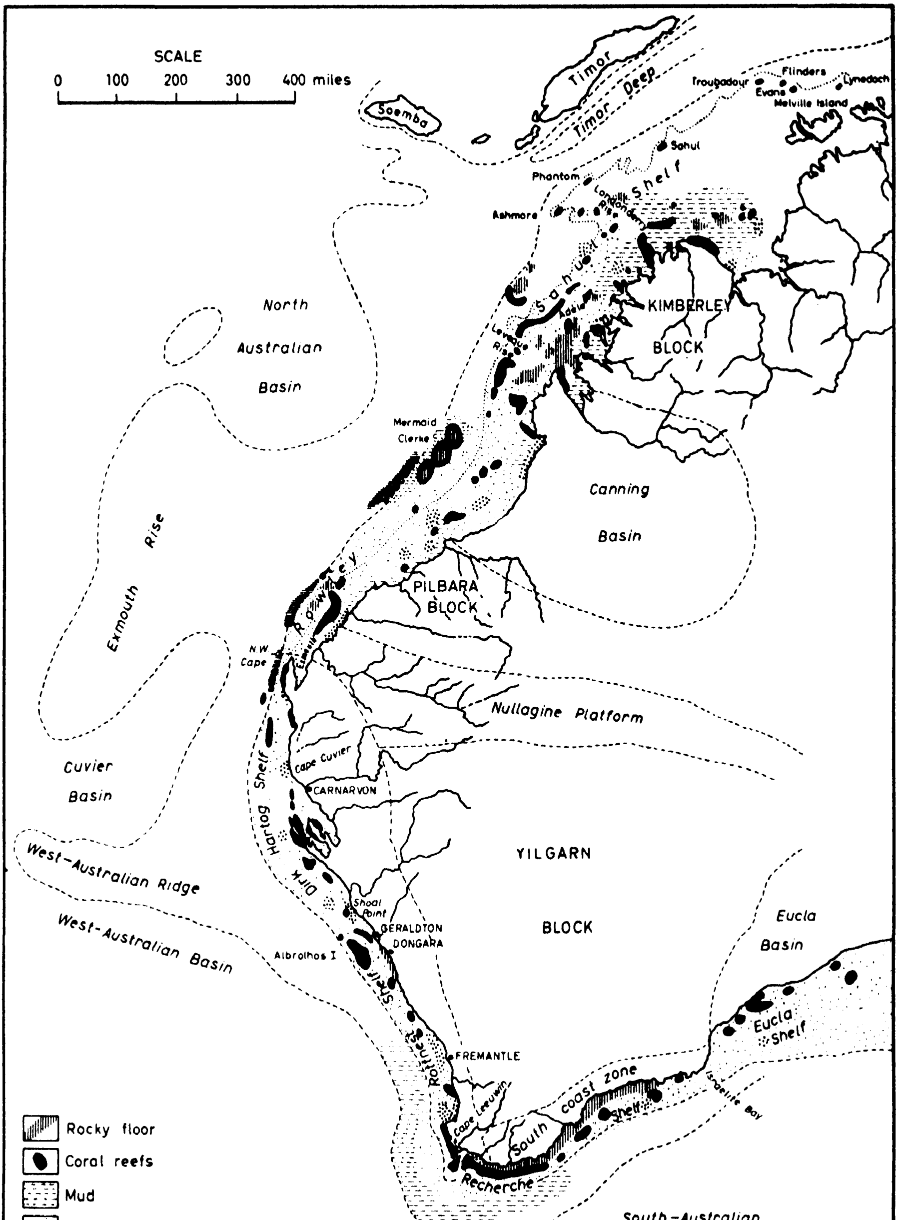


Fig. 156. Sedimentation on the West Coast of Australia (after Fairbridge)

In Legend, ● = reefs of coral and bryozoa.

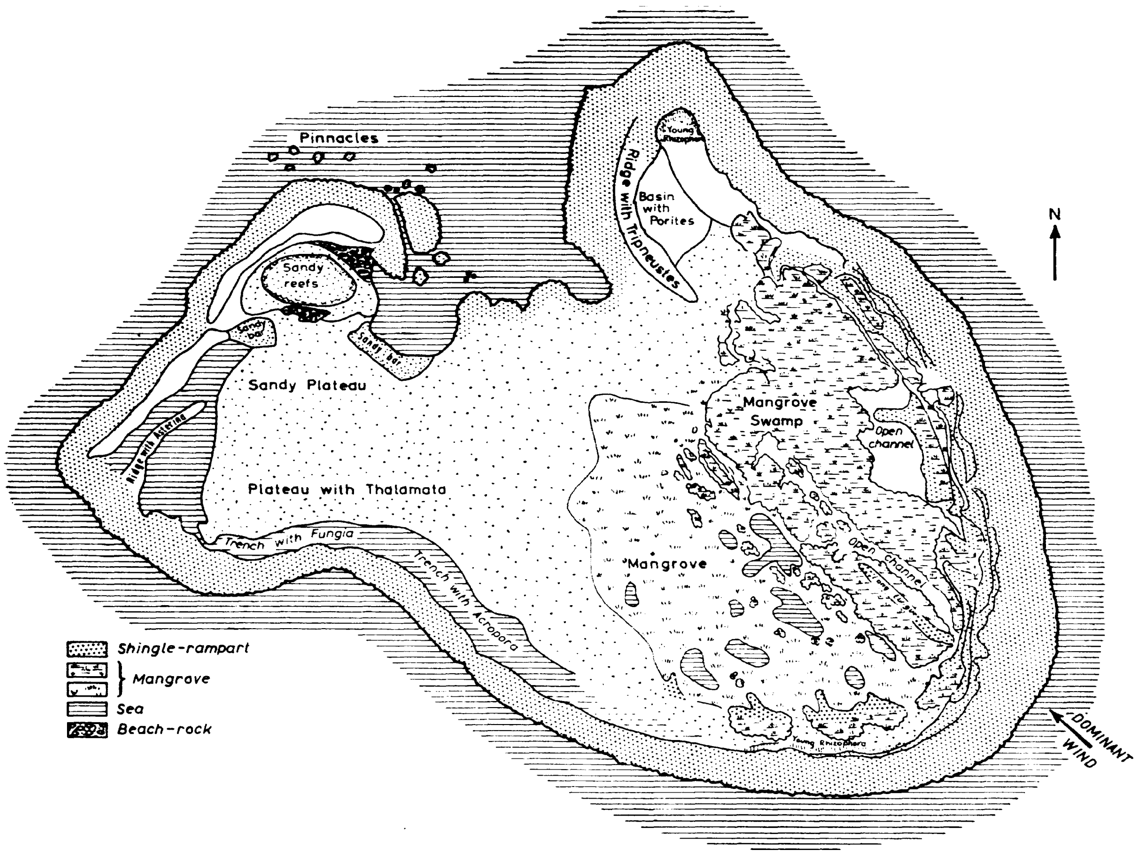


FIG. 157. THE LOW ISLANDS OF THE GREAT BARRIER OF AUSTRALIA (after Spender, 1929; Moorhouse, 1934; Fairbridge and Teichert, 1945) [*cf.* fig. 155]

FIG. 158. THE LOW ISLANDS OF THE GREAT BARRIER OF AUSTRALIA (Photograph: Fairbridge) [*cf.* "Low Island", fig. 155]

FIG. 159. TWO ISLES, IN THE LAGOON OF THE GREAT BARRIER REEF (Photograph: Fairbridge)

The two islands are situated on a submarine platform which can be seen below the surface of the sea.

cause devastation principally between 20° and 22° S. Occasionally the hurricanes cross the land and almost reach the Great Australian Bight.

Two ocean currents are known. One follows the west coast northward and divides into two at the North West Cape. One branch joins the subequatorial current, while the other continues along the northwest coast, across the Sahul Shelf and then passes into the Arafura Sea. The second current flows westward along the southern coast during July and August, but is reversed in February and March.

On this Western Australian continental shelf, the coral reefs are arranged in archipelagoes, related to the distribution of structural ridges and platforms. Shelf atolls coincide with subsiding sedimentary basins (Fairbridge, 1950).

East Coast: the Great Barrier.—This reef barrier is the most important in the world, and is entirely different from the reefs of the west coast. It

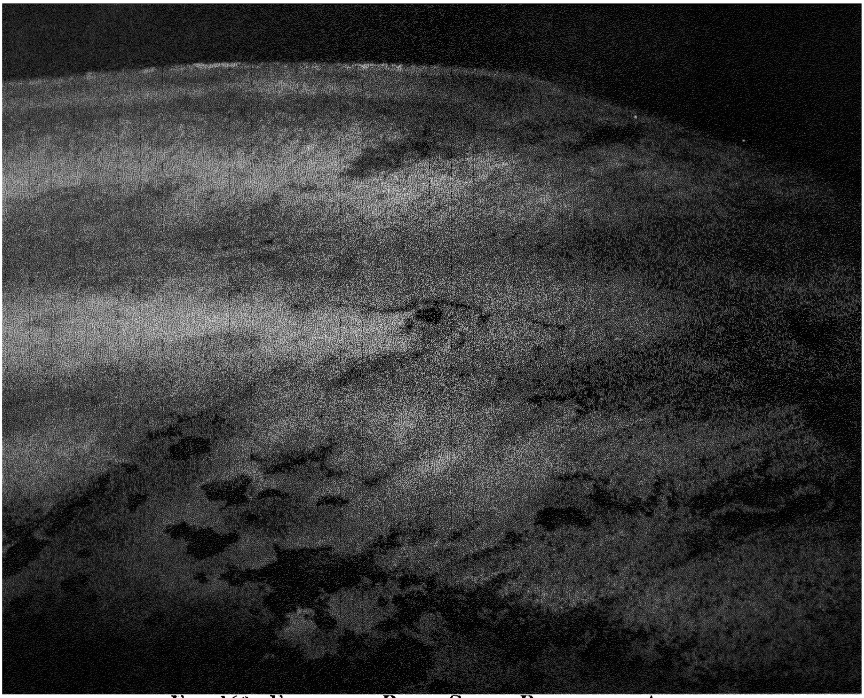


FIG. 160. ENDEAVOR REEF, GREAT BARRIER OF AUSTRALIA

Reef platform with shallow coral basins, living corals (dark) and coral sand (light) [*cf.* fig. 155]. (Photograph: Fairbridge.)

cuts off a zone of calm water protected from the open ocean in which are numerous reef platforms, mostly evolved from shelf atolls, and superimposed by sand cays and mangrove (such as the Low Isles) (figs. 155 and 157–162).

The Coral Reefs of the East Indies (fig. 163).—Among the East Indian reefs, there are 21 groups of atolls and 17 reef-barriers. There are very few of them on the Sunda Shelf (see p. 50) but they are numerous in the regions of the archipelago where the crust is unstable; that is, in the mobile zones in the course of orogenesis. Thus the East Indies is a particularly interesting region for the study of the complex relationships which occur between almost independent phenomena such as the growth of corals, erosion, orogenic movements, the movements of blocks and eustasy (Umbgrove, 1947).

The reefs of the Sunda Shelf are exceptional because they are established on muddy floors. Thus, their early development differed from that of reefs which established themselves on rock. In the Thousand Islands, and in the Bay of Jakarta (= Batavia), corals initiate on hard objects. The first forms ("pioneers") are well-defined genera, usually branching (*Acropora*, *Montipora*) or massive (*Porites*). These particular corals are those which flourish on the outer margins of reefs. Reefs on muddy floors also occur southwest of the Celebes in the barrier reef of the Spermonde Shelf, at

FIG. 161. THE RIBBON REEF, PART OF THE EXTREME OUTER EDGE OF THE GREAT BARRIER OF AUSTRALIA

Note the very calm water inside the reef, i.e. to the left of the photograph. (Photograph: Fairbridge.)

Krakatoa and at Emmahaven (west of Sumatra). Similar reefs also occurred in the Early Pliocene (Gunning hingga padang) and in the Late Miocene (Tji Lanang) of Java.

Once the reefs have been established, they develop at the same rate that alluvium, brought by rivers, is deposited. The muddy floor compacts and the reef sinks. The amount of sinking is about one-third to one-half of the total thickness of the reef (in the Emmahaven reef sinking totals about 13 feet). Thus a true *facies change* takes place.

The reefs of the Bay of Jakarta show the following facies, which are similar to those shown by reefs on rocky floors:

1. In the shallow, warm water (c. 96° F.) of the "moat" (inside the rampart) *Montipora ramosa* thrives on a sandy bottom.

2. The "rampart" formed of coral debris (shingle) represents the Rhodophyceae facies (the Lithothamnium rim) which encloses branching corals (*Acropora*).

3. On the outer slope of this rampart to the northwest and to the southeast of the island lives *Montipora pliosa*.

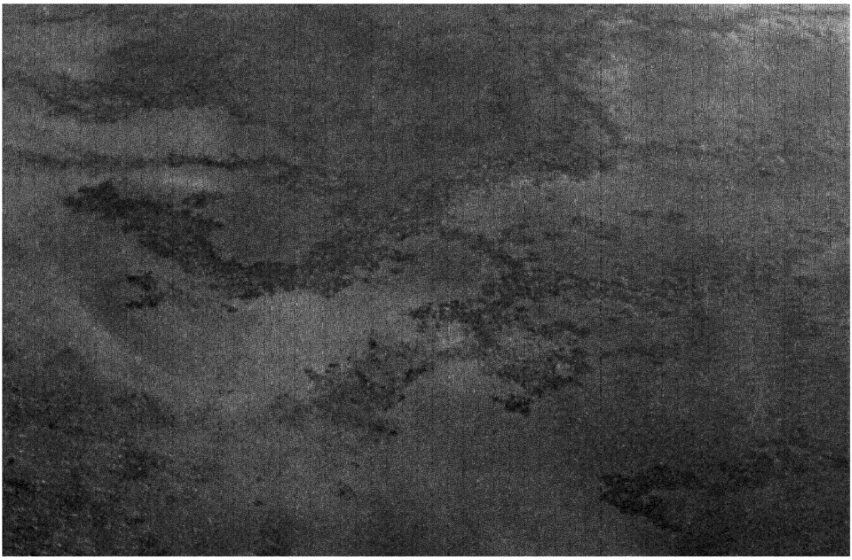


FIG. 162. BATT REEF, QUEENSLAND

Showing the isolated patches of reef (dark) on a broad, sandy reef platform (light). The whole of the area is slightly below sea level. (Photograph: Fairbridge.) [*cf.* "Batt Reef", fig. 155.]

4. From the west round to the south, the reef facies is impoverished, because it is covered with coral sand.

The wind largely controls the growth and the position of the coral reefs in the open sea of the Indonesian archipelago, but in the calmer waters of the Sunda Shelf it has less influence. The East Indies is in the region of the Asian monsoons and the erosion due to these winds balances the growth of the reef. Thus the monsoons shape and orient the reefs.

Like any rocky coast, the reef is subject to marine abrasion. It is even possible, as has been suggested by Spender, Crossland and Gardiner, that this may be responsible for the formation of reef-platforms. Most vigorous reefs grew up to the 10-foot eustatic sea-level maximum in mid-Holocene time; since then there has been a progressive planing down of the older reef

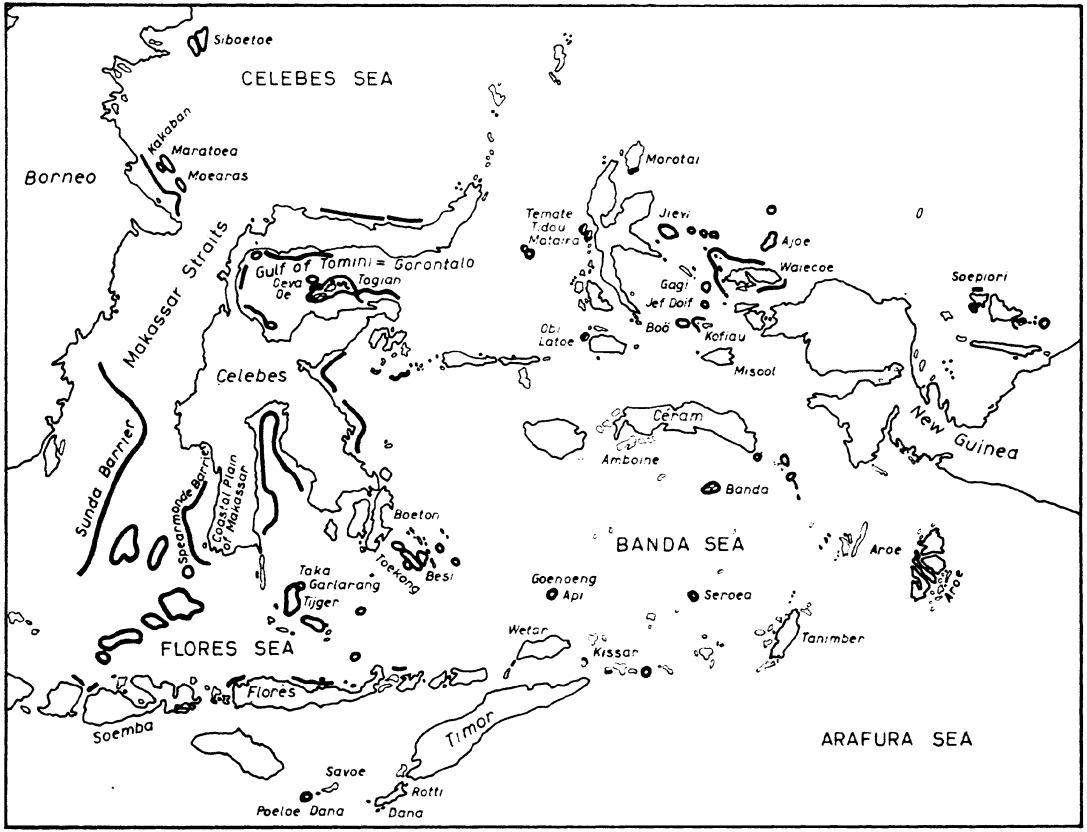


Fig. 163. Distribution of Coral Reefs in the East Indies (after Umbgrove)

The reefs are shown by heavy lines.

material by erosion and a filling-in of the lagoons and shallow pools. Thus most "reef-flats" are compound: partly eroded, partly constructed, partly sedimented. The associated sediments are the coral sands and coral shingle (represented here by the Lithothamnium rim). It is here that the orientation of the wind and the occurrence of the monsoon is important. The coral sands form cusps opening away from the prevailing wind. Umbgrove (1947) refers to sandy underwater reefs formed by two cusps, whose horns point in opposite directions; each one corresponding to the two directions of the monsoon (e.g. Takat Bloekoeran, in the Straits of Moedera).

The winds and the surf from the east-southeast influence the orientation of the Thousand Islands of the Java Sea, and the monsoon from the southeast makes itself felt in the Spermonde archipelago (southwest of the Celebes) but is deflected by the high mountains of the South Celebes, and does not affect the Straits of Macassar. The northwesterly monsoon, however, is felt there and in the Java Sea.

The orientation of the crescentic sandy reefs convex to the prevailing wind is practically the marine equivalent of the arrangement of sand dunes with respect to the wind that built them. The "horse-shoe reef" is thus matched by the barchan dune.

The part played by other winds, especially those which blow from the land toward the sea, or vice versa, should not be overlooked. These apply to fringing and barrier reefs but not to oceanic reefs.

Reefs which are formed in the calm zones do not show the same sedimentological characteristics as those which are subjected to the monsoons. A good example noted by Umbgrove (1947) is that of the barrier reef of the Togian Islands situated in the deep gulf of Tomini (Celebes) where it is sheltered from all winds. In this instance, there is no rampart composed of coral shingle, no living coral on the slopes, and no sandy reef. Moreover, Melobesiae are rare.

In the hurricane belts, which lie north of latitude 5° N. and south of latitude 5° S., the erosion of the reefs by the waves is particularly violent. Reefs are broken into blocks, the large mushroom-shaped fragments are thrown up onto the beaches, where they form "negroheads". The Malay archipelago, being situated between 6° N. and 9° S., is only subjected to these hurricanes in the southeastern part (Timor, Banda, Rotti, Kisar, Leti, Damar, Kei). Kuenen has observed "negroheads" much farther to the north, in the northwest of Morotai, in the Nanoesa Islands and on the east coast of Karakelong (Talaud).

Erosion, similar to that due to cyclones, is produced by the tidal waves accompanying paroxysmal eruptions, such as those of Krakatoa and Paloeweh.

The role of sea water as a solvent in the erosion of coral reefs has been suggested (Murray and Gardiner) although tropical seas are almost certainly saturated with calcium carbonate. This solvent action has already been noted (p. 100) in relation to the shaping of the supralittoral stage. It has also been observed by Kuenen (1933) and by Umbgrove (1947) in some regions sheltered from the waves, that is, in places where erosion by wave action can be excluded. Such solvent action is limited to the tidal zone. Boring organisms, especially *Echinometra mathaei* (Blainville) also have a destructive effect (pp. 89–91).

The general morphology of reefs raises a number of problems which various hypotheses have attempted to resolve.

According to Daly, "glacial-control", that is, eustatism (eustasy) linked to Quaternary glaciations and deglaciations, is the principal cause which has shaped the forms of reefs. This theory, however, cannot explain all the features of the reefs in the East Indies and the Sunda Islands. These regions have been subjected not only to eustatic changes, but also to orogenic movements. Some of the most recent reefs rest concordantly on their substratum (Maroekoe), whereas others are discordant (Padang, Siboetoe, Morotai, Kisar and Votap). Moreover, because the enormous thickness of certain reefs cannot be adequately explained by the "glacial-eustatic" theory, Ladd and Hoffmeister (1936) proposed the theory of the "antecedent platform". According to this theory, corals in the subequatorial zone should be able to develop on any foundation situated at a suitable depth, and should be able to grow without requiring any variation in sea level. However, a lowering of the oceanic surface allows erosion by the action of waves and atmospheric agents, and will cause degradation of the lagoon floor, while a rising sea level during interglacial and postglacial periods will favor the coral growth.

Reefs are of several types. Some have lagoons commonly 400 feet, and exceptionally 600 feet, in depth, as in the Togian Islands (Celebes). According to Umbgrove (1947), these lagoons are ancient gorges belonging to a submerged relief formed during glacial times. The reefs occur on the summits of this old surface, and the lagoons tend to be filled with sediment.

But the influence of orogenic movements seems to have a direct effect on the growth and behavior of reefs in these areas.

The role of *subsidence* has been demonstrated in this region by Kuenen (1933). The atolls grow on a subsiding marine floor. The Tijger islands (fig. 164), situated to the south of the Celebes, rise from the edge of a submerged crater, which, in turn, rises from a sea bed 6,500 feet below sea level. The Toekang Besi Islands to the southeast are arranged along tectonic lines in the substratum. These atolls slope very steeply (40°–50°) down to 1,600–2,000 feet, but may be vertical down to 650 feet, and finally, between 3,000 and 6,500 feet, they join an undulating submarine plateau. According to Kuenen, true reefs originated near the end of the Tertiary, around the anticlines of an undulating plateau which was at that time subaerial. This plateau has since sunk several hundred feet, in places as much as 3,000 feet. Block faulting has led to the intermittent uplifting of some of them, or may have caused tilting (Lintea Atoll). To the northeast of Borneo, the atolls of Kakaban, Maratoea and Moearas rest on a plateau only a thousand feet deep. Kuenen (1933) has invoked subsidence as the origin of the barrier reefs of Kofian, the islands of Boë, Gagi, Waigeoe, of the discontinuous barrier of the northeast coast of Borneo, the barrier reef of the Banggai archipelago and nearly all the atolls of Jef Doil, Poeloe Jiew and Ajoe.

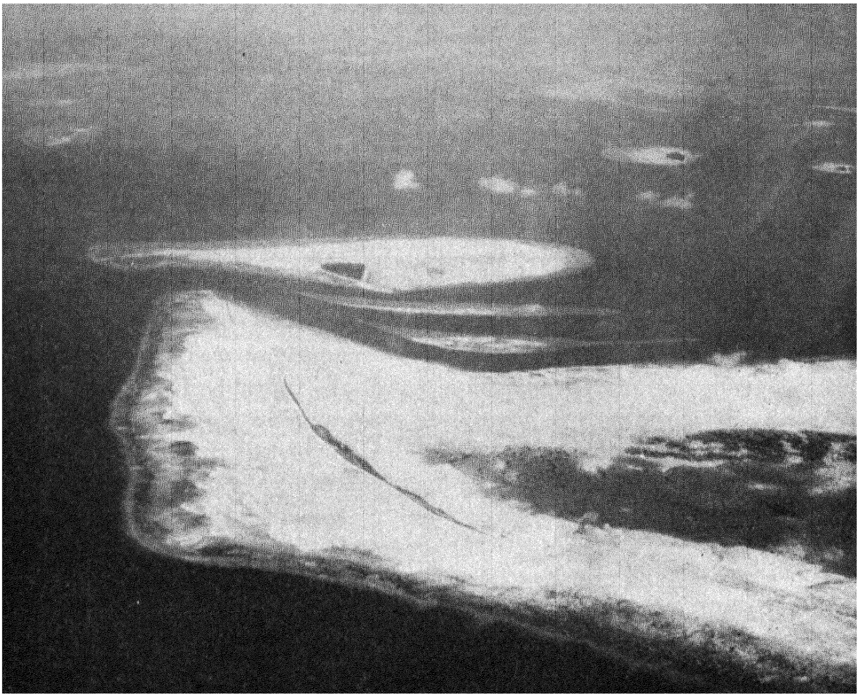


FIG. 164.—ATOLLS OF THE TIJGER GROUP, EAST INDIES, SOUTH OF CELEBES

Note the numerous reefs, rising from an antecedent platform. The sandy zones are light-colored. These atolls are of the same compound type as those of the Maldive Islands. (Photograph: Fairbridge.) [*cf.* "Tijger", fig. 163.]

Dead reefs were found by dredging in the Ceram Sea by the Siboga Expedition at depths between 4,280 and 5,358 feet, far below the limit of living reef corals. The subsidence must have been much faster than the upward growth of the reef (Umbgrove, 1947).

The movement of crustal blocks also explains the development of the reef-barrier in the Togian Islands (Tomini Gulf), because no negative movement of the shore is apparent. Among eroded volcanoes in the central part of the islands, there are uplifted and planed reefs of Late Neogene, or

Pleistocene, age which enclose calcareous material probably deposited in a lagoon. The whole of the Togian Islands forms a block, delineated by recent faults, which has been uplifted while adjacent blocks have sunk and produced the deep sea now surrounding the islands. Movements of similar type are common in the East Indies; they are particularly well known in the Celebes itself.

The "geanticlinal" uplift of reefs has been used by Brouwer (1918) to account for the occurrence of "reef-caps". Brouwer noted that, in the eastern part of the East Indies, reefs (dating at most from the end of the Tertiary) were uplifted nearly 4,000 feet, and concluded that they were situated on the geanticlinal axes of island arcs which had been raised since the Plio-Pleistocene to different heights independent of their respective ages. On the island of Dana, south of Rotti, uplift has raised the reef limestone to 118 feet above sea level, and it is now surrounded by a plain representing the old lagoon. This uplift has also caused extensive and well-defined dislocations in the present topography (fig. 165). The atolls of Maratoea, Kakaban and Moearas also have uplifted reefs. The terraces of these reefs are locally termed "karangs". In the island of Kissar, the "reef-cap" is formed of five terraces, the highest of which is 500 feet above sea level. It is possible, as has been supposed by Martin (1896–1903), that the uplifts accompanied the subsidence of the seas of Banda, Celebes and Soeloe. At Binongko, in the Toekang Besi islands, Kuenen (1933) has shown that the terraces, numbering 14 in all, are asymmetrical (that is, they are not continuous round the island). This island has risen 650 feet in several stages and has remained a solid block, without buckling.

Finally, according to Umbgrove (1947), the history of the coral reefs of the Togian Islands, for example, is as follows: (1) the reefs were formed in the Late Neogene or Early Pleistocene; (2) they were uplifted several hundreds of feet above the level of the sea and have been eroded, while the adjacent fault blocks sank to form the deep sea areas (fringing reefs were formed at this time); (3) the platform supporting the Togian Islands has sunk and the fringing reefs formed in the preceding stage have developed into barrier reefs. Superimposed on these orogenic movements have been the eustatic. The latter have generally been more rapid than the former. However, the orogenic displacements have the greatest amplitude. In the eustatic class the rapid oscillations of mid- and late-Holocene time are very striking (Kuenen, 1933).

Coral Reefs of the Red Sea.—The Red Sea is the result of earth movements which took place during the late Tertiary. It is a true "rift valley" in which the faults have remained active since its formation. The temperature and salinity of the Red Sea favor the development of coral reefs. They are, in fact, very numerous, and occur as large and small peaks on the surface of foundered blocks. The bases of these reefs are 1,500 to 3,000 feet below sea level. Since reef corals cannot live so far down, the existence of peaks capped with living corals near the surface shows that the subsidence of the sea floor was very slow. The thickness of coral rock thus formed varies between a few feet in the littoral zone, to more than 1,500 feet far from the coast. There is no doubt that this thickness has been partly controlled by variations in sea level. Pleistocene oscillations ("Glacial control") on the one hand, and the minor Holocene oscillations (10, 5 and 2 feet) on the other, have left many emerged reef platforms (Nesteroff, 1955). Thus, on many of the peaks, the Red Sea reefs form atolls.

The coral sands and muds are mainly formed by the digestive activity of those fish which feed on the coral (see p. 90).

The living Madrepora can be found as deep as 200 feet, but they do not grow well below 80 feet. Below this the Sponges and, in particular. the Alcyonarians are abundant.

Reefs of Northwest Madagascar.—The coast of northwest Madagascar (Guilcher, 1956) is very irregular, with some high cliffs, and includes the two large deltas of the Mahavavy and the Sambirano. It is washed by a warm sea which averages 84° F. in February and 77° F. in August. Consequently, it is an ideal environment for the development of corals. The tidal range is considerable (15·7 feet at Majunga) and low tides expose the living top of the reef. At the heads of bays where mangroves grow profusely, the sediment carried in suspension makes the sea turbid. Elsewhere, the sea is clear. The distribution of the reefs is directly influenced by this difference in purity of the water.

Uplifted fringing reefs are common (Orangea peninsula, Nosy Vaha). These were built when the sea level was 16 to 20 feet higher than at present, and were separated from the land by a channel 6 to 10 feet deep. Behind this channel, another (older) fringing reef is capped by consolidated fossil dunes and by unconsolidated younger dunes. The reefs may be as much as 50 feet above sea level, and the dunes may be 300 feet high. The latter appear to be elongated toward the southeast.

Present-day erosion has cut a littoral platform dominated by an overhang in the fossil reef. This platform may be 100 feet wide.

Dunes continue to be formed on the east coast of the Bobaomby massif to the south of Cape Ambre, and are gradually covering the two fossil reefs. These reefs are separated by a discontinuous longitudinal depression contain-

FIG. 165. THE ISLET OF DANA, TO THE SOUTH OF ROTTI, SOUTHWEST OF TIMOR. AERIAL VIEW

This islet is formed, in part, of an old coral limestone reef covered with vegetation, and has been fractured in several places. Some of the fractures are filled with water; others have been overgrown by the forest. This is an excellent demonstration of the hypothesis put forward by Brouwer that many East Indian reefs have formed on the crests of moving geanticlines. (Photograph: Fairbridge, 1943.) [*cf.* fig. 163.]

ing water courses into which the tide flows. The mangroves which live there have probably done so since the Flandrian transgression.

The modern reefs resemble those of Queensland in that a sandy platform (*cay, key*) is present which is sometimes cemented to form a beachrock. The common occurrence of living reefs in parallel ridges seems to be linked with the orientation of the dominant swell (for example, Nosy Foty, fig. 166). Most of the islands rest on a basaltic substratum formed by lava flows of Cretaceous or Tertiary age.

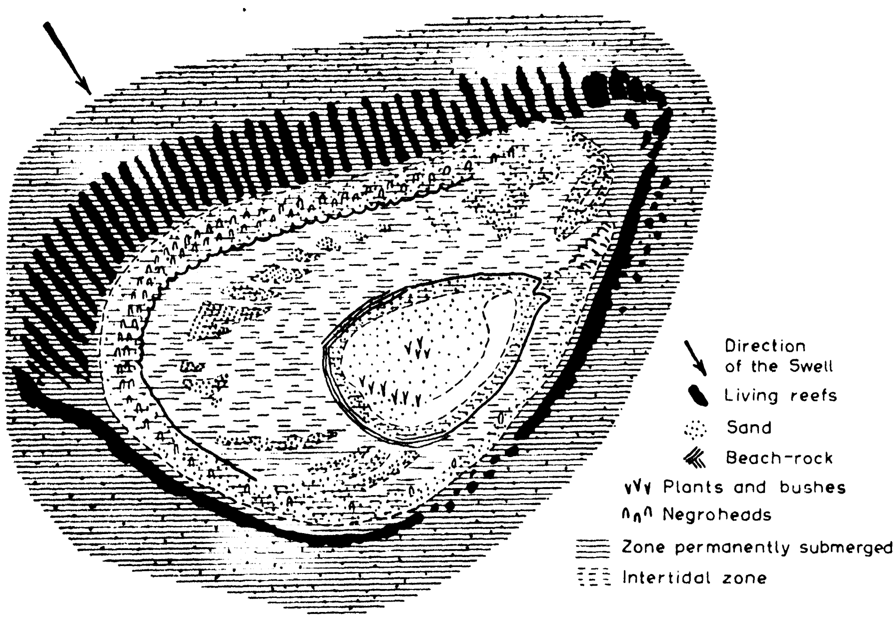


FIG. 166. THE NOSY-FOTY REEF NORTHWEST OF MADAGASCAR (after Guilcher, 1956)
Note the "horns" of sand and reef debris driven in across the reef-flat by waves under the prevailing wind. The general pattern is identical with the Low Isle type of the Great Barrier Reef of Australia.

Apart from these island reefs, there are a number of fringing reefs (Analalava, Nosy Bé, Ampasindava peninsula, Nosy Ankarea and Nosy Mitsio).

In front of Analalava, a fossil reef resting on Cretaceous rocks is raised almost 5 feet above sea level. The modern reef has taken the place of its predecessor. The living reefs are covered by Cymodocea, and the ridges by oysters.

Pools are frequently observed on the surface of the reef-flats. They are often crescentic in shape, resembling the "negative" of a barchan dune. These pools contain micro-atolls which have overhanging margins, indicating a reaction against choking muds.

From the northern tip of Madagascar a barrier some 20 to 40 feet below the surface extends for 220 miles southwestward to Analalava and is separated from the land by water 60 to 160 feet deep. It is cut by passages 160 to 260 feet deep which probably correspond to the prolongation of great rivers. This barrier is almost entirely covered by sand, and few of the coral polyps are still living, in spite of the generally favorable conditions. It is possible that this is, in fact, a noncoralline submerged ridge, such as a consolidated Pleistocene dune.

In most cases, except in the Orangea-Nosy Vaha region, the Madagascar reefs seem to be very recent.

Reefs of the Marshall Islands and the Atolls of the Pacific.—Most of the Pacific atolls, and especially Bikini (Marshall Islands), occur on ancient basaltic volcanoes. The atolls of the Marshalls cap two parallel volcanic chains. In the same region, there are guyots, that is, submarine volcanic cones truncated at their summits (see p. 58) which are between 3,000 and 5,000 feet below the surface of the ocean. These platforms sometimes bear Late Cretaceous coralline limestones, ripple-marks, rolled pebbles, or vesicular lavas, which are all characteristic of shallow depths and indicate that these volcanoes were eroded by waves at the end of the Cretaceous. Furthermore, the platforms have since sunk progressively from the surface down to their present depths. This is definite proof of subsidence. It is probable that only those which sank slowly were able to keep their population of living corals, in the form of atolls.

Examples of Fossil Reefs

Although a certain number of modern reef-forming corals are known also in the fossil state, many "bioherms" contain organisms which have no living descendants. Their ecological interpetation then becomes exceedingly difficult (Termier, 1952, p. 223–229). Moreover, there is undoubted and recurrent evidence of a close link between paleogeographic conditions, movements of the earth's crust, and the ecology of living organisms. The reef, an organic complex, has had its morphology molded by these three factors.

The Niagaran Barrier Reef of the Michigan Basin.—The paleogeographic map of North America during the Niagaran (middle Silurian) shows a ring of bioherms round the Michigan basin. They are well known in the western part of New York State, in Ontario, Michigan, Ohio, western Indiana, Illinois, Iowa and in eastern Wisconsin. To the north they appear to link up with the reefs of the Arctic.

Each bioherm is lens- or dome-shaped and forms a feature in the topography (sometimes known as "Klintars"). The centre is composed of dolomite or of nonstratified limestone containing abundant Stromatopora, Tabulata, Cephalopoda and *Stromatactis*. The outer part is formed of thin-bedded limestone or of well-bedded shale dipping steeply away from the dome. These beds form the flanks of the reef and result partly from the formation of limestone *in situ* and partly from the accumulation of reef debris. They pass laterally into the normally stratified, generally horizontal beds of limestone with chert of the surrounding area, which are interpreted as deposits of a calm sea.

The flanks of the reefs of northeastern Illinois have been closely studied by Lowenstam (1948). They consist of porous dolomite or dolomitized limestone which passes into green clay and contains a fauna of robust organisms, together with much debris. This consists of compound corals (*Favosites*), the colonies of which are joined by corallites of *Synaptophyllum*; large solitary corals; trilobites such as the Illaenid *Bumastus* and crinoids (Crotalocrinoidea). In the shales between the reefs there are fragments of crinoids (*Pisocrinus*) and Eucalyptacrinoidea, small isolated corals (*Diaphorostoma*), and fragments of sponges, brachiopods and trilobites.

Some of the reefs, for example those of Marine Pool, contain oil; the detrital limestones are very porous and constitute the "reservoir-rock". The "source-rock" appears to have been of different age, although the reefs themselves and the organic-rich shales between the reefs could have produced some oil.

The Frasnian Reefs of the Ardennes (fig. 167).—The Frasnian reefs on the edge of the Dinant basin were described by Mailleux and subsequently examined and reinterpreted by M. Lecompte (1954 and 1956). Two types of constructional organisms appear to have existed concurrently: on the one hand, there were the Stromatopora, which are sensitive to mud and live in agitated shallow water, and on the other hand, the Tabulata and Tetracoralla, which are able to live in calmer and deeper water. It is possible to distinguish massive Stromatopora reefs built in the zone of turbulence, coral reefs constructed below them, and "mixed" reefs which began as the second type and terminated as the first.

The fossil reefs of the Frasnian take different forms according to their position in the Dinant basin. *On the south and west borders of the basin*, they are almost hemispherical lenses, built up of successive caps higher in the center (due to more rapid growth) than at the periphery. Their contacts are sharp, but the marginal zones are less clearly defined and pass gradually into the surrounding argillaceous sediments. There is *no detrital talus or any central concentration of sand* as in modern coral reefs. The role of algae is apparently subordinate, although *Sphaerocodium* and *Girvanella* may be present in the middle zone of the reef, indicating that the reefs were formed in very shallow water. "Stromatactis", a calcareous efflorescence which forms a cement at certain levels seems to be formed by the activity of Cyanophyceae. These lenses suggest a generally calm environment (perhaps analogous to that of the pinnacles of lagoons of atolls). According to Lecompte (1954) the reef rhythms of the Frasnian of the Ardennes correspond to alternations of subsidence and stability, or even of uplift. The domed form of the reefs shows that they developed as subsidence continued. Moreover, the absence of necrosis of the polyps and the lack of beachrock suggests that at no time did these "reefs" reach the surface of the water.

On the *northern border of the Dinant basin* the "reefs" are no longer bioherms, but a well-stratified biostrome. This region seems to have undergone slow, uninterrupted subsidence.

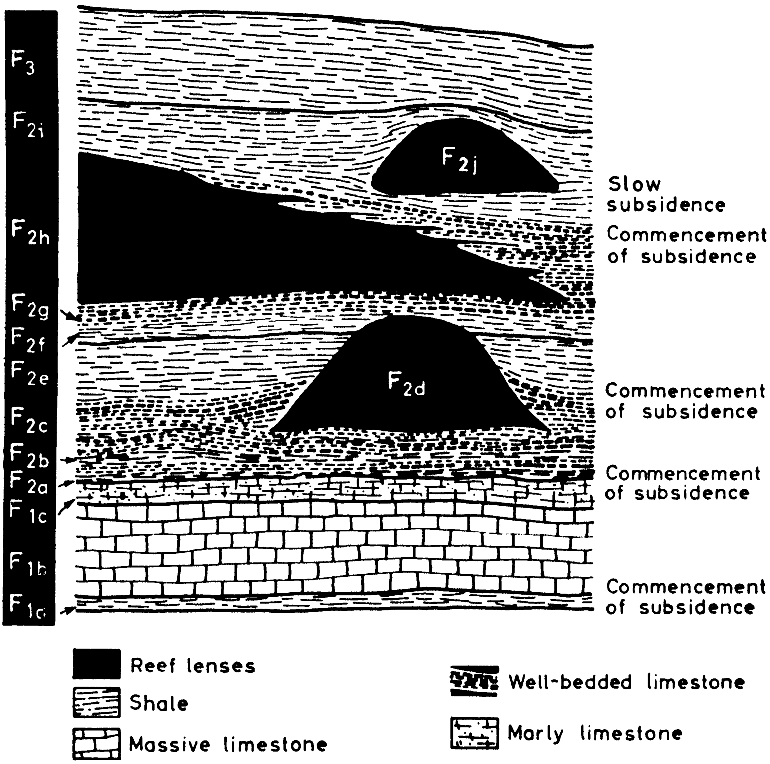


FIG. 167. DIAGRAMMATIC SECTION OF THE FRASNIAN (DEVONIAN), NEAR FRASNES, BELGIUM (after Rutten, 1956, with some information from Lecompte, 1954 and 1956)

In the intermediate zone on the *eastern border of the Dinant basin*, subsidence is reduced by the proximity of the littoral zone and has given rise to organically formed masses intermediate in type between the bioherm and the biostrome.

Thus the bioherms of the Frasnian form a *subsiding series* up to 1,300 feet thick and consisting of shales, calcareous shales and limestones.

The "Kess-Kess" of the Sahara (figs. 168 and 169).—At Hamar Laghdad (Tafilelt, Morocco) protuberances occur which are named by the local inhabitants "kess-kess" (from the name of the African copper vessel used for preparing couscous). E. Roch (1934) has ascribed some forty of these to the work of corals during the Devonian. Their form and their size varies from mounds 6 to 10 feet high to cones 100 feet high. Other groups are also known in the Tafilelt. C. Pareyn (1957) has shown that they are bioherms eroded out from the surrounding sediments: "isolated reefs, pinnacles, coral patches, imperfectly formed reefs, reef barriers formed by the joining together of several individual reefs, etc." The first are, without doubt, of Coblenzian age. Most of the others are Eifelian, and the reef barrier seems to have persisted almost to the Late Devonian. The corals are always Tabulates

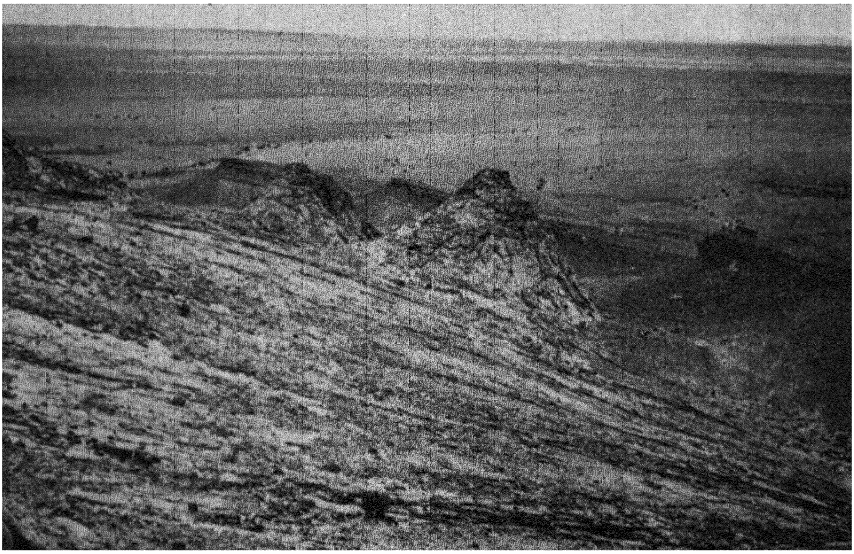


FIG. 168. REEF-KNOLLS ("KESS-KESS") IN THE DEVONIAN OF ERFOUD, MOROCCO
(Photograph: Pareyn)

(*Favosites*, *Thamnopora*, *Alveolites*, *Heliophyllum*) accompanied by accumulations of crystals ("*Ptylostroma*"). Very rapid subsidence appears to have determined the conical form of these reefs, which grew in height to avoid being buried in mud.

In the Givetian of the Sahara, J. P. Lefranc (oral communication) has seen ten similar cones at Azzel Matti (60 miles southeast of Reggan). These, however, are formed of bedded, foetid limestones containing goniatites, crinoids, gastropods and brachiopods. They enclose nodules of bituminous limestone. Elongated ridges 1,000 to 1,500 feet long are associated with the cones.

Structures comparable to the "kess-kess" are known in Russia in the Upper Carboniferous and the Permian of Tcherlitamak (185 miles south of Ufa). They are composed of masses of limestones containing brachiopods and fenestellids.

"Reef-knolls", which play a similar topographic role in the present landscape have long been known in the Carboniferous of Yorkshire (England). They appear to have formed along the edge of a subsiding reef platform, and thus tend to mark the boundary between shallow and deep-water facies.

Carboniferous Reefs of the Great Western Sand Sea (Grand Erg Occidental) of Algeria.—In the Upper Visean south of Jebel Mezarif, C. Pareyn

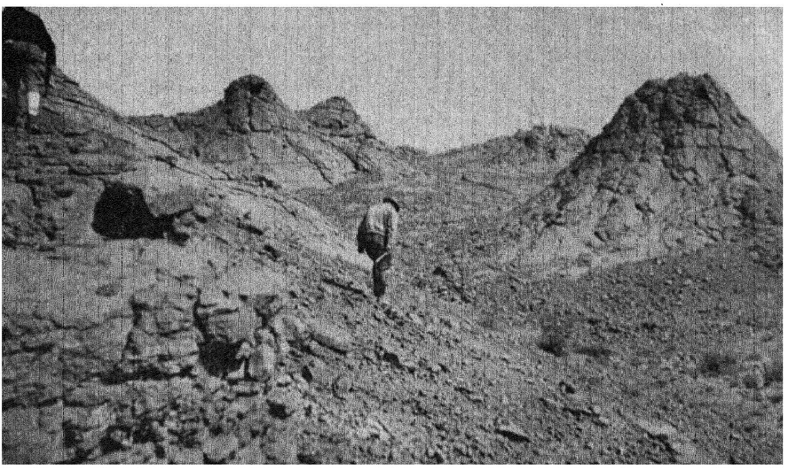


FIG. 169. FURTHER EXAMPLES OF DEVONIAN REEF-KNOLLS, ETC., ERFOUD, MOROCCO (Photograph: Clariond)

(1959) has described a series of reefs which become younger as they are traced toward the north. These are bioherms which are independent of each other. The extension of the reef domain at each level is not more than ten miles and is limited toward the north by green terriginous clays which were associated with the erosion of the fold of Ben Zireg (see p. 208). These dome-shaped reefs are 30 to 150 feet high and have a core of fine-grained limestone which is flanked by a detrital layer composed of abraded reef organisms and remnants of creatures living close to the reef. They are separated laterally and stratigraphically by organo-detrital limestones containing many specimens of *Productus giganteus*. In the subreef facies there are fenestellids and sponges. As in modern reefs, the role of algae was important; *Sphaerocodium*, it appears, was situated in the zones affected by turbulence, while Girvanellas were widely dispersed. Their disintegration seems to have provided the mud which forms the matrix of the reef. The crinoids, being more tolerant than most of the other fixed invertebrates, are abundant in the areas surrounding the reefs.

The bioherms fall into several biological categories. In the first group are the reefs containing fenestellids, sponges and "Ptylostroma" (granular calcite probably resulting from algae during their decomposition). These reefs are similar to those which, in France, form the Waulsortian facies (p. 264) (where the fenestellids act as sediment traps). They are 60 to 100 feet high and are covered by clays containing crinoids or by detrital limestones which appear to be bedded. The second group are reefs comparable to those of the Ardennes (p. 288) and characterized by "Stromatactis", which are accumulations of calcite particles produced by algae around thin branching corals of the genus *Heterophyllia.* They are associated with *Lithostrotion*, which is generally branching (sometimes massive), *Dibunophyllum* and sponges. The vertical biological succession is commonly as follows: (1) "*Stromatactis*", sponges, *Heterophyllia*, crinoids and gastropods, marking a calm period; (2) Microdetrital limestone, fenestellids and sponges; (3) *Lithostrotion.* Mixed reefs exist, commencing with fenestellids and sponges and continuing with a *Lithostrotion* phase, which suggests competition between the two associations. There are also reefs formed of beds of *Lithostrotion* alternating with beds of crinoids, which correspond to periods of turbulence. The *Lithostrotion* often forms conical mounds 8 inches to 6 feet high.

The physical conditions which underlie the formation of these bioherms appear to be well established. In a zone which tends to subside, normal detrital sedimentation favors the growth of crinoids, brachiopods and sometimes sponges and fenestellids. During periods when subsidence ceases or is greatly reduced, corals are able to develop in calm water. Abundance of algae suggests shallow water.

By the Namurian, subsidence had practically finished and the last reef phase passed into large biostromes.

The Permian Reefs of Guadalupian Age in the Southwestern United States.—The Pennsylvanian and Permian are present in the southwestern United States, in the Delaware Basin. The detrital deposits of the basin outcrop in the high ground in the southeast of New Mexico and adjacent Texas. To the east the basin is bordered by a continental upland area into which it passes by way of a zone of marginal flexures. In this zone, which is of Guadalupian age, limestones, dolomites and reefs such as those of Goat Seep and Capitan, have been formed. The dip of the talus-slopes toward the Delaware basin is 25° to 35°. The Goat Seep reef, for example, is a mass of dolomite about 1 mile wide and 1,500 feet thick. The lower half is built up of massive banks, while the upper part is unstratified and forms a true bioherm.

It is interesting that this reef and its contemporaries enclose a number of very fragile coelenterates (*Cladopora*, *Cladochonus*, *Lophophylidium*, *Lindstroemia*) in close association with sponges, bryozoans, molluscs and brachiopods, as well as Fusilinids.

These reefs are not, therefore, biologically comparable to the present-day hermatypic coral reefs. Their position in the marginal zone between a land mass and a deep basin is, nevertheless, similar to that of the Sunda Barrier (fig. 163). There is thus some indication that the occurrence of the bioherms was linked to conditions independent of those such as temperature, salinity and purity of the sea, which govern the existence of hermatypic corals.

Upper Jurassic Reefs in the Salève.—On the hill of Salève, about 4 miles south of Geneva, A. Carozzi (1955) studied a series of beds (nearly 750 feet thick) showing sedimentary "rhythms" (see p. 354). Each "rhythm" ends with the appearance of large lenticular coral reefs, with polyps in the position of growth. These masses are less than 30 feet high and are about 50 to 60 feet long. They are preceded and followed by fossiliferous clastic limestones.

The sedimentary rhythm adjacent to these reefs indicates a *progressive diminution in the depth of sedimentation*. This is apparent from the ecological succession in the rocks which comprise:

(1) *friable limestones*, formed below 150 feet and consisting of fine debris, mainly pelagic ostracodes;

(2) *pseudo-oolitic limestones*, formed at about 150 feet, containing few organic remains other than rare Dasycladaceae and a few annelids. These limestones occur as rounded fragments cemented together by cryptocrystalline calcareous "paste" formed by the reworking of interstitial reef deposits;

(3) *fossiliferous and pseudo-oolitic limestones*, deposited between 80 and 50 feet, and corresponding to areas of maximum development of benthonic foraminifers (Textularids and Miliolids), Dasycladaceae, and annelids;

(4) *reef limestones* formed between 50 and 15 feet. Following the reef limestone, a new series begins with stage 1 (above).

The rhythmic cycle thus defined is due to a regular uplift, which reaches its maximum with the production of the reef limestones. This is followed by subsidence which brings the series back to the beginning of a new sequence.

The Aptian Reefs of the Djebel Ouenza (Algeria) (figs. 170–172).—In the Aptian marls of the Djebel Ouenza and the mountains of Mellegue, G. Dubourdieu (1957) has observed lenticular accumulations of "reef-like" limestone. The sediments are predominantly marly, and their total thickness varies from 300 to 1,500 feet (Ouenza mountain).

In Aptian times the sea spread into Algeria. Its shore line lay to the north of the salt plains (shotts) of Melrhir, el Rharsa and Djerid. In Tunisia, between Kairouan and Gafsa, much detrital material of continental origin was poured into the sea and the sediments were of lagoonal type (red dolomitic and gypsiferous marls). There were, almost certainly, islands in this region.

In the beds of the mountains of Mellegue, the reefs occur on elongated folds. The lowest coastal zone, at Sidi Emmbarka, has no reefs, but has beds of oysters instead. The thickness of the series indicates that considerable subsidence occurred. Some detrital quartz coming from neighboring land masses (to the south, near the Saharan continent) was carried in by currents and mixed with the organically formed sediments.

The sedimentary series preceding the reefs include banks of small, worn fragments of calcareous rocks and fossils, with some oolites. There are also fragments of green algae and corals, but entire fossils are rare. All the debris is rounded, which indicates the action of currents and waves in very shallow water. The formation of these shoals is undoubtedly due to orogenic movements. At a much lower level alternations of marls and sublithographic limestone pass upward into oyster "gravels" and oolitic beds. The oolites, which are imperfect, "represent the final stage of transportation of more or less rolled debris". These zones are certainly the calmest and contain thick-walled foraminifers and rare corals heralding the reefs.

The unstable conditions which presided at the beginning of the Aptian were not favorable for the establishment of true bioherms. Only when the floor was stable were organisms able to build strong reefs in regions where folds previously elevated the sea bed. Here again is the concept of "reef-caps" put forward by Brouwer (see p. 283).

The limestones at the base of the Aptian contain green algae, which occur where there are few rounded fragments and almost no oolites. They appear, therefore, to have been deposited in comparatively still water. Corals and foraminifers (*Orbitolina*) are often present.

The true reefs contain several facies: fine-grained limestones resulting from the chemical or biochemical precipitation of calcium carbonate; breccias and microbreccias formed of fragments of calcareous organisms; oyster limestones; rudaceous or oolitic facies; and intercalations of marl. All these are of shallow water origin. The fine-grained limestones, often rich in *Miliola*, predominate in the central part of the bioherms and often contain rudistids. Most of the polyps are fragmentary and are rarely found *in situ*. They are commonly difficult to distinguish in the massive limestones, but they generally occur on the periphery of reef complexes. The corals are absent in those zones of limestone accumulation where rounded fragments of other organisms are abundant. Both coral limestones and oyster limestones are absent from the central regions.

The limestone accumulations show indentations into which the marls penetrate. These indentations are analogous to the "terraces" of the Belgian Frasnian. The indentations seem to have formed in the shallow sea bottom and to have resulted from the disintegration of early reefs. The lenticles which are now seen are the "ruins in which the superstructure has been battered down". They are, in fact, atolls, either subannular or elongated according to the form of the ridge beneath them. The central part, composed of calcareous mud and containing Miliolas, green algae, and rudistids, forms the largest part of the complex. It represents a zone of calm water analogous to a lagoon.

G. Dubourdieu has attempted to reconstruct these atolls (figs. 170–172).

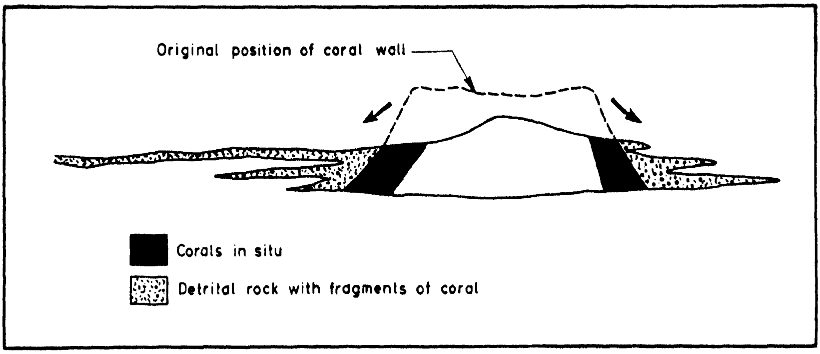


FIG. 170. RECONSTRUCTION OF THE VARIOUS FACIES OF A REEF COMPLEX IN THE CRETACEOUS OF MELLÈGUE MOUNTAINS, ALGERIA (after G. Dubourdieu, 1957)

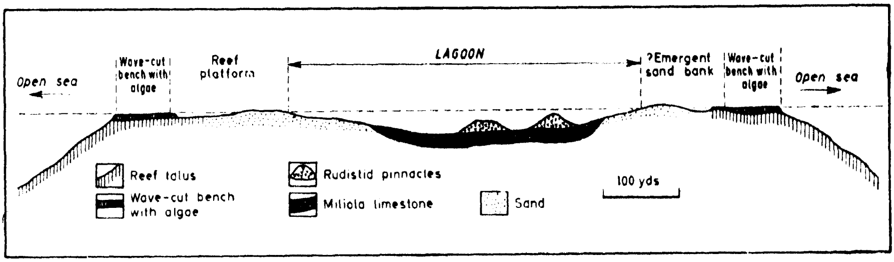


FIG. 171. HYPOTHETICAL SECTION ACROSS AN APTIAN (CRETACEOUS) ATOLL DURING GROWTH (after G. Dubourdieu, 1957)

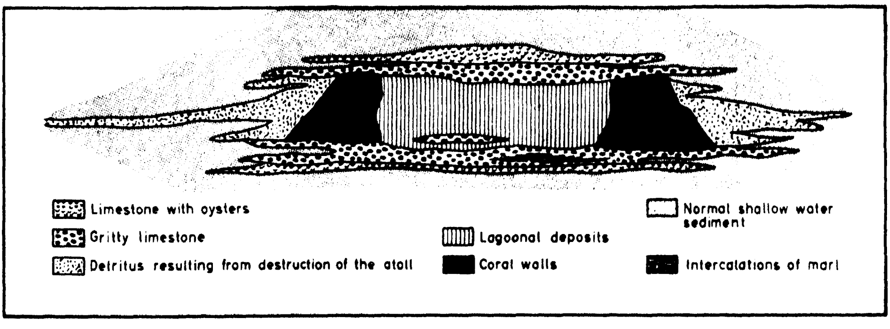


FIG. 172. STRATIGRAPHIC SECTION OF AN APTIAN (CRETACEOUS) REEF OF THE MELLÈGUE MOUNTAINS, ALGERIA (after G. Dubourdieu, 1957)

Debris covered their outer slope whose irregular surface was colonized by foraminifers, sea urchins and pelecypods (including rudistids), corals, sponges, bryozoa, crinoids and brachiopods. The worn-down ring of the atoll was covered on its outer edge by red algae related to Lithothamnium, in a position analogous to that on modern atolls. The reef-platform which corresponds to the rest of the atoll ring was composed of coarse, or poorly microbrecciated, rock fragments, and was often rich in green algae, while large rudistids appear to have lived on the edge of the lagoon. It is possible that the atolls may have been covered at times with "waterbloom", and some dunes may have formed. The lagoon supported Dasycladaceae in its shallow parts, but where the algae were absent, the water was undoubtedly deeper. Small rudistids associated with green algae formed "pinnacles" on the floor of the lagoon.

The death of the atolls, which did not survive into the Albian, seems to have been due to a renewal of tectonic activity affecting the submarine floor. Well-bedded rocks composed of rounded fragments were deposited on the limestone complexes which remained after the death of the reefs. These were followed by oyster or Orbitolina-limestones, together with intercalations of marl. Conditions similar to those preceding the appearance of the atolls thus returned.

The Urgonian Reefs of Northern Spain.—In the Asturian and Pyrenean region, the Urgonian facies extends from the Aptian to the Lower Albian. Between Bilbao and Santander, it consists of stratified beds (biostromes) and lenses (bioherms). These masses are very different from atolls and fringing reefs, but resemble coral platforms. They are characterized by their large proportion of calcareous cement which is homogeneous and crystalline. This cement surrounds the constructional organisms, which are essentially madreporians and rudistids (*Toucasia*). Algae are almost absent. P. Rat (1957) believes that the cement is formed by chemical precipitation, but it has been shown (p. 248) that the disintegration of calcareous algae can produce a similar result. In all cases the Urgonian masses began in clear water and were not brought to an end by emergence, but by burial under terrigenous debris coming from the Castilian delta.

14

Some Limestone Peculiarities and Karst

NODULAR AND "GRIOTTE" LIMESTONES

The fine-grained limestones formed from a homogeneous calcareous ooze may have had their crystallization disturbed during diagenesis. This can explain the formation of pseudo-oolitic limestones, pseudo-breccias, or nodular limestones, which can be recognized by the morphological characters of their component parts. These components are bound together in the fresh coherent rock by a thin ferruginous or argillaceous film, but, after weathering, they readily fall apart.

Nodular Chalk

Rocks of this type are known in the Paris Basin and in England and have been described by L. Cayeux (1936, 1941). They are either interbedded or discordant with the normal Upper Cretaceous chalk and are evidence of the disturbance of the sea bed where it has been raised almost to the point of emergence far from the coast. These rocks are chalky, variable in hardness, and contain glauconite, calcium phosphate, and hematite, which were incorporated in the rock at the time of the uplift (after the deposition of the chalk). There was thus a temporary interruption in sedimentation and a period during which currents may or may not have eroded the surface of the chalky ooze.

"Griotte" Limestones

The term *griotte* was defined in 1837 by Leymerie as a red, "knotty" marble in which the nodules had formed around goniatites and Clymenias. The angular limestone fragments are very similar to the nodules of nodular limestones and are held together by a ferruginous or clayey cement. Most of the griotte marbles which have been described have come from the Upper Devonian (Fammenian) of Europe and North Africa: The Black Mountains (France), Mouthoumet, the Pyrenees, Spain, Morocco and the Sahara. It is, in fact, a lithofacies closely comparable to the nodular limestones of the same age and showing, moreover, similar fauna (goniatites, orthoceratids). These fossils are not always present, and the formation of griottes cannot be attributed to them. The presence of pyrites suggests a nearby "sulphuretum" (sulfur source) and there is sufficient iron oxide to give the rocks a characteristic red tint. Several examples occur in Morocco and in the south of France.

In *central Morocco* on the southwest flank of the Gara of Mrirt (gara = a rock isolated by wind erosion) a bed of griotte limestone crops out in banks 2 to 24 inches high, sometimes forming a homogeneous mass and sometimes alternating with brittle shales. The succession is 65 to 80 feet thick and certainly spans the Frasnian and the Fammenian, and probably the Strunian stages of the Devonian.

This mass is formed of a "brecciated rock with a very irregular surface covered in minute oxide ridges, brought into relief by rainwash. The fractured surface reveals sometimes a light brown sublithographic limestone and sometimes a more or less friable sediment which is generally gray in color" (H. Termier, 1936, p. 374).

About 7 miles northwest of Mrirt, in the region of Dechra Aït Abdallah, it is possible to study the history of the sedimentation of a nodular griotte limestone. The Eifelian includes limestones, generally in the form of flagstones, containing *Tentaculites*, which are believed to be pelagic organisms. A nodular limestone containing goniatites, orthoceratids, and trilobites, several feet thick, has served as a marker band in mapping the area on a scale of 1/10,000. Above this, some of the flaggy limestones contain algae and, becoming conglomeratic in places, resemble the griottes. A little higher, the flagstones have yielded an abundant flora of Psilophytales (H. and G. Termier, 1948, 1950).

The Givetian, which succeeds the Eifelian, is a reef facies, which has been almost entirely reworked. The corals form more or less rounded pebbles which occur in several of the overlying horizons (see p. 208). Locally the Fammenian is distinct and always occurs in the form of griotte limestones alternating with shales which sometimes contain ostracodes. The griottes of this region "are nodular limestones with a clayey ferruginous cement and may be considered to be monogenetic breccias. The angular pieces of limestone are believed to have been picked up from unconsolidated sediments by currents and waves, and dropped at some distance from the shore in calmer water where clays are being deposited. The nodules very often contain goniatites or clymenids" (J. Agard, P. Morin, G. Termier and H. Termier, 1955) (fig. 173).

Associated with the griottes are limestone "nests" almost entirely composed of brachiopods or rhynchonellids (*Halorella*), representing a more sheltered facies. The Devonian is followed by the conglomerates of the Strunian (which forms part of the same cycle, but is earlier than the great gap corresponding to the Tournaisian and the Lower Visean). These conglomerates are rather similar to the griottes of the Fammenian and the Givetian since they contain, almost exclusively, limestone fragments and rolled corals derived from these stages. The fragments are, however, eroded from an already indurated limestone and are not derived (as in the earlier stages) from unconsolidated sediments in the process of formation. Well-bedded sandy limestones occur in the Strunian. This stage can be sandy for considerable thicknesses. The cement of the Strunian conglomerates contains rounded grains of quartz, and the rock as a whole indicates that considerable material has been derived from the land. The fauna enclosed in the

FIG. 173. FAMMENIAN "GRIOTTE" LIMESTONE (DEVONIAN) IN CENTRAL MOROCCO
Note the sections of goniatites. (Photograph: G. Termier.)

cement is composed essentially of crinoids, brachiopods and rare, simple corals. It is a fauna of sandy limestones, and differs from that of nodular limestones and griottes.

A. Ovtracht and L. Fournié (1956) examined the griottes of the Pyrenees, the Corbières, and the Black Mountains (France) and have distinguished three principal facies:

(*a*) *Intraformational conglomerates* in lenses not more than 15 feet thick and a few hundred feet long. These are monogenetic breccias consisting of limestone fragments which are only slightly rounded, and sometimes are angular, with a cement of purplish-red marly clay.

(*b*) *Griottes "sensu stricto"* composed of alternate beds of shale and limestones, irregularly corrugated. These griottes swell into nodules which often enclose goniatites, orthocerids, pelecypods or crinoids. This facies contains little detrital material, but is sometimes rich in limonite, and authigenic chlorite and sericite. It also contains finely divided plant debris and spores. A griotte limestone at Couflens has calcareous nodules in the form of drawn-out almonds, of which the extremities have a tendency to curl over, as do the more or less anastomosing filaments of the shale of the matrix. These illustrate the extreme plasticity of the material and the phenomenon of sliding on the sea floor.

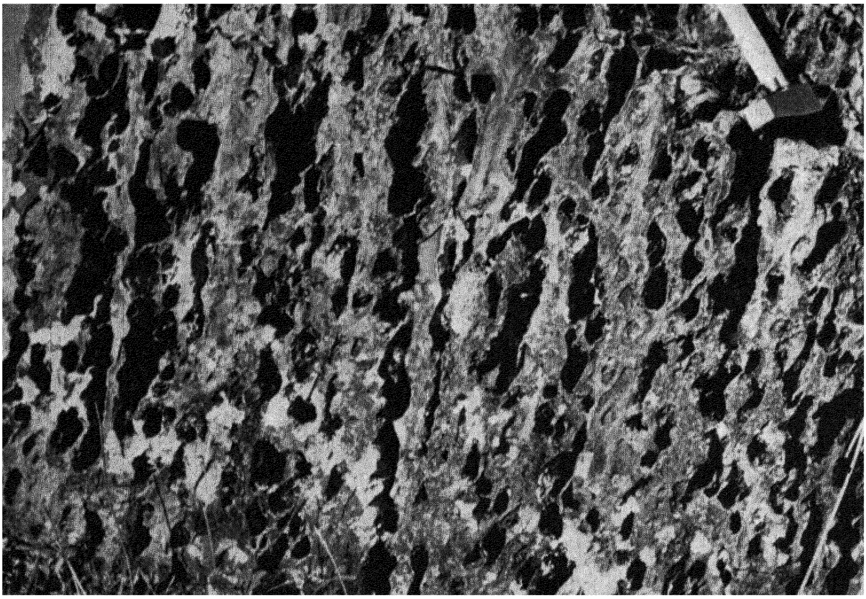


FIG. 174. A SILICIFIED LIMESTONE, MISSISSIPPIAN OF MOROCCO, IN WHICH THE LIMESTONE NODULES HAVE, IN PART, BEEN REMOVED BY ALVEOLAR EROSION

The northwest side of the Jebel Aouam, to the north of the big bend in the Oued Akerkour-Norma (Central Morocco). (Photograph: G. Termier.)

(c) *Spotted limestones*, compact and red or green in color, occurring in the Black Mountains and in the Courbières. The closely packed limestone fragments are ovoid or fusiform and are separated by a thin argillaceous film.

The griottes of the Pyrenees contain nearly 2% of manganese and some of them grade into dolomites.

The oldest known griotte limestones are the "scoriaceous limestones" of the Lower Cambrian of Morocco (G. Choubert, 1952) in which the nodules have been dissolved out by meteoric water, leaving the cement upstanding. Similar limestones are known also in the Visean of Central Morocco (H. Termier, 1936) where the etching of the rock has been favored by silicification of the matrix (fig. 174).

The Guillestre Marble

Among griottes of different age may be noted the Guillestre marbles of the Briançon and sub-Briançon zones of the French Alps. This bed is a pseudo-brecciated layer of Early Malm (Late Jurassic) age resting on a Jurassic succession of variable thickness and facies. It is transgressive over the Triassic and rests in some places on the Middle Jurassic (Blanchet, 1934, p. 78). It is a bed of kidney-shaped nodules set in a shale matrix. Its color, like that of the griottes, is red or green. It contains ammonites. This facies was formed in shallow water where the shoals of the Briançon cordillera originated.

Conclusions Relating to Griottes

In general, the griottes appear to be limited to those zones involved in orogenic movements, for example on the flanks of cordilleras. As a result of this localization and on account of their structure they can be considered as *evidence of reworking at the beginning of diagenesis* which is almost simultaneous with sedimentation. They thus differ from conglomerates wherein diagenesis only begins after the rock is indurated.

The griottes have probably been formed in very shallow seas and often occur on an unstable floor where slides within the muddy sediment can take place during deposition.

The "Calcare Ammonitico Rosso" (Red Ammonite Limestone)

This is a marly, nodular limestone facies, red, pink, greenish or gray in color, which is well developed in Italy in the Lombardy and Venetian Alps. It occurs principally at two levels: in the Toarcian (*Phylloceras* and *Lytoceras* limestones) and in the Dogger (*Aptychus* limestones).

Almost identical facies are found in several Mediterranean countries; in Andalusia (Sinemurian, Callovian, Oxfordian, Kimmeridgian), in Morocco (Middle and Upper Lias, Callovian), in Algeria (Toarcian, Bajocian, Callovian, Oxfordian) and in Tunisia (Oxfordian, Tithonian).

The nodules, up to 3 inches across, are almond-shaped and consist of fine-grained pelagic limestones, almost free from detrital material. They often contain ammonites. The matrix is more or less a paper-shale enclosing fragments of echinoderms.

According to G. Lucas (1955) the rock was formed from a single uniform sediment, without large blocks. He believes that the nodules resulted from local cementation where the environment was reducing. The environment of the matrix, on the other hand, was oxidizing, and iron from the terriginous laterites was transformed into hematites. The compaction of the sediment was due to migration of "imbibed" water, and friction striae can sometimes be observed in the surface of the nodules. This compaction is probably associated with the pseudo-brecciation.

This conclusion is very different from that put forward earlier and it is evident that further research is required to reconcile these diverse opinions.

GEOMORPHIC DEVELOPMENT OF SOLUBLE ROCKS: THE EVOLUTION OF KARST TOPOGRAPHY (figs. 175–183)

The weathering of evaporites and limestones consists essentially of their solution in water. This is in contrast to the erosion of crystalline and detrital rocks (which is chiefly abrasive) and much more simple than the process of laterization which is also of a chemical nature.

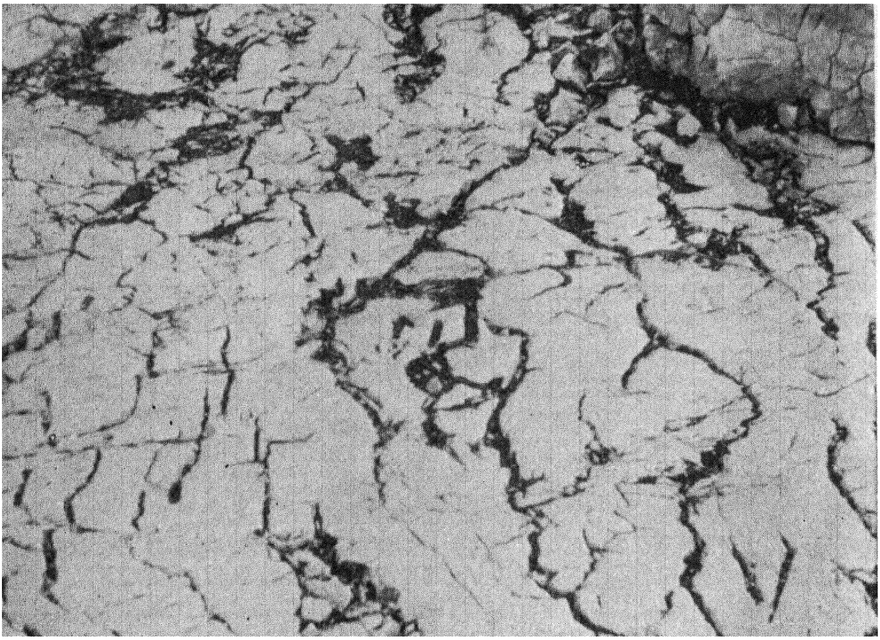


FIG. 175. FISSURING OF A LIMESTONE SURFACE IN THE JEBEL AOUJGAL, REGION OF MRIRT, CENTRAL MOROCCO (Photograph: G. Termier)

The evaporites rarely form important masses, but when they do they show the effects of solution most markedly. This is so in the salt mountains of North Africa (e.g. Djelfa in Algeria) and the outcrops of gypsum in the French Alps (near Pralognan and the Izoard and the Galibier passes). Similarly, the gypsiferous shales of the Permian Irwin basin (Western Australia) have given rise to sink holes, miniature canyons, and an underground drainage system, that is, a region of "badlands".

In contrast, the great tablelands of *limestone* have permitted the study of the morphological evolution of landscape resulting from the effects of solution. To such areas, the name *karst* has been given after the region in Istria which furnishes the finest examples.

The evolution of karst topography has been described in detail in many treatises on geomorphology and only a very brief outline will be given here.

In this evolution, water acts not only at the surface but also at subterranean levels, where it is situated above an impermeable horizon. On the surface of an extensively exposed bed of limestone the simplest solution phenomena are the small swallow holes (fig. 176) which rapidly coalesce and give rise to small channels which penetrate deeper and deeper to form *clints* (fig. 177). Often the limestone ridges themselves are cut again by being hollowed out in a different direction, and ultimately isolated mounds and pillars are produced. This type of sculpture is sometimes called *karrenfeld*, and there is a particularly fine example of it in the Causses region of the Massif Central (France) at "Montpellier-le-Vieux" near to the Roque-Sainte-Marguerite (Aveyron). In most cases, as one would expect, clints follow the direction of run-off. This can be seen near the head of some rivers, for example, in the Baumes cirque in the Tarn Gorge (France). This process may result in the removal of the greater part of a bed of limestone and leave only residual hills, called *hums*, which themselves have been hollowed out by water.

At depth, water dissolves the limestone as it follows joints and fissures. Thus it erodes *underground* at the level of the water table and forms a subterranean network of shafts, caves, siphons and passages which attract speleologists and tourists. The caves at Han (Belgium), the Causses grotto and the celebrated potholes at Padirac are well known (fig. 178).

When the limestone is very thick, surface rivers may carve out deep canyons, such as the gorge of the Tarn (fig. 179). Sometimes the surface water plunges down into the underground system by way of a *sink hole.* (One of the most famous examples in Europe is that of the Rhone at Bellegarde.) The river, however, retains its individuality and may return to the surface through a *resurgence* or *spring* (e.g. the "Source" of the Loue in the Jura, fig. 180).

The development of karst can thus deprive areas of their surface streams (e.g. Yucatan, figs. 182 and 183). They are, in fact, regions of dead (or dry) valleys. Communication between the surface run-off and the underground system is maintained by swallow holes, which may be formed in several ways (Maksimovitch and Goloubeva, 1952). They vary very much in size, being mere depressions or large *dolinas.* The smaller ones may expose bare rock, but the dolinas may support a much thicker cover of vegetation than their surroundings. Continued development may lead to the coalescence of several sink holes and the formation of vast depressions or *uvalas* several acres in extent.

The *false craters* of Morocco are a consequence of the formation of dolinas.

The Causse of the Central Atlas mountains is essentially a plateau of

Fig. 176. Solution Hollows—the Initial Stage in the Formation of Clints in a Silurian Limestone at Lake Mjøsen, Norway (Photograph: G. Termier)

dolomitic limestone of Early and Middle Liassic age. The plateau shows karst development and has a number of solution swallow holes. Between Azrou and Timhadit, the plateau has been pierced by some fifty volcanoes which have poured out thick lava flows, mainly of basalt. The development of the karst has occurred since this volcanism, and has caused false craters to form. Close examination of one of these holes shows that it has been produced by the solution of the limestone beneath the basalt crust and has been followed by the collapse of the lava roof over the dolina. All the stages of this process can be seen, and the structure produced simulates a true

FIG. 177. THE BEGINNING OF CLINT FORMATION AT THE SUMMIT OF MT. SALÈVE, FRENCH ALPS (Photograph: G. Termier)

crater. The cavities are 30 to 500 feet across and 30 to 100 feet deep. The largest are occupied by extremely dense vegetation (H. Termier, 1936, p. 170).

The base level of a karst is formed by the water table, which is itself dependent on the position of impermeable horizons. At the end of the evolution of the karst the water courses have reached the level of the water table and are therefore, underground. The surfaces of such basins generally form an enclosed karst plain (or *polje*) on the limestone massif.

There are five typical phases in the development of the karst cycle:

(1) the emergence of a limestone region on which a stream system develops;

(2) abrasion and peneplanation leads to the removal of the noncalcareous terrain, and the denudation of the limestone;

(3) further uplift of the region causes the stream system to cut deeper: this is the *initial stage of the karst*;

(4) the surface drainage disappears completely and an underground drainage system develops: this is the *stage of karst maturity*;

(5) removal of the greater part of the limestone which lies above the water table: this is the *stage of old age of the karst.*

FIG. 178. THE PADIRAC CAVE (SOUTHWESTERN FRANCE), LOOKING TOWARD THE EXIT (Photograph: G. Termier)

The karst erosion ends only when all the limestone is removed, or a marine transgression invades the area. In fact, erosion levels are well preserved on the limestone (Baulig).

Maucci (1953) gave the name "*castelnuovan*" to a stage subordinate to stage four above. The term was used to describe the stage when tributaries on noncalcareous terrain continue to run after the disappearance of the principal water courses. The finest examples are found in Istria (see pp. 310–312).

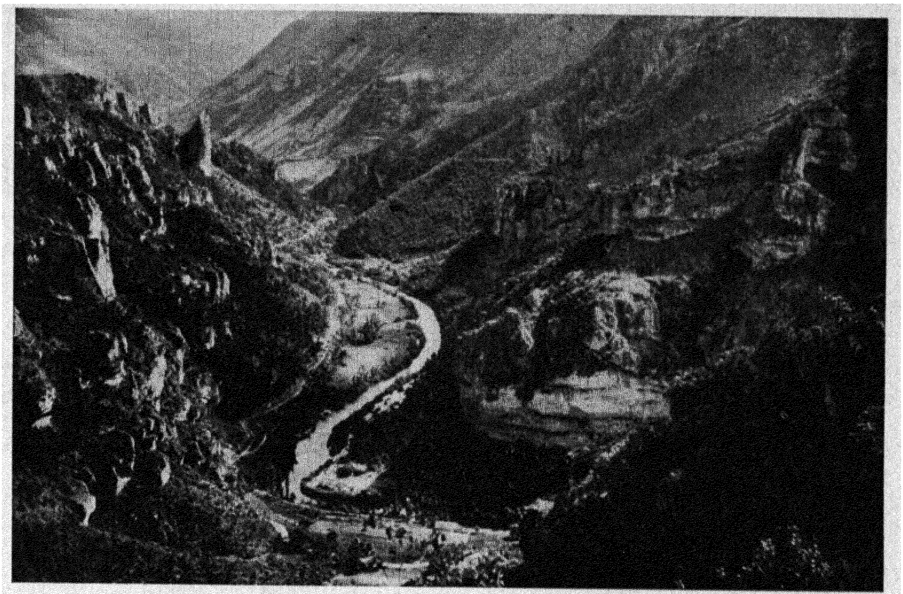


FIG. 179. A KARST IN A LIMESTONE PLATEAU IN THE SOUTH OF FRANCE. THE GORGE OF THE TARN, SEEN FROM THE POINT SUBLIME (Photograph: G. Termier)

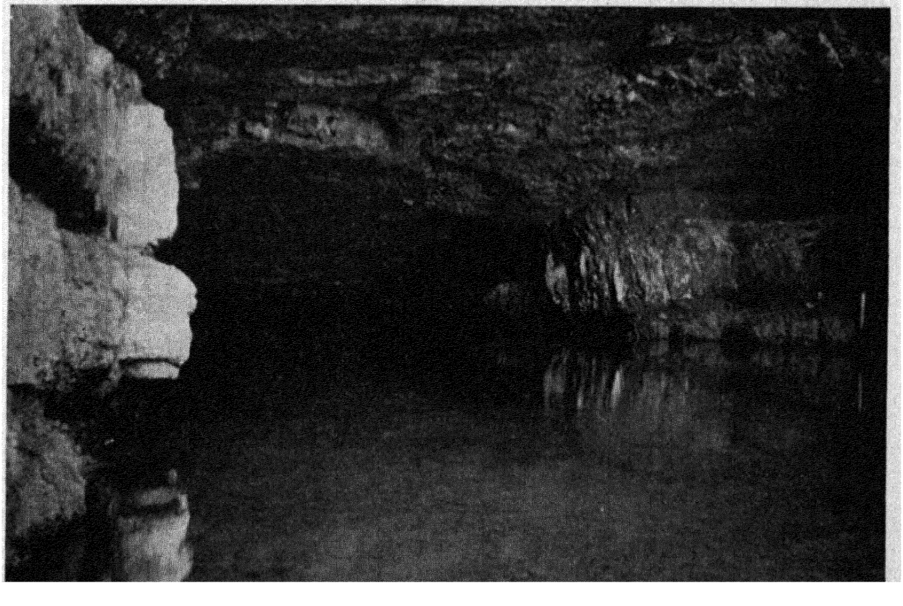


FIG. 180. THE REAPPEARANCE OF THE RIVER LOUE AFTER AN UNDERGROUND SECTION, FRENCH JURA MOUNTAINS (Photograph: G. Termier)

Yucatan (fig. 183)

Situated in a tropical zone, Yucatan demonstrates a simple type of karst since it occurs in *an unfolded region.*

It is a low-lying plateau (less than 650 feet high) formed by the coastal plains resting on a platform in the southern part of the Gulf of Mexico and bordered, as is the whole of this area, by lagoons which form behind sandy bars. It is composed chiefly of coral limestones (Miocene in the south, Pliocene in the north). The orogenic relief is practically nil, and is broken

FIG. 181. A DOLINA IN THE VERCOURS, 6,500 FEET ABOVE SEA LEVEL, NEAR THE TERMINUS OF THE MOUNTAIN RAILWAY ABOVE VAILLARS-DE-LANS, ISÈRE, FRANCE (Photograph: G. Termier)

only by the line of the Ticul Hills which rise to 460 feet to the north of Uxmal. When seen from the air, Yucatan appears to be a *region without surface water courses*, because all the drainage is underground. The surface has depressions known as *aguadas*, which fill with rain water during the wet season. There are also marshes or *akalches*. The forms most typical of karst are the *cenotes*, which are natural wells formed by the collapse of the roof of one or several caverns, and which are at the level of the water table. The underground water concentrates in these to form lakes which may be very deep (fig. 183). They were regarded as sacred by the Mayas who constructed their temples close to them.

The Dinaric Region

The mainland region of Dalmatia is subject to considerable variations in level of the water table; basins are transformed intermittently into lakes and the dolinas into ponds. Like nearly all modern coast lines, the Dalmatian littoral zone is a coast of transgression (submergence). It is part of a

FIG. 182. AERIAL VIEW IN THE PENINSULA OF YUCATAN, MEXICO, WHICH HAS BEEN SUBJECTED TO KARST EROSION

About thirty small, light-colored circles, locally known as "cenotes", can be picked out in the forest. (Trimetrogon photograph: Mexican Military Cartographic Department. By permission of the Secretary of National Defence.)

region which was folded at the time of the Alpine movements. Consisting principally of limestones, it is well suited to the development of karst topography. The rivers which are often underground, generally terminate in rias.

By way of subterranean channels the sea penetrates into the karsts situated in the interior of the country. This happens in the case of the Lake of Scutari, among others (Baulig, 1930).

Istria (figs. 184 and 185)

Since the Late Tithonian, Istria has been subjected to an alternation of emergences and submergences. Consequently, it has developed, in turn, as an area of karst erosion and as an area of limestone deposition. The succession of events there, shows: (1) a Neocomian karst topography, followed by (2) the Cenomanian transgression, (3) a Senonian karst landscape, followed by (4) a transgression which commenced in the upper Senonian but which continued almost to the end of the Oligocene, (5) emergence and new peneplanation in the Miocene. The last stage has given

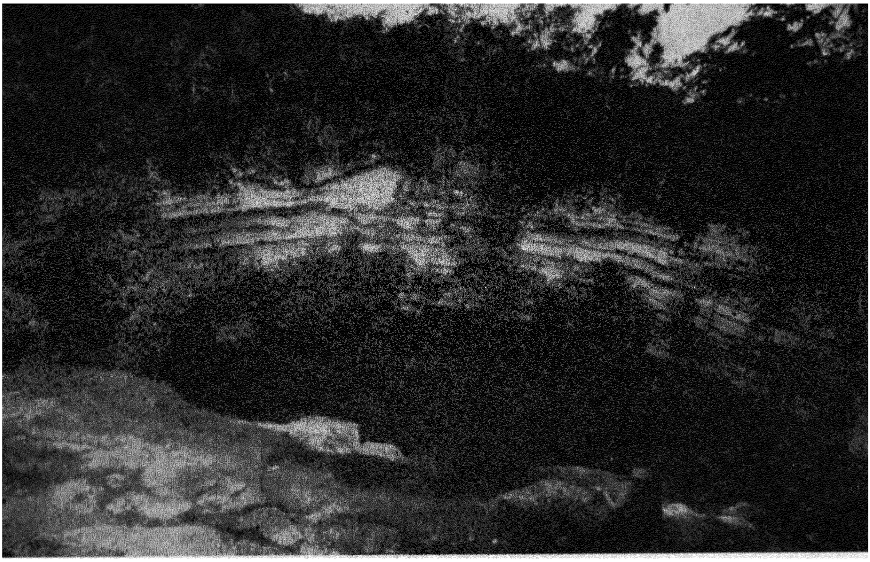


FIG. 183. THE "SACRED CENOTE" OF CHICHEN ITZA, YUCATAN, MEXICO (Photograph: G. Termier)

rise to a stream pattern termed "prekarst" which was established between the Early Miocene (Lupolano) and the Pontian. Rivers dating from this time include the *Paleobogliuno*, the *Paleofoiba*, the *Paleorisano* and an early course of the *Castelnuovo* (fig. 184). Then, perhaps due to an uplift of the region, there was a renewal of the development of the karst landscape (C. d'Ambrosi, 1954) and the stream pattern began again to cut down into the limestone. The rise of sea level due to the Flandrian transgression checked this process. Traces of karst topography can now be found from 5,000 feet above sea level down to 300 feet below sea level (in the Gulf of Fiume). Today the only evidence of the courses of the ancient streams are the numerous slots of their mouths which are similar to fjords. These are *paleo-rias* (fossil rias) (as for example, the port of Fionana). Some of the ancient rivers have been captured (the *Paleobogliuno* by the *Arsa*, the *Paleo-fiumara* by the *Quieto*). The main courses of the Dragona and the Castel-nuovo are barely functional, but their tributaries continue to flow (Castel-nuovan phase, see p. 306). Several of the "resurgences" of streams became submarine after the last transgression.

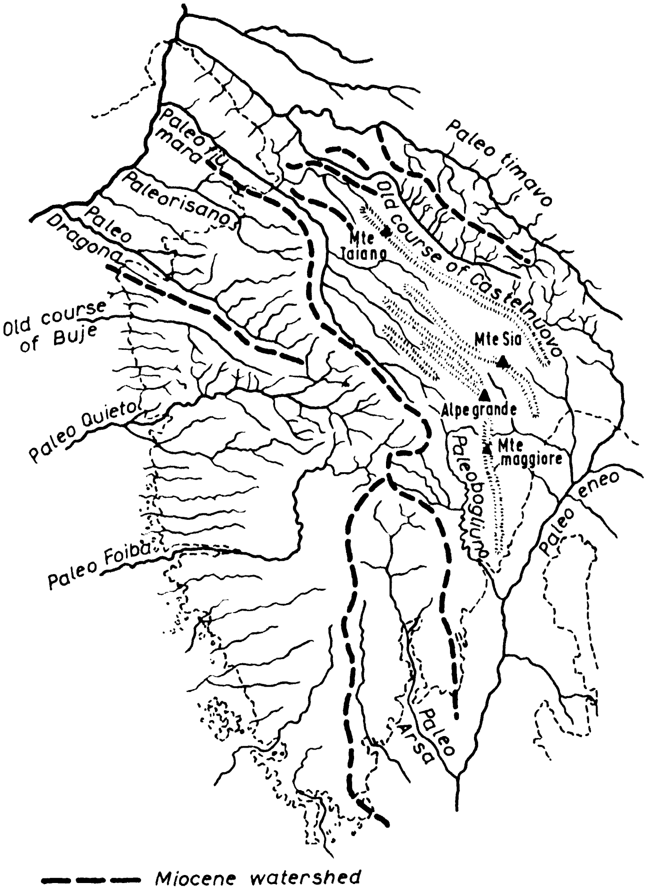


FIG. 184. THE KARST EVOLUTION OF ISTRIA, YUGOSLAVIA (after d'Ambrosi)
1. During the Quaternary, before the Flandrian transgression.

Provence

The broad surfaces of the limestones of the high plateau of Vaucluse are affected by karst erosion which give rise to the typical dolinas, potholes, clints and "stone-fields". Part, at least, of this karst scenery (for example, that at the foot of Ventoux) was formed during the Eocene, but has since become blocked and then rejuvenated. There are many resurgences, of which one is the Fountain of Vaucluse. The plateaus of the Ardeche and the

Garrigues form benches which have also been subjected to ancient karst erosion. There are few clints or dolinas, but there is a vast amount of rubble. Resurgences also occur. They are called *boulidous* if intermittent, and *fontaines* if permanent. In all these regions the rivers form canyons. The Garrigues is an area covered with Hermes oaks, holly oaks (Ilex), junipers,

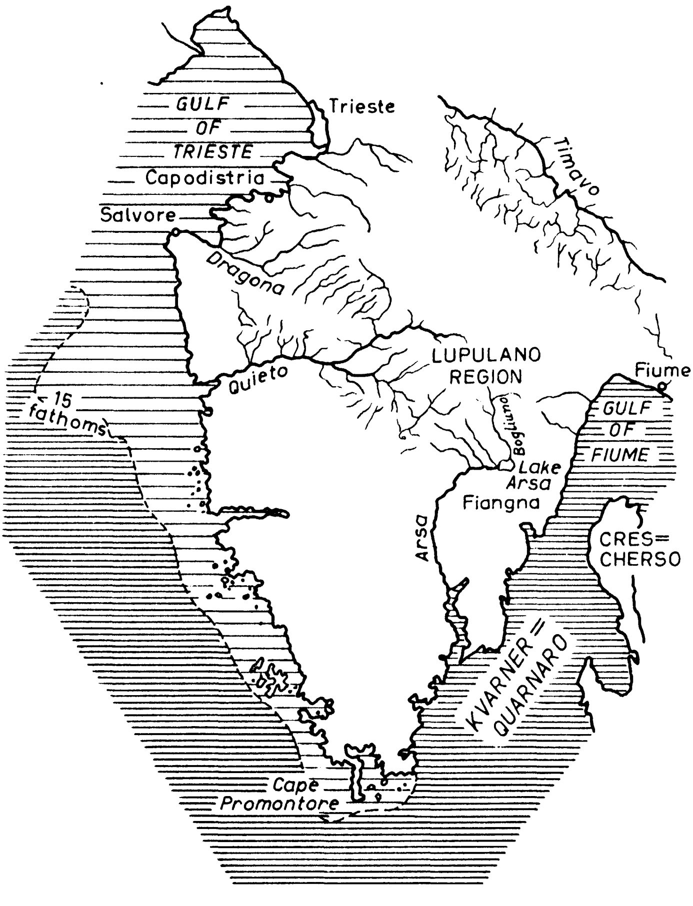


FIG. 185. THE KARST EVOLUTION OF ISTRIA

2. At the present day.

cistus, box, arbutus and mastic trees. The valleys are progressively deepened by a falling base level and karsts are formed only locally. Other notable limestone surfaces are the Plains of Orgon and the Mouriès Plateau.

There are also enclosed depressions, analogous to the Dinaric poljes: the Gard depression, the Baux depression, and the pools on the Istres-Miramas massif are examples. Most are occupied by small lakes (e.g. at Pujuat, Saze, etc.).

In the Montpellier (Hérault) region, there are low-lying limestone plateaus containing clints. An example of fossil karst topography has been brought to light by the quarrying of bauxite at Combecave-Pins. Here, the bauxite floor is a Bathonian limestone clint, whose grooves have been filled with bauxite (A. F. de Lapparent, 1956) (see figs. 76–77, p. 151).

FIG. 186. THE OUTLET OF A TRAVERTINE ("TUFA") ENCRUSTING SPRING, ALGERIA (Photograph kindly supplied by the Direction de l'Hydraulique et de l'Equipement rural, Algiers)

Travertine Deposits (figs. 186–189)

One of the incidental features of karstic erosion is the deposition of travertine around springs, the formation of stalactites in caves, and the occurrence of travertine "curtains" (for example, Hamman Meskoutine). Such deposits are often found in the fossil state. The role of algae in the precipitation of limestones has already been discussed.

Finally, there is a similar process which affects silica, as well as carbonates. Hot water charged with gas, such as that coming from geysers, forms surface deposits of material derived from the rocks traversed by the hot water. In Yellowstone Park the deposit is siliceous and soluble in water at high temperature. Geyserite is a siliceous precipitate with a composition similar to that of opal (fig. 189).

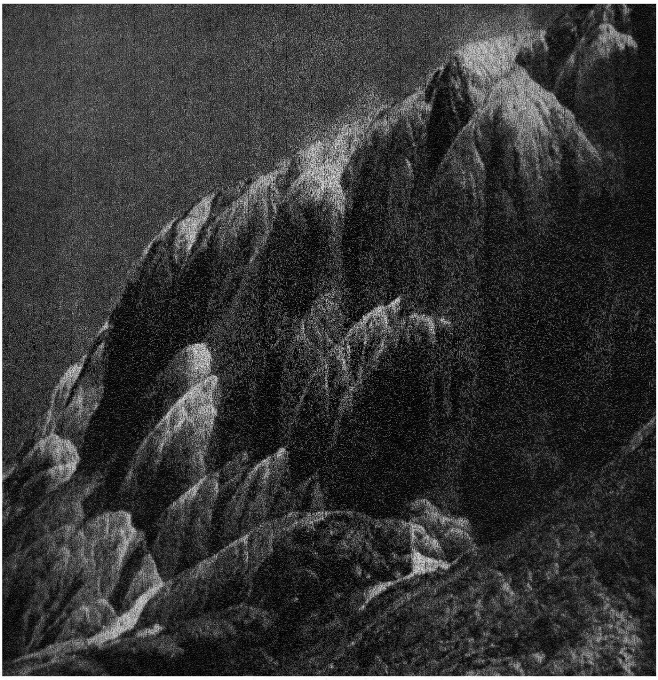


FIG. 187. TRAVERTINE DEPOSITS AT THE SOURCE OF HAMMAM MESKOUTINE, CONSTANTINE, ALGERIA

Water at a temperature near boiling point dissolves limestone from the nearby hills and deposits it in the form of travertine. (Photograph: *Service Photographique du Gouvernement Général de l'Algérie.*)

SOILS ASSOCIATED WITH LIMESTONES

The soils resting on limestones fall into two principal categories:

1. The *rendzinas*, gray or grayish-brown, which contain from 3 to 12% of organic matter and a variable quantity of calcium carbonate. Various varieties can be recognized: proto-rendzinas, which are thin and without earthworms (Kubiena, 1943); and the mull-rendzinas, full of earthworms (Kubiena, 1943). Similar types are associated with gypsum terrains (Miklaszewski, 1924).

2. Red or brown soils, *red earth* or *terra rossa*. These are abundant around the Mediterranean particularly in Carniola and in Istria (on the karst limestone) on the floor of dolinas. They are well known in Charente-Maritime on the Senonian limestone and also in the Paris basin and in the north of France, where they cover the Chalk-with-flints.

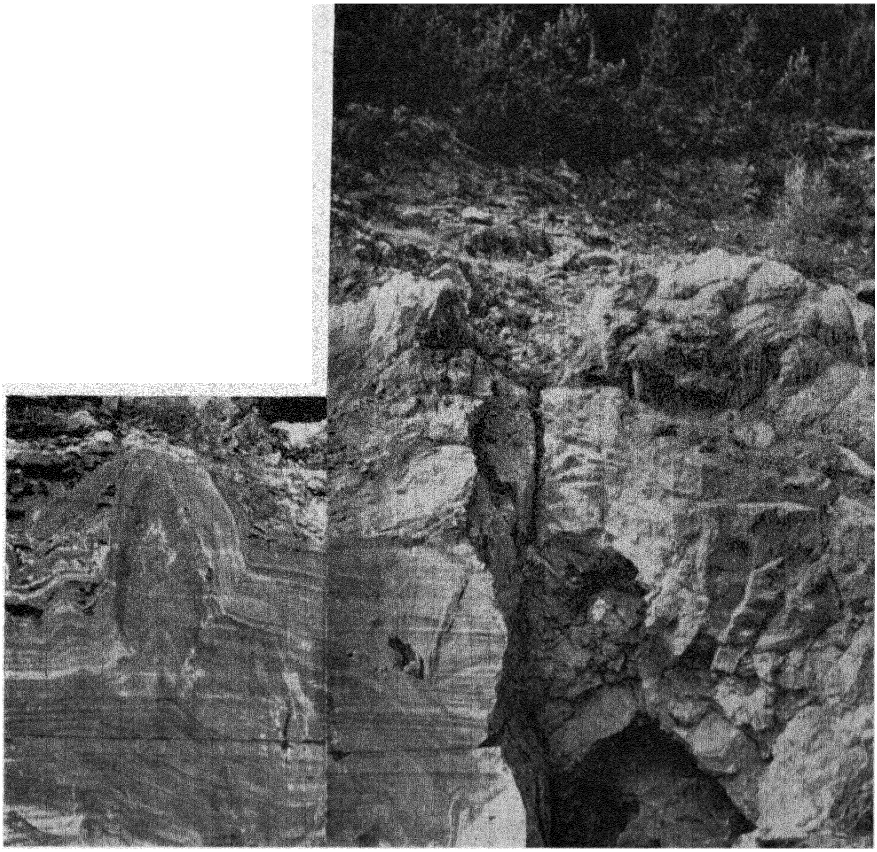


FIG. 188. THE ONYX-MARBLE QUARRY AT FONTRABIOUSE, AUDE, FRANCE: a Devonian karst

In the layers of onyx can be seen the pattern of fossil stalagmites and stalactites. (Photograph: G. Termier.)

It is generally accepted that their origin is *autochthonous* and that they are "decalcification clays" formed *in situ* at the expense of the underlying limestone. In effect, an argillaceous limestone cropping out in an emergent region and long exposed to the action of atmospheric agents will slowly be eroded and finally dissolved. The clay remains because it is insoluble. If it is not transported from the area by running water it will remain in place or accumulate in depressions and pockets in the limestone. The composition of this soil, where the argillaceous component ranges from 32 to 59%, suggests that sometimes it has been derived from impurities in the limestone (residual clay) or sometimes from some source other than the rocks on which it rests.

Thus, in certain cases these soils are undoubtedly *allochthonous*, since some *terra rossa* soils rest on very pure limestones without a transition zone.

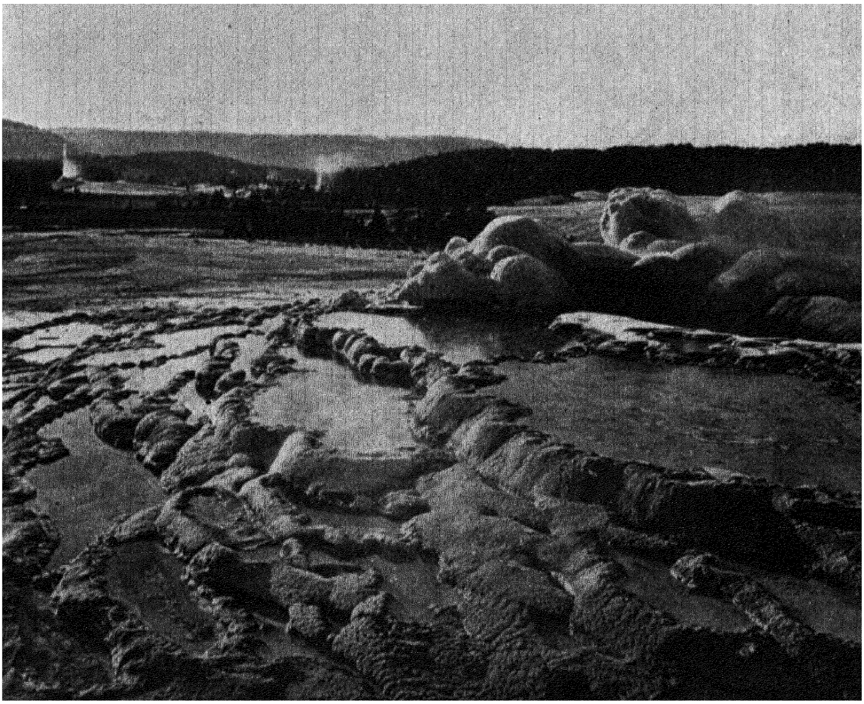


FIG. 189. YELLOWSTONE PARK. GEYSERS AND HOT SPRINGS FORMING SILICEOUS CONCRETIONS AND RIMMED TERRACES OF GEYSERITE

In these instances the boundary between the red soil and the rock is very sharp and outlines a karst-eroded surface. It is even possible that the iron of the red soil has been derived from volcanic dust.

The origin of the *terra rossa* must therefore be judged on field relations.

The bauxites have been compared to the *terra rossa* (see p. 148).

Finally, it may be noted that the Cenozoic iron ores of the Jura (*depôts sidérolithiques*) have been thought to be residual red earths which have been reworked, and in which concretions of various forms have developed (for example, pisolites) (see pp. 147 and 152).

15

Saline Sedimentation

THE ORIGIN OF SALINE SEDIMENTS

There are several ways in which saline sediments can accumulate. In the first place, there are those deposits which are of detrital origin and which are attributed to the transport of halite and gypsum by the wind. These may occur in playas and in coastal lagoons. For these to be recognizable in the geological column they must have retained their sandy texture or at least their cross-bedded dune structure, or ripple-marked surfaces. An example of this appears to be the saccharoidal ("sucrosic") gypsum of the Ledian (Eocene) of Cormeilles-en-Parisis, which according to Bourcart and Ricour (1954) is a sand often showing ripple-marks.

Two examples of deposits formed *in situ* can be cited:

Sand roses (=desert roses) which are gypsum incrustations produced in the Saharan dunes by evaporation. Very beautiful examples are found at El-Goléa and at Souf (ENE of Taggourt) in Algeria.

Gypsum crystals (baguettes) which have been observed by J. Avias (1953) in the clays of drained swamps in the coastal zone of Moindon, in New Caledonia (see p. 347).

The second of these cases is especially important in connection with the pyritic sediments which are altered to gypsiferous sediments by oxidation. In closed salt lakes, such as in coastal lagoons, most of the sedimentary deposits are black clays containing pyrite, that is, a "sulphuretum", in which the deposition of abundant organic matter is assisted by bacteria, and, in particular, by sulfur bacteria. This results in the formation of hydrogen sulfide and iron pyrite.

By atmospheric weathering and by the action of ground water, the sulfides are oxidized and give rise to sulfates (gypsum and alum) sometimes accompanied by sulfuric acid. The best example of these reactions is that furnished by the clays which have been deposited around the Scandinavian Shield since the beginning of the Middle Cambrian. The alum shales of the Middle and Upper Cambrian and of the Lower Tremadoc are so named because of the presence of large amounts of alum and gypsum formed from the sulfides which they originally contained (p. 233). In the Pleistocene, the black clays which the Littorina Sea deposited on the borders of the Gulf of Bothnia contained sulfides. These were oxidized to sulfates of aluminum, magnesium and calcium where the water table was near the surface of the ground. These alums give rise to some acid, saline soils in Finland (p. 322, fig. 192).

In the presence of calcium carbonate, the chemical reaction can be written thus:

$$2FeS_2 + O_2 + H_2O + CaCO_3 = CaSO_4 + Fe_2O_3.nH_2O + CO_2 + 4S.$$

The high salt content of Lake Eyre (estimated at 4,000 million tons of gypsum, 400 million tons of NaCl, and 7 million tons of K and Mg) raises the problem of the origin of such playa salts. Gypsum forms dunes on the borders of the lakes.

Four hypotheses have been postulated: (1) that Lake Eyre is a relict sea or salt lake; (2) the salts came from weathered marine sediments; (3) they owe their origin to substances dissolved in the water issuing from a large artesian basin; (4) they are oceanic salts carried by the wind and deposited by rain in the drainage basin of Lake Eyre. Bonython (1956) supported the fourth of these hypotheses, and estimated that one pound of sodium chloride per acre is thus deposited in the basin every year. This is approximately equivalent to 150,000 tons per year. The 400 million tons of salt which the basin contains could therefore be deposited in about 3,000 years. In connection with the second and third hypotheses it should be noted that the rivers supplying Lake Eyre contain about 15 mg./liter of NaCl and during the floods of 1949–1950 deposited 450,000 tons of salt. The weathering of marine sediments which contain 1% of salt could give rise to 400 million tons of salt in 6,000 years. Artesian waters containing 1 gm. of NaCl per liter could give rise to the same amount of salt in 25,000 years. The sodium chloride seems to have been constantly reworked, since the total tonnage is remarkably small.

The river waters carry principally calcium carbonate, together with minor quantities of sodium chloride and calcium sulfate. The amount of magnesium is very small. In rain water, only the amount of sea salt can be used as an indication of an oceanic origin. Vegetation, and the exchange of cations in the soil, seems to account for the small amounts of K^+ and Mg^{++} in the lake water. These are absorbed by clay minerals in preference to sodium (Na^+). It is possible also, although this has not been observed in the Great Salt Lake of Utah, that the calcium carbonate carried by the rivers has been precipitated in large part through the mixing of the river waters with the salt lake waters.

In Lake Eyre the gypsum is always solid, and the sodium chloride is generally solid (only a small part occurs in solution) whereas the salts of K and Mg are always in solution. It is possible that the magnesium is retained by the calcium carbonate precipitated from the river waters in the form of dolomite, while the potassium may enter into the composition of the clay minerals (jarosite, illite).[1]

[1] Jarosite, a mineral of the alum group: $KFe^{+++}(SO_4)_2(OH)_6$.
Illite, an alumino-silicate of potassium: $K_{2-3}Al_{11}Si_{12-13}O_{35-36}(OH)_{12-13}$.

According to the theory proposed by Bonython (1956) all the calcium carbonate carried into Lake Eyre will remain there. This allows an estimate of 500,000 to 20,000 years for the formation of the deposit, on the basis of the figures given above.

In general, evaporite deposits indicate saline conditions harmful to life, and contain no fossils. There are, however, cases where organisms which have come from nonsaline lagoons have been preserved in gypsum. *Protocardia tikechkachensis*, for instance, has been found in thin limestones interbedded with gypsiferous sandstones and marls in the Oued Guigou (Morocco) syncline at Sidi Saïd and at Sidi Malah (H. Termier, 1936, p. 852). To the east, near the Immouzer of Marmoucha, between Aït Tabet and the great gypsum outcrop of Aït Bazza, the same species can be found in marls containing many crystals of gypsum (H. Termier, 1936, p. 869). Three species of Modiola are associated with *Protocardia* in several exposures of this gypsiferous facies.

In the green gypsiferous clays of the Turkestanian (Paleocene) of Ferghana, O. S. Vialov (1946) has found oysters (*Fatina* and *Flemingostrea*), and a pecten (*Chlamys*) preserved in gypsum in a bed of gypsum several inches thick, which is stained brown by iron oxides. However, it appears that, in this case, the calcium sulfate is the product of the decomposition of pyrite contained in the blue clay, and does not result from primary precipitation.

The great masses of salt included in sediments, such as those of salt domes, are associated with water-bearing zones. The circulating brine may come to the surface and carve out karsts, but solution also occurs around and beneath these deposits.

Recent Evaporite Deposits

There are a number of ways in which evaporites can be formed. These can be exemplified by some modern deposits. Lake Eyre, studied by Bonython, has already been mentioned.

The principal salts of Lake Eyre are sodium chloride and gypsum. The proportion of Mg and K salts to those of Na is much less than in sea water. Nodules of native sulfur, with a crust of gypsum, are found in a laminated clay, and are believed to be of organic origin. Palygorskite (or attapulgite) is known to occur in a dolomitic mudstone.

The concentration of salts in the waters of Lake Eyre has been progressively increased by evaporation; and when the saturation point is reached, deposition can occur. During the period studied, saturation by sodium chloride (320 g./liter) occurred in January, 1952, and saturation by gypsum in December, 1951.

The salt itself forms a crust, which is often broken by small cracks from which emerge small accumulations of pure (99%) sodium chloride. The formation of this crust is irregular, since the supersaturation of the brine increases toward the surface where salt is deposited, and the top of the salt bed is raised in places. The small heaps of salt so formed are transported by the wind to form *salt islands*. On occasion, rain water dissolves some of the salt. The salt may be colored various shades of pink, due to the presence of organic substances produced by a flagellate (*Dunaliella*). This organic compound is a carotenoid which smells of violets. Some of the salt at the surface may be colored by a ferruginous dust, while some just below the surface is purple due to a mucilaginous substance probably of bacterial origin (Bonython, 1956).

A section 12 feet deep shows, from the bottom upward: a dolomite containing a layer of gypsum crystals, then a layer of multicolored clay with a band of gypsum crystals. Above this comes an important bed of white granular gypsum, 5½ feet thick, in which two layers of crystalline gypsum occur. Then follows a thin bed of hard salt, which appears to be fine-grained gypsum. Next is a black clay, and finally there is the top crust of halite (NaCl) about one foot thick. Within the crust there is a purple-colored layer. The black clay is composed of gypsum, kaolin, quartz, palygorskite and jarosite. The dolomite contains several impurities (silica, alumina, iron oxide, kaolin and illite).

Another example of salt deposition occurs in desert terrains. In the Sahara, 43 miles east of Zegdou, the Cenomanian-Turonian calcareous flagstones form a *sebkha* (=playa) filled with evaporites 16 feet thick. The deposit is composed of gypsum and contains two layers of thenardite (Na_2SO_4).

It is possible that some salt deposits are due to the concentration and crystallization of solutions in marine muds. This hypothesis has been put forward by Bourcart and Ricour (1954) to explain the salt horizons of the Triassic of Europe and North Africa.

The evaporation of sea water gives rise successively to Fe_2O_3, $CaCO_3$, $CaSO_4.2H_2O$, NaCl, $MgSO_4$, $MgCl_2$, NaBr and KCl. It seems probable that the very soluble salts (chlorides and sulfates) are concentrated where the solutions have migrated through fissures formed by desiccation in the drying muds. At the same time the sulfides are altered to sulfates by the processes outlined above.

The hypothesis is not completely satisfactory, because it does not explain why the Triassic is the system most rich in evaporites, whereas other systems containing sediments of the type described above contain no evaporites. For example, in the Cambrian of Scandinavia there is an abundance of alum shales, but there are no salt deposits. In contrast, however, the sequence of modern sediments in Lake Eyre is directly comparable to many saliferous horizons of the Triassic.

The principal types of evaporites can be divided into three groups: chlorides (salt, halite, rock-salt), sulfates (gypsum), and carbonates.

There are also salts of the same composition as the evaporites which are of direct volcanic origin, such as natron, $Na_2CO_3.10H_2O$. This is deposited, for example, in the great solfataric crater at the foot of Toussidé (in the Tibesti Mountains of the Central Sahara). This crater bears a strong resemblance in many ways to a playa (fig. 190). The same salt occurs in a playa associated with *Lake Natron* in East Africa.

In general, it seems that the chlorides owe their origin to the sea, and the sulfates to the continents, but the origin of the carbonates is complex.

Sodium chloride is generally found associated with calcium sulfate, more or less hydrated, and with magnesium carbonate.

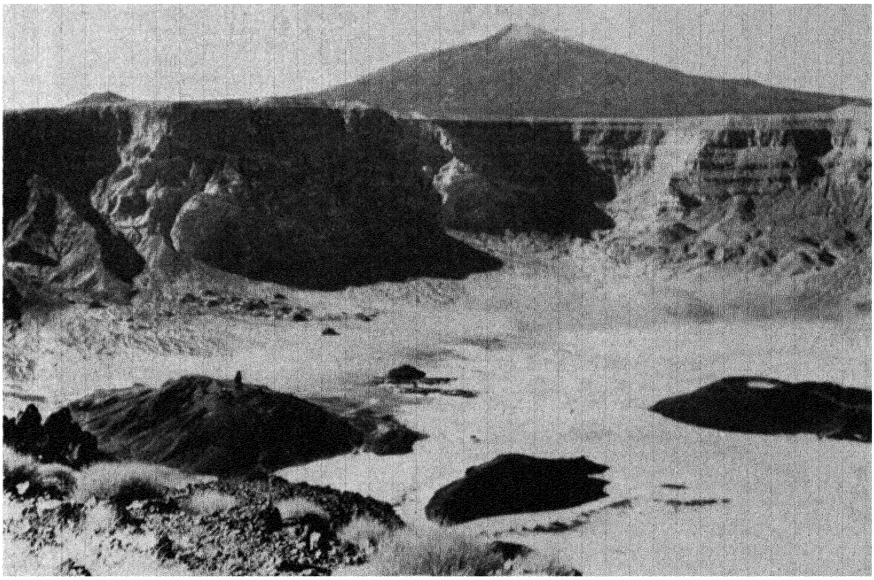


FIG. 190. AN EXAMPLE OF A PLAYA OF VOLCANIC ORIGIN: THE TROU AU NATRON ("NATRON" CRATER) IN THE TIBESTI MOUNTAINS, CENTRAL SAHARA

In the background, Mt. Toussidé, 10,712 feet. In the foreground a vast crater, the floor of which is occupied by a playa. The white crust covering it is formed of natron (sodium carbonate). The small, dark volcanic craters have broken through the salt crust. (Photograph: Freulon.)

The Origin of Sedimentary Sulfur.—Under certain conditions which occur in the course of diagenesis, gypsum may be reduced to pure sulfur, as in the sulfur found in Italy in the "gessoso-solfifera" beds. A similar origin has been proposed for the sulfur of the Gulf of Mexico. The active agents are the sulfate-reducing bacteria, with metabolic energy derived from petroleum hydrocarbons.

Saline and Alkaline Soils (figs. 191 and 192)

These soils occur frequently in arid and semiarid climates. They are characterized by an excess of sodium salts with occasional traces of potassium salts.

The saline and alkaline soils generally owe their surplus sodium to the ascent and infiltration of phreatic water, where the water table is near the topographic surface, as in the prairie soils or in the endorheic basins. They frequently represent the relics of ancient seas or of salt lakes. The soils around the Great Salt Lake, Lake Bonneville (see fig. 191) and the Caspian Sea are of this type. Moreover, an elevation, even artificial, of the water table is liable to make the soil alkaline by replacing the exchangeable calcium with sodium (Robinson, 1949). A similar change resulting from the presence of sodium also occurs near oceanic coasts, particularly under arid climates, by the action of sea spray and rain. Thus, according to L. J. H. Teakle (1937) rain carries 330 lb. of sodium chloride per acre per year over the west coast of Australia. The salt of certain soils of the Yugoslavian islands in the Adriatic Sea is believed to be carried by the wind (Gračanin, 1935). The role of the wind is proved (see p. 170) by the transport of gypsum and salt from the sebkhas and shotts of North Africa.

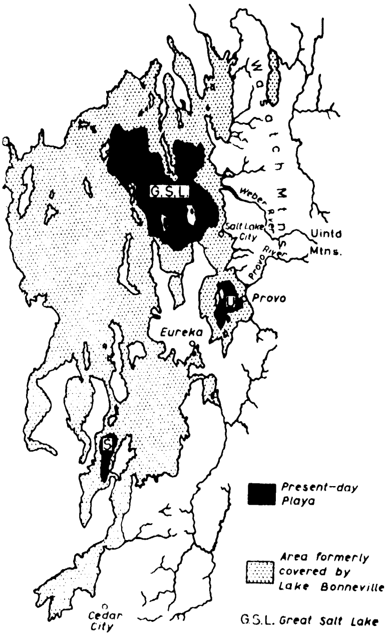


FIG. 191. THE AREA FORMERLY COVERED BY LAKE BONNEVILLE AND THE PRESENT-DAY PLAYAS (simplified after Hunt, Varnes and Thomas, 1953)

G.S.L. = Great Salt Lake, U = Utah Lake, S = Sevier Lake.

In Finland, the sulfate soils on the Gulf of Bothnia are restricted to the sediment deposited by the Littorina Sea. These correspond to the most saline of the early stages of the present Baltic, when the salinity does not appear to have exceeded 1·2% (0·8% in the Baltic). The sediments are black clays showing the following composition: 54·1–61·33% SiO_2; 13·47–14·81% Al_2O_3; 2·54–8·47% Fe_2O_3; 1·21–2·96% CaO; 2·19–3·12% MgO; 2·42–4·63% K_2O; 1·27–2·24% Na_2O; 0·16–0·96% P_2O_5; 0·23–0·93% SO_3; 0·07–0·76% S; 0–0·02% Cl; 1·37–12·53% H_2O and organic matter. TiO_2, FeO and MnO are present in traces. The dark color is due to iron sulfide undoubtedly formed by the action of anaerobic bacteria. In contact with air the sulfides are oxidized to sulfates of aluminum, magnesium and calcium, in the regions where the water table is near the surface of the ground. Drainage of the area causes the disappearance of the saline soils (Aarnio, 1924, 1930; Kivinen) (fig. 192).

In arid countries river waters which are normally rich in calcium and magnesium salts and which may contain salts of sodium, deposit calcium and magnesium carbonates upon evaporation, while the sodium remains in

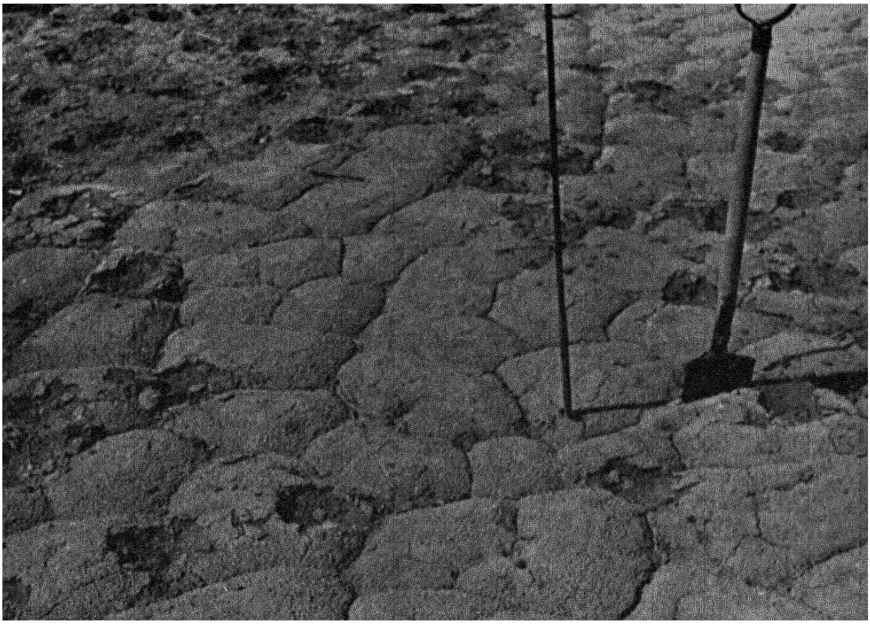


FIG. 192. SULFATE SOIL OF THE SOLVA REGION, FINLAND, SHOWING A POLYGONAL SURFACE (Photograph: Kivinen)

solution in the ionic form. These ions pass either into the phreatic waters, or are adsorbed by the argillaceous soil particles, where they cause deflocculation which leads to impermeability and sterility of the soil.

Drainage and irrigation of these soils is difficult. In ancient Mesopotamia, the phreatophytes (*Proserpina stephanis* and *Alhagi maurorum*) were used to aerate a deep dry zone and stop the capillary rise of saline waters. The land was also allowed to lie fallow for long periods (see p. 136).

The general increase in aridity can be seen at the present time. Moreover, as has already been shown while discussing the oueds and pluvial lakes (pp. 112, 18), it is also evident in the extension of the saline soils. This can be observed in Languedoc (southern France) as well as in Mesopotamia. On the edge of littoral pools, saline soils can often be found. Patches of "saltings", characterized by chlorides in the soil, have also been forming during the past fifteen years or so, near the borders of the Massif Central in regions where saliferous rocks occur beneath the subsoil. These rocks include gypsiferous and saline Triassic marls, Aquitanian (Oligocene) marls and particularly, Helvetian (Middle Miocene) marls. The appearance of this salinity in the soil is attributable to the decrease in rainfall in the area (Gèze and Servat, 1950).

B. Aarnio (1930) has distinguished several soil types according to the composition of the salts which they contain. There are *neutral saline soils* in which chlorides and alkaline sulfates predominate and from which carbonates are absent; *alkaline saline* soils characterized by alkaline carbonates and bicarbonates; and *acid saline soils* in which aluminum sulfate and ferric sulfate are dominant.

The *acid saline soils* such as those of the Gulf of Bothnia can be considered separately because they are the product of very special origin and climate. They are rendered almost sterile by the sulfuric acid which is liberated by hydrolysis. *Spergularia salina* is, however, occasionally found on them.

Russian pedologists distinguish, in place of the *alkaline* or *neutral saline soils*, the *solonchak*, the *solonetz* and the *soloti soils*. The *solonchaks* are soils in which sodium chloride and sodium sulfate predominate, and which are flocculated; these are the neutral soils of Aarnio. In arid countries saline efflorescences often make them appear white. In humid countries, they are generally black due to the presence of organic matter and are often boggy, as in Finland (Aarnio, 1924). Sometimes such soils are completely devoid of vegetation, as around the Great Salt Lake and on the shores of the Dead Sea. Yet in hot countries halophyte bushes (*Tamarix*, *Coton*, *Hibiscus*) seem to thrive on them.

The second category is that of the *solonetz*, or alkaline saline soils, characterized by the presence of sodium carbonate (not less than 20% Na). They are deflocculated, and at depth show a prismatic structure. They are frequently black. They generally develop from solonchaks by a process known as *solonization* which corresponds to the leaching and removal of excess sodium in the preceding type. In a region of saline soils, the areas undergoing solonization become depressions where structural change brings about compaction of the soil. In these depressions rain water produces alkaline solutions in which humic matter accumulates, and these, on drying out, give the soil its black color. These processes are known to occur in the Ukraine (in the basin of the Dnieper), in Hungary, in the arid west of the United States, and in North Africa (the Chélif plain, among others).

Moreover, the structure of a solonetz can be acquired by soils which do not have originally the exact solonetz composition; they may, in fact, vary with the Ca/Mg ratio (N. I. Usov, 1939).

Finally, the *soloti* are the degraded alkaline soils derived from *solonetz* *soils* by *solotization*, a process analogous to podsolization. The hydrolysis of sodium to sodium hydroxide in the presence of calcium carbonate forms a calcic soil. In the absence of salt, the hydrogenated soil becomes enriched in silica and sesquioxides. When these latter are leached out, the soil passes into a bleached eluvium, rich in SiO_2. Solotis are known in the U.S.S.R. and in the western United States. They are always associated with the two other types of saline soils.

An application of the knowledge of saline soils has been made in the plain of the Lower Chélif in Algeria (J. H. Durand). At the research station at Hamadena, the soils are dominantly solonchak, secondarily characterized by being sodic, magnesian and calcic. It has been observed that: (1) the *Na solonchaks* are associated with the *Na,Mg solonetz* soils (in which the Mg is secondary); (2) *Na,Mg solonchaks* are associated with the *Na,Mg solonetz* or with *Mg solonetz*; (3) the *Mg solonchaks* with *Mg solonetz* are less difficult to cultivate than the preceding types; (4) *Na,Ca solonchaks* with *Na,Mg solonetz* or with *Mg solonetz* can be leached; (5) *Ca solonchaks* can pass into *Mg solonetz*; (6) the *Mg solonetz* contain less than 3% of soluble salts to a depth of 18 inches. The salts contained in the ground water and the soluble salts of the soil are different. This means that part of these salts is derived from the alluvium of the oueds, which come from the hills to the south. The salts reach the surface by thermodialysis.

16

Some Examples of Complex Marine Sedimentation

Recent Marine Sedimentation in the Gulf of Lions

Off the Mediterranean coast of France, in the Gulf of Lions, the continental shelf forms a broad platform, less than 650 feet deep, descending in a gentle slope almost as far as an imaginary line from Banyuls to Marseilles. The sea floor of the deep zones of the Mediterranean is not considered here: it is mainly mud populated by gorgonian corals, echinoids and foraminifers.

Beyond the littoral zone (see pp. 257–262) the greater part of the continental shelf is occupied by neritic mud with sponges, alcyonarians, echinoids, (*Eledone* and *Sepia*), ascidians, fish, etc., which are extensively collected by trawl fishing. However, between 300 and 650 feet there are large patches of sand and gravel, with a margin of muddy sand.

As elsewhere, the greatest variations in sedimentation occur in the littoral zone. The *Cardium* sands and the muddy sands with *Donax* occur along the low coasts with beaches, and in front of lagoons and deltas, whereas bare rock, occurring most often where the coast is high, is covered with *Chtamalus*, *Littorina*, *Pachygrapsus*, *Lygia*, ascidians, mussels and sea urchins (*Paracentrotus*). As in the Gulf of Marseilles, the assemblage, which includes Melobesiae and bryozoans (Pérès and Picard, 1952) occurs on a wave-cut bench.

Sedimentation in the Black Sea

The sedimentological characteristics of the Black Sea are of varied nature and seem to have remained constant over the past 2,500 years or so.

On the continental shelf in the northwest, between sea level and about 100 feet down, there are sands which are colonized by *Zostera* down to about 20 feet and banks of oysters (*Ostrea taurica* and *O. lamellosa*) to about 100 feet. Between 50 and 200 feet in the north and west there are silts containing *Mytilus galloprovincialis*. This silt is generally yellowish-gray and contains 7·31 to 39·18% of calcium carbonate. Near Odessa, where it is very near to the surface of the sea, this silt is very rich in organic matter but poor in calcium carbonate, and is almost black. To the north of latitude

45° N. the silts are covered by the alga *Phyllophora rubens* over an area of about 4,000 square miles. All round the margin of the Black Sea the deep part of the continental shelf between 200 feet and 500 feet is covered with *Modiola phaseolina* (a species imported very recently) resting on very clayey

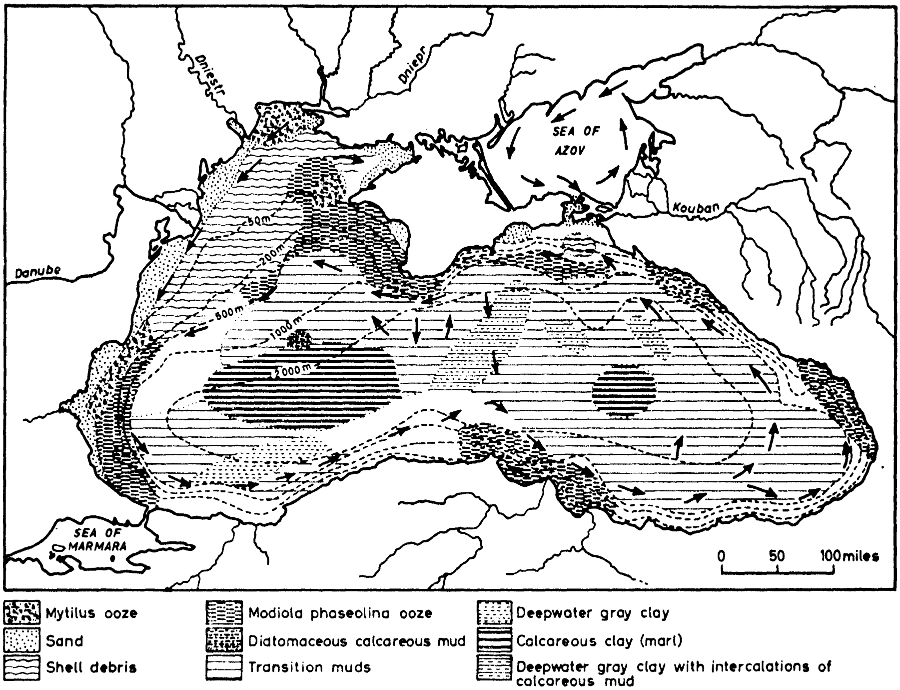


FIG. 193. DISTRIBUTION OF SEDIMENTS IN THE BLACK SEA AT THE PRESENT TIME (after Raüpach, 1952, and Erünal-Erentöz, 1956)

silts. These silts are white to gray, with 7·33 to 47·88% calcium carbonate. In the northwest and to the south of the Straits of Kertch, the *Mytilus* and *Modiola* silts are almost the only sediment.

The *sediments of the deep sea are*:

(1) A gray microstratified clay with 75% of terrigenous material, 7·21 to 33·34% of $CaCO_3$, and 3% of organic matter. In 1 mm. of this clay there are five beds of light calcareous clay and five dark sapropelic layers. This clay is localized on the coast of Anatolia where it accumulates very rapidly.

(2) A marl with 28·3% of terrigenous matter, 50·37 to 72·47% of $CaCO_3$ and 8% of organic matter. The marl consists of alternating fine-grained white layers of calcium carbonate, and dark sapropelic clay layers (70 to 100 layers in 1 cm.); it is not associated with any particular depth.

(3) There is often a passage or alternation between gray clays, calcareous muds, and fine sands.

(4) Very locally, at a depth of 5030 feet there is a 5-inch layer of calcareous ooze containing diatoms. This bed has a calcium carbonate content varying from 20·68 to 60·68% and also contains some organic matter.

(5) Sands and clayey sands less than 1 inch thick are found in the deep water in the southeast. The grain size is between 0·05 and 0·01 mm. The sands contain 10·74% of $CaCO_3$ and 77·9% of terrigenous matter. According to Strakhov, this material is brought from the mountains by the Chorokh and then redistributed by the superficial circular current.

The rate of sedimentation in the Black Sea averages 0·5 mm. per year, but reaches 4 cm. to 10 m. at certain places in the silts, 0·5 to 1·5 mm. on the continental shelf and 0·04 to 0·08 mm. in the depths.

THE SEDIMENTARY HISTORY OF BASINS

In order to describe the sedimentary history of a region, it is necessary to consider several important factors, in particular those of *facies*, the *sedimentary cycle*, *rhythms of sedimentation* and the *thickness of beds*.

FACIES

The idea of a facies is extremely rich in meaning and fruitful in its applications, and can be utilized in a variety of ways.

"The sedimentary aspect of a lithological entity", without regard to its age, is the *lithofacies* (R. C. Moore, 1949). Fossils which may have assisted in the formation of the sediment are not included. In a lithofacies, two subdivisions can be recognized: the *physiofacies* which relates to the "physical aspects of the environment" and the *biofacies* relating to "biological aspects of the environment". Moore tends to reserve the general term *facies* for contemporaneous formations.

According to Caster (1934), it is possible to distinguish a *magnafacies* which results from the deposition of a sediment over a long period of time (heterochronism) and which is displaced geographically during that time. A good example is that furnished by the Upper Jurassic reefs in the east of France (Bourgeat, 1887). Each magnafacies is made up of *parvafacies* which are limited in time and correspond to the usual conception of a facies.

Sloss, Krumbein and Dapples (1949) define *tectofacies* according to their relationship with orogenesis. This term, which R. C. Moore (1953) rejects, could strictly be applied to facies including orogenic sediments such as Flysch and Molasse.

In fact, the term *facies* should be used to define the whole of a sedimentary assemblage or a faunal assemblage, the latter being equivalent to a biotope or a modern thanatocoenosis, and should not be limited in space or time. The variations of facies of a given epoch depend on the geomorphological conditions. If the time factor is considered, the variations in facies over a prolonged period are dependent upon climatic variations, orogenesis and epeirogenesis.

It is also necessary to consider the changes in the regime of currents which may be under the influence of climate.

TIME OF DEPOSITION, SUBSIDENCE, THICKNESS

Slow and periodic subsidence gives rise to monotonous series of sediments which are generally horizontal. It is accompanied in most cases by a regular supply of sediment.

Detrital sediments are undoubtedly deposited much more rapidly than organically formed sediments since erosion depends on only two factors, rainwash and gravity, whereas the deposition of limestones, for example, necessitates the coexistence of numerous conditions (temperature, clarity, salinity) which are those required for vigorous organic growth. A given thickness of a clastic sediment does not, therefore, normally represent the same duration of time as a series of the same thickness formed of limestone or of coal. In the latter case especially, compaction results in a considerable reduction in thickness of the resultant rock during diagenesis (p. 337).

Thus caution seems necessary in the interpretation of age determinations or correlations based on the thickness of sedimentary rocks. M. Kay (1955) has distinguished between the *rate of deposition* and the *rate of subsidence* which may, or may not, coincide. On the one hand, sedimentation may take place in nonsubsiding areas or on the other, subsidence may be less than the thickness of the sediment. For example, the Pliocene and Pleistocene Ventura Basin (California) has received 16,000 feet of sediment, of which the oldest beds are marine and contain foraminifers, indicating a depth of deposition of 5,000 feet. By contrast the most recent beds are continental. The amount of subsidence has thus been about 11,000 feet (Natland and Kuenen, 1951). Generally, however, the thickness of a series indicates the total subsidence of the basin in which they have been deposited.

Whether or not subsidence occurred during deposition, it is necessary to consider the continuity of the supply of sediment and its location. Kay, on the basis of the very limited available evidence, has estimated the average rate of subsidence in geosynclines and on continents. In the former, subsidence hardly exceeded 500 feet in a million years, and was rarely more than 1,000 feet in an era. At the maximum, it reaches 2,000 feet per million years in the deeps of orthogeosynclines, in trenches, and in foredeeps. In continental basins, subsidence amounts to 500 feet in a million years, but generally it is not so prolonged. Other causes of variation are eustatic movements (see pp. 22–25).

In some cases (see pp. 159–163), as for example in varves, it is possible to calculate the annual or even seasonal rate of deposition of the deposit.

Shepard (1948) has given some interesting information on oceanic deposits, far from the coasts. The mean rate of deposition of a *Globigerina* ooze is 632 years per centimeter, but one centimeter of the compacted rock represents 1,320 years. The red clays of the deep oceans are formed at the

FIG. 194

FIG. 195

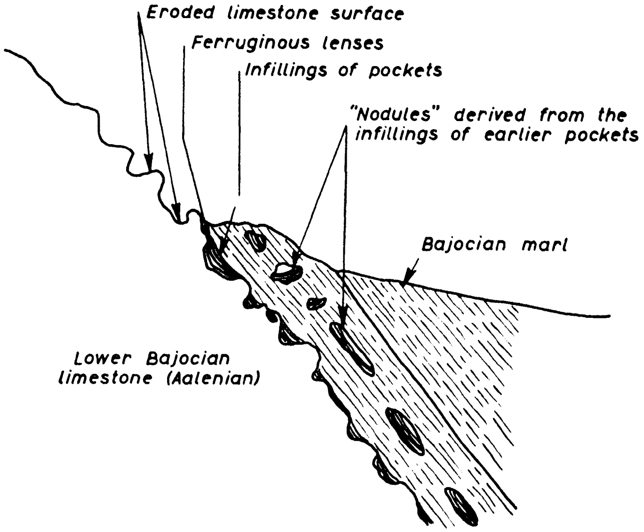


FIG. 196

FIGS. 194 to 196. THE EVIDENCE OF AN INTERRUPTION IN SEDIMENTATION IN THE JURASSIC, BAJOCIAN STAGE, IN THE NORTHERN PART OF THE CENTRAL ATLAS MOUNTAINS, MOROCCO

The Aalenian limestone (below) has an irregular surface whose hollows have been filled with "paper" shale. This surface also has small concretions of iron oxide in the cavities. The paper shales were reworked into nodules during the deposition of the marls (Boulemane marls with Cadomites), which belong to the younger Bajocian. There was thus an interruption in sedimentation, either due to emergence, or to submarine currents, which eroded the limestone (remaining during sedimentation as "hard-ground") for a brief period within the Bajocian. This type of erosion into hollows is reminiscent of the supralittoral stage (p. 63, fig. 23). (Photographs taken at Tazrou Tamrabet on the southeastern limb of the Tichoukt anticline, Morocco, H. Termier.) (In Fig. 194, the limestone appears dark and the marl is light-colored, whereas in Fig. 195, the limestone is light and the paper shales and marl appear dark.—Translator.)

rate of 962 years for one centimeter of sediment and 2,000 years for one centimeter of rock (or 5 m. in one million years).

Zones impoverished in sediment, even where the deposits are fairly thick, are always produced by phases of arrest of the subsidence. On tidal coasts where a fairly considerable width of the strand may remain dry for more than half the year, the sediments will be subjected to atmospheric erosion. Even thick series of sediments, where the surfaces of beds are covered by ripple-marks, show evidence of interruptions in sedimentation, which may be annual or semiannual (see p. 209).

Some sea floors such as that of the English Channel, are affected by marine currents which hinder the settling of all deposits (Dangeard, 1927).

Finally it may be noted that part of a bed may be carried away without leaving any trace.

LACUNAS (NONSEQUENCES)

(figs. 194–197)

The result of erosion is the formation of more or less coarse material which is then transported a long or short distance and finally deposited in various associations with other materials. For the geologist (stratigrapher, petrographer, paleogeographer, or paleoecologist), the study of sediments provides an important "*clew*"[1] in reconstructing the conditions of erosion and deposition. Great care is needed, however, for in this domain, the phenomena of convergence are frequent.

The places of deposition are generally different from those of erosion. The maximum intensity of erosion occurs on unprotected summits and on slopes, whereas the sediments accumulate in depressions. The movements of uplift and subsidence are thus, respectively, the causes of erosion and deposition. Considered on a large scale, it can be said that erosion affects the whole surface of the continents, while the sediments accumulate in the sea where erosion and transportation are relatively weak. Naturally, this is an oversimplification, since even mountains may be covered with a veneer of detrital sediments and there are marine sediments which have not been derived from the continents. It follows that the evolution of the surface of the lithosphere tends invariably towards a leveling of the surface, that is towards a pediplain or a peneplain.

Moreover, it is rare that a particular sediment has been formed by only a single geological phenomenon. In most cases, sediments result from a multitude of agents. Undoubtedly, there exist a number of simple deposits which can be identified as glacial, fluvial, eolian or marine. But, most often, they are of mixed origin such as fluvio-glacial, fluvio-eolian, fluvio-volcanic, fluvio-marine, etc. This does not include soils which are elaborated by the actions of very many diverse factors.

Sediments represent only a part of the time period to which they are attributed. Even where a sediment is apparently continuous, it is preceded either by a period of nondeposition, or by a period of denudation, and, if this occurs intermittently, the phases of deposition are separated by periods without deposition which may be accompanied by denudation. Thus, solely from a sedimentological standpoint, it is practically impossible to establish rigorous correlations between basins. This implies a further consideration: because most phases of movement of the earth's crust have been disclosed by interruptions or changes in sedimentation, it is hardly possible to establish exact correlations of movements which have occurred in basins remote from each other.

Formations having identical sedimentary compositions are deposited at different rates, dependent upon the ecological or physical (e.g. hydro-

[1] The authors use *fil conducteur*, i.e. a *clew*, in allusion to the ball of thread used to guide Theseus through the labyrinth in the mythological story.—Translator.

dynamic) conditions. The speed of growth of organisms, like the destructive power of waves, varies from place to place.

However, there are certain lithologic or faunistic peculiarities that appear from time to time, which, if used with great care, constitute marker horizons ("microfacies"). The products of enormous, but rather infrequent, ash-showers in the Mesozoic of the Midwest is a special example.

It is possible to establish an ideal profile for sedimentary deposits extending from the highest ridges of the continents to the deepest zones of

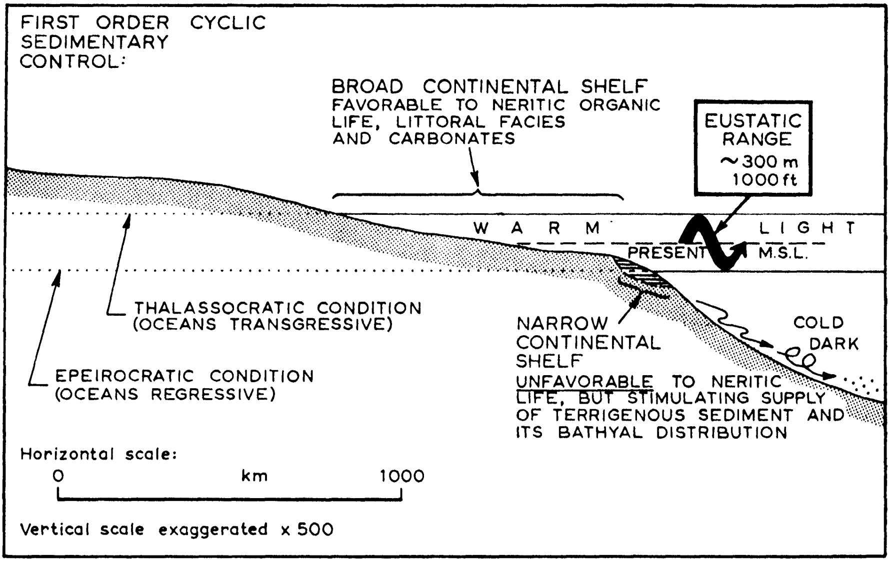


FIG. 197. DIAGRAM TO ILLUSTRATE HOW THE MAJOR SEDIMENTARY CYCLE IS EFFECTED BY LARGE OSCILLATIONS OF RELATIVE SEA LEVEL (prepared by R. W. Fairbridge)

If those changes are world-wide (eustatic), or even continent-wide (geodetic), then a first order control of cyclic sedimentation is introduced. During transgressive phases [so-called "thalassocratic" condition, i.e. oceans dominant], there will be broad continental shelves, where warmth and light will favor widespread organic activity, carbonate sedimentation and widespread littoral facies. In contrast, in the regressive phases [so-called "epeirocratic" condition, i.e. continents dominant], the continental shelves will be greatly reduced in width, neritic organic life squeezed into a very narrow habitat, or forced to adapt to deeper conditions on the dark bathyal slopes, where slumps and turbidity flows further discourage colonization; lowering of base level accelerates stream flow and stimulates clastic terrigenous sedimentation, and tends to reduce or obscure carbonates.

the oceans. The thickness of the deposits clearly reaches a maximum at the level of basins near to the mouths of rivers: on the borders of continents in the general exorheic case, and in closed depressions in the case of endorheic basins.

The shifting mouths of many meandering rivers bring about changes in topography by their aggradation. Likewise, coastal marine currents modify their courses according to geological and climatological changes.

Breaks in Sedimentation

The existence of uncomformities, which may be disconformities or discordances, implies that interruptions or breaks in sedimentation occur during phase of erosion.[1]

Sedimentation is not a continuous phenomenon. Even in the absence of emergence or orogenesis it is often possible to observe "omissions of sedimentation" in a stratigraphic series.

The general term *lacuna* also known as a *diastem*, *hiatus*[2] or *gap* is used to indicate the absence of beds, which may be due to nondeposition or to subaerial or marine erosion.

"Hard-grounds"

"Hard-grounds" are parts of the sea floor composed of indurated rock which have been perforated by lithophagic organisms. These sea beds, generally on the continental shelf, do not accumulate sediments because they are swept clear by currents, as, for example, parts of the English Channel at the present time (L. Dangeard, 1927).

[1] The Anglo-Saxon use of the word *unconformity* implies the occurrence of horizontal beds on an eroded surface which has been tilted and planed off, or in the case of the igneous rocks, stripped of their original cover; an *angular discordance* implies an unconformity on beds which have been folded or tilted, then eroded; in contrast, a *disconformity* is an unconformity on horizontal beds, which have been subjected only to denudation.

[2] Editor's note: The term hiatus is used in reference to the "*time* that is not represented at an unconformity" by J. H. Weller (1960, p. 383).

17

Diagenesis—The Transformation of Sediments after Their Deposition

The term *diagenesis* was introduced by Gümbel (1888, p. 383) to describe the action of warm water on clastic sediments. According to him, this action led to the formation of crystalline schists. This original meaning has been discarded by J. Walther (1893–1894, pp. 692–711) and replaced by that of *transformation of sediments* after their deposition, independently of orogenic pressure and volcanic heat. This transformation results in *lithification*, or the passage of the sediment to the coherent state. Rigorous limits cannot be applied to the term since there is an insensible passage between the phenomena of sedimentation and diagenesis, and to some extent between diagenesis and metamorphism. It would seem possible to distinguish several superimposed levels of diagenesis, in the same way that several zones of regional metamorphism can be recognized.

K. Hummel (1922) has given the name *halmyrolysis* to the whole of the rearrangements and replacements that take place in the sediment while it is still on the sea bed. P. Kessler has used the term *metharmosis* to describe the chemical changes which are due to atmospheric weathering and which occur after the sediment has been removed from the direct action of the sea water. In the authors' opinion, ground water, which contains mineral salts and which fluctuates in level, is important in the final stages of the evolution of sediments of all origins in continental environments (see pp. 79–81). Finally, it may be noted that Wetzel has used the term *thololysis* for the sublacustrine evolution of a sediment.

DIAGENETIC PHENOMENA

At the instant of deposition, a sediment may be more or less homogeneous. It is composed either of fragments of other rocks (detrital fragments and particles) or of small crystals which are formed *in situ* (authigenic minerals) or of colloidal gels or of organic matter or even of a mixture of all of them. This sediment forms a bed, often called the *interface*, on a more or less solidified base, and is under the mass of the water which constitutes the environment of sedimentation. Thus, this bed is saturated with water.

Due to the continuity of deposition, the interface itself is covered by successive layers until it is buried beneath sediment which may be several thousand feet thick. This bed is composed of "solids", consisting of the sedimentary material and "holes" (or pores) occupied by water.

As soon as a bed is removed from the direct influence of water by the accumulation of succeeding layers, it is subject to diagenetic alteration. This is due to biological and chemical changes in the environment, to the rapid increase in pressure and to the very slow rise in temperature.

The principal phenomena are grouped together under the general term "diagenesis" and will be reviewed below; they are as follows: biochemical alteration, compaction, solution, cementation, recrystallization, metasomatism and authigenesis. It is often difficult to determine the *precise instant* at which the phenomena take place, but it is certain that they can occur very rapidly. For example, in the Clinton Group (Silurian) the replacement of certain beds by iron minerals occurred before the deposition of the limestone beds which overlie them. Also, in the Franco-Belgian coal basin, pebbles of coal are found in the roof-rock of coal seams (P. Pruvost). However, there is nothing to prove that the pebbles were not derived from much older coal seams.

In contrast, ancient sediments are known which have undergone practically no diagenesis, such as the blue clays of Tallin and Leningrad, which are of Early Cambrian age.

BIOCHEMICAL ALTERATION

The initial processes of diagenesis take place in the order: biological, chemical and physical.

MacGinitie (1935) has shown that organisms abound in the mud of the swamps of the Gulf of Monterrey, in California. ZoBell (1938) has shown the importance of a bacterial flora and has given (1946) the following details: on a muddy marine floor, the *number of bacteria* per gram of mud is 63×10^6 in the top 5 cm., while at a depth of 1·5 m. it is less than 1,000 per gram; the *oxidation-reduction potential, Eh,* changes from $-0{\cdot}07$ volt near the interface to $-0{\cdot}28$ at a depth of 2·4 m., but the graph of the *reduction capacity* shows that reduction occurs chiefly near the interface; and finally, the *hydrogen ion concentration, pH,* passes from 7·8 to 7·9 in the same interval.

Biochemical Horizons

The work of Tromifov (1943), Emery and Rittenberg (1952) and Debyser (1952, 1954) has shown that a succession of biochemical horizons exists in modern sediments.

In the case of a homogeneous and continuous mud, there is a pH gradient (about 8·2 at the surface, below 7 within the sediment and above 7 at still greater depths) and also an Eh gradient (strongly positive at the surface, strongly negative in the first foot or so and again positive below this). In a heterogeneous sediment, such as that of the continental shelf, variations in pH are periodic between, for example, argillaceous beds and sandy beds.

Debyser stresses the importance of this zonation as it affects the solution or precipitation of minerals such as silica, carbonates, phosphates and pyrites.

When no sediment is deposited for a long time, the mud behaves as a soil, as previously noted (p. 222 and Termier, 1952, pp. 113–120). Silica then concentrates in the acid layers, and carbonates concentrate in the alkaline ones. There is thus a truly contemporaneous diagenesis of the sediment. The characteristics of biochemical horizons "result principally from the activity of micro-organisms". It appears, therefore, that living organisms and organic matter are paramount in the early stages of diagenesis.

COMPACTION

Compaction is a physical change, and consists of a *reduction in the volume* of a sediment through *diminution of the porosity* and *expulsion of the water* which is present in the pores. It varies considerably with the sediment: 3% for sand; 90% for a layer of vegetable material which is being transformed into coal. This reduction in volume is due to compression of the bed by the weight of the overlying sediment ("gravitational compaction") which acts at right angles to the stratification and can cause cohesion in such materials as clays. In orogenic zones, compaction is augmented by folding. This results in pressures being exerted which may be in any direction and which may cause crushing of the rocks.

Athy (1930) has shown that the porosity of sands decreases from approximately 42% at the surface to 32% at a depth of 6,000 feet, while that of shale decreases from about 47 to 4·5% in the same interval.

SOLUTION

Proof that original crystals have been removed by solution is furnished by the existence of geometrically shaped cavities (negative crystals). Salt crystals and even hopper crystals of salt in limestone and clays (H. and G. Termier, 1948, p. 402) are often so dissolved, and sometimes replaced by calcite, quartz, etc., as pseudomorphs.

CEMENTATION

Cementation is due to the precipitation of mineral matter in the pores and interstices of a sediment. It may occur either during sedimentation or after it, and in particular affects the most soluble rocks. The mineral precipitates may or may not be of the same composition as the rock. Cementation results in a reduction of the porosity of the rock and thus increases its coherence.

The principal cements of sedimentary rocks are calcite, dolomite, siderite, iron oxides and silica. Calcite often forms the cement of recent sandstones (in eolianites, the lower part of sand dunes is cemented by $CaCO_3$ derived from the upper part), while dolomite occurs in the older sandstones.

Silica, however, is more frequent than calcite. Amongst sandstones with a silica cement, it is found that opal and chalcedony occur in the younger ones, whereas in the older ones, the silica is nearly always present as quartz.

In arid countries, water near the surface of the soil can rise by capillarity and form carbonate crusts (caliche, etc., pp. 156–157). This phenomenon can contribute to the cementation of outcropping rocks.

Finally, cementation is so widespread in the phreatic zone that hydrologists and ore-geologists distinguish a zone of deposition or cementation (see p. 81).

RECRYSTALLIZATION

Marine and lacustrine sediments begin to solidify when they still contain a large amount of water. Sediments of nonaquatic origin, such as those of deserts, can receive rain water. Hydrological studies have shown that meteoric water can circulate to considerable depths, and there is no doubt of the existence, in certain places, of sources of juvenile water. Thus it is usual to find that beds of sedimentary rock always contain more or less large quantities of water.

The constituents of a sedimentary rock are thus liable to be partially or completely dissolved according to their solubility, size, and environmental conditions (pressure, temperature, etc.). Differential solution ("intrastratal solution" of Pettijohn, 1941) and the internal redistribution of mineral substances is therefore considered to be of great importance.

Certain rocks are dissolved and recrystallized so readily under the effect of differential pressures, that they can flow as plastic substances toward zones where the pressure is lowest. This is particularly true of halite (rock salt) which forms domes and diapirs rising up through sedimentary layers of greater density. To a lesser degree, gypsum and anhydrite behave similarly.

The recrystallization of limestone and dolomite leads to the formation of marble. This may result from fissuring and crushing by orogenic movements or from the commencement of metamorphism. It seems, however, that in most limestones and dolomites, even those of fine grain, recrystallization has played a part in their formation from carbonate muds or coarser debris. Carbonate crystals increase in size by the deposition on them of material from interstitial solution in the sediment itself. Growth of the grains thus results from a redistribution of the material. Similarly, siliceous sandstones can be converted into quartzites (see below).

METASOMATISM AND AUTHIGENESIS

The term *metasomatism*[1] implies the chemical modification of rocks by the "almost simultaneous" transport and substitution of mineral matter

[1] The authors specifically use the term *métasomatisme* in relation to metamorphic and igneous rocks. In relation to sedimentary rocks, they prefer the term *métasomatose*. In English no such distinction is made and both terms are translated as *metasomatism*.—Translator.

"whose composition may differ partly or wholly from that of the host rock provided that the original volume is conserved."[1] The most frequent examples are dolomitization and silicification, which very often preserve the original structures. But there are many others which can convert sedimentary rocks into ore-minerals. They are described in textbooks of ore-geology.

Replacement of Limestone by Iron Salts

The effects of substitution by iron salts can be seen in places in the Clinton Group (Middle Silurian) of Pennsylvania and New York. The calcium carbonate of fossils contained in the Clinton shales has been replaced, while ferruginous oolites in the limestone also appear to be due to replacement. On the other hand, primary ferruginous oolites are known, bacterially precipitated, and this may be the source of the iron replacing original carbonate fossils, etc.

Authigenesis

Kalkowsky (1880, p. 41) has given the name *authigenic* to those minerals which are formed *in situ* in rocks, and especially in sedimentary rocks, with which they become integrated. This term is in contrast to *allothigenic* (Kalkowsky, 1880) which describes the original minerals formed by the crystallization of igneous or metamorphic rocks, and *allogenic* which refers to transported sedimentary detritus.

Tester and Atwater (1934) distinguish *authigenic* minerals formed *in place* around a nucleus in sedimentary rocks, from *secondary* minerals which are growths of the *same nature or of an isomorphous species* around pre-existing minerals.

The principal substances forming authigenic minerals are the carbonates, silica, silicates (feldspars, mica, chlorite) and sulfides.

Carbonates.—The carbonates formed by living organisms are calcite, aragonite and vaterite, the latter being completely unstable. The calcite is often formed in fossiliferous sediments from aragonite which constitutes the shells of many molluscs (gastropods), the spicules of alcyonarians, etc. Inversion of aragonite may be very rapid (twelve months), or may take many thousand years. When a quartz sandstone becomes cemented by calcite, under certain conditions (of the ground water, probably) the cement assumes a crystal continuity, independent of the grains. The phenomenon is known as "Fontainebleau sandstone crystallization", although it is actually rare, and most of the original Fontainebleau sandstone of the Paris Basin is uncemented.

Dolomite is present in limestones as rhombic "metacrysts" which can

[1] The authors' definition has been expanded by the addition of the words in quotation marks.—Translator.

cut across original structures (for example, oolites) and fossils. The conditions of dolomitization will be discussed later.

Siderite is very rare, either as a replacement or authigenic mineral.

Silica.—A good example of an authigenic mineral is provided by the bipyramidal quartz crystals found in the redbeds of the Permo-Triassic of Morocco and Algeria, and also in the Triassic marls and gypsum deposits of the Pyrenees (Lacroix; 1893, pp. 109–113). In sandstones the grains of quartz often have a layer of secondary silica which has the same optical orientation as the nuclear grain, and which tends to develop the symmetry and crystal form of quartz. The surface which separates the secondary quartz from the original grain can be seen in thin section under the microscope. When the silicification is incomplete and some spaces between the grains are filled with another cement, the rock is a quartzitic sandstone; when complete or nearly complete, the sandstone is converted into a quartzite.

In fossiliferous rocks, plant and animal remains are often well preserved by silicification. Even the finest structures are retained (Psilophytales in Scotland, tree trunks in Autun, the Sahara and Arizona, and the branchial apparatus of Permian brachiopods in Cambodia and the Glass Mountains, Texas). Unfortunately, in the last-mentioned case, beekite, in the scrobicular and chalcedonic varieties, often covers the surface of shells with a thick crust which destroys all ornamentation.

Colloidal silica is released in the migratory phase of lateritic erosion. Thus the formation of concretions or siliceous crusts in basins receiving the migratory phase can reasonably be attributed to lateritic erosion. Concretions of this type are found, for example, in the opaline kaolinitic clays of the Chad Basin. This transportation has locally favored organisms which utilize silica, such as diatoms, which have then multiplied so rapidly that diatomites have been formed. It is highly probable that the pre-Pliocene siliceous crust of the basin of Lake Eyre (Australia) represents the deposition of a migratory phase released from the Miocene laterites of the peneplaned area (p. 152).

Evidence of silicification is also to be found in the deposits of the continental shelf. The Moroccan phosphate basin, studied by Salvan (1955), can be taken as a good example of all the silicified formations associated with phosphate series (e.g. Phosphoria beds of the Permian in North America, Senonian phosphates in northern France and England and the Maestrichtian phosphates in the Negev). The Moroccan basin existed from the Maestrichtian until the Early Lutetian; siliceous horizons are frequent especially outside the phosphate levels.

The siliceous formations of the Moroccan phosphates are chiefly dark phosphatic cherts, in banks which enclose light-colored pseudo-ooliths. There are also thin beds of dark chert, without pseudo-ooliths, alternating with marls occurring at the top of the beds. Less often, massive light-colored chert beds several feet thick are developed, which contain grains of calcite. Associated with these are subspherical or flattened "cannon-balls" of chert in sandy beds, "menilites" of various forms, and geodes.

On the whole, it seems that the biosphere plays only a small part in the development of siliceous formations associated with phosphates or with other sediments. Most of these formations are derived from the mechanical alteration of crystalline rocks of the continents, or of ash and submarine lavas.

Radiolarites associated with the rocks of the ophiolitic suite (of geosynclines), demonstrate the subordinate role of living organisms in comparison to that played by transported material.

Feldspars.—Many types are represented by the authigenic crystals in limestones, for example, orthoclase (Lacroix, 1893, pp. 108–109), microcline (Lacroix, 1893, pp. 818–819), and the albite of Roc Tourné, near Bourget, Savoy (Lacroix, 1893, pp. 162–168). The chief occurrences of authigenic feldspar have been listed by Boswell (1933).

Sulfides.—Cubes of pyrite and nodules of marcasite often occur in black shales and in slate, and, as we have seen, sulfides are normally formed in anaerobic marine muds by bacterial action.

Mica and Chlorite.—The formation of mica and chlorite in sediments and rocks has already been reviewed (pp. 224–225).

Diagenesis and Metasomatism of Calcareous Sediments

Carbonates are relatively unstable, and for this reason are readily transformed by diagenesis. Vaterite ($CaCO_3$) hardly ever occurs in rocks. Aragonite, which is 3 to 9% more soluble than calcite (Chilingar, 1956) readily recrystallizes as the more stable calcite. On the Funafuti atoll and on the Great Barrier Reef of Australia, borings have shown that below about 100 feet, all the aragonite is transformed into calcite (Fairbridge, 1950).

Calcite itself is often metasomatized with the formation of dolomite. Isomorphism of the two minerals explains why they are found in solid solution with each other. Calcite may also contain traces of strontium, barium and lead in solid solution. Chave (1954) has shown that the amount of magnesium in calcite is greater in the lower organisms (foraminifers and calcareous algae) than in the higher ones (molluscs and arthropods). It is also increased if the temperature of the sea water is high. There are neritic sediments which are *initially* rich in magnesium, and others, which, in similar conditions, are mainly composed of aragonite transformed into calcite. It seems that an alternation between these two types of carbonate sediments can be produced by climatic cycles or changes of sea level. Calcite rich in magnesium is hardly more stable than aragonite (Jamieson, 1953), and for this reason magnesium-rich calcites are probably the most important "hosts" for penecontemporaneous dolomitization (Fairbridge, 1957).

The chief magnesian limestones are the algal limestones. These are rapidly attacked by metasomatism and develop into dolomite. This causes the precipitation of new dolomite directly from the surrounding water, which is saturated with calcium and magnesium.

High pressure also favors the formation of dolomite at depth. It is probably this property which has led to the dolomitization of the base of the Funafuti atoll below a depth of about 650 feet. This problem is referred to later (pp. 343–345).

Concretions.—These are aggregates which result from the accumulation of mineral matter round a center of attraction and grow from the interior outward. They are often nodular, or may be in part mamillated or botryoidal (with small cavities), though they may take on any form. Their dimen-

FIG. 198. SEPTARIAN NODULES IN THE SILURIAN OF THE MRIRT REGION, CENTRAL MOROCCO (Photograph: G. Termier)

sions vary from less than 0·04 inches to several feet. In structure, they are generally concentric, occasionally radial (marcasite) and, more rarely, "cone-in-cone". Their distribution seems haphazard because it results from many causes. Concretions are, however, generally aligned parallel to the bedding.

In clays and shales, calcareous nodules are often found. Some have a sandy nucleus; others display an internal fracture system (septaria, fig. 198). In limestones, flint or chert nodules and marcasite nodules are common.

Concretions are classified as: (1) *syngenetic*, or developed contemporaneously with deposition of the rocks in which they are found; (2) *epigenetic*, or formed after the deposition of the enclosing rock.

It is often very difficult to know how to classify a concretion when in the field. It is necessary, therefore, to examine closely the form, size, nature, structure, partitioning, and particularly the relationship with the enclosing rock. Epigenetic concretions are frequent in porous rocks, such as sands and sandstones, and are due to the infiltration of water carrying soluble salts and colloidal material.

The formation of these concretions results from several interacting physico-chemical processes: water circulation, transport of material, solution, substitution, cementation, force of crystallization, etc. It may also be noted that concretions very often have a core or nucleus consisting of one or more fossils: the organic matter has served as a center of attraction for certain chemical elements (Termier, 1956, pp. 182–184).

The Problem of Dolomitization.—There is no known example of the precipitation of dolomite among modern sediments. Only isolated crystals of dolomite in deepwater muds, together with glauconite, have been found (Leinz, 1937). Precipitated dolomites are also very rare in the stratigraphic column. In France, the best known are those of the Keuper. These are composed of very fine-grained dolomite crystals, which alternate with marls or illitic clays containing lenticles of rock salt, anhydrite or gypsum.

Most dolomitic limestones (in which the proportion of dolomite is 10 to 50%) and calcareous dolomite (50 to 90% dolomite) result from a metasomatic enrichment in magnesium.

The carbonates of calcium and magnesium form an isomorphous series. Dolomite contains an equal proportion of $CaCO_3$ and $MgCO_3$. In the sea, organisms which have a magnesian test contain magnesium ions in solid solution in the calcite (Spotts, 1952; Chave, 1952). The presence of very high concentrations in rocks which have undergone diagenesis indicates that a slow reaction has enriched them in $MgCO_3$. Chilingar (1956) has calculated that in modern sediments, the Ca/Mg ratio is 40, and that the annual precipitation of magnesium is 13 million tons. On the other hand, Clarke (1924) calculated that the oceans receive 93 million tons of Mg^{++} annually. The oceans contain at the present time 17×10^{14} tons of magnesium, and, at the present rate, this quantity could be accumulated in 18 million years. There is thus an excess of 80 million tons of magnesium per year which could allow localized establishment of supersaturation, and hence lead to natural precipitation.

In fact, dolomitization is largely the result of diagenesis. It seems that limestones which already contain magnesium (equivalent to 5 to 10% of dolomite) are more readily transformed. This appears to confirm a tendency, already indicated by the presence of magnesium in the interstices of the calcite lattice, toward a final more stable equilibrium represented by dolomite. Limestones of algal origin are important among magnesian limestones (H. Termier and G. Termier, 1951). The former include the stromatolithic limestones of the Precambrian, the Alpine Trias rich in Dasycladaceae and in Cyanophyceae symbiotes (*Sphaerocodium*) and also the Lias of North

Africa containing Dasycladaceae, Codiaceae and *Sphaerocodium*. These belong to *coastal facies, often sublagoonal in a warm climate* and containing a fauna of molluscs with thick shells, foraminifers and echinoderms which are rich in magnesium. These magnesian limestones are susceptible to magnesium metasomatism which generally induces recrystallization.

The Lithothamnium, which are the richest in magnesium of the calcareous algae, seem to have become the most abundant type from the Cretaceous onward.

It has been noted that a very large part of modern coral reefs has been formed by calcareous algae: for example, *Halimeda* in Bermuda, and Lithothamnium in many other places. The magnesian limestones thus formed are very good material for dolomitization. Modern dolomitized reefs show a transition from magnesian limestones containing 5% $MgCO_3$ to dolomites containing more than 40% $MgCO_3$, which confirms this view.

The substitution of magnesium for calcium in the lattice of carbonates can also be explained by the instability of aragonite, which very readily takes up strontium carbonate in solid solution.

The conditions which favor the rapid dolomitization (in several thousand years) of marine limestones, reef or nonreef, are: the concentration of Mg^{++} ions in the sea water, a high salinity (about 4%), a moderately high temperature, a carbon dioxide pressure higher or lower than normal, a high pH, reducing conditions, and finally, the presence of organic matter, hydrocarbons and ammonium compounds (Fairbridge, 1957).

These conditions, which occur in coastal lagoons containing algae, have already been noted. They also occur at the bottom of coral reefs, which explains the dolomitization of the older part of the Funafuti atoll. In the latter case, the amount of $MgCO_3$ is generally about 5%, down to 633 feet. However, within this zone, there are two levels in which maxima of 16% are reached. These correspond to low sea levels at the beginning of the present epoch. Below 633 feet, almost the whole of the reef has been dolomitized. This depth seems to be within about 23 feet or so of the depth most favorable for dolomitization. Very occasionally, at levels which correspond to phases of accelerated subsidence (Judd, 1904; Reuling, 1934; Fairbridge, 1957), the limestone is unaltered. Dolomitization does not, however, affect all the great reefs; exceptions exist, for example, between 4,230 and 4,560 feet on Eniwetok, on the upper part of Kita-Daito-Jima in Japan, and on numerous other atolls. At Atiu in the Cook archipelago, dolomitization decreases from the center to the exterior of the old atoll. At Bikini there is hardly any dolomitization.

The enrichment of magnesium in deepwater clays, recorded by the Challenger expedition (Högbom, 1894), and in the shallow-water calcareous algae of the Bay of Naples (Magdefrau, 1933, 1942) indicates that this process, which can be interpreted as the beginning of diagenesis, is widespread in the sea.

It is also apparent that there have been periods during which the seas particularly favored dolomitization. At such times they were salty, but not oversaturated. Their waters were neritic and abounded with calcareous algae. The latter absorbed large amounts of magnesium, and also, as a result of photosynthesis, liberated considerable quantities of carbon dioxide.

In contrast to these, there are the less alkaline deep waters, far from the coasts, where limestones containing chert are formed.

It follows that dolomitic series show a less regular rate of sedimentation than limestone series deposited far from the shore. Moreover, they are often broken up by lacunae (nonsequences).

Diagenesis of Saline Deposits

The saline deposits undergo some of the most important diagenetic modifications in sedimentary petrogenesis. They are the most soluble rocks and the most easily recrystallized. Moreover, their water content is usually high, though they can be partially dehydrated. They can also yield to the pressure of other beds (diapirs, domes, etc., see p. 201) and are particularly plastic. Water plays a very active role: either the salt is contained in a clay where the salt water causes deflocculation, or the water is trapped in the joints where the salt recrystallizes in large crystals or in fibers (*static* zone of Fournier, 1925), or it circulates (*dynamic* zone of Fournier, 1925) and carries away most of the rock salt. These transformations, which are really metasomatic and analogous to those studied by ore-geologists, result from the circulation of phreatic water. Thus, Bonte (1955) believes that the replacement of red nodules of polyhalite in the Lower Keuper by gypsum, takes place in the dynamic zone.

At depth, calcium sulfate is present in the form of anhydrite. If the sedimentary cover is thin, this is converted into gypsum. In the Keuper of the Jura, the limit of transformation is at about 230 feet depth (Bonte, 1955), although fibrous gypsum may occur as deep as 650 feet.

Meulerization[1]

The formation of "meulieres" in the Pliocene and Pleistocene of the Sahara has been studied by Alimen and Deicha (1958). The principal example quoted comes from the upper part of the Pliocene of the Hamada of Guir.

Meuliere is a rock very similar to a sandstone. It contains intact eolian grains and also detrital quartz decayed by solution. The dissolved portion of the latter is replaced either by calcite or by secondary silica (quartzitic, chalcedonic or opaline) or by an intimate mixture of the two. In the final stage of this replacement, all the quartz grains are destroyed. The cement of

[1] A *meulière* is, literally, a stone suitable for millstones, but is used in a geological context to describe a peculiar type of cherty rock. The Anglicized forms "meuliere" and "meulerization" for the process are used here.—Translator.

the true meuliere is always calcite with traces of iron oxide. The rock is usually associated with sandstones and can only be distinguished by the unique cement of secondary silica. There is always a transition between the two rock types. There are also exceptional cases of meulieres in which silica has replaced gypsum and the crystal form of the latter has been retained.

It appears that the process of meulerization is one of "consolidation on the surface, immediately after the deposition of the sediment". It is of "pedological type following the evaporation of layers of moisture in a mixed sediment of limestone and detrital quartz". The fundamental agent of these modifications, which are characterized by a rapid oxidation of the iron and an attack on the quartz (possibly due to the presence of gypsum), may be the intense insolation which still occurs in the Sahara and which has been manifest in all its arid phases. The surface temperature of rocks may reach 80° C. (176° F.). The process resembles "hard pan" formation.

The meulieres of the Paris Basin (Sannoisian meulieres of Brie and the Stampian meulieres of Beauce) occur as lenses in lacustrine limestone. These meulieres are former limestones in which the carbonate has locally been replaced by chalcedony and quartz. Atmospheric weathering attacks the rest of the limestone, leaving the rock pitted and reddened by oxidation of the iron. In fact, it seems that an origin comparable to that of the Saharan meulieres can be envisaged. In this case the silica has been carried to the lakes by rivers coming from the Massif Central, partly as detrital sand and partly as colloidal gels, or in true solution.

Recent Diagenesis in New Caledonia

Particularly rapid diagenesis has occurred in the recent sediments of New Caledonia. These sediments are:

1. Rocks of subaerial origin associated with rock outcrops, especially peridotites and serpentines. The type of climate favors the development of tropical forest. Although the great period of laterite formation has long ended, the formation of an iron crust can be seen, for example, on the surface of mine cuttings made less than 80 years ago where the laterite crust is $\frac{1}{4}$ inch thick (Avias).

2. Rocks of littoral and sublittoral marine origin, which are more or less associated with coral reefs, or with transported terrigenous material.

From the open sea toward the land, there is first a barrier reef, then a lagoon, and then a fringing reef. Coincident with the river deltas there are mangrove swamps.

Following the recent eustatic movements, the sea has fallen nearly 6 feet, causing the emergence of part of the swamp, which now only becomes flooded during the high tides of autumn. On these parts a crust of calcium carbonate 12 inches thick has formed. The heads of dead reefs buried in mud are, in part, silicified, and contain well-formed crystals of quartz with pyramidal terminations. Avias has attributed this silicification to the migratory phase of the laterization. In clays formed *in situ*, translucent crystals of gypsum 3 inches long occur. This process may explain the alternation of gypsum and marls in fossil deposits.

That part of the mangrove swamp which is still marine has channels (creeks) in which nodules are forming at the present day. They are of very hard, blue limestone, and are being deposited round organisms such as crabs, teredo tubes, and vegetation. Ammonium compounds formed by putrefaction seem to be mainly responsible for the formation of these nodules.

Very rapid alteration by diagenesis is occurring in New Caledonia, where Avias (1949) observed that American garbage dumps left in 1942 have been converted into a very hard rock cemented with calcite and iron oxide. In the same way, in the cliffs of calcareous dune sand terraces, ferruginous patches have appeared due to the diffusion of iron from shell fragments which have been almost entirely absorbed. Modern organisms such as crabs are sometimes calcified.

CONDITIONS OF DEPOSITION OF CERTAIN ROCK TYPES

Reconstruction of the Original Conditions.—Sedimentary rocks show evidence of several types of diagenetic evolution.

In some cases this evolution has been almost nonexistent, as for example, the blue clays of Tallinn which are still plastic, although they date from the Early Cambrian and thus are at least 500 million years old. At the other extreme, certain sediments have been totally altered and it is very difficult to recognize in them original structure and texture.

One of the objects of sedimentary petrography is the reconstruction of the physiography, the biotopes, and the environment which existed during the deposition of each type of sediment.

THE SEDIMENTARY IRON MINERALS

Part at least of the sedimentary iron minerals were deposited at a well-defined moment in geological history (Termier, 1954 and 1956, pp. 215–222). This comment appears to be due to L. Cayeux (1931) who stressed the fact that the demolition of a mountain chain is a source of iron and a starting point for the formation of oolitic minerals. In this case the iron minerals are retained in the molasse sediments. The mode of erosion and sedimentation of iron differs from that of detrital rocks since they are precipitates, which implies a preferential separation. According to Bichelonne and Angot (1939) this may result from the mixing of terrestrial waters containing iron salts with strongly saline, sea water, not far from the coast. The theory of biorhexistasy (Erhart, 1956) is important here: the release of the iron occurs in the forest soils of pedalfer and lateritic types. If these types of soils are absent, the iron is retained. Following the destruction of the forest, these soils are carried by running water and the deposition of their components takes place as described earlier.

Van Leckwijck and Ancion (1956) found a good example of sedimentary iron minerals in the Paleozoic deposits of Belgium. These occurred at six horizons arranged *en echelon* from south to north, from the oldest to the most recent, and seemed to follow the transgression of the Devonian sea. The first of these horizons is hematite, rich in phosphorus and silica, in the Gedinnian (Lower Devonian) of the north slopes of the Ardennes anticlinorium. The second is a hematitic and sideritic oolite at the base of the Couvinian (Middle Devonian) on the southern flank of the Dinant synclinorium. The third is a hematitic oolite, accompanying Givetian (Middle Devonian) shales of the south flank of the Vesdre massif. The fourth is an oolite at the base of the Frasnian (Upper Devonian) on the north slope of the Dinant synclinorium; it is composed of hematitic ooliths passing westward into chamosite ooliths with a calcareous or dolomitic cement. The fifth is a hematitic oolite with a silico-argillaceous matrix at the base of the Fammenian (Upper Devonian) on the south flank of the Namur synclinorium, and also in the western part of the northern border of this region. Finally, the sixth horizon is an oolitic ore at the base of the Strunian (Upper Devonian) consisting of oolites of hematite, chamosite and siderite in a calcite paste.

Comparing these facts with the theory of biorhexistasy, a number of questions arise: 1. Was the vegetation sufficiently dense between the end of the Silurian and the beginning of the Carboniferous to give rise to pedalfers? 2. Why did the dense vegetation of the coal forests not lead to the formation of ferruginous deposits during the Westphalian transgression? 3. Is the existence of ferruginous oolites (which result from a rhexistasic phase) compatible with a cement often calcareous or dolomitic (resulting from a biostasic phase)? The hypothesis of a hard pan seems better fitted to the facts.

PHOSPHATE DEPOSITS

The greater part of the phosphorus in the hydrosphere has come from the erosion of the continents. It is carried by rivers to the sea in the form of phosphate ions. However, it is the biosphere that extracts the phosphorus as phosphate, which is one of the principal nutrients of plankton. Phosphates are also important in the building of the skeletons of vertebrates and the shells of brachiopods, and are constituents of nervous tissue and of certain diastases (phosphatases of molluscs, for example). Consequently, phosphates become concentrated in vertebrate bone-beds and in masses of phytoplankton controlled by upwelling currents rich in nutrients (p. 234).

The major phosphate deposits are more common in marine series than in continental ones. Undoubtedly the latter do contain them: certain Fenno-Scandinavian lakes include among their deposits, phosphorous compounds (together with bituminous substances and limonite) which alternate with silica, calcareous muds, clays and manganese oxide ore deposits. Massive phosphorite deposits such as those of Quercy which have been formed in limestone caves, and guano deposits are well known. But the characteristic horizons, with their typical structures are all found in shallow water on the continental shelf.

In general, two types of marine phosphates are distinguished: those of platforms and those of subsiding basins (Visse, 1953).

The *platform* deposits occur in oxygenated open seas and are light in color. They generally indicate trangressions.

Thus, on the Russian platform they occur associated with the great marine advances of the Upper Jurassic, the Lower Cretaceous, the Upper Cretaceous and the Lower Tertiary. A series often begins with a conglomerate, followed by pre-phosphate sands and ending with a bed of phosphate (Kazakoff, 1937).

Visse (1953) has shown that, generally, the beginning of a phosphate series is marked by a conglomerate; or, in a lagoonal facies it may begin with a "bone-bed". This is followed by the detrital pre-phosphatic sandy clays, and then the phosphate formation in which the amount of phosphate reaches a maximum. Finally, the uppermost layers poor in phosphate are associated with limestones, cherts, conglomerates or with redbeds.

The phosphates of Morocco, light in color and deposited in a broad basin, belong to this category of platform phosphates. They contain 45 to 78% $Ca_3(PO_4)_2$ and were deposited from the Maestrichtian to the Lower Lutetian as small pseudo-oolites forming a sand or the components of a sandstone with a calcareous cement.

In the *subsiding basins* the phosphates are darker in color because they contain more organic matter. They were associated with gypsiferous lagoonal episodes and seem to be linked with regressions. The beds of the Algerian-Tunisian frontier at Gafsa are of this type. These phosphate beds show cross-bedding which indicates transport by currents.

Marine phosphates occur mainly as nodules, pseudo-oolites or coprolites accompanied by abundant plankton (siliceous skeletons of diatoms and radiolaria and calcareous tests of foraminifers), by organic matter (humic acids, cellulose, sulfur compounds, amines and organic phosphorus), by pyrite, and by gypsum formed by the alteration of pyrite. Very often glauconite is present in the transgressive facies. Traces of iodine (0·012%), fluorine, vanadium, and uranium are occasionally found. Besides these substances which are of organic origin, there also occur quartz, montmorillonite, sometimes illite, and less frequently sepiolite, which have been derived from the underlying rocks.

The phosphates which are rich in silica (cherts) or in calcite, occupy an intermediate place in the sedimentary sequence between the fine detrital sediments and the limestones (Visse, 1953). This is well shown by the Permian phosphates of North America, especially the Phosphoria Formation. In the littoral zone of these deposits the phosphates are rich in calcium carbonate and even pass into limestones, whereas toward the open sea they are rich in chert and pass into clays.

Phosphate Nodule Horizons

Apart from the great phosphate deposits economically important, phosphate nodule horizons often occur in the sedimentary series. They usually indicate a transgression. These levels are frequently accompanied by glauconite. For example, the first beds of the Cambrian transgression over the old platforms consist, as has already been shown, of glauconitic detrital sediments and limestones which contain phosphate nodules (Goldschmidt and Störmer, 1923). In France, in the Black Mountains, and in the Pyrenees the basal beds of the transgressive Visean enclose phosphate nodules.

18

Conclusions—Cycles and Causes

(Table VII)

The object of the present work has been the study of the phenomena which control the distribution of materials on the surface of the globe according to cycles of erosion and sedimentation. These problems have been discussed from a geological standpoint, noting the major events and the constant modifications which have occurred throughout the Earth's history: changes of climate and altitude, foldings, subsidences, volcanism, transgressions and regressions, marine facies and continental facies.

Stratigraphic geology teaches the sequential nature of these phenomena. Before closing we must consider the form of cycles and rhythms within such sequences.

LITHOLOGICAL SEQUENCES

A lithological sequence is "a series of two or more lithological types (lithotopes) forming a natural series, without important interruptions other than those of stratification joints" (Lombard, 1953). These sequences may be *cyclical*, *rhythmic* or *arhythmic*. Their sizes vary considerably: they may be as large as a stage, an outcrop, a hand-specimen, or they may be microscopic.

Lombard has defined an ideal series containing all the possible types, in an ideal order, which he has called a *fundamental series*. It comprises from bottom to top: coarse detrital sediments, finer sediments, colloids, limestones, and evaporites. Each sequence may contain positive or negative parts (if the order of the types is inverted).

SEDIMENTARY CYCLES

The idea of a *sedimentary cycle* was introduced by M. Gignoux (1913) in relation to the Italian "Pliocene *sensu lato*" which, he considered, consisted of the Plaisancian marls, the Astian sands and ended with the Calabrian (now considered Early Pleistocene, see Table I).

In actual fact, "every series of marine formations in a given region, which is limited between two regressions, constitutes a sedimentary cycle". Generally, a sedimentary cycle begins with coarse debris (basal conglomerate) and becomes progressively finer grained upward. This can only occur

TABLE VII.—EVOLUTION OF NONDETRITAL

	LIMESTONES AND DOLOMITES	PHOSPHATES	
PLIOCENE AND QUATERNARY	Coral reefs and deep coral patches		
MIOCENE			
OLIGOCENE		Quercy phosphorites	
EOCENE	Lacustrine limestones Nummulitic limestones	*Phosphates of North Africa*, Quercy phosphorites	
CRETACEOUS	*First Pelagic limestones* (*Chalk*) Rudistid bioherms	First phosphates of Morocco Phosphatic chalk Lower Cretaceous nodules	
JURASSIC	Oolitic limestones and coral bioherms	Phosphate nodules	
TRIASSIC	Alpine Trias (Dolomites)		
PERMIAN	Fusulina limestones of Tethys	*Phosphoria Formation* (U.S.A.)	
PENNSYLVANIAN			
MISSISSIPIAN	Carboniferous limestone, bioherms	Phosphate nodules (Pyrenees, Black mountains)	Black phosphates
DEVONIAN upper	*Griotte limestones*	Tennessee Phosphates (with pyrites) of Arkansas	
DEVONIAN middle	Many limestones and bioherms		
SILURIAN		Podolie nodules	
ORDOVICIAN	*First Algal and coelenterate bioherms*	Graphitic nodules of Wales	
CAMBRIAN			
—500 m. yrs	*First animals with calcareous skeletons* (e.g. *Archaeocyathus*)	Conglomerates with phosphatic nodules	
PRECAMBRIAN			
—600 m. yrs	Development of Stromatolites		
—800 m. yrs			
—1,800 m. yrs		Lacustrine phosphates of Scandinavia	
—2,650 m. yrs	First Stromatolites in Southern Rhodesia		

in regions where the sea has invaded a relatively high coast. The sea has often, however, transgressed across low-lying coasts of a physiographically mature continent. This has happened in the case of the Silurian, Visean (Upper Mississippian) and Middle Cretaceous transgressions over North Africa and the Sahara.

R. C. Moore (1953) lays stress on "the regular and progressive modifications of the environment" and on the return to the initial conditions which represent a complete sedimentary cycle. Several cycles may succeed one another.

The ideal cycle rarely occurs, since some phases are usually absent. The conditions occurring in a region can be very different according to the period in time which is considered. Sedimentary cycles are not wholly marine and detrital. The most spectacular are those which are mixed, and include continental, brackish, and marine horizons. Among these may be noted the coal seams of the paralic basins of the Carboniferous in Europe and North America. In such series it is usual to consider that the beginning of a cycle

SEDIMENTS THROUGHOUT GEOLOGICAL TIME

EVAPORITES	CARBONACEOUS ROCKS	SULFIDES	LATERITES
Playas and Salinas	Peats and lignites	....	Rare
Sicily (Sulfur series)	....	....	Mainly redistributed
Potash of Alsace, etc.	....	....	....
Gypsum of Paris	....	....	Africa, central Europe
....	Europe, Asia	Pyritized fossil deposits	Mediterranean zone, America, Asia
....	Coals of Asia	Pyritized fossil deposits	Mainly redistributed
Keuper deposits, etc.	....	....	Rare
Zechstein of Europe, American deposits	....	German Kupferschiefer	
....	Maximum of coal formation	....	Lateritic clays largely reworked
Gypsum beds (U.S.A.)			
....	....	....	*First forests;* Bauxites of Oural
Salina Group (U.S.A.)			
....	Kuckersite of Esthonia (Cyanophycea boghead)	Alum shales of Scandinavia	
....	Kulm (Sweden)		
First deposits of halite			
....	Shungite (Karelia)	Beginning of bacterial sulfur cycle	
....	Algal "coals" of Michigan		
....	First organic carbon		

is marked by a marine transgression and that the end occurs at the top of the coal seams, or, if they are absent, by that part of the cycle which comes closest to emergence.

However, in the case of some marine limestones there are entirely marine cycles, as has been shown by W. D. Brückner (1953). These limestones may undergo changes in their proportion of calcium carbonate, either through a variation in the amount of transported continental material, or through an increase or reduction in the solubility resulting from temperature changes (of climatic origin) in the water, or possibly from changes in depth. According to Brückner, the second explanation is the most likely. For example, in the Helvetic zone of the Alps there is a cycle which begins with an alternation of marls and limestones, and ends with more or less massive limestones. The boundary between limestone and marl is often clear and is sometimes marked by a thin layer of glauconite, phosphate, pyrite or sand. Some beds are often missing. In incomplete cycles in which there is no return to the initial conditions, it seems probable that changes in sedimentation are the consequences of variations in the temperature of the water. It dissolves more limestone when it is cold, and less when it is warm. According to Carozzi (1955), these cycles are linked to tectonic phases, and Brückner's hypothesis does not explain the observed facts.

Sedimentary cycles denote alternations (pulsations) of transgressions and regressions repeated many times over vast continental areas. In these cases subsidence and uplift are balanced against each other. The uplift or positive movement is responsible for renewing the supply of sediment (Dreyfuss, 1954). Eustatic variations in sea level are also probably involved. Moreover, sedimentary cycles occur over more or less extended zones which possess a certain degree of structural uniformity. Even toward the center of a relatively homogeneous assemblage, all variations in facies modify the beds, and in practice it is rarely possible to use them to establish stratigraphic correlations.

R. C. Moore (1948) stressed the fact that the upper third of the Pennsylvanian of Kansas, shows 35 successive marine invasions, belonging to as many cycles. But Moore was able to show that, on the whole, this series consists of regularly ordered successions of different types of cycles, which he has called "cycles of cycles".

Finally, a stage or series of stages may often constitute a *large-scale cycle* as does part of the Lower Jurassic of Morocco (Prerif, Central Atlas). This is a marine series in which the marine transgression occurs in the Toarcian between the Triassic continental phase and the regression of the Middle and Upper Jurassic. This is called a *geologic cycle.*

RHYTHMS OF SEDIMENTATION

Rhythms of sedimentation or *cyclothems* are lithological sequences which are repeated in a regular fashion. They are distinguished from sedimentary cycles in that, instead of returning to the initial state by passing through all the original stages in the reverse order, they return abruptly.

All sedimentary rocks are likely to enter into rhythms. One of the most typical examples of a sedimentary rhythm is that of "graded-bedding" in the detrital series (p. 205). In general, the negative sequence is not found, but the rhythmic repetition of the normal sequence is common.

The origin of the rhythms of sedimentation again seems linked to the pulsations of the earth's crust in the form of subsidence and positive epeirogenic movements. It is possible that rhythmic marine currents (Bersier, 1948) or benches formed by "creep" (see p. 77) may locally play some part. M. G. Rutten (1951) has emphasized the close relationship between subsidence and erosion as it effects sedimentary rhythms: "the fine-grained sediments indicate a slow erosion in the hinterland and in consequence, uplift during this period was feeble. On the other hand, coarse-grained sediments indicate active erosion and strong movements in the hinterland." The depth of the sea seems to be relatively unimportant, according to Tercier (1939).

Sequences composed of very fine-grained almost colloidal components, such as varves, are frequently banded. In these, the detrital material decreases upward, the later phases being organic colloids or lime muds. In the case of varves, it is certain that seasonal variations are important even though they act indirectly on the erosion areas supplying the detritus.

The detrital limestone successions found in certain flysch (calcareous flysch) deposits, contain calcareous sandstones and calcareous sandy shales, with limestone nodules in the shales, overlain by limestones. The succession of detrital-colloidal-calcareous rocks is found on continental areas as well as in the flysch and the molasse. The sequence of detrital-colloidal-calcareous-carbonaceous rocks corresponds to rhythms of the same order: it appears that it is to this that R. C. Moore has applied the term *sedimentary cycle*. These sequences pass into carbonaceous shale sequences. The succession of colloidal calcareous rocks results from changes in depth, and from currents and climatic variations which affect the amount of calcium carbonate (Brückner, see p. 353).

Pararhythmic successions are concerned essentially with those series in which limestones are present; the rhythm of the succeeding beds is modified. They are frequent in alternations of marls and limestones or of shales and limestones. They also are common in the flysch[1] and the molasse[1] and are known in the Cretaceous Chalk. A. Lombard (1953) has supposed it to be due to "a process of unstable equilibrium between the calcium ions and the clay ions in a state of suspension from the time of the calcium precipitation". In the Chalk, the deposition of calcium carbonate seems to have been continuous, while the transport of detrital material may have been temporary.

There are also evaporite sequences given by Lombard as, KCl, NaCl, anhydrite, clay; at times with limestone or dolomitic limestone intercalated between the anhydrite and the clay. These minor cycles can occur on a very small scale. The rhythm seems to depend on the degree of saturation of the water by each salt, and to follow essentially the laws of van't Hoff.

An explanation of the formation of rhythmic series has been put forward by A. Lombard (1953). He distinguishes the "imponderables" which are the salts in process of precipitation, particles of colloidal dimensions and organic residues which remain a long time in suspension, and the "ponderables" which sink rather quickly: coarse or medium detrital sediments, and benthonic and nektonic organisms. Naturally, there are all

[1] *Flysch:* an orogenic sediment formed during the uplift of a mountain chain, usually marine and chiefly composed of shales with rhythmic intercalations of sandstone and limestone, sharply alternating.

Molasse: postorogenic sediments, detrital and continental, marking the beginning of the erosion of the chain after its completion. Typically fluvial and conglomeratic at the piedmont, passing distally to feldspathic silts, marls and lignites in the molasse basins. Both are tectofacies in the sense of Sloss *et al.* (1949).

gradations between the two types and the boundary between them varies with the degree of agitation of the environment. But, on the whole, the "imponderables" are deposited slowly, with a clear tendency toward a leveling of the floor, while the "ponderables" are spread over the bottom under the action of gravity and of currents until they reach a position of equilibrium. They tend to show "graded-bedding" in a horizontal direction as well as in the vertical direction. An *active phase* can be distinguished during which sediment is regularly provided during subsidence. A *passive phase* is that in which only the "imponderables" are deposited, forming "stratification joints" (bedding planes). There is thus a true *geomorphological control over the sea bed.* This idea permits the reconstruction of the conditions which occurred during the deposition of the series and links up with the view of F. F. Grout who noted a "control of sedimentation by the adjustments of the crust" (1932, p. 344).

It can be said that the "positive sequences correspond to *progressive* conditions of sedimentation" while the "negative sequences indicate *regressive* conditions of sedimentation". This is the principle of the sedimentary cycle which can be divided into a positive sequence and a negative sequence.

This explanation is readily adaptable to the formation of rhythmic detrital series and also to the seasonal rhythms shown by varves (see pp. 162–163).

Cycles and rhythms may be prolonged. H. Erhart believes that they can be explained by his theory of biorhexistasy (see p. 155) according to which the vegetation on continents controls the formation of marine sediments close to coasts. This theory, which scarcely accounts for the origin of marine deposits, is well adapted to paralic/lagoonal sediments which are directly related to rivers. The normal sequence is given by many authors (e.g. Grout, p. 342): (1) conglomerate (not always present); (2) sand; (3) clay; (4) limestone.

The views of G. Millot (1957) can be readily accepted. This author distinguishes two fundamentally different types of marine sedimentary cycles:

1. The cycle of "general" type (V. M. Goldschmidt, 1937, 1945; A. Lombard, 1956) which begins after an orogenic phase with the deposition of sands and coarse detrital material, and is followed, after the reduction of surface relief, by the deposition of detrital argillaceous sediments. If the climate permits, a phase of laterization ensues which gives rise to ferruginous deposits; and finally, when the continental surface is completely reduced to base level, the deposits consist solely of soluble materials (carbonates and salts).

2. The biorhexistasic cycle (Erhart, 1955, 1956) which applies to stable regions such as shields, commences with the deposition of dissolved substances (the migratory phase of laterization) as carbonates, salts, and colloidal silica; then after the destruction of the forests, the components

of the residual phase (for example, iron and kaolinitic clays) are deposited. Finally, if there is a rejuvenation of the relief, sands and coarser detrital materials are laid down.

An example of a succession which is the reverse of the normal one occurs in Morocco (H. Termier, 1930 and 1936). At Bon Achouch the folded shales and hard sandstones of the Upper Visean are overlain by a stratigraphically discordant succession of Autunian beds consisting of (*a*) fissile blue-gray *shales* containing plant fragments, white mica, ilmenite, zircon, iron oxide, carbonaceous and argillaceous matter, 7 to 10 feet thick; (*b*) a yellowish-white, fine-grained arkosic *sandstone* composed of angular quartz, microcline, plagioclase and muscovite, and containing an abundant flora 3 to 7 feet thick; (*c*) a gray *conglomerate*, containing a sandy cement and pebbles of hard sandstone, quartz and quartzite, many feet thick. It is apparent therefore that the continental detritus becomes progressively coarser.

Examples of Rhythmic Series

THE RHYTHMIC SEDIMENTATION OF THE ENGLISH JURASSIC was noted as early as 1822 by Conybeare and Phillips. The succession, clay, "grit" and calcareous sandstone, which may be oolitic or marly, reoccurs nine times in the System (Arkell, 1933); the Rhaetic and the Lower Lias include several more cycles.

THE RHAETIC-LIASSIC BASIN OF NORTHWEST SCANIA, SWEDEN, which is of paralic type, has been partly supplied by rivers, and partly by the sea (p. 192). It provides an example of rhythmic sedimentation comprising at least twelve cycles in the Rhaetic and the Hettangian totalling 820 feet (Troedsson, 1948). Each cycle shows three rock types: (1) more or less coarse sandstone; (2) clays with carbonaceous partings containing terrestrial vegetation; (3) calcareous sandstones with siderite, accompanied perhaps by clay which contains banks of brackish pelecypods (*Ostrea*, *Mytilus*, *Modiola*, *Gervilleia*) and "bone-beds". This cycle is readily explained by the relative movements of the Scandinavian Shield and of the intermediate Fenno-Scandinavian zone of which Scania is a part. Applying the ideas noted earlier (p. 329) it can be concluded that the Rhaetic and the Hettangian represent about 2 million years, and an effective subsidence in the basin of 820 feet. Each cycle averages about 160,000 years. Also, on the average, each of the three phases forming the cycle represents approximately 50,000 years.

The fine-grained sandstone (0·05 to 0·1 mm.) is composed of quartz, feldspars and micas and often shows cross-bedding and ripple-marks. Toward the top, the sandstones become banded due to the intercalation of argillaceous laminae. The clays are black, gray or white and often contain fine quartzitic sand. The argillaceous minerals are kaolinite and montmorillonite. The thin carbonaceous layers are generally associated with a bituminous clay. Calcium carbonate is rare, occurring only in "cone-in-cone" structures, or as a cement in the sandstone, or mixed with siderite. Ferruginous sandstones and clays and nodules of clay ironstone are also present. There are also oolitic iron ores similar to those occurring in southeast Scania at the same horizon.

THE CAMBRIAN SANDSTONE OF NEXÖ on the Island of Bornholm is derived from the disintegration of a Precambrian granite. At the base there is a poorly sorted arkose which has angular grains consisting of quartz and feldspar with a brownish argillaceous cement. Toward the top the grains are better rounded and graded, while the mean size decreases. The amount of feldspar decreases and the cement becomes lighter in color and is composed of kaolin and silica. The highest beds are well-stratified sandstones about 30 inches thick to the north of Nexö. These beds become very thin and very hard toward the top. The Nexö sandstone is a littoral sandstone throughout. The intermediate stage containing the kaolinitic and siliceous cement is a good example of a convergence between the evolution of an arkose produced by a purely detrital process and that of pure chemical laterization.

"Reef" Rhythms

M. Lecompte (1954, 1956) has applied the ideas of rhythm to the study of Devonian reefs and at the same time has related these to the movement of the sea bed. In the Belgian Frasnian (see p. 288) the cyclothem is composed of the following sequence:

(1) Pure limestone, either massive or stratified, containing globular Stromatopora, formed in shallow agitated water (at the center of the bioherm).

(2) Slightly argillaceous limestones with lamellar Stromatopora and reef-building corals; formed in relatively calm water.

(3) Very argillaceous or subnodular limestone and nodular shales with brachiopods, simple and branching corals, bryozoans, pelecypods, echinoderms, intermediate in the sequence between (2) and (4) (corresponding to the top of the bioherm).

(4) Shale with small or dwarf fauna or none.

(5) Fine shales with *Buchiola* and goniatites.

Beds (4) and (5) are situated between two superimposed bioherms. According to Lecompte (1956) the character of these rocks, the absence of nodules, and the relative absence of fauna indicate that they were formed in the deepest water. They probably represent a phase during which there was transport of more abundant terrigenous detritus produced by climatic changes or to the uplift and erosion of cordilleras in this zone of the Variscan chain.

In each case, the succession occurs in the order of 1 to 5 above, from the center of the top of the bioherm outward, and in the inverse order of 5 to 1 from the top of one reef up to the one overlying it. Some zones may be absent, and this is interpreted as the result of the acceleration, for example, in the rate of subsidence.

Lecompte stresses the relationship between the formation of the reef and the almost perfect peneplanation of the neighboring continent. In fact, at the same time, many reefs were forming near high land as well as near low-lying (peneplaned) land masses.

CAUSES OF EROSION AND SEDIMENTATION

Before concluding, it seems desirable to review once more those problems which are affected by all branches of physical and natural science. The causes of erosion and sedimentation can be classified as follows:

1. **The climatic factors** which are independent of internal and external geological phenomena and which are of extraterrestrial origin. Thus, the degree of insolation received by the Earth from the sun varies between maxima and minima (fig. 1, p. 5). On the other hand, variations in the density of cosmic rays striking the surface of the globe may have influenced the physico-chemical processes which affect living organisms.

2. **The internal factors** which are orogenesis and epeirogenesis, and to a lesser degree, volcanism. Epeirogenic movements which affect vast areas, perhaps even most of the planet, act on the surface of the earth and cause differences in altitude which allow the effects of gravity to operate. Thus streams can erode in uplifted areas, while basins receive enormous accumulations of debris: the detrital sediments. Such differences in level occur not only relative to the surface of the sea or to the local base level (of endorheic basins) but also to the greatest depths of the seas. Detrital materials tend generally to move toward the lowest points even on the abyssal plains.

3. **The superficial factors** resulting from the activities of living organisms. Bacteria attack and transform the minerals of rocks, soils and muds, while the chlorophyll of plants affects directly the amount of oxygen in an environment where sediments are deposited. Finally, animals and plants are the principal source of the natural fuels and those rocks which contain carbon, for carbon is the essential element of living organisms and occurs in the form of coal, hydrocarbons and carbonates.

Among the three great groups of causes which control erosion and sedimentation, those that are of cosmic origin are mainly dependent on the solar cycle, while the internal causes are inherent in the Earth, insofar as the planet possesses a mass and a well-defined equilibrium. The tendencies of the continents to rise and the oceans to sink are not, perhaps, peculiar to the Earth, but they are dependent on its mechanics. The fact that the hydrosphere is very important in the formation of oceans and river networks does seem peculiar to the Earth, but it is possible that water vapor exists in the atmosphere of other planets. Precipitation on their surfaces, perhaps produces erosion and detrital sedimentation comparable to that on the Earth. It is highly probable that aqueous erosion is not limited to our "terraqueous" globe, but it is known, for example, that it does not occur on the moon. The disaggregation into fine powder of the surface of the moon is nowadays attributed to the diurnal alternation of very high and very low temperatures. Aqueous erosion thus places the Earth into a well-defined category among the planets. It allows the accumulation and grading of detrital material in basins which thus have a complex history and which will subsequently be subject to such characteristic phenomena as subsidence and regional metamorphism, which are stages in the Earth's history (H. and G. Termier, 1954).

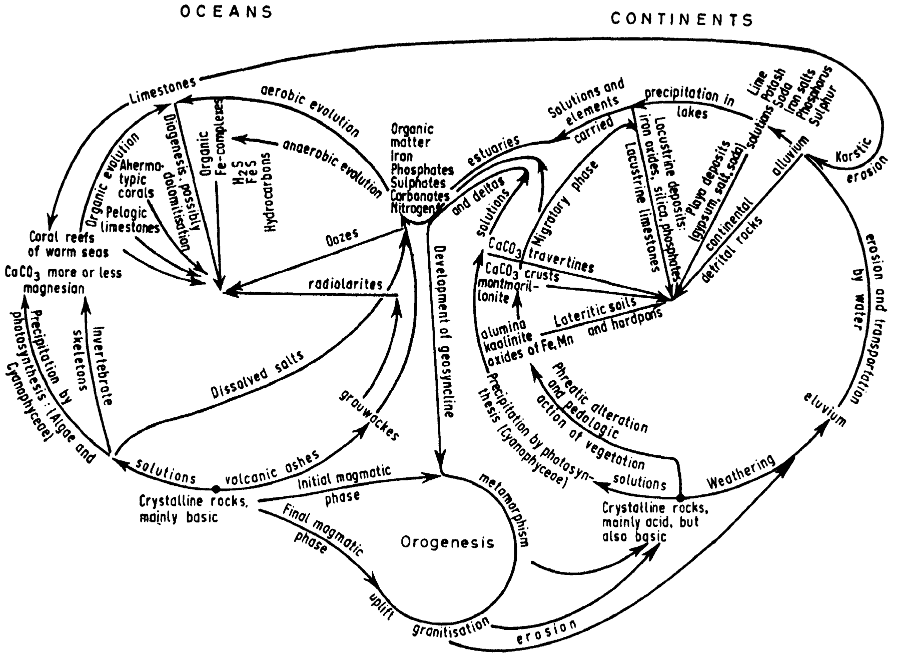


FIG. 199. SIMPLIFIED DIAGRAM ILLUSTRATING THE "PANTOCYCLE"

The superficial causes seem to be still more closely integrated with the character of the Earth. While the Earth's biosphere may not be the only assemblage of living organisms in the Universe, it is probable that it is different, even chemically, from those which may exist elsewhere. The biosphere is characterized by an irreversible phenomenon: evolution. It is this which provides the essential link between the structure of plants and animals. As geological history is unraveled, it is found that the biosphere plays an increasingly important role in erosion and sedimentation and that the evolution of organisms progressively modifies certain characteristics of their chemical affinities. The increasing abundance of organisms has played a part in the superficial geochemistry of the Earth.

It is customary to describe erosion and sedimentation in terms of characteristic cycles, either of detrital sedimentation, or of chemical elements (C, Ca, Al, Si, Fe, P, etc.).

Each individual geochemical cycle should not be considered immutable. The only permanent features of the cycle of an element are, in fact, those common to the detrital cycle. All those characteristics which link the cycle to the biosphere are progressively modified during the course of organic evolution, that is, during geological time.

THE PANTOCYCLE
(Fig. 199)

The circulation of the materials of the surface of the Earth thus comprises, not a single unalterable stereotyped cycle, but a collection of cycles, a *pantocycle* (H. and G. Termier, 1958) in which each part is subject to a suite of changes which occur as an ordered series of irreversible phenomena. The principal factors are *geological history* which governs *orogenesis*, and *evolution* which governs the *biosphere*. All are interlinked with the changing circumstances accompanying epeirogenesis and the variations in the climate.

Superficially, erosion and sedimentation represent only the leveling of the surface of the earth. However, due to various effects of the pantocycle, the proportions of the different types of sedimentary rocks formed during the course of geological time have gradually been modified and an increasingly important role has been played by the carbonate rocks and the evaporites. These are products of the hydrosphere which, despite the abundance of material going into solution, is only rarely saturated. They are also formed, in part, by the biosphere which has progressively assisted in the formation of the rocks.

Table VII illustrates the *geochemical evolution* which the surface of the Earth has undergone since Precambrian times. The evolution of the characteristic cycles of some of the more common chemical elements is shown in Figures 200 to 206.

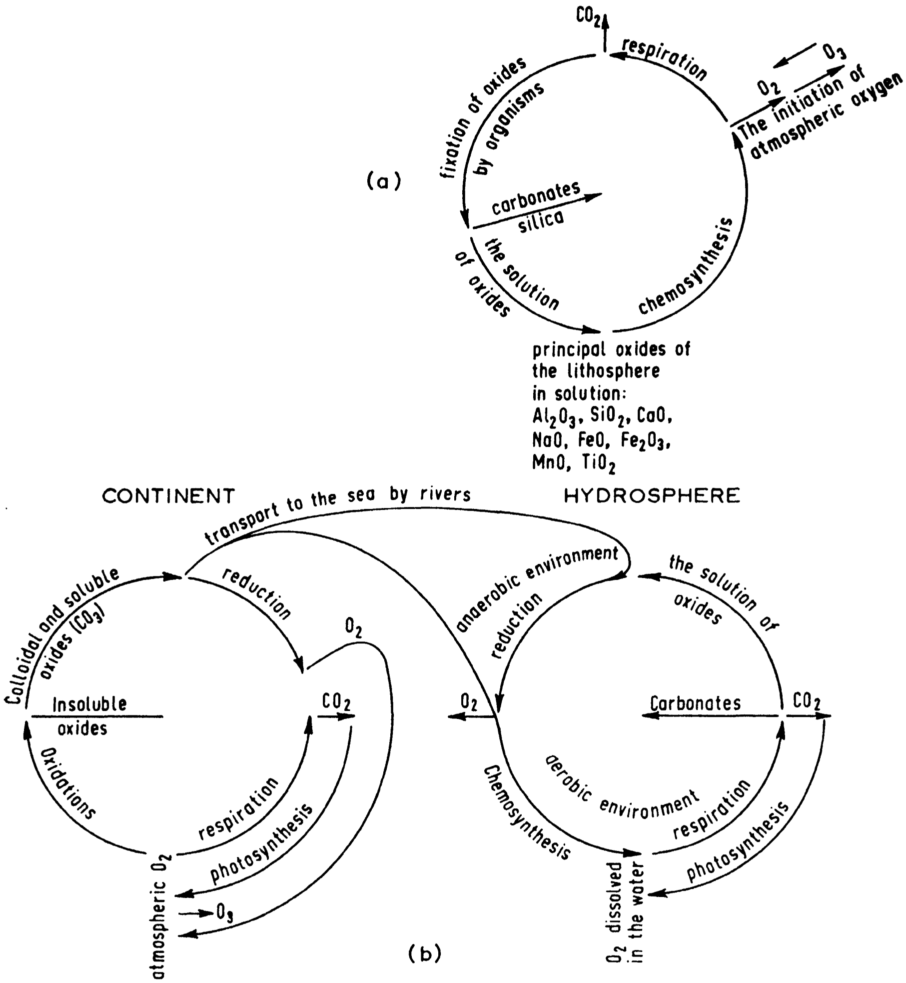


FIG. 200. EVOLUTION OF THE OXYGEN CYCLE, FROM TERMIER: GLYPTOGENESE, 1961

(*a*) The Hydrosphere, before the atmosphere became oxidized, during the first 2,000 million years.

(*b*) Present Epoch (since the formation of an oxidized atmosphere, after the first 2,000 million years).

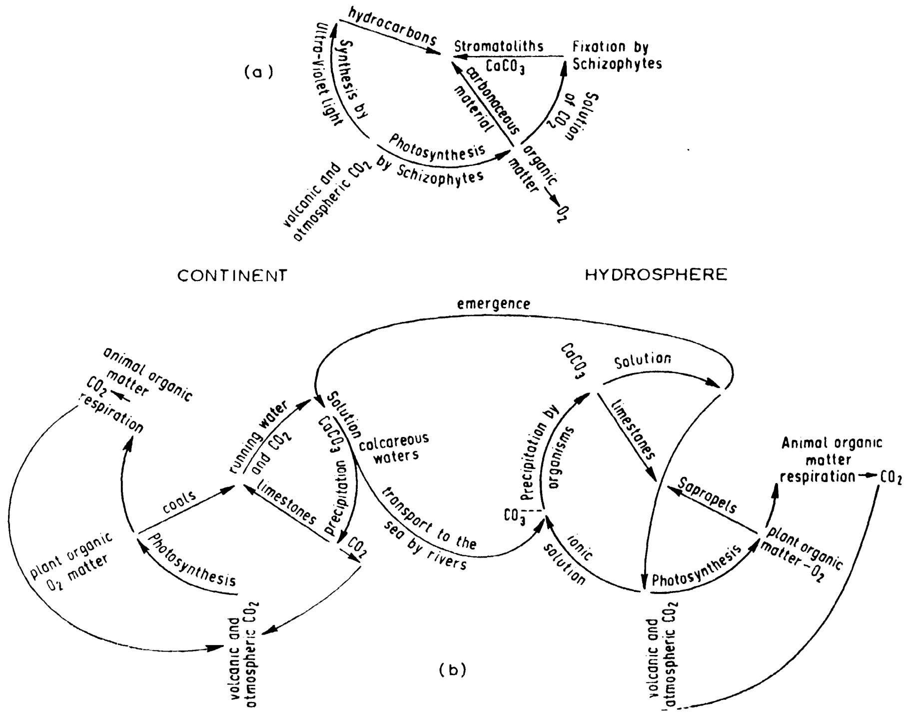


FIG. 201. EVOLUTION OF THE CARBON CYCLE (*N.B.* $CaCO_3$ is generally accompanied by Mg), FROM TERMIER: GLYPTOGENESE, 1961

(*a*) Before the appearance of land vegetation, when the atmosphere was still "primitive".

(*b*) Present Epoch (since the appearance of land vegetation).

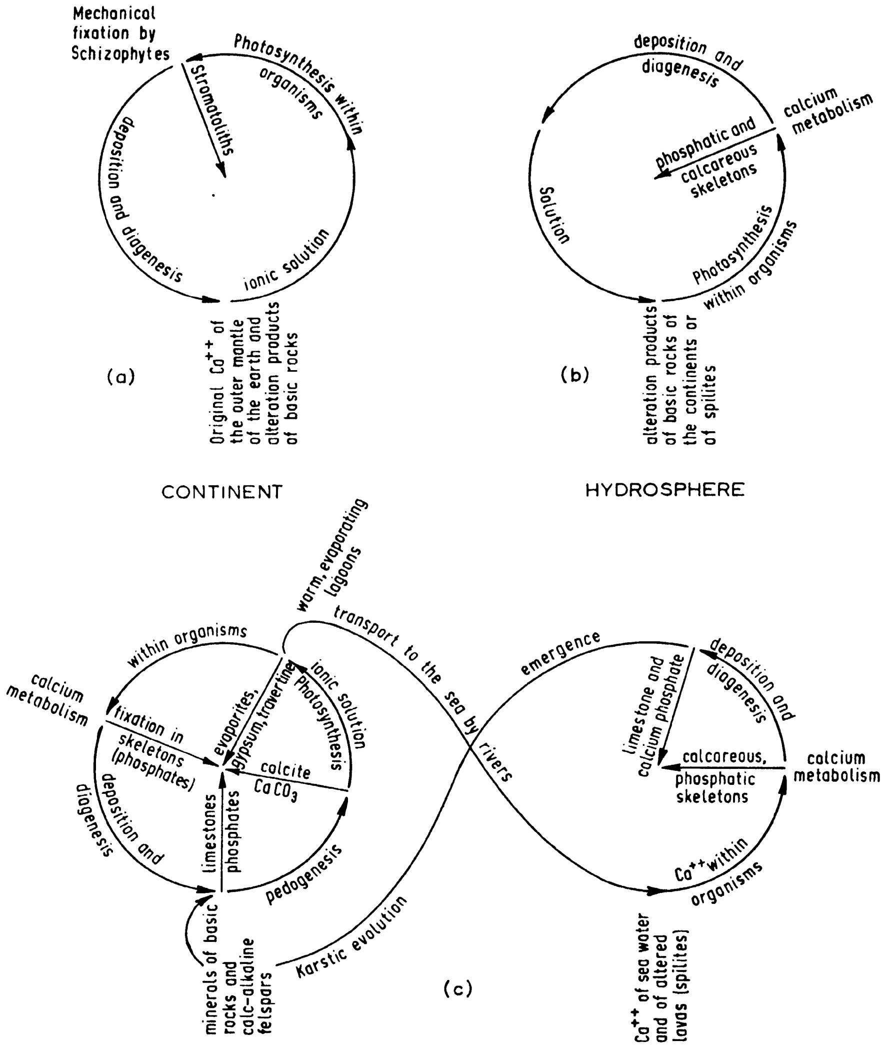


FIG. 202. EVOLUTION OF THE CALCIUM CYCLE, FROM TERMIER, GLYPTOGENESE, 1961

(*a*) Precambrian (the precipitation of limestone in the presence of chlorophyll).

(*b*) Cambrian (animals with calcareous skeletons).

(*c*) Present Epoch (since the appearance of life on the continents).

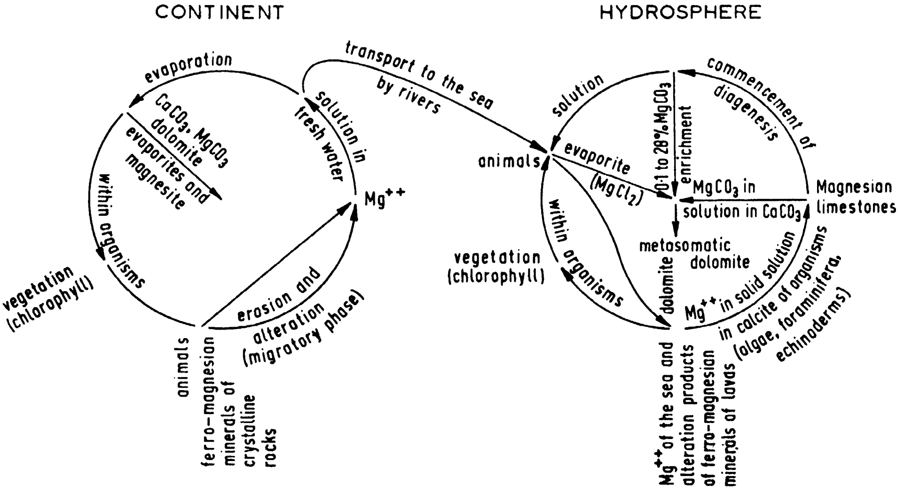


FIG. 203. PRESENT CYCLE OF MAGNESIUM, FROM TERMIER: GLYPTOGENESE, 1961

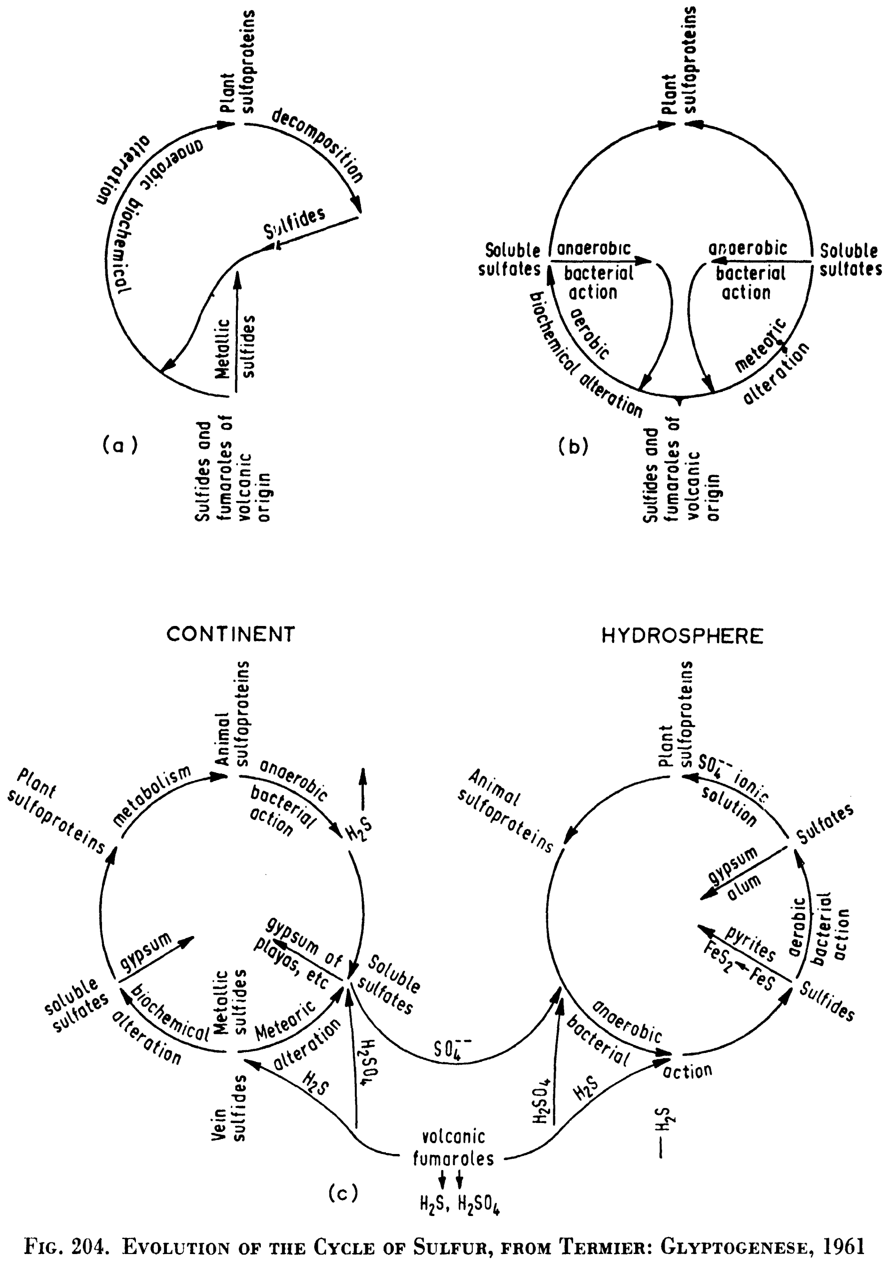


FIG. 204. EVOLUTION OF THE CYCLE OF SULFUR, FROM TERMIER: GLYPTOGENESE, 1961

(*a*) Primitive reducing atmosphere.

(*b*) The commencement of isotopic fractionation resulting from the occurrence of living organisms (after 800 million years).

(*c*) Present Epoch (since the appearance of life on the continents).

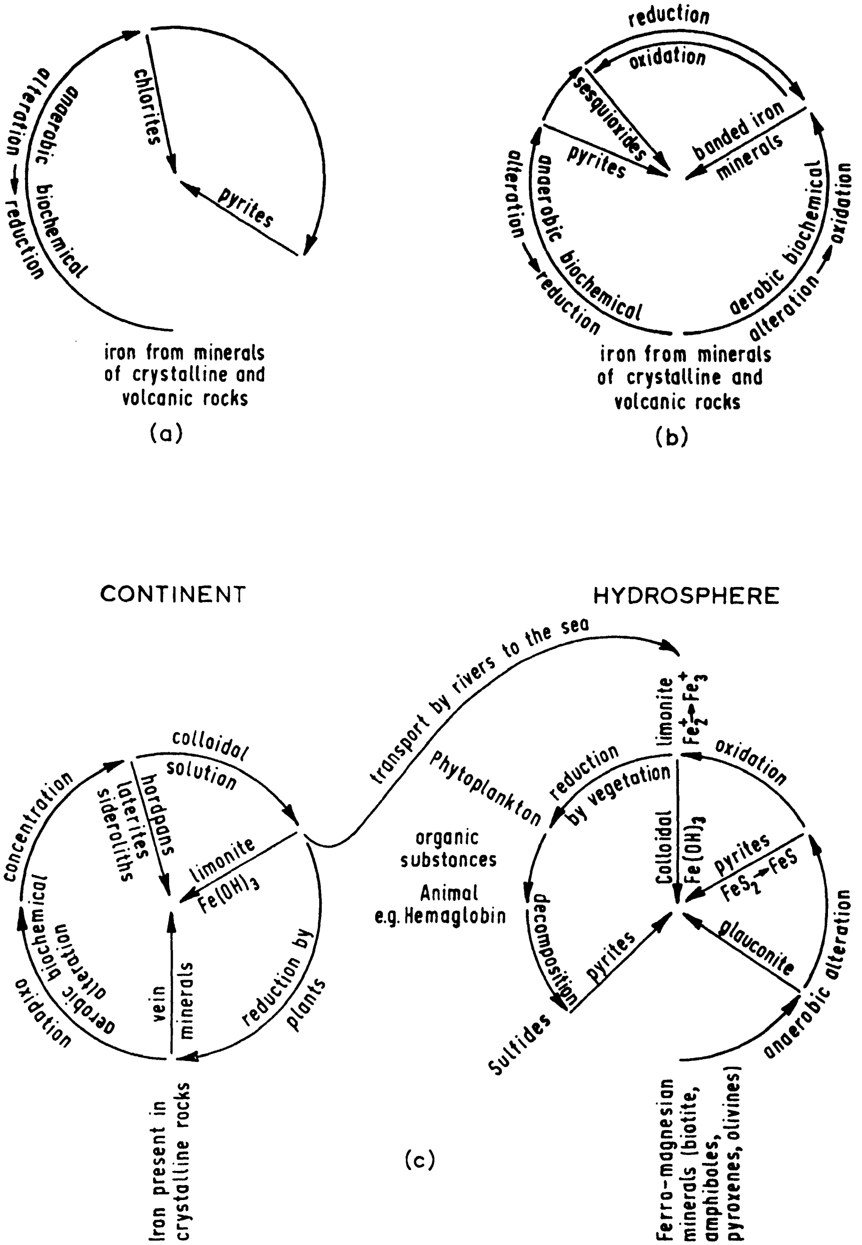


Fig. 205. Evolution of the Cycle of Iron, from Termier: Glyptogenese, 1961

(*a*) At a very early period, when the atmosphere was reducing.

(*b*) During the Precambrian, when the atmosphere was become oxidizing.

(*c*) The Present Epoch (since the appearance of forests).

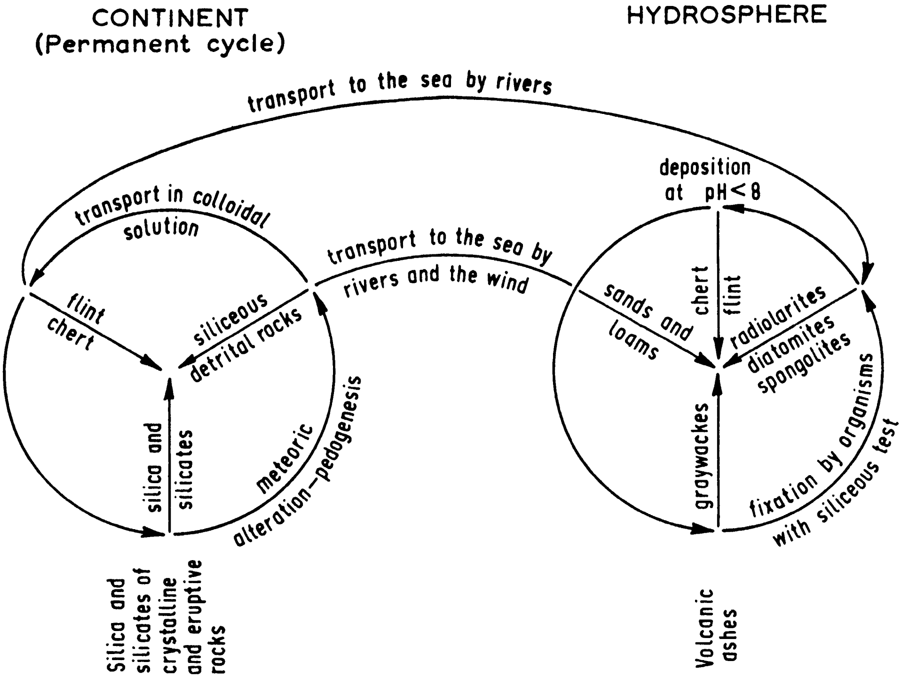


FIG. 206. CYCLE OF SILICA, FROM TERMIER: GLYPTOGENESE, 1961

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Chapters 12–14

Carbonate Sedimentation

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Glossary

(Compiled by H. and G. Termier, R. W. Fairbridge and D. W. Humphries)

Active capture — capture of one river by another, where headward erosion reduces the divide between two drainage basins, one becoming deeper than the other (*cf.* passive).

adret — a French dialect term used in the French Alps, to designate the south side of mountains exposed to the sun. The opposite, north side, is called the *ubac.*

agouni — a Moroccan (Berber) term designating a rather broad gully carved by a torrent. These gullies are generally dry on the surface, but have an underground water supply.

aguada — a small superficial depression in the karst country of Yucatan, Mexico. *Aguadas* fill with water in the rainy season.

ahermatypic coral — coral which does not contain symbiotic zooxanthellae *q.v.* (*cf.* hermatypic corals).

akalche — a bog or depression in the karst country of Yucatan, Mexico.

allophanite — a Russian pedological term indicating a black hydromorphic soil, formed on volcanic, often basaltic, rocks. (From Allophane, an amorphous gel of the clay family.)

alluvium — fluvial or torrential sediments transported some distance from their source.

anauxite (ROSS & KERR, 1931) — a variety of silica-rich kaolinitic clay $Al_2O_3.2$ to $3(SiO_2).2H_2O$, present in lateritic soils.

angular discordance — a type of unconformity (also termed angular unconformity) in which a sedimentary formation lies on beds previously folded and subsequently planed off; thus the two units show a variable angle between one another.

anteclise (BOGDANOFF, 1958) — a broad upward flexure of the basement on a continental scale (*cf.* syneclise).

anthraconite — a coal-black bituminous marble or limestone usually emitting a fetid smell when rubbed (= Stinkstone).

anticlinal river — a river which flows along the axis of an anticline; similarly a synclinal river is one which flows along a syncline.

apatotrophic (SWAIN & MEADER, 1958) — refers to lakes which are brackish and contain living organisms.

argille scagliose — an Italian stratigraphic term which describes a complex Jurassic to Oligocene group composed of *scaly* (squamous) looking clay and including mixed boulders of various dimensions. It is a typical orogenic megafacies.

arkose (TWENHOFEL, 1937) — a special type of sandstone; a coarse detrital sediment, often continental, in which there is more than 25% of feldspars, usually derived from the disintegration of granitic rocks.

arroyo (BRYAN, 1923) — a Spanish term for a normally dry stream bed, in semiarid or arid regions, e.g. in southwestern U.S.A. and northern Mexico. It is equivalent to the term *wadi.*

asif — a Moroccan (Berber) term for a large valley in a mountainous region like the Atlas. It is generally superficially dry but has underground drainage (see also *agouni*).

Badlands — an American term for regions occupied by soft sediments, in which streamcutting has produced deeply carved grooves, ridges, pinnacles, etc. These often make beautiful scenery but are useless for agriculture and bad for travelers.

bahr — an Arabic term for a river; also a local term in the Lake Tchad region (Central Africa), where it refers to the channels near the edge of the lake. These divide the shore into a series of islands, and sometimes look like rivers.

bajada — a Spanish geographic term indicating the complex of overlapping alluvial fans at the edge of an arid (endorheic) basin. The top of the *bajada* merges with the pediment.

barchan — a Turkestani word for a desert dune shaped like a crescent or horse-shoe, whose "wings" point downwind.

bathygenic movements — negative tectonic movements, i.e. the subsidence which characterizes oceanic basins. The term was conceived as the opposite of epeirogenic movements, which are uplifts and apply rather more logically to the continents.

beachrock — an English expression describing consolidated sands formed on the shores of tropical and subtropical regions. Beachrock is generally a calcarenite and is often rather similar to an eolianite, i.e. a wind-carried sand. Its rapid consolidation is derived from an abundance of calcite or aragonite and is probably helped by micro-organisms such as bacteria and unicellular algae.

bioherm (CUMINGS, 1932)	a massive accumulation built by organisms (animals and/or plants) on the sea floor. In the stratigraphic column, it does not conform to usual horizontal bedding. The term is a loose equivalent to "reef", for example "coral reef" or "organic reef".
biorhexistasy (ERHART, 1955)	a recent theory of sediment production related to variations in the vegetational cover of the land surface. The theory uses pedogenesis (soil formation) to explain geological phenomena. (From the Greek = biological stabilization.)
biostrome (CUMINGS, 1932)	a bedded structure, such as a shell bed, crinoid bed, coral bed, etc. Biostromes consist of, and are built mainly by sedentary organisms. They do not swell into lens-like or mound-like forms.
biscuits (water-, algal)	calcareous concretions found in playas or temporarily dry lakes and rivers. They are round cake-like structures of $CaCO_3$ which form around blue-green or green algae, as a result of photosynthesis (average size 1×10 cm.).
boghead (QUECKET, 1853; THIESSEN, 1925)	a compact coal formed by microscopic algae in lakes or seas after a period of intense organic productivity (= torbanite).
bolsón (pl. bolsones) (WILSON H. M., 1900)	a Spanish term meaning a bag. It is used in the arid regions of Mexico to describe a closed, endorheic depression (e.g. the Bolsón of Mapimi).
botn (WERENSKIOLD, 1915)	a Norwegian word meaning the bottom of a glacial lake or of a fjord. It is almost equivalent to a glacial valley cirque, cf. a *fjeldbotn* which is a glacial cirque in an icefield.
boulidou	a French (Provençal) dialect term for a resurgent intermittent spring in karst country.
breccia	a coarse clastic sediment containing angular elements of any origin; the equivalent with rounded pebbles is a conglomerate.
bult	originally a Dutch word meaning a hump; it is now used to describe the ancient, stabilized sand dunes of the Kalahari Desert, which are covered with scattered trees. *Bults* may be 100 feet high and are equivalent to the *Qoz* dunes in the Sudan.
burozem	a Russian pedological term for brown forest soils.
butte-témoin	a French term describing a hill or hillock of horizontal beds cut off from their lateral connection by stream erosion. Also sometimes translated into English as *witness butte*. An outlier.
Cadoule	a French dialect term (from Languedoc in southern France), indicating certain biological associations in the brackish pools of the Gulf of Lions. *Cadoules* are

Cadoule—*cont.* essentially Ostrea bioherms or reef knolls, with *Serpula*, *Hydroides*, *Anomia* and rare *Pecten*. They occur, for example, in the Etang de Thau.

calcare ammonitico rosso an Italian stratigraphic term borrowed from quarrymen. It describes a red, nodular, fine-grained Jurassic limestone, often rich in Ammonites.

calcarenite (GRABAU) a limestone or dolomite composed of coral or shell or of sand derived from the erosion of other limestones.

calcilutite a term used in Grabau's classification of limestones. It was recommended by Pettijohn to describe those limestones that are exceptionally fine-grained (clay-size particles) and homogeneous, with conchoidal fractures. They are commonly named "lithographic limestones". A calcilutite is a consolidated calcareous mud.

caliche a pedological term of Spanish origin for the crusts that are concentrated in some arid soils of the pedocal type which have been incompletely leached. Caliches seem to form *per ascendum* by capillarity. Although generally calcareous, they may also be gypseous or salty; in the arid region of Chile they are brecciated, the cement being especially rich in nitrates.

cataclinal river (POWELL, 1875) a stream which flows down the dip of the strata.

cenote a Mexican (Yucatan) word meaning sink holes. They are found in limestones, and arise from the collapse of one or several karst caves. Underground water can be seen at the bottom of these holes.

cerozem a Russian pedological term for a subtropical gray soil, somewhat salty or crusted, and developed in steppe (or savanna) country.

chernozem a Russian pedological term for a black hydromorph meadow soil, rich in organic matter, and hence very fertile.

cirque the French word for circus. Used by geomorphologists for the crescent-like head of glacial valleys; sometimes used in nonglacial country to describe a semicircular escarpment.

clint a bare, level surface developed on horizontal beds of limestone. The vertical fissures formed by solution along the joints are termed *grikes*.

colluvium coarse and detrital sediments of torrential origin deposited just at the foot of slopes, but not carried away.

consequent valley (DAVIS, 1889–90) a valley which follows the primary slope.

coorongite (THISELTON-DYER, 1872; MORRIS, 1877) (from the Coorong lagoon, in South Australia): a dolomitic marly sediment, rich in organic matter, and derived from the growth of microalgae.

copropel — an organic lacustrine sediment, originating essentially from animal excrement.

coquina — a limestone composed of loosely aggregated shells and shell fragments; an accumulation of shell debris.

crasnozem — a Russian pedological term for a red subtropical forest soil; a pedalfer not so well developed as ferralite (*q.v.*).

crons — a local Belgian term for travertine precipitated by the photosynthesis of blue-green algae and mosses below springs and in forest streams.

cryoplanation (BRYAN, 1946) — the erosion of land surfaces by processes associated with frost action.

Culm — a local term from the Harz Mountains (Germany) which has gained a stratigraphic and sedimentological significance. Stratigraphically it may be an equivalent of part of the Mississippian and may have been deposited during the Variscan Orogeny. These are mainly conglomerates and grits, often with graded-bedding, which may be interpreted as orogenic deposits largely transported by turbidity currents.

Dalbotn — the same as *botn* (*q.v.*).

dallol — a very broad dry valley, specifically a part of an old drainage system on the left bank of the Niger River (West Africa).

daya — an Arabic word from North Africa, meaning a depression whose origin may or may not be karst. It may be temporarily water-filled during the rainy season.

dess — a Moroccan term for silt deposited during "embryonic" stream flow in an arid country.

diaclinal river (POWELL, 1875) — a river which crosses the direction of folding.

diastem — the time interval during which sedimentation in an area temporarily ceases.

dimictic lakes (HUTCHINSON, 1957) — deep fresh-water lakes in temperate climates. The water in these lakes turns over twice a year, during spring and autumn.

disconformity (GRABAU) — a break in a stratigraphic sequence in which two horizontal superposed series are separated by an irregular surface, the lower having undergone erosion before the deposition of the second.

dolina (CVIJIC, 1894) — a Slav word meaning a valley; geomorphologically it refers to a dry valley in karst country, generally a closed depression.

dreikanter — a German term for a faceted pebble or boulder shaped by the wind, characteristically with three flattened surfaces.

drewite (FIELD, 1919)	white, fine-grained aragonite lime mud, first described from the Bahamas, and named after Drew, who studied the marine bacteria associated with it.
dunite	an ultrabasic rock essentially formed of olivine. It was first described from New Zealand.
dy (VON POST, 1861; NAUMANN, 1930)	a Swedish word for silt. It is used in limnic sedimentology to mean a deposit rich in colloidal organic matter.
Edeyen	an Arabic dialect term (from the Tamahäg, central Sahara and Fezzan, southern Libya) describing complex dunes, which are usually called *ergs* (*q.v.*).
eluvium	superficial weathering products, generally coarse clastic, which have remained in the same place. Eluvium is really a form of soil.
endorheism (DE MARTONNE, 1928)	inland drainage, drainage toward the center of a land mass (cf. exorheism).
enneri	a term used in southern Libya and the Niger Republic for a dry river valley (= *wadi*).
eolianite (SAYLES, 1931)	a consolidated eolian sand, generally a calcareous dune sand, with a calcite cement. (Beachrock is often lithologically similar, but initially has an aragonite cement which later turns to calcite. Eolianite is characterized by 30–32° primary dips, beachrock by 10° dips and offshore bar sands by 20–25° dips.)
epeirogenic movements (SCOTT, 1907)	the broad uplift or depression of large areas of the land or ocean floor unaccompanied by major folding or fracturing of the rocks.
epilimnion	the upper layer of the water in a dimictic lake.
erg	the Arabic term for a large area of complex sand dunes in the Sahara.
esker (DANA, 1895)	a Scandinavian word for fluvio-glacial hills occupying lake basins. Eskers are poorly stratified and often occur as long, tortuous ridges.
eustatic change in sea level	a world-wide change of sea level, due to the growth and decay of ice sheets, and to the displacement of water by accumulating sediments, tectonism, etc.
eutrophic	describes lakes which have little oxygen in the bottom waters and much nutrient matter.
exorheism (DE MARTONNE, 1928)	drainage toward the oceans surrounding a land mass.
Ferralite	a pedological term for eluvial soils originating from basic crystalline rocks which undergo chemical erosion. The resultant soil is a mixture of iron, aluminum, manganese and titanium hydrates (*cf.* laterite).

firth — a Scottish name for a long estuary, similar to a fjord. Firths originate from river or glacial valleys invaded by the sea during the Flandrian transgression.

fjeldbotn — a Norwegian glacial cirque carved by an ice field.

flysch (HERITSCH, 1929) — a German–Swiss word for "crumbly or friable material". An orogenic sediment first used for Cretaceous and Eocene rocks in Switzerland. It comprises all sorts of detrital sediments, generally marine, deposited often by turbidity currents, or slumping, and displaying a variety of tracks and trails. It has therefore also a sedimentological meaning without time-stratigraphic connotation.

fontaine — a French word for a spring. It is used by geomorphologists to describe a resurgent permanent spring. Opposite of *boulidou* (*q.v.*).

foum — in Arabic, the mouth. Used locally in the Atlas Mountains and by geomorphologists to designate the point of emergence of a river into a plain.

frozen soil — in arctic countries, a soil which remains frozen all the year and displays some peculiar morphologic features (permafrost).

Gara — an Arabic term for *butte-témoin* (witness butte).

geode — a druse or cavity lined by a mineral precipitate in which the minerals are often beautifully crystallized.

Gipfelflur (PENCK, 1919) — a German expression meaning *summit floor*, used by geomorphologists when describing young mountains, where the highest peaks are approximately of the same height. It results from erosion of a once more or less uniform uplifted surface.

gley soil — a variety of meadow soil in which the water table is so high that the lower part of the soil, lacking aeration, becomes rich in iron sulfide and phosphate.

glyptogenesis — a term used by E. Haug (1904) to mean the geomorphological carving of the earth's surface. The cycle of geological phenomena comprises lithogenesis or petrogenesis, orogenesis, then glyptogenesis. The authors used these three terms for three parts of their Treatise on Geology. The translator generally uses glyptogenesis to indicate the active sculpturing aspect of morphology.

graben (SUESS, 1875) — a German word for a trench or rift valley, formed by collapse between two faults or series of faults (opposite of *horst*).

graywacke — a very confusing term. (1) For English-speaking geologists (Krynine Classification, 1945), it is generally a dark-colored detrital rock, a medium-grained type of sandstone, the components of which are quartz,

graywacke—*cont.* chert, micas and chlorite. Generally it originates among flysch facies (*q.v.*) in geosynclinal seas, often associated with basic volcanism. In this work, it is used in that sense. (2) For German geologists, it is more often a detrital rock with graded-bedding, typically the Culm graywacke (*q.v.*). (3) For French and Belgian geologists, it is frequently a calcareo-siliceous sandstone, often very fossiliferous and particularly abundant in the Devonian and Lower Carboniferous (Mississippian) of the Ardennes.

griotte the French quarryman's term for fine-grained limestone, nodular, red and green, while the nodules recall a certain variety (griotte) of cherry. This facies, frequent in the upper Devonian of Europe and North Africa, includes Goniatites and is often used as an ornamental building stone.

growler (L. KOCH, 1945) an English term to designate a certain variety of broken-up drift ice, blocks of which rise more than 5 ft. above sea level.

guano a Peruvian term referring to extensive accumulations of sea-bird excrement in South America and in some Pacific Islands. It is used as a fertilizer, and is rich in phosphates and nitrogenous matter.

Guillestre marble a nodular red and green fine-grained limestone recalling griotte (*q.v.*). It comes from the Upper Jurassic of the town of Guillestre in the French Alps.

guyot (HESS, 1946) an American oceanographic term for a flat-topped seamount or *tablemount*, best known in the Pacific and Atlantic oceans. The flat top of the guyot is interpreted as a marine and subaerial planation surface (named after the Swiss scientist Guyot, who was closely associated with Princeton University).

gyttja (VON POST, 1861; WASMUND, 1930) a Swedish word meaning mud. Gyttja occurs at the bottom of lakes and is a rich organic deposit in which aerobic life is still possible.

Hamada (or *hammada*) an Arabic term used in the Sahara to describe a bare stony plain, from which the fine soil and sand is removed under wind action. It is a typical feature of the desertic climate.

hard-ground an English nautical term for the sea floor where currents are so strong that no sediment accumulates. The bottom consists of quite consolidated sediment or hard rock (= *hard bottom* of Twenhofel).

hardpan an English agricultural term (used mainly in the U.S.A., Africa and Australia) for a horizon in podsolic and lateritic soils hardened by precipitation and cementation. Also known in modified form, following

hardpan—*cont.*	secondary cementation and exposure as *duricrust*, siliceous, ferruginous or calcareous (Woolnough, 1927), and thus *silcrete*, *ferricrete* and *calcrete* (Lamplugh, 1902).
hermatypic coral (VAUGHAN and WELLS, 1943)	a term used for reef-builders, mainly reef corals, which are characterized by symbiotic zooxanthellae (*q.v.*) in their endodermal tissues. In shallow tropical seas these permit a vigorous photosynthesis that favors the rapid and extensive growth of organic reefs. (Also *ahermatypic coral* for those without this modification.)
heterochronism (NABHOLZ, 1951)	the phenomenon by which two analogous geological deposits may not be of the same age, although the process is similar, e.g. a continuous lithofacies may be developed in a sequence of successively younger stages, crossing time planes obliquely.
hiatus	any break, or the time value of any break in the sedimentation, without any implied folding or weathering. (See also *diastem*, a time interruption.)
homoclinal river (POWELL, 1875)	a river which flows along the strike of the beds on the limb of a fold.
horst (SUESS, 1875)	a German term used in geology to mean an uplift either between two faults or bounded by fault complexes.
hum	a Croatian word for a conical hill; a limestone variety of *butte-têmoin* (witness butte) commonly associated with the close of a karst evolution, and representing the last remains of an overlying calcareous series (also *cockpit country*).
hummock (L. KOCH, 1945)	an English word for small hill. In glaciology it is used to describe broken drift ice.
hydrolaccolith (MÜLLER, 1959)	a lens of ice which expands gradually in a frozen soil, tending to lift the upper layers, commonly to form a pingo (*q.v.*).
hydromorph (NEUSTREUE, 1926)	a variety of soil characterized by the high level of the water table, for example in river meadows.
hypolimnion	the stagnant lower part of the waters in a dimictic lake.
Illite (GRIM)	$(OH)_4K_y(Al_4Fe_4.Mg_4.Mg_6)Si_{6-y}.Al_yO_{20}$ with y varying from 1 to 1·5; a clay mineral generally of marine origin rather close to muscovite mica. (Named after Illinois.)
imi	a Berber term from North Africa used in the same sense as *foum*: the emergence of a mountain river (*oued*) on to a plain.
inselberg (PASSARGE, 1904)	a German term for a rocky hill or mount isolated by erosion in otherwise rather flat arid scenery.

irhzer — a Berber term from North Africa (Atlas Mountains) for a straight groove carved in the mountains by a stream.

ironpan — same as *hardpan* (*q.v.*) with a ferruginous composition. (Also known as *ferricrete*—Lamplugh, 1902.)

Jarosite — a yellow mineral close to alum: $KFe_3(SO_4)_2(OH)_6$ found with clays in some dry lakes and other lacustrine or lagoonal facies, associated with gypsum.

jetlozem — a Russian word for a yellow soil formed in a subtropical climate.

Kar — a German–Swiss word for a glacial cirque in the Alps.

karang — a Malay word for an emerged terrace of old fringing coral reef material, in Indonesia. Also used simply for the coral limestone itself.

karrenfeld — a German term for the *lapies* of karst scenery where small ridges or flutes of limestone are dissected and isolated by a surficial flow of water.

karst (CVIJIC, 1894) — is the name of a region of Yugoslavia (in German, *Karst*; in Slav, *Kras*) where the entire region consists of Mesozoic limestones which have undergone the special geomorphic evolution of soluble rocks. Extended now to all limestone regions which have undergone a similar evolution.

kess-kess — an Arabic term employed in Tafilelt (Morocco) for biohermal reef knolls rising above the general landscape, having been isolated by erosion.

kevir — a Persian word for a closed desertic depression, which has a salty crust, i.e. a true playa. *Kevirs* occur in the central basins of Iran.

kheneg — an Arabic term for a canyon carved in the Atlas Mountains, under an arid climate.

klippe — a German term (meaning *cliff*) used in structural geology for a tectonic outlier and in geomorphology where the isolated thrust block is of resistant material, e.g. limestone.

Kolm (TORNQVIST, 1883) — concretions of uranium-rich organic nature modified by diagenetic changes; included in the early Paleozoic Alum Shales in southern Sweden. (The English word *Culm* is sometimes used as the lithologic equivalent.)

kopje — a South African word meaning a hillock, rising above a pediment surface. Kopjes (typically granite) are of the same geomorphic family as inselbergs and monadnocks.

koris	a North African term for a water-worn, but now completely dry valley, as in the Aïr Massif (Central Sahara).
kukersite (ZAKESSKY, 1916)	an organic sediment resembling a boghead coal, found in the Ordovician of Estonia.
Kupferschiefer	a German term for very fossiliferous Triassic shales containing large concentrations of sulfides—notably of copper, but also of many other metals.
Lacullan	Anthraconite, stinkstone (*q.v.*).
lacuna	a Latin term equivalent to hiatus. Time equivalent of unconformity.
laterite; laterization (BUCHANAN, 1807; modified)	a nonsaturated red soil (pedalfer) found in humid tropical countries. It contains iron, aluminum, manganese and titanium hydrates, and is characteristically formed from basic or alkaline crystalline rocks by the leaching of silica.
lateritoid (FERMOR)	a red tropical soil, superficially limonitic or manganiferous, concentrated from acid (quartz-rich) crystalline or sedimentary rocks.
liman (GREGORY, 1913)	a branching lagoon or estuary, characteristic of the north shore of the Black Sea. It is cut off by barrier islands which are deposited by littoral currents. The lagoons thus isolated receive deltaic fluvial sedimentation.
loess	a German term for a peridesertic (generally periglacial) loam, easily carried by the wind as dust. It is typically yellow or tan-colored, homogeneous and unstratified, except where interrupted by soil layers; it is usually calcareous when fresh, but often develops vertical stringers or nodules of $CaCO_3$, or may become completely leached. Widespread in China, Central Europe and the Midwest of the U.S.A.
Marron soil	a variety of pedalfer formed under forests in subtropical mediterranean climatic zones. It is very clayey at depth.
menilite	an irregular nodular chert, found in Tertiary marls, the type locality being Menilmontant in one of the suburbs of Paris, France.
metasomatism	a chemical exchange of material (with deposition and solution) in rocks. It often contributes to the formation of ore deposits or to diagenesis.
meulière	a French word for *millstone*; sedimentary rock greatly modified by silification from ground water. It may be a sandstone or limestone, with eolian or detrital quartz grains, replaced in part by calcium carbonate

meulière—*cont.* or chalcedony. Frequently it is a cavernous, poorly bedded rock occuring in lenticles. Its hardness and texture favor its use as a millstone.

molasse an old Swiss name for a special greenish sandstone-marl facies; formed after the elevation of a mountain system and including conglomerates, feldspathic sandstones, and other detrital deposits, often lacustrine. The type area is in Switzerland and Bavaria, along the northern edge of the Alps (mid-Tertiary in age).

monadnock (DAVIS, 1893) a geomorphologic term, derived from Mount Monadnock (southern New Hampshire) which rises above the New England peneplaned upland. Its significance is close to *inselberg* but it occurs in a temperate climate, i.e. it is a residual hill or mountain which is a remnant of an eroded massif that was originally much larger.

monomictic lake (HUTCHINSON, 1957) a subtropical lake with only one hydraulic overturn, which takes place in the winter season (cf. dimictic).

Nappe (LUGEON, 1903) a tectonic term for a thrust sheet. Originally the French term *nappe de recouvrement.*

névé granular ice, snow that is partly melted and refrozen; the upper part of a glacier.

nodular chalk a variety of white limestone or chalk, commonly intercalated in Cretaceous *White Chalk* in the Paris Basin and in England. It displays not only a nodular structure, but features also glauconite, calcium phosphate and hematite. It was possibly deposited during disturbed bottom conditions.

nuée ardente (LACROIX, 1902) literally *glowing cloud* in French. It is a very fluid and mobile mixture of incandescent ash and rock fragments with hot gases, which is transported very quickly down slopes of volcanoes. The type area for such an eruption is Mont Pelée in Martinique (West Indies).

nunatak (NORDENSKIOLD; WRIGHT, G. F., 1889) an Eskimo term for mountainous peaks which rise above ice fields and which consequently have not generally been subjected to glacial erosion or peneplanation. It is a phenomenon somewhat comparable to inselbergs of arid countries and monadnocks of the humid regions.

Obsequent Valley (DAVIS, 1895) a valley (structurally controlled as a rule) sloping in the opposite direction to the general dip of the strata.

olistolith (FLORES, 1955; BENEO, 1956) an exotic boulder or even a huge block enveloped in an olistostrome (i.e. a gravity-operated slump deposit).

olistostrome (Flores, 1955; Beneo, 1956)	a sedimentary accumulation transported by slumping: the *argille scagliose* of Italy contain type examples.
omuramba (pl. omirimbi)	In South Africa, the beds of intermittent rivers; in many cases these become so choked by wind-drifted sand or silt, that the original gradients are lost, and a series of shallow lakes or *vleis* now appears after rains instead of a flowing river (L. C. King).
os	a Scandinavian term for deltaic accumulations which are transported by subglacial streams and deposited just at the snout of glaciers. (Approximately the same as *esker*, = *osar*, pl.)
oued	see *wadi*.
Paleic surface (Reusch, 1900)	a smooth preglacially eroded surface, as in Norway; type area probably early Tertiary in age.
passive capture	a stream diversion, due either to diastrophic forces, or sometimes to aggradation, resulting in the stream spilling out of its valley into a lower basin.
pedalfer (Marbut, 1927)	a warm, humid soil type, which is associated with the accumulation of sesquioxides (Fe, Al) in certain horizons of the clay complex, but which contains no accumulations of calcium carbonate.
pedocal (Marbut, 1927)	a semiarid (low-temperature) soil type, including a zone of calcium carbonate accumulation.
pelagic	appertaining to the open ocean: a term used to describe communities of marine organisms which live free from direct dependence on bottom or shore; the two types are the nektonic, or free-swimming forms and the planktonic, or floating forms.
pelite	an equivalent term, favoured by French-speaking geologists, for lutite, a rock of clay-sized particles. It is often a fine-grained, silty shale, in which clay minerals may or may not be dominant. Since argillaceous rocks (i.e. rocks composed of clay minerals) are normally designated "clay" or "shale", pelite becomes, by implication, a fine-grained rock of very fine quartz or rock flour; also "pelitic tuff" for one volcanically derived.
peridotite	an ultrabasic (melanocratic) crystalline rock, in which there is no quartz or feldspar, consisting essentially of olivine (= *peridot*), and possibly pyroxenes, amphiboles, magnetite and ilmenite.
periglacial	1. the area adjacent to the border of the Pleistocene ice sheets; 2. the climate of this area; 3. the phenomena induced by this climate.
permafrost	an arctic soil in a permanently (perenially) frozen condition.

phreatophyte — a plant that obtains its water from the zone of saturation.

pingo (PORSILD, 1938) — an Eskimo term for a conical hill. An arctic soil phenomenon, generally a hill which may reach 150 ft. in height, formed in the permafrost by a hydrolaccolithic process. A lenticle of ice grows in volume and eventually assumes the appearance of a pseudo-volcano by the top bursting.

pisolite — a concretionary sediment or soil of pea-size grains.

planosol — an intrazonal soil with a well-developed hardpan; such a soil forms much more rapidly than it is eroded.

platière — a local French term approximately equivalent to *cadoule* (*q.v.*).

playa — a Spanish word for an ephemeral lake or sheet of water, often with high saline concentration, located in a land-locked endorheic basin.

podsol — a deeply leached type of soil in eastern Europe and northern Asia. It was first described from Russia, and is found in cool humid (forest-covered) regions south of the tundra zone. It is a pedalfer which has undergone a downward migration of clays and colloidal and sesquioxide components.

polje (CVIJIC, 1894) — a Yugoslav term for a small karst plain, deeply incised in limestone country, when the base level of streams coincides with the ground water level.

puddingstone — a conglomerate with generally well-rounded pebbles, reminiscent of an old-fashioned pudding.

Radiolarite — a very fine-grained siliceous rock of homogeneous texture, originating as a radiolarian ooze, and composed primarily of the lattice-like skeletal framework of Radiolaria. Frequently, but not always, such rocks accompany oceanic volcanics like basaltic pillow-lavas and serpentines (*ophiolites*) and other deep-sea facies, in the same stratigraphic sequence.

rasskar — a hanging cirque, characteristic of Norway, which is interpreted as an old scree channel having been initiated at the end of Glacial times, and carved upward by weathering.

reg — a Saharan (Arabic) term, rather similar to *hamada* (*q.v.*). Principally used for low stony plains undergoing deflation: winds carry away the sand and leave only polished and patinated pebbles.

regolith (MERRILL, 1897) — a residual fragmental soil derived either from underlying rock as eluvium (lower part) or partly from alluvium (upper part).

regur — a Hindustani word for the black *cotton soil* typical of some parts of India (and Africa) which occupies a

regur—*cont.* large area in the southern (basaltic) part of the Indian Peninsula. It is a hydromorphic soil formed in the grassy savanna plains of the tropical humid zone.

rendzina a gray or brown soil, overlying calcareous rocks, frequently found in central Europe, thus typical of temperate climates which have both summer and winter rain.

ria (DAVIS, 1895) a Spanish word for the estuary or mouth of a river deeply invaded by the sea. A ria coast is suggestive of deep-stream erosion drowned by a transgression. Many present-day coasts, particularly along the Atlantic Ocean, exhibit rias formed during the Flandrian (Holocene) transgression.

roche moutonnée (DE SAUSSURE, 1796) a French expression (literally *sheep rock*). Refers to the polished, striated and rounded surfaces of large boulders or rocky outcrops shaped by glaciers (sometimes called also in English *sheepback rock* or *gray wether* because they look like sheep on a distant hillside).

Saekkedaler (HELLUND, 1875) an equivalent of *botn* (*q.v.*).

saï a piedmont plain, characteristic of the Tarim basin (central Asia), covered by patinated pebbles derived from the disintegration of larger cemented blocks.

sansouire a dialect term of southern France for a salty lagoon found in a temporarily submerged alluvial plain. Its vegetation reminds one of that of a *schorre.*

sapropel (POTONIÉ, 1906) a black, stinking organic mud which liberates hydrogen sulfide into the overlying water, and produces an anaerobic environment.

scaglia rossa a special red calcareous shale lithology: in particular a stratigraphic term for a Cretaceous formation in northern Italy.

schorre a Dutch word meaning the part of a shore that the sea covers only during spring tides; generally mud flats vegetated by halophytic grasses, etc.

scoriaceous limestone a former nodular limestone in which nodules have been dissolved, so that the surface has become pitted and irregular like that of volcanic scoria (i.e. *cellular*).

sebkha or **sabkha** an Arabic term for a depression, generally close to the water table, and covered with a salt crust. Characteristic feature of North African and Arabian coastal lands.

serir an Arabic term (= *dry*), also an equivalent of *reg* (*q.v.*).

serpentine a green variegated rock mainly of hydrated silicates of the form $Mg_3Si_2O_5(OH)_4$, derived from *peridotites* (*q.v.*) by hydration or sometimes from dolomitic limestone.

shott (chott) — an Arabic word for a type of playa occupying the lowest part of an endorheic basin. It is a flat and salty depression, always moist, but rarely has flowing water. The shotts are fed by ground water or artesian water, which rises to the surface by capillary action, through the sandy clay floor of the depression.

sial (SUESS, 1906) — the upper part of the earth's crust which is mainly composed of *si*lica and *al*umina. The "ideal" rock type is granitic in character, and restricted to continents.

sima (SUESS, 1906) — the lower part of the earth's crust which is mainly composed of *si*lica and *ma*gnesium. The "ideal" rock type is basalt or periodite, common in ocean basins.

slikke — from the Dutch *slik* or mud. It is a part of the shore marked by intertidal mud flats which are rich in decaying organic matter mixed with sand and cut by tidal channels. There is no vegetation.

smolnitz — a Bulgarian variety of chernozem soil, which is poor in humus and which is produced under a Mediterranean climate.

soengei (*also* sungei) (FAIRBRIDGE, 1951) — a Malayan word for a river, also with a special use for marine channels in the Aroe Islands. These cut right across the islands and are the remains of a Quaternary antecedent fluvial system, invaded by the recent Flandrian transgression.

solonchak — a Russian pedological term for a saline soil, notable for its sodium chloride and sulfate content. In the Caspian Sea area, it applies also to salty depressions like North African sebkhas.

solonetz — a Russian pedological term for a saline soil dominated by sodium carbonate.

soloti — a Russian term for a solonetz soil which has undergone hydrolysis so that it becomes a calcic soil with silica and sesquioxides.

stinkstone (JAMESON, 1804) — a fetid bituminous limestone.

subsequent valley (JUKES, 1862) — a structurally controlled valley, developed after a consequent valley into which it passes.

sulphuretum (GALLIHER, 1933) — a mud in which hydrogen sulfide is produced by the action of anaerobic bacteria (cf. *sapropel*).

synclinal river (DANA, 1863) — a river flowing along the axis of a syncline.

syneclise (BOGDANOFF, 1958) — a slowly formed, deep structural depression of the basement that is filled with thick sediments.

Taffoni — a Corsican dialect term for honeycomb cavities developed in the vertical faces of crystalline rocks. They always occur on the sides facing toward the south.

takyr — a Russian term for an ephemeral lake or depression temporarily converted into a rain-water pond.

talet — a Berber term of the High Atlas, meaning a dried-out torrential gully.

tangue (BOURCART and CHARLIER, 1959) — a local term for the complex calcareous mud of the shallow bays in the north of Brittany (N.W. France) which is in part a fluvio-lacustrine silt reworked by recent marine transgressions and mixed with finely powdered molluscan shell material. It contains 25 to 60% calcium carbonate.

tepee structure (NEWELL and others, 1953) — a North American Indian name for a tent, used to describe disharmonic sedimentary structures, comprising a symmetric ridge between two horizontal beds. These structures were probably formed during diagenesis (possibly expansion of anhydrite to gypsum).

terra rossa — an Italian name for red soil, a ferruginous residual clay, resting on limestones from which it originates under Mediterranean climatic conditions.

thalweg — the German term for a valley floor occupied by a river and its alluvium: the longitudinal valley profile.

thanatocenosis — a group of organisms brought together after death, e.g. by current action.

tillite — a consolidated till or old boulder clay, i.e. a glacial deposit left by the melting-out from a ground moraine.

tirs — an Arabic term used in pedology to describe some black Moroccan soils that are rather similar to chernozems.

tjäle — a Swedish word for *permafrost* (*q.v.*).

tombolo — an Italian term for an offshore sand bar that has connected an island to the mainland.

trachyphonolite — a volcanic lava whose composition lies somewhere between trachyte and phonolite; generally of high viscosity.

travertine — from the Italian travertino, or tivertino: a calcareous concretionary sediment deposited by stream or spring waters which have become saturated with $CaCO_3$ by flowing through limestones. The classic locality is the famous Tivoli spring, near Rome.

trogschluss — a German term for a type of glacial cirque similar to *botn* (*q.v.*).

tsunami — a Japanese term for a strong tidal wave generally caused by submarine earthquakes and mass slides. Often of ocean-wide effect and potentially very dangerous to coastal structures.

tufa — a calcareous concretionary rock similar to *travertine* (*q.v.*), generally restricted to specially spongy varieties associated with some springs and waterfalls.

tuff a rock composed of volcanic ash laid down in fresh or sea water, or subaerially. If not cemented, it is often very soft.

Ubac a dialect term from the French Alps, for the northern slope of a hill or mountain (the shady side of the mountain).

unconformity a break in sedimentation, marked by a structural non-sequence; also applies to igneous-sedimentary contacts.

uvula (Cvijic, 1894) a Slav term for a collapse in karst basin country, produced from several coalesced dolinas.

Varve a Swedish term for bed. Employed by de Geer for thinly banded dark clay and light-colored sandy layers, deposited in lakes, and usually derived from glacial or fluvio-glacial sedimentation. Each pair represents a single year; the coarse-grained light band is formed in summer and the dark organic layer in winter.

vey a term used in Normandy (France) for a relatively extensive area of unconsolidated sediments, the greater part of which is alternately covered by the tides (= tidal-flat area, *slikke*, *q.v.*).

vlei in South Africa, an intermittent swamp or marsh in a river valley, similar to an elongated playa or sebkha, evaporating in the sun. On the Great West Australian Plateau there are examples of these "salt lakes", which are remnants of dismembered river systems.

Wadi an Arabic term widely used in arid or subtropical countries for an intermittent river, often completely dried out; also applies to the valley itself.

water bloom the name given to the occurence of prodigious quantities of unicellular algae (usually blue-green algae) in marine and fresh waters which give rise to a visible coloration of the water.

wave-cut bench an erosional feature common to sea coasts and major lakes, for a narrow shore platform approximating low-tide level; generally produced by a combination of subaerial erosion (down to base level) and the mechanical removal of this debris by waves.

wildflysch (Kaufman, 1886) a clastic sedimentary association, displaying large, irregularly sorted boulders and slumped material such as olistostromes. Like ordinary *flysch* (*q.v.*) it is an orogenic association, but the beds are often wildly contorted (hence the name).

Xerophyte — a plant which can tolerate a considerable degree of desiccation.

Yardang (HEDIN, 1904) — a geomorphological term for ridges, alternating with rounded troughs, carved by the wind in piedmont sediments and terrace deposits; they were described for the first time from the Tarim Desert (Central Asia).

Zooxanthellae — unicellular green algae, i.e. flagellates with chlorophyll which live by symbiosis in many marine animals (Radiolaria, Foraminifera, Siphonora, hermatypic corals, Tridacna, etc.). By their photosynthesis, CO_2 is removed and oxygen produced. In combination these gases play an important role in the host's metabolism.

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For the most complete treatment of geological terms, see *Glossary of Geology and Related Sciences*, Washington, American Geological Institute, 2 ed. (incl. supplement), 1960, 325 + lxxii pp., and *A Glossary of Geographical Terms* (ed L. Dudley Stamp), London, Longmans, Green, 1961, 539 pp.

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Index

Figures in **bold type** indicate places where topics are discussed in some detail; figures in *italics* refer to illustrations; * denotes an entry in the Glossary, pp. 399–417.