PART D

LARGE-SCALE PHENOMENA

This section deals with oceanic phenomena with horizontal scales from approximately 100 km up to the widths of the oceans themselves. Their very size makes these phenomena difficult to monitor by surface-based methods, so that satellite-mounted active microwave techniques are especially valuable and may be the only way of gathering the information needed for scientific progress.

An important class of large-scale features consists of those concerned either directly or indirectly with the vertical topography of the ocean surface, estimated from the center of the Earth. This class includes the shape of the geoid, which is determined by the internal mass structure of Earth; the quasi-stationary anomalies due to spatial variations in sea density and steady current systems; and the time-dependent variations due to tidal and meteorological forces and to varying currents. These features are discussed in subsequent sections. All these features make use of the techniques of radar altimetry, although currents can also be studied by satellite tracking of floats.

Certain large-scale phenomena could, in principle, be usefully observed by satellite altimeters; however, it is unlikely that the necessary precision (approximately 1 cm) could be obtained in the near future. Steric variations in sea level, due to time-dependent changes in water density, notably at annual and semiannual periods, are known from shore-based tide gages to have amplitudes of less than 10 cm in most parts of the ocean. Tsunamis sometime reach meters of amplitude at a coastline, but in the open ocean where they would be most useful to measure, they merely consist of long waves of a few centimeters' amplitude and some tens of kilometers' wavelength. Prospects for detecting tsunamis by altimetry are discussed by Greenwood et al. (ref. 3–43).

MARINE GEODESY

Background

The primary function of geodesy is to determine the shape of the Earth by carefully measuring the geometric distance between points on its surface. For years, this approach seemed to be satisfactory because the surveyed areas were limited to portions of continental crusts and the measurements were not required to consider the micromovements of the solid Earth crust, which was considered as a perfectly rigid body. Problems related to the choice of a common reference datum for all those local determinations were approached through the use of a theoretical surface called the geoid, which is an equipotential of the Earth gravity field.

The geoid on land is not a physical surface; it is determined through a numerical process that involves the computation of an integral of gravity anomalies called the Stokes formula. The geoid on the oceans was long considered to be in coincidence with the physical surface of the sea (refs. 3–44 and 3–45).

During the last 10 yr, the problems of geometric geodesy have been extended to the open oceans mainly through the needs of accurate navigation and positioning. Marine geodesy has appeared as a separate scientific endeavor because most of the hypotheses of practical "land" geodesy were found to poorly match the physical properties of the ocean environment. At the same time, the accuracies of the available instruments for distance measurements were drastically improved so that movements of the crust of very small amplitude had to be considered.

Presently, serious difficulty is encountered in coping with the shape of the Earth per se, and marine geodesy must be concerned with four important surfaces (ref. 3–46): the physical surface of the Earth crust, the physical surface of seawater, the geoid, and the mean sea level.

The physical surfaces are experimentally accessible but the others are concepts with definitions that unfortunately tend to vary slightly from one author to another. Following is a brief overview of the problems encountered when dealing with those surfaces.

The physical surface of the Earth crust.—The Earth crust is the solid surface of Earth, including that portion underlying the seas and the oceans. Long regarded as perfectly rigid, the Earth crust is, in fact, the outer shell of a "pulsating" elastic medium in everlasting slow movement. Those displacements are essentially as follows:

1. The solid Earth tides (20 to 40 cm, vertical).
2. The continental drift (a few centimeters a year, horizontal).
3. Movements caused by earthquakes and fault displacements (a few centimeters to a few meters in any direction).
4. Tilt of coastal areas caused by the rising and falling of ocean tidal motions (a few centimeters in any direction).
5. Movements caused by the modification of loading conditions on the crust (erosion-sediments piling).

Some of those effects would have to be considered very carefully when using a radar altimeter with a 10-cm resolution capability.

The physical surface of the sea water.—The water surface changes position for many reasons, some of which are ocean tides; air pressure; winds; temperature, density, or pressure gradients; movements caused by sudden variations in the physical surface of the ocean floor (earthquakes, fault displacements); and geologic changes associated with the melting of glaciers (ref. 3–47).

This fast-moving surface is obviously difficult to match with the present concept of the geoid over the open seas; a fresh look at the problem is badly needed. Other important interactions that may occur between the two surfaces are the cross-coupling of ocean tides and solid Earth tides and the movements of the sea surface due to movements of the ocean floor. Both these interactions tend to be very difficult to decipher in the coastal zone areas.

The geoid.—The only satisfactory definition of the geoid from the mathematical viewpoint is that of an equipotential of the Earth gravity field with a certain absolute potential. However, an empirical approach has been imposed by practical considerations so that local geoids have been defined by each national geodetic agency as equipotential surfaces passing through a defined point (datum point), usually chosen in the countries having oceanic borders near the mean sea level. Unfortunately, those local geoids do not coincide over the total surface of the Earth because the datum points have not been chosen on the same equipotential.

One of the greatest contributions of satellite altimetry (using coast-to-coast tracking across the Atlantic Ocean) will be to provide experimental data ultimately useful in the choice of a common datum for North America and Europe. This datum should be chosen as a reference for the North Atlantic marine geoid.

Three ways of mapping the shape of the geoid are available today.

1. Measurement at sea of gravity and integration through the Stokes formula.
2. Measurement at sea of the vertical deflection of gravity and integration through the Veining-Meinez formula.
3. Measurement of the altitude of a satellite over the physical surface of the sea leading to the shape of the geoid with the use of appropriate correction and filtering of the data.

Gravity at sea cannot be measured, in the foreseeable future, to an accuracy better than 0.01 m/sec^2, thus leading to uncertainties in the geoidal heights computed from the sea of at least a few meters. Deflections of the vertical could be measured to an accuracy of 0.5 sec of arc with the GEON system (ref. 3–48) and to a few seconds of arc using an inertial navigation system (ref. 3–49).

The wealth of geoidal information available from radar altimetry has been adequately demonstrated by data from the Skylab S193 altimeter experiment (ref. 3–50). Figures 3–25 and 3–26 are two examples. Figure 3–25 shows the profile of the Puerto Rican trench area, which represents a mass deficiency, and figure 3–26 shows the effect of a sea mount, which corresponds to a mass concentration. Because the geophysical value of such information is well known, this discussion will be directed toward geoidal feature resolution available from altimetry.

To discuss the large-scale resolution capabilities of radar altimetry, it is first necessary to examine geoid spectrum and altimeter-related characteristics. Figure 3–27 shows a geoidal power spectral density of the Puerto Rican trench area, computed by using Skylab S193 data from McGoogan et al. (ref. 3–50) and Miller and Brown (ref. 3–51). This area was selected because it should contain short-wavelength components of substantial magnitude; therefore, the analysis should yield an approximate upper boundary on geoidal data processing requirements for GEOS–C. Figure 3–28 shows the spatial filter transfer function of the altimeter, which was computed based on the specific tracking system used in the GEOS–C altimeter. The mini-

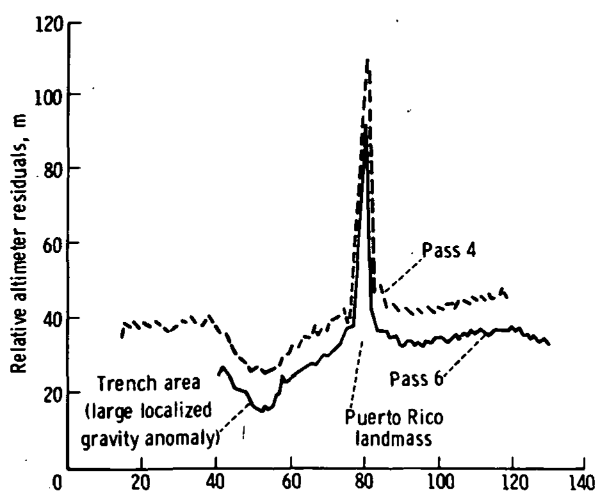


FIGURE 3–25.—Geoidal profile of the Puerto Rican trench area from Skylab passes 4 and 6 (ref. 3–48).

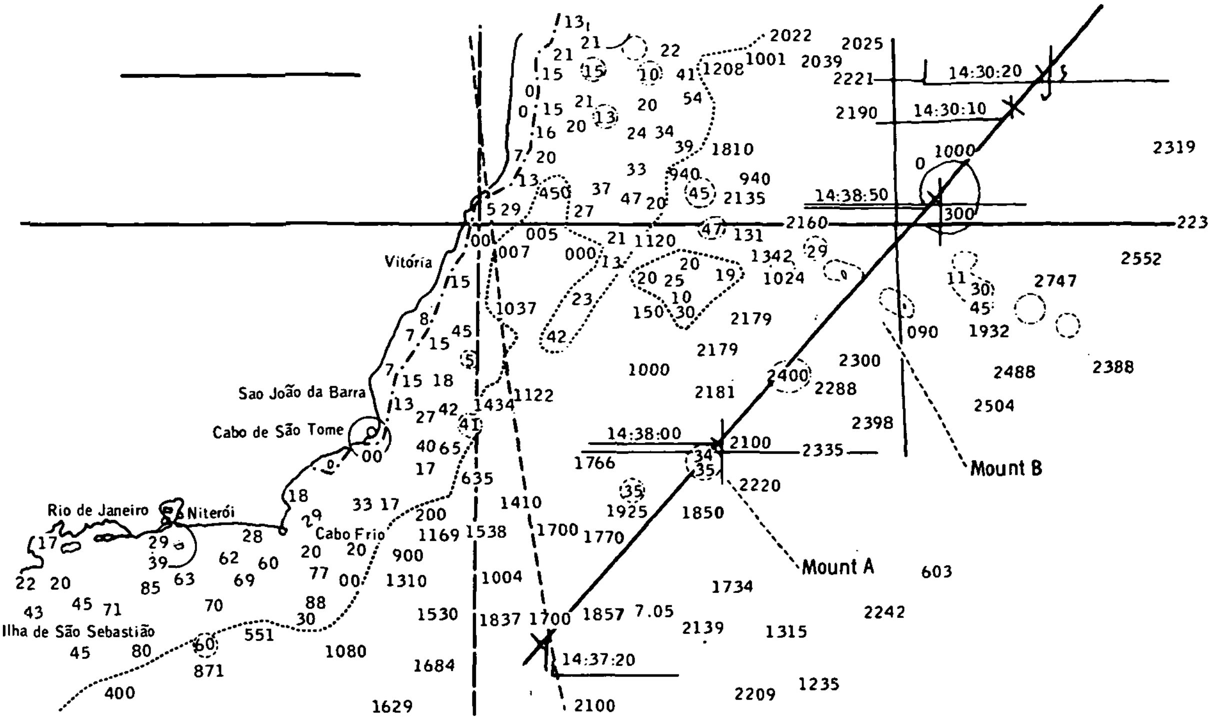
mum-mean-square linear filter for processing altimeter data may be derived based on the power spectrum and spatial filter models discussed and on known altimeter random error characteristics. Optimum impulse response functions are shown in figure 3–29. Referring again to figure 3–27, the half-power bandwidth of the optimum filter is seen to be 40 km for the GEOS–C design.

These findings indicate that future altimeter systems will be able to provide substantial improvement in resolution of features to wavelengths as short as 5 km.

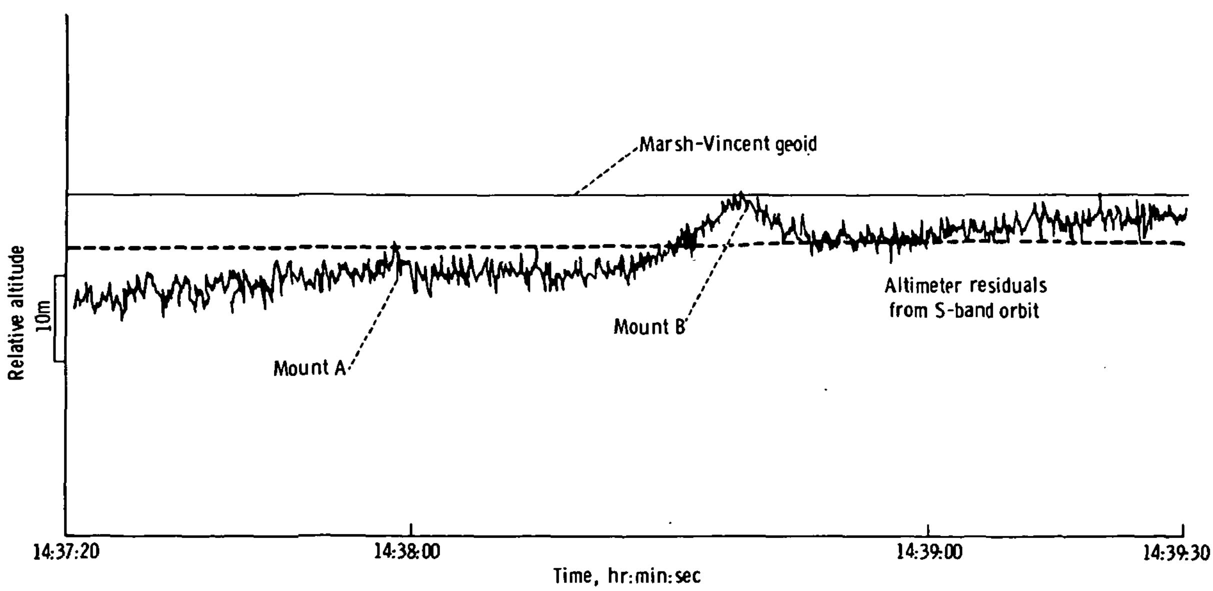
The mean sea level.—The geodesist, looking for a reference surface from which land heights could be measured, adopted the average height of the ocean surface in coastal areas as defining segments of its reference surface, the hypothesis being that the geoid and the mean sea level coincide all along the coastal lines.

The basic idea was that proper averaging in time of tide gage logs taken in ports along the coast would provide reliable estimates of the true position of the geoid. Coastal geodesists have developed complex schemes for filtering out all tidal influences. Unfortunately, other systematic effects are still present in their computations. It is well known that a 10-m discrepancy exists between the two ends of the eastern coast of the United States. High correlations of the residuals with atmospheric pressure variations (ref. 3–52) have been noted by oceanographers all along the Mediterranean coast.

The mean-sea-level computations could be corrected with radar altimeter measurements, but this is a very difficult task indeed. Radar altimeters take instantaneous profiles of the physical surface of the sea, whereas mean sea level is computed from time logs taken at fixed points. The only method of comparison would be by the averaging of many profiles measured by the altimeter. Such a time-consuming task would only be fulfilled many years from now after many successful launches of radar altimeters aboard satellites.



(a)



(b)

FIGURE 3–26.—Effects of subsurface features on altimeter data; Skylab 3, pass 11. (*a*) Geographic area and ocean bottom topography (numbers indicate water depth). (*b*) Altimeter relative profile.

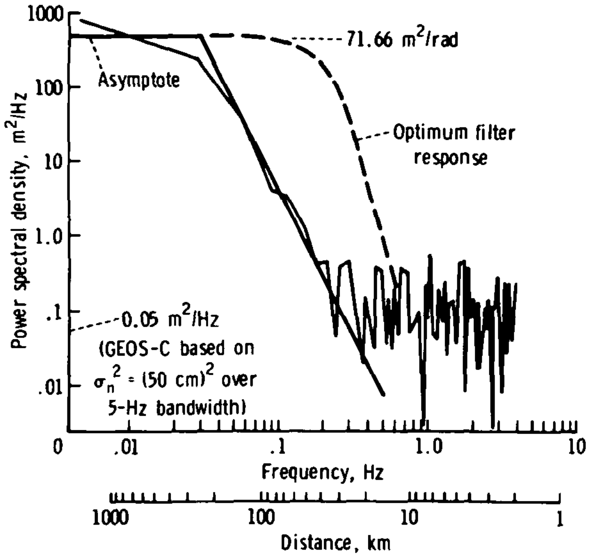


FIGURE 3–27.—Geoid undulation spectrum of Puerto Rican trench and Wiener filter transfer function.

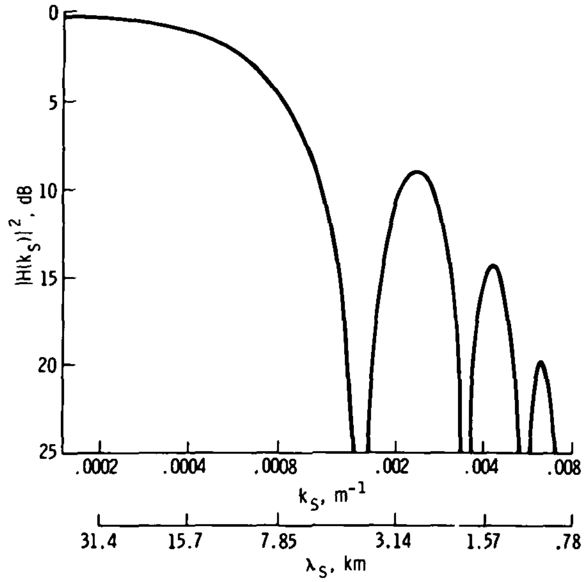


FIGURE 3–28.—Spatial filter transfer function, where altimeter altitude is 440 km and antenna beamwidth is 1.5° (rectangular gate). (From ref. 3–50.)

Figure 3–30 presents ideas on how an integrated system for marine geodesy could be built around the short-arc coast-to-coast tracking of a radar altimeter. Signals T_1, T_2, and T_3 are time references transmitted by the satellite to synchronize all the ocean and Earth tides logs.

Applications

The need for high-accuracy geoidal information to support studies such as crust deformation and solid-Earth models is well known (ref. 3–53). Marine geodesy is concerned with two main applications: the positions of objects on the Earth solid surface in a marine environment, and the average position of an object on the water surface.

Requirements of users for each category are summarized in tables 3–V and 3–VI (ref. 3–54) and are expressed in terms of needed precisions in geocentric (ϕ, λ) position and height (h) above the geoid. Many of the specifications can only be fulfilled if a marine geoid has been determined with a relative precision of better than 1 m ($\approx$50 cm) and an absolute precision of a few meters ($\approx$5 m).

A secondary requirement for high-accuracy ($<$10 cm) geoidal information is that of supporting ocean topography activities. If oceanographic applications of remote-sensing activities are to achieve their full potential, it will be necessary to define the geoidal, or *permanent, contours to a resolution comparable* to the magnitude of oceanographic effetcs.

OCEANIC TIDES

Historically, tidal research has concentrated on the immediately practical problems of computing tide tables for the major ship-

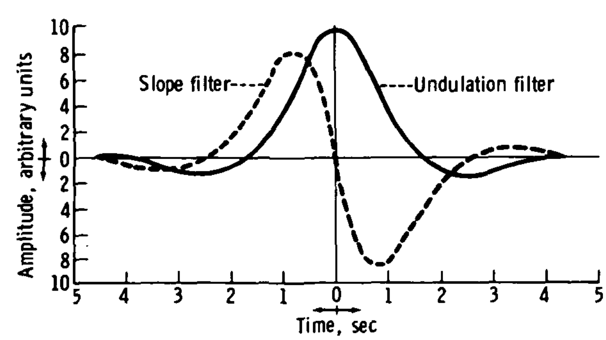


FIGURE 3–29.—Derived weighting function for geoidal data processing.

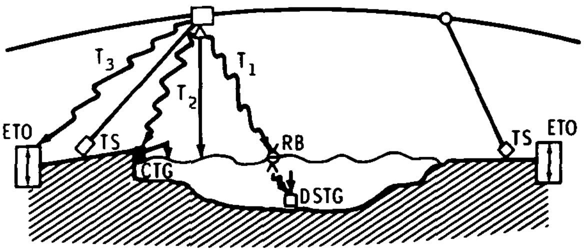


FIGURE 3-30.—Concepts of an integral coast-to-coast radar altimeter system for marine geodesy.

ping ports. Unfortunately, this can be done quite accurately by a purely empirical process that teaches one almost nothing about tidal dynamics. Since the work of the 18th- and 19th-century French mathematician, Laplace, it has been clear that the proper understanding of tidal dynamics demands a knowledge of the tides in the open ocean, where they are generated by the Newtonian gravity field of the Moon and the Sun. Such knowledge would not only be a fundamental contribution to oceanography but would also, in principle, enable the computation of the propagation of tidal energy into the shallow seas and estuaries and, hence, give predictions of both tidal elevations and streams in any location that may be required for future practical purposes.

No one has yet succeeded in attaining this knowledge. Computing the disturbing gravitational field that drives the system is fairly easy, and Laplace's field equations have been written in countless learned scientific papers, but their solution for the complex geometry and topography of the actual ocean presents great difficulties. The difficulties arise partly from the large number of possible normal modes of oscillation, partly from the fact that all the ocean basins interact (so that partial solutions for a single ocean are impossible without independent tidal measurements around its entire boundary), and partly from the necessity to model the dissipation that is concentrated in the intricate network of shallow seas. Analytical solutions in geometrically defined basins are only of qualitative value. Numerical solutions for actual ocean topography tax even the largest modern computers and ultimately only show the extreme sensitivity to fine topographical detail and to the yielding of the Earth crust, factors previously considered negligible.

The lack of definitive global maps for the oceanic tides is a major lacuna in modern geophysical knowledge. Apart from the intrinsic interest in defining the response of the ocean to the known gravitational forces and the driving mechanism for the co-oscillating tides in all the shallow seas, reliable tidal maps are fundamental to the solution of a

TABLE 3-V.—*Positions of Motionless Objects on the Solid Surface of the Earth*

Activity	Precision, m: ϕ, λ	Precision, m: h	Absolute (A) or Relative (R)
Coastal landmarks . .	1	0.5	R
Island positioning . .	20	20	A
Deap sea mining . . .	10	3	A
Drilling platforms . .	20	—	R
Shallow water topography	20	—	R
Geologic mapping . .	50	—	R

TABLE 3-VI.—*Average Position of an Object on the Surface of the Water*

Activity	Precision, m: ϕ, λ	Precision, m: h	Absolute (A) or Relative (R)
Oceanographic buoys	70	—	A
Oceanographic ship on station	35	—	A
Drifting buoys . .	50	—	R
	200	—	A
Oceanographic vessel en route	100	—	R
Submarine cable operation	100	—	R
Hydrographic vessel on station	5	—	R
	70	—	A

number of other geophysical problems. Researchers on tides of the Earth and of the atmosphere, on geomagnetisms, and on the momentum balance of the Earth rotation are increasingly finding their work hinging on the oceanic tides. They turn to the oceanographers, who are unable to supply the required data.

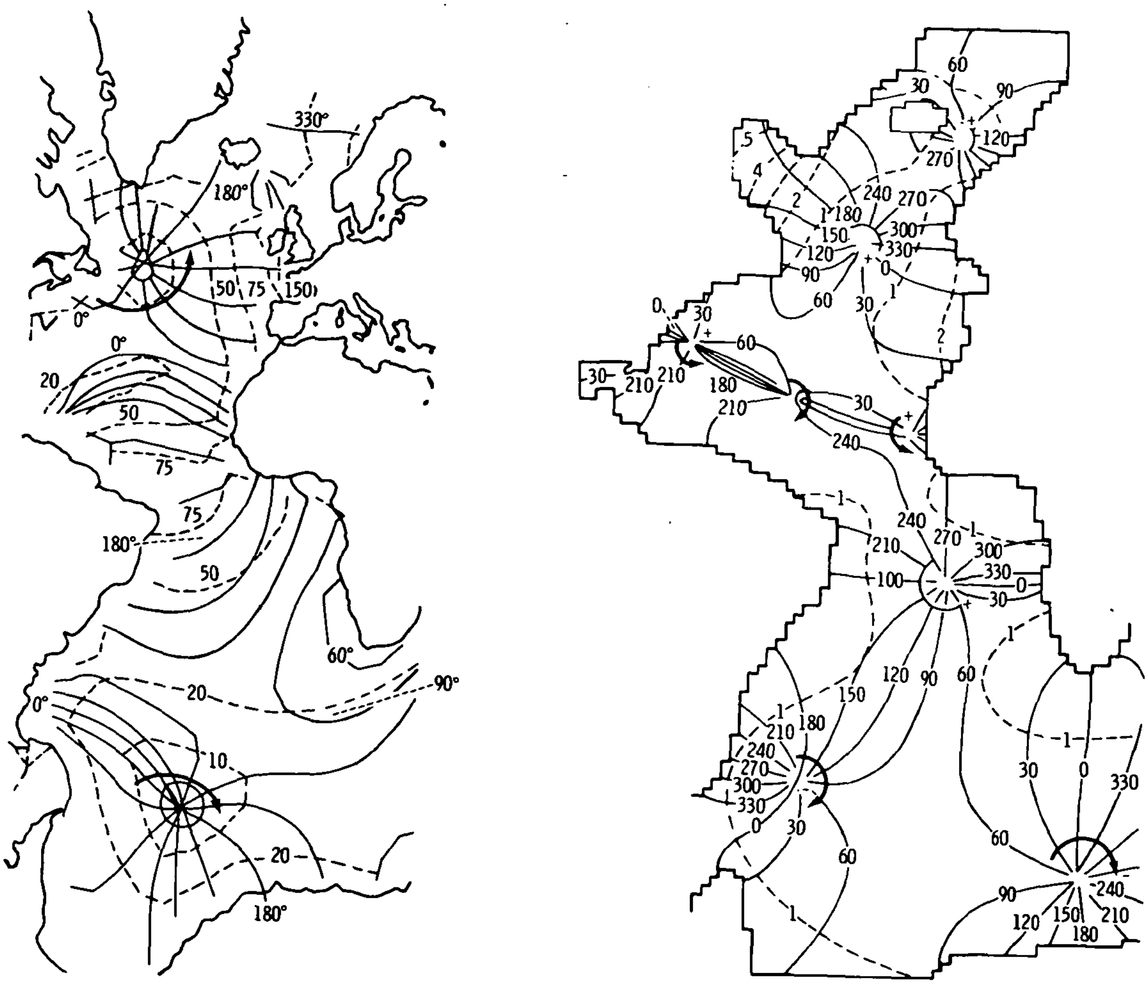
State of the Art

Approximately six maps of the M_2 tide (lunar semidiurnal constituent) exist that have been computed for most of the world's oceans in recent years by various methods and assumptions. Assumptions vary, but the most fundamental divisions occur (1) between those who force the solutions to agree with assumed continental boundary tidal conditions and those who eschew the use of all empirical data, and (2) between those who attempt to allow for crustal yielding and those who assume a rigid Earth. (For a modern review, see Hendershott (ref. 3–55).) Two of the most authoritative and advanced results are compared in figure 3–31. They are in fair agreement on certain features, such as the tidal configuration in the North Atlantic, but their results for the South Atlantic are quite different. Such disagreement, here and in other oceans, is typical of all cotidal maps to date. Ironically, it is extremely difficult even to obtain the measurements necessary to find out which interpretation, if any, is correct. Maps of the diurnal tides are rarer, but those that are published show a similar lack of agreement.

Much of the controversy about global tidal computations centers on ocean/land interaction, which is rather complicated, but there are also limitations due to the lack of proper information to be fed into the computer model from the boundaries bordering shallow seas. A typically crucial area for the Atlantic Ocean appears to be the Patagonian Shelf, but reliable data from here and many other places are sparse. Computer models would also benefit from measurements across transoceanic sections, which would provide boundary conditions for separate evaluation of smaller zones. Isolated spot measurements at certain places where the tidal amplitude is known to be large would also be valuable.

Seabed pressure sensors for measuring tides in the open sea have been increasing in numbers over the last decade, but they are expensive devices with a high risk of loss and require expert supervision and many ship deployments. The sensors are supported only by a few nations that have well-developed oceanographic technologies; thus, the chances of obtaining data in the foreseeable future, from most of the Southern Hemisphere at least, are remote. To date, the only tidal data that have been taken from the open sea (apart from islands) have been from within a few hundred kilometers of North America, Northwestern Europe, and South Australia. Monitoring the tides in all the world's oceans by direct measurement, or even obtaining the minimal data that might enable computation, is not possible at this time.

Radar altimetry from spacecraft will transform the measurement situation. The difference between satellite range and altimeter height above the sea surface, suitably corrected for various biases and errors, will vary from a variety of causes discussed elsewhere in this report (ref. 3–45). The variables may be divided into (1) permanent or quasi-permanent features of ocean topography, (2) features that vary in time at any given Earth location, and (3) tracking and other errors. Variables of type (1), including the geoid and steady dynamic features, although approximately 1 to 100 m, can in principle be extracted either from independent knowledge or, in due course, by averaging from repeated satellite passes. Variables of type (2) include the tides, up to approximately ± 2 m in amplitude, and also include random variations caused by the weather, on the order of 0.1 to 0.3 m. (Transients such as surges and tsunamis may be ignored in this context because their amplitudes are appreciable only in shelf seas.) A large proportion of the weather influence is a direct reflection of atmospheric pressure at 1 m of sea level per N/m^2. This can be removed by means of



(a) (b)

FIGURE 3-31.—Cotidal maps of the Atlantic Ocean using different methods and boundary conditions. (*a*) Cotidal maps computed by Pekeris and Accad, 1969 (ref. 3-56). (*b*) Cotidal maps computed by Hendershott, 1972 (ref. 3-55).

synoptic meteorological data computed on land. The residual signal will therefore be considerably tide dependent, provided variables of type (3) are well below 1 m. This dependence is further enhanced by using the strong coherence that exists between any oceanic tidal variable and the perfectly calculable tide-generating potential of the Moon and Sun (ref. 3-57).

Requirements

As in all applications of satellite altimetry, the only basic requirement demanded is accuracy of the interpretable signal. Because modern trends in altimeter design enable its accuracy to be reduced to a few centimeters, the major requirement is in the accuracy of orbital tracking over as wide a global coverage as possible. Stating minimal requirements outside the context of discussion of approaches to various problems is difficult, but the situation may be summarized as "given any reasonable accuracy, one can expect to learn so much tidal information and no more." With 1- to 2-m accuracy, there is some hope of gaining useful information about the oceanic regions with largest tides, such as parts of the North Atlantic. With 0.5-m accuracy, one should certainly obtain good information from such regions and could hopefully consider wider oceanic fields. Complete tidal definition over the world's oceans would require 5- to 10-cm accuracy. In any case, continuous signals for many months or years are assumed.

An altimeter "footprint" of approximately 100 km should be adequate for most tidal work in the deep ocean. To resolve the smaller horizontal scale of tides in most shelf seas, resolution better than 20 km would be needed.

Technical Approach

At some stage in the interpretation of the altimeter tidal signal, one will have to allow for the fact that the signal differs fundamentally from the usual land-based concept of the ocean tide. If z' is the anomaly in sea level sensed by a tide gage, that sensed by the altimeter is $z = z' + z''$, where z'' is the tidal elevation of the Earth crust, approximately 0.2 m. (See ref. 3–55 for a full dynamical discussion.)

A large part of z'' consists of a direct yielding of the body of the Earth to the gravitational forces, which is easily calculable, but there is a smaller residue of approximately 0.1 m that can only be calculated in terms of global integrals of the ocean tide itself (i.e., the "loading tide"). Clearly, for some purposes, one can interpret z in terms of z' merely by approximating to z'' by the simple body tide, with small loss of accuracy. For higher precision, a fair approximation to the loading tide could be computed from one of the existing rough cotidal maps. The entire situation procedure is, in any case, one of successive approximations.

Two possible approaches to data analysis are possible. One method relies on obtaining many traverses of a given track zone; that is, for example, 500 km wide. After correcting for the geoid (which preferably should not vary too steeply in the chosen area), for atmospheric pressure, and for any other known anomalies, the successive altimeter readings in any 500-km-square section are treated as a signal that is partly coherent with both the diurnal and semidiurnal parts of the tide-generating potential, calculated at the appropriate times, together with a random noise due to tracking errors. One would attempt to extract perhaps as many as eight arbitrary constants to describe the coherent signals in terms of "admittances" to the potential, with one further arbitrary constant to absorb a residual error in the geoid.

Applying numerical filters aimed at isolating the M_2 tide is similar in principle (ref. 3–58). With orbital height errors of approximately 0.5 m and tidal amplitudes of approximately 1 m, one could obtain solutions of reasonable accuracy from about 50 passes. Larger numbers of passes give greater accuracy or permit more refined definition of admittances. Several possible variants could be considered. The admittances could be treated as arbitrary functions of distance along an entire oceanic track, or as localized functions of two dimensions, constrained to satisfy the tidal dynamical equations. In some places, the diurnal tide could be ignored. The ultimate result would be empirical tidal measurements in a network of grid areas covering a chosen zone.

The other approach would be to compare every traverse of an ocean, on whatever track, with the corresponding simultaneous solution of a given computer model of oceanic tides, again after removing the major part of the geoid and other variations assumed known. The residual variance about space-located averages would be compounded because of tracking errors and errors in the computer model. The computations would be repeated for the same data with variation of frictional and other uncertain parameters in the model until the residual variance was minimized. One could then say that the model output was "correct" for the ocean area considered, and a refinement of the geoid would also result.

In any approach to tidal analysis of altimeter signals, the accuracy of the solution will obviously be a function of the tracking error variance divided by the number of track repetitions. Any means of reducing this quotient is beneficial and will extend the scope of any investigation; for example, to regions of smaller tidal range. With current practices, vertical errors can apparently be as little as 0.2 m in the vicinity of tracking stations, worsening to 1 to 2 m in areas remote from these stations. Increasing the number of tracking stations would have obvious advantages. However, oceanic zones exist where the tide is already known to an accuracy of 0.5 m or better, either from pelagic measurements or by approximate calculations where the tidal range is known to be very small. These zones could be used as "surface truth" to identify the tracking errors where they might otherwise be large. A program can also be envisioned of pelagic tidal measurements in oceanic zones selected for this purpose. (The surface measurements do not necessarily have to be taken simultaneously with the altimetry.) However, 0.5-m accuracy is hardly sufficient to resolve tidal signals in the controversial South Atlantic (fig. 3–31), where, according to island data, the amplitudes are approximately 0.3 m (ref. 3–59). Such areas would have to be computed from altimeter measurements along boundary zones where the tidal range is larger.

Applications

Finally, assuming that all errors can be reduced to useful levels and all the analysis has been done, what applications would knowledge of ocean tides have, both to science and to the practical world of mankind? The scientific uses have been outlined by Munk and Zetler (ref. 3–60), and their conclusions are restated here.

1. The knowledge of oceanic tides on a global basis is necessary to properly interpret nearly all geophysical measurements at tidal frequencies, including measurement of gravity, magnetic field, etc.
2. Measurements of oceanic tides will lead to improved tidal models and ultimately to improved calculations of tidal dissipation in the ocean, which, when subtracted from dissipation obtained from changes in the length of day, will yield the dissipation in the solid Earth. These measurements will provide important information concerning the plastic properties of the solid Earth.
3. Oceanic tidal currents, when coupled with electric and magnetic field measurements on the ocean bottom, can probe the conductivity (and thus temperature) within the upper mantle as a function of frequency (depth). Horizontal gradients of temperature should exist near volcanic belts, in areas of downthrusting plates, and at postulated upper mantle "hotspots."
4. Finally, the response of the Earth to the shifting weight of the oceans, as measured by Earth strain or tilt meters, gives information about the local elastic structure of the Earth. This, too, is expected to vary near margins of lithospheric plates.

Knowledge of oceanic tides at the boundaries of shallow land-bounded seas enables better computation of the tides and tidal streams in such seas for purposes of navigation of ships, tidal power-generating schemes, and flood warning (see the following section). The inland Earth tide, which requires the oceanic tide for proper computation, is of value to geodetic leveling, surveying procedures, and the accuracy of astronomical siting.

STORM SURGES AND WIND SETUP

Storm surges may be defined as transient anomalies in sea-surface elevation (and currents) caused by storms, with time scales in the region of 10^4 to 10^5 sec, horizontal scales of approximately 10^2 to 10^3 km, and vertical excursions as high as approximately 6 m. The associated currents may be regarded as determinable from the gradients of elevation and will not be considered here as independent variables. Storm surges are largely confined to shelf seas, partly because the dynamic effect of wind stress is inversely proportional to water depth, and also because critical wave speeds in shallow water are comparable with speeds of propagation of weather systems, causing more efficient coupling with the atmosphere than with the deep ocean. Storm surges have been intensely studied in certain sea areas where they present a frequent danger of flooding to adjacent lowland, such as in the southern part of the North Sea, the Adriatic Sea near Venice, and the shelf seas bordering the Eastern and Southern United States. Flood danger also depends on the coincident state of the tide, and, to some extent, study of storm surges and local tides are inseparable.

An important ingredient of some storm surges is a quasi-static pileup of water toward the coastline, caused by the onshore wind stress and the radiation stress in the associated waves. This effect is usually called "wind setup." In some areas, where wind strength and direction persist for many days and coastal topography prevents the formation of steady currents, wind setup may form a steady deformation of sea level of some decimeters. However, although such a setup may not be thought of as a storm surge, its dynamics are not essentially different.

When tidal predictions were sufficiently well developed for sea level to be approximately partitioned into "astronomical tide" and "surge residual," researchers observed that the latter was to some extent a direct function of local weather and tried to express it as a simple linear regression of wind components and local atmospheric pressure. Results varied from passable to poor, and obviously neglected the spatial as well as the time history of the weather patterns and the transmission of energy from other parts of the shelf in the form of a free wave (sometimes called an "external surge"). More complex response operators, which take into account the spatial and temporal wind and pressure fields, have been developed with some success, but scientific practice has tended to veer away from such empirical approaches, whereby "prediction constants" are estimated from data by a least-squares calculation. The modern tendency is to predict storm surges by numerical solution of the wave equations over a network of small areas covering the shelf sea concerned; at each area, the three components of atmospheric stress are specified at regular increments of time.

State of the Art

Three examples of disastrous surges are shown in figures 3–32 and 3–33. Figure 3–32 shows elevations in the North Sea computed over a numerical grid from the dynamical equations. The surge of February 17, 1962, reached more than 3 m above tide level and caused severe flooding in the Cuxhaven-Hamburg area of Germany. Figure 3–33 shows the surge height, exceeding 7 m, along the U.S. gulf coast during the passage of Hurricane Camille.

The principal features and causes of storm surges are fairly well understood and, within certain limits, can be roughly reproduced by integrating the wave equations with driving stresses from the wind and pressure fields. But the demand continues for further research and information to help understand the finer scale features, such as their interaction with the tide and with coastal topography, and the response to unusual weather conditions. Because wind stress is the most important factor, success in predicting or understanding storm surges depends on the accuracy with which the winds over the sea can be specified. A serious hindrance to research is the lack of observational data of surface elevations offshore. With few exceptions, nearly all the observations of surges come from coastal tide gages, but these observations can give a distorted and/or limited picture of their main structure in the open sea because waves and currents cause steep gradients near the shoreline. Like the ocean tide problem, the logistics of offshore recording are difficult. Stormy shallow seas present a particularly hostile environment for recording equipment and data transmission systems.

Active microwave technology could help research on storm surges in two ways. Wind determination over the sea by the "scatter-

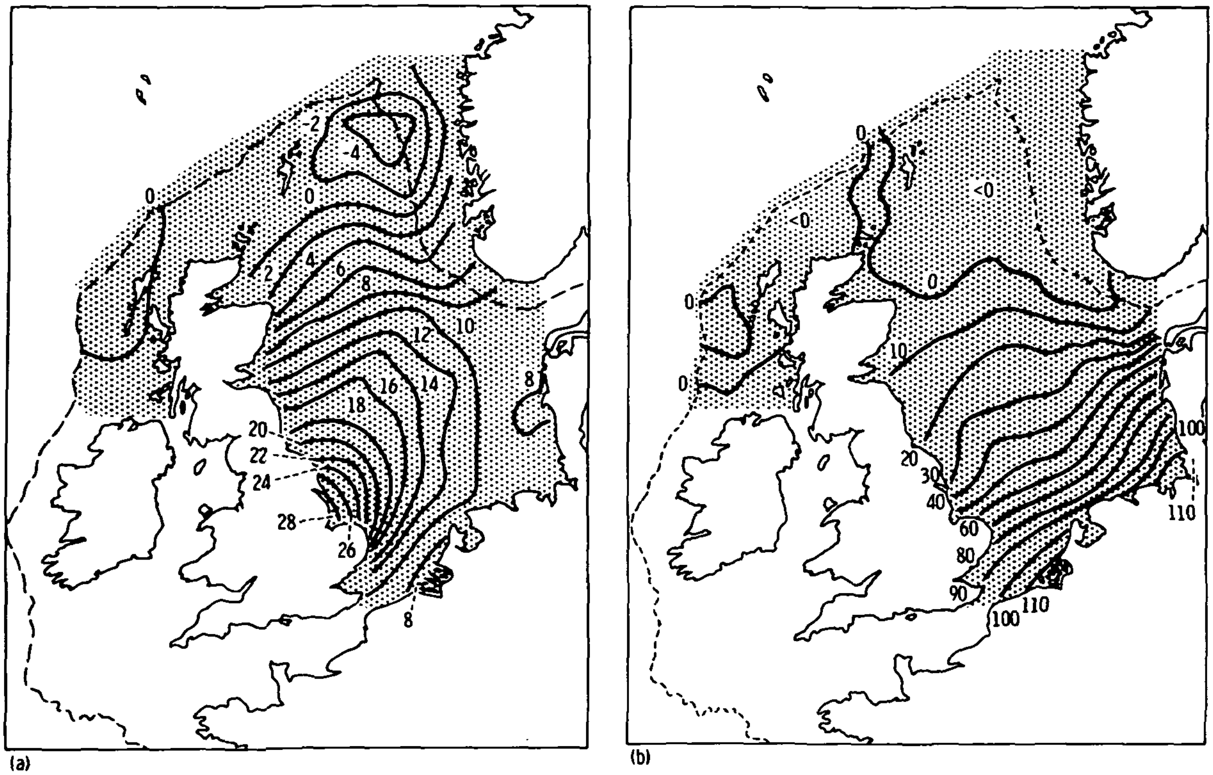


FIGURE 3-32.—North Sea storm surge. (*a*) September 14, 1962. (*b*) September 17, 1962.

ometer" technique would give a more accurate picture of the driving forces than is usually obtainable from computations based on the pressure field, and altimeter tracks over surge-ridden seas would give valuable information on the open-sea structure of the surge itself.

The scatterometer data would be more frequently available, because the wide swath of the instrument would include a given sea area on a large number of tracks. By the nature of altimeter tracks, the chance of sensing a large storm surge, which may occur only once or twice in a year, would be correspondingly less. However, as in the case of tsunamis, a single fortuitous track could provide useful data for checking a computer model that would be virtually unobtainable by conventional methods. Any track across the sea in calm weather would naturally provide useful information about the tides.

Requirements

Requirements of the scatterometer would be at least twice-daily coverage of a given sea area with dimensions of approximately 1000 km, with interpretation to windspeed over a grid spacing of approximately 100 km. To improve on surface-based estimates, windspeed would require precision to approximately 2 to 3 m/sec. Because, in storm conditions, the wind field may change considerably in a few hours, the winds measured by the satellite would not in themselves be sufficient to follow the full time history of a storm surge, but the data for intermediate times could be supplied from a computed weather model, which would itself be improved by the scatterometer data.

The altimeter "footprint" would need to be smaller than that required for oceanic tidal research. A diameter of approximately 20 km would be satisfactory. Vertical precision

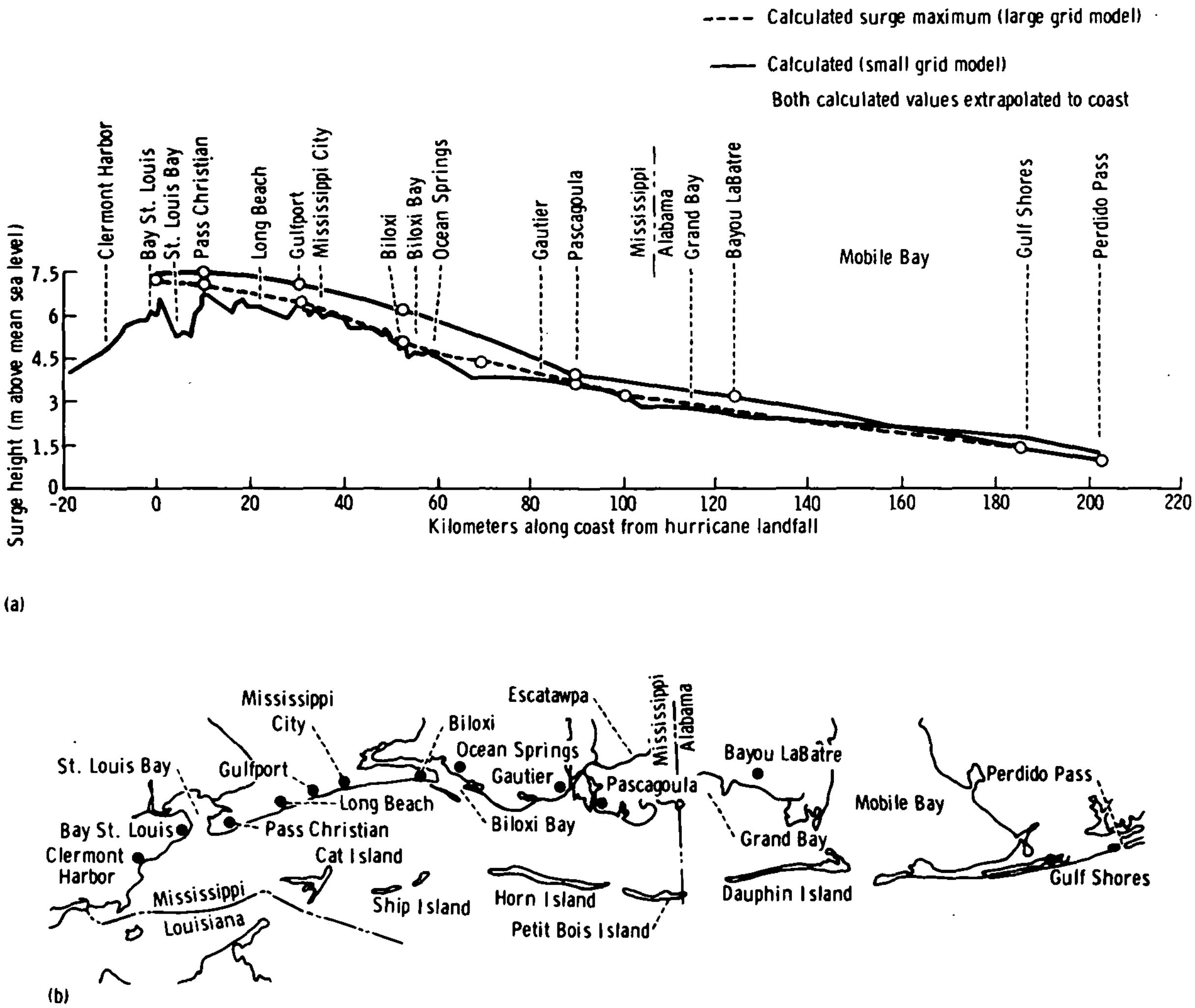


FIGURE 3-33.—Comparison of observed and calculated surge height maximums along the Gulf Coast during Hurricane Camille. (*a*) Surge height compared to distance along coast. (*b*) Geographical map.

of 0.5 m in the altimeter signal would be useful for measuring a large surge, but because the chances of encountering a large surge are slight, one would like in any case to record the more frequent small surges with an accuracy of 0.2 m. At the same time, orbital tracking errors and other systematic errors that do not vary by more than a few centimeters in 100 km can be neglected, because the seas of interest are always well provided with coastal tide gages and only the gradients of the sea surface are really required from the altimeter.

Applications

One application of these data would be to improve computer models designed to forecast storm surges. These models in turn are used as a basis for warning the public of danger from floods and for alerting sea defense operators. Another use of these models, of growing importance, is in warning large ships of unusual shoaling caused by "negative surges"; that is, depressions in sea level associated with the more familiar positive surges.

CURRENTS

Until recently, the methods of study in oceanography were limited to the standard shipbound tools. By comparing the area that can be covered by a ship in a reasonable amount of time to the total ocean, trying to understand the entire ocean by using the standard methods is equivalent to trying to understand the nature of a very large forest by painstakingly examining each tree. Many of these methods and their shortcomings are discussed in detail by Fomin (ref. 3–61) and Neumann (ref. 3–62). This does not imply that the traditional methods are useless; in understanding some particular aspects of the ocean, they are actually indispensable. However, as far as the global ocean model or large-scale phenomena are concerned, the traditional methods are not effective enough to produce a synoptic view. This shortcoming can be compensated to a large extent by the newly developed remote-sensing techniques.

Remote-sensing techniques are the only observation methods that are capable of fast scanning over a vast area. Because of this unique capability, these techniques will become the most important tools in large-scale physical oceanographic research. However, because of the way data are collected, all the information thus obtained is confined to the surface. But the environment of the ocean is a complicated one; all the physical processes in the ocean are controlled by both surface and subsurface parameters. These parameters act and interact to produce the phenomena actually observed. Remote-sensing techniques are only effective in measuring those phenomena controlled by surface parameters. Fortunately, such phenomena include most of the crucial problems in physical oceanography. Furthermore, some subsurface phenomena such as internal waves can also be inferred indirectly from microwave measurements of the surface.

The most severe deficiency of the traditional methods is the time factor. All ocean phenomena change with time as well as space. The traditional methods, because they are shipbound, are highly time constrained. They cannot produce any synoptic data of global scale. As a result, the so-called global data are actually patched-up works from different times and different cruises. For example, the meandering of the Gulf Stream is a well-known feature of the motion. If, as shown in figure 3–34, the stream path from A to B is P_1 at time t_1 and a ship surveys the area in time interval t_2-t_1, the path changes continuously to P_2 in the same time interval. The result of the survey will produce a path P that is not a path of the event at any time.

Another view of this deficiency of time factor is the spatial restriction. To cover a reasonable area in a given time interval, the ship will have to move with a minimal amount of delay, thus occupying fewer stations. Therefore, the gaps will be wide. This discrete set of data will create problems when interpolation is used in processing the data.

The combined effect of the space and time deficiencies is to eliminate the possibility of getting field data to improve the accuracy of global condition prediction, which depends not only on a local system but also on the interrelation between different systems. The Gulf Stream, for example (fig. 3–35), is dominantly controlled by the wind field of the northern Atlantic as a whole, rather than by local winds. However, events in the south-

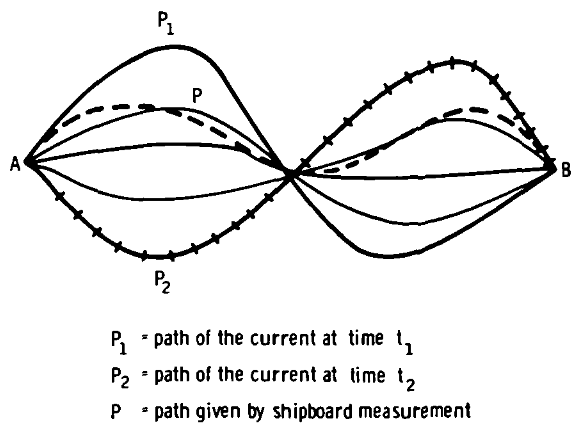


FIGURE 3–34.—Sketch illustrating problem with traditional methods of data gathering. Time for a ship to travel from A to B is t_2-t_1.

ern Atlantic will eventually feed to the Gulf Stream through current systems (nos. 15, 2, 20, etc., in fig. 3–35).

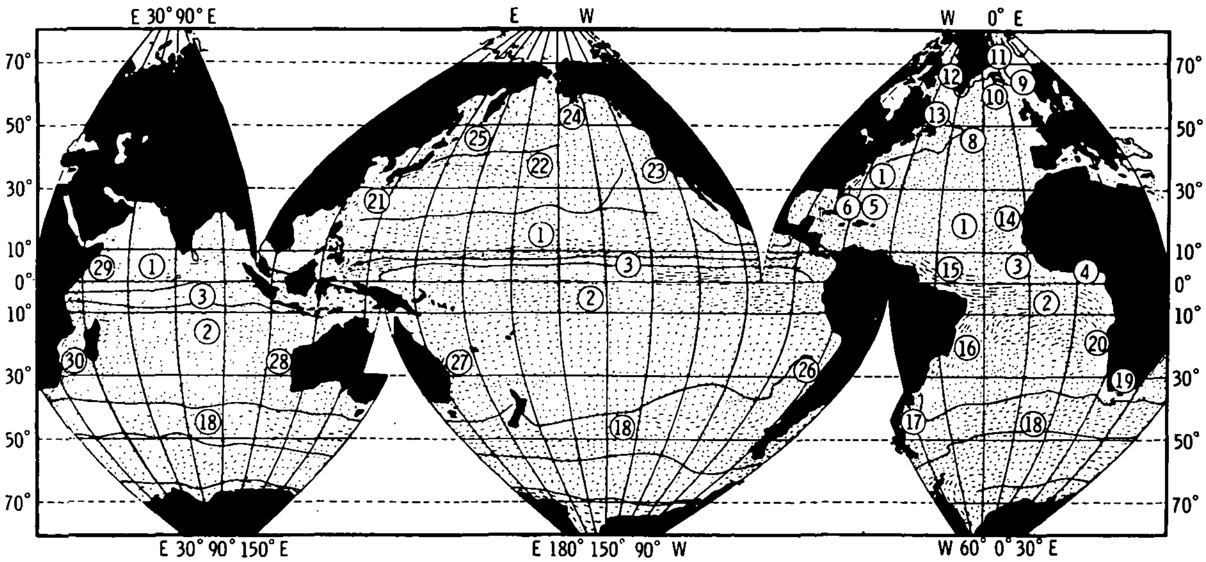
Although severely limited, the traditional method can produce data on the water column under the surface. Such information is important in studies on interval waves and baroclinic current, etc.

Advantages of Remote Sensing

Because remote-sensing techniques are capable of rapidly scanning large areas, they can be used to obtain continuous data on a global scale. Such data are essential to the predictive models on weather and sea state. From remote sensing, useful data can be obtained concerning the mean sea level, the sea state, and the surface temperature field. (Also see fig. 3–36.)

The first method of measuring ocean currents by remote sensing was by satellite tracking of drifting buoys. This was the French meteorological EOLE satellite system (Cooperative Application Satellite CAS–A) that was launched from the NASA Wallops Island Station on August 16, 1971. The EOLE data collection and positioning system has been operational since early September 1971. The satellite tracks and collects data from drifting buoys during each pass by interrogating the buoy transponders. The stored range and range-rate observations are later transmitted to ground telemetry stations and processed at the Centre National d'Etudes Spatiales (CNES) computing center in Brétigny, France. In the future, the observations will be processed at the NASA Goddard Space Flight Center.

Although originally designed for balloon tracking, the EOLE system has proved to be an ideal tool for applications such as drafting objects positioning (icebergs and



Note: 1 = North Equatorial Current, 2 = South Equatorial Current, 3 = Equatorial Countercurrent, 4 = Guinea Current, 5 = Antilles Current, 6 = Florida Current, 7 = Gulf Stream, 8 = North Atlantic Current, 9 = Norwegian Current, 10 = Irminger Current, 11 = East Greenland Current, 12 = West Greenland Current, 13 = Labrador Current, 14 = Canary Current, 15 = Guiana Current, 16 = Brazil Current, 17 = Falkland Current, 18 = Antarctic Circumpolar Current, 19 = Agulhas Current, 20 = Benguela Current, 21 = Kuroshio, 22 = North Pacific Current, 23 = California Current, 24 = Aleutian Current, 25 = Oya Shio, 26 = Peru Current, 27 = East Australian Current, 28 = West Australian Current, 29 = Somali Current, 30 = Mozambique Current, 31 = (Indian) Monsoon Current (Northern Hemisphere summer).

FIGURE 3–35.—World chart of ocean currents during Northern Hemisphere winter (ref. 3–62).

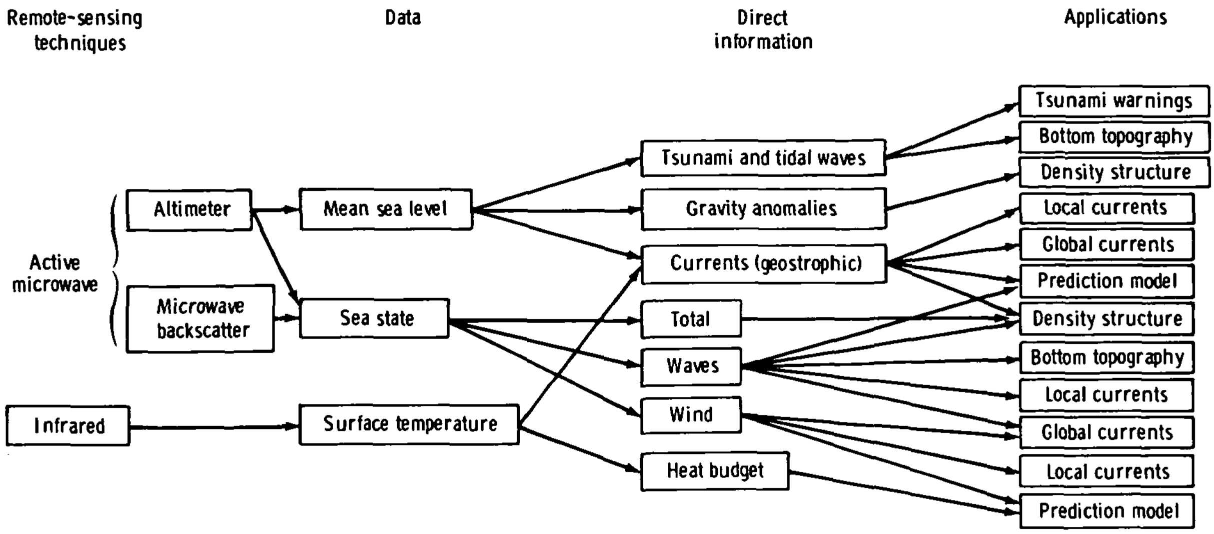


FIGURE 3–36.—Uses of remote-sensing techniques.

buoys). Routine operation can provide positions with accuracies of approximately 2 to 3 km.

The Virginia Institute of Marine Sciences started drifting buoy experiments using the EOLE positioning system in September 1972. The main objective is a study of currents along the coast of Virginia, outside the mouth of the Chesapeake Bay. The buoys are round disks carrying the radio and EOLE antennas, under which is attached a cubic case containing the electronics. A 5- to 20-m cable links it to a submarine crosslike sail.

Although the drifting buoy system uses satellites for data acquisition, it has many of the disadvantages of the traditional methods discussed previously. In this sense, this system does not offer many of the real advantages achieved by remote sensing.

The most promising method for measurement of ocean-surface currents has been developed from increasingly accurate radar or active microwave altimeters. The basis for current measurement using satellite-borne altimeters is that most important ocean currents are driven by changes in the ocean-surface elevation.

By measuring the surface slope alone, the surface-current velocity can be determined. This calculation is not new, but the practical implications of this approach have only recently been realized through the possibility of accurate measurements of ocean-surface slope by satellite-borne radar altimeters. The accuracy to which the slope is measured directly determines the accuracy of the surface current.

Measurement of surface current seems simple; however, difficulties arise because data obtained from satellite altimeters still contain substantial amounts of noise. To analyze the data and to calibrate the instrument, the satellite measurements must be related to models of the ocean-surface relief. To do this, one must rely on in situ hydrographical data available in the literature. The most useful data are measurements that lead to surface relief by means of dynamic height.

Requirements

The expected change in surface elevation due to geostrophic currents for areas of well-defined current integrity is approximately 0.5 to 1.0 m change in elevation over distances of 25 km or larger. Current features in these regions, such as eddies or weaker subcurrent systems, cause a height variation of approximately 25 cm. To detect the transient topographic signatures of these magnitudes, the permanent topographic features must first be determined to somewhat higher resolution. On the basis of defining permanent features to within 10 cm with 80 percent confidence (over a 25-km spatial extent), using nominal satellite groundtrack velocities, the required resolution translates to 12 cm rms for altitude samples that are statistically independent for a 1-per-second data rate. In some special situations, the height change may occur over distances less than 25 km, in which case a resolution (footprint) of approximately 1 to 3 km is recommended.

Technical Approach

The present technical approach is to measure the surface elevation and use this measurement to compute surface currents. Computed currents can be compared with currents from in situ measurements of dynamic height or measured elevations compared with measured values of dynamic height. At places where the density profile is known, the subsurface currents may also be computed. The accuracy of the surface current computed by this method depends on the accuracy of the measured height and the geostrophic assumption. In most situations, the geostrophic assumption is good and the accuracy of the surface elevation is the only parameter affecting accuracy.

If subsurface currents are desired, then the density profile is also required. The density profile cannot be measured remotely; therefore, any computation of subsurface currents will require local measurements.

The present approach is based on data obtained solely from an altimeter. The increased development of remote sensing will permit the measurement of many parts of the wave spectrum. Presently, the high-frequency or high-wave number part of the spectrum can be measured by active microwave backscatter as shown by Krishen (ref. 3–63), Daley (ref. 3–64), and Valenzuela et al. (ref. 3–65). In addition, part of the low wave number spectrum can be obtained for an altimeter by the measurement of significant wave height $H_{1/3}$ or by imaging radar. Huang et al. (ref. 3–66) have shown that there is a relation between nondirectional wave spectrum, surface wind, and current. Moreover, they have determined the change in wave spectrum with surface current using windspeed as a parameter. This was done by using the Pierson-Moskowitz-Kitaigorodskii (ref. 3–67) spectrum for the zero current condition. Figure 3–37 shows the wave number spectrum as a function of surface current u for a windspeed W of 10 m/sec. If the surface wind and the spectrum can be obtained, this would provide an alternate method of current measurement.

It may not be possible, at least in the near future, to remotely sense the entire wave spectrum. Because of this, Huang et al. (ref. 3–66) have computed the relation between rms surface roughness and current. Surface roughness also depends on surface wind, and figure 3–38 shows the rms slope as a function of windspeed with current speed u as a parameter. In principle, surface roughness could be used for current measurements. Figure 3–39 contains the same information as figure 3–38, but with windspeed as the parameter.

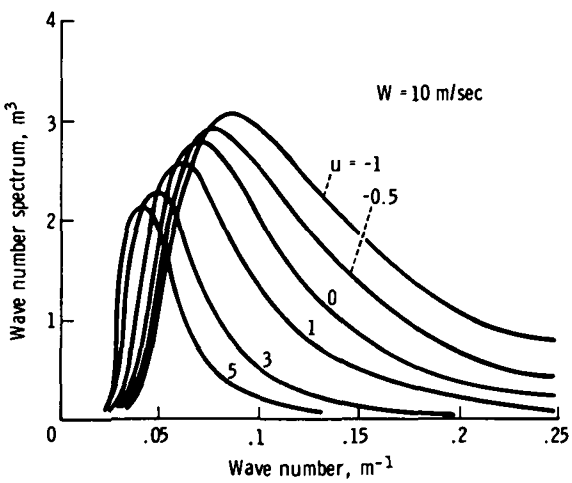


FIGURE 3–37.—Changes of wave number spectra for a 10-m/sec windspeed under different current conditions.

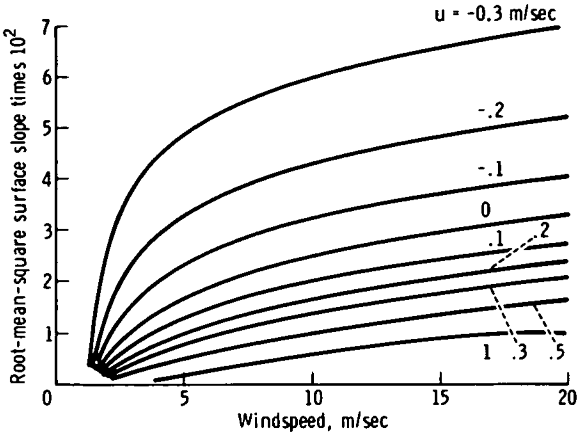


FIGURE 3-38.—Variation of rms surface slope with windspeed, using current speed as a parameter.

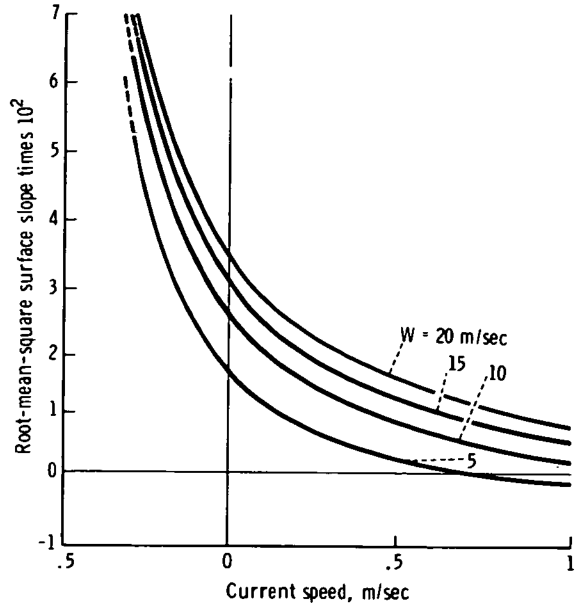


FIGURE 3-39.—Variation of rms surface slope with current speed, using windspeed as a parameter.

The surface roughness or the energy in the high wave number waves increases quite rapidly with wind (see figs. 3-38 and 3-39). Thus, an increase in the energy in the high wave number region can be caused by either adverse currents or the wind. At high windspeeds, the effect caused by the wind will probably overshadow any change in wave spectra caused by currents. Therefore, it is postulated that only in low winds can surface roughness be used to measure current. At this time, however, additional efforts must be made to separate the two effects. This alternate method of current measurement, at least at low windspeeds, can be used as a check (also obtained by remote sensing) on the currents obtained from the satellite altimeter.

The ultimate goal must be remote sensing of the complete directional wave-height spectrum. This is slightly different from the present approach of measuring the high-frequency spectrum by means of backscatter and of inferring wind by a relation between wind and the wave-height spectrum at specific (high) wave numbers. If, by using various remote-sensing systems and possibly models of the spectra, the total wave-height spectrum can be obtained, then the high-frequency portion could be used to infer wind; and the rest of the spectrum, including the high-frequency portion, would be used to measure currents and other parameters of interest (e.g., wave energy, significant wave heights, etc.).

Applications

Virtually all ocean waves are influenced by ocean currents and, thus, information is required on these currents. Any activity using ships will be able to use surface current information to reduce time en route, to lower costs or fuel consumption, and to avoid bad weather. Some examples are shipping and naval operations, fishing operations, and search-and-rescue operations. Access to measured or predicted ocean-surface currents will be of immediate benefit to all these users.

Long-range weather forecasting and/or weather models also require a global picture of the surface currents. The proposed current information will benefit both. As the ocean becomes more crowded, pollution and the ability to predict pollution becomes increasingly important. Surface currents are also needed for predictions of pollution from any known source. Moreover, slicks detected by a measurement of the high wave number spectrum can reveal pollution or salt intrusion.

GLOBAL WAVE STATISTICS

Any dynamic phenomena on the ocean surface, such as ship motion and forces on anchored or stationary platforms, are governed by the actual and/or expected sea state. Because the ocean surface is random, the only way to define it is by statistical description. The usual method used to describe wave motion is by the wave-number or wave-frequency spectrum and the expected variation of this spectrum under different conditions. Therefore, models to predict the expected wave spectra for given geographical and meteorological conditions have received much attention.

The primary mechanism by which waves are generated is the surface wind. Moreover, in the open ocean (i.e., without boundaries) with no currents, the wind uniquely determines the wave spectrum. However, the relationship between wind conditions and the sea state is still not completely resolved. When boundaries or currents are present, the prediction of wave statistics becomes increasingly difficult.

Because of the inability to uniquely predict the wave spectrum, the approach must be to measure the wave spectrum, or wave statistics, so that mathematical models may be refined by observations and thus yield the desired data. The measurements must be made on a global scale because some of the factors (e.g., storms, geostrophic currents, etc.) that affect the spectrum extend for hundreds of kilometers. Because of these large-scale processes that influence the sea state, measurements at a single point or at several points at different times are difficult to interpret. Therefore, remote sensing offers a great many advantages in developing global wave statistics. Such statistics have two immediate and widely useful purposes:

1. To establish reference conditions to be used for design and planning, and for evaluating global wave models.
2. To monitor any changes from the reference condition, such as those caused by storms, currents, topography, pollution, etc. The change in wave spectra can be used to evaluate the size and nature of these phenomena.

A less-immediate goal is the generation of improved methods of predicting and modeling ocean-wave spectra.

State of the Art

The first work on a model to predict ocean-wave characteristics was undertaken by Cornish (ref. 3–68). Little work followed this until Sverdrup and Munk (ref. 3–40), motivated by the necessity of wave information for naval operations, developed a model to predict wave characteristics. Their analysis was concerned with a relation between wave heights and periods and did not deal with the concept of a wave spectrum. Neumann (ref. 3–69) was the first to deal with the concept of a wave-frequency spectrum from which statistical properties could be obtained. Using Neumann's concepts, Pierson et al. (ref. 3–70) developed a model for the prediction of wind-driven sea and ocean swell spectra. The model has experienced considerable modification and improvement and still remains the most widely used model for forecasting ocean wave characteristics or statistics. The concept of an equilibrium spectrum was introduced by Phillips (ref. 3–71). This provided additional insight into the parameters that must be included in the gravity wave spectra and on the shape of the high-frequency part of the spectrum. A recent summary of wave spectra models for a complete range of wave numbers has been given by Pierson and Stacy (ref. 3–15).

Generally, wave spectra models have been quite successful. For example, a NOAA National Data Buoy Center report (ref. 3–72) provides analyses of ocean-wave spectra as determined by hindcasts. The development of these models has required the acquisition of a large amount of information about waves. Data concerning waves are routinely collected by ships at sea. Wave heights, periods, and dominant direction of travel are recorded and radioed ashore in a special code by many ships at 00:00, 06:00, 12:00, and 18:00 G.m.t. These measurements are collected, stored at national data centers, and used to produce statistical summaries of the waves for the various areas of the oceans. The cost of collecting, storing, and analyzing data of this kind amounts to millions of dollars annually.

Also, a very small number of weather ships have wave recorders that measure the rise and fall of the sea surface as a function of time at a point. These spectra have been studied and used to develop wave-forecasting techniques and to aid in the design of ships. Wave records of this nature are obtained routinely at four or five locations in the North Atlantic and at one location in the North Pacific.

A consortium of oil companies has used wave recorders on their drilling rigs to collect a large amount of wave data during hurricanes in the Gulf of Mexico. The substantial effort to obtain wave data in these ways is a measure of its present value. However, this level of effort is inadequate for the future needs of the United States and the other nations of the world.

The reports from ships are estimates, not measurements, of the wave conditions. They are also concentrated along shipping lanes and do not adequately describe conditions on a uniformly spaced grid over the oceans. Moreover, they tend to be substantially in error, and the data obtained before a few years ago had built-in biases that tended to group reported wave heights near 5 and 10 m and to report high waves inadequately. In one study, in which these estimates of wave heights were compared with measured wave heights, the estimates frequently differed by a factor of 2 (either too high or too low) from the measured values. Approximately 1200 ships report wave conditions in this way every 6 hr, primarily in the Northern Hemisphere.

The shipborne wave recorder data are of high quality and have proved invaluable for many scientific investigations. However, the number of ships is too few and their locations too random to provide an adequate data base.

In addition, predictions of wave spectra are based on local conditions, primarily because this is all the information that is available. However, very-large-scale phenomena (storms, global currents, etc.) alter or determine the wave spectrum. For this reason, recourse must be continually made to observations of measured wave characteristics typical of those given by Hogben and Lumb (ref. 3–73). Moreover, in coastal regions, the open ocean waves must be considered in combination with local topography to develop realistic estimates of the wave spectra.

If global meteorological data were available, more accurate models for the prediction of wave spectra could possibly be developed. Such models would need to be verified by global wave statistics, however. Therefore, remote sensing appears to not only offer the opportunity to evaluate and improve wave spectra but to be the only way presently available to do so.

Requirements

The complete wave frequency or number spectrum need not be measured. Various parts of the spectrum must be measured, however, so that present models can be used to infer the complete spectra. In addition to the measurements of wave spectra or wave statistics, data on meteorological conditions are also required. These data are needed not only at the point of wave measurement but also globally, because of the large-scale phenomena that govern local wave spectra.

Technical Approach

The acquisition of sufficient data to develop a global picture of the local wave spectrum and its expected extreme is required. By using presently available models (e.g., Pierson et al. (ref. 3–70) and Pierson and Stacy (ref. 3–15)), a complete wave spectrum can be developed from the significant wave height $H_{1/3}$ or the surface wind. Both together permit some checking of the models. The significant wave height can be obtained by the present satellite altimeter, although the accuracy may not be as great as desired. Surface wind can be obtained from microwave backscatter, but a higher accuracy may be needed.

Imaging radar can also be used to generate wave statistics. Imaging radar gives a picture of wave patterns that is at present a qualitative picture of the wave field. Therefore, the fact that waves can be seen makes the imaging radar a powerful tool for the oceanographer. The most exciting prospect is that of mapping wave patterns and wave buildup during large storms. This wave buildup occurs under cloud cover that only active microwave sensors can penetrate. It is in these large storms that the bigger, more damaging waves are generated.

In addition to mapping storm waves, the mapping of swell over continental shelves will permit the study of wave refraction in coastal regions. The continental shelves will be valuable real estate in the next few decades as a place to put man's machinery, such as offshore airports, powerplants, and oil rigs. The ability to predict wave climates in these areas and to verify these predictions by observations is another important contribution that imaging microwave sensors can make to oceanography. Again, the ability of active microwave sensors to penetrate clouds during storms is a most important attribute.

The capability for global monitoring of sea state and ocean waves, which will be provided by spaceborne imaging radar, is also a very important contribution. This is particularly true for oceans in the Southern Hemisphere, where wave measurements in the open ocean are so widely scattered that many areas of the ocean are essentially unknown. Little shipping and almost no land observation points are present in the Southern Hemisphere oceans between latitudes 20° and 40°, so that the interaction between wind and waves and the buildup of waves is practically unmonitored.

All of this information can be used to generate a global picture of the expected wave spectra or the wave statistics characteristic of these spectra. These data would establish a reference state for the ocean. This state could be considerably more accurate and detailed and with better resolution than the present tabulation by Hogben and Lumb (ref. 3–73) or those from the U.S. Department of Commerce (ref. 3-72).

With the acquisition of a reference condition, continued monitoring of the oceans can be used to reveal any changes that occur because of storms or other large-scale phenomena. These two goals, the development of more accurate global wave statistics and the monitoring of the magnitude of their changes, should be ample justification for the program.

The many advantages provide motivation to attempt to sense more completely the total wave spectrum. At present Krishen (ref. 3–63), Valenzuela et al. (ref. 3–65), and Daley (ref. 3–64) have used microwave backscatter to measure the high wave number (above 0.1 cm^{-1}) part of the wave spectrum. The use of the longer wavelength radar, or perhaps imaging radar, to sense the lower frequency portions of the spectrum should be considered. This is because the ultimate goal must be a tabulation of the complete wave-number spectrum and, in the future, possibly the directional wave-number spectrum, on a global basis. Active microwave systems are now the only sensing technique that can be used to measure wave spectra directly.

Applicability

The use of remote sensing to establish a reference picture of global wave statistics would provide data needed to plan shipping routes, design ships and/or offshore structures, and aid in the planning of future ports. This reference condition is especially needed in costal regions where the influence of the local topography on incoming open ocean waves makes prediction of sea state quite difficult.

The monitoring of local variations from an established reference condition probably has its most immediate applicability in the routing of ships. The financial benefits obtained by routing ships around areas of heavy seas are quite large and can be easily accomplished with present remote-sensing technology. Other advantages would be in the monitoring of large-scale storms and their influence and the tracing of various surface pollutants, such as oil spills or the intrusion of saltwater into freshwater. Over longer periods, the compilation of wave statistics and their variation can reveal any unusual or significant changes in topography. In addition, there are a number of specific coastal problems that these data can help resolve.

When the depth of the water becomes less than half the wavelength of a spectral component in a wave spectrum, the speed of the component is affected by the depth. The shallower the water, the slower the wave travels. Complex offshore submarine topography turns the waves away from deeper regions and focuses them at shallower regions. Shallow areas off the coasts of parts of the continents, called continental shelves, are as much as several hundred kilometers wide. The depths in these shallow areas are often complex, and the waves are refracted in ways both difficult to describe and to compute.

All structures to be built in such shallow-water areas require design wave data so that the structures can withstand the forces on them produced by the high waves during a storm. Also, the continued action of the lower waves can erode the material around the structure base and cause it to collapse. With the many proposed offshore structures around the coasts, the problem of adequate designs for them will become increasingly more pressing during the next few decades.

One area studied in the past and under renewed theoretical analysis is the New York Bight, especially as affected by the Hudson Submarine Canyon. Offshore sites just 10 or 20 km apart in this area can, at times, be exposed to waves 10 times higher at one site than at another.

During selected conditions with appropriate offshore deepwater waves, data can be obtained that will provide verification for wave refraction studies in shoal water. Moreover, potential sites all over the world can be surveyed concerning wave refraction effects for preliminary, and perhaps even final, design considerations. The actual images will have to be recovered because the waves become shorter and change direction as the water becomes shallower, so that the concept of a spectrum representative of an entire area is not applicable.

A scientist familiar with the needs of the user community could rather easily select approximately 500 sites on a global basis for obtaining daily data, some over the deep ocean and others at coasts, so that in the course of each year a global data base of approximately 180 000 coastal refraction patterns and deep-water wave-number spectra could be obtained.

Finally, advantages will be gained by improved predictive models to forecast wave spectra. The present numerical spectral wave forecasting models are quite good. They incorporate many physical features of the generation and propagation of waves but not all the variously proposed theoretical features of waves. However, any numerical model of a physical phenomenon of the oceans can be improved because there are always increasing levels of complexity that need to be modeled. At present, the problem is lack of an adequate verification data base on which to build improved numerical spectral forecasting models. The measurements to be made by satellites would provide the kind of data needed to develop greatly improved numerical wave-forecasting models by providing a means to determine the errors in the present models. The need will always exist for forecasting wave conditions for as many days as possible into the future. The forecast problem involves wave spectra and weather, but improvement in wave forecasts is a necessary requisite for improvement in weather forecasts.

POLAR ICE FIELDS

The polar regions are a fundamental part of the Earth heat engine, yet the way in which they interact with the other parts of the atmosphere/ocean/ice/land system is poorly known because they have existed behind what has been called an "observational barrier." Most polar geophysical studies are based on a paucity of data; relatively short-time series of measurements acquired at a few points spread over vast distances. The state and behavior of the Earth surface is one aspect of the interaction-exchange problem that can be said to be the most important in polar regions, where changes of phase of water into snow and ice occur over large areas at small times scales, and where large permanent floating and grounded ice masses are involved in a complex feedback process with atmospheric and oceanic circulation. This problem can be studied only by using satellite remote-sensing techniques.

In the vast array of remote-sensing techniques available for polar studies, microwave remote sensing, both passive and active, promises to be the most useful tool. The polar regions are in the dark for a large part of each year and when they are in their sunlit periods they are normally cloud covered. Therefore, sensors that can observe the polar surface at all times, such as microwave sensors, are needed to provide the sequential synoptic imagery needed for elucidation of the complex cause and effect processes of these vast regions.

Scope

Polar ice includes three distinct forms of ice that have widely divergent properties:

1. Sea ice, which covers large parts of polar oceans and some subpolar seas, contains brine, averages one to several meters in thickness, and ranges in age from hours to several years.
2. Icecaps and glaciers, which are composed of freshwater ice that averages several kilometers in thickness for Antarctica and Greenland and is tens of thousands of years old.
3. Lake ice and estuary ice, which is fresh and brackish water ice of a wide variety of thicknesses and ranges in age from hours to several months

In the following section, the three forms of ice are discussed in relation to active microwave remote sensing.

Sea Ice

Of all the forms of frozen water that exist on Earth, the one about which least is known is the sea ice that covers vast areas of the oceans; 10 percent in the Northern and 13 percent in the Southern Hemisphere. A good example of the paucity of knowledge is that, for the International Geophysical Year (IGY) period, essentially nothing is known about the extent and morphology of the winter sea ice surrounding Antarctica.

Individual ice floes have been observed to move 50 km in a day, and speeds of 10 to 20 km/day are common. Leads (cracks) and polynyas (large irregular openings) open and close at all times. The seasonal variations in areal extent of the ice canopies are large; approximately 15 percent for the Arctic and 80 percent for the Antarctic. In short, sea ice is the most rapidly varying solid on the Earth surface.

Although the feedback mechanism that exists within the air/ice/water system is not well defined, intuition and the few facts available suggest that the variation of the edge of the icepack must be of prime importance. On the time scale of months, air/sea interaction at the edge of the pack would cause an intensification of the atmospheric baroclinicity, which in turn would alter the surface ocean regime.

To monitor the seasonal patterns of surface heat exchange over polar oceans, one must systematically observe the location and extent of pack ice and the location and duration of large polynyas within the pack. Although meteorological satellites such as Tiros, Itos, and NOAA have given (and will continue to give) highly useful ice data, an alltime, all-weather capability of observing sea ice did not exist until the launching of Nimbus 5 in December 1972.

The electronically scanning microwave radiometer (ESMR) on Nimbus 5, operating at a frequency of 19 GHz (1.55 cm), is providing the first daily synoptic view of polar sea ice distribution. The NASA remote-sensing flights during the 1971 and 1972 Arctic Ice Dynamics Joint Experiment (AIDJEX) pilot experiments, which flew an ESMR identical with that aboard Nimbus 5, have provided data that allow the interpretation of ice types shown on the microwave brightness temperature maps obtained from space. The difference in brightness temperature between seawater and ice at 19 GHz is approximately 120 K; therefore, the ice edge on the maps shows up clearly and can be positioned geographically with an accuracy of 30 km, the nadir resolution of the 1.4° beamwidth of the ESMR.

The ice-edge positions shown in ESMR images differ considerably from the averages shown for winter months in the U.S.S.R. and U.S. Antarctic atlases. The actual ice edges are more irregular than the ones deduced from aircraft and ship reports shown in the atlases and are more extensive (farther north) in several areas, such as the Ross Sea. Of special note are the large polynyas revealed in the Ross and Bellingshausen Seas, which are not shown at all in the atlases.

Gloersen et al. (ref. 3–74) have used aircraft microwave images to show that it is possible to distinguish first-year sea ice (1.15 m thick) from multiyear sea ice (1 to 3 m thick) and to estimate amounts in mixtures of the two. Campbell et al. (ref. 3–75) have compared ESMR and aircraft microwave images of the Arctic ice canopy to delineate its gross morphology.

Although ESMR is a prime tool to study gross characteristics of sea ice morphology and dynamics, it does not provide the kind of high-resolution data needed for a wide variety of scientific and commercial purposes. To test the existing and developing numerical models for sea ice dynamics and thermodynamics, high-resolution sequential imagery of select areas is badly needed. The optimum sensors for this program would appear to be active microwave sensors.

American radar imaging of the polar sea ice dates back to the early 1960's (ref. 3–76). In a flight by a military aircraft from the North American Continent to the vicinity of the North Pole, a long strip of imagery was obtained with an X-band real aperture system. The ability of the radar to distinguish various classes and ages of ice was first shown in Anderson's analysis of these images. In 1967, the NASA Earth Resources Aircraft Program (ERAP) P–3A aircraft flew north of Point Barrow with the 13.3-GHz scatterometer. Although ground parties were on the ice to make thickness measurements, their lack of mobility prevented measurements from being more than of marginal utility, and the ice types had to be determined by analysis of aerial photographs made simultaneously with the scatterometer runs. Unfortunately, the more interesting ice morphologies occured some miles from the ice party's location. By comparing the photographs with ice atlas photographs and information gleaned from conversation with ice observers, Rouse (ref. 3–77) was able to make tentative identification of ice types and to show that the multiangle scatterometer observations could be well correlated with ice type.

In 1970, another NASA ERAP mission was conducted north of Point Barrow, with the ice party restricted to the uninteresting fast ice within a few miles of Point Barrow. In this mission, a series of scatterometer lines were flown with both 13.3- and 0.4-GHz instruments; the same area was imaged with the 16.5-GHz DPD–2 multipolarized imaging radar by flying the imager in a box around the area covered by the scatterometer. Parashar et al. (ref. 3–78) were able to make excellent identification of ice types by detailed stereoscopic analysis of photographs of the area. The classification by ice type was then converted into effective ice thickness, using the average ice thickness appropriate to each kind of ice. The results were then interpreted in terms of ability to measure ice thickness with the various radars, but most of the work concentrated on what could be done with individual incident angles appropriate to an imaging radar that might be used operationally. Figure 3–40 shows the encouraging result at 13.3 GHz. Note that there is an ambiguity between the very thinnest ice category (new ice, less than 5 cm thick) and ice approximately 1 m thick. Fortunately, this ambiguity can be resolved by image interpretation, for the new ice has a characteristic texture quite different from that of the older first-year ice. It was found that the 0.4-GHz data could resolve this ambiguity if needed, but that the lower frequency by itself could not distinguish between the thicker multiyear ice and ice approximately 0.5 m thick. Indications from image analysis of the DPD–2 were that four categories (open water, ice less than 18 cm thick, ice 18 to 90 cm thick, and thicker ice) could be readily identified on the images, even though the images were somewhat saturated. The identification was an indication that the ambiguity in intensity between new ice and meter-thick ice did not exist on the cross-polarized image, but saturation on the film prevented assurance that this conclusion was correct. Parashar et al. (ref. 3–78) have also developed a theory that seems to explain the observations.

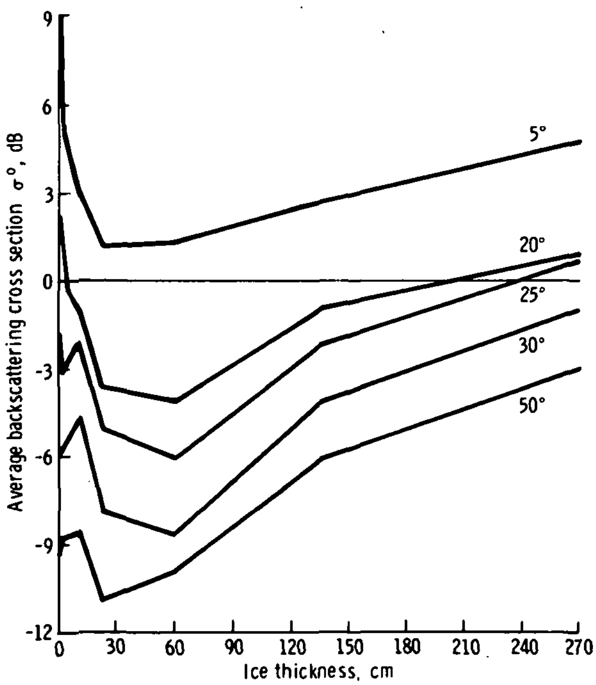


FIGURE 3–40.—Experimental σ° compared to ice thickness for different incident angles (frequency of 13.3 GHz, vertical transmit/vertical receive polarization).

The Arctic and Antarctic Institute of Leningrad has engaged in imaging of sea ice for at least 6 yr (ref. 3–79). The Soviets believe that they can clearly distinguish thin ice, thick first-year ice, and multiyear ice. In the late spring of 1973, an ice map of the entire shipping region across the north of the Eurasian Continent was prepared using the Toros 16-GHz real aperture radar imager. The system is presumably being used operationally in connection with directing shipping convoys along this route. In fact, the Soviet participants in the 1973 joint Bering Sea passive microwave experiment used the Toros images for "ground truth" to identify whether the passive microwave was able to distinguish water and ice and one ice type from the other. The Soviets have stated that the radar definitely distinguished more different ice types than did the passive microwave sensors.

The first American active microwave experiment on sea ice in which sequential images of ice were obtained took place north of Point Barrow during April and May 1973 when a joint U.S. Geological Survey (USGS)/Cold Regions Research and Engineering Laboratory (CRREL) team used an X-band side-looking airborne radar (SLAR) mounted in a Mohawk aircraft. In 2 weeks, nine flights were made over selected large areas of sea ice (200 by 50 km) to the east, north, and west of Point Barrow.

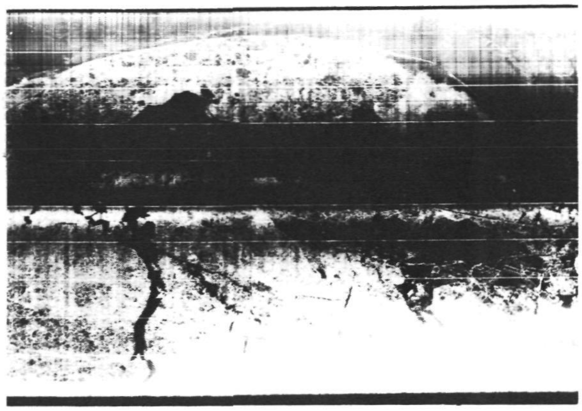


FIGURE 3–41.—An X-band SLAR image from the April 28, 1973, mission west of Point Barrow.

Figure 3–41 shows a portion of the April 28 mission data from west of Point Barrow. The coast can be seen with Wainwright at the upper left and Icy Cape at the upper right. The zone of shorefast ice shows clearly, as does the coastal lead. The pack ice is clearly seen to be composed of numerous fragmented floes and leads and has undergone strong recent deformation. The pack was made up of mostly first-year ice with some multiyear ice.

Figure 3–42 shows the same area 5 days later on May 3. The shorefast ice remained the same during this period, but the pack ice can be seen to have undergone strong deformation. Major lead changes occurred while the pack was translated only a few kilometers to the west.

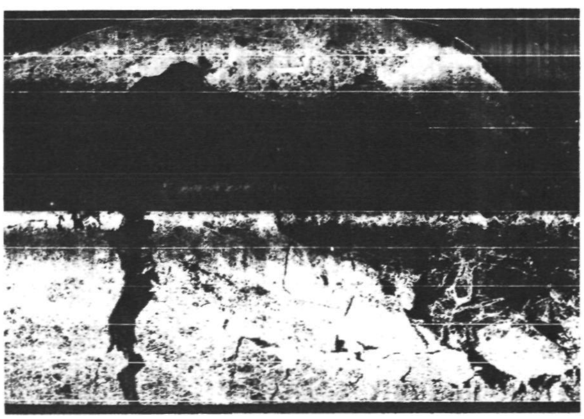


FIGURE 3–42.—An X-band SLAR image from the May 3, 1973, mission west of Point Barrow.

A comparison of figures 3–41 and 3–42 with figure 3–43 reveals that SLAR shows the difference in ice types. Figure 3–43 shows a segment of the pack ice on May 2 approximately 190 km north of Harrison Bay (240 km northeast of Point Barrow). In this image, numerous undisturbed large ice floes can be seen, and the ridges appear as white linear features. The ice in this area was far less deformed than the ice west of Point Barrow.

Although much work remains to be done on these data, SLAR remote sensing of sea ice provides discernment of the following features.

1. Leads and polynyas can clearly be distinguished from ice.

2. Thin ice (pancake, frazil, gray) can

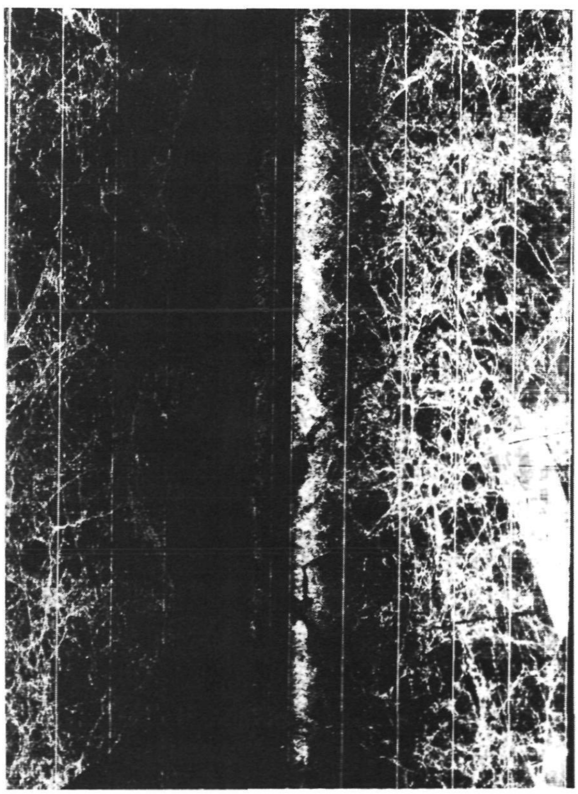


FIGURE 3–43.—An X-band SLAR image showing Arctic Ocean pack ice.

be distinguished from open water in leads and polynyas.

3. Ridges can generally be distinguished from leads, especially when the leads are large.

4. Ice floe shape and size can clearly be observed.

5. Land (permafrost) can clearly be distinguished from shorefast ice, water, and pack ice.

6. Ice-type distinction is good enough to permit distinctions between water, thin ice, thicker first-year ice, and the thicker multiyear ice.

Icecaps and Glaciers

Approximately 85 percent of the freshwater on Earth exists as ice in Antarctica and Greenland. Glaciological research in those areas has been aided greatly by the development of radio echo sounding techniques, which has made possible the delineation of bedrock topography and the measurement of ice thickness. Both these variables are needed to test recently developed numerical models. However, if the complex interaction between icecaps and glaciers and climate are to be understood, the ice/air interface events must be understood, especially the accumulation rate.

The ESMR images of the Arctic and Antarctic from Gloersen et al. (ref. 3–80) show very interesting signature variations on icecaps. For Greenland, brightness temperature differences of 323 K occur across the icecap, with the highest brightness emittance corresponding roughly to the highest elevation of the icecap. Although the cause of these variations is not fully understood, they are apparently not due to variation of the physical temperature of the ice, but to the emissivity variations probably induced by variations of the small-scale structure. Such crystal metamorphosis could be caused not only by temperature variations but also by pressure gradients within the ice.

In the ESMR images of the Antarctic icecap, no such correlation with elevation occurs; instead, the lowest brightness temperatures occur near the center of the continent, which would be expected if the variation in brightness temperatures was predominantly due to the variations in the physical temperature of the ice. Thawing and recrystallization of the surface ice does not occur on the Antarctic cap as it does in Greenland.

As interesting as these passive microwave measurements of icecaps are, they will be useful only for gross structure studies, whereas an urgent glaciological need exists for detailed measurements of the surface structure of icecaps and glaciers. It is precisely in this area that SLAR techniques could prove highly useful.

The Jet Propulsion Laboratory (JPL) has recently obtained L-band sounder profiles on Alaskan glaciers and on Greenland. The data show signal returns from structural features below the surface. The Greenland profiles show a tilted, layered structure that looks as if it had been formed by surface flow.

The only two-dimensional SLAR images of glaciers and icecaps are those obtained by V. V. Bogorodsky and V. S. Loshchilov of the Arctic and Antarctic Research Institute of Leningrad.[2] These Soviet scientists have obtained numerous SLAR (S-band) images of Siberian glaciers. All were obtained through a dry snow cover over nontemperate glaciers (those having temperatures below 273 K within the ice mass). In the great majority of images, the foliation planes formed by the annual accumulation processes were clearly visible. Most of the images were of glaciers with a snow cover thicker than several meters.

From the glaciological point of view, these data are exciting because one of the most difficult things to measure in glaciers is the rate and pattern of accumulation and ablation. As a glacier flows, the deposited layers of snow metamorphose into ice, which slowly flows downslope. Therefore, the pattern of foliation planes observed is due to two processes: accumulation/ablation and flow. Thus, to deduce the recent accumulation history of

[2] Personal communication, J. W. Campbell.

a glacier from the foliation planes, one must allow for the dynamics. This may not be too difficult for slow-moving nontemperate glaciers.

Bogorodsky and Loshchilov believe that SLAR measurements of glaciers can be used for accumulation studies. They have also obtained images of Arctic ice islands (pieces of icecaps that have calved off an ice shelf and drifted about within the matrix of sea ice floes). These images all showed the ice to have a parallel banded structure, with the bands having a spacing on the order of several hundred meters. They do not know what these bands are, but surmise that they were formed by the accumulation process existing on the icecap where the island was formed. The structural differences between ice island and sea were so great that distinguishing between the two was always possible.

Lake and Estuary Ice

Several of the world's major shipping routes are covered by lake or brackish ice for several months a year (i.e., Gulf of St. Lawrence and Baltic Sea), and great efforts are being made to extend the navigable season in these areas. To do this, two uses of sequential radar imagery are necessary: (1) as input into numerical models of ice dynamics and thermodynamics, and (2) as routing maps to direct ship traffic through the thinnest ice in an area.

A considerable part of the Asian and North American tundra is lake covered, and during the winter these lakes freeze both fully and partly (not to the bottom). Measurement of the ice cover on these lakes is important for many biological studies.

During the Skylab 4 overflights, SLAR imagery was obtained of the ice covers in Lake Ontario and the Gulf of St. Lawrence. These data show that radar remote sensing of lake and brackish ice can make it possible to distinguish the following features:

1. Leads and polynyas can clearly be distinguished from all forms of ice.
2. Rafted ice can be distinguished from undisturbed ice.
3. Shorefast ice can be distinguished from moving ice.
4. Ice floe size and shape can clearly be observed.
5. The approximate age of ice can be observed in terms of gray, gray-white, and white ice.
6. Ridges can generally be distinguished from small leads.

Bogorodsky and Loschilov, who have made SLAR X-band observations of lake ice in the Soviet Union, agree with these six tentative conclusions.

A very interesting phenomenon was observed on the tundra lakes east of Point Barrow during the USGS/CRREL experiment. In figure 3–44, a SLAR image of a series of thaw lakes to the east of Point Barrow is shown. Some of the lakes appear white (strong return), whereas others appear dark (weak return). After comparing this and other SLAR images with surface measurements, it was concluded that the lakes that appear white in the images are frozen all the way to the bottom (to the

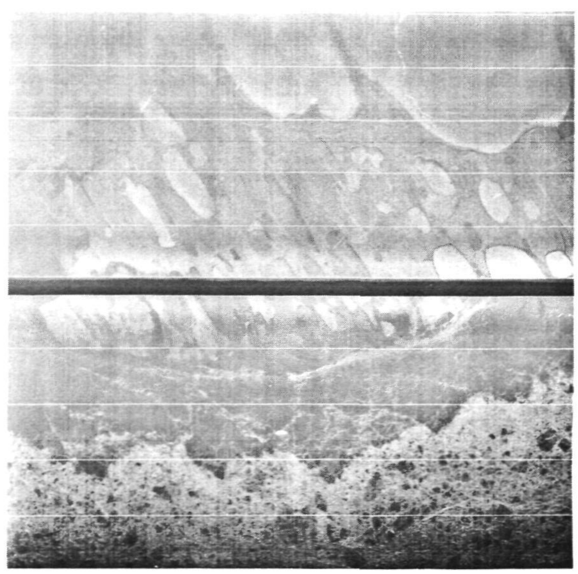


FIGURE 3–44.—An X-band SLAR image showing frozen thaw lakes east of Point Barrow.

permafrost table) and the lakes that appear dark have a water layer beneath the ice cover. This finding is important from several points of view. Such images could provide information to local communities concerning which lakes to tap for a winter water supply. The biological regime between completely and partly frozen lakes is different, the latter having far more fish.

Conclusions

Satellite-borne active microwave sensors offer the only means of obtaining high-resolution imagery of polar ice during all times, thus penetrating the "observational barrier" that has hampered much-needed research. Sequential synoptic active microwave imagery of sea ice is urgently needed to test and help develop numerical models of air/ice/ocean interaction. Such data will greatly enhance the usefulness of ESMR imagery of the gross characteristics of the sea ice canopies.

The only remote-sensing means capable of observing subsurface structures of ice-caps and glaciers is active microwave. The SLAR images of glaciers clearly delineate the accumulation foliation planes through several meters of snow cover. Those of ice-caps reveal a subsurface parallel banded structure. Such images promise to provide needed data on accumulation amounts, and patterns may help in determining surface flow patterns. Satellite radar is the best way to track and study the metamorphosis of ice island (bergs).

Active microwave imaging of lake and estuary ice will provide high-resolution, sequential, synoptic data needed to extend the navigable season in busy waterways such as the Gulf of St. Lawrence and the Baltic Sea.

The SLAR can be used to determine which tundra lakes are frozen to the bottom and which are not, an important logistical and ecological technique.

Although the potential of active microwave sensors that currently exist has been demonstrated as ice sensors, neither U.S. nor U.S.S.R. scientists yet know what frequency (or combination of frequencies) or what polarization (or combination of polarizations) is best for monitoring either sea ice or lake ice. Although earlier theories might be used to permit some extrapolation or interpolation, these too need improvement and further verification. Hence, in addition to further flights over the ice, both theoretical research and research with a microwave spectrometer that can be transported onto the ice seems to be needed if the optimum system parameters are to be established.

PART E

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TECHNICAL APPROACHES

INSTRUMENTATION

Radar altimeters, scatterometers, and imaging radar are the major instruments needed to accomplish the applications discussed earlier in this chapter. This section describes the instrumentation in terms of its functions, future developments, constraints, and applications.

Altimetry

Introduction.—Satellite altimetry has demonstrated the capability for determining mean sea level. Although past efforts have not achieved the ultimate goal in spatial and height resolution, it is attainable in the next decade. On attaining this goal, satellite altimetry will enable improved knowledge of