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WOODS HOLE OCEANOGRAPHIC INSTITUTION
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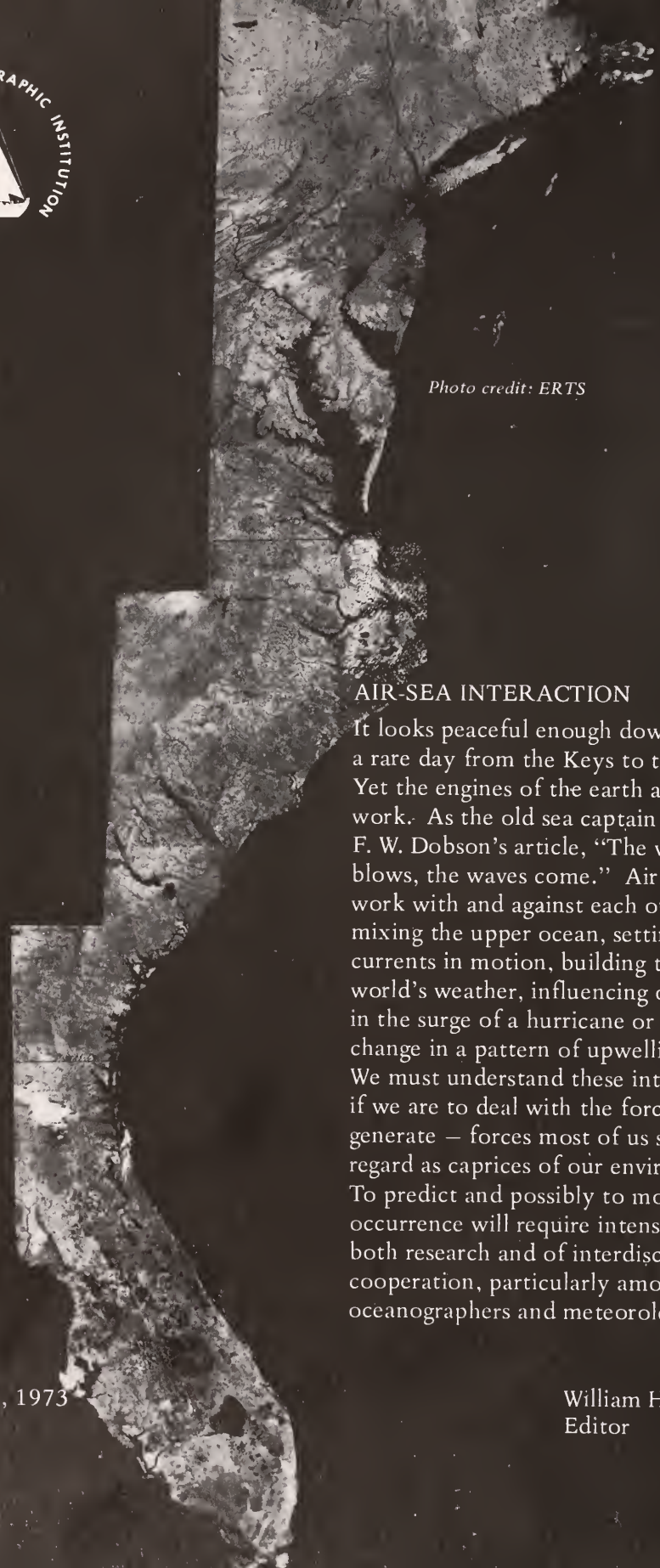

Photo credit: ERTS

AIR-SEA INTERACTION

It looks peaceful enough down there—a rare day from the Keys to the Cape. Yet the engines of the earth are at work. As the old sea captain says in F. W. Dobson's article, "The wind blows, the waves come." Air and sea work with and against each other, mixing the upper ocean, setting currents in motion, building the world's weather, influencing our lives in the surge of a hurricane or the change in a pattern of upwelling. We must understand these interactions if we are to deal with the forces they generate — forces most of us still regard as caprices of our environment. To predict and possibly to modify their occurrence will require intensification of both research and of interdisciplinary cooperation, particularly among oceanographers and meteorologists.

Cover Photo:
Hurricane Ava, June, 1973
Credit: NOAA

William H. MacLeish
Editor

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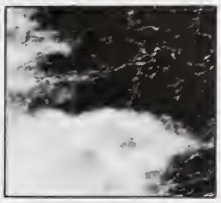
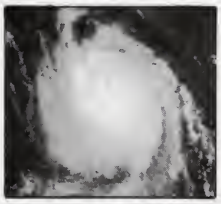
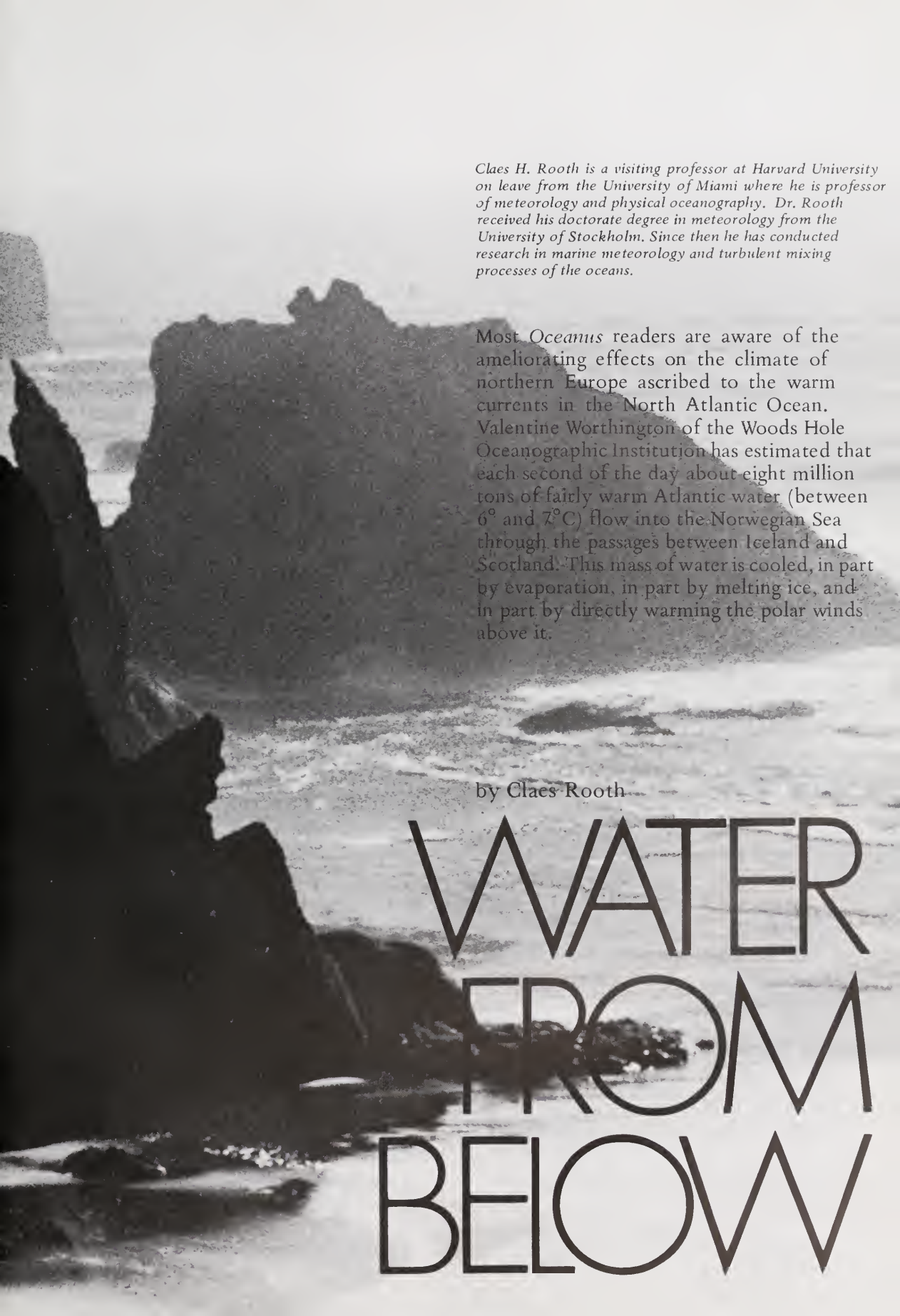
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts





Fog along Oregon shore can be traced to upwelling off coast. (Photo credit: F. B. Grunzweig, Photo Researchers)



Claes H. Rooth is a visiting professor at Harvard University on leave from the University of Miami where he is professor of meteorology and physical oceanography. Dr. Rooth received his doctorate degree in meteorology from the University of Stockholm. Since then he has conducted research in marine meteorology and turbulent mixing processes of the oceans.

Most *Oceanus* readers are aware of the ameliorating effects on the climate of northern Europe ascribed to the warm currents in the North Atlantic Ocean. Valentine Worthington of the Woods Hole Oceanographic Institution has estimated that each second of the day about eight million tons of fairly warm Atlantic water (between 6° and 7°C) flow into the Norwegian Sea through the passages between Iceland and Scotland. This mass of water is cooled, in part by evaporation, in part by melting ice, and in part by directly warming the polar winds above it.

by Claes Rooth

WATER FROM BELOW

The total amount of heat involved is roughly equivalent to the total solar radiation that would fall on the Scandinavian peninsula if it were placed under cloud-free skies in the tropics. Since most of this heat is given off along the shores of Scandinavia in winter, its effect on her weather is easy to appreciate.

When it returns to the Atlantic, as it eventually must, the water has cooled off by about 5°C. A small fraction has been diluted by melt water or coastal runoff and so does not sink, but the bulk of the return flow occurs at substantial depth in the ocean. Yet this cold, dense tongue of water does not always remain where the more familiar laws of physics put it. Under certain circumstances, it reappears at or near the surface at places often thousands of miles from where it sank. Such reappearances, termed upwelling events, have profound effects upon climate, resources, and thus upon man himself.

To understand why the upwelling phenomenon is concentrated within quite limited portions of the total ocean areas, we must first review a few basic principles in ocean dynamics.

In the gamut of physical phenomena associated with large-scale motions in the atmosphere and in the oceans, one stands out as universally important. It is the deflection of moving objects or fluid elements from linear motion – really from great circle paths – which arises as a consequence of the earth's rotation.

A ship might make the passage from Miami, Florida to the mouth of the English Channel along a great circle route in about 16 days. To us this is the most direct path, and it is perceived as straight. But viewed from space, the ship traces a spiral with 16 turns (one for each revolution of the earth) drawn out by the earth's orbital motion.

Thus, an ocean current which appears to flow in a straight path is actually describing a spiral in space. It is not surprising, then, that just as a centrifugal reaction is generated by twirling a stone at the end of a string, so an apparently straight-line motion on earth leads to a reaction force associated with the curvature of the actual path through space.

Although not extremely difficult, the detailed analysis of this phenomenon falls outside the scope of our present discussion. What is important to us here is simply the fact that objects moving along the earth's surface seem to be acted upon by a force dependent upon the motion itself. Its magnitude, like the inertial reaction in Newton's law for accelerated motion, is proportional to the mass of the object. But rather than being dependent on acceleration, this rotational reaction effect is proportional to the speed of motion, and, like the centrifugal reaction, it acts in a direction perpendicular to the direction of motion.

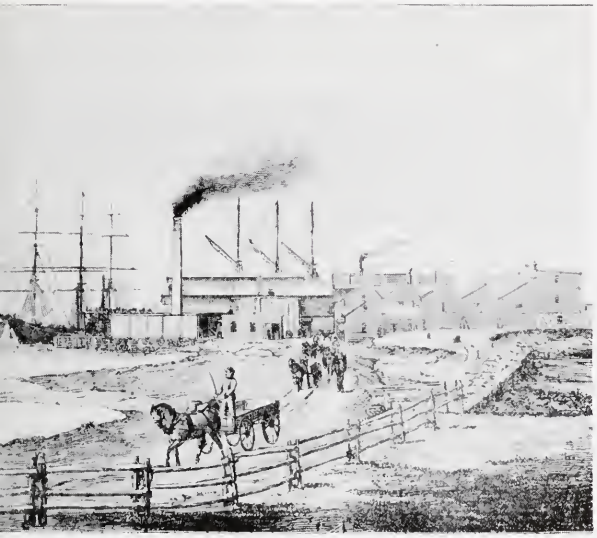
Mary Sears

Guanay birds of Peru feed in waters enriched by upwelling. Seabird droppings were shipped to plants such as the

This apparent force is known as the Coriolis force, and the associated proportionality factor is generally called the Coriolis parameter. The magnitude of the Coriolis parameter varies with location on the earth as the sine of the latitude angle. The deflection of motions due to the Coriolis effect occurs to the right in the northern hemisphere and to the left in the southern hemisphere.

When winds blow over the oceans, they generate a drag force on the surface, which is usually called the wind stress. If we suppose the sea to be calm initially, the near-surface waters will first be accelerated in the direction of the applied wind stress. But as downwind current speed increases, a deflection to the right (assuming northern hemisphere conditions) develops due to the Coriolis effect. If the wind is maintained for several days, a balanced state tends to develop where the wind stress is completely balanced by the Coriolis force. V. Walfrid Ekman, an early

National Marine Fisheries Service



acific Guano Company, which produced fertilizer and offended oses in Woods Hole between 1863 and 1889.

twentieth century mathematician, was the first to analyze this problem satisfactorily. Hence the aggregate transverse water flow associated with the sustained application of a surface stress on the ocean is generally known as the Ekman transport. Variations in Ekman transport are a key element in the generation of upwelling phenomena.

Away from the immediate vicinity of either surface or bottom, a different primary balance of forces dominates the dynamics of winds and currents. The Coriolis effect is here kept in check by pressure gradient forces acting between regions of high and low hydrostatic pressure. This state of affairs is known as geostrophic balance, and the geostrophic currents associated with it flow around the pressure anomaly centers, approximately following the isobaric (constant pressure) curves in horizontal surfaces.

The general ocean circulation regime encompasses the combination of Ekman transport patterns and the geostrophic current fields in a consistent global motion field. We will now select for attention those special situations where this coupling leads to an upward motion into the surface layers of cold interior waters from the deep ocean.

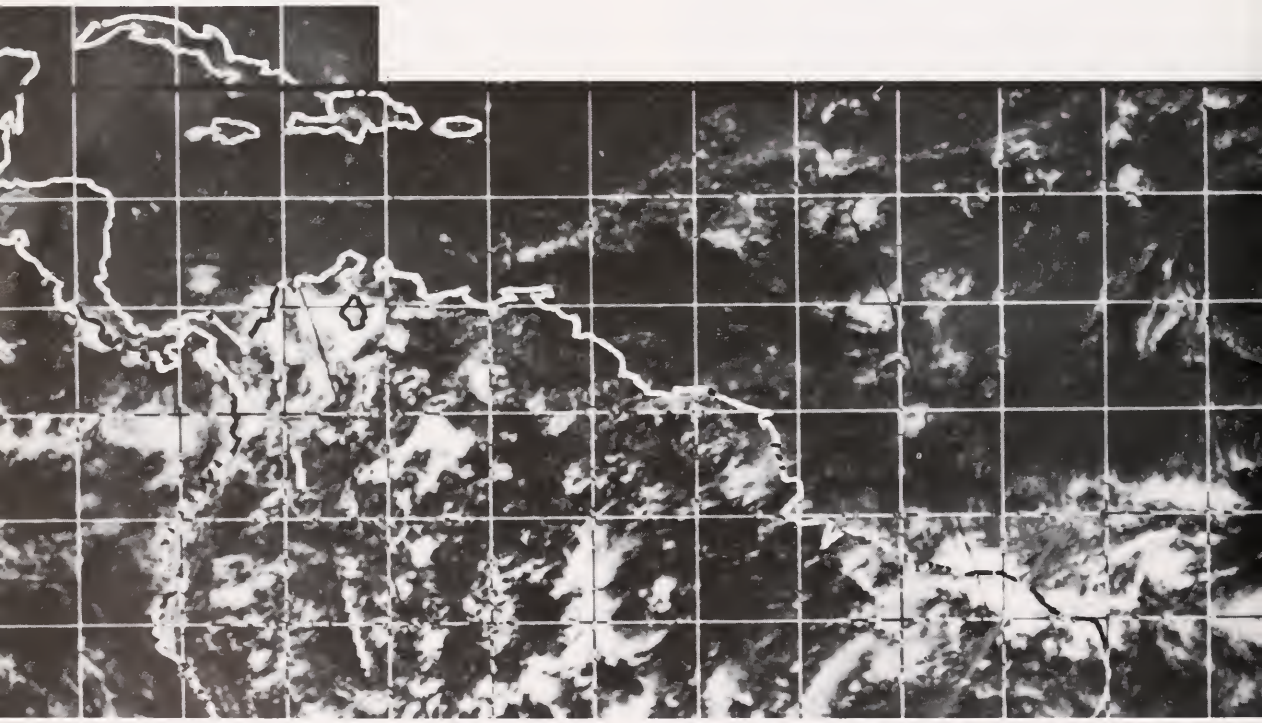
The diagnosis of upwelling situations depends on the basic principle of mass conservation: currents move water around, but over time the same volume must enter a region as leaves it. It is true that for a while a net inflow or outflow could occur, leading to a change in sea level, but if such imbalances are maintained for extended times, large sea-surface slopes would result. The associated pressure fields would then tend to cause modifications of the motion patterns until balance in transport was achieved.

Now, when a wind blows parallel to a coastline, the tendency is to set up an Ekman drift in a direction perpendicular to the coast. But near the coast, water can only be supplied to feed this horizontal motion by drawing it up from below. The consequences are sometimes dramatic, as they are along the Oregon coastline in summer. Here, northerly winds generally prevail, driving the warm Pacific surface waters offshore, and thus creating a cool ribbon of upwelled water along extensive stretches of the coastline. Low banks of fog usually blanket the cold water region, particularly near the transition between cold and warm waters – the upwelling front. By reflecting much of the incoming sunlight, this fog slows down appreciably the warming of the upwelled water and thus helps stabilize the resulting cold conditions.

Another important stabilizing effect arises in the upwelling mechanics through

the development of a coastal current jet (see page 48) associated with the upwelling front. Since the upwelling cold water is much denser than the surface water it displaces, it must be actively lifted by some mechanical force. This force is supplied by a pressure drop at the coast which arises as the offshore Ekman drift begins. Since water cannot flow through the coast line, the sea surface level is lowered at the coast, causing a shoreward pressure gradient to form in the entire water column below.

Near the surface, this pressure gradient inhibits the veering of the wind-accelerated surface waters, and the acceleration proceeds until a current develops where the Coriolis force balances the coastward pressure drop. But once it is established, this balance does not depend at all on the surface wind. Therefore, if the wind were suddenly to stop, the coastal current provides a barrier to the shoreward return of the warm water which was pushed offshore by the Ekman drift. It takes a wind reversal to break down this barrier by destroying the current and by setting up an onshore Ekman drift of warm water. Otherwise, only the heat of the sun can act to eliminate the cold coastal



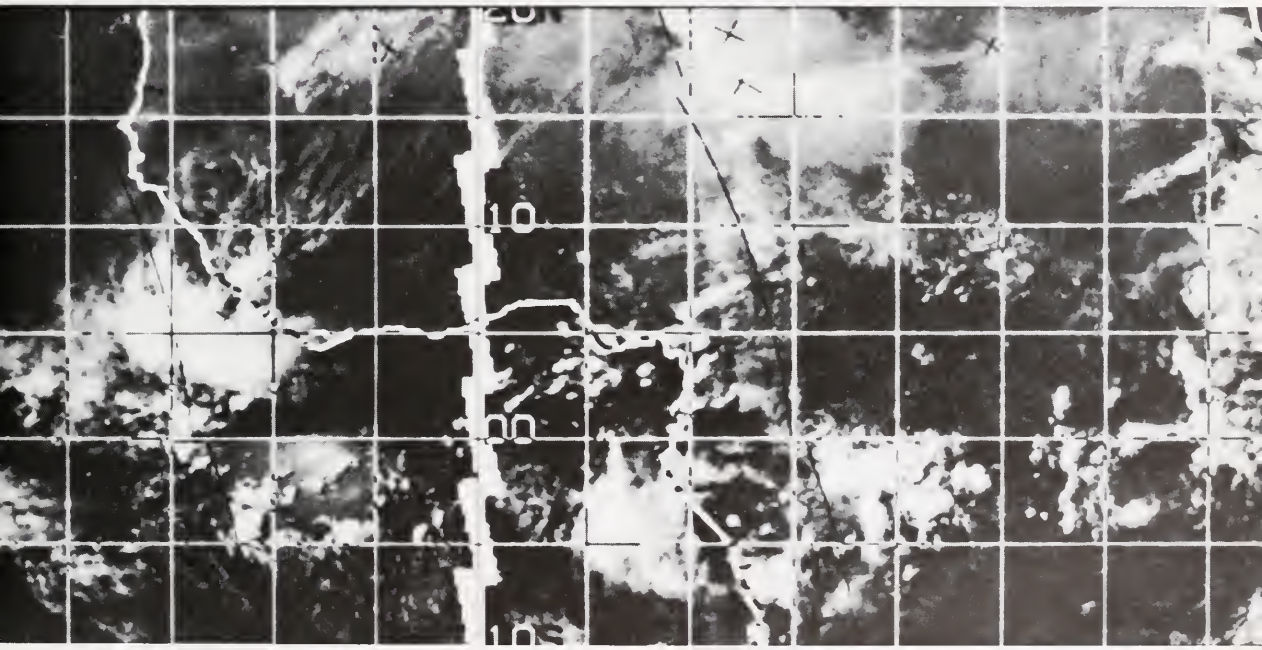
condition after the upwelling-causing winds subside.

To the occasional visitor, the most dramatic effect of this coastal upwelling regime along the Oregon coastline probably is the degree to which its climatic effect has discouraged the dense coastal settlement which otherwise characterizes the shorelines of the United States. But offshore, along the upwelling front, a biological scenario unfolds which makes the study of the upwelling processes one of the most important areas in current marine research. The oceanic ecological system depends critically on the recirculation of nutrient matter which has been used up in the growth of organisms within the near-surface waters. As they and the various groups of animals feeding on them defecate and die, a steady rain of the by-products of life settles into the deep ocean waters. As this matter settles, it is attacked by bacteria and broken down like

refuse in a compost heap. But while the weekend gardener can return the nutrients to the soil by spreading his compost, the oceans must rely on the upwelling processes to return to the surface the nutrients which have been dissolved into the water during the decay process. The upwelling circulation acts as a giant nutrient conveyor, causing riotous biological activity along the intensive upwelling zones.

A famous example is the anchovy fisheries off the Peruvian coast. There, through the intermediary action of sea birds, some of the nutrients end up on land in the form of mountains of guano. In the days of the great whaling ships, these guano deposits were the last resort to save the economy of an unsuccessful whaling voyage by bringing home fertilizer for New England gardens.

Other important coastal upwelling regions, with associated productive fisheries are located in areas of steady trade winds or monsoons, such as the northwest and southwest coasts of Africa, and along the northern boundary of the Arabian Sea. In contrast to the humid conditions along the cool Oregon coastline, these more tropical regions are desert-like all the way to the shore line. Several factors, such as large-scale trajectories of the air masses and higher rate of solar heating of the coastline, contribute to the aridity. But in addition, the temperature dependence of the thermodynamic properties of humid air leads to great differences in the relative significance of evaporation and direct sensible heat exchange as heat transfer mechanisms at the sea surface. The heat capacity of a kilogram of air does not change with temperature within the normal range of surface temperatures, but the capacity to hold water vapor and its associated latent heat of condensation grows exponentially with temperature; it approximately doubles with a warming of 10°C. Consequently, one finds that in the warmer climates most of the solar heat absorbed by the oceans is not used to



Satellite photograph of the area in which the GATE experiment will take place. Continents are outlined for emphasis. (Photo credit: GATE)

warm the atmosphere directly, but instead serves to humidify the oceanic air masses. This provides an enormous latent energy store, which is responsible for the violence of the tropical weather phenomena. By competing with the evaporation process for the incoming solar energy, the upwelling cold water is therefore interfering at a fundamental level with the energetics of the atmospheric convection processes.

Meteorological satellites have given us a breathtaking overview of weather patterns unfolding over the entire earth. Daily picture sequences are used in the weather forecasting services to help interpret the state of the atmosphere, and the model computations are performed on giant data processing machines. As hundreds and thousands of these pictures of cloud distributions accumulate, it also becomes possible to combine them in order to give us information about long-term average properties of the weather, even in the most remote regions of the globe.

A feature which stands out very clearly in such composite pictures, particularly when they are prepared on a seasonal basis, is a narrow band of cloudiness somewhat off the equator – generally in the northern hemisphere. This band marks the predominant propagation path of tropical weather disturbances associated with extensive cloud development. It is called the Intertropical Convergence Zone, or the ITCZ for short. Strong east winds characterize conditions on the poleward side of the ITCZ, while wind conditions on the equator side are generally quite confused. This zone is therefore an area of rapid variation of the poleward Ekman transport in the surface layer, and thus a zone of intensive upwelling. The last conclusion is, for the moment at least, essentially theoretical, since we do not have adequate data in hand to establish its applicability to the actual situation in the oceans.

This presentation has touched lightly upon many things, and some difficult problems have been necessarily much simplified. Above all, the link between the sinking of cold waters to abyssal depths and the near-surface upwelling process, which draws directly on water from the first few hundred meters, has been passed over completely. In between these areas, there must occur some other forms of upwelling processes which are not dependent on local weather phenomena, being inaccessible to them, but on the interior dynamics of the ocean itself. While consistent theoretical models have been proposed, lack of critical experimental information leaves us ignorant of the real mode of functioning of this, the main thermocline region.

The upwelling process represents, from the meteorological point of view, a feedback effect in the dynamics of the atmosphere, where oceanic weather phenomena modify the upper ocean conditions in ways which in turn exert influence on the weather itself. Such interactions are assuming increased significance in our attempts to understand the meteorological processes, as both fundamental research and weather prediction development tend to become concentrated on understanding longer time scale phenomena. The important role played by ocean fisheries as a source of food and as an economic mainstay for many countries with otherwise scarce natural resources lends further importance to research in this area. We have no assurance at present that severe perturbations in an upwelling system, such as that caused by the El Niño condition off Peru, are predictable in terms of their onset or persistence, but the stakes are high in terms of finding a rational basis for human resource investment as well as for resource management.

The magnitude of the experimental challenges in generating the requisite information to test our ideas about oceanic dynamics and transport processes has forced individual researchers and their institutions to band together in cooperative ventures known by acronyms like CUEA, GEOSECS, MODE, and NORPAC[1]. These are all essentially oriented toward marine science

and bear in various degrees on the problems discussed here.

A major meteorological observation project aimed at understanding the dynamics of the perturbation systems that jointly constitute the intertropical convergence is scheduled to take place this summer and early fall. Part of a long-term project to improve our understanding of global meteorological processes, it is known as the Global Atmospheric Research Programme, or GARP, which is jointly sponsored by both the governmental and the scientific international meteorological organizations. GARP has been stressing to an increasing degree the significance of joint studies of the atmospheric and the oceanic processes. Accordingly, the upcoming experiment, which will focus on the tropical North Atlantic, has within it a substantial oceanographic component aimed at studying the response of the upper ocean layers to the individual atmospheric perturbations as the latter are being studied by the meteorologists. Known as the GARP Atlantic Tropical Experiment or GATE, this effort will marshall the greatest concentration of observational resources ever applied in a single experiment in environmental dynamics. Over 30 ships, 9 of them equipped with weather and tracking radar sets, a dozen aircraft, special satellite coverage and a substantial number of instrumented oceanographic buoy systems are involved. Ten countries have contributed ships and four have contributed aircraft to the program. Scientists in many countries are committed to assist in the evaluation of the observations after the completion of the experiment.

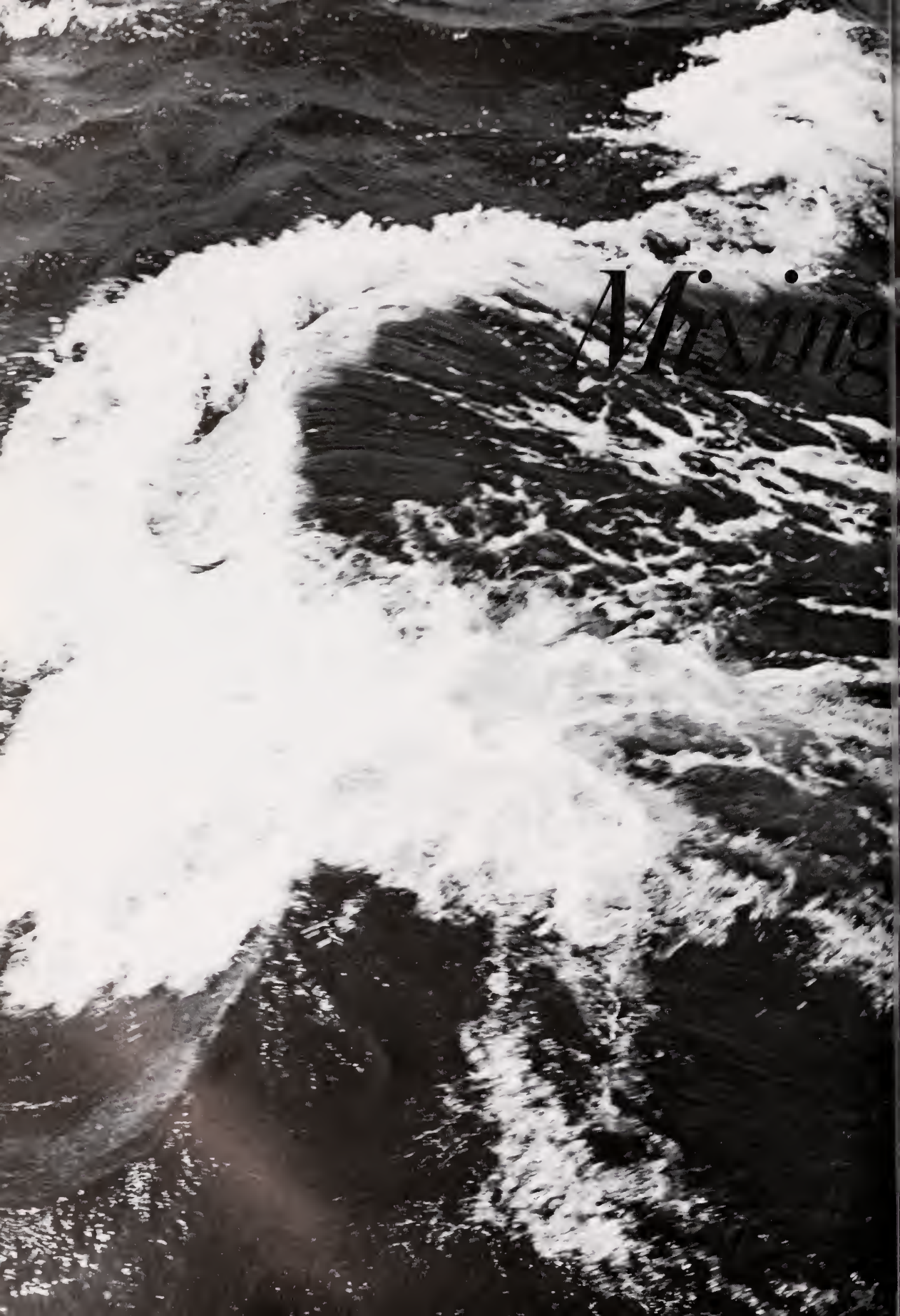
The oceanographic observations have three major focal points: direct modification of surface properties by meteorological influences; transient structure of ocean currents and upwelling due to local stress perturbations; and propagation of wave-like perturbations with special emphasis on the equatorial region. In addition, a routine hydrographic observation program on all participating vessels will provide, for the first time in the history of oceanography, a data set comparable to a meteorological synoptic map series.

Severe problems will undoubtedly appear in any program of this size; we can expect at best only partial success. But it still seems a fairly safe thing to predict that the last half of the Nineteen Seventies will see substantial advances made in our understanding of tropical air-sea interaction processes, and thus of a major link in the global heat transport mechanism – largely based on both the successes and the inevitable failures of GATE.

William H. MacLeish

Cold, upwelling water provides nutrients resulting in an abundance of marine life.

[1] Coastal Upwelling Ecosystems Analysis, Geochemical Sections Study, Mid-Ocean Dynamics Experiment and North Pacific Experiment.


Mining

Peter M. Saunders holds a doctorate in meteorology from Imperial College in London. He is an associate scientist in the Institution's department of physical oceanography specializing in air-sea interaction.

in the Upper Ocean

by Peter M. Saunders

Most of the sun's radiant energy penetrates only a meter or so into the upper ocean. Without wind the heat is retained there, producing a shallow warm layer. In summer, after several calm days, a layer one meter thick may have warmed up 5°C. (You have probably encountered this phenomenon while swimming in a lake.) A series of windy days with the same heat input but with attendant turbulent mixing in the upper ocean produces a warming trend of 0.2°C over a depth of 25 meters. The same arguments hold on a seasonal time scale: the annual range of surface temperature – the difference between winter minimum and summer maximum – depends critically on the extent and intensity of turbulent mixing in the upper ocean. Far from the margins of continents, the annual range, though only 2–10°C, has a profound effect on both marine life and the seasonal cycle in the atmospheric circulation. In winter, when the temperature contrast between Pole and Equator is a maximum, the atmospheric flow becomes intense and stormy. In the summer both the thermal contrast and the flow weaken.

Though by no means exhaustive, these considerations suggest the importance of answers to several questions: how is turbulence generated in the upper ocean? how intense is it? how deep does it penetrate? – questions about to become, if not already so, some of the most actively pursued in the field of air-sea interaction.

The existence of turbulent mixing in the upper ocean is illustrated in two simple ways – by visual observation of the diffusion of dye or natural debris, and by the lowering of a recording thermometer. The latter generally reveals the presence of the "mixed layer", an isothermal column of water whose depths vary from less than a meter to over one thousand meters, depending on location and season. Beyond these statements very little can be added because few observations have been made. This is in striking contrast to the regions just *above* the sea-surface where turbulence has been quite intensively investigated, especially in conjunction with studies of waves. The experimental problems are more severe below the interface, but until recently it has been the lack of critical inquiry which, in my judgment, has discouraged experimentalists from the formidable task.

How is turbulence generated in the upper ocean? I believe at least four processes are at work: wave breaking, convection, and two kinds of 'shear instability' arising from the currents produced by the drag of the wind on the sea surface. I shall try and describe each in turn. When the crest of a wave becomes steep enough, water begins to plunge down the leading edge of the wave and rapidly buries itself below the surface. As the now

Breaking waves are one of several processes at work in generating turbulence in the upper ocean. Water plunges down the leading edge of steep waves, burying itself below the surface. (Photo Credit: Jan Hahn)

NOAA

Sulphur dust, released over ocean by low-flying plane, settles on surface to form a reticular network of lines in a wind blowing from right to left. After some 20 minutes windrows develop about 20 meters apart.

smooth crest moves on, it leaves behind a mass of mixing water and entrapped air. Wave-breaking first occurs when the wind exceeds 10 to 12 knots and becomes increasingly frequent with increasing wind speed. Such mixing is undoubtedly very important close to the surface. My guess is, however, that it does not penetrate more than a few meters down, even in quite strong winds. Convective mixing, on the other hand, commonly penetrates tens of meters and sometimes over one thousand meters. Its principal cause is the evaporation of a few millimeters of water per day, a process which extracts heat from the skin of the ocean, producing cooler, saltier water there. Under gravity this dense water sinks, first in thin filaments that subsequently aggregate into clumps or plumes of turbulent mixing fluid that grow with depth. The more rapid the cooling (rapidity is increased by strong, cold winds), the more intense the convective turbulence and the deeper it penetrates.

Next we come to the mixing which arises from currents produced by the drag of the wind on water. The drag decreases downward from the sea surface, and so does the speed of the current produced by it. This condition, a change in current with depth or 'current shear', gives rise to a kind of turbulent mixing manifest by windrows. Windrows (also called Langmuir cells) can be visualized as pairs of counterrotating helical vortices pointing in the direction of the current shear (see page 45); the vortices carry surface water down perhaps tens of meters and simultaneously produce upwelling of water to the surface from these depths. William McLeish of NOAA, Miami, has shown that windrows are the big brothers of similar circulations as small as a few centimeters apart. The motions are spatially irregular, change with time, and can be properly described as turbulence.

At the bottom of the mixed layer, not only does turbulent mixing change abruptly, perhaps within a vertical distance of only a meter, so also do the density and the mean current. These conditions favor the generation of "billow turbulence", a process first discovered in the relatively quiet region just below the mixed layer by John Woods of Southhampton University. Billow turbulence can be visualized as horizontal vortices (much as Langmuir cells) with their axes normal to the current shear – that is across the current rather than down-current. Many oceanographers believe billow turbulence to be the principal mixing process in the bulk of the ocean, and it may even be important at the base of the mixed layer. (In the atmosphere, billow turbulence is frequently encountered by high flying aircraft and described as clear-air turbulence).

Does billow turbulence control the depth of the mixed layer? Does convection invariably occur when the mixed layer deepens? Does wave-breaking or down-shear turbulence penetrate to the bottom of the mixed layer? These questions were amongst the considerations which led some English colleagues to organize a series of oceanographic experiments in the Eastern Atlantic called the Joint Air-Sea Interaction

Experiment (JASIN). In the fall of 1972, I joined the second experiment of the series. For the rest of this article, I will describe some results of measuring mean currents in the mixed layer during that experiment. These findings are an essential step in understanding 'shear instability'.

Conventional current meters consist of a rotor (a paddle wheel on a vertical axis) to measure speed and a vane to measure direction. Data is commonly recorded internally, since in use the instrument is left unattended for weeks or months. Its design presupposes that currents are nearly horizontal and change slowly with time. In the mixed layer, neither of these assumptions is realistic because of the effects of waves on the ocean surface – sea and swell. As all students of oceanography know, water particles travel under waves in almost circular orbits in a vertical plane: the amplitude decreases with depth, becoming negligible when the depth exceeds one half of the wave length (or distance between the wave crests). A current meter, whether it be on a fixed support in shallow water or on a mooring under a buoy, is surrounded by orbital motion in the water. It is a characteristic of the upper 15 meters of the ocean that the orbital motion is larger than the average current velocity. Thus, the instantaneous current, the sum of average current and the orbital motion, not only reverses periodically but also becomes vertical flow, sometimes up and sometimes down. To measure the current accurately, not only must the rotor respond to the horizontal part of the instanteous current, but the vane must respond rapidly as the current changes direction.

Simple tests show that neither of these conditions are met by conventional devices. We already have evidence that because of the improper response of the rotor, currents are overestimated when measured at mid-depth by instruments hung below a surface mooring: the heaving of the buoy is transmitted down the mooring line almost unattenuated, so that the average current becomes all but impossible to separate from the vertical motion. In measurements made near the surface, this appears to be a less severe problem than that of the vane response. The JASIN experiment provided evidence that a current meter with a large vane (one meter long in an Aanderaa meter) cannot respond to current reversals in the mixed layer (the vane tends to get stuck in the direction of the mean current). Because the rotor is not responsive to reversals, currents computed from it may be 50 centimeters per second faster than those measured with a small vane (10 centimeters long in a vector-averaging meter). But is even the small vane responsive enough? Are its currents overestimates too?

An alternative method of current measurement is provided by a drogue. In the JASIN experiment Dr. John Swallow of Britain's Institute of Ocean Sciences constructed drogues consisting of a canvas screen (1.5 meters by 3 meters) stretched over a frame in the form of a cross and suspended below a floating buoy. The canvas cross is thus held at fixed depth while the buoy is tracked from an attendant ship and the currents derived. The tendency of the wind and waves to drag the surface buoy is resisted by the great drag of the canvas, so that the drogue comes quite close to following the water at its depth.

Encouragingly, the intercomparison between measurements made with drogues and small-vane current meters show close agreement. The agreement is not perfect, but we have determined that at least some of the variation is due not to the instruments but to the ocean – to the spatial separation of the current meter and the drogue. When average currents are computed over a 12½-hour tidal cycle, the influence of spatial variations is much reduced. This agreement is over an oceanographically important range of current speeds; it gives support to my assertion that the measurements we need to understand turbulence in the upper ocean are possible though arduous and strengthens my belief that they will be made shortly.


G
DANGER KEEP AWAY

Persistence in the Pacific

by Robert R. Dickson

Robert R. Dickson is a fisheries specialist from Lowestoft Laboratories in Great Britain. He has recently spent six months at the Scripps Institution of Oceanography working with Dr. Jerome Namias and others in connection with the North Pacific Experiment (NORPAC), a study of large-scale interactions of ocean and atmosphere.

Further reading in this field would include Sawyer, J.S., "Notes of the possible physical causes of long-term weather anomalies." W.M.O. Technical Note No. 66, pp. 227-248, 1965; and Namais, Jerome, "Large-scale and long-term fluctuations in some atmospheric and oceanic variables". pp. 27-48 in "The Changing Chemistry of the Oceans," Nobel Symposium 20, *Almquist and Wiksell, 1972.*

Most of the total variability of the atmosphere is accounted for by weather anomalies[1] with periods of less than one month. Nevertheless, a significant part of the total variability appears to be due to longer-period changes. Thus, when we mask the influence of the rapidly changing, short-term weather systems by averaging meteorological data over periods of months or seasons, we do not merely arrive at the long-term average or "normal" state. Instead these monthly or seasonal averages show major deviations from the normal situation and these features – pressure anomaly cells for example – tend to be large (5000 kilometers across).

Of course, these time-averaged anomaly patterns are statistical features: the presence of a low-pressure anomaly cell on a seasonal pressure anomaly chart does not imply that the cell was continuously present in that location during each day of that particular

[1] In the context of this article a departure from the long-term average situation in ocean or atmosphere is referred to as an "anomaly" or, in shorthand form "D.M." (departure from the mean).

season. Instead, these cells are thought to reflect the persistent reappearance of short-lived systems in relatively fixed locations. In the example just described, it is probable that the low-pressure anomaly cell reflects the passage of a number of depressions which merely reached their greatest development or were abnormally delayed in that location. When one considers a sequence of these time-averaged anomaly charts, the main cells of each pattern may be traced from one chart to the next and are clearly developing or moving in a slow but systematic manner. One therefore suspects that whatever agency is responsible for forcing the persistent recurrence of fast-moving weather systems to produce these dominant, long-term circulation anomalies is itself capable of evolution and may slowly move or change in effectiveness with time.

Much thought has been given in recent years to what the agency or agencies might be. One expert, John S. Sawyer of the Meteorological Office, Bracknell, England, has pointed out that since friction and viscosity appear capable of wiping out the total circulation energy of the atmosphere over a period of about a week, circulation momentum can be ruled out as a generator of the climatic anomalies. Whatever is responsible, must fulfill three criteria: it must be comparable in scale with the observed climatic anomalies, *i.e.*, more than 1,000 kilometers across; it must persist for periods of at least one month; and it must be capable of giving up heat at a rate of at least 50 langleys per day (one langley representing an energy transfer of one calorie per square centimeter of surface). After reviewing likely candidates – including the extent of snow cover and sea ice, evaporation from vegetation, and anomalies of soil and sea surface temperature – Sawyer concludes that only the latter, known as SST anomalies, meet the requisite conditions.

For reasons such as these, SST anomalies are now regarded as having an important moderating influence on climate at time scales of a few months to a few years in duration, and a good deal of effort is being expended in attempts to describe the details of the coupling mechanisms. The work is being channelled along two main avenues of research, the first involving the construction and refinement of large-scale numerical models, and the second using statistical techniques to identify the principal patterns of interactive behavior in ocean and atmosphere. Both approaches have their shortcomings in the absence of physical measurements capable of describing the full complexity of the interaction processes. Yet both have significantly advanced our understanding of these processes from mere description toward the ultimate goal of prediction. This article is concerned only with the latter (statistical) technique and has the aim of reviewing some recent studies which give grounds for optimism that prediction is a realistic goal for the not-too-distant future. These studies also illustrate the rationale behind the statistical approach.

The furtherance of the statistical technique in air-sea interaction research has owed much to a wide range of illuminating studies by Jerome Namias at Scripps Institution of Oceanography. Using new basic techniques, Namias has identified many of the factors which will be important to the planning of future physical experiments – the characteristic space and time scales of change in ocean and atmosphere, the sensitive areas for heat transfer between ocean and atmosphere, and the more distant climatic effects associated with major abnormalities of the ocean temperature field.

Namias has also developed a "screening" technique for the Pacific sector which demonstrates predictive properties in its own right, and it is this method which best illustrates the mutual interdependence of oceanic and atmospheric behavior. Using twenty-year files of sea surface temperature (SST), sea level pressure (SLP) and 700 mb^2 height for five-degree intersection points over the Pacific north of 20°N, multiple regression

The monster buoy "Alpha", shown above and on page 14, during offloading operations at Pearl Harbor. The device was used during the NORPAX "Pole" experiment earlier this year. (Photo credit: U.S. Navy)

analysis is used to generate multiterm equations which express any one of these three variables (at each of 59 "predictand" points) as a function of either of the remaining variables (using 225 "predictor" points). Once generated, these equations may be used either to *specify* the geographical distribution of one variable from the distribution of another in the same month or season or to *predict* the future state of one field from the observed behavior of another. To date, Namias' experiments in specification have met with considerable success. In Figure 1, for example, the average SST anomaly patterns observed during the winters, of 1969–70 and 1971–72 are compared with those specified from the patterns of 700 mb height in the same seasons; it is clear that both the patterns and the amplitudes of anomaly are strikingly familiar in each case.

Compared with specification experiments of this type, attempts at prediction using the screening technique are at an early stage of development, and clearly there are several possible approaches to be explored. We could develop a second set of predictive equations which express each of our three variables as a function of the others in the preceding season (winter SLP as a function of fall SST, for example). Alternatively, in view of the great persistence of thermal anomalies in the ocean, it may prove more convenient, say, to *predict* winter SST from fall SST, and to use the predicted winter temperature anomaly pattern to specify the winter distribution of SLP or 700 mb height.

Namias is now working with these and other versions of the screening technique in an attempt to exploit its full potential in specification and prediction and to explore its limitations. Some of these limitations are now apparent, and others may be guessed at. For example, it is likely that the technique will be most successful *within* either the warm season or the cold season; the major disruptions of the surface temperature field that occur with the establishment or breakdown of the seasonal thermocline are likely to reduce our ability to predict events at these times of year. Secondly, as between summer and winter, it is probable that the best results will be obtained during the latter, when thermal communication between ocean and atmosphere is maximal. Finally, at any season there will be anomalous atmospheric situations (hopefully a limited number) when the normal specification or prediction techniques will break down for no apparent reason. It is already clear, for example, that the success of the screening technique in the Pacific sector during the cold season is greatly diminished when persistant blocking anticyclones (high-pressure cells) dominate the northwest Atlantic in the vicinity of the Davis Strait during the same cold season. Other similar situations probably await identification. Nevertheless, even with these limitations, the screening technique is proving to be of value, not only in the improvement of our predictive skills but also in providing a greater

[2] An upper air pressure surface (such as the 700-millibar level) is frequently used in these studies since it tends to provide a useful simplication of the pressure patterns found on the sea level charts.

understanding of the causes of persistence in the atmosphere and its relation to oceanic events.

Aside from persistence, our climatic and hydrographic records show clear evidence of a second, related factor basic to the further development of our skills. This is the phenomenon of recurrence, whereby certain patterns of behavior in ocean and atmosphere are repeated in the same seasons of successive years with much greater regularity than might be expected from pure chance. At present, although we are able to show that the ocean and atmosphere may share this type of behavior, our data base is again insufficiently detailed to describe the complex interactions involved and to provide us with a true physical explanation. Equally, with no clear idea as to the cause, we are unsure of the best location to make such measurements. It is in this type of situation that relatively simple statistical techniques may provide some initial clue as to the processes involved, and indicate where we might look and what we might measure if a physical explanation is to be achieved. In fact, the problem of recurrence offers a good example of the use of these techniques, and for this reason some of the initial results are described below.

As a first step, we may show that recurrence of oceanic and atmospheric behavior is primarily a cold-season event. Correlating the seasonal *patterns* of Pacific SST anomaly in successive pairs of years, we find that during the period 1947–70, the winter-to-winter and spring-to-spring correlation coefficients were 0.50 or greater on 11 occasions (*i.e.*, 11 pairs of seasons) while only four such cases were observed in summer and fall. (A grid of 102 five-degree squares was used in each pattern correlation.) Secondly, we may

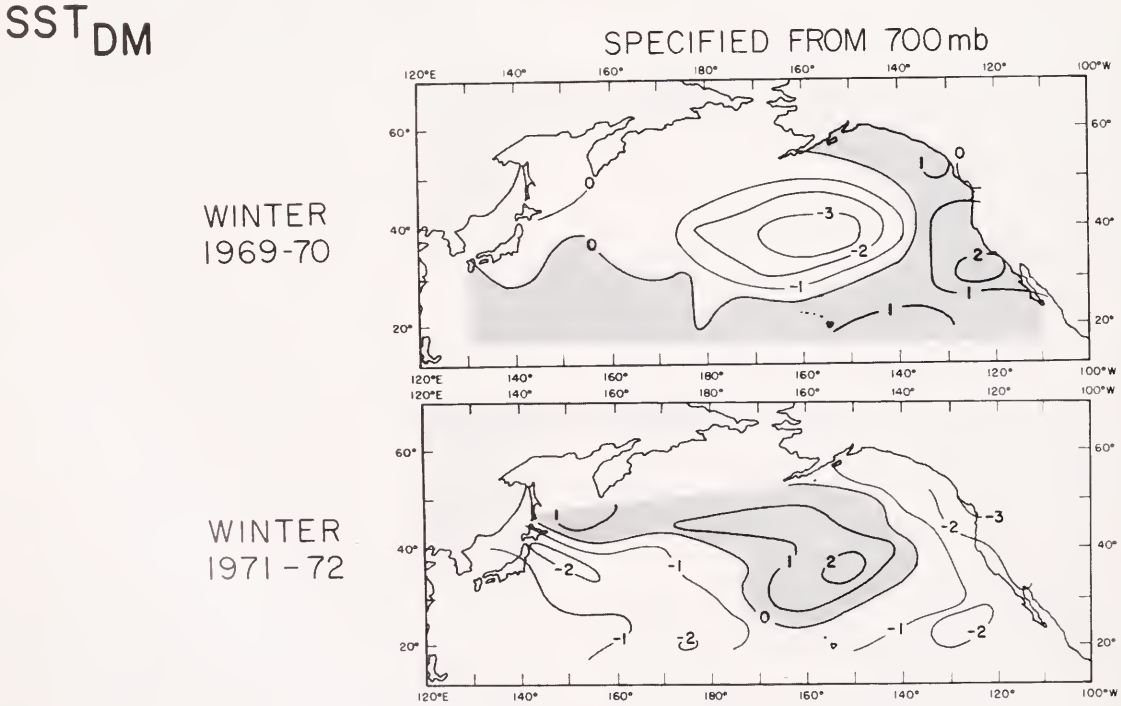
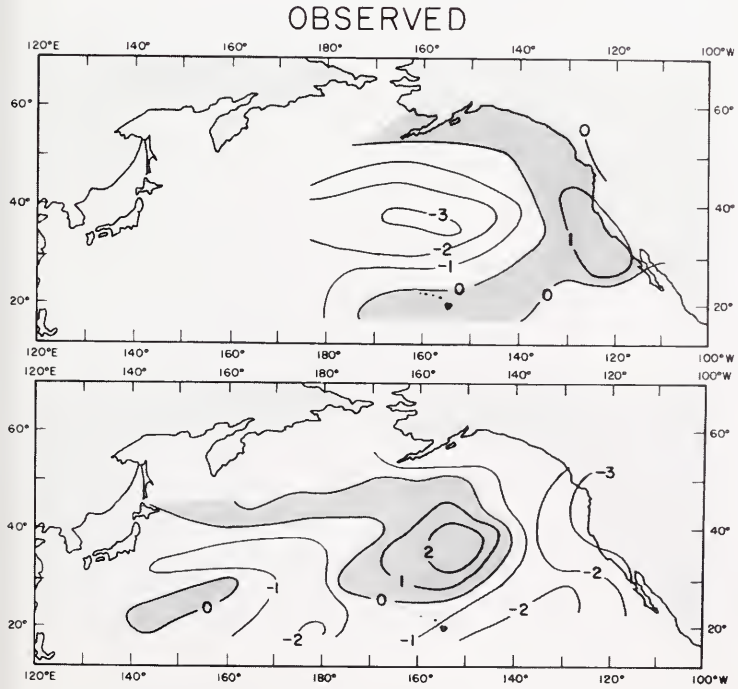


FIGURE 1: Examples of the specification technique. The charts above show the patterns of sea surface temperature anomaly (in degrees Fahrenheit) as specified from the distribution of 700-millibar height in the winters of 1969-1970 and 1971-1972.

show that SST pattern recurrence is associated not with a haphazard variety of patterns but with one particular distribution of Pacific SST anomaly in winter and spring, and with a second pattern in summer and fall. The patterns themselves are shown in Figure 2, a and b, contoured at intervals of 0.5°F. Here we have merely identified and averaged all cases where the seasonal SST fields of adjacent years showed correlation coefficients greater than 0.60 (within each season the SST anomaly patterns were essentially similar in each case).

Since the seasonal patterns of 700 mb height anomaly show much the same recurrence tendencies as those of sea surface temperature, it is perhaps expected that the causes of SST pattern recurrence are in some way bound up with atmospheric behavior, and Figure 2, c and d (due to Namias) provide some initial indication that this is the case.

For each 5° square in the eastern Pacific, Namias has correlated the 24-year time series of mean winter and mean summer SST's (1947–71) with the corresponding series of 1000–700 millibar thickness values, and the resultant correlation coefficients have been contoured and shaded wherever the correlations exceed the 5% level of significance. Since 1000–700 millibar thickness is a measure of the mean virtual temperature in the lower layers of the atmosphere, the areas of peak correlation indicate sites where the thermal states of ocean and atmosphere are in close adjustment and are held to represent regions of strong thermal interaction. At present, our actual measurements of heat exchange between ocean and atmosphere are not sufficiently detailed or extensive to confirm the validity of these inferred patterns, but the principal zones of interaction shown on these charts are at



The charts above show the temperature anomaly patterns actually observed. "DM" refers to an anomaly, a departure from the mean. (Courtesy, J. Namias)

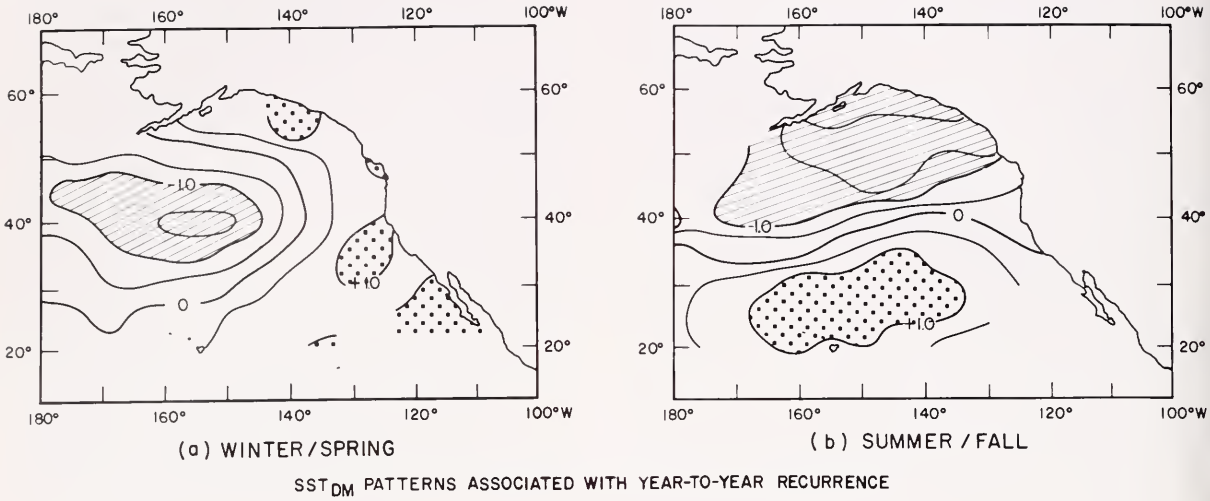
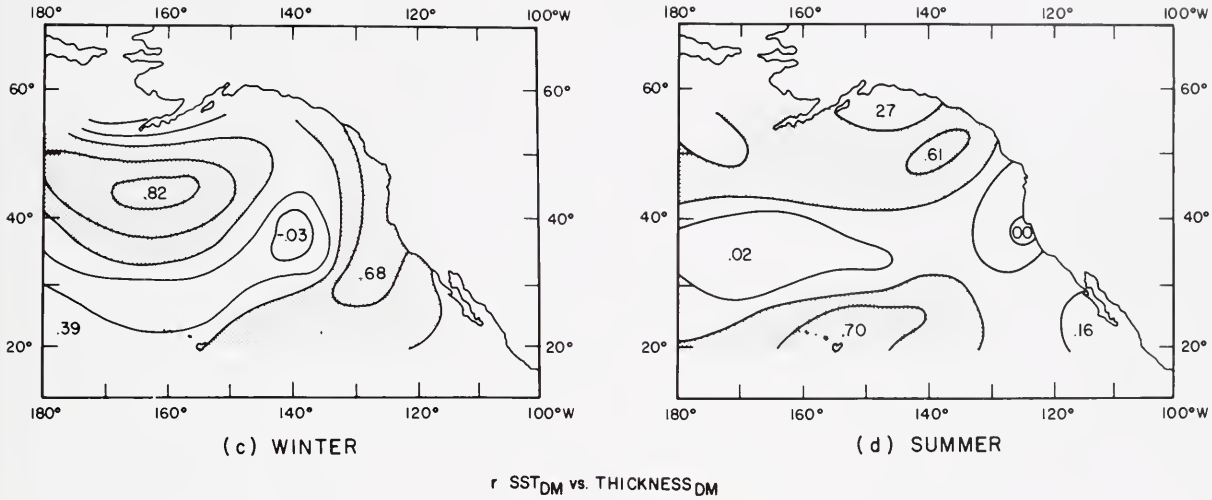


FIGURE 2: Charts a *and* b *indicate patterns of sea surface temperature anomaly (in degrees Fahrenheit) in the eastern Pacific most closely associated with year-to-year recurrence during the seasons of winter-spring and summer-fall respectively.*

least explicable in terms of known atmospheric behavior. For example, in winter, one such zone lies to the south of the Alaskan Peninsula, where outbreaks of polar air are subjected to rapid warming from below, resulting in intense convective instability and in heat exchange rates of up to 2000 ly/day. A second zone curves southwestward from the west coast of Canada to a point east of the Hawaiian Islands, and in this zone the destruction or reinforcement of the west coast inversion in winter provides the potential for major variations in heat exchange.

In the context of our discussion, the main point of interest centers on the similarity between these inferred patterns of heat exchange in winter and summer and the SST distributions most closely associated with year-to-year recurrence during the cold and warm seasons (compare Figure 2, a and c, b and d). This similarity suggests the idea that the recurrence of a particular SST anomaly pattern between the winter (summer) seasons of successive years requires the formation of an anomaly pattern which is sufficiently intense to retain its identity throughout the intervening warm (cold) season. In turn, the formation of intense SST anomaly centers requires the potential for intense heat exchange between ocean and atmosphere, and this potential for heat exchange is not ubiquitous but shows geographical and seasonal variations. By this reasoning, the SST anomaly distributions associated with recurrence might be expected to reflect these variations. A further point is also apparent from this study. The available



Charts c *and* d *show the coefficients of correlation between sea surface temperature and 1000-700 millibar thickness in the eastern Pacific during the winter and summer seasons respectively. The shaded areas represent zones where the ocean and atmosphere are apparently in close thermal adjustment.*

results show that although a given SST anomaly pattern may *recur* in the same season of successive years, it does not appear to *persist* in the surface layers of the ocean throughout the intervening seasons. The most obvious conclusion is that the "memory" for recurrence resides elsewhere, and that the subsurface oceanic layers must also be considered.

The results of this ongoing study certainly do not resolve all of the problems associated with recurrent oceanic and atmospheric behavior. Indeed they raise a number of new questions. Does the recurrence of an SST anomaly pattern in successive winters merely indicate the resurgence of a pattern which has persisted below the thermocline throughout the intervening summer? How are summer SST patterns able to recur occasionally without being destroyed during the intervening cold season? We must bear in mind, however, that these statistical techniques were not intended as a substitute for the physical approach but as a means of providing the conceptual framework on which a physical research program might be based; in this they have been successful. Within the context of the current North Pacific Experiment, for example, a number of physical experiments ("POLE", "SECTION" and "BOX") are now planned which will test and quantify the concepts developed from statistical studies such as those discussed above. To many, it is this type of dual approach which would appear to offer the greatest potential for advancing the science of air-sea interaction from the empirical to the physical levels of understanding.

The Complex Killer

by R. H. Simpson

Robert H. Simpson received his Ph.D. in meteorology from the University of Chicago. He has just retired as Director of the National Hurricane Center of NOAA in Miami and has accepted a position as research professor in the department of environmental sciences at the University of Virginia.

For more detailed reading on this subject: Dunn, Gordon E. and Banner I. Miller, Atlantic Hurricanes. *Louisiana State University Press, 1960.*

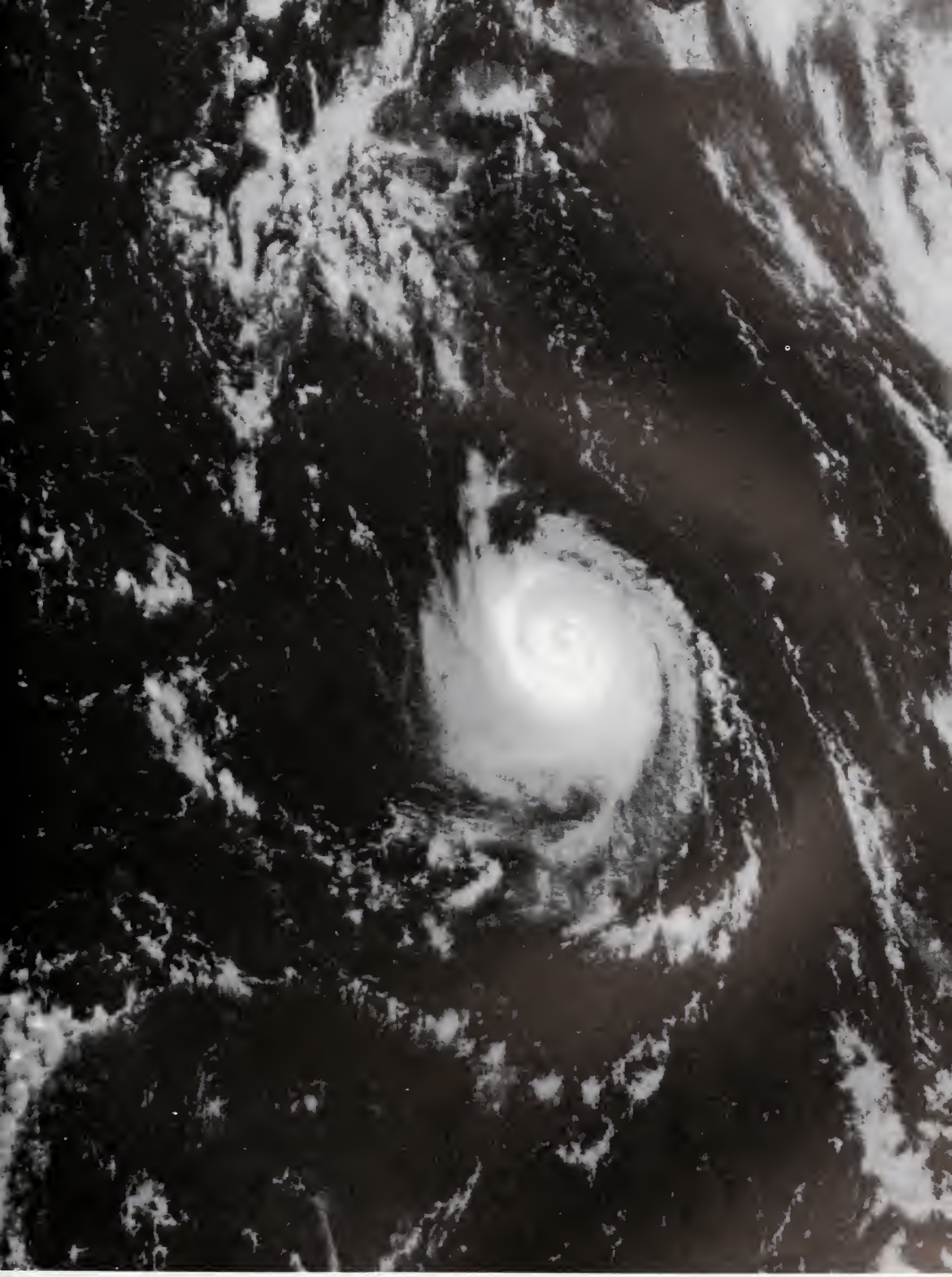
For some years following the turn of the century, scientists alluded to the hurricane as one of the simplest of atmospheric storms – a circular, symmetric circulation of an ideal vortex in a homogeneous (barotropic) environment. Aside from its awesome destructiveness it was felt by many to be of only trivial academic interest.

Today, after several decades of observing the hurricane from research aircraft, after nearly a decade of monitoring its environment with weather satellites, and after numerous experiments with numerical simulation models of its circulation, scientists have established the hurricane as one of the more challenging problems in meteorology. It is *not* a simple or ideal vortex. Its circulation properties involve important asymmetries and – perhaps most important of all – its destiny is clearly linked to the intimate coupling of the atmospheric circulation to the ocean from the sea surface through the thermocline layer. It is certainly the most dramatic, and probably the most intricate, example of air-sea interaction that has been observed.

The couple between ocean and atmosphere is less important in the development phase of the tropical cyclone than it is in the maturing process. Of the 100 tropical disturbances, or hurricane seedlings, which march across the tropical Atlantic each year with remarkable dependability, only one in ten develops sufficient strength to sustain gale force winds, and the total number achieving this level varies by a factor of two or three from year to year.

These seedlings are clusters of cumulonimbus clouds or cloud lines, each living and breathing independently as regards the exchange of mass with the environment. Under favorable circumstances, they may be transformed into an atmospheric heat pump in which a number of penetrative cumulus are persuaded to share a common large-scale spiraling inflow at cloud base level and a common outflow at the cloud tops. The conditions for effecting this transformation appear to depend mainly upon the atmospheric environment; the ocean environment plays only a passive role, namely that of supplying surface temperatures high enough to support convective mixing in the lower troposphere. Sufficient fuel, in the form of latent heat of condensation, is available from the influx of moist tropical air to place the heat pump in operation, to cause surface pressure in the core of the disturbance to fall 10–15 millibars, and to support sustained winds of gale force.

It is at this point, as the organized winds of the tropical storm begin to rough up the seas and disturb the thermocline, that coupling of the atmosphere and seas plays a determining role in the destiny of the tropical cyclone. The storm system cannot acquire hurricane strength and contract the circle of maximum winds tightly about the storm center unless the heat content of the in-spiraling tropical atmosphere is augmented by 12–15%. And this augmentation can only come from a flux of



Hurricane Ginger, 1971: high-resolution picture of eye, rain clouds and cirrus clouds just after seeding. (Photo credit: T. Fujita, University of Chicago)

heat directly from the ocean reservoir. That this can and does occur has been established by innumerable flights through hurricanes in highly instrumented research aircraft at elevations ranging from a few hundred feet to the top of the storm.

The establishment of this energetic coupling is of course dependent upon sea surface temperatures sufficiently warm to enable the flux to occur effectively. However, once it is established, a subtle but important dynamic coupling ensues which apparently involves a substantial reordering of the thermocline layer beneath the hurricane. Interesting observations of cold water in the wake of hurricanes in the 1950's led to a succession of oceanographic surveys in the 1960's to study thermocline anomalies induced by hurricane wind stresses. The results, reported in the work of Leipper and of Reid at Texas A & M, laid the groundwork for many theoretical investigations. Among these was an important study by J.E. Geisler at the University of Miami proposing a mechanism by which the dynamic sea-air coupling in a moving hurricane can produce a hierarchy of thermoclinal disturbances, including a condition for upwelling, which relate not only to the wind stresses on the sea but may vary explicitly as a function of speed of the hurricane center. Hurricane data from airborne expendable bathythermographs, studied recently by Peter Black of the National Oceanic and Atmospheric Administration (NOAA), support the theoretical finding of Geisler and graphically demonstrate the extent of the dynamic coupling between the hurricane circulation and the ocean thermocline.

The thermocline anomalies in the wake of the hurricane have been observed to persist at times for weeks and pose a broad spectrum of interesting questions and problems in oceanology. However, still another aspect of sea-air coupling is of more practical and dramatic importance. This is the generation and maintenance of the potential for storm surge, the phenomenon that brings disastrous inundation and is responsible for most deaths along coastal areas as a hurricane moves inland. The potential for the storm surge is generated over the open seas by a unique and intricate combination of the so-called "inverted barometer" – a hydrostatic rise in sea level due to the atmospheric pressure drop in the hurricane – and the large-scale swirl or vorticity of the water generated by the stresses of the in-spiraling winds on the water surface.

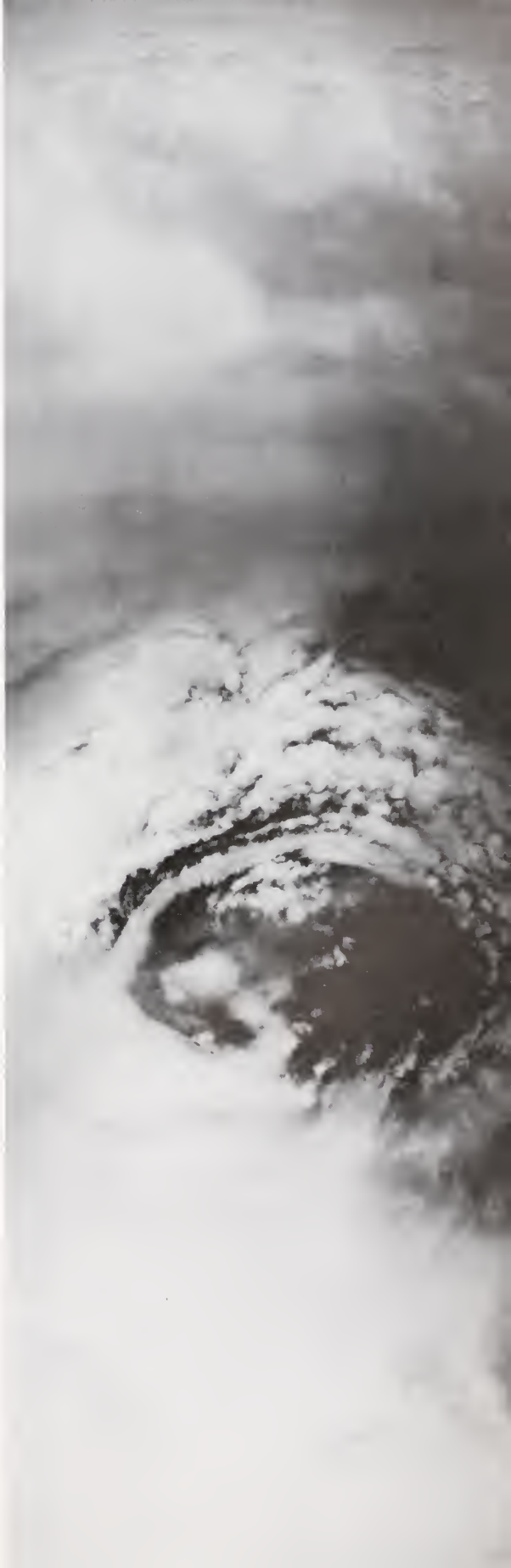
In deep water the uplift of the sea surface is measured only in tens of centimeters over a distance of kilometers. Due to the assymetries of the wind stresses, the maximum vorticity or swirly transport of water occurs to the right and abreast of the moving storm center. In deep water this vorticity extends its influence deep into the thermocline to trigger internal waves and the thermal adjustments described by Geisler, but it does not contribute to a significant rise in sea level until the system approaches shoal water. When water depths decrease to 50 fathoms or less, the deep column of swirling water placed in motion by the wind stresses begins to scrape bottom, and the column tends to shrink. Since vorticity is a conservative property, there must be either a divergence of mass or an uplift of the sea surface. As bottom friction retards divergence, the dramatic result is an uplift of the sea surface, which, combined with the inverted barometer effect, reaches a peak height to the right of the moving center near the position of maximum sustained winds. The mound of water which is generated in this manner is what is known as the storm surge.

Thus, a hurricane arriving at an island where deep water extends virtually to the shoreline does not produce significant inundation due to storm surge. However, when it passes over an extensive shelf of water with depths of less than 50 fathoms, surge height at the open coastline may reach more than five meters. In 1969, during Hurricane Camille, an extreme case, the peak surge reached nearly eight meters.

In the last five years, great progress has been made in the modeling and numerical

simulation of the complex interactions in the sea-air couple that generates storm surge. A prognostic model developed by Chester Jelesnianski of NOAA has not only become a primary tool in predicting coastal inundations due to hurricanes, but has explained the large variability in inundation from hurricane to hurricane, and the conditions which may cause two coastal sectors equally vulnerable to hurricanes to have quite different vulnerabilities to storm surge inundation. This model, known as SPLASH (Special Program to List the Amplitude of Surge Heights), presents the profile of storm surge at an open coast as a function of a coastal shoaling factor, premapped and stored in the program, and of the hurricane characteristics which comprise the input data. The model does not deal with the more complex problem of distributing the surge heights over complex embayments and estuaries or the distribution of the water inland. Such special cases as estuaries can be handled only by models tailored to the three-dimensional bathymetry of the individual estuary.

The SPLASH program has two parts: SPLASH I, applied to hurricanes which cross the coast at a known angle; and SPLASH II for hurricanes which parallel but do not cross the coastline. The model demonstrates that the storm surge height at an open coastline will vary as a function of: 1) the central pressure of the hurricane; 2) the shoaling factor or seaward extent of shoal water from the coast; 3) the radial distance from the storm center at which the maximum wind occurs; and 4) the speed at which the hurricane center approaches the coast. For maximum storm surge heights, shoal water must extend seaward for a distance not less than the radius of maximum winds (*i.e.*, radial distance from storm center to the circle of maximum winds). This means that while an extreme hurricane such as Camille can generate a storm surge of nearly eight meters on the



Hurricane Betsy, 1965: photo was taken by reconnaissance aircraft at an altitude of 11 miles north of Grand Turk Island in the West Indies. (Photo credit: U. S. Air Force)

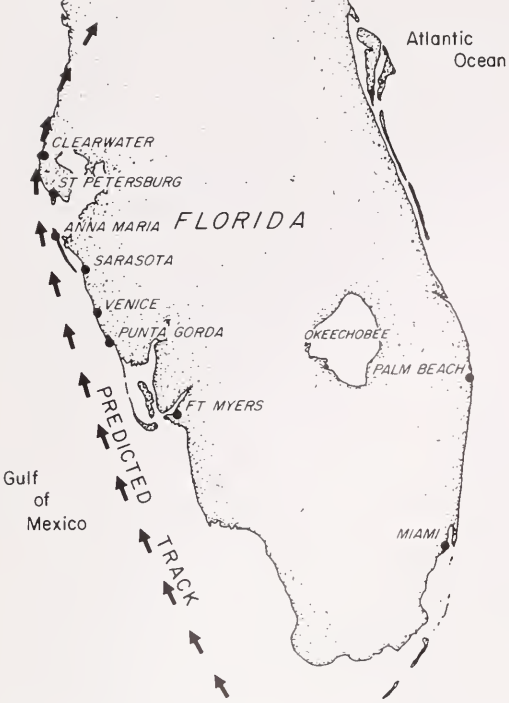
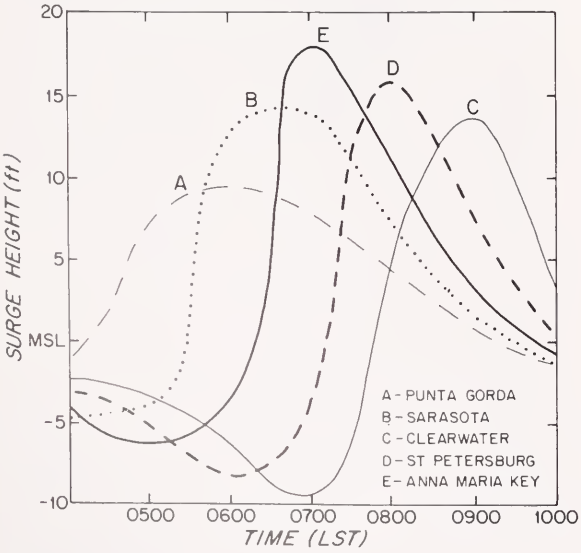


FIGURE 1: When hurricanes move parallel to and near a coastline to the right of the storm track, coastal residents may experience the storm surge after the center passes abreast. With tides substantially below normal as the center approaches, water levels may rise rapidly – a foot per minute or more as it departs.

Storm profiles as a function of time are shown below for a series of stations on the Florida west coast marked on map. These are computed by SPLASH (see text) for the path which would have been followed by Hurricane Donna (1960) if it had entered the coast in accordance with the track predicted at the time Donna was approaching the Florida Keys. Donna actually crossed the coast south of Fort Myers.



Mississippi coast, a storm with the same characteristics moving inland along the mid-Atlantic coast – where the shoaling factor is much less – would be expected to generate little more than half that surge height. Conversely, the same storm crossing the broad shoals off St. Marks, Florida, could generate an even higher storm surge.

SPLASH states that a hurricane crossing the coastline at, say, 12 knots and producing storm surge of four meters would produce a surge of five meters or more (other factors being equal) if its transit speed across that same coast were increased to 18 knots. It also states that if maximum winds occur at a greater radial distance from the center, the storm surge peak will increase slightly, and the lateral scope of inundation will expand. This particular factor raises an important question concerning the net impact of modifying hurricane wind strength by cloud seeding methods, a subject to which we shall return later.

SPLASH II provides a number of surprising possibilities for inundation from storms moving parallel to the coast. This simulation shows that for hurricanes moving "up the coastline" (*i.e.*, with the coastline to the left of the track) the inundation potential remains minimal, while for a hurricane moving down the coastline (the coast to the right of the track) and at a considerable speed (20 knots or more), the rise in water levels at any open coast occurs with great rapidity and follows passage of the storm center, with deceptively low water before the center arrives. In this case the water levels at the coast may increase at the rate of more than a foot a minute. Finally, the SPLASH formulation demonstrates that in some embayments, and even over open shoal water, it is possible for the hurricane to move at a resonant speed in relation to the depth of underlying shoal water; this may trigger a series of seiche-like waves, which spread outward rapidly, bringing inundation to the coastal plain in sudden increments.

As our knowledge of hurricanes expanded, considerable interest centered around the

possibilities of using cloud-seeding procedures to alter some aspect of the energy releases in the hurricane to reduce the destructive capacity of the system. The national project known as STORMFURY sponsored by NOAA with substantial assistance from the Department of Defense, has been dedicated to this objective for more than a decade. Underlying the work is the knowledge that in the hurricane, as in many other severe storms, nature is not very effective in releasing latent heat of fusion in supercooled clouds, and that the introduction of silver iodide crystals into supercooled clouds will cause freezing and the release of latent heat of fusion. Numerical simulation experiments support the hypothesis that strategic seeding in the eye wall radially outward from the point of maximum winds can cause the eye wall to migrate outward or reform at a greater radial distance from the center. As a result, the in-spiraling winds in the lower layers, accelerating towards lower pressure, cannot move as close to the pressure center and do not acquire as great tangential speeds before reaching the eye wall (where acceleration ceases).

There has been evidence from experiments in each of four hurricanes seeded by Project STORMFURY (Esther, 1961; Beulah, 1963; Debbie, 1969; and Ginger, 1971) that the adjustments in the wind system observed after seeding followed the hypothesis. While it has not been established that these adjustments could *not* have occurred naturally rather than in response to the seeding, the results from the 1969 Debbie experiment, where five successive seedings at 90-minute intervals were conducted, bordered on the spectacular. In this case, the maximum winds after the fifth seeding were observed to be less than 70% of those measured just before the seeding began. Since the wind force, as a measure of destructiveness, is proportional to the square of the wind speed, this reduction in maximum winds represents a reduction of more than 50% in the static wind forces at the point of maximum force. However, viewing the experiment in the context of a sea-air coupled system, a few additional questions need to be asked to evaluate the operational implications and net benefits to be expected. The sea is at least as important as the wind in its destructive potential, clearly more so in the destruction of life. One then must ask how the adjustments in the hurricane circulations predicted by the hypothesis will change the storm surge potential within the hurricane. First, the hypothesis predicts an increase in radius of maximum wind with little or no change in central pressure. Other things being equal, the simulation model would indicate that the storm surge would slightly increase due to the seeding. However, the hypothesis anticipates a distinct drop in maximum winds, which should reduce the height of the surge.

Other things are rarely equal, however, and at this juncture we do not know enough about the interactions between the various adjustments which might occur due to seeding to predict whether the procedure would produce any significant change in the potential storm surge. Perhaps this enigma best attests to the intricacy of the coupling which links the ocean and the atmosphere in the hurricane.

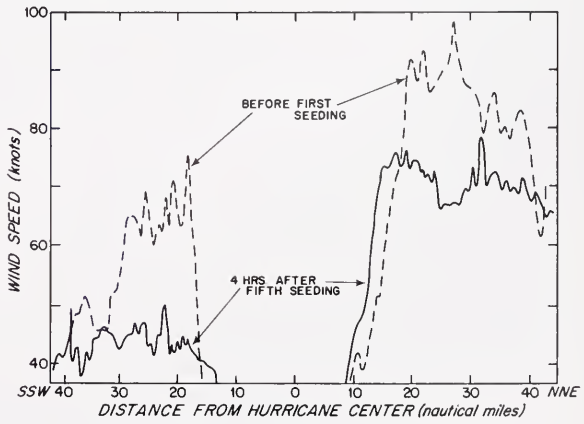


FIGURE 2: (After Gentry) Wind speeds in Hurricane Debbie (1969) reached a maximum of 98 knots before the first of five seeding flights spaced ninety minutes apart. Four hours after the last seeding, the highest speed was 78 knots; there was significant general reduction in winds throughout the storm core. Southwest of the storm center, winds were all less than 50 knots, although hurricane-force winds had extended outward 30 miles from the center before the seeding.

"The Wind blows, The Waves come"

by F. W. Dobson

Fred W. Dobson is a research scientist in the air-sea interaction group at Bedford Institute of Oceanography where he has been conducting his research since 1969, when he received his Ph.D. in oceanography and physics at the University of British Columbia.

A young oceanographer studying the problem of wave generation had an old uncle, a retired Yarmouth sea captain, who had made his start as a cabin boy in his father's bark.

Hawaiian Surf (Photo credit: Ron Church, Photo Researchers)

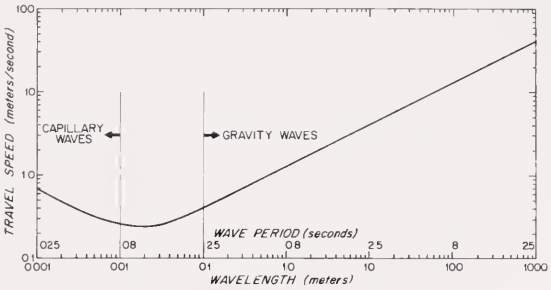


FIGURE 1: Deep-water waves have a nonlinear dispersion relation between wave length and speed of travel. Logarithmic scales show the complete wave length range of sea waves, from one millimeter to one kilometer. At right are the long, fast-moving "gravity" waves of the open sea, wind-driven waves whose lengths commonly are in the 100-400 meter range and whose speeds most often run between 25 and 50 knots. At left are the short "capillary" waves, which roughen the surface to form "catspaws".

When the two got together, the talk usually turned seaward. The old man would often ask, "Nephew, tell me again what it is you are doing." The nephew would reply, "Uncle, I am trying to understand how the sea waves grow as the wind blows over them." The uncle would look puzzled, and the reply would come, getting ever testier with repetition: "But boy! The wind blows, the waves come. What more is there to bother about?" There are two answers to the question (neither one would have satisfied the uncle one whit). First, wave generation remains a mystery, a fascinating hydrodynamical problem. The second answer, better suited to proposals for funding, is that waves do interfere with man's passage on the sea (any true oceanographer will attest to that), and, because their growth process is not understood, their occurrence cannot be well predicted.

The physics of gravity-capillary wave propagation on the surface of a liquid was worked out in the late 1800's by Green, Airy, Stokes, Rayleigh, Helmholtz, and Kelvin. Water waves have a 'dispersion' relation – that is, a relation between wave length[1] and speed of travel – which is not linear (Figure 1).

[1] Wave length is the horizontal distance from crest to crest or hollow to hollow.

The winds that generate the waves have a structure, too. They are invariably turbulent – gusty, in other words. The wind speed is low near the sea surface and increases with height; if the surface underneath is rough, this increase is more gradual than it is when the surface is smooth. Except for a viscosity-dominated region within a millimeter or two of the sea, the increase is approximately logarithmic: wind speed at any height is proportional to the logarithm of the ratio of that height to some reference height, commonly called the 'roughness height' in the jargon of the trade.

Because the wind speed varies with height, there is always a height (a 'critical level') below which the waves are actually moving faster than the wind. In our wind profile, for example, the critical level for a component of the wave field of a length of 25 meters, which travels at a speed of 6 meters per second, would be one centimeter. An understanding of the air flow in the immediate vicinity of this critical layer is crucial to an understanding of the generation process for gravity waves at sea.

One of the interesting points about the air flow is the difficulty that has been encountered in finding wave-induced effects in measurements made over the water. The wave-induced wind fluctuations are so small that they are completely masked by the wind turbulence itself. This makes things difficult, since it means that sensors must be placed within centimeters of the sea surface before wave-induced air speed fluctuations can be measured. No one has yet devised a wind speed probe that can give reliable results when it is fixed at a level so low that it is often struck by the waves. It is from this problem that wave followers have sprung; the sensor must be mounted on a wave-following device if it is to stay close enough to the surface.

As one travels seaward from a coast in an offshore wind, the composition of the waves changes with distance from shore (fetch). Nearest the shore there is nothing but ripples. With increasing fetch, wave height

and wave length increase. At any given fetch the highest waves, the 'dominant' waves, are those with the longest wave length. At long fetches the waves become enormous rolling hills of water, moving at high speed. There are many records for high waves: off the mouth of Halifax Harbour, our accelerometer buoy measured one 45 feet in height going by at about 40 knots; the North Sea oil rigs have reported wave heights in excess of 100 feet. When the speed of the dominant waves reaches the wind speed, the sea is said to be 'fully developed'. If the growth of a given wave length (that is, of one component of the wave spectrum) is followed as it travels downwind, an interesting result emerges. Near shore the waves are very small and grow slowly, more or less linearly, with fetch. At some 'critical fetch', they suddenly begin to grow very quickly – in fact, exponentially – with fetch. Their height increases in this manner until it reaches a 'saturation' level – and then *decreases,* finally reaching an 'equilibrium' level beyond which there is no further growth. Why the wave components 'overshoot' their final equilibrium value is a fascinating question as yet not completely understood. The final composition of a fully developed sea depends on the wind speed, with the longest and largest wave components being those whose speed approximates the mean wind speed.

There is no paucity of hypotheses for explaining the generation and growth of sea waves. I will describe only those which seem the most promising and go on to discuss recent efforts to sort out the experimental evidence.

The only theory to be discussed here that deals with wave generation starting with a

Wave-follower developed by the Bedford Institute of Oceanography to measure fluctuations of wind, temperature, and pressure at two heights. Mast is attached to large undersea tower in North Sea. Sonic anemometer is on top. Below and to the left is vertically mounted wave-measuring staff. The two bomb-like devices are pressure sensors. Wave follower is the large cylinder projecting from water (it is lowered to make measurements). Measurements were recorded aboard ship in background, VWS Atair, *tethered to instrument by signal and control cables.*

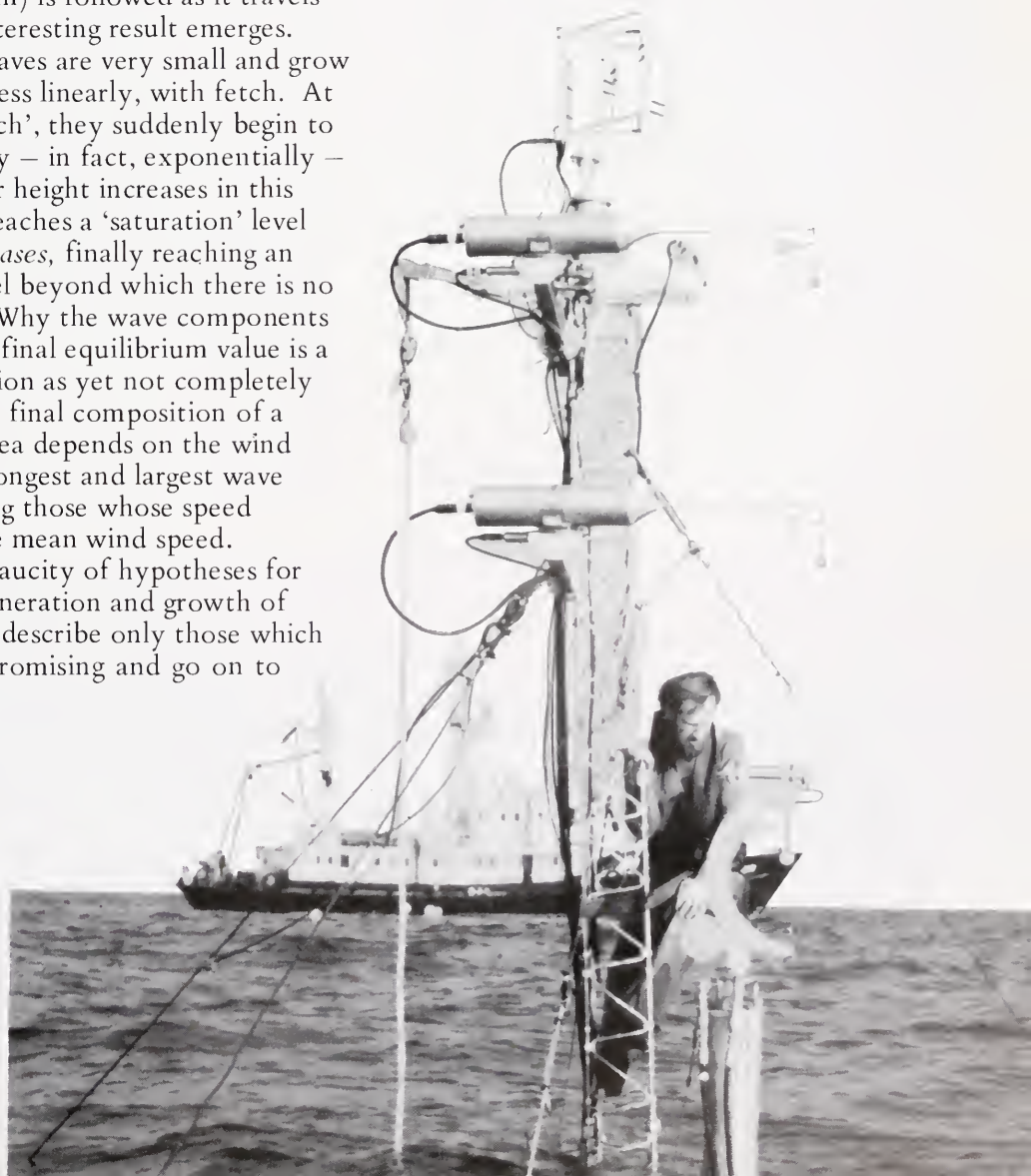


FIGURE 2: Typical vertical profile of mean wind speed over the sea. Wind speed varies little above one meter, enormously below.

smooth surface is one developed by O. M. Phillips at Cambridge. Turbulent eddies in the air stream are carried along at about wind speed. Associated with the eddies are pressure fluctuations which, though random in nature, retain their characteristics for a finite length of time. It is these moving pressure fields which generate the waves. The energy of each wave component increases linearly with time. A given wave is excited if it moves at the same speed as the pressure field – that is, a given moving pressure field will produce a wave-pattern similar to a ship's wake, with the relatively slow (short) waves traveling at small angles to the wind, and the longer, faster-moving waves traveling at angles which make their downwind velocity component equal to the wind speed. Because pressure-producing eddies exhibit a broad range of sizes, they generate waves over a broad region of the spectrum. Observed linear growth rates fit the predictions of the theory tolerably well.

Another way in which waves may grow is via interactions among themselves. This subject has been extensively but separately treated by O.M. Phillips and by K. Hasselmann of the University of Hamburg. An existing wave field consists of a mixture of components of different wave lengths, and these components interact to produce energy at other wave lengths. The latest field results indicate that such interactions can cause energy transfers which are as large as or larger than the energy input from the wind. The theory apparently has been successful in explaining the 'overshoot' phenomenon. An easily observed example of interactions of waves of different lengths is the production of capillary waves on the forward face of longer waves which are about to break. The energy transfer can go from short waves to long ones, too, and in fact acts over the whole spectrum of wave lengths.

We now turn to theories involving positive 'feedback'. In these theories the process causing the waves to grow is linked with the growth itself, so that as a wave increases in height it increases its capacity for further growth. This leads to an exponential increase of wave energy with time.

One such theory is the 'separation'

hypothesis of Sir Harold Jeffreys, published in 1925 and considered for 30 years thereafter to be the final answer to the wave generation question. According to Jeffreys, the air over the waves stalls at the wave crests – that is, flow separation occurs there, causing zones of low pressure just downwind of the crests. These low pressure regions work on the waves, causing them to grow. The theory was debunked by F. Ursell in 1956 and went completely out of fashion. It has now been resurrected to explain measurements of wind speed fluctuations made in laboratory tunnels, where the wind speed is much greater than the wave speed. It may explain some of the growth of short waves in the presence of strong winds (ten or more times greater than the wave speed).

Another feedback theory, due to John Miles of Scripps Institution of Oceanography and published in 1962, acts in the same regime as does flow separation – high winds, short, slow waves. When the wind speed is so high that the 'critical height' is within the viscosity-dominated region of the flow (within millimeters of the sea surface), a certain class of resonance occurs between already-present waves in the air flow and water, which causes short waves to grow exponentially with time; the predicted growth rates are relatively large but drop off quickly with decreasing wind-to-wave-speed ratio.

A third feedback theory, also due to John Miles (it was published in 1957), is probably the most important for the growth of large gravity waves at sea, and I will confine further discussion to it and its successors. The existence of a logarithmic wind profile, maintained by the turbulence, leads to an instability in the strongly sheared flow at the 'critical level' where wind and wave speeds are equal. At this level, the air transfers energy (and hence horizontally-directed momentum) to the waves by means of the working of pressure forces. The rate of energy transfer is proportional to the existing wave height, and so exponential growth ensues. The pressure forces, normally low over the wave crests and high over the troughs, are shifted downwind by the interaction of the wavy surface below and the air flow above, so that work is done on the waves. The theory is 'tight' – that is, it has no free parameters and therefore is ideally suited for testing. The tests, however, have been far from simple to perform, and results to date have been contradictory.

Miles' theory states that the wind transfers its energy to the waves through the action of pressure forces, which are produced in the air by the waves themselves. These wave-induced

Jan Hahn

pressures are predicted by theory, and so measurements of pressure over waves provide not only a direct measurement of wind-induced growth rates but also a direct comparison with theory. Further, wave-induced effects in the pressure field are larger than, and therefore stand out above, the turbulent 'background' pressure fluctuations up to large heights. This makes it possible to use pressure sensors to study wave-induced flow structure at heights where the sensors are not too likely to be wiped out by an extra-large wave – a practical constraint not to be overlooked by the experimentalist.

The problem was that until recently (1972) no one had developed a pressure sensor which would adequately measure small fluctuations in air pressure without introducing significant errors caused by its own distortion of the air flow (by the Bernoulli effect). J. Elliott of the University of British Columbia has now developed such a sensor, and so, independently, has R. Snyder of Nova University. They have used these sensors to provide a first look at the pressure field over wind-generated waves, and the results are interesting – but more on that later.

Recently, experimentalists and theoreticians alike have been preoccupied with the problem of explaining the discrepancies between theory and experiment. Vertical profiles above the waves of fluctuations in pressure and wind velocity, measured both in the field and in wind-wave tunnels, always differ in some significant way from the Miles theory, and also among themselves. Perhaps the most perplexing anomalies exist in the measurements of wave growth rates. Direct measurements of waves in their exponential growth phase give rates of growth ten times those predicted by Miles' theory. Measurements of wave-induced pressures, on the other hand, give growth rates which variously are ten times greater than, are twice as large as, and are slightly less than, the Miles' predictions!

All of these measurements indicate in one way or another that Miles' theory falls short of describing the wave generation process. While the measurements were appearing, W. C. Reynolds of Stanford, R. Davis at Scripps, M. J. Manton at UBC, R. Long at Nova University and A. A. Townsend of Cambridge have produced wave generation theories having one thing in common: they begin with the Miles' model as an approximate solution, and then use various relations between the turbulent-wavy part and the mean part of the flow over the waves to solve more complete equations, which include interactions between the turbulent and wavy motions in the air. They all predict one or another of the observations tolerably well. In fact, it is difficult to choose among them.

In 1968 and 1969, in an international wave-measuring experiment called JONSWAP (Joint North Sea Wave Project), measurements were made with a variety of wave sensors of the variation in time and space of a field of growing wind waves. The site was a 150-kilometer line running ENE from the island of Sylt, just south of the western end of the German-Danish border. The JONSWAP results, through some rather complex inferences, provided an estimate of the fraction of the total downward transfer of horizontally directed momentum which was being absorbed by the waves. The fraction turned out to be quite large, typically 80%, and this agreed with my findings, obtained from growth rates due to the working of pressures on the waves. An estimated 90% of the energy input from the wind was found to be transferred from the growing waves in their 'overshoot' phases to shorter wave lengths, and 5% to longer wave lengths via wave-wave interactions. The 5% of the wind input transferred to longer waves was found to go directly into the waves which were in their 'exponential growth' phase, and accounted for a whopping 80% of their growth rate! At long fetches, less could be said about wave development; the *minimum* ratio of wave-induced to total downward transfer of horizontally directed momentum (wind stress) was 0.1 for large fetches.

In the fall of 1973, a second JONSWAP experiment was run, this time with the

hope of obtaining measurements of the energy input from the wind to the wave field using pressure measurements from two wave followers (one of the Bedford Institute of Oceanography and another from the University of Florida), as well as estimates of wave dissipation and a more extensive set of wave measurements, including aircraft profiling data. The BIO device is shown on Pg. 31. Unfortunately, our first try at an intercomparison of wave growth measurements from pressure sensors failed because of a combination of poor weather and instrumentation difficulties. In spite of our problems JONSWAP II has given us a considerable amount of data to analyze. We are embarking on further experiments, during which we hope to make a detailed study of the structure of the pressure and wind speed fields above the waves. The first hurdle – the development of a measuring device – is now passed, and the really exciting part – making the measurements and interpreting the results – is under way.

Wolf Rock Lighthouse off Lands' End on England's southern coast. (Photo credit: UPI)

ICE

OCEAN, ATMOSPHERE

by Kenneth L. Hunkins

Kenneth L. Hunkins received his Ph.D. in marine geophysics from Stanford University in 1960. He is now an adjunct professor at Lamont-Doherty Geological Observatory at Columbia University.

For further reading in this field: Sater, John (ed.), The Arctic Basin, *The Arctic Institute of North America, Washington, D.C., 1969.*

A film of floating ice made up of floes ranging from meters to kilometers in diameter covers most of the Arctic Ocean. This ice cover, which is only two to three meters thick, is in continual motion, drifting with the winds and currents. As the floes drift, they grind together, buckling into pressure ridges, and crack apart, leaving leads of open water Although movement may appear erratic over short distances, there are general patterns to the drift streams involving a circular gyre on the Alaskan side and an exodus into the Greenland Sea from the European side.

Sea ice is remarkably sensitive to climatic conditions, varying over a year from almost total coverage of the Arctic Ocean in the cold months to about 60% coverage at the end of the summer. During the same period, the pack loses about half a meter in thickness, which it regains again the following winter.

The polar regions are the heat sinks for the earth's thermodynamic engine; heat is transported by the global ocean-atmosphere system from the warm tropics to the cold polar regions. Sea ice is the most important factor governing the intensity of the polar heat sink. Nearly three-fourths of the incoming radiation from the sun is reflected by the ice, while only about one-tenth is reflected from open water. The ice cover thus drastically reduces heat exchange as well as water exchange between ocean and atmosphere: in summer, its high reflectivity inhibits warming of the Arctic Ocean; in winter its insulation effect is such that during the month of January heat loss from the ocean to the air is small, even though there is a temperature difference of 40°C across the few meters of ice separating air and water.

Sea ice beginning to form. (Photo credit: Jan Hahn)

The fragility of the ice, combined with its variations in extent and its importance in the thermal forcing of the ocean-atmosphere system, give it a large but not yet readily predictable role in shaping the climate of the arctic regions. Further, some have suggested that arctic pack ice exerts a significant influence on global climate, its presence or absence being linked with oscillations between ice ages and interglacial periods. According to some theories, the ice-atmosphere-ocean system is inherently bistable – able to maintain either icy or ice-free conditions over time. Others suggest that the present ice pack is stable: if it were to be removed by some means, natural or artificial, it would refreeze again. These strong differences of opinion highlight our present lack of knowledge of fluid earth processes in the polar regions.

Increasing our understanding of these problems is the goal of the Arctic Ice Dynamics Joint Experiment (AIDJEX), a seven-year program which began in 1970. The concept of AIDJEX grew out of the realization that the isolated drifting research stations, which both the United States and Soviet Union have maintained in the Arctic Ocean over the past quarter-century, were not adequate for answering questions about large-scale sea ice deformation. An array of stations would be needed to measure the strain of the ice pack under the forces of winds and ocean currents.

Arctic Institute of North America

AIDJEX

Drifting sea ice separates to form leads and grinds together, buckling into pressure ridges.

AIDJEX is a cooperative effort of research groups from universities, government agencies, and industry, coordinated by a central staff with headquarters at the University of Washington. An information bulletin is published at irregular intervals, and a data bank is maintained in Seattle, as is the primary numerical modeling effort. Funding comes primarily from the National Science Foundation and the Office of Naval Research, with additional smaller amounts from other government agencies. In addition to the United States, Canada has participated heavily in both preliminary scientific studies and logistic operations. Japanese scientists have taken part in a pilot program, and the Soviet Union has sent visitors to a pilot station.

The Arctic Ice Dynamics Joint Experiment will be integrated with two other large-scale experiments to monitor the dynamics of global circulation. These are the forthcoming Global Atmospheric Research Program (GARP) with tropical "experiments" in the earth's heat source region (see page 9), and the Polar Experiment (POLEX) put forward by the USSR as a comprehensive program for monitoring ocean-atmosphere interaction over the polar heat sinks.

The main AIDJEX project, which begins in early 1975 and lasts for one year, will use an array of manned and unmanned stations to monitor the relationship between the stresses exerted on the top and bottom of the ice and its subsequent movement. The positions of the stations as well as the atmospheric and oceanographic forces on them will be determined in order to find, among other things, an empirical relationship between the forces and movements in pack ice. The field data will be tested concurrently against numerical models incorporating various factors which govern behavior of this complex material. In final form, the new information should shed light on a number of unknowns which cannot be measured directly. One such is the "internal ice stress", produced in the pack ice when wind and water stresses vary from place to place. The "internal ice stress" occurs only in fields of pack ice, not in single isolated floes, and its description should be one of the most important results of AIDJEX.

Another objective of the AIDJEX program is the description of wind and water stress in terms of simple parameters which can be incorporated into computer models. The nature of ridging and cracking has been investigated in the pilot experiments. There is an interaction between the stress fields and ridges: winds can produce ridging and increased roughness, which in turn increase the wind stress. The limits of this feedback process are still under study. The ultimate goal is to develop a model of sea ice in which simple observations of position and barometric pressure from a future grid of telemetering buoys on the ice will be adequate for predicting the behavior of the ice.

Beyond these goals, the unique conditions in the Arctic present an opportunity for fundamental oceanography and meteorological experiments. It was in the Arctic Ocean that Nansen, in his expedition on the *Fram*, (1893–96) first observed that ice drifts to the right of the wind direction. These observations stimulated Ekman's theory of boundary layers in which both friction and the earth's

rotation are important (see page 5). The theory is still one of the cornerstones of oceanography. Internal waves were also first observed by Nansen on the same expedition. More recently, detailed observations of turbulence, microstructure and eddy motions have all been made possible by the ice platform from which instruments may be suspended without the interference of wave action. It seems reasonable to expect that future observations in the Arctic Ocean will provide further insight into basic oceanographic processes. In the atmosphere conditions are also appropriate for fundamental experiments. The upper ice surface is one of the earth's largest level surfaces on which roughness does not vary with wind speed, as it does on the open ocean. Strong inversions characterize the lowest atmosphere much of the time in the Arctic and provide stable boundary layer conditions.

Three AIDJEX pilot studies have been conducted in the Arctic Ocean, each for a period of one or two months, in 1970, 1971 and 1972. The 1972 pilot program involved over 80 persons in the largest and most complex scientific project on drifting ice ever undertaken by the United States. Three manned stations were situated in a triangle 100 kilometers on a side, positioned by satellite navigation and acoustic bottom transponders. The stations moved generally westward with the average ice drift in this region, covering about 1000 kilometers in seven weeks. The peak drift speed observed was 22 kilometers per day. Other scales of ice motion were monitored near the main camp by a 10-kilometer laser strain net and a 100-meter optical strain net. The atmosphere was monitored by a number of micrometeorological programs as well as by tethered balloon ascents at the main station and surface weather observations at all three manned stations.

The large-scale ice pack motions were tracked with a network of automatic data buoys established in a 1000-kilometer net around the manned array. The data buoys were of two types. One type was interrogated and located by a Nimbus Satellite with the Interrogation Reconnaissance Location Satellite (IRLS) system. The pack ice in the AIDJEX region was surveyed several times by the NASA "Galileo" aircraft with remote sensors to collect microwave, infrared and visible images of the ice for correlation with other data. Information on ice properties was collected at the main station to help evaluate the remote-sensing images through correlation with actual surface data.

The 1972 program also involved current, temperature and salinity observations on varying scales and with different techniques. One group from the University of Washington took hydrographic profiles with water bottles and reversing thermometers twice daily at the three manned stations, and also deployed deep current meters for interior flow measurements. Another group from Washington used fast response current meters and a CTD (conductivity-temperature-depth) recorder to measure turbulent properties in the boundary layer beneath the ice. Data from the current meters gave a direct measure of skin friction between water and ice by a technique involving correlation

FIGURE 1: Drift patterns of arctic pack ice.

of turbulent eddy motion. The CTD results showed the influence of brine produced in the upper water column; the stratification of the upper layers by salt has an important influence on the water drag.

Scientists from Lamont-Doherty Geological Observatory mounted an array of current meters rigidly on fixed masts hanging below the ice to continuously monitor currents in the Ekman layer, a 10- to 20-meter-thick layer below the ice. Water stress in the Ekman layer was estimated from these observations by a technique of summing momentum in the layer. The ridges on the underside of the ice were found to be deeper in relation to the thickness of the Ekman layer in the water than are the ridges on the ice surface in relation to the thickness of the Ekman layer in the atmosphere.

It is to be expected that in a large and carefully planned experiment such as AIDJEX, interesting and unusual phenomena will occur from time to time which had not been fully anticipated by the planners. Such were the transient undercurrents, attaining speeds of 40 centimeters per second at a depth of 150 meters, which were noted on certain occasions. Although similar motions apparently have been observed a few times before in the Arctic Ocean, they were not noted in the 1970, nor in the 1971 programs. The 1972 work clearly showed them to be subsurface eddies. Eddy diameters of 10 to 20 kilometers were found in the depth range of 50 to 300 meters.

The Arctic eddies contrast with those in other oceans having generally larger diameter and a surface rather than subsurface maximum horizontal velocity. The differing properties of the Arctic eddies may be associated with the ice cover and with the steeper density gradient there. If so, the Arctic Ocean provides an opportunity on a geophysical scale to study eddies under altered conditions. The origin of these eddies and their part in the exchange of momentum, heat, and salt are not known. It may be that they are formed in the oceanic front north of Alaska, which separates the more saline water entering from the Pacific via the Bering Strait from the less saline surface water of the Arctic Ocean. If this is the case, the eddies must play an important role in the transfer of properties between polar and temperate oceans in the northern hemisphere.

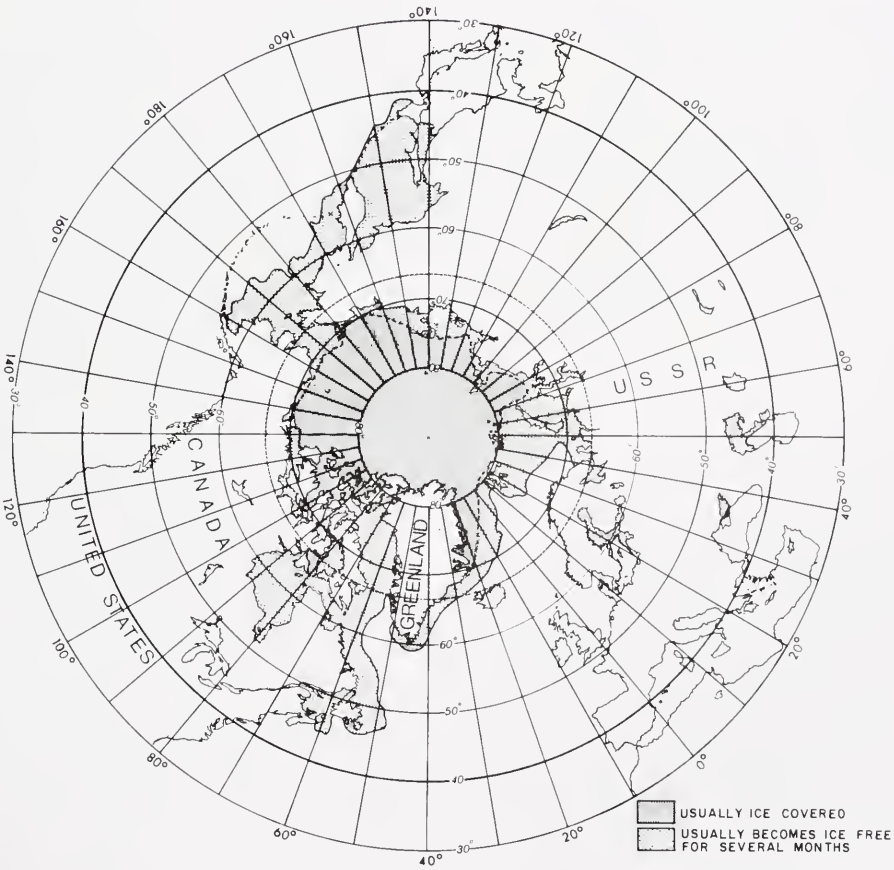
The main AIDJEX field program begins in February, 1975. An array of four manned stations with 100-kilometer spacing will be deployed in the same area as that used two years ago. Positioning again will be done by satellite navigation. The programs at the four camps will involve surface weather observations and oceanographic observations in the upper layers. Special studies of shorter duration will be restricted to the main camp. Ocean currents will be monitored to a depth of 200 meters with fixed and profiling current meters. Salinity and temperature will be monitored frequently to 1000-meter depth with CTD recorders at each station. The data from the various sensors will be recorded on a magnetic tape and flown at frequent intervals to the main camp, where they will be reduced with the aid of a minicomputer system, providing results on atmospheric and oceanographic variables soon after they are taken. A new type of data buoy has been designed jointly by the National Data Buoy Office, Polar Research Laboratories and the AIDJEX Office with funding from NOAA. The buoys will be deployed in a 300 to 400-kilometer radius around the manned array, transmitting information on position, atmospheric pressure, temperature and wind to the main camp.

Sea ice in the Antarctic is only beginning to receive interest and attention. During its maximum extent, the ice covers eight percent of the southern hemisphere, shrinking to about one-fifth of this area during summer. The wide seasonal variation suggests a large potential for climatic influence. When AIDJEX draws to a close after 1977, it is expected that its results and techniques will be turned southward to the problems of sea ice in the ocean surrounding Antarctica. This should give us basic information needed for an understanding of the important role of sea

ice in both polar heat sinks. With understanding, we may eventually acquire the ability to predict. Prediction of natural and human influences on sea ice and the effects of the ice on climate will permit evaluation of certain engineering plans which have been proposed for climate control.

Two specific schemes for influencing climate on a large scale are the removal of ice cover by spreading heat-absorbing substances and the stoppage of water exchange between the Arctic and Pacific by a dam across the Bering Strait. Both plans are expected by their advocates to improve northern climate and evidently are technically feasible today. However, it is doubtful that we presently have an understanding of the interaction of sea ice, ocean and atmosphere adequate to predict the outcome of such enterprises. Hopefully the studies described here and future work on ice-air-water interaction will provide a firmer dynamical basis for prediction.

FIGURE 2: Extent of sea ice cover in Northern Hemisphere. (Credit: Arctic Institute of North America)



Wind on the

Gabriel T. Csanady received his Ph.D. in mechanical engineering from the University of New South Wales. Dr. Csanady was chairman of the department of mechanical engineering for four years at the University of Waterloo and is now a senior scientist in the physical oceanography department at the Institution. He first became interested in great lakes dynamics while working on pollution problems in Lake Huron.

We all learn in elementary geography that the Great Lakes influence the climate of midwestern states and of Ontario, mitigating seasons and increasing rainfall. The interaction is not all one way: the atmosphere affects these lakes (and others of similar size) in a number of ways, some of which are quite important to those of us who enjoy swimming and boating or who drink lake water. Specifically, in lakes, as in oceans, winds are the major cause of water movements or "circulation", the nature and intensity of which largely determines the biological "health" of all natural bodies of water.

Some of the more spectacular effects of winds on large lakes have been known for some time. Within this category fall the "seiches" or large-scale oscillations similar to the sloshing of water in a bathtub. At Buffalo, for example, the level of Lake Erie fluctuates by as much as five feet in response to storms. This is of some practical importance, because the resulting change of hydraulic head appreciably affects the output of the Niagara electric power plant, not to mention the inundation problems at the opposite (western) end of the lake, which have received public attention lately.

Some more recent work on the Great Lakes, in which I have been involved, has been prompted by public concern over pollution problems. The reader may have come across statements in the press to the effect that Lake Erie is either "dying" or already "dead", and perhaps wondered what a live lake did. Phrases of this sort are, of course, picturesque exaggerations. The underlying truth is far less dramatic but quite serious enough: for a few weeks in midsummer some pockets of anoxic (oxygen-less) water, in which usual forms of aquatic life cannot survive, do form at the bottom of Lake Erie. This is evidence of serious pollution in the sediments, and it is fortunate that increased funds for research on the Great Lakes were made available subsequent to this discovery. As a result, several large-scale studies have been carried out in the past few years on different Great Lakes. Some of these (notably the International Field Year on the Great Lakes, IFYGL: 1972–1973, a cooperative study of Lake Ontario) required the cooperation of many government agencies, university scientists and others, both in the United States and Canada. Fifteen years ago, we knew less about water movements in the Great Lakes than in the North Atlantic or the Baltic Sea. Today we know a great deal more. Indeed, the knowledge gained in the course of this work is already proving useful in understanding complex processes in the sea, where controlled observations are harder to carry out. The oceanography of the continental shelves is likely to benefit especially, because the depths and widths of the shelves are similar to those of large lakes.

As the wind blows over a water surface, it drags along the top layer of the water, generating waves and turbulence. The

Fox Point, Rhode Island. Behavior of marine effluents is better understood, thanks to research in the Great Lakes.

Lakes

by G.T. Csanady

NERBC, U. S. Army Corps of Engineers

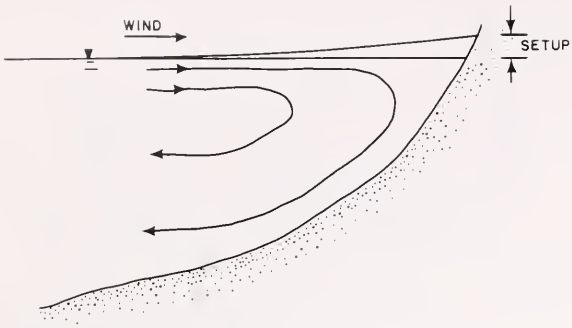


FIGURE 1: Wind drift against the shore. Diagram illustrates the vertical pattern in which top layers move with the wind – raising or "setting up" the downwind water level – and the bottom layers move against the wind.

horizontal drag force per unit area is the wind stress, which accelerates the water until some counterforce balances it. One such counterforce arises from the piling up of water against the windward shore. As the lake level "sets up", *i.e.,* becomes higher at the downwind end, a horizontal force is exerted on each unit mass of water, equal to the lake surface slope times the acceleration of gravity. In deep water, the surface slope need be only one centimeter of rise in ten kilometers of distance to balance even very strong winds. This is because the gravity force produced by the surface slope acts equally on each particle in the water column, so that the total force per unit surface area of the lake (the force countering the wind stress) is proportional to water depth. Assuming the wind stress to be constant, the slope will thus be large where the water is shallow, and vice versa.

The wind accelerates the water in the nearshore, shallow zone in a downwind direction. Return flow is produced by the "setup" in the deep portions of the lake. The total amount of water going one way is approximately the same as the returning amount. In shallower water, the cross section available to the flow is smaller and higher velocities develop. Wind-driven currents are therefore strongest in shallow water.

According to the above argument, a persistent wind would keep on accelerating the water in the shallow coastal zone forever. This is clearly not the case. The second important counterforce to wind stress is the frictional drag the lake bed exerts on water moving over it. This force is much like the resistance to flow in pipes and conduits and it varies with the square of the current velocity.

So far we have been discussing the flow pattern in a horizontal plane. Superimposed on this we also find circulation in the vertical plane, especially at windward or leeward shores. Figure 1 illustrates the vertical pattern, in which the top layers move with the wind, the bottom layers against the wind. The total flow, under equilibrium conditions, is zero, reflecting a balance of wind stress and gravity force due to the raised water level. In the top layers, the effects of the wind overwhelm those of gravity, and in the bottom layers gravity wins out, a situation analogous in some ways to the horizontal flow pattern. Qualitative evidence for this kind of flow pattern is there for everyone to see on a wind-blown lake. Quantitative evidence, however, is difficult to obtain under field conditions. In laboratory flumes, such flow patterns have been simulated and studied, but our understanding of them is far from complete.

Watching a wind-blown lake, the reader may have observed long, straight streaks of foam aligned with the wind, known as "windrows". These have been used by pilots to ascertain the direction of the surface wind. They were first described scientifically 35 years ago by Irving Langmuir, but their physical mechanism remains a mystery. We know that the accumulation of foam (or of other floating debris) occurs at a confluence line between two counterrotating vortices, as sketched in Figure 2, but how the wind can produce long vortices with their axes parallel to it is not so easy to explain. The current view is that these are part of the big-eddy structure of turbulent wind drift, and that the circulations associated with them extend to the bottom of the lake or to a "false bottom" provided by the thermocline, a boundary we will discuss further on.

The wind sets up circulations of different patterns in the horizontal plane, the vertical plane parallel to the wind, and that perpendicular to the wind. Each of these affects pollutants released near shore in its own characteristic way. Coastal currents paralleling the shore have the highest velocities and tend to produce long plumes of effluent traveling alongshore for considerable distances. The worst effects are produced by the onshore component of the wind, through the circulation depicted in Figure 1, which can effectively transport pollutants to beaches. Generally speaking, the Langmuir circulations of Figure 2 are beneficial, because they promote vertical mixing and thereby supply oxygen to the deep layers of the lake (and contribute to the dispersal of pollutants).

Density differences in a lake, as in the ocean, produce effects seemingly quite out of proportion to their magnitude. As the sun heats the top layers of a lake in the summer, the water becomes a little lighter at the top than at the bottom – by something like one or two parts in one thousand. Normally, we can safely ignore much greater changes than this in material properties, but in the case of density, the force of gravity makes the difference. In connection with surface level slope, we have already seen that one millionth of the force of gravity can balance an ordinary wind stress in water about 100 meters deep. It therefore should not come as too much of a surprise that one or two thousandths of the force of gravity (the difference in the weights of a top, warm water layer and a bottom, cold one) can overwhelm the effects of the wind completely. Vertical mixing in the summer (by Langmuir circulations and similar big, wind-driven eddies) penetrates only to the depth where turbulence is suppressed by the force of gravity – about 25 meters in the Great Lakes. This well-mixed layer (called the "epilimnion") is bounded below by the "thermocline", a region of rapid temperature drop and density increase beneath which lies a cold and largely homogeneous water mass (the "hypolimnion"), effectively isolated from the surface by the thermocline. The absence of turbulence in the neighborhood of the thermocline indicates that here the vertical transfer of any substances (oxygen or nutrients) as well as of heat and momentum becomes exceedingly slow, quite negligible in the time scale of a few weeks.

The yearly thermal cycle of a large lake at mid-latitudes may be conceived of as beginning in April, when the water is more or less completely homogeneous and somewhat colder than the temperature of its maximum density (4°C or 39°F). At this time of the year, heating of the surface by the sun is intense and produces a slight increase in density – enough to produce sinking motions and good vertical mixing ("spring turnover"). In the shallow waters near shore, the entire water column soon heats up past 4°C; further surface heating produces a slightly lighter fluid which would stay at the surface were it not for mixing by wind. As we have seen, wind mixing penetrates only so far and then a thermocline begins to form. However, the deeper waters take much longer to reach 4°C and become significantly warmer at the

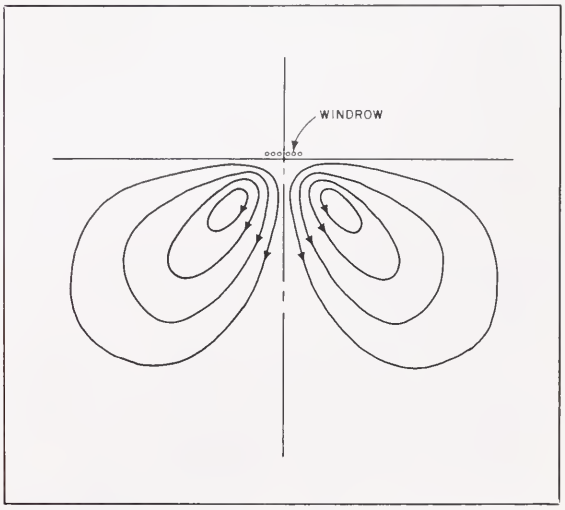


FIGURE 2: Diagram in vertical plane showing windrow-forming Langmuir circulation pattern. In as yet unexplained fashion, wind produces long, counter-rotating vortices with their axes parallel to it. Foam or floating debris accumulates at confluence line between vortices.

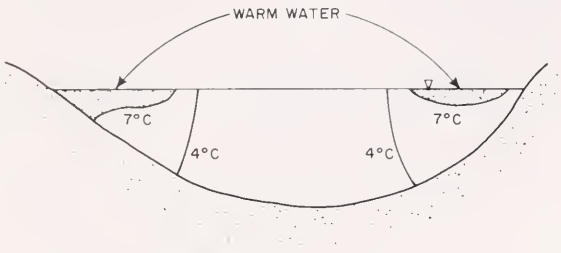
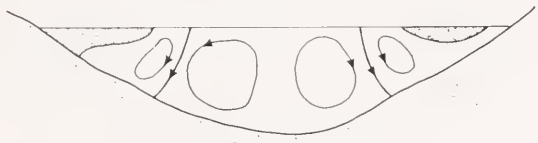


FIGURE 3: Typical spring thermal structure in Lake Ontario. The 7°C isotherm can be regarded as the center of the thermocline. When winds blow the warm water offshore, it forms a lens at the surface; blown onshore, the water above the thermocline takes on a wedge shape.

Downward leg of a lakewide circulation pattern is established on the offshore side of the spring thermocline illustrated above. Called the "thermal bar", the leg apparently acts as a separating streamline between a nearshore and an offshore circulation cell. Bar is formed by water which has achieved maximum density (4°C) through horizontal mixing of slightly colder and slightly warmer water.



surface, so that the spring thermocline at first only exists over the shallowest regions near shore. Here, considerable horizontal temperature contrasts remain in existence for some six weeks, the shallow waters warming up quite rapidly in places. In late May, it can be quite pleasant to swim in Lake Huron or Lake Michigan, while a month later, when the temperatures have equalized horizontally, the water is usually quite cold again.

The spring thermocline rings the lakes and, being located at a generally shallow depth, is strongly influenced in its dimensions by winds. Figure 3, a cross section of Lake Ontario, illustrates the typical spring thermal structure. The 7°C isotherm[1] may be taken to be the center of the thermocline. When winds blow the warm water offshore, it forms a lens at the surface (evident along one shore in Figure 3). Blown onshore, the warm water above the thermocline takes on a wedge shape.

Strong offshore winds would blow the warm water well into the lake and eventually break down its stratification. However, at this time of the year, low-level winds over the cold Great Lakes tend to be light, due to the same physical conditions which produce the stability of the thermocline. Atmospheric air in contact with the cold water surface becomes significantly heavier than the air above, so that the force of gravity reduces and possibly suppresses vertical mixing in the air. This means that the lower layers of air are not effectively dragged along by the stronger winds above; the wind stress exerted on the water surface becomes lower on the average than at other times of the year. This no doubt contributes to the long survival of shallow spring thermoclines.

As greater volumes of warm water are produced by solar heating, the spring thermocline moves offshore. Usually with some assistance from the winds (a major storm), significant horizontal temperature and density contrasts are finally eliminated by late June. A summer thermocline becomes established at an average depth ranging from 10 to 25 meters. In the larger lakes, most of the water lies below the summer thermocline and remains quite cold, close to 4°C. However, in relatively shallow Lake Erie the hypolimnion volume is small and is moreover confined to a few distinct pockets. These small volumes of isolated hypolimnion water must provide all the oxygen necessary for chemical and biological processes in the sediments. Because the sediments consist at least partly of pollutants with an active oxygen demand, the oxygen content of these deep water pockets is soon depleted. For biological reasons this is apparently quite undesirable.

The "equilibrium" summer thermocline and the anoxic pockets in Lake Erie persist only for the four to six weeks of midsummer.

[1] A line on a chart connecting points having equal temperature.

By late summer the top layers of the Lakes begin to lose significant amounts of heat, mainly at night. The cold water forming at the surface sinks and helps agitate the layer above the thermocline. When coupled with the mechanical stirring of a stronger wind, surface cooling leads to a deepening of the thermocline. Eventually, the thermocline descends to the bottom – in Lake Erie quite rapidly, in deep lakes over the span of the early fall months. By late fall the winds mix the lake top to bottom ("fall overturn"). In the winter, a weak stratification is usually in evidence, surface water between 0° and 4°C being somewhat lighter than deeper water, the latter staying close to its temperature of maximum density. However, stronger winds easily break down this weak stratification; it may be that wind mixing penetrates hundreds of meters. The deep portions of the Great Lakes, therefore, do not usually freeze over. The heat content of their deep water mass is large, and winds tend to mix them vertically.

The wind effects we have so far discussed are by and large pretty much in accord with one's physical intuition. Some quite unexpected phenomena arise, however, in connection with the "Coriolis force", the apparent force due to the rotation of the earth (see page 4). Because the earth turns around its axis once a day, its rotation only affects slow motions, of a time scale comparable to a day or longer. In the Great Lakes, motions of this kind are prominent only in periods of strong stratification, *i.e.*, when a thermocline is present. One way for the lake waters to move under such circumstances is for the top layer to slide one way, the bottom layer in the opposite direction, with zero depth-averaged flow. Such "internal mode" motions, or readjustments of the internal mass distributions, generally proceed slowly and are very noticeably influenced by the Coriolis force.

The Coriolis force is proportional to the velocity of any moving object or particle and, in the Northern Hemisphere, points 90° to the *right* of the motion. As the water begins to move with the wind, the Coriolis force gradually deflects it to the right. Far from any shores, the water comes to move toward the right of the wind in a matter of six hours (at the latitude of the Great Lakes). Once the water is moving in this direction, the Coriolis force can exactly balance the wind stress. Such equilibrium motion is called "Ekman drift" (see page 5). However, water impulsively set in motion by the wind overshoots the equilibrium Ekman drift in the manner of a pendulum; its direction of motion keeps changing periodically, with a period of about 16 hours in the Great Lakes.

To those not acquainted with the theory of rotating fluids, the most surprising consequence of the earth's rotation is that as the wind begins to blow, the initial piling up of the water does not occur at the down-wind end of the lake, but on the right-hand shore, looking along wind. This of course is a consequence of the Ekman drift, which is stopped by the right-hand shore. On the opposite, left-hand shore, cold water is piling up in the bottom layer, while the top waters are being depleted. In fact, it only takes a modest storm to remove all the warm water from the left-hand shore, allowing cold bottom water to well up. Such upwellings generally have beneficial consequences for marine life, because with the deep water come mineral nutrients (marine fertilizers). However, the attractiveness of the lake for swimming declines markedly when the water temperature suddenly drops from 65°F to 40°F.

The prevailing winds in the summer around the Great Lakes are southwesterlies. "Right-hand" shores to these winds are the eastern shores of Lakes Michigan and Huron, and the south shore of Lake Ontario. These are generally the warm shores in the summer, the opposite ones being cold more often than not. Of course, northerly or easterly winds also occur and they produce upwellings on the so-called warm shores as well (at our cottage on the eastern shore of Lake Huron we usually get these when we have houseguests).

The warm water piled up on the right-

hand shore by a storm escapes in the coastal boundary layer, always flowing with the shore to its right, *i.e.*, with the shore pressure balancing the Coriolis force. Similarly, the cold water from the left-hand shore flows around the lake in a counterclockwise direction. Actually, individual particles do not move much further than a few tens of kilometers in this mass readjustment. This motion is a long internal wave which propagates on the thermocline around the perimeter of the lake, always counterclockwise in the Northern hemisphere. Such a wave is known as an (internal) "Kelvin wave" after its British discoverer, Lord Kelvin. We may say that the internal seiche in large lakes takes the form of a Kelvin wave which is "trapped" within a coastal boundary layer of some 5–10 kilometer width.

With the seiche-like motion being confined to a nearshore band, some high velocities may be expected there. The resulting flow structures are known as "coastal jets", bands of high speed warm water centered a few kilometers offshore. Strong currents of this sort are important in carrying pollutants away from their point of introduction. They are essentially temporary structures, set up irregularly by passing storms. As upwellings and downwellings alternate in the course of summer (even if one or the other usually predominates in a given location), the coastal zone is effectively flushed out with water coming from the depths of the lake. This helps the deeper lakes to digest much of what we throw into them. Lake Erie is too shallow for significant upwellings to occur – a fact which adds both to swimming enjoyment and pollution.

Recently, proposals have been made for more intense human exploitation of the continental shelves – for example, the location of nuclear power stations a few miles offshore. To assess potential pollution problems arising from such projects, we should know more about coastal currents and about the circulation over continental shelves. From one point of view, the shelves are large lakes with one shore removed: along the remaining shore we should still expect to find such phenomena as coastal jets, for example. The depths and widths of continental shelves are very similar to those in large lakes, and we should find basically similar dynamic factors in operation.

Ron Winch, Photo Researchers

The ledges of Lake Superior.

Vicky Briscoe

Spring at the Oceanographic: Lulu's *"windowbox."*

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