I I
Low-Frequency
Variability of the
Sea

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11.1 Introduction

The purposeful study of the time-dependent motion of the sea having periods longer than about 1 day is comparatively recent. In the classic *Handbuch* of the early 1940s, Sverdrup, Johnson, and Fleming (1942), one searches in vain for more than the most peripheral reference to temporal changes on the large scale (one of the few examples is their figure 110 showing the California Current at two different times). Until very recently, the ocean was treated as though it had an unchanging climate with no large-scale temporal variability. The reason for this is compelling and plain: until the electronics revolution of the past 30 years, the major oceanographic observational tool was the Nansen bottle; using slow, uncomfortable ships, it took essentially 100 years to develop a picture of the gross characteristics of the mean ocean. The more recent period, 1947 (Sverdrup, 1947) through about 1970 (Stommel, 1965; Veronis, 1973b; and see chapter 5), was one of the intensive development of the theory of large-scale, steady models of the ocean circulation. The methods were initially analytic, later numerical. Most of these models were essentially low-Reynolds-number, steady, sluggish, sticky, climatic oceans. In them, the role (if any) of small-scale, time-dependent processes is simply parameterized by a positive eddy coefficient (*Austauch*) implying a down-the-mean-gradient flow of energy, momentum, heat, etc. The westward-intensification theories (Stommel, 1948; Munk, 1950) imply that any strong influence of such eddy coefficients would be confined to the western boundaries and could be ignored in the interior ocean, except possibly in the immediate vicinity of the eastward-moving free-jet "Gulf Stream" (see Morgan, 1956). The resulting models bear a remarkable resemblance to many of the gross features of the large-scale mean ocean circulation (see chapter 5).

The culmination of these analytic and numerical models of the large-scale circulation coincided with a number of developments that ultimately undermined the momentary confidence that the models represented the correct dynamics of the ocean circulation. These developments were of two kinds—instrumental and intellectual.

By 1970 instruments had been developed that made it possible to obtain *time-series* measurements in the open sea for periods far longer than a ship could possibly remain in one location. These instruments included moored current meters, drifting neutrally buoyant floats, pressure gauges, and many others (Gould, 1976; and see chapter 14). An additional "instrument" was the computer, which made it possible both to handle the large data sets generated by time-series in-

struments and to explore new ideas by nonanalytic means. This computer impact has been felt, of course, in most branches of science.

The intellectual developments that shifted the focus from the mean circulation to the time-dependent part were also of various kinds. The analytic models seemingly had reached a plateau at which their increasingly intricate features [e.g., essentially laminar boundary layers of higher and higher order as in Moore and Niiler (1975)] seemed untestable and intuitively implausible outside the laboratory. Physical oceanography is also to some extent a mirror of meteorology; by 1970 most oceanographers were at least vaguely familiar with the picture of the atmosphere that had emerged over the previous decades. In that fluid system, the view of the role of eddies had shifted from a passive means of dissipating the mean flows (through purely down-gradient fluxes of momentum, energy, etc.) to a much more interesting and subtle dynamic linkage in which the mean flows (the climate) were in at least some parts of the system *driven* by the eddy fluxes (Jeffreys, 1926; Starr, 1968; Lorenz, 1967). Because many of the meteorological results would apply to any turbulent fluid, there was reason to believe that the ocean could also exhibit such intimate dynamic linkages. But we should note that even now much work is still directed at studying the mean circulation by essentially classical (though improved) means, as if the variability were not dynamically important (e.g., Schott and Stommel, 1978; Wunsch, 1978a; Reid, 1978). The extent to which such pictures of the mean circulation of the large-scale tracers will survive complete understanding of variability dynamics is not now clear.

In this chapter we shall review what is known about the variability of the ocean. The expression "low-frequency variability," which is part of the title of this chapter, is a vague one used in a variety of ways by oceanographers, and encompassing a wide range of things. Here we mean by it anything with a time scale longer than a day out to the age of the earth, although we cannot really study by instrumental means phenomena with time scales longer than about 100 years. In spatial scale, it means phenomena ranging from some tens of kilometers to the largest possible global ocean oscillations. We shall, in common with recent practice, also refer to the "eddy" field in the ocean. This word is often prefixed by "mesoscale" and is used loosely to denote the subclass of variability encompassing motions occurring on scale of hundreds of kilometers with time scales of months and longer. It is a convenient shorthand and is meant to imply neither any particular dynamics nor only flows with closed streamlines. (The equivalent Soviet term is "synoptic scale").

There is little doubt that oceanographers were quite aware, from the very beginning, of time variability in the ocean. Maury (1855, p. 358) remarked that in drawing his charts he had disregarded "numerous eddies and local currents which are found at sea." He also notes in particular (p. 188) the highly variable equatorial currents of the Pacific Ocean. Even earlier, Rennel (1832) had quoted another observer (C. Blagden), as referring to North Atlantic currents as "casual" (Swallow, 1976).

Most of the astute observers who worked at sea since Maury were very conscious of the difficulties of drawing conclusions about the mean circulation in the presence of a highly time-dependent field. Figure 11.1, taken from Helland-Hansen and Nansen (1909), clearly depicts what one suspects to be a time-dependent eddy field. Sverdrup et al. (1942) make the statement that determining the mean is difficult in the presence of the time variations, and that the closer the station pairs are together, the greater is the requirement of simultaneity in hydrographic measurements. This is, of course, a statement about the frequency-wavenumber character of the baroclinic variability.

It is possible to give many instances of references to ocean variability and eddies throughout the history of observational oceanography. But it is also fair to say

Figure 11.1 Chart of Norwegian Sea surface currents as constructed by Helland-Hansen and Nansen (1909); reproduced by Sverdrup et al. (1942). One presumes the small-scale currents are in fact time-dependent features.

that little attention was paid to the phenomenon per se; it was a nuisance—a noise-contaminating determination of the time mean flow. There are some major exceptions, including Pillsbury's (1891) heroic efforts in the Florida Current, the 400-page work by Helland-Hansen and Nansen (1920), and somewhat later, Iselin's (1940a) attempts at monitoring the western North Atlantic. The question of the physical significance of a weak mean flow in the presence of strong variability has rarely been addressed even now.

11.1.1 Early Theory

The first theoretical attempts to study the purely time-dependent oceanic motions at low frequency seem to be outgrowths of the papers by Rossby and collaborators (1939) and by Haurwitz (1940a). These two studies, while directed primarily at the atmosphere, nonetheless addressed themselves to the large-scale time-dependent wave motions of a rotating hydrostatic fluid—a characterization applying equally well to the ocean. The Rossby paper in particular introduced the β-plane approximation. These early efforts, and the large number that followed, examined the wave motions known much earlier. Indeed Laplace (1775) had discussed motions that we now would call Rossby' or planetary waves (or in Hough's terminology, "tidal motions of the second class"). The study of these motions has a long and distinguished history (e.g., Darwin, 1886; Rayleigh, 1903; Poincaré, 1910), culminating in Hough's (1897, 1898) remarkable study of the solutions of the Laplace tidal equations on a sphere. [Lamb (1932), in his chapter on tides gives a good summary of this work. He also thoroughly discusses (§206 and §212) what we call "topographic Rossby waves" in which topographic gradients play a role analogous to the variation of the Coriolis parameter with latitude on a sphere.]

But it was Rossby's β-plane that demonstrated the physics in the simplest form and permitted an escape from the geometrical complexities of spherical coordinates. Arons and Stommel (1956), Veronis and Stommel (1956), Rattray (1964), Rattray and Charnell (1966), and others made explicit attempts to understand the possible role of Rossby waves in the ocean. Longuet-Higgins in a series of papers (1964, 1965) justified the β-plane approximation and carried out a modern exhaustive search of the solutions on a sphere for a complete range of parameters far beyond what Hough could do in his time (Longuet-Higgins, 1968a, Longuet-Higgins and Pond, 1970). Most of this work was done in the absence of any direct observational base in the ocean. (For further discussion of these waves, see chapters 10 and 18).

Observations, which will be discussed at length below, suggest that linear wave models are inadequate to describe much of the actual time-dependent motion in the ocean. Nonetheless, as in the atmosphere (Holton, 1975), many of the features of the observations are qualitatively similar to those deducible from the linear theories. That is, the physics is modified by the nonlinearity, but many of the linear features persist into the nonlinear range. The precise extent to which this is true is a function of the periods and spatial scales of the motions and is not really understood. As a generalization, it may be safe to assert that the largest oceanic scales of fluctuation are most likely to be dominated by linear dynamics. A linear description becomes increasingly doubtful for smaller scales, and barotropic motions ought to be more nearly linear than baroclinic ones (Rhines, 1979).

The postulate of a time-dependent field in the interior ocean immediately calls into question (Stommel, 1965, p. 221) one of the fundamental deductions of the steady-ocean models—that a Sverdrup balance applies in the interior ocean.

Consider, for example, Stommel's (1948) model of a homogeneous flat-bottom ocean on a β-plane. Let ψ_0 be the time-mean transport streamfunction and let ψ_1 be the time-dependent part. Then the time average vorticity balance may be written

$$J(\nabla^2\psi_0, \psi_0) + \langle J(\nabla^2\psi_1, \psi_1)\rangle$$

$$+ \beta\frac{\partial\psi_0}{\partial x} + R\,\nabla^2\psi_0 = -\hat{\mathbf{k}}\cdot\nabla\times\tau, \tag{11.1}$$

where the bracket denotes a temporal average, $\hat{\mathbf{k}}\cdot\nabla\times\tau$ is the vertical component of the mean wind-stress curl, J denotes the Jacobian operator, and R is the coefficient of bottom friction. Let us assume that the mean field varies over scales of 10^4 km, and that the time-dependent eddy field varies over 10^2 km. Let both the mean flows and the time-dependent part have magnitude $10\,\mathrm{cm\,s^{-1}}$. Scaling, we obtain roughly, in nondimensional form,

$$10^{-4}\,\nabla^2\psi_0 + \frac{\partial\psi_0}{\partial x} + 10^{-3}J(\nabla^2\psi_0, \psi_0)$$

$$+ 10^1\langle J(\nabla^2\psi_1, \psi_1)\rangle = -\hat{\mathbf{k}}\cdot\nabla\times\tau. \tag{11.2}$$

Away from the western wall, the first term is negligible; hence if we ignore both nonlinear terms, an interior balance is

$$\frac{\partial\psi_0}{\partial x} = -\hat{\mathbf{k}}\cdot\nabla\times\tau, \tag{11.3}$$

which is the conventional Sverdrup balance. But the nonlinear term

$$10^1\langle J(\nabla^2\psi_1, \psi_1)\rangle \tag{11.4}$$

will be of the same order as the Sverdrup terms if the

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correlation in the bracket is no greater than 0.1, and hence the Sverdrup balance would be upset. The ease with which one could destroy the simple relation (11.3) has motivated many of the recent studies of mesoscale eddies. It is fair to state, however, that we are still not in a position to compute terms like (11.4) (or comparable terms in more sophisticated models) with sufficient accuracy to assess the adequacy of the Sverdrup relation. There is some evidence (Leetmaa, Niiler and Stommel, 1977) that (11.3) is qualitatively correct away from the eddy-rich area near the Gulf Stream itself, but no quantitative assessment has yet been possible. One would anticipate, based upon the known wide variability in eddy energy levels (see discussion below) that there is a wide geographical variability in (11.4).

Constructs such as equation (11.2), which suggest that eddies may be important in the open sea, are, in effect, a reopening of the question that appeared to have been answered by Stommel (1948) and Munk (1950), where the first-order effects of eddies—in the guise of eddy viscosities—were confined to the western boundaries of the ocean. Webster (1961a, 1965), following Starr's lead, showed that at least in some regions the sign of the eddy flux of momentum might be opposite to that assumed in the viscous models, and now we are at the stage of being unsure even whether the regional confinement to the west, which had seemed so clear in 1950, is valid.

Equation (11.2) and more realistic formulations imply that an energetic eddy field could upset the lowest-order open-ocean mean-vorticity balance. But eddies also can carry mass, momentum, heat, and other variables. It is not difficult to show the potential importance and confusion that can arise from the presence of a strong time-dependent flow field possessing small scales. A simple example was presented by Longuet-Higgins (1969c), who considered a weakly nonlinear wave. In the weak-interaction limit, one can write the time-mean particle velocities as

$$\langle U_E \rangle = \langle U_L \rangle + \langle U_S \rangle,$$

where U_E is the Eulerian velocity, U_L the Lagrangian velocity, and U_S the Stokes velocity, which derives from the wave Reynolds stresses. The Eulerian velocity is the value that would be measured by a current meter at a fixed point; the Lagrangian velocity would be measured by tracking a dyed particle. In the absence of imposed exterior flows, Longuet-Higgins (1969c) demonstrated that the Eulerian and Lagrangian flows need not have the same magnitude or direction, and indeed can yield values differing by 180°. In the highly nonlinear limit it can be extremely difficult to find any simple relation between the Eulerian and Lagrangian flow fields. Such possibilities call into question the entire notion that there is some unique "general circulation"

of the ocean. Presumably one must carefully define the quantity whose overall circulation is desired, be it heat, mass, momentum, energy, passive tracer, etc., and seek the dynamic balance that will govern its flux. Doing this represents one of the most important problems facing oceanographers, who only recently have come to grips with the existence of oceanic time variability. One anticipates that over the next decade the problem will be solved, but it is impossible at the present time to perceive the details of the actual dynamic and kinematic balances in the ocean.

In the past, there have been some attempts to study special situations in which time variability was modeled in simple ways in order to seek an understanding of its potential role in the mean circulation. In particular, Pedlosky (1965c), Veronis (1966c), and Munk and Moore (1968) all examined the possible role of weakly nonlinear Rossby waves in generating large-scale mean flows. All of these models used an Eulerian frame; as noted above, obtaining an Eulerian means does not necessarily imply the presence of an actual net water movement. Often one can demonstrate the actual impossibility of a Lagrangian mean (e.g., Moore, 1970) and the question often hangs on subtle questions of dissipation (Eliassen and Palm, 1960; Charney and Drazin, 1961) and the existence of critical levels (Andrews and McIntyre, 1978a).

11.1.2 More Recent Theory

Many of the second-order effects of linear Rossby-wave motions have been studied. Longuet-Higgins and Gill (1967), Lorenz (1972), Kim (1978), and Jones (1979) have shown that the waves themselves are unstable. The scattering of the waves by random currents was examined by Keller and Veronis (1969), and scattering from topography was studied by Hall (1976) among others. McKee (1972) examined the diffraction limit of the waves. Interaction of Rossby waves with mean shears has been studied primarily in a meteorological context by Charney and Drazin (1961), Holton (1975, chapter 4), and finite-amplitude waves in steady flows were analyzed by Pedlosky (1970).

Finite-amplitude "soliton" solutions have been constructed by Flierl (1979b) and Redekopp (1977) as an effort to model the extreme form of ocean variability represented by Gulf Stream rings. These models have usually been based upon some form of Korteweg-de Vries equation and share many of the same properties as other known solitary waves (Whitham, 1974, chapters 16, 17; also see chapter 18, this volume).

Partly in response to the observational data base, which suggests that linear dynamics cannot be wholly adequate, there is a recent and growing literature of the opposite extreme, that is, based on the assumption of a completely turbulent motion. In the weak-interaction theories of Rossby waves (e.g., Longuet-Higgins

and Gill, 1967) the energy transformation of a wave occurs on a time scale long compared to a wave period. In the turbulence models, this is no longer true; the interactions are rapid and strong and one is forced into a statistical framework. An important step was taken by Charney (1971a), who showed that "geostrophic turbulence" in a stratified fluid would behave much like strictly two-dimensional turbulence known in other contexts (Fjørtoft, 1953; Kraichnan, 1967). This follows from the requirement that the fluid satisfy two independent quadratic constraints—conservation of energy and conservation of potential vorticity (enstrophy).

These ideas have been much elaborated since then (Rhines, 1979; Holloway and Hendershott, 1977; Salmon, Holloway, and Hendershott, 1976; Salmon, 1978). The problems are subtle and difficult and the dominant mechanisms not yet completely sorted out. Numerical experiments have been done to study several different physical processes that can govern the evolution of given initial conditions. The pure barotropic nonlinear cascade process moves energy toward larger spatial scales. Another nonlinear process transfers baroclinic energy toward the Rossby radius of deformation and thence to the barotropic mode. Topographic influences can scatter energy toward either larger or smaller scales, depending upon the wavenumber spectrum of the initial conditions and of the bottom topography. The relative importance of each of these processes depends on the energy level and spatial structure of the initial conditions. The balances are often subtle and no single process seems to dominate throughout the plausible range of initial conditions.

The advent of observations of mesoscale motions in the early 1970s also stimulated attempts to make numerical oceanic general-circulation models that could resolve an interior eddy field rather than parameterizing it simply as a constant, positive eddy viscosity. By restricting the calculations to limited ocean areas, it was possible to have many forms of time variability appear explicitly in the models (Holland and Lin, 1975a,b; Robinson, Harrison, Mintz, and Semtner, 1977). As computers have become larger and faster, the models have become ever more complex and realistic, including all of the physical mechanisms of the simpler-process models. However, it is fair to say that even the most sophisticated models now extant cannot fully model all of the observed physical complexity of the ocean (Schmitz annd Owens, 1979). But this is not to denigrate the models. They are beginning to show qualitatively many of the observed features of the oceans and doubtless rapidly will become much more realistic. Nonetheless, the ocean is very complex; numerical models that are intended to be realistic need also to be complex. As more and more physics is added to the computer codes the results become increasingly diffi-cult to understand. Indeed it may be that understanding a fully realistic numerical model of the ocean requires nearly as much time, effort, and ingenuity as understanding the real ocean.

We shall not dwell further here on the modern theories because there have been a number of recent reviews of the state of the art. Rhines (1977, 1979) has discussed many of the basic ideas; the numerical models have been examined and compared by Robinson, Harrison, and Haidvogel (1979) and Harrison (1979b). As we proceed to discuss the observations, we shall introduce additional theoretical hypotheses as needed.

11.2 The Field of Variability of the Ocean

Perhaps the most sweeping generalization that can be made about the known time variability in the ocean is this: the ocean is filled with time-varying features, with all space and time scales, whose energy levels vary by orders of magnitude over the ocean basins. To state it slightly differently, the field of variability is locally representable by a continuous frequency-wave-number spectrum; but the underlying process is not spatially stationary in the statistical sense and this vitiates much of the utility of the spectral description. The use of the word "continuous" for the spectrum is deliberate. Through much of the history of oceanography, as in many fields, there has been a search for simple line processes, "cycles," which simply do not exist.

We know from the past decade of observation that simple universal parameterizations of the variability are not valid. The upper ocean is different (at least superficially) from the lower ocean and the gyre centers are different from the gyre boundaries. Eastern walls differ in their variability from western walls. It is probably also true that the dynamics, as well as the kinematics, of these regions differ. At this time, it is difficult to give more than a fragmentary picture of open-ocean low-frequency variability because our observational tools are still not adequate for the job of measuring a global ocean.

11.2.1 The Meteorological Forcing Function
The ocean is driven by the heating of the atmosphere and by direct momentum transfer from the winds. The details of these processes are not completely clear (e.g., Phillips, 1977a; Kraus, 1977), involving as they do small-scale turbulent transfer processes within the marine atmospheric and oceanic boundary layers. Nonetheless it is probable that the larger scales of these forcing functions are able to communicate themselves from the atmosphere to the ocean. That is, the details of the transfer of momentum to the ocean from the
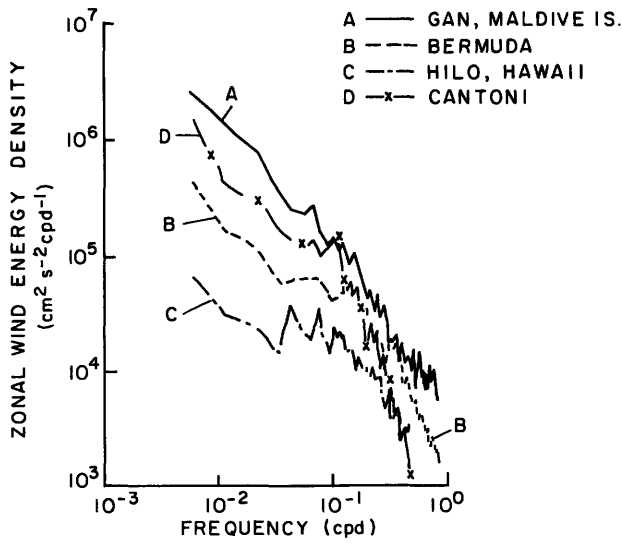
Figure 11.2A Zonal wind spectra from a few representative open-ocean islands.
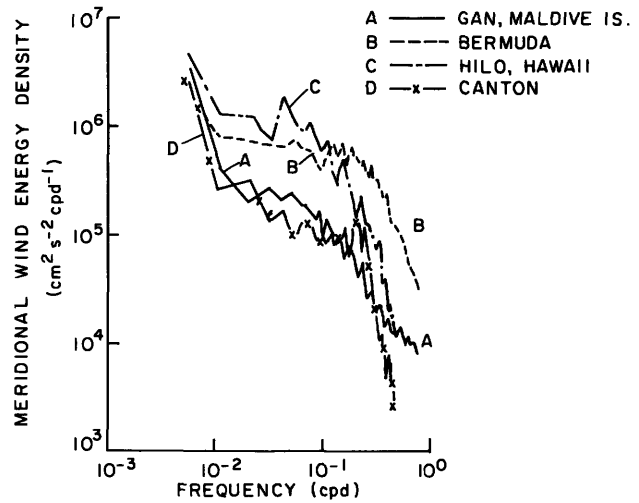

Figure 11.2B Meridional wind spectra from open-ocean islands.

winds involves such small-scale processes as ordinary ripples, but we anticipate that if the wind varies over a 1000-km scale, then it is this variability scale that is relevant to understanding the oceanic response to winds, and it is thus meaningful to seek a description of the forcing function in frequency-wavenumber space.

It is comparatively easy (Wunsch and Gill, 1976; Philander, 1978) to show that direct atmospheric-pressure forcing is a very inefficient process compared to wind-stress forcing. Direct thermal forcing is likely also (Frankignoul and Müller, 1979) to be comparatively weak except on the very largest time scales that determine the mean thermohaline general circulation.

The description of how the atmosphere forces the ocean is of considerable interest in studying the field of variability, but it may not be a decisive factor. The reason is that theoretically one can drive oceanic variability indirectly through instabilities of the strong "mean" current systems (Lipps, 1963), and of the interior ocean (Gill, Green, and Simmons, 1974; Robinson and McWilliams, 1974). Nonetheless in some regions at least—the continental shelves (see chapters 7 and 10), in the open sea far from intense currents (Brown et al., 1975) and in regions like the Florida Current (Düing, Mooers, and Lee, 1977; Wunsch and Wimbush, 1977)—there does seem to be some direct response to atmospheric forcing, although it seems to be at the short-period end of the spectrum, namely, periods shorter than about 10 days.

Some representative wind spectra are displayed in figure 11.2. It should be noted that stress is usually
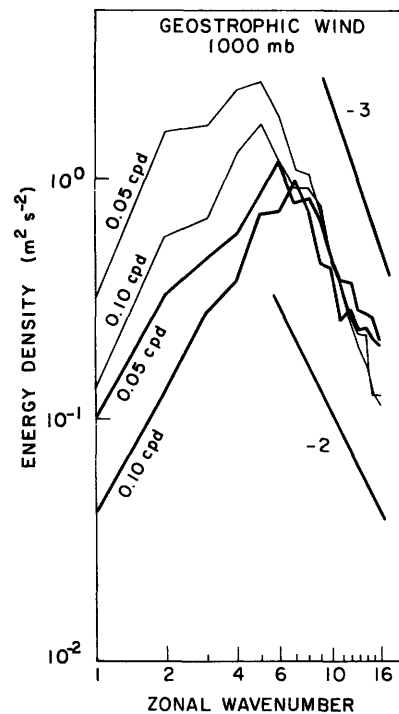

Figure 11.2C Estimated wavenumber spectrum of meridional geostrophic wind at 30°N (heavy line) and 50°N (light line) for two frequencies at 1000 mb. Notice steep slope at high wavenumbers. (Frankignoul and Müller, 1979.)

computed from the two winds components, (w_x, w_y) by the formula

$$(\tau_x, \tau_y) = C(w_x^2 + w_y^2)^{1/2}(w_x, w_y), \qquad (11.5)$$

where C is a parameter depending upon the drag coefficient and which in general (Bunker, 1976) depends upon the air–sea temperature difference and possibly, upon the wind speed itself. But the spectrum of stress will strongly resemble that of the wind because the expression (11.5) preserves the zero crossings of the wind components. The spectrum is distorted relative to that of the wind by the amplitude modulation factor $(w_x^2 + w_y^2)^{1/2}$.

The spectra tend to become white (or less red) at periods longer than a few days, reflecting the unpredictability of weather (by linear methods at least), and then redden again at the longer (and here unresolved) periods (but see figure 11.9A). The spectra show a great variety of geographical effects, viz., latitude changes, proximity to continental influences, topography (the Hilo spectrum in particular seems greatly affected by the presence of high mountains, being highly anisotropic at low frequency), and sea–land contrasts. In the mid-latitude spectra, most of the energy is found in the 4–10-day band characteristic of the weather systems. At periods longer than those displayed, the wind spectra tend to become white (see Willebrand, 1978). Frankignoul and Müller (1978) have computed estimates of the 1000-mb zonal wavenumber spectra for a variety of latitudes, displayed here in figure 11.2C. One sees a distinct concentration in the low wavenumber bands. Extrapolation to the sea surface is not straightforward, however.

The response of the ocean to forcing by fluctuating wind fields has been considered by Phillips (1966), Frankignoul and Müller (1979), Leetmaa (1978), and Harrison (1979a) in the period range of days to months. Although the final word has not been spoken, it appears that over most of the ocean direct wind forcing is unlikely to compete with internal instability processes. The weak seafloor pressure fluctuations measured by Brown et al. (1975) are spatially coherent on the large scale only at periods of 10 days and shorter, where current meter records show very little energy. These may in fact be wind-forced barotropic modes but they are energetically unimportant. In what follows the reader may want to compare the shape of the wind spectra displayed in figure 11.2 with those of the other variables discussed later.

Frequency spectra of atmospheric-pressure fluctuations (not shown) also tend to show a whitening at low frequency although Madden and Julian (1972) and Luther (1980) have found some large-scale organized motions at long periods, circa 50 days.

Sea-level measurements of atmospheric variability are sparse and inadequate for making definitive statements. Frankignoul and Müller (1979) attempted to construct a model spectrum by synthesizing the available data. They assumed spatial homogeneity and isotropy in the wind field. The two assumptions are ultimately untenable, but their model is probably the best that can be constructed at the present time.

11.2.2 Interannual Fluctuations in the Ocean

These are changes with periods longer than 1 year. Our major emphasis will be on those motions deduced in the modern era of instrumentation, excluding periods accessible only through essentially geological methods. Thus with one exception (see below) we will not treat what is best called paleo-oceanography, which deserves a full treatment by itself.

Before attempting to describe what is known about the very long-period changes in the ocean, there are two points to be made. Consider first figure 11.3. Figure 11.3A is a section made by the vessel *Challenger* from New York to Bermuda to St. Thomas, Virgin Islands, in 1873. The figure at bottom is a section from the Grand Banks to Bermuda to the Mona Passage obtained in 1954 and 1958 (Fuglister, 1960). The qualitative resemblance is very close; clearly the fundamental assumption of large-scale physical oceanography—that at least some aspects of the overall circulation, in particular the large-scale baroclinic structure, have remained stable for 100 years—is correct. That is consistent with the statement of Sverdrup et al. (1942) noted above about combining nonsimultaneous stations if they are sufficiently widely spaced, and is a (usually) unstated assumption in discussions of the mean fields to this day. It is well justified, for example, by comparing the high-quality *Meteor* sections in the South Atlantic made in 1925–1927 and reported by Wüst and Defant (1936) with those made in 1954–1958 and reported by Fuglister (1960). One is hard pressed to detect any significant differences on the large scale. (Of course, from these data one is able to say nothing regarding barotropic changes.)

One result of the paleo-oceanographic studies is useful here too. Figure 11.4 displays the reconstruction from paleontological data of sea-surface temperature 18,000 years ago, and the annual mean today (from Gates, 1976). Eighteen thousand years BP was the height of last major glaciation (CLIMAP Project Members, 1976). What is so striking is how *little* the ocean surface changed away from the immediate proximity of the edge of the ice sheet. Indeed the change appears to be less than the present seasonal range (Fuglister, 1947). That changes in the ocean under the impact of such a large disturbance as a glaciation are so minor suggests that seeking changes in the ocean owing to present
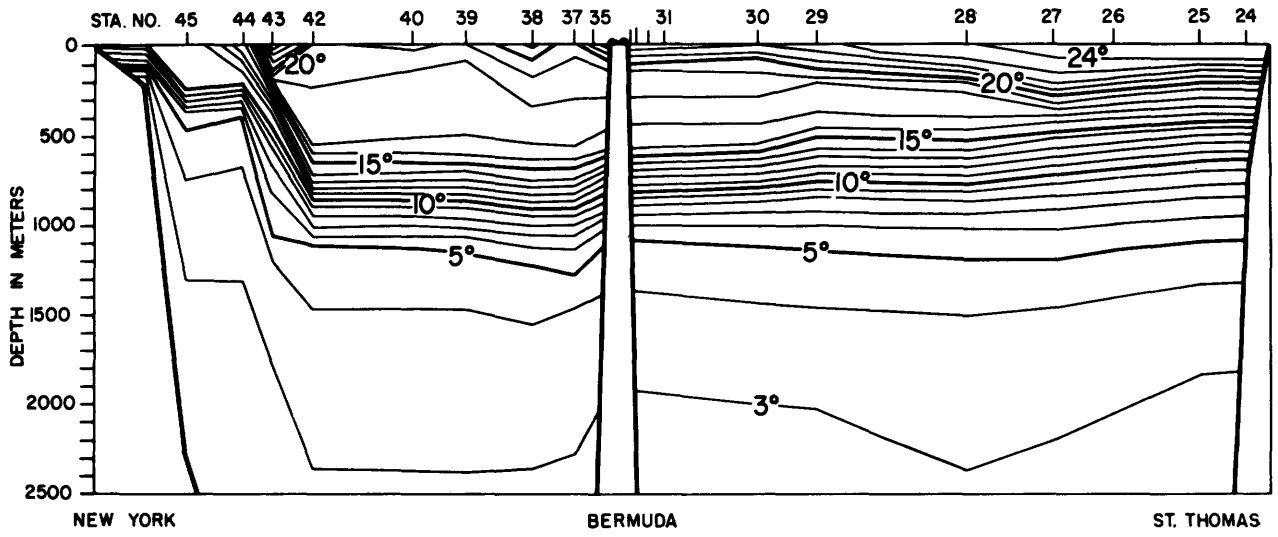
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Figure 11.3A *Challenger* section (1873) from New York to Bermuda to Virgin Islands. (Tizard et al., 1885b, p. 135.)
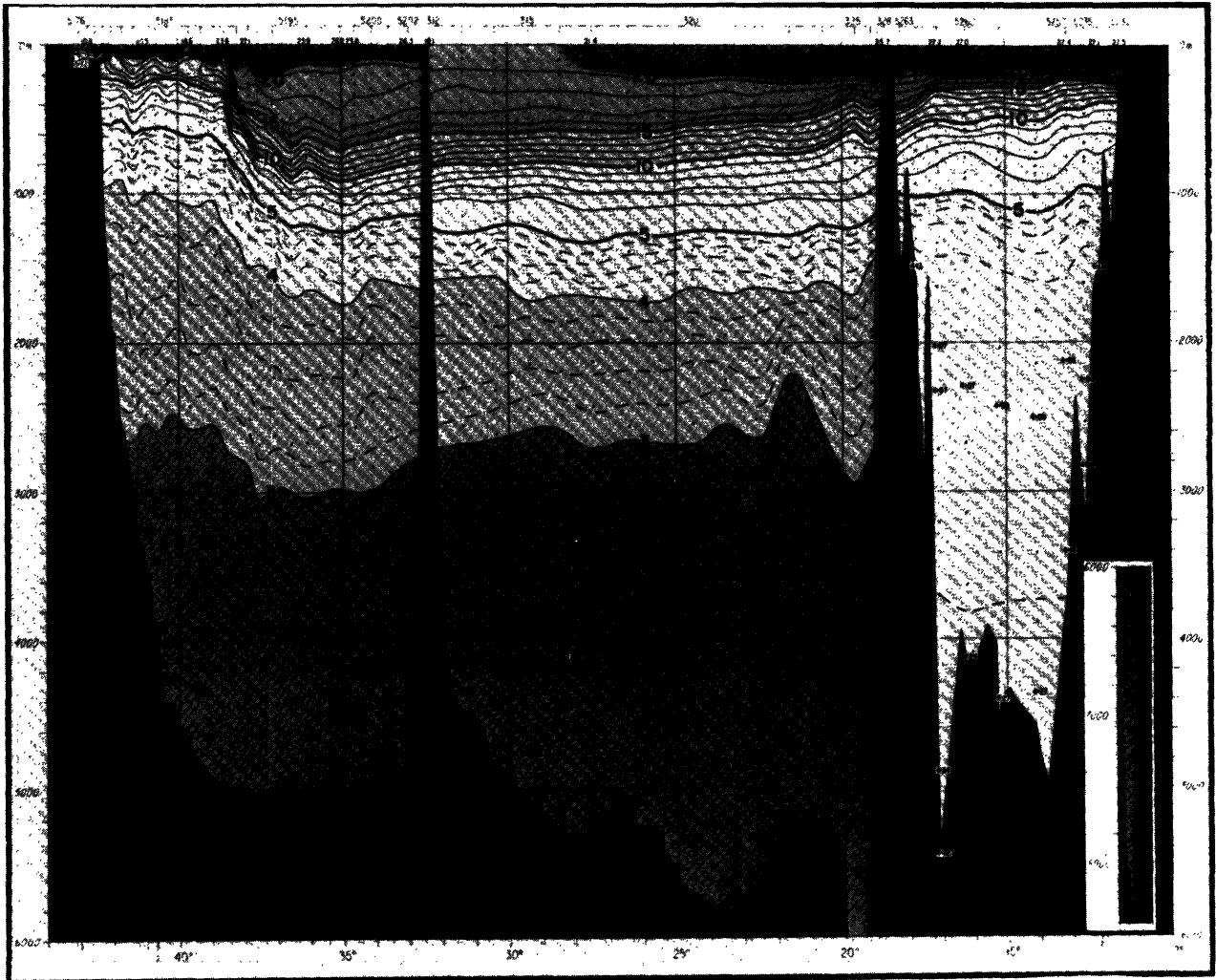

Figure 11.3B Section made in the fall, 1954 (from Nova Scotia to Bermuda and Mona Passage to South America), and in winter, 1958 (from Bermuda to Mona Passage). (Fuglister, 1960.)
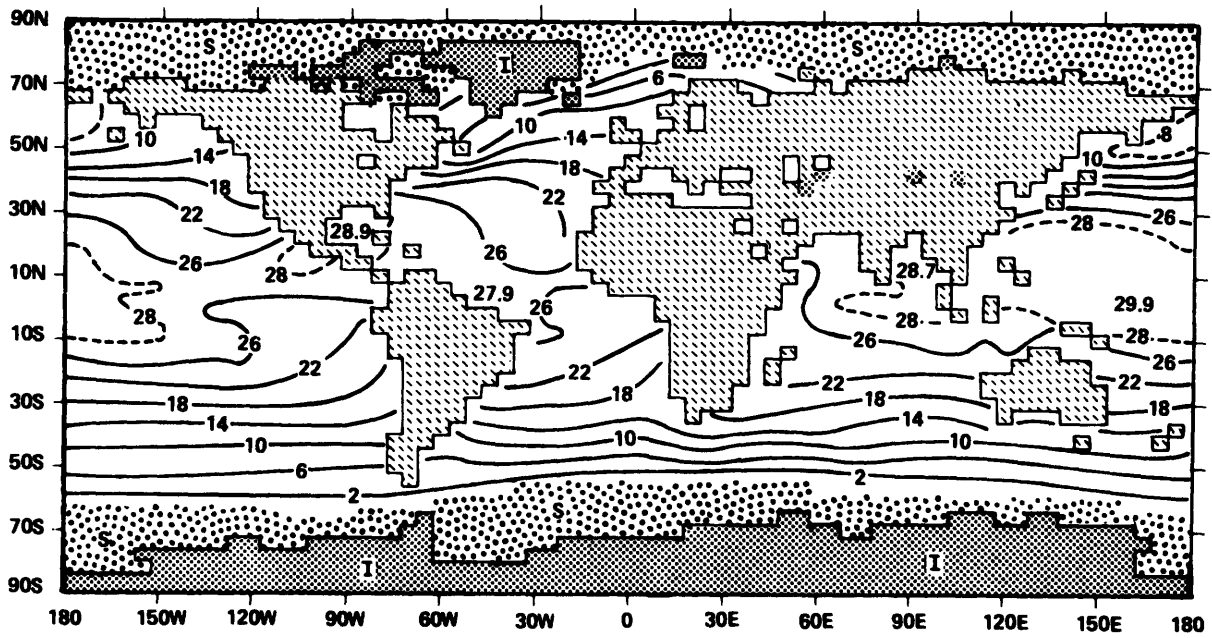
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Figure 11.4A Present annual mean sea-surface temperature in North Atlantic. (Gates, 1976.)
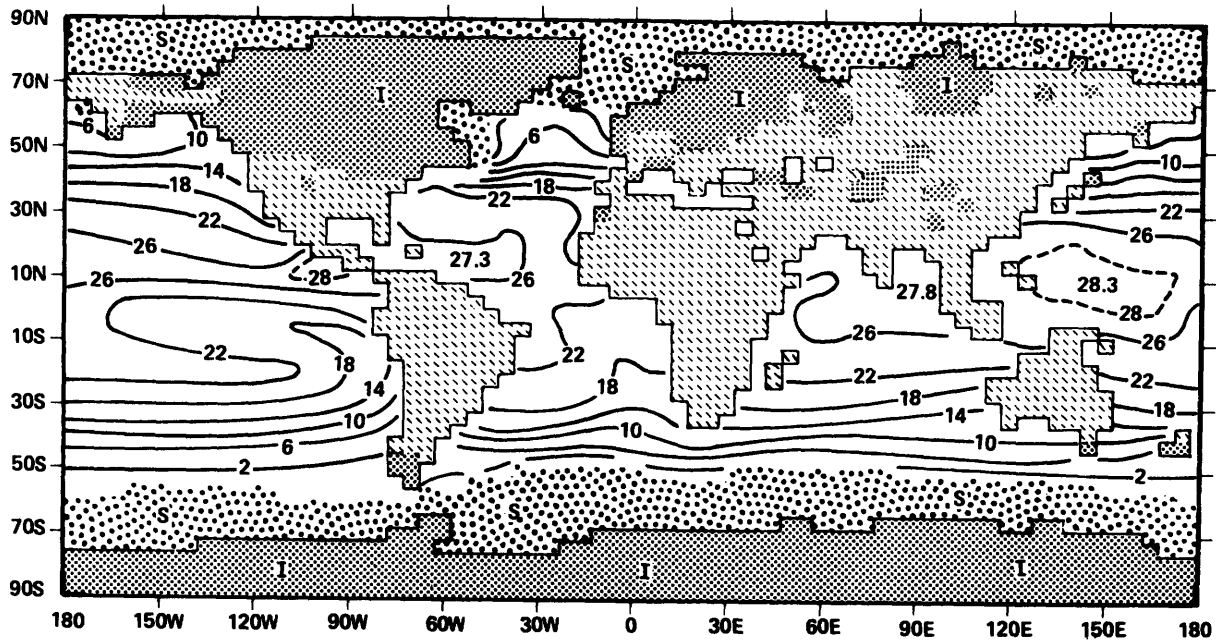
Figure 11.4B Estimated sea-surface temperature 18,000 years ago at height of glaciation as determined by CLIMAP study of fossil assemblages in deep-sea cores.

minor climatic fluctuations may be fruitless. The changes 18,000 years ago could have been much greater at depth than at the surface; whether this is indeed true is now the subject of study.

The success of oceanographers in obtaining at least qualitative descriptions of the large-scale circulation by combining observations made many years, or even decades, apart in time is, as noted, a reflection of some underlying truth about the frequency–wavenumber content of the baroclinic variability in the ocean. The most intense fluctuations seem to occur on a spatial scale small compared to the large-scale mean gyres. On the other hand, if one is to seek the role of the ocean in climatic changes, it is likely to be reflected in very large-scale low-frequency oceanic fluctuations, on the intuitive assumption that perturbations to the atmosphere owing to small-scale fluctuations of the ocean will tend to be "integrated out" by the atmosphere. In a nonlinear system like the atmosphere, the validity of this assumption is by no means obvious, but it is a reasonable initial hypothesis. One can then attempt to search for very large-scale changes in ocean circulation on time scales of years and space scales of thousands of kilometers.

For a number of reasons, most of the work on this hypothesis has been conducted in the Pacific Ocean. First, it has long been hypothesized (Bjerknes, 1969, Namias, 1972) that the U.S. continental weather may be sensibly modified by large-scale thermal anomalies over the Pacific Ocean. Bjerknes provided some specific hypotheses now generally called "teleconnections." Beginning in the early 1960s investigators at the Scripps Institution of Oceanography began a series of investigations to attempt to define, and ultimately to understand, both the apparent anomalies themselves and the role they might play in U.S. climate and weather. It seems clear that the anomalies are real; but it also seems fair to state at many of the links between the anomalies and weather are the result of wishful thinking rather than evidence (e.g., Davis, 1976, 1978a).

Sea Level Few extended time series for studying very long-period motions are available; the only real data consists of sea-level measurements. The idea of using tide-gauge records for studying fluctuations of geostrophically balanced currents evidently dates back to Sandström (1903), although Montgomery (1938b) seems to have been the first to actually attempt it.

The spectra presented here in figure 11.5 (see also figure 11.9), in Wyrtki (1979), in Munk and Cartwright (1966) and in other places, of the longest available records [circa 100 years—the longest record may be the one from Brest analyzed by Cartwright (1972), which runs from 1856 to date] are red, although decreasingly so beyond periods of 1 year (see Figure 11.9).

The limits of this increasing power with decreasing frequency are unknown. In some regions, one may be seeing slight fluctuations in the geodetic levels of the tidal gauges; in other regions this seems implausible. Taken at face value, the red spectra suggest that there indeed may be barotropic, large-scale fluctuations of the oceanic gyres. One infers that they are barotropic because of the decrease in temperature variance with lengthening period appearing in the temperature spectra (see section 11.2.4), and large scale because of the decreasing variance in the moored-current spectra at long periods. But the evidence is ambiguous.

A number of attempts to understand the physics of long-period fluctuations in sea-level records have been made (Wunsch, 1972c, Groves and Hannan, 1968; Groves and Zetler, 1964; Shaw and Donn, 1964; Schroeder and Stommel, 1969). The major difficulty is that not only are long records few, but the number of long simultaneous records, which are vital for understanding spatial correlations and possible propagation, are even rarer.

Sea-level fluctuations at a point are a complex summation of many different physical phenomena. Much of the work cited above was dedicated to attempting to unravel the role of local weather variables in sea-level fluctuations. In the open sea, using island data, Groves and Hannan (1968), Wunsch (1972c), and others found an inverted barometer response at periods longer than about a day. At periods of months and longer, the effects of wind tend to dominate. The procedure for deducing the relative role of the two wind components and pressure is not straightforward because the weather variables are themselves coherent; a multiple regression procedure described in the cited papers is required. With *localized* weather effects removed one can then ask whether fluctuations from location to location are coherent and can be ascribable to any particular known physics. A major problem has been (e.g., Groves and Hannan, 1968) the absence of measurable coherences between the few simultaneous island records available. The entire procedure is made very difficult because the prior removal of the fluctuations coherent with local weather may remove global phenomena that are forced by the meteorology.

The clearest picture of large-scale fluctuations of sea level stems from the work of Wyrtki (1974, 1975a) in the Pacific. Previous work in the Atlantic by Schroeder and Stommel (1969) and Wunsch (1972c) had suggested that on the time scale of months that sea level was to a large extent responding to fluctuations in dynamic height relative to reference levels at about 1500 db (decibars). Wyrtki (1979) has shown in the tropical and equatorial Pacific Ocean that he could find large-scale coherent patterns of dynamic height variation that also
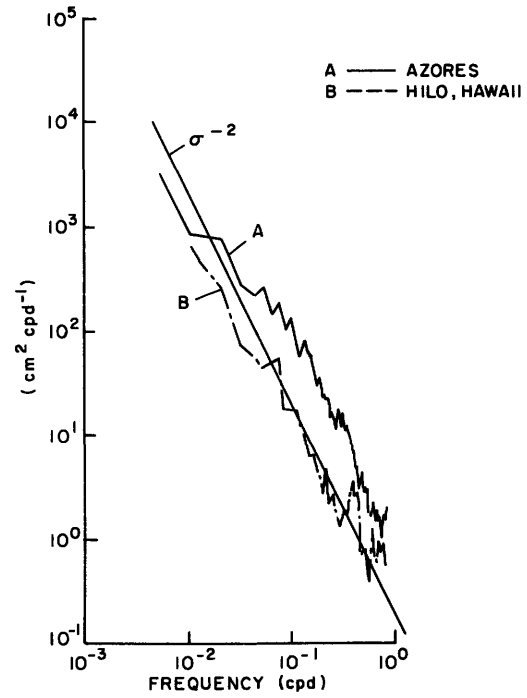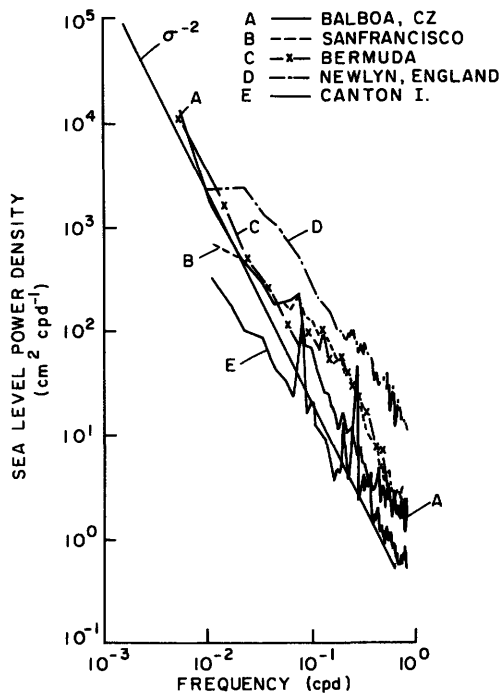
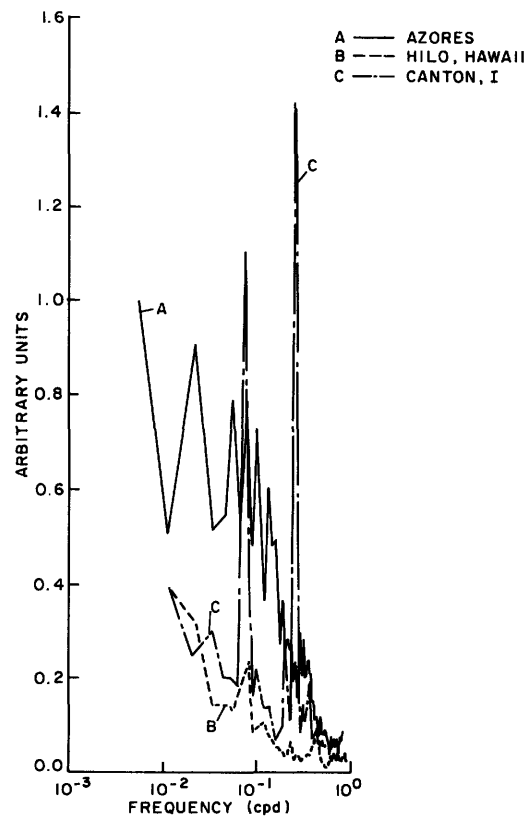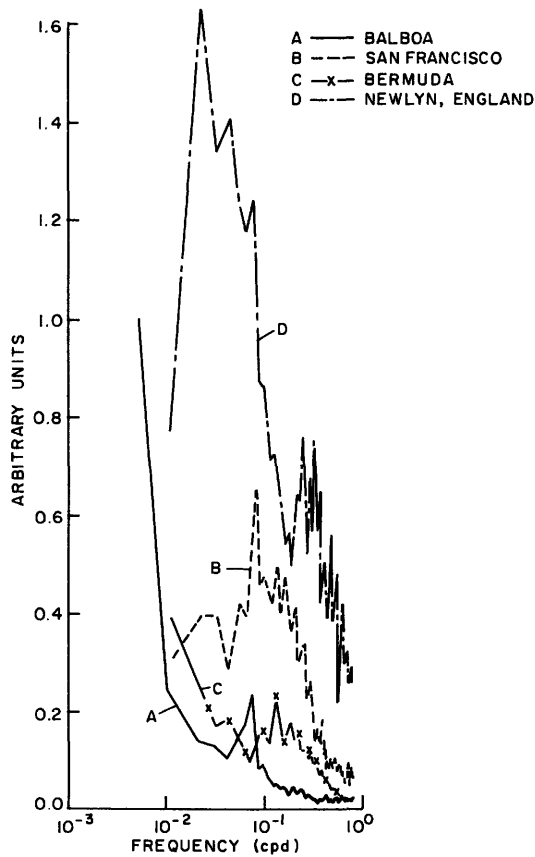Figure 11.5A Spectra from a few representative sea-level records (log-log form).

Figure 11.5B Spectra from sea-level records in an energy- (variance-) conserving form; notice that units are arbitrary—only relative spectral shapes are comparable. Canton I spectrum contains sharp peaks at 4 and 5 days, described by Wunsch and Gill (1976). The fortnightly tide is also apparent in the spectrum.

shows up convincingly in the Pacific tide gauges. He has been able to relate his observations to fluctuations of the large-scale tropical current systems of the Pacific. Much of this work has been directed toward understanding of the relationship between the ocean fluctuations and the catastrophic economic and climatic effects of the El Niño phenomenon (see chapter 6).

There does seem to be a link (or at least a correlation) between trans-Pacific sea-level fluctuations, the occurrence of warm water on the coast of Peru, and gross changes in the wind field over the Pacific. These latter changes are supposed to be part of the so-called southern oscillation and the Walker cell circulation (Bjerknes, 1969) in the atmosphere. Thus there is some indication of an actual coupling of large-scale oceanic and atmospheric fluctuations, but the global extent of the phenomenon and the causal links are very obscure. The data base is inadequate to be truly definitive, but the apparent patterns are plausible and are a highly promising line for future research into gyre-wide fluctuations, at least in the near-equatorial oceans.

For the purely oceanic phenomenon one needs ultimately to understand the extent to which one is seeing dynamic topography variations relative to a fixed reference level (a difficult idea to rationalize) or fluctuations possibly representable as vertical normal modes. In this latter case, the changes in sea-surface topography imply equivalent deep-water fluctuations, which are, however, at this time totally unknown.

Presumably similar fluctuations occur in the Atlantic and Indian Oceans. But the absence of many islands in these oceans has largely precluded the determination

of sea-level fluctuations there. The Azores and Bermuda records were examined by Wunsch (1972c) and the Iceland record by Donn, Patullo, and Shaw (1964). Most of the low-frequency variability of the Indian Ocean, especially in the western portions, is obscured by the very large monsoonal signals.

Thermal Record The most conspicuous low-frequency phenomenon in the ocean is the sea-surface temperature anomalies of the Pacific that have been studied intensively the past two decades. Barnett (1978) has reviewed this work. A primary motive for the interest was the Bjerknes (1969) teleconnection hypothesis. There is little doubt that extensive changes in sea-surface temperature do exist and can persist for months and years. Figure 11.6, taken from Barnett (1978), is a multiple-year record of temperature (fluctuations about the long-term mean) at Talara, Peru, and Christmas Island. These "anomalous" temperatures occupy major areas of the Pacific Ocean. Figure 11.7 displays the first three empirical normal modes of Pacific sea-surface temperature (from Barnett and Davis, 1975). These modes describe slightly under half the total variance of the Pacific sea-surface temperature fluctuations; their immense scale is apparent.

Much of the interest in the anomalies has been in the possibility that in changing the lower boundary condition of the atmosphere, fluctuations in the atmosphere might be induced on the long oceanic time scale rather than on the intrinsically short atmospheric time scale. Studies of the question have been hampered
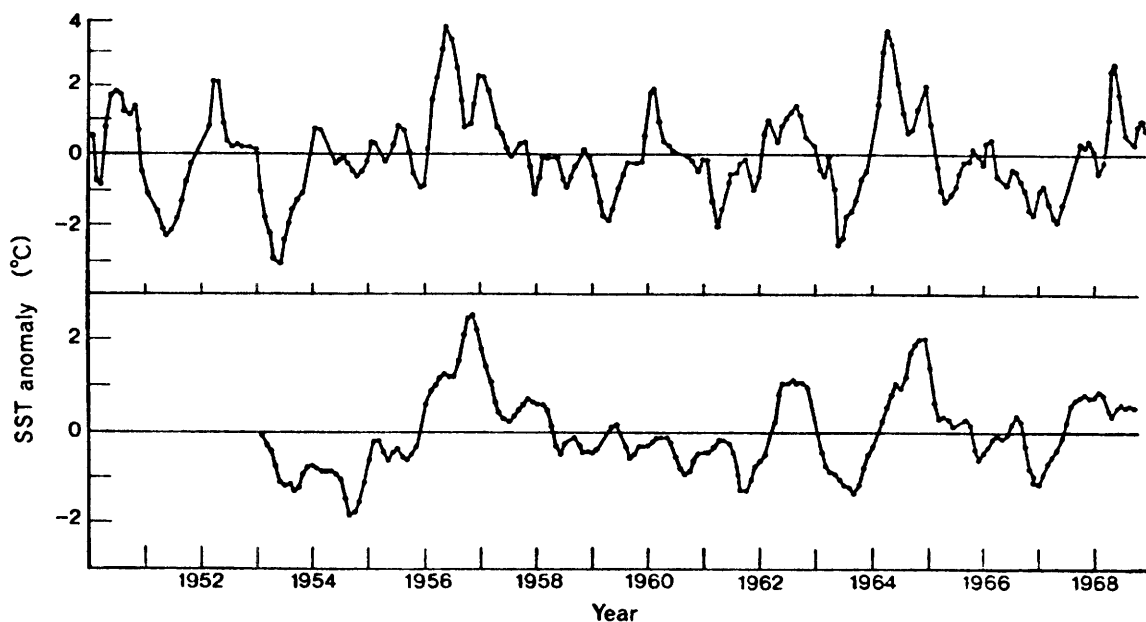

Figure 11.6 Sea-surface temperature anomalies at Talara, Peru (upper) and at Christmas Island (lower). (Barnett, 1978.)
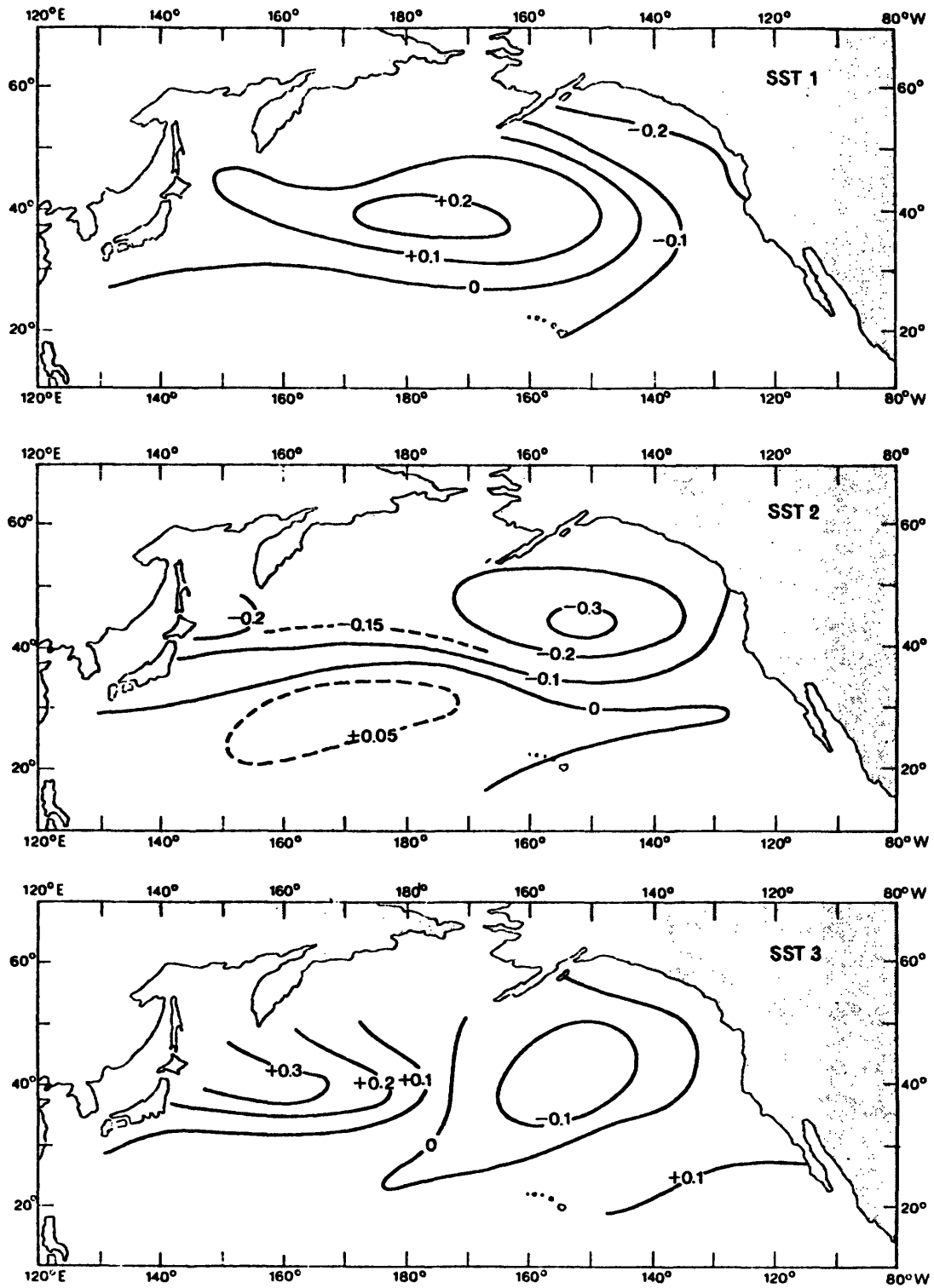
Figure 11.7 Three lowest empirical normal modes of sea-surface temperature anomaly field in North Pacific Ocean. Notice very large scales involved. (Barnett and Davis, 1975.)

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by the inherent noisiness of the atmosphere. But Davis (1976) showed that changes in the ocean tended to lag those in the atmosphere, thus implying that the anomalies were being driven by the atmosphere rather than the reverse. In a later paper, Davis (1978a) obtained some atmospheric predictability from sea-surface temperature anomalies stratified by season. But the same predictability was found using atmospheric sea-surface pressure anomalies. The cause-and-effect relationships thus remain unknown. Rowntree (1972) and Kutzbach, Chervin, and Houghton (1977) have studied the reaction of the atmosphere to imposed sea-surface temperature anomalies. With unrealistically large anomalous values, an atmospheric reaction can be detected, but its significance is still not understood.

Frankignoul and Hasselmann (1977) have shown that purely random forcing of the ocean by the atmosphere can plausibly generate anomalies. The degree to which the surface features represent anomalies of heat content, that is, represent subsurface features as well, also is not clear. White and Walker (1974) have displayed time–depth diagrams for temperature anomaly at three positions in the Pacific Ocean (figure 11.8). Gill (1975b) shows similar data and perhaps the simplest conclusion to be drawn is that the physics governing sea-surface temperature anomalies is a combination of interaction with the atmosphere and with deeper ocean dynamics in a form that varies in space and time. It is difficult to relate the gyre scale fluctuations described by Wyrtki, to the anomalies, except in the case of El Niño.

Observations in the Atlantic similar to those made in the Pacific have been described by Rodewald (1972). On a much longer time scale, there are variations in the extent of the pack ice in the vicinity of Iceland that seem relatable to atmospheric changes. Oceanic changes that may accompany the atmospheric fluctuations beyond those locally involved with the sea ice are unknown.

The significance and meaning of apparent large-scale gyre-wide fluctuations is obscure. Attempts at demonstrating massive fluctuations in western boundary current flows that might be of real climatic significance have generally not been convincing. The wild variability in estimates of the volume transport of the Gulf Stream north of Cape Hatteras (see Worthington, 1976) has generally been due to varying methods of estimating reference levels, and there are no convincing demonstrations of nonseasonal volume-transport fluctuations. Similar remarks apply to the Antarctic Circumpolar Current, where 100% changes in estimates of the volume transport essentially disappear when appropriate reference levels are used (Nowlin, Whitworth, and Pillsbury, 1977).

Much of the problem of documenting real low-frequency large-scale fluctuations in the major ocean gyres may be looked upon as a classical problem of aliasing. Momentary fluctuations in the mass field, which owe their existence to short-period time variability, may inappropriately be interpreted as representing much lower-frequency phenomena (Worthington, 1977b). The large-scale baroclinic structure of the ocean is remarkably stable and one cannot store large amounts of water anywhere in the system without raising havoc with sea level. Real low-frequency oceanic changes on large scales appear to be subtle and difficult to detect in the presence of the much stronger short-period fluctuations discussed below in section 11.2.4.

11.2.3 Annual Variability
The entire atmosphere-ocean system is subject to a large thermal forcing with an annual period as the sun migrates meridionially throughout the year. In the atmosphere the seasonal change in character is of course enormous at mid- and high-latitudes. The ocean would respond both to the direct thermal forcing of the sun and to the indirect thermal forcing due to the change in air–sea temperature difference. One anticipates, from calculations like those of Gill and Niiler (1973) and Frankignoul and Müller (1979), that except in the mixed layer and on the equator the response to both would be quite small. Even in the region of intense wintertime air–sea heat exchange that occurs in the northwest regions of the northern hemispheric subtropical gyres and that leads to the formation of 18°C water in the Atlantic and 16°C water in the Pacific, the oceanic response seems localized and nearly passive (Warren, 1972; Worthington, 1959). More generally Gill and Niiler (1973) show that the heat budget is essentially locally controlled with lateral advection comparatively unimportant. Only the upper 100–200 m of ocean is involved.

It is the oceanic response to the annual cycle in the wind field that is likely to dominate the annual period. The annual line in the wind field at the ocean surface is superimposed upon a background continuum, and is surprisingly weak (figure 11.9) even in the monsoon regions of the Indian Ocean. The annual line at mid-latitudes, e.g., at the location of Bermuda, seems to carry no more than about 1% of the total wind-field variance in periods between 10 years and 1 day.

Patullo, Munk, Revelle, and Strong (1955) analyzed the available tide-gauge records for the annual variability in sea level and provide an interesting global map. Most of the inverted barometer-corrected sea-level signal between 40°N and 40°S is in "steric" height, i.e., the essentially passive change resulting from the warming or cooling of the upper water col-
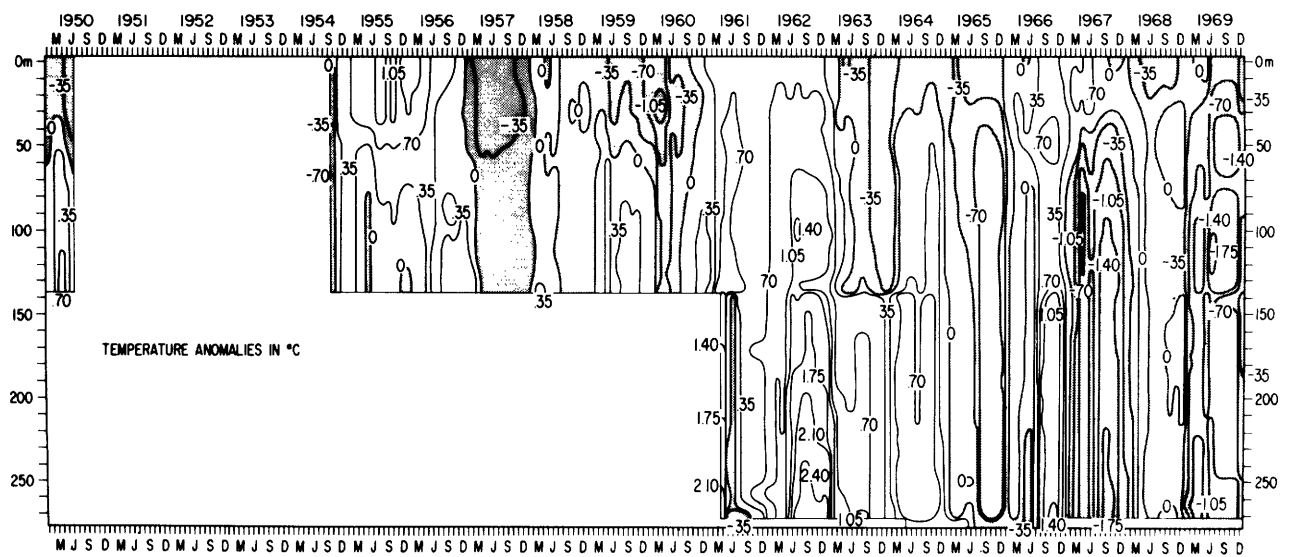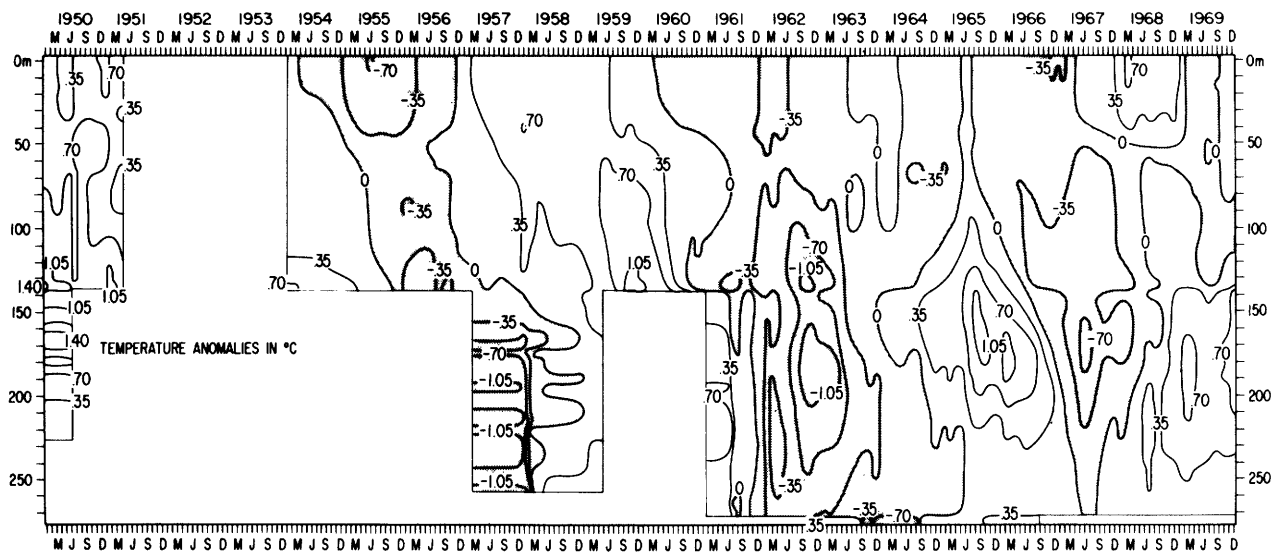
Figure 11.8 Time-depth diagram of subsurface anomalies of temperature at ocean station N (upper at 30°N, 140°W) and at station V (lower at 34°N, 164°E). (White and Walker, 1974.)

umn. The International Geophysical Year (IGY) data were subsequently examined by Donn, Patullo, and Shaw (1964), resulting in much the same conclusions. The effects of the direct steric changes do not appear to be visible below about 300 m in subtropical latitudes (Wunsch, 1972c).

The change in elevation of the sea surface owing to these direct heating effects should give rise to a weak annual surface circulation; this circulation would be weak because of the very large scale over which the steric effects are induced.

In their analysis of the annual variability, Gill and Niiler (1973) concluded that over most of the ocean the major baroclinic effects of the wind would be to induce an Ekman suction at the annual period with a consequent response of the main thermocline. This signal seems to be lost in the background movement of the thermocline in response to other forces. By and large, deep current meter or temperature records of a few years duration do not display any recognizable annual variability above the continuum [although White (1977) claims he sees annual-period baroclinic Rossby waves generated at the eastern boundary of the Pacific].

The western boundary currents do show some form of annual cycle; in the Florida Current the cycle manifests itself over the entire water column from top to bottom and has been convincingly documented by Niiler and Richardson (1973) as an essentially barotropic change. Farther north, Fuglister (1947, 1951) documented a cycle in the surface temperatures and flow by using ship-drift reports. The extent to which this annual cycle penetrates into the deep water north of the Florida Straits is unknown. Much earlier, Iselin (1940a) had attempted to study the annual cycle of the Gulf Stream between Woods Hole and Bermuda. Over a 3-year period, he found transport fluctuations between 76 and 93 $\times$ 10^6 m^{3-1}, with the system strongest in summer. In view of the problems of aliasing referred to above, the necessity for a fixed reference level, and the short record duration, it is difficult to assess the significance of this result. Wyrtki (1974) has shown an annual cycle in the surface circulation of the equatorial Pacific in response to the annual trade-wind changes. Elsewhere (Wyrtki, 1975a) he finds little annual signal except in the immediate vicinity of the Kuroshio.

There is a fundamental problem in using the annual cycle to understand the dynamics of the ocean. Because virtually all physical variables (wind, temperature, sea level, etc.) show an annual line, it is difficult to sort out cause and effect. Narrow-band processes tend to contain much less useful information than random broad-band ones.

11.2.4 The Mesoscale

The ocean appears to be in a turbulent state with a very high Reynolds number, and as that fact has come
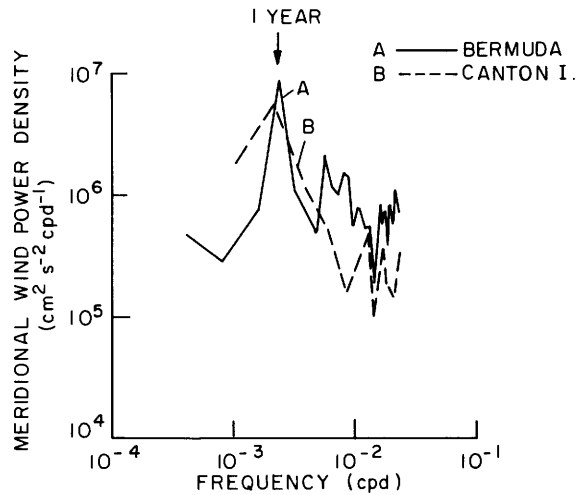

Figure 11.9A Spectra of wind records (displayed in figure 11.2 but at much higher resolution), showing the annual peak.
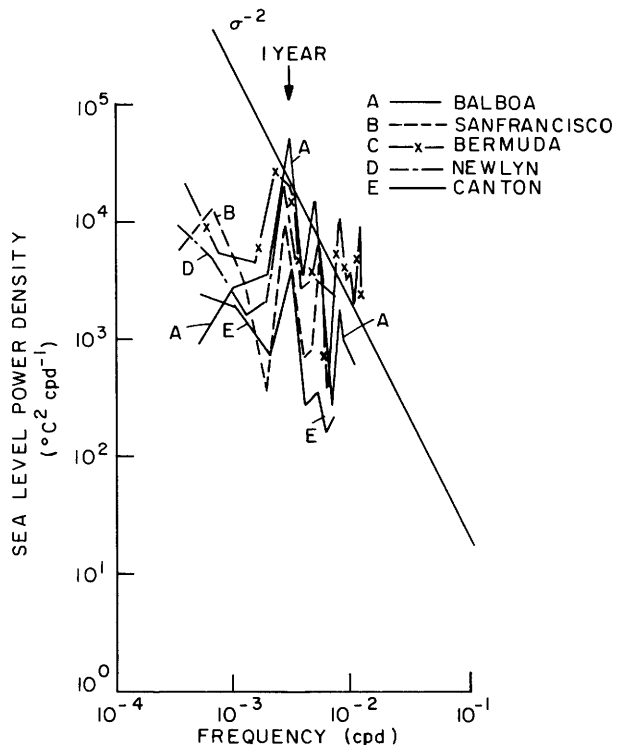

Figure 11.9B Spectra of long sea-level records (displayed in figure 11.5 but at much higher resolution), showing resolved annual peak.

to be more widely appreciated attention has shifted from the very large-scale mean flows and large-scale variability discussed above, to the nature of the turbulence itself. Much of this shift in focus has come about because of technical innovations. These made possible multimonth *in situ* time series independent of ship endurance limitations. The presence of disturbances of spatial scales of O(100 km) is obvious in conventional hydrographic sections (figure 3b from Fuglister, 1960) but studies of these undulations in the isopycnals were impossible before about 1970. In discussing the mesoscale, we include all fluctuations with the appropriate space and time scales, including continental shelf waves even though these latter motions are essentially nonturbulent in character.

Shelf Waves These waves were discovered by Hamon (1962), who found that sea-level fluctuations on the east coast of Australia were not in isostatic balance with atmospheric pressure (i.e., the inverted barometer effect failed to apply). Robinson (1964) pointed out that Hamon's results could be rationalized in terms of topographic Rossby waves excited by atmospheric pressure forces [the basic physics of the phenomenon is discussed by Lamb (1932, §212)]. Subsequently Adams and Buchwald (1969) demonstrated that wind stress was a much more efficient driving mechanism and obtained reasonable agreement between theory and observation.

Since the original observations of Hamon and the later theoretical work much attention has been paid to these waves for a number of reasons. Unlike most open-ocean variability, linear theory seems to apply reasonably accurately, the shallow water of the continental shelves is comparatively easily accessible, conventional harbor tide gauges can often be used, and the waves presumably play a major role in the time-dependent response of the low-frequency shelf circulation to external forces.

We shall not dwell on the subject here because a number of comprehensive reviews have appeared recently (e.g., LeBlond and Mysak, 1977; Munk, Snodgrass, and Wimbush, 1970; Mysak, 1980; and see chapter 10). The detailed connection between open-ocean variability and shelf waves is unknown. It is possible, however, that, in some situations, the waves generate open-ocean effects rather than merely being the response to local winds or forcing from the open sea. An example of this may be in the Gulf Stream System (Düing, Mooers, and Lee, 1977; Wunsch and Wimbush, 1977), where waves of several days period are observed in the Florida Current region. A number of authors have suggested that these waves may trigger meanders and other motions of the Gulf Stream System in the region north and east of Cape Hatteras. No actual cause and effect has ever been demonstrated; the topic is discussed in chapter 4.

Gill and Clarke (1974) made plausible the hypothesis that shelf waves (they specifically emphasized the baroclinic Kelvin wave) play a fundamental role in the special subclass of mesoscale variability called coastal upwelling. If this is correct (and it permits remote forcing to give rise to strong localized effects) there can be a dramatic influence of shelf waves on local climatology. In some cases (e.g., the El Niño phenomenon), coastal upwelling may be part of an ocean-wide, and perhaps even global, system involving both the ocean and atmosphere. The extent to which such events linking shelf waves (which may be thought of as a local focusing and amplification of comparatively weak forcing) are part of a global system is the subject of much work at the present time. That such links exist seems clear; their extent is obscure. Upwelling deserves a special discussion of its own; the reader is referred to O'Brien et al. (1977).

In our more general context, shelf waves provide a simple (i.e., nearly linear) example of mesoscale variability that in some cases exhibits the combined effect of rotation, topography, and baroclinicity. The literature contains a large number of special cases of topography for which analytical solutions are available. The basic physical phenomena are Kelvin waves trapped against the coast, double Kelvin waves at the shelf edge (Longuet-Higgins, 1968c), barotropic topographic Rossby waves (Rhines, 1969a), and baroclinic topographic Rossby waves (Rhines, 1970). In the presence of stratification finite bottom slopes couple the different wave types together and one is generally driven to numerical solutions (Wang and Mooers, 1976), but the underlying physics may be understood from the simpler physical situations. Ou (1979) has significantly advanced the analytic treatment of the problem. Hendershott in chapter 10 discusses these topics further.

Very similar effects occur at mid-ocean islands (Longuet-Higgins, 1969b; Wunsch, 1972b; Hogg, 1980); both barotropic and baroclinic trapped waves can occur and particular bottom-slope configurations can modify qualitatively the dispersion relations. The barotropic case with varying topography had been treated by Lamb (1932, §212), for the *inside* of a cylinder, rather than for the outside, as is appropriate to the oceanic case (the mathematical differences are slight).

Mesoscale Eddies Much of what we now call the *mesoscale variability*, or the *eddy field*, is plainly what early investigators like Maury had in mind when they described their difficulties with time-dependent motion. In a modern context, these motions were first quantitatively measured in 1959 (Crease, 1962; Swallow, 1971) in a noteworthy experiment. Using the

newly developed neutrally buoyant floats (Swallow, 1955) they detected motions west of Bermuda with velocities of up to 40 cm s^{-1} and apparent time scales of roughly 40 days. Their measurements were analyzed in an important paper by N. A. Phillips (1966b), who suggested that they could be treated as a sum of linear Rossby waves driven by the wind. Phillips's model could not reproduce the high velocities (but see Harrison, 1979a). Later, Rhines (1971b) suggested that they were intensified by bottom trapping on the Bermuda Rise, but this too seems inadequate.

By 1970, the combination of the knowledge of the existence (at least in the area studied by Swallow and Crease) of very intense time-dependent motions, of the theoretical work on turbulent interaction, and of the meteorological picture alluded to, led to a series of experiments aimed at elucidating the nature of the mesoscale in the ocean [Koshlyakov and Monin (1978) discuss the history of parallel Soviet efforts, which seemingly had a very different intellectual content]. It seems fair to call the 1970s the "decade of the mesoscale" in physical oceanography both for the intensive efforts that went into it and for the striking advances in knowledge that occurred. The initial tentative observational programs (e.g., Gould, Schmitz, and Wunsch, 1974) had first to convince the investigators of the reality of the phenomenon; it was only then that serious efforts could be mounted to understand mesoscale physics. Although 100-day time-scale fluctuations are much more humanly accessible than the interannual frequencies described above, it is still important to note that to obtain 20 degrees of freedom at a resolved 100-day period for a spectrum of velocity (or anything else) from a point measurement requires 1000 days of data. Thus studies of the mesoscale are necessarily in their infancy.

It is probably a mistake to overemphasize the mesoscale as an isolated phenomenon. There is reason to believe that the very large-scale phenomena described previously and the mesoscale are to a great extent simply manifestations of the extremes of a continuum. Ocean variability occupies a broad space-time spectrum; for purely experimental reasons it has been convenient to study the 100-km scale separately from the 5000-km scale. But in the long term, it will be necessary to understand their linkages.

On the other hand, there does appear to be a "bulge" or plateauing of low-frequency spectral variability in the very roughly defined period range of 50–150 days. It is apparent even in the sea-level spectra, and as nearly as one can tell this plateau is a global phenomenon. We shall follow Richman, Wunsch, and Hogg (1977) in calling this the *eddy-containing* band. At the low-frequency end this band fuzzily merges into the annual and interannual variability. We shall refer to

frequencies above the eddy-containing band as the *isotropic band* because of the energy isotropy there. For the moment this distinction into separate bands is kinematically descriptive; there may also be a dynamic distinction.

Much of the work on mesoscale variability was focussed in the Mid-Ocean Dynamics Experiment (MODE-1) which ran from about 1971 to 1973. Many of the results have been summarized elsewhere (MODE Group, 1978; Mode-1 Atlas Group, 1977; Richman, Wunsch, and Hogg, 1977). The effort was continued in POLYMODE—as an amalgam of the early Soviet efforts Polygons (Brekhovskikh, et al., 1971) and the MODE experiment. By the time of POLYMODE, however, the mesoscale field was widely recognized as a universal, important phenomenon and was being studied as part of many aspects of oceanography, including the biological ones.

The detailed results of these experiments are too extensive to be described here, and, indeed, their full implications are not understood. But it appears that an eddylike field of variability is nearly universal in the ocean (MODE Group, 1978; Schmitz, 1976, 1978; Bryden, 1979; Bernstein and White, 1974, 1977; Hunkins, 1974) and is normally much more energetic than the local mean flow (see figure 11.10).

The time variability of the North Atlantic ocean on the large scale is represented by figure 11.11, taken from Dantzler (1977) [see Bernstein and White (1977) for a similar discussion of the North Pacific and Lutjeharms and Baker (1979) for the Southern Ocean]. It is a compilation of the temperature variance in the North Atlantic from expendable bathythermograph (XBT) measurements. The variance of temperature, which may be thought of as a crude measure of potential energy of the fluctuation field, shows a large-scale midocean trough, with steep gradients toward the boundaries, and especially toward the Gulf Stream System. This figure, while extremely suggestive, also illustrates several of the caveats that must be applied to discussions of the field of variability.

The variability field is commonly referred to as the *eddy problem*; the implication is that eddies have an identifiable scale, nearly unit aspect ratios and possibly even closed streamlines. But figure 11.11 is simply a representation of the time variability of the North Atlantic irrespective of its source. For example, it is well known that the Gulf Stream moves and meanders (Fuglister, 1963; and see chapter 4). Such variability will appear in figure 11.11 and presumably represents much of the energy appearing in the northwest region. The large-scale, low-frequency variability described in the previous sections would also be included but probably is only a small fraction of the variance. It is more than a semantic distinction to wish to distinguish the variability of migrating jets from that of features with
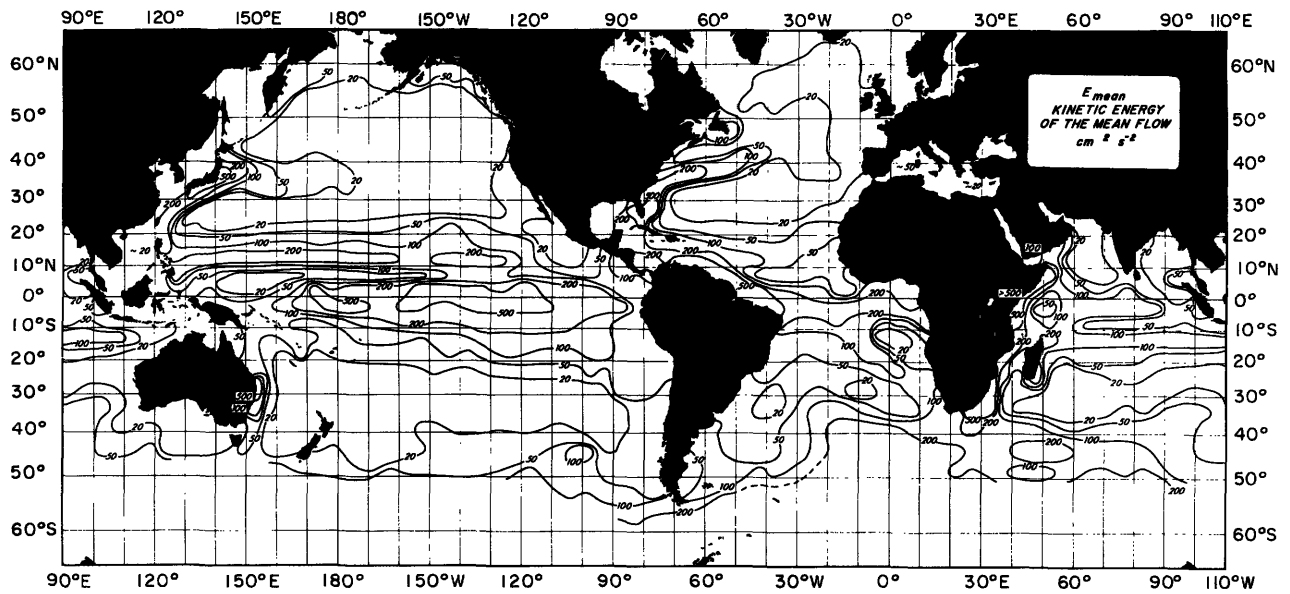
Figure 11.10A Five-degree-average chart, based upon ship-drift reports of mean surface kinetic energy. (Wyrtki, Magaard, and Huyer, 1976.)
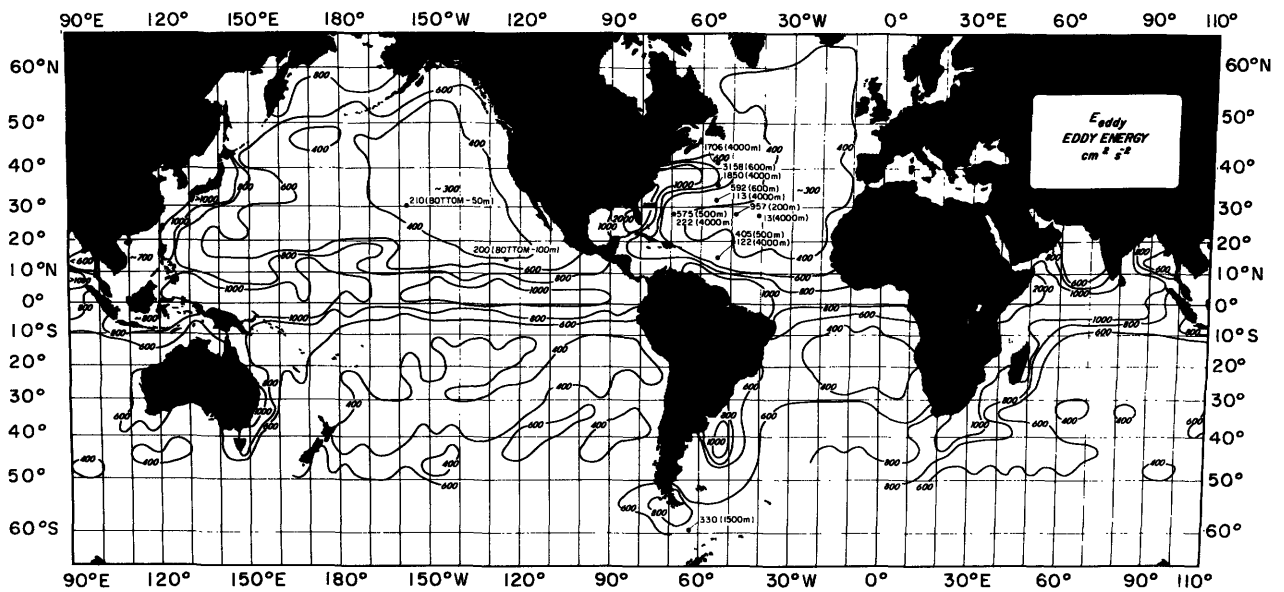

Figure 11.10B Five-degree-average chart, based upon ship-drift reports of total fluctuation kinetic energy (Wyrtki, Magaard, and Hager, 1976). Superimposed upon the chart are kinetic energy densities (cm^2 s^{-2} cpd^{-1}) from moored current meters, at a 50-day period at nominal depths of 500 m and 400 m. Actual depths are shown.
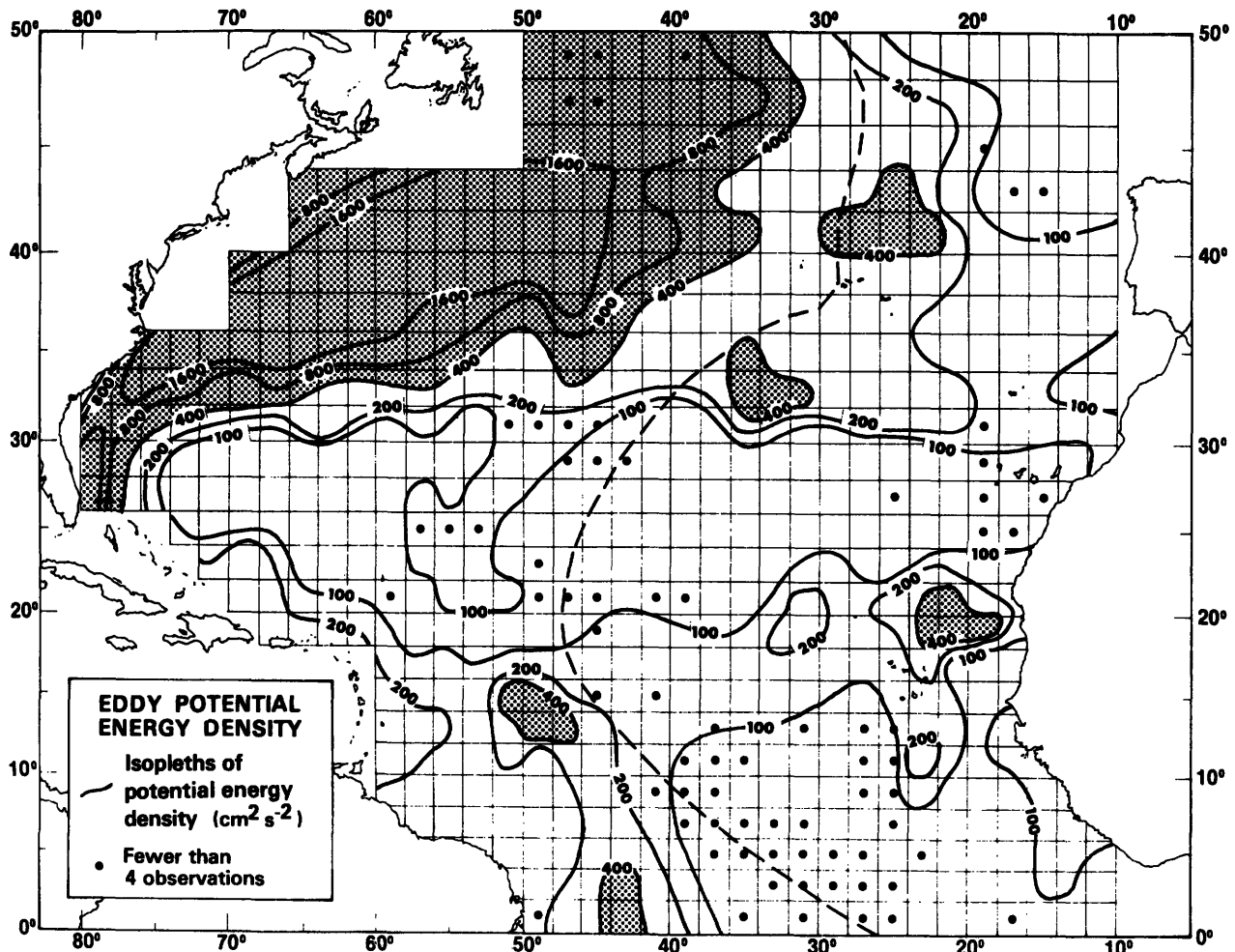
Figure 11.11 North Atlantic eddy potential energy as compiled from data by Dantzler (1976). Notice strong intensification in Gulf Stream region and general increase toward all boundaries.

relative isotropy of scale. Another difficulty is that XBTs measure only the baroclinic contribution to the eddy field. If there is a significant velocity field that might be described as barotropic, it will be missed by this form of data. Finally, XBTs reach only to some hundreds of meters depth. There is every reason to believe that the variability of the layers below the thermocline can be very different from that above. Unfortunately, we do not have sufficient data to construct for the deep ocean a figure analogous to figure 11.11. In the absence of such data there is an understandable tendency to interpret the upper layers as representative of the entire water column; there is little evidence to support that hypothesis, although it seems plausible that there should be a significant correlation between upper and lower oceans.

Another, global but still crude, depiction of the time variability of the surface ocean is figure 11.10 (from Wyrtki, Magaard, and Hager, 1976). It displays the mean kinetic energy and the variability about the mean based upon ship-drift observations averaged over 5°-squares.

The pitfalls of using such a data base are many and obvious; nonetheless, on a global basis, figure 11.10B is probably the best that can be constructed at the present time. Again, the motions represented there will be a sum over all time and space scales and are not confined to the mesoscale. In the North Atlantic there is a considrable qualitative resemblance to Dantzler's chart. One is probably safe in concluding that in a gross sense the largest total variability tends to be found on the western sides of the oceans, with secondary maxima along the other boundaries (including the equator). These regions are, of course, the ones associated with the apparently strongest mean currents (notice the resemblance between figures 11.10A and 11.10B), and one might surmise that there is some difficulty in distinguishing the variability as a boundary-intensified phenomenon from one associated with strong mean flows (an attempt at a distinction may be specious in any case).

Some feeling for the underlying nature of the fluctuations that contribute to (and perhaps dominate) figures 11.10 and 11.11 is displayed in figure 11.12. Here
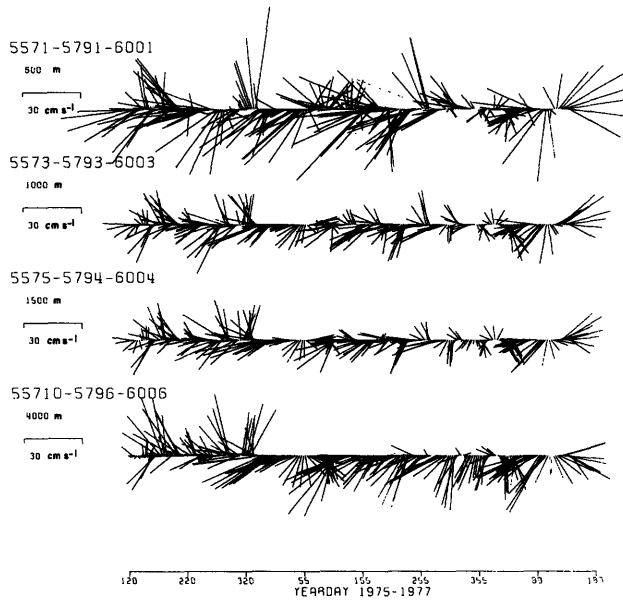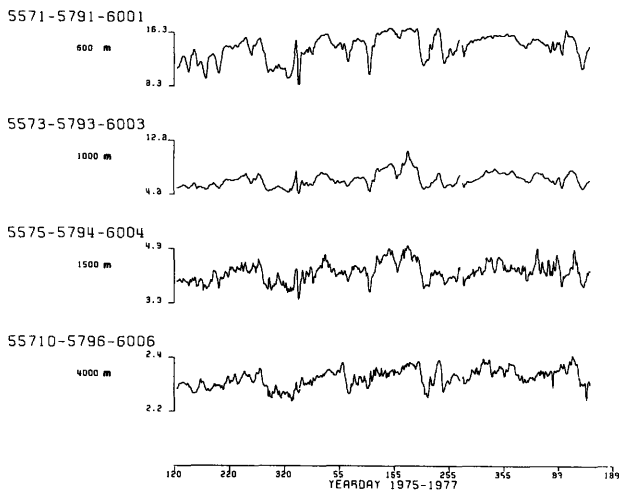
(11.12A)

(11.12B)

Figure 11.12A,B Three-day average-velocity vectors from two locations (35°56'N, 55°06'W and 31°36'N, 55°05'W), respectively, in the North Atlantic (see figure 11.11B). Note large
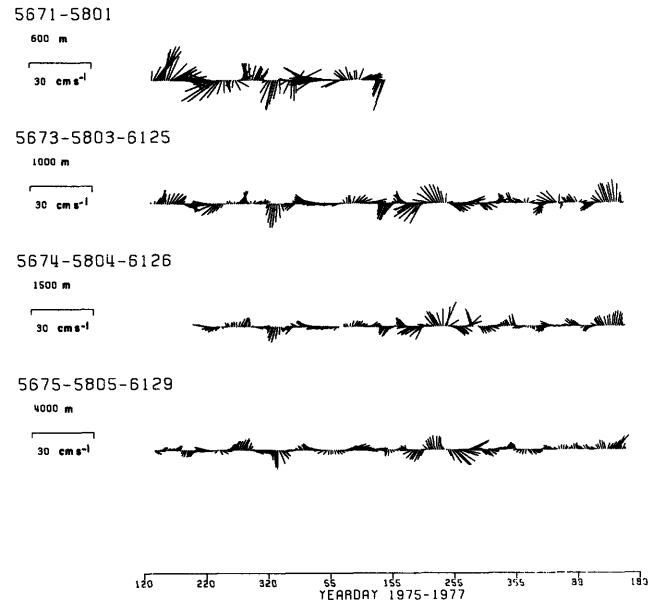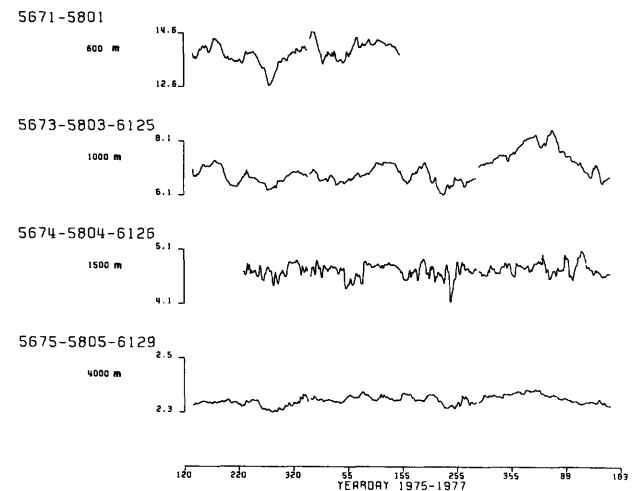
change in energy level over comparatively small separation in position between the two moorings. The difficulty of defining a mean flow from such measurements should be obvious.

(11.12C)

(11.12D)

Figure 11.12C,D Temperature records from same moorings as in figures 11.12A and 11.12B. Note change in character and amplitudes of fluctuations.

are the velocity vectors and temperature measurements (averaged over 24 hours) from two moorings in the north Atlantic at positions 35°56′N, 55°06′W and 31°36′N, 55°05′W. One sees a very complex and energetic time variability over the record lengths (nearly 27 months in the longest records), a considerable resemblance between different levels in the vertical, and a very marked decline in energy levels over the 400-km separation between the moorings. This remarkable change in energies is roughly consistent with that implied by figure 11.11, although the change is much steeper because of the 5° averages in figure 11.10.

Another view of the variability is shown in figure 11.13 depicting the spectrum of temperature fluctuations near Bermuda in the main thermocline (from Wunsch, 1972b). One sees a process with a broad spectral peak in the vicinity of 100 days with a relative decline at longer periods. As noted by Wunsch (1972b), the decline might have been anticipated because there is a stable large-scale mean-temperature field, at least in the 100-year historical record.

In an effort to display some of the characteristics of the low-frequency variability, I have collected here (figures 11.14 and 11.15) a number of spectra of long time series of different variables. Insofar as possible, these are displayed on common scales with common resolution to simplify the direct comparison of one region with another. In all cases, I display the logarithm of the spectral density as a function of the logarithm of the frequency. A number of the spectra are also plotted as frequency times power density against the log of the frequency (the so-called *variance-preserving form*). A word is in order about these different displays. The log–log forms tend to emphasize the very lowest frequencies in any spectrum that follows roughly a σ^{-p} form, where σ is the frequency and $p > 0$ simply because the value of the energy density–unit frequency band is increasing. With increasing frequency, the logarithmic transformation compresses an increasing number of equally spaced frequencies into a fixed distance along the abscissa.

It is often the low frequencies we are most interested in (and long records are the most difficult to obtain), and dynamic models often produce power-law spectra that show up most clearly (as straight lines) on log–log graphs.

But a spectral representation actually gives equal weight to evenly spaced frequencies, and the Parseval relation between a function and its Fourier transform treats all frequencies alike. Thus, a log–log plot of a record with a frequency spectrum σ^{-p} will give the impression that most of the energy is in the very lowest frequencies. But because the number of equally spaced frequencies in a uniform logarithmic-frequency interval increases uniformly with frequency, a σ^{-1} spectrum

actually has as much energy in any fixed logarithmic-frequency distance, centered at an arbitrary frequency, as at any other frequency. This fact has led many investigators to plot frequency times energy density against log frequency, thus compensating for the increasing number of points–unit log frequency. One then interprets the energy in any given log frequency interval as the *area* under the curve. Thus a σ^{-1} spectrum would plot as a horizontal line. This, however, is *not* a white spectrum because it is still true that in any fixed *linear* frequency interval the total energy decreases with frequency as σ^{-1}. This form of the display can also give rise to very sharp spectral peaks (some may be seen in the figures); in many cases, however, the area under these sharp peaks is rather small and hence they may in fact represent little excess energy overall. Reader beware.

Because of the long time scales of the mesoscale phenomena, there are few quantitative measurements of energy levels or time and space scales. Most of the existing measurements are in the western North Atlantic as the result of MODE/POLYMODE but there are a few notable exceptions. Superimposed upon figure 11.10B are the locations of those records for which spectra are displayed here. As a crude measure of energy level, the kinetic-energy density at 50-day periods is also shown; this measure was chosen because it is independent of record duration and is resolved by a substantial number of measurements. Total record variance, which has been computed by many authors, is dependent upon the length of the measurement. A nominal shallow (circa 500–600-m) and deep (circa 4000-m) value is displayed in figure 11.10B. The variety of specific depths shown in figure 11.10B is one measure of the difficulty of making comparisons using the existing data base. Note also that to the extent that the local stratification changes from region to region, there can be a change in record kinetic energies due to the quasi-kinematic effects of wave refraction in a nonhomogeneous medium. Fortunately, these changes are rarely very large.

In the spectral displays, a line with slope −2 at a fixed energy level is drawn on the log–log form of the plots to make easier a comparison of energy levels and spectral shapes from one location to another. Most of the available long records (figure 11.14) are from the western North Atlantic [for additional displays see Schmitz (1976, 1978) and Richman, Wunsch, and Hogg (1977)]. An eastern Pacific record is displayed in figure 11.15A (Hayes, 1979a), the values from another Pacific record (B. Taft, private communication) are shown in figure 11.10B, and one from the Drake Passage area is shown in figure 11.15B (from Bryden and Pillsbury, personal communication). Although the energy levels
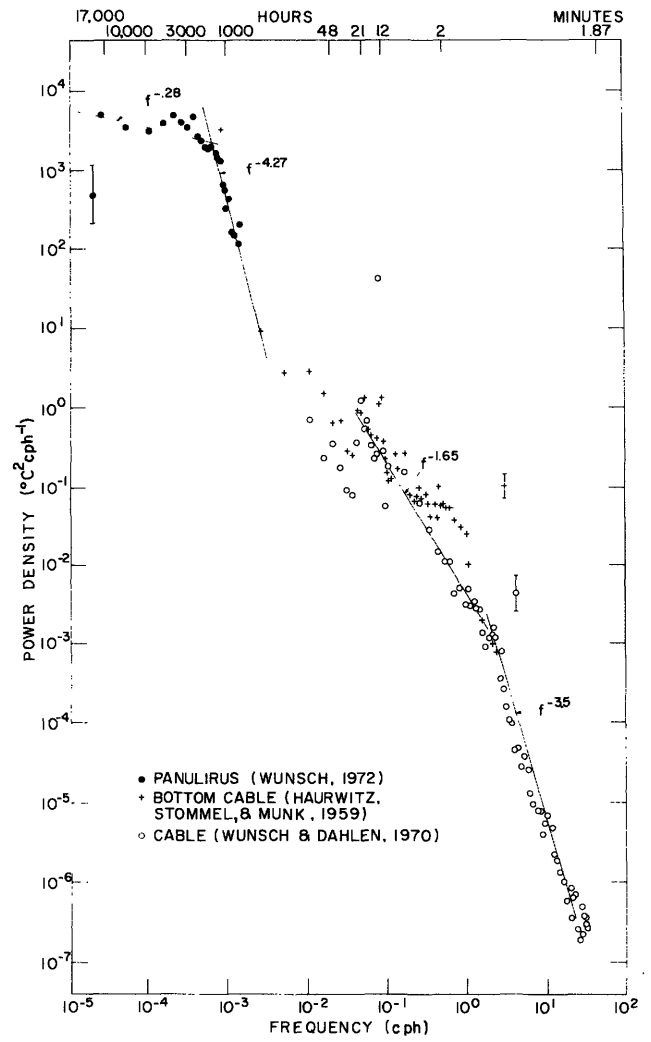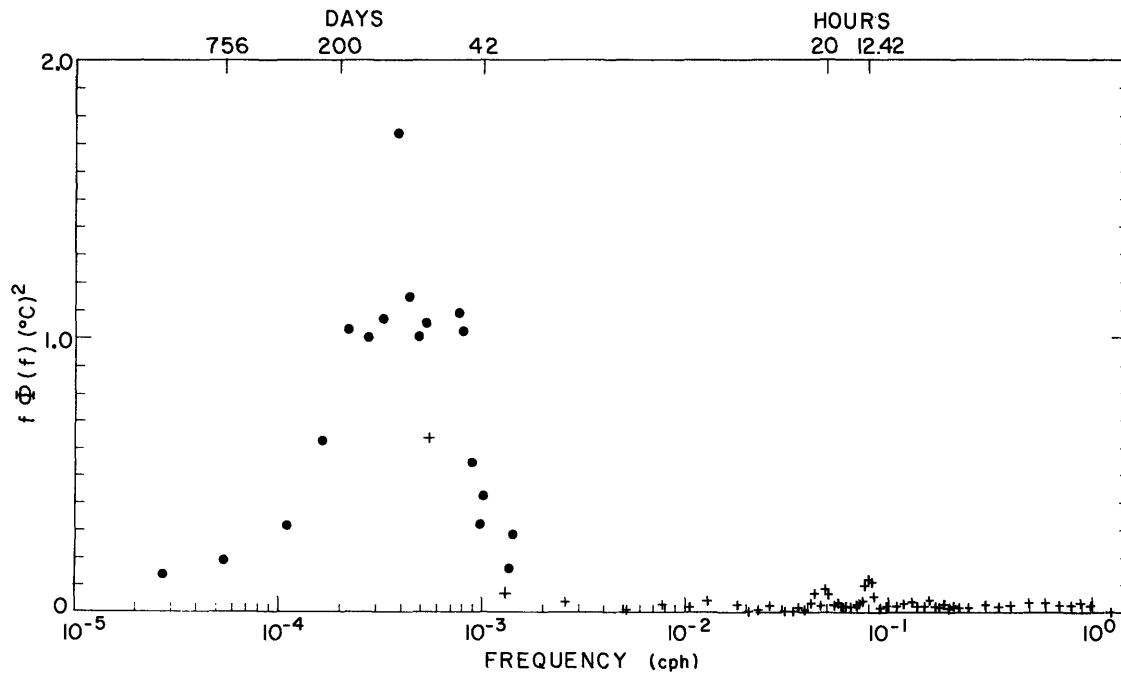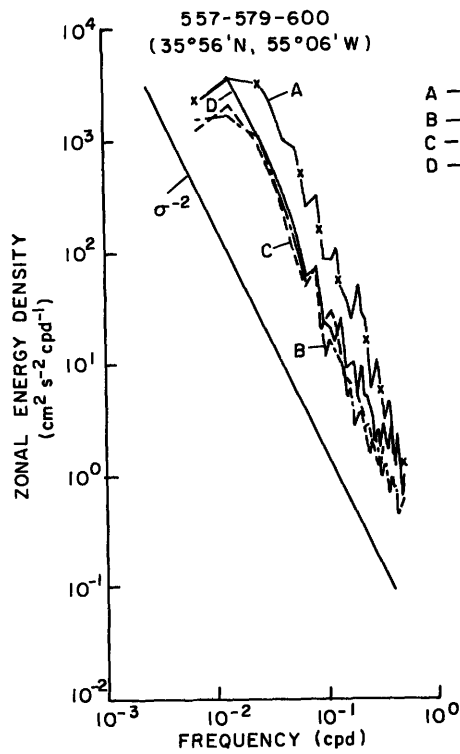
Figure 11.13 Spectrum of temperature at Bermuda in the main thermocline from record duration of 13 years. (A) is log-log form, (B) variance-preserving form. Bulk of energy is around 100-day periods. (Wunsch, 1972b.)
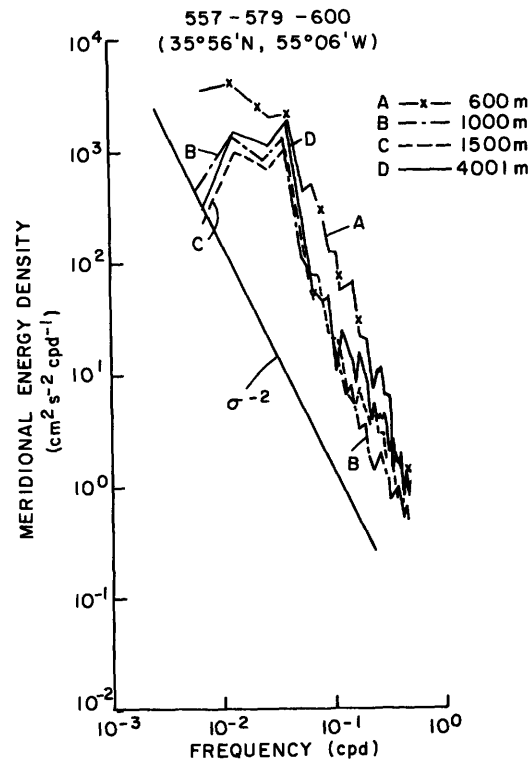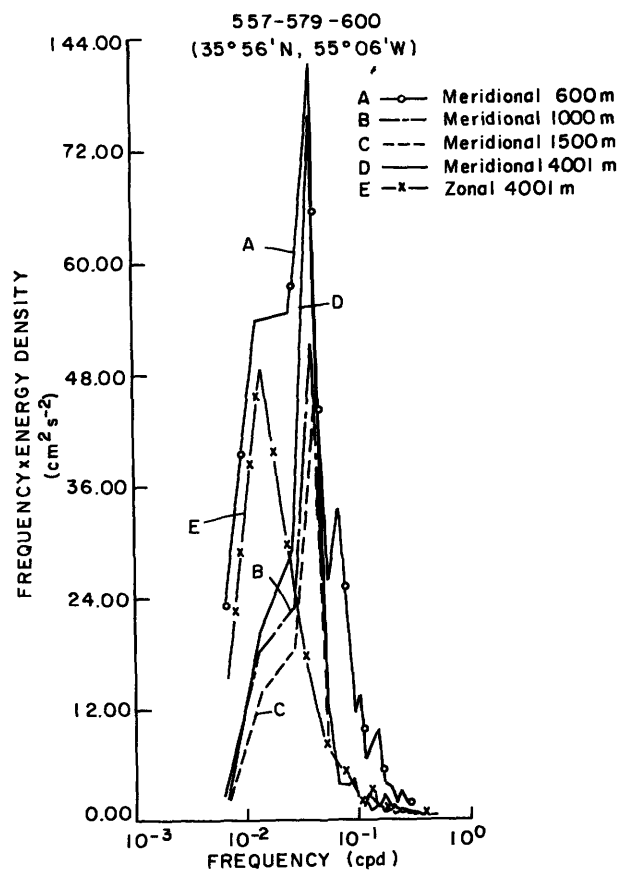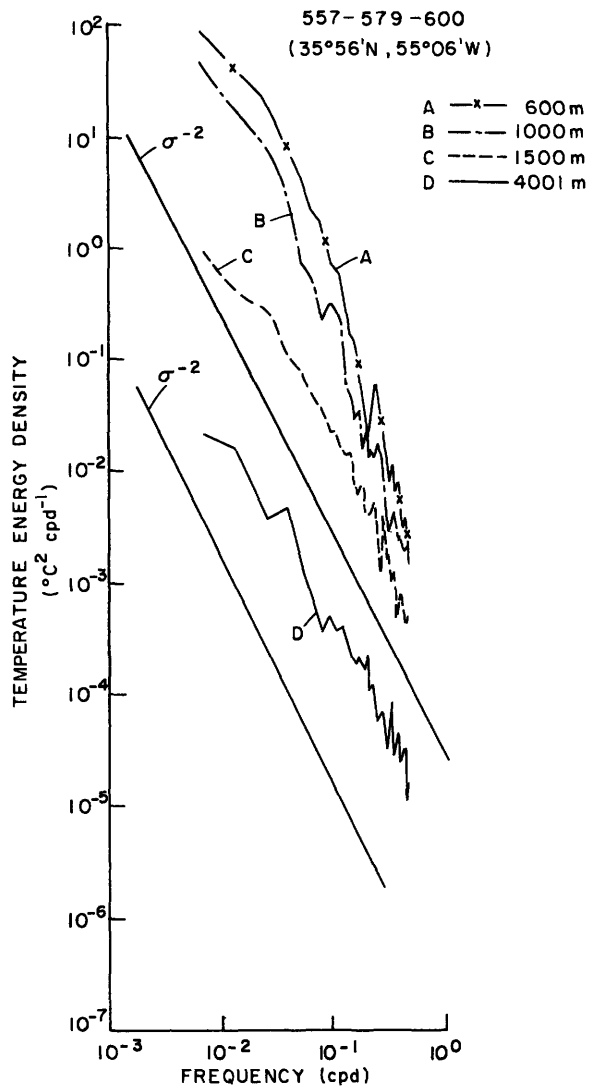
(11.13A)

(11.13B)

Carl Wunsch

(11.14A)

557-579-600
(35°56'N, 55°06'W)

ZONAL ENERGY DENSITY ($cm^2 s^{-2} cpd^{-1}$)

σ^{-2}

FREQUENCY (cpd)

A —x— 600 m
B —·— 1000 m
C —— 1500 m
D —— 4001 m

(11.14B)

557-579-600
(35°56'N, 55°06'W)

MERIDIONAL ENERGY DENSITY ($cm^2 s^{-2} cpd^{-1}$)

σ^{-2}

FREQUENCY (cpd)

A —x— 600 m
B —·— 1000 m
C —— 1500 m
D —— 4001 m

(11.14C)

557-579-600
(35°56'N, 55°06'W)

FREQUENCY×ENERGY DENSITY ($cm^2 s^{-2}$)

144.00
72.00
60.00
48.00
36.00
24.00
12.00
0.00

FREQUENCY (cpd)

A —o— Meridional 600 m
B —— Meridional 1000 m
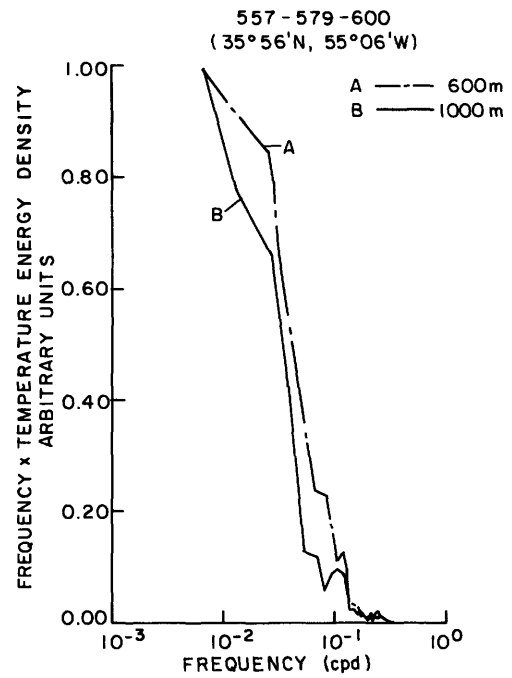C —— Meridional 1500 m
D —— Meridional 4001 m
E —x— Zonal 4001 m

Figure 11.14 Spectra of velocity and temperature from a variety of positions in the western North Atlantic, log–log forms are always shown. In a number of cases, the variance-preserving plots are also displayed. On log–log forms a straight line of slope 2 and fixed energy level is shown for comparison purposes. Variance-preserving plots for temperature are on an arbitrary scale because large range of values defeats a linear-scale display.

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Low-Frequency Variability of the Sea

557-579-600
(35°56'N, 55°06'W)

A —x— 600 m
B —·—· 1000 m
C ----- 1500 m
D ——— 4001 m

TEMPERATURE ENERGY DENSITY (°C² cpd⁻¹)

σ^{-2}

σ^{-2}

FREQUENCY (cpd)

(11.14D)

557-579-600
(35°56'N, 55°06'W)

A —·—· 600 m
B ——— 1000 m

FREQUENCY x TEMPERATURE ENERGY DENSITY ARBITRARY UNITS

FREQUENCY (cpd)

(11.14E)

567-580-612
(31°36'N, 55°05'W)

A —·—· 628 m
B ----- 1028 m
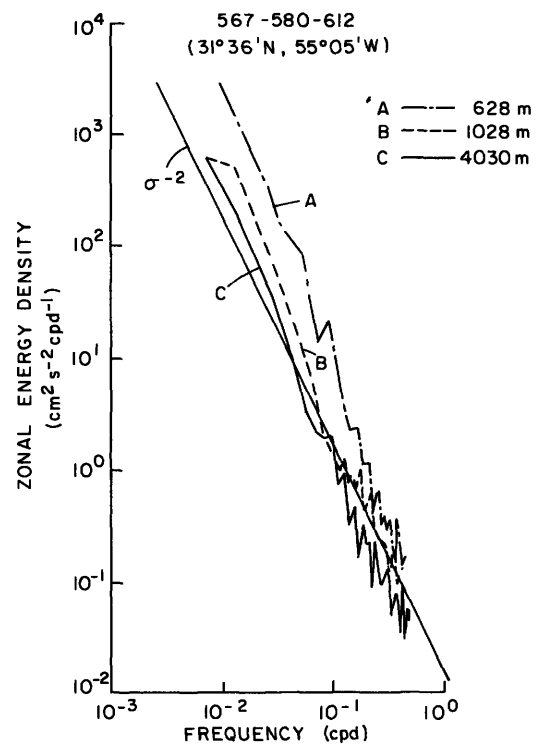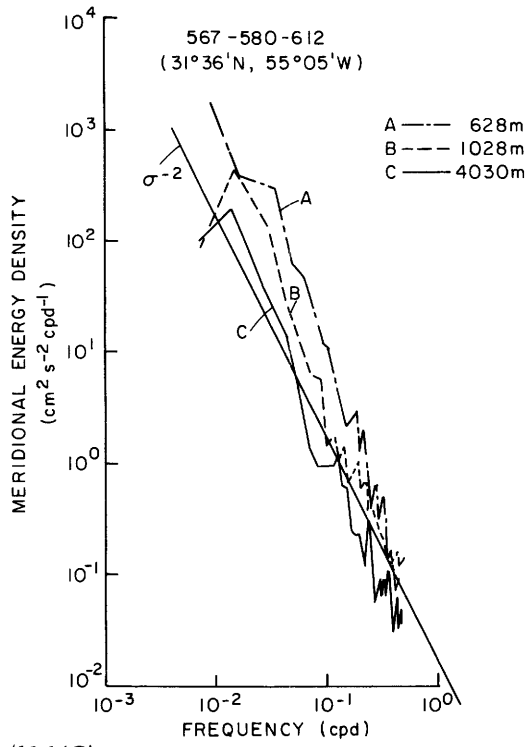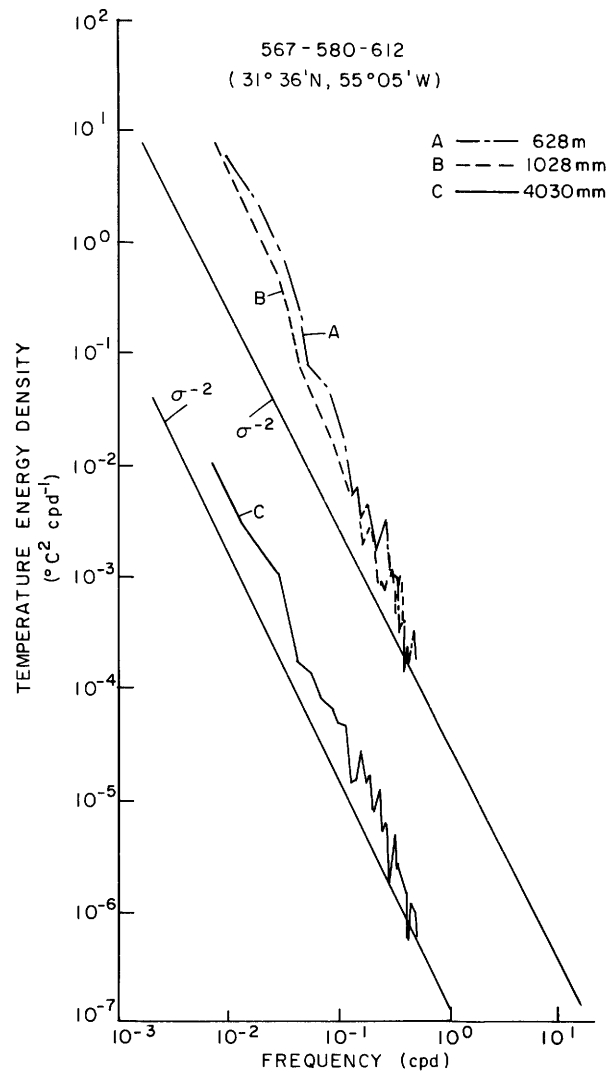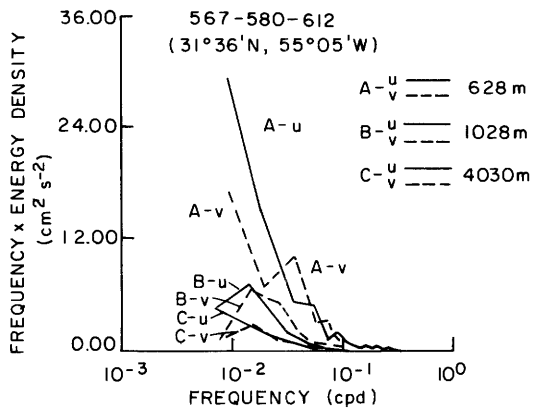C ——— 4030 m

ZONAL ENERGY DENSITY (cm² s⁻² cpd⁻¹)

σ^{-2}

FREQUENCY (cpd)

(11.14F)

(11.14G)

(11.14H)

(11.14I)

(11.14J)

Low-Frequency Variability of the Sea

542
(28°01'N, 69°39'W)

A ——— 495 m
B - - - 1499 m
C -·-·- 4000 m

ZONAL ENERGY DENSITY (cm² s⁻² cpd⁻¹)

FREQUENCY (cpd)

(11.14K)

542
(28°01'N, 69°39'W)

A ——— 495 m
B - - - 1499 m
C -··- 4000 m

MERIDIONAL KINETIC ENERGY (cm² s⁻² cpd⁻¹)

FREQUENCY (cpd)

(11.14L)

542
(28°01'N, 69°89'W)

A ——— 495 m
B - - - 1499 m

TEMPERATURE POWER DENSITY (°C² cpd⁻¹)

FREQUENCY (cpd)

(11.14M)

542
(28°01'N, 69°39'W)

A – u/v   495 m
C – u/v   4000 m

FREQUENCY x ENERGY DENSITY (cm² s⁻²)

42.00

36.00

24.00

12.00

0.00

FREQUENCY (cpd)

(11.14N)

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Carl Wunsch

623
(27°25'N, 41°08'W)

A ———— 128 m
B - - - - 1426m
C —·—·— 3927m

σ^{-2}

MERIDIONAL ENERGY DENSITY (cm^{2}s^{-2}cpd^{-1})

10^4
10^3
10^2
10^1
10^0
10^{-1}
10^{-2}

A
B
C

FREQUENCY (cpd)
10^{-3} 10^{-2} 10^{-1} 10^0

(11.14O)

623
(27°25' N, 41°00'W)

A - $\frac{u}{v}$ ═══ 128m
B - $\frac{u}{v}$ - - - 1426m

FREQUENCY×ENERGY DENSITY (cm^2 s^{-2})

36.00
24.00
12.00
0.00

A-u
A-v
B-v
B-u

FREQUENCY (cpd)
10^{-3} 10^{-2} 10^{-1} 10^0

(11.14Q)

623
(27°25'N, 41°00'W)

A ———— 128 m
B - - - - 1426m
C —·—·— 3927m

σ^{-2}

ZONAL ENERGY DENSITY (cm^{2}s^{-2}cpd^{-1})

10^4
10^3
10^2
10^1
10^0
10^{-1}
10^{-2}

A
B
C

FREQUENCY (cpd)
10^{-3} 10^{-2} 10^{-1} 10^0

(11.14P)

CLUSTER C
(15°02'N, 54°13'W)

A ———— Zonal 150m
B - - - - Meridional

ENERGY DENSITY (cm^{2}s^{-2}cpd^{-1})

10^4
10^3
10^2
10^1
10^0
10^{-1}
10^{-2}

A
B
σ^{-2}

FREQUENCY (cpd)
10^{-3} 10^{-2} 10^{-1} 10^0

(11.14R)

623
(27°25'N, 41°08'W)

A ——— 128 m
B — — 1426 m

TEMPERATURE ENERGY DENSITY
(°C² cpd⁻¹)

FREQUENCY (cpd)

(11.14S)

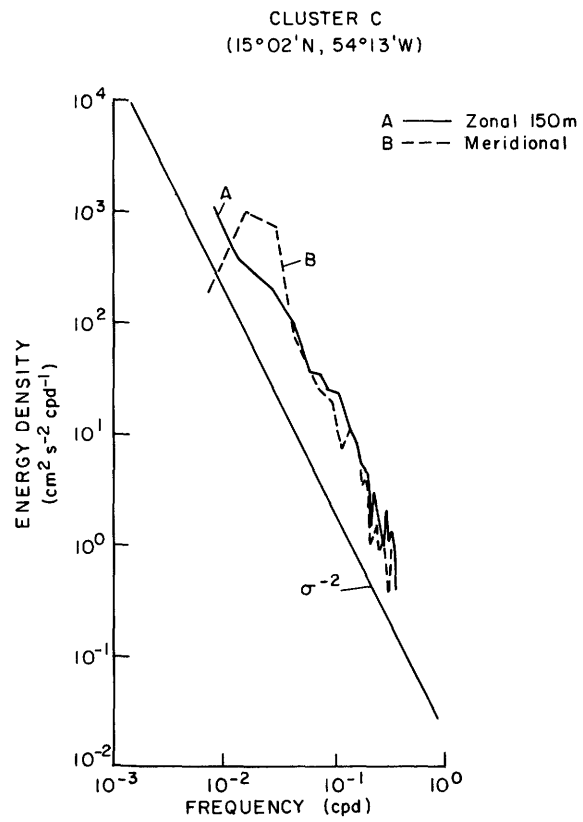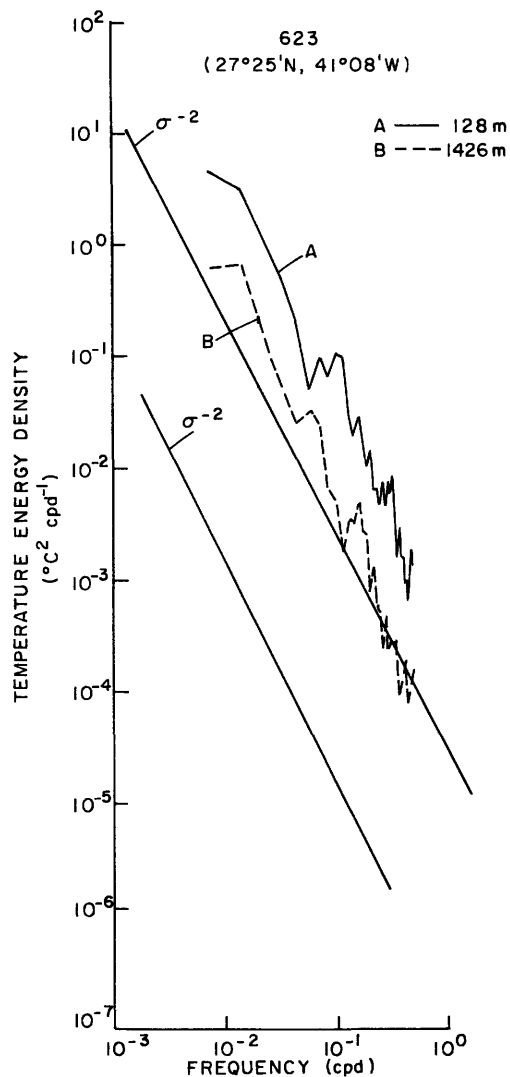vary in the resolved band by an order of magnitude or more (compare figures 11.14A, 11.14F, and 11.15A), the spectra all seem to display some common characteristics. The high-frequency band is never far from σ^{-2}, and there is a tendency to zonal dominance as one enters the low-frequency band. The velocity spectra mostly show some energy enhancement in the eddy-containing band—usually most apparent in the variance-preserving plots, and in common with the sea-level spectra shown above. Overall, the temperature spectra are redder than the velocity spectra and tend not to show the eddy-containing band as clearly. But the much longer record (13 years) used to produce figure 11.13 suggests that these spectra would ultimately drop at lower frequencies as well [the spectra at MODE Center shown by Richman et al. (1977) do display this drop].

Figures 11.14A–11.14E display the spectra of records obtained in the near proximity to the Gulf Stream. As noted by Schmitz, the motion is much more barotropic in character than in the records obtained elsewhere (figure 11.14F). This result is consistent with the observation by Richman et al. (1977) that the fluctuation kinetic-energy density increases much faster toward the Gulf Stream than does the potential energy. The near-Gulf Stream records exhibit a strong peak in the 25–30-day range for meridional velocity; but in the zonal velocity, the peak is shifted toward lower frequency (see figure 11.14C but notice that this peak occurs in the variance-conserving plot). The temperature spectra here are weak and red.

At the site 31°N, 55°W (figures 11.14F–11.14J) the velocity spectra in the thermocline do not show a clear eddy-containing band; there is a definite tendency toward zonal dominance in these records. The red nature of the thermocline velocity spectra had previously been noted in the MODE area by Richman et al. (1977), and by Schmitz (1976).

Overall, it seems clear that the time scales in the thermocline are longer than in the deep water. This is probably a topographic effect, with the bottom gradients serving to provide a stronger effective beta. Theory (e.g., Huppert and Bryan, 1976; Hogg, 1976; Rhines and Bretherton, 1974) suggests that topography can be a source of eddy motions through its interaction with larger-scale flows both in a wave-generation mode, and through destabilization of the larger scales. But there does seem to be a real reduction in the overall energy levels at depth as one enters regions of rougher topography (Schmitz, 1978; Fu and Wunsch, 1979), as though the flow had been "spun down" with the topography extracting energy from the eddy-containing band perhaps through a scattering process. The mechanism is evidently highly baroclinic because the energy levels
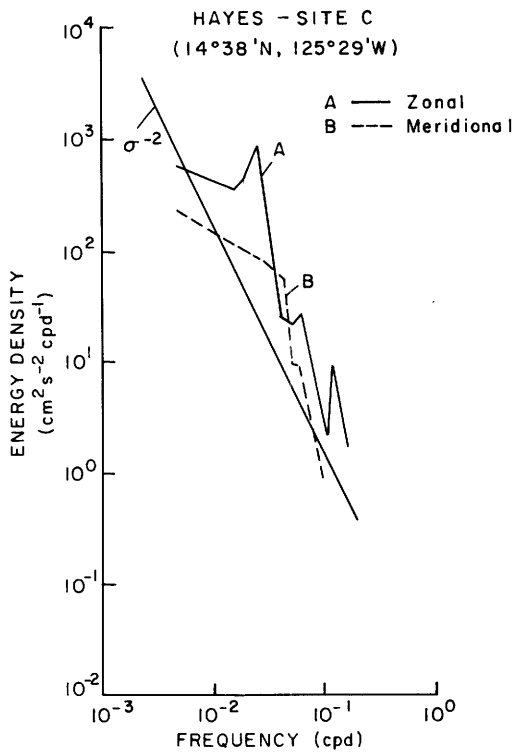
Carl Wunsch

HAYES – SITE C
(14°38'N, 125°29'W)

A —— Zonal
B ---- Meridional

Figure 11.15A Spectrum from eastern Pacific Ocean, replotted from Hayes (1979a).

DRAKE PASSAGE
1519 M

A — Zonal
B -- Meridional

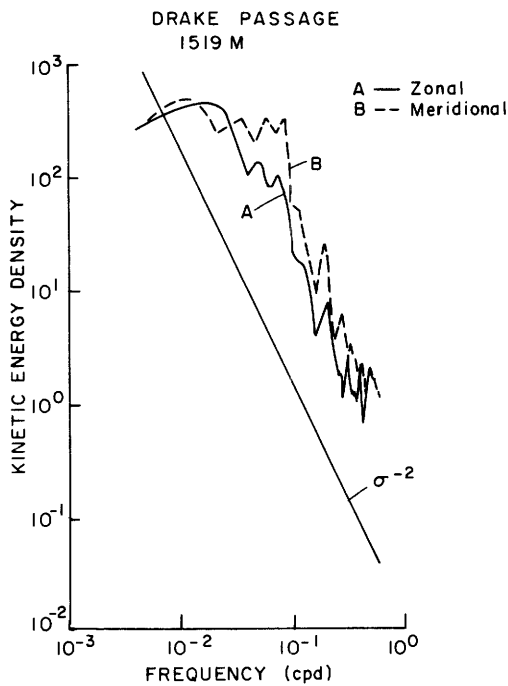Figure 11.15B Spectrum of velocity from Drake Passage, replotted from Bryden and Pillsbury (1979, personal communication).

in the thermocline seem relatively unaffected by topography (compare figures 11.14K and 11.14O, one over the Hatteras abyssal plane, the other over the mid-Atlantic ridge). There is no adequate theory of this process at the present time; even numerical models do not yet adequately handle topographic effects.

Rhines (1977) and Salmon (1978) have suggested that as energy levels increase there is a tendency for motions to become more barotropic, and this seems to be true in most of the records to date (especially near the Gulf Stream; see figures 11.14A–11.14E). Unfortunately, however, most of the energetic records have also been obtained over smooth topography, and it is not yet possible to separate in the observations an internal dynamic mechanism that is purely energy-level dependent from the direct effects of topography.

At any given location, a detailed discussion of the mesoscale becomes intricate (e.g., Richman, Wunsch, and Hogg, 1977; Freeland, Rhines, and Rossby, 1975). The motions vary with depth, location, and frequency even within comparatively small geographical regions. At this time, it is not really clear what the important characteristics of the mesoscale are. The enormous spatial variability indicated in figure 11.10B suggests that spectral representations in the wavenumber domain—which are equivalent to an assertion that the process is spatially stationary—may not be appropriate.

To a good first approximation, most of the baroclinic energy can be found in a form in which the thermocline simply moves up and down, the entire water column moving together (e.g., MODE Group, 1978; Richman et al., 1977). This dominance of the "lowest mode" is in striking contrast to the mixture of high modes required to describe internal-wave observations (chapter 9). The simplest explanation of this lowest-mode character of the observations is in the tendency of quasi-geostrophic nonlinear interactions to drive the motion toward larger scales both in the vertical and horizontal (Charney, 1971a; Rhines, 1977; Fu and Flierl, 1979). There are some features (figure 11.12) for which this characterization does not apply, but they seem uncommon and short-lived except in the Artic Sea (Hunkins, 1974). In the Arctic the eddies Hunkins describes seem closer to a second baroclinic mode in structure. But unlike much of the North Atlantic variability, they also include anomalous water carried along in the core as a kind of "bubble." They may be generated through the overflow process whereby Bering Sea water enters the Arctic Sea, and their physics may resemble the Mediterranean water bubble described by McDowell and Rossby (1978).

In the open ocean, the more common form of first-mode motion is confined to periods of about 100 days and longer. At the shorter periods in the isotropic band along the high-frequency sloping "face" of the velocity

spectra, there is little coherence in the vertical, suggesting very short vertical scales, and the motion has much of the character one would expect of the geostrophic turbulence models [i.e., the horizontal kinetic energies are isotropic, the energy is partitioned equally between potential energy and the two kinetic energy components, and in the vertical the energy varies with the buoyancy frequency $N(z)$ (Charney, 1971a)].

The simple vertical movement of the thermocline superficially resembles the lowest dynamic mode on a linear β-plane. But a quantitative examination of its form suggests a distortion, perhaps due to a larger scale shear, and a coupling of several modes characteristic of nonlinear motions (Davis, 1976; Richman et al., 1977). In the western North Atlantic, it appears that the time scales are too short to be describable by linear dynamics, and any hope that one could describe the low-frequency variability by a simple superposition of Rossby waves seems slight, although this has not prevented attempts at doing so.

On the other hand, Bernstein and White (1974) claim that in the eastern Pacific XBT observations of the upper ocean give baroclinic signals that are consistent with elementary Rossby-wave dispersion curves. Their result is agreement with the observation (Fu and Wunsch, 1979) that in moored measurements in the central Atlantic near the mid-Atlantic Ridge the time scales appear to be longer and begin to approach those to which we might apply linear theory. There is a suggestion here that the westward intensification in the eddy field—an analog of the westward intensification of the mean circulation—also may drive the eddy flow into a more nonlinear range, much as one can postulate a nonlinear Gulf Stream in a more nearly linear interior mean ocean flow.

The resources required to study the detailed local dynamics of the eddy field are very great and such studies have rarely been attempted. With the possible exception of Gulf Stream rings, and other features in the immediate proximity of the Gulf Stream (and by presumption near other western boundary currents), the motions are indistinguishable from geostrophic balance. This near-geostrophic balance, which one anticipates on the basis of simple scale analysis (N. A. Phillips, 1963), dominates the momentum balance and precludes direct study of the nongeostrophic terms in the governing equations that could lend insight into the source and sinks of the motions. To proceed, one is forced to study the vorticity balance that eliminates the lowest-order relation. Attempts along these lines have been made for the MODE-1 data by Bryden and Fofonoff (1977) and by McWilliams (1976). For that place at that time a specific balance can be identified in the quasi-geostrophic vorticity equation, although

McWilliams showed that on the average (as opposed to the instantaneous balance) a linearized form seems to work well.

The question raised in the introduction [equations (11.1)-(11.4)] concerning the effects of variability, in particular the mesoscale eddy field on the mean circulation, is difficult to answer. The large gradients of energy density described by Schmitz (and made obvious in the figures 11.11B and 11.14) suggest the presence of large Reynolds-stress gradient terms of the form

$$\frac{\partial}{\partial x}\langle u'^2\rangle, \quad \frac{\partial}{\partial y}\langle v'^2\rangle,$$

where u', v' are the fluctuation-velocity components. The dynamic implications of these terms, while clear in principal, are obscure in practice because the complete suite of Reynolds stresses has not been measured. In the immediate vicinity of the Gulf Stream, Thompson (1977) has argued that the off-diagonal terms $\langle u'v'\rangle$ show a convergence of momentum into the Gulf Stream consistent with a Rossby-wave energy flux in the opposite direction. Away from the immediate vicinity of the Gulf Stream, the significance of these off-diagonal terms is not understood (a related question is the meaning of the varying ratios of meridional and zonal energies visible in the spectra displayed here).

Eddies may also carry a heat flux through Reynolds terms of the form $\langle u'T'\rangle$, $\langle v'T'\rangle$, etc. Except for Bryden (1979), who reports a significant poleward eddy heat flux in the Drake Passage and similar reports from the Florida Straits, no one has succeeded in measuring statistically significant eddy heat fluxes. The correlations between temperature fluctuations and velocity fluctuations in the regions where measurements exist are sufficiently weak that it appears that much longer records will be needed to measure significant values—which, however, may be so small as to be dynamically meaningless. Both the Drake Passage and Florida Current results may be typical of strong jetlike currents over topography rather than of the open sea. Many of the features of these two regions are similar [compare Wunsch and Wimbush (1977) and Baker et al. (1977)].

The absence of significant eddy heat-flux terms also seems to rule out open-sea baroclinic instability as an important source of eddy energy (Gill, Green, and Simmons, 1974), although the upper ocean has not been sampled adequately. Such an instability requires that the eddies extract available potential energy from the mean flow, thus generating an eddy heat flux down the mean temperature gradient. Even in the region of the Atlantic North Equatorial Current (Keffer and Niiler, 1978), which had seemed a strong candidate for such instability (Gill, Green, and Simmons, 1974), the eddy heat-flux terms are small. One presumes such instabilities (or more likely a mixed barotropic–baroclinic

instability) must occur in the near-Gulf Stream region but the difficulties of measurement in the strong velocities that occur there has precluded quantitative estimates of the heat flux-terms. (Strong mooring motions induced by the high velocities produce fictitious temperature signals coherent with the velocity field.)

If we rule out open-sea generation by meteorology and open-sea baroclinic instability as mesoscale eddy sources, we are left only with topographic generation and generation and radiation from strong boundary currents as possible significant eddy sources. Both are difficult to evaluate quantitatively. Schmitz (1978) and Fu and Wunsch (1979) showed that the major effect of topography seems to be the spin-down of the deep ocean layers relative to the thermocline. Topography would thus more closely resemble a sink of eddy energy than its source. At the present time, then, the strong boundary currents seem the most likely *major* source of eddy energy.

These general results, coupled with the intense concentration of low-frequency energy near the western boundary currents, suggests that the eddy field may indeed be dynamically unimportant except in the immediate vicinity of the boundary currents. The presence of a strong recirculation on the flanks of the Gulf Stream System (Worthington, 1976; Stommel, Niiler, and Anati, 1978; Wunsch, 1978a; Reid, 1978; and see chapter 1) is associated with an intense barotropic eddy field (Schmitz, 1978) and a large resident population of Gulf Stream rings (Richardson, Cheney and Worthington, 1978). A discussion of the Gulf Stream System variability may be found in chapter 4 and will thus not be pursued here. But at this time the only strong evidence for the importance of the variability in the dynamics of the mean flow is in the western boundary current regions. Theories (Pedlosky, 1977) show that the interior field could be largely radiated from the Gulf Stream System and from moving Gulf Stream rings (Flierl, 1977; see also chapter 18).

The mesoscale and other spatial scales of variability become bound up with almost all other aspects of oceanography, including biological, geological, and chemical as well as physical oceanography. To the extent that the variability transports properties, nutrients and other tracers can be expected to be carried along. Eddies will scour the bottom, confusing the interpretation of sediments, and indeed will carry sediments with them from one region to another, possibly quite contrary to the overall time-mean flows. On the physical side, the variability will interact with internal waves and fine structure on the short-wavelength end of the spectrum, and there is probably an interaction between the mesoscale variability and the interannual fluctuations at the other end. We have not dealt directly here with any of these questions; to do so would require a complete discussion of almost all aspects of oceanography and would be premature in any case.

11.3 Summary and Conclusions

The extent to which the ocean undergoes large-scale climatological fluctuations remains uncertain in the face of difficult sampling problems and the short duration over which appropriate instrumentation has existed. Given that large fluctuations exist at least in the sea-surface temperature field, we cannot really distinguish changes imposed by the atmosphere with a static ocean response from those in which large-scale oceanic dynamics are directly involved. Study of the atmospheric spectra displayed here and elsewhere suggests that the forcing and response by the atmosphere is a complex function of position and time scale. With all of the current emphasis on the question of climatic changes, we have been able to do little more than define what we do not know.

Study of the more accessible mesoscale is very much in its infancy and generalizations are dangerous because so little of the ocean has actually been appropriately sampled. At present, one can draw a few qualitative conclusions that may survive future observational programs. The mesoscale eddy field is highly inhomogeneous spatially—both in the horizontal and the vertical. Direct wind forcing and instabilities of the interior ocean are unlikely to provide much energy to the mesoscale. Much of the total mesoscale energy is found in the immediate proximity to the western boundary currents in the region where they are going seaward. It seems, therefore, that these regions are the generators of much of the eddy field energy. Detailed mechanisms are not yet known. The intense recirculations that exist on both sides of the Gulf Stream, the underlying eddy field (Luyten, 1977), and the formation and decay of Gulf Stream rings there are probably related. They suggest an intimate relation between the generation of the eddy field, the maintenance of the eastward going jet, and the transition of the jet into the comparatively quiescent interior in a complex dynamic linkage that we can only vaguely understand. One would guess that the next 10 years will find many of the keys to these puzzles.

The spectra that are displayed throughout this paper suggest that oceanic low-frequency variability does have some characteristic features independent of geography. An eddy-containing band between 50 and 150 days is a feature of most regions. Almost all the spectra show an isotropic band at frequencies higher than the eddy-containing band with a slope not far from −2. (In the immediate vicinity of topography the isotropic band may show more structure.) The interannual band has a distinct tendency to zonality, especially in the