OCEANOGRAPHY OF ANTARCTIC WATERS

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Abstract. The physical oceanography for the southwest Atlantic and Pacific sectors of antarctic waters is investigated with particular reference to the water structure and meridional circulation. The cyclonic gyres of the Weddell Sea and area to the north and northeast of the Ross Sea are regions of intense deep water upwelling. Water at 400 meters within these gyres occurs at depths below 2000 meters before entering the gyral circulation. The northern boundary for the Weddell gyre is the Weddell-Scotia Confluence, and that for the gyre near the Ross Sea is the secondary polar front zone.

The major region for production of Antarctic Bottom Water is the Weddell Sea, whereas minor sources are found in the Ross Sea region and perhaps in the Indian Ocean sector in the vicinity of the Amery Ice Shelf. The Ross Sea Shelf Water contains, in part, water related to a freezing process at the base of the Ross Ice Shelf. The mechanism may be of local importance in bottom water production.

The salt balance within the Antarctic Surface Water indicates approximately 60×10^6 m^3/sec of deep water upwells into the surface layer during the summer. This value is also found from Ekman divergence calculation. In winter, only one half of this value remains with the surface water; the other half sinks in the production of bottom water. An equal part of deep water is entrained by the sinking water, making the total southward migration of deep water 10^8 m^3/sec during the winter. On averaging over a period of a year, it is found that the deep water meridional transport is approximately 77×10^6 m^3/sec. The ratio of zonal to meridional transport is, therefore, between 3:1 to 2:1.

The recirculation of water between the antarctic water masses and Circumpolar Deep Water is large. The volume of water introduced by the inflow of North Atlantic Deep Water is only a fraction of the recirculation transport but is essential in that its high salinity maintains the steady state salinity condition of antarctic waters.

INTRODUCTION

The three oceans are the Atlantic, Pacific, and Indian [*Fleurieu,* 1798–1800; *Krümmel,* 1879]. The north polar extremity of the Atlantic is occasionally considered to be a separate ocean, the Arctic Ocean, but more often it is recognized as a marginal sea of the Atlantic, in much the same manner as are the Caribbean and Mediterranean seas. The major oceans are bounded on the south by the shores of Antarctica. The zone between Antarctica and the southern coasts of Australia, South America, and Africa permits free interocean circulation. Such a zone allows processes that equalize the characteristics of the major oceans. Because of the importance of these processes and the obviously similar climatic conditions of this circumpolar zone, it is often given a separate name: the 'Antarctic Ocean,' 'Antarctic Seas,' or 'Southern Ocean (or Oceans).' However, this 'ocean' lacks a northern boundary in the classical sense, and so it is usually set at an arbitrary latitude or some oceanographic boundary as a convergence or divergence.

For the purpose of this study, it is sufficient simply to call the waters within this zone the antarctic waters. This general term is used to denote the waters from the coast of Antarctica northward to the Antarctic Convergence. The water within the zone between this convergence and the Subtropical Convergence is subantarctic water. This term is best applied only to the surface layers of this zone, since the intermediate, deep, and bottom layers are more or less continuous across the Antarctic Convergence.

REVIEW OF ANTARCTIC OCEANOGRAPHY

The basic structure and circulation of antarctic and subantarctic waters have been extensively studied by *Dean* [1937] and other members of the Discovery Expeditions. Table 1 lists the major general antarctic oceanographic references. The purpose of the present study is to: (1) elucidate the characteristics of water masses and transition zones making up antarctic waters; (2) further investigate bottom water formation and possible regions of formation; and (3) esti-

TABLE 1. General Antarctic Oceanographic Studies

Author	Date of Publication	Antarctic Area
Brennecke	1921	Atlantic
Drygalski	1926	Atlantic
Deacon	1933	Atlantic
Sverdrup	1933	Atlantic
Wüst	1933	Atlantic
Mosby	1934	Atlantic
Wüst	1935–1936	Atlantic
Deacon	1937	Circumpolar
Mackintosh	1946	Circumpolar
Midttun and Natvig	1957	Pacific
Model	1957–1958	Atlantic
Burling	1961	SW Pacific
Deacon	1963	Circumpolar
Ishino	1963	Circumpolar
Kort	1964	Circumpolar
Brodie	1965	Circumpolar
Tolstikov	1966	Circumpolar
Gordon	1967*a*	SW Atlantic; SE Pacific

mate the meridional water transport. The following is a very brief account of the oceanography of antarctic waters. The reader is referred to *Deacon* [1937, 1963], *Gordon* [1967*a*], and other authors shown in Table 1 for a more complete description.

Figure 1 is a schematic representation of the water mass structure, core layers, and meridional components of motion. The main flow is zonal, westward south of the Antarctic Divergence and eastward north of the divergence. The surface and bottom water masses are antarctic in origin, in that their characteristics are acquired south of the Antarctic Convergence. Their northward and downward component of motion is compensated by a southward and upward flowing deep water mass. The results of this important meridional exchange are that heat and nutrients are supplied to the surface of antarctic waters from lower latitudes, and the oxygen content of the deep water of the world is replenished. Such a process allows a steady heat flux from ocean to atmosphere and a high biological productivity rate in the photic zone, maintaining the proper environment for life in the deep water and the low temperatures of the deep ocean. The ratio of meridional zonal flow is discussed in a later section. The structure shown in Figure 1 is found around Antarctica. However, important variations with longitude occur within the water column. These variations can be correlated with the asymmetry of Antarctica and submarine ridges and basins.

Antarctic Convergence (Polar Front Zone)

The Antarctic Convergence, which can be considered as an oceanic polar front zone, is the region separating the antarctic and subantarctic water masses. It has characteristics that alternately suggest convergence, divergence, or a combination of both [*Wexler*, 1959]. The positions of the polar front found by the *Eltanin* and those determined by *Mackintosh* [1946] and *Houtman* [1964] are shown in *Gordon* [1967*a*, pl. 13]. Two expressions of the front are defined: the surface expression (large surface temperature gradients) and a subsurface frontal expression (the location where the temperature minimum begins to increase in depth toward the north at a relatively rapid rate). The polar front zone is found to be 2°–4° of latitude in width. The surface and subsurface expressions are many times separated by a number of kilometers; the more common case is a more southerly surface expression. In the western Southeast Pacific Basin, a double frontal system is found. In this region, the fairly stable T_{min} layer extends from Antarctica to the secondary frontal zone. Between the secondary and the more northern primary zone is a region of a weak and broken T_{min} layer and suggestions of divergences. Occasionally the T_{min} layer descends slightly at the secondary front, but the major descent occurs at the primary front. *Eltanin* BT data from cruises 25 and 27 (along longitudes 127° and 157°W) indicate that such a double structure also exists to the west of the Mid-Oceanic Ridge (see 'Antarctic Polar Front Zone,' this volume). *Houtman* [1967] shows a similar structure south of New Zealand. The occurrence of a double frontal system with combination of convergence and divergence may be fairly common.

Some of the Antarctic Surface Water sinking in the region of the frontal zone contributes to the formation of the low saline intermediate water of the world oceans. The rest mixes into the Subantarctic Surface Water contributing to the warm water sphere of the world ocean. The descent of the Antarctic Surface Water begins when the T_{min} layer is between 200 to 300 meters (see *Gordon* [1967*a*, pl. 3]). Although the T_{min} is quickly destroyed by mixing, the low salinity is maintained and is used as the identifying characteristic of the intermediate water masses of the world ocean. *Ostapoff* [1962] shows surface water cutting across the 200-meter level as a band of low salinity water. Figure 2 is the salinity at 200 meters constructed from *Eltanin* stations. The low salinity band shows some more detail than that of Ostapoff; it suggests that sinking is not uniform throughout the area.

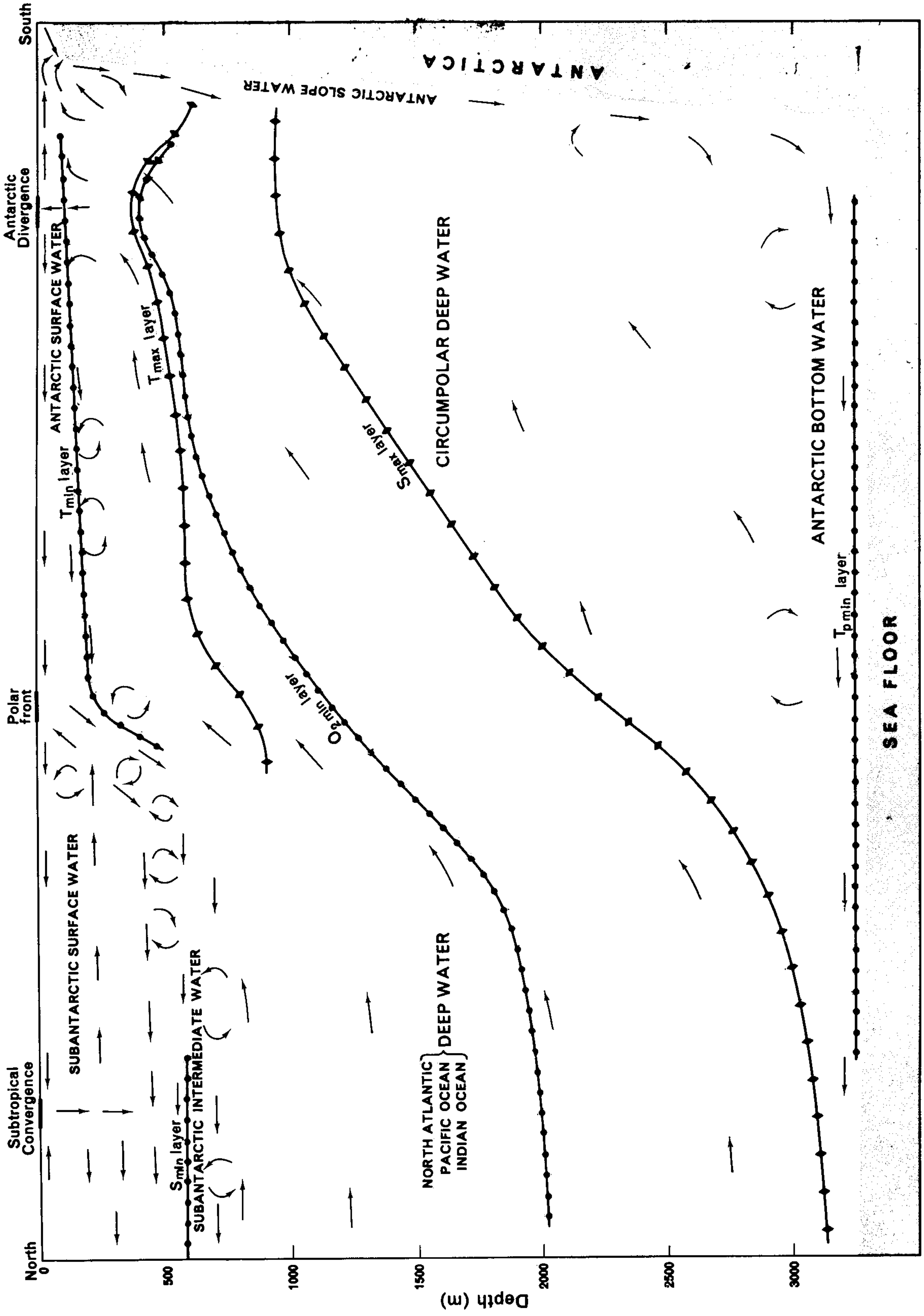

Fig. 1. Schematic representation of water masses and core layers in antarctic waters (from *Gordon* [1967*a*]).

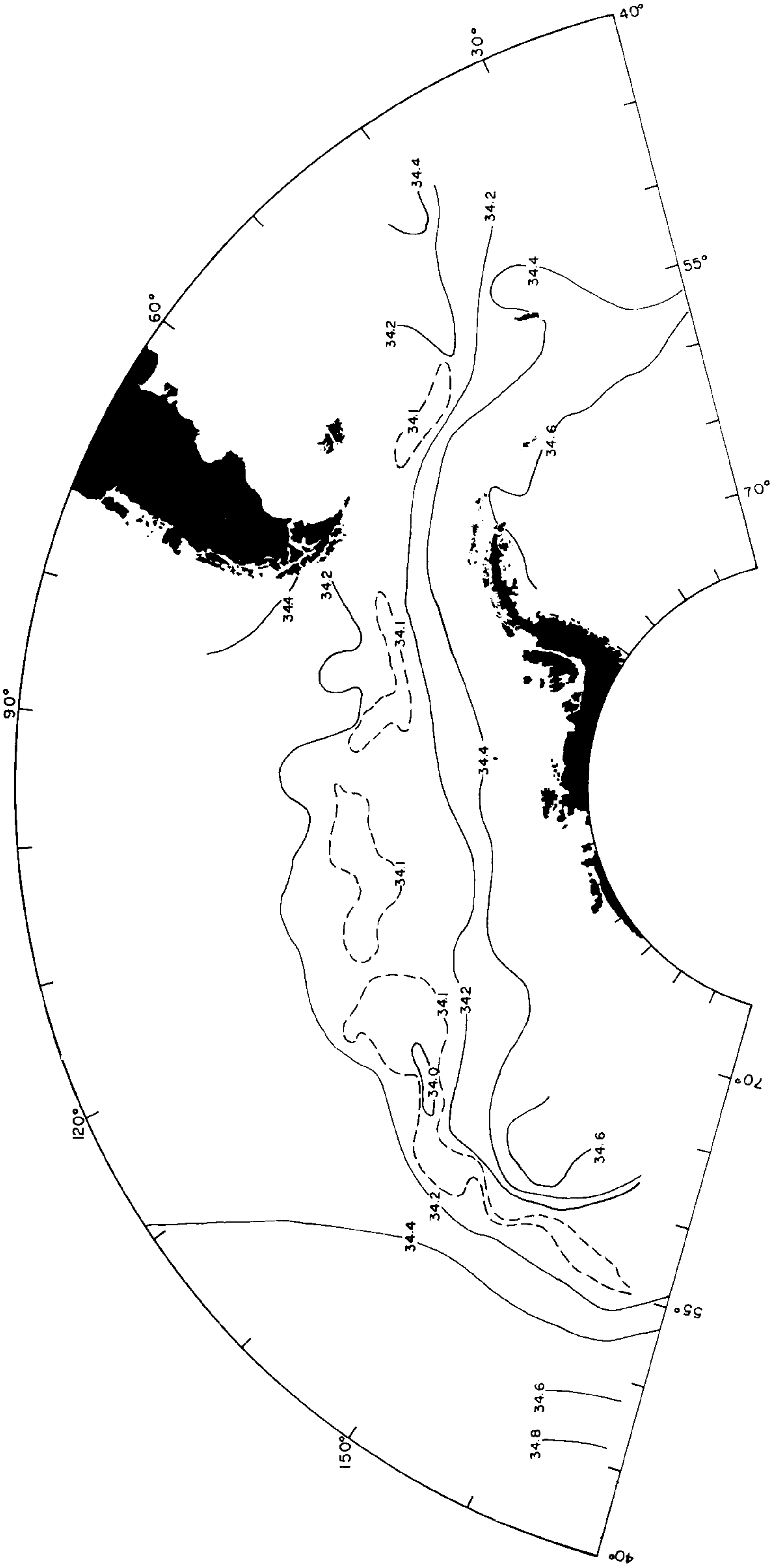

Fig. 2. Salinity distribution, 200 meters, *Eltanin* observations.

On comparing it with *Gordon* [1967*a*, pl. 14], it is found that the surface water descent compares well with the primary frontal zone. The temperature at 200 meters is shown in Figure 3. The isotherm distribution is extended to the continent by using the data in *Tolstikov* [1966]. The warmer band of water in the southern Southeast Pacific Basin and Weddell Sea shows the position of the Antarctic Divergence. The colder band just to its north is the Antarctic Surface Water. The convergence occurs in the large temperature gradient region. The mean position can be approximated by the 2° isotherm. *Botnikov* [1964] used the 2°C isotherm of the temperature minimum layer to define the summer position of the convergence.

Antarctic Divergence

The Antarctic Divergence (average latitude of 65°S) is essentially a wind-produced feature. The westerlies north of the divergence transport surface water eastward with a small component to the north, whereas the coastal easterlies cause a westerly flow with a southward component to the surface water. The resulting divergence and deep water upwelling, which may be of a diffusive nature or perhaps occur within limited regions over short time periods, are the most important oceanographic processes of antarctic waters. They allow the deep water contact with the antarctic atmosphere and the associated sea-air interaction. The upwelling necessitates a southward flow of the deep water. This flow on conserving its angular momentum may help create the zonal currents of antarctic waters and the small amount of attenuation of the current with depth.

Along the antarctic coasts, cold, relatively saline shelf water forms, owing to the intense cold and freezing of sea water. Shelf water on mixing with roughly equal proportions of deep water forms Antarctic Bottom Water, which flows northward. This northward motion initiates a westward component of motion along the continental rise of Antarctica and accumulation of bottom water to the east of the main submarine ridges.

Weddell-Scotia Confluence

The Weddell-Scotia Confluence [*Gordon,* 1967*a*] or Bellingshausen Front [*Model,* 1957, 1958] is the line separating the water derived from the Weddell Sea and that derived from the Pacific Ocean. It is a line extending from the Bransfield Strait through the central Scotia Sea and north of the South Sandwich Islands. It is most pronounced and stationary in the Circumpolar Deep Water. The small change in position and character from 400 to 4000 meters suggests only minor amounts of convergence at the Weddell-Scotia Confluence. In the surface layer, large horizontal variations are found. Here the boundary of the Weddell and Southeast Pacific surface water is much more turbulent than the deep water boundary. However, the separation is obvious on the *T/S* diagram of the area (Figure 4).

WATER MASSES

Temperature-Salinity Relation

Water masses can be conveniently defined by their particular relationship of temperature and salinity. For this purpose, a plot of temperature against salinity of all data points is made. Such a plot, the *T/S* diagram, was introduced to oceanography by *Helland-Hansen* [1916].

To construct a *T/S* diagram for the region investigated by *Eltanin,* the hydrographic data from most of cruises 7 to 27 were used. Figure 4 is the group *T/S* plot for *Eltanin* cruises 7, 8, 9, 12, and 22, which represent the Drake Passage, the Scotia and Weddell seas, and the areas immediately north and east of these seas.

The water masses were identified in the following manner: The Antarctic Surface Water includes the water within and above the temperature minimum layer (100–300 meters). The upper and lower deep water are identified by a temperature maximum and salinity maximum, respectively. The Weddell Deep Water (WDW) is that deep water south of the Weddell-Scotia Confluence. The deep water found in the Pacific sector and north of the Weddell-Scotia Confluence in the Scotia Sea is called the Southeast Pacific Deep Water (SPDW). The water between the Antarctic Surface Water and Circumpolar Deep Water and the water between the Circumpolar Deep Water and Antarctic Bottom Water constitute transitional zones. Within each of these zones, a layer of zero meridional motion exists. The bottom water boundary with the deep water is not well defined and is arbitrarily set at slightly below the 0°C isotherm in the Weddell Sea and +0.5°C in the Pacific sector. The Subantarctic Surface Water is the thick isohaline layer north of the polar front zone. At its base is the Antarctic or Subantarctic Intermediate Water, defined by a weak salinity minimum (it is much more obvious north of the Subtropical Convergence). A transitional zone between the Subantarctic Surface Water and the deep water is the relatively isothermal layer between these two water masses.

Fig. 3. Temperature distribution, 200 meters, *Eltanin* and IGY observations. Isotherms near Antarctica are taken from *Tolstikov* [1966].

Besides the general usefulness of such a diagram in identifying water masses and defining their T/S region, three important facts are found:

1. The deep water of the Weddell Sea is in the same T/S region as the bottom water of the Southeast Pacific Basin. Therefore, it is reasonable to conclude that the WDW at 400 meters is derived from that bottom water which occurs below 3000 meters in the northern Drake Passage. The Southeast Pacific Bottom Water leaves the sea floor in the northern Scotia Sea and the area to the northwest of South Georgia [*Gordon*, 1966, 1967*a*]. Therefore, great amounts of upwelling are associated with the transfer of water into the cyclonic Weddell gyre. The oxygen of the WDW is between 4.5–4.7 ml/l compared to the 4.7–4.9 ml/l range of the bottom water passing through the northern Drake Passage, indicating some oxygen consumption in transit from the Drake Passage to the Weddell Sea. It is interesting to point out that the bottom water formed in the Weddell Sea includes some of the warmer bottom water of the Southeast Pacific Basin, which may be derived from the area of the Ross Sea. The volume transport of the water below 3000 meters in the northern Drake Passage is 25×10^6 m^3/sec (calculated from *Gordon* [1967*b*, fig. 3]). The probable transfer is in the vicinity of 30°E [*Model*, 1957], where a broad southward penetration of warm water occurs.

The WDW is, in part, converted to surface water by sea-air exchange and, in part, entrained with the sinking shelf water in the production to AABW. In addition, some WDW may exit from the gyre directly. Therefore, it is not possible to arrive at a rate of AABW production; however, if all the WDW exits as a 50% component of the AABW, implying that the surface water of the Weddell gyre is derived from outside the Weddell Sea, an upper limit of 50 sv is placed on the AABW outflow. The actual value is probably much less than this value.

2. The warm water occasionally found to override the temperature minimum layer south of the subsurface expression of the polar front [*Gordon*, 1967*a*] is either Subantarctic Surface Water, as in the case of *Eltanin* cruise 8, or warmed (modified) Antarctic Surface Water, cruises 7 and 22, which represent January and February conditions. The warmed Antarctic Surface Water has salinities similar to the Antarctic Surface Water, that is, 0.2 to 0.3‰ lower than the Subantarctic Surface Water. The warming obscures the temperature surface expression of the polar front, and the surface boundaries of the Antarctic Surface and Subantarctic Surface Water, in these cases, are found only by a change in salinity. The water to the south is slightly less dense than the Subantarctic Water. In such a case, some Subantarctic Surface Water may sink at the front to join with water of the temperature minimum layer.

3. The Antarctic Surface Water can be subdivided into the surface water of the Weddell Sea and that of the Scotia Sea. The line of separation is the Weddell-Scotia Confluence. The surface water of the Weddell Sea is colder and spans a greater salinity range.

The T/S distribution of the *Eltanin* stations of the South Pacific sector of the antarctic waters is shown in Figure 5. The following cruises are included: 10, 11, 13, 14, 15, 17, 19, 20, 23, 27. The excluded cruises were either in the Tasman Sea, or mainly north of 50°S.

Basically, the same water masses shown in Figure 4 are found in Figure 5. It is possible to divide the Antarctic Surface Water and the underlying transition zone into two sections, as is done in the Weddell and Scotia seas. The division is the secondary polar front zone. Besides the sharp difference in the surface T/S points north and south of the secondary front, some difference is found in the deep water. The upper deep water south of the secondary front (that is, in the southwestern Southeast Pacific Basin and the area immediately northeast of the Ross Sea) is in the same T/S region as the lower SPDW and transition into the Southeast Pacific bottom water, indicating a return flow of the deep water to the southwest in the same manner as the flow into the Weddell Sea. From the core layer maps of *Gordon* [1967*a*] this return occurs between 130° to 140°W. Similar to the WDW, it represents an upwelling of deep water from between 2000 and 3000 meters to about 400 meters. The gradients across the secondary front zone in the deep water are not as intense as they are across the Weddell-Scotia Confluence.

The phenomenon suggested by the T/S diagrams of intense upwelling in the cyclonic gyres of the Weddell Sea and southwestern Southeast Pacific Basin may be the primary method in which deep water is carried upward. Similar conditions may exist near the Kerguelen Plateau.

The Subantarctic Intermediate Water is much more pronounced in the Pacific sector, covering a T/S region of slightly lower salinity, perhaps due to the lower salinity of the Antarctic or Subantarctic Surface Water. Warmed or modified Antarctic Surface Water is found on cruises 19 and 23, again indicating that the more

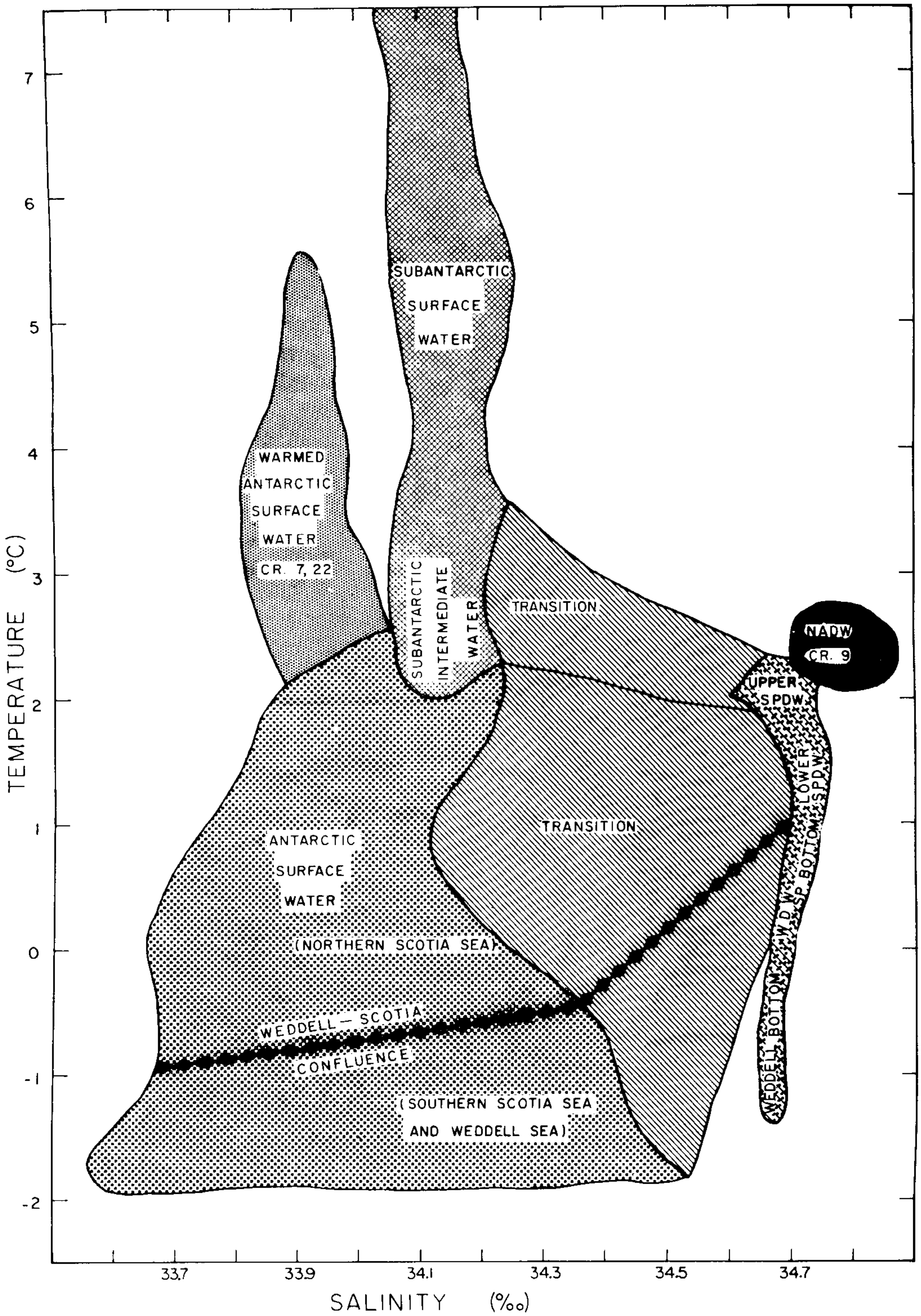

Fig. 4. Temperature-salinity diagram for Drake Passage, Scotia Sea, and Weddell Sea. SPDW = deep water of the Southeast Pacific Basin; SP Bottom = bottom water of the Southeast Pacific Basin; WDW = deep water of the Weddell Sea; NADW = North Atlantic Deep Water.

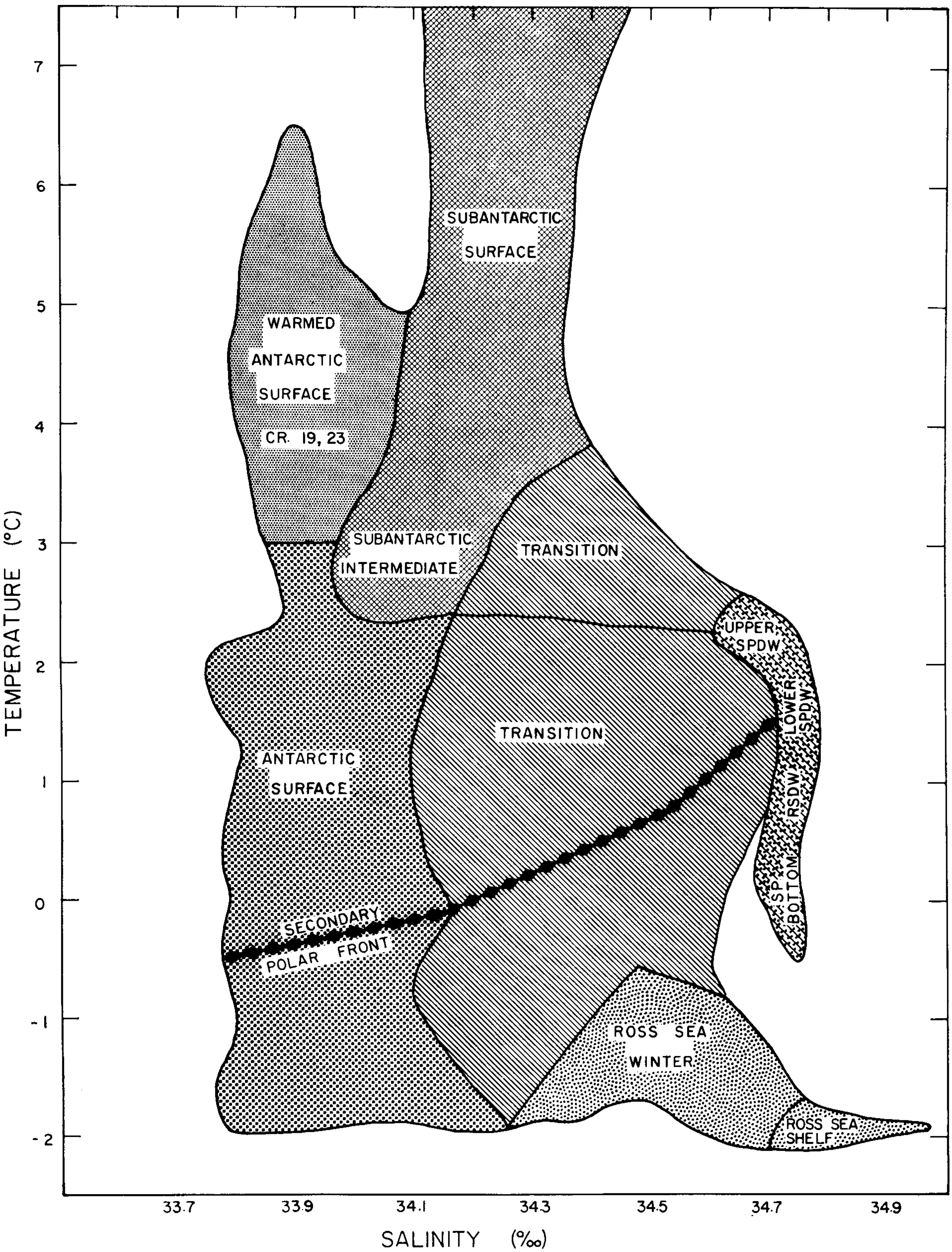

Fig. 5. Temperature-salinity diagram for Pacific Ocean sector of antarctic waters. Abbreviations as in Figure 4 caption; RSDW = deep water in the ocean region north and northeast of the Ross Sea, but south of the secondary polar front zone.

southern surface expression of the polar front zone relative to the subsurface expression is not always due to overriding of Subantarctic Surface Water.

Water masses associated with the Ross Sea are shown in Figure 5. These are the Ross Sea Winter Water and the dense Ross Sea Shelf Water, which is discussed below. Both of these water masses represent a protrusion on the T/S diagram to colder, more saline water and, therefore, must have an origin related to ice formation: either sea ice or freezing beneath the Ross Ice Shelf.

Oxygen Saturation of Antarctic Waters

The degree of saturation for all *Eltanin* oxygen data was calculated based on the solubility of oxygen in sea water given by *Green and Carritt* [1967]. The surface water has saturations between 95–100% with occasional observation of saturations above 100%. Surface values as high as 108% are found (station 522). Supersaturated values may be caused by rapid surface heating, observed to occur under calm summer conditions, which decreases the solubility of the surface water, causing the initial 100% saturation level to increase.

The vertical convection within the Antarctic Surface Water causes a high degree of oxygen saturation to the base of the surface water. The degree of saturation is fairly uniform from the surface to the 95% level; a sharp drop in saturation occurs below this. The 90% saturation level occurs only meters below the 95% level and can be taken as the base of free vertical convection. Figure 6 shows the depth of this surface. The depth of the convective layer south of the polar front zone rarely exceeds 200 meters. Depths of less than 100 meters are found in the stable surface water of the southern section of the Southeast Pacific Basin and the northern Weddell Sea.

The winter values tend to be slightly greater than those found in summer. In the polar front zone, large variations of depth of the 90% surface are found, owing, no doubt, to the fluctuating convergence and divergence within this zone. Within the Subantarctic Surface Water, the depths are greater than 300 meters, with values as high as 740 meters.

The lowest degree of saturation (between 50 and 60%) occurs in the vicinity of the temperature maximum and oxygen minimum. Below this oxygen saturation minimum, the saturation increases to values above 60%. Where a high percentage of Antarctic Bottom Water exists, the saturation reaches 80%. The high saturation of the bottom water, roughly intermediate between deep and surface water, indicates the recent contact with the sea surface of a substantial component. The ratio of shelf or surface water to deep water within the Antarctic Bottom Water cannot be far from unity for the bottom water leaving the Weddell Sea; this agrees with the conclusions of *Baranov* and *Botnikov* [1964] and other considerations discussed later in this study.

The upper boundary of the Antarctic Bottom Water can be seen by a rapid oxygen saturation increase close to the sea floor. Figure 7 shows the saturation versus depth of selected regions of antarctic water between 20°W and 170°W. Sections *a* to *d* are along the path of flow of the bottom water from the Weddell Sea [*Gordon*, 1966]. The rapid increase of oxygen saturation occurs approximately 1000 meters from the sea floor. Substantial amounts of bottom water are evident in sections *a*, *c*, and *d*, indicating that the main flow is east of the South Sandwich Trench. The bottom water of the trench first decreases in the degree of saturation by approximately 10% before an increase is observed to the bottom. Adiabatic warming would increase the bottom oxygen saturation level but not by the amount observed near the trench floor. Therefore some bottom water renewal occurs, but the trench floor appears not to be the major avenue of AABW flow.

Sections *e* to *g* are of stations located in the Scotia Sea. In the central Scotia Sea and northern Drake Passage where potential temperatures show little Antarctic Bottom Water, the saturation of the lower 1000 meters increases only slightly or decreases with depth. In the southern areas of the Scotia Sea which receive a fresh supply of bottom water from the passage between the Bruce and South Orkney ridges, the bottom saturation shows a marked increase. This increase begins about 800 meters above the bottom, which agrees well with the upper limits of the bottom water in the southern Drake Passage discussed by *Gordon* [1966] and used in geostrophic calculations [*Gordon*, 1967*b*].

Sections *h* to *j* in the Southeast Pacific Basin show only small increases in bottom oxygen saturation and, in the case of the southeastern region of the basin, a marked decrease. Such a decrease indicates that the bottom oxygen values are significantly influenced by organic decay and that renewal of highly oxygenated Antarctic Bottom Water is small. Section *j* in the northern Southeast Pacific Basin shows a steady increase in saturation from the minimum to the bottom. A higher percentage of Antarctic Bottom Water

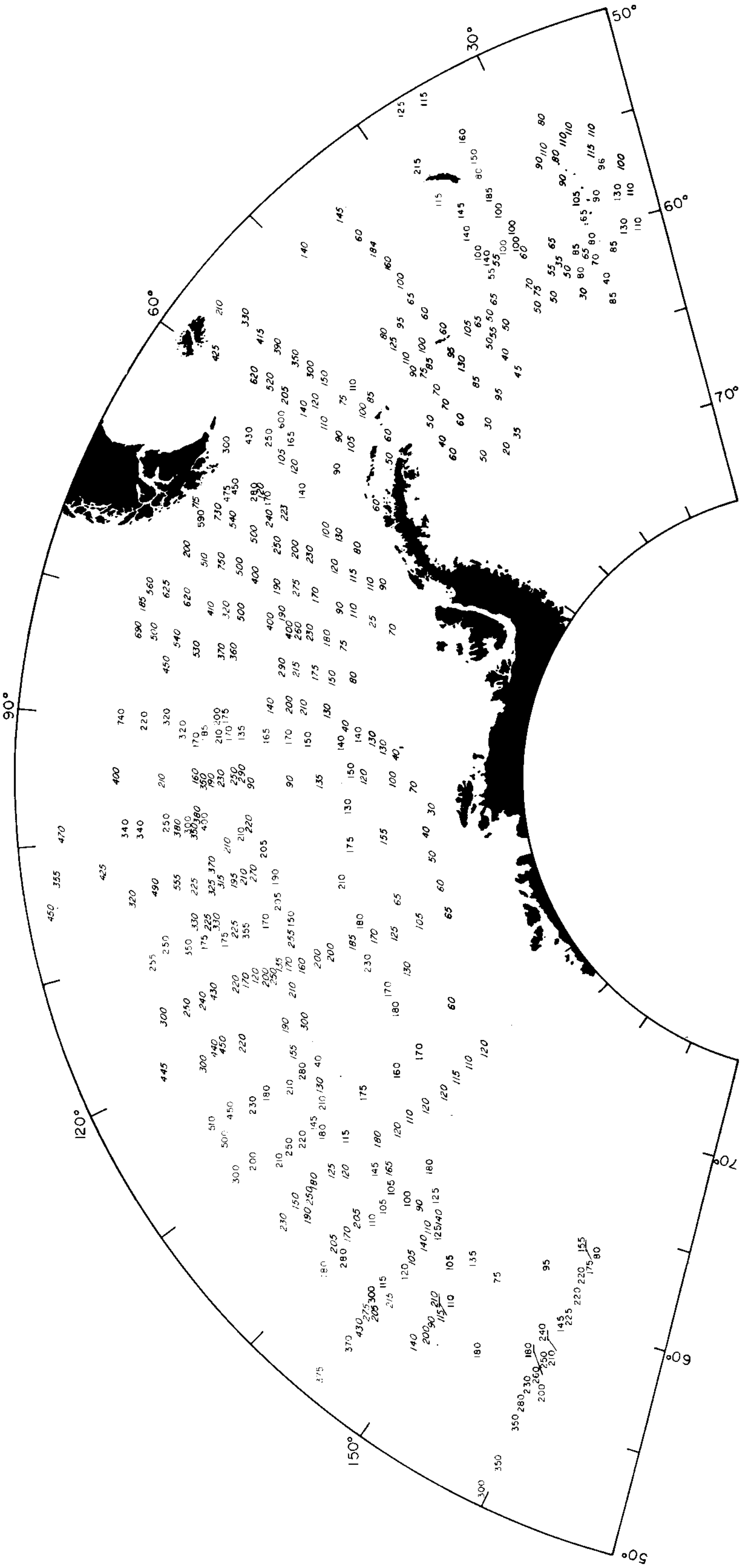

Fig. 6. Depth of 90% oxygen saturation level (values in slant letters represent summer conditions).

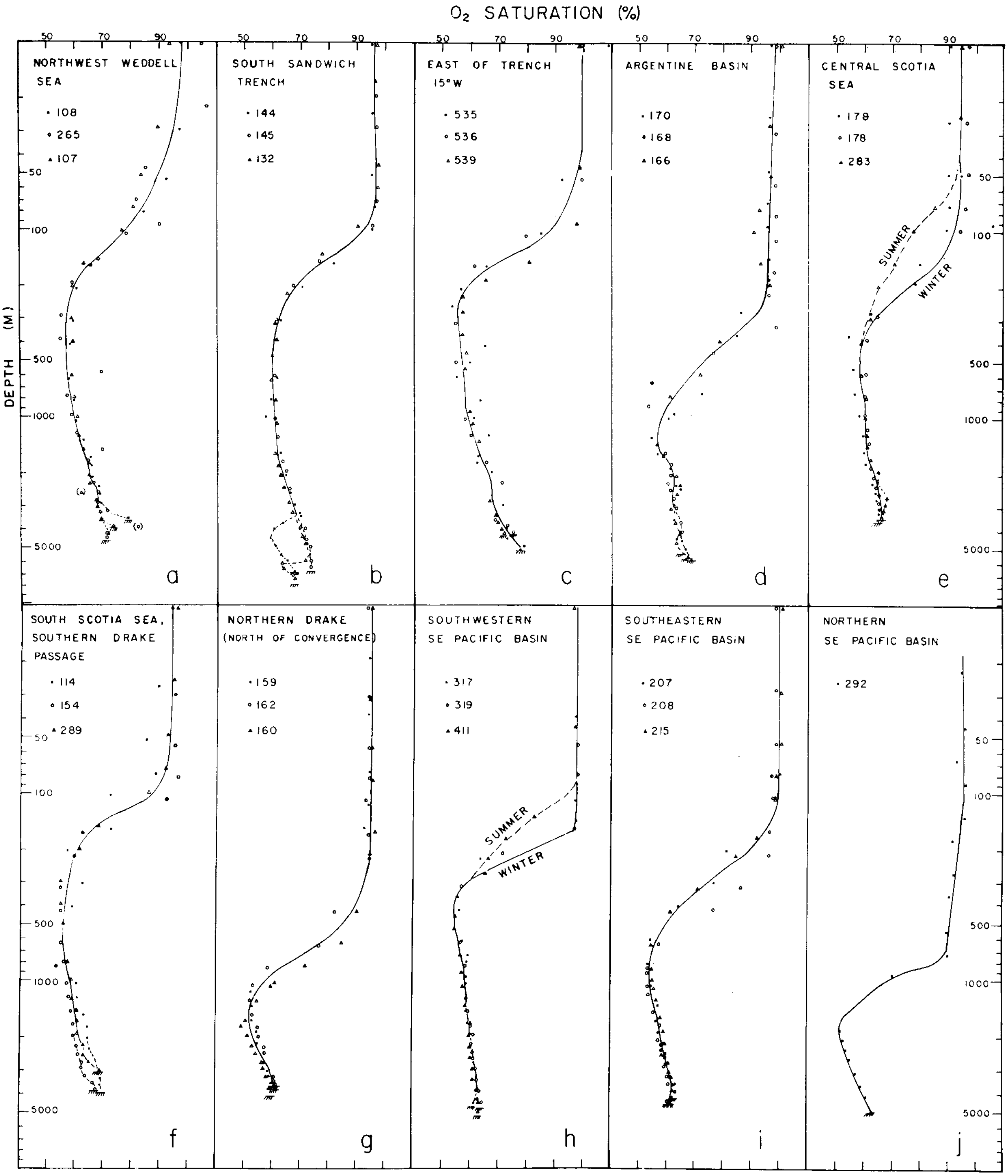

Fig. 7. Oxygen saturation versus depth for select areas of antarctic waters.

must reach this area than reaches the southeast regions of the basin. This is consistent with the bottom flow deduced by *Gordon* [1966].

Time Variation of Water Structure

Only in a general sense is the ocean in a steady state. Variability in the structure of antarctic waters must occur. It is likely that the polar front zone has periodic and nonperiodic fluctuations and may possess wavelike disturbances similar to the atmospheric polar front. Seasonal fluctuations of temperature, salinity, and circulation are also obvious. There are other variations: (1) tidal and inertia period fluctuations occur; (2) from temperature variations in the Antarctic Surface Water, it appears that a large horizontal translation of the Weddell-Scotia Confluence must exist in the surface layer; (3) it is generally accepted that the sinking of shelf water in bottom water production is a sporadic process [*Mosby*, 1968]; and (4) there are internal waves within the main pycnocline. The proper way to investigate such processes is through use of multiship operations and long-term recording sensors placed in numerous positions around Antarctica.

On September 30, 1965 (the same day as the *Eltanin* hydrographic station 473 of cruise 20), a series of bathythermograph observations was taken at approximately 15-minute intervals. The position of the observation and the times are shown in Figure 8. The ship was south of the Antarctic Convergence, which it crossed at 55°41′S and 144°32′W. The time plot of temperature versus depth is shown in Figure 9. (It is possible that the absolute temperatures are 0.3° too low.) The variations observed are no doubt a combination of time and space changes. In the first section, before the trawl, vertical migrations of isotherms, mostly just below the temperature minimum layer and at 250 meters, are evident. These may be associated with internal waves or turbulence. After the trawl, the characteristics of the structure change. The temperature minimum is not as extreme, and a fairly strong inverted thermocline occurs at 150 to 200 meters. The thermocline decreases in intensity during the last hour of observations. Between local time 1650 to 1730 and 1815 to 1845, a secondary temperature minimum layer exists above the main layer.

Repeated lowerings, such as the above experiment, would be more useful if they extended to deeper levels. This can be accomplished through the use of the continuously in situ salinity temperature-depth recorder (STD). The comparison of the up and down trace of the STD (a few hours separation) shows changes in the microstructure [*Gordon*, 1967*c*].

A study of standard level temperature, salinity, and oxygen data of relocated *Eltanin* stations at and below 2000 meters was carried out by *Jacobs* [1966]. On comparing 16 station pairs (see *Jacobs* [1966, table 2]), the stations of cruises 17–21 show a systematic difference from those of cruises 14 and 15. The average of these changes is slightly above the precision of the instruments, but the spread is too great to be significant.

A number of *Discovery* stations and *Eltanin* stations were taken along 79°W. The temperature and salinity above 1000 meters show large changes (Figure 10); below 1000 meters the variations are small and are about the equivalent to the error of the measurement. During the time of *Discovery* stations, the higher temperatures and salinities at 600 meters indicate that smaller amounts of sinking occurred at the polar front zone than at the time of the *Eltanin* stations. The 30-year spacing of the data is, of course, no indication of the periodicity of the process; however, it does show that such variations do occur.

Supercooled Water

The freezing point of water decreases with salinity; this has been shown experimentally by *Knudsen* [1903] and *Miyake* [1939]. The values found by Miyake's equation are approximately 0.08°C lower than those found using Knudsen's. The *Eltanin* data of cruises 7 to 27 were inspected for supercooled water (water below the freezing point) using both equations. There are 44 temperature observations below Knudsen's freezing point; their depths range from the surface to 890 meters, mostly between the surface and 200 meters. They are in the Ross Sea area and the waters south of the Weddell-Scotia Confluence. Using Miyake's equation, only three of these observations are below freezing. Because of the elimination of the majority of the Knudsen supercooled points using Miyake's equation, it is assumed that the latter equation is correct. At worst, it will indicate only the extreme case of supercooled water. Miyake's equation is as follows:

$$\Delta T = 0.056903\ (S - 0.030)$$

where ΔT is the freezing point depression and S is salinity in parts per thousand.

In addition to the *Eltanin* data, the hydrographic observations in the National Oceanographic Data Center marsden square files from 10°W westward to 180° and south of 50°S were inspected, as well as the Deep Freeze '63 and '64 data in the Ross Sea. The position of

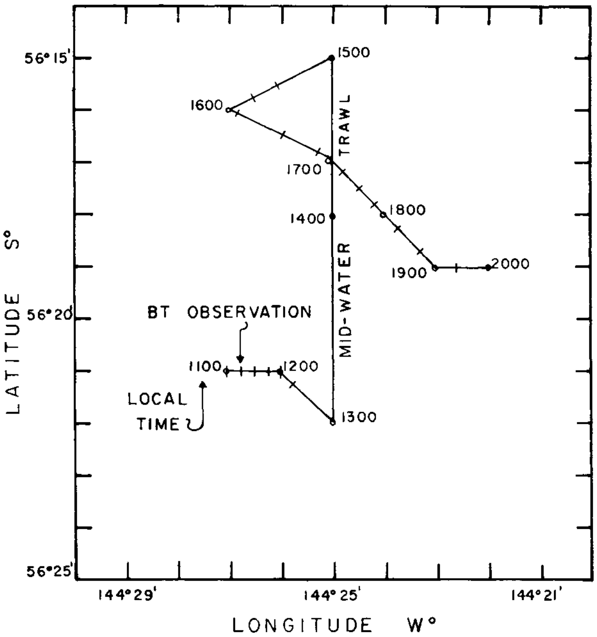

Fig. 8. Hourly *Eltanin* positions during repeated bathythermograph observations on September 30, 1965.

the stations showing supercooled water and the depths at which this water was found are shown in Figure 11. (The data of *Littlepage* [1965] in the McMurdo Sound are not included.) The data are clustered in the southern Weddell Sea and Ross Sea. Only the two *Eltanin* stations in the northern Weddell Sea are anomalous both in position and depth of the observation. They are assumed to be in error and are not included in the discussion below.

The *T/S* points of the supercooled temperature observations are shown in Figure 12. The salinities are fairly high, equivalent to that of the upper deep water. Therefore, if these temperatures were achieved at the sea surface where salinities are low, they must have been accompanied by freezing. In this case, it is difficult to imagine why supercooled water would occur, since it is probable that there would be sufficient nuclei and turbulent motion to produce freezing. It appears reasonable to assume that the water was not produced at the sea surface.

The depression of the freezing point of pure water with pressure is 0.075°C per 100 meters of water depth [*Lusquinos*, 1963]. Assuming that sea water also is affected in the same magnitude, it is found that none of the temperature observations shown in Figure 11 are supercooled. The temperature versus depth of the observations and depression of the freezing point with depth (34.65‰) is shown in Figure 13. It is reason-

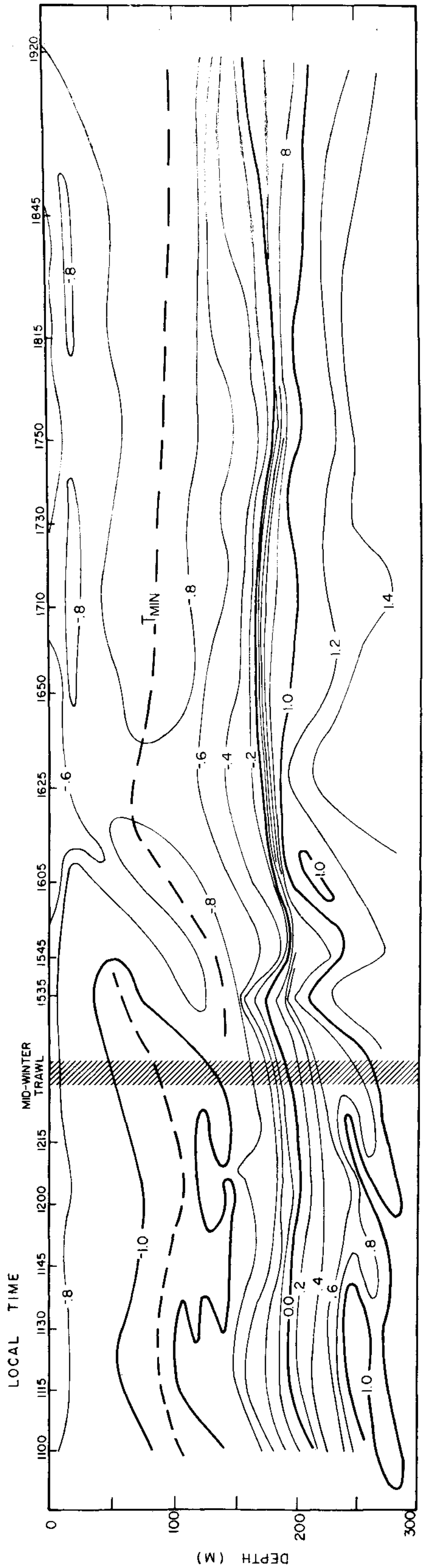

Fig. 9. Temperature versus depth of repeated bathythermograph observations along the ship's track shown in Figure 8.

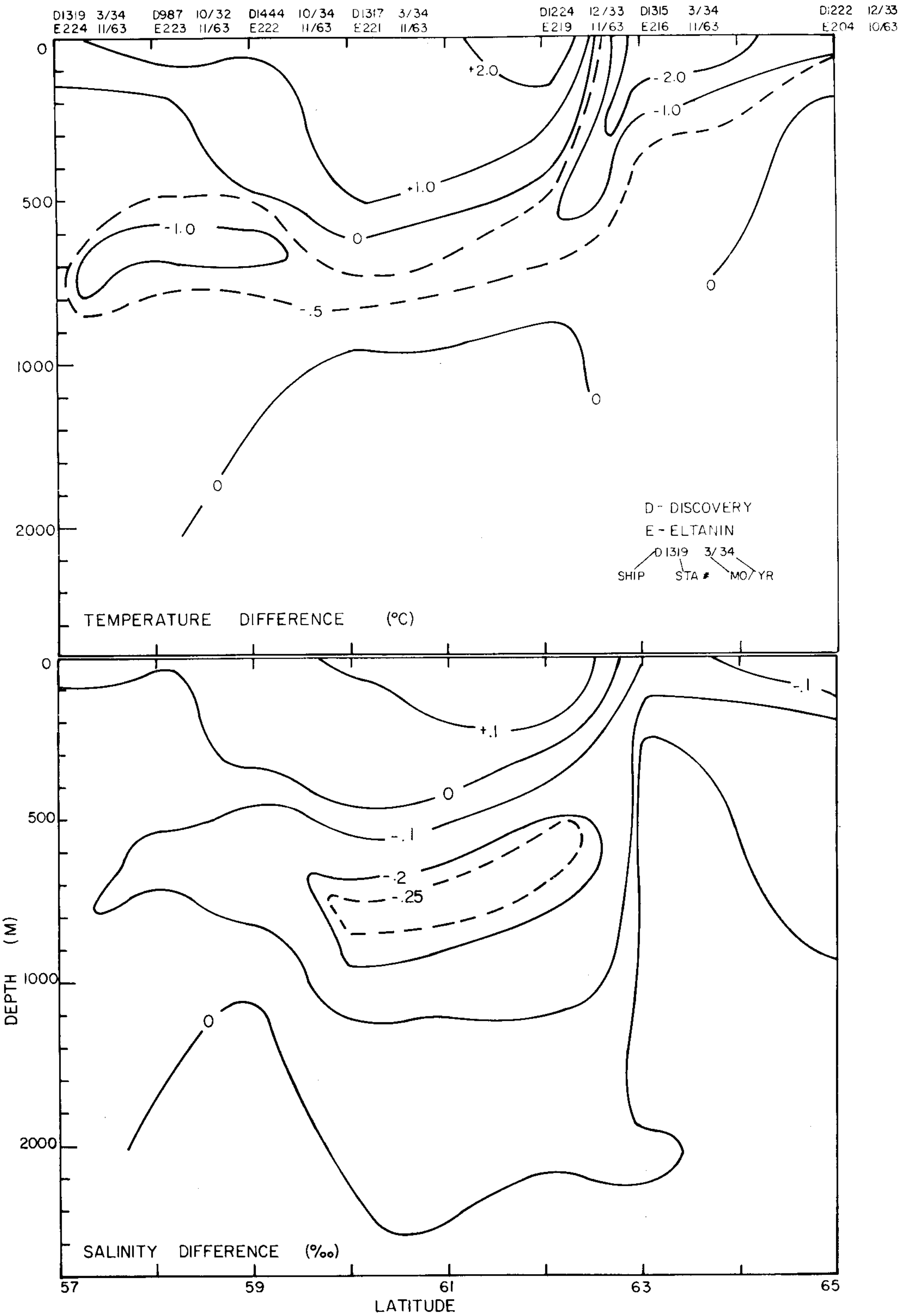

Fig. 10. Salinity and temperature differences of *Discovery* and *Eltanin* data in Southeast Pacific Basin (upper 1000 meters).

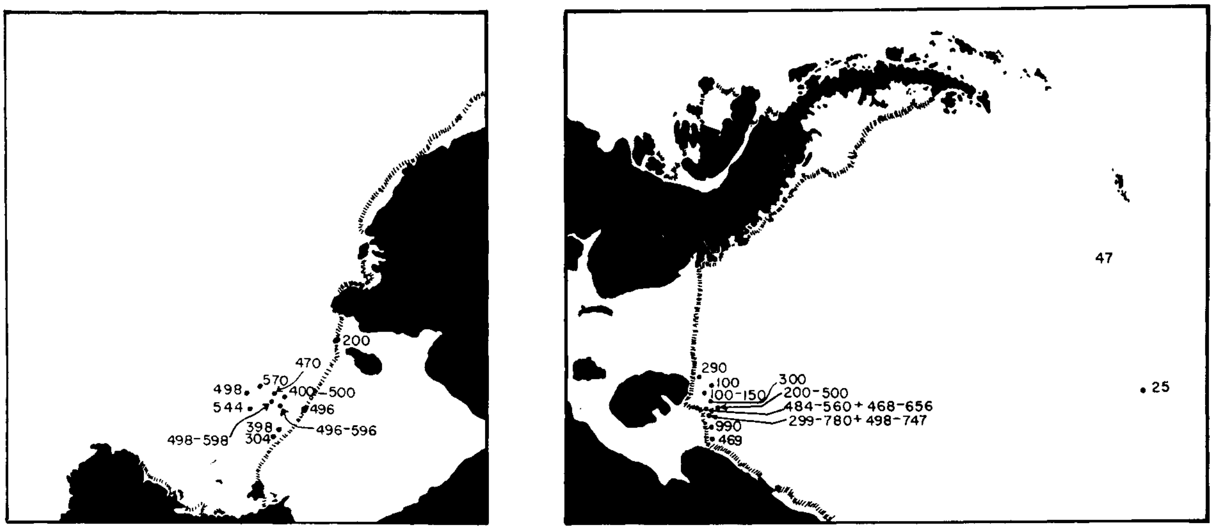

Fig. 11. Position and depth of supercooled sea water.

able to assume, as Lusquinos does, that the cold temperatures are produced below the sea surface and that no supercooling occurs. It has been suggested [*Littlepage*, 1965] that the salinity of the water is higher than that determined on the deck of the ship because of the suspension of small ice crystals that melt before the salinity of the sea water is determined. In this case, in situ measurements of salinity must be used. However, data of the STD aboard the *Eltanin* show that the in situ salinity of supercooled water is the same as the shipboard determined salinity. (See stations 648 and 649 of *Jacobs and Amos* [1967].) There is a pos-

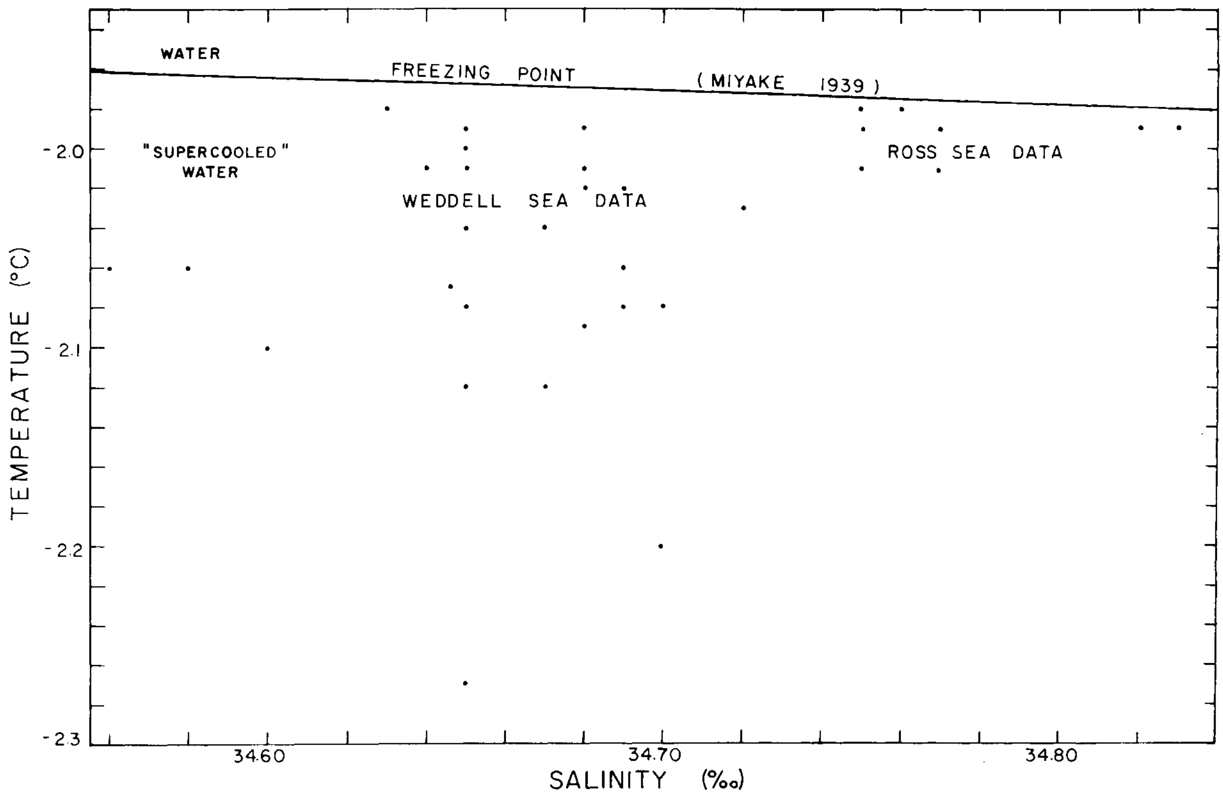

Fig. 12. Temperature-salinity diagram of supercooled water.

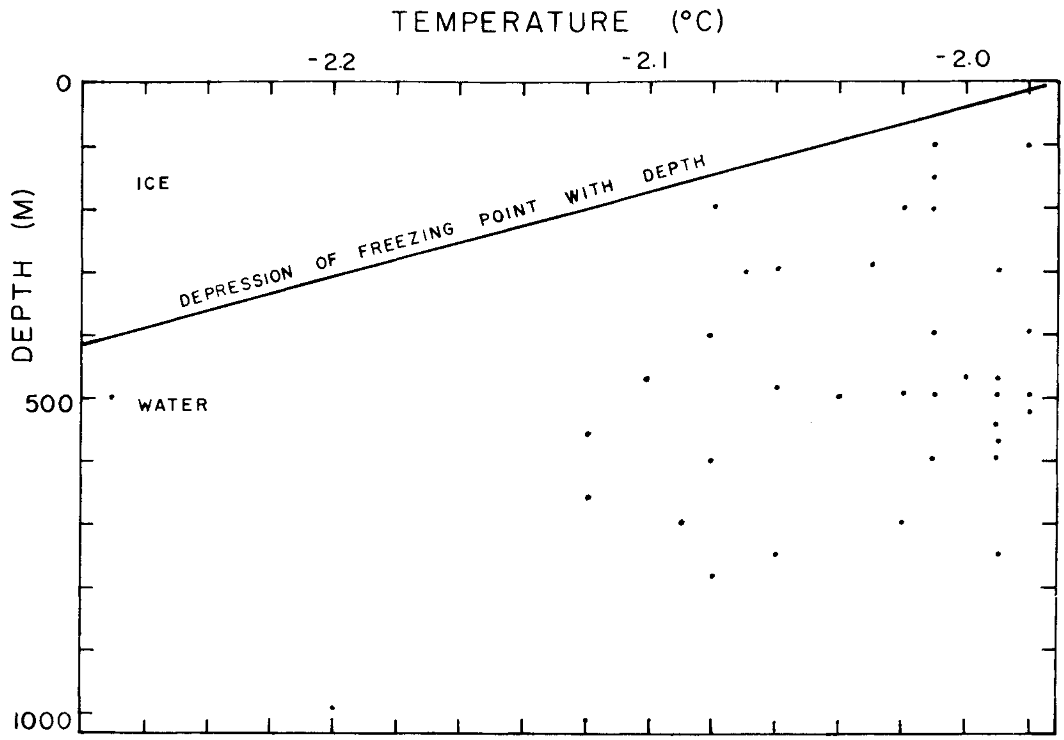

Fig. 13. Sea water freezing point versus depth and depth of observed supercooled water.

sibility that the STD is affected by the ice crystals in such a way as to give results similar to the shipboard value, but this is unlikely.

The striking similarity of the cold temperatures is their proximity to an ice shelf, and a minimum depth of 100 meters, with most of the observations at 400 to 600 meters. The undersides of the edge of the Ronne and Ross Ice Shelves are approximately 200 meters below sea level, increasing to over 500 meters before the ice contacts the sea floor [*Zumberge and Swithinbank,* 1965]. Both melting and freezing would yield water below the one atmosphere pressure freezing point of sea water. However, the salinity of the melted water would be lower than that salinity of the brine released in freezing. The salinity of the upper 200 meters of sea water along the ice front is only slightly above 34.0‰, which is below that of the supercooled water. Therefore, it is reasonable to conclude that the cold water is derived from freezing at the underside of the ice shelves. It is interesting to speculate that the melted water from the lower part of the ice shelf (15–20 g/cm^2/yr [*Shumsky and Zotikov,* 1963]) would rise to the sea surface (owing to its lower density), where because of the lower freezing point it would freeze. However, it is not known if there is upwelling or sinking along the ice front; *Thomas* [1966] believes that sinking occurs. Sinking would supply the water that freezes to the bottom of the shelf ice. The resulting higher saline water drops to the sea floor under the ice and flows northward, contributing to the Ross Sea Shelf Water.

Ross Sea Shelf Water

Of special interest, partly related to the above, is the saline water that fills the topographic depressions of the Ross Sea. (See the topographic map of Antarctica (1965) prepared by the American Geographical Society.) A dense network of hydrographic stations was tested during Deep Freeze '63 and '64 by the USS *Edisto* and USS *Atka* [*Countryman and Gsell,* 1966]. These data will be used in this discussion. In the *T/S* diagrams of the Ross Sea (see Figure 5) the shelf water is found in the *T/S* region from 34.75 to 35.00‰ and −1.8°C to slightly below −2.0°C [*Countryman and Gsell,* 1966]. The density (σ_t units) is above 28.0, which is greater than that of the bottom water of the open ocean. The interval from 34.75 to 34.80‰ is a transition with the shelf water; therefore, 34.80‰ will be taken as the upper boundary. The depth of this surface and the temperature on this surface are shown in Figure 14. The depths to the 34.80‰ isohaline surface are smallest in the extreme

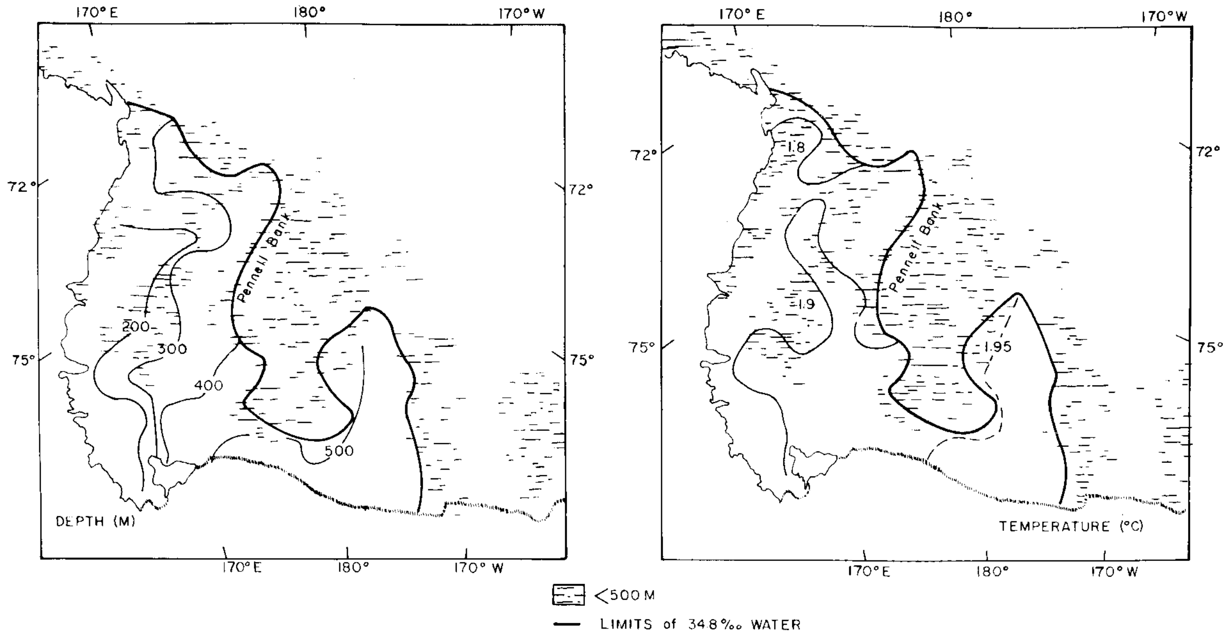

Fig. 14. Depth and temperature of upper boundary of the Ross Sea Shelf Water.

southwestern corner of the Ross Sea, coming to within 100 meters at the sea surface. The depth increases rapidly toward the east. The Pennell Bank protrudes through the shelf water. The thickest section of shelf water is over the three north-south oriented troughs of the southwestern Ross Sea. Since the 34.8‰ isohaline surface reaches bottom at the northern end of these troughs, it is probable that a blocking sill exists here. The slope in the shelf water may be in equilibrium with a northwest flowing geostrophic current.

The temperatures on the 34.8‰ surface are mostly near −1.9°C, although the coldest temperatures (near −2.0°C) are found at its southeastern limits. These temperatures are for the most part at or slightly below the *Miyake* [1939] freezing point values.

In the area shown in Figure 15 a temperature inversion is found in the lower hundred meters (average is 190 meters above the sea floor). The T/S diagram of four sample stations in this zone is shown in Figure 16. The in situ temperature is near −1.95°C (or in some stations −2.0°C) approximately 200 meters above the sea floor. Below this, the temperature increases by as much as 0.1°C, with an average increase of 0.05°C. The salinity also increases to the bottom in all cases. It is possible that the warmer bottom water represents a segment of the shelf water which is trapped in a topographic depression and is warmed by the geothermal heat flux. The volume of the shelf water below the temperature minimum layer is 7.7×10^{18} cm^3. Assuming normal continental shelf heat flow of 40 cal/cm^2/yr (M. Langseth, personal communication) and no water influx or discharge, this water must have been trapped in the depression for a period of ten years. Therefore, the lower part of the Ross Sea Shelf Water may be confined in the shelf depression for a number of years, while the upper part of the shelf water and the temperature minimum layer most likely have a much more active rate of renewal.

The formation of the highly saline shelf water is no doubt due to freezing, either at the sea surface or from below the Ross Ice Shelf. The presence of water with temperatures near −2.0°C indicates (as discussed above) an ice shelf origin. However, most of the shelf water is near or higher than the freezing point (the freezing point after Miyake for water of 34.85‰ is −1.98°C); therefore, a sea ice origin for at least part of the shelf water is not ruled out. A schematic of the possible origin for the bulk of the Ross Sea Shelf Water is shown in Figure 17.

Antarctic Bottom Water Production

The method of production of Antarctic Bottom Water is not exactly known, although it is generally accepted that the characteristics of the bottom water are attained by a mixture of surface water and deep water [*Brennecke*, 1921; *Mosby*, 1934; *Defant*, 1961]. It is not known if the surface component comes from the continental shelf regions, where owing to the shallow bottom, sufficiently dense surface water can form [*Brennecke*, 1921; *Mosby*, 1934, 1968] or it is de-

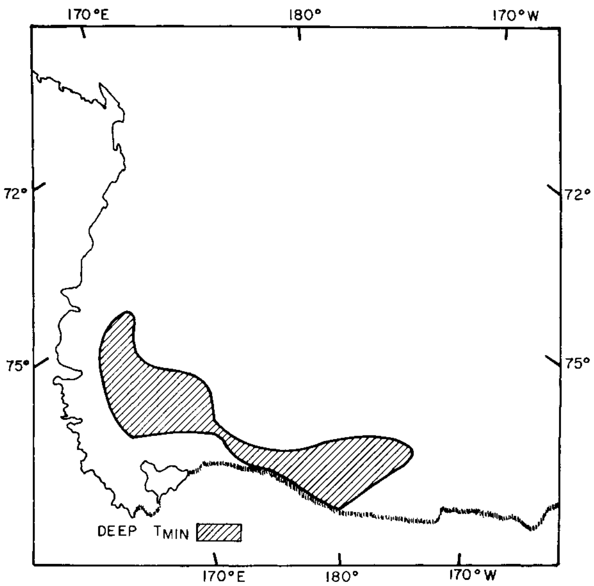

Fig. 15. Areal extent of deep T_{min} layer within Ross Sea Shelf Water.

rived directly from deeper oceanic areas (shown in schematic diagram by *Munk* [1966]). Either way the surface water temperature must decrease to near freezing and the salinity must increase, creating a surface density high enough to initiate sinking. Such conditions exist during ice formation periods. The period

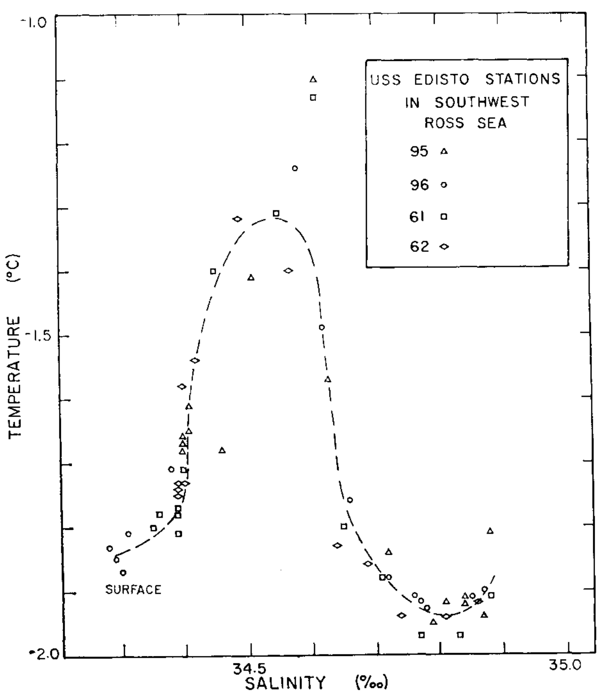

Fig. 16. Temperature-salinity diagram of Ross Sea Shelf Water.

of most rapid formation is early winter (April to June) [*Neuman and Pierson*, 1966, fig. 4.10]. After June, the limit of the pack ice continues to advance slowly until September. As the ice thickens, the lower heat conduction of the ice slows the rate of thickening. However, breaks in the ice cover caused by icebergs flowing through the ice field or divergence in the wind or currents expose the water surface and ice forms rapidly. Such breaking of the ice field is fairly common in the Weddell Sea [*Heap*, 1964]. Therefore, the most rapid cooling and salinity increase of the surface water occurs in early to mid-winter, but the accumulating effect is to produce a late winter density maximum of the surface water; hence the bulk of bottom water production may occur in late winter [*Mosby*, 1934], when the critical salinity of 34.63‰ is attained on the shelf [*Fofonoff*, 1956].

Alternative methods for formation of bottom water involve evaporation of surface water, which leads to effects similar to ice formation [*Ledenev*, 1961] or freezing of sea water onto the bottom of the ice shelves. It is unlikely that either of these methods can account for the vast quantities of Antarctic Bottom Water observed through much of the world ocean (see below).

Fofonoff [1956] (also see *Mosby* [1968]) suggests that nonlinearity of the equation of state of sea water can result in the formation of water denser than either component and may explain Antarctic Bottom Water production from shelf water and deep water components. He shows that surface water of the Weddell Sea with salinities between 34.51 to 34.63‰ on mixing with the Weddell Deep Water will form water denser than either component and, therefore, surface water of these salinities is not present except where the surface water is not directly above the warmer deep water. On the continental shelf the water attains the high salinity necessary to produce the dense Antarctic Bottom Water. Other areas around Antarctica can be investigated by a similar technique.

For select areas around Antarctica (Figure 18) the observed T/S points within the temperature minimum layer and temperature maximum layer are plotted.

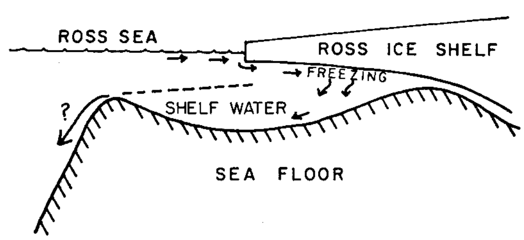

Fig. 17. Schematic of origin of Ross Sea Shelf Water.

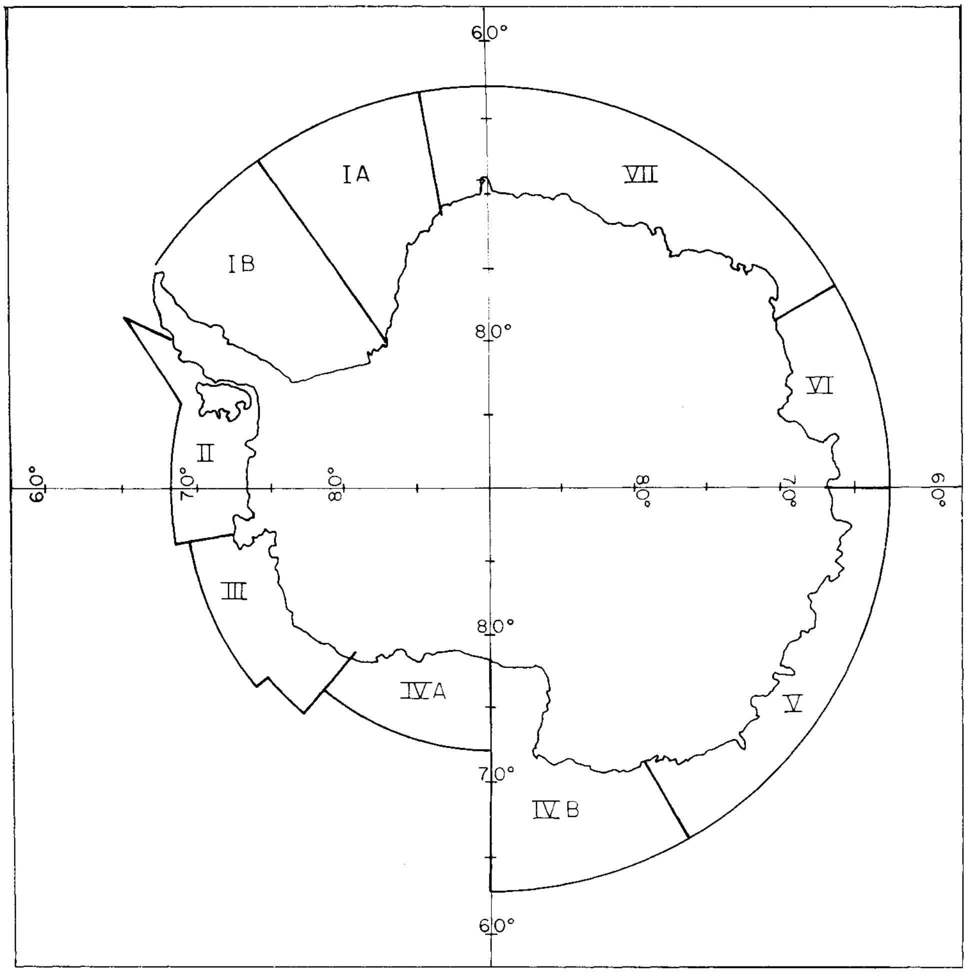

Fig. 18. Select areas for T_{max}/salinity and T_{min}/salinity diagrams.

The water in the temperature minimum layer is assumed to form the previous winter and so represents the densest surface water which has remained at the surface. Surface water which reached a higher density has dropped out of the surface layer and contributed in part to the Antarctic Bottom Water, and deeper transition zones. A tangent is drawn to the sigma-t at the mean T/S point of the T_{max} layer; in addition, a line connecting this point to a point on the freezing line with the same density is drawn. When salinities of the surface water are less than that at the intersection of the first line with the freezing point line, no mixture with the deep water can result in a denser water. Between this salinity and the salinity at the intersection of the second line with the freezing point, certain mixtures are denser than either component. As the salinity increases toward the second point, a smaller component of deep water is necessary to attain a denser solution. At salinities higher than the second point, sinking must occur. A decision can be made as to the probability of production of bottom water in the areas shown in Figure 18 by observing how closely the salinity of the temperature minimum layer comes to the first critical salinity.

Figure 19 shows the T/S plots and the two critical salinities for each area. Only areas I B (the western Weddell Sea) and IV B (the area northwest of the Ross Sea) show a high likelihood of bottom water production. Areas II and III (the Bellingshausen and Amundsen seas) show a very low likelihood of bottom water production. The other areas show fair possibilities.

When the salinity of the surface water at the freezing point increases above the first critical salinity value, only intense vertical mixing would cause conditions permitting production of water denser than either the surface or deep water. As the salinity increases further, the necessary vertical mixing decreases. There-

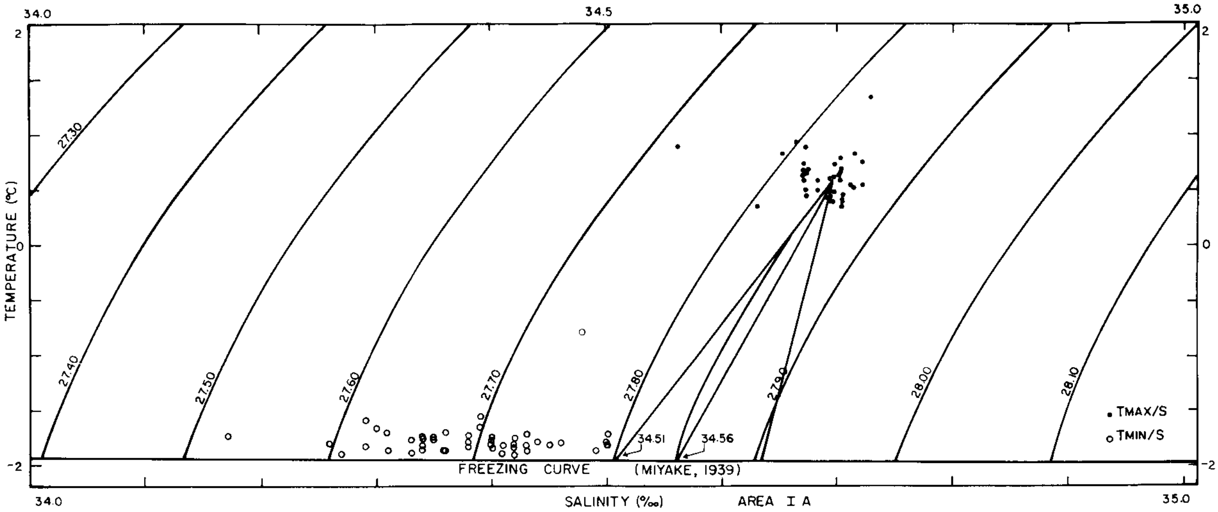

Fig. 19*a*. T_{max}/salinity and T_{min}/salinity diagrams.

fore, in the open sea where turbulence and vertical mixing in the surface layer are probably high owing to waves and wind, the surface water is quickly mixed, and high salinities in the T_{min} layer are not attained. However, where vertical mixing is small, the surface water can attain high salinities before sinking. For this reason, bottom water may be more likely to occur in areas somewhat sheltered from the open ocean turbulence, such as the Weddell Sea, the Ross Sea, and the indented bays around Antarctica.

Antarctic Bottom Water may have some contribution from shelf water produced by freezing at the bottom of the ice shelves, as discussed in the previous section. *Mellor* [1963] believes that 7.8×10^{17} g/yr of ice is discharged by the ice shelves of Antarctica. If the ice shelves are in equilibrium and the total volume is produced by bottom freezing, 1.2×10^6 m³/sec of shelf water would be produced (−2.0°C and 34.9‰). This water would have to be diluted by at least 2 parts of the water found at the depth of the continental shelf break around Antarctica to produce water similar in temperature and salinity to Antarctic Bottom Water. Therefore, less than 4×10^6 m³/sec of bottom water would be formed. This value would increase as it flows northward, owing to mixing with deep water. However, it is not probable that such a mechanism can explain the total amount of Antarctic Bottom Water, since 20×10^6 m³/sec leaves the Weddell Sea [*Baronov and Botnikov*, 1964], and it is not reasonable to assume that all the ice shelves are produced by

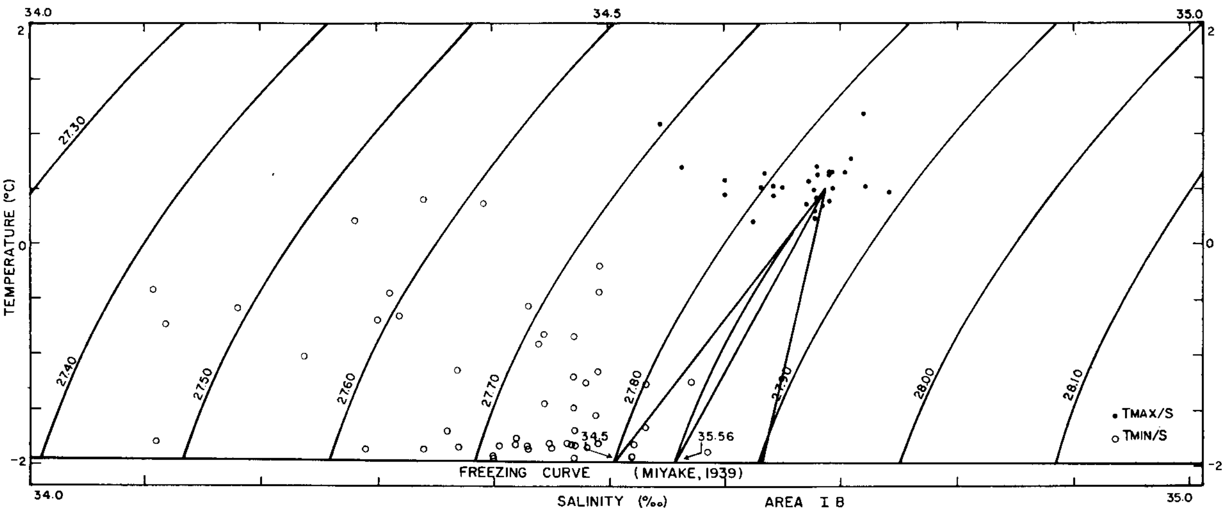

Fig. 19*b*. T_{max}/salinity and T_{min}/salinity diagrams.

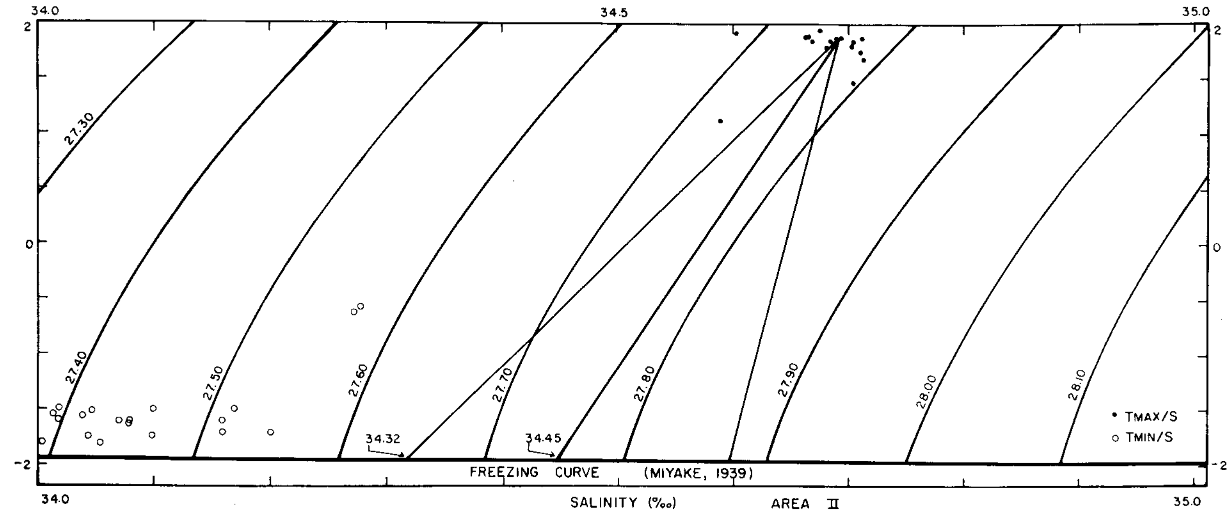

Fig. 19*c*. T_{max}/salinity and T_{min}/salinity diagrams.

freezing at the underneath side; however, such a mechanism may be of local importance, as in the Ross Sea.

Deacon [1937] has recognized the Ross Sea depression as a trap for catching the densest water produced at the sea surface; for this reason, the Ross Sea may be a poor source of Antarctic Bottom Water. *Mosby* [1968] suggests that the dense shelf water may act as an artificial sea floor, and the cold Ross Sea water may override it to reach the open ocean and contribute to bottom water.

It is conceivable that some of the dense shelf water does exist at the mouth of the troughs. This is expected during the winter when more shelf water may be formed or during periods when internal waves may cause 'blobs' of shelf water to escape in a manner similar to the formation of North Atlantic Deep Water [*Steele et al.*, 1962]. A likely place for this escape is the deep canyon cut into the continental slope near 178°E. However, no indication is found in the Deep Freeze '63 and '64 data. The non-steady-state character of such an overflow may be the reason. This canyon is a good location to place continuously recording bottom current meters.

There is evidence in the deep ocean that the high saline shelf water does escape and contributes to Antarctic Bottom Water. The bottom water of the southwest Pacific Ocean is a secondary salinity maximum, i.e., the salinity below the lower Circumpolar Deep Water first decreases, but on approaching the bottom

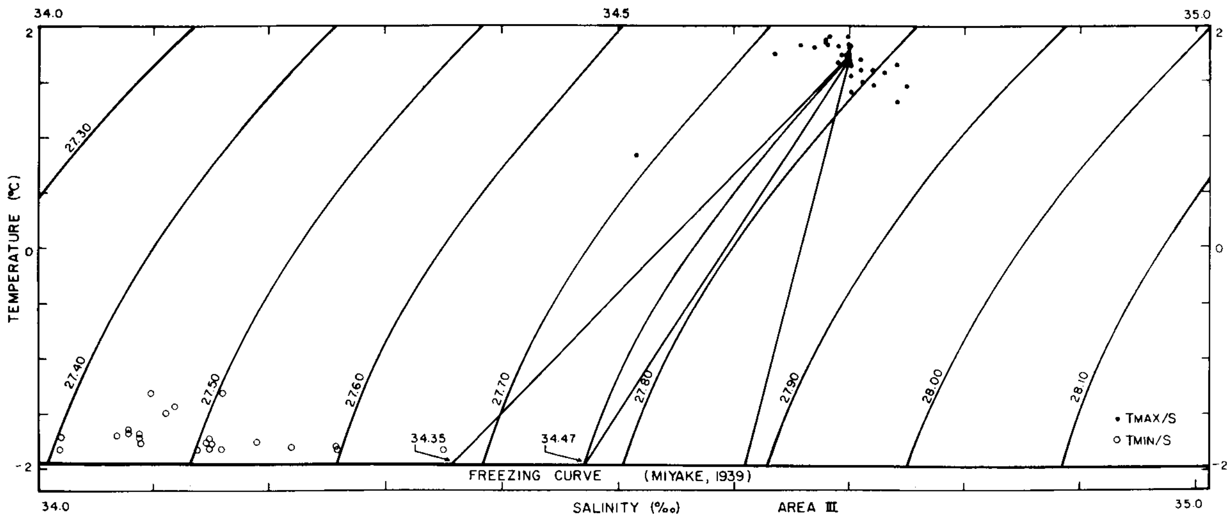

Fig. 19*d*. T_{max}/salinity and T_{min}/salinity diagrams.

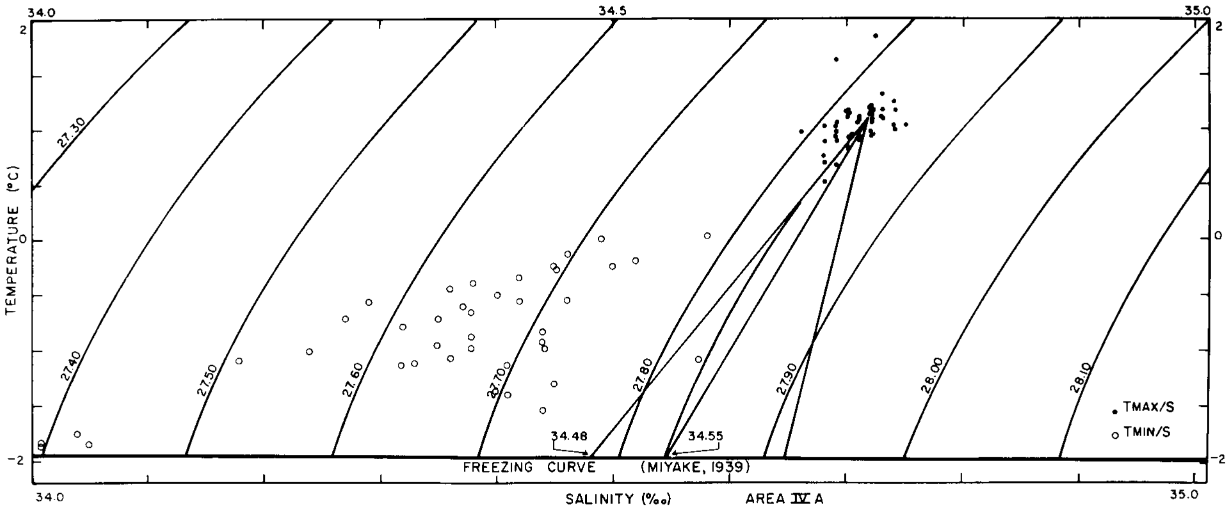

Fig. 19e. T_{max}/salinity and T_{min}/salinity diagrams.

increases. In the Weddell Sea, the salinity decreases slowly from the lower Circumpolar Deep Water to the sea floor. This difference is indicated on the curves of the deep water and bottom water shown in Figures 4 and 5. The curve of Figure 5 has an inflection point at temperatures about 0.3°C; this does not occur in Figure 4. The bottom salinity maximum and the Ross Sea Shelf Water intrusion to the high salinity part of Figure 5 suggest that the two water masses have a relationship. The USNS *Eltanin* cruise 37 (January–February, 1969) along the coastal area west of the Balleny Islands supports the view of a high salinity (> 34.72‰) Ross Sea outflow along the sea floor. This cruise will be discussed in future reports. The oceanography of the Ross Sea is discussed in a study by S. S. Jacobs and others (in preparation).

DEEP WATER MERIDIONAL TRANSPORT

Terms of Salt Balance Equation

The fresh water is introduced to Antarctic Surface Water from two sources: the excess of precipitation over evaporation and the continental runoff. To maintain the salt content of the surface water, it is necessary for deep water to upwell to the surface layer. Part of the fresh water input is seasonal in that the continental runoff is accomplished during the summer months. This, in addition to the alternating freezing and melting of the antarctic pack ice cover, causes seasonal surface salinity variation. If no deep water upwelled, the seasonal variations of surface salinity would fluctuate about a base line of decreasing salinity.

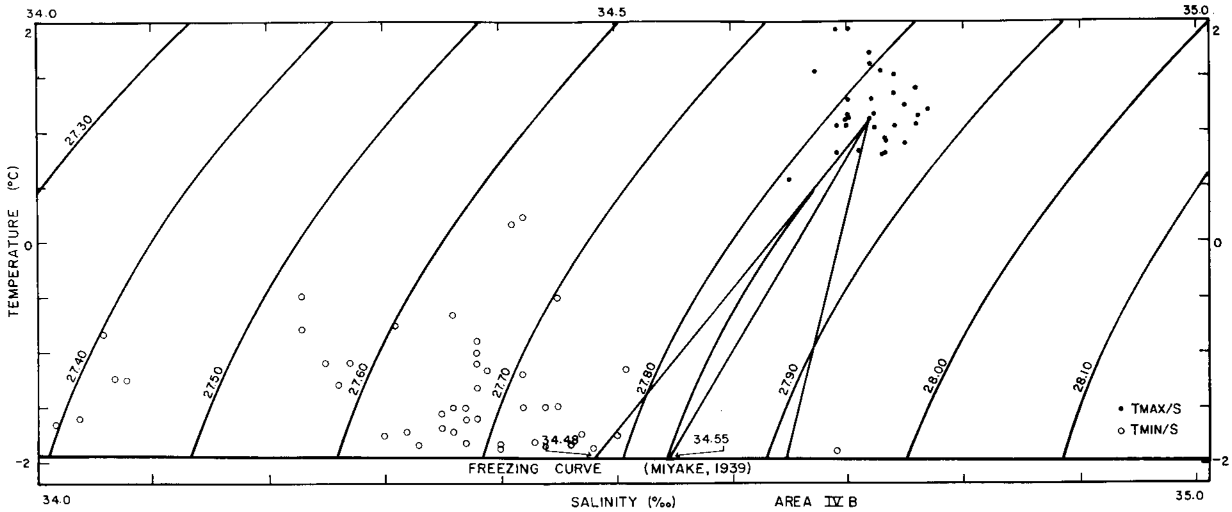

Fig. 19f. T_{max}/salinity and T_{min}/salinity diagrams.

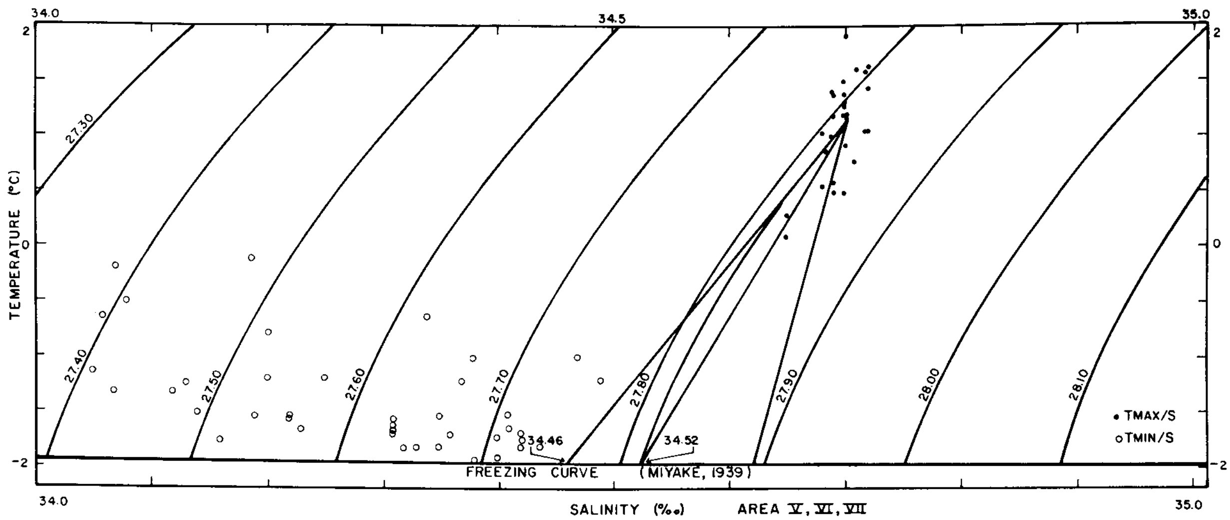

Fig. 19g. T_{max}/salinity and T_{min}/salinity diagrams.

The excess of precipitation over evaporation for all latitudinal belts has been estimated by *Wüst et al.* [1954]. From these data, the average value over antarctic waters is found to be approximately 40 cm for a period of one year. The area of the Antarctic Surface Water from the southern limits of the Antarctic Convergence near 56°S to the continent is 20×10^{16} cm^2; therefore, 8.0×10^{18} g of fresh water is introduced to the sea from the atmosphere each year.

The continental runoff includes ice calving, melting of ice shelves, blown snow, and fresh water runoff. An estimate can be made based on the annual precipitation minus evaporation and sublimation (net accumulation) and the assumption that the volume of the antarctic ice sheet remains constant. This appears to be a reasonable assumption [*Robin and Adie,* 1964; *Giovinetto,* 1964; *Gow,* 1965]. The net accumulation estimated by Giovinetto is $(2.1 \pm 0.4) \times 10^{18}$ g/yr. His calculations are based on measurement within ten drainage systems (including ice shelves), totaling 13.6×10^6 km^2, or 97% of the total areal extent of Antarctica. Therefore, approximately 2.1×10^{18} grams of water are transported directly from Antarctica to antarctic waters each year. *Mellor* [1963] estimates the iceberg discharge of ice shelves and continental ice as approximately 0.8×10^{18} g/yr (7.8×10^{17} from ice shelves and 0.37×10^{17} from unchanneled continental ice). *Shumsky et al.* [1964] use the value of $(1.63 \pm 0.1) \times 10^{18}$ g/yr, or twice that of Mellor. Various earlier estimates of the amount of calving [*Wexler,* 1961, table 1] range from 0.04×10^{18} to 1.21×10^{18} g/yr. A value of 1×10^{18} g/yr for iceberg discharge leaves 1.1×10^{18} g/yr to be discharged by melting at the ice-sea boundary, fresh water runoff, and blowing of snow off the continent. Most of the 10^{18} g/yr of ice that leaves Antarctica melts before reaching the Antarctic Convergence [*Tolstikov,* 1966, pl. 124]. For the purpose of these calculations, it is assumed that all the ice melts within the Antarctic Surface Water. In addition, all the ice-to-water phase change of the continental runoff is accomplished during the six summer months.

The amount of ice melting and freezing each year is estimated from variation in pack ice boundaries observed from ships and Nimbus 1 satellite views to be 2.3×10^{19} g [*Munk,* 1966]. Unlike the precipitation and runoff estimates, this value is fairly well agreed upon.

Summer Months

The water balance equation for the period October to March is as follows:

$$M_1S_1 + M_2S_2 + M_{ice}S_{ice} + M_{ASW}S_{AW} = [M_1 + M_2 + M_{ice} + M_{ASW}]\, S_{AS} \qquad (1)$$

where

M_1, S_1 = mass and salinity of the fresh water input;

M_2, S_2 = mass and salinity of deep water entering the Antarctic Surface Water;

M_{ice}, S_{ice} = mass and salinity of melting water from pack ice;

M_{ASW} = mass of Antarctic Surface Water;

S_{AW} = salinity of Antarctic Surface Water at the end of September;

S_{AS} = salinity of Antarctic Surface Water at the end of March.

The value of M_{ASW} used for these calculations is based on a surface water areal extent south of the southern limits of the convergence of 20×10^{16} cm^2 and a depth of 100 meters, which is a reasonable average found from the depth of the 90% O_2 saturation surface. The S_{AW} and S_{AS} values are average values in the upper 100 meters of antarctic waters. During the summer months, it is expected that the entire annual continental runoff and seasonal melting of pack ice are accomplished. A further assumption is made: the precipitation over evaporation is constant throughout the year.

A further consideration is that during the initiation of pack ice melting in spring small pockets of brine drop out, enriching the salt content of the surface water. This process is counteracted by the accumulation on the pack ice of snow accumulated by precipitation and blown snow off Antarctica. The snow melts in spring when the pack ice becomes porous or 'rotten.' For the purpose of the salt budget approximation, it is assumed that these two processes cancel each other.

Solving equation 1 for M_2:

$$M_2 = \frac{M_{ASW}(S_{AS} - S_{AW}) \quad + M_{ice}(S_{AS} - S_{ice}) + M_1 S_{AS}}{S_2 - S_{AS}} \quad (2)$$

Using $S_1 = 0$, $S_2 = 34.6‰$ (approximate salinity of the upper Circumpolar Deep Water) an ice salinity of 4.0‰ [*Serikov*, 1965; *Neumann and Pierson*, 1966], M_2 can be found as a function of ΔS, the sea sonal variation of salinity of the upper 100 meters of the Antarctic Surface Water (Figure 20). Where S_{AS} or S_{AW} appear separately, the values 33.8 and 34.0 are

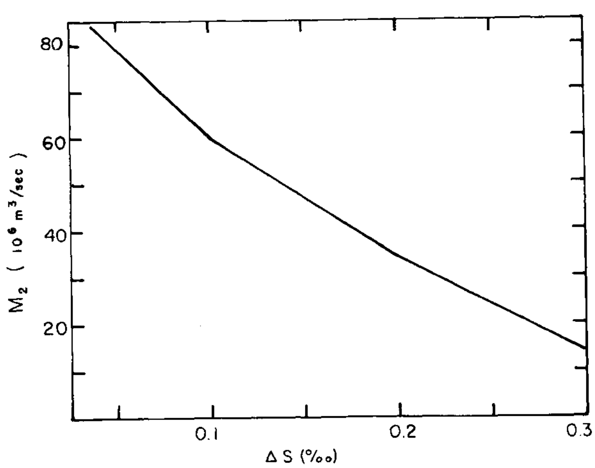

Fig. 20. Amount of necessary upwelling, M_2 versus salinity, seasonal variations of antarctic surrounding water, Δs.

used, respectively. Since ΔS is most sensitive in determining M_2, the relation shown in Figure 20 is a good approximation. Obviously, the value of ΔS is very critical to the outcome, in that the necessary upwelling of deep water into the surface layer varies from 34 to 80×10^6 m^3/sec, as the surface salinity difference decreases from 0.2‰ to 0.05‰.

The average salinity in the upper 100 meters of Antarctic Surface Water for all hydrographic stations in the western hemisphere south of the convergence is 33.9‰ for March and 34.0‰ for September. Therefore, ΔS is 0.1‰, and M_2 is 60×10^6 m^3/sec.

The average March and September salinities for the upper 100 meters show considerable regional variation. The Weddell Sea values are generally 0.2‰ to 0.3‰ higher than the Pacific values. However, the seasonal change at any particular geographic area can best be approximated by 0.1‰. The value calculated for M_2 is, of course, a rough approximation; estimate of the error based solely on the uncertainty in the ΔS value would be $\pm$ 20×10^6 m^3/sec. This, in addition to uncertainties in the value for precipitation minus evaporation, continental runoff, and pack ice fluctuations, indicates that 60×10^6 m^3/sec should be used with reservation. However, it does indicate that the upwelling of deep water into the surface layer is most probably between 40 to 80×10^6 m^3/sec.

The value of M_2 is not the entire southward transport of deep water. Additional quantities are needed to balance that which is entrained in the northward flowing bottom water, and that which mixes with antarctic shelf water in bottom water formation.

It is not expected that the deep water upwells into the surface layer around Antarctica. In certain regions, more intense upwelling would occur. Here surface waters would contain the highest percentage of deep water; hence, they would be relatively high in salinity and nutrients. Figure 21 shows the surface trace of the 34.0‰ isohaline [*Deacon*, 1937]. In general, the surface salinity is above 34.0‰ south of 65°S. In three regions, this isohaline penetrates northward: the Weddell Sea, Ross Sea, and Kerguelen Island regions. From the higher salinities, it is probable that upwelling is concentrated in these regions. (From the group *T/S* diagrams, it is found that upwelling is intense in the Weddell gyre and the area northeast of the Ross Sea.) It is interesting to point out that in these regions meridionally oriented ridges occur, vast ice shelves exist along the coast, and the first two are bottom water sources; the third is suspected as being a source area. Such an occurrence is

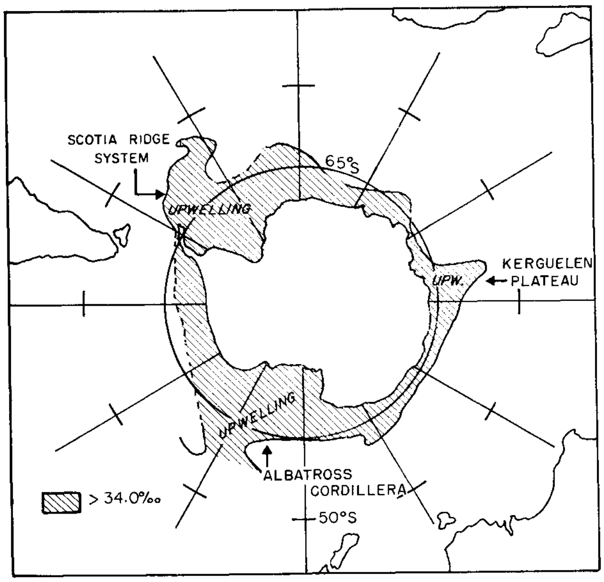

Fig. 21. Position of surface 34.0‰ isohaline (from *Deacon* [1937]).

no doubt due to the ridges that prohibit the warm circumpolar water from penetrating southward and permit extensive pack ice fields to form.

Upwelling in the vicinity of ridge crests would produce higher productivity of surface water within and downstream of these regions. Lower productivity is expected over the central and eastern portions of the basins.

Ekman Surface Divergence

It is generally accepted that the divergence of the Antarctic Surface Water and the associated deep water upwelling are mostly a result of the wind [*Sverdrup*, 1933]. Thermohaline effects no doubt play an important role, since the sinking shelf water must be replaced by deep water. However, in summer the wind is more important, since only minor sinking due to thermohaline alterations occurs. It is possible to calculate the wind-produced divergence by the following procedure.

In an unbounded homogeneous ocean, the stress of the wind on the sea surface would create a surface current system as derived by *Ekman* [1905]. The current vectors decrease in magnitude and rotate as the depth increases, forming the classical Ekman spiral. The surface current vector is directed at 45° to the left of the wind in the southern hemisphere. At a depth D, the current is directed in the opposite direction of the surface flow with a magnitude of $1/e$ of the surface value. The depth of D depends on the vertical eddy viscosity and the latitude. It usually occurs between 100 to 300 meters below the sea surface. The net total transport of the wind drift current is found to be directed at right angles to the wind, to the wind's left in the southern hemisphere, and given by the following expressions assuming that the wind is solely a zonal wind [*Sverdrup et al.*, 1942, 1946]:

$$T_\theta = +\tau_\lambda/\rho f \tag{3}$$

where

T_θ = total meridional transport of the wind drift current across one centimeter of a parallel;
τ_λ = zonal wind stress, dynes/cm^2;
ρ = sea water density;
f = Coriolis parameter = $2\,\omega \sin\theta$;
ω = earth's angular momentum;
θ = latitude.

The total transport across a parallel T_θ' is found by multiplying equation 3 by the number of centimeters of that parallel given by $L = 2\,\pi r \cos\theta$, where r is the earth's radius. Therefore

$$T_\theta' = LT_\theta = \pi r \cos\theta\ (\tau_\lambda/\rho f) \tag{4}$$

From the continuity equation, it is possible to calculate the amount of vertical motion between latitudes θ_1 and θ_2, $T_{vt(\theta 1\,-\,\theta 2)}$, at the base of the wind drift current, since

$$T_{vt(\theta 1\,-\,\theta 2)} = T'_{\theta 1} - T'_{\theta 2} \tag{5}$$

Therefore

$$T_{vt(\theta 1\,-\,\theta 2)} = (\pi r/\omega\rho) \cdot [\tau_\lambda\ (\theta_1)\ \cot\theta_1 - \tau_\lambda\ (\theta_2)\ \cot\theta_2] \tag{6}$$

The vertical transport can be calculated if the zonal stress, as a function of latitude, is known. To establish a realistic function, the following conditions must be met: (1) zero stress at the Antarctic Divergence at 65°S; (2) westward directed stress (negative) south of the divergence; and (3) the eastward stress, north of the divergence, must be at a maximum value in the vicinity of the central antarctic polar front zone, 53°S. For the purpose of calculation of the upward transport, a value of + 2 dynes/cm^2 is taken at 53°S, and a sinusoidal function is assumed. The expression for τ_λ (θ becomes

$$\tau_\lambda\ (\theta) = 2 \sin\ (-75\ (\theta) + 487.5) \tag{7}$$

Since the stress increases toward the north, the vertical transport will be upward, i.e., a surface divergence with upwelling.

Table 2 shows the wind stress at each degree of latitude from 53°S to 70°S and the upward transport at the base of the wind drift current in 1° bands. The total upwelling is 54 × 10^6 m^3/sec. However, 20.2 ×

10^6 m^3/sec or 37% of the total value is accomplished in the 4° band between 58° and 62°S, and the maximum upward transport is found between 60° and 61°S or 4° north of the divergence. This is due to the increase in area of latitudinal bands as the latitude decreases.

The choice of +2 dynes/cm^2 at 53°S is open to debate.

The relation of stress (τ) to wind velocity (W) for neutral temperature profile of the air (neutral stability) just above the sea surface [*Malkus*, 1962] is

$$\tau = c\rho W^2 \tag{8}$$

where $c = 2 \times 10^{-3}$ and ρ = air density = 1.27×10^{-3}. The zonal wind necessary to give 2 dynes/cm^2 stress is 20 knots.

The *Marine Climatic Atlas of the World* [*U.S. Navy*, 1965] charts on surface wind indicate that in the ocean region between 50°S and 60°S, the winds are most frequently from the northwest, west, or southwest in the range of Beaufort force 5 and 6 or 17 to 27 knots. Since the wind many times has a meridional component, a 20-knot value for the average zonal wind is reasonable, and the τ (θ) function used in the above calculation is satisfactory, at least to a first approximation.

The upwelling into the surface layer as found by the Ekman transport divergence agrees very well with that found by water budget considerations for the summer months. It is expected that owing to friction and other deviations from the original assumptions made in the calculations of the pure Ekman spiral, the wind-produced upwelling may deviate somewhat from that calculated above.

TABLE 2. Upwelling Transport in 1° Bands around Antarctica

Latitude θ	Dynes/cm^2	T_{vt} $(\theta_1 - \theta_2)$ m^3/sec
53	2.00	1.9×10^6
54	1.98	2.4
55	1.93	2.9
56	1.85	3.4
57	1.73	3.6
58	1.59	4.0
59	1.41	4.0
60	1.22	4.2
61	1.00	4.0
62	0.77	4.0
63	0.52	3.8
64	0.26	3.5
65	0	3.2
66	−0.26	2.9
67	−0.52	2.5
68	−0.77	2.0
69	−1.00	1.7
70	−1.22	
53–70		54.0

WINTER MONTHS

During the winter months, April to September, the continental runoff ceases and pack ice is forming. There is an additional loss of salt through the sinking of cold saline surface water, which eventually becomes part of the Antarctic Bottom Water. The water balance equation for the winter months is

$$M_1S_1 + M_2S_2 - M_{\text{ice}}S_{\text{ice}} - M_3S_3 + M_{ASW}S_{AS} = (M_1 + M_2 - M_{\text{ice}} - M_3 + M_{ASW})S_{AW} \tag{9}$$

where M_3, S_3 are the mass and salinity of the surface water which sinks to depths greater than 100 meters. Solving for M_3 with $S_1 = 0$ gives

$$M_3 = [M_1S_{AW} + M_2(S_{AW} - S_2) + M_{\text{ice}}(S_{\text{ice}} - S_{AW}) + M_{ASW}(S_{AW} - S_{AS})] / S_{AW} - S_3 \tag{10}$$

From the $(T_{\min} - T_{\max})/S$ diagrams (Figure 19), it is found that sinking of surface water occurs at salinities above 34.6‰. Using 34.6‰ for S_3, 33.9 and 34.0‰ for S_{AS}, S_{AW}, respectively, and other values as given in the previous section, then M_3 is calculated to be 35×10^6 m^3/sec, assuming M_2 as 60×10^6 m^3/sec. Therefore, during the winter months about one-half of the upwelling deep water contributes to water that sinks to depths greater than 100 meters. Only the densest of this water eventually becomes Antarctic Bottom Water. Part sinks to the transition zones between the surface and upper deep water and between the lower deep water and the bottom water. On sinking, this water entrains more deep water. This entrained deep water does not reach the surface and is not included in M_2. The amount of entrained water would be about the same magnitude as the sinking surface water, since bottom water temperatures and salinities are about midway between that of the surface water and the main mass of Circumpolar Deep Water. Thus, the total amount of southward transport of Circumpolar Deep Water during the winter months is about 10^8 m^3/sec, which is much less than the 8×10^8 m^3/sec estimated by *Kort* [1962, 1964]. Since the zonal transport of the circumpolar current is about 2×10^8 m^3/sec [*Gordon*, 1967*b*], during the winter months an approximate 2:1 ratio exists between zonal and meridional transport in antarctic waters. The Ekman Divergence of about 50×10^6 m^3/sec suggests that approximately one-half of the meridional transport is wind induced. The other half is initiated by the thermohaline effect of bottom water production.

During the summer, the meridional transport is 60 ×

10^6 m^3/sec, all of which is M_2 and contributes to the Antarctic Surface Water. In winter, only half of this amount remains in the surface water. This assessment rests on the assumption made earlier that M_2 is constant throughout the year. A seasonal variation of divergence of the surface wind [*Tolstikov,* 1966, pl. 108], indicates that M_2 would vary somewhat. Variations in the value of M_2 and the southward migration of the deep water would be reflected in variations of the zonal transport. This is expected because of the relationship between zonal and meridional flow based on conservation of vorticity [*Kaplan,* 1967]: a southward transport initiates an easterly current. The Antarctic Circumpolar Current is largest during the months of March and April [*Yeskin,* 1962]. Therefore, it is probable that during these months M_2 is at a maximum. A large seasonal effect in the production of bottom water and renewal of surface water must occur.

World Ocean Meridional Transport

Owing to the strong stratification within the upper water column in the tropical and subtropical regions of the oceans, the deep ocean has only limited interaction with the surface layers, mainly through eddy diffusion. At high latitudes, the cold climate and lack of strong stratification permit deep vertical convection. The water masses below the main thermocline of the lower latitudes are derived from these polar and subpolar regions. Figure 22 depicts, in schematic form, the basic meridional ocean circulation. Sinking occurs at high northern latitudes, producing deep water; additional deep water is derived from the warm and salty subtropical marginal seas of the northern hemisphere. The loss of surface water is compensated by a flow from the warm to cold sphere by a northward current in the northeastern North Atlantic Ocean. The warm water sphere is, in turn, replenished by a transfer of water across the antarctic polar front zone and upwelling through the main thermocline. The upwelling is also necessary to compensate partially for the downward flux of heat by eddy diffusion [*Robinson and Stommel,* 1959]. The deep water flows south with increasing volume transport, owing to the addition of Antarctic Intermediate Water (with some Arctic or Subarctic Intermediate Water) and Antarctic Bottom Water. South of the polar front zone, the bottom water transport is increased by entrainment of deep water, but to the north the transfer is from bottom to deep water.

The water balance equation may be written as follows:

$$N_3 = N_1 + N_2 + A_3 + (A_2 - U) \tag{11}$$

$$W = N_1 + N_2 + N_4 = A_1 + U + N_4 \tag{12}$$

therefore

$$N_3 = A_1 + A_2 + A_3 \tag{13}$$

where

N_3 = upwelling Circumpolar Deep Water;
N_2 = subtropical marginal sea produced deep water;
N_3 = upwelling Circumpolar Deep Water;
N_4 = intermediate water formed in the northern hemisphere (that part of northern hemisphere intermediate water which becomes incorporated into the deep water is included directly in the N_1 term);
A_1 = Antarctic Surface Water mixing across the polar front zone into the warm water sphere;
A_2 = intermediate water formation, southern hemisphere;
A_3 = bottom water;
U = water upwelling through main thermocline; all of this water is derived from the intermediate water, A_2 and N_4;
W = transfer from warm to cold water sphere.

The source region for the N_1 water is the North Atlantic. Some of the intermediate water of the North Pacific may become incorporated into the deep water and thus be included in N_1, but it is probably only a small amount. The N_2 water is derived mainly from the Mediterranean and Red seas and Persian Gulf. Intermediate water is formed in both hemispheres, but all of N_4 re-enters the warm water sphere before reaching the southern hemisphere (by definition of the N_4 term). The upwelling U water occurs throughout the tropical and subtropical regions of the world ocean with some variation with longitude [*Wyrtki,* 1961]. Since a portion of the W water is from the southern hemisphere, there must be a net northward transport across the equator. This transport would best develop in the Atlantic Ocean, where most of the deep water is produced. The northward transport prohibits a well-defined tropical or equatorial water mass from forming in the Atlantic unlike the other oceans [*Sverdrup et al.,* 1942].

It is possible to place values on some of the terms of equation 11. These values are considered to be only approximations. Of more importance is the relative magnitude of the terms.

The N_3 term from the salt budget considerations above is 100×10^6 m^3/sec. Unknown is the amount

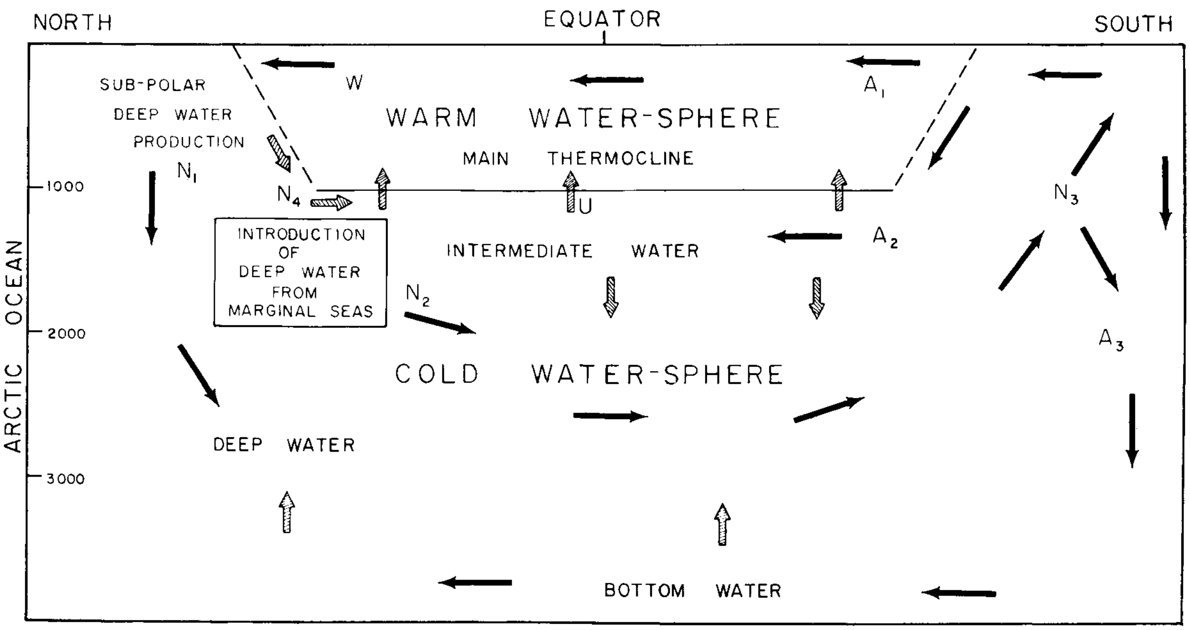

Fig. 22. Schematic of meridional transport of world ocean. See page 196 for key to abbreviations.

of water which is recirculated between A_2, A_3, and N_3, that is, the value of $A_2 - U$ and A_3. These may be calculated if, in addition to N_3, the terms U, N_1, N_2, and A_1 are known. The first three of these terms can be approximated. From heat budget consideration in the case of U, an average upward velocity of 1 to 5 $\times$ 10^{-5} cm/sec is expected in the main thermocline [*Robinson and Stommel*, 1959; *Wyrtki*, 1961]. The N_1 and N_2 terms can be approximated from recent measurements and calculations. The Atlantic-produced portion of N_1 from the Norwegian Sea overflow is most likely between 6 to 12 $\times$ 10^6 m^3/sec [*Worthington and Volkmann*, 1965], since 5 to 6 $\times$ 10^6 m^3/sec overflows to either side of Iceland on the ridge system from Greenland to the Faroes. However, it is not known how much of the flow to the west is entrained Iceland-Faroes overflow. The amount of deep water produced along the southern coast of Greenland [*Wüst*, 1935) is unknown, although it is expected to be less than the Norwegian Sea overflow. *Arons and Stommel* [1967] calculate a value of 6 $\times 10^6 \times$ m^3/sec for the Atlantic part of N_1, based on radiocarbon data in the North Atlantic. The values of N_2 for the Atlantic (outflow from Mediterranean) have been approximated as 1.5 $\times$ 10^6 m^3/sec by *Boyum* [1967]. *Ovchinnikov* [1966] states that the Atlantic inflow into the Mediterranean is between 0.84 to 1.05 $\times$ 10^6 m^3/sec. Since the outflow must equal this difference of river runoff and precipitation with the water lost by evaporation within the Mediterranean, N_2 is most likely less than 1 $\times$ 10^6 m^3/sec. The sum of N_1 and N_2 for the Atlantic Ocean, which must make up most of these components, is probably between 10 to 15 $\times$ 10^6 m^3/sec. Since N_3 is of order 10^8 m^3/sec, the values of A_3 and $A_2 - U$ are quite large, about 90 $\times$ 10^6 m^3/sec together. From the salt budget considerations, the ratio of these terms is most likely 1:1, with the possibility of A_3 being slightly larger. The recirculation of antarctic water masses into the deep water occurs all around Antarctica and is not restricted to the Atlantic Ocean; however, some longitudinal variation is natural. The deep water of Atlantic origin does not fit directly on the T/S curve of the Circumpolar Deep Water (see Figure 4). Therefore, the Circumpolar Deep Water is made of modified North Atlantic Deep Water and is to a large extent a closed system involving only southern hemisphere circulation. Perhaps only 10–20% of the transport of the Antarctic Circumpolar Current is due to additions from the northern hemisphere.

If the value of A_1 is known, it would be possible to calculate the above ratio. The A_1 term can be found using equation 12; however, the error in estimating N_1 and N_2 and especially the uncertainty of the value of U makes such a calculation impractical. It is best to arrive at A_1 directly from ocean observations. For

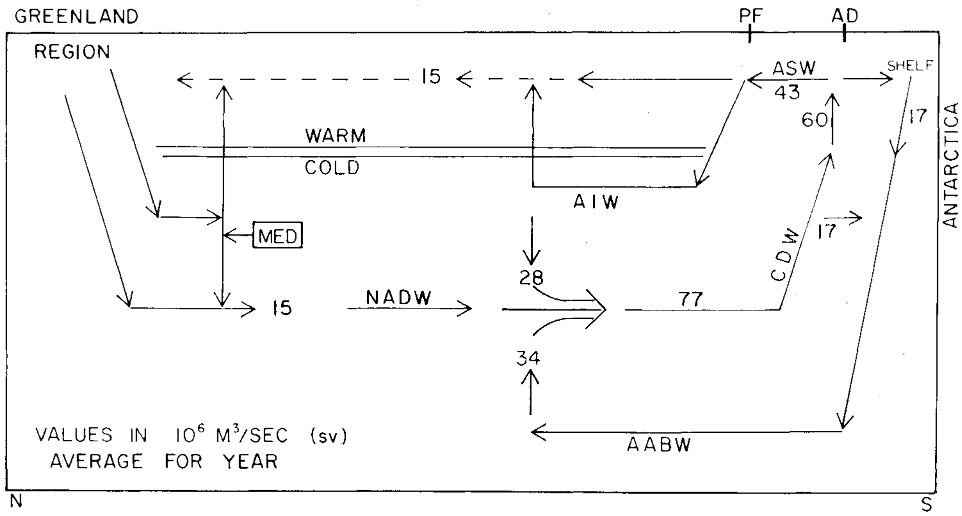

Fig. 23. Mean annual values of the meridional transport of the world ocean.

this purpose, radioisotope data are useful, as is, possibly, the ratio of antarctic to subantarctic fauna and flora in the surface water.

Figure 23 shows the magnitude of the net meridional transport at various levels for the world ocean. The values represent yearly averages from the salt budget considerations and estimates of the transport of the North Atlantic Deep Water. The relative values are of more significance than the absolute values, owing to the large uncertainties involved with such calculations. The annual average ratio of zonal to meridional transport in antarctic waters is between 3:1 to 2:1.

CONCLUDING REMARKS

The oceanographic processes that occur in antarctic waters are essential to: (1) maintenance of the intermediate, deep, and bottom water of the world ocean as an aerobic environment; (2) the removal of heat and addition of fresh water necessary for a steady-state character of the deep water; (3) renewal of the warm water sphere, which is depleted in the production of North Atlantic Deep Water; (4) the equalization of water characteristics of the three major oceans. The results of these processes lead to high biological productivity in antarctic waters and a net heat flux from the ocean to the atmosphere. The driving mechanism in antarctic circulation is the wind and the thermohaline effect along the coast of Antarctica. Both cause upwelling of deep water, which can only be replaced by a southward migration of the Circumpolar Deep Water.

The deep water that reaches the sea surface in the vicinity of the divergence is transformed into Antarctic Surface Water by sea-air interaction and vertical diffusion. Most of this water flows northward; part of it becomes incorporated into the Subantarctic Surface Water and the rest eventually sinks at the Antarctic Convergence, contributing to the formation of Antarctic or Subantarctic Intermediate Water. Some of the surface water flows southward to the shores of Antarctica, where intense climatic conditions produce a very cold, salty shelf water. This water sinks, mixing with equal proportions of deep water to produce Antarctic Bottom Water.

Within the Antarctic Circumpolar Current, an axis of flow can be found at the depth of the salinity maximum layer [*Goldberg*, 1967]. The axis occurs slightly south of the polar front zone. Such a situation is very similar to the atmospheric conditions of the midlatitudes, where a polar front separates the cold polar air masses from the tropical or subtropical air masses. Directly above or slightly poleward of the atmospheric polar front is the axis of flow of the westerlies. This axis is called the Jet Stream [*Hare*, 1963].

Perhaps the analogy can be drawn further: The mean flow within the axis of the Antarctic Circumpolar Current is maintained by a transfer of energy from large eddies in a fashion similar to the energy transfer by the high and low pressure systems of the midlatitudes into the Jet Stream [*Lorentz*, 1966]. Such speculation may be evaluated by the study of time variations within antarctic waters.

The above model depicts only the average conditions; there are strong variations with longitude, mainly due to the asymmetry of Antarctica and the submarine topography. The Antarctic Circumpolar Current is diverted northward and intensifies upon approaching a north-south oriented ridge. After passing the ridge, it turns southward and becomes more diffuse. On passing over the fracture zones through the ridges, the flow is constricted at all depths.

From the group T/S diagrams of the southwest Atlantic and Pacific sectors of antarctic waters, it is found that two cyclonic gyres exist in which water found at levels below 2000 meters to the northwest upwell to approximately 400 meters in the central parts of the gyre. These gyres bring higher salinities to the surface waters, as shown in Figure 22 and in Figure 2 (note the waters above 34.6‰ in the Weddell Sea and northeast of the Ross Sea).

The meridional transport is of the order of 10^8 m^3/sec or less. It is probable that the transport is not equally distributed about Antarctica, but that it is confined to the regions of the cyclonic gyres. It is expected that this transport has a seasonal variation because of such variations in the wind field and thermohaline alterations along the coast of Antarctica. That the zonal transport is greatest during the months of March and April indicates that meridional transport is greatest during these months. April begins the period of rapid ice production. Perhaps the maximum in meridional transport during April is due to a maximum in bottom water production.

Since the introduction of deep water from the north is most likely between 10 and 15 $\times$ 10^{+6} m^3/sec, most of the Antarctic Circumpolar Current is water which is recirculated between the antarctic water masses and the deep water. Such recirculation can occur in all three oceans. However, the introduction of deep water from more northern sources is confined mostly to the Atlantic Ocean.

The Antarctic Surface Water and the Antarctic Bottom Water both occur in the frictional boundaries of the ocean. The surface water receives vorticity from the wind. The Antarctic Bottom Water on flowing northward and conserving its vorticity tends to flow to the west along the continental margins of Antarctica. However, a circumpolar westward flow of bottom water is prevented by the presence of submarine ridges. The bottom water on reaching these ridges turns to the north.

SPECULATIONS OF PALEOCEANOGRAPHY

In the following speculations the ice cover of Antarctica is assumed to vary. Also, it is assumed that the continent has not moved relative to the South Pole. These assumptions may conflict with each other and should be noted when considering the following remarks.

If Antarctica were ice free, the air pressure over the continent would be expected to vary greatly with season, being low during the summer and high in winter, similar to seasonal variations over Siberia. In this case, the wind produced upwelling of deep water would be much reduced owing to the reduction in the offshore pressure gradient and associated winds. During the summer the monsoonlike antarctic winds would be westerly over the water and land areas with a slight onshore component. In winter, the atmospheric high over the land area would cause a zone of polar easterlies, but this zone would not extend seaward. The westerlies over the water areas would be more intense than in the summer but substantially less than at present. Therefore, the nonglaciated condition would initiate a strongly seasonal or monsoonal character in the wind and associated fluctuation in the divergence of surface water.

An additional effect would be the absence of the very cold dry katabatic winds that characterize ice-covered regions. These winds in the present condition of Antarctica can extend to the water, where they cause intense thermohaline alterations. In nonglaciated or partially glaciated conditions, offshore winds will occur during the winter; however, they will not be as cold as the present winds, since, in the absence of the very cold ice surface, some warming by the long wave ground radiation would occur. Therefore, thermohaline alterations of the shelf waters would be absent or minimal during nonglaciated or partly glaciated periods.

During periods of floating ice shelves, representing extensive glaciation, freezing and/or melting to the underside may produce additional thermohaline effects.

Both of these changes (decreased wind divergence and thermohaline alterations) would result in a marked decrease in the meridional transport of the deep water during summer and a lesser decrease during the winter months. Such a decrease would be reflected in a decreased Antarctic Circumpolar Current and a less vigorous vertical circulation. If deep water is still produced in the North Atlantic at times of an ice-free Antarctica, it is possible that the entire zonal transport of the Antarctic Circumpolar Current would not be greater than the transport of the North Atlantic Deep Water.

If Antarctica is only glaciated in the interior, the

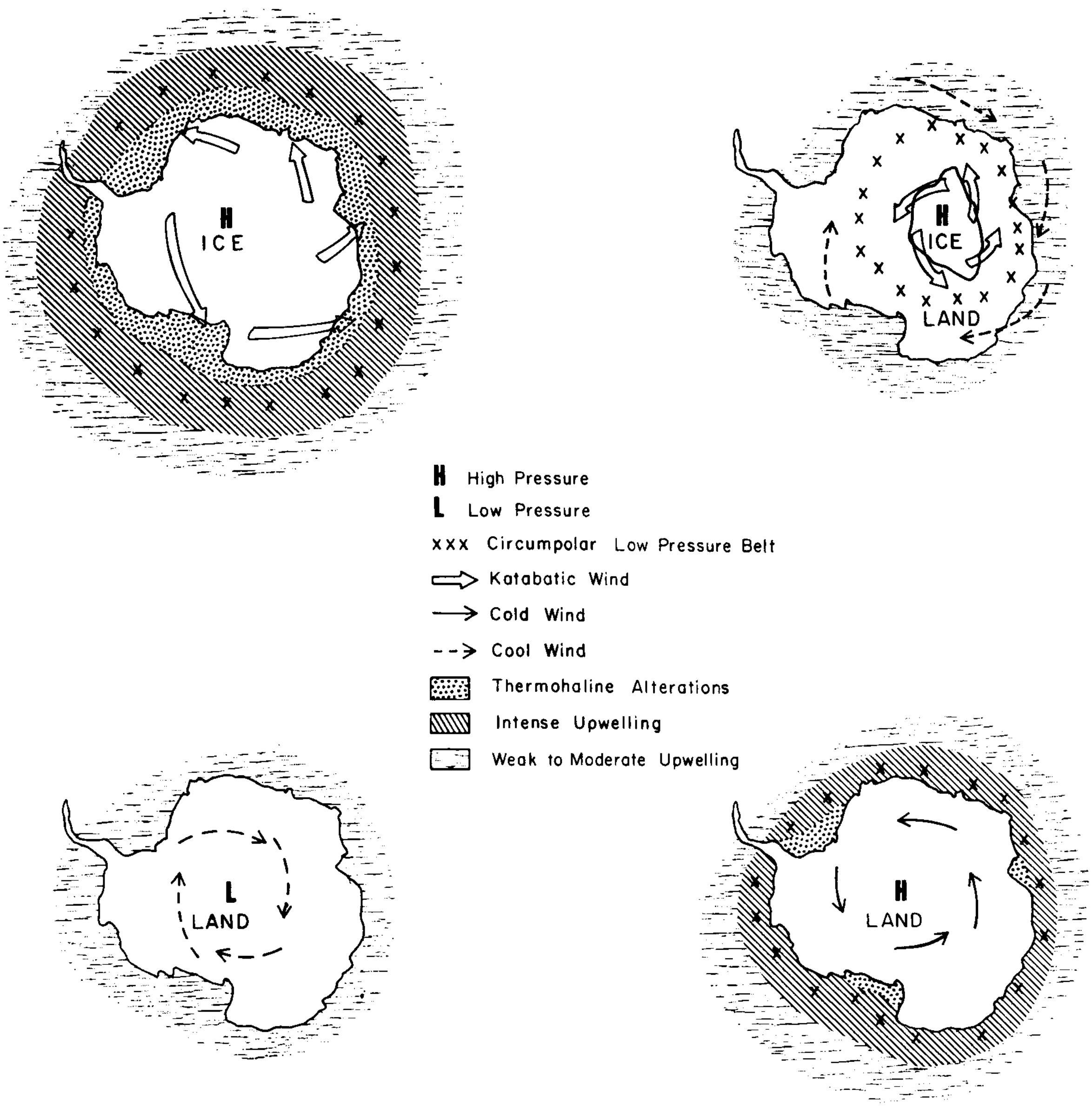

Fig. 24. Schematic representation of atmosphere sea level pressure and oceanographic conditions during times when Antarctica is (*a*) fully glaciated; (*b*) partially glaciated (summer); (*c*) ice free (summer); and (*d*) ice free or partially glaciated (winter).

pressure over the ice field would be high and the belt of low pressure surrounding the ice may not occur over the water area, but be entirely situated over land. In this case, the wind divergence over the sea would be small. In addition, the thermohaline effects at sea would not be as large owing to the absence of the katabatic winds and floating ice shelves. Therefore, the present antarctic circulation is not only dependent on the presence of a glaciated continent, but this glaciation must extend to the shores of the continent; that is, Antarctica must be fully glaciated. This condition has been mentioned by *Hays* [1967], based on a study of the sediment cores around Antarctica. Figure 24 shows in schematic form the three conditions: fully glaciated, partly glaciated, and ice free.

The results of the decreased circulation in Antarctica during periods of nonglaciation or partial glaciation would have certain major effects on the deep and bottom water of the world ocean and on the processes within antarctic waters: (1) reduction of the biological productivity of antarctic waters; (2) reduction in the renewal of oxygen of the cold water sphere; (3) decrease in the magnitude of bottom circulation, at least in the southern hemisphere; and (4) most likely an increase in the temperatures of the deep water.

In addition, changes would occur in the warm water sphere. The decrease of the volume transport of the Antarctic Circumpolar Current may cause surface conditions of the three oceans to differ more than they do at present. Now the oceans are equalized to a great extent by the large transport about Antarctica. On decreasing this transport, the larger rates of evaporation over the Atlantic may cause the Atlantic to be saltier, i.e., they may cause a growth in the size of the high saline 'Sargasso' gyres in the North and South Atlantic. An additional effect would be an increase in the depth of the main thermocline, due to an increase in the temperature of the deep water. This would deepen the oxygen minimum and nutrient rich layers that occur just below the thermocline. It is possible that such a deepening would decrease the amount of nutrients and associated high productivity in the surface waters of regions of subtropical upwelling, i.e., off the northwestern and southwestern coasts of Africa and off Peru and California.

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REFERENCES

Arons, A. B., and H. Stommel, On the abyssal circulation of the world ocean, 3, *Deep-Sea Res., 14,* 441–457, 1967.

Baronov, G. I., and V. N. Botnikov, Surface of no motion and water masses in the Weddell Sea, *Soviet Antarctic Exped. Inform. Bull.,* English Transl., *5*(6), 385–388, 1964.

Botnikov, V. N., Geographic position of the Antarctic Convergence Zone in the Pacific Ocean, *Soviet Antarctic Exped. Inform. Bull.,* English Transl., *4*(6), 324–327, 1964.

Boyum, G., Hydrology and currents in the area west of Gibraltar, 19 pp., 25 figs., *Tech. Rept. 36,* Mediterranean Outflow Project, NATO Subcommittee on Oceanographic Research, Bergen, Norway, 1967.

Brennecke, W., Die ozeanographischen Arbeiten der Deutschen Antarktischen Expedition, 1911–1912, *Arch. Deutsche Siewarte, 39*(1), 1–216, 1921.

Brodie, J. W., Oceanography, Antarctica, in *New Zealand Antarctic Society Survey,* edited by T. Hatherton, pp. 101–127, Reed, Wellington, New Zealand, 1965.

Burling, R. W., Hydrology of circumpolar waters south of New Zealand, *New Zealand Dept. Sci. Ind. Res. Bull., 143, Mem. 10,* 9–61, 1961.

Countryman, K. A., and W. L. Gsell, Operations Deep Freeze '63 and '64 summer oceanographic features of the Ross Sea, *Tech. Rept. TR-190,* U. S. Navy Oceanographic Office, 1966.

Deacon, G. E. R., The hydrology of the Southern Ocean, *Discovery Rept., 15,* 1–24, 1937.

Deacon, G. E. R., A general account of the hydrology of the South Atlantic Ocean, *Discovery Rept., 7,* 171–238, 1933.

Deacon, G. E. R., The Southern Ocean, Ideas and observations on progress in the study of the seas, in *The Sea,* vol. 2, edited by M. N. Hill, pp. 281–296, Interscience, New York, 1963.

Defant, A., *Physical Oceanography,* vol. 1, 729 pp., Pergamon, New York, 1961.

Drygalski, E. V., Ozean und Antarktis, *Deutsche Sudpolar Exped., 1901–1903, 7,* 1926.

Ekman, V. W., On the influence of the earth's rotation on ocean currents, *Roy. Swed. Acad. Sci., 2*(11), 1905.

Fleurieu, C. P. Claret de, Observations sur la division hydrographique du globe, et changements proposes dans la nomenclature generale et particuliere de l'hydrographie, in Marchand, E., *Voyage Autour du Monde, 4,* 1–74, Paris, Imprimmerie de la Republique, VIII, 1798–1800.

Fofonoff, N. P., Some properties of sea water influencing the formation of Antarctic bottom water, *Deep-Sea Res., 4,* 32–35, 1956.

Giovinetto, M. B., Estimate of mass flux in Antarctica, *Geol. Soc. Am. Spec. Paper 76,* 309, 1964.

Goldberg, R., Mean oceanographic conditions within the Southeast Pacific Basin, 100 pp., Masters thesis, Columbia University, New York, 1967.

Gordon, A. L., Potential temperature, oxygen and circulation of bottom water in the Southern Ocean, *Deep-Sea Res., 13,* 1125–1138, 1966.

Gordon, A. L., Structure of Antarctic Waters between 20°W and 170°W, *Antarctic Map Folio Series, Folio 6,* edited by V. Bushnell, American Geographical Society, New York, 1967*a*.

Gordon, A. L., Geostrophic transport through the Drake Passage, *Science, 156*(3783), 1732–1734, 1967*b*.

Gordon, A. L., Physical oceanography aboard the *Eltanin, Antarctic J., 3*(5), 1967*c*.

Gow, A. J., The ice sheet, in *Antarctica,* edited by T. Hatherton, ch. 9, pp. 221–258, Praeger, New York, 1965.

Green, E. J., and D. E. Carritt, New tables for oxygen saturation of seawater, *J. Marine Res., 25*(2), 1967.

Hare, F. K., *The Restless Atmosphere, An Introduction to Chemistry,* p. 192, Harper Torchbooks, Harper and Row, New York, 1963.

Hays, J. D., Quaternary sediments of the Antarctic Ocean, *Progress in Oceanography,* vol. 4, pp. 117–131, Pergamon, New York, 1967.

Heap, J. A., Pack ice, in *Antarctic Research,* edited by R. Priestley, pp. 308–317, Butterworth, London, 1964.

Helland-Hansen, Bj., Nogen Hydrografisk Metoder., *Förh. Ved. Skand. Naturf., 16,* 357–359, Oslo, Juli 1916.

Houtman, T. J., Surface temperature gradients at the Antarctic Convergence, *New Zealand J. Geol. Geophys.*, *7*, 245–270, 1964.

Houtman, T. J., Water masses and fronts in the Southern Ocean south of New Zealand, *New Zealand Dept. Sci. Ind. Res. Bull.*, *174*, *Mem. 36*, 1967.

Ishino, M., Studies on the oceanography of the Antarctic Circumpolar Waters, *J. Tokyo Univ. Fisheries*, *49*(2), 73–181, 1963.

Jacobs, S. S., Physical and chemical oceanographic observations in the Southern Ocean, USNS *Eltanin* Cruises 16–21, 128 pp., *Tech. Rept. 1-CU-1-66*, Lamont-Doherty Geological Observatory of Columbia University, Palisades, New York, 1966.

Jacobs, S. S., and A. Amos, Physical and chemical oceanographic observations in the Southern Oceans, USNS *Eltanin* Cruises 22–27, 287 pp., *Tech. Rept. 1-CU-1-67*, Lamont-Doherty Geological Observatory of Columbia University, Palisades, New York, 1967.

Kaplan, S., A prolegomena to the dynamics of the Antarctic Circumpolar Current, 86 pp., Master's thesis, Columbia University, New York, 1967.

Knudsen, M., Gefrierpunkttabellen für Meerwasser, *Publ. Circ.* *5*, Copenhagen, 1903.

Kort, V. G., The Antarctic Ocean, *Sci. Am.*, *207*(3), 113–128, 1962.

Kort, V. G., Antarctic Oceanography, in *Research in Geophysics*, edited by H. Odishaw, vol. 2, chap. 13, pp. 309–333, J. Engelhorn, Massachusetts Institute of Technology Press, Cambridge, 1964.

Krümmel, O., *Handbuch der Ozeanographie*, 1st ed., 2 vols., Stuttgart, 1897.

Ledenev, V. G., Contribution to the study of surface currents in the seas of the Pacific sector of the Antarctic, *Soviet Antarctic Exped. Inform. Bull.*, English Transl., *3*, 257–262, 1961.

Littlepage, J. L., Oceanographic investigations in McMurdo Sound, Antarctica, *Biology of the Antarctic Seas, II*, *Antarctic Res. Ser.*, *5*, edited by George A. Llano, pp. 1–37, American Geophysical Union, Washington, D.C., 1965.

Lorentz, E. N., The circulation of the atmosphere, *Am. Sci.*, *54*(4), 402–420, 1966.

Lusquinos, J. L., Extreme temperatures in the Weddell Sea, *Acta Univ. Bergen. Ser. Math. Rerumque Nat.*, *23*, 1963.

Mackintosh, N. A., The Antarctic Convergence and the distribution of surface temperatures in Antarctic Waters, *Discovery Rept.*, *23*, 177–180, 1946.

Malkus, J. S., Large-scale interaction, in *The Sea*, vol. 1, edited by M. N. Hill, pp. 88–294, Interscience, New York, 1962.

Mellor, M., Remarks concerning the Antarctic mass balance, *Polarforschung*, *Ser. 5*, *33*(1/2), 179–180, 1963.

Midttun, L., and J. Natvig, Pacific Antarctic waters, 130 pp., *Sci. Res. Brategg Exped. 1947–1948*, *3*, 1957.

Miyake, Y., Chemical studies of the western Pacific Ocean, 3, *J. Chem. Soc. Japan*, *14*(3), 239–242, 1939.

Model, F., Ein Beitrag zur regionalen ozeanographic der Weddellsee, *Deutsche Antarctic Exped. 1938–1939 Wiss. Ergeb.*, *2*(2), 63–96, Georg Kartographers Anstalt 'Mundus' Helmut Striedieck, Hamburg, 1957–1958.

Mosby, H., The waters of the Atlantic Antarctic Ocean, *Sci. Res. Norwegian Antarctic Exped. 1927–1928*, *11*, 1–131, 1934.

Mosby, H., Bottom water formation, in *Symposium on Antarctic Oceanography, September 13–16, 1966, Santiago, Chile*, Scott Polar Research Institute, Cambridge, England, 1968.

Munk, W. H., Abyssal recipes, *Deep-Sea Res.*, *13*, 707–730, 1966.

Neumann, G., and W. J. Pierson, Jr., *Principles of Physical Oceanography*, 545 pp., Prentice-Hall, Englewood Cliffs, New Jersey, 1966.

Ostapoff, F., The salinity distribution at 200 meters and the antarctic frontal zones, *Deut. Hydrograph. Z.*, *15*, 133–142, 1962.

Ovchinnikov, I. M., Circulation in the surface and intermediate layers of the Mediterranean, *Oceanology*, English Transl., *6*(1), 48–59, November 1966.

Robin, G. de Q., and R. J. Adie, The ice cover, *Antarctic Research*, edited by R. Priestley, pp. 100–117, Butterworth, London, 1964.

Robinson, A. R., and H. Stommel, The oceanic thermocline and associated thermohaline circulation, *Tellus*, *11*(3), 295–308, 1959.

Serikov, M. I., Density and salinity of Antarctic sea ice, *Soviet Antarctic Exped. Inform. Bull.*, English Transl., *3*, 262–264, 1965.

Shumsky, P. and I. A. Zotikov, On bottom melting of the antarctic ice shelves, in *Snow and Ice Publication*, pp. 225–237, UGGI Association of Hydrologic Scientists, Berkeley, California, 1963.

Shumsky, P., A. N. Krenke, and I. A. Zotikov, Ice and its changes, in *Research in Geophysics*, vol. 2, *Solid Earth and Interface Phenomena*, edited by H. Odishaw, pp. 425–460, Massachusetts Institute of Technology Press, Cambridge, 1964.

Steele, J. H., L. V. Worthington, and J. R. Barrett, Deep currents south of Iceland, *Deep-Sea Res.*, *9*, 465–474, 1962.

Sverdrup, H. U., On vertical circulation in the ocean due to the action of wind with application to conditions within an Antarctic Circumpolar Current, *Discovery Rept.*, *7*, 139–170, 1933.

Sverdrup, H. U., M. W. Johnson, and R. H. Fleming, *The Oceans*, 1st ed., Prentice-Hall, Englewood Cliffs, New Jersey, 1942.

Thomas, C. W., Vertical circulation off the Ross Ice Shelf, *Pacific Sci.*, *20*(2), 239–245, 1966.

Tolstikov, E. I., Ed., *Atlas Antarktiki*, vol. 1, G.U.C.K., Moscow, 1966. (English Transl. in *Soviet Geography: Rev. Transl.*, *8*(5-6), American Geographical Society, New York, 1967.)

U. S. Navy, Oceanographic atlas of the Polar Seas, part 1, Antarctic, *Hydrograph. Off. Publ. 705*, 1957.

U. S. Navy, *Marine Climatic Atlas of the World*, vol. 7, NAVWEPS 50-1C-50, 1965.

Wexler, H., Ice budgets for Antarctica and changes in sea level, *J. Glaciol.*, *3*(29), 867–872, 1961.

Worthington, L. V., and G. H. Volkmann, The volume transport of the Norwegian Sea overflow water in the North Atlantic, *Deep-Sea Res.*, *12*(5), 667–676, 1965.

Wüst, G., Meteor Reports, *Wiss, Ergeb. Deut. Atlant. Exp. Meteor 1925–27*, Band *6*, Teil 1: Schichtung and Zirkulation des Atlantischen Ozeans. Lieferung: Das Bodenwasser und die Gliederung der Atlantischen Tiefsee, 1933. Lieferung: Die Stratosphäre, 1935. Band *6*, Atlas Schichtung und Zirkulation des Atlantischen Ozeans, Teil A and B: Stratosphäre, 1936.

Wüst, G., W. Brogmus, and E. Noodt, Die Zonale Verteilung von Salzgehalt, Niederschlag, Verdunstrung, Temperatur, und Dichte an der Oberfläche der Ozeane, *Kiel. Meersforsch., 10,* 137–161, 1954.

Wyrtki, K., The thermohaline circulation in relation to the general circulation in the oceans, *Deep-Sea Res., 8,* 39–64, 1961.

Yeskin, L. I., Water balance of the Drake Passage, *Tr. Sov. Antark. Exsp., 20,* 1962.

Zumberge, J. H., and C. Swithinbank, The ice shelves, in *Antarctica,* edited by T. Hatherton, chap. 8, pp. 199–220, Praeger, New York, 1965.