Part One

General Ocean Circulation

I
Deep Circulation of the World Ocean

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1.1 Introduction

Historically, the deep circulation of the ocean has been viewed from the perspective of property fields, mainly the distributions of temperature, salinity, density, and dissolved-oxgyen concentration. The practical reason for not considering velocity measurements as well, of course, was a technical incapacity for making them until very recently. On the whole, this was probably not a bad thing: not merely because the property distributions are as interesting in themselves as the motion field, but also because the scalar fields are so much more stable than the velocity vectors—allowing spot measurements from different areas even years apart to be combined into coherent pictures that tell a good deal about general patterns of deep flow, albeit indirectly. The slight differences between corresponding hydrographic sections in the atlases by Fuglister (1960) and by Wüst and Defant (1936), when compared with the fluctuations much larger than the means of deep velocities observed by the MODE Group (1978), for example, demonstrate how much easier it is to obtain statistically significant information pertinent to the overall global deep circulation from water-property data than from current measurements.

On the other hand, the information gained from the property fields allows only a limited view of the deep motions, at very best some kind of long-term average. Although oceanographers have usually been mindful of variability in the deep flow, even if only to accomplish eddy mixing, it seems extremely unlikely that anyone imagined the highly energetic low-frequency mesoscale motions that current records have revealed. Instead, because of the stability of the property fields, it was stationarity rather than variability that was emphasized, however implicitly, in the circulation pictures derived from them. That stability also was surely the basis for the conceptual structure of water types and masses that has been so enormously useful in summarizing and comprehending the temperature-salinity structure of the ocean and in identifying features in the property fields that can be exploited as tracers for the flow. Without velocity information, though, such descriptions of oceans have sometimes degenerated into taxonomic sterility (naming something doesn't explain it), and perhaps sometimes there has been too elemental a character ascribed to water masses (as if they were truly building blocks rather than names for features), leading to pictures of the ocean more suggestive of rigid geological strata than of the real motion field that forms the distributions.

Plainly there cannot be a satisfying description of the deep ocean circulation that does not meld station data with current records. It does not seem to me, though, that such a description is yet possible. Far too few current records have been obtained to describe the deep, low-frequency motions in a global sense; it is only in the western North Atlantic that one can even contemplate making a basin-wide description. Moreover, we simply have not learned how to combine the stable station data with the fluctuating-velocity records to tell a story that is both consistent and informative. For example, Reid, Nowlin, and Patzert (1977) reported a record (Cato 2) from a current meter moored on the South American continental slope in the core of the North Atlantic Deep Water: for 2 weeks the daily-averaged velocity vectors were directed southwestward, parallel to the isobaths, as one would have expected in this particular deep western boundary current; but then the flow abruptly changed direction and went eastward for nearly 2 weeks. Nevertheless, with due regard for different density and accuracy of observations, the high-salinity core of the current looked very much as had been depicted by Fuglister (1960) and Wüst and Defant (1936). How are we to approach these two different sets of data, to reconcile the variability of the one to the steadiness of the other, and to learn something significant from their combination about that boundary current?

Finally, it is not at all clear what effect the low-frequency velocity fluctuations have on the long-term mean flow. There is enough theoretical reason (e.g., Rhines, 1977) to suspect that their role in its dynamics may be substantial, but measurements of deep Reynolds stresses are meager. In fact, values reported by Schmitz (1977) from the Sargasso Sea well south of the Gulf Stream actually favor a negligible contribution to the vorticity balance there, but those measurements are far too few to give a general characterization of the deep open ocean.

Consequently, although I recognize its incompleteness, the following account of the deep circulation of the world ocean is undertaken mainly from the traditional perspective of hydrographic station data, with reference to current measurements only where they seem helpful in estimating velocities and transports of the prominent currents. The emphasis is on mean thermohaline circulation. What has been learned about the low-frequency motions is described in detail by Wunsch in chapter 11 of this volume.

In section 1.2, I have attempted a historical review of what seem to me to be the important events and dates in the development of ideas about the deep circulation, from the first deep temperature measurements through Sverdrup's comprehensive synthesis in chapter XV of *The Oceans* (Sverdrup, Johnson, and Fleming, 1942). In section 1.3, I have discussed the dynamical ideas of Stommel and his colleagues that led to the overthrow of a substantial part of Sverdrup's picture, and its revision in contemporary thinking with dynamically consistent models of circulation. Section 1.4 is an account of the sinking processes that supply water to the deep ocean from the surface layer. Section 1.5 is a consideration of how well the kinds of deep-circulation patterns envisioned in dynamical theory stand up to observation; it is necessarily mainly a digest of the evidence for deep western boundary currents in the world ocean. Finally, in section 1.6, I have speculated about some fundamental aspects of the deep circulation that seem to me to be not very well understood at this time.

The focus throughout is more on the circulation of deep water than on the complementary problem of its properties, because the subject of deep-water characteristics has been treated in detail by Worthington in chapter 2 of this volume. There is, of course, some overlap with that chapter, as well as with Reid's general discussion of the mid-depth circulation (chapter 3). To the extent that our opinions are in conflict, we hope that readers will recognize subjects for further observation and thought.

1.2 Historical Development of Ideas about the Deep Circulation

In 1751 Henry Ellis, captain of the British slavetrader *Earl of Halifax*, wrote to the Reverend Stephen Hales to describe some deep temperature measurements that he had made at lat. 25°13′N, long. 25°12′W, with a "bucket sea-gage" devised and provided for him by Hales. This instrument, to be attached to a sounding line, was a "common houshold pail" covered at top and bottom by valves that would be forced open during descent and pushed shut by drawing the bucket back to the surface; it was furnished with a thermometer so that the temperature of the water sample thus trapped could be read when the bucket was returned to the ship. In his letter, which Hales transmitted to the Royal Society of London, Ellis (1751) reported:

> Upon the passage, I made several trials with the bucket sea-gage, in latitude 25′-13″ north; longitude 25′-12″ west. I charged it and let it down to different depths, from 360 feet to 5346 feet; when I discovered, by a small thermometer of Fahrenheit's, made by Mr. Bird, which went down in it, that the cold increased regularly, in proportion to the depths, till it descended to 3900 feet: from whence the mercury in the thermometer came up at 53 degrees; and tho' I afterwards sunk it to the depth of 5346 feet, that is a mile and 66 feet, it came up no lower. The warmth of the water upon the surface, and that of the air, was at that time by the thermometer 84 degrees. I doubt not but that the water was a degree or two colder, when it enter'd the bucket, at the greatest depth, but in coming up had acquired some warmth.

Ellis's guess of one or two degrees warming during the ascent was based on changes observed in the temperature of the sample while on deck. Modern data, however, indicate that Ellis's values were some 10–

13°F too high. Perhaps he underestimated the conduction through the walls of the sampler when it was in the ocean, or perhaps the valves were not tight enough to prevent exchange of water while it was being raised (especially if raised unevenly).

Nevertheless, these are the earliest recorded subsurface temperature measurements in the open ocean—indeed, the first anywhere to a substantial depth—and they pointed to what must be the most fundamental and striking physical feature of the ocean: that deep water is all cold, and warm water is confined to a relatively thin layer near the surface in the tropics and subtropics.

Ellis himself did not seem to realize the far-reaching significance of his data, for he remarked later in his letter to Hales:

> This experiment, which seem'd at first but mere food for curiosity, became in the interim very useful to us. By its means we supplied our cold bath, and cooled our wines or water at pleasure; which is vastly agreeable to us in this burning climate.

He does not seem to have made any further measurements, nor does his work appear to have stimulated immediate exploration of the deep water by others, for, according to Prestwich's (1875) tabulation, it was more than 60 years before any additional temperature measurements were made to the depth that Ellis reached (though others were made during this time at lesser depths).

What Ellis's observations meant, of course, was that deep water in the tropics must derive from polar regions, and that, accordingly, there must be a meridional circulation system in the ocean to carry deep water equatorward. Evidently (M. Deacon, 1971), the first recorded recognition of this profound implication was by Count Rumford in his essay, "The Propagation of Heat in Fluids," first published in 1797.[1] Rumford had made the experimental discovery of convection currents in liquids (the basic subject of this essay), and, in considering their possible role in nature, he reasoned (Rumford, 1800):

> But if the water of the ocean, which, on being deprived of a great part of its Heat by cold winds, descends to the bottom of the sea, cannot be warmed *where it descends,* as its specific gravity is greater than that of water at the same depth in warmer latitudes, it will immediately begin to spread on the bottom of the sea, and to flow towards the equator, and this must necessarily produce a current at the surface in an opposite direction.

In advancing evidence for such a deep flow, he drew the correct inference from Ellis's measurements, in as clear and straightforward a fashion as one could want:

> But a still more striking, and I might, I believe, say, an incontrovertible proof of the existence of currents of cold water at the bottom of the sea, setting from the poles towards the equator, is the very remarkable difference that has been found to subsist between the temperature of the sea at the surface and at great depth, at the tropic—though the temperature of the atmosphere *there* is so constant that the greatest changes produced in it by the seasons seldom amounts to more than five or six degrees; yet the difference between the Heat of the water at the surface of the sea, and that at the depth of 3600 [*sic*] feet, has been found to amount to no less than 31 degrees; the temperature above or at the surface being 84°, and at the given depth below no more than 53°.
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> It appears to me to be extremely difficult, if not quite impossible, to account for this degree of cold at the bottom of the sea in the torrid zone, on any other supposition than that of *cold currents from the poles;* and the utility of these currents in tempering the excessive heats of these climates is too evident to require any illustration.[2]

During the early nineteenth century, Humboldt (1814, 1831, 1845) popularized the notion of deep currents flowing from polar regions toward the equator. He mentioned no sources for the idea, but he did cite Ellis's measurements and later ones of better quality as supporting data.[3]

The physicist Lenz (1845) had interested himself in how the subsurface vertical temperature gradient varied with latitude. In the course of his investigation he discovered the shoaling of the thermocline at the equator in the Atlantic; and the much sparser data available from the Pacific suggested the same phenomenon to him there. He regarded that shoaling as evidence for upwelling of deep water to the sea surface, and, acquainted with Humboldt's work, he proposed a more specific scheme of closed meridional circulations than had been given hitherto. This scheme involved great convection cells symmetric about the equator, with water sinking in high latitudes, flowing equatorward at depth, and rising in the tropics to return poleward near the surface. He considered that this upwelling also contributed to the reduction in surface salinity that he had observed at the equator in the Atlantic. His understanding of the driving mechanism was not more profound than Rumford's, but he did recognize that wind stress and the earth's rotation would distort the motions substantially from the simple cellular form. He made no explicit reference to the Indian Ocean, but he discussed the symmetric cells in terms of general validity. Variants on Lenz's construction, some perhaps conceived independently, were to influence interpretations of the deep circulation into the first decade of the twentieth century (e.g., Schott, 1902; Brennecke, 1909).

Dr. Carpenter, for one (Carpenter, Jeffreys, and Thomson, 1869), invoked the general idea to account for differences in deep temperatures observed in the northern North Atlantic, and his resulting prolonged and contentious controversy with Croll attracted a great deal of attention to it (M. Deacon, 1971).[4] Carpenter had a more elastic conception of the convection cells than did Lenz, though, believing that their warm, poleward-directed limbs could touch bottom in places; nor was he rigid about the equatorial symmetry, considering that the northern cells would be weaker and less extended laterally than the southern ones, in conformity with the oceans being much more open to polar water in the south than in the north—whereby deep water from the Antarctic could be expected to penetrate into the northern hemisphere. He referred explicitly to recent subsurface temperature measurements in the Arabian Sea to argue that all deep water in the Indian Ocean derived from the Antarctic, and he implied a belief that most of the deep water in the Pacific would also be found to come from the south.

Nevertheless, Prestwich (1875) seems to have been the first to demonstrate from observations the nonexistence in the Pacific of relatively strong deep upwelling near the equator and the lack of a northern source for the deep water there. These facts showed that deep water in the North Pacific, as well as that in the South Pacific, was supplied from the Antarctic, and thus invalidated Lenz's picture of symmetric convection cells. Prestwich believed, however, that it was still a correct description of the meridional circulation in the Atlantic.

Following the *Challenger* expedition, Buchan (1895) produced world-ocean maps of the distribution of temperature at various levels, based mainly on the *Challenger* data. His map for 2200 fathoms illustrated the spreading of deep water from the Antarctic into the three oceans, including details of how its course is shaped by submarine ridges:

It is also to be noted that the lowest deep-sea temperatures are found in those parts of the ocean which lie in the Southern hemisphere, and that, on the whole, higher temperatures are encountered as we recede from the Antarctic region. It may also be pointed out that the lower deep-sea temperatures extend farther to the north from the Southern Ocean, just over those depths of the sea which appear to have, and probably do have, a direct communication with the south; that is, are not cut off by any intervening submarine ridge separating them from the cold waters of Antarctica.

. . . There can be no doubt that these very low deep-sea temperatures have their origin in the Southern or Antarctic Ocean, the icy cold waters of which are propagated northward, the rate of propagation being so slight as to be regarded rather as a slow creep than as a distinctly recognizable movement of the water.

Buchan (1895) also inferred a southward movement of deep water from the North Atlantic into the South Atlantic. As Merz and Wüst (1922) pointed out later, this perception, together with Buchanan's (1884) specific-gravity profile that showed the intermediate-water salinity minimum extending well north of the equator from high southern latitudes, would have been sufficient to disprove Lenz's equatorial upwelling from great depth to the sea surface—and thereby his symmetric convection cells for the Atlantic; but neither Buchan nor Buchanan attempted a comprehensive discussion of the meridional circulation.

Buchan formed his idea of southward flow after noticing in the *Challenger* data that the vertical temperature gradients at depths from 800 fathoms to 1500 fathoms and deeper (how much deeper not stated) were much smaller in the South Atlantic than in the North Atlantic. He linked this difference to the distribution of salinity (specific gravity):

The specific gravities at the bottom of the ocean afford a ready explanation for this remarkable distribution of temperature. Owing to the higher specific gravities of the North Atlantic, an extensive deep-sea current from the North to the South Atlantic, carrying a higher temperature with it, sets in at depths at which the influence of the surface currents is no longer felt, and becomes more pronounced as the depth below 1000 fathoms is increased. Hence the North Atlantic receives large accessions to its salinity from the surface currents, which the deep-sea currents again return to the South Atlantic.

This was probably the first recognition of deep southward flow across the equator, but unfortunately it is all that Buchan (1895) said on the subject. One cannot tell whether he also realized that the deep flow was layered, with North Atlantic water overriding the bottom water from the Antarctic, or whether he viewed "deep water" in the terms of his day as a single unit moving uniformly northward or southward, whereby the oppositely directed North Atlantic and Antarctic flows would simply collide and upwell at some middle southern latitude.

Whatever Buchan may have believed, the first unambiguous, well-documented statement of deep southward flow from the North Atlantic between the bottom and intermediate waters from the Antarctic was made by Brennecke (1911) on the basis of stations occupied by the *Deutschland* along the western rim of the South Atlantic:

Das Hauptergebnis unserer Reihenmessungen ist die Feststellung eines Tiefenstromes in etwa 1500 bis 3000 m Tiefe der vom Nordatlantischen Ozean nach Süden vordringt und durch hohe Temperatur und hohen Salzgehalt sich von der über- und unterlagernder Schicht abhebt. Dieser Tiefenstrom konnte bei allen Rehenmessungen von 5°S-Br. bis 40°S-Br. klar erkannt werden. Wenn bislang noch nicht erkannt worden ist, so liegt dies einerseits an der geringen Zahl der Messungen, die im diesen Schichten tatsächlich ausgeführt worden sind, andererseits an den frühen vielfach benutzten Maxima-Mimima-Thermometern, die eine Temperaturumkehr in der Tiefe nicht anzeigen. Soweit unserer Messungen Aufschluss geben, wird dieser Tiefenstrom bei seinem Vordringen nach Süden in grössere Tiefe gedrängt, bzw. in seiner Oberschicht mehr und mehr durch den schon von früheren Forschungen her bekannten, entgegengesetzt gerichteten, d.h. nordwärts vordringenden Tiefenstrom in 1000 m (ausgezeichnet durch das Minimum des Salzgehalts) gemischt.

The *Challenger* thermometers having been of the minimum-type, Buchan could not have detected the temperature inversion. Another factor that contributed to Brennecke's success was that, although he could not have known it, he actually made his stations in the swiftest part of the southward flow, where the inversion is much more markedly developed than to the east, and where, happily, the strongest evidence for the flow was available.

In the early 1920s, Merz amassed a file of all available deep observations of temperature and salinity, and began a program of systematic reexamination of these data to delineate the deep circulation of the whole world ocean. In the course of this study Merz and Wüst (1922) and Merz (1925) explicitly refuted the existence of Lenz's symmetric convection cells in the Atlantic, on the evidence already cited, and they proposed a comprehensive new picture of the meridional circulation there, whose essential feature was hemispheric exchange of water: ocean-wide northward flow across the equator in the upper kilometer, compensated by southward flow below to about 4000 m, with basin-wide northward and southward flows of bottom water in the western and eastern basins, respectively.[5] Their working materials were north-south sections of temperature and salinity, and they regarded the depression of isotherms and isohalines in the upper 2 km in the northern subtropics as indicating sinking from the surface in those latitudes, and thus formation of the North Atlantic deep water there. In making this misinterpretation of that feature, they evidently did not fully consider either the baroclinic character of the Gulf Stream gyre or the development of anomalously high temperatures and salinities at mid-depth from mixing of the Mediterranean outflow into the gyre.[6]

Following Merz's death in 1925 on the *Meteor* Expedition, Möller and Wüst undertook to complete his program for the Indian and Pacific Oceans, respectively. In her study of the Indian Ocean data set Möller (1929) joined with Schott (1926) to advance the Merz scheme of meridional circulation, emphasizing hemispheric exchange, for that ocean too—northward flow of intermediate and bottom water, southward flow of deep water in between—with the deep water sinking from the surface in the Arabian Sea, and receiving contributions from the Red Sea and Persian Gulf outflows. Thomsen (1933), however, found systematic error in some of these salinity values, and new data from the *Dana* showed much less extensive southward penetration of high-salinity deep water from the North Indian Ocean than Möller had described, implying a different character of deep-layer flow from that in the Atlantic: certainly weaker, and if directed across the equator at all, only in response to the northward movement of intermediate and bottom water. Implicit in this conclusion was rejection of the idea of deep water sinking from the surface of the Arabian Sea, although the influx of Red Sea water was recognized. From the distribution of bottom potential temperature, Wüst (1934) clarified the role of the Central Indian Ridge in dividing the bottom flow into at least two separate regimes (the Ninetyeast Ridge being then unknown), with bottom water entering from the south into both the western and the eastern basins of the Indian Ocean, as Schott (1902) had suggested in his report on the *Valdivia* observations.

For the Pacific, Wüst (1929) found that the general quality of data available was inferior to that for the other two oceans, and, with deep property differences being much smaller there anyway, he did not reach such definite conclusions as for the Atlantic. Nevertheless, his analysis suggested to him that the Pacific was different from the Atlantic, in that hemispheric exchange was insignificant, and that, in both the North and South Pacific, intermediate and bottom water spread equatorward and deep water poleward (i.e., that deep water sank from the surface layer near the equator, rather than in middle northern latitudes, as he thought it did in the Atlantic). In one respect, Wüst's (1929, 1938) work on the deep Pacific in this decade was actually retrogressive, because some temperature measurements made by the *Tuscarora* in 1874 that are incompatible with modern data led him to infer a northern source for bottom water in the Sea of Okhotsk, contrary to Prestwich's (1875) correct finding that sinking to great depth does not occur anywhere in the North Pacific.

From his study of the *Carnegie* observations, Sverdrup (1931) realized that there was not, in fact, any equatorial downwelling in the Pacific. He also recognized that Pacific bottom water did not derive from sinking directly to the south near Antarctica, as is the case for the Atlantic, but that all water below the low-salinity intermediate layers in the Pacific was, in effect, carried to it by the Circumpolar Current. He thought it likely that at least in the South Pacific there was some southward return flow of deep water, as in the other two oceans, but the very low oxygen concentration in the deep North Pacific indicated only small hemispheric exchange.

The idea of a deep-water connection among the Atlantic, Indian, and Pacific Oceans via the Circumpolar Current was reinforced by the systematic cruises of the *Discovery* in the Antarctic, which obtained evidence for the extension of high-salinity deep water all the way from the Atlantic to the Pacific (Deacon, 1937).

By the time Wüst began his comprehensive treatment of the *Meteor* results, Helland-Hansen and Nansen (1926) had shown clearly—and Defant (1931) had reaffirmed—that the high salinities at mid-depths in the subtropical North Atlantic were due to the Mediterranean outflow; and Wattenberg (1929) had traced the maximum in oxygen concentration below to high northern latitudes. Wüst (1935) therefore abandoned the broad mid-latitude source for the deep water that Merz and he (1922) had proposed, and that he (1928) had later modified somewhat. Instead, he distinguished three layers within the deep water, all spreading southward, and each with its own "formation" site: Upper North Atlantic Deep Water, characterized by the deep salinity maximum attributed to the Mediterranean outflow; Middle North Atlantic Deep Water, identified by a maximum in dissolved-oxygen concentration, and traced to the Labrador Sea; and Lower North Atlantic Deep Water, identical to Wüst's (1928) earlier "North Atlantic Bottom Water," defined by a second and deeper oxygen maximum, and thought to be formed somewhere in the waters south or southeast of Greenland by deep convection in winter.

The picture of deep circulation for the world ocean that Sverdrup (Sverdrup, Johnson, and Fleming, 1942, chapter XV) then constructed was broadly Merzian: ocean-wide or basin-wide flows in deep and bottom layers, the bottom water moving generally northward, the deep water southward—with the greatest hemispheric exchange in the Atlantic and the least in the Pacific. Sverdrup departed qualitatively from the Merzian schematic only in inferring that the principal movement of deep (and bottom) water in the North Pacific was a slow, nearly closed, clockwise circulation. By this time, though, the areal extent of regions where deep sinking was thought to occur had been much reduced from what Merz had supposed: sinking in the southern hemisphere limited mainly to the Antarctic continental slope in the Weddell Sea (Brennecke, 1915, 1921; Mosby, 1934), and sinking in the northern hemisphere confined to high latitudes in the western basin of the Atlantic, with probably smaller, subsurface influxes of relatively dense water from such marginal seas as the Mediterranean and Red Seas.

Even then, however, Wüst had already realized that the idea of deep and bottom flows in the Atlantic that were uniform in a basin-wide sense was oversimplified to the point of being misleading. His maps of tracer distributions from the *Meteor* data (Wüst, 1935) demonstrated more intense southward propagation of North Atlantic Deep Water in the western than in the eastern South Atlantic, and geostrophic velocity calculations (Wüst, 1938) showed relatively strong northward flow of Antarctic Bottom Water (at speeds of several centimeters per second) only close to the South American continental slope. In discussing property distributions, Wüst (1935) was careful to speak of the "spreading" ("*Ausbreitung*") of water types, stating that such spreading had a current-like character only on the western side of the Atlantic, and that elsewhere it was accomplished, in effect, by eddy fluxes. He regarded this distribution of currents as an extremely strange phenomenon, for which he was unable to give a satisfactory explanation; and, indeed, the real significance of these "deep western boundary currents" was not to be appreciated until dynamically consistent models of the deep circulation were developed.

It should be acknowledged, however, that, as penetrating as Wüst's insights were, it was not, in fact, he who first discovered the bottom-water boundary current, but Buchanan, who was presumably the chief contributor to chapter XXII of the *Challenger Report Narrative*. The *Challenger* stations were the first set that by virtue of their spacing could possibly have revealed such a feature, and Buchanan (Tizard, Mosely, Buchanan, and Murray, 1885a) did not fail to see it:

> Another equally remarkable current is that which brings the cold water from the south polar regions up along the South American coast at the bottom as far as the Equator. . . . That this does take place is shown by the low bottom temperatures observed at every Station along the western side of the South Atlantic almost to the Equator, and that there is here a sensible current of cold water along the bottom is shown by the sudden change in the rate of decrease of temperature with increasing depth which is observed at depths of from 2000 to 2200 fathoms.
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> . . . The temperature of the water from 2400 fathoms to the bottom was uniform, the mean result of six observations being 32.43°. This water underruns the body of the Atlantic water, which at 1500 to 2000 fathoms has here a temperature of 37°, producing a temperature gradient of about 1.3° per hundred fathoms at the steepest. For the preservation of this gradient, a considerable supply of cold water is requisite, and it must be drawn from higher latitudes. But any motion of the water towards the Equator will be accompanied by a strong deflection to the westward (proportional to the change of the cosine of the latitude). A measure of this deflecting force is furnished by the rise of this cold water at the more inshore Station on the 28TH, where the maximum gradient is at about 1750 fathoms, while on the 29TH, at a distance of 120 miles,[7] it is at about 2100 fathoms.

In the last two sentences Buchanan seems to have been groping toward some notion of the thermal-wind relation, and to have correctly associated the lateral scale of the slope in the maximum-gradient surface with that of the northward flow beneath. He had bad luck with his discoveries, though. Not only was this one ignored [except for an allusion by Wüst (1933)], but, as will be recalled, his almost simultaneous discovery of the Atlantic Equatorial Undercurrent was completely forgotten after three decades.

1.3 A Dynamical Framework

Paralleling the intuitively appealing concept of ocean-wide or basin-wide meridional flows were two other ideas that seemed equally reasonable. One is that, in compensation for the known sinking of near-surface water in small regions, there is a general slow upward movement of deep water over most of the rest of the ocean, which balances the downward diffusion of heat from the upper layer and thus accounts for the existence of the main thermocline. The other is that the mean horizontal deep flow in the open ocean is strictly geostrophic, in the sense that streamlines parallel isobars, and the speed is inversely proportional to their spacing and to the sine of the latitude. As indicated in section 1.1, the latter may not seem so obvious now as it once did, but there is yet no proof to the contrary. Stommel (1965, manuscript of first edition completed in 1955) perceived the dynamical implication of these ideas—that the vertically integrated deep flow should be directed poleward—with the bizarre corollary that deep water in the open ocean of known polar origin should be moving on the whole *toward* its sources rather than away from them. Specifically, broad, deep equatorward flows in the North Atlantic, South Indian, and South Pacific Oceans, as envisaged in the Merz-Sverdrup picture, do not seem consistent with the peculiar dynamics of slow motions on a rotating sphere.

The physical argument is simple. The geostrophic vorticity balance is $\beta v = f w_z$, where f is the Coriolis parameter, β its meridional derivative, v the northward velocity component and w the vertical component, and z the vertical coordinate (positive upward). Integrating the equation vertically from the ocean bottom to the "top" of the deep water (however defined), and taking the floor of the ocean to be level in the large-scale mean, so that the vertical velocity at the bottom can be taken to be zero in a large-scale sense, then yields the result that, with upward velocities at the top of the deep water, the meridional component of flow is poleward, in both the northern and the southern hemispheres. This can also be seen in more rudimentary terms: because geostrophic flow cannot cross isobars, the transport per unit depth between any two streamlines of the horizontal flow increases downstream in flow toward lower latitudes, and decreases downstream in flow toward higher latitudes. *If* water is also moving upward out of the deep layer, then the associated horizontal flow must be the convergent one—directed poleward.

Clearly, however, there must be regions where this constraint breaks down, so that deep water of polar origin can reach the low latitudes where it is found. For the large-scale circulation generally, the regions where the rules of the open ocean are relaxed are the western boundaries of basins, where relatively narrow, intense currents exist to satisfy whatever boundary or continuity conditions are not fulfilled by the interior flow (see section 1.5.4). Stommel (1958) therefore hypothesized deep western boundary currents for all the oceans to carry deep water away from its sources, to supply it to the interiors of oceans from their western sides (rather than from their polar margins) and thus feed the poleward interior flows, and to correct ocean-wide continuity imbalances between the upward flux and the meridional flow.

The flow patterns thus constructed were so contrary to intuition that, to test their consistency, Stommel, Arons, and Faller (1958) developed a corresponding theory for circulation systems forced by a variety of prescribed distributions of sources and sinks in a rotating tank. The excellent agreement with the experiments [continued further by Faller (1960)] encouraged Stommel and Arons to rework the theory in spherical coordinates (1960a), and they used that theory to develop a model of the global deep circulation (1960b), including estimates of velocities and volume transports. The three oceans were represented as flat-bottom basins bounded by either parallels or meridians, and connected in the south by an analogue to the Antarctic Circumpolar Current. The flow was taken to be barotropic, driven by sources in the model counterparts of the Weddell Sea and northern North Atlantic (representing sinking from the surface layer), and by the compensatory upwelling, distributed uniformly over the rest of the world ocean. The resulting circulation pattern in a slightly modified version of the model (Kuo and Veronis, 1973) is sketched in figure 1.1. The analogue to the Circumpolar Current is indicated along the southern margin; its transport through the counterpart of the Drake Passage is not determined by the model dynamics, and was estimated roughly from observations. The layer thickness is 3 km, the strength of each source is about $17 \times 10^6\ m^3\,s^{-1}$, and the upwelling rate is $1.5 \times 10^{-5}\ cm\,s^{-1}$; the horizontal speeds in the interior are of order $10^{-2}\ cm\,s^{-1}$, and the transports of the western boundary currents, of order $10 \times 10^6\ m^3\,s^{-1}$.

The interior flows are everywhere poleward and eastward, fed by intense currents along the western boundaries. In the Atlantic, the boundary current emanates from the northern source and flows to the southern limit of the South Atlantic, where it, the interior flow, and the Weddell Sea source increase the transport of the deep Circumpolar Current. That in turn is the source for northward-flowing western boundary currents in the Indian and Pacific Oceans, entering at their southwestern corners. At the northern margins of those two oceans the interior flow forced by the upwelling feeds into northern boundary currents, which flow westward into southward-directed western boundary currents.

This model was never intended as a realistic *description* of the deep-ocean circulation, and in at least two respects it would be a qualitatively bad one. By treating the ocean floor as level, the model gives single circulation systems in the northern and southern hemispheres of each ocean, whereas the mid-ocean ridges divide the oceans into separate basins with multiple circulation systems and multiple entrance points for deep water from the Antarctic. Also, because the model is barotropic, it cannot allow layered deep flow, with opposed boundary currents one above the other, as occur most spectacularly in the South Atlantic, where Antarctic Bottom Water flows northward beneath the southward current of North Atlantic Deep Water. (In fact, there could be no direct, boundary-current flow of Antarctic Bottom Water from the Weddell Sea into the South Atlantic anyway were it not that the Drake Passage sill is substantially shallower than the floor of the Atlantic to the east.)

Nevertheless, it was a dynamically consistent scheme, which offered some drastically different ideas about the deep circulation from those in the Merz-Sverdrup picture of basin-wide meridional flows, especially the new physical concept of deep western boundary currents. A search was immediately made for the predicted deep current off the east coast of the United States, by means of direct current measurements with neutrally buoyant floats; its discovery was reported by Swallow and Worthington (1957), and provided dramatic support for the essential truth of the new conceptual scheme. Considering what has been learned since then about the prevalence of strong, low-frequency motions in the deep water, one would probably question now how sound a "proof" of the current those measurements were, given their restricted geographical scope and brevity of duration (1 to 4 days); but they were consistent among themselves and with later observations (see section 1.5).

Actually, as pointed out in section 1.2, the relatively intense deep meridional flows adjacent to South America had been recognized long before, and Stommel (1950a) had already suggested that they might be dynamically akin to the Gulf Stream. Moreover, as the theory was being developed, Wüst (1955) constructed geostrophic velocity sections for two of the *Meteor* transects, one of which vividly illustrated these two currents, the one directed northward, the other southward. Thus the general idea of deep western boundary currents was not without observational support even before its confirmation by Swallow and Worthington (1957).

It was only for simplicity and clarity that the Stommel-Arons model was presented as a barotropic one. It can be brought somewhat closer to the real stratified ocean by regarding it instead as a theory for vertically integrated flow. Equatorward interior flow at some levels is therefore not excluded, so long as the vertical integral is poleward. Layered interior flows might be difficult to detect in tracer distributions, though, because mixing of properties can mask the flow patterns (see below, this section).

The theory is clear, however, in allowing no hemispheric exchange of water (of the sort that Merz emphasized) through the interior flow, because at all levels the meridional component of flow vanishes with the Coriolis parameter at the equator: hemispheric ex-

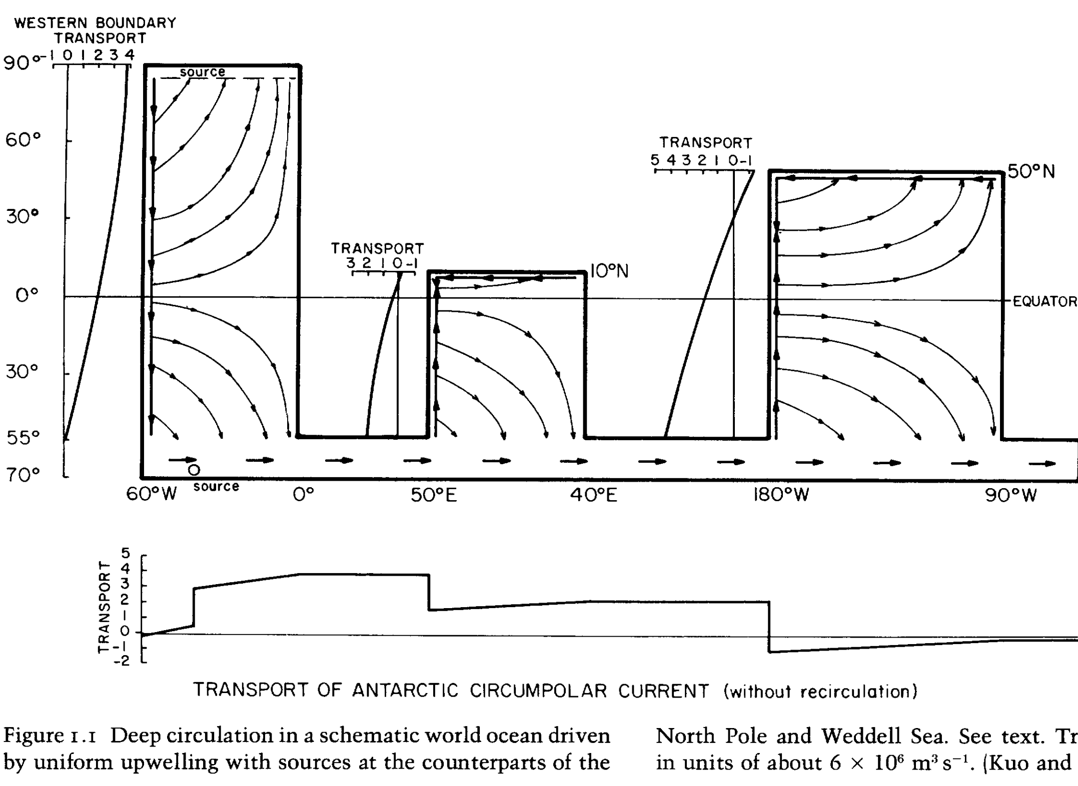


Figure 1.1 Deep circulation in a schematic world ocean driven by uniform upwelling with sources at the counterparts of the North Pole and Weddell Sea. See text. Transports measured in units of about 6×10^6 $m^3 s^{-1}$. (Kuo and Veronis, 1973.)

change by the mean flow is limited to the boundary currents. There can still be hemispheric exchange of *properties* all along the equator, of course, through lateral mixing.

A more fundamental novelty of the Stommel-Arons model is in its looser connection between specific horizontal flows and specific sinking phenomena than was implied in the earlier schemes. The net interior flow at any position is directly coupled to the local upward movement of the deep water, but the amplitude of that upwelling (as distinguished from its geographical variation) is related to the world-ocean integral of the rate of sinking into the deep water, rather than to the downward flux in any one sinking region. Nor is there any one-to-one relation between the transport of a deep western boundary current and the rate of sinking in a region from which it may emanate, even after subtracting losses to interior upwelling—as may be seen by considering the water budget for the portion of a deep ocean north of an arbitrary parallel of latitude. The boundary-current transport T out of that volume (positive southward) can be calculated as

$$T = \frac{f}{\beta} \int w \, dl - \int w \, d\sigma + S,$$

where S is the rate of sinking into that volume, and w is the vertical velocity at the top of the deep water; the first term on the right, expressing the net interior transport into the volume, includes an integral of w along the arbitrary latitude, and the second term is an integral over the top of the volume. Not only can T be less than S, as is obvious without the formula, but T can also be *greater* than S, signifying recirculation in an ocean or basin, because the boundary current must compensate not only the sinking but also any excess of interior transport into the volume over upward flux out of it. Extreme examples of this effect are illustrated in the northern Pacific and Indian areas of figure 1.1, where southward flowing boundary currents are required even though no sinking occurs to the north.

Despite the physical consistency of the deep-circulation model, the Merz-Sverdrup picture still looks intuitively more believable, in that poleward interior flows are not obvious in the property distributions. Stommel and Arons (1960b) suggested that lateral mixing of tracer properties might be so intense as to mask the patterns of such slow mean interior flow; to explore that effect, Kuo and Veronis (1973) made calculations of dissolved-oxygen distributions based on the circulation field of figure 1.1 and ranges of values for the mixing coefficient and oxygen-consumption rate. They prescribed the oxygen concentration at the two sinking points, and compared the resulting fields with the observed distribution at 4 km.

Two extreme cases are instructive. With no mixing, high oxygen concentrations are introduced to the oceans from their western boundaries, and the values diminish eastward on account of consumption, with the effect that the lowest values are found along the eastern boundaries; the western boundary currents are discernible in the field, as well as the eastward component of interior flow, but the meridional component of the latter is not particularly evident. With mixing but no advection, however, the oxygen concentration is simply diffused zonally in the circumpolar belt, and relatively high values are introduced to the Indian and Pacific Oceans all along their southern boundaries; the concentrations decrease northward and the lowest values in those oceans are obtained at their northern boundaries. (In the Atlantic, with northern and southern sources, the lowest values are found in the intertropical zone.)

The best match with observation (figure 1.2) was an intermediate case, representing approximately "equal" effects of advection and mixing. The mixing coefficient in this case was 6×10^6 cm^2 s^{-1}, and the oxygen consumption rate, 2×10^{-3} ml l^{-1} yr^{-1}—values roughly consistent with those from other studies. The patterns were found to depend on the speed of the Circumpolar Current, since that affected the "source" values for the Indian and Pacific Oceans; the optimal transport value for the Drake Passage was 35×10^6 m^3 s^{-1}. That figure from the model is difficult to compare precisely with observation because, given that the "top" of what ought to be considered deep water is much shallower in the Drake Passage than in the tropics, it is not clear what is the most appropriate depth interval to choose for comparison; density sections combined with yearlong current measurements made during the ISOS program gave estimates of the net transport below 1000 m of about 47×10^6 m^3 s^{-1}, and below 2000 m of 12×10^6 m^3 s^{-1} (H. Bryden and R. D. Pillsbury, personal communication).

Certainly there is nothing in figure 1.2 to indicate the character of the underlying flow field, except perhaps some hint of the western boundary currents. The most obvious interpretation of the pattern—and one quite wrong—would be of slow, ocean-wide, transequatorial movement northward from the Antarctic and southward from the northern North Atlantic, in the manner of Merz and Sverdrup, though perhaps somewhat stronger in the west than in the east. Even though, as Kuo and Veronis (1973) pointed out, these calculations were based on a highly simplified flow scheme (especially as not constrained by the ridge system), the results do suggest that, with recognition of a moderate degree of lateral mixing in the ocean, observed tracer distributions can probably be rationalized in terms of more realistic flow fields constructed from the Stommel-Arons dynamics.

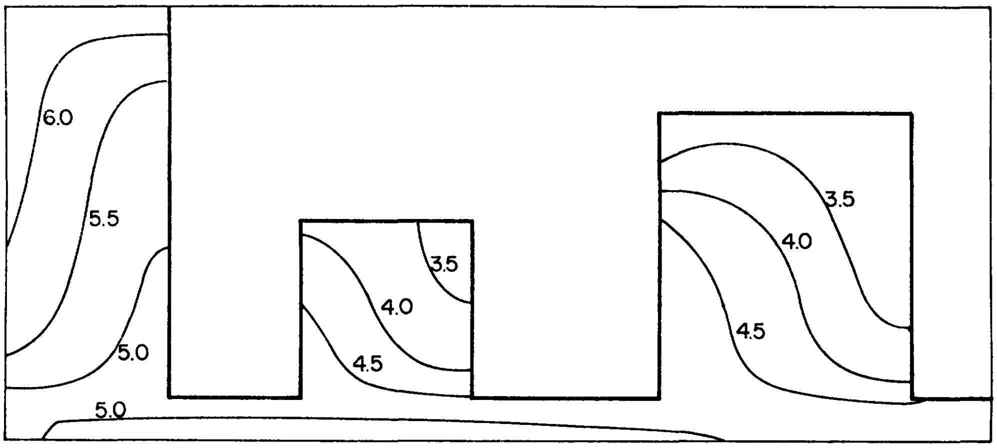


Figure 1.2 Model distribution of dissolved-oxygen concentration ($ml\,l^{-1}$) associated with the flow field in figure 1.1. See text for "best-fit" values of controlling parameters. (Kuo and Veronis, 1973.)

Although the Stommel-Arons (1960b) model itself is not a theory for the actual mean deep circulation, the model does provide a dynamical framework in which to think about aspects of the circulation. Evidence bearing on the verification of elements of the model as applied to the real ocean is discussed in section 1.5.

1.4 Sources of Deep Water

Even though the individual sites where water sinks to great depth do not seem so important for driving the overall deep circulation as they once did, it is desirable to know the rates at which sinking occurs, because the compensating upwelling from the rest of the deep water is thought to force its mean horizontal circulation. Moreover, the climatic conditions at the specific sinking regions determine the characteristics of the descending water, and thereby the properties and layering of deep water through the world ocean. The sinking phenomena are an integral part of the deep circulation as well, and the various processes by which near-surface water is brought to depth are of considerable interest in themselves.

As noted in section 1.2, in the southern hemisphere deep sinking is limited to the waters around Antarctica. In the northern hemisphere it occurs only in the northern North Atlantic, but there are also outflows from marginal seas: the Mediterranean Sea, the Red Sea, and the Persian Gulf. These outflows descend principally to mid-depths, but affect the water characteristics at deeper levels too.

1.4.1 Sinking around Antarctica

The only location in the Southern Ocean where sinking to the bottom through convective overturning has been identified is the Bransfield Strait (figure 1.3), which separates the northern tip of the Antarctic Peninsula from the South Shetland Islands (Clowes, 1934; Gordon and Nowlin, 1978). There a somewhat isolated trough of depth 1100-2800 m is filled with a mass of nearly homogeneous water very different from the surrounding Circumpolar Deep Water: much higher oxygen concentration, and notably lower temperature, salinity, and nutrient concentrations. Values of tritium concentration, increasing toward the bottom below 300 m in 1975, are further evidence for recent contact between the bottom water and the sea surface. Although no observations have been made there in the winter season, these circumstances strongly imply that local winter convection renews the water column through its full depth. So far as one can tell from the property distributions, however, none of this bottom water flows out of the trough into the rest of the ocean, probably because of the topographic barriers.

Reduced vertical stratification northeast of the Bransfield Strait, in a zone separating the Scotia and Weddell Seas, raises the possibility of some deep convection there too (G. E. R. Deacon and Moorey, 1975), but the evidence to date is less clear-cut than that from the Bransfield Strait.

The bottom water that does enter the rest of the ocean originates in several areas of the Antarctic continental shelf, where water is made sufficiently cold and saline that, in flowing down the continental slope and mixing with the surrounding deep water, it is dense enough to reach the floor of the ocean. In order of decreasing amount and extent of influence on the deep-water property distributions—and presumably, therefore, of rate of bottom-water production—these regions are the Weddell Sea, the Ross Sea, and the Adélie Coast (figure 1.3; A. L. Gordon, 1974); perhaps there is some production off Enderby Land too.

Because of its predominant role, the Weddell Sea has attracted the most attention, although the extreme harshness of working conditions there has discouraged

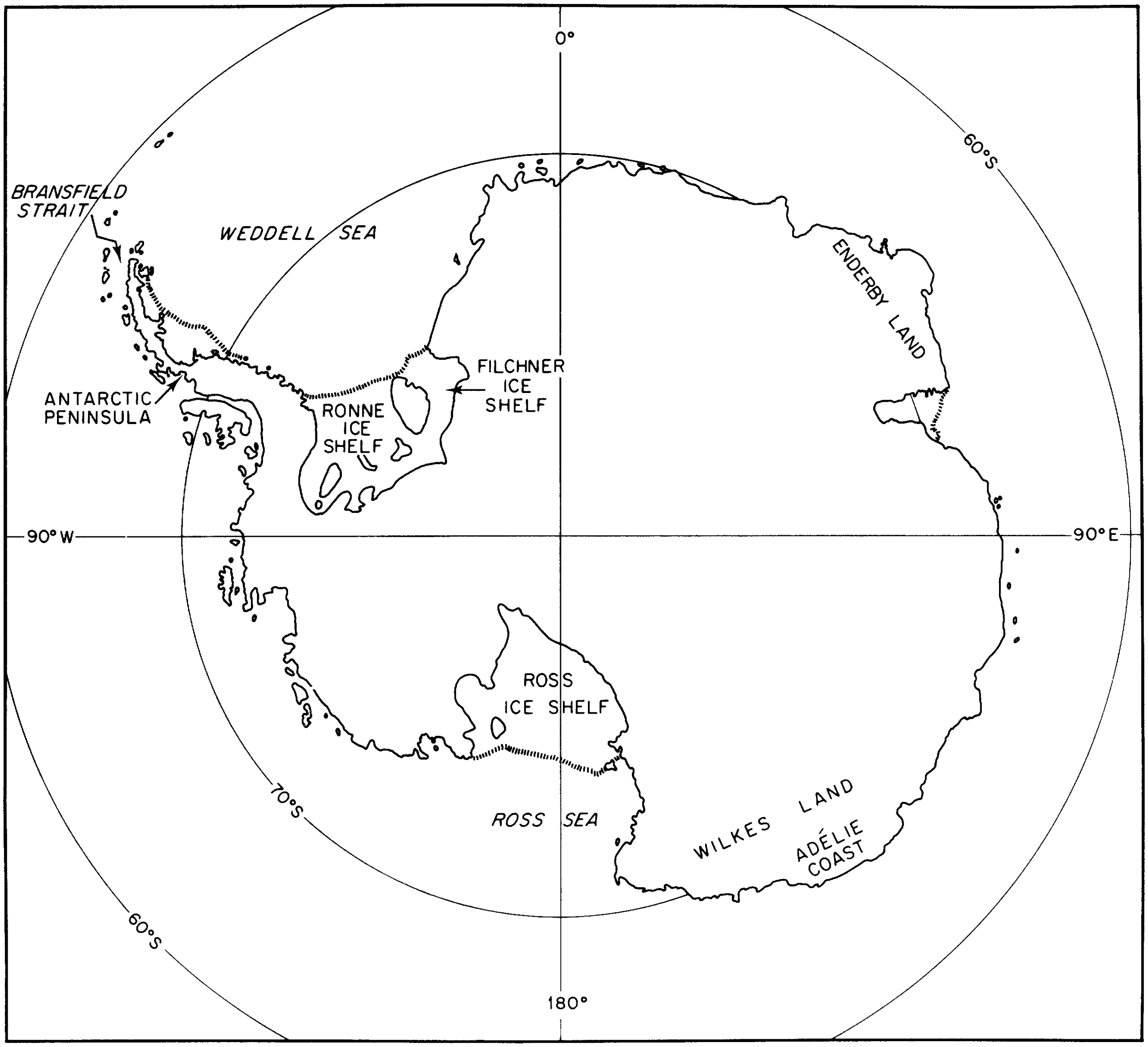


Figure 1.3 Index map identifying Antarctic place names mentioned in text.

any wintertime observations since the *Deutschland* was frozen into the pack ice in 1912 (Brennecke, 1921). Nevertheless, summer data suggest that flow down the continental slope takes place throughout the year. Sections of temperature, salinity, and dissolved-silica concentration, as constructed from observations made in 1968 on a station line running eastward from the Antarctic Peninsula (figure 1.5; see figure 1.4 for station positions) show a 200-m thick layer of relatively cold, fresh, low-silica bottom water on the slope, extending onto the floor of the Weddell Sea (Carmack, 1973). Although the station line did not reach the continental shelf (ice cover has prevented any shelf-water observations in that region), the evidence points plainly to a flow of shelf water down the slope, probably quite obliquely, entraining, and being diluted by, the surrounding water along the way. Similar sections near longs. 40°W and 29°W (Foster and Carmack, 1976a), as well as 10-20°W (Carmack and Foster, 1975), show diminishing evidence for downslope flow to the eastward, and the distribution of bottom potential temperature in the Weddell Sea (figure 1.6; Foster and Carmack, 1976a) demonstrates that the newly formed bottom water leaves the continental slope mainly at the northern tip of the Antarctic Peninsula, in lats. 63-65°S.

This Weddell Sea Bottom Water—defined by Carmack and Foster (1975) as of potential temperature $\theta < -0.7$°C—is widespread in the Weddell Sea, but since

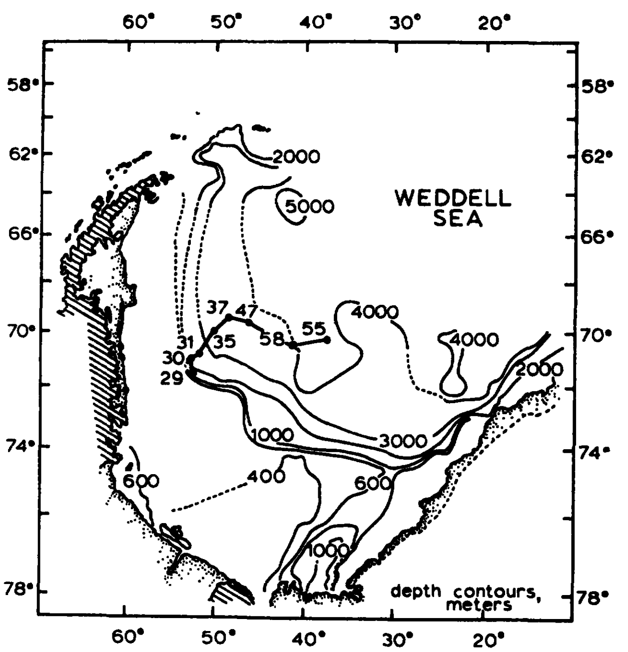
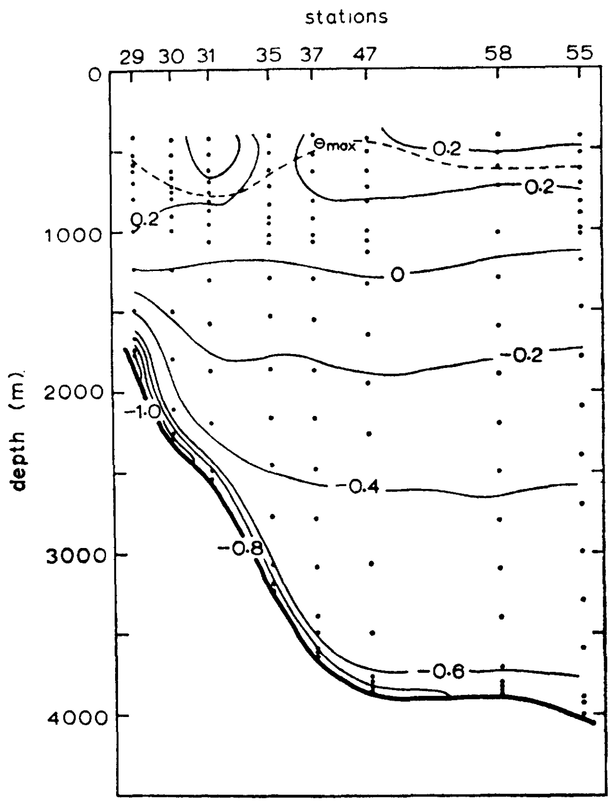


Figure 1.4 Positions of stations occupied by U.S.C.G.C. *Glacier* in the Weddell Sea, 26 February-13 March 1968, that are used for the construction of the sections in figure 1.5. Isobaths labeled in meters. (After Carmack, 1973.)

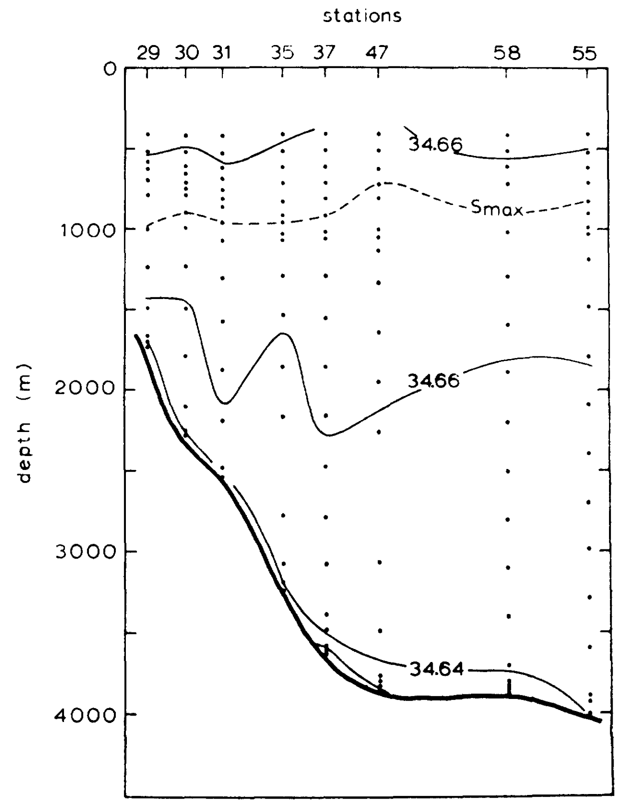
water of such extreme characteristics is not found to the north in the Argentine Basin (Reid, Nowlin, and Patzert, 1977), it must be mixed fairly rapidly into the water above. That overlying water in the Weddell Sea is itself also "Antarctic Bottom Water," but "older" water, whose properties have been modified by mixing elsewhere; it flows westward into the Weddell Sea close to the continent in the clockwise Weddell gyre (e.g., Brennecke, 1918; G. E. R. Deacon, 1976).

By combining station data with current records of duration between 3 and 4 weeks, Carmack and Foster (1975) estimated that the net flow of Weddell Sea Bottom Water out of the Weddell Sea is about $16 \times 10^6\ m^3 s^{-1}$. At their stations, that water had a mean potential temperature of -0.786°C, brought about, presumably, by mixing overlying water of $\theta = -0.6$ to -0.7°C into water that initially, at the edge of the continental shelf, was of $\theta = -1.2$ to -1.4°C. They estimated, therefore, that the rate of sinking from the shelf in the Weddell Sea is $2\text{-}5 \times 10^6\ m^3 s^{-1}$.

In accounting for the formation of Weddell Sea Bottom Water, three distinct processes need to be treated: (1) how the density of the shelf water is increased to the point where it can sink through the relatively warm, saline water at mid-depths; (2) how the water moves off the shelf; and (3) how it descends with entrainment to the floor of the Weddell Sea. The first problem has been considered at length by Gill (1973). Although near-surface water in the open Weddell Sea has a temperature close to the freezing point, it is generally too fresh to be dense enough to sink through the water below. On the shelf, however, the salinity can be increased by salt release during ice formation, and there more saline water is, in fact, found, which is dense enough to sink to depth. By itself the annual cycle of freezing and melting at the sea surface cannot account for the greater salinity of the water on the shelf, because whatever salt were added to the water during winter freezing would be mixed back into the melt water during summer. Nor can it be supposed that the salt-enriched water formed during winter immediately leaves the shelf, because it is found during summer both on the shelf and flowing down the continental slope. Seabrooke, Hufford, and Elder (1971) therefore suggested that the salinity enhancement takes place year-round through freezing at the base of the Filchner and Ronne Ice Shelves (figure 1.3). Gill (1973) showed that this was unlikely, however: because the heat flux through the ice shelf is too small to allow enough salt release for a significant rate of bottom-water production; because the evidence is for melting rather than freezing under the ice shelf anyway; and because the temperature of the most saline shelf water is commonly well above the freezing point at the depth of the ice-shelf base. It seems instead that the higher



(1.5A)



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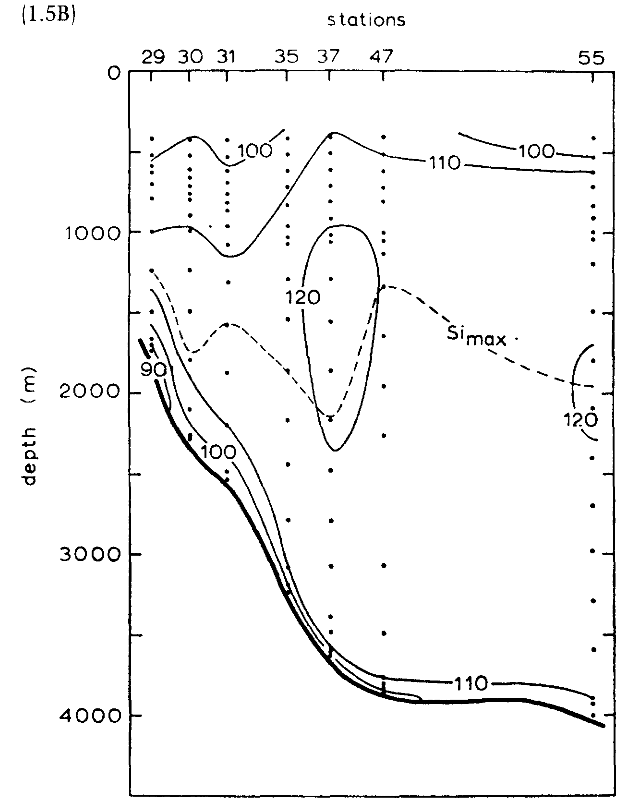


(1.5C)

Figure 1.5 Sections of (A) potential temperature (°C), (B) salinity (‰), and (C) dissolved silica concentration ($\mu M\, l^{-1}$) along a line running eastward from the Antarctic Peninsula (left) in the southwestern Weddell Sea, illustrating the descent of dense shelf water along the Antarctic continental slope. (Carmack, 1973.) See figure 1.4 for positions of stations.

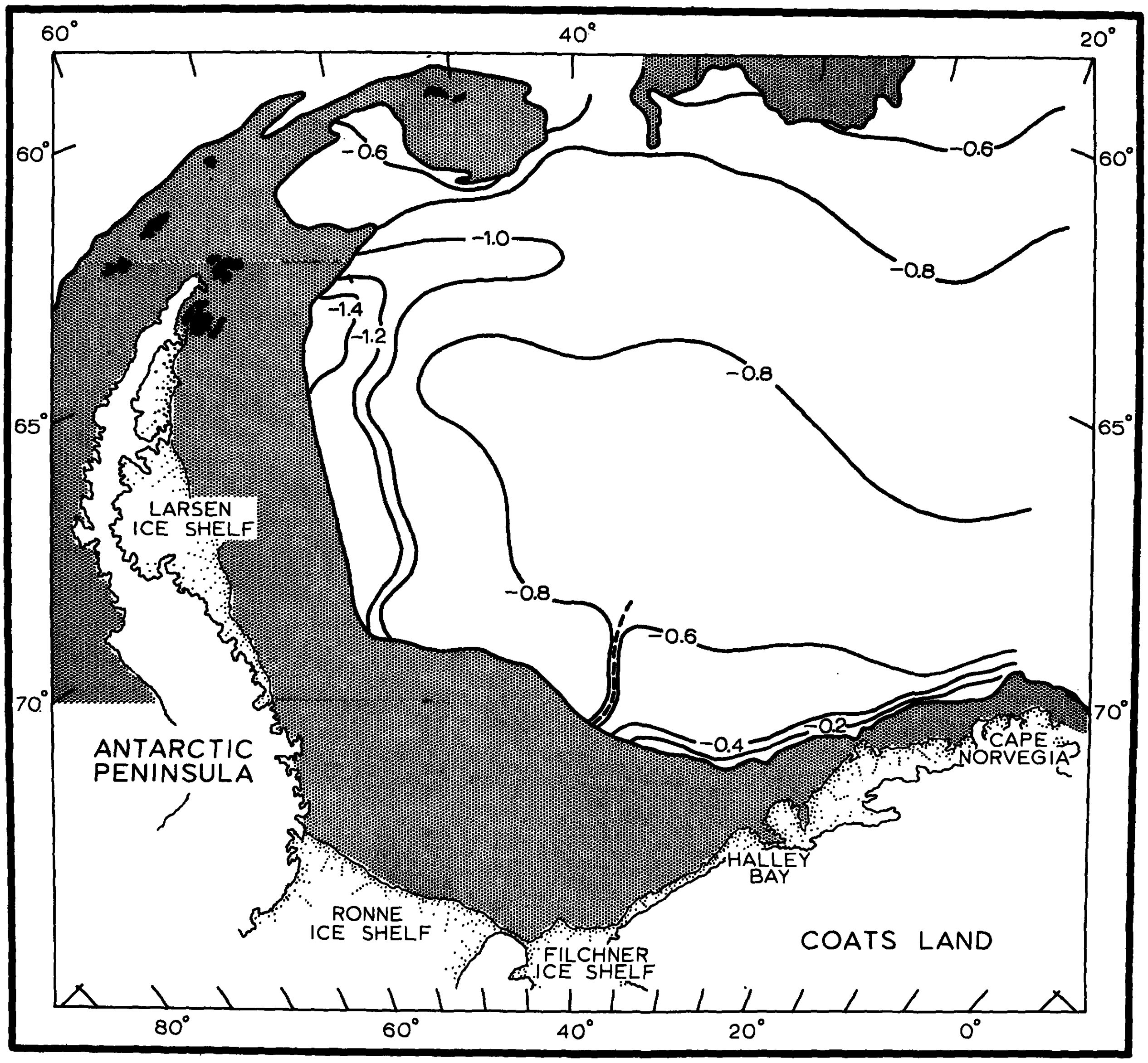


Figure 1.6 Distribution of bottom potential temperatures in the Weddell Sea, illustrating the eastward spreading of newly formed bottom water from the northern tip of the Antarctic Peninsula. (Foster and Carmack, 1976a.)

salinities are indeed due to the seasonal freezing-melting cycle at the sea surface: evidently there is a prevailing offshore movement of the pack ice that transports potential melt water locked up in ice out of the region, thus allowing a net brine production there. That movement also increases the wintertime brine release by opening up leads in the ice field where much more rapid freezing occurs. A plausibly estimated value for the net *effective* rate of ice formation on the continental shelf, through the complete seasonal cycle, is about 1 $m\,yr^{-1}$ (Gill, 1973).[8]

Given denser water on the shelf than offshore, it is still not immediately obvious why it moves toward the continental slope, because even if the density difference were imparted impulsively, the water should only advance offshore a distance of order of the deformation radius (10–20 km), and the ensuing steady motion should be geostrophic flow *along* the isobaths, rather than across them. In zonally symmetric conditions without longshore pressure gradients, the only possible steady efflux of bottom water from the shelf would be in a frictional boundary layer, but the likely transport there is much too small to be realistic (Gill, 1973). Antarctica is far from zonally symmetric, of course, and the existence of large-scale indentations in the coastline, with much broadened continental shelves, as in the Weddell Sea, allows much more substantial transport off the shelf.

In particular, the salinity of the shelf water in the Weddell Sea increases markedly from east to west, by about 0.4‰, and the associated longshore density gradient implies a thermal-wind shear, with surface water moving onshore and the deeper shelf water moving offshore. Combining these density data with a value for the onshore Ekman flux associated with easterly winds, Gill (1973) estimated the offshore transport of dense water along the length of the shelf to be about $1 \times 10^6\ m^3\,s^{-1}$, of which half is compensation for the surface Ekman flux. The net offshore salt transport associated with this vertical circulation is only about half that required to remove the amount of salt added at the surface by net ice formation of 1 $m\,yr^{-1}$. It seems likely, though, that the longshore salinity gradient is itself maintained by a slow westward flow along the shelf—part of the Antarctic coastal current—whereby the salinity of the water is gradually raised to the westward by the widespread freezing there. There are no direct measurements of mean velocities on the shelf, but a longshore flow of $1 \times 10^6\ m^3\,s^{-1}$, carrying salt-enriched water off the shelf at its western end on the Antarctic Peninsula, would be adequate to remove the remaining salt produced during the annual ice-formation cycle (Gill, 1973). These estimates, derived entirely from conditioning on the shelf, thus give a figure for the total flux of dense water off the shelf of about $2 \times 10^6\ m^3\,s^{-1}$, which is in fairly good agreement with the value of $2–5 \times 10^6\ m^3\,s^{-1}$ obtained by Carmack and Foster (1975) from direct measurements of the outflow of Weddell Sea Bottom Water.

The offshore-moving shelf water encounters the waters of the open Weddell Sea in a frontal zone at the edge of the continental shelf. Detailed studies of this zone demonstrate a rather complicated mixing process, involving both the warm deep water of the Weddell Sea and the winter-cooled near-surface water, through which the characteristics of the shelf water are modified to those of the descending Weddell Sea Bottom Water (Foster and Carmack, 1976a).

As noted, there are subtleties as well in the dynamics of the descent on account of the rotational constraint, which tends to orient the flow westward along the isobaths rather than across them. In order to identify the conditions that bring about flow down the slope to the floor of the Weddell Sea, with the amount of entrainment observed, Killworth (1977) considered a series of simplified models of turbulent plumes in a stably stratified environment. Bottom friction was essential, of course, for any downslope flow at all, but the mere fact of shelf water being denser than its surroundings was not in itself sufficient for flow to reach the sea floor with the right entrainment, whether the plume was two-dimensional, three-dimensional, or intermittent. Because the thermal-expansion coefficient for sea water increases with pressure, however, very cold descending water can draw on its own *internal* energy to increase its kinetic energy; thus if a relatively dense falling parcel is sufficiently colder than its surroundings, the buoyancy force acting on it can *increase* during the descent, even though the parcel is falling in stably stratified water. With this effect taken into account, Killworth (1977) was able to construct model plumes (three-dimensional, steady as well as intermittent) that both descended far enough to reach the floor of the Weddell Sea and entrained roughly the observed amount of surrounding water along the way. These had a substantial component of flow along the isobaths too, consistent with the bulk of the bottom water entering onto the sea floor in the northwestern corner of the Weddell Sea.

The Weddell Sea Bottom Water is fresher than the overlying water, but in the southwestern Pacific—specifically, north of the Ross Sea (figure 1.3) to the mid-ocean ridge (about lat. 65°S), and close to Antarctica south of Australia—the salinity increases with depth near the bottom to values >34.72‰ (Gordon, 1971, 1975a; Gordon and Molinelli, 1975). This high-salinity bottom water must have a different source, and it has been traced to the Ross Sea (e.g., A. L. Gordon, 1974), where cold, saline shelf water has been observed to descend the western continental slope in a manner similar to that of the downslope flow in the Weddell Sea (Gordon, 1975a). Although the effect of the Ross Sea bottom water on the deep temperature distribution is barely noticeable, and its effect on the salinity distribution only became clear through the precision afforded by conductivity measurements, its production rate is not necessarily a great deal less than that of Weddell Sea Bottom Water, because the water with which the shelf water mixes to form bottom water is both more saline and much warmer than in the Weddell Sea. No transport measurements like those of Carmack and Foster (1975) have been made, however.

The oceanographic conditions on the Ross Sea continental shelf are roughly similar to those in the Weddell Sea, in that there is westward flow along the shelf, and the salinity of the shelf water increases to the west by 0.3–0.4‰ (Jacobs, Amos, and Bruchhausen, 1970; Jacobs, Gordon, and Ardai, 1979). This correspondence suggests that the mechanisms sketched by Gill (1973) are also operative in the Ross Sea in moving dense water off the shelf onto the continental slope. A vertical circulation (involving the Ekman transport and geostrophic shear) about 60% as large as that on the Weddell Sea shelf is indicated (Gill, 1973); that is, an offshore flux along the length of the Ross Sea shelf of roughly $0.6 \times 10^6\ m^3\,s^{-1}$. No estimate has been made of the rate of removal of shelf water by the longshore flow.

The water deeper than 300–500 m on the western part of the Ross Sea shelf is the most saline (34.8–35.0‰) and the most dense (σ_t = 28.0–28.1) water found in the Antarctic (Jacobs, Amos, and Bruchhausen, 1970; Gordon, 1971). One wonders to what extent it may be involved in the formation of bottom water in the Ross Sea, but no very satisfactory account seems to have been given for the existence of this very saline water. Because the high salinities of the layer are highly correlated with low temperatures, and because its thickness and breadth increase southward to the Ross Ice Shelf, it is tempting to think that the characteristics of the layer may be imparted through freezing at the base of the ice shelf. Near the front of the ice shelf, though, the evidence is of melting at the base (Jacobs, Amos, and Bruchhausen, 1970). Moreover, vertical variations of temperature and salinity observed beneath the ice shelf far to the south at 82°S, 169°W demonstrated melting there too at the time of observation, but because the lowest 6 m of ice was of marine origin rather than terrestrial, it is not certain what the prevailing condition is at that point (Gilmour, 1979; Jacobs, Gordon, and Ardai, 1979). Even if freezing is occurring, the maximum possible rate at the base of the ice shelf is very small (3.5 cm yr^{-1}; Jacobs, Gordon, and Ardai, 1979), and any outflow from beneath the ice shelf consistent with the required salinity enhancement would be much less than the estimated offshore transport of water on the continental shelf. The bulk of this very saline water is found in a large topographic depression on the shelf, whose sill depths near the edge of the shelf are not well known (Jacobs, Amos, and Bruchhausen, 1970); perhaps most of that dense layer is indeed renewed only slowly through freezing under the western Ross Ice Shelf, is trapped in the depression, and does not contribute significantly to the local bottom-water formation (Gordon, 1971).

On the other hand, the high salinity of this western shelf water might be due entirely to annual sea-ice freezing, as proposed by Gill (1973). Unusually dense water formed in that process could contribute substantially to bottom water, or it might form infrequently and merely accumulate in the topographic depression if the sill is actually shallow enough to allow much containment. The situation is unclear.

Several similar deep depressions—though of much smaller area—exist on the continental shelf off the Adélie Coast of Antarctica (figure 1.3). They are filled below their sill depths with water of temperature close to the freezing point, and of salinity 34.4–34.7‰, probably as a result of deep convection associated with winter sea-ice formation (Gordon and Tchernia, 1972). Evidently such water spills over the sills, because it is found as a layer a few tens of meters thick on the nearby continental slope, underlying the high-salinity bottom water from the Ross Sea. This low-salinity bottom water mixes into the Ross Sea water rapidly enough that it has not been detected far from the continental rise near its points of origin. No estimates of the rate of production of the Adélie Coast water have been made, though the rate must be relatively small, given the small volumes of such water that have been found away from the shelf. The flow down the continental slope appears intermittent, inasmuch as duplicated stations in February 1969 and December 1971 showed that the comparatively fresh layer was more widespread at the latter time (Gordon and Tchernia, 1972).

Evidence has also been cited by Jacobs and Georgi (1977) for some small production of bottom water off Enderby Land. They suggest that new source regions are revealed almost with each new exploration of the Antarctic continental slope.

Bottom water from these several sources—the Weddell Sea being the paramount one—mixes with the warmer, more saline water above to form the "Antarctic Bottom Water" of the world ocean. Since all the subtypes originate through freezing at the sea surface, it is the coldest, and thereby the densest, deep water in the open ocean, detectable into northern latitudes by low temperatures close to the bottom.

Not all deep water formed in the Antarctic is bottom water, however. Carmack and Killworth (1978) have described a layer of relatively cold, fresh water at depths of about 2 km off Wilkes Land (figure 1.3; long. 150–170°E), which they demonstrated through Killworth's (1977) plume theory to have originated most probably from shelf water that is not quite dense enough to sink to the bottom. Instead, it sinks to a depth where its density matches that of the surrounding water, in the manner of the Mediterranean and Red Sea outflows. This particular instance of sinking to mid-depths was discernible because the shelf water intruded into saline deep water from the Ross Sea, and hence could be identified by its relatively low salinity. Such phenomena seem possible elsewhere around Antarctica, but in regions where the surrounding deep water is fresher—more like that in the Weddell Sea—they would be far more difficult to detect in distributions of temperature and salinity, though perhaps less so in distributions of nutrients (e.g., silica). As Carmack and Killworth (1978) pointed out, if the salinity of shelf water really is increased almost everywhere around Antarctica by annual freezing, then some process must operate to remove the extra salt from the shelf; that process may be sinking of water to mid-depths, in the manner that was observed off Wilkes Land. If so, then the rate of deep sinking in the Antarctic might be much greater than that involved in the immediate production of bottom water; but there is no basis at present for estimating *how* much greater.

Furthermore, a set of closely spaced observations in the eastern Weddell Sea in February 1977 revealed a column of relatively cold, fresh water some 30 km in diameter, extending from near the surface through the warm, saline deep water to about 4000 m (Gordon, 1978). It suggests a relic of deep convective overturning, such as has been well documented in the northwestern Mediterranean Sea in regions of similarly small area (MEDOC Group, 1970). It may represent yet another mode of deep sinking in the Antarctic, if not to the bottom at least to mid-depths, that had been missed in the past, perhaps because of difficulty of access to critical areas or too coarse a spacing of stations. The specific conditions for overturning in the Weddell Sea have been carefully considered by Killworth (1979), but again there is unsufficient basis at present for estimating how significant a contribution such events may make.

1.4.2 Sinking in the Northern North Atlantic

Next to the Antarctic Bottom Water, the densest water in the open ocean is what Wüst (1935) called Lower North Atlantic Deep Water. It does not originate, however, as Nansen (1912) first proposed, and Wüst (1935) and Sverdrup (Sverdrup, Johnson, and Fleming, 1942) maintained, from wintertime convection to the bottom of the Irminger or Labrador Seas,[9] but from dense Norwegian Sea water that flows into the North Atlantic over three sills on the ridge connecting Greenland and the British Isles (figure 1.7). These overflows entrain resident North Atlantic water in the course of their descent, and join together to form the bottom water of the northern North Atlantic.

It is curious that the significance of these overflows for the large-scale circulation was not appreciated until the late 1950s. Dr. Carpenter (Carpenter, Jeffreys, and Thomson, 1869) had hypothesized long before that overflow east of the Faroe Islands was the source of bottom water in the northern North Atlantic, and Tizard (1883) actually observed some such flow over the Wyville Thomson Ridge, which is the southwestern boundary of the Faroe Bank Channel (figure 1.7). Knudsen (1899) discovered the western overflow, and also found overflow water on the southern side of the Iceland–Faroe Ridge, which he supposed had come south across that ridge; in hindsight, his station positions suggest that this water was really outflow from the Faroe Bank Channel, and had thus derived from the eastern, rather than the central, overflow. In any case, with better station coverage, Nielsen (1904) unambiguously established the existence of the central overflow, and Nansen (1912), with additional observations, discussed all three of them, presenting illustrative sections of water properties across the ridge. Yet somehow all four observers, with the possible exception of Nielsen, believed that the overflows could make only a minor contribution to the deep water of the North Atlantic—an opinion also expressed by Wüst (1935). Although Jacobsen (1916), Brennecke (1921), and Defant (1938) all briefly suggested somewhat greater significance for the overflows, they were generally ignored until Cooper (1952, 1955a,b) and Dietrich (1956, 1957a,b), more or less independently, recalled them to attention.[10] [Evidence for overflow was also noted by Vinogradova, Kislyakov, Litvin, and Ponomarenko (1959) in new data obtained at this time, but they, like the writers at the turn of the century, did not recognize its large-scale significance.]

The deepest passage across the ridge is the Faroe Bank Channel, which has a sill depth of about 800 m. Evidently some small amounts of Norweigan Sea water in this passage do spill intermittently over the Wyville Thomson Ridge (Ellett and Roberts, 1973), but most of this overflow continues northwestward through the channel to its sill, where the measured shear of geostrophic velocity in combination with analysis of water characteristics indicates an outflow of Norwegian Sea water of roughly 1×10^6 m^3 s^{-1} (Crease, 1965). After exiting from the Faroe Bank Channel, this current joins a second overflow, which passes over the ridge between the Faroe Islands and Iceland, where the sill depth is

300-400 m (figure 1.7).[11] By referencing geostrophic velocity calculations for sections southeast of Iceland according to the water characteristics, Steele, Barrett, and Worthington (1962) estimated the volume transport of this combined Iceland-Scotland overflow, after it has left the ridge, to be about 5×10^6 m^3 s^{-1}. Analysis of that overflow water in terms of its component water types showed that about 3×10^6 m^3 s^{-1} of the net flow is entrained Atlantic water, leaving 1×10^6 m^3 s^{-1} for the transport of "pure" Norwegian Sea water across the Iceland-Faroe sill (Worthington, 1970). With different data, Hermann (1967) also estimated roughly 2×10^6 m^3 s^{-1} for the net outflow of Norwegian Sea water between Iceland and Scotland.

This overflow continues southward against the eastern side of the Mid-Atlantic Ridge, and passes through the Gibbs Fracture Zone near lat. 53°N into the Labrador Basin (figure 1.7). Its westward volume transport there was calculated by Worthington and Volkmann (1965)—on the same basis as by Steele, Barrett, and Worthington (1962) farther north—to be still about 5×10^6 m^3 s^{-1}, suggesting little direct flow into low latitudes of the eastern North Atlantic, although Lee and Ellett (1965) have detected overflow influence on water characteristics as far south as lat. 47°N in the eastern Atlantic. This current then flows northward in the eastern Labrador Basin, and, south of Greenland, joins the third overflow from the Norwegian Sea. That overflow comes through the Denmark Strait (Dietrich, 1957b; Mann, 1969), where the sill depth is about 600 m, and is easily identified as cold water banked against Greenland (figure 1.8).

Geostrophic velocity calculations, referenced by water characteristics as for the Iceland-Scotland overflow, gave a figure of about 10×10^6 m^3 s^{-1} for the volume transport of the full overflow current south of Greenland (Swallow and Worthington, 1969), thereby indicating a contribution of 5×10^6 m^3 s^{-1} from the

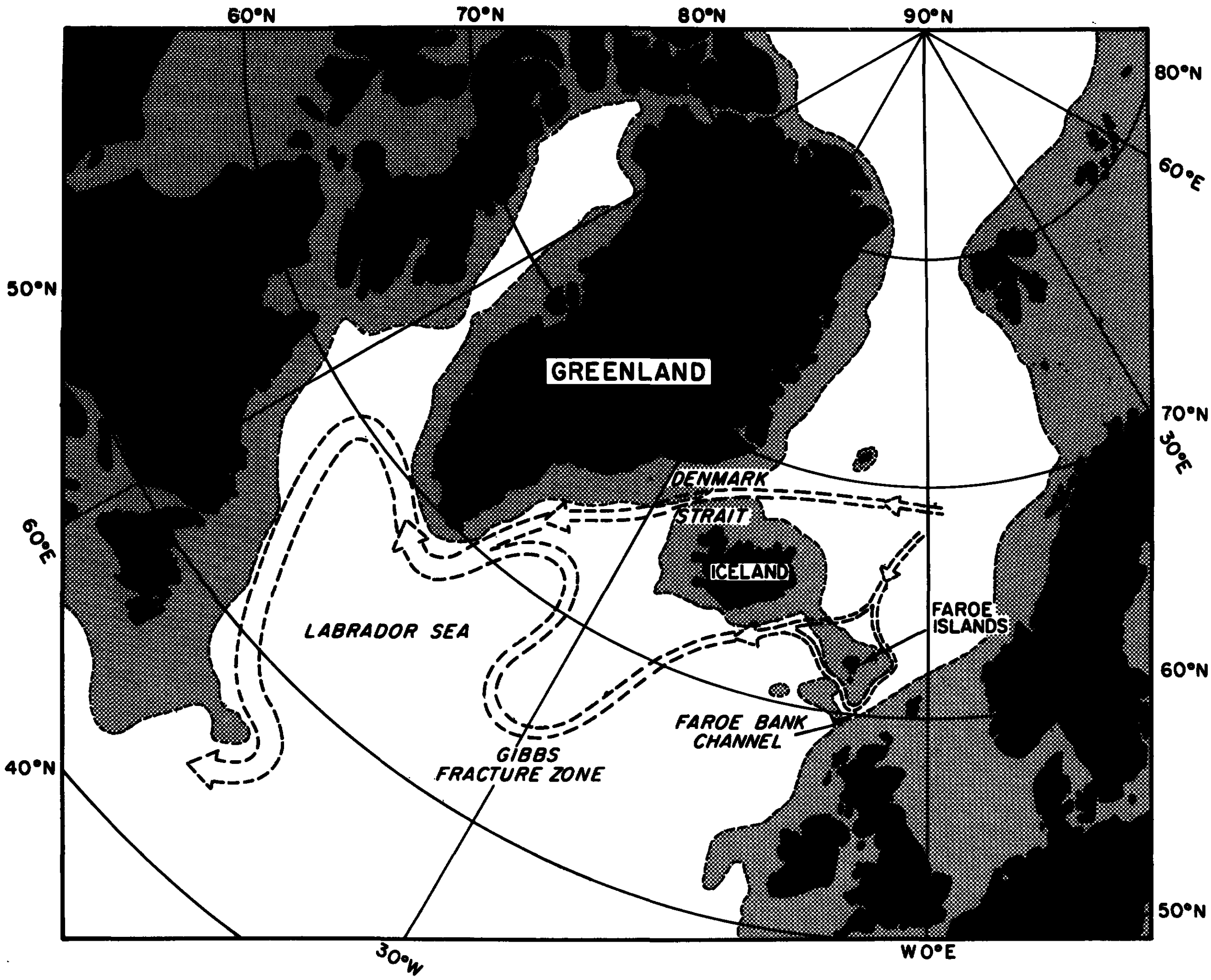


Figure 1.7 Index map identifying the overflow currents from the Norwegian Sea and local place names mentioned in the text. (After Worthington, 1970.)

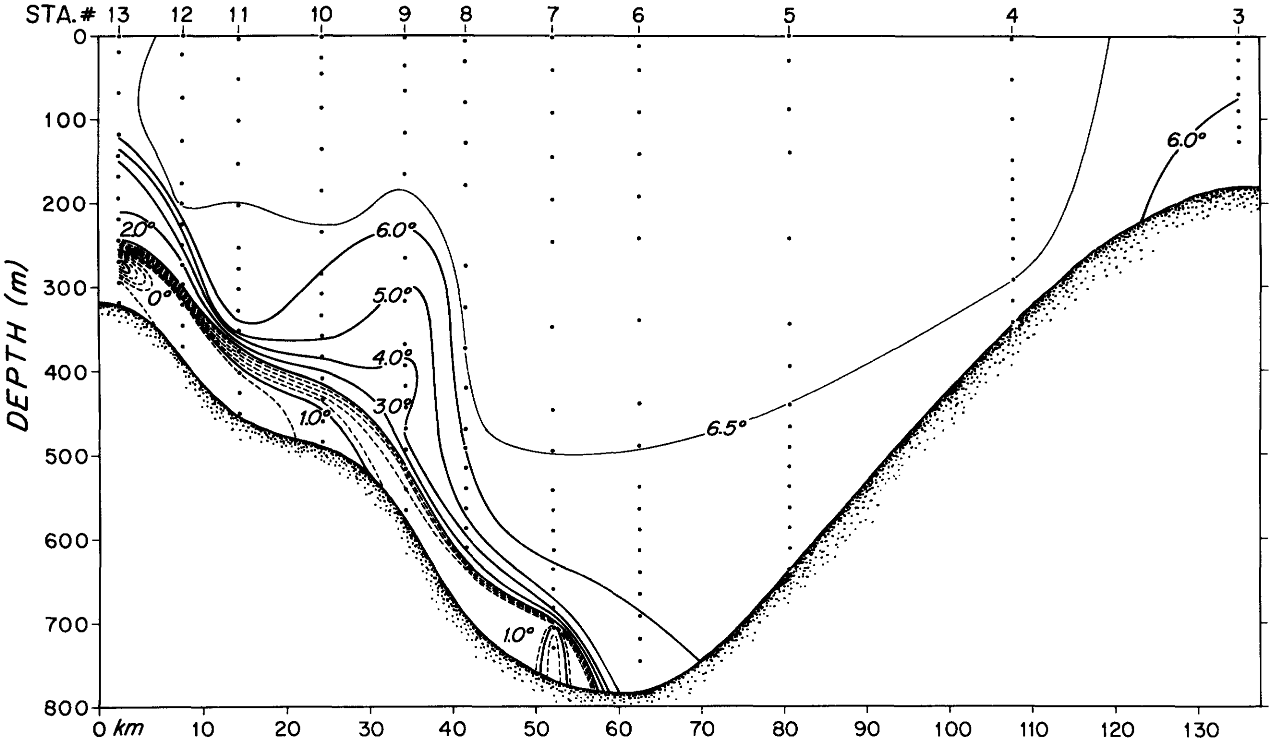


Figure 1.8 Temperature (°C) section across the Denmark Strait in lats. 65–66°N (left, west; see figure 1.7), illustrating the southward flow of cold water from the Norwegian Sea. C.S.S. *Hudson* cruise BI 0267, stations 3–13, 28–29 January 1967. (Worthington, 1969.)

Denmark Strait overflow. Worthington (1970) has estimated that about one-fifth of this is Atlantic water entrained by the overflow as it descends the Greenland continental slope, but the irregularity of the overflow is so great that the fraction may generally be much larger. For example, from a series of hydrographic sections reported by Mann (1969), Smith (1975) calculated an overflow transport of $1.3 \times 10^6\ m^3 s^{-1}$ near the sill, which increased fourfold toward the southern tip of Greenland. Even though the overflows on either side of Iceland originate from the same Norwegian Sea water, the Denmark Strait overflow is distinctly colder and fresher than the Iceland–Scotland overflow (potential temperature 0.0–2.0°C and salinity 34.88–34.93‰, as contrasted with 1.8–3.0°C and 34.98–35.03‰; Worthington, 1976), mainly because the upper-kilometer Atlantic water that they entrain is warmer and more saline to the east.

Although the gross dimensions, course, and entrainment of the descending Denmark Strait overflow can be fairly well rationalized in terms of a steady turbulent plume model (Smith, 1975), it is, in fact, a highly variable flow: its water properties differ markedly from one time to another (e.g., Stefánsson, 1968; Lachenbruch and Marshall, 1968; Mann, 1969), and month-long near-bottom records of velocity and temperature (not taken, it was thought, from the core of the current) showed that overflow was occurring in bursts 1 or 2 days long, with speeds up to 140 cm s^{-1}, separated by intervals of 1 to several days (Worthington, 1969). Variability in water characteristics suggests that the Iceland–Scotland overflow is equally unsteady (Tait, 1967; Lee, 1967).

The process by which dense water is supplied to these overflows has not been investigated in as much detail as the formation of Antarctic Bottom Water. The Norwegian Sea appears to be an area of large mean-annual heat loss to the atmosphere, however, and Worthington (1970) worked out a heat budget for it in which he characterized it as a mediterranean basin where relatively light incoming water is made dense by strong local winter cooling, then sinks, and runs out beneath the inflow.[12] The overflow water is not *deep* Norwegian Sea water, which is colder and more dense still; tritium concentrations show that the overflow water comes from depths no greater than 1000 m, and probably from levels still shallower in the pycnocline (Peterson and Rooth, 1976). The outflow process thus seems to be separate from that of deep-water renewal; perhaps the overflow phenomenon is akin to the shallow buoyancy-driven flow over a sill modeled by O. M. Phillips (1966a) for the Red Sea. As with the saline water on the Antarctic continental shelf, the rotational constraint would tend to confine newly formed dense water to the Norwegian Sea, so that it would be expected to leak out throughout the year(s), rather than just when cooling was most intense.

The other location in the northern North Atlantic where deep sinking occurs is the Labrador Sea. North of roughly lat. 40°N there is a distinct salinity minimum at a potential temperature of about 3.5°C (Worthington and Metcalf, 1961), which grows more pronounced with approach to the Labrador Sea, where nearly homogeneous water (temperature $\approx$3.5°C, salinity $\approx$34.9‰) is found in winter from the sea surface to about 1500 m (e.g., Worthington and Wright, 1970). The salinity minimum is associated with an oxygen maximum, which was Wüst's (1935) diagnostic for his Middle North Atlantic Deep Water, and these characteristics must be imparted in the Labrador Sea, presumably through deep convection driven by intense winter cooling at the surface. Despite extensive wintertime surveys in the area, however, no well-documented deep-overturning events have yet been reported, but Lazier (1973a) has found a few examples in the records of Weather Ship *Bravo* (56°30'N, 51°00'W) of deep isopycnals rising to the surface layer for brief intervals of time, in association with virtually homogeneous water columns 1500 m deep. Most likely, the convection occurs like that in the northwestern Mediterranean Sea (MEDOC Group, 1970), in small patches and very intermittently (perhaps not even every winter), so that the convection could escape notice unless the observations were frequent and closely spaced.[13]

Because of these small space and time scales, it is difficult to make much more than a guess at the long-term mean rate of deep sinking in the Labrador Sea. From an estimate of the net annual heat flux across the sea surface, Wright (1972) suggested $3.5 \times 10^6\ m^3\,s^{-1}$.

Both the Labrador Sea water and the overflow water are more saline at their sources than the Antarctic Bottom Water, and their salinities are increased somewhat in middle latitudes of the North Atlantic through mixing with the outflow from the Mediterranean Sea (see below, this section). Although they are both made dense by cooling, their temperatures are not reduced to the freezing point before sinking, but remain high enough that, despite the salinity difference, they are less dense *in situ* than the Antarctic Bottom Water, and therefore override it.[14] This high-salinity water spreads throughout the South Atlantic, is carried eastward around Antarctica in the Circumpolar Current, and its elevated salinities can be followed into low latitudes of the South Indian and South Pacific Oceans (Reid and Lynn, 1971). The effect of North Atlantic Deep Water in maintaining high oxygen concentrations in the deep ocean is familiar, but its source waters are also exceptionally low in dissolved silica: $<8\ \mu M\,l^{-1}$ in the Norwegian Sea overflow (Stefánsson, 1968), roughly $10\ \mu M\,l^{-1}$ in the Labrador Sea water (Mann, Coote, and Garner, 1973), and about $6\ \mu M\,l^{-1}$ in the Mediterranean outflow (Schink, 1967). Consequently, the deep North Atlantic has the lowest silica concentrations of all the oceans (Metcalf, 1969), there is a pronounced deep silica minimum in the South Atlantic between the Antarctic Intermediate and Bottom Waters (Mann, Coote, and Garner, 1973), and there is even a weak silica minimum in the deep southwestern Pacific due to the influence of North Atlantic source waters (Warren, 1973).

1.4.3 Outflows from Marginal Seas

Excess evaporation makes water in the Mediterranean Sea, the Red Sea, and the Persian Gulf more saline and thereby more dense than that in the outlying ocean, and circulations are set up in the shallow connecting straits in which near-surface water flows in, and dense, saline water flows out beneath. The outflows do not fall to the floor of the open ocean, however, because in descending the continental slopes they entrain enough light thermocline water to become less dense than the deep water of the open ocean, and they therefore spread out at mid-depths. They are thus not important sources for the *deep* circulation, but through vertical mixing they do at least influence the properties of the deep water; hence they are noted briefly here.

The salinity maximum marking the core of the Mediterranean outflow [Wüst's (1935) Upper North Atlantic Deep Water] is found at a depth of about 1200 m in the southeastern North Atlantic (e.g., Fuglister, 1960), but relatively high salinities resulting from the outflow can be detected at least as deep as 3000 m (Worthington and Wright, 1970), and the effect of the outflow in raising the salinity of the Labrador Sea water and Norwegian Sea overflow water was remarked above. The rate of outflow through the Straits of Gibraltar has been estimated from the overall salt budget for the Mediterranean Sea and by current measurements of duration a few days to 2 weeks; both methods agree on a mean value of about $1 \times 10^6\ m^3\,s^{-1}$ (Lacombe, 1971). Calculations by Smith (1975) show that by the time the outflow current completes its descent and departs from the continental slope, its volume transport has increased, by entrainment of Atlantic water, to roughly $10 \times 10^6\ m^3\,s^{-1}$. To what extent any of this transport contributes to the flow of the deep water below—as distinct from its properties—is not known.

The salinity maximum associated with the Red Sea outflow occurs at depths of 500–600 m in the northern Arabian Sea, but relatively high salinities, due presumably to mixing with the outflow water, are found there at least as deep as 2500 m (Wyrtki, 1971). A salt-budget calculation based on climatological data cited by Siedler (1968) gives a mean outflow through Bab-el-Mandeb of $0.4 \times 10^6\ m^3\,s^{-1}$, which is the same value that he derived from direct current measurements there of 2.5

days duration. Such close agreement was probably fortuitous, however, because his directly measured inflow value was 40% greater. The entrainment during the descent into the Gulf of Aden has not been estimated.

The Persian Gulf outflow (salinity maximum at depths of 300-400 m in the Gulf of Oman; Wyrtki, 1971) has a small, patchy effect on intermediate-depth property fields in the North Indian Ocean (Rochford, 1964), but any influence on the deeper distributions seems to be masked by that of the Red Sea outflow (Wyrtki, 1971). It is probably only slight, however, because the outflow through the Strait of Hormuz is very small: only $0.1 \times 10^6\ m^3\ s^{-1}$ according to the salt budget (Koske, 1972).

1.5 Deep Western Boundary Currents in the World Ocean

Many more hydrographic stations have been occupied since Sverdrup wrote chapter XV of *The Oceans.* They have filled in blank areas in the world-ocean coverage, added details to property distributions, and sharpened the definition of features, but our broad picture of the property fields has not changed qualitatively. What *has* changed greatly is the conception of their basis: that the underlying mean deep circulation, largely masked by lateral mixing of the water characteristics, is a system of western boundary currents and poleward interior flows, as in the dynamically consistent models of Stommel and Arons (1960a,b)—rather than the Merzian system of basin-wide, slow meridional flows from source regions.

Although this conception makes good sense, it is not so easy to tell from observation how correct it is. The key element in the circulation theory is the supposed general upward movement of deep water, but the hypothesized mean rate ($\sim 10^{-5}$ cm s^{-1}) is so small as to be utterly beyond present measurement capability. Probably measurement of the deep interior flow, to see whether it is directed poleward in vertical average, *is* technically feasible, but certainly not economically and socially so, given the immensity of the current-meter program that would be required. Large-scale tracer fields are not of much help in this respect since mixing effects can make them ambiguous with regard to circulation patterns for the small mean speeds ($\sim 10^{-2}$ cm s^{-1}) of the interior flow (section 1.3; Kuo and Veronis, 1973).

What *can* be observed, through current measurements and short lines of hydrographic stations, are the narrow, relatively swift, deep western boundary currents predicted by the circulation theory. To be sure, some other theory, with different forcing, might also require such boundary currents, but they are a necessary element of the Stommel-Arons dynamics, and demonstrations of their existence in much of the world ocean have contributed strong support—indeed, the primary observational support—for its essential ideas. This section will therefore assess the evidence for deep western boundary currents in the world ocean, with comment on a few related features of the circulation.

1.5.1 Atlantic Ocean

There are two deep boundary currents in the western Atlantic, the one flowing southward from the Norwegian and Labrador Seas, and the other flowing northward from the Antarctic [the latter not included in the barotropic model of Stommel and Arons (1960b) north of the counterpart of the Antarctic Peninsula].

After the Norwegian Sea overflow current passes the southern tip of Greenland, it takes a counterclockwise course around the Labrador Sea (figure 1.7) and flows southeastward along the North American continental slope by Labrador and Newfoundland (Swallow and Worthington, 1969). Figure 1.9, for example, is a temperature section running from southern Labrador northeast to the southern tip of Greenland. The cold overflow water may be seen pressed up against the lower Greenland slope, and the associated density gradient (mainly determined by temperature there) is consistent geostrophically with northwestward flow increasing in strength from mid-depths to the bottom. Adjacent to the North American continental slope, the southeastward-flowing deep boundary current is evident in similar fashion, 200-300 km wide, probably including Labrador Sea water above the overflow water. Three neutrally buoyant floats tracked nearby at depths of 1600-2400 m all registered southeastward flow and, in combination with station data, suggested extension of such flow up to about 1200 m. Though the measurements were of insufficient duration (12-31 hours) for very satisfactory results, nevertheless the indicated volume transport for the deep boundary current was roughly $10 \times 10^6\ m^3\ s^{-1}$ (Swallow and Worthington, 1969).

High-oxygen and low-silica values close to the continental slope in a similar depth interval show that this current continues around the Grand Banks of Newfoundland, at least in the long-term mean, but records from moored current meters in the area have not yet given clear, direct evidence for such flow (Clarke and Reiniger, 1973; Clarke, Hill, Reiniger, and Warren, 1980).

Westward and southward in the North Atlantic, pronounced tracer characteristics marking the deep boundary current are lost, and its cross-stream density gradient is not readily distinguishable from that of the nearby Gulf Stream; hence evidence for its existence comes mainly from velocity measurements. On the continental slope north of the Gulf Stream near long.

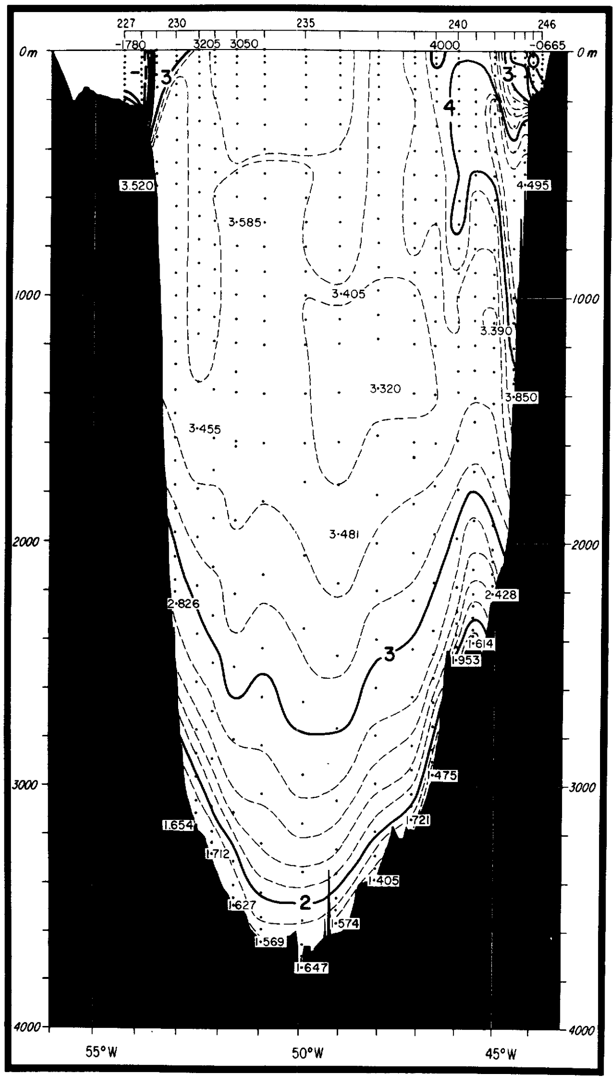


Figure 1.9 Temperature (°C) section along a line running northeastward across the Labrador Basin from southern Labrador (left) to the southern tip of Greenland (see figure 1.7), illustrating the deep western boundary current of the North Atlantic, flowing southwestward against the continental slope of Labrador. M.S. *Erika Dan* stations 227-246, 21-28 February 1962. (Worthington and Wright, 1970.)

70°W, current records of several months duration from five levels at a single site (Webster, 1969), deep float tracks of duration 1 to 2 days at several locations (Volkmann, 1962), and near-bottom current records of length 7-25 days, also from several locations (Zimmerman, 1971), all suggest a prevailing westward flow, surface to bottom, between the Stream and the continental shelf. What part of that should be considered the deep boundary current is unclear.

Clear evidence for the passage of the deep current under the Gulf Stream near Cape Hatteras (lat. 35°N) has been obtained from float tracks of 1 to 3 days duration (Barrett, 1965), from transport-float sections (Richardson and Knauss, 1971), and from current records of length 3 to 8 weeks (Richardson, 1977). These last were from six current meters moored 100 m above the bottom at depths of 1-4 km on a line crossing the mean axis of the Gulf Stream, and they showed a striking persistence of southwestward flow with speeds typically 10 $\mathrm{cm\,s^{-1}}$. The mean velocities in combination with station data indicated a volume transport for the deep western boundary current of $24 \times 10^6\ \mathrm{m^3\,s^{-1}}$.

The discovery of this current by means of neutrally buoyant floats farther south (lat. 33°N), offshore of the Gulf Stream, was mentioned in section 1.3; the width of the current appeared to be about 100 km there, and its volume transport was computed to be $7 \times 10^6\ \mathrm{m^3\,s^{-1}}$ (Swallow and Worthington, 1961). Using a similar zero-velocity surface to reference geostrophic calculations, Amos, Gordon, and Schneider (1971) estimated the transport of the current at lat. 30°N, near the Blake Bahama Outer Ridge, to be $22 \times 10^6\ \mathrm{m^3\,s^{-1}}$.

Whether the differences in the cited transport estimates imply spatial variation is unclear. Probably they are related more to differences in methods and definitions of the current than to anything else.

Three SOFAR floats at depths of 1500-2000 m in the western Sargasso Sea were observed to drift westward in lats. 28-30°N, and then move rapidly southward (speeds roughly 10 $\mathrm{cm\,s^{-1}}$) along the continental slope (Riser, Freeland, and Rossby, 1978). These tracks are consistent with the idea of a deep boundary current, but not with the specific Stommel-Arons (1960b) model, in which water flows *eastward* from the boundary current into the interior. Their significance for the basic ideas of the model is unclear (see Fofonoff's discussion in chapter 4 of this volume).

Some indications have been found for the continuation of the southward flow to about lat. 23°N, north of Hispaniola (Tucholke, Wright, and Hollister, 1973), but no clear evidence for a boundary current flow is available between there and the equator. [That there is some sort of deep southward movement in this latitude belt has long been recognized, of course, from large-scale property distributions (e.g., Wüst, 1935).]

Deep water from the North Atlantic is easily identified in the South Atlantic by its relatively high salinity at depths exceeding 1500 m, and relatively strong southward flow along the entire western boundary of the South Atlantic is evident from intensification close to South America of the salinity maximum (depth increasing southward from 1600 m to about 2500 m) and its associated temperature inversion (Wüst and Defant, 1936; Fuglister, 1960). The breadth of this zone of increased salinity maximum is 500-1000 km, which, however, is probably greater than the width of the boundary current because of lateral-mixing effects. Clear examples of the North Atlantic tracer characteristics are shown in figure 1.10: sections of potential temperature, salinity, and the concentrations of dissolved oxygen and silica on a line crossing the southern Brazilian Basin, roughly along lat. 30°S (Reid, Nowlin, and Patzert, 1977). The temperature inversion is seen at depths of 1600-1800 m within about 1000 km of South America; the strong salinity maximum and silica minimum are found at about 2000 m within 500 km of the continental slope, and the oxygen maximum lies deeper (2000-3000 m) and extends somewhat farther eastward with high values. The impression is of a concentration of southward flow near South America, from the "top" of the Antarctic Bottom Water at about 3500 m to roughly 1500 m.

Probably the most satisfactory estimates of volume transports for the deep meridional flows in the western South Atlantic are those by Wright (1970). He performed geostrophic calculations for the IGY sections (Fuglister, 1960) by assuming zero meridional velocity at the boundary between North Atlantic Deep Water and Antarctic Bottom Water, as defined by a break in the temperature-depth and salinity-depth curves near 2°C and 34.89‰, found at levels of 3400-4000 m [the same break that had been discovered by Buchanan (see section 1.2), and exploited also for velocity calculations by Wüst (1938)]. As an upper boundary for the North Atlantic Deep Water, he adopted the 1600-m level at 16°N and 8°S and the 1800-m level at 32°S (Wright, personal communication), and he derived thereby a figure of $9 \times 10^6\ m^3\ s^{-1}$ for the southward transport of North Atlantic Deep Water (including both boundary current and interior flow in the western trough).

Beneath the North Atlantic Deep Water, the cold, northward-flowing Antarctic Bottom Water is pressed up against the South American continental slope along the entire length of the South Atlantic, except close to the equator (Wüst and Defant, 1936; Fuglister, 1960). It is indicated in figure 1.10 by the relatively low temperature, salinity, and oxygen concentration and the high silica concentration (cf. figure 1.5) below 3500 m, the extreme values being found close against the western boundary. The isotherms (and isopycnals) slope downward to the east, consistent geostrophically with a northward-flowing current increasing in speed toward the bottom. The width of this current, as defined by the zone of sloping isotherms (Cato 6 stations 7-11), is about 500 km. Wright's (1970) calculations give a volume transport for the current of about $6 \times 10^6\ m^3\ s^{-1}$ in middle latitudes of the South Atlantic, diminishing to about $2 \times 10^6\ m^3\ s^{-1}$ near the equator.

The current continues northward across the equator, but in lats. 8-16°N it is found not beside the western boundary but against the Mid-Atlantic Ridge, which is the *eastern* boundary of the basin (Fuglister, 1960). Why this transposition occurs is not wholly clear, but perhaps it is related to the sharp northward increase in the depth of the western basin there, from about 4200 m at lat. 2°N to about 5500 m at lat. 16°N. The idea is familiar that meridional bottom slope has an effect on ocean circulation analogous to that of variation in the Coriolis parameter; poleward increase in depth tends to counteract the β-effect, and, if the rate of increase is large enough, boundary currents can be shifted from the western to the eastern sides of basins (e.g., Fandry and Leslie, 1972). The transposition should occur if, roughly speaking, the meridional bottom slope $S > H\beta/f$, where H is the thickness of the bottom current. At lat. 10°N, $f = 2.5 \times 10^{-5}\ s^{-1}$ and $\beta = 2.3 \times 10^{-13}\ cm^{-1}\ s^{-1}$; for $H \approx 700$ m at lats. 8°N and 16°N (Fuglister, 1960), the critical value of the slope is 6.4×10^{-4}, which is less than the actual slope of about 8.4×10^{-4}.

Whatever the reason for the transposition, there is no indication of this boundary current on transatlantic sections at and north of lat. 24°N (Fuglister, 1960). Apparently, the water from it spreads out somehow over the floor of the basin (Worthington and Wright, 1970), and such a poleward interior flow would be qualitatively consistent with the level-bottom Stommel-Arons dynamics. Traces of Antarctic Bottom Water have been found in the lowest few hundred meters, for example, south of the Gulf Stream on the 50th meridian (e.g., Clarke *et al.*, 1980), in the vicinity of the Blake-Bahama Outer Ridge near lat. 30°N (Amos, Gordon, and Schneider, 1971), and north of the Bahama Banks and Hispaniola (Tucholke, Wright, and Hollister, 1973). In the latter two instances these traces had fairly definitely been incorporated into the southward-flowing deep western boundary current, suggesting a general counterclockwise spreading and movement of Antarctic Bottom Water in the western North Atlantic.

The coldest water in the eastern trough of the Atlantic (north of the Walvis Ridge) is observed at the equator (potential temperature <1.7°C; Wüst and Defant, 1936; Fuglister, 1960). Minimum temperatures increase gradually both northward and southward, showing that the principal inflow of water below the general crest of the Mid-Atlantic Ridge (3500 m depth, say) occurs from the western Atlantic through the Romanche Fracture Zone, on the equator (e.g., Drygalski, 1904; Wüst, 1933; Metcalf, Heezen, and Stalcup, 1964). No estimate of the rate of this inflow has been made. Silica values on the surface where potential temperature is 2°C increase northward (Metcalf, 1969), as well as southward (Chan, Drummond, Edmond, and Grant, 1977), indicating further that the deep water "ages" poleward.

Many additional fracture zones cut across the Mid-Atlantic Ridge, but the existing evidence suggests that there are only three others through which noticeable deep-water transport may take place. As described in section 1.4, the Iceland-Scotland overflow passes westward into the Labrador Basin through the Gibbs Fracture Zone at lat. 53°N (Steele, Barrett, and Worthington, 1962). Indications of eastward flow through the Vema Fracture Zone near lat. 11°N were shown by Heezen, Gerard, and Tharp (1964), but the flux must be much less than that through the Romanche Fracture Zone, given its much smaller influence on property distributions in the eastern Atlantic. Perhaps there is also some slight westward flow through the Kane Fracture Zone near lat. 24°N (Purdy, Rabinowitz, and Velterop, 1979): high silica concentrations appear to extend westward across the Mid-Atlantic Ridge there (Metcalf, 1969), but the accuracy of the measurements is not entirely certain.

As noted in section 1.2, the Walvis Ridge blocks most direct deep connection between the Antarctic and the eastern trough of the Atlantic, but bottom temperatures nearly as low as on the equator at lat. 32°S in the southwestern corner of the Angola Basin (Fuglister, 1960) undoubtedly signify a small leakage of water from the south across the Walvis Ridge, probably through the Walvis Passage near long. 7°W (Connary and Ewing, 1974).

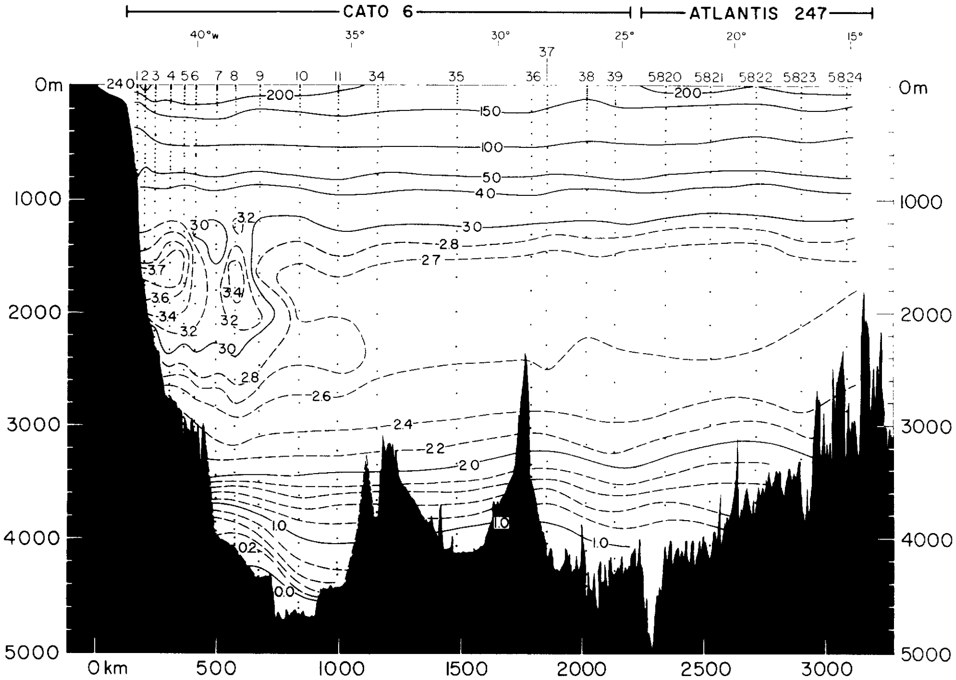
The dynamical fact that poleward movement from an equatorial source can be accomplished by an interior flow may explain to some extent why no deep boundary currents along the western margin of the eastern trough (Mid-Atlantic Ridge) have been observed in transatlantic hydrographic sections (Fuglister, 1960). On the other hand, as pointed out in section 1.3, the budgetary function of western boundary currents is not only to transport deep water into low latitudes, but also to correct imbalances between interior flows and upward fluxes; and it is not easy to see how continuity can be maintained in the eastern trough of the Atlantic without recirculation boundary currents. In the North Pacific and North Indian Oceans, where recirculation boundary currents are essential elements of the Stommel-Arons (1960b) model, convincing evidence for such currents has not been found either (see below, this section); but the observational material there is not really adequate to demonstrate whether or not they exist. In any case, there is no match yet between theory and observation in regard to recirculation currents.

1.5.2 Indian Ocean

The geometry of the Indian Ocean is more complicated than that of the other oceans, because several major ridges divide it into a multiplicity of separate basins (figure 1.11). The course of deep flow is correspondingly more tortuous and divided than in the other oceans, and further removed from the simple idealizations (Stommel and Arons, 1960b).

Since the Mozambique Basin (see figure 1.11 for geographical names) is closed off to the north at depths greater than 2500–3000 m, the main point where deep water enters the western Indian Ocean from the Circumpolar Current is the passage between the Crozet Plateau and the Kerguelan Ridge (Jacobs and Georgi, 1977). The pattern of potential temperature at 4000 m (Wyrtki, 1971) demonstrates equatorward flow on through the Crozet Basin, and this flow has been observed to pass through the Southwest Indian Ridge via the Atlantis II and Melville Fracture Zones (longs. 57°E and 61°E, respectively) into the Madagascar Basin beyond (Warren, 1978). On a broader scale, continuity of flow between the Crozet and Madagascar Basins is supported by the distributions of bottom potential temperature (Kolla, Sullivan, Streeter, and Langseth, 1976) and of clay-mineral fractions in the sediments of the two basins (Kolla, Henderson, and Biscaye, 1976).

This flow extends northward as a relatively narrow, intense, near-bottom current close against Madagascar, the effective western boundary for the deep South Indian Ocean (Warren, 1974). The distribution of temperature along lat. 12°S, from the northern tip of Madagascar to the Central Indian Ridge, is illustrated in figure 1.12. The coldest bottom water is seen pressed against the slope of Madagascar, and below 3600 m the isotherms slope downward to the east in a zone 400–500 km wide, indicative of the breadth of the boundary current. The sign of the associated horizontal density gradient is consistent with a northward geostrophic velocity increasing toward the bottom. Unlike the South Atlantic, water characteristics offer no clear demarcation here between southward flow above and northward flow below, but the volume transport of the current was estimated geostrophically, with reference to 3600 m, to be about 4×10^6 m^3 s^{-1} at this section (Warren, 1974). A comparable estimate, 5×10^6 m^3 s^{-1}, was obtained at a similar section extending eastward from Madagascar along lat. 23°S. (No direct velocity measurements have ever been reported in deep boundary currents of the Indian Ocean.)



CATO 6
ATLANTIS 247
40°W
35°
30°
25°
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0m
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0 km
500
1000
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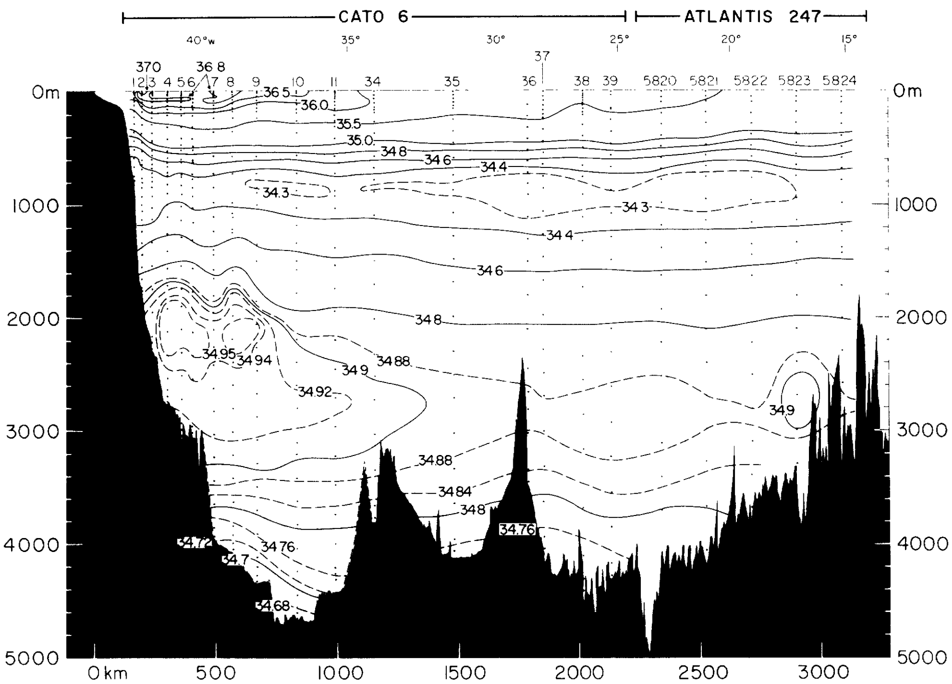
2500
3000

(1.10A)



CATO 6
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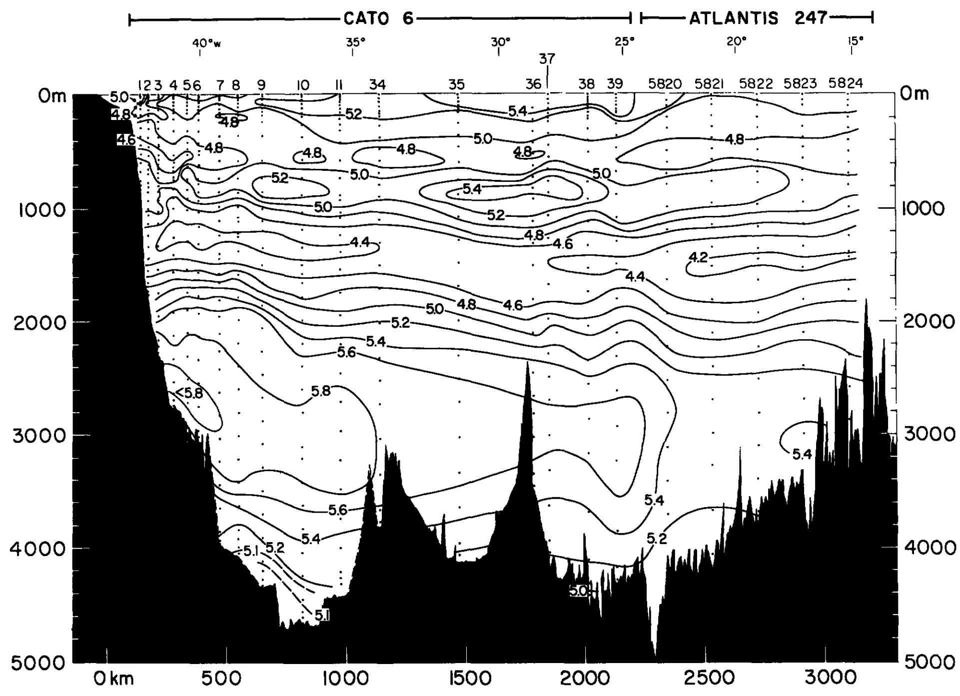
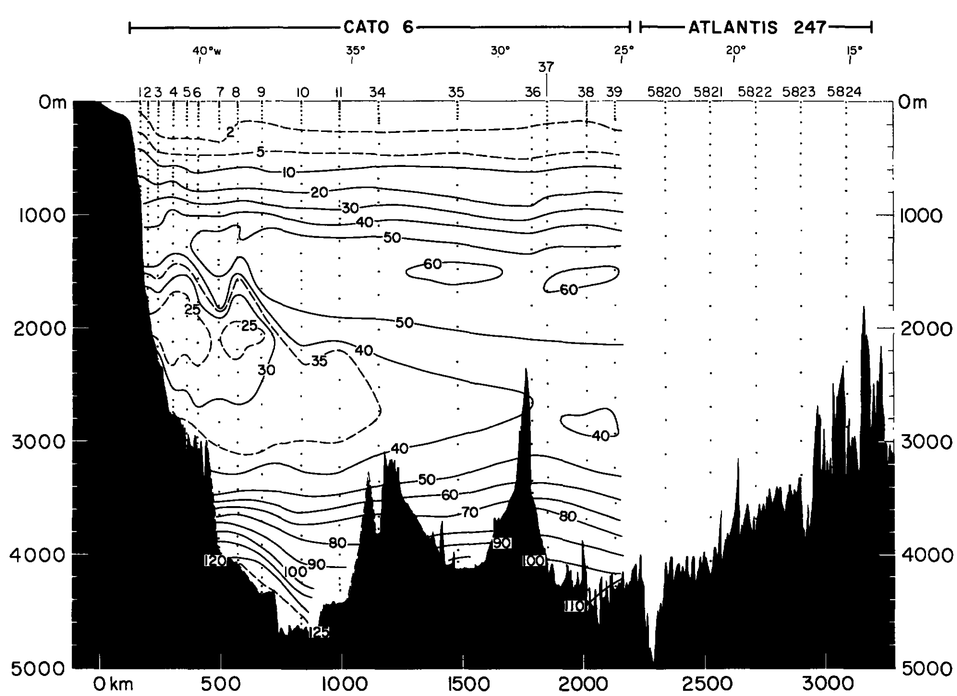
(1.10B)



(1.10C)



(1.10D)

Figure 1.10 Sections of (A) potential temperature (°C), (B) salinity (‰), and the concentrations of (C) dissolved oxygen (ml l^{-1}) and (D) silica ($\mu M l^{-1}$) along roughly lat. 30°S from South America (left) to the Mid-Atlantic Ridge, illustrating the two deep western boundary currents of the South Atlantic, namely, the northward-flowing Antarctic Bottom Water and the southward-flowing North Atlantic Deep Water above. Cato 6 (R.V. *Melville*) stations 1-11, 8-12 November 1972 and stations 34-39, 25-29 November 1972; R.V. *Atlantis* stations 5820-5824, 5-9 May 1959. (Reid, Nowlin, and Patzert, 1977.)

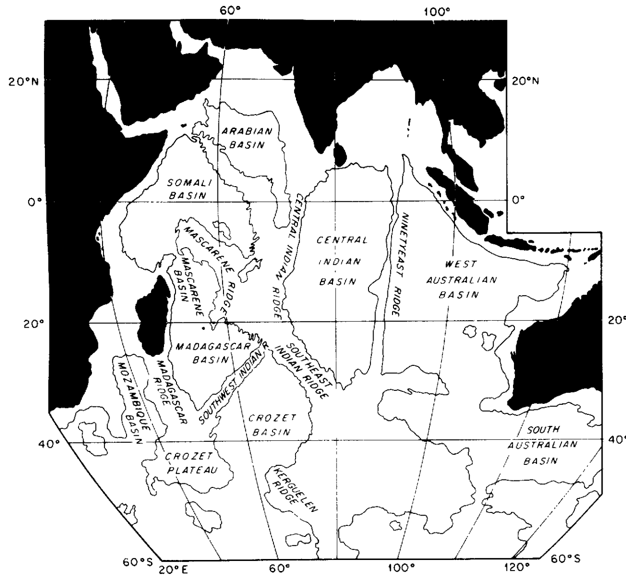


Figure 1.11 Index map identifying basins and ridges in the Indian Ocean, including an approximate representation of the 4-km isobath. (After Wyrtki, 1971.)

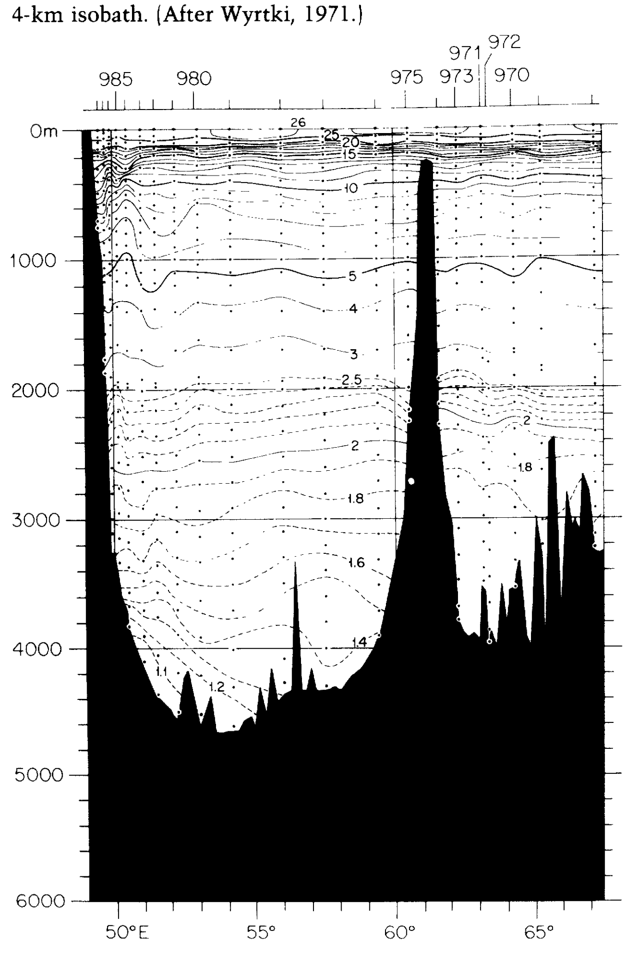


Figure 1.12 Temperature (°C) section along lat. 12°S between Madagascar (left) and the Central Indian Ridge, illustrating the deep western boundary current adjacent to Madagascar in the Mascarene Basin (see figure 1.11). R.V. *Chain* stations 968-988, 20-26 July 1970. (Warren, 1974.)

Evidently, some part of the current continues northward through the Amirante Passage (near 9°S, 52°E) into the Somali Basin (Johnson and Damuth, 1979), but the rate of inflow has not been estimated. During the southwest monsoon of 1964, deep salinity values near the continental slope off Somalia were found to be slightly lower at given potential temperatures than in the central Somali Basin, suggestive of a deep, northward-flowing boundary current (Warren, Stommel, and Swallow, 1966). The evidence was marginal, however; the observations have never been expanded or even repeated, and it is at least conceivable that the flow, if real, was related more to the seasonal Somali Current in the water above than to the global deep circulation.

In the Arabian Basin to the north, nothing is known about the deep circulation, except that the deep water must be renewed relatively slowly, given that its oxygen concentration is the lowest in the deep Indian Ocean (3.6 ml l^{-1} at 4000 m, as contrasted with 4.0-4.2 ml l^{-1} in the Somali Basin; Wyrtki, 1971).

A salinity maximum is found at about 2500 m in the southwestern Indian Ocean, and it is clearly due to North Atlantic Deep Water carried eastward by the Circumpolar Current. The maximum values decrease northward, but north of lat. 15°S, roughly, the salinity at these levels (and isopycnals) *increases* northward into the Arabian Sea (Wyrtki, 1971; see also figure 3.16B in this volume), plainly an effect of the salt source formed by the Red Sea outflow. Mixing thus seems to mask the large-scale field of motion at these levels, and it is uncertain what the sense and strength of the meridional flow are.

The western sequence of basins is separated from the Central Basin and West Australian Basin by the Central Indian Ridge and the Ninetyeast Ridge (figure 1.11). The deep Indian Ocean is open to the Antarctic, however, not only south of the Crozet Basin, but also just west of Australia; and the fact that 4000-m temperatures in the Central Basin and West Australian Basin are 0.2-0.3°C lower than in the Somali Basin and Arabian Basin (Wyrtki, 1971) indicates that the deep water in the former two basins is supplied through the eastern passage, as suggested long since by Schott (1902) and Wüst (1934). Consistent with the Stommel-Arons dynamics, this source water flows northward in the West Australian Basin as a deep western boundary current along the eastern flank of the Ninetyeast Ridge, observed at lat. 18° S (Warren, 1977). The field of specific volume anomaly (figure 1.13) shows a zone of sloping isopycnals at depths of 3000-4500 m, with the densest water against the Ninetyeast Ridge, and the breadth of the zone—indicating the width of the boundary current—being some 500-700 km. At depths of 3000-4000 m there is a slight silica maximum (concentrations >130 μM l^{-1}) in the basin, with values generally increasing toward the west; the maximum layer is separated from the Ninetyeast Ridge by a narrow zone about 30 km wide, however, where the values are lower and the maximum is absent. This distribution suggests: (a) northward flow of low-silica water from the Antarctic within the full boundary-current region deeper than 4 km, but only immediately adjacent to the Ninetyeast Ridge at lesser depths; and (b) southward flow of the high-silica water of the North Indian Ocean (Wyrtki, 1971) elsewhere in the boundary-current region. With reference to a zero-velocity surface constructed on that basis, the northward volume transport of the Ninetyeast Ridge current was estimated geostrophically to be about $4 \times 10^6\ m^3 s^{-1}$ (Warren, 1977), like that of the deep Madagascar current. The course of the current north of lat. 18°S has not been observed.

In the distribution of dissolved-oxgyen concentration calculated by Kuo and Veronis (1973) for their model of the Indian Ocean, the lowest values occurred in the northeastern corner, because that was the region farthest removed, in the sense of combined advection-diffusion time scale, from the oxygen sources in their model: the circumpolar water in the south, and the single boundary current along the western margin of the ocean. The fact noted above that the lowest deep values are actually found in the Arabian Sea becomes intelligible in light of this second deep boundary current, as an oxygen source in the eastern Indian Ocean.

Only very tentative remarks can be made at this time about the deep water in the Central Basin. *Atlantis II* stations 2288-2306, occupied along lat. 18°S, show a temperature-minimum layer centered near 4000 m, most markedly developed toward the east (figure 1.14). This layer is also characterized by a salinity minimum and an oxygen maximum, both of which also decay westward from the Ninetyeast Ridge. Whether there are passages through the southern boundary of the Central Basin that permit exchange of deep water from high latitudes is uncertain, but it seems unlikely that this water with Southern Ocean characteristics derives directly from the south, because the dynamics requires *poleward* interior flow. The properties of the water below 3500 m are very much like those in the Ninetyeast Ridge current at 3500-4000 m on lat. 18°S, however, and it seems possible that the very deep water in the Central Basin is supplied across the Ninetyeast Ridge over known sills of depth 3500-4000 m at lats. 3°S and 10°S (Sclater and Fisher, 1974). Thus at lat. 18°S this water could be circulating in a southward interior flow. On a line of stations occupied by U.S.N.S. *Wilkes* in April 1979 along the western flank of the Ninetyeast Ridge between lat. 12°S and the equator, such an overflow was, in fact, observed near the 10°S sill, but no clear evidence for overflow was found at 3°S. Whether the 10°S

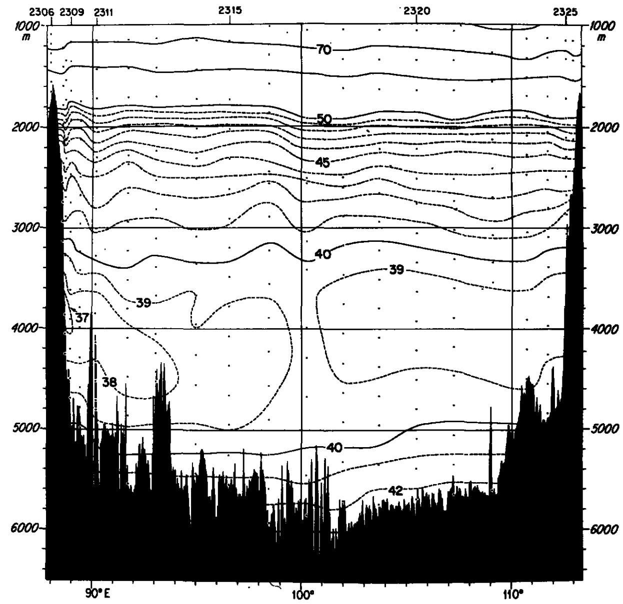


Figure 1.13 Section of specific volume anomaly ($10^5\ cm^3 g^{-1}$) below 1000 m along lat. 18°S between the Ninetyeast Ridge (left) and Australia, illustrating the deep western boundary current of the West Australian Basin (see figure 1.11). R.V. *Atlantis II* stations 2306-2326, 7-17 August 1976. (Johnson and Warren, 1979.)

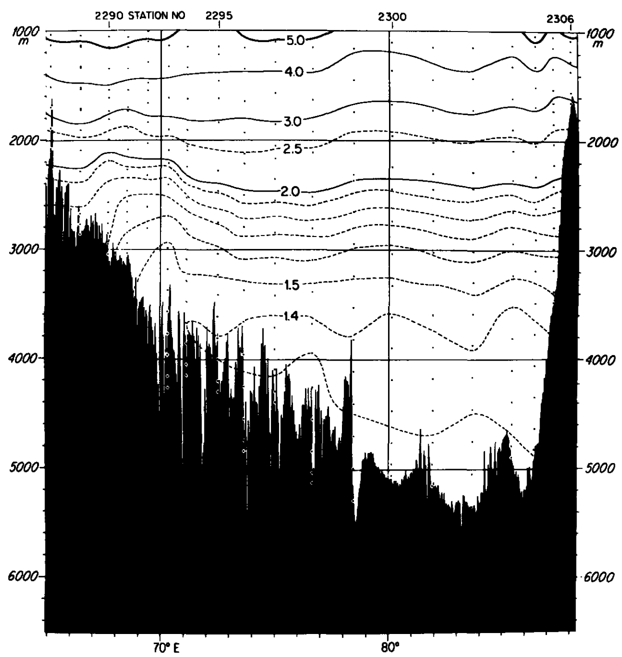


Figure 1.14 Temperature (°C) section across the Central Indian Basin along lat. 18°S from the crest of the Central Indian Ridge (left) to the Ninetyeast Ridge, illustrating the westward increase of temperature below 3500 m and the possible northward flow above the flank of the Central Indian Ridge at depths of 2000-3000 m. R.V. *Atlantis II* stations 2288-2306, 27 July-7 August 1976.

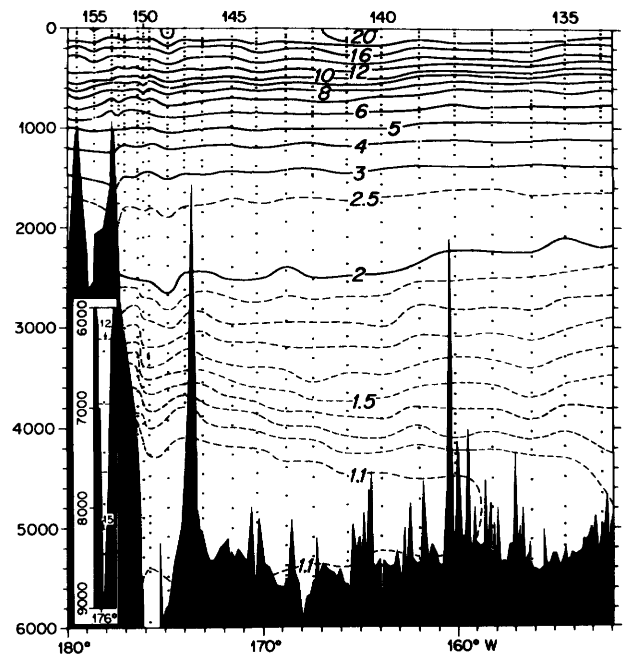
overflow is steady or intermittent is not certain, nor of course does this single set of observations establish that overflow never takes place across the 3°S sill.

At shallower levels, 2000-3000 m, isotherms sloping downward to the east above the flank of the Central Indian Ridge (figure 1.14; stations 2291-2296) suggest a third deep western boundary current in the Indian Ocean. The southern boundary of the Central Basin is deep enough to allow direct northward flow from the Antarctic at these levels, and the fact that, for given potential temperatures in this depth interval, oxygen values are higher by about 0.2 ml l^{-1}, and silica values lower by about 3 μM l^{-1}, at this group of stations than at those farther east hints of a more recent southern origin for the water in this current, consistent with northward flow there. The property differences are small, however, and observations are not yet numerous enough to define the deep flow pattern in the Central Basin with certainty.

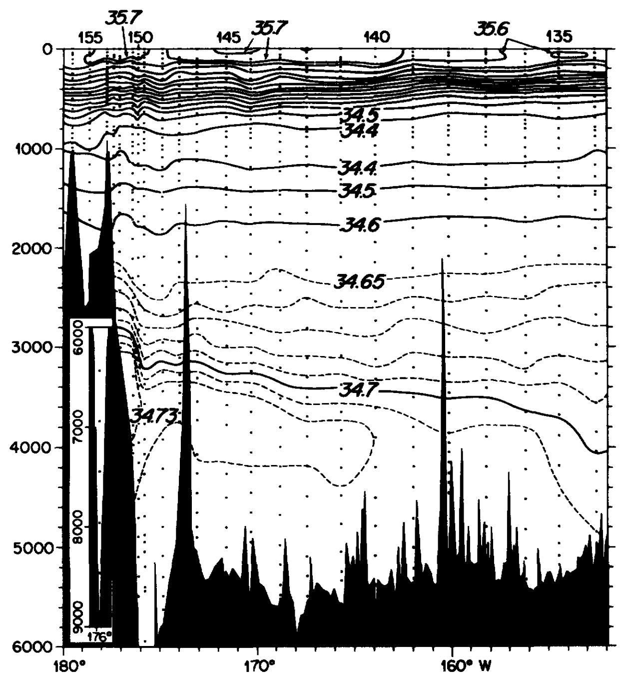
1.5.3 Pacific Ocean

The Tasman Basin just east of Australia is the Pacific counterpart of the Mozambique Basin, in that it is closed off to the north at depths greater than 2850 m (Wyrtki, 1961a), and thus the effective western boundary for the deep South Pacific is not Australia but New Zealand and the Tonga-Kermadec Ridge. The deep current flowing northward along this boundary has been observed at lats. 28°S and 43°S on two transpacific hydrographic sections (Stommel, Stroup, Reid, and Warren, 1973). Portions of the temperature and salinity sections along lat. 28°S between the Tonga-Kermadec Ridge and the foot of the East Pacific Rise are illustrated in figure 1.15. The coldest water is again found toward the west, and below about 2500 m the isotherms slope downward to the east in a zone that, below 3500 m, is some 1000 km wide. At depths of 3000-4000 m a faint salinity maximum represents the last traces of deep water from the North Atlantic carried southward through the South Atlantic, eastward around Antarctica in the Circumpolar Current, and northward here into the Pacific. These traces are also registered as a slight minimum in the corresponding silica section (Warren, 1973). At both lats. 28°S and 43°S the mid-depth oxygen minimum is stronger and somewhat deeper (depths of 2000-2500 m) within 2000-3000 km of the effective western boundary than in the central South Pacific, indicating southward flow from the North Pacific at those levels, concentrated toward the western side of the basin (Reid, 1973a). It is perhaps analogous to the flow of North Atlantic Deep Water in the South Atlantic, though much weaker relative to the northward flow below.

The oxygen minimum is associated loosely with a silica maximum near the 2500-m level, but the silica-maximum layer is separated from the western bound-



(1.15A)



(1.15B)

Figure 1.15 Sections of (A) temperature (°C) and (B) salinity (‰) along lat. 28°S between the Tonga-Kermadec Ridge (left) and the base of the East Pacific Rise, illustrating the principal deep western boundary current of the South Pacific; the Tonga-Kermadec Trench is continued in the inset. Depth in meters. *Scorpio* stations 134-156, 3-18 July 1967, U.S.N.S. *Eltanin* cruise 29. Reproduced from *Scientific Exploration of the South Pacific*, ed. W. S. Wooster (Washington, D.C., National Academy of Sciences, 1970), pp. 45-46.

ary by a zone of low silica concentration about 100 km wide at lat. 28°S and some 1000 km wide at lat. 43°S (Stommel, Stroup, Reid, and Warren, 1973). The pattern of silica variation is strikingly like that at lat. 18°S in the West Australian Basin, but the features of the distribution in the South Pacific are much more markedly and clearly developed. Those features suggest northward flow close to the western boundary from about 2000 m to the bottom as well as in a much broader zone deeper than 3000-3500 m, with southward flow of low-oxygen, high-silica water from the North Pacific somewhat removed from the boundary at depths of 2000-3000 m. The southward flow seems to carry about 3×10^6 $m^3 s^{-1}$ at lat. 28°S, and estimates for the volume transport of the northward current are roughly 20×10^6 $m^3 s^{-1}$ at both latitudes (Warren, 1973, 1976). The latter is probably the largest of all the deep western boundary currents, consistent with its being the principal supplier of deep water to the largest ocean.

At lat. 22°S, nine neutrally buoyant floats were tracked for 4-6 days each in the boundary current (Warren and Voorhis, 1970). The averaged velocities (a few centimeters per second), in combination with thermal-wind calculations, suggested a surface of zero meridional velocity that descended eastward from 3100 m at the Tonga-Kermadec Ridge to 3600 m, and a volume transport for the underlying northward flow of 13×10^6 $m^3 s^{-1}$. On account of the brevity of the velocity measurements, it is not clear just how representative these figures are of the prevailing flow.

The only other velocity measurements in this current—lasting from a few hours to a few days—were made with current meters near the Samoan Passage (9°S, 169°W), whose sill (depth 4500-5000 m) separates the principal basins of the North and South Pacific (Reid and Lonsdale, 1974). The current measurements and density field indicated northward flow through the passage below roughly 3800 m and southward flow above.

The East Pacific Rise separates the main basin of the South Pacific from a sequence of smaller ones in the east. Water characteristics in the Southeast Pacific Basin at lat. 43°S (relatively low temperature and silica, high salinity and oxygen) demonstrate inflow of deep water to this sequence from the south, with a volume transport of perhaps 5×10^6 $m^3 s^{-1}$ (Warren, 1973). The property fields do not give much impression of a *boundary* current, however, and while the characteristics farther north at lat. 28°S are still indicative of a relatively recent southern origin for deep water in the east, the flow must be very diffuse, and it certainly does not look "current-like." Probably the rugged topography in the region disperses the flow. From the temperature distribution Lonsdale (1976) has shown that the bottom water passes north from the Southeast Pacific Basin into the Chile Basin through faults in the Chile Rise near longs. 85-90°W (sill depth 3800-4000 m); further northward flow, into the Peru Basin, occurs via the Peru-Chile Trench and across a sill near long. 89°W. Bottom water also flows eastward across the East Pacific Rise, through several passages in lats. 4-10°S and at lats. 2°N and 8°N. The principal passage from the Peru Basin into the Panama Basin to the north is the northern extension of the Peru-Chile Trench; a 17-day record of bottom current in this passage gave an average velocity, northeastward, of 33 $cm\,s^{-1}$, implying a volume transport of inflowing bottom water to the basin (deeper than about 2500 m) of about 0.35×10^6 $m^3 s^{-1}$ (Lonsdale, 1977). Such extraordinarily rapid inflow is thought to be forced, at least in part, by strong local geothermal heating in the Panama Basin (Detrick, Williams, Mudie, and Sclater, 1974).

Although, as noted, the Tasman Basin on the other side of the South Pacific is closed off by a sill of depth 2850 m, the temperature-salinity characteristics of water deeper than 3000 m in the Coral Sea Basin and Solomon Basin to the north demonstrate that this water is supplied from the Tasman Basin (Wyrtki, 1961a). By a heat-budget calculation, however, Wyrtki (1961a) estimated that this rate of inflow was roughly 0.04×10^6 $m^3 s^{-1}$, a trivial value in comparison with those for the deep northward flows farther east in the South Pacific.

From similar evidence Wyrtki (1961a) argued that deep water in the nearby New Hebrides Basin, New Caledonia Trough, and South Fiji Basin was renewed from the central Pacific to the north. Correspondence in water properties, however, suggests that the deep water of the South Fiji Basin might derive from the deep boundary current east of New Zealand and the Tonga-Kermadec Ridge, by spillage over a sill of depth 2500-3000 m just north of New Zealand (Warren, 1973).

In the northward course of the principal deep boundary current, the salinity maximum that is useful as a tracer characteristic is gradually diminished by mixing, and it disappears just north of the Samoan Passage (Reid and Lonsdale, 1974). South and east of Hawaii, though, Edmond, Chung, and Sclater (1971) identified a bottom layer a few hundred meters thick with temperature and silica concentration distinctly lower, and salinity and oxygen concentration distinctly higher, than in the water immediately above; this layer is undoubtedly sustained by a northeastward extension to the east of Hawaii of part of the flow through the Samoan Passage, that extension at levels of 4500 m and deeper having been demonstrated in broad terms by Knauss (1962a) and Mantyla (1975). The sharp vertical gradients defining the top of the layer might also mark a zero-velocity surface, but that is not self-evident: such gradients certainly do not do so in analogous features observed in the inflow to the Panama Basin (Lonsdale, 1977) and on the Blake-Bahama Outer Ridge in the western North Atlantic (Amos, Gordon, and Schneider, 1971). In any case, 4-day current records demonstrating eastward flow of bottom water at speeds of a few centimeters per second along the northern side of the Clipperton Fracture Zone—southeast of Hawaii—support the general course of water movement indicated by the large-scale property fields (Johnson, 1972). That movement extends eastward across the East Pacific Rise through the low-latitude faults noted above, and it can also be traced at least as far north as lat. 30°N (Mantyla, 1975). Such a poleward interior flow is consistent with the Stommel-Arons dynamics, and it is different from the southward flow in the eastern North Pacific hypothesized by Sverdrup (Sverdrup, Johnson, and Fleming, 1942, chapter XV; see also section 1.2).

Part of the flow through the Samoan Passage also extends northwestward into the North Pacific (e.g., Wooster and Volkmann, 1960; Knauss, 1962a; Mantyla, 1975), but there is no conclusive evidence for deep western boundary currents anywhere in that ocean. In low latitudes this lack of evidence is due at least in part to the island chains and undulating rises that make it difficult even to define a clear-cut western boundary; perhaps the very concept of interior flow plus boundary current is not a useful one there. In middle and high latitudes, Japan and the Kuriles form a distinct western boundary, but, because the deep North Pacific is so far removed from near-surface sources, the water is nearly homogeneous, and deep property maps near the western margin show mostly observational noise, rather than indications of prevailing currents (Moriyasu, 1972).

Nan'niti and Akamatsu (1966) have described deep velocity measurements made with neutrally buoyant floats tracked for 1-4 days from three sites along the Japan Trench near lats. 32, 38, and 40°N. Motions observed at 2-3 km were predominantly southward, with speeds of a few centimeters per second; floats at 1000 and 1500 m moved somewhat faster, and, near 32°N, more toward the west. Worthington and Kawai (1972) have reported additional float measurements, lasting about 1 day each, made just east of Honshu in lats. 34-35°N; two floats set to a depth of 1 km inshore of the Japan Trench moved westsouthwestward at speeds of order 10 $cm\,s^{-1}$, consistent with the observation above near 32°N, while three other floats, released offshore of the Japan Trench at depths of 1-3 km, moved northeastward. Each of the measurements cited, taken individually, is of much too short a duration for a statistically significant estimate of mean flow, but the ensemble hints of a deep, southward-flowing recirculation current close against Japan, as predicted by the Stommel-Arons model. Velocity measurements of very much greater duration are badly needed, however, to establish the existence of that current. Longer-term records have been obtained from current meters moored south of Honshu, but those results probably reflect the local deep flow in the Shikoku Basin, rather than the circulation in the open North Pacific, and the connection between the former and the latter has not been delineated (Taft, 1978).

In the central North Pacific, temperatures at 3500 m are slightly lower than those generally found at the same level in the South Pacific, suggesting widespread deep upwelling (Knauss, 1962a), and thus supporting that assumption of the Stommel-Arons model.

Along the northern rim of the deep North Pacific, near the Aleutian Trench, there is a band of water 100-200 km wide that is colder by about 0.05°C than the water to the south (Knauss, 1962a). It is unlikely that this feature is due to upwelling of deeper water because the anomaly persists all the way to the bottom (Reed, 1969). Reed (1970a) and Mantyla (1975) showed further that this band is continuous with cold water to the west, and inferred therefrom a small eastward flow of deep water along the northern margin. The inference seems unexceptionable, but it is puzzling in terms of the Stommel-Arons circulation model: as described in section 1.3, the model requires northern boundary currents, but they flow *westward* in consequence of being fed by the poleward interior flow; an eastward-flowing current diminishing to the east should be supplying an *equatorward* interior flow. It is conceivable that in this subpolar region the vertical velocity at mid-depths is downward rather than upward, so that the interior flow *would* be equatorward rather than poleward, but if so, then the local density-depth curves might be expected to have curvatures of opposite sign from those to the south, where the vertical velocity is thought to be upward. That the observed curvatures are in fact of the same negative sign suggests (but does not prove) that the upwelling is general. Reconciliation of this eastward current with a southward-flowing western boundary current (if the latter indeed exists) would also be problematical for any simple flow scheme. The cold band is a perplexing feature of the deep circulation.

In summary, when oceans have been reconnoitered with appropriate observational strategies, deep western boundary currents have been found where anticipated along the western margin of the North and South Atlantic, along the eastern sides of Madagascar and the Ninetyeast Ridge in the South Indian Ocean (perhaps also along the eastern flank of the Central Indian Ridge), and beside the New Zealand platform and Tonga-Kermadec Ridge in the South Pacific. Northward flow east of the East Pacific Rise is less like a boundary current than might have been imagined, perhaps because of the topographic complexity, and in the Atlantic the northward-flowing Antarctic Bottom Water current actually becomes an *eastern* boundary current in the southern North Atlantic, possibly on account of meridional bottom slope. The existence of deep western boundary currents in the North Pacific and North Indian Oceans has not been definitely established, probably for lack of suitable measurements. Finally, how eastward flow along the northern rim of the Pacific and the apparent *absence* of western boundary currents in the deep eastern Atlantic are to be reconciled with circulation dynamics is not obvious.

1.5.4 Structure of Deep Western Boundary Currents

When Stommel and Arons (1960b) constructed their model of a global deep circulation, they did not inquire into the character and dynamics of the deep western boundary currents, but simply hypothesized them as a closure device. It is clear from hydrographic sections that these currents differ among themselves, and differ considerably in form from the much swifter western boundary currents of the upper water like the Gulf Stream and Kuroshio. Most notably, perhaps, the deep currents are much wider, except in the North Atlantic. In the South Atlantic, Stommel and Arons (1972) recognized a striking parallelism between isopycnals in the Antarctic Bottom Water current and the bottom profile of the broad South American continental rise. They developed a two-layer model to show how such parallelism—and the associated greatly enhanced width of the current—could be consistent with uniform potential vorticity across the current. Undoubtedly a gentle bottom slope must act to broaden bottom currents, if only because geostrophic flow tends to follow isobaths, but in the vicinity of other deep currents that are at least as wide, particularly those in the South Pacific and the West Australian Basin of the Indian Ocean (figures 1.15 and 1.13), there is no broad continental rise, and the currents flow over ocean floor that is level (apart from small-scale features). Some other physics must be responsible for the large widths there.

In a different theory based on linear dynamics and the idea of lateral mixing of density (Warren, 1976), the velocity and water-property distributions are decomposed into interior and western-boundary fields, and the western-boundary density field is governed by a balance between lateral (zonal) diffusion of boundary-field density and vertical advection of the total density by the boundary current:

$$-w\rho_0 E = K\rho_{xx},$$

where ρ and w are the boundary fields of density and vertical velocity, E the static stability associated with the full density field, K the horizontal diffusion coefficient, and ρ_0 a mean density. Linear dynamics is appropriate to the deep currents because their speeds are small (in contrast to the Gulf Stream, say), but the extent to which *density* mixes horizontally in different parts of the ocean, as distinct from temperature and salinity individually, is uncertain. The theory is incomplete, moreover, in several respects; but, with application of upper and lower boundary conditions, it produces eigenfunctions that fairly successfully rationalize the observed structure of deep boundary currents.

Combining this density balance with the geostrophic vorticity equation and the thermal-wind relation gives a basic zonal scale for each eigenfunction [the "western scale" cataloged by Blumsack (1973), *q.v.* for parameter restrictions] of $l = (Kf^2)/(\beta g E_0 h^2)$, where h is the corresponding vertical scale and E_0 a typical value of E. In deep water E_0 is generally $10^{-9}\ \mathrm{cm^{-1}}$; for $K = 10^7\ \mathrm{cm^2\,s^{-1}}$, and for an eigenfunction of vertical scale 1500 m, in midlatitudes ($f = 10^{-4}\ \mathrm{s^{-1}}$, $\beta = 2 \times 10^{-13}\ \mathrm{cm^{-1}\,s^{-1}}$, $g = 10^3\ \mathrm{cm\,s^{-2}}$) l would be 222 km, which is about right for the observed current widths, considered as two or three horizontal scales (e-folding distances).

The key property that l varies *inversely* with h (the eigenfunctions are tall and thin or short and broad) accounts for several features in the form of the currents. In the South Pacific, for example (Warren, 1976), a combination of two eigenfunctions is necessary for a good fit to the observed variation of density with depth at the western boundary on lat. 28°S (figure 1.15); the one generates the narrow zone of northward flow close to the boundary between 2000 m and the bottom that was inferred from the silica distribution, and the other requires the broad zone to the east with weak southward flow above 3500 m and northward flow below, as indicated by the oxygen, salinity, and silica distributions. A comparable construction (Johnson and Warren, 1979) accounts moderately well for the similar property distributions in the Ninetyeast Ridge current.

The scale relation also helps to explain why the deep boundary current of the North Atlantic (figure 1.9) is narrower than those in the southern hemisphere. The former is directed equatorward uniformly from 1000 m or shallower to the bottom, implying a large vertical scale, while the latter, being composed of equatorward flows near the bottom and poleward flows at mid-depths (some short distance from the western boundary, at least), have rather smaller vertical scales. For the North Atlantic current, with $f = 1.2 \times 10^{-4}\ \mathrm{s^{-1}}$ (lat. 55°N) and $h = 3000$ m, l is 80 km; while in the South Atlantic (figure 1.10), with $f = 0.7 \times 10^{-4}\ \mathrm{s^{-1}}$ (lat. 30°S) and $h = 1000$ m (quarter-wavelength of vertical variation), $l = 250$ km. It is not obvious, though, why poleward flow is part of the boundary-current systems in the southern hemisphere. It seems natural enough for the South Atlantic, where there is a northern source for deep water, but there are no northern sources in the Indian and Pacific Oceans. It must be a consequence somehow of the density stratification of deep water, in which case the existence of poleward flow along the boundary in the South Atlantic may not be linked so tightly to the northern source as intuition would suggest.

The boundary layer described by l is strictly geostrophic, in the sense that $\beta v = f w_z$. For example, by the density balance above, where isopycnals slope downward to the east near the bottom in the boundary region, the curvature of the density field demands a local downward vertical velocity, and thereby a negative value for w_z, which, through the geostrophic vorticity balance, necessitates the equatorward current. It is *this* departure from the Stommel-Arons open-ocean regime—downward vertical velocity—rather than a relaxation of geostrophy that allows this particular kind of equatorward boundary current.

In order to bring the diffusive density flux to zero at the western boundary, however, a nongeostrophic inner boundary layer, the so-called "hydrostatic layer," was fitted to the geostrophic layer (Warren, 1976). The scale of the inner layer $l_H = [(AgE_0h^2)/(Kf^2)]^{1/2}$, where A is the viscosity coefficient. Generally in the deep ocean $l_H \sim 10$ km and is less than l; when the scales are comparable, the two layers merge into a single layer of different form, not interpretable in so simple a fashion. [If $l_H \gg l$, it is a Munk boundary layer, of scale $(A/\beta)^{1/3}$, but for the moderate stratification and the vertical scales of the deep ocean, Munk layers do not apply.] The hydrostatic layer also completes the local vertical flow circuit by supplying and absorbing the vertical flow in the geostrophic layer. In addition, the inner layer sets the meridional velocity to zero at the western boundary, thereby providing a viscous shear-stress force to balance the meridional pressure gradient caused by the meridional variation in Coriolis parameter.

On the other hand, considering the thinness of the hydrostatic layer in relation to the large mixing coefficients on which it is based, the layer looks like an artificial construction, a concept that is probably internally inconsistent. Moreover, even apart from the generally questionable nature of the mixing-coefficient parameterization of diffusive flux, extrapolation of constant values of the coefficient into the coast is surely wrong. Processes somewhat different from those in the open ocean must be at work immediately adjacent to the boundary, and the hydrostatic layer is probably more a mathematical closure device than something one is likely to find at sea. Nonetheless, the layer is conceptually useful in that it illustrates specifically the incompleteness of the geostrophic layer, and shows what conditions must be fulfilled by the immediate boundary processes.

Boundary layers based on the same dynamics and density balance can also be constructed on eastern sides of oceans, but the eigenfunctions would have monotonic vertical variation, which is not compatible with realistic upper and lower boundary conditions. Furthermore, the condition that $w = 0$ at the floor of the ocean would require that the magnitude of the boundary fields increase upward to the sea surface, so that such layers could not be depth-intensified anyway. Hence the boundary layers that close deep circulations through these physical balances are necessarily *western* boundary currents.

The density-diffusive model requires that the fluctuating motions in the deep sea mix density horizontally, but that they do not, through divergence of Reynolds stress, drive the mean boundary currents directly. How true these assumptions are remains to be seen.

1.6 Why Is There a Deep Thermohaline Circulation At All?

Having discovered convection currents in the laboratory, Count Rumford argued that high-latitude cooling should force analogous currents, of global scale, in the ocean. With this motivation, he examined oceanographic data and disclosed the polar origin of deep water. In hindsight after nearly two centuries, however, Rumford's idea as to *why* there should be a deep meridional circulation seems too simple: too much an extrapolation of "everyday" experience, and too little informed both of how buoyancy flux is effected near the sea surface and of how the earth's rotation controls slow motions of such large scale.

To be specific, it is not really obvious that there should be a buoyancy flux at the sea surface to force sinking in high latitudes. The salinity of surface water is increased through evaporation and freezing, and its temperature is governed by short-wave solar radiation, outgoing long-wave radiation, sensible heat flux, and latent heat flux. The last three fluxes are determined as much by the sea-surface temperature as by external conditions, and it is easy to imagine situations in which the temperature of the water column adjusts to those conditions so that there is *no* net annual buoyancy flux across the surface, even in polar latitudes. To be sure, different temperatures in different latitudes would imply different densities, and therefore a field of meridional pressure gradient, but that need not drive a substantial meridional flow. As Ekman (1923) stressed, on the rotating earth the zeroth-order momentum balance would be geostrophic, and the pressure-gradient field would be associated wholly with *zonal* flow (except in the frictional boundary layers).

For an example, consider the earth to be entirely covered with an ocean of uniform depth and, to include only Rumford's essentials, disregard wind stress and salinity flux. Assume further that the external parameters controlling the components of heat flux depend only on latitude. Then an equilibrium field is possible in which the temperature is independent of depth and longitude, and is adjusted to the external conditions of each latitude to bring about zero net annual heat (and density) flux across the sea surface everywhere. In steady state the implied meridional density gradient is then balanced by a zonal thermal-wind shear, and the only meridional flow occurs in surface and bottom Ekman layers, as required by the conditions of no stress at the surface and no slip at the bottom. Demanding that the meridional fluxes in the two Ekman layers balance gives an eastward geostrophic flow (for density increasing poleward) that decreases linearly with depth from the surface to zero within the bottom Ekman layer. The surface speed u is typically $(gD\,\Delta\rho)/(\rho_0 fL)$, where $\Delta\rho$ is the meridional density difference, L the distance from equator to pole, and D the ocean depth; the Ekman-layer speeds are of order $u\alpha/D$, and the Ekman depth $\alpha \equiv (\nu/f)^{1/2}$, where ν is the vertical viscosity. If $\Delta\rho = 6 \times 10^{-3}$ g cm^{-3} (typical meridional difference in surface density), $D = 5$ km, $L = 9000$ km, and $f = 10^{-4}$ s^{-1}, then $u = 33$ cm s^{-1}, and the eastward volume transport, pole to pole, is about 1.5×10^{10} m^3 s^{-1}. For $\nu = 10^2$ cm^2 s^{-1}, the Ekman speeds are roughly 7×10^{-2} cm s^{-1}, and the meridional transports across the 45th parallels in the Ekman layers are merely 0.1×10^6 m^3 s^{-1}, poleward at the surface, equatorward at the bottom. This boundary-layer flow does tend to upset the imposed density field with which the geostrophic flow is associated, but the speeds are so small that the horizontal advection of density is easily balanced by vertical diffusion within the Ekman layer, without noticeably disturbing the interior density field.

This flow is *not* the response to meridional density forcing that Rumford envisaged, nor, of course, is it anything like the meridional circulation that actually occurs in the ocean. The circumstances of the model are far removed from reality in many respects, but it is informative to ask what are the *significant* differences that lead to such a different oceanic circulation system. Most fundamental, perhaps, is the existence of continents, which impose meridional barriers to zonal flow. Consequently, meridional pressure gradients cannot be balanced everywhere by Coriolis forces, and they must force meridional flow somewhere in the system. Meridional barriers are not *small* perturbations to the water-covered-globe model, however, and it is not clear what the different circulation pattern would be.

For example, merely introducing meridional barriers need not lead to *widespread* rising or sinking motions at a much greater rate than required by the upper and lower frictional boundary layers. It is conceivable, in fact, that a stationary density field, with associated zonal flows, could be achieved essentially through lateral and vertical diffusion, with no substantial vertical motion except in thin meridional-boundary layers [a variation without wind stress of a model developed by Rattray and Welander (1975)].

On the other hand, the horizontal circulation in the actual deep ocean is thought to be a consequence of localized sinking and general upwelling. The sinking that is known to take place, moreover, seems not to be merely a concomitant of the overall meridional density gradient, because most of it occurs from sheltered, semienclosed regions (Antarctic continental shelf, Norwegian Sea, low-latitude marginal seas) where near-surface water is driven in, contained long enough to become exceptionally dense, and then is forced back to the open ocean, sinking to depth because of its high density. Wind stress probably contributes to the forcing (e.g., on the Antarctic continental shelf), and certainly salinity enhancement through freezing and evaporation is the principal agent of densification in some cases. Nevertheless, the existence of such embayments where negative buoyancy flux can be sustained against the tendency for adjustment to a no-flux condition at the sea surface—a second geometric departure from the water-covered-globe model—appears to be important for "production" of deep water in the amounts observed.

In the Labrador Sea, sinking occurs through a different process, deep convective overturning, but that must be highly intermittent, associated with weather anomalies, because one would expect the effect of an annual heating and cooling cycle constant from one year to the next to be simply formation and destruction of a seasonal thermocline, with water properties just below adjusted to the winter conditions. Owing to the weakness of the stratification in the central Labrador Sea, however, severe winters or severe weather events within a single winter can, apparently, generate occasional convective overturn to great depth. Climatic unsteadiness thus seems to be another significant way in which the real world differs from the simple model sketched.

Given local, externally forced sinking in the deep ocean, there must be a compensating rising of deep water elsewhere. If this upwelling were confined to regions as small as those of the sinking, it should be discernible in property distributions; since it is not, one supposes that it is widespread over most of the rest of the ocean. This assumed large-scale character of deep upwelling is crucial to ideas about the horizontal circulation of deep water, but the physical basis for it has not been elaborated. Laboratory experiments (nonrotating) on convection forced by heating and cooling at the same level show an analogous asymmetry between the sizes of sinking and rising regions, the asymmetry being attributed to the relative efficiencies of advective and diffusive buoyancy flux (Rossby, 1965). The idea (in oceanic terms) is that although all density forcing occurs at the sea surface, density is added through vertical advection and withdrawn by vertical diffusion, and the total density flux across any level must be zero in the climatological mean. The advective flux varies essentially with the transport of the vertical circulation, while the total diffusive flux, depending as it does on the vertical density gradient, is proportional to the area across which it occurs. Consequently, to equalize the two fluxes, the area over which the upwelling limb of the vertical circulation feeds the upward diffusive flux of density may need to be very much greater than the area in which the downwelling limb occurs, and could, in fact, occupy most of the ocean. This is in the nature of an energy argument, however, and the dynamics of how the forcing generates the vertical motion has not been elucidated even for the laboratory experiments, let alone for the somewhat different oceanic problem.

These concluding remarks have been general and quite speculative, with little possibility of developing them to any satisfying conclusion. They point, rather, to several quite fundamental aspects of the deep ocean circulation that seem, even after 200 years of study, to be still only dimly understood. These basic questions, in contrast to those raised in sections 1.4 and 1.5, seem not likely to be answered by new observational programs, but by hard physical thought applied to data in hand.

Notes

1. I am indebted to Sanborn Brown, Rumford's editor and biographer, for explaining (personal communication) the circumstances in which this essay was first published: "In the late 18th century it was customary to publish the same paper in several journals, and, since the concept of science as an international effort was not yet accepted, it was also customary to publish articles in several different languages. The paper was first published as a separate pamphlet in London by his usual publisher, Cadell & Davies. Rumford sent the manuscript to Cadell & Davies on May 14, 1797 and it was published in July. On May 21 he sent a copy of the same paper to his physicist friend, Professor Auguste Pictet in Geneva, which at that time was part of France. Pictet was the editor of the *Bibliothèque Britannique*, and he translated the paper into French and published it later in that same year 1797. Count Rumford was at that time considered to be a German physicist since he was permanently settled in Munich and was a general in the Bavarian army. The paper was therefore translated into German, probably by one of his favorite mistresses, Countess Nogarola. It was published as a whole in the *Neues Journal der Physik* and in little pieces in the *Chemische Annalen für die Freunde der Naturlehre, Arzney, Gelahrtheit, Haushaltungskunst und Manufacturen*, also both in 1797. It was subsequently published a number of times both in English and in German. All of these publications have the same text."

2. Zöppritz and Krümmel (Krümmel, 1911) attributed the earliest concept of a polar origin for deep water to J. F. W. Otto (1800). In fact, Otto's discussion of the subject is a word-for-word transcription of selected passages from a German translation of the third chapter of Rumford's essay, published in 1799 in the *Annalen der Physik*—even to the extent of reproducing Rumford's erroneous figure of 3600 feet. Insasmuch as Otto did not even hint that he was quoting someone else, it is not surprising that Zöppritz and Krümmel failed to give Rumford the proper credit.

3. Although Humboldt (1831) asserted that he had proved in 1812 the existence of the deep equatorward flow, M. Deacon (1971) points out that there was enough stylistic similarity between Rumford's discussion and those of Humboldt and other later writers to justify the belief that Rumford was their ultimate source. Humboldt was certainly acquainted with Rumford's general idea of convective heat transfer [e.g., *Annalen der Physik* **24** (1803), 17], and it is difficult *not* to believe that he had also noticed Rumford's inference about deep currents, because three of his own papers were printed in the same number of the *Annalen der Physik* that contained the German translation of the pertinent section of Rumford's essay (see footnote 2).

4. Carpenter's source was partly Humboldt and partly Buff (1850), the latter not citing any specific antecedents for it. His discussion is reminiscent of Rumford's, however, in that he illustrated the convection phenomenon by a laboratory experiment: heating a glass vessel from below, which had been filled with water that had powder mixed into it to make motions visible. Rumford had stumbled onto convection currents by setting aside a large thermometer, strongly heated during the course of an experiment, on his window sill to cool; the "spirits of wine" in the thermometer had been contaminated with dust particles, which on being illuminated by the sunlight, revealed by their motion the ascending and descending currents set up in the thermometer as it cooled.

5. Thomson (1877) had long before recognized that bottom water in the South Atlantic was substantially warmer on the eastern side of the Mid-Atlantic Ridge than on the western, and, invoking his "doctrine of continuous barriers," he had hypothesized the existence of the then unknown Walvis Ridge to isolate the eastern basin from direct Antarctic influence.

6. That misinterpretation was originated by Schott (1902); earlier writers had put the sinking region farther north, where the surface temperature was closer to that of the deep water. A complementary misinterpretation must have been the root of Lenz's (1845) idea of strong equatorial upwelling of deep water.

7. *Challenger* stations 323 (28 February 1876, 35°19′S, 50°47′W) and 324 (29 February 1876, 36°09′S, 48°22′W).

8. An unresolved problem in Gill's (1973) analysis is how the ice is moved offshore, because the prevailing winds around the Antarctic coast are from the east, and thus tend to drive the pack ice together with the near-surface water onshore (Solomon, 1974). Yet the ice does seem to move away (Gill, 1973), and such movement appears essential for net annual brine production. Perhaps the coastal current helps to carry ice out of its formation regions.

9. Cooper (1952) pointed out that Nansen (1912) formed his hypothesis on the basis of observations that were not his own,

and were of uncertain quality; moreover, that his hypothesis was never actually vindicated, and no evidence can be found in modern data for convection to the bottom in that region—although observations made due south of Greenland suggested overturning to 2500 m in March 1935 (Dietrich, 1957a), and convection to mid-depths must certainly occur in the Labrador Sea. Cooper (1952) did leave open the possibility of unusually severe conditions when the data used by Nansen (1912) were collected, so that convection all the way to the bottom might conceivably have been occurring at that time.

10. My impression of the reason why the earlier writers underrated the importance of the overflows is that they found hardly any trace of overflow water in the open ocean just south of the Greenland-Scotland ridge. They do not seem to have suspected that such overflows should join together to form a narrow current along the northern and western boundaries of the North Atlantic (figure 1.7) rather than spread directly southward, and the early observations were much too sparse to reveal the existence of that current. Cooper (1955a), on the other hand, had access both to the general concept of western boundary currents and to more comprehensive observational coverage.

11. See the many reports of The Iceland-Faroe Ridge International (ICES) "Overflow" Expedition, May-June, 1960 (1967), *Rapports et Procès-Verbaux des Réunions. Conseil Permanent International pour l'Exploration de la Mer* 157, 274 pp.

12. Tizard (1883) had previously suggested such an analogy, but he abandoned it after concluding that the surface inflow was compensated mainly by outflow at the surface rather than at depth.

13. At the XVII General Assembly of the IUGG in Canberra, December 1979, Clarke and Gascard (1979) described a direct observation in March 1976 of convection penetrating to depths greater than 2000 m, with downward velocities as high as 9 cm s^{-1}, in an area of diameter 10 km near the western side of the Labrador Sea.

14. In much of the western Atlantic, the potential density referenced to the sea surface is actually *less* for the Antarctic Bottom Water than for the Lower North Atlantic Deep Water above it (e.g., Lynn and Reid, 1968). Because the thermal expansion coefficient for water increases with pressure, however, the potential density referenced to some appropriate deep level is indeed greater for the colder Antarctic Bottom Water than for the water above—a point first explained, apparently, by Ekman (1934).