geostrophic flow, and isopycnal analysis, was carried out by Riley (1951). Although his purpose was to study the nonconservative concentrations and to derive rates of oxygen utilization at various depths, he cast his study in the framework of the general circulation and provided flow patterns along various σ_t-surfaces (from 26.5 through 27.7), the deepest extending to about 2100 m in the central South Atlantic. These flow patterns, while not carried as far as they might have been had they been the principal purpose of investigation, are based upon salinity, oxygen and nutrient data (gridded by 10° intervals of latitude and longitude), as well as the density field, and show some remarkable features.

The scale of the grid he used eliminated some features, of course, and he missed the Gulf Stream return flow except in his deepest layers, and for the upper circulation in the subtropical zone he found only the large anticyclonic gyre that Sverdrup et al. (1942) had mapped. He did show that a branch of the subarctic cyclonic gyre extends southward from the Labrador Sea, forming a substantial Gulf Stream undercurrent as far south as Cape Hatteras, carrying waters of high-oxygen and low-nutrient concentration from the Labrador Sea along the western boundary. He found a poleward subsurface flow along the eastern boundary of the North Atlantic carrying northward waters of high salinity from the Mediterranean outlfow and of lower oxygen from the eastern tropical zone. He found that some part of the southward-flowing North Atlantic Deep Water turns eastward near the equator, carrying waters of higher oxygen content eastward between the two eastern tropical zones of low oxygen. Much of this, of course, was quite similar to the earlier results of Wüst and Defant. He did not attempt to carry their studies far forward, but to array them better for his particular study. His principal interests were in estimating the utilization of oxygen and the regeneration of nutrients, and the depth ranges and rates at which these processes occur. He found, using his estimated circulation patterns, that the total oxygen consumption and phosphate regeneration below the σ_t-surface 26.5 (average depth about 200 m) represent the utilization of about one-tenth of the surface production of organic matter by phytoplankton. This is consonant with later findings of Menzel and Ryther (1968), using measurements of dissolved organic carbon, that nearly all regeneration of nutrients takes place above about 500 m, and that in the deeper waters oxygen and nutrients are much more nearly conservative characteristics than in the upper levels. They are not entirely conservative, of course, even at great depth, as Fiadeiro and Craig (1978) have emphasized.

It is, perhaps, unfortunate that Riley's (1951) work came later than Munk's (1950) study. Otherwise, it might have stimulated a more thorough investigation, even with those limited data, of the variation of flow patterns with depth that might have been carried out concurrently with the studies of total transport. Both approaches merited further investigation, but it appears that the impact of Munk's very exciting paper, using one approach, had already engaged the attention of many investigators, and Riley's approach was not so quickly followed, even as more adequate data and methods became available. It has not been ignored or forgotten, of course; one very intensive continuation of his study, attempting to derive rates of deeper circulation, is the GEOSECS program.

It is also worthwhile to note that the differences in approach were not only conceptual but also practical. The Sverdrup transport concept allowed investigators to perform complex studies upon an idealized (homogeneous, steady, two-layer, flat-bottomed, etc.) ocean under an idealized or realistic wind field and to achieve important results in terms of total transport. Riley's sort of approach required the assimilation and manipulation of large quantities of data (though still perhaps too few and of uncertain quality) in order to perform the calculations, and to achieve quite different sorts of results (subsurface flow patterns, for example, instead of total transport).

Wyrtki (1961b), in a study of the thermohaline circulation and its relation to the general circulation, emphasized more clearly the density stratification of the ocean and the necessity that models should include not just surface and abyssal flow, but at least two additional layers, the intermediate and deep waters. These two layers have circulations quite different from the other layers and from each other, and their flow patterns obviously cannot be derived from the various assumptions of purely wind-driven homogeneous or two-layer oceans. And even the four layers discussed by Wyrtki (1961b) are simplifications, as he recognized.

3.6 Mid-Depth Studies Using Isopycnal Analysis

The concept that buoyancy forces in a stratified fluid may influence flow and mixing to conserve density more than other characeristics has been a topic of interest for a long time. Examination of characteristics along surfaces defined by various density-related parameters began in the 1930s, both for the atmosphere and the oceans. Various quantities ($\sigma_t, \sigma_\theta, \delta_T, \delta_\theta$, and σ_1, σ_2, σ_3, . . . , referring the density to 1000, 2000, 3000 db), . . . have been employed, and the method has been called "isentropic," "isosteric," "isanosteric," "isopycnic," and "isopycnal." (Hereafter, I shall refer to all the investigations as isopycnal and to mixing along any of the surfaces defined by these parameters as lateral mixing.) None of these quantities is entirely satisfactory because surfaces so defined can represent mixing or spreading surfaces only in various approximations. The problem, of course, is that while such spreading may take place predominantly along such definable surfaces it need not, and indeed cannot, exactly preserve any chosen density parameter. Density is also altered by mixing processes, as examination of the characteristics along such isopycnals makes obvious. This assumption of maximum mixing and flow along such isopycnals remains an assumption, but it has been accepted as one of the useful concepts in studying the ocean.

Isopycnal analysis can lead to some understanding of the mid-depth flow, but until recently most such studies were of the upper waters. The first major studies using the methods of isopycnal analysis were those of Montgomery (1938a) and Parr (1938). Both of these dealt with the upper levels of the ocean, where a simple density parameter, such as σ_t, could be used. At greater depths the choice becomes more difficult. It is interesting to note that in Wüst's (1933) study of the deep Atlantic he showed a vertical section of potential density σ_θ in the deep water. Was he led to do this by some consideration of isopycnal analysis? In any case, he found an inversion—a maximum in σ_θ well above the bottom—and supposed that this would imply a hydrostatic instability. He concluded that either the equation of state or the salinity-chlorinity ratio was not correct and never again dealt with potential density. Instead, he used (Wüst, 1935) the core-layer method in his analysis of the Atlantic stratosphere.

Would the course of investigation of the deep waters of the world ocean have taken a different turn if Wüst had found a way around the inversion in potential density? Ekman immediately provided the proper resolution of the σ_θ inversion in his review (1934) of Wüst's paper. He showed that the vertical gradient of σ_θ is not identical with stability and is not even a useful approximation (in some cases different in sign) below a depth of a few hundred meters. He first proposed the use of different sorts of potential density referred to pressures of 1000, 2000, 3000, . . . db, designated σ_0, σ_1, σ_2, ... This is the concept and notation that Kawai (1966) and Reid and Lynn (1971) have used. Veronis (1972) has provided a thorough exposition of the problems in dealing with density.

Montgomery (1938a) derived a method for calculating the geostrophic shear between a surface of constant specific volume anomaly and a deeper isobaric surface (or a deeper surface of constant specific volume anomaly). This has been used in the upper ocean by various investigators (Reid, 1965; Tsuchiya, 1968; Buscaglia, 1971). At greater depths, however, the specific volume anomaly term (which is in itself not a significant physical quantity, but an offset from an arbitrary standard that varies with pressure) does not correspond to any of the various quantities used in isopycnal interpretations and cannot be used below the upper few hundred meters.

Montgomery's (1938a) study of the upper layers (mostly above 500 m) of the Atlantic Ocean between the equator and 30°N was the first attempt to discuss the concept of isopycnal analysis and its practical application, and to implement it over a substantial area of the ocean. It was based upon almost the same data set that Defant (1936) had used to derive circulation from the salinity maximum and the vertical density gradients (used to determine "discontinuity levels"). Both of these studies were hampered by the lack of north–south lines of stations in a predominantly zonal circulation system (still a limitation in much of the ocean), but the isopycnal method seemed to be the more fruitful. Later work, with a more complete data set (Cochrane, 1963, 1969) appears to support Montgomery's interpretations. It is interesting to compare Montgomery's maps with those recently produced in Merle's (1978) atlas, which uses the larger data base now available.

Clowes (1950), in a study of the waters surrounding southern Africa, used maps on σ_t-surfaces reaching as deep as 1200 m to identify saline waters flowing southward from the Indian Ocean with the Agulhas Current and then westward into the Atlantic. At the deeper surfaces he used, such as σ_t of 26.5 to 27.25, the contrast between the high-salinity west-Indian source and the lower-salinity eastern Atlantic (Intermediate) waters is particularly marked: both the inflow to the Atlantic and the mingling eddy patterns where the Agulhas Current meets the West Wind Drift are well delineated.

Taft (1963) used isopycnal δ_θ-surfaces to discuss the distribution of salinity and oxygen south of the equator in all the oceans. The greatest depth reached by his isopycnals was about 1500 m. Although the data set he used was sparse, it was well chosen and representative, and he was able to show, among other features, the southward extension from the Arabian Sea of high-salinity, low-oxygen water through the Mozambique Channel, with a part entering the Atlantic. In particular, the maps of the depth of the isopycnals are the first to show for all three southern oceans that the great lens of low-density water corresponding to the subtropical anticyclonic gyre at the surface shifts poleward at greater depths. This had been apparent for both the North and South Atlantic from the *Meteor* atlas and for the North Pacific from the NORPAC atlas (NORPAC Committee, 1960), but had not been mapped for the South Pacific and Indian Oceans. He remarked also upon the differences in salinity (and thus temperature) between the various oceans on the isopycnals chosen. The influence of low-salinity water from the Pacific, entering the Indian Ocean north of

Australia, is clearly evident in the upper 300 m, and some effect may be detected as deep as 1200 m in the eastern area. Wyrtki's (1971) atlas, with additional data, confirms this effect to at least 1000 m in the eastern area, and Sharma (1972) has carried out a more detailed study of this low salinity in the upper waters (at a σ_t of 26.02).

Kirwan (1963), in a study of the circulation of the Antarctic Intermediate Water of the South Atlantic, calculated relative geostrophic flow along isopycnal σ_t-surfaces relative to 2000 db. He showed not only that the subtropical anticyclonic gyre shifts poleward at greater depth, but also that an eastward flow appears north of it, at about 10°S, in the entire depth range he considered (about 300 to 1200 m). While this is consonant with the results of Defant (1941b), Defant did not explicitly accept such a flow pattern as real. Riley (1951) has maps of flow on various σ_t-surfaces, some of which indicate such a flow, but his others do not, and he does not remark upon it. Not surprisingly, the three investigators, using the same data base, though with different methods, derived somewhat similar results in that area. But only Kirwan was concerned with details of flow in that area and explicitly pointed out the eastward flow beneath the surface. Later, Reid (1964a) noted it at the surface across one meridian, and Mazeika (1968) mapped the surface flow over a larger area. Lemasson and Rébert (1973) found it both in the geostrophic shear and by direct measurement. The most recent and comprehensive maps are those in the atlas prepared by Merle (1978).

A similar feature was noted in the Pacific Ocean (Reid, 1961a) when the data were being arrayed to perform a larger-scale isopycnal analysis of the intermediate-depth low-salinity water. Investigations using isopycnal surfaces have nearly always led to consideration of geostrophic shear as well, as the depth patterns of the isopycnals seem to be defined largely by the geostrophic balance to the flow.

Such isopycnal studies have also led to a consideration of vertical diffusion as well, as nearly all large-scale examinations show some obvious evidence of its effect. A particular case is that of the salinity minimum that lies within the subtropical anticyclonic gyre of the North Pacific. Kuksa (1962) had examined the salinity minimum of the North Pacific and used the geostrophic shear between 400 and 1000 db in studying the circulation. Later (Kuksa, 1963), he examined the salinity along an isopycnal ($\sigma_t = 26.75$) and concluded that the origin of the low-salinity water coincides with the shallow subsurface temperature minimum found in the northwestern Pacific in summer (that is, from the winter mixed layer, of which the temperature minimum in summer is a remnant). Reid (1965) has concluded, however, that while the isopycnal range near the core of the minimum does not outcrop in the North Pacific, yet the source of the minimum is at the sea surface there. Vertical diffusion from the low-salinity waters in the mixed layer north of 45°N, through the pycnocline and into the higher-density underlying water, is the only process that can account for the salinity pattern to the south (Reid, 1965).

There have been more isopycnal studies of the Pacific Ocean than of the other oceans, though most of them have dealt with the waters above a depth of 1000 m. A series of papers began with the work of Austin (1960) and Bennett (1963) in the eastern tropical area and continued with Reid's (1965) study of the Intermediate Waters, Cannon's (1966) study of the Tropical Waters, Barkley's (1968) atlas on a set of surfaces from σ_t-values 23.00 to 27.70, and Tsuchiya's (1968) analysis of the upper waters of the intertropical zone. Tsuchiya's study was particularly important in that it investigated the longitudinal extent and depth range of the Equatorial Undercurrent and revealed the sources of its waters. Employing isopycnal distributions and geostrophy, he showed that the major source of the waters of the Equatorial Undercurrent, which are highly saline and oxygen rich, is the South Pacific, though the Undercurrent and North Equatorial Countercurrent in the western area are not separated by a westward flow. It is in the west that the South Pacific waters appear to cross the equator, as Sverdrup et al. (1942) had suggested, though this pattern is stronger at greater depths. He examined also the North Equatorial and South Equatorial Countercurrents, pointing out that Yoshida (1961) had given some explanation of the latter in terms of the curl of the wind stress. Tsuchiya described the flow as lying near 5°S at depths of 200 to 300 m and shifting toward 10°S at shallower depths. The feature had been examined, and a few direct measurements made, by Burkov and Ovchinnikov (1960) and Koshlyakov and Neiman (1965). Tsuchiya (1975) has presented additional information from the more recent EASTROPAC expedition (Love, 1972).

In a study of the Subtropical Mode Water of the Pacific, analogous to the 18° water of the North Atlantic described by Worthington (1959), the characteristics were examined on isopycnal surfaces from the equator to 45°N by Masuzawa (1969), who made some use of the deeper density field in discussing the circulation of this slowly moving central-ocean water mass.

An attempt to approximate an isopycnal surface over a wide pressure range and to map the characteristics along such a surface over the world ocean was made by Reid and Lynn (1971). The spreading of a layer of water formed in the North Atlantic Ocean from a mixture of the Denmark Strait overflow water with the ambient, more saline waters south of Greenland was traced along such an approximated surface by changing the reference pressure wherever the depth of the isopycnal changed by some chosen amount. In this case the reference pressures were 0, 2000, and 4000 db of hydrostatic pressure. The warm, saline water found at the chosen isopycnal in the North Atlantic could be traced southward in the Atlantic, along the Circumpolar Current into the Indian and Pacific Oceans, and back into the southwestern Atlantic. The effect of vertical mixing in altering the density as well as other characteristics is particularly obvious in the deep North Pacific, where salinity and temperature decrease northward along a deep isopycnal ($\sigma_4 = 45.92$) in the absence of any new lateral source of cold water.

Callahan (1972) used a pair of isopycnals (δ_θ values of 50 and 30 cl ton^{-1}, corresponding to σ_θ-values of 27.60 and 27.81) lying somewhat shallower than that chosen by Reid and Lynn (1971) for examining the ocean south of about 40°S. His upper surface corresponds closely to the vertical oxygen minimum of the upper Circumpolar Deep Water, the lower surface approximately to the vertical salinity maximum associated with the Warm Deep Water of the Antarctic that originates from the North Atlantic. He accounted for the low oxygen of the upper circumpolar waters partly by lateral exchange with the very low-oxygen waters of the western Indian Ocean and southeastern Pacific Ocean and found evidence of a western boundary current flowing poleward at depths of 1600–1800 m east of New Zealand.

The isopycnal method has also been used in studies of the Atlantic Ocean. Buscaglia (1971) examined the Intermediate Water in the western South Atlantic using isopycnal distributions and relative geostrophic flow and concluded, as had Martineau (1953), that the low-salinity water from the Antarctic does not extend northward all along South America, as Wüst (1935) had supposed. He found, instead, that it extends along the coast only to about 40°S with the Malviñas (Falkland) Current, and then turns eastward and flows around the anticyclonic gyre, whose axis at the depth of the Intermediate Water lies at about 35°S. The concept of thermohaline flow as a western boundary current does not apply at the depth range of the Intermediate Water in either the South Atlantic or the Pacific. Instead, the circulation seems more like that of the wind-driven anticyclonic gyres recognized in the upper layer, though they appear to be shifted poleward in this depth range.

Similarly, Ortega (1972) used both isopycnal distributions and relative geostrophic flow in his study of the Caribbean. He was able to show that the Intermediate Waters enter only through the southern passages, where the isopycnals lie shallower, in balance with a generally westward transport, and that the warmer and more saline, less dense waters are the dominant part of the incoming water farther north, where the isopycnals lie deeper.

Lazier (1973a) examined the renewal of Labrador Sea Water and showed that renewal takes place in the central part of the sea in winter, and that the cooled, less saline waters spread outward along isopycnals that rise to meet the sea surface near the axis of the cyclonic circulation that obtains in the northern North Atlantic.

Pingree (1972) first studied mixing processes in the deep ocean and found evidence of isopycnal mixing in the small-scale structures. He defined a potential density over a small depth range in terms of one intermediate pressure and called such isopycnals neutral surfaces. Later, Pingree (1973) examined larger patterns and showed that cold, low-salinity waters from the Labrador Sea may extend along such neutral surfaces into the Bay of Biscay. Pingree and Morrison (1973) examined the northward extension of Mediterranean outflow water along isopycnals variously defined and found that this water has maxima in buoyancy frequency both above and below it. [This sort of structure was examined later in the South Atlantic by Reid, Nowlin, and Patzert (1977).]

Ivers (1975) used a variant of Pingree's neutral-surface concept in examining the deep circulation of the northern North Atlantic. His neutral surfaces are normal at every point to the gradient of potential density, such potential density being referred to the pressure at the point in question. He examined five such neutral surfaces, all of which outcrop in the north. The shallowest outcrops in the Labrador Sea and extends to about 900 m near 38°N. The deepest outcrops only in the Norwegian–Greenland Sea and lies as deep as 3000 m along 40°N. From this examination he concluded that there is a cyclonic gyre to the north of the Gulf Stream–North Atlantic Current, and that this gyre, though distorted by the Rockall Bank, Reykjanes Ridge, and Greenland, extends throughout the northern North Atlantic down to 3000 m at least. He found that most of the features of the salinity distribution can be explained by processes of lateral flow and mixing: the notable exceptions are the vertical mixing induced by cooling at the outcrops and by the rapid flow over the sills and through the narrow channels of the Greenland–Scotland Ridge. He attempted an interpretation of the geostrophic shear, not assuming a level or even continuous layer of zero flow, but by entering various constraints upon the velocity wherever measurements of speed were available or where qualitative reasoning, based upon the characteristics along the neutral surfaces, indicated a sense of flow. The result is a qualitative, but internally consistent, circulation field. In addition to deriving flow patterns within the four embayments of the North Atlantic (Labrador Sea, the troughs east and west of the Reykjanes Ridge, and the Rockall Channel), he shows a strong return flow south of the Gulf Stream and a subsurface poleward eastern boundary current.

More recently, Clarke, Hill, Reiniger, and Warren, (1980) have examined the area just south and east of the Grand Banks of Newfoundland, using isopycnal distributions, geostrophic shear, and current measurements. Their purpose was to determine whether there is a branching of the Gulf Stream near the Grand Banks, with a part flowing into the Newfoundland Basin, or whether a separate anticyclonic gyre exists within the Newfoundland Basin, as Worthington (1976) had proposed. They concluded, on the basis of the distribution of characteristics and the relative geostrophic flow, that the branching does occur.

3.7 Comparison of Relative Geostrophic Flow at Mid-Depth with Numerical Models of Transport

There is an interesting correspondence between some of the newer patterns of mid-depth circulation based upon geostrophic flow relative to some deep isobaric surface and the total transports calculated in some of the recent numerical models. This seems especially significant in that the newer pattern seems to have been arrived at quite independently through the two different methods of investigation.

3.7.1 The Density Field

The density field in the North Atlantic Ocean can be represented in part by a map of the depth of a σ_t-surface (figure 3.10). On this map the major trough has the shape of the letter C, with the two arms extending eastward from the western boundary. An analagous pattern exists in the South Atlantic but was not apparent in the zonal data array of the *Meteor* expedition, from which this map was made.

The density pattern in the North Atlantic has about the same shape as that in figure 3.10 over a substantial depth range, as can be seen in the *Meteor* atlas, and is reflected in the appropriate maps of steric height (Leetmaa et al., 1977; Reid, 1978; Stommel, Niiler and Anati, 1978). The trough in figure 3.10 appears as a continuous ridge on the maps of steric height. This C-shaped pattern thus suggests the Gulf Stream and its westward return flow, which turns southward near 70°W, eastward along about 30°N, and finally westward along about 25°N.

The poleward arm of the C is much more clearly defined in these data than the equatorward limb. It was suggested in Jacobsen's (1929) maps (figure 3.6A) of steric height and was very clear in Defant's (1941a) maps of relative flow, but apparently, it was not accepted by most investigators as a real feature until the recent work of Worthington (1976).

A similar pattern appeared marginally in the Pacific from the *Carnegie* maps (Fleming et al., 1945), though the data were sparse. It has already been pointed out that Taft (1963) showed for all three southern oceans that the great lens of low-density water corresponding to the subtropical anticyclonic gyre shifts poleward at greater depths, and that Kirwan (1963) mapped it in detail in the South Atlantic and noted an eastward flow appearing on the equatorward side. All subsequent treatments of the data (Muromtsev, 1958; Reid, 1965; Barkley, 1968; Reed, 1970b; Burkov, Bulatov, and Neiman, 1973; Reid and Arthur, 1975) have shown this shift.

It is illustrated clearly in the world maps of Burkov et al. (1973) of geopotential anomaly at the sea surface and 500 db relative to 1500 db (reproduced here as figures 3.11A and 3.11B). They note that the maximum values of steric height along the axes of the subtropical anticyclonic gyres are farther poleward in the deeper field in all oceans.

The equatorward limb of the C-shape is not such a strong feature and has been much harder to detect. Of course, it must appear to some extent in surface maps that show the return flows of the Gulf Stream (Defant, 1941a; Reid et al., 1977), Kuroshio (Koenuma, 1939; Wyrtki, 1975a; Reid and Arthur, 1975), East Australian Current (Wyrtki, 1975a), Brazil Current (Reid et al., 1977) and Agulhas Current (Wyrtki, 1971). Its eastward part was discussed by Yoshida and Kidokoro (1967) and Hasunuma and Yoshida (1978) for the Pacific.

At depths below the sea surface, the split of the anticyclonic high cell into two parts, connected in the west into a C-shape, can be seen in the South Pacific Ocean on the map of steric height at 800 db relative to 1500 db, reproduced as figure 3.12 from the publication of the Academy of Sciences of the U.S.S.R. (Kort, 1968). At the sea surface the axes of the subtropical gyres in the North and South Pacific lie near 20°N and 20°S in mid-ocean (figure 3.11A). At the 500-db surface they lie near 30°N and 35°S. At the 800-db surface the northern axis is near 35°N and the southern axis near 40–45°S. A new high lies along about 30°S, connected to the other feature along about 180°: This is the C-shape, which is defined by the present data bank better in the South Pacific than in any other ocean.

The C-shaped deep pattern has also been shown in the North Atlantic by Leetmaa et al. (1977). They take into account not only the westward Gulf Stream recirculation but its southward component across 24°N; from their map of steric height, this would appear to be southeastward, as part of the equatorward arm of the C.

Reid (1978) used a somewhat more detailed interpretation of the relative geostrophic flow field in an attempt to account for the distribution of the Mediterranean outflow water in the North Atlantic, and Reid

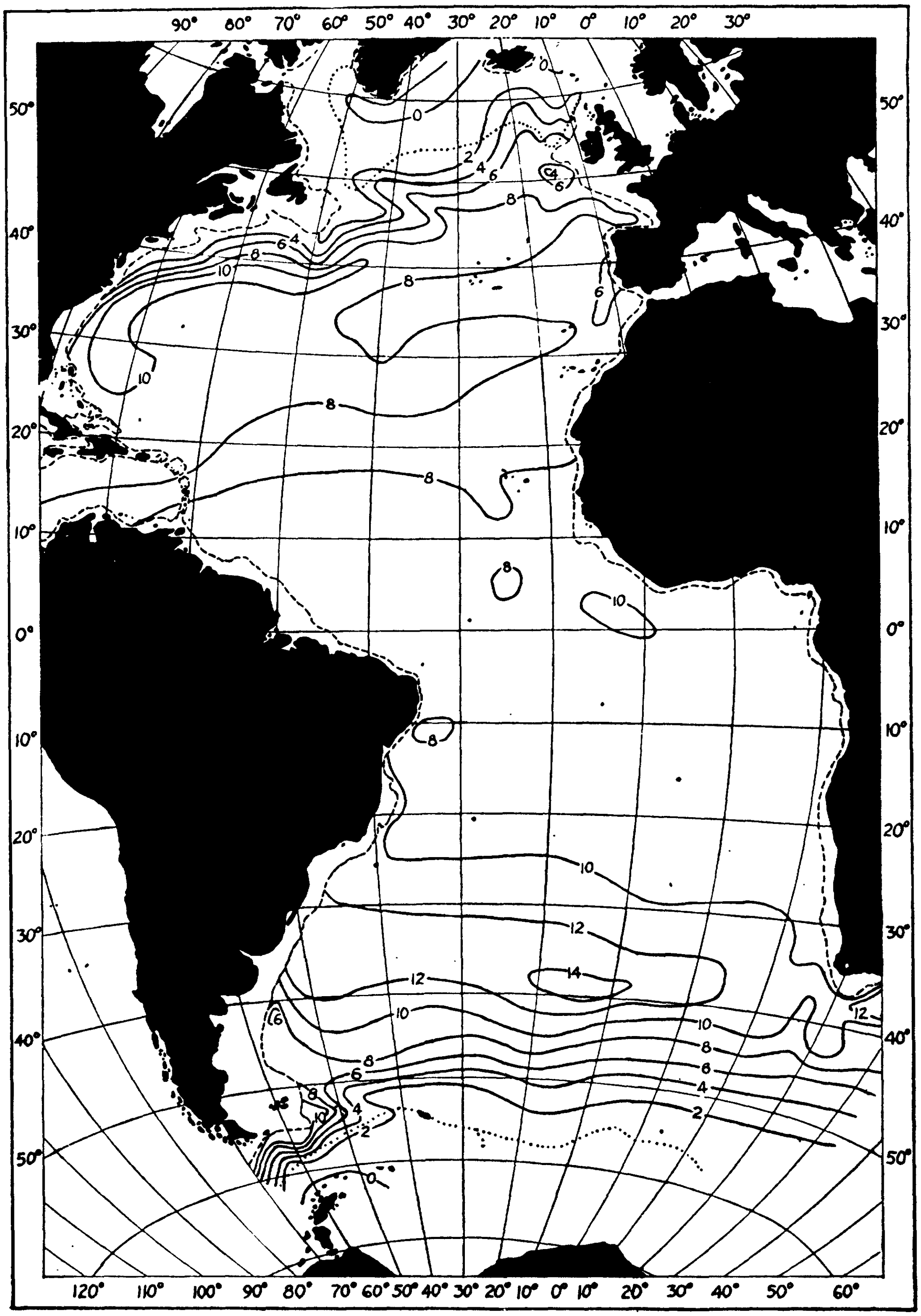
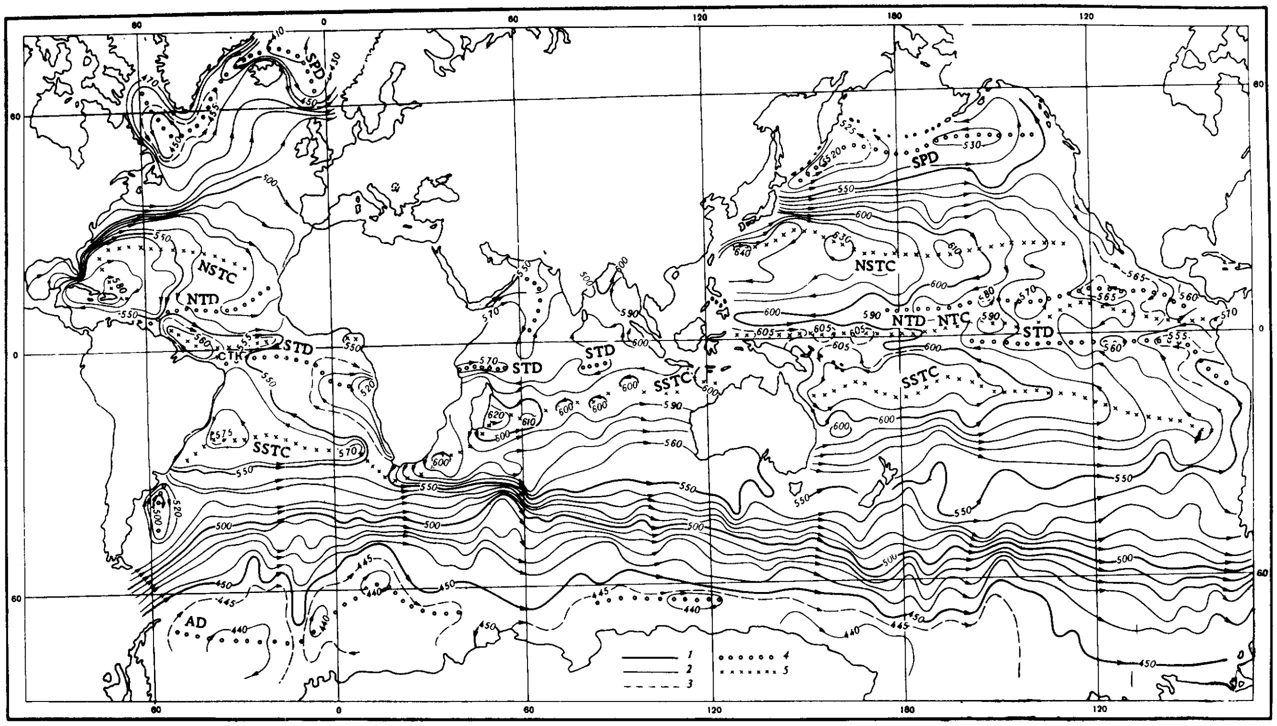
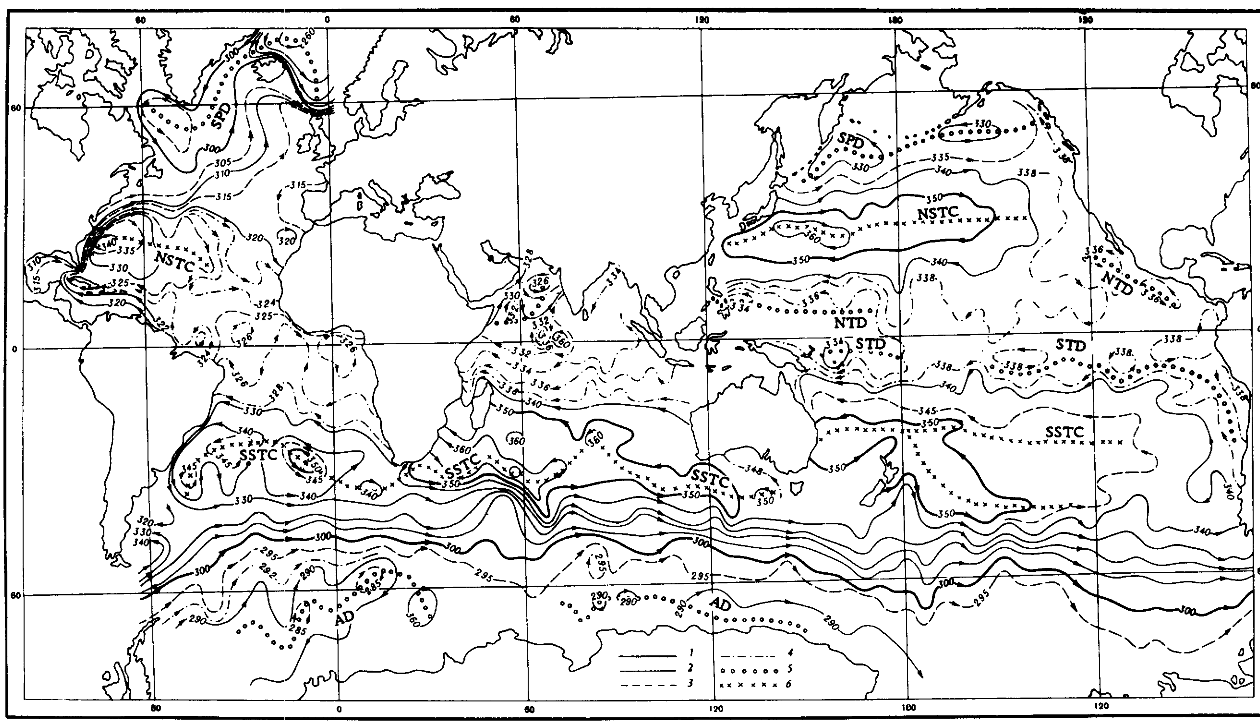


Figure 3.10 Depth in hektometers of the surface where σ_t = 27.4. (Montgomery and Pollak, 1942.)



(3.11A)



(3.11B)

Figure 3.11 Steric height (dynamic cm) (A) at the sea surface, and (B) at 500 db, with respect to the 1500-db surface. (Burkov et al., 1973.)

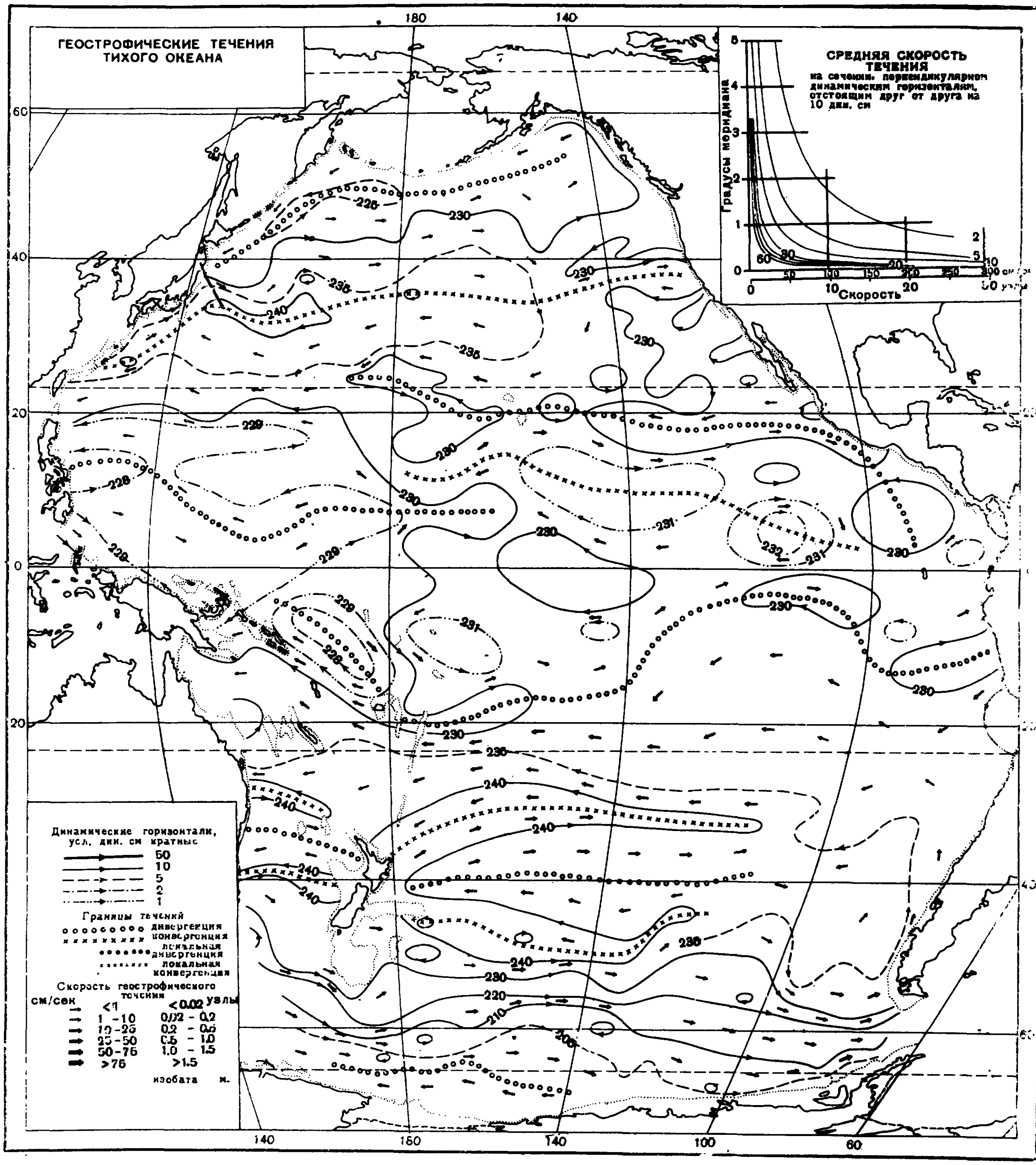


Figure 3.12 Steric height (dynamic cm) at 800–1500 db. (Kort, 1968.)

and Mantyla (1978) presented a map of steric height in examining the mid-depth oxygen pattern in the North Pacific: in both cases the C-shape is clear, and is consonant with the distribution of those characteristics.

3.7.2 Numerical Models

Most of the flow patterns illustrated by the numerical models are of total transport or of surface flow. Models of flow fields at mid-depth (1000–2500 m) are presented but rarely. This may be because there has been so little information with which such results could be compared. Total transports need not correspond precisely to the mid-depth flow, but are probably weighted toward the near-surface flow, which is usually much stronger. It is still useful to compare them with the steric maps, as there seems to be some correspondence between the patterns.

The first of the numerical models to give some indication of this C-shaped pattern was that of Bryan (1963), followed by such studies as those of Veronis (1966b), Blandford (1971), Gill and Bryan (1971), Holland and Lin (1975a), Semtner and Mintz (1977), and Robinson, Harrison, Mintz, and Semtner (1977). The first and last of these are reproduced as figures 3.13 and 3.14.

In the first of these (figure 3.13), only the beginning of the deformation is seen, but it very quickly expands (Veronis, 1966b) and the pattern of figure 3.14 is typical of most of the subsequent studies. I shall not try to describe the bases of such models, but shall borrow from Robinson et al. (1977, p. 191), who stated that their streamfunction (figure 3.14) is "qualitatively similar to nonlinear single gyre experiments by Veronis (1966b) and Holland and Lin (1975a). . . . The interior circulation is qualitatively similar to the traditional linear Sverdrup interior." However, the poleward shift does not require a nonlinear model; for example, it appears in Kuo's (1978) model of a nonhomogeneous ocean, both with and without consideration of bottom topography.

Only Gill and Bryan (1971), who modeled the South Pacific Ocean, noted that the poleward shift at increasing depth bore some resemblance to the actual fields of density and relative geostrophic flow in the South Pacific and cited some supporting evidence. Their model showed an eastward flow equatorward of the shifted anticyclone and connected to it in the west, but did not complete the C-shape with a westward flow farther equatorward.

From an examination of figures 3.13 and 3.14 it is unclear what to call these features. Is what has been referred to as a poleward shift really a shift, or is it that only the higher-latitude part of the C-shaped pattern was recognized at first and a phrase prematurely applied? Do we really have a poleward shift of the anticyclonic gyre with a new but weaker gyre added on the equatorward side and connected in the west, or is it merely that one large gyre, thought at first to be elliptical, now is seen to have a huge and asymmetric dent in its eastward side?

Two studies of the pattern in relation to the distribution of characteristics bear upon this. In one (Reid, 1978), the position of the highly saline waters of the Mediterranean outflow at a depth of 1000 m is interpreted as extending westward across the North Atlantic between the two arms of the C-shaped pattern. Lower values of salinity are seen to the north, west, and south of the Mediterranean outflow. From the salinity pattern alone, one might imagine that there are two separate anticyclonic gyres, that in the north containing Labrador Sea water and that in the south made up of low-salinity water from the South Atlantic. There is no immediate reason from the salinity pattern to suppose that the two anticyclonc gyres are connected in the west, though the pattern of relative geostrophic flow suggests that they are and that the C-shaped steric high surrounds the Mediterranean outflow.

In a similar treatment of the North Pacific (Reid and Mantyla, 1978), however, the oxygen pattern does require a connection in the west between the two anticyclonic arms, and the C-shaped pattern must be complete. The higher oxygen values in the far north must derive from a lower latitude and extend continuously through both arms of the C-shaped pattern.

3.7.3 The Diagnostic Models

Various investigators have used compilations of data to describe the density field and used these as inputs to their models. Arraying a proper set from the available data bank has not been an easy task. Substantial editing of the materials would be required to eliminate biases and random errors from a data base that extends over a period of more than 50 years, and has been collected with some differences in techniques and instrumentation, with most expeditions based on an ad hoc plan to address a particular problem rather than to contribute to a coherent ocean-wide program. Accuracy is most limiting at the greater depths, where both horizontal and vertical gradients of salinity are small, and errors can often exceed the local time-and-space variation.

As a result, most of the diagnostic models have used simplified or smoothed fields and have not yet made extensive use of the density field at great depths. If they do use unedited materials at great depths, a few measurement errors may lead them seriously astray. Work is under way on the preparation of such an edited data set by, for example, Levitus and Oort (1977).

Kozlov (1971), using density at only two levels in the Pacific, found only the poleward shift in the North Pacific, but found the full C-shape in the South Pacific. Holland and Hirschman (1972) used a partly diagnostic

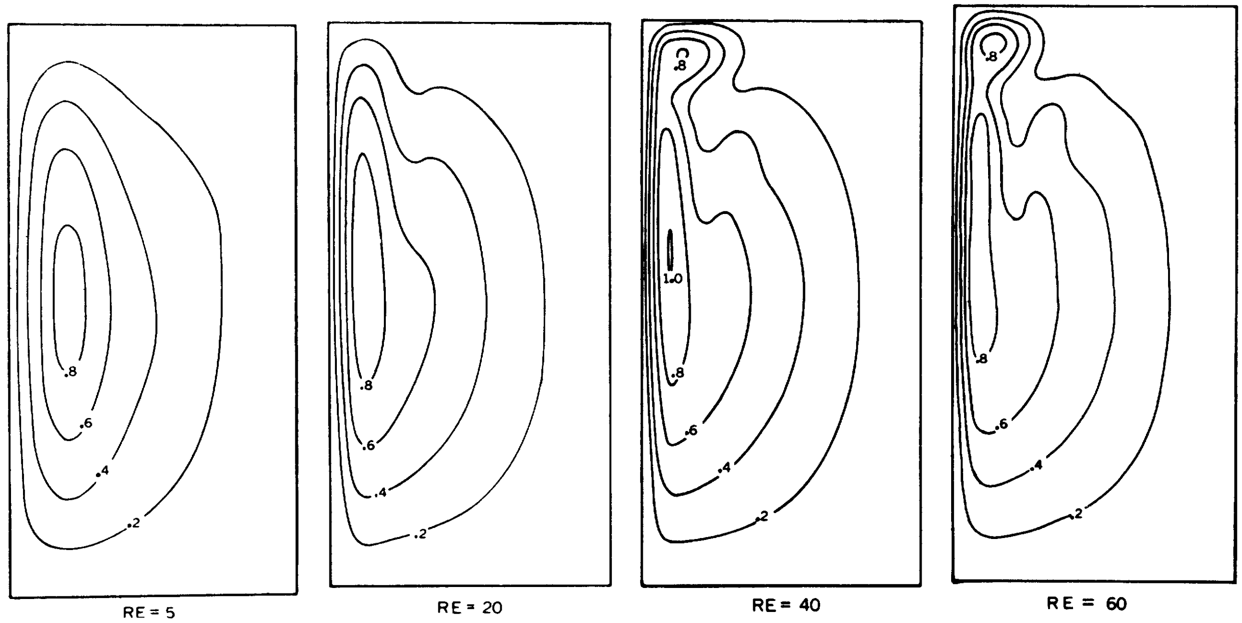


Figure 3.13 Patterns of transport stream functions for various values of Reynolds number: 5, 20, 40, 60. (Bryan, 1963.)

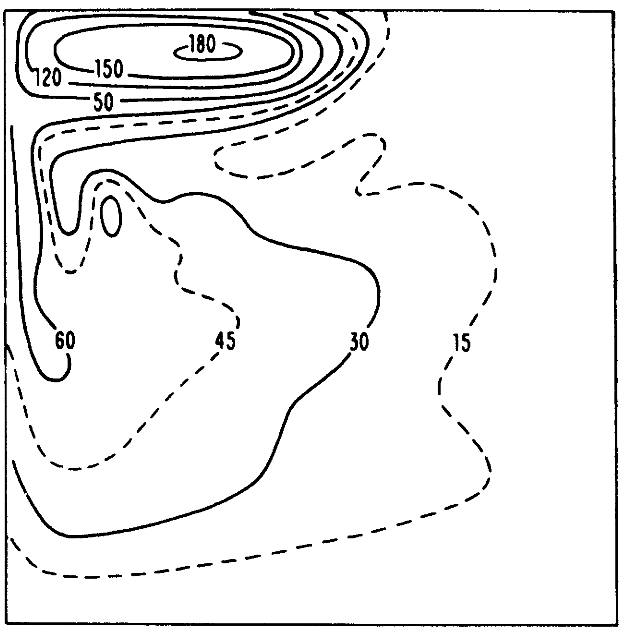


Figure 3.14 Mean transport stream function, in 10^6 m^3 s^{-1}. (Robinson et al., 1977.)

model for the North Atlantic. Their data base for the upper 1000 m matched fairly well the patterns of Defant (1941a), Reid et al. (1977) and Stommel et al. (1978) and showed the C-shape in the density field. The derived surface flow of the diagnostic model showed it somewhat better, but their mass transport showed it less clearly. Their deep flow field (their figure 11) appears to have too many separate gyres to allow for interpretations of the C-shape, but clearly shows the poleward shift. (The feature in their figures 11 and 12 along 15–25°N at about 40°W, referred to as a complex set of currents west of the Mid-Atlantic Ridge, is remarkably like the *Carnegie* effect referred to earlier that obscured Defant's maps, but not Wüst's.)

Cox (1975) used a data set averaged or interpolated to a 1° × 1° grid for the world ocean. His flow field indicates a weak poleward shift, but only in the southern oceans; in only the South Pacific is there a suggestion of the C-shape.

Sarkisyan and Keonjiyan (1975) produced various diagnostic models of total transport and free-surface elevation (surface flow) in the North Atlantic, both with and without bottom relief, and emphasized that consideration of the effects of baroclinicity and bottom relief is essential. All of their results give the poleward shift, or Gulf Stream recirculation, with an eastward flow south of the recirculation, but only one of their maps (surface topography) suggests the complete C-shape.

3.7.4 Other Approaches

It has been supposed that over much of the ocean the speed of flow varies with depth, usually decreasing. The poleward shift with depth and the other limb of the C-shape indicate that direction also changes with depth. Stommel and Schott (1977) also find a change in direction with depth, which they call the β-spiral, when applying their method of calculation of absolute velocity from the density field. The results of their calculations from data surrounding 28°N, 36°W give a flow at 200 m directed about south by west. This is consonant with the typical large elliptical anticyclonic flow at the surface and with the various maps of relative geostrophic flow (Defant, 1941a; Reid et al., 1977; Leetmaa et al., 1977; Reid, 1978). Their calculated flow spirals clockwise with increasing depth and is toward about east by north at 1000 m. This is quite consonant with Reid's (1978) map of flow at 1000 m (geostrophic flow at 1000 db relative to 2000 db): it corresponds to the interior side of the southern arm of the C-shaped gyre.

In a later study, Schott and Stommel (1978) used this method to calculate the absolute geostrophic flow from the density field at several other positions. At 20°N, 54°W they find a weak southwestward flow at 100 m depth, where Leetmaa et al. (1977) and Reid et al. (1977) find a weak southeastward flow. This is very near the axis of the equatorward arm of the C and may not be significantly different. Near 1000 m depth they find a stronger flow toward east by north, matching fairly well Reid's (1978) relative-flow map. In the Pacific, at 22°N, 151°W, they calculate a weak south-southwestward flow at 100 m, consonant with the various maps of surface flow relative to 1000 or 500 db (Reid, 1961a; Wyrtki, 1975a). At 1000 m their calculated flow is directed east by north, consonant with the 1000–3500-db map of the North Pacific prepared by Reid and Mantyla (1978).

A similar calculation made in the North Pacific from data along 35°N and 155°W has been made by Coats (1979). He points out that this is farther north than the 28°N Atlantic position used by Stommel and Schott (1977) and in a different part of the circulation system. He finds the near-surface flow directed east-southeast, turning clockwise downward, and toward west-southwest at 1000 m; he notes that these directions are in agreement with the maps of flow at the surface relative to 1000 db (Reid, 1961a) and at 1000 db relative to 3500 db (Reid and Mantyla, 1978). These agreements may signify only that in these areas the velocity decreases monotonically with depth through much of the depth range considered. If the unknown absolute values at the reference surfaces for the relative-flow patterns, while not zero, are small, then the relative flows are not offset very much from the absolute. In such particular cases, the two methods would agree closely, as they use the same geostrophic shear field.

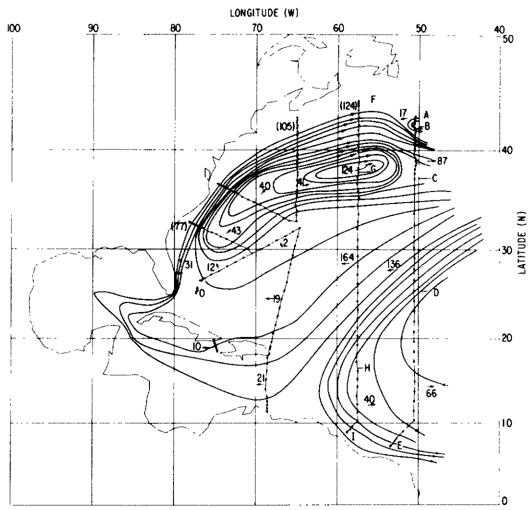
This could, perhaps, partly explain the discrepancy in the South Atlantic between the calculations of Stommel and Schott (1977) and the relative maps. They all (between 20 and 30°S, 10 and 20°W) indicate northwestward flow at the surface, consonant with the 0–2000-db map (Reid et al., 1977), but southeastward flow near 1000 m, quite different from the westward relative flow at 1000–2000 db (Reid et al., 1977). This implies that if there is no error in either the data bank or in its use, there must be a strong eastward flow at 2000 m depth. Unfortunately, there are no north–south data sets in that area that allow this to be examined properly through the density field, though both Wüst and Defant (figures 3.7 and 3.8) found some evidence for such an eastward flow.

Wunsch (1978a) has used selected data to examine the general circulation of the North Atlantic west of 50°W. He has used linear, geostrophic, and mass-conserving models and inverse methods. As in the diagnostic and the β-spiral models, the geostrophic shear is preserved, with some smoothing. In the areas where the shear is strongest, it remains monotonic with depth, and his particular constraints include very low velocity at the bottom. It is not surprising, then, that his total transport pattern (figure 3.15A) shows some resemblance to the maps of relative flow. His map (figure 3.15B) of the transport in the depth range of 17 to 12°C (about 400 to 700 m southeast of the Gulf Stream) shows the return flow westward but not the eastward flow near 25°N seen on the relative maps at 100–1500 db (Stommel et al., 1978) or at 0–1000, 0–2000, or 1000–2000 db (Reid et al., 1977; Reid, 1978). He does show this feature in his map (figure 3.15D) in the range of 4 to 7°C (about 1000 to 1600 m), though it is nearer 20°N. His deepest map (figure 3.15E), at temperatures less than 4°C (below 1600 m), indicates no Gulf Stream south of 35°N and no flow west of 70°W, but a southward turn of the return current east of 70°W extending past 10°N. This is remarkably like Defant's (1941b) map of "absolute" topography of the 2000-db surface (figure 3.8), except north of 35°N in the west. There Defant had inserted a zero surface sloping southeastward from 1000 m to 1900 m across the Gulf Stream, and at greater depths this provided a southwestward flow from the Labrador Sea as far south as Cape Hatteras.

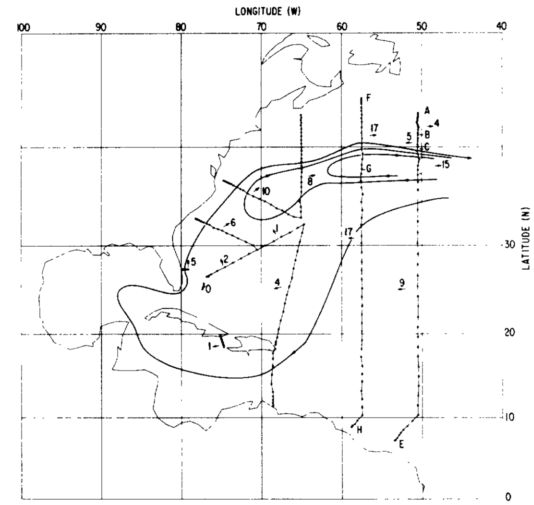
None of Wunsch's maps indicates the full C-shaped pattern, including a westward flow along the southern arm, except possibly in the range from 4 to 7°C, which indicates such flow at 10°N. This is farther south, however, than the relative flow maps of Stommel et al. (1978) and Reid (1978).

3.8 Mid-Depth Patterns in the World Ocean

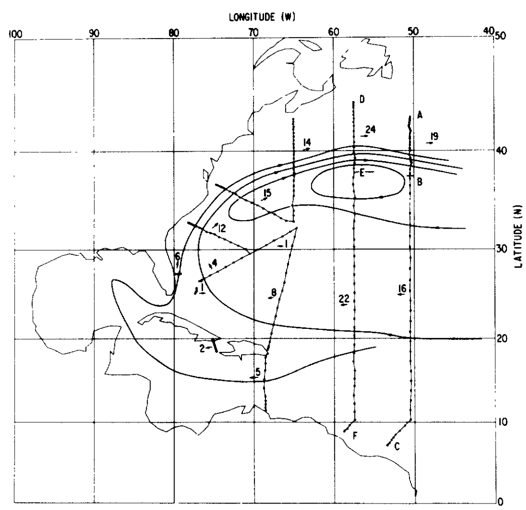
In the hope of bringing some coherence to the preceding descriptions of the various expositions, I have pre-



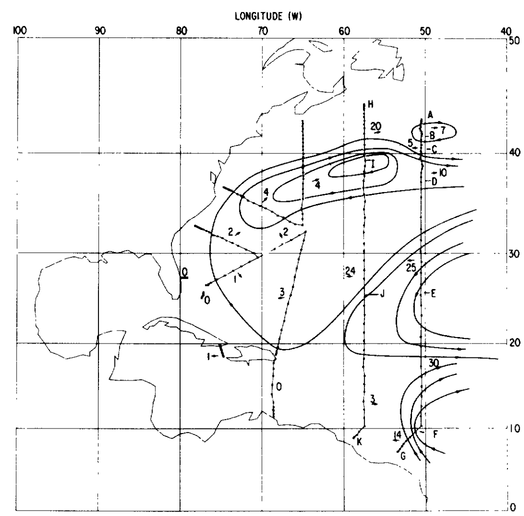
(3.15A)



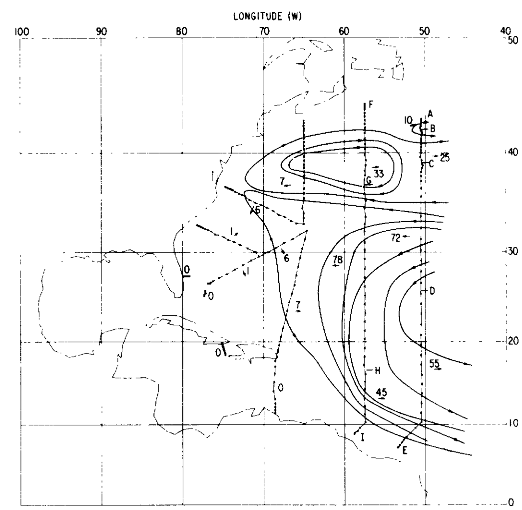
(3.15C)



(3.15B)



(3.15D)



(3.15E)

Figure 3.15 Circulation diagrams: (A) Total transport; (B) transport, 12-17°C level; (C) transport, 7-12°C level; (D) transport, 4-7°C level; (E) transport, <4°C level. (Wunsch, 1978a.)

pared a set of maps (figures 3.16A–3.16E, 3.17). They are meant to show the patterns of characteristics in the world ocean that are produced by the various sources and the processes of advection, diffusion, consumption, and regeneration.

The data set used is not necessarily the best selection possible from the present data bank, but represents the present stage in such a selection. Station positions are indicated by dots (though along some tracks the stations were so close together that I omitted alternate dots to avoid creating a solid line on the figure). The various fields are not equally well represented because the measurements vary in both quantity and quality. The selection is best in the Pacific because I am more familiar with the data there. Also, oxygen and nutrient data are more numerous there and perhaps of better quality. In the Atlantic, both oxygen and nutrients appear to show some biases among different expeditions. I have taken some liberties in adjusting the values for these maps: perhaps new data sets, or data I have not yet considered, will make substantial changes. The maps of the Indian Ocean show a denser station coverage in mid-ocean than the others. This may be because I have begun the selection there only recently and have not yet had time to identify those of highest quality.

For substantial areas of the ocean, these maps must be recognized as preliminary, and some alterations of the patterns will be made, perhaps from further consideration of these data alone, and certainly as other data are made available. Indeed, the mapping may be premature, but we must start somewhere.

3.8.1 Distributions on an Isopycnal Surface

The maps prepared consist of various characteristics along a chosen density parameter and of steric height 2000–3500 db. The surface represented here was chosen, almost at random, from the range of density that lies near 2000 m in low latitudes and extends throughout most of the ocean, intersecting the sea surface only in very high latitudes. It was not meant to represent any particular source, but merely the general patterns and processes that obtain throughout the world ocean. The density parameter (henceforward called isopycnal) is defined by a value of 37.0 in σ_2 at depths below 1500 m. Where it rises to the north in the Atlantic, it is defined in the manner of Reid and Lynn (1971) by a value of 32.47 in σ_1 between 1500 and 500 m and by a value of 27.845 in σ_0 where it lies above 500 m. Toward the Antarctic the values are 32.44 in σ_1 and 27.76 in σ_0: these differ from those in the North Atlantic because of the lower temperature and salinity. The surface does not extend as shallow as 1500 m in the North Indian or North Pacific Oceans, and the original parameter of 37.0 in σ_2 applies there.

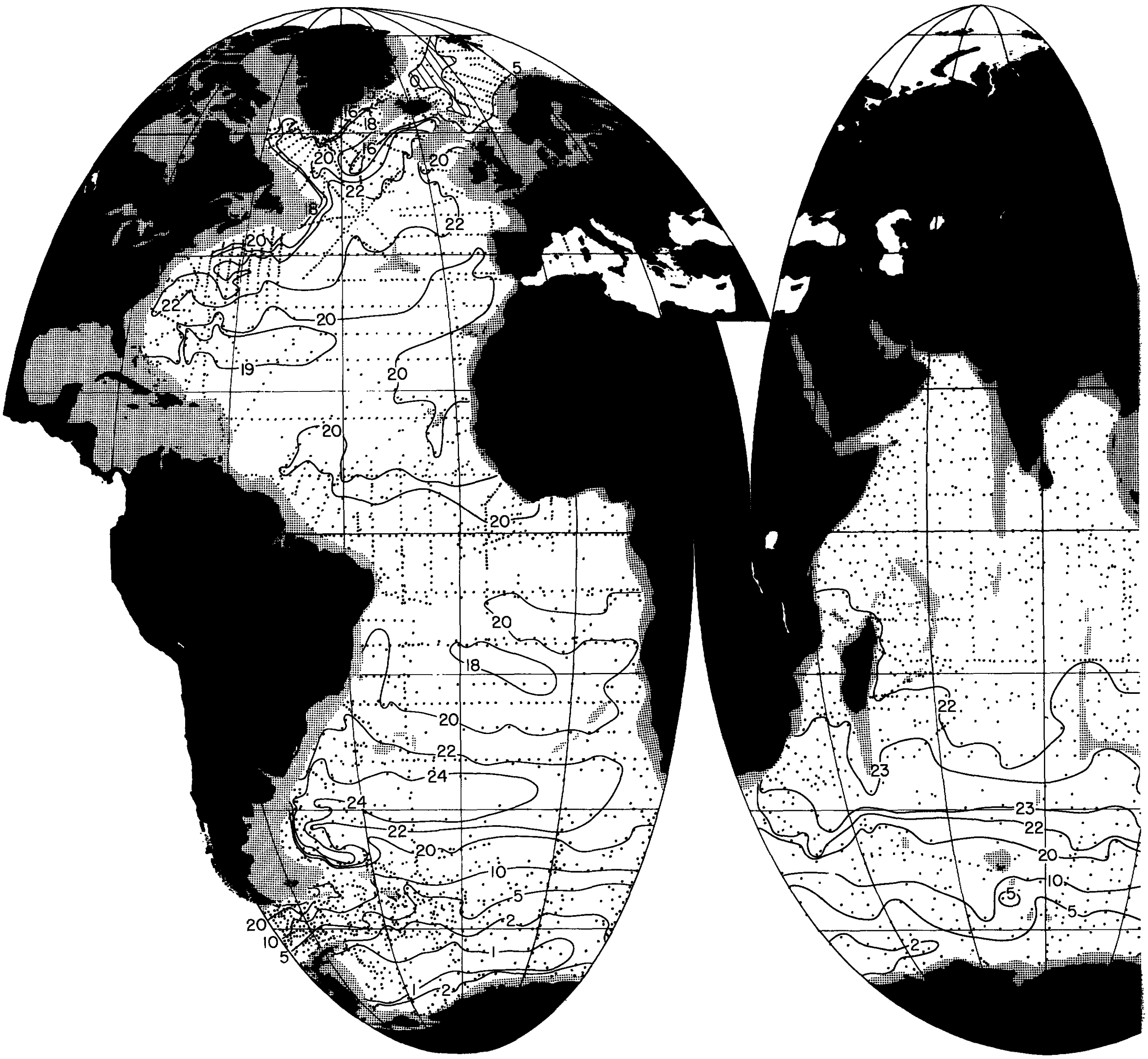
Although the isopycnal was not chosen to correspond to any particular layer, its position does correspond to, or lie near, some layers that have been named or discussed. In the northwestern Atlantic, it lies within the lower part of the Labrador Sea Water. Farther south, it lies near the layers that Wüst (1935) had called the "Upper and Middle North Atlantic Deep Water." It lies beneath the salinity minimum of the Subantarctic Intermediate Water and in the Antarctic it is included within the Circumpolar Water. In the Indian and Pacific Oceans, it lies well beneath the Intermediate Water but above the Common Water described by Montgomery (1958).

Depth The depth of this isopycnal (figure 3.16A) is greatest in the Pacific and least in the Atlantic. Values near the equator are above 1900 m in the Atlantic, 2250 m in the Indian, and 2600 m in the Pacific: the Pacific is the least dense of the oceans. In the Atlantic, the isopycnal lies deepest in the northwest and near 30 to 40°S, consonant with a poleward shift of the deep anticyclonic circulation. In the North Atlantic, it extends across the Faroe-Iceland Ridge and through the Faroe Channel and the Denmark Strait, into the Norwegian-Greenland Sea, where it outcrops. In the south, it lies at depths less than 100 m in the central Weddell Sea gyre and probably outcrops there in winter. It lies shallow all around Antarctica and probably outcrops at various places on the shelf.

In the Indian Ocean it lies deepest between 30 and 40°S in a long trough that extends eastward south of Australia and into the Tasman Sea. In the Pacific, which has a better set of north-south expedition tracks than the other oceans, the shape can be defined better, and it shows a series of zonal troughs and ridges reflecting the vertical shear patterns that will be seen on the map of steric height.

Salinity The highest salinity values (figure 3.16B) are seen in the central North Atlantic, directly beneath the vertical salinity maximum originating from the Mediterranean outflow. Another high is seen just south of Iceland, stemming from the flow from the Norwegian Sea over the Iceland-Scotland Ridge (Worthington and Wright, 1970; Ivers, 1975). Low values are found within the Labrador Sea, where the overturn and vertical diffusion provide a low salinity whose lateral extension separates vertically the Mediterranean and Norwegian sources, as seen in the atlases of Fuglister (1960) and Worthington and Wright (1970). The Denmark Strait outflow does not provide an extreme in salinity at this density. In the North Atlantic, the depth of this isopycnal lies near the depth of Wüst's Middle North Atlantic Deep Water, which he defined by an intermediate oxygen maximum, and the salinity pattern is very similar. In the South Atlantic, the isopyc-

Figure 3.16A (Facing pages) Depth (hm) of the isopycnal on which the characteristics are shown on figures 3.16B–3.16E. In the shaded areas, all the water is less dense than the isopycnal chosen. The isopycnal outcrops around the hatched area in the Norwegian-Greenland Sea, and in the following figures, the sea-surface values are contoured there but the area is not outlined.



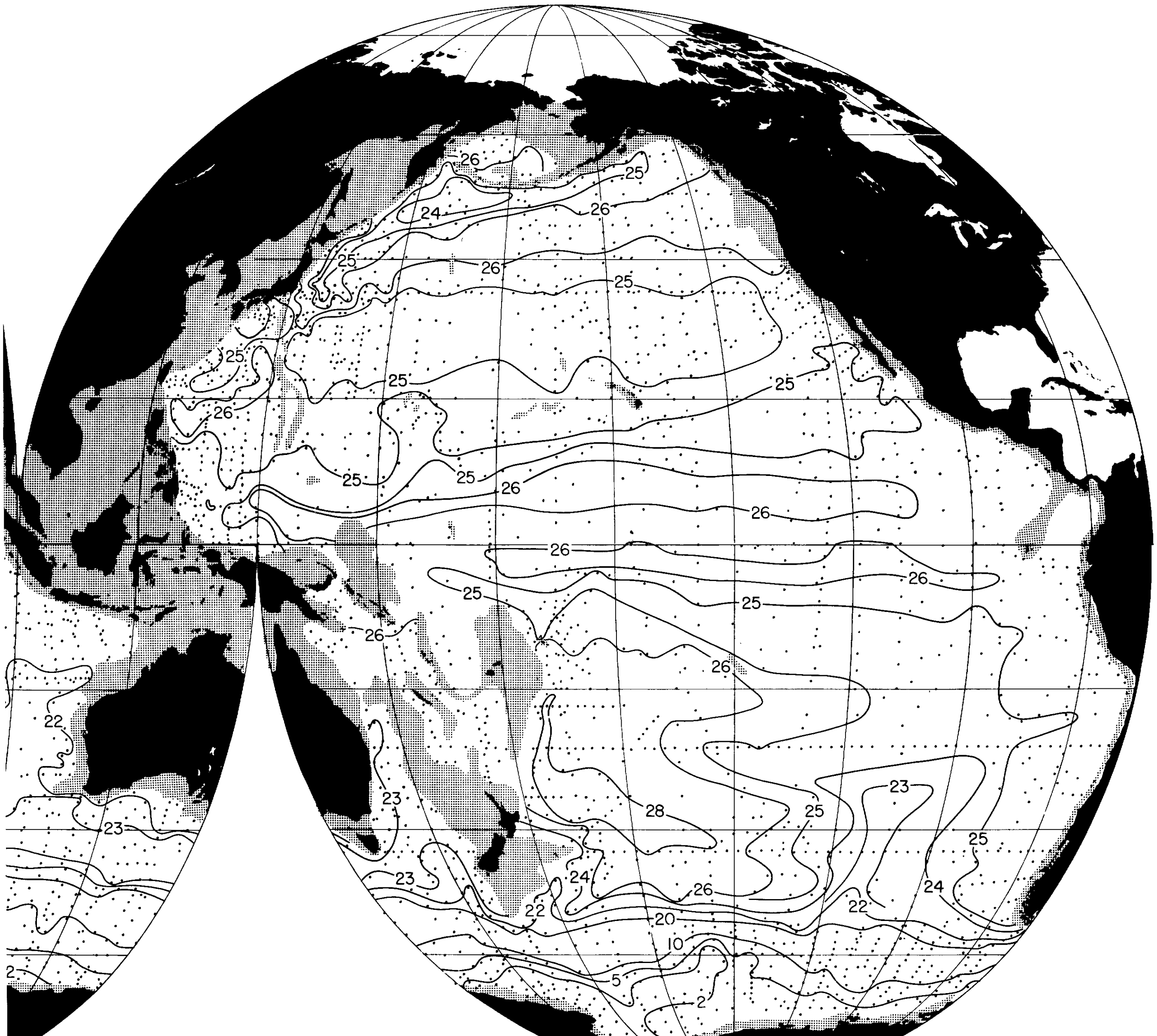

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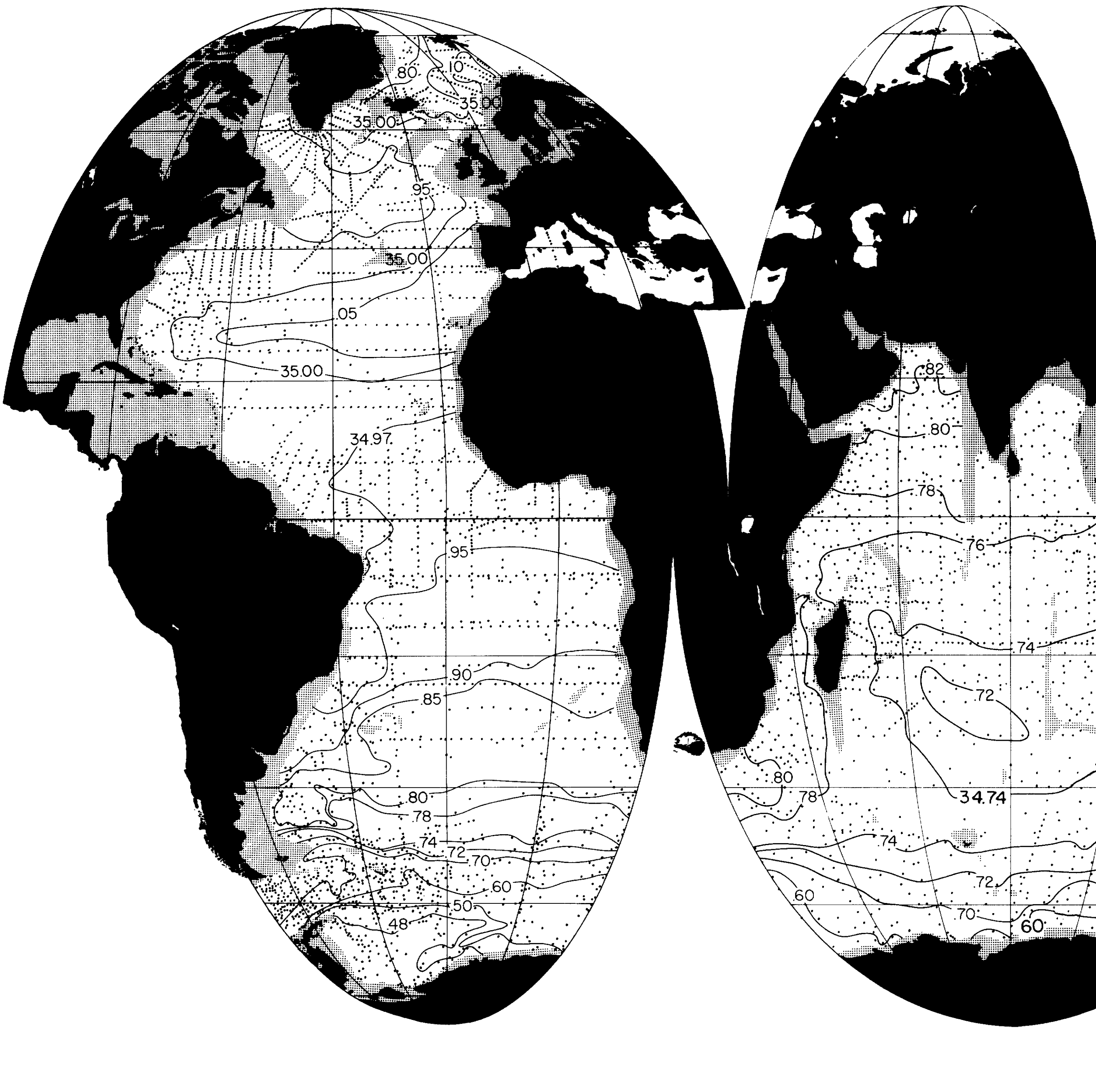
Figure 3.16B (Facing pages) Salinity (‰) on the isopycnal.



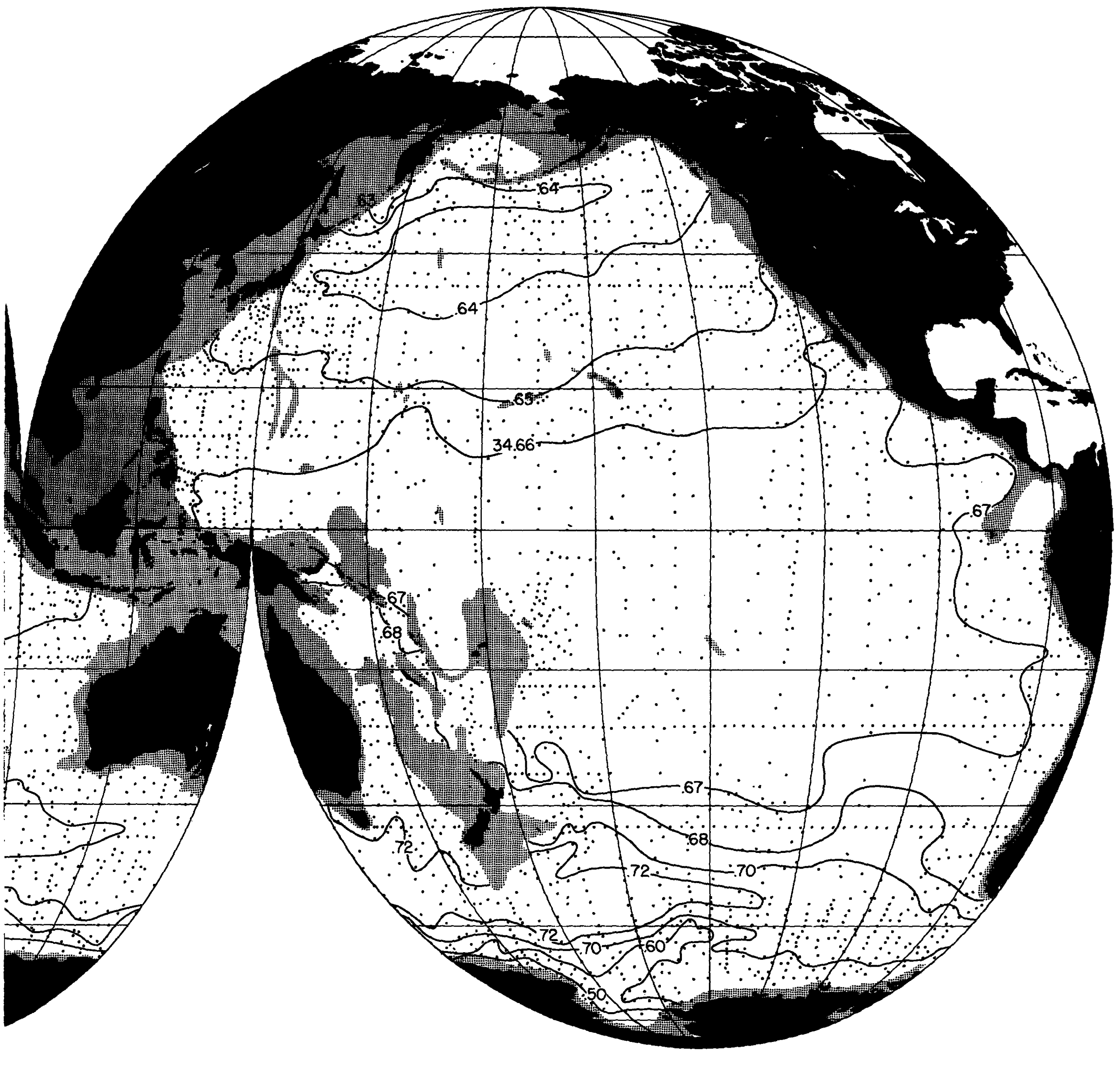

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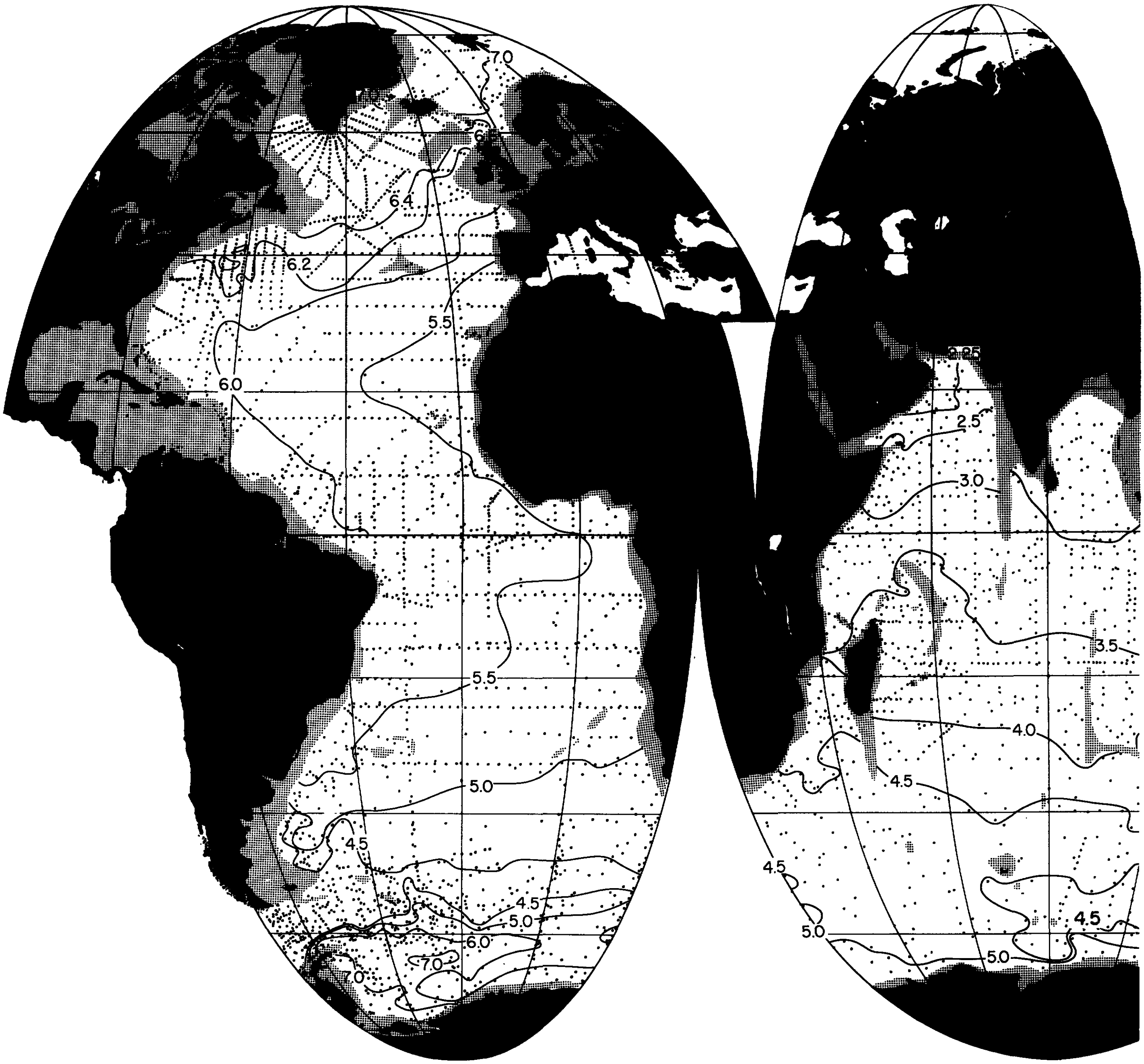
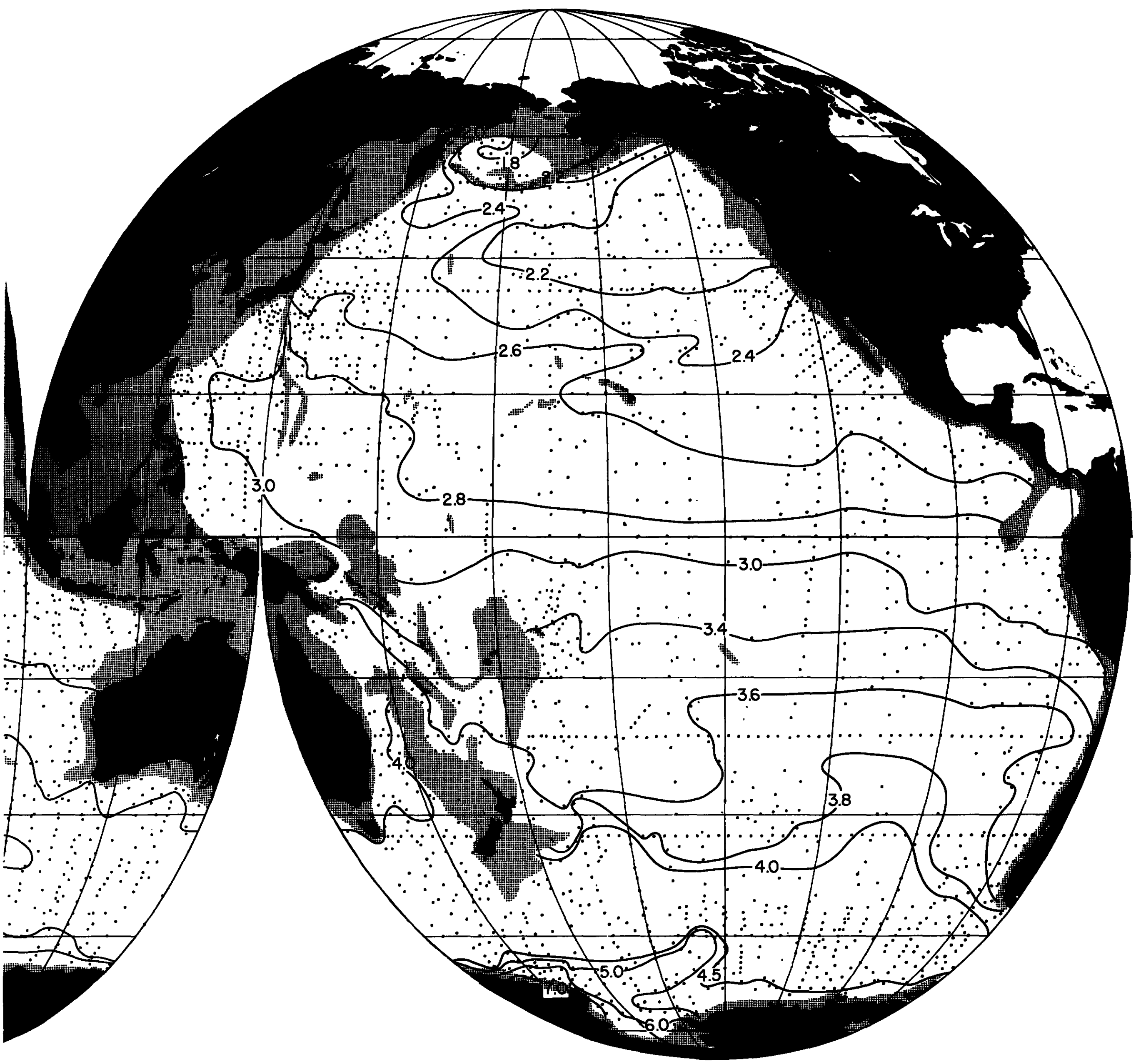
Figure 3.16C (Facing pages) Dissolved oxygen ($ml\,l^{-1}$) on the isopycnal.

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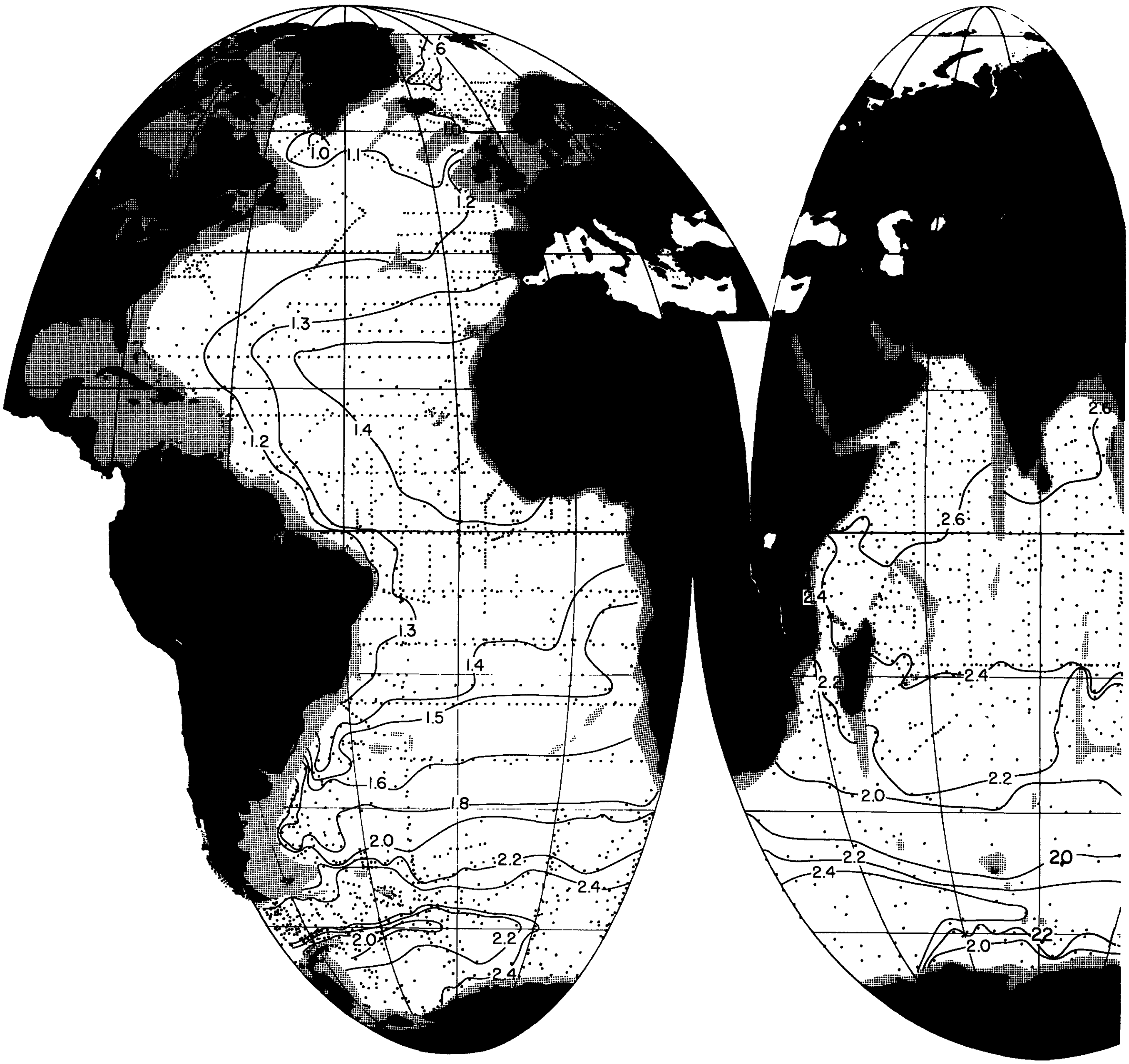
5.0
4.5
6.0

Figure 3.16D (Facing pages) Phosphate ($\mu M\, l^{-1}$) on the isopycnal.



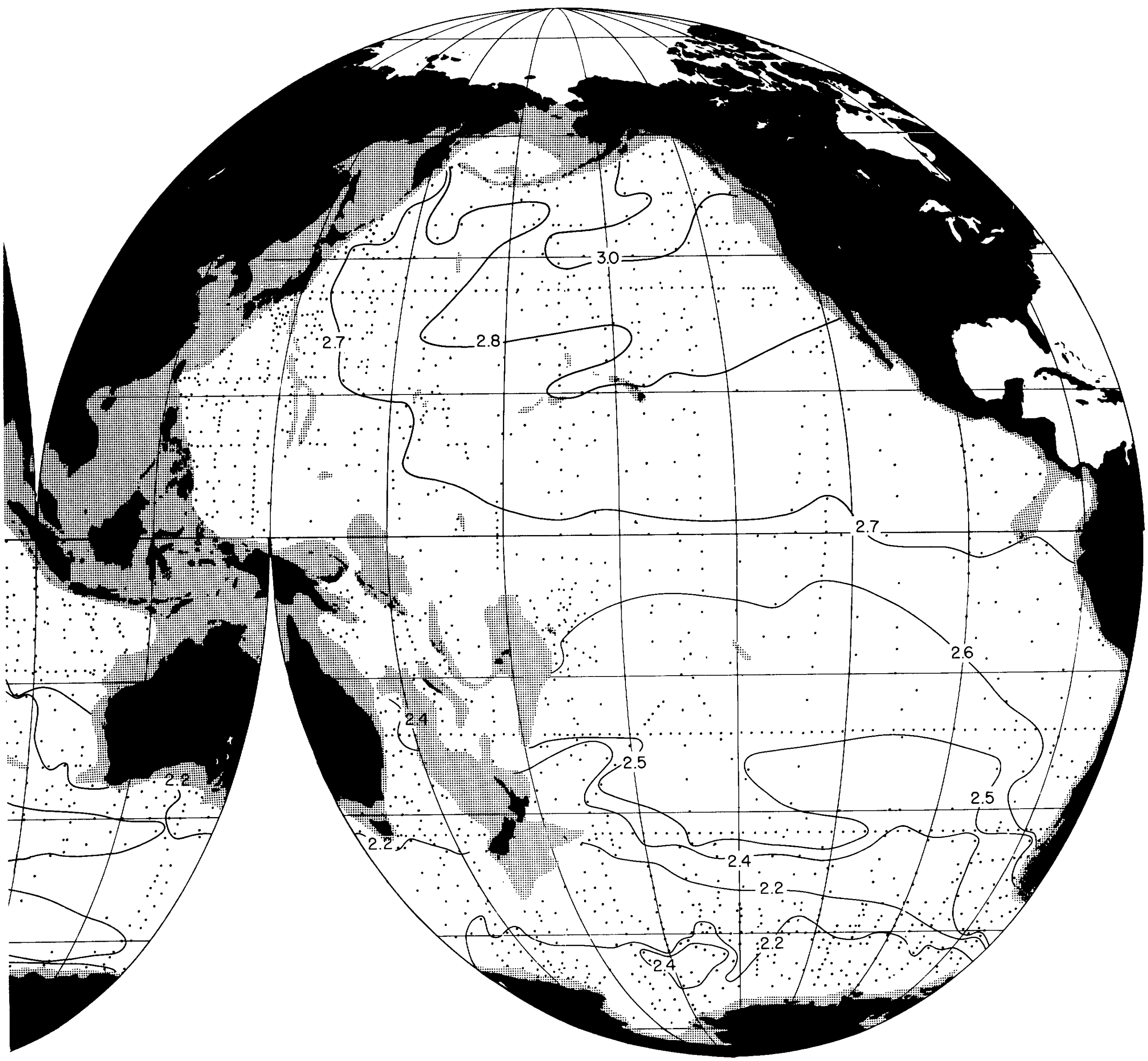

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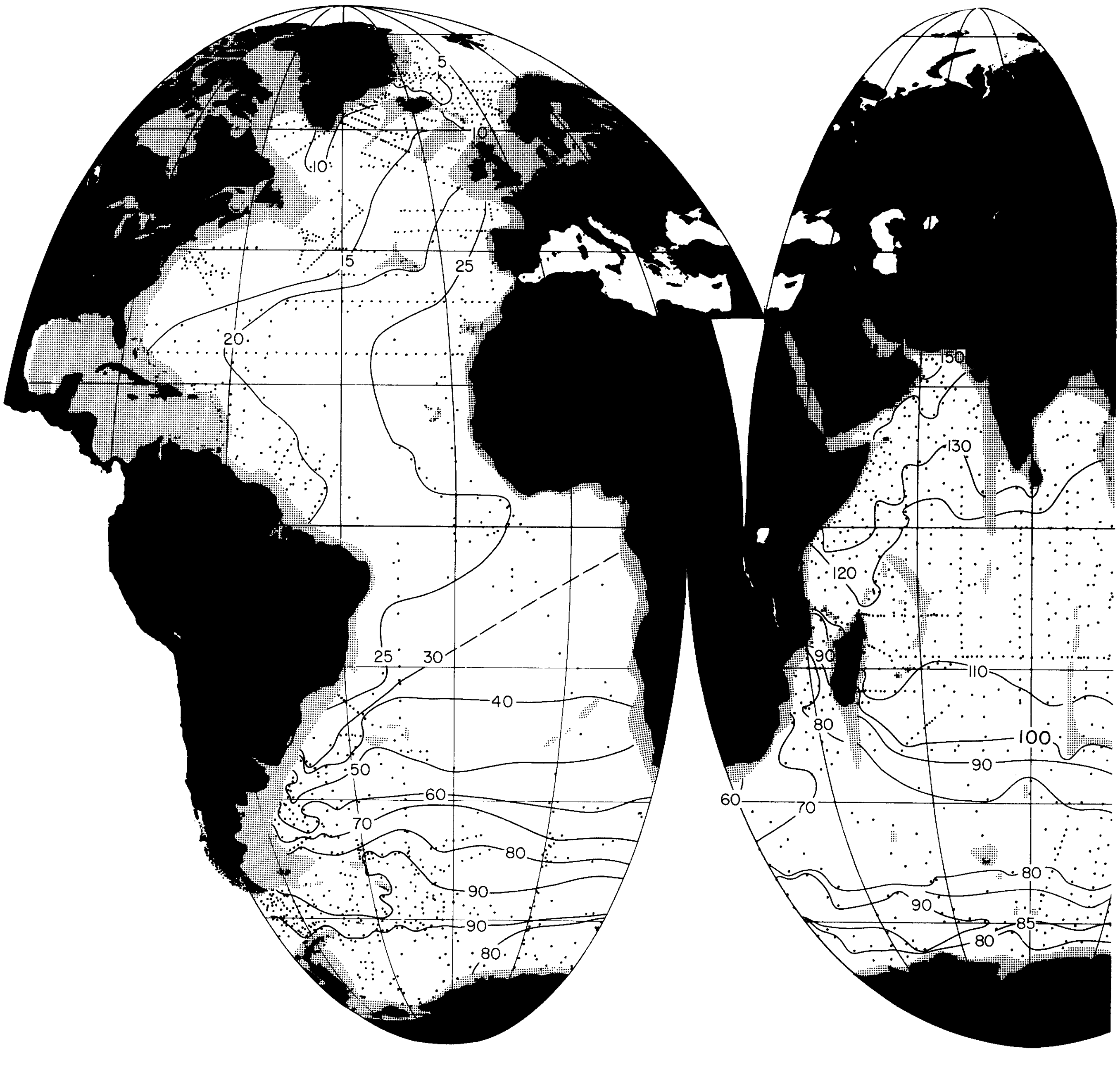
Figure 3.16E (Facing pages) Silica ($\mu M l^{-1}$) on the isopycnal.



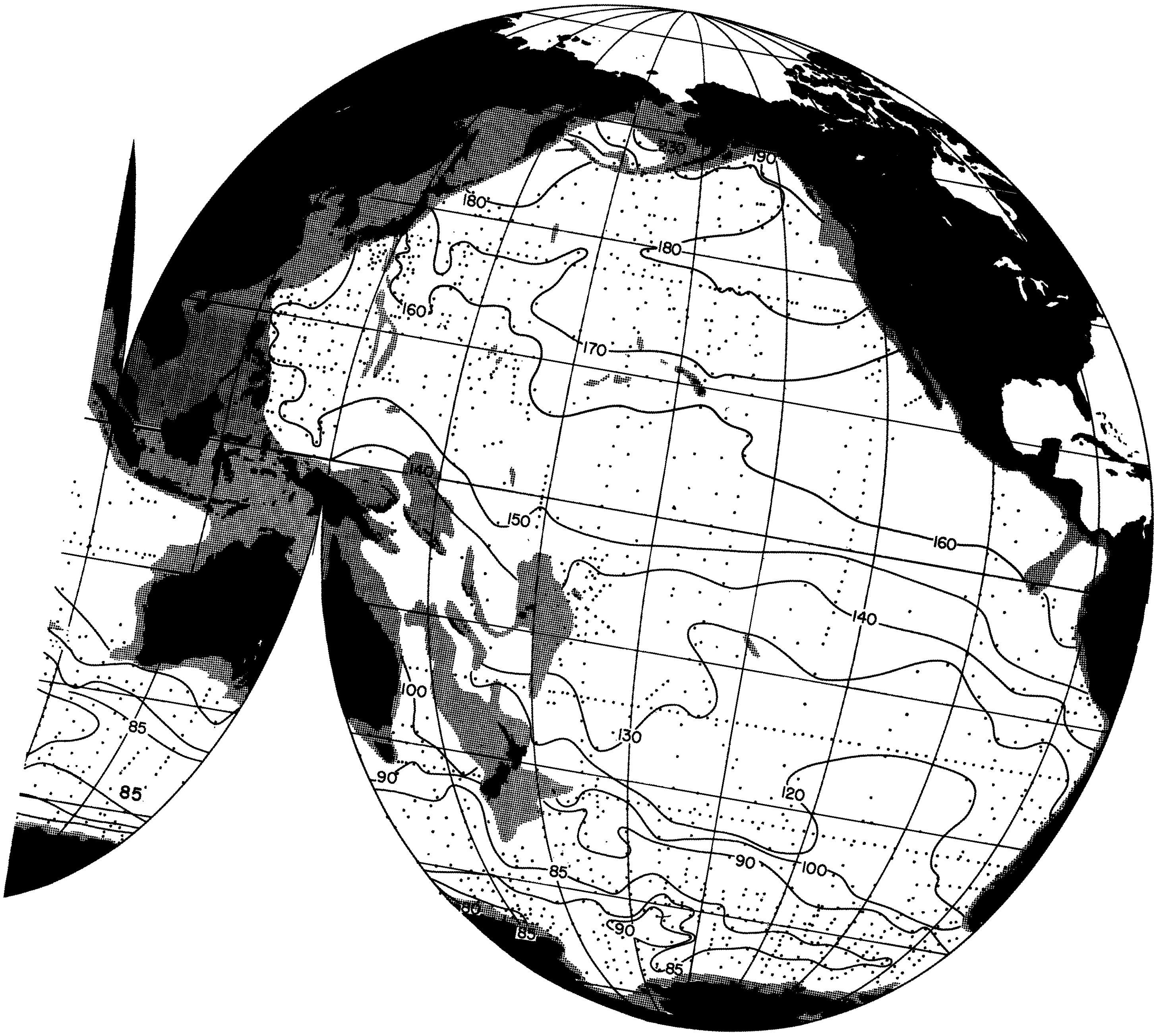


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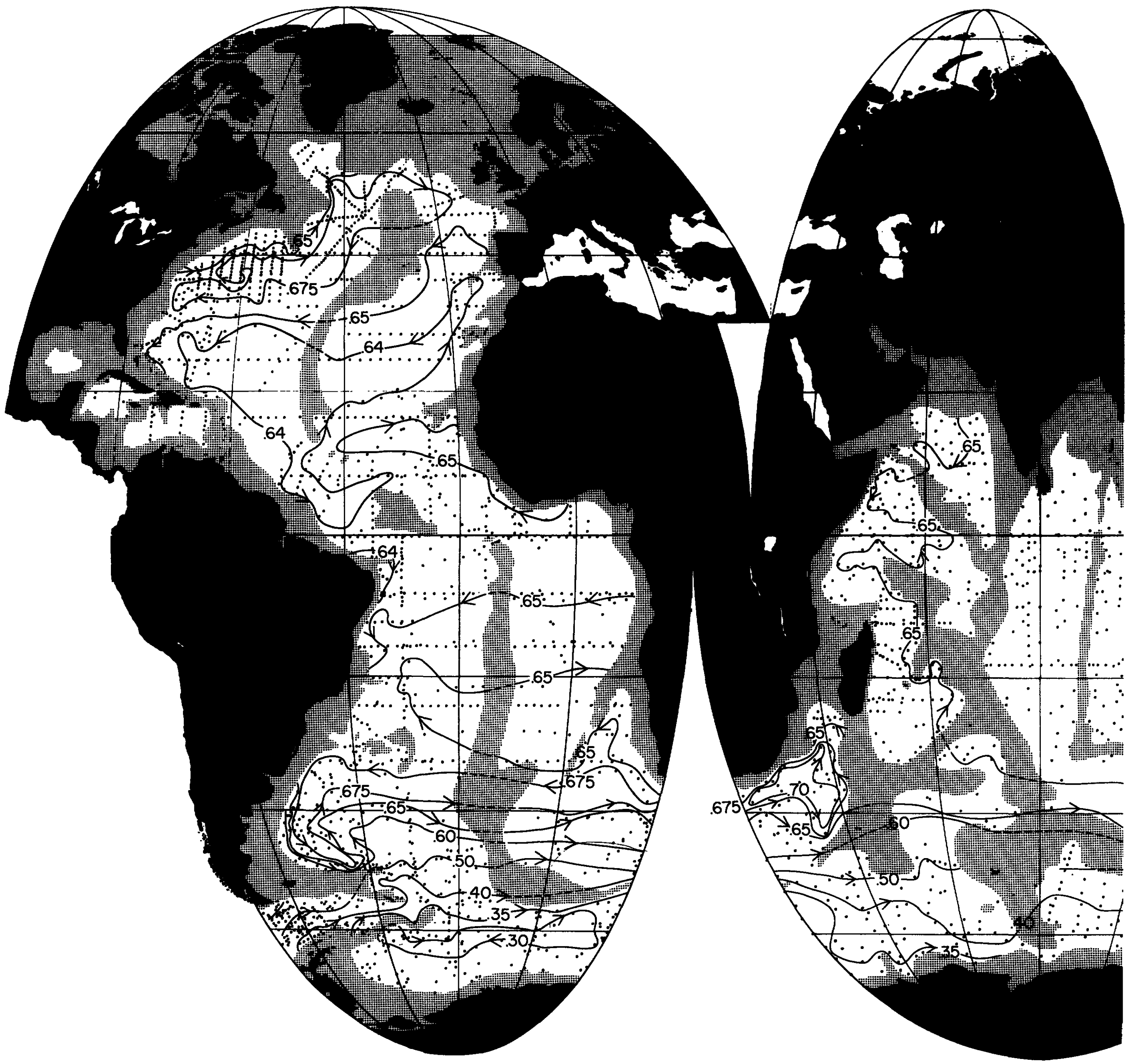
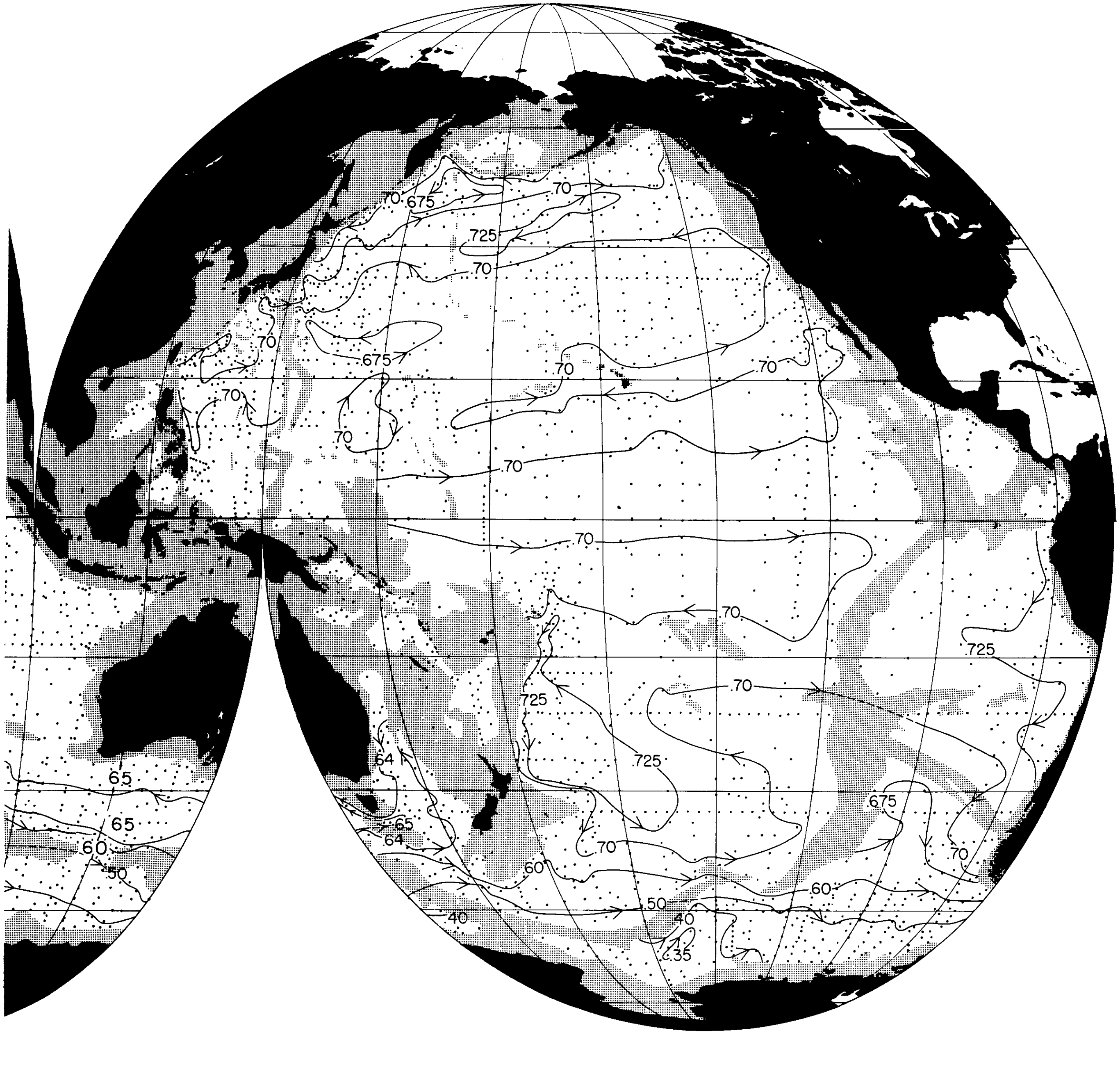
Figure 3.17 (Facing pages) Steric height at 2000–3500 db [dynamic m ($10\,m^2\,s^{-2}$ or $10\,J\,kg^{-1}$)]. Shaded area is less than 3500 m depth.

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nal lies below but near the depth of his Upper North Atlantic Deep Water, defined by an intermediate salinity maximum, and the salinity pattern here is similar. The three maps of salinity [Wüst's (1935) plates XIII and XIX and figure 3.16B herein] show roughly similar patterns because three of the principal sources of salinity extrema—the Mediterranean outflow, the Labrador Sea, and the Antarctic—extend through a thick depth range. The core method, however, which follows extrema, did not show the effect of the Iceland-Scotland overflow seen in figure 3.16B. The vertical extremum in salinity associated with this overflow lies deeper and is not connected with that of the Mediterranean outflow. The vertical oxygen maximum rises to the surface in the north, and Wüst ascribed the high values east of the Reykjanes Ridge to overturn and vertical mixing south of Iceland. This figure (3.16B) and that of oxygen (3.16C) suggest an outflow from the Norwegian Sea rather than overturn and mixing alone.

The saline mid-depth waters of the North Atlantic extend as a subsurface salinity maximum through most of the world ocean (Wüst, 1951; Stommel and Arons, 1960b; Reid and Lynn, 1971). This maximum lies beneath the less saline Intermediate Water and above the less saline but denser Antarctic waters wherever they obtain; in the North Pacific it provides the bottom layer. In the North Indian Ocean, which lacks the Intermediate Water salinity minimum, the North Atlantic waters lie beneath the more saline waters derived from the Red Sea and the Persian Gulf, which provide a maximum in salinity at lower density and shallower depth.

The southward extension of the North Atlantic high salinity is seen along the western boundary, turning eastward near 40–50°S and extending as a tongue of high salinity between the less saline Antarctic waters to the south and the less saline waters, influenced by the overlying Antarctic Intermediate Water, to the north. The tongue extends through the Indian and Pacific Oceans into the Drake Passage. As it emerges from the Pacific, it flows between the more saline North Atlantic Water and the less saline Weddell Sea Water and is no longer a lateral maximum. Some part of it turns back with the Weddell Sea gyre, into the southern Weddell Sea.

In the Indian Ocean, the area of high salinity is seen clearly in the Arabian Sea, resulting from vertical diffusion from the very saline outflows from the Red Sea and Persian Gulf, that affect this deep layer in the same way as does the Mediterranean outflow in the Atlantic. The vertical salinity maxima of these sources, reflecting the depth of their direct input, are much shallower, well above 1000 m; yet there is a clear contribution of high salinity to much greater depths.

The source of the low salinity to the north of the eastward-extending maximum near 40 to 60°S is also demonstrated by the isolated lateral salinity minimum in the central South Indian Ocean: there is no source possible for this minimum other than vertical diffusion from the overlying low-salinity Intermediate Water. This process creates the low salinity of the central Indian Ocean that separates the Atlantic high salinity at the tip of Africa from the Arabian Sea high.

In the South Pacific, there is a northward excursion of higher salinity across 40°S in the east. There is also a suggestion of some water of higher salinity flowing westward south of 60°S as part of the Ross Sea gyre.

Over most of the Pacific, the salinity on this isopycnal decreases to the north. As in the Indian Ocean, there is no other source for this freshening than vertical exchange with the overlying low-salinity Intermediate Water. The very lowest values are found near the Kuril Islands.

Oxygen The highest values of dissolved oxygen concentration (figure 3.16C) are found at the outcrop in the Norwegian-Greenland Sea. Values only a little lower are found in the Weddell Sea gyre and probably along the Antarctic coast in winter. The major source of oxygen to this isopycnal in the Atlantic is in the north, where vertical diffusion in the Labrador Sea and overflow from the waters north of the Greenland-Scotland Ridge provide concentrations of more than 6.4 $ml\,l^{-1}$. High values from this area extend southward along the western boundary, much as Wüst (1935) showed for the core of the Middle North Atlantic Deep Water, which he defined by an intermediate-depth oxygen maximum, at depths slightly greater than this isopycnal. As the vertical maximum ends near 50 to 55°S, he was unable to follow the water further. Along this isopycnal, a tongue of high oxygen extends eastward across the Indian Ocean, as Callahan (1972) showed for a slightly deeper isopycnal. Farther east, its value is intermediate between the high values around Antarctica and the lower values of the Pacific to the north, and its final features in the south are the excursion into the Southeast Pacific Basin and around the Ross Sea gyre.

During the passage around Antarctica, the concentration of oxygen has been reduced so much, in spite of the Antarctic source, that in its reentry into the Atlantic through the Drake Passage, it appears as a lateral minimum, between the waters from the North Atlantic and the fresher waters of the Weddell Sea gyre. Its decrease can be the result of respiration and decay at this density or by vertical diffusion to the overlying strong oxygen minima of the Indian and Pacific Oceans, which are generally both deeper and of lower concentration than those in the Atlantic. As Callahan (1972) has shown, the lower values of oxygen within the Indian and Pacific Oceans contribute substantially to the decrease of oxygen along the Circumpolar Current by lateral exchange. It is noteworthy that the high salinity introduced by vertical diffusion in the Arabian Sea is not accompanied by an increased oxygen there.

Phosphate and Silica The other nonconservative concentrations (figures 3.16D, 3.16E) show slightly more detail than the oxygen. They are lowest in the Labrador and Norwegian-Greenland Seas and highest in the Arabian Sea and northeastern Pacific. Phosphate differs by a factor of about 2.5 between the northwestern Atlantic and northeastern Pacific, silica by a factor of about 12. These ranges are large enough to provide patterns more detailed than, as well as different from, those of salinity.

Following the pattern of salinity and oxygen, a simple picture emerges of low nutrient values in the Labrador Sea extending southward along the western boundary to the West Wind Drift, or Circumpolar Current. A tongue of low nutrients extends eastward from the South Atlantic all across the Indian and Pacific Oceans and through the Drake Passage. Concentrations within the lateral minimum increase to the east, and as the waters enter the Atlantic, they appear as a high in nutrients between the nutrient-poor waters of the western South Atlantic and the shallower, somewhat depleted waters of the Weddell Sea gyre. That the highest values in the Indian Ocean are found in the Arabian Sea may be a consequence of the ridges; the narrow gaps may limit the lateral exchange.

A somewhat clearer pattern is seen in the North Pacific. In both phosphate and silica, the highest open-ocean values (excluding the almost enclosed Bering Sea) are in the northeast, with slightly lower values to the north. The resulting tongue of low nutrient extending northeastward from the western boundary suggests an advective feature. In phosphate, there are two of these tongues (possibly two in silica also but the data are too few to be certain), consonant with the shallower flow pattern proposed by Reid (1978).

Sources of the Characteristics The maps make fairly clear the sources of the high and low values of the characteristics that appear in various areas. The lateral sources are the Norwegian-Greenland Sea, which provides shallower waters of this density that extend downward into the open Atlantic, and the long zone around Antarctica, where this isopycnal lies shallow or outcrops. Both lateral sources provide high oxygen and low nutrients (though the lowest-nutrient waters are from the north), but the northern sources are warm and saline, the southern sources colder and fresher.

A combination of convection and vertical diffusion in the Labrador Sea extends high-oxygen and low-nutrient concentrations down to this layer. Over most of the ocean, where this layer lies at mid-depth, the characteristics appear to be modified only by vertical diffusion and by consumption of oxygen and regeneration of nutrients. In the northeastern Atlantic and northwestern Indian Oceans, the salinity is made high by vertical diffusion from the overlying layers of outflow from the Mediterranean Sea and the Red Sea and Persian Gulf. In the southeastern Indian Ocean and the North Pacific, the low-salinity values must be the consequence of vertical diffusion from the overlying Intermediate Water; in neither case is there a lateral source. Over most of the Indian and Pacific Oceans, this isopycnal lies beneath a thick oxygen minimum and a thick maximum in phosphate, and such vertical diffusion as takes place does not raise the oxygen concentration or lower that of the nutrients on this isopycnal.

3.8.2 The Steric Height at 2000 db Relative to 3500 db

It seems worthwhile to examine the density field to inquire whether the geostrophic vertical shear is consistent in any simple way with the mid-depth patterns that have been illustrated herein.

The upper surface of 2000 db was selected to correspond roughly to the depth of the chosen isopycnal over most of the ocean. The 3500-db surface was selected not because of any special assumptions of minimal flow near that pressure, but because the shear field is weak over much of the area in this depth range and it seemed useful to consider a thick layer. A substantial part of the western boundary flow takes place in areas shallower than 3500 m and cannot be represented on such a map, but we already have some information about the boundary flow there, at least in some cases.

This map (figure 3.17) has been made from a selection of stations rather than from a set of averages. It is, of course, not synoptic, even to season, and though most of the data are from the later period when salinometers have been available, it has been necessary to include some earlier, less accurate measurements as well. As in the set used in figure 3.16, to which it corresponds closely, it does not represent the best possible selection, but the present stage of progress toward it.

The map has been made not only to illustrate such features as it can of the deep density field but also to indicate what is lacking in the bank of good-quality data. One striking feature is that although the contours extend mostly east-west, indicating a zonal flow that requires north-south sections of data for proper resolution, most of the major track lines (IGY, Scorpio, INDOPAC, etc.) extend east-west, except in the Antarctic zone. This has certainly compounded the difficulty in interpreting the density field and limited my own confidence in many of the features I have drawn. We must, however, start somewhere. Though the earlier interest in abyssal meridional flow has dictated the east-west station arrays, there seems to be enough zonal pattern in the density field to merit investigation. A few more north-south lines would certainly help.

The Range of Steric Height The range of values of steric height is very small. A difference of about 30-35 dynamic cm (hereafter cm) is seen across the Antarctic Circumpolar Current, but north of 40°S the range is about 7 cm in the Atlantic Ocean, 6 cm in the Indian, and 7 cm in the Pacific. The western boundary currents and the Antarctic Current have the strongest gradients, and variability will be strongest there, but in much of the ocean the data indicate that in a 10° × 10° area, for example, the combination of time and space variability must be very small—little more than 3 cm, and much less than Wyrtki (1975a) has shown in the upper 500 m alone.

The surface of the Pacific Ocean stands on the average about 40 cm higher than the Atlantic with respect to the 1000-db surface, and the North Atlantic and North Pacific stand, respectively, about 14 and 17 cm higher than the South Atlantic and South Pacific; referred to 4000 db, the Pacific stands about 68 cm higher than the Atlantic (Reid, 1961b; Lisitzin, 1974). The upper-level differences are reflected in figure 3.11A (from Burkov et al., 1973), relative to 1500 db, and the Indian Ocean seems to be intermediate in steric height.

These differences are not only reduced in the 2000-3500-db layer, but in some cases reversed. The 2000-db surface is about 5 cm higher relative to 3500 db in the North Pacific than in the North Atlantic, but it stands highest in the South Pacific (excluding the areas poleward of 50°). The North and South Atlantic and the Indian Ocean values do not appear to differ very much from each other.

The Flow Interpreting the pattern of steric height as representing relative geostrophic shear, the most striking feature is the predominance of zonal relative flow, except in the eastern Indian Ocean, where no clear pattern can be seen in the present data set.

This pattern, of course, does not necessarily represent the sense and magnitude of the absolute flow, but only the geostrophic shear between two isobaric surfaces. Where a deep or abyssal boundary current is moving faster than the flow at 2000 m, the relative flow may be in the wrong sense. Fairly rapid flow may occur at 2000 m in areas where the water depth is less than 3500 m: flow in these areas will of course not appear on the map. The deep current from the northern North Atlantic that flows cyclonically around the Labrador Sea and southwestward along the continental slope of North America does not appear on this map. This is because most of the 2000-m flow of that current takes place in water depths less than 3500 m (Worthington, 1970; Swallow and Worthington, 1961; Luyten, 1977; Richardson, 1977). The map begins, instead, with the deep Gulf Stream, which it portrays clearly, and the broader return current (Defant, 1941b; Worthington, 1976). This is quite consistent with Richardson's (1977) interpretation of the direction of the flow at 2000 m over deep water and is remarkably consistent with the float trajectories near that depth reported by Riser, Freeland, and Rossby (1978) near 20°N, 74°W, which extended westward and southward, paralleling the 0.65-m contour in that area.

It is also true, as Defant (1941a) pointed out, that along much of the western boundary of the North Atlantic the vertical shear field is such that the calculated flow will be northward relative to the deeper water. If a surface shallower than 3500 db had been chosen, values could be calculated in shallower water but would not have shown a southward flow relative to the underlying water (Swallow and Worthington, 1969; Ivers, 1975).

Along the western boundary of the South Pacific (just east of the Tonga-Kermadec Ridge, near long. 180°), a deep northward flow of Antarctic water has been detected (Warren and Voorhis, 1970) with a southward flow above it as part of the subtropical anticyclonic gyre (Reid and Lonsdale, 1974). At about 22°S, Warren and Voorhis (1970) propose a surface of zero meridional velocity that varied from 3100 to 4300 m, and in this region the sense of flow given by the shear map appears to be correct and the relative speed might not be biased very much. Farther south, Warren (this volume, chapter 1) finds the northward flow extending up to about 2000 m near 43°S. The shear map, referred to the stronger flow at 3500 db, extends the southward flow too far south before turning eastward. The characteristics along the isopycnal, which lies near 2700 m at 43°S, suggest a northward extension of the circumpolar waters at this depth to about 30°S before turning southeastward and around the anticyclonic gyre.

A case of steep continental slope and broad western boundary currents is seen in the South Atlantic near 40°S (Reid et al., 1977), where the strong shear between the equatorward abyssal flow and the poleward mid-depth flow is reflected in figure 3.17. Warren's studies of the Indian Ocean were directed toward the deep western boundary currents, but he also provides information about the choice of references and the overlying flow in most of the areas he considered. With measurements along 12°S and 23°S in the western Indian Ocean, Warren (1974) found a deep (3000-3500 m) northward western boundary current flowing close against Madagascar. He found no boundary current near 2000 m but proposed the weak northward flow between 2000 and 3000 m that appears in figure 3.17.

Later, Warren (1977), with the section along 18°S in the eastern Indian Ocean, showed evidence for a deep northward flow just east of the Ninetyeast Ridge, and indicated that the deeper flow is somewhat stronger than that at 2000 m. The shear field is weak and the other data are confusing, so it has not been possible to provide contours there. Across the Southwest Indian Ridge (near 30°S, 60°E), he tentatively set 3500 m as a zero reference level and calculated northward flow below it (Warren, 1978). This reference surface also provides northward flow at 2000 m.

3.9 Comparison of the Maps of Shear Field and Characteristics

Over those parts of the ocean where direct measurements or other sorts of information are available, the shear map (figure 3.17) is at least not in severe disagreement as to the direction of flow. Over the rest of the area, particularly the central ocean, we must simply recognize that this map can at best represent only the vertical geostrophic shear, and inquire how it relates to the patterns of figure 3.16.

3.9.1 Flow across the Equator

The transequatorial flows will not be seen in the geostrophic shear field, but the characteristics suggest a southward extension across the equator in the western Atlantic and a northward extension in the western Pacific. A corresponding case might be made for northward extension in the eastern Atlantic and southward extension in the eastern Pacific, but this is not so clear. In the Indian Ocean, the characteristics suggest a southward flow across the equator in the west.

3.9.2 Atlantic Ocean

In the North Atlantic, the shear field shows westward flow between 30 and 50°N in the eastern area, and eastward flow near 15–25°N. This is consonant with the pattern of high salinity extending westward across the North Atlantic. The lower-salinity waters from the north may simply flow southwestward to the western boundary near 25°N, where they turn southeastward, surrounding the area of high salinity. These low-salinity waters may also move southwestward along the continental slope from Greenland to Cape Hatteras, in waters less than 3500 m deep and hence not represented on the shear map. The Gulf Stream at this depth appears here as simply part of the anticyclonic flow, with its return circulation perhaps carrying southward a major part of the lower-salinity northern waters.

This scheme is also consistent with the distribution of oxygen and nutrients. Their patterns differ from that of salinity only in that the source of the high salinity is clearly the Mediterranean outflow, but the low oxygen and high nutrients have sources at a lower latitude, in the equatorial and tropical eastern Atlantic as well as from the Mediterranean, and their westward extensions cover a broader zone. They are all consistent with a westward flow north of 25°N in mid-ocean, as the shear map suggests, and of course with the southwestward recirculation of the Gulf Stream at this depth.

In the South Atlantic, the shear map suggests some segments of a western boundary current, and all of the characteristics extend extrema southward along the boundary. Just north of the Falkland Plateau (50°S, 50°W), where the abyssal flow is strongly northwestward, the relative flow (figure 3.17) may appear much stronger than the actual flow. At the confluence of the western boundary current on this map with the Circumpolar Current, they both turn back northwestward and then eastward at about 40°S. All of the characteristics suggest such a bight, with Antarctic waters turning sharply northwestward near 40°W.

The zonal flow patterns between the equator and 40°S in the eastern Atlantic in figure 3.17 are very poorly defined by the available data, which are all from east-west tracks. Likewise, the characteristics are ill-defined there, and no useful comparison can be made.

3.9.3 Indian Ocean

In the Indian Ocean, the shear field as interpreted here is too weak north of 30°S for a useful comparison. Some suggestion of a northward western boundary current is seen in the anticyclonic feature at the tip of Africa, and perhaps of a northward flow between Madagascar and the Mascarene Ridge. The oxygen and nutrients partly support this, but their dominant feature is a southward extension along the western boundary from the Arabian Sea toward Madagascar. The major extension, however, of the Atlantic characteristics seen entering at the tip of Africa is clearly eastward, and this is consonant with the shear field.

3.9.4 Pacific Ocean

It is in the Pacific that data are most nearly satisfactory for this study. They allow a much better resolution of the patterns than in any other area than the Circumpolar Current. The shear field in the South Pacific shows an anticyclonic gyre near 35–40°S, an eastward flow leaving this gyre along 25°S in the east, a westward flow along 15°S, and an eastward flow along the equator: this last feature is weak and must remain uncertain. This is the complete C-shape, roughly circumscribed by the 0.70-dynamic-m contour. There is also a southward flow along the eastern boundary.

All of these features are matched in the North Pacific—an anticyclonic gyre near 40–50°N, an eastward flow near 20–30°N, a westward flow from 20°N in the east to 15°N in the west, and the weak eastward flow near the equator. The 0.70-dynamic-m contour that circumscribes the C-shape in the South Pacific circumscribes only the two arms of the C; it is broken apart in the west, perhaps by the bottom topography. Whether the Philippine Sea is a major part of this pattern or is largely cut off at this depth, as are the Caribbean Sea and Gulf of Mexico from the open Atlantic, is uncertain.

There is a suggestion of a northward extension of characteristics near the western boundary of the North Pacific. A tongue of high salinity extends from Japan northeastward to mid-ocean, and this is even clearer in some of the nonconservative characteristics. The extension is consonant with the eastward flow of the subarctic cyclonic gyre and the adjacent anticyclonic gyre. The nonconservative characteristics are also consonant with a return flow southwestward across 40°N in mid-ocean. Oxygen and phosphate also indicate an eastward flow at about 20-25°N in mid-ocean, corresponding to the shear field. For silica, the pattern is less clear there.

The correspondence between the shear field and the characteristics is clearest in the South Pacific. From the Circumpolar Current, tongues of high salinity and oxygen, and low-nutrient concentration, extend northward around the eastern part of the anticyclonic gyre of the shear field. From the equatorial zone, low-oxygen and high-nutrient concentrations extend southward around the western part of the gyre. The shear field suggests that the eastward flow from mid-ocean at about 20-30°S carries the water from both these sources eastward toward South America. From there, they are carried southward to the Circumpolar Current and through the Drake Passage. Over this area the contours of the nonconservative characteristics and of steric height are nearly parallel.

It appears to be along the route defined by the shear field in the South Pacific that the extreme characteristics created in the North Pacific—the thick, mid-depth oxygen-minimum and nutrient-maximum layers described by Reid (1973a)—leave the Pacific and enter the circumpolar system.

3.9.5 Antarctic Ocean

In the Antarctic Circumpolar Current, all of the fields are fairly well defined and in simple consonance. The shear is eastward, with excursions only into the South Atlantic around the Falkland Plateau (50°S, 40-50°W), in the Weddell and Ross sea gyres, and the Southeast Pacific (Bellingshausen) Basin. It carries the saline, high-oxygen, low-nutrient water from the western Atlantic all around Antarctica and back into the Atlantic. Salinity is still a lateral maximum as it emerges from the Drake Passage, but the nonconservative characteristics have changed from one extreme to the other during their long subsurface passage.

3.9.6 Westward Flow in the Deep Anticyclonic Gyres

There is one other noteworthy feature of the poleward shift of these gyres. They may shift, at some depths, south of Africa and Australia. The westward return flow of the West Wind Drift currents can be seen clearly in all oceans, and it seems to be returning from all across the ocean. Indeed, part of the return flow (if that is a correct term) in the South Indian Ocean may originate from the Tasman Sea, and some part of the Agulhas Current system may extend across the Atlantic. South of Australia, the characteristics rather suggest that such flow may occur at depths near this isopycnal. South of Africa, only the oxygen and phosphate might suggest such a flow. It would be interesting to explore this possibility at greater depths.

3.10 Conclusion

Mid-depth circulation has received much less attention than the upper wind-driven layer (of uncertain or at least disputed thickness) and the abyssal, presumed thermohaline flow. Some evidence for the direction of these flows was available from ship's drift at the surface and such quantities as potential temperature patterns at the bottom, but little evidence of the details of any mid-depth patterns was available to the early investigators other than the long meridional extensions of salinity and oxygen extrema. Theoretical, numerical, and descriptive studies have dealt mostly with surface or abyssal flow, or with total transport.

That there might be flow patterns at mid-depth that are significantly different from these has been considered possible, at least since the work of Prestwich (1875), but the means of working on these were few. The major attempts to date have been made through the descriptive studies.

I have cited numerous studies here that have involved mid-depth circulation through the examination of the distribution of conservative and nonconservative characteristics along surfaces of constant depth or along extrema in the characteristics, or along some sort of density surface, or by consideration of relative geostrophic flow, or through some combination of these. (The list is of course incomplete, partly because of space, but also because I am not as familiar as I could wish to be with the work of many of the investigators, especially those in Japan and the Soviet Union: I am sure that I have left out some important studies.)

I have made some remarks also about the trends—the abrupt change in emphasis from flow patterns at depth to models of total transport that occurred in the early 1950s, and the more recent trend back toward models of nonhomogeneous oceans with irregular bottoms. This has brought the density field back into the foreground; with all its limitations, it still seems to be one of the strong ocean signals and one that can no longer be neglected.

I cannot claim to have established the correct mid-depth flow pattern from the materials cited or from the new maps presented here. I have, however, made a preliminary case for a general pattern of mid-depth circulation. It involves substantial zonal flow in mid-latitudes, and is perhaps more analogous to the overlying wind-driven gyres than to the recognized thermohaline deep and abyssal flow. The poleward shift in the density field cannot be questioned: it is a real feature of all the oceans and is seen clearly on every density field prepared from data. The equatorward arm of the C-shape appears in some of these maps, and in some models, both prognostic and diagnostic.

The flow field considered was calculated, quite arbitrarily, to the 3500-db surface. It is strong in the higher latitudes, but within about 20° of the equator, the field is weak and the shear uncertain. This is especially so in the South Atlantic, where data are lacking, and in the Indian Ocean, where the present selection of data gives no resolution. Taking the sense of flow from the shear map does not make for any obvious difficulties, except for the narrow zone near New Zealand that has been discussed. This does not mean that it is therefore correct everywhere else, but it shows, I believe, that such considerations of the shear field may serve as a useful way to start work on the problem. Over the greater part of the ocean, where no direct information is available, the only tests we can apply are to compare it with the distribution of the characteristics and with the various models.

The shear field, treated as flow, seems to be supported fairly well by the patterns of characteristics. This was to be expected in the Circumpolar Current, which is deep, broad, and in high latitudes, where the shear signal is well defined. It might not have been expected that the patterns of characteristics and the shear field would match (at least qualitatively) to the degree that may be seen in the middle latitudes of the North Atlantic and the Pacific. In the South Pacific, in particular, the fields are defined clearly, and the coherence between the patterns of shear and characteristics is remarkably good.

On the deeper isopycnal chosen to represent the Lower North Atlantic Deep Water (Reid and Lynn, 1971) only the depth, salinity, and potential temperature were illustrated. It lay in greater depths (3–4 km in middle and low latitudes) than the isopycnal mapped herein, and, with only the conservative characteristics illustrated, little suggestion of zonal flow was suggested except in the Circumpolar Current. Western boundary currents at these depths were suggested in the southern Indian and Pacific Oceans, and a mid-ocean southwestward flow across 40°N in the Atlantic; this latter seems to correspond to the deep return flow of the Gulf Stream.

Including the nonconservative characteristics on the isopycnal presented herein has added considerable detail to the pattern and allowed for a broader interpretation. Some support is given to Prestwich's (1875) concept of poleward extensions of warmer waters well beneath the sea surface, even down to the depths of the isopycnal illustrated here. In the North Atlantic, there is a northward extension of highly saline water along the eastern boundary (on the isopycnal, of course, this is also warmer water). The deep shear field shown here does not show this flow, but the earlier work of Helland-Hansen and Nansen (1926) and Reid (1978) at depths near 1000 m does show it. A more recent study (Reid, 1979) discusses the extension of some of the deeper warmer waters through the Faroe–Shetland Channel into the Norwegian Sea, where they contribute to the warm layer of the eastern and northern Norwegian Sea and the Arctic Ocean.

The warm water (the high-salinity tongue) extending along the western boundary of the South Atlantic into the Circumpolar Current is well known, of course. And along this isopycnal, an extension of the warmer waters into the North Pacific, at both the eastern and western boundaries, is seen, and is roughly consistent with the shear field. In neither case, however, is the flow only meridional: if the oxygen and nutrient patterns and the shear field have been interpreted correctly, this water may reach the higher latitudes, at least in part, by a gyre-to-gyre transport, involving substantial zonal flow.

If diagnostic models are to proceed usefully, then a better density field must be provided for much of the ocean. Substantial improvements over the selections shown here can certainly be made, given enough time, but a set of north–south station lines must also be obtained if the density field is to be made more useful.