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 0.1×10^6 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 10×10^6 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.

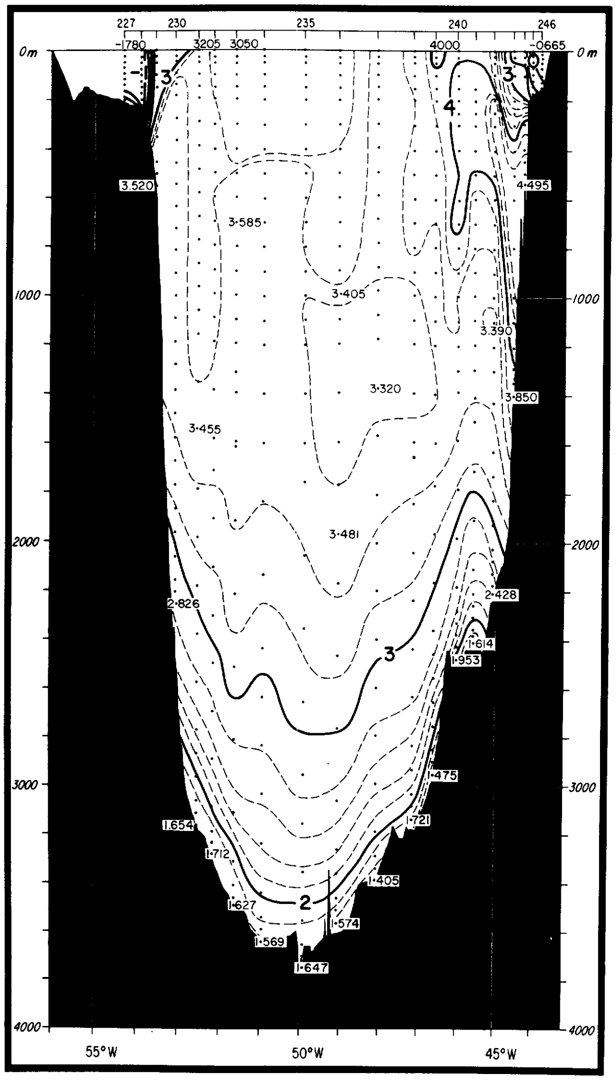


Figure 1.9 Temperature (°C) section along a line running northeastward across the Labrador Basin from southern Labrador (left) to the southern tip of Greenland (see figure 1.7), illustrating the deep western boundary current of the North Atlantic, flowing southwestward against the continental slope of 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 $cm\,s^{-1}$. The mean velocities in combination with station data indicated a volume transport for the deep western boundary current of $24 \times 10^6\ m^3\,s^{-1}$.

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 $7 \times 10^6\ m^3\,s^{-1}$ (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 $22 \times 10^6\ m^3\,s^{-1}$.

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 $cm\,s^{-1}$) 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 9×10^6 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 6×10^6 m^3 s^{-1} in middle latitudes of the South Atlantic, diminishing to about 2×10^6 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).

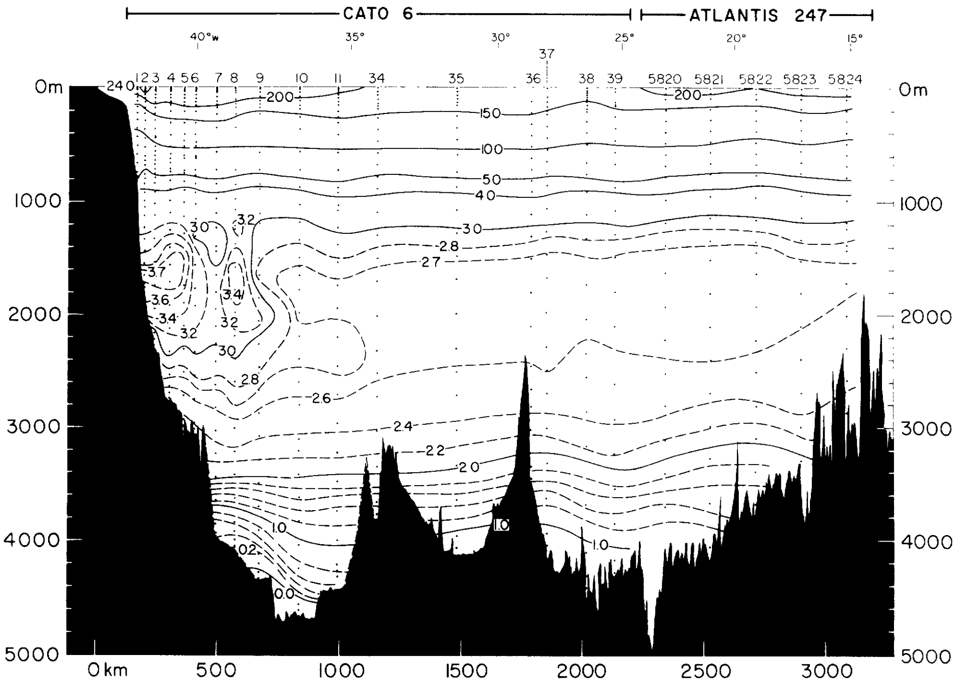
The dynamical fact that poleward movement from an equatorial source can be accomplished by an interior flow may explain to some extent why no deep boundary currents along the western margin of the eastern trough (Mid-Atlantic Ridge) have been observed in transatlantic hydrographic sections (Fuglister, 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.)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ATLANTIS 247
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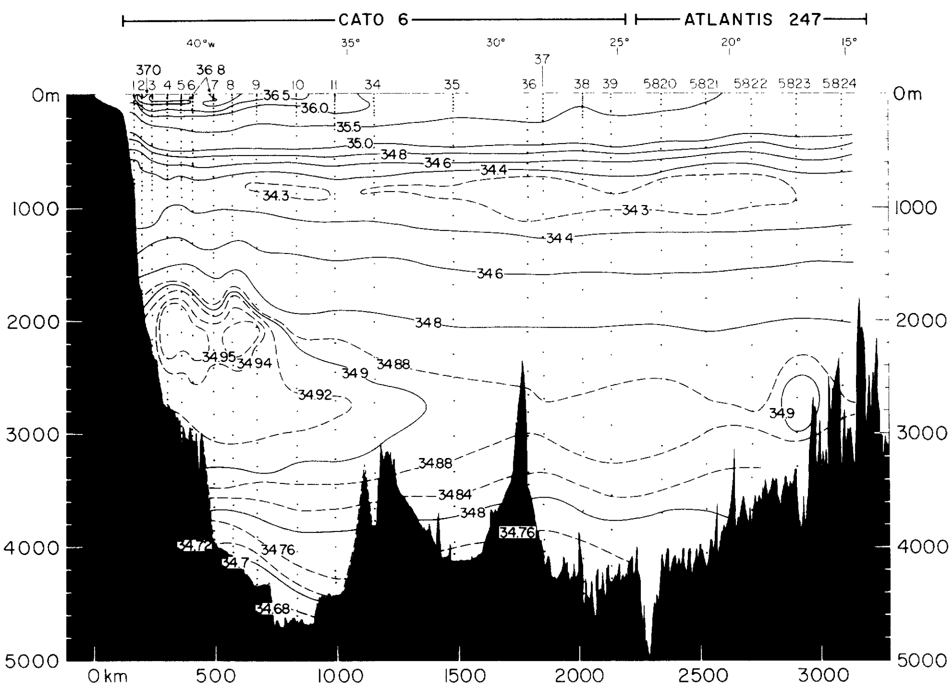
(1.10A)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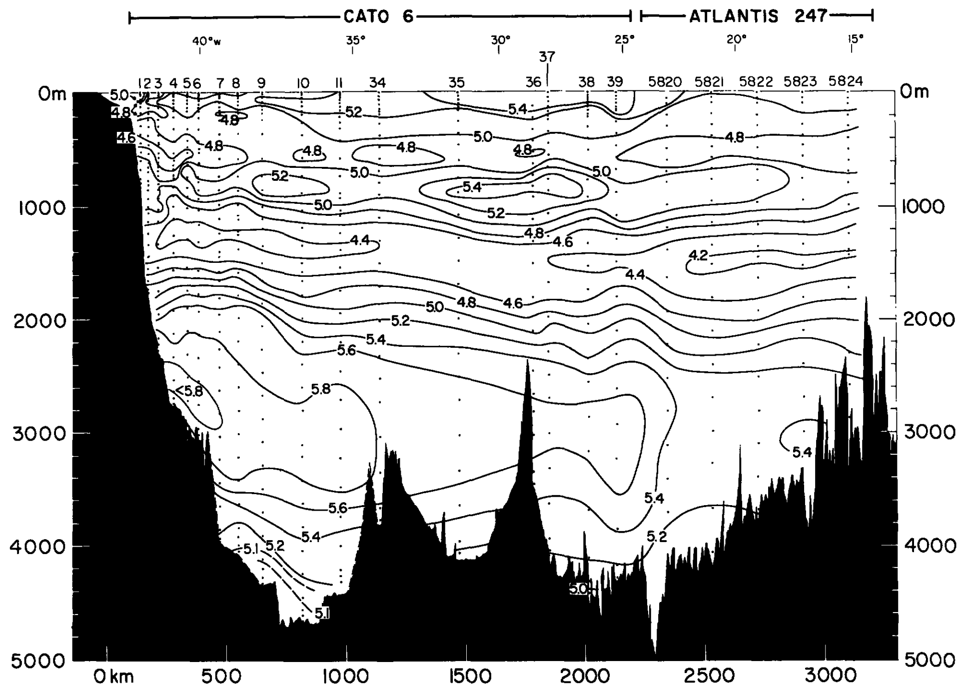
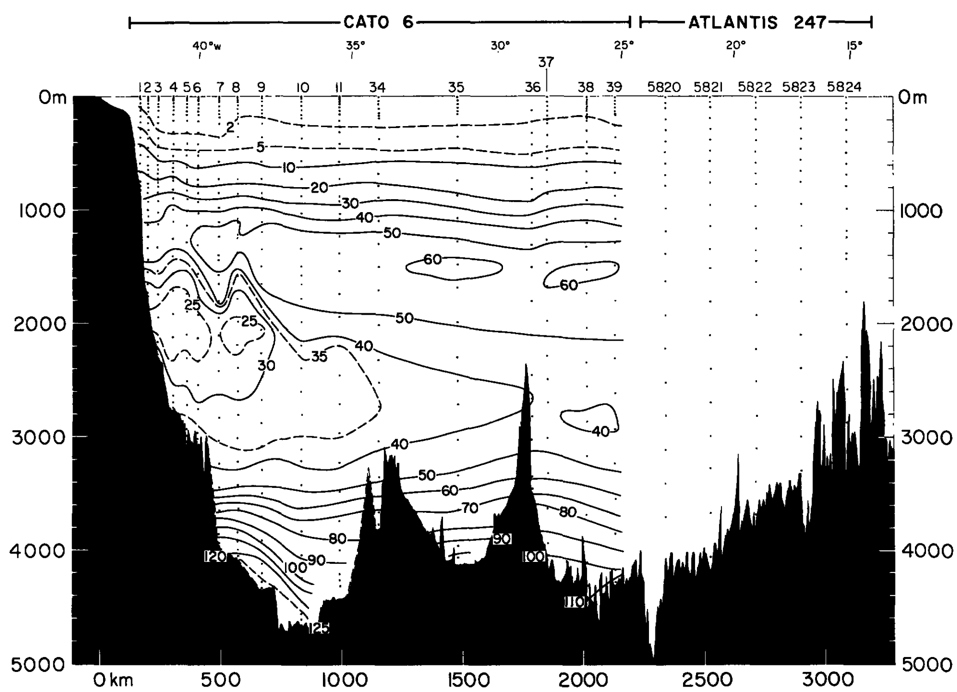


(1.10C)



(1.10D)

Figure 1.10 Sections of (A) potential temperature (°C), (B) salinity (‰), and the concentrations of (C) dissolved oxygen ($\mathrm{ml\,l^{-1}}$) and (D) silica ($\mu\mathrm{M\,l^{-1}}$) 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.)

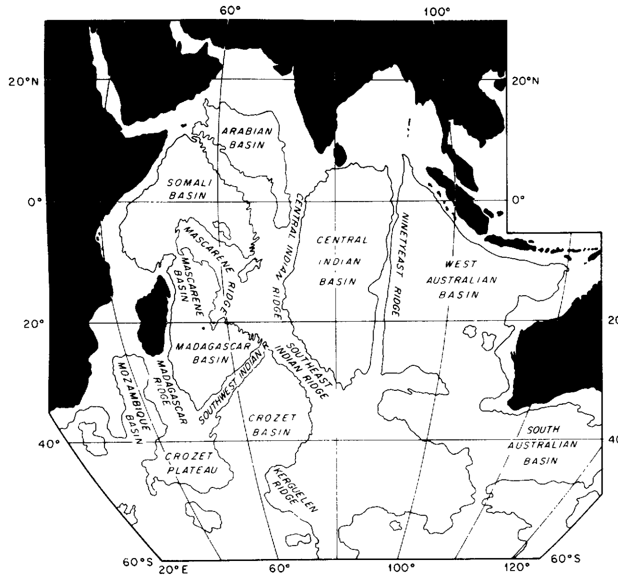


Figure 1.11 Index map identifying basins and ridges in the Indian Ocean, including an approximate representation of the 4-km isobath. (After Wyrtki, 1971.)

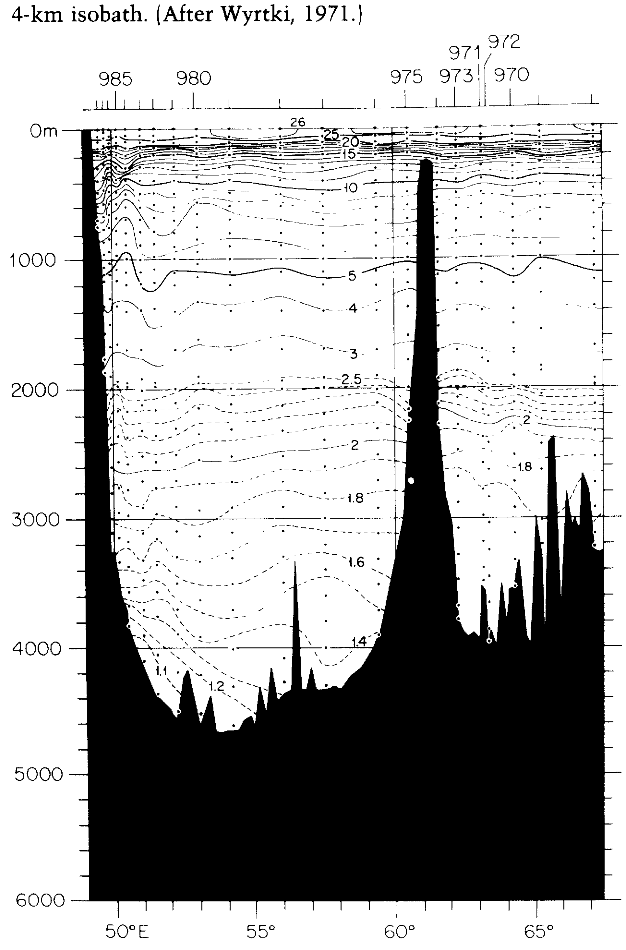


Figure 1.12 Temperature (°C) section along lat. 12°S between Madagascar (left) and the Central Indian Ridge, illustrating the deep western boundary current adjacent to Madagascar in the Mascarene Basin (see figure 1.11). R.V. *Chain* stations 968–988, 20–26 July 1970. (Warren, 1974.)

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

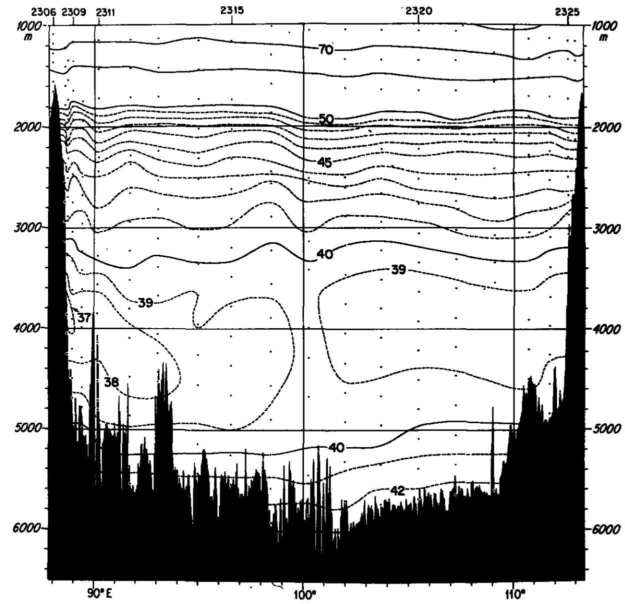


Figure 1.13 Section of specific volume anomaly ($10^5\ cm^3 g^{-1}$) below 1000 m along lat. 18°S between the Ninetyeast Ridge (left) and Australia, illustrating the deep western boundary current of the West Australian Basin (see figure 1.11). R.V. *Atlantis II* stations 2306–2326, 7–17 August 1976.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

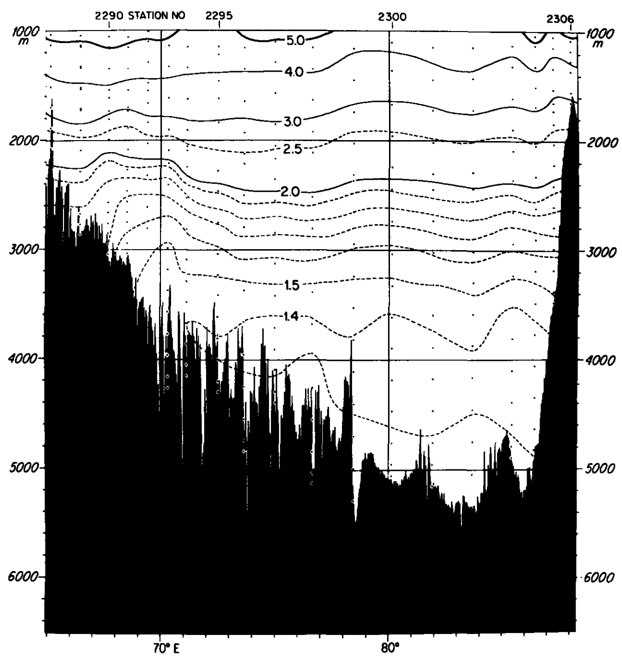


Figure 1.14 Temperature (°C) section across the Central Indian Basin along lat. 18°S from the crest of the Central Indian Ridge (left) to the Ninetyeast Ridge, illustrating the westward increase of temperature below 3500 m and the possible northward flow above the flank of the Central Indian Ridge at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

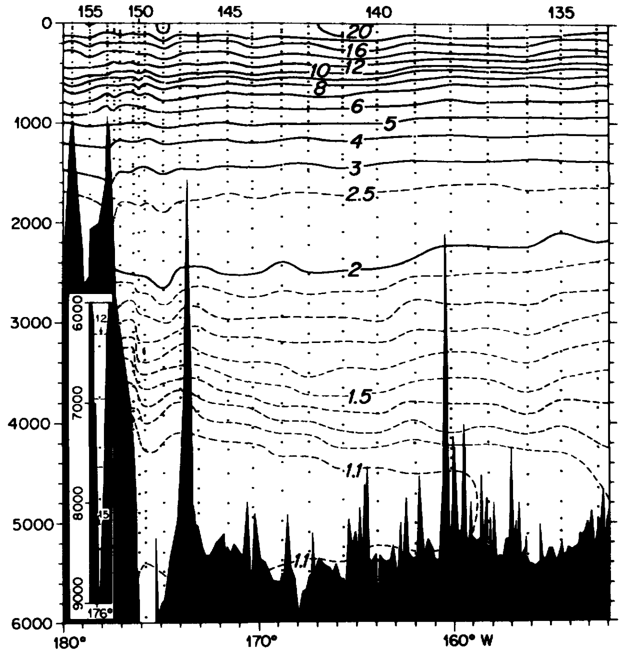
overflow is steady or intermittent is not certain, nor of course does this single set of observations establish that overflow never takes place across the 3°S sill.

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 $\mu M\ l^{-1}$, 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.

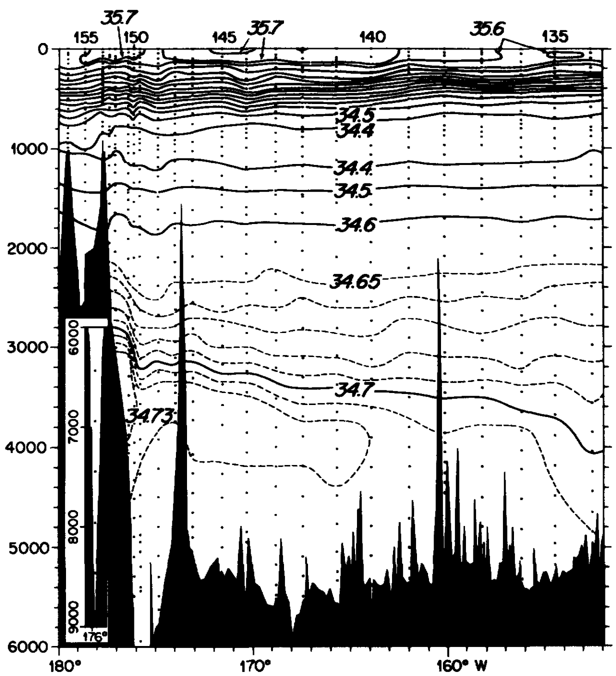
1.5.3 Pacific Ocean

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 $\mathrm{cm\,s^{-1}}$, 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 $K = 10^7\ \mathrm{cm}^2\,\mathrm{s}^{-1}$, and for an eigenfunction of vertical scale 1500 m, in midlatitudes ($f = 10^{-4}\ \mathrm{s}^{-1}$, $\beta = 2 \times 10^{-13}\ \mathrm{cm}^{-1}\,\mathrm{s}^{-1}$, $g = 10^3\ \mathrm{cm\,s}^{-2}$) 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 $\Delta\rho = 6 \times 10^{-3}\ \mathrm{g\,cm^{-3}}$ (typical meridional difference in surface density), $D = 5$ km, $L = 9000$ km, and $f = 10^{-4}\ \mathrm{s^{-1}}$, then $u = 33\ \mathrm{cm\,s^{-1}}$, and the eastward volume transport, pole to pole, is about $1.5 \times 10^{10}\ \mathrm{m^3\,s^{-1}}$. For $\nu = 10^2\ \mathrm{cm^2\,s^{-1}}$, the Ekman speeds are roughly $7 \times 10^{-2}\ \mathrm{cm\,s^{-1}}$, and the meridional transports across the 45th parallels in the Ekman layers are merely $0.1 \times 10^6\ \mathrm{m^3\,s^{-1}}$, 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).