Deep Sea Research, 1971, Vol. 18, pp. 1063 to 1088. Pergamon Press. Printed in Great Britain.

On the influence of the Norwegian–Greenland and Weddell seas upon the bottom waters of the Indian and Pacific oceans*

JOSEPH L. REID† and RONALD J. LYNN‡

(*Received* 23 *November* 1970; *in revised form* 5 *March* 1971; *accepted* 10 *March* 1971)

Abstract—The bottom waters of the North Pacific and North Indian oceans have temperature and salinity distributions that suggest origins from the extreme waters of the Norwegian–Greenland and Weddell seas. We attempt to trace these waters from their sources to the abyssal Pacific and Indian oceans by examining distributions of temperature and salinity along a stratum defined by density parameters. We assume that the major flow and mixing will take place along such surfaces, though the results make plain that vertical mixing is also important. The density stratum we have chosen to examine extends from the sea surface in the Norwegian–Greenland Sea and from near the surface in the Weddell Sea to depths of about 3500 m in the central oceans and below 4000 m in the North Pacific.

The cold and saline water of the Norwegian–Greenland Sea is traced along the density stratum through the Denmark Strait, where vertical mixing raises both temperature and salinity to their maximum values in the central North Atlantic. From there the temperature and salinity decrease monotonically southward toward the Weddell Sea, partly by lateral mixing with the cold, low-salinity waters on this stratum where it lies near the sea surface in the Weddell Sea, and partly by vertical mixing with the underlying Antarctic Bottom Water. From the southern South Atlantic the high values of temperature and salinity (the stratum now lies close to the vertical maximum in salinity) extend eastward with the Antarctic Circumpolar Current into the Indian and Pacific oceans, with monotonically decreasing temperature and salinity as further vertical mixing erodes the maximum in salinity, until the salinity maximum is found at the bottom in the North Pacific Ocean.

The stratum we have defined terminates at abyssal depths in the northern Indian and Pacific oceans; since water must rise somewhere to balance the sinking in regions of bottom-water formation, there must be upward flow across the stratum elsewhere. The tremendous areal extent of the salinity maximum, however, suggests that the upward flow through the stratum must be minimal except in the North Indian and North Pacific oceans, where stability is shown to be very low at the depth of the stratum.

PURPOSE

THE PURPOSE of this study is to examine the relation of the bottom waters of the North Pacific and North Indian oceans to the various supposed sources of deep water in the World Ocean. Earlier studies (WÜST, 1929, 1938, 1951; SVERDRUP, 1931; COCHRANE, 1958; MONTGOMERY, 1958; POLLAK, 1958; STOMMEL and ARONS, 1960; BOLIN and STOMMEL, 1961; MUNK, 1966) have examined the problem through simple distributions of temperature, salinity, and oxygen, through budgets of such characteristics and through mathematical models based more or less on the distribution of such characteristics.

One characteristic that has not been used explicitly in such studies is density. There are obvious difficulties both in defining such a parameter and in making valid interpretations. Although density parameters have been used to illustrate oceanic features

*Contribution from Scripps Institution of Oceanography.
†Scripps Institution of Oceanography, La Jolla, Calif.
‡National Marine Fisheries Service, La Jolla, Calif.

and to obtain interesting and useful results (MONTGOMERY, 1938; TAFT, 1963; KUKSA, 1963; REID, 1965; CANNON, 1966; BARKLEY, 1968; TSUCHIYA, 1968; RUPP, 1969), they have mostly been limited to the use of potential density in the upper layers (0–1500 m). At greater depths potential density may have inversions (WÜST, 1933; KAWAI, 1966; LYNN and REID, 1968) and therefore cannot be taken to represent surfaces of flow or mixing. We have attempted to use a method that may allow us to use density parameters, in at least an approximate fashion, over a wider range of depth (pressure), and to investigate the deep and abyssal waters on the usual assumptions of isopycnal flow and mixing.

METHOD

LYNN and REID (1968) have suggested that the usefulness of isopycnal analysis may be extended to deep waters by calculating density referred to an appropriate pressure; and that, where water masses and density surfaces extend through wide depth ranges, mixing paths may be constructed of short segments of such density surfaces, each referred to appropriate pressures. It is not immediately clear that this method can yield useful results; therefore, some of the concepts are reviewed below.

Isopycnal analysis

Isopycnal surfaces (referred to a sea pressure of zero decibars) have frequently been used to investigate lateral mixing and circulation in the ocean. The assumption is made that exchange can take place more readily along isopycnal surfaces than across them, since such exchange does not significantly alter the field of density, and little work need be done to effect the exchange. Vertical mixing must be assumed either to be negligible or to take place in such a fashion that a parcel must mix with both deeper and shallower layers so that its own density is not changed by the mixing, though its other characteristics may be. The latter assumption is particularly awkward if one is dealing with water near the bottom.

The effect of pressure on mixing

Pressure has been assumed constant in such studies. Although the various assumptions are not entirely fulfilled, the method has been useful in studies of mixing and circulation in the upper few hundred meters. Even at shallow depths the pressure variation can be troublesome, but, when mixing and flow through the entire depth range of the ocean are considered, the pressure effects are so severe that they cannot be ignored.

An ordinary θ–S diagram (potential temperature versus salinity) is shown in Fig. 1a with isopycnals drawn corresponding to zero decibars of sea pressure. Figs. 1b and 1c are similar θ–S diagrams, but the isopycnals are drawn for pressures of 2000 and 4000 decibars. They allow one to find the *in situ* density a water parcel would attain if moved adiabatically to depths where the pressure is 2000 or 4000 decibars. The density increases substantially, of course, but the critical difference is that it does not increase uniformly: the slopes of the isopycnals decrease with increasing pressure. This is mostly the effect of the variation of compressibility with temperature: the colder water is more compressible.

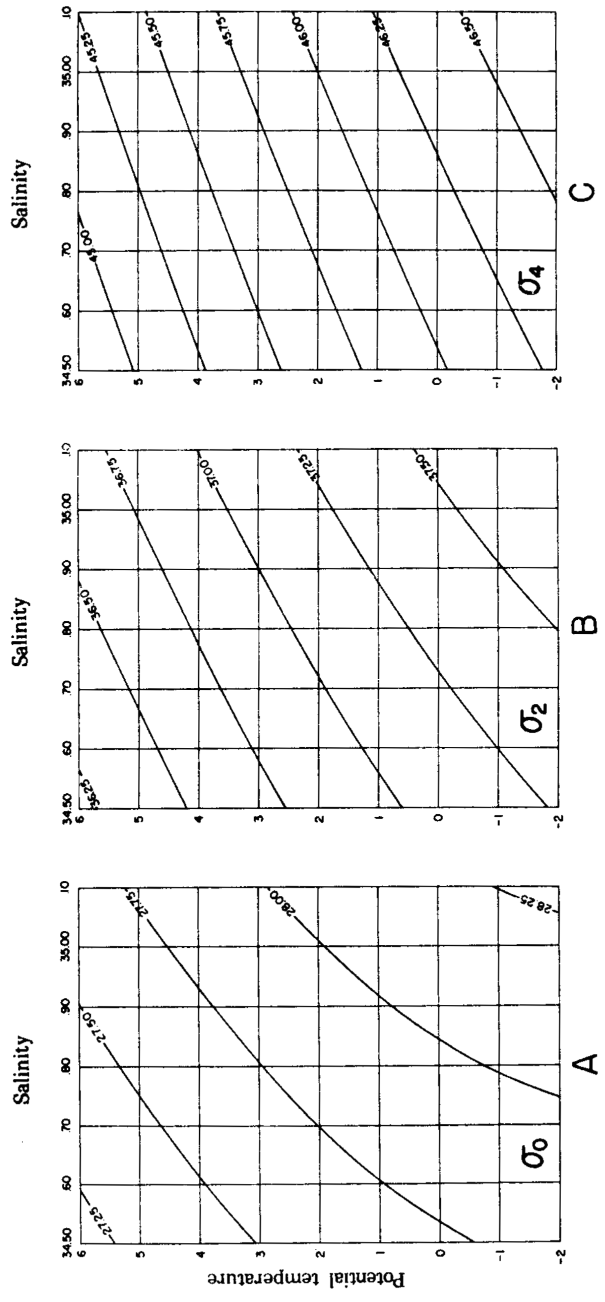

Fig. 1. Density at sea-surface pressure (σ_0), at 2000 decibars pressure (σ_2) and at 4000 decibars pressure (σ_4) as functions of salinity and potential temperature.

One simple interpretation of such diagrams is that a water sample, whose characteristics are θ, S, and p (pressure), can be more usefully represented on a θ–S diagram if a segment of the isopycnal appropriate to its *in situ* pressure is drawn through the point. Under the original assumptions, mixing will occur most freely with those other water parcels which lie adjacent to it along the same isopycnal, so long as the pressure does not vary. If a water parcel moves to a different depth, the slope and value of the isopycnal segment will change and it should mix with water parcels along a different curve.

However, this does not lead to a uniquely determined set of lines or surfaces along which all water parcels are miscible under the desired criterion of no change in the density field, or a minimum of work. This can be demonstrated in a simple way by considering the following exchanges.

Extreme examples

Consider a water parcel at the sea surface with $\theta = 2°C$ and $S = 34{\cdot}95‰$. Its σ_0 (or density* determined from its potential temperature and salinity at zero decibars sea pressure; this has been called potential density† and its usual symbol is σ_θ) is about 27·95. If this parcel is moved adiabatically to a depth where the pressure is 4000 decibars, its density (in σ notation, here called σ_4) will be about 45·96. If it mixes at this pressure along the 45·96 isopycnal and achieves a θ of 0·93°C and an S of 34·70‰, and is then moved adiabatically back to the sea surface, its density (σ_0) will be about 27·83. The net change through the adiabatic and lateral paths has been from a σ_0 of 27·95 to a σ_0 of 27·83. Lateral (isopycnal) mixing at one depth (pressure) has thus changed the *in situ* density that the parcel would have at another depth (pressure). If the lateral mixing with cooler, less saline water had taken place at the sea surface, and then the parcel had been moved adiabatically to 4000 decibars, the σ_4 would have been greater than 45·96. Thus it seems impossible to consider lateral mixing without taking into account the pressure at which the mixing occurs. There is no reason to think that exchanges of this sort may not occur in the ocean: a water stratum may be formed near the sea surface with a particular *in situ* density, sink to another pressure, mix laterally, and rise again in another area, arriving at its original pressure level with an *in situ* density different from its original value. (All of this may take place without violating any of the assumptions about lateral mixing.) Conversely, surface waters with different potential densities may freely (without violating these assumptions) mix with (or supply) the same deep-water density stratum.

However, the extreme example given above is not from the real ocean. There are only a few restricted areas where water sinks to great depths, and the temperature and salinity values it encounters there are not so widely different from its own. It thus seems worthwhile to investigate whether the actual circumstances of the ocean are such that the formation and flow of the major water masses can usefully be studied by

*The density parameter we use is actually $\sigma = 10^3\,(\rho\text{–}1)$ where ρ is specific gravity, not density. Density in g/cm^3 is slightly greater than ρ, but we shall use 'density' instead of 'specific gravity' for brevity of expression and to remain compatible with earlier usage.

†We use the ordinary definitions of 'potential density' and 'potential temperature' (attained after moving the parcel adiabatically to the sea surface), though they might be generalized to refer to any pressure (i.e. 4000 db, as RUPP (1969) has done).

Fig. 2. Positions of the meridional vertical sections shown in Figs. 3a, 3b, 3c and 9. Darkened areas represent the major elevations separating the principal basins.

[*facing p.* 1066]

consideration of exchanges along isopycnal surfaces in which the density is referred to appropriate pressures.

Application of the method

In this exploratory study we have used θ–S diagrams at only three pressures (0, 2000 and 4000 decibars) in following the water, but through examination of the data we conclude that closer spacing would not have significantly altered these results.

We take the surface defined by the density parameters we have chosen as representing a stratum of water of finite thickness. For example, the temperature–salinity characteristics that will be seen along this surface at depths greater than 3000 m fall into only five of the temperature–salinity classes (0·5°C by 0·10‰) whose volumes were calculated by MONTGOMERY (1958) for the World Ocean, but these five classes account for about one-quarter of the ocean volume. Our surface thus can be taken to represent a substantial part of the ocean, and we prefer to refer to it as a stratum.

THE CHOICE OF THE STRATUM

We have chosen to examine a stratum that lies within the layer of high salinity that extends continuously from the North Atlantic to the abyssal Indian and Pacific oceans. SVERDRUP (1931) and WÜST (1951) have discussed this layer, and STOMMEL and ARONS (1960) have based one of their models upon its continuity. In the Indian and Pacific oceans the abyssal water has σ_4 values near 45·92, and this value is found also in the deep waters of the North Atlantic. The continuity of this water mass is clear not only from the distribution of salinity but also from that of σ_0: this density is found at or near the σ_0 maximum* that extends above the bottom in the western Atlantic from about 45°N to 32°S, and beyond there it is either beneath or within the principal salinity maximum of the deep water.

We use this parameter to define the stratum at depths below 3000 m. Where the stratum rises above this depth (in high latitudes), density parameters referred to lower pressures are chosen to define the stratum.

The stratum chosen extends to the sea surface in the Norwegian–Greenland and Barents seas and in the Weddell Sea near the ice shelf, corresponds closely to the salinity maximum of the Antarctic Circumpolar Water and southern Indian and Pacific oceans, and illustrates most of the deepest waters of the Indian and North Pacific oceans. The whole is intended to illustrate the general circulation of a substantial part of the ocean and the nature of the flow and exchanges that take place along this stratum between waters whose extreme characteristics are formed and maintained at distant sources.

*LYNN and REID (1968) suggested that the σ_0 maximum lay along about σ_4 = 45·93; we have used 45·92, which also closely follows the σ_0 maximum, because it is better defined by the available data in the abyssal Indian and Pacific Oceans.

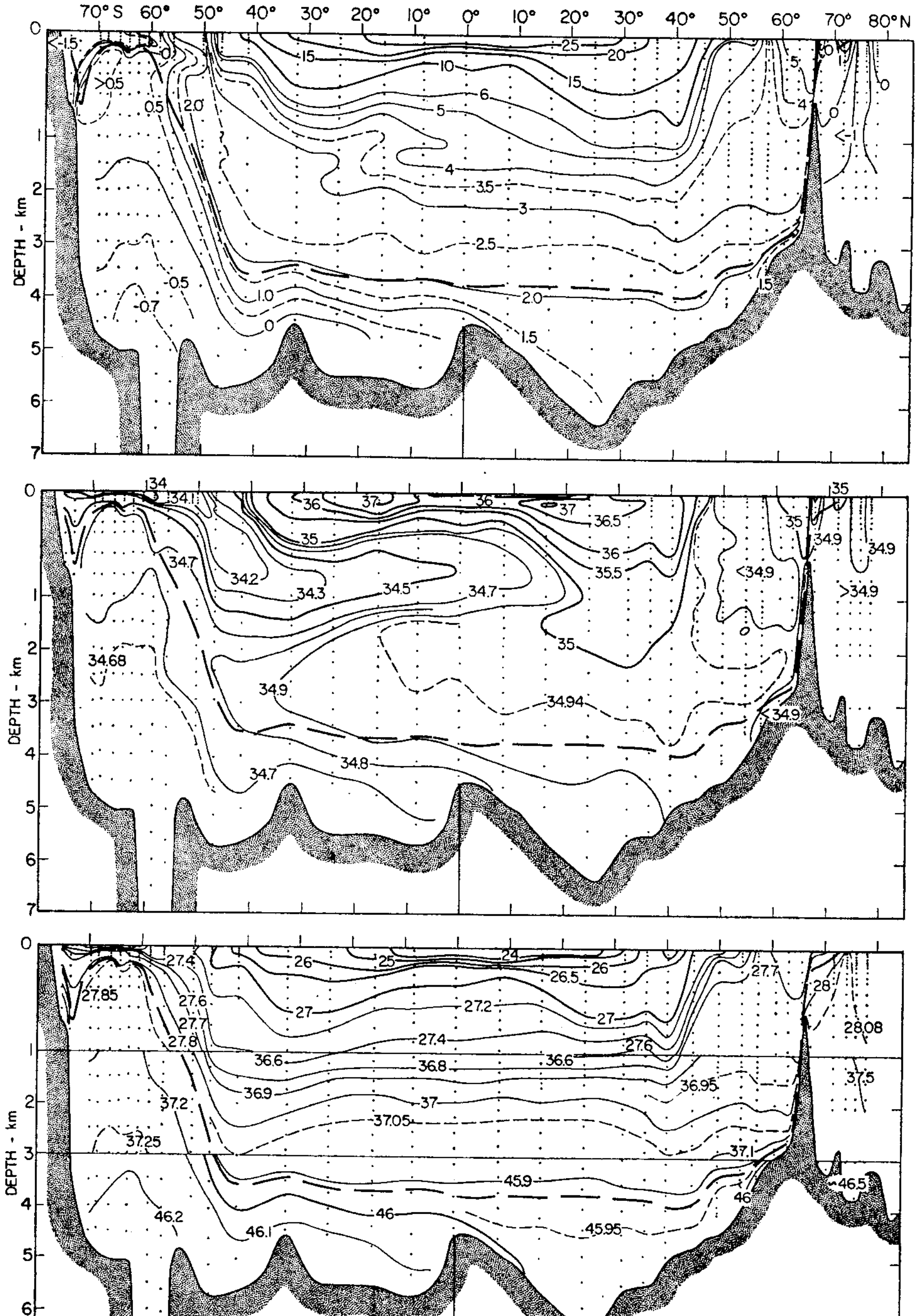

Fig. 3a. Potential temperature, salinity, and σ_0 (0–1000 m), σ_2 (1000–3000 m), and σ_4 (> 3000 m) on a meridional vertical section in the western Atlantic Ocean. On Figs. 3a, 3b, 3c and 11 the vertical exaggeration is $2{\cdot}22 \times 10^3$ above 1000 m depth and $1{\cdot}11 \times 10^3$ below. The heavy dashed line indicates the stratum defined in the text. Stations are identified in the Appendix.

THE LOWER NORTH ATLANTIC DEEP WATER

In the North Atlantic

In describing the waters of the North Atlantic Ocean WÜST (1935) identified two warm, highly saline layers as Upper and Middle North Atlantic Deep Water. Beneath these layers but distinct from the Bottom Water lies a cooler, less saline layer which he called the Lower North Atlantic Deep Water (LNADW). He noted (1933) that this water has a higher potential density than the waters above and below but regarded this as an artifact of errors in the constants used in calculations. In fact, a proper calculation (SCHUBERT, 1935) shows stability despite the σ_0 maximum.

In the North Atlantic south of 50°N (Figs. 2 and 3a) this water lies beneath the warm saline layer and above the cold, less saline water at the bottom. It could not have been formed from a vertical mixture of these two layers, however: it is either colder or more saline than any possible mixture of these layers. It does, therefore, represent a stratum with distinct characteristics (one of which is a potential-density maximum) which must have a distinct origin. We believe that this origin is at the overflow area from the Norwegian–Greenland Sea into the western North Atlantic, and principally in the area between Greenland and Iceland.

The depth where σ_4 equals 45·92 is greater than 3000 m in low and middle latitudes (Figs. 3a, 3b, 3c). Where the surface of σ_4 equal to 45·92 rises to 3000 decibars, we desire to change to a surface defined by a value (or values) of σ_2. If the range of θ and S along the intersection is too broad, then there will be a range of isopycnals (of σ_2, for example) which may be too broad for any single value to be representative. The intersection of the 3000 m surface with the surface where σ_4 equals 45·92 in the North Atlantic occurs only west of the Mid-Atlantic Ridge and in latitudes 47°–63°N, just south of Greenland (Fig. 4a). The length of the intersection is about 2700 km. The θ–S range of the intersection (Fig. 4b) is from 2·04°C, 34·904‰ to 2·12°C, 34·925‰. This range of θ–S, when plotted on a θ–S diagram with isopycnals for 2000 decibars, covers a range of σ_2 from about 37·135 to 37·140. We feel that this range is sufficiently small to justify choosing a single σ_2 value of 37·14 to examine the stratum at depths above 3000 m.

The stratum continues to rise northward beyond the 3000 m level. It intersects the 1000 m surface in the Denmark Strait where at that depth the Strait is no more than 50 km wide (near 65·5°N and 30·5°W). Data are sparse in the region of the intersection and it is not possible to make a good estimate of the θ–S range. Interpolation between stations gives 1·95°C and 34·885‰. A σ_0 value of 27·905 has been chosen, although at this shallower level the range may not be as narrow as at the first transition. The stratum is continuous through the Denmark Straits and extends throughout parts of the Norwegian, Greenland, Barents and Arctic seas at 600 m or less.

Along the stratum that we have defined (Figs. 4a, 4b), the salinity (and temperature) decreases northward through the Denmark Strait, although a wide range of values, including warmer and more saline water, is found in the eastern parts of the Norwegian and Greenland seas. In the Greenland Sea (on the surface where σ_0 equals 27·905) the temperature and salinity range from −0·65°C, 34·68‰ to 3·10°C, 35·01‰, though most of the values are below 1°C and 34·80‰. In the Norwegian Sea the values at 27·905 range from 2·00°C, 34·89‰ to 4·10°C, 35·14‰: part of this warm saline water is flowing northward past Spitzbergen into the Arctic Ocean and eastward

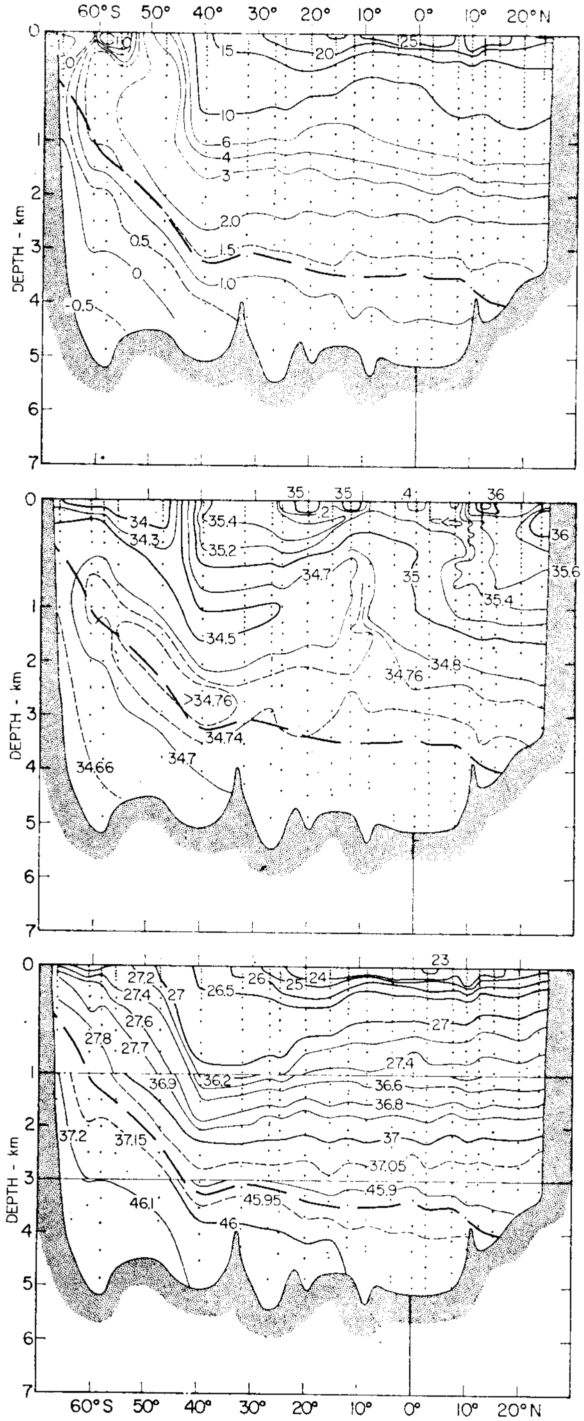

Fig. 3b. Potential temperature, salinity, and σ_0 (0–1000 m), σ_2 (1000–3000 m), and σ_4 (> 3000 m) on a meridional vertical section in the western Indian Ocean. The heavy dashed line indicates the stratum defined in the text. Stations are identified in the Appendix.

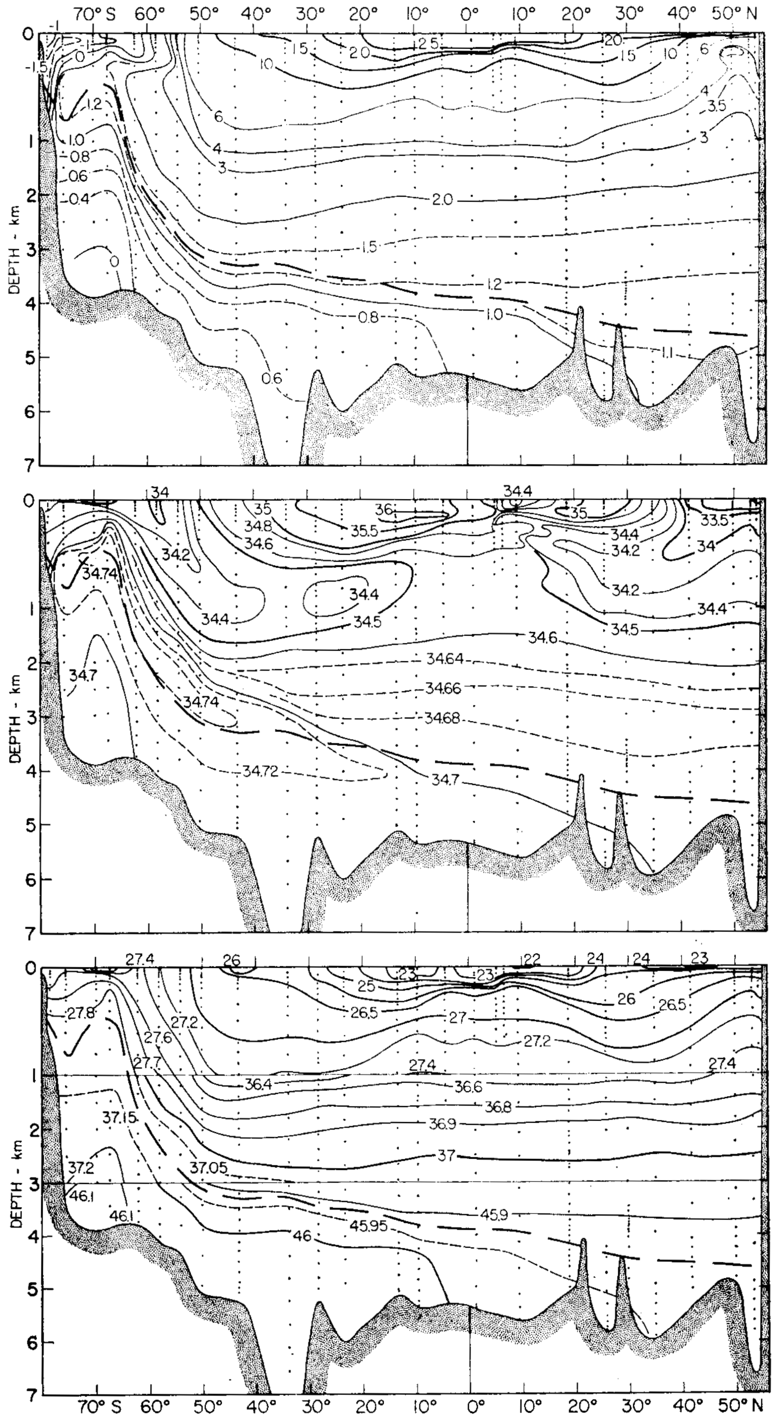

Fig. 3c. Potential temperature, salinity, and σ_0 (0–1000 m), σ_2 (1000–3000 m), and σ_4 (> 3000 m) on a meridional vertical section in the western South Pacific and central North Pacific Ocean. The heavy dashed line indicates the stratum defined in the text. Stations are identified in the Appendix.

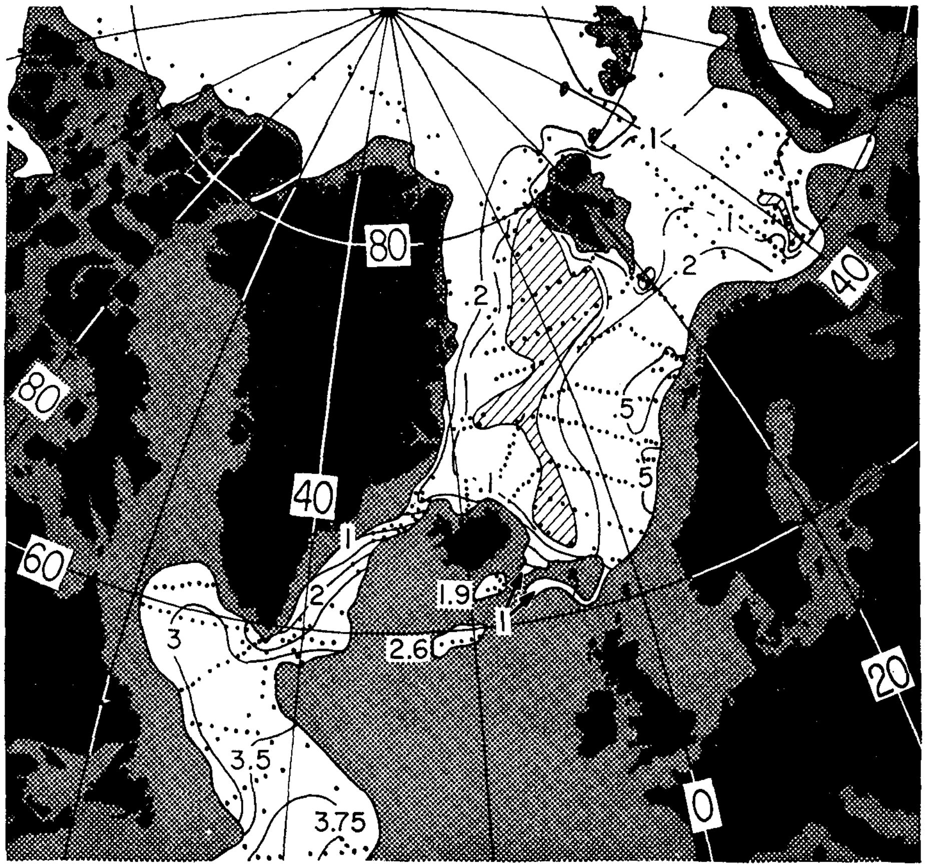

Fig. 4a. Depth (in km) of the stratum where σ_0 equals 27·905 above 1 km, σ_2 equals 37·14 between 1 and 3 km and σ_4 equals 45·92 below 3 km in the northern North Atlantic Ocean and the Norwegian and Greenland seas. Station positions in the Faroe–Shetland Channel are omitted from the figure. Shading indicates areas where the entire column is less dense than this stratum; the area in the Norwegian–Greenland Sea where the entire water column is more dense than this stratum is indicated by hatching. Data are identified in the Appendix.

into the Barents Sea. In addition, north of 62°N this warmer water is separated from the other water of this density in the Greenland Sea: an area near Mohns Ridge contains (at least in winter) no water of σ_0 as low as 27·905. The areal separation and the difference in direction of flow (Norwegian Current flowing north for the warm water, East Greenland Current flowing south for the colder water) suggest that it is the colder waters near Greenland (mostly below 1°C and 34·80‰) that are most available to contribute to the Lower North Atlantic Deep Water if it is to be formed in the Denmark Strait. However, these waters must be made both warmer and more saline to reach the values of 2·12°C, 34·925‰ and σ_2 equal to 37·14 encountered south of the Strait at the 3000 m level.

We cannot clearly establish continuity along the stratum from the high θ and S in the Norwegian–Greenland Sea to the other high values south of Greenland (Fig. 4b).

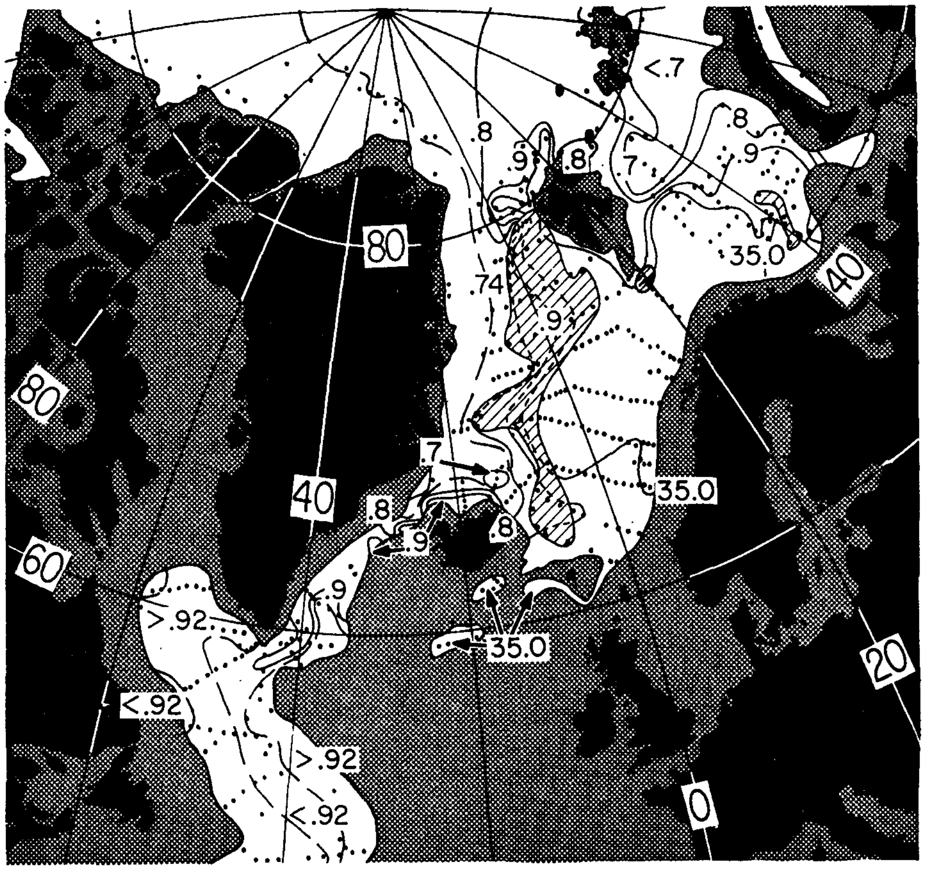

Fig. 4b. Salinity (per mille) along the stratum whose depth is shown in Fig. 4a. The first two digits have been omitted where the salinity value is between 34·0 and 35·0‰.

We therefore seek to account for the latter values by vertical mixing. Some of the warm and saline waters of the Norwegian–Greenland Sea flow southward over the sill east of Iceland along the Reykjanes Ridge, and then turn westward and northward toward the Denmark Strait. This sort of flow has been demonstrated by WORTHINGTON and VOLKMANN (1965) for water of 3·2°C (somewhat shallower than the stratum we have defined). TAIT, LEE, STEFÁNSSON and HERMANN (1967) show that relatively warm and saline water crosses the Iceland–Faroe–Shetland Ridge and lies at depths near 1500 m at the foot of the Ridge; our stratum lies within this water, which is presumably a mixture of Norwegian Sea water with the warmer water south of the Ridge.

CREASE (1965) has shown water of σ_t as great as 28·05 within the Faroe–Shetland Channel. STEELE, BARRETT and WORTHINGTON (1962) find water with σ_t as high as 27·96 (or σ_0 about 27·98) at depths of 2000 m on the slope southeast of Iceland. They calculate this to be Norwegian–Greenland Sea water mixed with warmer water from the south in about equal parts, which flows southwestward past Iceland. However, these high values of σ_0 (27·98) do not appear to be continuous across the Reykjanes Ridge: the water that does cross appears to be less dense.

LEE and ELLETT (1965 and 1967) have investigated the distribution of the Northeast Atlantic Deep Water and pointed out that it crosses the Mid-Atlantic Ridge south of Iceland and extends throughout the northwestern basin. In this basin the Northeast Atlantic Deep Water includes a salinity maximum at a potential temperature of about 3·0°C; on *Gauss* Sta. 222 the θ and S at 2980 m are about 3·0°C and 34·97‰: the σ_2 of this water is 37·05. This is about the densest water that crosses the Ridge, and, after it has crossed, it lies above the water of σ_2 equal to 37·14 which has passed through the Denmark Strait.

Since we are dealing with a density stratum that has a wide depth range, it is more convenient to illustrate this flow by using a density parameter (in this case σ_2) instead of depth as the ordinate in Fig. 5* and to represent depth in the same manner as θ and S. Isobaths and bottom depths are shown at the appropriate σ_2 values. Along 59°30′N, water of our stratum (here at σ_2 equals 37·14) is seen east of the Reykjanes Ridge with salinity as great as 35‰ and θ above 2·5°C. This water has come southward east of Iceland. The highest values of salinity at the bottom occur here and as the water continues to move south, without another source of such saline water, the salinity and density both decrease by vertical mixing. At 53°30′N no values of σ_2 as great as 37·14 are found east of the Ridge, and only a small remnant of the bottom water (against the Ridge) is still above 35‰. South of 53°30′N the water moves westward across the Ridge (where the sill depth is greater) and turns northward again on the west side of the Ridge.

The colder, less saline, but denser waters at 59°30′N from the Greenland Sea have passed southward through the Denmark Strait; they pass southward west of the Ridge beneath the warmer, more saline and less dense waters from the eastern side that are now flowing northward. The vertical mixing that has taken place has cooled and freshened the Iceland–Scotland overflow water and made the Denmark Strait overflow water warmer and more saline. In particular, vertical mixing of waters above and below the stratum has produced waters at the surface where σ_2 equals 37·14 that are more saline and warmer than are found at this density in the Denmark Strait. Characteristics on the stratum have been raised from values of about 2·0°C and 34·89‰ at 59°30′N to values as high as 2·1°C and 34·93‰ at 53°30′N; some observations near this section show salinity as high as 34·94‰ where σ_2 equals 37·14. The lateral maxima in salinity and potential temperature on this stratum in the northwestern Atlantic thus appear to be the product of vertical mixing (at depths of 2·5–3·5 km) of two water masses from shallower levels in the Norwegian and Greenland seas.

Western Atlantic

South of 50°N the depth of this stratum (Fig. 6) on the western side of the Mid-Atlantic Ridge is near 3800 m as far as the equator, except along the coast of North America, where it slopes upward to about 3200 m. The variation of salinity and

*WORTHINGTON and WRIGHT (1970) have used depth as the ordinate in their illustrations of these data and others in the immediate region. MANN, GRANT and FOOTE (1965) and GRANT (1968) also present atlases in which the pertinent details of the temperature and salinity structure in this area can be examined.

potential temperature* is very slight in the North Atlantic south of 40°N, and the tongue of saline, warm water (Figs. 7 and 8) is not defined south of about 35°N. The water there is nearly homogeneous in the lateral sense, and, unlike WÜST's (1933) description of the overlying Upper and Middle North Atlantic Deep Water, we cannot claim evidence of a stronger extension along the western boundary.

The extreme values in the northern part of the western basin are modified gradually toward the south, and the gradients increase. The sharpest increase occurs south of 30°S, probably because of the restriction imposed by the Rio Grande Rise and the Walvis Ridge. South of 40°S the stratum slopes upward again and the Mid-Atlantic Ridge is no longer a major obstacle to exchange between the western and eastern basins.

Eastern basins

High values of S and θ are seen also on the eastern side of the Mid-Atlantic Ridge (Figs. 7 and 8). They reflect the influence of the overlying more saline waters, though the depth of 4600–5000 m is much too great and the temperature (θ) of 2·06°C much too low to be accounted for by direct Mediterranean influence: more likely they also represent the influence of the overflow from the Iceland–Scotland area. The great depth of the stratum in the eastern basins is related to the generally less dense waters found there: only limited amounts of cold Antarctic Bottom Water cross the Ridge through the Romanche Trench. METCALF, HEEZEN and STALCUP (1964) estimate that the controlling depth is about 3750 m, which is within 100 m of our density surface in the area of the Romanche Trench. The greater part of the volume of the eastern Atlantic is filled by warmer and less dense waters. These are principally WÜST's (1935) Upper and Middle North Atlantic Deep Water, which take their characteristics from the Mediterranean and the Labrador–Greenland–Scotland area, respectively. Likewise, the Walvis Ridge severely limits the influx of Antarctic Bottom Water to the eastern basin from the south, and the various components of the North Atlantic Deep Water fill the greater part of this basin. The depth of our stratum lies at nearly 5000 m at 20°S.

Southwestern Atlantic

As the surface where $\sigma_4 = 45{\cdot}92$ extends into the South Atlantic and enters the Antarctic Circumpolar Current, its depth decreases (Fig. 6). A geostrophic balance to this current requires strong horizontal density gradients, and all of the isopycnals slope upward to the south. The surface where $\sigma_4 = 45{\cdot}92$ rises from about 3600 m near 20°S to 3000 m near 45°S, not only in the Atlantic but also in the Indian and Pacific oceans. This radial near-symmetry is important to the method we are using: not only does the intersection of the σ_4 - equals - 45·92 surface with the 3000 m level lie in a very narrow zone around Antarctica, but the characteristics of the water along the intersection have a narrow range, and this means that the shift from our σ_4 surface

*On a surface defined by a single density parameter the potential temperature is uniquely defined by the salinity, but we have used three density parameters (Fig. 6). A salinity of 34·70‰ on the surface where σ_0 equals 27·83 defines a θ of 0·95°C; where σ_4 equals 45·92, it defines a θ of 1·18°C. Therefore, charts of both S and θ are presented (Figs. 7 and 8), though the features are not very different. We have not resolved all of the minor differences in contouring that may occur in the interpolation of such related fields.

to a σ_2 surface is reasonably well defined. The total range of salinity along this intersection is from 34·72 to 34·82‰ and the corresponding values of σ_2 are from 37·09 to 37·11; a value of σ_2 equal to 37·10 has been chosen to define the stratum farther south, to depths of 1000 m.

Indian and Pacific oceans

Warm saline water extends eastward from the southwestern Atlantic into the Indian Ocean; north of about 50°S the depth of the stratum is greater than 2 km and it falls rapidly to 3200–3400 m near 30°S (Fig. 6). In the far north the depth is near 4 km. Salinity and temperature decrease northward from about 40°S (Figs. 7 and 8), reflecting the influence of the underlying Antarctic Bottom Water (Fig. 3b). The distribution of temperature and salinity suggests the existence of a western boundary current on the eastern side of Madagascar.

The tongue of warm saline water extends into the Pacific at slightly lower values (cf. GORDON, 1967). There is also a suggestion of a western boundary current near 180° longitude as far as 20°S (WARREN, 1970). Northward from there the stratum lies between the Intermediate Water and the remnant of the Antarctic Bottom Water (Fig. 3c), and the temperature and salinity decrease.

In the North Pacific it is confined to the central area and its depth is greater than 4000 m; vertical gradients are weaker there than elsewhere and its depth is not precisely defined by the data. Below 4000 m the potential temperature is almost constant within the measurement limitations and salinity increases only slightly. Minor errors in salinity measurements can account for vertical displacements of several hundred meters. Indeed, the values of σ_4 below 4000 m in the North Pacific vary but little from the 45·92 that defines this surface; in the North Pacific this stratum represents, in effect, the bottom water.

Around Antarctica

The effect of the warm saline water from the North Atlantic can be seen clearly in the northern Indian and Pacific oceans, but there is substantial evidence of Antarctic characteristics along the southern edge of the warm saline tongue.

The depth of the stratum continues to rise toward the south: the intersection with the 1000 m level lies near 60°S except south of Africa, where it is at about 50°S and in the southeastern Pacific, where it is nearer Antarctica, at about 70°S. Along this intersection, the salinity varies from 34·70‰ to 34·75‰ (Figs. 6 and 7). These S and corresponding θ values are very much like those at 3000 m, and the range of σ_0 they define is from 27·82 to 27·84; a value of 27·83 has been chosen to extend the stratum farther upward and southward. (The lowest values of salinity and temperature along the 1000 m intersection of the σ_2 - equals - 37·10 surface occur in the Scotia Sea. They are about 34·704‰ and 1·11°C, and correspond to a σ_0 value of 27·82. In this region we might have chosen slightly lower values of S and θ, where σ_0 is 27·82, but we believe the differences are not significant.)

In the South Atlantic the lower North Atlantic Deep Water meets the Antarctic Circumpolar Water (SVERDRUP, JOHNSON and FLEMING, 1942), which is a mixture of the Lower North Atlantic Deep Water and water from the colder, less saline Antarctic, principally from the Weddell Sea. This Circumpolar Water extends along the Circum-

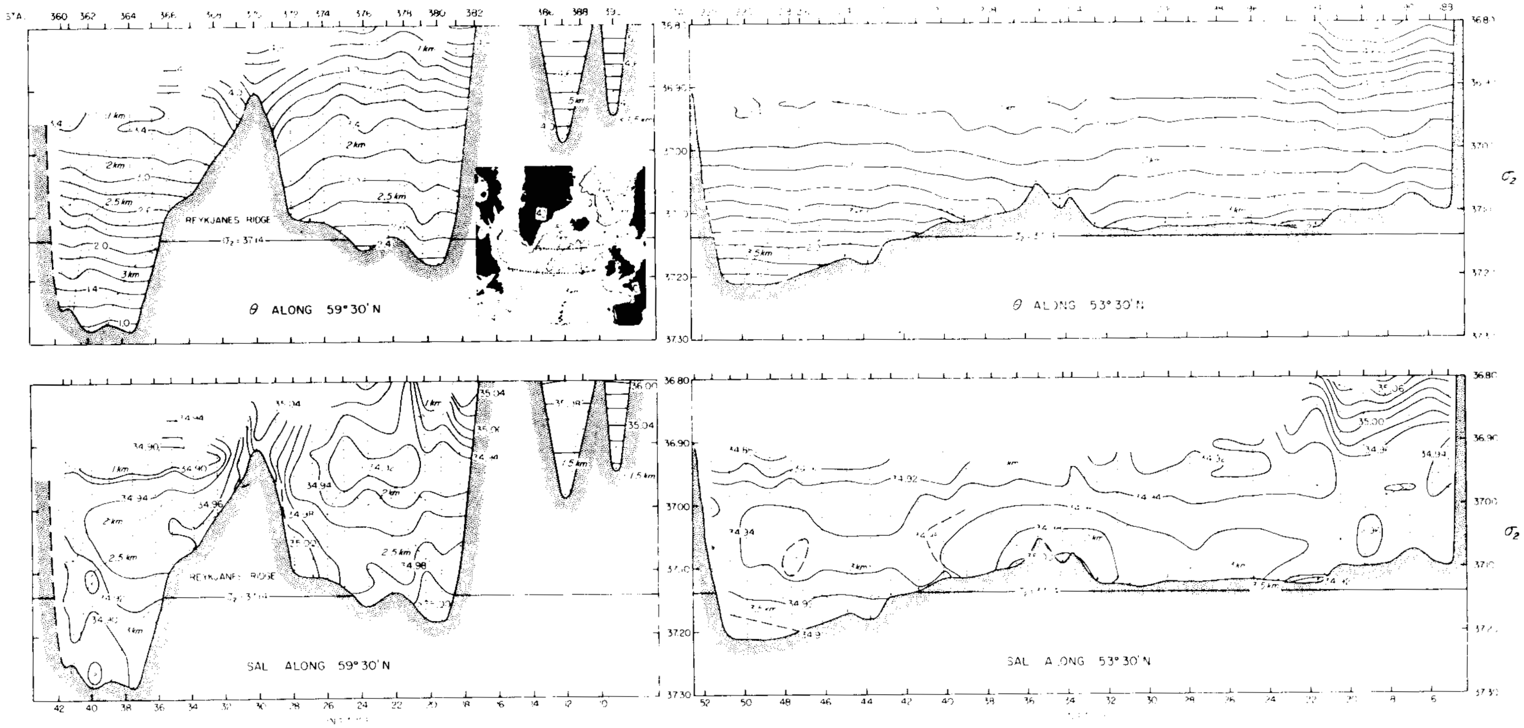

Fig. 5. Vertical sections along 59°30'N and 53°30'N of θ, S, and depth; the ordinate is σ_2. Values at depths greater than 1000 m only are shown.

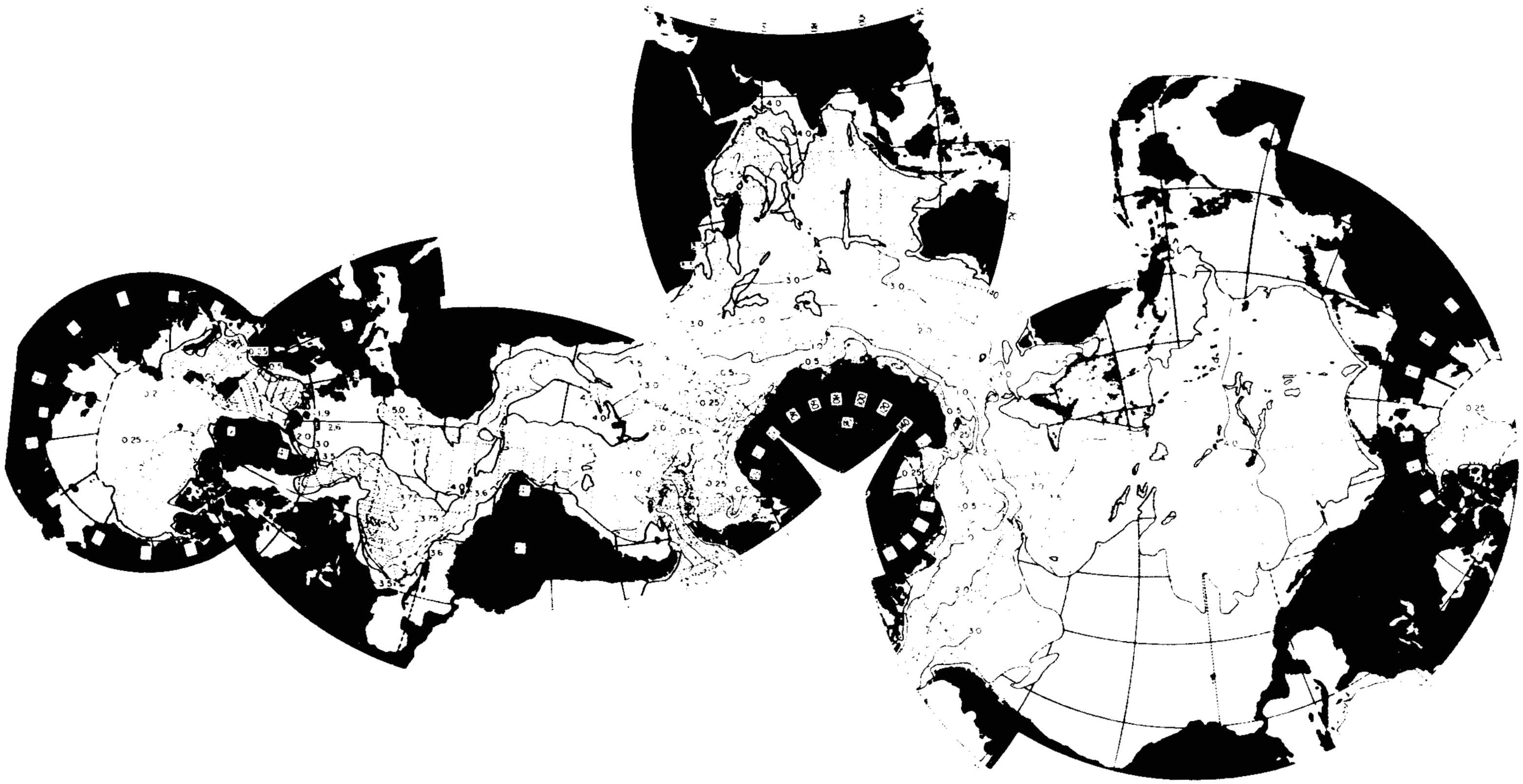

Fig. 6 Depth (in km) of the stratum where [illegible] equals 45.92 (below 3 km) and the corresponding [illegible] and [illegible] values above 3 km. Data are identified in the Appendix.

[1076b]

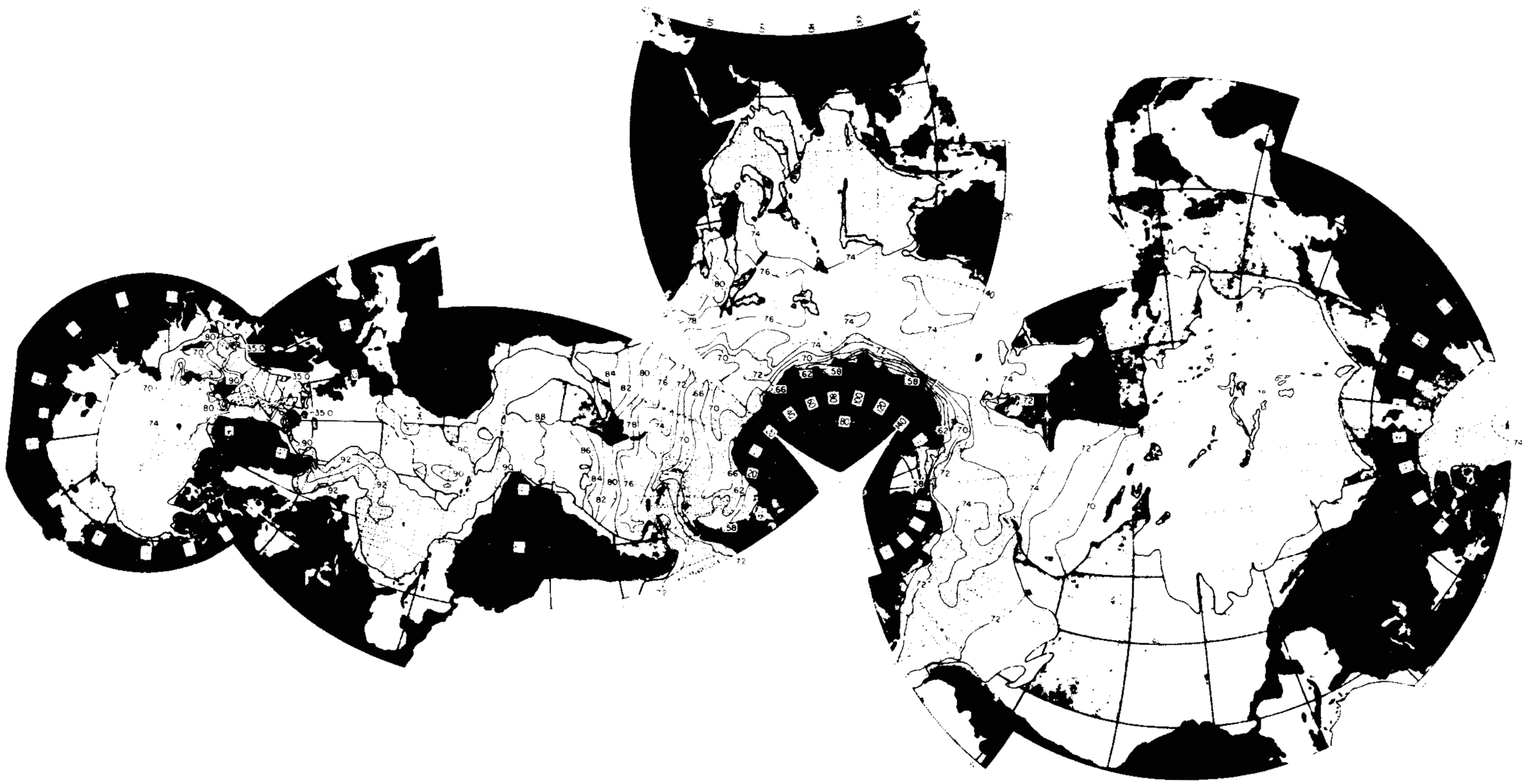

Fig. 7. Salinity (per mille) along the stratum where σ_4 equals 45.92 (below 3 km) and the corresponding σ_2 and σ_3 values above 3 km.

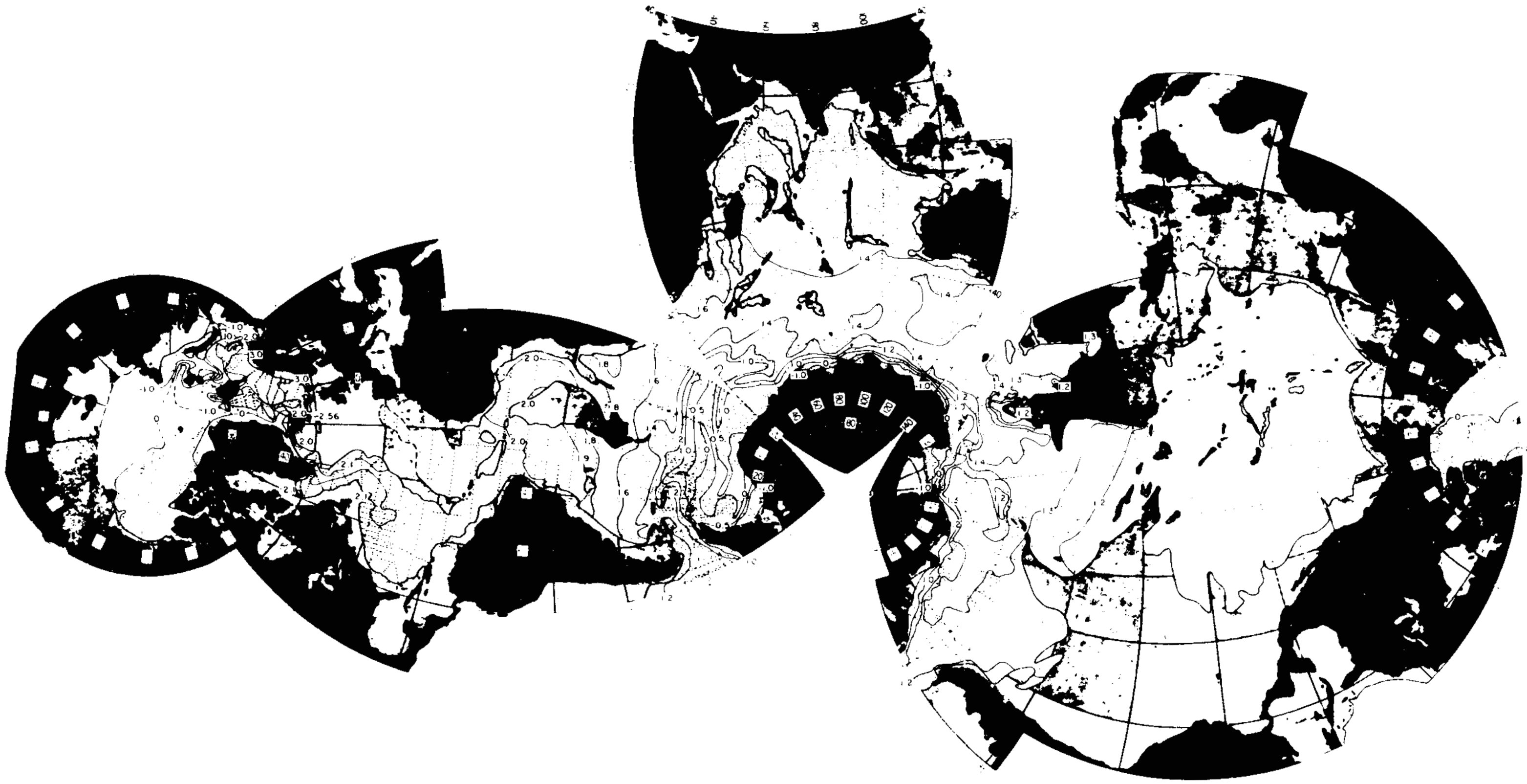

Fig. 8 Potential temperature (°C) along the stratum where σ_4 equals 45.92 (below 3 km) and the corresponding σ_2 and σ values above 3 km

[1076d]

polar Current through the Indian and Pacific oceans as a tongue of water warmer and more saline than the waters on either side (Figs. 7 and 8). It returns to the Atlantic through the Drake Passage (CLOWES, 1933; GORDON, 1971). It is this returning water, with S and θ near 34·72‰ and 1·2°C, rather than the colder, fresher waters of the Weddell Sea, that the Lower North Atlantic Deep Water first encounters at the northern limb of the Weddell Sea cyclonic gyre.

The stratum has colder, lower-salinity values where it lies near the surface around Antarctica. In some areas near the coast these values approach the freezing point (about −1·90°C at 34·55‰). However, it is remarkable that the low values do not extend to still lower latitudes, but are interrupted by the high-salinity, warm tongue that extends eastward from the South Atlantic all around Antarctica.

Indeed, a tongue of warm saline water returns westward from the Indian Ocean into the Weddell Sea; the area of less than 250 m along about 60–65°S from 15°E to 50°W longitude appears to be the axis of the Weddell Sea cyclonic gyre, with the shoaling related to the geostrophic circulation. Temperature and salinity are correspondingly low along this axis as a consequence of vertical mixing with the overlying water. However, no salinity less than 34·59‰ or θ less than −0·90°C has been observed on the stratum in that area. The overlying surface salinity there is quite low (mostly less than 34·10‰) and the temperature minimum lies above the 27·83 surface. A general winter overturn to this depth in the gyre seems unlikely.

The stratum intersects the sea surface in a small area near the ice shelf in the southwestern Weddell Sea; at several stations nearby it lies at depths less than 150 m underlying surface water of relatively high salinity (about 34·50‰). It seems likely that in this region (where bottom depth is 500–800 m) a substantial area may have no water with σ_0 as low as 27·83 in winter.

This stratum is shallow north of the Ross Sea in what appears to be a feature comparable to the Weddell Sea gyre. Otherwise, near Antarctica it is slightly shallower than 500 m, except in the Southeast Pacific, where it lies between 500 and 1000 m.

THE PATH OF THE WATERS THAT FILL THE ABYSSAL INDIAN AND PACIFIC OCEANS

The extension of the Lower North Atlantic Deep Water toward the North Pacific is illustrated on a vertical section (Fig. 9) which follows the lateral salinity maximum (Fig. 7) from the North Atlantic through the southern Indian Ocean to the South Pacific and then extends northward with the western boundary current along the Tonga–Kermadec Ridge, and then northward through the North Pacific. The section (Fig. 9) is meant to indicate the source, path, and modification of the more saline waters that fill the abyssal Indian and Pacific oceans. Since the stratum varies in depth but is nearly everywhere deeper than 3 km along this section, the ordinate chosen is σ_4. Since our stratum acquires its extreme characteristics only after two water masses from the Norwegian–Greenland Sea have mixed south of the Denmark Strait, there is not a unique path to follow north of the Strait. The path north of *Erika Dan* Sta. 316 (numbered at the top of the figure near the left) is taken from Fig. 3a, and in the Norwegian–Greenland Sea passes through the area which contains no water of density as low as that of the parameter we use.

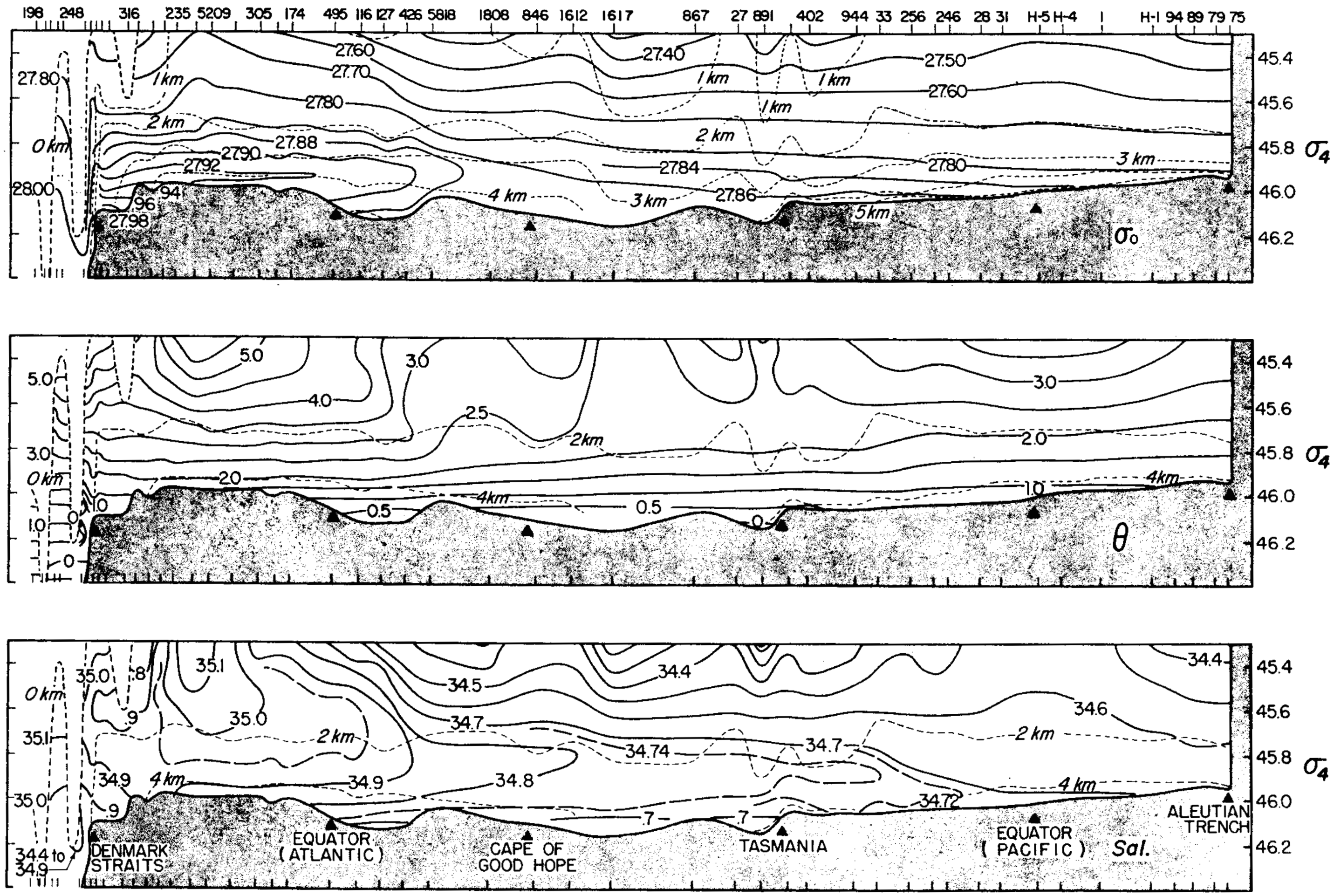

Fig. 9. Potential density, potential temperature, and salinity along a vertical section from the Norwegian Sea to the North Pacific (position of scetion shown in Fig. 2). The ordinate is σ_4, the *in situ* density a parcel of water would have if moved adiabatically to a pressure of 4000 decibars. Depth is shown by the short dashed lines.

Potential density

It is clear that the high potential density near the bottom south of the Denmark Strait has originated in the Norwegian–Greenland Sea. This maximum in σ_0 is seen to be above the bottom (Fig. 9) from about 45°N to the central South Atlantic. As a conservative quantity σ_0 can directly trace the effect of the Norwegian–Greenland Sea water at least this far.

Beyond the Atlantic Ocean σ_0 increases monotonically with depth. There is no comparable source of warm (2°C), saline (34·9‰) water at great depths to maintain this inversion, and vertical mixing along the path results in a general decrease of σ_0 not only at the bottom, but also along the stratum.

Potential temperature

The warmest water of $\sigma_4 = 45{\cdot}92$ is found just south of the Denmark Strait as a consequence of the vertical mixing discussed in a previous section. Everywhere along the path the potential temperature is decreasing with depth at the depth of the stratum. The coldest bottom waters encountered (< 0°C) are at the southernmost segment of the path, south of Australia; from there the path turns abruptly northward, and the bottom temperatures rise monotonically to about 1·1°C in the North Pacific. The potential temperature decreases monotonically along the stratum from the North Atlantic to the North Pacific; this decrease reflects the influence of mixing with the underlying colder bottom water of Antarctic origin.

Salinity

In the Atlantic Ocean the surface where σ_4 equals 45·92 is seen to lie between the great salinity maximum and the less saline bottom waters. Only the deeper, cooler waters of the salinity maximum extend beyond the Atlantic Ocean where the stratum lies close to, but just below, the maximum. North of the Tokelau Trough (near 10°S in the Pacific) the maximum salinity is found at the bottom, and the σ_4 of the bottom water is very nearly 45·92.

DISCUSSION

Lateral effects

There exist two areas where water of the chosen density parameters can be formed at or near the sea surface. One is the Norwegian–Greenland Sea, whose products are vertically mixed at depth southwest of Greenland to form the saline and warm extremes of the stratum. The other area (cold and of low salinity) includes the Weddell Sea and regions along the Antarctic Continent where the stratum lies near the sea surface. The direct effect of the Weddell Sea is seen in the Weddell Sea gyre and its extension into the Indian Ocean. At its extreme the Antarctic influence appears to extend to 30°S in the Atlantic where the lateral gradients steepen. The characteristics of the relatively warm saline waters found in the west central South Atlantic (34·86‰ and 1·9°C) decrease to 34·75‰ and 1·5°C in the mid-Indian Ocean. South of 60°S in the Weddell Sea gyre there is a return flow of moderately warm and saline Circumpolar Water from the Indian Ocean. Elsewhere around Antarctica the strongest lateral gradients are quite close to the continent, indicating that the lateral admixture of the extreme Antarctic water is apparently very limited. In the Indian and Pacific oceans the Circumpolar Current is dominated by the saline, warm water derived from the North Atlantic. The zone of this water extends eastward between 40°S and 60°S in the Indian Ocean and 50°S to 70°S in the Pacific.

It is not clear why the lateral Antarctic influence is so small along this stratum and the North Atlantic influence so great. It has long been recognized that the North Atlantic provides warm, saline water of a particular density range that is found over most of the ocean between the two major products of the Antarctic—the Bottom Water, which is cold and of moderate salinity, and the Intermediate Water, which is of low salinity. This does not mean, of course, that the Weddell Sea does not produce water of the North Atlantic density range, but that the bulk of the water in that range has the characteristics of the North Atlantic rather than of the Antarctic. This warm saline water from the North Atlantic extends laterally upward through the Antarctic Circumpolar Current (where the isopycnals rise abruptly to the south) and mixes with near-surface Weddell Sea water of lower salinity. FOFONOFF (1956) has suggested that the characteristics are such that the mixtures may yield only surface water, shelf water, or Bottom Water, all of different density from the North Atlantic Deep Water: in this sense the Weddell Sea may serve as a sink for some of the saline North Atlantic Deep Water and, in particular, may reduce the mass of water in this density range.

Vertical mixing

In the northern Indian and Pacific oceans the temperature and salinity on this stratum decrease to lower values than in the Circumpolar Water. Since there is no lateral source for this cooling and freshening (no water of this density is formed or found near the sea surface in either area), it must be a result of vertical mixing.

The nature of this vertical mixing, and its relation to lateral flow, can be examined through the differences in the θ–S curves. Vertical mixing is required in all of the northern oceans to account for the characteristics, but it is not immediately obvious that water flowing between an overlying high-salinity layer and an underlying low-salinity layer should become more saline in one area and less saline in another.

The explanation for the sense of change lies in the curvature of the θ–S curve at the stratum. Just south of the Denmark Strait $(\delta^2 S)/(\delta\theta^2)$ is positive; i.e. the curve is concave toward higher salinities (Fig. 10, curve A), and vertical mixing should tend to increase salinity near our density value (and hence potential temperature). Southward from the maximum values on the stratum (about 40°N to 60°N; Figs. 7 and 8) the curvature changes sense: the bottom water is much colder and less saline (as the Antarctic source is approached) and the overlying tremendous salinity maximum decreases through mixing with the Subantarctic Intermediate Water. The water along the stratum meets no lateral source of colder, lower-salinity water and does not change so markedly, hence the change in sense of curvature (Fig. 10, Curve B) that occurs just north of the equator in the Atlantic. The remainder of the figure illustrates the further changes in the θ–S curves along the path. The curves remain concave toward lower salinities, as both the underlying and overlying waters become less saline, until ultimately a nearly straight line is achieved near the equator in the Pacific. At this point, and northward in the Pacific, the water is a nearly linear mixture of the overlying Intermediate Water and the modified remnant of Antarctic Bottom Water. The

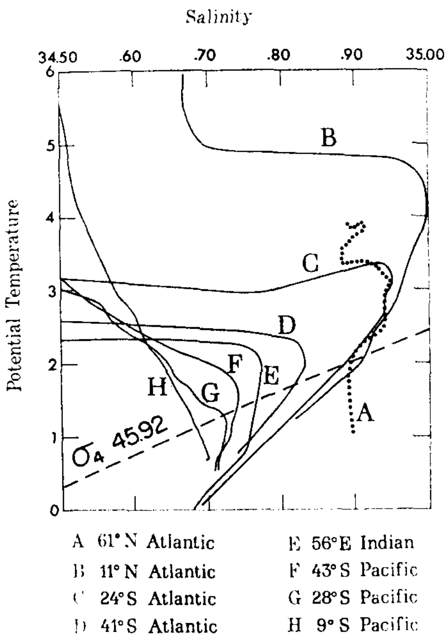

Fig. 10. Temperature–salinity (θ–S) curves at positions along the path of the section shown in Fig. 9.

curvature is gone. This does not mean necessarily that the water in the stratum from the North Atlantic does not extend this far, but that mixing has finally altered its characteristics, and those of the waters above and below, to the extent that a linear θ–S relation exists among the three. Further vertical mixing in the North Pacific cannot destroy the linear relation in the deeper water: its two effects in the northern regions are to mix away the deep end of the line until the final bottom values of about 1·06°C and 34·698‰ are attained, and to swing the upper end of the line toward lower salinities.

Upward flux

If water from the Norwegian–Greenland and Weddell seas sinks from the surface and fills the abyssal Indian and Pacific oceans, then water must rise somewhere, and somewhere there must be vertical flow across the stratum we have defined. It does not seem likely that much upward transport can take place through the salinity tongue. At successively greater distances from the source, the maximum value of salinity is found at higher densities, hence the major erosion of the salinity maximum is by mixing with the overlying lower-salinity water; the deeper values are not so markedly changed.

Where substantial vertical flux takes place, the salinity maximum cannot be maintained. In the northern Indian and Pacific oceans the maximum has disappeared, and these seem to be the most likely places for steady-state upward flux to be important. In the Indian Ocean north of 10°S a source of more saline water lies above the stratum,

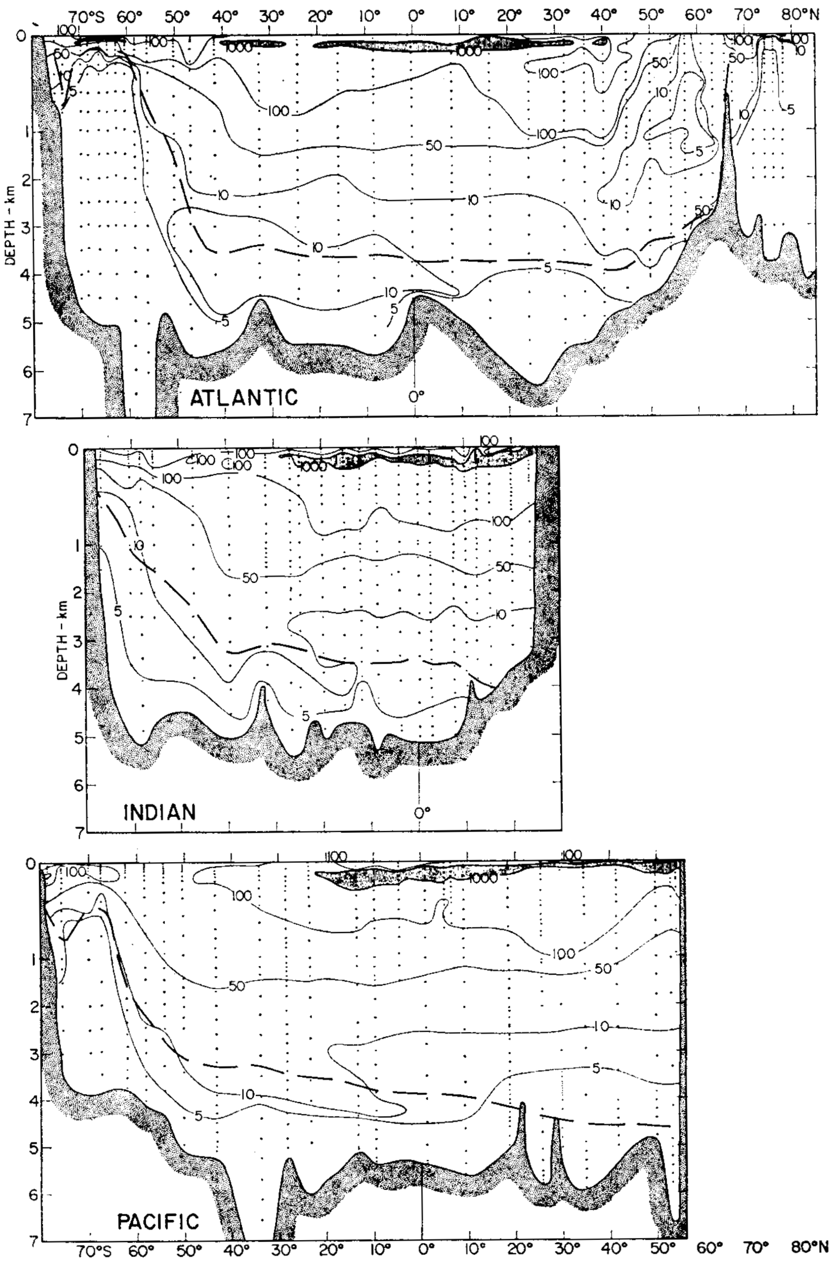

Fig. 11. Stability along the three north–south lines in the Atlantic, Pacific, and Indian oceans. The quantity is $E = \Delta'\rho/\Delta Z$ (10^{-8} g/cm^3/m), after HESSELBERG and SVERDRUP (1914). The scale of the figure does not allow the details of the upper few hundred meters to be represented; some values near zero occur in the mixed layer and some values much greater than 1000×10^{-8} occur within the pycnocline maximum.

and an upward flux of water is required to balance the downward flux of salt. In the North Pacific Ocean the highest salinity is at the bottom, and an upward flux of more saline water is required to balance the excess precipitation at the surface; these areas have already been shown to require vertical mixing to account for the changes in characteristics on the stratum. Additional vertical mixing with the low-salinity Intermediate Water, with a lateral flow of the mixture southward toward or across the equator, may bring this water to shallower levels and start the return path.

Stability at the depth of the stratum

Upward flow and mixing might be expected to take place most effectively in those areas where they are least opposed by the hydrostatic stability. The quantity examined (Fig. 11) is (after HESSELBERG and SVERDRUP, 1914)

$$E = \frac{\Delta'\rho}{\Delta Z}$$

where the $\Delta'\rho$ (g/cm^3) indicates the density difference that would exist between a pair of vertically adjacent samples if the deeper sample were moved adiabatically to the level of the upper sample (a distance of ΔZ, in meters). If one assumes an accuracy of about 10^{-5} g/cm^3 for the calculations of density, an error in E of about 10^{-5} g/cm$^3/\Delta Z$ must be expected. In the deeper parts of the ocean the vertical spacing of samples is typically 300–500 m; the error in E is thus about 2 or 3×10^{-8} g/cm^3/m in typical data observed at this interval. (If the stability were calculated from two parcels 2 km apart vertically, the resolution would be improved and would show the Antarctic to be much more stable than the Norwegian–Greenland Sea, as the stratification in S and θ (Fig. 3a) would suggest.) The values within the 5×10^{-8} contour have no coherent pattern, but merely represent measurement error about some low value of E. From Fig. 11 it appears that the thickest layers of low-stability water are found in the Antarctic, the Norwegian–Greenland Sea, and the North Atlantic between 40° and 60°N and above 3000 m. The first two are recognized as areas where relatively low-density water from near the surface is made colder and more saline, resulting in the formation of dense bottom water; the net vertical transport of water in those areas is downward. In the third, substantial stirring and vertical flow may occur, but the layer does not extend deep enough to remove the bottom water.

The Atlantic section is not significantly different from SCHUBERT'S (1935) section, except that it includes the Norwegian–Greenland Sea and more Weddell Sea daat, The stratum we are investigating is in a fairly stable part of the column as it extends southward from the Denmark Strait; its deep-water maximum in stability is at 60°N at 3000 m depth, where the Norwegian–Greenland Sea water has cascaded over the ridge beneath the less dense waters.

The remaining layers of low stability are found in the deeper parts of the North Indian and the North Pacific oceans. Of these the North Indian is slightly more stable. possibly because of its limited latitudinal extent and the higher temperature in the upper 2000 m. In the Pacific north of 20°N, the very weak vertical temperature gradient (in spite of the opposing salinity gradient) provides the thickest layer of low stability ($< 5 \times 10^{-8}$) surrounding our stratum.

The sudden decrease in stability at about 10°S in the Pacific, through the North

Tokelau Trough (REID, 1969), might be considered a consequence of a blocking of the deepest, densest water. Some possible effect of this is seen (Fig. 3c), but the saline water in the bottom of the North Pacific ($> 34 \cdot 70‰$) shows that the water has been influenced by the North Atlantic; its low temperature ($< 1 \cdot 0$°C in potential temperature) shows a clear effect of the Antarctic Bottom Water. Thus, both North Atlantic Deep Water and Antarctic Bottom Water must enter this region of low stability in the North Pacific. Whether the Trough actually precludes the movement of the densest part of this mixture or merely enforces further vertical mixing to weaken the vertical gradients has not yet been examined. Higher densities are observed in the South Pacific ($\sigma_4 > 46 \cdot 10$; Fig. 3c) than appear in the eastward-flowing water in the Drake Passage (REID and NOWLIN, 1971), indicating that the densest water entering the South Pacific south of Australia must be subject to vertical mixing and must depart, either northward or eastward, at a reduced density.

CONCLUSION

Water originating in a zone of high potential density forms a lateral extension of σ_0 (a conservative quantity) as a maximum above the bottom. Vertical mixing of the waters above and below cannot create or maintain such a feature, but can only dissipate it. The feature extends laterally for (at least) 10,000 km along the western Atlantic before the mixing has sufficiently smoothed θ and S to eliminate the σ_0 maximum. A salinity maximum in the vertical distribution is found in the Atlantic, and it clearly extends, below the surface, with no other source of high salinity, from the North Atlantic to the equatorial Pacific. Vertical mixing or flow can only dissipate this maximum: its maintenance against vertical dissipation is by lateral processes, which maintain the maximum over a distance of 36,000 km.

Although the stratum occurs in the water column all the way from the North Atlantic to the North Indian and North Pacific, it does not necessarily follow that we are dealing with the same water for the entire distance. The bottom water becomes less dense as it extends northward in all oceans (except the North Atlantic north of 30° to 40°, where a new source is met) (LYNN and REID, 1968). This is accomplished by vertical mixing of the bottom waters with those above them. Actually a net vertical flow is required upward from the bottom water to account for the continuous input of bottom water from the Weddell and Norwegian–Greenland seas. Water must somewhere be moving upward across our stratum, and, since the density distribution appears to be steady, a downward diffusion of the characteristics leading to lower density is required.

It can be argued that what we have done by extending this density parameter to the North Indian and North Pacific oceans is to intercept the mixture of North Atlantic Deep Water with Antarctic Bottom Water at the point where its density, decreasing with time, has just achieved the value we have chosen. We believe, however, that the long extensions of the potential-density maximum and the salinity maximum, and the related change in curvature of the θ–S diagrams, may be taken to imply that this interception of new water is limited to the northern parts of the Indian and Pacific oceans. To the extent that this is true, this exercise is useful in examining the deeper circulation.

While vertical processes ultimately limit the concentrations, the very broad-scale lateral extensions of various water masses lead us to believe that lateral processes play a very significant role; vertical processes alone could not create such features nor do they immediately eliminate them. Lateral processes cannot create all of the characteristics of such features, nor dissipate those which terminate at the ocean bottom. What we seem to see is local formation through either vertical or lateral processes, extension over wide areas by lateral movement and mixing, and ultimate dissipation by both lateral and vertical mixing and flow. There is some reason to believe that the northern Indian and (particularly) the northern Pacific are regions where vertical dissipation is most important.

Acknowledgements—This paper represents one of the results of research conducted under the Marine Life Research Program, the Scripps Institution's component of the California Cooperative Fisheries Investigations, a project sponsored by the Marine Research Committee of the State of California, and research conducted under contract with the Office of Naval Research.

REFERENCES

BARKLEY R. A. (1968) *Oceanographic atlas of the Pacific Ocean.* Honolulu, 20 pp., 156 Figs.

BOLIN B. and H. STOMMEL (1961) On the abyssal circulation of the World Ocean—IV. Origin and rate of circulation of deep ocean water as determined with the aid of tracers. *Deep-Sea Res.*, **8** (2), 95–110.

CANNON G. A. (1966) Tropical waters in the western Pacific Ocean, August–September 1957. *Deep-Sea Res.*, **13** (6), 1139–1148.

CLOWES A. J. (1933) Influence of the Pacific on the circulation in the southwest Atlantic Ocean. *Nature, Lond.*, **131**, 189–191.

COCHRANE J. D. (1958) The frequency distribution of water characteristics in the Pacific Ocean. *Deep-Sea Res.*, **5** (2), 111–127.

CREASE J. (1965) The flow of Norwegian Sea Water through the Faroe Bank Channel. *Deep-Sea Res.*, **12** (2), 143–150.

FOFONOFF N. P. (1956) Some properties of seawater influencing the formation of Antarctic bottom water. *Deep-Sea Res.*, **4** (1), 32–35.

FUGLISTER F. C. (1960) Atlantic Ocean atlas of temperature and salinity profiles and data from the International Geophysical Year of 1957–1958. *Woods Hole Oceanogr. Inst. Atlas Series*, **1**, 209 pp.

FUGLISTER F. C. (1963) Gulf Stream '60, Chap. 5, pp. 265–373. In: *Progress in oceanography*, SEARS, editor, **1**.

GORDON A. L. (1967) Structure of Antarctic waters between 20°W and 170°W, *Antarctic Map Folio Series, Am. geogr. Soc.*, Folio 6: 10 pp. 14 plates.

GORDON A. L. (1971) Oceanography of Antarctic waters. *Antarctic Research Series*, **15** (Antarctic Oceanology I), 169–203.

GRANT A. B. (1968) Atlas of oceanographic sections, temperature–salinity–dissolved oxygen–silica, Davis Strait–Labrador Basin, 1965–1967. *Rep. Atlant. Oceanogr. Lab.*, 68-5: 80 pp.

HESSELBERG T. and H. U. SVERDRUP (1914) Die Stabilitätsverhältnisse des Seewassers bei vertikalen Verschiebungen. *Aarb. Bergens Mus.*, No. 14, 17 pp.

KAWAI H. (1966) A generalized potential vorticity in the ocean. *Spec. Contr., Geophys. Inst., Kyoto Univ.*, **6**, 79–93.

KUKSA V. I. (1963) Basic regularities in formation and distribution of intermediate waters in the northern part of the Pacific Ocean. (In Russian) *Okeanologiia*, **3**, 30–43.

LEE A. and D. ELLETT (1965) On the contribution of overflow water from the Norwegian Sea to the hydrographic structure of the North Atlantic Ocean. *Deep-Sea Res.*, **12** (2), 129–142.

LEE A. and D. ELLETT (1967) On the water masses of the Northwest Atlantic Ocean. *Deep-Sea Res.*, **14** (2), 183–190.

LYNN R. J. and J. L. REID (1968) Characteristics and circulation of deep and abyssal waters. *Deep-Sea Res.*, **15** (5), 577–598.

MANN C. R., A. B. GRANT and T. R. FOOTE (1965) Atlas of oceanographic sections, temperature–salinity–dissolved oxygen, Northwest Atlantic Ocean, Newfoundland Basin and Gulf Stream, February 1962–July 1964. *Rep. Bedford Inst. Oceanogr.*, 65-16: 55 pp.

METCALF W. G., B. C. HEEZEN and M. C. STALCUP (1964) The sill depth of the Mid-Atlantic Ridge in the equatorial region. *Deep-Sea Res.*, **11** (1), 1–10.

MONTGOMERY R. B. (1938) Circulation in the upper layers of southern North Atlantic deduced with use of isentropic analysis. *Pap. phys. Oceanogr. Met.*, **6** (2), 55 pp.

MONTGOMERY R. B. (1958) Water characteristics of Atlantic Ocean and of world ocean. *Deep-Sea Res.*, **5** (2), 134–148.

MUNK W. H. (1966) Abyssal recipes. *Deep-Sea Res.*, **13** (4), 707–730.

POLLAK M. J. (1958) Frequency distribution of potential temperatures and salinities in the Indian Ocean. *Deep-Sea Res.*, **5** (2), 128–133.

REID J. L., JR. (1965) Intermediate waters of the Pacific Ocean. *Johns Hopkins oceanogr. Stud.*, No. 2, 85 pp.

REID J. L. (1969) Preliminary results of measurements of deep currents in the Pacific Ocean. *Nature, Lond.*, **221** (5183), 848.

REID J. L. and W. D. NOWLIN, JR. (1971) Transport of water through the Drake Passage. *Deep-Sea Res.*, **18** (1), 51–64.

RUPP S. M. (1969) Temporal variations of the deep water at Ocean Station P. *Tech. Rep. Chesapeake Bay Inst.*, **57** (Ref. 69-10): 38 pp. (Unpublished manuscript).

SCHUBERT O. VON (1935) Die Stabilitätsverhältnisse im Atlantischen Ozean. *Wiss. Ergebn. dt. atlant. Exped. Meteor* 1925-27, **6** (2) (1), 54 pp.

STEELE J. H., J. R. BARRETT and L. V. WORTHINGTON (1962) Deep currents south of Iceland. *Deep-Sea Res.*, **9** (2), 465–474.

STOMMEL H. and A. B. ARONS (1960) On the abyssal circulation of the world ocean—II. An idealized model of the circulation pattern and amplitude in oceanic basins. *Deep-Sea Res.*, **6** (3), 217–233.

SVERDRUP H. U. (1931) The origin of the deep-water of the Pacific Ocean as indicated by the oceanographic work of the *Carnegie*. *Beitr. Geophys.*, **29**, 95–105.

SVERDRUP H. U., M. W. JOHNSON and R. H. FLEMING (1942) *The oceans: their physics, chemistry, and general biology*. Prentice-Hall, 1087 pp., 7 charts.

TAFT B. A. (1963) Distribution of salinity and dissolved oxygen on surfaces of uniform potential specific volume in the South Atlantic, South Pacific, and Indian Oceans. *J. mar. Res.*, **21** (2), 129–146.

TAIT J. B., A. J. LEE, U. STEFÁNSSON and F. HERMANN (1967) Temperature and salinity distributions and water-masses of the region. (Chapter 3). *Rapp. P.-v. Réun. Cons. perm. int. Explor. Mer*, **157**, 38–149.

TSUCHIYA M. (1968) Upper waters of the intertropical Pacific Ocean. *Johns Hopkins oceanogr. Stud.*, No. 4, 50 pp.

WARREN B. A. (1970) General Circulation of the South Pacific. In: *Scientific exploration of the South Pacific*, W. S. WOOSTER, editor., Nat. Acad. Sci., Washington, D.C., 33–49.

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

WORTHINGTON L. V. and W. R. WRIGHT (1970) North Atlantic Ocean atlas of potential temperature and salinity in the deep water including temperature, salinity and oxygen profiles from the *Erika Dan* Cruise of 1962. *Woods Hole Oceanogr. Inst. Atlas Series*, **2**: 24 pp., 58 plates.

WÜST G. (1929) Schichtung und Tiefenzirkulation des Pazifischen Ozeans. *Veröff. Inst. Meeresk. Univ. Berl., N.F.*, (*A. Geogr.-Naturwiss.*) **20**, 1–64.

WÜST G. (1933) Schichtung und Zirkulation des Atlantischen Ozeans—I Lieferung. Das Bodenwasser und die Gliederung der Atlantischen Tiefsee. *Wiss. Ergebn. dt. atlant. Exped. Meteor*, **6** (1) (1), 1–106.

WÜST G. (1935) Schichtung und Zirkulation des Atlantischen Ozeans—II Lieferung. Die Stratosphäre. *Wiss. Ergebn. dt. atlant. Exped. Meteor*, **6** (1) (2), 109–288.

WÜST G. (1938) Das Problem des Antarktischen Bodenstroms in Weltmeere. *Compt. Rend., Int. Geogr. Congr., 15th Amsterdam* 1938, **2** (2b), 45–48.

WÜST G. (1951) Über die Fernwirkungen Antarktischer und nordatlantischer Wassermassen in den Tiefen des Weltmeeres. *Naturw. Rdsch.*, **3**, 97–108.

APPENDIX

We have used about 3700 hydrographic stations out of more than twice that many we considered. Our choices favored observations using salinometers, observations from large expeditions, and winter data (in high latitudes), though it was necessary to use many data not fitting these criteria.

The data were obtained through the U.S. National Oceanographic Data Center. The stations used are too numerous to list in detail, but the great majority are identified below. (Numbers in parentheses are N.O.D.C.'s index numbers.)

Arctic Ocean: Almost all data available through N.O.D.C. archives to 1969 that reached the depth of the stratum.

Norwegian and Greenland seas: Anton Dohrn 1957, 1958, 1959 (060001, 9, 918); *Erika Dan* 1962 (310170); *Gauss* 1958 (580005, 13); *Helland-Hansen* 1958 (58908, 41); *Johan Hjort* 1958 (580035, 957).

Atlantic Ocean north of 40°S*:* All of the I.G.Y. expeditions used in FUGLISTER (1960); Gulf Stream '60 (FUGLISTER, 1963) (310974), *Meteor* (060004); and from the EQUALANT expeditions the *Crawford* (310993), *Explorer* (310990), *Geronimo* (310995) and *Pillsbury* (310996).

Indian Ocean north of 40°S*: Africana* (91050); *Anton Bruun* (31918); *Argo* (310181, 184, 269); *Atlantis* (31197), *Diamantina* (09001, 2, 3, 4, 7, 29); *Discovery* (74007, 30, 37, 40); *Meteor* (06007); *Natal* (91040, 44), *Pioneer* (31201); *Serrano* (31090); *Vema* (31834); *Vityaz* (90010, 32, 34).

Pacific Ocean north of 40°S*: Alexander Agassiz* (311219); *Argo* (311181, 2); *Atlantis* (311219); *Brown Bear* (310558); *Dana* (26009); *Discovery* (740702); *Eltanin* (311212); *Gascoyne* (09005, 33); *George B. Kelez* (310852): *Horizon* (310802); NORPAC (310346); *Pioneer* (310099, 945); *Rehoboth* (310685); *Ryofu Maru* (491219); *Spencer F. Baird* (311217); *Vityaz* (900589, 861, 862); and as yet unpublished Scripps Institution data from NOVA, STYX, SPHERES, and KIKI expeditions.

Southern Ocean south of 40°S*: Argo* (31181); DEEP FREEZE II (31561, 90), III (31593), '61 (31672), '62 (31867, 951); *Deutschland* (06044); *Discovery* (74037, 38, 39, 40, 702); *Eltanin* (311212); *Glacier* (318057); *Meteor* (06004); *Norvegia* (58006); *Vityaz* (90004, 830); and unpublished Scripps Institution data from the PIQUERO expedition.

Because the data are from various times, methods, and institutions, it has been necessary to do some smoothing in order to achieve the contours seen on the maps. In the Norwegian and Greenland seas, where the stratum lies at shallow depths and time variations may occur, we have selected mostly winter and spring data from 1957–59. Beneath the Gulf Stream the data show more fine-scale variation than the figure can illustrate, though the variations are of marginal significance. In the tropical Indian Ocean the data showed a broader range than we have mapped, but no coherent patterns were suggested and we believe that the broad range stems from diversity of method and quality, and that the values shown are not misleading, though the shapes of the contours may be. Between 40°S and 60°S in the eastern Indian Ocean two areas of low θ and S ($<$ 1·4°C, $<$ 34·74‰) are shown in the midst of the saline, warm tongue (Figs. 7 and 8). This is the best resolution we could make of the available data; it is likely to be altered as newer data appear.

Additional and connecting stations to complete the World Ocean section.

Ship	*Originator's Sta. #*	*Lat.*	*Long.*	*NODC Id. #*
Atlantis	5462	11°27′N	50°34′W	310751
Crawford	426	24°20′S	29°38′W	310831
Atlantis	5818	32°35′S	26°45′W	310837
Discovery	1808	41°36′S	0°29′E	740040
	846	40°41′S	23°02′E	740038
	1612	40°28′S	39°03′E	740040
	867	49°26′S	98°22′E	740038
Diamantina	27	44°04′S	122°34′E	090007
Discovery	891	56°03′S	135°11′E	740038
Eltanin	682	51°51′S	150°23′E	318035
	402	54°00′S	159°54′E	310790
Discovery	944	47°42′S	178°16′W	740038

Station list for vertical sections.

Ship	*Originator's Sta. #*	*Lat.*	*NODC Id. #*
		Atlantic Ocean	
G. O. Sars	198	78°16′N	580013
	185	76°18′N	
	179	74°30′N	
	318	72°00′N	580050
Johan Hjort	248	70°45′N	580957*
Anton Dohrn	1595	67°33′N	060001
	2268	65°50′N	060918
	2283	65°15′N	*
	2313	62°19′N	
	2317	61°00′N	
Erika Dan	316	57°26′N	310170
	320	54°26′N	
	339	49°27′N	
Discovery	3520	45°17′N	740622
Crawford	235	40°15′N	310623
Chain	44	36°17′N	310207
Atlantis	5209	32°01′N	310566
Discovery	3615	24°31′N	740622
Crawford	305	16°14′N	310623
	174	8°14′N	310583
	495	0°15′S	310831
	116	8°18′S	310583
	127	15°45′S	
	423	24°16′S	310831
Atlantis	5812	32°26′S	310837
Discovery	72	41°43′S	740035
Eltanin	170	46°50′S	310707
	176	50°21′S	
	131	55°54′S	
	145	58°34′S	
Discovery	1994	60°36′S	740040
	1996	62°32′S	
	1998	64°16′S	
	2000	66°00′S	
	2002	68°19′S	
	2004	69°50′S	
Deutschland	64	73°34′S	060044
	61	75°22′S	
		Indian Ocean	
Anton Bruun	191	23°57′N	310372
Discovery	5067	20°13′N	740007
Atlantis	64	15°18′N	310197
Discovery	5387	12°42′N	740030
Argo	62	10°14′N	310269
	57	7°39′N	310269
	31	2°59′N	310184
		Indian Ocean (cont.)	
Atlantis	139	0°13′N	310197
	144	3°43′S	
	165	8°30′S	
Argo	86	12°00′S	310269
Atlantis	205	15°28′S	310197
	735	20°01′S	310247
	741	24°56′S	
Argo	109	26°51′S	310184
Atlantis	767	31°59′S	310247
	766	32°00′S	
Argo	111	39°45′S	310184
Discovery	1617	47°22′S	740040
	1622	55°37′S	
	2113	58°23′S	
	857	60°40′S	740038
Thorshavn	6	66°34′S	580812
		Pacific Ocean	
Brown Bear	52	54°03′N	310748
Pioneer	75	53°11′N	310099
	79	49°51′N	
	89	41°32′N	
	94	34°46′N	
Argo	H–1	29°36′N	(Zetes)
	H–3	25°44′N	(Nova)
Horizon	1	18°45′N	(Spheres)
Argo	H–4	9°17′N	(Nova)
H. M. Smith	33	8°03′N	310465
	30	5°06′N	
Argo	H–5	1°05′N	(Nova)
Stranger	25	4°00′S	310724
A. Agassiz	31	9°33′S	(*Styx*)
	28	13°25′S	
Gascoyne	246	23°25′S	090033
Eltanin	145	28°19′S	311212
Gascoyne	256	33°52′S	090033
Eltanin	33	43°12′S	311212
	327	50°01′S	310707
	329	54°00′S	
	337	58°03′S	
	341	62°01′S	
Discovery	2232	67°10′S	740040
	1267	69°49′S	740039
Staten Is.	8	75°25′S	310672
Discovery	1644	78°25′S	740040

*Only bottom values were used.