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GENERAL OCEAN CIRCULATION

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INTRODUCTION

Ocean water is induced to move in response to the action of the wind, thermohaline alterations at the sea surface, tidal forces, and, to a lesser degree, thermal changes at the lower boundary (geothermal heat flux). The time variability of the first three of these driving mechanisms induces a time dependance on ocean circulation. The mean circulation over a period T would be considered to be the general ocean circulation if the mean varies little as T is increased. Recent measurements of ocean currents (Schwartzlose and Isaacs, 1969; Schmitz *et al.*, 1970; Rossby and Webb, 1971; Webster, 1971; Irish and Snodgrass, 1972) indicate that variability is rather large on a time scale of hours to weeks. Often major changes in current follow extended time periods of only short (order of a day) variations. Hence it is not possible to construct a picture of the general circulation based on "direct" current measurements, except in very local areas, where the average current is strong compared to the variability, or where great quantities of data are available over the long time period, as in the case of ships' drift data in reference to the immediate surface circulation (though such data may be considered as "indirect" current evidence).

Our knowledge of the general circulation is derived mainly from the distribution of seawater characteristics. The large-scale distribution of the temperature and salinity fields are analogous to a long-period, integrating, current meter. They respond to the "average" advection-diffusion condition of the ocean. Because the distributions of temperature and salinity fields are in a near-steady state, one can assume they respond to a circulation pattern nearly invariant in time and hence represent the general circulation. Comparing distributions derived from numerical models to the observed distribution of temperature and salinity fields, as well as other traces, is an important test for the models (Kuo and Veronis, 1970).

Paleo-oceanographic circulation patterns during the geological past can be studied from the sediment record (vertical and horizontal sediment distribution) obtained in piston cores. The sediment distribution can be used in much the same way as distributions of seawater parameters, since the sediment, by its terrigenous, biogenic, and isotopic composition reflects the physical oceanographic conditions at the time of deposition. Depending on the sediment thickness and rates of sedimentation, one can obtain in a few regions sediments deposited as long ago as the Cretaceous period (64 million years ago), although the paleo-oceanography of the last few million years can be studied over extensive sections of the ocean floor. The much longer drill cores obtained through the JOIDES project provide information on the sediment distribution of extensive ocean regions for earlier periods.

Recent investigations of the paleomagnetism of the seafloor have allowed reconstruction of past geography (Dietz and Holden, 1970). This information, coupled with the sediment record, gives a basis for applying to numerical models of ocean circulation realistic boundary conditions that existed in the geologic past. Numerical modeling of the paleo-ocean circulation is feasible and presents an interesting application of these methods.

The National Oceanographic Data Center (NODC) reports in their Inventory of Archived Data (1969) a total of 3.45×10^5 hydrographic stations in the world ocean. This total has increased during the last few years. Inspection of these data may very well indicate long-term drifts in the temperature and salinity fields as noted by Worthington (1954, 1956, 1966) in reference to the North Atlantic Ocean. Interesting variability in surface salinity and temperature patterns in the western, tropical Pacific Ocean are discussed by Rochford (1972). Only in recent years have we developed a long enough sequence of accurate hydrographic data to study year- or decade-long trends in the ocean. Such information is invaluable to the study of variability of the earth's weather patterns (Bjerknes, 1959; Naimas, 1970; Naimas and Born, 1970).

In order to "handle" the great amount of oceanographic data for the study of the general circulation and secular changes, an up-to-date data bank is needed in an easy-to-use format. Although the NODC efforts have aided, a more intense drive is needed to organize such a data bank and provide digitized, continuously updated hydrographic data on magnetic tapes in a standard format.

In recent years, the study of water-mass distribution with associated current meter observations has enlightened us about the abyssal circulation of the deep western boundary currents of the Pacific and Atlantic Oceans and about the magnitude of the Antarctic circumpolar current. Significant to some of these studies is the use of silicate (SiO_3) as an important indicator of circulation patterns. Silicate concentrations range from near zero in some surface waters to 130–140 μgm·at/liter, and are believed to be of greater conservation (smaller relative "decay-generation") than oxygen, which is also an often-used parameter.

ABYSSAL CIRCULATION

The major regions of formation of abyssal waters are the Norwegian-Greenland and the Weddell Seas. High-saline abyssal water is derived from the outflows of the Mediterranean and Red Seas, and a low-saline water that marks the top of the abyssal water column is formed along the polar and subpolar oceanic frontal regions of the Antarctic and northwestern Pacific and Atlantic Oceans. These regions are producers of the abyssal layer of the ocean. The slow warming of the abyssal waters by various processes allows these regions to remain source areas without a secular increase in surface density. The abyssal circulation pattern that spreads these waters is induced by a combination of thermohaline and wind influences. It is generally considered that the wind is the predominant force (Wyrtki, 1961), although the formation of these waters is a thermohaline process.

The source of abyssal waters in the Southern Hemisphere is mainly in the Weddell Sea, where the bulk of Antarctic bottom water is produced. All of this important water mass is not formed in the Weddell Sea, nor is all Antarctic bottom water of identical properties (Gordon, 1974; Gordon and Tchernia, 1972). Table 1 (Gordon, 1974) shows the basic varieties and ranges in salinity of bottom water thus far observed.

The Weddell Sea, because of its very cold bottom water, is likely the main producer of Antarctic bottom water. The most significant outward flow of bottom water is in the form of a "contour-following current" along the periphery of the Weddell basin (Hollister and Elder, 1969; Gordon, 1974). The rate of production is not known, although estimates of near 20×10^6 m^3/s to a probable upper limit of 50×10^6 m^3/s have been made (Gordon, 1971a). The bulk of Antarctic bottom water has an average potential temperature of near -1 °C. As it flows northward it mixes with warmer overlying water and is warmed from below by the geothermal flux (Olsen, 1968). Most of the bottom water enters the Atlantic and returns southward combined with the North Atlantic deep water. This combination, with additional Antarctic bottom water and circumpolar deep water, flows eastward via the Antarctic circumpolar current to enter the Indian and Pacific Oceans. Reid and Lynn (1971) trace the spreading of abyssal waters on isopycnal surfaces (the sigma value of 45.92 relative to a 4,000 decibar level is used to trace spreading of abyssal water in the world ocean). Their study and that of Lynn and Reid (1968) serve as useful guides to abyssal-water spreading.

TABLE 1 Characteristics of Observed Antarctic Bottom Water

Variety	Location	Potential Temperature (°C)	Salinity (‰)	Oxygen (ml/liter)	Silicate (ug·at/liter)
Low salinity	Western periphery of Weddell basin	−1.4	34.634–34.674	6.7	87
	Deepest parts of Weddell basin	−0.7	34.634–34.674	5.9	110
	Adelie Coast	−0.7	34.650	5.9	110
High salinity	Deep ocean adjacent to the Ross Sea	−0.5	34.738–34.754	5.6	104

The formation of North Atlantic deep water involves a transfer of warm surface water into the abyssal layers. The relatively warm, salty North Atlantic central water mass (Sverdrup *et al.*, 1942) enters the Norwegian-Greenland seas, where heat is lost to the atmosphere at the extremely large rate of 75 Kcal/cm^2 · yr (Worthington, 1970). Worthington considers the Norwegian Sea as a mediterranean basin because low-density water enters the surface layers and high density exits at depth. A water budget suggests that 9×10^6 m^3/s is the rate of surface-water inflow to the Norwegian Sea and 6×10^6 m^3/s is the rate of overflow across the Greenland-Iceland-Faroe ridge. The 6×10^6 m^3/s overflow combines with 4×10^6 m^3/s of water outside of the ridge to produce North Atlantic deep water at rate of 10×10^6 m^3/s, with equal amounts produced on both sides of Iceland, although Reid and Lynn (1971) consider that the principal overflow occurs between Iceland and Greenland (Denmark Strait). The overflow is found to be sporadic (Lachenbruch and Marshall, 1968; Worthington, 1969). In April 1965 Lachenbruch and Marshall found a "boluses" (after Cooper, 1955) of cold water overflowing the Denmark Strait and suggest that these are common, but of short duration and hence rarely observed. The temperature of the April 1965 "boluses" was near −0.45 °C. Worthington (1969) found temperatures near 0.0 °C during periods of maximum overflow. The sporadic overflow may be related to the highly variable water structure of the North Atlantic polar front and to the variable productivity of the fishing grounds in the Iceland region. Lee and Ellett (1967) showed that the volume of overflow and the characteristics of the overflow water may vary.

The temperature-salinity relationships of the waters of the northern North Atlantic are shown in schematic form in Figure 1. The overflow west of Iceland (Denmark Strait) varies between nearly pure Norwegian Sea water to mixtures of this water with Labrador Sea water. The resultant overflow produces the northwest Atlantic bottom water (Lee and Ellett, 1967). The potential temperature is near 1 °C with a salinity near 34.90 ‰. East of Iceland the overflow is warmer and saltier and represents a mixture of Norwegian Sea water and warm salty central North Atlantic water. Lee and Ellett call this the northeast Atlantic deep water. This latter water mass crosses the mid-ocean ridge in the vicinity of the Gibbs fracture zone at 53° N (Worthington and Volkmann, 1965) to enter the western Atlantic and mix with the Denmark Strait overflow to form North Atlantic deep water. Reid and Lynn (1971) state this as follows:

> The colder, less saline, but denser waters at 59°30′ N from the Greenland Sea have passed southward through the Denmark Strait; they pass southward west of the Ridge beneath the warmer, more saline and less dense waters from the eastern side that are now flowing northward. The vertical mixing that has taken place has cooled and freshened the Iceland-Scotland overflow water and made the Denmark Strait overflow water warmer and more saline.

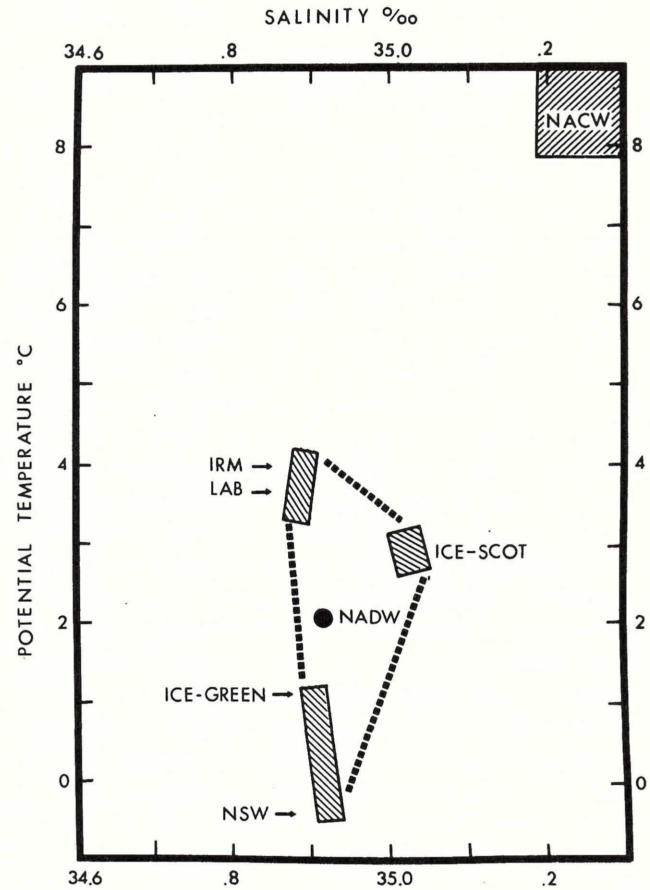

FIGURE 1 Schematic representation of potential temperature-salinity relationships of the water masses involved in North Atlantic deep water production.

The division of North Atlantic deep water into middle and lower components (the upper being the well-defined outflow from the Mediterranean Sea), as done by Wüst (1933, 1935) and mentioned by Edmond and Anderson (1971) and Reid and Lynn (1971), is a fine division. In light of the apparent time variability in the northern North Atlantic, it is not clear what these two slightly different subspecies of North Atlantic deep water signify; the two overflows may not mix entirely and the denser northwest Atlantic bottom water may become the lower North Atlantic deep water mentioned by Wüst (1935).

The North Atlantic deep water has a potential temperature of 2 °C, a salinity of approximately 34.91 ‰, and is produced (sporadically) at an average rate of 10×10^6 m^3/s. Because the transport of water within the deep western boundary current of the South Atlantic is roughly twice this value (Wright, 1969), additional water must be entrained, perhaps a return flow of the Antarctic bottom water (Amos, *et al.*, 1971; see Atlantic Ocean section of this chapter).

The Mediterranean outflow adds approximately 1×10^6

m^3/s (Ovchennikov, 1966) to the deep water. It is relatively warm and salty, initial values being 12 °C and 36.6 ‰, respectively (Wüst, 1935). The high temperatures represent a significant heat input to the abyssal waters.

In a steady-state temperature regime, the input of new cold abyssal water (Antarctic bottom water and North Atlantic deep water) must balance the input of heat by a geothermal heat flux, downward convection of relatively warm water (mainly derived from the Mediterranean Sea), and downward diffusion of heat across the thermocline.

The potential temperature-salinity-volume diagram of Montgomery (1958) shows the average potential temperature of abyssal water (defined as the water below the intermediate water masses, less than 4 °C) to be 1.70 °C for the world ocean. This would be the temperature of the waters colder than 4 °C if they were mixed. This modal temperature will be used as the reference to determine the heat input or removal by convective motion. Water masses introduced to the abyssal layer with a potential temperature above the modal temperature add calories to the abyssal waters, while addition of colder water removes calories. Of the four concentrated water masses entering the abyssal layer three add heat; these are: North Atlantic deep water, Mediterranean Sea overflow, and Red Sea overflow. The Antarctic bottom water removes the heat introduced by the three "warm" water masses, as well as the heat introduced by geothermal heat flux and downward heat diffusion.

The North Atlantic deep water is produced at a rate of 10×10^6 m^3/s and is 0.30 °C warmer than the modal temperature. Using a specific heat of 0.94 the North Atlantic deep water is found to introduce 0.28×10^{13} cal/s. The Mediterranean overflow is 10.30 °C warmer than the modal temperature and flows at a rate of 1×10^6 m^3/s, hence introduces 0.97×10^{13} cal/s. The Red Sea overflow is very warm, being near 20 °C (Düing and Schwill, 1967) with a production rate of 0.2×10^6 m^3/s (Siedler, 1968), which leads to a heat input of 0.34×10^{13} cal/s. The inflow of −1 °C Antarctic bottom water to thermally balance the convective heat input is 6.25×10^6 m^3/s (see Table 2), the largest component being the Mediterranean Sea overflow.

The heat input by geothermal heat can be calculated from the average geothermal heat flux of 1.6×10^{-6} $cal/cm^2 \cdot s$ (including the mid-ocean ridge, Bullard, 1963) and the total area of the deep-sea floor (deeper than 2,000 m) which is 3.05×10^{18} cm^2. The deep-sea geothermal heat flux is 0.5×10^{13} cal/s, which requires an inflow of 1.97×10^6 m^3/s for balance.

Additional Antarctic bottom water is needed to balance the heat input by the downward diffusion of heat. This may be approximated indirectly. Estimate of the heat loss to the atmosphere for the areas between the Antarctic polar front and Antarctica is 15,000 $cal/cm^2 \cdot yr$ (Gordon, 1971b) or 9.5×10^{13} cal/s. This difference may be made up by heat flux across the polar front and/or a greater downward diffusive heat flux in the world ocean (i.e., more Antarctic bottom water produced). If the lateral diffusion across the front is of minor importance, a total 38×10^6 m^3/s of Antarctic bottom water is needed to balance the heat budget and arrive at 15,000 $cal/cm^2 \cdot yr$ heat loss from the Antarctic waters to atmosphere (see Figure 3, and Table 2). The heat flux by diffusion across the thermocline is 7.46×10^{13} cal/s, the major heat input to abyssal waters.

TABLE 2 Heat Input Relative to the Average Abyssal Water Potential Temperature, 1.70°C

Process	Heat Input (cal/s)	Antarctic Bottom Water (−1 °C) Needed for Thermal Balance (m^3/s)
Convective		
a: North Atlantic deep water	0.28×10^{13}	1.11×10^6
b: Mediterranean overflow	0.97×10^{13}	3.81×10^6
c: Red Sea overflow	0.34×10^{13}	1.33×10^6
Geothermal heat flux	0.50×10^{13}	1.97×10^6
Diffusive heating	7.46×10^{13}	29.40×10^6
TOTAL HEAT INPUT	9.55×10^{13}	37.62×10^6

Using the abyssal waters of 3 °C or less to calculate the modal temperature (and in so doing assume the heat input from the Mediterranean and Red Sea outflow goes in total into this colder reservoir) a larger flow of Antarctic bottom water is needed for balance. The average temperature is 1.5 °C and the necessary Antarctic bottom water transport is 40.5×10^6 m^3/s. It therefore is reasonable to deduce that for thermal balance of abyssal waters an Antarctic bottom water transport of 35–40 $\times 10^6$ m^3/s is necessary.

It is clear that factors other than downward heat diffusion into abyssal waters is of significance in "driving" the abyssal circulation. The convective heating terms varied in the geological past so one would expect some variability in Antarctic bottom water production (or whatever other water mass was responsible for heat removal from abyssal waters). It is also interesting to point out that the present water masses that introduce heat to abyssal layers are mainly in the North Atlantic Ocean, where one also finds a large inflow of the Antarctic bottom water.

The total production of Antarctic bottom water of 38×10^6 m^3/s is set by the amount of sea-to-air heat exchange south of the polar front. The downward heat diffusion across the lower thermocline is introduced for balance. A higher rate of North Atlantic deep water production would add more heat to abyssal waters but it would still be small compared to the probable diffusive heat flux. Increasing the North Atlantic deep water by four times would introduce 1.12×10^{13} cal/s into abyssal waters. While this would roughly give a 1:1 ratio of Antarctic bottom water to North Atlantic deep water production rates, as suggested by Craig and Gordon (1965) on the basis of oxygen-isotope data, it requires an abyssal water renewal of 79×10^6 m^3/s. The residence time of such an ocean would be 560 years. This value is perhaps half of the expected residence time (Broecker and Li, 1970). The residence time determined using the transport values of Table 2 (a 4:1 ratio of Antarctic bottom water to North Atlantic deep water) is 890 years, closer but still below the expected value. A residence time of 1,000 years requires an input of 43.5×10^6 m^3/s.

It is possible that heat loss from sea to air south of the polar front may be partially made up by heat flux across the polar front, as mentioned above. A horizontal coefficient of turbulent mixing of 10^6 cm^2/s would introduce only 6×10^9 cal/s, far below the heat loss to the atmosphere. Thus it seems likely that most of the heat loss is balanced by a southward advection of heat by the deep water.

The 1:1 ratio arrived at by Craig and Gordon is based on an Antarctic bottom water sample in the northwestern Weddell Sea. It is probable that Antarctic bottom water is formed elsewhere around Antarctica with less intense characteristics than the variety observed flowing around the periphery of the Weddell Sea (Gordon, 1972). The less intense bottom water may have higher δO^{18} values, and so more Antarctic bottom water may be included in the world ocean abyssal waters.

The downward diffusion of heat across the thermocline may be estimated. Using a value of 1 cm^2/s for the vertical mixing coefficient and a temperature gradient of 1°/100 m for the lower part of the thermocline the downward heat flux is 0.94×10^{-4} cal/cm$^2 \cdot$s for the ocean area between 40°N to 40°S (240×10^6 km^2 calculated from values given on page 9 of Sverdrup *et al.*, 1942) roughly. The poleward limits of the main thermocline total heat flux is found to be 22.6×10^{13} cal/s. Much of this heat is used in warming the upwelling waters. The calculations suggest that only one-third enters the abyssal waters to be returned to the surface in Antarctica.

Pacific Ocean

The abyssal circulation model of Stommel and Arons (1960a,b) shows that abyssal waters enter the Pacific Ocean from the Antarctic region, south of the Campbell Plateau, and extend into the Pacific via a narrow western boundary flow. The data of *Eltanin* cruises 28 and 29 (Scorpio Report, 1970) show that the lower section of the water column immediately east of the Campbell Plateau and the Kermadec-Tonga Ridge is of Antarctic origin. Figure 2 shows the temperature-salinity relations of stations south and east of the Campbell Plateau. The similarity of the lower section of the water column suggests a continuous path of Antarctic water northward in the western boundary of the Pacific abyssal waters. Warren and Voorhis (1970) determine the northward transport of the deep western boundary current to be 12.9×10^6 m^3/s.

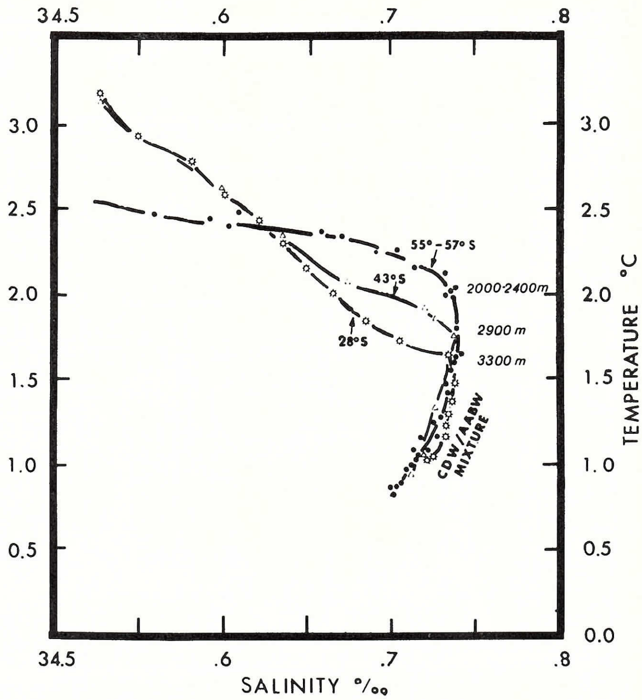

FIGURE 2 Temperature-salinity diagram of hydrographic stations of the Campbell Plateau and the deep western boundary current of the South Pacific at 43° and 28° S. *Eltanin* stations used in this figure: Cruise 28–31 43° 16′ S, 168° 30′ W; Cruise 29–151 28° 00′ S, 176° 30′ W; Cruise 50–1527 57° S, 170° E; 1529 56° S 170° E; 1531 55° S 170° E.

The rather sharp break in the temperature-salinity curve marks the "roof" of the undisturbed Antarctic waters in the water column. Proceeding northward, it deepens in a manner that can be considered as erosion of its upper boundary. The salinity and oxygen of the water at shallower depths than the ". . . sharp discontinuity or cusp in both potential temperature and salinity profiles . . ." (Chung, 1971) drop off rapidly with distance from the cusp, while concentration of nutrients increases. At the cusp, the salinity is at maximum, which is also associated with a silicate minimum (Figure 3) (Craig, 1972). This minimum silicate layer is carried northward with the maximum salinity from Antarctic waters by the deep western boundary current.

The increased silicate between 2,000-3,000 m (associated with a maximum in phosphate-phosphorus and nitrate-nitrogen and a minimum in oxygen) suggests that this water is not being carried northward from Antarctic waters. One would expect the general trend of decreasing nutrient concentration with decreasing depth to occur across the 2,000-

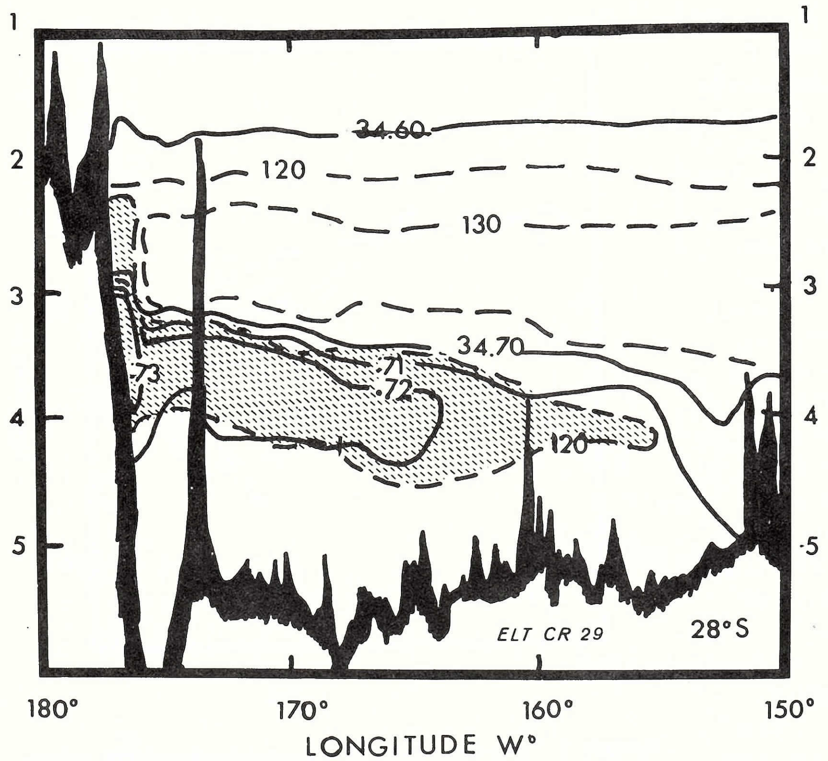

FIGURE 3 Salinity and silicate distribution in the deep western boundary current of the South Pacific at 28° S. The stippled area has silicate of less than 120 μgm·at/liter. (*Scorpio Reports*, 1970).

3,000-m interval, if that were the case. Craig (1972) points out that the maximum nutrient layer represents a southerly flow. This is supported by the direct current measurement obtained during *Eltanin* cruise 40, which indicated a zero meridional velocity "barely above the deep salinity maximum" (Warren and Voorhis, 1970), and is suggested in the Figure 5 of Reid (1965), which shows the 2,000-m maximum of inorganic phosphate-phosphorus in the Pacific along 160° W to be "connected" to the massive region of high phosphate-phosphorus water in the North Pacific.

Above the northward-flowing bottom water in the western boundary region, there is southward-flowing water, not unlike the situation found in the South Atlantic Ocean (Wüst, 1957). What about the deep western boundary flow in the Indian Ocean? The hydrographic section along 32° S given by Wyrtki (1971) shows a similar pattern in salinity, oxygen, and nutrients (the reverse in silicate gradient occurs at lower silicate values than is the case in the South Pacific, and only in the extreme western region) as shown in the South Pacific sections of *Eltanin* cruises 28 and 29. A double-layered, deep, western boundary current may be common to all the southern oceans. In the Atlantic, however, the southerly-flowing water is the North Atlantic deep water, i.e., a relatively "young" water mass rather than an oxygen-depleted, nutrient-rich water mass, as in the Pacific and Indian Oceans.

The continued flow of abyssal water northward in the Pacific Ocean has been studied in a number of recent papers (Reed, 1969; Reid, 1969; Gordon and Gerard, 1970; Chung, 1971; and Edmond *et al.*, 1971). The pattern is one of slow northern migration through the many deep passages in the western tropical Pacific, mainly via a passage immediately southeast of the Phoenix Islands, called the Tokelau trough (see Reid, 1969, and Plate 2 of Gordon and Gerard, 1970). The flow bifurcates near 15° N. One arm of the flow extends into the North Pacific in the western boundary region, and the other flows eastward through a deep passage south of the Hawaiian Islands, near 13° N and 166° W. This water then turns northward after passing the 155° W meridian. The upper boundary of the northward-flowing bottom waters in the South Pacific and parts of the North Pacific is marked by an increased temperature gradient and can be considered as a "benthic thermocline" (Chung, 1971). A benthic thermocline is also observed in the Atlantic Ocean (Amos *et al.*, 1971).

The bottom waters enter the North Pacific, north of Hawaii, from the southeast and west, where they upwell and return to the south, part of the western boundary flow. The slow warming and increasing thickness of the adiabatic temperature layer of the bottom waters (Olsen, 1968), when proceeding northward into the Alaskan basin, suggests an influence of geothermal heating. From the degree of warming of the lower kilometer (below the *in situ* temperature minimum) and a geothermal heat flux of 64 cal/cm^2·yr, it is calculated by Gordon and Gerard that a period of 750 years is necessary for flow from the western tropical Pacific to reach the upwelling zone of the North Pacific. This period is also calculated on the basis of oxygen depletion at a rate of 2×10^{-3} ml/liter·yr (Arons and Stommel, 1967).

Atlantic Ocean

The abyssal circulation of the Atlantic Ocean is more vigorous than the Pacific and Indian Oceans' abyssal circulation, the primary reason being the presence in the Atlantic of the source for both major abyssal water masses. The North Atlantic deep water flows southward along the western margins of the North Atlantic Ocean as a contour-following current (Heezen *et al.*, 1966), i.e., parallel to the isobaths, with the shallower topography to the right of the vector. A number of papers deal with the flow patterns south of the Greenland-Iceland Ridge, all confirming the contour-following model (Worthington and Volkmann, 1965; Swallow and Worthington, 1961; Jones *et al.*, 1970; Amos *et al.*, 1971).

Along the East Coast of the United States, a number of hydrographic and current-meter measurements have been obtained. These are summarized in Figure 4 (taken from Figure 12 of Amos *et al.*, 1971). The total transport of the deep western boundary current determined at each section varies from a low of $4 \times 10^6 m^3/s$ to 50×10^6 m^3/s. A recent estimation by Richardson and Knauss (1971) is 12×10^6 m^3/s, which is in agreement with Barrett (1965). The intersection of the southward-flowing deep current with the northeastward-flowing Gulf Stream in the Cape Hatteras region has been studied by Barrett (1965), Rowe and Menzies (1968), and Richardson and Knauss (1971). They observed that the southward flow is fragmented and interspersed with filaments of northward-flowing, or tranquil, water. The time variability in this region is probably large, as suggested by the time series obtained farther north (but in similar situation) by Schmitz *et al.* (1970). A sharp reversal in bottom current that perhaps was associated with a Gulf Stream meander was observed. Amos *et al.* (1971) suggest, on the basis of water-mass characteristics, a confluence of Antarctic bottom water with North Atlantic deep water in the deep western boundary current, as indicated in Figure 4. These two water sources remain distinct from one another in the Blake-Bahama region, but blend further to the south.

The southward flow of the mixture of North Atlantic deep water and Antarctic bottom water continues into the South Atlantic, incorporating more Antarctic bottom water from below, Antarctic intermediate water from above, and Mediterranean water from the side. Wright (1969) estimates a total southward flow of this combination across 32° S at 20×10^6 m^3/s, i.e., a doubling from the initial volume transport of pure North Atlantic deep water. The southward-flowing deep water eventually abuts with the less-saline circumpolar deep water that is injected into the Atlantic Ocean by the Antarctic circumpolar current via the Drake Passage. The North Atlantic deep water and the circumpolar deep water flow eastward into the Indian and Pacific Oceans (Wüst, 1933, 1935, 1936; Reid and Lynn, 1971). The Atlantic Ocean contribution to the circumpolar water remains distinct from the Drake Passage flow eastward to the Kerguelen Island area. To the east of Kerguelen, these two water masses become homogeneous (Gordon, 1971b). The Atlantic Ocean input to the circumpolar flow is small compared to the Drake Passage transport, but it introduces to Antarctic waters the salt necessary to balance the freshwater inflow into Antarctic waters (Gordon, 1971a).

Antarctic bottom water flows northward from the Weddell Sea into the Argentine basin via a gap in the Falkland fracture zone (Le Pichon *et al.*, 1971) near 50° S and 35° W. The Antarctic bottom water flows northward as a deep western boundary current over the continental rise of the Argentine coast (Wüst, 1933). It exits from the Argentine basin by way of the Rio Grande Passage and flows into the Brazil basin. It continues to spread northward and remains a distinct water mass but in decreasing degrees of concentration, as far north as 40° N in the North American basin.

The eastern Atlantic Ocean does not receive much Antarctic bottom water from the south, because of the obstruction of the mid-ocean and Walvis ridges. However, a deep passage in the mid-ocean ridge in the vicinity of the equator, the Romanche Trench, allows Antarctic bottom water to enter the eastern Atlantic Ocean (Wüst, 1933), where it spreads both north and south and may even extend in trace amounts to latitudes of the British Isles (Lee and Ellett, 1967).

The volume transport of the Antarctic bottom water is estimated by Wright (1970) to decrease from $5\text{–}6 \times 10^6$ m^3/s across 32° S to 1×10^6 m^3/s across 16° N.

The northward-flowing deep western boundary current in the South Atlantic differs from that of the South Pacific. It is composed of more concentrated Antarctic bottom water, rather than a mixture of this with circumpolar deep water. In the Indian Ocean, the deep western boundary current would most likely contain more concentrated forms of North Atlantic deep water in its upper layers and a mixture of circumpolar deep water and Antarctic bottom water in the lower layer (Ivanenkov and Gubin, 1960). Hence, the existence of the deep western boundary flow may depend on the planetary wind systems (Stommel and Arons, 1960a,b), but the actual composition would depend on the relative positions of water-mass source regions.

SURFACE CIRCULATION

The mean surface currents respond mainly to the climatic wind field. The depth to which surface circulation patterns penetrate is dependent upon the water column stratification. In tropical areas, the slope of the strong thermocline efficiently compensates for the sea-surface, dynamic-topog-

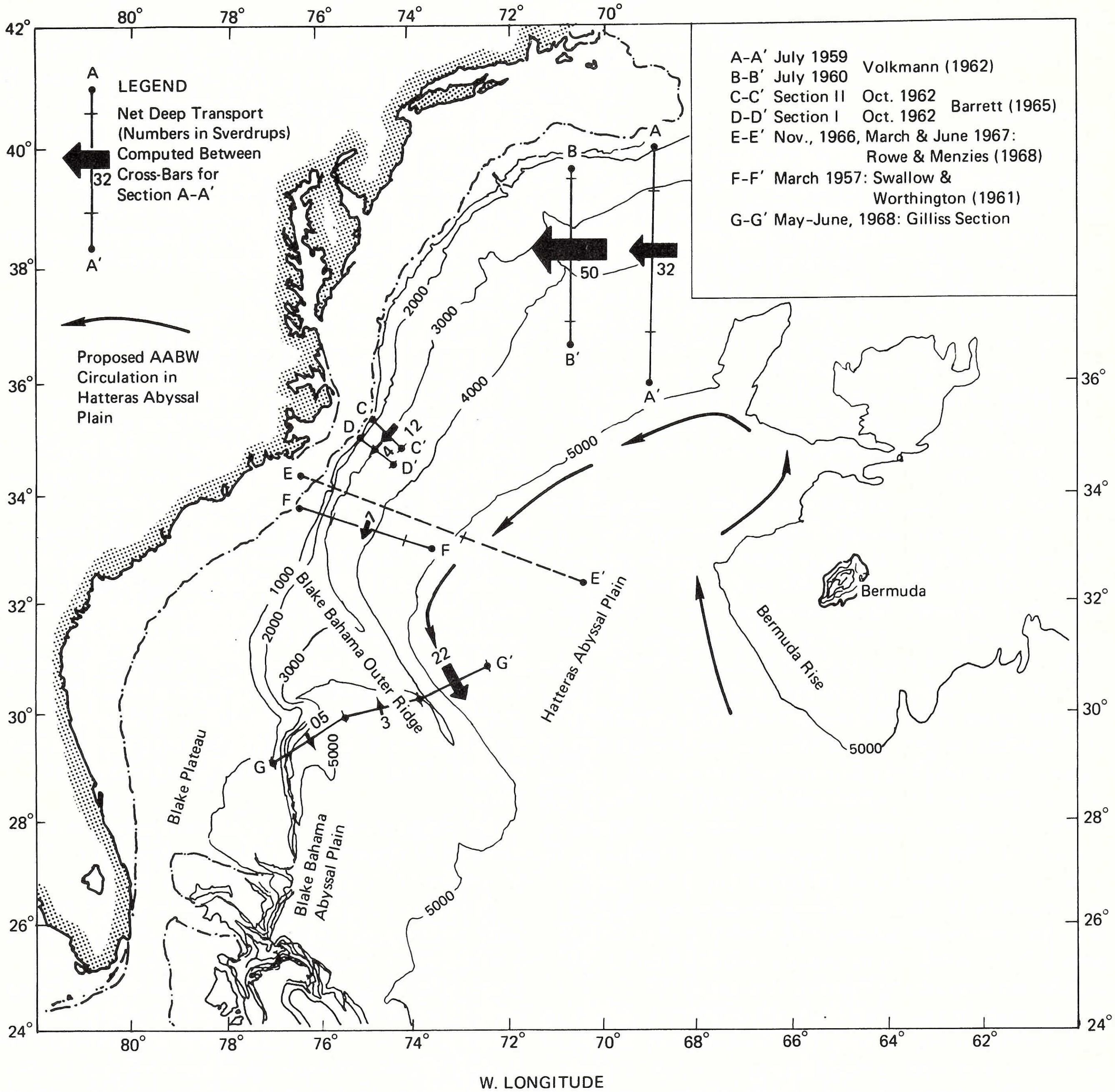

FIGURE 4 Estimate of volume transports and flow of Antarctic bottom water in the western North Atlantic (From Amos *et al.*, 1971).

raphy slope, so rapid attenuation of currents with depth occurs. Current meter data suggest that the equatorial surface circulation pattern extends to only 300-500 m. In polar regions, where waters are less stratified, the attenuation of the currents with depth is small. Maksimov and Vorob'yev (1962) suggest attenuation in Antarctic waters to the sea floor of only 40 percent. Based on water-mass characteristics, current evidence from oriented bottom photographs and the few existing near-bottom current measurements in the Arctic and Antarctic, the surface circulation pattern extends to the sea floor.

In strong pycnocline regions, a surface circulation pattern is often separated from the abyssal circulation pattern by a layer of tranquil motion. Such a zone exists in the equatorial regions and is marked by an oxygen minimum, although the latter is not necessarily a product of weak horizontal motion because the consumption rate is not constant in space (Wyrtki, 1962).

The surface circulation can conveniently be divided into three basic components: equatorial, subtropical gyre, and subpolar gyre. There is also the special case of the polar sea in the Arctic.

Equatorial Circulation

The equatorial circulation (Knauss, 1962; Metcalf *et al.*, 1962; Montgomery, 1962; Metcalf and Stalcup, 1967; Wyrtki, 1967; Tsuchiya, 1968) consists of four basic components: the north and south equatorial currents that flow westward; the surface countercurrent that occurs below the zone of minimum wind stress, generally north of the equator, although a countercurrent exists south of the equator in the western tropical Pacific; and the equatorial undercurrent, found directly on the equator below the sea surface. The equator wind system produces upwelling along the equator and along the northern fringes of the surface countercurrent.

The equatorial upwelling has a negative effect on increasing wind: As the equatorial Hadley cell increases, the upwelling also increases. This, in turn, lowers the equatorial surface temperature and has an effect of decreasing the Hadley cell (Bjerknes, 1966; Flohn, 1972). Sea-surface temperature in the equatorial Pacific Ocean shows cycles of 5–8 years in length and a strong correlation with tropical rainfall (Allison *et al.,* 1971). Such relationship of sea-surface temperature to large-scale weather patterns offers an opportunity for long-range weather prediction, if we understand the complex feedback mechanisms.

An interesting series of hydrographic and current meter observations have been made in the central and western tropical Pacific Ocean (by the group from Centre ORSTOM in Noumea, New Caledonia). Colin *et al.* (1971, 1972) discovered that along 170° E the equatorial undercurrent is divided into two cells, one at 100 m located at the bottom of the mixed layer and one at 200 m within the thermocline. A westward-flowing current, the intermediate equatorial current, is observed below the deeper undercurrent. The undercurrent extends downward to flank the intermediate equatorial current both to the north and south, at 2°30′ N and 2°30′ S.

The equatorial undercurrent at 100 m does not seem to be in geostrophic balance and reverses itself, as does the westward-flowing surface water when the trade winds are replaced by a westwind (a wind from west to east). The lower equatorial undercurrent apparently is a steady geostrophic flow composed of north equatorial countercurrent water, subtropical South Pacific water, and Coral Sea water. The geostrophically balanced intermediate equatorial current transports westward roughly the same volume of water as the deeper undercurrent. The intermediate current exists at least from 140° E to the Galapagos Islands at the depth of the isanosteric surface of 125 cl/t. There is evidence that this current exists all through the world ocean.

The double cells of the undercurrent exist from New Guinea to 150° W or 160° W. To the east, only one core near 100 m is observed.

The Pacific Ocean equatorial undercurrent (Cromwell current) has a transport in the central Pacific of 30–40 $\times 10^6$ m^3/s (Knauss, 1962). Towards the eastern end (Galapagos Islands), the transport reduces rapidly to 2-3 $\times 10^6$ m^3/s (Christensen, 1971), and variations in depth, width, and distance off the equator are observed. The equatorial undercurrent is also a persistent feature in the Atlantic Ocean from 38° W to 6° E, where the waters are of higher salinity than those in the Pacific but are derived from the Southern Hemisphere, in the same manner as the Cromwell current. The Indian Ocean has no steady equatorial undercurrent. It appears to occur only in the middle and late northeastern monsoon, when the east wind is acting on the tropical sea surface.

Subtropical Gyre

The subtropical anticyclonic gyre extends from northern equatorial regions to 50–60° latitude. The relief of the sea-surface topography is approximately 1 m. The density field lies in hydrostatic equilibrium, and the isothermal surfaces obtain their deepest depths below the most elevated portion of the subtropical gyre, not in the center of the gyre but strongly shifted to the western boundary near the 30th parallel (Stommel, 1965). The gyre is geneally considered to be wind-driven. However, purely wind-driven models predict volume transports of water about one-half of the geostrophically estimated transports available to Munk (1950) and perhaps one-third of current estimations. A thermohaline driving mechanism may be of some significance (Worthington, 1972; Shaw and Wyrtki, 1972).

The shape of the warm surface layer in a two-layer ocean was inspected by Shaw and Wyrtki (1972). Because we know the shape of the isothermal surfaces of the subtropical gyre better than we know parameters such as wind stress and internal viscosity of the water, they used the maximum depth of the warm-water layer to determine the frictional coefficient of 2.3×10^{-6} s^{-1} and a transport of 71×10^6 m^3/s for the subtropical gyre. The wind stress necessary to explain the shape of the subtropical gyre was 3 dyn/cm^2. The authors believe this is too large and "consequently an additional driving mechanism should be in operation, and it might be argued that this driving mechanism is given by the continuous addition of thermal energy into the warm upper layer." The northwestern and northern boundary of the subtropical gyre separate from the "wall" when the discontinuity between the warm and cold layer outcrops. This surfacing of the cold water is forced by the horizontal flow of the gyre, as the increased Coriolis parameter at higher latitudes requires increased slope of the discontinuity. Shaw and Wyrtki (1972) showed that the southernmost point of surfacing of cold water along the western boundary is critically dependent on the volume of the warm-water layer and total transport, i.e., increasing the volume shifts the separation to the north. Some warm water is possibly lost to the cold water north of the separation in a process of "overspilling," which Shaw and Wyrtki (1972) estimated to be 16×10^6 m^3/s.

Exchange of waters across the northwestern margins of the subtropical gyres occurs when the large meanders of the Gulf Stream (Stommel, 1965) become detached to produce eddies, generally in the region from 58-68° W and 35-40° S (Parker, 1971). This process is common to the south of the Gulf Stream and represents an injection of relatively cold slope water into the Sargasso Sea. These eddies or cyclonic rings (Fuglister, 1972), appear as thermal domes with cold cores, often with a surface temperature differential of 8 °C with ambient water, and have been observed to have a diameter of 110 km (after an initial elliptical shape), with a surface current around the periphery of cold water of 3 knots. The rings' lifetime apparently is near 12 months, and they slowly migrate to the southwest. At first they touch the sea floor but quickly lose contact (Barrett, 1971). Fuglister (1972) estimates a total of five to eight cyclonic rings, with an equal number of anticyclonic rings (warm Sargasso Sea water cores injected into the slope water) forming each year. This would be necessary to satisfy continuity, as the mean path of the Gulf Stream shifts seasonally, i.e., cyclonic rings are produced from April to November when the Gulf Stream shifts northward, and anticyclonic rings are produced in the period of southward movement.

The cold-water, cyclonic rings are very plentiful in the northwestern and western Sargasso Sea, although their diameter is about one-half of that observed near the area of formation. Parker (1971), on the basis of temperature, has identified 62 rings between 1932 and 1970. Barrett (1971) points out that ". . . although the effect of the ring decay is to decrease APE (available potential energy) (raise the thermocline) of the western Sargasso Sea, there must also be a renewing agency sufficient not only to balance this effect (ring decay) but also to supply in this region sufficient energy to maintain the thermocline at its maximum depth against the general upward tendency." This source is most likely the intense thermohaline activity in the western Sargasso Sea during periods of polar air mass outbreaks (Worthington, 1972).

The warm-water anticyclonic eddies have recently been discussed by Saunders (1971). These eddies are formed near 70° W and 39° N. They are slowly destroyed by heat loss to the atmosphere and by mixing. Warm-water eddies should be more common in slope waters. If this is not the case, it may indicate (and available evidence supports this) that they coalesce with the Gulf Stream.

Upstream of the large-meander and eddy-shedding region, the Gulf Stream has a higher degree of stability and shows a steady growth in transport along its route. Knauss (1969) points out that the volume transport increases from 33×10^6 m^3/s in the Florida Straits to 147×10^6 m^3/s at 64°30′ W, an increase of 7 percent per 100 km over a distance of 2,000 km downstream of the Florida Straits.

Subpolar Gyres

The subpolar gyres are best developed in the Atlantic Ocean and in the Norwegian-Greenland and Weddell Seas. In the North Atlantic, there is a transfer of warm subtropical water (or central water mass) into the Norwegian Sea. This warm, relatively high-saline water is cooled rapidly, but with only moderate freshing; and deep convection ensues to begin the process of North Atlantic deep water formation. In the North Atlantic, there is a loss of warm surface water into the abyssal layers. This must be made up elsewhere by a return of abyssal waters to the warm subtropical waters. This is accomplished by slow upwelling in the thermocline and/or a transfer of water across the oceanic polar-front zones.

In the Southern Hemisphere of the subpolar gyre, there is a separation from the subtropical gyre by the circumpolar belt of water with its strong zonal characteristics. Therefore, the Southern Hemisphere subpolar gyres are isolated from warm water and are colder than their Northern Hemisphere counterparts. The low temperature, the relatively low precipitation, the continental runoff, and the tendency of rapid mixing of melt water, causes the Antarctic surface waters to have relatively high salinity. After a reasonable amount of ice formation, deep convection is induced to form Antarctic borrom water.

In the Pacific Ocean, the northern subpolar gyre has very low surface salinity and deep convection does not occur, except in the northwestern region where some intermediate water forms (Reid, 1965). The southern subpolar gyre in the Pacific Ocean occurs in the Ross Sea region, where high-salinity (34.74°/$_{oo}$) Antarctic bottom water is formed. The Indian Ocean, of course, has no northern subpolar gyre and has only a small southern subpolar gyre in the Amery Ice Shelf region.

It is in the cyclonic gyres of the subpolar regions where deep convection takes place. Although these gyres are wind-induced, it is the thermohaline action that determines whether sinking will occur and determines the depth of convection. In these subpolar regions, the surface salinity is most important; if the salinity does not rise to levels above 34.6 °/$_{oo}$, even freezing temperatures will not allow deep convection into the abyssal layers (below the intermediate waters) of the open ocean. At higher salinities, convection will occur even at the higher temperatures, i.e., the case for the outflow of the Mediterranean Sea. In a stratified ocean, an Ekman convergence is not sufficient to induce deep convection; the thermohaline alterations resulting from sea–air–ice interaction must accompany the wind-induced convergence.

The subpolar gyre of the Weddell Sea has a very marked northern boundary. This boundary, called the Weddell-Scotia Confluence by Gordon (1967), extends from the Antarctic Peninsula into the central Scotia Sea. It has the effect of blocking the Antarctic circumpolar current transversing the Drake Passage. The strength of the Weddell Sea gyre and its northern boundary may be an important consideration in the dynamics of the Drake Passage and entire Antarctic circumpolar current.

In the northern polar sea, the high continental runoff, the confinement of the low-salinity surface water by the polar easterlies, and the surrounding land masses do not permit deep convection. The deep and bottom waters of the Arctic are formed at more southerly latitudes.

CONCLUSION

The influence of bottom topography in ocean circulation, especially in high latitudes, is well recognized. However, often the restrictions placed on flow by the primary features of the bottom topography are negated by secondary features, i.e., narrow passages. A few of these were mentioned in the above discussion: the Romanche, Gibbs, and Falkland fracture zones in the Atlantic Ocean; the Tokelau Passage and the passage south of Hawaii in the Pacific; and many secondary features in the high southern latitudes. These features play important roles in the spreading of abyssal and Antarctic waters. Yet they are not usually included in the approximation of oceanwide bottom bathymetry used in numerical modeling.

It may be desirable to model important individual features separately. This was done by Boyer (see pp. 327-339). The numerical models have yielded circulation patterns consistent with observations on a scale not previously compared.

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REVIEW: B. A. WARREN

Arnold Gordon has given a detailed overview of recent observations of the general circulation, including some rough estimates of heat fluxes. His discussion is comprehensive, and I do not think that other oceanographers would wish to take exception to much that he has said. There are a few points, however, that might be worth emphasizing or amplifying further.

Deep Circulation

Atlantic Ocean

Twenty years ago oceanographers believed that surface water sank to great depth during winter convection in the Labrador and Irminger Seas to form the deep and bottom water of the North Atlantic. Extensive wintertime hydrographic stations, however, have shown conclusively that sinking from the surface in these areas does not reach to depths greater than about 1,500 m (Grant, 1968; Worthington and Wright, 1970). This water from the Labrador Sea does move southward between the Gulf Stream and North America, but it is *not* the North Atlantic deep water. The latter, it has been found, derives from several overflows from the Norwegian Sea (Gordon lists the pertinent references), and this discovery represents a major revision to ideas of the deep circulation in the North Atlantic.

As Gordon points out, the flow of this water away from the Norwegian Sea in a narrow boundary current has been traced through velocity measurements and water-property extrema along the northern boundary of the deep Atlantic and through velocity measurements along the East Coast of the United States. Similarly, the southward flow of North Atlantic deep water along the eastern coast of South America was recognized many years ago in contrast in water characteristics. It may be worth noting, however, that there is little direct evidence to date in the tropical North Atlantic to connect these two boundary currents, despite compelling theoretical reasons to believe that they are connected. This is not to suggest that the connection does not exist, only that the property contrasts here are too slight to show flow very clearly and that the few current measurements yet made in the area are ambiguous.

Far to the south, the manner in which surface water sinks to the bottom of the Weddell Sea to form Antarctic bottom water has been controversial and has raised some questions, including whether the sinking takes place over most of the surface of the Weddell Sea (through one mechanism or another) or is confined locally. The hydrographic sections occupied by the *Glacier* in 1968, however, show fairly convincingly that the bottom water sinks only from the edge of the continental shelf around the Weddell Sea (Seabrooke *et al.,* 1971).

Indian Ocean

Although on theoretical grounds one expects to find bottom water carried northward into the Indian Ocean from the Antarctic in a western boundary current, observations made before and during the International Indian Ocean Expedition (Wyrtki, 1971) were not adequate, because of station spacing, to show its existence. Since then, two hydrographic sections occupied to the east of Madagascar have clearly shown the current, about 400 km wide, at depths greater than 3 km with a probable volume transport of about 5 × 10^6 m^3/s (Warren, 1971). The course of this current to the north is not known. Summer observations in 1964 gave marginal evidence for a deep northward current along the Somali continental slope (Warren *et al.*, 1966), but this flow, if real, may have been a deep penetration of the seasonal Somali current, rather than a northward extension of the deep western boundary current.

It is a moot point whether, as Gordon conjectures, the deep boundary current in the South Indian Ocean is overlain by a second boundary current, flowing southward. Certainly the observations reported by Warren (1971) do not indicate this.

Pacific Ocean

The full results of the *Scorpio* Expedition (Stommel *et al.*, in press) were not available to Gordon at the time he wrote his review article. Dynamic computations, referred to zero velocity at 2 km, give a volume transport for the deep, northward-flowing boundary current near New Zealand and the Kermadec Ridge of about 19 × 10^6 m^3/s at both latitude 43° S and latitude 28° S; and nearly all the southward return flow appears to be confined to the Southwest Pacific basin, between the boundary current and the East Pacific rise, rather than being distributed fairly uniformly over the full breadth of the South Pacific (Warren, in press). On the eastern flank of the East Pacific rise there is a secondary northward-flowing current at latitude 43° S (transport about 5 × 10^6 m^3/s, estimated as above), but only a broad poorly defined net northward flow in the deep eastern Pacific at latitude 28° S.

As Gordon describes, a portion of the principal deep western boundary current has been traced northward into the tropical Pacific, where the flow evidently bifurcates (Reed, 1969), one part continuing northward into the western North Pacific, the other (probably the lesser) turning eastward south of Hawaii. The resulting deep flow pattern in the North Pacific seems unresolved to me, however. Reed (1969) clearly shows that deep water south of the Aleutians must enter the area from the west, rather than the south or east, yet deep water maps prepared by Moriyasu (1972) for the western North Pacific do not show a clear continuous northward flow from the tropics to the Aleutians. Nor is a "western boundary current" structure well-established for the deep North Pacific: Direct velocity measurements of a few days duration (Worthington and Kawai, 1972) revealed a southward current at depths of 500–3,500 m on the continental slope east of Honshu (latitude 35° N), but suggested northward flow at depth further to the east. Deep observations in the North Pacific are simply too sparse at the present time to tell a very convincing story as to how the water is moving around.

It is certainly arguable whether, as Gordon suggests, there is a southward-flowing current above the deep western boundary current of the South Pacific. Permanent ocean currents are associated with pronounced cross-stream density gradients, yet there is little to indicate such structure at these levels in the *Scorpio* sections (Stommel *et al.*, in press). Furthermore, Wyrtki's (1962) discussion of the deep layer of oxygen minimum and nutrient maxima in terms of negligible horizontal flow seems so promising to me that I have difficulty envisioning a boundary current in that layer. Indeed, there is no reason to expect the boundary current structure of the Pacific to be symmetrical to that of the Atlantic: The Atlantic has a northern source of deep water, while the Pacific does not. It might be informative to investigate through extensive numerical modeling the kinds of flow fields that are consistent in detail with the deep oxygen minima and nutrient maxima.

Surface Circulation

Gordon's discussion emphasizes the equatorial currents and the Gulf Stream. Some brief reference to work on others of the great near-surface currents may be worthwhile.

Pacific Ocean

Oceanographers will benefit greatly from a book just published, *Kuroshio* (Moriyasu, 1972), which brings together into one treatise an exhaustive and comprehensive account of virtually everything that has been learned about the Kuroshio and the general circulation of the western North Pacific. It would be futile to try to digest this wealth of material in a few sentences (or thus to say anything at all about the Kuroshio here); but it would be derelict not to hail the publication of this book as an immensely valuable research source in regional oceanography.

The analogous western boundary current of the South Pacific, the East Australian current, has been known for over a century, but it seems much less readily defined than the Kuroshio. Various geostrophic estimates of its transport, relative to 1,300 m, average 28 × 10^6 m^3/s (Hamon, 1965), but later work, revealing numerous eddy-like features and considerable time variability, raises questions concerning the continuity of the flow along the coast of Australia and its connection to the movement in the interior South Pacific (Boland and Hamon, 1970). Recalling that the Gulf Stream, at one time in the history of its exploration, also appeared filamentous and fragmentary, one wonders to what extent more detailed current-tracking will "restore" the integrity of the East Australian current.

Other features of South Pacific circulation have been reviewed recently in *Scientific Exploration of the South Pacific* (Wooster, ed.).

Indian Ocean

The major western boundary current of the Indian Ocean, the Somali current, was mapped for the first time in any detail during the International Indian Ocean Expedition (Swallow and Bruce, 1966; Bruce, 1968; Düing, 1970). Its existence as a narrow, swift current was confirmed, with surface speeds up to 350 cm/s, and directly measured volume transports in the upper 200 m of up to 62×10^6 m^3/s (although the current surely reaches deeper into the ocean than 200 m). Its subsequent flow to the east has been pictured as a large wave train, but the observations are not sufficient to assure the correctness of this interpretation. The Somali current is especially interesting, of course, because of its seasonal reversal, particularly with regard to its time of response to the southwest monsoon. Off northern Somalia, where the current achieves its greatest intensity, there are no series of observations adequate to gauge this time, but surface current measurements near latitude 2° S showed the development of intense northward flow within 10 days of the onset of strong local southerly winds (Leetmaa and Truesdale, 1972). This development occurred 3–4 weeks *before* the monsoon set in over the interior of the North Indian Ocean, however, reflecting the earlier commencement of the monsoon south of the equator and indicating greater regional heterogeneity in the formation of the Somali current than has been envisaged.

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DISCUSSION

ROBINSON: Can the diffusive character of the flow be dis-associated from the dynamics of the variability?

BRETHERTON: How confident are we in the magnitude of the transport calculations of the abyssal circulation?

GORDON: The bulk of the transport probably occurs in strong currents whose width is less than 100 km. The velocity in these currents may be on the order of 10 cm/s. There could be large errors associated with geostrophic calculations for the velocity in mid-ocean regions where the velocity is on the order of mm/s.

NIILER: In warm-water circulation, direct observations of the transport indicate that the natural variability is perhaps on the order of the transport itself.

VERONIS: Thermocline theory normally requires a slow poleward flow in mid-ocean regions as a result of upwelling. On the other hand, studies of distribution of properties and Gordon's own studies suggest a flow toward the equator for these regions.

GORDON: Simple one-dimensional thermocline models cannot account for the circulation and the horizontal mixing that no doubt is taking place in the real ocean.

GATES: There's a strong possibility that transient eddies are an important mechanism in the transport of heat and salt in the ocean. Because of their transient nature, the eddy effects won't be seen in average meridional circulations. Therefore, calculations based on the average meridional circulation may be in serious error. The situation is similar to that which exists in the atmosphere where it is well known that average meridional flow cannot account for the transport of heat in the atmosphere. This is due to the presence of very vigorous mesoscale eddies.