CHAPTER XV

The Water Masses and Currents of the Oceans

Within any given region, the character of the water masses depends mainly upon three factors—the latitude of the region, the degree of isolation, and the types of currents—the relative importance of which will be brought out in the course of the discussion.

It is rational to deal first with the simpler conditions and gradually to enter upon the more complicated. This can be done by first discussing the Antarctic Ocean, where the latitude effect is easily explained, where the waters are in free communication with those of the major oceans, and where the system of prevailing currents is unusually clear-cut. After a description of conditions within the Antarctic Ocean we shall consider the Atlantic, Indian, and Pacific Oceans, the adjacent seas of these oceans, and in conclusion we shall discuss the deep-water circulation between the large ocean basins.

The Antarctic Circumpolar Ocean

BOUNDARIES AND GENERAL CHARACTERISTICS OF WATER MASSES. It is difficult to assign a northern boundary to the Antarctic Ocean because it is in open communication with the three major oceans: the Atlantic, the Indian, and the Pacific Oceans. In some instances the antarctic waters are dealt with as parts of the adjacent oceans and are designated the Atlantic Antarctic Ocean, or the Indian or Pacific Antarctic Ocean, whereas in other instances the antarctic waters must be considered an integral part of all oceans. The latter case, for instance, applies when dealing with the deep-water circulation, which is of such a nature that the intercommunication between different regions must be taken into account. The antarctic waters, on the other hand, can be discussed without entering upon details of the conditions in neighboring oceans, in which case one has to establish oceanographic boundaries between the Antarctic Ocean and the neighboring areas.

On the basis of observations of surface temperatures only, it is possible to divide the Antarctic Ocean into two separate regions and to establish its approximate northern boundary. Near the Antarctic Continent the surface temperature is low, but with increasing distance from the coasts it increases slowly until a region is reached within which an increase of two or three degrees takes place in a very short distance.

This region of sudden increase of surface temperature is one at which part of the surface water sinks, and is called the *Antarctic Convergence*, which has been found to exist all around the Antarctic Continent. Proceeding further to the north, the surface temperature again rises

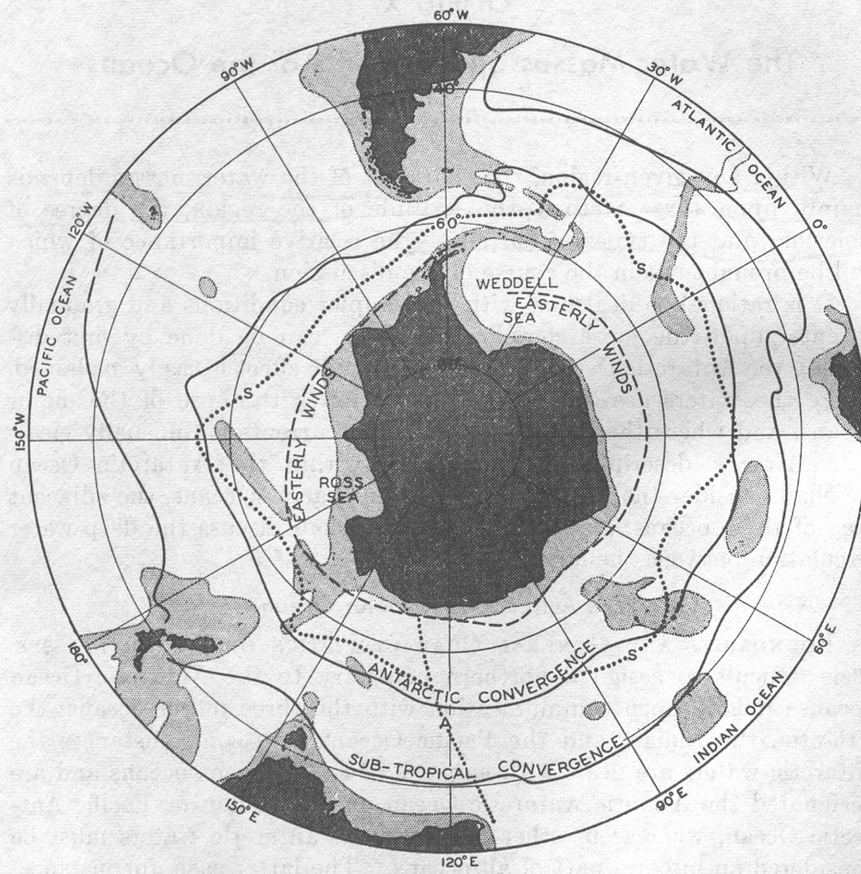

Fig. 158. Locations of the Antarctic and Subtropical Convergences. Locations of the vertical sections shown in figures 159 and 160, and limit between westerly and easterly winds (*dashed line*). *Light shading:* depths less than 3000 m.

slowly until a second region of rapid increase is encountered at the *Subtropical Convergence,* which also has been traced all around the earth except in the eastern South Pacific, where its location is not established. The area extending from the Antarctic Continent to the Antarctic Convergence is called the *antarctic region* and the area between the two convergences is called the *subantarctic region.* From the oceanographic point of view it is logical to consider an Antarctic Ocean extending from the Antarctic Continent to the Subtropical Convergence except in the eastern South Pacific, where an arbitrary limit has to be selected.

Figure 158 shows the location of the two convergences according to the results of the *Discovery* Expedition (Deacon, 1937a). The location of the Antarctic Convergence is, in part, dependent upon the distribution of land and sea and upon the bottom topography. The southward displacement of the Antarctic Convergence to the south of Australia and New Zealand can probably be ascribed to the relative narrowness of the passage between these regions and Antarctica, and the similar displacement off South America can be attributed to the southerly location of Drake's Passage separating South America from Graham Land. The bends of the Antarctic Convergence are to a great extent related to the bottom topography. On the western side of the submarine ridges, the more conspicuous of which are indicated in the figure by the 3000-m contour, the convergence is deflected to the north, and on the eastern side to the south. The reason for this relation to the bottom topography was explained when dealing with the currents (p. 466).

The location of the Subtropical Convergence is similarly related to the distribution of land and sea, but appears to be less dependent upon the bottom configuration. It is less well defined than the Antarctic Convergence, and instead of referring to a *line* of convergence it is, according to Defant (1938), more correct to refer to a *region* of convergence.

Within the antarctic and subantarctic regions the water masses can be classified according to their temperature-salinity characteristics (p. 141). Such a classification is here conveniently based on a study of the distribution of temperature and salinity in a vertical section running at right angles to the Antarctic Continent. Figure 159 is taken from Deacon's discussion (1937a) and shows the distribution of temperature and salinity along a section running south-southwest from Cape Leeuwin, Western Australia, to the ice edge south of Australia in 63°41′S. The two convergences are indicated which divide the area into the antarctic and subantarctic regions. Within the antarctic region a surface layer of low temperature and low salinity is present. Below this surface layer one recognizes a transition layer which increases in thickness towards the north and within which the temperature rapidly increases to values higher than 2°, and the salinity gradually increases to values higher than 34.5 ‰. Below the transition layer one encounters the *Antarctic Circumpolar Water*, the greater mass of which has a salinity a little above 34.70 ‰ and a temperature between 2° and 0°. Water of similar characteristics is found within the subantarctic region below a depth of about 2000 m. That is, the deep water within the subantarctic region has the same characteristics as the Antarctic Circumpolar Water which rises towards the surface near the Antarctic Continent. At station 887, close to the continent, a salinity of 34.67 ‰ is met with at a depth of 200 m, but at station 879 south of the Subtropical Convergence the same salinity is found in the deep water below 2000 m. Close to the Antarctic Continent the temperature of the deepest water is less than 0° and the salinity is less than 34.7 ‰. This cold water represents the *Antarctic Bottom Water.*

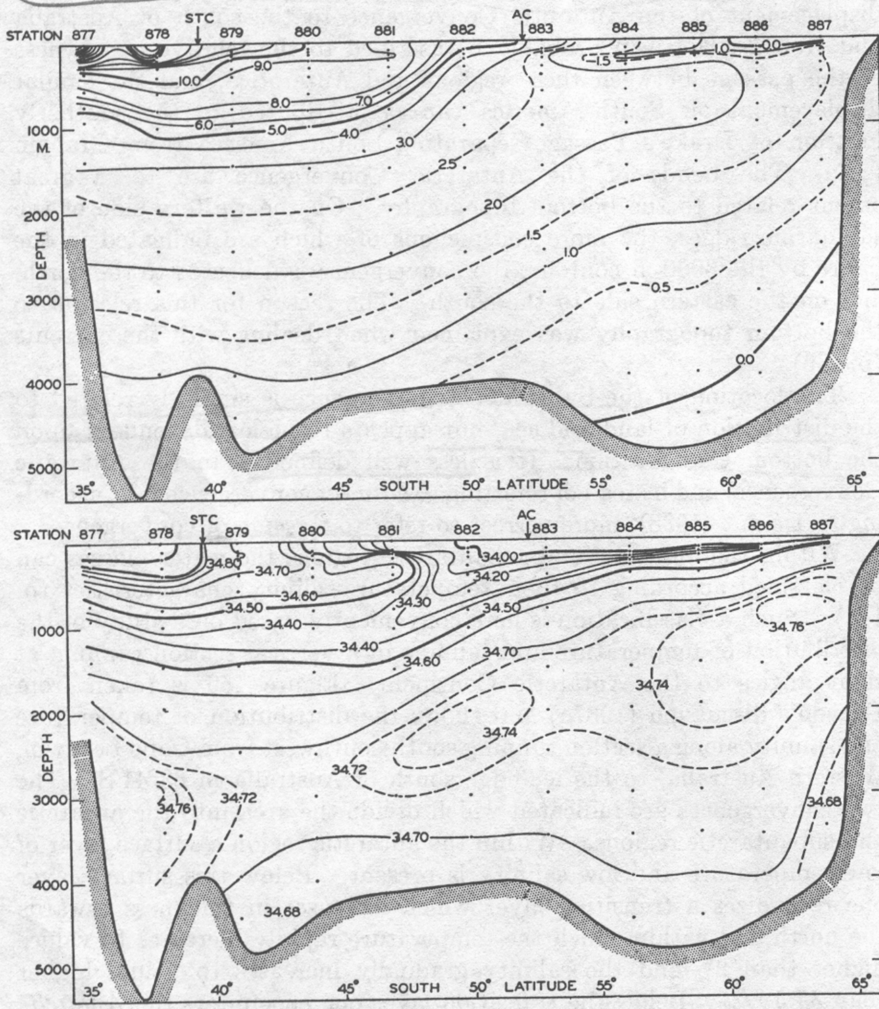

Fig. 159. Distribution of temperature and salinity in a vertical section from Cape Leeuwin, Australia, to the Antarctic Continent (after Deacon). Location of section shown by line *A* in fig. 158.

Within the subantarctic region one can conveniently distinguish between two water masses above the deep water, the *Subantarctic Upper Water* of a uniform temperature between 8° and 9° and of a relatively high salinity, appearing in the salinity section as a tongue extending towards the south, and the *Antarctic Intermediate Water* of a temperature between

3° and 7° and of low salinity, extending towards the north like a tongue that appears to form a continuation of the water in the antarctic layer of transition. The deep water, which is encountered below a depth of about 1500 m, consists of two water masses, the upper deep water which has salinities up to 34.80 ‰ and the lower deep water which is of the same character as the bulk of the Antarctic Circumpolar Water. These different water masses, which will be discussed in greater detail, can be recognized in the southern part of the vertical section of the Atlantic Ocean (fig. 210, p. 748) in which, however, the bottom water extends further north and in which the upper deep water in the subantarctic region has a higher salinity and a higher temperature.

THE ANTARCTIC SURFACE WATER. In winter the southern half of the antarctic region is covered by a thin layer of nearly homogeneous surface water of a thickness of 100 m or less which has a temperature about freezing point and a salinity that varies between 34.0 and 34.5 ‰, the highest value being found in the Weddell Sea. At a station occupied by the *Deutschland* in the Weddell Sea (lat. 65°32′S, long. 43°00′W) on August 26, 1912 (Brennecke, 1921), a nearly homogeneous layer was found between the surface and a depth of 100 m having a temperature of −1.83° and a salinity of 34.47 ‰. With increasing distance from the Antarctic Continent the temperature rises to 0° where the Antarctic Convergence is far south, and to about 1° where this convergence is far north.

In summer the salinity at the surface is decreased owing to the melting of ice. By far the greater amount of the radiation surplus of the season is used for melting ice, but a small amount is used for raising the temperature of the water above freezing point. In the vicinity of the pack ice the surface temperature and salinity vary greatly from one locality to another, because the nearly fresh water produced by melting of ice is not immediately mixed with the surrounding or deeper water. In ice-free areas a more thorough mixing takes place, particularly where strong winds are frequent, and there a well-mixed layer may extend to depths of 55 to 85 m. The temperature of this water is a few degrees above freezing, owing to absorption of heat, but below this warmed layer colder water is found. This colder water may have been formed *in situ* during the preceding winter, but it is also possible that a northward flow of the colder stratum takes place (Mosby, 1934).

On the continental shelves, particularly in the Weddell Sea, the winter cooling and the increase in salinity associated with freezing of ice may extend to the bottom, leading to formation of homogeneous water which, from the surface to the bottom, may have a temperature as low as −1.95° and a salinity as high as 34.70 ‰. The importance of this cold and saline water to the formation of the bottom water will be discussed on page 611.

Antarctic Circumpolar Water. At a short distance below the surface water the Antarctic Circumpolar Water is encountered, which is characterized, in general, by a temperature higher than 0.5° and a salinity slightly above 34.7 ‰. Within this water the temperature generally is at a maximum at a depth of 500 to 600 m, below which it decreases slowly towards the bottom. The salinity is at a maximum at a somewhat greater depth that varies between 700 and 1300 m and below which it decreases towards the bottom. The great uniformity of this water which, all around Antarctica, has nearly the same temperature and the same salinity, is illustrated in fig. 160, which shows a

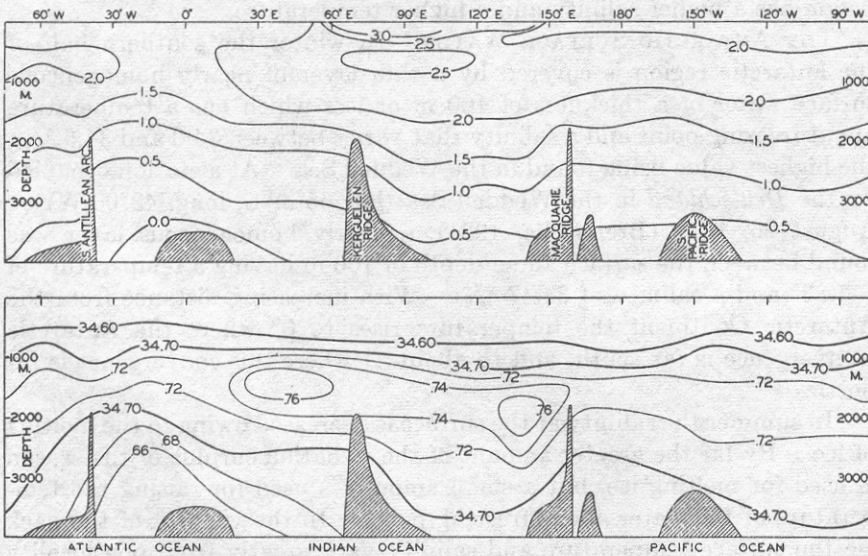

Fig. 160. Distribution of temperature and salinity in a vertical section around the Antarctic Continent. Location of section is shown by line *S* in fig. 158.

vertical section around Antarctica. The section, the location of which is indicated by the line marked *S* in fig. 158, has been placed between two transport lines of the circumpolar current (fig. 163, p. 615) and follows therefore approximately the same water mass. Disregarding the surface water, one finds that within this enormous body of water the differences in temperature hardly exceed two degrees, and below a depth of about 800 m the differences in salinity do not exceed 0.1 ‰. It should be observed, however, that if the section had been placed closer to the Antarctic Continent the temperatures and salinities would have been lower, and if it had been placed at greater distances from the continent the temperatures and salinities would have been higher.

The water which enters the Drake Passage from the west (left margin of section), between 1000 and 4000 m, has a salinity a little higher than

34.70 ‰, the maximum values being slightly above 34.72 ‰. Proceeding to the east, the Weddell Sea bottom water shows up to the east of the South Antilles Arc with salinities as low as 34.66 ‰. The maximum salinity remains at about 34.72 ‰ as far as to long. 20°E, where it increases rapidly, reaching values above 34.76 ‰ in long. 30°E. This increase must be due to admixture of the saline upper deep water from the Atlantic Ocean. Relatively high salinities are found to the south of the Indian Ocean, as indicated by a second maximum of 34.76 ‰ in longitude 140°E to the south of Australia, but when entering into the Pacific area the salinity decreases in the direction of flow, owing to the admixture of the low-salinity Pacific Deep Water. On the whole, the salinity is remarkably uniform and the differences which have been discussed here approach the limit of accuracy of the observations. Along the bottom the isohaline 34.70 ‰ follows closely the bottom configuration, indicating that the water actually flows across the submarine ridges.

The temperature section shows that the water which enters Drake Passage from the west has maximum temperatures only slightly above 2°C and minimum temperatures near the bottom somewhat below 0.5°. Towards the east of the South Antilles Arc the cold bottom water of the Weddell Sea appears, and at higher levels the maximum temperature drops below 2°, probably owing to admixture of this cold water. In long. 10°E the maximum temperature increases and, between long. 60° and 120°E, lies above 2.5°. These high temperatures are due to admixture of relatively warm water from the Atlantic and Indian Oceans. When the water passes into the Pacific area, the maximum temperature again decreases owing to admixture of the colder Pacific Deep Water. The isotherms of 1° and 0.5° do not exactly follow the bottom topography as should be expected if the water crossed the different submarine ridges without mixing taking place, for which reason it appears necessary to assume that processes of mixing exercise some influence upon the distribution of temperature and therefore, also, upon the distribution of salinity.

THE ANTARCTIC BOTTOM WATER. The most extreme type of bottom water is found in the Weddell Sea area, where below a depth of about 4000 m the temperature is about −0.4° and the salinity is about 34.66 ‰. This water is probably formed by the mixing in approximately equal parts of water flowing down from the continental shelf and of Antarctic Circumpolar Water (Mosby, 1934). The former has a temperature, on an average, of about −1.9° and a salinity of about 34.62 ‰, whereas the circumpolar water near the Antarctic Continent has a temperature of 0.5° and a salinity of 34.68 ‰. Expressed by means of σ_t, the densities of these two types of water are 27.89 and 27.84, respectively, and the density of the bottom water is 27.86. The water on the continental shelf, the salinity of which may be increased up to 34.62 ‰ or slightly higher by freezing of ice, has the greatest density and sinks down along the continental slope; but while sinking it is mixed with the warmer and more saline circumpolar water of somewhat lower density, and a bottom water is formed which has a slightly greater density than the deep water.

According to this explanation, first offered by Brennecke (1921) and later substantiated by Mosby (1934), the freezing of ice on the Antarctic continental shelves is of primary importance to the formation of the Antarctic Bottom Water. On a small scale, cooling of water of relatively high salinity may contribute to the formation of bottom water by a process similar to the one which takes place in the waters around Greenland (see p. 656), but this process appears to be limited to some areas in the Bransfield Strait (Clowes, 1934).

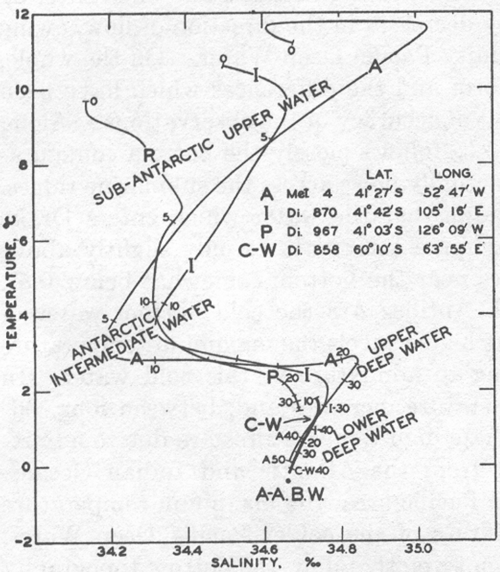

Fig. 161. Temperature-salinity curves at three stations in the subantarctic regions in the Atlantic (*A*), the Indian (*I*), and the Pacific (*P*) sectors of the Antarctic Ocean, and at a station within the region of circumpolar water (*C-W*). Antarctic Bottom Water shown by *A-A.B.W*.

By far the major part of the Antarctic Bottom Water is formed in the Weddell Sea, as is evident from the section in fig. 160, according to which water of a salinity of 34.66 ‰ is present only between 30°W and 30°E. Intermittently, bottom water is also formed south of the Indian Ocean between long. 30°E and 140°E (Sverdrup, 1940), but no evidence exists for such formation to the south of the Pacific Ocean. These features have bearing on the deep-water circulation of the different oceans and we shall therefore return to them later (p. 745).

THE WATERS OF THE SUBANTARCTIC REGION. These waters differ much more both regionally and in a vertical direction, for which reasons a well-defined classification cannot be undertaken. The curves in the *T-S* diagram in fig. 161 illustrate the waters present and show that generally one can distinguish between five different water masses, the Subantarctic Upper Water, the Antarctic Intermediate Water, the Upper Deep Water, the Lower Deep Water, and the Bottom Water.

The three curves representing conditions in the subantarctic region are based on observations in the Atlantic, Indian, and Pacific sectors, respectively. The curve marked *A* represents conditions at *Meteor* station 4 in the western part of the South Atlantic Ocean, the curve marked *I* is based on the observations at *Discovery* station 870 in the Indian Ocean southwest of Australia, and the curve marked *P* on observations at *Discovery* station 967 near the mid-line of the South Pacific Ocean. For comparison, the curve *C-W* has been added to show conditions at a typical station within the Antarctic Circumpolar Water and the point *A-A.B.W.* to show the extreme Antarctic Bottom Water. On the curves the depths are indicated in hectometers.

Within all areas the upper water is characterized by relatively high temperatures and a salinity maximum at some distance below the surface, the highest salinities occurring in the Atlantic area. The Antarctic Intermediate Water is present in all areas. In the Atlantic it is of somewhat lower salinity and is found at a somewhat lesser depth below the surface, but these features are partly related to the distances of the stations from the Antarctic Convergence. At the stations in the Atlantic and Indian Ocean areas the Upper Deep Water shows a relatively high salinity and high temperature, whereas the Lower Deep Water is of the same character as the Antarctic Circumpolar Water, but is found at a greater depth. In the Pacific Ocean the Upper Deep Water is of lower temperature and lower salinity, but the Lower Deep Water approaches the character of the Antarctic Circumpolar Water.

The reason for these regional differences will be discussed when dealing with the deep-water circulation and it is sufficient here to emphasize that the subantarctic waters differ within the three oceans, whereas the antarctic waters are of similar character all around the Antarctic Continent.

THE CURRENTS OF THE ANTARCTIC OCEAN. The more or less permanent currents of the Antarctic Ocean consist principally of two types, the relative currents which are associated with the distribution of mass, and the wind drifts of the surface layers.

The character of the relative currents can be derived from fig. 162 showing the distribution of the specific volume anomalies in the section which was used for demonstrating the character of the water masses (see fig. 159). At any level the water of the smallest specific volume, that is, the greatest density, is found near the Antarctic Continent and, in general, the lines of equal anomaly slope downward from south to north. As a consequence of this distribution of mass, the isobaric surfaces all rise towards the north if the 4000-decibar surface is considered horizontal. This rise towards the north is evident from the upper part of fig. 162, in which are shown the profiles of a series of isobaric surfaces relative to the 4000-decibar surface. The current which is associated with the distribution of mass must be directed away from the reader, that is, from west to east, because in the Southern Hemisphere the denser water is on the right-hand side when looking in the direction of flow. In the section the vertical scales have been greatly exaggerated, for which reason the slopes of the anomaly lines appear to be very great, but actually the slopes are small and the slopes of the isobaric surfaces are much smaller. The accompanying velocities are quite small, as indicated in the upper part of fig. 162, where the numbers entered show

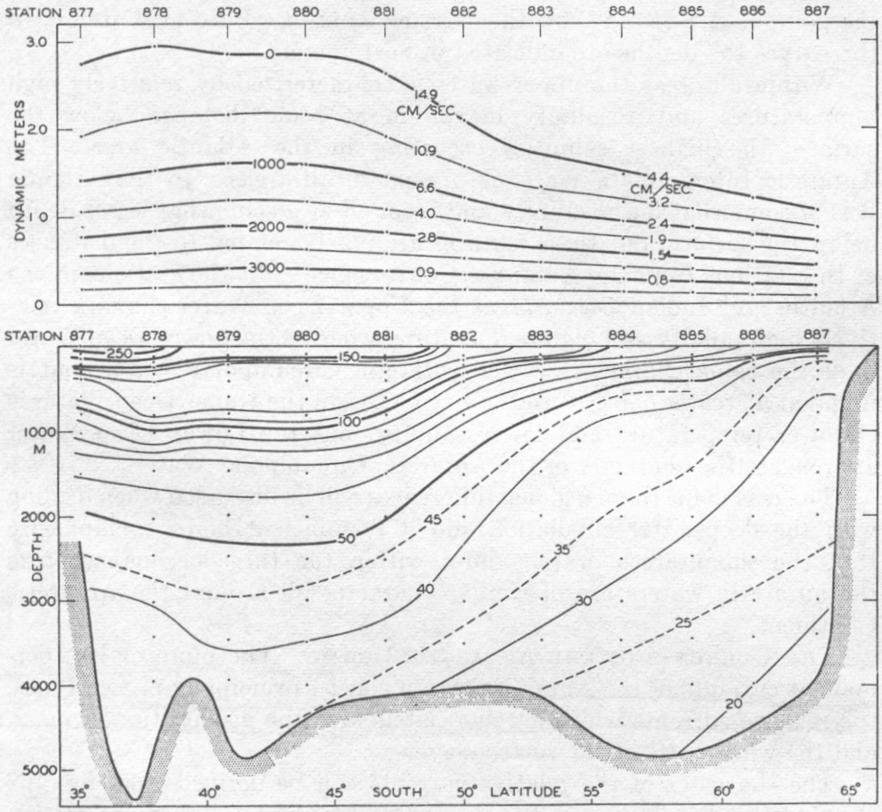

Fig. 162. *Lower:* Distribution of the anomaly of specific volume, $10^5\delta$, in a vertical section from Cape Leeuwin, Australia, to the Antarctic Continent (see fig. 158). *Upper:* Profiles of the isobaric surfaces relative to the 4000-decibar surface.

the horizontal velocities in centimeters per second, assuming that at 4000 m the water is at rest. At the surface a velocity of nearly 15 cm/sec is found to the north of the Antarctic Convergence, but within the antarctic region the surface velocity is only 4.4 cm/sec. At the Antarctic Convergence the isobaric surfaces appear to change their slope abruptly, and in the figure this feature has intentionally been exaggerated because such a break must be present where a wedge of lighter water is found over a body of water of greater density (p. 445).

In general, the current runs from west to east around the Antarctic Continent, but is locally deflected from its course, partly by the distribution of land and sea and partly by the submarine topography. The effect of the submarine topography is evident in Deacon's charts (1937b) of the dynamic topography of the surface or the 600-decibar surface relative to the 3000-decibar surface, and it is also seen from fig. 163,

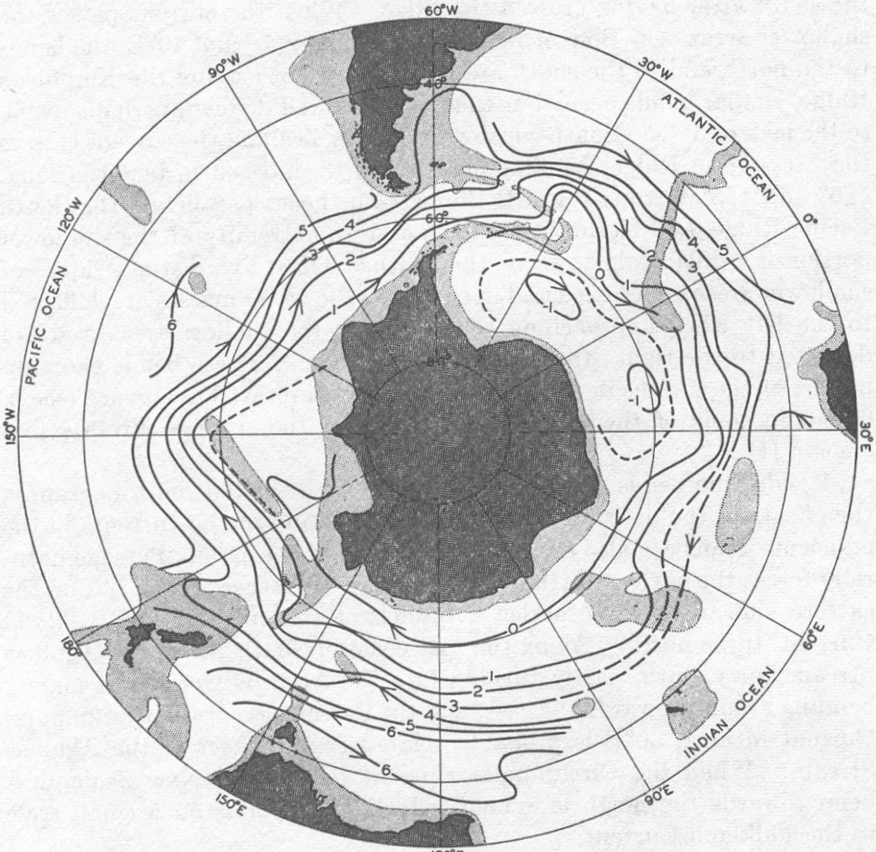

Fig. 163. Transport lines around the Antarctic Continent. Between two lines the transport relative to the 3000-decibar surface is about 20 million cubic meters per second.

which shows the total volume transport relative to the 3000-decibar surface as computed by means of the data which Deacon has used for his presentation, adding the results from a number of *Meteor* stations in the South Atlantic. In the figure the volume transport is shown by lines which are approximately in the direction of transport and which have been drawn with such intervals that the transport between two lines equals 20×10^6 m^3/sec.

Following the current around the Antarctic Continent, it is seen that when approaching a submarine ridge the current bends to the left and after having passed the ridge it bends to the right, in accordance with our preceding discussion of the effect on a deep current (p. 466). Having passed through Drake Passage, the Circumpolar Current approaches the South Antilles Arc, where it first bends to the north and then to the south after having crossed the ridge. When the current passes the shallower areas near Bouvet Island between long. 0° and 10°E, the bends to the north and to the south are repeated; when passing the Kerguelen Ridge, similar bends occur but these are less well defined, perhaps owing to the lack of data. South-southwest of New Zealand the current crosses the Macquarie Ridge, bending sharply north and south, between long. 180° and 155°W, runs towards the east-northeast paralleling the South Pacific Ridge, and having passed the eastern extremity of the shallower portion it bends again towards the south. Thus, five distinct ridges or shallower areas are crossed and in each case the water masses are deflected to the left when approaching the ridges or the shallow areas and are deflected to the right after having passed them. These bends show up in part at the surface in the location of the Antarctic Convergence (see p. 607) and some of them appear on charts of the surface currents (see chart VII).

Besides the bends which are associated with the bottom topography, the effects of the distribution of land and sea and of the currents in the adjacent oceans are also seen. The location of the Drake Passage naturally forces the current further south than in any other region, but on the eastern side of South America a branch of the current, the Falkland Current, turns north. Along the east coast of South Africa the Agulhas Stream flows south, partly turning into the Atlantic Ocean but mainly bending around towards the east, and the narrowness of the Circumpolar Current in long. 30°E is probably related to the effect of the Agulhas Stream. When the Circumpolar Current has passed New Zealand, a bend towards the north is evident which corresponds on a small scale to the Falkland Current.

The transport line marked "zero" is not continuous around the Antarctic Continent, but begins in the Weddell Sea area and ends against the coast of the Antarctic Continent in the region of the Ross Sea. This feature indicates that our representation is not entirely correct and that the transport does not take place exactly along the lines which have been entered, because actual lines of equal transport must be closed lines and cannot end against the shores. Some transport must take place across the lines, and the fact that our line "zero" stops against the continent to the west of Drake Passage indicates that in this narrow passage a certain amount of piling up of water takes place and that transport occurs here across the contour lines. Referring the flow to the topography of the isobaric surfaces, this means that in Drake Passage and the Scotia Sea the water does not flow parallel to the contours but flows uphill. Within the Antarctic Ocean this region is the only one in which there is evidence of a marked discrepancy between the lines of equal transport and the actual flow of water, and on the whole it may be assumed that the transport lines give a fairly correct picture of the character of the Circumpolar Current.

A flow to the west near the Antarctic Continent is evident only in the Weddell Sea area, where an extensive cyclonic motion occurs to the south of the Circumpolar Current. The water masses which take part in this cyclonic movement, however, are small compared to those of the Circumpolar Current, and their velocities are small.

Within the subantarctic region the current is also directed in general from west to east but only the southern portion of the waters close to the Antarctic Convergence circulates around the Antarctic Continent and forms part of the Circumpolar Current. In the Pacific Ocean the northern portion belongs to the current system of that ocean and must therefore be dealt with again when describing the conditions in the South Pacific. Thus, a strict northern limit of the Antarctic Ocean cannot be established on oceanographic principles, but a boundary region has to be considered which, depending upon the point of view, may be assigned either to the Antarctic Ocean or the adjacent oceans.

The total transport of the Antarctic Circumpolar Current must be greater than is apparent from fig. 163, which shows only the transport relative to the 3000-decibar surface. According to the figure, the transport through Drake Passage is about 90 million m^3/sec, whereas Clowes' computations (1933) gave 110 million above the 3500-decibar surface. It is therefore probable that the absolute transport between the lines in fig. 163 is not 20 million, but at least 25 million m^3/sec.

This discussion of the Circumpolar Current is based entirely on the distribution of mass as derived from deep-sea oceanographic observations. The surface currents have been determined independently by means of ships' records, and show that the flow of the surface water is governed partly by the distribution of mass and partly by the effect of the prevailing winds. Near the Antarctic Continent easterly and southeasterly winds blow away from the large land masses, but between lat. 60° and 40°S strong westerly winds prevail. The dashed line in fig. 158, p. 606, shows (according to Deacon, 1937a) the approximate boundary between the regimes of the easterly and westerly winds. Correspondingly one finds, as seen on chart VII, westward surface currents prevailing near the border of the Antarctic Continent and eastward surface currents at some distance from the coast. In the Southern Hemisphere the wind drift deviates to the left of the direction of the wind and consequently the eastward surface current shows a component towards the north. A divergence must be present between the westward and the eastward surface currents, drawing deep water towards the surface, and in the sections (fig. 159, p. 608) this divergence shows up close to the continent by the high temperature and high salinity of the water at a depth of 100 m. Between the Antarctic and Subtropical Convergences the surface current is generally directed towards the east, but on the northern side of the Antarctic Convergence it shows a component to the south, and at the southern boundary of the Subtropical Convergence, a component to the north. The component to the south must be associated with the development of the Antarctic Convergence, which has not yet been satisfactorily explained.

Deacon attributes the Antarctic Convergence to the character of the deep-water circulation. He points out that the convergence is found where the relatively warm but saline deep water rises, and he reasons that within the antarctic region the upper water, being heavier than the warm surface water further north, would sink if it were not prevented by the highly saline deep water. He writes (1937a, p. 23):

> As soon, however, as the northward current has passed the point where the deep water climbs steeply towards the surface it is no longer prevented from sinking, and the sections suggest that the antarctic water flows over the steep ascent of the warm deep water like a stream over a waterfall.

Sverdrup (1934), on the other hand, suggests that the convergence is brought about because in the subantarctic region a thermohaline circulation dominates, which would carry the light surface waters to the south, whereas in the antarctic region a wind circulation dominates, which would carry the surface water to the north. It remains to be seen whether either of these explanations is acceptable or whether a combination of them, or perhaps an entirely new point of view, must be introduced.

As a consequence of the divergence and convergence at the surface, a transverse circulation must be present which is superimposed upon the general flow from west to east. Such a transverse circulation is evident from the vertical sections in fig. 159, p. 608, and fig. 210, p. 748. In these, between 45° and 63°S, the deep water seems to climb from a depth of about 3000 meters to within 200 meters of the surface. The deep water does not appear to reach the very surface, but within the entire antarctic region a considerable amount of deep water of high salinity must be added to the surface layer, because otherwise the salinity would be lowered owing to excess of precipitation and the addition of fresh water by the melting of antarctic icebergs. A southward flow of deep water is also required for balancing the transport of the surface water to the north by the prevailing winds. In the section shown in fig. 159, sinking of water near the Antarctic Continent probably does not take place, but in the Weddell Sea (fig. 210) this sinking is of great importance, and a movement of the deep water towards the continent is also necessary there to compensate for the formation of bottom water. Part of this bottom water flows away from the Antarctic Continent, but part mixes with the deep water and returns with this water to the antarctic regions.

At the Antarctic Convergence, water of relatively low salinity and low temperature sinks. A small portion of the sinking water appears in some areas to return towards the south at a depth of a few hundred meters, but the greater part continues towards the north, forming the tongues of Antarctic Intermediate Water which in all oceans can be traced to great distances from the antarctic region. It is probable that this water gradually mixes with the underlying deep water and returns to the Antarctic with the deep water (p. 750). Within the subantarctic regions the upper water appears to flow towards the south, but the nature of this flow is not fully understood.

The antarctic water which leaves the surface has a temperature of 2.2° and a salinity of 33.80 $^0/_{00}$ in the Atlantic area (Wüst, 1935). In the Indian Ocean area and to the south of Australia, similar values are found (Sverdrup, 1940) which appear applicable to the Pacific area as well, because the Antarctic Convergence follows the isotherms and isohalines of the sea surface (Deacon, 1937a). Thus, water of temperature 2.2°, salinity 33.80 $^0/_{00}$, and $\sigma_t = 27.02$, represents a *water type* (p. 143) that is continuously or intermittently formed near the Convergence, where it sinks. When sinking, it is rapidly mixed with surrounding waters and a *water mass*, the Antarctic Intermediate Water, is formed which mainly spreads towards the north, being characterized at its core by a salinity minimum. Owing to continued mixing, the characteristic T-S relation of the intermediate water changes as the distance from the Convergence increases, and layers of gradual transition are found both above and below the water mass, for which reason no distinct boundaries can be introduced. The best method for following the spreading out of the intermediate water and the results of mixing is the "core method" which was introduced by Wüst (p. 146), who, in the Atlantic Ocean, examined the temperature and salinity at the core of the intermediate water and determined the depth of the core.

On the basis of these considerations one arrives at the schematic picture of the transverse circulation which is shown in the block diagram in fig. 164. In the diagram the formation of bottom water has been taken into account and, furthermore, it is indicated that the deep-water flow towards the Antarctic Continent is strengthened by addition of both intermediate and bottom water. The main features of the transverse circulation are represented by the sinking of cold water near the Antarctic Continent, the climb of the deep water towards the surface, the northward flow of surface water, and the sinking of the Antarctic Intermediate Water.

The climb of the deep water towards the surface is of great biological importance, because through that process the amount of plant nutrients in the surface layer is being constantly renewed; the enormous numbers of organisms found in the antarctic water is therefore closely related to the type of transverse circulation which is present.

From the foregoing it is evident that one cannot consider the Antarctic Circumpolar Water mass as a body of water which circulates around and

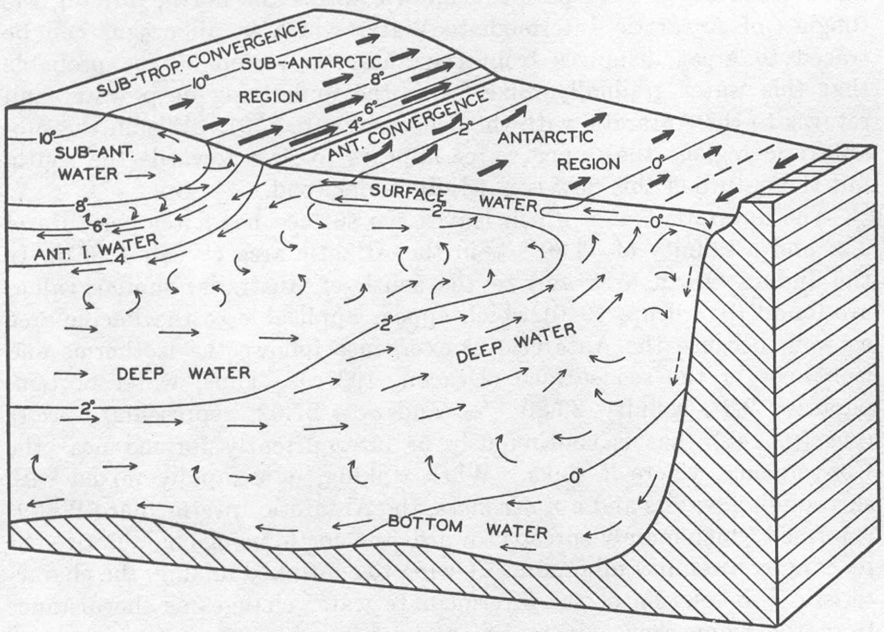

Fig. 164. Schematic representation of the currents and water masses of the Antarctic regions and of the distribution of temperature.

around the Antarctic Continent without renewal. On the contrary, one has to bear in mind that water from the antarctic region is carried towards the north and out of the region both near the surface and near the bottom, and that deep water from lower latitudes is drawn into the system in order to replace the lost portions. Through external processes, cooling and heating, evaporation and precipitation, freezing and melting of ice, and through processes of mixing, the temperature and salinity in a given locality remain nearly unchanged over a long period, except for such changes as are associated with displacements of the currents. Thus, the stationary distribution of conditions which characterize the entire Antarctic Ocean represents a delicate balance between a number of factors which tend to alter the conditions. On the other hand, an individual water particle, which describes a most complicated path, is subjected to great changes. It may lose its identity by mixing with adjacent water masses or it may have its temperature and salinity radically altered if brought to the surface.

When one examines the dynamics of the Circumpolar Current, frictional forces have to be taken into consideration. The stress which the wind exerts at the sea surface cannot be balanced by a piling-up of water in the direction of flow (p. 488) because the current is continuous around the earth, and therefore it must be balanced by other frictional stresses. The velocities along the bottom are too small to give rise to bottom friction, and one is therefore led to the conclusion that great stresses exist at the boundaries of the current due to lateral mixing. According to Sverdrup (1939a) such lateral mixing takes place. The

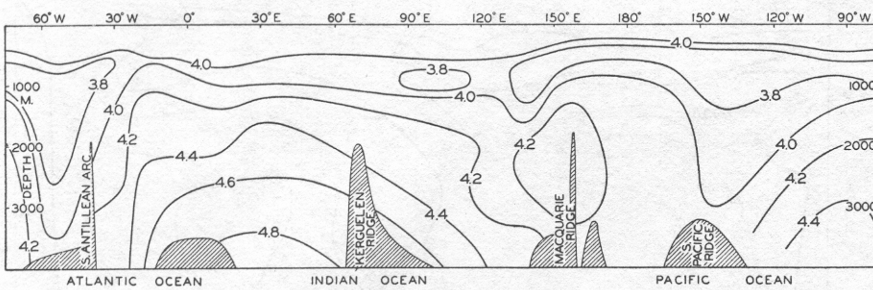

Fig. 165. Distribution of oxygen (ml/L) in a vertical section around the Antarctic Continent. Location of section is shown by line S in fig. 158. See also fig. 160.

torque exerted by these stresses along the quasi-vertical boundary surfaces of the current must balance the torque exerted by the wind on the surface, but so far no attempt has been made at an analysis of the dynamics of the current, taking friction into account.

Oxygen Distribution. The distribution of oxygen around the Antarctic Continent is illustrated by the section in fig. 165, which corresponds to the sections in fig. 160. The water which enters the Drake Passage from the west has a low and uniform oxygen content below a depth of 600 m. A minimum oxygen content of about 3.7 ml/L is found at a depth of about 1200 m, and near the bottom the oxygen content is only slightly higher than 4.0 ml/L. To the east of the South Antilles Arc the oxygen content is considerably higher, evidently because the bottom water of the Weddell Sea contains a relatively large amount of oxygen. In the section, values higher than 4.8 ml/L are shown and the minimum value, which is here encountered above a depth of 1000 m, is only slightly less than 4.0 ml/L. As the water proceeds further towards the east, the oxygen content generally decreases both within the layer of minimum oxygen content and within the deep and bottom water. The decrease, however, is irregular, and more data are needed in order to establish definitely that such a decrease takes place and, if so, the order of magnitude of the decrease (Sverdrup, 1940).

The variation of oxygen content in a north-south direction is illustrated in fig. 166, representing the *Discovery* section to the south of Western Australia, and in the southern parts of the sections in fig. 210, p. 748, and fig. 212, p. 753. In fig. 166 the high oxygen content of the surface waters decreases towards the north, owing to the increasing temperature. The water which sinks at the Antarctic Convergence has a relatively high oxygen content, but the content decreases so rapidly towards the north that no intermediate maximum develops. A layer of minimum oxygen content, within the subantarctic zone, is found at a

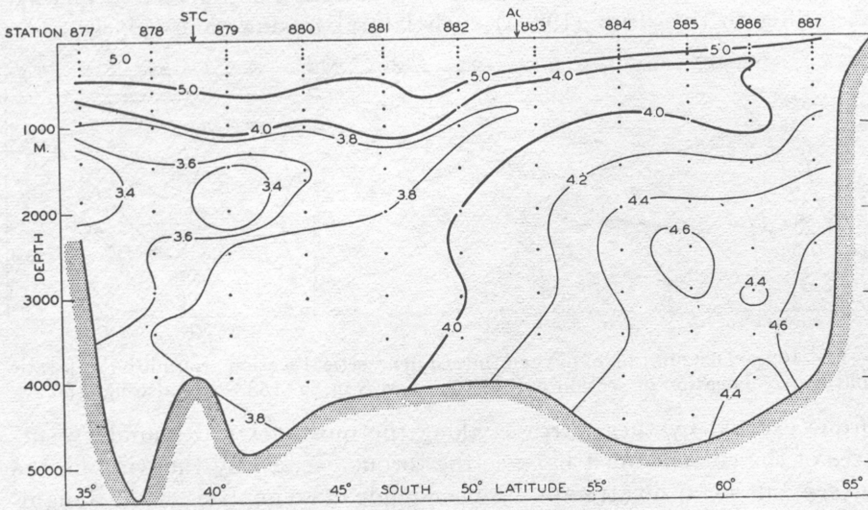

Fig. 166. Distribution of oxygen in a vertical section from Cape Leeuwin, Australia, to the Antarctic Continent (after Deacon). Location of section shown by line *A* in fig. 158. See also figs. 159 and 162.

depth of nearly 2000 m and this layer rises to the south. In the antarctic region, the layer of minimum oxygen content is encountered at a depth of about 600 m. The Antarctic Circumpolar Water has a higher oxygen content than the deep water further north, and the Antarctic Bottom Water which is found near the Antarctic Continent shows the highest oxygen content with values above 4.6 ml/L.

In the Weddell Sea area which is shown in fig. 210, p. 748, the bottom water has an oxygen content higher than 5.5 ml/L, and close to the continental slope values up to 6.5 are encountered. These high values confirm the conclusion that the greater amount of the Antarctic Bottom Water is formed in the Weddell Sea region. Mosby (1934) points out that the shelf water which contributes to the formation of the bottom water has an oxygen content of about 7 ml/L, whereas the deep water with which the shelf water is being mixed when sinking has an oxygen content of about 4.5 ml/L. If these two types of water are mixed in approximately equal proportions, the resulting water, the bottom water, should have an oxygen content of about 5.75 ml/L, in good agreement with the observed values which are higher than 5.5 ml/L.

To the north of the Weddell Sea the Antarctic Intermediate Water is characterized by a maximum of oxygen below which is present a layer of minimum oxygen content which rises towards the south. The deep water shows a higher oxygen content than that in the same latitude to the west of Australia. Within the subantarctic waters, then, regional differences are present not only in temperature and salinity (p. 612) but in oxygen content as well.

ICE AND ICEBERGS. Two forms of ice are encountered in the Antarctic Ocean, *sea ice* which is formed by freezing of sea water and *icebergs* which represent broken-off pieces of glaciers. The appearance of both sea ice and icebergs varies widely. Several classifications have been proposed (Transehe, 1928, Smith, 1932, Maurstad, 1935, Zukriegel, 1935) and several codes for reporting ice have been introduced, one of which (Maurstad, 1935) has been adopted by the International Meteorological Organization. The classification of sea ice deals with forms produced under different conditions of freezing, forms which are due to the tearing apart and packing together of the ice under the action of winds and tides, and forms which result from disintegration and melting of the ice. The classification of icebergs is based on the shape of the bergs, which depends partly upon the type of glacier from which they originate and partly upon the processes of melting and destruction to which they have been subjected.

The great mass of sea ice around the Antarctic Continent is in general described by the somewhat loose term "pack ice," by which is meant drifting fields of close ice, mainly consisting of relatively flat ice floes, the edges of which are broken and hummocked. The pack ice is kept in motion by winds and currents. The effect of the wind is conspicuous when observations are made from a vessel drifting with the ice, because every change in wind direction and velocity brings about a corresponding change in the drift of the ice. Careful observations of the ice drift were conducted by Brennecke (1921) during the drift of the *Deutschland* in the Weddell Sea in 1911–1912. He found that the direction of the ice drift deviated on an average 34° from the wind direction and not 45° as required by the Ekman theory of wind currents (p. 493). The discrepancy was explained by Rossby and Montgomery (1935) as due to a layer of shearing motion in the water directly below the ice; but it may be due in part to the resistance offered by the ice, because the wind does not blow uniformly over large areas, wherefore the ice is packed together in some areas and torn asunder in others. The ice resistance in the Antarctic appears, however, to be small compared to that in the Arctic, because in the Antarctic the drift of the ice is not impeded by land barriers on all sides. The antarctic pack ice consists, therefore, of larger ice floes than the arctic pack ice and is less broken and piled up.

The antarctic pack ice has in all months of the year a well-defined northern boundary beyond which no great amounts of scattered ice are

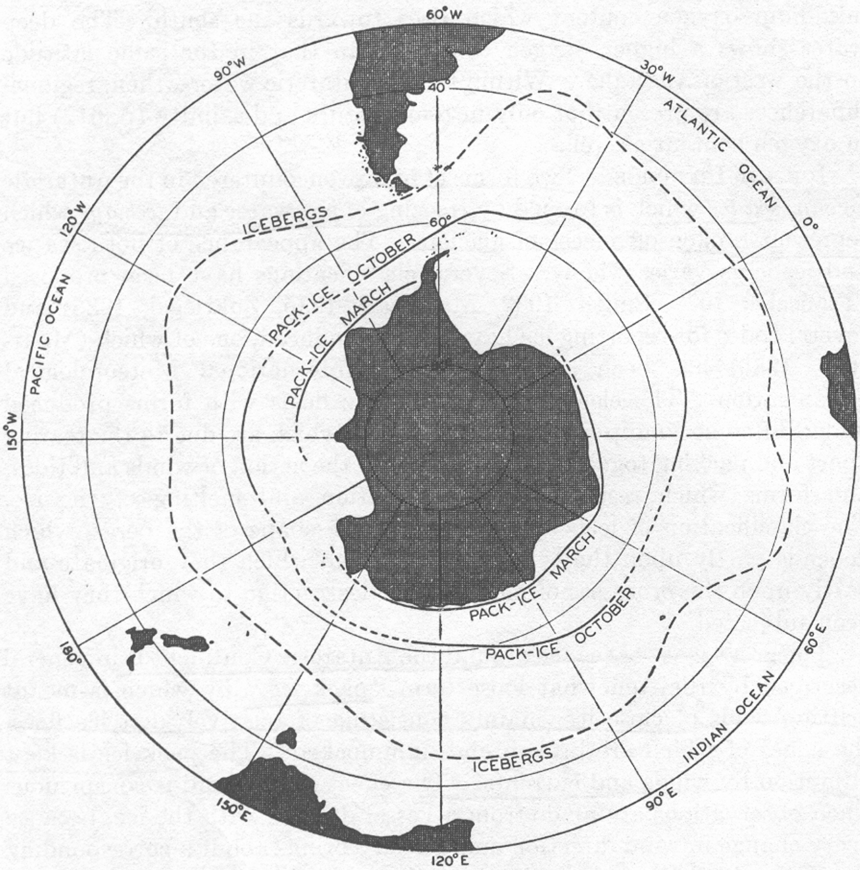

Fig. 167. Average northern limits of pack-ice around the Antarctic Continent in March and October (after Mackintosh and Herdman), and average northern limit of icebergs according to British Admiralty Chart No. 1241.

encountered. The northern limits of the ice at the end of the winter (October) and the end of the summer (March) are shown in fig. 167 (Mackintosh and Herdman, 1940). Parts of the antarctic coast are always ice-free in summer, such as the Pacific side of Graham Land, and other parts are often ice-free, such as the coasts to the south of Australia and Africa. The eastern areas of the Weddell Sea and the

Ross Sea are always ice-free. The Weddell Sea can often be entered from the east without encountering pack ice, but in order to enter the Ross Sea it is always necessary to pass through a broad belt of pack ice.

The great icebergs of the Antarctic originate mainly from the *shelf ice*, which represents the direct continuation of the antarctic ice cap where it extends into the shallow waters surrounding the continent. The shelf ice is partly afloat, and when it is pushed far out it breaks off in enormous pieces which may be tens of kilometers wide and up to 100 km long, and may rise 90 m out of the water, which corresponds to a thickness of about 800 m. These giant bergs, which have occasionally been mistaken for islands, drift through the pack ice and often move in an opposite direction to the pack because they are carried by currents and are less influenced by wind. The melting of these bergs takes a long time, for which reason they drift to greater distances from the continent than the sea ice. Figure 167 also shows the average northern boundary of icebergs according to British Admiralty chart No. 1241, but icebergs or remains of icebergs have been reported much further north than is indicated by this average limit. On April 30, 1894, a small piece of floating ice was sighted in lat. 26°30′S, long. 25°40′W, but otherwise remains of icebergs are rarely reported from localities north of 35°S in the Atlantic Ocean, north of 45°S in the Indian Ocean, and north of 50°S in the Pacific Ocean.

The South Atlantic Ocean

The three major oceans, the Atlantic, the Indian, and the Pacific, can be considered as deep bays that are in open communication with the Antarctic Ocean to the south but are closed at their northern ends. Of these oceans the Atlantic extends farthest to the north, and several large adjacent seas, the Mediterranean, the Caribbean, the Gulf of Mexico, Baffin Bay, the Norwegian Sea, and the North Polar Sea, exercise characteristic effects on the waters of the North Atlantic. The communication between the North Atlantic Ocean and the Antarctic takes place through the South Atlantic Ocean, which is of nearly constant width and without adjacent seas. This wide and short channel is divided lengthwise by the South Atlantic Ridge, which rises from depths exceeding 5000 m to within 2000 m from the surface. The two deep troughs on either side of the ridge are divided into basins by ridges running more or less in an east-west direction, the most conspicuous of which is the Walfish Ridge which closes the eastern trough off to the south.

THE WATER MASSES OF THE SOUTH ATLANTIC OCEAN. In order to illustrate the character of the water masses in the South Atlantic Ocean, seven *Meteor* stations have been selected, the locations of which are shown in the inset map in fig. 168. The stations are equally distributed over the entire South Atlantic Ocean between latitudes 41°S and the Equator.

In spite of the great distance between these stations, the character of the water is nearly the same, as is evident from the T-S values in fig. 168, according to which a fairly well defined water mass present in the upper layers is characterized by a nearly linear T-S relationship between the points $T = 6°$, $S = 34.5\ ^0/_{00}$ and $T = 18°$, $S = 36.0\ ^0/_{00}$. The vertical temperature-salinity relationship at these stations is similar to the horizontal temperature-salinity relationship that is found in the region of the Subtropical Convergence between latitudes 30° and 40°S, and

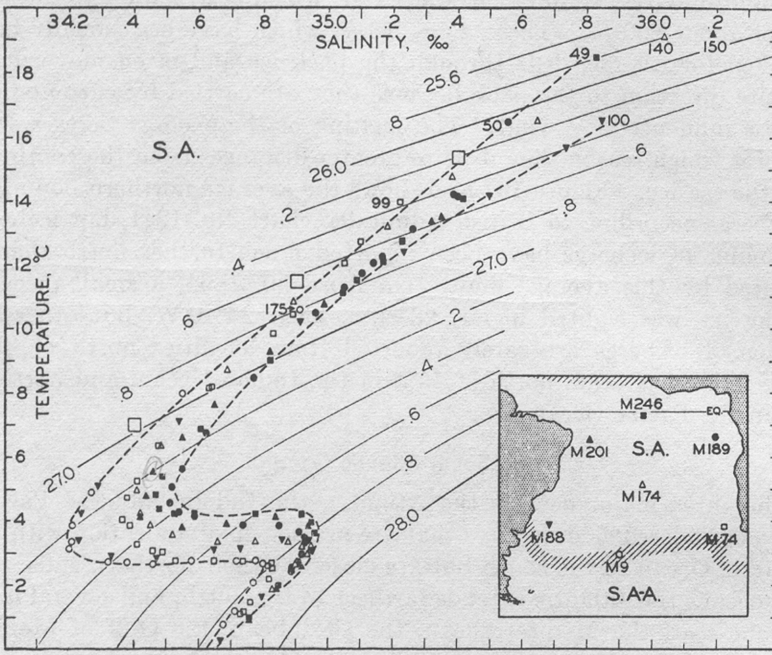

Fig. 168. Temperature-salinity relation in the South Atlantic Ocean. Depths of shallowest values are shown. Winter surface values in latitudes 30°S to 40°S indicated by large squares. Inset map shows location of *Meteor* stations used. S.A. indicates South Atlantic region; S.A-A., subantarctic region.

shown by large squares, for which reason it is probable that this entire water mass has been formed by sinking within the Subtropical Convergence area and subsequent spreading along the proper σ_t surface (p. 145). This water mass will be called the South Atlantic Central Water.

Below the Central Water the Antarctic Intermediate Water shows up by the characteristic salinity minimum which is found over the greater part of the ocean at a depth of approximately 800 m. The Antarctic Intermediate Water is most typically present at the southern stations and at the stations close to South America, and appears to be considerably mixed with other water masses at the northern stations off the African coast. Below the intermediate water, deep and bottom water is present, the character of which will be discussed when dealing with the deep-water circulation of the oceans.

The *T-S* data in fig. 168 give no information as to the vertical extension of the different water masses because indications of depths have been omitted in order to avoid making the presentation too complicated. The vertical extension can be seen from fig. 169, showing the distribution of temperature and salinity in the *Meteor* profile VII (Wüst and Defant, 1936), which nearly follows the parallel of lat. 22°S. The central water, which is approximately bounded by the isohalines 36 ‰ and 34.65 ‰, is found in a layer which is about 450 m thick; the intermediate water, of salinity less than 34.65 ‰, is about 750 m thick, whereas the deep and

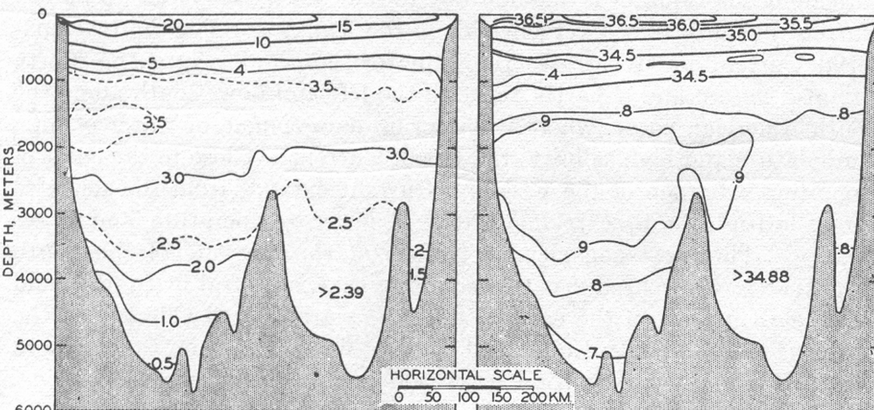

Fig. 169. Distribution of temperature and salinity in vertical sections across the South Atlantic Ocean in about latitude 22°S (after Wüst.) Sections are seen from the south.

bottom water is of greater thickness than the two other water masses together. The figure also demonstrates the presence of a thin layer of surface water which in this particular profile is of high temperature and high salinity. The upper water and the surface layer, as shown in the figure, together comprise the troposphere according to Defant's definition (p. 141). In these layers the strongest currents are encountered.

THE CURRENTS OF THE SOUTH ATLANTIC OCEAN. The most outstanding current of the South Atlantic Ocean is the Benguela Current, which flows north along the west coast of South Africa and is particularly conspicuous between the south point of Africa and lat. 17° to 18°S. In agreement with the dynamics of currents in the Southern Hemisphere, the denser water of low temperature is found on the right-hand side of the current, that is, close to the African coast. Under the influence of the prevailing southerly and southeasterly winds the surface layers are carried away from the coast, and upwelling of water from moderate depths takes place in most seasons of the year (Defant, 1936a). As a consequence of this upwelling, a band of water of low temperature and relatively low salinity is found along the coast extending to a distance of approximately 200 km. This upwelling is of great importance to the biological conditions, because the upwelled water brings plant nutrients into the euphotic zone and thus makes possible the development of large populations of plants and of higher organisms (see fig. 216, p. 786).

Proceeding towards the Equator the Benguela Current gradually leaves the coast and continues as the northern portion of the South Equatorial Current, which flows towards the west across the Atlantic Ocean between lat. 0° and 20°S. In the charts showing the distribution of surface temperatures (charts II and III), the tongue of low temperature along the Equator is due not only to the cold waters of the Benguela Current but also to a divergence along the Equator. The South Equatorial Current partly crosses the Equator, continuing into the North Atlantic Ocean, and it partly turns to the left and flows south along the South American coast, where it shows up as a tongue of water of high temperature and high salinity, the Brazil Current. Close to the coast of Argentina a branch of the Falkland Current extends from the south to about latitude 30°S, carrying water of lower temperature and lower salinity. The most conspicuous feature of the currents of the South Atlantic is represented by the counterclockwise gyral with the cold Benguela Current on the eastern side, the warm Brazil Current on the western side, the South Equatorial Current flowing west on the northern side, and the South Atlantic Current flowing east on the southern side. It is a system of shallow currents because the entire circulation takes place above the Antarctic Intermediate Water or within the troposphere, and near the Equator it is probably limited to a depth of less than 200 m.

In agreement with this circulation, the water within the southern and eastern parts of the current system is of somewhat lower salinity than that of the northern and western parts (see fig. 168, p. 626). In the south, admixture of subantarctic water lowers the salinity, but in the north admixture of saline surface water increases the salinity.

A quantitative study of the currents of the South Atlantic Ocean on the basis of the *Meteor* data (Defant, 1941) leads to results similar to those that have been described here. Defant computed the absolute geopotential topographies of a series of isobaric surfaces by the third method described on p. 457. His conclusions are essentially in agreement with those which can be derived by making use of the equation of continuity (the fourth method, p. 457).

For calculations of the transport through vertical sections between South Africa and South America (p. 465), the *Meteor* data lead to the result that the transport above a depth of 4000 m is directed towards the north, but this is obviously impossible because the *net* transport through such a section must be nearly zero. It is therefore necessary to assume a layer of no motion at some intermediate depth, above which the transport will be directed towards the north and below which it will be directed towards the south. Examination of the *Meteor* profiles to the south of lat. 20°S shows that the layer of no motion is found at the boundary between the Antarctic Intermediate Water and the deep water. The depth of the layer in lat. 20°S is about 1100 m and increases somewhat toward the south. The Antarctic Intermediate Water and the upper water move, then, on the whole, to the north, whereas the deep water moves south, but close to the bottom there is a flow of Antarctic Bottom Water towards the north. The net transport to the south below 1400 m amounts approximately to 15 million m^3 per second.

TABLE 76

TRANSPORT OF WATER ACROSS LAT. 30°S AND ACROSS THE EQUATOR

Latitude	Water mass	Current	Transport in million m^3/sec toward	
			North	South
30°S	Upper water	Benguela Current	16	
		Brazil Current		10
		Central part of South Atlantic gyral	7	7
	Intermediate water		9	
	Deep water			18
	Bottom water		3	
0°	Upper water		6	
	Intermediate water		2	
	Deep water			9
	Bottom water		1	

The transport through vertical sections in the South Atlantic can be studied in greater detail, and approximate values can be computed for the amounts of the different types of water which are carried to the north and to the south, taking the continuity of the system into account. In table 76 are shown results of such computations, as referred to a section in 30°S. The table also contains values for the transport across the Equator, but these have been obtained in a different manner. The value of a northward transport of about 6 million m^3 per second of upper water is mainly based on a comparison between the waters of the Caribbean Sea and the Sargasso Sea, and the values of the northward transports of intermediate and bottom water are estimated.

The approximate correctness of the figures can be checked by considering that the net transport of salt must be zero. A calculation of the net transport of salt requires knowledge of the velocity distribution within the different parts of the current system, but a rough computation which is based on average values confirms the above conclusions. We shall have to return to some of these matters when discussing the deep-water circulation, but here it will be emphasized that a large quantity of South Atlantic Central and Intermediate Water crosses the Equator and enters the North Atlantic Ocean, where it exercises an influence on the character of the waters along the coast of South America and in the Caribbean Sea and in the Gulf of Mexico.

In his discussion of the currents of the South Atlantic, Defant (1941) shows that in lat. 30°S to 45°S the stream lines of the currents are wavelike, the apparent wave length being 850 to 880 km. He points out that according to a theory of Rossby, extended by Haurwitz (1940), stationary wave patterns in a primary current are possible, provided that a relation exists between the velocity and width of the primary current and the wave length of the disturbance. Using this relation and introducing measured values of the width of the current across the South Atlantic and the wave length of the disturbances, Defant obtains a velocity of about 27 cm/sec for the primary current, in good agreement with values derived from the geopotential topography of the sea surface.

The wave pattern must be quasi-stationary because it is derived from observations in different years. Defant suggests that it is related to the topography of the bottom, but if this is true the deflections caused by the bottom topography must be in the opposite direction from those which were described on p. 466 (see fig. 116, p. 467 and fig. 117, p. 468). It is perhaps more likely, as also suggested by Defant, that the wavelike disturbance originates where the Brazil Current and the Falkland Current meet.

Oxygen Distribution. The major features of the oxygen distribution are shown in the longitudinal sections in fig. 43, p. 210, and fig. 210, p. 748. According to these sections the oxygen content of the central water is low near the Equator where, off west Africa, minimum values below 0.5 ml/L are found at a depth of about 350 m. This minimum is less pronounced off the coast of South America, where the minimum values are only slightly below 3 ml/L. The great depletion of oxygen along the African west coast is probably due to the oxidation of remains of organisms which fall towards the bottom from the productive region of the Benguela Current.

On the American side (fig. 210, p. 748), the Antarctic Intermediate Water is characterized by high oxygen content and this water mass can therefore be traced by means of a tongue of maximum oxygen values which extends nearly to the Equator. In the transition layer, below the intermediate layer within which the direction of flow reverses, the oxygen content shows a minimum of less than 4.5 ml/L but in the deep water values above 5 ml/L occur, decreasing to the south.

The Equatorial Region of the Atlantic Ocean

THE WATER MASSES OF THE EQUATORIAL REGION OF THE ATLANTIC OCEAN. Between depths of 100–200 m and 600–700 m, the water masses of the equatorial region are mainly of South Atlantic origin, but in the northern parts there is evidence of admixture of large quantities of North

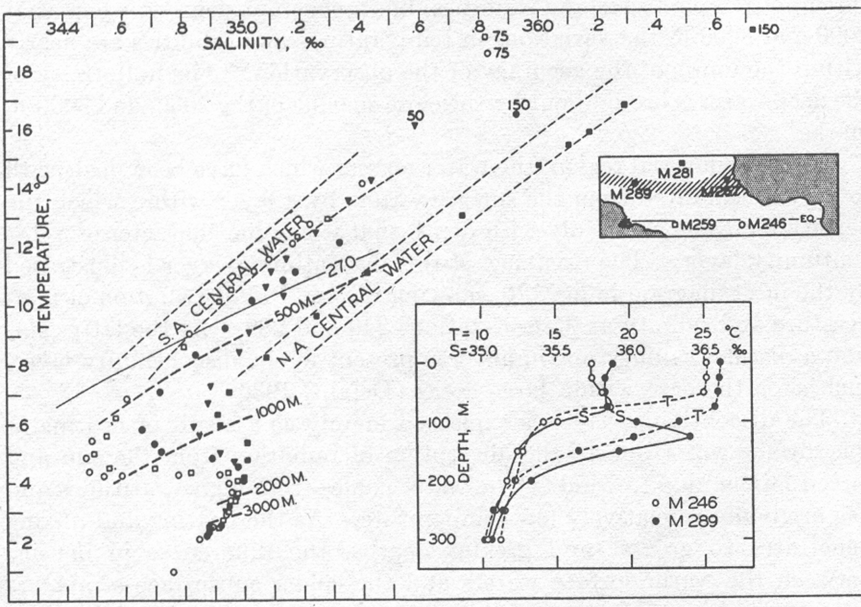

Fig. 170. Temperature-salinity relation in the equatorial region of the Atlantic Ocean. Inset map shows location of *Meteor* stations used, and inset diagram shows vertical distribution of temperature and salinity at *Meteor* station 246 in 2°40′S, 16°37′W, and at *Meteor* station 289 in 11°02′N, 49°33′W.

Atlantic Central Water which, as will be seen later, at temperatures between 8° and 15°C has a salinity which is nearly 0.5 ‰ higher than that of the South Atlantic Central Water. Figure 170 contains *T-S* values at selected equatorial stations, the locations of which are given on the inset map, and also shows the characteristic *T-S* relationships of the central water masses. At the two stations close to the Equator, stations M246 and M259, nearly typical South Atlantic water occurred, but at station M289 in 11°N and 50°W the water contained a considerable admixture of North Atlantic water and the same was true of station M267 in lat. 14°N off the African coast, where North Atlantic water appeared conspicuously at temperatures below 10.5°. At the most northern station, M281 in lat. 19°N, nearly pure North Atlantic water was encountered.

In sum, there exists in the Atlantic Ocean no equatorial water mass, but South Atlantic Central Water flows across the Equator and becomes mixed with the North Atlantic Central Water to the north of lat. 10°N.

In the figure are entered lines joining the *T-S* points at the different stations, which represent conditions at depths of 500, 1000, 2000, and 3000 m, and the line $\sigma_t = 27.0$ is shown. At 500 m there is a wide range in the temperature and salinity of the water but the density is practically constant, as is evident from the fact that the 500-m line in the diagram nearly coincides with a σ_t curve. At 1000 m there is still a considerable spread of temperatures and salinities, but a constant density, whereas at 2000 and 3000 m the variations in temperatures and salinities are nearly within the limits of the accuracy of the observation. The uniformity of the deep water is brought out by the close spacing of the 2000- and 3000-m marks.

In the equatorial region, the water masses which have been dealt with so far are separated from the surface waters by a layer within which the density increases so rapidly with depth that it has the character of a discontinuity layer. The existence of this discontinuity layer is illustrated by the inset diagram in fig. 170, showing the vertical distribution of temperature and salinity at *Meteor* stations 246 and 289. At the latter station a distinct salinity maximum was present in the discontinuity layer, such as is the case within large areas (Defant, 1936b).

The discontinuity layer is explained mainly as a result of heating of the surface water due to the absorption of radiation from the sun and forced mixing due to wind. The water comes from higher latitudes and has originally a relatively low temperature. As the heating and mixing penetrates to greater and greater depths, the differences in density between the warm surface waters and the colder subsurface strata are concentrated in a thinner and thinner layer within which the stratification becomes more and more stable. When the stratification has become so stable that the layer takes the character of a discontinuity surface, further mixing is inhibited because the eddy conductivity is greatly reduced by the exceedingly great stability (p. 476) and the amounts of heat which are conducted downwards become so small that they are carried away by weak currents. Additional absorption of solar energy is used mainly for evaporation, and the final temperature which is attained will depend upon the rate of evaporation and is therefore determined by the interaction between the atmosphere and the ocean.

THE CURRENTS OF THE EQUATORIAL REGION OF THE ATLANTIC OCEAN. In a given locality the thickness of the upper warm layer depends not only upon the intensity of mixing but also upon the character of the circulation in the top layers, because in the presence of currents the discontinuity surface cannot remain horizontal but must slope (p. 444). On the basis of the *Meteor* data and all other observations available from the tropical parts of the Atlantic Ocean, Defant (1936b) has been able to construct a chart showing the topography of the discontinuity surface between latitudes 25°N and 25°S. Figure 171A is based on this chart. In the presence of a sloping discontinuity surface the flow of the water above that surface, relative to the underlying water masses, must, in the Northern Hemisphere, be in such direction that the surface rises to the left of an observer looking in the direction of flow, and in the Southern Hemisphere such that the surface rises to the right. On the basis of these rules, arrows have been entered showing the direction of flow which, in general, agrees well with the observed average surface currents in the tropical regions of the Atlantic. This picture shows the gyral in the South Atlantic Ocean which has already been discussed and demonstrates the complicated character of the currents near the Equator where a

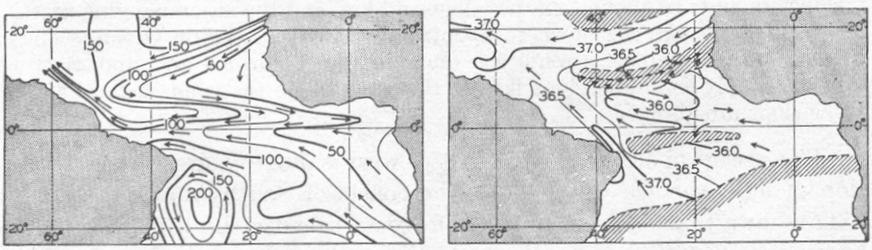

Fig. 171. (A) Topography of the discontinuity surface in the Equatorial region of the Atlantic, shown by depth contours in meters, and corresponding currents. (B) Salinity in the layer of maximum salinity below the discontinuity layer and assumed currents. Regions without salinity maximum shaded. (Both representations after Defant.)

countercurrent towards the east, the Equatorial Countercurrent, is imbedded between the equatorial currents of the two hemispheres.

Evidently the countercurrent is related to the distribution of mass as demonstrated by the slope of the discontinuity surface. In a profile from south to north the discontinuity surface rises towards the Equator, reaches a maximum elevation at the Equator, and drops towards a minimum in 2° to 3°N latitude, at the northern boundary of the South Equatorial Current. Between 2° to 3°N and 6° to 8°N the discontinuity surface rises towards the north, reaching a second and more pronounced maximum in the latter latitude. This is the region of the countercurrent, and to the north of the second maximum, where the surface slopes downwards again, the North Equatorial Current is found. This distribution of mass was first recognized by means of the *Carnegie* observations in the Pacific (fig. 198, p. 710). On the basis of these, Sverdrup (1932) suggested that the countercurrent appeared because the trade winds maintaining the equatorial currents are not symmetrical to the Equator, the calm belt between them lying in the Northern Hemisphere, whereas the effect of the rotation of the earth is symmetrical in respect to the Equator.

He thought that in these circumstances a stable distribution of mass could exist only in the presence of a countercurrent, and this idea was followed up by Defant (1936b), who demonstrated that in the equatorial part of the Atlantic Ocean the distribution of mass was similar to that shown by the *Carnegie* observations in the Pacific. Later on Sverdrup (1939b) pointed out that these data only show that the countercurrent is associated with a typical distribution of mass but do not account for the dynamics of the current.

The dynamics of the countercurrent have recently been discussed by Montgomery (1940) and by Montgomery and Palmén (1940). Montgomery writes:

> The trade winds, by continually exerting a westward stress on the sea surface, produce a westward ascent of sea level along the Equator. This slope amounts to about 4 cm per thousand km, and the accompanying pressure gradient extends down to about 150 m in the Atlantic Ocean. The Equatorial Counter Currents are found in the doldrums and apparently result simply as a downslope flow in this zone where the winds maintaining the slope are absent.

The ascent of sea level from east to west is due to a greater accumulation of light surface water along the coasts of South America. Figure 171A shows that such accumulation exists because the thickness of the homogeneous surface layer above the discontinuity surface increases from less than 40 m in the east to about 140 m in the west. Above a depth of 150 m the isosteric surfaces, therefore, slope downwards from east to west, but below 150 m they are practically horizontal. Accordingly, Montgomery and Palmén find that at the Equator the geopotential height of the free surface referred to the 1000-decibar surface is 14 dyn cm greater in the west than in the east, but the 150-decibar surface is parallel to the 1000-decibar surface. The countercurrent is therefore a very shallow current which is confined to the surface layer above the discontinuity.

A swift current which is embedded between water masses moving in the opposite direction must be subject to a considerable retardation owing to friction. A certain amount of energy is therefore needed for maintaining such a current, and this energy is, according to Montgomery and Palmén, derived from the trade winds which maintain the slope of the sea surface. Montgomery and Palmén assume that the retarding friction may be due to lateral mixing with the westward-flowing equatorial currents, and estimate the coefficient of lateral eddy viscosity to be 7×10^7 g/cm/sec. This concept leads to the important conclusion that on both sides of the countercurrent the equatorial currents are subjected to great lateral frictional stresses which are directed opposite to the horizontal stresses exerted by the trade winds on the surface. Similar lateral stresses must be directed opposite to the general flow along the continental boundaries of the two large counterclockwise gyrals of the two hemispheres and the torque exerted by these stresses perhaps balances the torque exerted by the stress of the wind on the sea surface (see p. 489). The lateral stresses between the equatorial currents and the countercurrent probably contribute very materially to the total torque and the countercurrent represents, therefore, a dynamically important link in the entire system of ocean currents.

Within the countercurrent and the adjacent equatorial currents, the frictional forces must lead to a transport of water across the isolines, that is, to a transverse circulation. This was examined theoretically by Defant (1936b), who considered friction due to vertical mixing, but the conclusions are probably applicable if the friction is due to lateral mixing. Defant's results are shown schematically in fig. 172, according to which

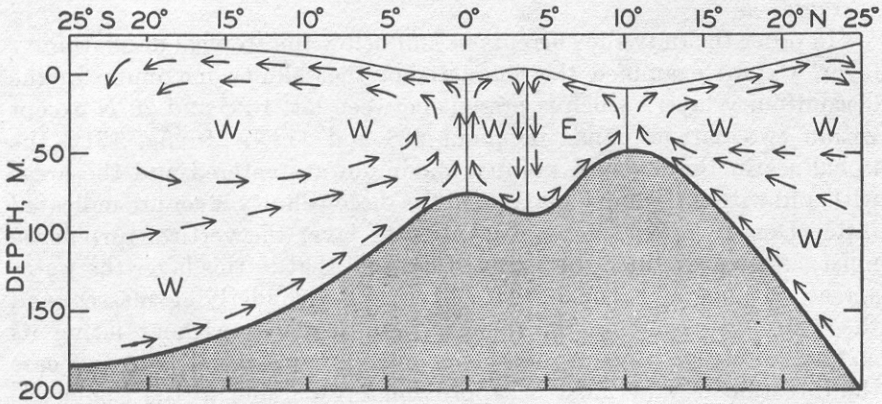

Fig. 172. Schematic representation of the vertical circulation within the equatorial region of the Atlantic. The direction of the currents is indicated by the letters *W* and *E*. The water below the discontinuity surface, which is supposed to be at rest, is shaded.

four "cells" are present, representing gyrals with horizontal axes, neighboring gyrals rotating in opposite directions. Within the southern cell the water sinks in the region of the Tropical Convergence and rises at the Equator, and within the next cell, located between the Equator and the southern boundary of the countercurrent, the water rises at the Equator and sinks at the boundary of the countercurrent. Within the countercurrent the water rises at the northern and sinks at the southern boundary, and within the northern cell sinking motion takes place at the Tropical Convergence and rising motion at the northern boundary of the countercurrent. The data from the Atlantic Ocean indicate the existence of these cells and the Carnegie section across the Pacific countercurrent (fig. 198, p. 710) demonstrates their presence convincingly. As a consequence of these transverse circulations, the northern boundary of the countercurrent and the Equator represent lines of *divergence*, whereas the southern boundary of the countercurrent is a line of *convergence* and individual water masses carried by the different currents follow complicated spiral-like trajectories.

In summer, when the countercurrent is best developed, the effect of the divergence at the Equator appears on the charts of surface temperatures (Böhnecke, 1936) as a tongue of low temperature, but no effect of the divergence at the northern boundary of the countercurrent is visible. The divergences also appear to be of biological importance because the ascending motion maintains a replenishment of nutrients to the surface layers, thus favoring the development of phytoplankton. The tongues of water of high phosphate content and of abundant phytoplankton which extend towards the west from the coast of Africa (see figs. 216 and 217, p. 786) are probably related to the structure of the currents.

In order to study the currents in and below the tropical discontinuity layer, Defant examined the character of the salinity maximum in the discontinuity layer, which is present between lat. 10°S and 20°N except within two narrow bands in about 3°S and 11°N. In fig. 171B the isohalines in the layer of salinity maximum are entered and the areas with and without salinity maxima in the discontinuity layer are indicated (after Defant). Within the discontinuity layer the vertical turbulence must, as already stated, be so much reduced that in this layer the water spread without being mixed with overlying or underlying masses, and the water can therefore move over long distances without losing its characteristics provided no horizontal mixing takes place. In this case the direction of flow must be approximately parallel to the isohalines, and from the chart showing these isohalines it is possible on these assumptions to derive the direction of flow *in* the discontinuity layer. The arrows in fig. 171B demonstrate the direction of these currents. It should be observed particularly that the belts without salinity maxima appear here as regions of convergence, whereas at the surface they are regions of divergence. This relation is in agreement with the conclusion as to the character of the vertical circulation (fig. 172).

Montgomery (1939) has approached the problem of the subsurface currents near the Equator in an entirely different manner, assuming that the subsurface flow takes place along σ_t surfaces and that besides the vertical mixing, lateral mixing along σ_t surfaces determines the distribution of salinity and temperature. He points out that the layer of maximum salinity approximately coincides with the surface $\sigma_t = 25.5$, and has prepared a chart showing that in this surface the isohalines have in general the same form as those in fig. 171B. He furthermore assumes that the principal water motion follows the *tongues* of low or high salinity, whereas Defant assumes that lines of flow are roughly parallel to the isohalines. The pictures of the current as derived by Defant and

Montgomery show considerable similarity, but the different currents appearing in the pictures are displaced relative to each other.

Montgomery points out that the observed distribution of salinity can remain stationary in the presence of the assumed currents if either vertical or lateral mixing dominates. He has computed the possible maximum values of the coefficients and obtained for the vertical one a value which was much greater than is consistent with the great stability of the stratification. If the vertical diffusion is small, the lateral diffusion must determine the character of the salinity distribution. The lateral diffusion was found to be about 4×10^7 cm^2/sec, in agreement with the value of the lateral eddy viscosity which was derived by an entirely different method (p. 485). At the present time it is not possible to decide which of the two interpretations is the correct one, but the dynamic consequences of Montgomery's concept appear to fit better into a large-scale picture of the ocean currents.

The chart of the surface currents (chart VII) shows that on an average a transport takes place from the Southern to the Northern Hemisphere in agreement with the conclusions which were based on the *Meteor* sections between lat. 20°S and 33°S, according to which an average transport to the north of about 8 million m^3/sec should take place above a depth of approximately 800 m. It appears that the greater part of this transport, about 6 million m^3/sec, takes place within the warm surface layers, for which reason it is probable that this process contributes towards displacing the thermal Equator towards the north. The fact that the surface temperature of the Atlantic Ocean is, in the Southern Hemisphere, lower, but in the Northern Hemisphere higher than that of the Pacific Ocean (see table 31, p. 127), is perhaps related to this transport.

The Adjacent Seas of the North Atlantic Ocean

The water masses and currents of the North Atlantic Ocean can be dealt with more precisely when the waters of the adjacent seas have been discussed, and we shall therefore turn first to the adjacent seas—the Caribbean Sea and the Gulf of Mexico, the Mediterranean and the Black Seas, the Norwegian Sea, the Baltic and the Polar Seas, and the Labrador Sea and Baffin Bay.

THE AMERICAN MEDITERRANEAN SEA. The Caribbean Sea and the Gulf of Mexico can be considered together under the name of the American Mediterranean Sea, a name which has been employed particularly by German oceanographers. The Caribbean Sea has the shape of a wide and somewhat irregular channel, the eastern end of which opens towards the tropical part of the North Atlantic (see fig. 6, p. 35). The opening is wide, but is partly closed by the submarine ridge on which the Lesser Antilles lie. In most localities the ridge rises to less than 1000 m below sea level and communication with the open ocean exists, therefore, down to this depth only. The northern boundary of the channel, which is represented by the islands of Cuba, Haiti, and Puerto Rico, is broken by several passages, some of which have greater sill depths than the sill depth of the Lesser Antilles ridge. The important passages are the Anegada and Jungfern Passages to the east of Puerto Rico, and the Windward Passage between Cuba and Hispaniola. The Jamaica Rise between Hispaniola and Honduras divides the Caribbean Sea crosswise in two regions, the eastern and the western Caribbean Seas, of which the latter is also called the Cayman Sea (Parr, 1937). To the northwest the Yucatan Channel opens into the Gulf of Mexico, which has a simpler bottom topography and which, towards the east, is in communication with the Atlantic Ocean through the Straits of Florida. The sill depths in the Yucatan Channel and in the Straits of Florida are approximately 1600 m and 800 m, respectively. The waters of the North Atlantic can therefore pass freely through the American Mediterranean Sea at depths less than about 800 m, but at greater depths communication with the large water bodies of the Atlantic Ocean is more or less restricted.

The water masses of the upper layers enter the Caribbean Sea from the east and show characteristics similar to those of the adjacent North Atlantic waters. Below a thin homogeneous top layer a nearly discontinuous decrease of temperature is found and in most localities a salinity maximum is present within the discontinuity layer. The water masses that enter between the Lesser Antilles show a streaky distribution of salinity, narrow bands of high salinity alternating with bands of low salinity, but these differences are rapidly smoothed out in the Caribbean Sea owing to such intense mixing that the waters flowing through the narrow passages of the Yucatan Channel and the Straits of Florida are of more uniform character.

The upper waters that flow into the Caribbean Sea are mainly of North Atlantic origin but contain a considerable admixture of South Atlantic water. The relative amounts of the two types of water can be approximately determined by a comparison between the temperatures and salinities of the South Atlantic water that crosses the Equator, the waters passing through the Yucatan Channel, and the water from the western Sargasso Sea, the *T*-*S* relations of which have been examined by Iselin (1936). Table 77 contains corresponding temperatures and salinities of these three water masses at different σ_t values. The last column in the table gives the ratio between the amounts of South Atlantic and North Atlantic water that pass through the Yucatan Channel. It is seen that at higher temperatures the current through the Channel carries about one part of South Atlantic and three and one half parts of North Atlantic water. The total transport is about 26 million m^3/sec, of which, accordingly, about 6 million m^3/sec represent South Atlantic water.

Below the upper water a considerable amount of Antarctic Intermediate Water enters the Caribbean Sea, such that the water at the intermediate salinity minimum is mainly of South Atlantic origin, as pointed out by Nielsen (1925) and Parr (1937). The intermediate salinity minimum in the Caribbean Sea decreases in intensity in the direction of flow. In the eastern part of the Caribbean Sea, in about longitude 68°W, the average minimum salinity is 34.73 $^0/_{00}$, whereas in the Yucatan Channel the average minimum salinity is 34.88 $^0/_{00}$. This

TABLE 77

CORRESPONDING TEMPERATURES AND SALINITIES AT STATED VALUES OF σ_t IN SOUTH ATLANTIC WATER AND IN THE WATERS OF THE WESTERN SARGASSO SEA AND YUCATAN CHANNEL

Value of σ_t	*A* South Atlantic		*B* West Sargasso Sea		Yucatan Channel		Ratio *A/B* in Yucatan Channel
	Temp. (°C)	S ($^0/_{00}$)	Temp. (°C)	S ($^0/_{00}$)	Temp. (°C)	S ($^0/_{00}$)	
26.4	15.8	35.64	18.3	36.55	17.8	36.38	1/4
26.6	13.6	35.39	16.8	36.32	16.1	36.11	1/3.5
26.8	11.3	35.08	14.8	35.98	13.9	35.74	1/3
27.0	9.0	34.82	12.4	35.61	11.6	35.40	1/3
27.2	6.0	34.52	10.0	35.29	8.7	35.03	1/2

increase in salinity is, according to Seiwell (1938), mainly due to vertical mixing within the moving water mass.

The deeper portion of the American Mediterranean Sea is divided into a series of basins, the most important being the Venezuela and Colombia basins in the eastern Caribbean Sea, the Cayman Trough and the Yucatan Basin in the western Caribbean Sea, and the Mexico Basin in the Gulf of Mexico. The character of the deep water in these basins depends upon their communications with the adjacent parts of the North Atlantic Ocean and upon the intercommunications between the basins. The greatest depth of the sills separating the basins of the eastern Caribbean from the North Atlantic are found in the Anegada and Jungfern Passages, between which lies the small St. Croix Basin. Figure 173 shows the distribution of potential temperature in a section passing from the Atlantic through the Anegada Passage, the St. Croix Basin, and the Jungfern Passage into the Venezuela Basin (Dietrich, 1939). It appears from this distribution that the sill depth of the Anegada Passage is somewhat greater than that of the Jungfern Passage, and that the waters in the St. Croix Basin therefore show a lower potential temperature, about 3.5°, as compared to 3.85° in the Venezuela Basin. The deep water in the Venezuela Basin is of lower density than that of the adjacent Atlantic Ocean, for which reason the deep water of the basin is renewed by inflow across the sill. The basin therefore belongs to the second type (see p. 148). The renewal of the deep water must be a relatively rapid process, because the oxygen content is high and has, according to Seiwell (1938), an average value of approximately 5.1 ml/L at a depth of 2500 m.

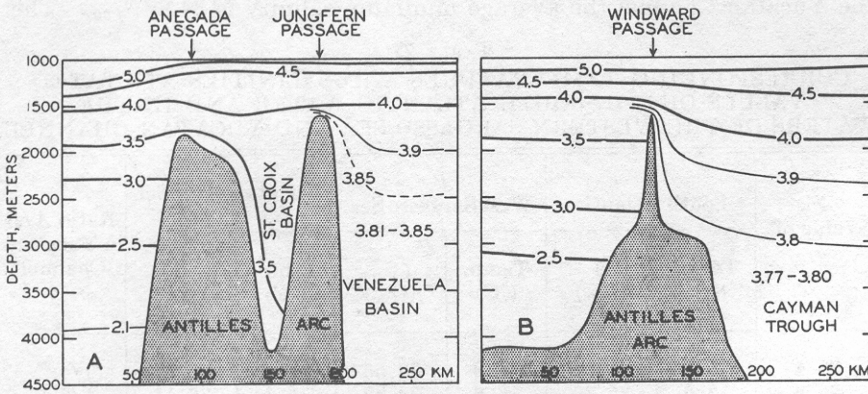

Fig. 173. Sill depths between the Atlantic Ocean and the Caribbean Sea according to temperature observations. (A) potential temperature in a vertical section through the Anegada and Jungfern Passages. (B) Potential temperature in a vertical section through the Windward passage (after Dietrich).

The sill depth of the Jamaica Rise, which separates the western Caribbean Sea from the eastern, is less than the sill depth of the Windward Passage. The deep water of Cayman Trough and the Yucatan Basin is therefore renewed through the Windward Passage which, according to the distribution of potential temperature (fig. 173), must have a sill depth of approximately 1600 m. The renewal of the deep water in these basins appears to be even more rapid because the average oxygen content in the Yucatan Basin and the Cayman Trough is, at 2500 m, between 5.5 and 6.0 ml/L.

The deepest connection between the Mexico Basin and the adjacent seas is found in the Yucatan Channel, and the deep water of the Mexico Basin is therefore renewed from the Yucatan Basin by flow across the sill in the Yucatan Channel. The sill depth is not exactly known but is probably somewhat less than the sill depth of the Windward Passage, because the potential temperature of the water in the Mexico Basin is about 3.95° as compared to 3.85° in the Yucatan Basin. The passages to the north of Cuba are much more shallow and no renewal of deep water can take place through them. Even in the case of the Mexico

Basin, renewal must be fairly rapid because the oxygen content at 2500 m averages about 5.0 ml/L.

The different sill depths have not yet been established by means of soundings, and the depths given above are all based on studies of the hydrographic conditions. These can be interpreted in several manners and different values of the sill depths have therefore been arrived at by different authors, but in the above we have followed Dietrich's presentation (1939), which appears to combine the known facts in the most satisfactory manner. Dietrich has also been able to plot the potential temperatures at depths greater than 2500 m, and these show a remarkable regularity in spite of the fact that the greatest differences observed in the single basins do not exceed 0.05°. The arrangement of the potential temperatures further supports the conclusions as to the localities at which inflow of deep water into the basins takes place.

The salinity of the deep water is very uniform. According to Parr (1937) the average salinity is 34.98 $^0/_{00}$ and the deviations from this value are within the probable errors of the salinity determinations.

It is evident from this summary that the American Mediterranean Sea does not exercise any influence upon the deep-water circulation of the Atlantic Ocean, but it may exercise a considerable influence on the circulation in the upper layers.

The surface currents in spring are shown in fig. 174. A strong current passes through the Caribbean Sea, continues with increased speed through the Yucatan Channel, bends sharply to the right, and flows with great velocity out through the Straits of Florida. On the flanks of the main current numerous eddies are present, of which the one in the wide bay between Nicaragua and Colombia and the one between Cuba and Jamaica are particularly conspicuous. In the Gulf of Mexico several large eddies exist, and all of these appear to be semipermanent features, the locations of which are determined by the contours of the coast and the configuration of the bottom.

The presentation of the currents in fig. 174 is based on ships' observations. When attempting a calculation of currents by means of numerous *Atlantis* data from the Caribbean Sea, Parr (1937) found that the flow is not directed parallel to the contours of the isobaric surfaces, but between the Lesser Antilles and the Yucatan Channel the current flows uphill. Parr suggested that this feature may be due to piling up of water in front of the narrow Yucatan Channel, caused by the stress exerted on the surface by the prevailing easterly winds. This idea was further examined by Sverdrup (1939b), who concluded that the piling up of the surface water can be fully explained as the effect of winds blowing with an average velocity of about 10 m/sec, which agrees well with the observed values in spring. A further consequence of this piling up is that in the Gulf of Mexico a higher sea level is maintained than along the adjacent coast of the United States facing the Atlantic Ocean. At Cedar Keys, on the western coast of Florida, the average sea level is 19 cm higher than the average sea level at St. Augustine, Florida, on the east coast. This indicates that the prevailing winds over the Caribbean Sea produce a hydrostatic head that may, according to Montgomery (1938), account for the major part of the energy of the Florida Current (p. 673).

Regardless of the effect of the stress of the wind or the existence of a hydrostatic head, the mass distribution must adjust itself to the major currents that are present. In general, the lighter water is therefore found on the right-hand side of the current and the sea surface rises to

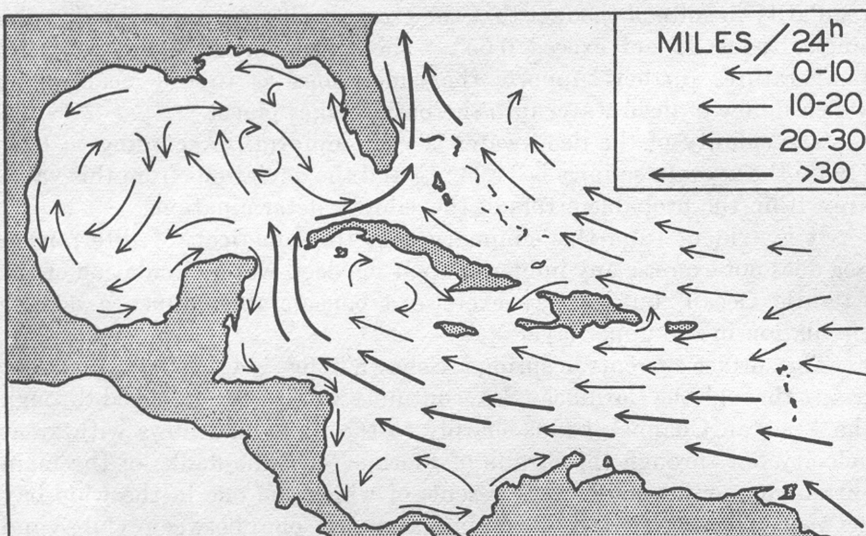

Fig. 174. Surface currents in spring in the American Mediterranean Sea (after Dietrich).

the right when looking in the direction of flow. This rise to the right is particularly conspicuous within a swift and narrow current and, in the case of the Florida Current, it has been computed that at the north coast of Cuba sea level must be about 45 cm higher than at Key West, Florida (Dietrich, 1936, Montgomery, 1938). The Florida Current will be further dealt with when discussing the Gulf Stream System of the North Atlantic, but in this place it should be pointed out that, according to Dietrich (1939), the Florida Current is essentially a direct continuation of the current through the Yucatan Channel and that the waters of the Gulf of Mexico mainly form independent eddies and are only to a small extent drawn into the Straits of Florida.

The European Mediterranean Sea. From the oceanographic point of view the waters of the Mediterranean Sea are basin water masses which are in communication with the Atlantic Ocean only through the narrow and shallow Strait of Gibraltar, the exchange of water through the Suez Canal being negligible (p. 688). The Mediterranean body of water can again be divided into several smaller ones, each of which has its specific characteristics, the most outstanding being the waters of the Black Sea, which are in restricted communication with the Mediterranean Sea proper through straits at the Dardanelles and the Bosporus. Geographically the Black Sea is not considered a part of the Mediterranean Sea, but oceanographically it should be regarded as such.

The Mediterranean proper is divided into a series of deep basins more or less isolated from each other. The most outstanding of these are, in the western Mediterranean, the Algiers-Provençal Basin to the west of Sardinia and Corsica and the Tyrrhenian Basin on the west side of Italy, and, in the eastern Mediterranean, the Ionian Basin to the south of Italy and Greece and the Levantine Basin to the south of Asia Minor (fig. 5, p. 34). The two former basins are separated from the two latter by shallow ridges between Tunisia and Sicily, whereas the two basins in the western Mediterranean Sea and the two basins in the eastern Mediterranean Sea are separated by deep sills.

In the Mediterranean Sea proper (not including the Black Sea), evaporation greatly exceeds precipitation and runoff, and in winter deep water of very high salinity is formed in different places by vertical convection currents. The Mediterranean proper belongs, therefore, to the first type of basin (p. 147), in which deep water of great density is formed with outflow over the sill and inflow at the surface.

The water masses of the Mediterranean are characteristically different from those of the adjacent North Atlantic, because, owing to the isolation, the deep water has a higher temperature. The Atlantic water which flows into the Mediterranean Sea is so rapidly mixed with Mediterranean surface water that it soon loses its Atlantic character. The surface salinity is higher than 37.00 ‰, except where the surface current flows to the east along the north coast of Africa as far east as Tunisia, and in the inner portion of the Adriatic Sea and the Aegean Sea, where there is a considerable addition of river water or surface water from the Black Sea. It is above 39 ‰ to the south of Asia Minor. The surface salinity is subjected to seasonal variations, but these are not known in detail. The surface temperature generally increases from the Strait of Gibraltar to the inner portions, except in winter, when the lowest surface temperatures are found in the most northerly portions, namely off the French and Italian Rivieras and in the inner portions of the Adriatic and Aegean Seas. The seasonal variation in temperature is great, the annual range everywhere exceeding 9° and reaching 13° to 14° off the Riviera and in the northern part of the Adriatic Sea.

Below the surface four different water masses are encountered, including a thick transition layer which separates the intermediate water from the deep water. The character of these water masses is illustrated in fig. 175, showing the vertical distribution of temperature and salinity at two *Dana* stations, station 4119 in the Tyrrhenian Sea and station 4070 in the Ionian Sea.

The surface water generally extends to a depth of 100 to 200 m. In the western Mediterranean the lower limit of the surface water is indicated by a temperature minimum, as seen at station 4119, whereas in the eastern Mediterranean the temperature minimum is generally absent and a layer of slow temperature decrease is found instead, as seen at station 4070. The depth of the temperature minimum or the depth of the layer of slow decrease represents the depth to which vertical convection currents ordinarily reach in winter, and the temperature and salinity at the depth of the minimum or in the layer of small decrease correspond to the surface values in winter.

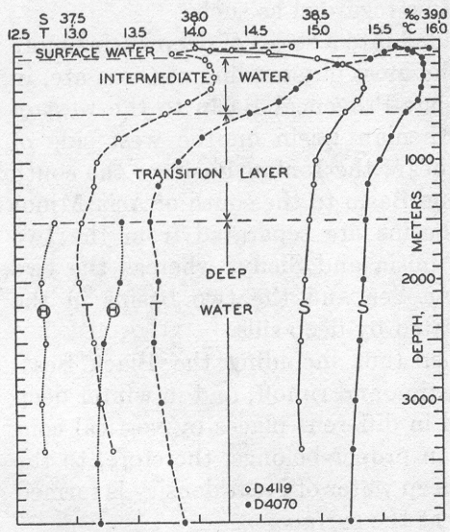

Fig. 175. Character of the water masses in the Mediterranean Sea illustrated by curves showing the vertical distribution of temperature (T), salinity (S), and potential temperature (Θ) at the *Dana* stations 4119 in 40°13′N, 12°06′E, and 4070 in 35°40.5′N, 21°54′E.

Below the surface layer one finds, in the greater part of the Mediterranean, an intermediate water which is characterized by a salinity maximum at a depth of 300 to 400 m. In the western sea a temperature maximum is present at nearly the same depth, but in the eastern sea only a second layer of small temperature decrease is found. This intermediate water, according to Nielsen (1912), is formed in the inner part of the Mediterranean, from where it spreads towards the west and finally flows out through the Strait of Gibraltar. At the *Thor* stations 178 and 160, in about long. 30°E, no subsurface salinity maximum was present but the salinity decreased regularly from the surface. Between 175 and 200 m the temperature remained constant, and this layer of constant temperature represents both the layer to which convection currents reach in winter and the layer of intermediate water which originates in that part of the sea. As this water spreads to the west, the salinity and the temperature at the core decrease owing to mixing processes.

The lower boundary of the intermediate water can be placed, somewhat arbitrarily, at about 600 m, where both temperature and salinity decrease rapidly. This decrease continues within the transition layer between the intermediate water and the deep and bottom water which, in the Mediterranean basins, is found below a depth of 1500 to 2000 m. At this depth a temperature minimum is encountered, and at greater depth the temperature increases adiabatically towards the bottom, as first shown by Nielsen. The potential temperatures at stations 4119 and 4070, at depths greater than 1500 m, are also plotted in fig. 175, from which it is seen that at station 4119 the potential temperature between 2000 and 3400 m was constant at 12.74° and at station 4070 the potential temperature between 2500 and 4200 m was constant at 13.19°. The salinity appeared to decrease slightly towards the bottom, but the decrease was nearly within the limits of accuracy of the observations. At station 4119 the average salinity below 2000 m was 38.42 ‰ and at station 4070 the average salinity below 2500 m was 38.63 ‰, the greatest deviation from the averages being 0.02 ‰. Therefore, the Mediterranean basins are filled by water of uniform character, although the type varies slightly from one basin to another. According to the *Thor* and *Dana* observations the average values of temperature and salinity at 2000 m are:

Basin	Temperature (°C)	Potential temperature (°C)	Salinity (‰)
Algiers-Provencal	13.00	12.69	38.39
Tyrrhenian	13.10	12.79	38.44
Ionian	13.57	13.25	38.65
Levantine	13.62	13.30	38.66

These data show that there is an appreciable difference between the western and eastern basins, which are separated by the rise between Tunisia and Sicily.

The deep and bottom water in the different basins is formed in localities in which the intermediate water is lacking and where convection currents in winter can reach from the surface to the bottom. Nielsen shows that the bottom water of the western basins originates mainly from the northern parts of the Balearic and Ligurian Seas, off the French and Italian Rivieras. At the *Thor* station 37, which was occupied on January 30, 1909, in lat. 41°56′N, long. 6°18′E, nearly uniform density was found between the surface and 2000 m. The average value of σ_t was 29.04, and the greatest deviations from the average were ± 0.04. The temperature at the surface was 12.40°, and the potential temperature at 2000 m was 12.70°, the salinities were 38.24 ‰ and 38.35 ‰, respectively. Even more extreme conditions were observed by Nathansohn

(Krümmel, 1911) off Monaco on April 7, 1909, when the temperature at the surface was 12.67°, the potential temperature at 2213 was 12.62°, and the salinities were 38.51 °/₀₀ and 38.49 °/₀₀, respectively. Probably no bottom water is formed in the Tyrrhenian Sea, and the slightly higher temperatures and salinities of the deep water are due to a small admixture of intermediate water. The deep and bottom water of the eastern basins is formed, according to Nielsen, in the Aegean Sea and in the southern part of the Adriatic Sea, from where it spreads to the south and west.

Nielsen believes that the formation of bottom water takes place intermittently and this concept is substantiated by the potential temperatures at station 4070, which were constant at 13.38° between 1500 and 2000 m and at 13.19° below 2500 m, indicating that slightly different water is formed under different conditions (fig. 175). Nielsen furthermore emphasizes that the uniform character of the deep and bottom

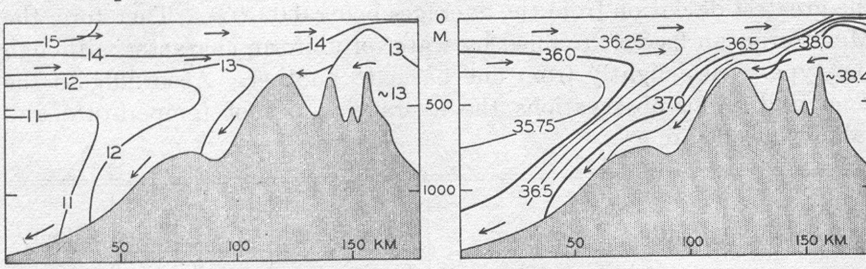

Fig. 176. Temperature and salinity in a vertical section through the Strait of Gibraltar.

water in a horizontal direction demonstrates active horizontal movement at great depth.

The water that sinks from the surface in order to renew the deep water must be replaced, and such replacement is accounted for by an inflow of Atlantic water. The exchange of water between the North Atlantic Ocean and the Mediterranean takes place through the Strait of Gibraltar, which is noted for its strong currents. The bottom topography of the Strait is complicated; in its shallow portion two or possibly three ridges are present with sill depths of about 320 m. The character of the water masses passing in and out through the Strait of Gibraltar is illustrated in fig. 176, showing isohalines and isotherms in a longitudinal section through the Strait. The observations used for perparing this section were made in late spring and early summer, in the months of May, June, and July of different years (Schott, 1928, Ramalho and Dentinho, 1931).

In the upper layers Atlantic water of a salinity somewhat higher than 36.00 °/₀₀ and a temperature higher than 13° flows in, whereas water of a salinity higher than 37.00 °/₀₀ and a temperature about 13° flows out along the bottom. The deeper water in the Mediterranean Sea inside the Strait of Gibraltar has a salinity of about 38.40 °/₀₀ and a temperature of 13°, but in the Strait intensive mixing takes place, whereby the salinity of the outflowing water is greatly reduced and that of the inflowing is increased. In the Strait the inflowing water has a thickness of approximately 125 m, but the boundary surface separating the in- and outflowing layers lies deeper on the African side of the Strait and the greater inflow therefore takes place on the southern side. This inclination of the boundary surface is due to the effect of the earth's rotation. The average velocity at the surface in some localities is in excess of 200 cm/sec (4 knots), and close to the African coast a narrow countercurrent with velocities up to 100 cm/sec (2 knots) is often encountered. Superimposed on the currents carrying water in and out of the Mediterranean Sea are strong tidal currents which greatly reduce the inflow when they are directed from the Mediterranean Sea to the Atlantic Ocean, and greatly increase the inflow when they are directed from the Atlantic to the Mediterranean. The inclination of the boundary surface probably varies with the speed of the inflowing current, and great vertical oscillations of tidal period therefore take place.

The average velocity of the total inflow, according to measurements on Danish expeditions, is approximately 100 cm/sec (2 knots). By means of this value Schott (1915) has computed the inflow to be approximately 1.75 million m^3/sec. The outflow can be computed by means of the relations on page 148. Such computations have been made by Nielsen (1912) and Schott (1915), but subsequent data indicate that they have assumed somewhat too high values for the salinity of the outflowing water. From the sections published by Schott in 1928 it appears that the average salinity of the inflowing water should be, in the Strait of Gibraltar, about 36.25 ‰ and that of the outflowing water should be not more than 37.75 ‰. With these values one obtains an outflow of 1.68 million m^3/sec and the difference between inflow and outflow, 70,000 m^3/sec, represents the excess of evaporation over precipitation and runoff. A fraction of this excess is made up, however, by a net inflow from the Black Sea, amounting to 6500 m^3/sec (p. 650).

The exchange of water through the Strait of Gibraltar presents a good example of how water from one region can be transformed by external influences and return as a different type of water, in this case as water of high salinity. The rapidity of the exchange is illustrated by stating that the in- and outflow is sufficient to provide for a complete renewal of the Mediterranean water in about seventy-five years.

Using the figures which Schott gives for precipitation and runoff, one arrives at the water budget of the Mediterranean Sea proper, which is summarized in table 78. From this it appears that the total evaporation amounts to 115,400 m^3/sec, corresponding to an annual evaporation of 145 cm, which is in fair agreement with the annual evaporation in these latitudes according to observations and computations (fig. 27, p.

121). As one might expect, the evaporation from the Mediterranean Sea is somewhat greater than that from the open ocean in the same latitude, which is 110 cm a year.

This discussion brings out a very characteristic difference between the Mediterranean Sea and the American Mediterranean Sea. The Mediterranean Sea exercises no great influence on the surface currents of the Atlantic Ocean because the amount of water that flows in through the Strait of Gibraltar represents a small fraction of the water masses that are transported by currents of the upper water layers; but the Mediter-

TABLE 78

WATER BUDGET OF THE MEDITERRANEAN SEA

Gains	m^3/sec	Losses	m^3/sec
Inflow from the Atlantic Ocean	1,750,000	Outflow to the Atlantic Ocean	1,680,000
Inflow from the Black Sea	12,600	Outflow to the Black Sea	6,100
Precipitation	31,600	Evaporation	115,400
Run-off	7,300		
	1,801,500		1,801,500

ranean exercises a widespread influence on the deep water of the North Atlantic by adding appreciable quantities of water of high salinity. The American Mediterranean Sea, on the other hand, exercises a great influence upon the currents of the upper layers because large water masses flow into the Caribbean Sea and out of the Gulf of Mexico, and the types of currents are greatly modified by the character of the passages; but the importance of the American Mediterranean Sea to the deep-water circulation is negligible.

Within the Mediterranean an exchange of water takes place across the submarine ridge between Tunisia and Sicily that is similar to the exchange through the Strait of Gibraltar. Surface water of Atlantic origin and of salinity about 37.20 ‰ flows towards the west through the channel which cuts the ridge between Tunis and Sicily at a depth of about 400 m. According to Nielsen about 4 per cent of the inflowing water is lost by excess evaporation in the eastern Mediterranean, whereas 96 per cent is carried out again by the intermediate current after the salinity has been increased by about 1.5 parts per mille.

In fig. 177 a schematic picture is given of the surface currents and the flow of the intermediate water (according to Nielsen, 1912). It is seen that both surface currents and intermediate currents have a tendency to circle the different areas in a counterclockwise direction.

The distribution of oxygen confirms the conclusions as to the origin of the different water masses of the Mediterranean and the flow of water.

The surface water, which extends to a depth of 100 to 200 m, shows a high oxygen content. The oxygen content of the intermediate water is relatively high in the eastern Mediterranean where this water is formed, but decreases toward the west, the lowest values being found in the western part of the Mediterranean area, the Balearic Sea. The transition layer between the intermediate and the deep water is characterized by an oxygen minimum that is more conspicuous in the eastern than in the western regions. The deep water has a somewhat higher oxygen content, the highest value being found in the Balearic Sea, where the formation of deep water is probably most rapid. These general features are shown in table 79, which is based upon the *Thor* and the *Dana* observations in the summers of 1910 and 1930. For three different areas, the Ionian, Tyrrhenian, and the Balearic Seas, the table contains mean

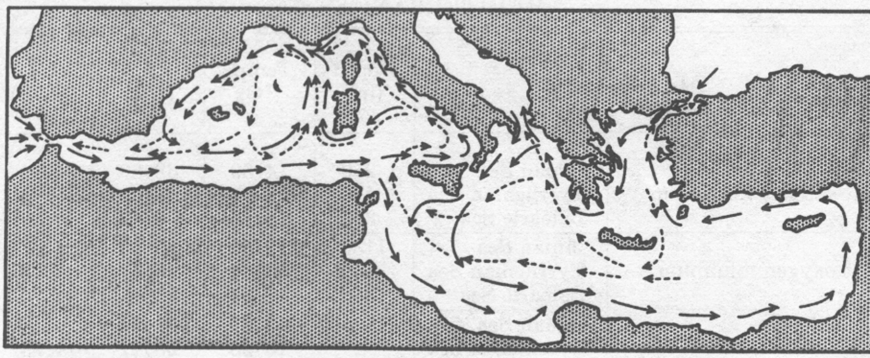

Fig. 177. Surface currents (*solid arrows*) and currents at intermediate depths (*dashed arrows*) in the Mediterranean Sea (after Nielsen).

values of temperature, salinity, and oxygen at the layer of salinity maximum, at the layer of oxygen minimum, and below 2000 m. It also gives average depths of the salinity maximum and the oxygen minimum within the different areas. The observations from 1910 and 1930 agree remarkably well in the western Mediterranean, but in the Ionian Sea the oxygen values and the salinity values were on the whole higher in 1930 than in 1910, indicating that in different years the formation of intermediate water may be more or less intense.

The water masses of the Black Sea are entirely different from those of the Mediterranean proper. In the Black Sea precipitation and runoff exceed evaporation, for which reason a surface layer is present of relatively low salinity and correspondingly low density. Vertical mixing tends to reduce the density of the deep water, which therefore has a lower density than that of the water in the Mediterranean Sea proper at the same depth. The character of the waters of the Black Sea is seen from table 80, containing some of the observations at *Thor* Station 172, about 100 m inside the Bosporus.

Owing to the distribution of density, the surface waters of the Black Sea flow out through the Bosporus and through the Dardanelles, and Mediterranean water flows in along the bottom. Intensive mixing takes place in these narrow straits, and the salinity of the inflowing water is therefore reduced from more than 38.50 ‰ at the entrance of the Dardanelles to between 35.00 and 30.00 ‰ where the bottom current enters the Black Sea at the northern end of the Bosporus. Similarly, the salinity of the outflowing water is increased from about 16.00 ‰ where it enters the Bosporus to nearly 30.00 ‰ where it leaves the Dardanelles. The out- and inflowing water masses are separated by a well-defined layer of transition which oscillates up and down according to the contours

TABLE 79

CHARACTERISTICS OF THE OXYGEN DISTRIBUTION IN THE MEDITERRANEAN

Location		Depth (m)	Temp. (°C)	S (‰)	O_2 (ml/L)
At salinity maximum	Ionian Sea	250	14.59	38.83	4.84
	Tyrrhenian Sea	420	13.91	38.65	4.30
	Balearic Sea	390	13.18	38.47	4.16
At oxygen minimum	Ionian Sea	1180	13.59	38.68	4.01
	Tyrrhenian Sea	960	13.30	38.51	4.12
	Balearic Sea	580	13.07	38.44	4.10
Below 2000 m	Ionian Sea		13.67	38.64	4.14
	Tyrrhenian Sea		13.23	38.41	4.25
	Balearic Sea		13.06	38.39	4.55

of the bottom. The currents through the Bosporus can well be compared to two rivers flowing one above the other in a river bed which has a width of about 4 km (2 miles) and a depth of 40 to 90 m.

Appreciable water masses pass through the Bosporus. According to current measurements and other observations by A. Merz (L. Möller, 1928), the most probable values of the out- and inflow through it are 12,600 m^3/sec and 6,100 m^3/sec, respectively. (The Mississippi River carries, on an average, 120,000 m^3/sec.) The difference between inflow and outflow, 6,500 m^3/sec, represents the excess of precipitation and runoff over the evaporation. Precipitation and runoff have been estimated at 7,600 and 10,400 m^3/sec, respectively, and the evaporation should therefore amount to 11,500 m^3/sec or 354 km^3/year. The area of the Black Sea is 420,000 km^2 and the above value therefore corresponds to an evaporation of 84 cm per year, in good agreement with observed and computed values for this latitude (p. 121).

For the Black Sea the out- and inflowing water masses are of the same order of magnitude as the difference between precipitation plus run-off and evaporation, whereas for the Mediterranean proper the in- and outflowing water masses are many times greater than the excess of evaporation. This is in agreement with the general considerations which were set forth on page 147. It follows that the renewal of the deep water of the Black Sea is a very slow process because the entire renewal has to be taken care of by the water which flows in along the bottom of the Bosporus. This inflow is so small in proportion to the total volume of water in the Black Sea that complete renewal of the water below a depth of 30 meters would take about 2500 years. For all practical purposes the deep water of the Black Sea is stagnant, as is evident from the

TABLE 80

HYDROGRAPHIC CONDITIONS IN THE BLACK SEA

(*Thor* Station 172, August 10, 1910, 41°32′N, 29°24′E, Sounding, 1090 m)

Depth (m)	Temp. (°C)	S (‰)	σ_t	O_2 (ml/L)	H_2S (ml/L)
0	24.1	17.59	10.56	5.14	
10	24.1	17.59	10.56	5.14	
25	12.73	18.22	13.54	7.40	
50	8.22	18.30	13.22	6.71	
75	7.44	18.69	14.62	5.51	
100	7.61	19.65	15.33	2.33	
150	8.31	20.75	16.12	0.17	
200	8.54	21.29	16.51		0.90
300	8.68	21.71	16.82		2.34
400	8.72	21.91	16.97		4.17
600	8.76	22.16	17.16		4.96
800	8.80	22.21	17.19		6.06
1000	8.85	22.27	17.24		6.04

fact that below a depth of 200 m the water contains no oxygen, but large quantities of hydrogen sulphide (table 80).

THE ARCTIC MEDITERRANEAN SEA. Under the name "Arctic Mediterranean Sea" are understood the areas to the north of the Wyville Thomson Ridge between the Orkneys and Faeroe Islands, the Faeroe Islands-Iceland Ridge, and the Iceland-Greenland Ridge. Exchange of water between the Atlantic Ocean and the Arctic Mediterranean Sea takes place across these ridges. In addition, the water masses of the Arctic Mediterranean are in restricted communication with those of the Atlantic through the English Channel and through the narrow sounds to the west of Greenland, and with the waters of the Pacific Ocean through Bering Strait. The Arctic Mediterranean is the largest of all adjacent seas and is subdivided into a number of distinct areas, of which the more important are the Norwegian Sea, the North Sea, the Baltic with its gulfs, the Barents Sea, and the North Polar Sea. The areas of some of these parts are shown in table 4 (p. 15).

Much the major exchange of water between the Arctic Mediterranean and the adjacent oceanic regions takes place through the straits between Scotland and Greenland. The character of the exchange is radically

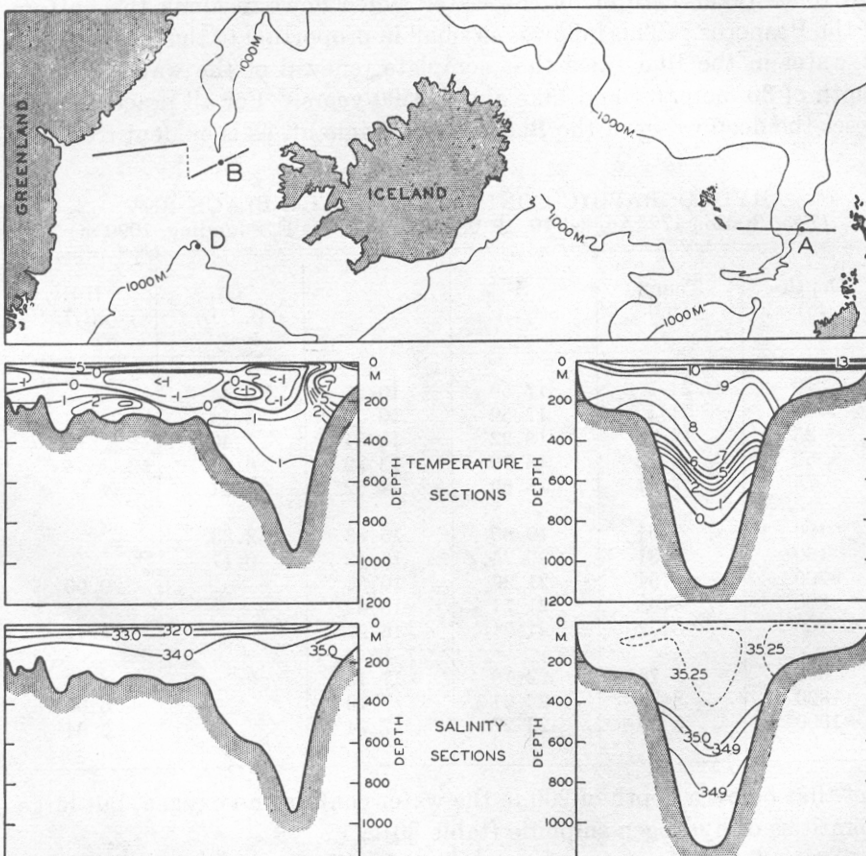

Fig. 178. Chart of the important openings between the Atlantic Ocean and the Arctic Mediterranean Sea, and vertical sections showing distribution of temperature and salinity in the Faeroe-Shetland Channel (*right*) and in the Denmark Strait (*left*). Stations A, B and D are shown, but station C lies outside of the region covered by the chart.

different from that between the Atlantic Ocean and the Mediterranean Sea, where inflow of Atlantic water takes place in the upper layers of the Strait of Gibraltar and outflow near the bottom. The opening between the Atlantic Ocean and the Arctic Mediterranean is so wide that the water flows in through the southeastern part of the opening and out through the northwestern. The character of the water masses is illustrated in fig. 178. The top part of the figure contains a chart of the area between northern Scotland and eastern Greenland on which the 1000-m bottom contour is shown. On the chart are indicated the location of two sections, one through the Faeroe-Shetland Channel to the north of the Wyville Thomson Ridge, and one through the Denmark Strait between Iceland and Greenland to the north of the Iceland-Greenland Ridge (fig. 3, p. 30). In the lower part of the figure are shown the distributions of temperature and salinity in these sections according to observations of the *Michael Sars* in 1910 (Helland-Hansen, 1930) and observations on board the *Heimdal* in 1933 (Helland-Hansen, 1936). The principal inflow takes place across the Wyville Thomson Ridge. The inflowing water masses are characterized by a salinity higher than 35.00 $^0/_{00}$ and a temperature higher than 4°, the larger part of the water masses having a salinity above 35.25 $^0/_{00}$ and a temperature above 8°. The section through Denmark Strait shows that a small amount of Atlantic water of salinity above 35.00 $^0/_{00}$ and temperature above 4° flows north along the west coast of Iceland, but by far the greater amount of the water masses shown in the left-hand sections of fig. 178 flow out, having temperatures between 2° and $-1.5°$, and salinities ranging from 34.90 $^0/_{00}$ to 31.00 $^0/_{00}$.

The great difference between the inflowing Atlantic water masses and the outflowing Arctic water is illustrated in fig. 179. In the main part of the figure, *T-S* curves are plotted showing the character of the waters at *Michael Sars* station 106 (A), which was occupied in the Faeroe-Shetland Channel on August 10 and 11, 1910, and at the *Heimdal* station 47 (B), which was occupied in the channel to the north of the Iceland-Greenland Ridge on July 22, 1933. The inset diagram shows *T-S* curves which are drawn on a much wider salinity scale, demonstrating the character of the waters at *Michael Sars* station 106 (A); at *Michael Sars* station 100 (C), which was occupied to the southwest of the Wyville Thomson Ridge on August 6, 1910; and at the *Heimdal* station 33 (D), which was occupied to the south of the Iceland-Greenland Ridge on July 17, 1933. The last two stations have been added in order to demonstrate the similarity between the waters in the Faeroe-Shetland Channel above 600 m and the water masses to the south of the Ridge on the entire distance from the Orkneys to Greenland. The inset diagram demonstrates that one has to deal here with exactly the same water mass, but water which to the north of the Wyville Thomson Ridge was found at a depth of 600 m, south of the ridge was present at 1200 m, and to the west of Iceland was present at 800 m.

The similarity of the water masses at *Michael Sars* station 100 and *Heimdal* station 33 shows that only part of the Atlantic water flows across the Wyville Thomson Ridge and that part bends towards the west and flows towards Greenland. This current will be dealt with later

(p. 682), but here it will be emphasized that the Orkney-Greenland Ridge forms a most effective barrier between the deeper waters of the Atlantic and those of the Arctic Mediterranean.

On the southern side of the barrier, Atlantic water is present to the greatest depths except along the continental shelf of Greenland, where the Polar Current flows south, but on the northern side of the barrier and below the sill depths the water masses have a uniform salinity of about

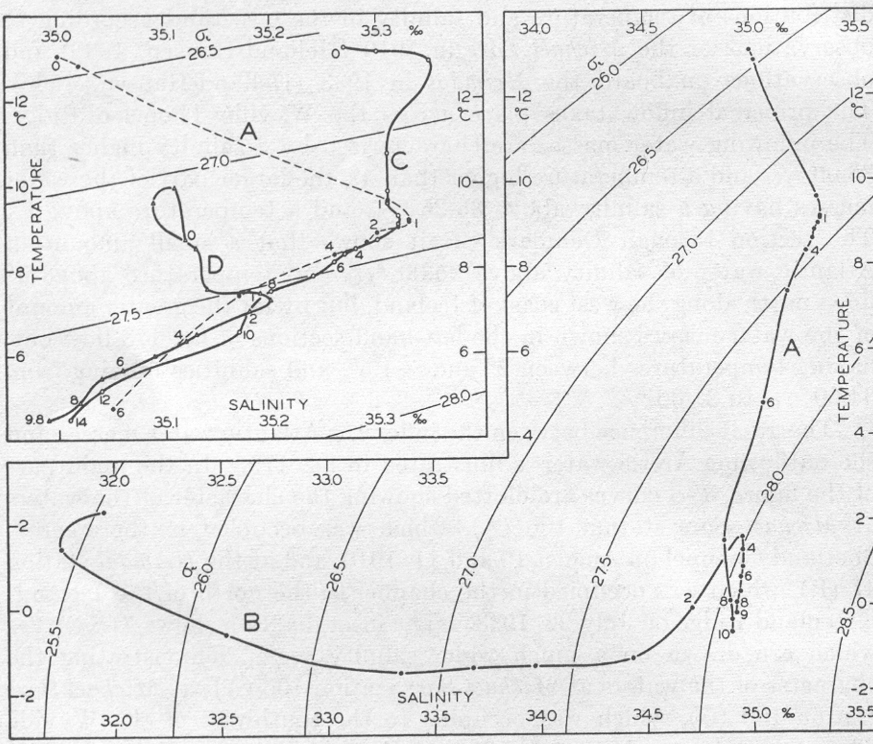

Fig. 179. Temperature-salinity relations at stations in the Faeroe-Shetland Channel (A) and in Denmark Strait (B). *Inset:* Temperature-salinity relations at stations southwest of the Wyville Thomson Ridge (C) and south of the Iceland-Greenland Ridge (D). Depths in hectometers are entered along the curves. Locations of stations A, B, and D are shown in fig. 178.

34.92 $^0/_{00}$ and a temperature below 0°. This deep water, formed in the Norwegian Sea, has a greater density than the Atlantic water at corresponding depths, but there is no evidence of an outflow of this water across the sills. At the Wyville Thomson Ridge the deep water of the Norwegian Sea does not reach up to the sill depth, whereas at the Iceland-Greenland Ridge it does, as is evident from the *Heimdal* observations, but no trace of that deep water was present at *Heimdal* station 33 on the southern side of the ridge (see curve D, fig. 179) nor at any of the *Meteor* stations in the same region (Defant *et al*, 1936).

Helland-Hansen (1934) has computed the volumes of Atlantic water that in May, 1927 and May, 1929, flowed north through a section from lat. 60°40′ on the coast of Norway to lat. 63°20′, long. 4°W, northwest of the Faeroe Islands. He found 2.78 and 3.31 million m^3/sec respectively, and the average inflow of Atlantic water may therefore be estimated at about 3 million m^3/sec.

In addition to the inflow of Atlantic water, an inflow of Bering Sea water takes place through Bering Strait. On August 1, 1934, the United States Coast Guard Cutter *Chelan* anchored in Bering Strait and measured currents between the surface and the bottom during 21 hours. From these measurements it was concluded (U. S. Coast Guard, 1936) that the transport inward through Bering Strait was 0.88 million m^3/sec, but the average annual transport is probably much smaller because the inflow is less regular and weaker in winter. It does not seem probable that the average inflow during the year exceeds 0.3 million m^3/sec.

The Arctic Mediterranean also receives a considerable amount of fresh water in the form of runoff from the great Siberian and Canadian rivers and from excess of precipitation over evaporation. The runoff from the Siberian rivers averages, according to Zubov (1940), about 0.16 million m^3/sec. The excess precipitation over the North Sea and the Norwegian Sea can be estimated at about 0.5 m per year, and over the Arctic Sea, where the precipitation is small, at about 0.12 m per year. On this basis one finds a total excess of precipitation of about 0.09 million m^3/sec.

The outflow from the Arctic Mediterranean Sea takes place principally through the Denmark Strait, for which reason the water budget of the region can be presented as follows:

Inflow northwest of Scotland	3.0 million m^3/sec
Inflow through Bering Strait	0.3 million m^3/sec
Runoff from rivers	0.16 million m^3/sec
Excess precipitation	0.09 million m^3/sec
Inflow and addition of fresh water	3.55 million m^3/sec
Outflow through Denmark Strait	3.55 million m^3/sec

On the basis of these figures one finds that a complete renewal of the waters of the Arctic Mediterranean Sea would take about 165 years.

A rough check on the relative correctness of the above values can be obtained by considering that the net salt transport into the Arctic Mediterranean must be zero. The average salinity of the inflowing Atlantic water is about 35.30 ‰ and that of the Bering Sea water about 32.00 ‰. Therefore $3.0 \times 35.3 + 0.3 \times 32.0 = 3.55\ S_A$ where S_A is the salinity of the outflowing Arctic water. This equation gives $S_A = 32.5$ ‰, in fair agreement with the salinity shown in the section in fig. 178. If the addition of fresh water were 0.1×10^6 m^3/sec or 0.4×10^6 m^3/sec, the salinity of the outflowing water would be 34.0 ‰

or 31.2 ‰, respectively, but of these values the first appears to be too high and the second appears to be too low, whereas the assumed values of 0.25×10^6 m^3/sec leads to a reasonable result.

The amount of heat given off by the water can also be estimated. The average temperatures of the in- and outflowing waters between Scotland and Greenland can be taken as 8°C and −1°C, respectively, and the average temperature of the water flowing through Bering Strait, the runoff, and the excess precipitation can be taken as 0°C. The total amount of heat given off by the waters in the Arctic Mediterranean Sea is then about 24×10^{12} g cal/sec. It is probable that at least half of this amount is given off where the Atlantic water flows north along the west coast of Norway, or over an area not greater than 2×10^{12} m^2. The average amount of heat given off in this area would then be 12 g cal/m^2/sec = 0.072 g cal/cm^2/min = 103 g cal/cm^2/day. This value, although uncertain, serves to demonstrate the important bearing of the Atlantic Current on the climate of the extreme northwestern part of Europe.

From the above discussion of the water, salt, and heat budgets of the Arctic Mediterranean Sea it is evident that the Atlantic water which contributes mostly to the inflow is diluted and cooled off to such an extent that the outflowing water is of entirely different character. We have here another striking example of local factors operating towards changing the inflowing water and producing a new water mass.

The influence of the local factors is more or less conspicuous in the different parts of the Arctic Mediterranean Sea, giving rise to other water masses than those mentioned, and to water masses produced by mixing. Every subdivision of the Arctic Mediterranean has its own characteristic water masses, which will be briefly described.

In the Norwegian Sea, Atlantic water is found off the west coast of Norway, where it flows to the north, losing some of its heat content to the atmosphere and being somewhat diluted by excess precipitation. On the right-hand side of the Atlantic water is the Norwegian coastal water, which has a lower salinity, owing to runoff, and a considerable annual range in surface salinity and temperature. On the left-hand side of the Atlantic water are found water masses which have been formed by mixing between the Atlantic water and the Arctic water which flows south along eastern Greenland. The latter is characterized by low salinity and temperatures below 0°C, as illustrated in the left-hand section in fig. 178. The mixed water in the central and western parts of the Norwegian Sea has a salinity around 34.90 ‰ and at the surface a temperature which varies considerably during the year. In winter the surface layers are cooled, but before reaching freezing point the waters attain a higher density than that of the deeper waters and therefore sink to the bottom. By this process, first described by Nansen (1906), the bottom water of the Norwegian Sea is renewed. As further evidence for the correctness of the explanation, Helland-Hansen and Nansen (1909) point out that surface samples, taken by sealing vessels to the northeast of Jan Mayen in March to May, show temperatures between $-1.2°$ and $-1.9°$, and salinities between 34.70 $^0/_{00}$ and 34.94 $^0/_{00}$. The bottom water has a salinity between 34.92 $^0/_{00}$ and 34.93 $^0/_{00}$ and in the northern part a temperature of about $-1.1°$ to $-1.2°$C, while in the southern part the temperature is about $-1.0°$C. This uniform bottom water fills all the basins of the Norwegian Sea at depths below 600 m, but above 1500 m the temperatures are somewhat higher than those mentioned.

The North Sea waters have, in general, salinities between 34.00 $^0/_{00}$ and 35.00 $^0/_{00}$, but Atlantic water of salinity above 35.00 $^0/_{00}$ is found in a tonguelike area to the south of a line from Scotland to the west coast of Norway and in another tonguelike area extending northwest from the English Channel. Norwegian coastal water of salinity as low as 30.00 $^0/_{00}$ is encountered in the northeastern part of the North Sea (Deutsche Seewarte, 1927).

The Baltic Sea, together with the Gulf of Finland and Gulf of Bothnia, represents a region which, oceanographically, has some features in common with the Black Sea. In the Baltic, as in the Black Sea, precipitation and runoff greatly exceed evaporation, for which reason a layer of brackish surface water is formed that flows out through the sounds between Sweden, the Danish Islands, and Jutland. Along the bottoms of these sounds, North Sea water that has been considerably diluted with Baltic water flows in. The sounds are narrow and shallow, for which reason intensive mixing takes place in and directly outside the sounds, and the influence of the Baltic water does not reach to any great distance. In this respect conditions are similar to those in the Black Sea and the Mediterranean, where the influence of the low-salinity water flowing out from the Black Sea is present only in the northern part of the Aegean Sea. There exists, however, one striking difference. The Black Sea has a deep basin filled by stagnant water of relatively high salinity, but in the Baltic only a few small basins exist, the deepest (with a maximum depth of about 210 m) being found off the island of Gotland. In these basins water of higher salinity is found, but the basins are so limited in extent and the difference in salinity between the bottom water and the surface water is so small that renewal of the deep water in the basins can take place, and stagnant water corresponding to that of the Black Sea is not found. Over large portions of the Baltic the salinity of the surface layer is about 7.00 $^0/_{00}$ but it drops, in the inner parts of the Gulfs of Finland and Bothnia, to less than 3.00 $^0/_{00}$, in spring even below 1.00 $^0/_{00}$ (Deutsche Seewarte, 1927). The salinity of the water in the different basins decreases from about 16.00 $^0/_{00}$ in the southern basins to about 12.00 $^0/_{00}$ in the basins further north. The surface temperatures show a large annual variation.

In the Gulfs of Finland and Bothnia ice forms during winter, and these gulfs remain ice-covered for periods between three and five months, depending upon the severity of the winter. In very severe winters the sounds between Sweden, the Danish Islands, and the Danish mainland also freeze over.

In the North Polar Sea three main water masses are encountered, the Arctic Surface Water, the Atlantic water, and the Arctic Deep Water. The Arctic Surface Water has a low salinity ranging from a few per mille at the mouths of the Siberian rivers to 32 ‰ or 33 ‰ to the north of Spitsbergen. Owing to the low salinity of the surface waters, no deep water is formed in the Arctic Sea itself, but in winter a top layer of homogeneous water is developed. On the North Siberian Shelf at a distance of about 400 km from the coast, the salinity of this homogeneous layer reached, in 1923, its maximum value, 29.67 ‰, in May (Sverdrup, 1929), but at the end of the summer a nearly fresh top layer was formed by the melting of ice, and as a result of admixture of this fresh water the salinity was somewhat reduced down to a depth of about 30 m. The temperature of this upper layer remained at freezing point during the winter and in summer was raised slightly above freezing point. It is probable that an upper layer of similar characteristics is found all over the Polar Sea, as indicated by Nansen's observations in 1893–1896, but data from later Russian expeditions are not yet available.

Below this upper layer and a transition layer is found the Atlantic water, which enters the North Polar Sea north of Spitsbergen as a subsurface flow. In 1931 (Mosby, 1938) water of a salinity of 35.10 ‰ and a temperature between 3° and 4° was found to the north of Spitsbergen in lat. 80°38′ and longitude 13°41′E, between depths of 75 and 400 m. Nansen's observations (1902) showed that this Atlantic water can be traced across the Polar Sea toward the region of theNew Siberian Islands, and his observations have recently been confirmed by those which were conducted during the drift of the *Sedov* in 1937–1940. Zubov (1940) states that both the temperature and the salinity of the Atlantic water were higher in 1937–1940 than they were in 1893–1896. In both instances admixture with Polar water led to a decrease of temperature and salinity in the direction in which the Atlantic water spreads out.

The deep water of the Polar Sea has a uniform salinity of about 34.93 ‰ and uniform temperature of about -0.85°. Nansen observed an increase of the temperature toward the bottom which can be explained as a result of adiabatic heating. This deep water cannot be formed anywhere in the Polar Sea, but it shows great similarity to the deep water in the northern part of the Norwegian Sea. Nansen therefore concluded that the deep water of the Polar Sea was formed in the Norwegian Sea and flowed in across the submarine ridge between Spitsbergen and Greenland, which supposedly has a sill depth of 1200 to 1500 m. These conclusions were confirmed by the observations made on board the *Nautilus* (Sverdrup, 1933). It is not yet known whether the observations on recent Russian expeditions will modify the earlier results.

The marginal areas of the North Polar Sea, the Barents Sea, the Kara Sea, and the areas between the islands of the Canadian Archipelago show complicated conditions owing to dominating local influences and inflow of different water masses. Atlantic water, for example, flows into the Barents Sea both to the north of Norway, where it branches out in different directions, and to the north of Spitsbergen, where part of the Atlantic water turns south between Northeastland and Franz Josef Land. The different basins in the Barents Sea therefore contain waters of slightly differing characteristics depending upon their origin.

The currents of the Arctic Mediterranean Sea have been repeatedly mentioned in the preceding discussion of the water masses. Later observations have not materially changed the conception of the surface currents of the Norwegian Sea which Helland-Hansen and Nansen presented in 1909. Their picture is reproduced in fig. 180, according to which the two main currents in that region are the Norwegian Current, representing a continuation of the North Atlantic Current, and the East Greenland Current.

The Norwegian Current, a part of the Gulf Stream system (p. 673), is flanked on the left-hand side by a series of whirls, some of which are probably stationary and related to the bottom topography, whereas others may be traveling eddies. Off northern Norway the Norwegian Current branches, one branch continuing into Barents Sea and another turning north toward Spitsbergen and bending around the northwest of the Spitsbergen islands. The waters of this current have a high salinity and a high temperature, the maximum salinity decreasing from about 35.3 ‰ north of Scotland to about 35.0 ‰ off Spitsbergen and the subsurface temperature decreasing from about 8° to less than 4° in the same distance.

The East Greenland Current flows on or directly off the East Greenland shelf, carrying water of low salinity and low temperature. The greater part of the East Greenland Current continues through Denmark Strait between Iceland and Greenland, but one branch, the East Iceland Arctic Current, turns to the east and forms a portion of the counterclockwise circulation in the southern part of the Norwegian Sea.

No velocities are entered on the figure, but within the Norwegian Current velocities up to 30 cm/sec, or about 0.5 knot, occur and within the East Greenland Current, where the water moves fastest directly off the shelf, velocities of 25 to 35 cm/sec are encountered. These values are based partly upon computation from the distribution of density and partly upon direct current measurements (Helland-Hansen and Nansen 1909, Jackhelln, 1936).

In the North Sea a counterclockwise circulation is also present, as has been demonstrated not only by the distribution of temperature and salinity but also by the results of large-scale experiments with drift bottles. The details of the currents are much more complicated than

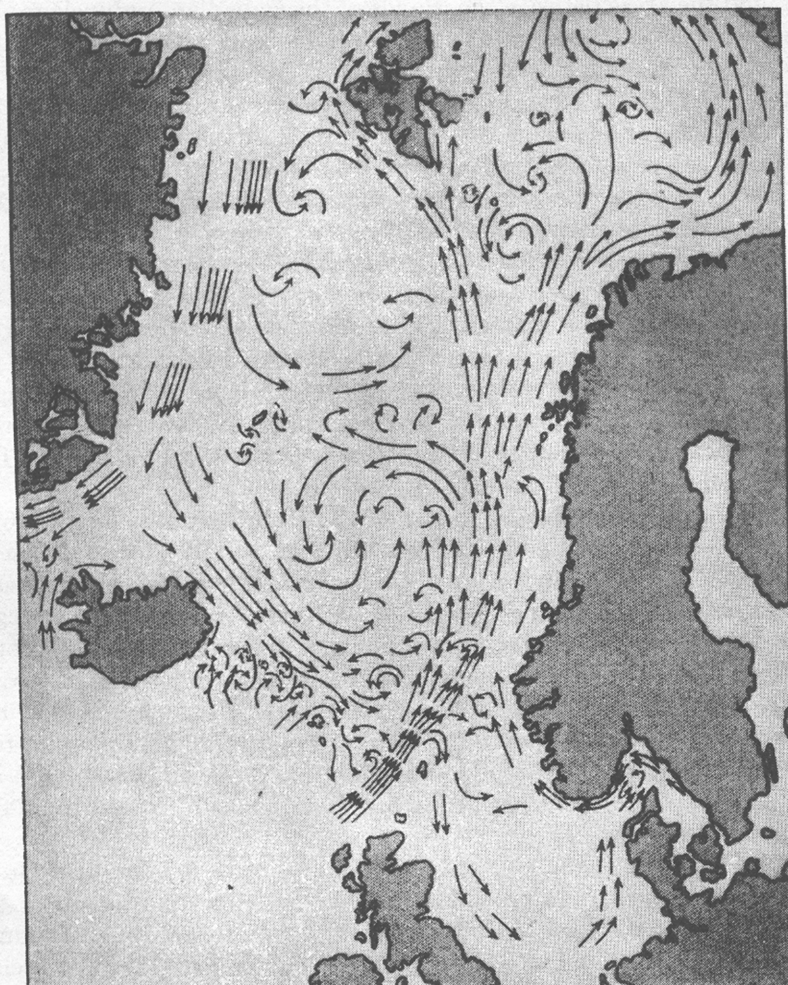

Fig. 180. Surface currents of the Norwegian Sea (after Helland-Hansen and Nansen).

can be shown in the figure, and several smaller but permanent eddies appear to be present. In the straits between Sweden and Denmark the surface current is directed in general from the Baltic to the North Sea, but along the bottom the water flows into the Baltic. In the Baltic and the adjacent gulfs the currents are so much governed by local wind conditions that no generalization is possible.

The outflowing water from the Baltic continues as a low-salinity current along the south and southwest coast of Norway, as shown in fig. 180. This Norwegian coastal current carries warm water in summer and cold water in winter, the development of the current being greatly influenced by the prevailing wind. A striking example of the advance of cold water masses along the coast in the winter of 1937 is discussed by Eggvin (1940), who traced the movements of a cold-water front by surface thermograph records and oceanographic observations at fixed localities.

In the North Polar Sea the surface currents are also greatly influenced by local winds. An independent current appears to be present only to the north of Spitsbergen and to the northeast of Greenland, where the surface waters flow south to feed the East Greenland Arctic Current. The greater part of the Atlantic water which reaches Spitsbergen as the northern branch of the Norwegian Current submerges below the Arctic surface water and spreads as an intermediate layer over large parts of the Polar Sea.

In the Barents Sea a counterclockwise circulation prevails with relatively warm water of Atlantic origin on the southern side and Arctic water on the northern, and with numerous eddies in the central portion. In the other marginal areas of the Polar Sea—the Kara Sea, the Laptev Sea, the North Siberian Sea, and the Chukotsk Sea—the currents are mainly determined by local winds and, in summer, by the discharge of large quantities of fresh water from the Siberian rivers or the Yukon River, but details of the currents are little known.

The Norwegian Current and its branches are subject to variations which are related to other phenomena. Helland-Hansen and Nansen (1909, 1920) have shown that the surface temperature of the Atlantic water off the Norwegian coast fluctuates considerably from one year to another and that high summer temperatures of the surface water are mostly followed by high air temperatures during the subsequent winter and spring. Besides such minor fluctuations major changes take place that lead to altered conditions over a number of years (Helland-Hansen, 1934). In 1901–1905 and in 1925 and 1927 the maximum salinity of the Atlantic water off western Norway at depths greater than 50 m was between 35.30 $^0/_{00}$ and 35.35 $^0/_{00}$, but in 1929 a maximum salinity of 35.45 $^0/_{00}$ was observed, and in 1928 and 1930 the highest values were 35.43 $^0/_{00}$ and 35.40 $^0/_{00}$ respectively. The temperatures in 1929 were higher than on any previous occasion and, in agreement with this observation, the air temperature in Norway from November, 1929, to April, 1930, was higher, on an average, than in any winter since 1900.

Helland-Hansen found that in 1929 the Atlantic water flowing into the Norwegian Sea was not only warmer and of higher salinity but that the volume was also greater than average. He points out that such an increased inflow must have had far-reaching consequences. Owing to the time needed for the water to reach the Barents Sea the effect of the large amount of warm water should appear in the Barents Sea two years after the water was observed off southwestern Norway, and the inflow of warmer water should lead to an increase of the ice-free areas in spring. In agreement with this reasoning, in May, 1929, the ice-free areas in the Barents Sea east of 20°E comprised 330 km^2; in May, 1931, they extended over 710 km^2. An effect on conditions to the northwest of Spitsbergen should also appear one or two years later. Continuous observations are not available, but Mosby (1938) points out that, whereas in the years 1910, 1912, 1922, and 1923 the maximum salinity of the Atlantic water varied between 34.95 ‰ and 35.06 ‰, it reached a value of 35.14 ‰ in 1931. In earlier years the maximum temperature varied between 2.57° and 4.48°, but in 1931 it reached 5.04°. The inflow of the warmer water led to the return into Spitsbergen waters of the cod, which had not been caught there in commercial quantities during the preceding fifty or sixty years.

The causes of the fluctuations which have been described are not known, nor is it known whether these fluctuations are periodic in character, and the same statements apply to fluctuations which occur in other regions of well-defined currents.

The large water masses of the Arctic Mediterranean have a high oxygen content, but within the many minor bodies of water which are in restricted communication with the waters of the Arctic Mediterranean the oxygen content varies within wide limits. In general, the Atlantic water of the Norwegian Current has a high oxygen content, but off the Norwegian coast minimum values somewhat below 5 ml/L have frequently been observed. The deep water of the Norwegian Sea contains large quantities of oxygen, in agreement with the concept that this water is rapidly renewed by vertical convection currents which in late winter extend from the surface to the bottom. At a temperature of $-1°$ values up to 7.2 ml/L have been observed, corresponding to a saturation of 88 per cent. Both the Atlantic water that flows into the Polar Sea to the north of Spitsbergen and the deep water that enters over the Spitsbergen-Greenland Ridge is rich in oxygen, as is evident from the observations by Mosby (1938) and Sverdrup (1933). In the deep water of the Polar Sea to the north of Spitsbergen the highest observed value was 6.73 ml/L at a depth of 3,000 m. On the other hand, in coastal areas where the bottom topography or other local conditions are important, very low oxygen values may be found. On the North Siberian shelf, where occasional intrusion of high-salinity water takes place, low oxygen values occur, probably owing to the consumption of oxygen by the organisms on the bottom of the shallow shelf or to decomposition of organic matter which has accumulated on the bottom. Values as low as 1.56 ml/L were observed at a distance of 5 m from the bottom, where the salinity was about 32.00 ‰, whereas at 15 m from the bottom, where the salinity was 29.00 ‰, the oxygen content was nearly 8 ml/L (Sverdrup, 1929). In numerous basins in Norwegian fjords the water contains no oxygen but does contain large amounts of hydrogen sulphide (Ström, 1936).

THE LABRADOR SEA AND BAFFIN BAY. The waters of the Labrador Sea are at all depths in free communication with those of the Atlantic Ocean, wherefore the Labrador Sea should be considered not an adjacent sea but a part of the Atlantic. The waters of Baffin Bay, on the other hand, are partly separated from those of the Labrador Sea by the submarine ridge across Davis Strait; but, because of the close relationship between the waters of Baffin Bay and Labrador Sea, and because of the strong influence of similar local factors in both regions, it appears desirable to treat them jointly.

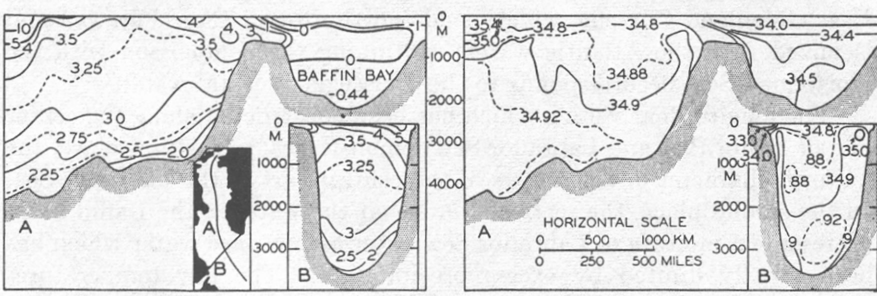

Fig. 181. Temperature and salinity distributions in a longitudinal section through the Labrador Sea and Baffin Bay, and in a cross-section from Labrador to Greenland. Location of sections shown in inset map.

The character of the water masses in the Labrador Sea and Baffin Bay is illustrated in fig. 181., based upon diagrams prepared by Smith *et al* (1937). The distributions of temperature and salinity are shown in a cross section from South Wolf Island off the coast of Labrador to Cape Farewell on Greenland and in a longitudinal section between latitudes 78°N and 55°N. The locations of the sections are indicated on the small inset map.

The cross section from Labrador to Greenland shows a distribution of temperature and salinity having considerable similarity to the distribution found in the Norwegian Sea. On the continental shelf off Labrador (to the left in the figure) is found water of a temperature which is mostly below 0°C and of salinity less than 34.00 ‰, the lowest values being less than 30.00 ‰. This water, which flows south, corresponds in its character to the waters of the East Greenland Current. The greater part of the section shows water of a uniform temperature which, except for a surface layer showing the effect of seasonal heating, is between 4° and 1.7°, by far the greater part of the water having a temperature between 3.5° and 3.0°. The salinity lies between 34.86 ‰ and 34.94 ‰, the highest salinity being found in the deep water at some distance from the bottom. At the surface the salinity is about 34.60 ‰, but the seasonal variations are great and higher values may be found in other seasons of the year. The uniform deep and bottom water of the Labrador Sea corresponds to the uniform deep water of the Norwegian Sea, but has a higher temperature, 3° as against −1°, and in part it has a somewhat higher salinity. Two different water masses are present off the coast of Greenland. Beyond the slope is a body of Atlantic water of salinity higher than 34.95 ‰, with maximum values above 35.00 ‰ and with temperatures higher than 4° and up to 6°. This body of water corresponds to the Atlantic water within the Norwegian Current. Close to the Greenland coast is water of a salinity between 34.00 ‰ and 31.00 ‰ and of a temperature, in summer, about 2°. This represents water of the East Greenland Current, which bends north around Cape Farewell and is mixed with the Atlantic water; continuing the comparison with the Norwegian Sea, it corresponds to the Norwegian coastal water.

The longitudinal section which has been constructed along the center line of Baffin Bay and Labrador Sea demonstrates, in the first place, the uniform character of the waters of the central part of the Labrador Sea. In the second place, the section shows that the water in the Baffin basin represents a mixture of Labrador Sea water and surface water which has been greatly diluted by excess precipitation. The low temperatures near the bottom indicate an admixture of surface water the salinity of which has been increased in winter by the freezing of ice.

The processes of cooling and of mixing which go on in the Labrador Sea are comparable to those in the Norwegian Sea. The large body of relatively uniform water in the central part of the Labrador Sea is formed by mixing of the two essentially different types of water, the Atlantic water and the Arctic water. In certain areas the mixed water at the surface will have a salinity in the neighborhood of 34.90 ‰ and, when cooled in winter to a temperature of 3° or lower, will sink and renew the deep and bottom water. That is, formation of deep and bottom water takes place in winter by vertical convection currents in the manner which was first described by Nansen when discussing the formation of the bottom water in the Norwegian Sea. The fact that the deep and bottom water has an oxygen content between 6.0 ml/L and 6.5 ml/L strongly supports the concept that this water is formed by vertical convection currents and is being renewed every winter.

In the Baffin basin entirely different conditions are encountered. The temperature and salinity of the bottom water in this basin indicate, as already stated, that here the bottom water represents a mixture of Labrador Sea deep water and surface water of the Baffin Bay region, the salinity of which has been increased sufficiently by freezing to cause the water to sink. It is probable that this sinking of cold surface water is a slow and intermittent process, because the oxygen content of the bottom water is as low as 3.5 ml/L. It is of interest to observe that the processes by means of which the bottom water of the Baffin basin is renewed are similar to those that take place in the Weddell Sea area, except that there sinking of water which is cooled on the continental shelf takes place more frequently, as is demonstrated by the higher oxygen content of the bottom water. A *T-S* curve from the central portion of the Baffin basin has a form similar to that of a *T-S* curve from the Weddell Sea, as should be expected if processes of similar character operate.

The surface currents of the Labrador Sea, which have been briefly mentioned, are shown in fig. 182, prepared by means of the charts of Smith *et al* (1937) and of Kiilerich (1939). These representations are the results of calculations based on the distribution of density. The outstanding features are the West Greenland Current that flows north along the west coast of Greenland and the Labrador Current that flows south off the coast of Labrador. Part of the West Greenland Current turns around when approaching Davis Strait and joins the Labrador Current, whereas part continues into Baffin Bay, where it rapidly loses its character as a warm current. Along the west side of Baffin Bay the Arctic waters flow south, having been partly reinforced by currents carrying Arctic water through the sounds between the islands to the west of Greenland. The central areas of the Labrador Sea and Baffin Bay both appear as areas in which numerous eddies occur, but nothing is known as to the permanency of the details shown in the picture. The similarity between the Labrador Sea and the Norwegian Sea is again striking, but it should be emphasized that, whereas the Norwegian Sea is in communication with the Atlantic Ocean in the upper 600 m only, so that the deep water is shut off, the Labrador Sea is in communication with the Atlantic at all depths and the deep water can freely flow to the south.

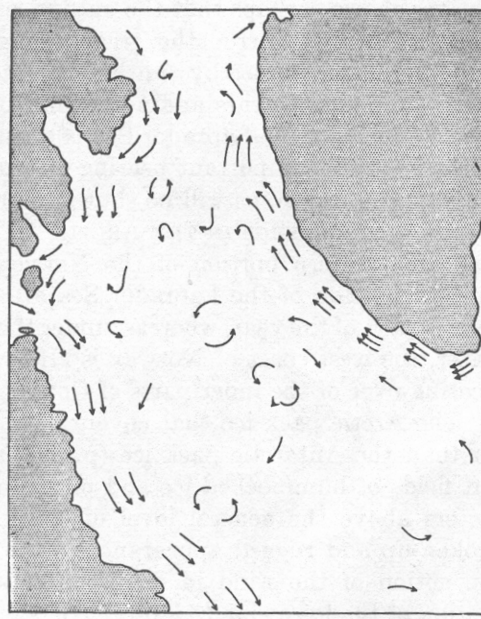

Fig. 182. Schematic representation of the surface currents in the Labrador Sea.

According to Smith *et al* (1937) the inflow in the Labrador Sea amounts to 7.5 million m^3/sec and the outflow along the coast of Labrador amounts to 5.6 million m^3/sec, both values referring to flow above a depth of 1500 m. From these figures Smith *et al* conclude that approximately 1.9 million m^3/sec sink and flow out from the Labrador Sea as deep water. This conclusion can be further examined by means of the available cross sections from Labrador to Greenland through which the net flow must be zero. The *Godthaab* section (Kiilerich, 1939) indicates, in August, 1928, an *inflow* of deep water amounting to about 0.5 million m^3/sec, whereas the *General Greene* section in August, 1935, indicates an *outflow* of deep water of 3.6 million m^3/sec, the outflow taking place below a depth of 1000 m. The discrepancy between the results from the two sections shows perhaps that the outflow of deep water is an intermittent process. If this is true the greater emphasis should be given to the above values calculated by Smith *et al*, which are based upon observations from a number of years and according to which the average outflow of deep water from the Labrador Sea is about 2 million m^3/sec. This flow to the south has important bearing on the entire deep-sea circulation of the Atlantic Ocean, as will be shown later on.

ICE AND ICEBERGS IN THE ARCTIC. The Polar Sea with its adjacent seas, the western portion of the Norwegian Sea, Baffin Bay, and the western portion of the Labrador Sea are covered by sea ice during the greater part of the year, whereas, under the influence of the warm Atlantic water, the west coast of Norway is always ice-free except for occasional freezing over of the inner parts of fjords.

The Arctic pack ice that covers most areas is more broken and piled up than the Antarctic pack ice (p. 623). Large, flat ice floes are rare, but fields of hummocked ice and pressure ridges rising up to five or six meters above the general level of the ice are frequently found. The broken-up and rugged appearance of the Arctic pack ice is ascribed to the action of the wind in conjunction with the restricted freedom of motion of the ice owing to land barriers on all sides.

The wind drift of the ice has been discussed by Nansen (1902) and Sverdrup (1928), who found that the direction of the drift deviated, on an average, 28° and 33°, respectively, from the direction of the wind, instead of 45° as required by Ekman's theory of wind currents. The discrepancy is due to the resistance against motion offered by the ice itself (Sverdrup, 1928, Rossby and Montgomery, 1935) and because this resistance is greatest at the end of the winter when the ice is most closely packed, the deviation of the ice drift from the wind direction is smallest at that time of the year. Similarly the wind factor, that is, the ratio between the velocity of the ice drift and the wind velocity, is smaller at the end of the winter than in summer, varying between 1.4×10^{-2} in April to 2.4×10^{-2} in September.

Under the influence of variable winds the ice is, in all seasons of the year, torn apart in some localities, where lanes of open water are formed, and is packed together in others. In winter the lanes are rapidly covered by young ice, which in a week or less may reach a thickness of 50 cm; but from the end of June to the middle of August no freezing takes place, because over the entire Polar Sea the air temperature remains at zero degrees or a little above zero. Where lanes are formed, the ice always breaks along a jagged line, and when the ice fields move apart they also are displaced laterally. In summer such lanes remain ice-free. When they close, as the ice is packed together, the two sides of the lane do not fit; corners meet corners, and between the corners openings of different shapes remain, many of them several hundred meters long. In summer the Arctic ice fields are therefore not continuous, but are honeycombed to such an extent that from no point can one advance as much as 10 km in any direction without striking a large opening in the ice. This characteristic makes possible the use of a submarine for exploration of the Polar Sea, as advocated by Sir Hubert Wilkins.

In winter the average thickness of the Arctic pack ice is probably 3 to 4 m and in summer 2 to 3 m, but under pressure ridges and great hummocks the thickness is greater. Hummocked ice is often found stranded where the depth is 8 to 9 m, and in exceptional localities where strong tidal currents prevail, masses of piled-up ice have been found stranded where the depth was 20 m.

The ice passes through a regular annual cycle. In summer melting takes place during 2 to 3 months, and on an average the upper 1 m of the ice melts. In winter ice forms on the underside of the floes, but the thicker the ice the slower the freezing. The average thickness of the ice depends mainly upon the rapidity of melting in summer and freezing in winter, and is therefore determined by climatic factors. Near shore, river water and warm offshore winds facilitate melting and the development of navigable lanes of open water along the coasts. In recent years the U.S.S.R. has been able to take advantage of such lanes along the north coast of Siberia for establishing shipping connections with the large Siberian rivers. Aerial surveys of ice conditions have preceded the operations and ice breakers have been used where necessary.

Within the greater part of the Polar Sea the ice is moving mainly under the action of the winds, but it is also carried slowly by currents toward the opening between Spitsbergen and Greenland. The speed of the current increases when approaching the opening, and great masses of ice are carried swiftly south by the East Greenland Current. Since 1894 detailed information as to the extent of the sea ice in the Norwegian Sea and in the Barents Sea has been annually compiled and published by the Danish Meteorological Institute.

The icebergs in the Arctic originate from glaciers, particularly on Northern Land, Franz Josef Land, Spitsbergen, and Greenland. No icebergs are encountered in the Polar Sea except near Northern Land, Franz Josef Land, and Spitsbergen because the Greenland glaciers do not terminate in the Polar Sea. The glaciers on the first three islands are small and produce only small icebergs. By far the greater number, and all large icebergs, originate from the Greenland glaciers and are carried south by the East Greenland and Labrador Currents. These icebergs are generally of irregular shape because the Greenland glaciers do not form thick shelf ice comparable to that of the Antarctic, but terminate in fjords where piece after piece breaks off as the glacier advances. Some fjords are closed by shallow sills on which the bergs strand.

Many of the icebergs carried south by the East Greenland Current disintegrate before they reach Cape Farewell, the southern cape of Greenland, but others are carried around the south end of Greenland and continue to the north along the west coast. They are joined by other icebergs from the West Greenland glaciers and together with these are finally carried south by the Labrador Current. The icebergs reach farthest south off the Grand Banks of Newfoundland in the months from March to June or July, when they represent a serious menace to shipping. After the *Titanic* disaster in 1911, the International Ice Patrol was established in order to safeguard the shipping by reporting icebergs and predicting their probable course. The ice patrol is conducted by the United States Coast Guard, the publications of which contain a large amount of information as to the number, distribution, and drift of icebergs in the region of the Grand Banks. Prediction of the ice drift has been based successfully on currents computed from the distribution of density as observed on cruises during which several lines of oceanographic stations have been occupied in about two weeks. This work of the Coast Guard is the outstanding example of practical application of the methods for computing ocean currents (p. 453).

The North Atlantic Ocean

Water Masses of the North Atlantic Ocean. The character of the water masses in the North Atlantic Ocean is illustrated in the T-S diagram in fig. 183. Ten stations scattered over the North Atlantic have been selected, the locations of which are shown in the inset map. Observations from the upper 100 m have been omitted. All stations are located at such distances from the coast that local waters are not considered.

The striking features of the presentation are that in the North Atlantic Ocean one has to deal principally with two typical water masses: the North Atlantic Central Water, characterized by a nearly straight T-S curve between the points $T = 8°$, $S = 35.10$ ‰, and $T = 19°$, $S = 36.70$ ‰; and a deep and bottom water, which is characterized by temperatures between 3.5° and 2.2°, and salinities between 34.97 ‰ and 34.90 ‰. Between these two typical water masses are found other water masses, most of which have not been formed in the North Atlantic Ocean but which exercise a considerable influence upon the distribution of temperature and salinity at mid-depths.

The North Atlantic Central Water evidently spreads over a very large area, because it is typically present at all stations shown in the map except at station G 1990 to the south of Cape Farewell on Greenland, and at station A 1175 in latitude 6°50′N. At station AH 14a, to the

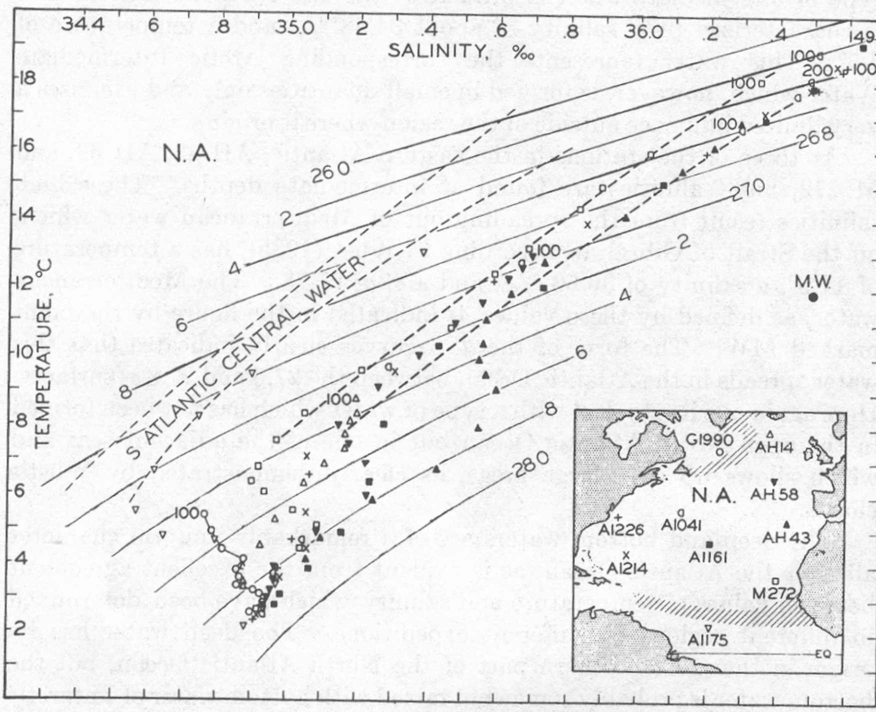

Fig. 183. Temperature-salinity relation in the North Atlantic Ocean. The depths of the shallowest values are indicated. Large squares represent winter surface values in the northwestern Atlantic. Inset map shows location of stations. *Abbreviations:* A = *Atlantis*, AH = *Armauer Hansen*, G = *General Greene*, M = *Meteor*, N.A. = North Atlantic Central Region, M.W. = Mediterranean Water.

south of Iceland, North Atlantic Central Water is found although no water occurs of a temperature above 9°. There is a remarkable agreement between the upper water at such widely separated stations as AH 58 to the west of the Bay of Biscay and A 1226 in the Gulf Stream region to the east of Cape Hatteras. At station A 1175 water of a different character is present, namely upper water which originates in the South Atlantic Ocean and has been carried across the Equator, but this water does not appear to exercise any great influence on the large body of North Atlantic water. The water at station G 1990 to the south of

Cape Farewell corresponds to the subantarctic water, but the subarctic water in the North Atlantic is evidently confined to a small region.

At intermediate depths, three other types can be pointed out. At station A 1175 Antarctic Intermediate Water is present which has its origin in the South Atlantic Ocean. Upon entering the North Atlantic, this water mixes with the North Atlantic masses and appears to exercise an influence on water which has a σ_t of approximately 27.4. Another type of intermediate water is present at the two northern stations and is characterized by a salinity of about 34.88 $^0/_{00}$ and a temperature of 3.5°. This water represents the corresponding Arctic Intermediate Water which, however, is formed in small quantities only and exercises a very limited influence outside of the region where it originates.

At three of the stations in the eastern Atlantic, AH 43, AH 58, and M 272, high salinities are found at intermediate depths. These high salinities result from the spreading out of Mediterranean water which, off the Strait of Gibraltar, according to Wüst (1935), has a temperature of 11.9°, a salinity of 36.50 $^0/_{00}$, and a σ_t of 27.78. The Mediterranean water, as defined by these values, is indicated in the figure by the point marked MW. The form of the *T-S* curves clearly indicates that this water spreads in the Atlantic Ocean between the 27.6 and 27.8 σ_t surfaces. Here again one has to deal with a type of water which has not been formed in the open North Atlantic Ocean but in one of the adjacent seas and which shows up over large areas, as clearly demonstrated by Wüst's charts.

The deep and bottom waters are of a remarkably uniform character all over the Atlantic Ocean, as is evident from the excellent agreement between values of temperature and salinity which have been determined in different regions by different expeditions. The deep water has its origin in the most northern part of the North Atlantic Ocean, but the bottom water is probably somewhat mixed with bottom water of antarctic origin. We shall return to that question when discussing the deep-water circulation.

The large body of North Atlantic Central Water must have been formed in the North Atlantic Ocean. Iselin (1939) has shown that this water has probably attained its characteristic temperatures and salinity when in contact with the atmosphere. Making use of Böhnecke's (1936) charts of surface salinity and surface temperature of the North Atlantic Ocean in late winter, Iselin plotted temperature-salinity values from the surface in a diagram, together with typical *T-S* curves in mid-ocean. These surface values, which represent the characteristic *T-S* relation in a horizontal direction, nearly coincide with the vertical *T-S* curves, and by selecting other corresponding values at the surface one could obtain complete coincidence. This agreement supports the conclusion that the central water mass is formed by sinking of surface water and lateral mixing, but, as explained on p. 145, vertical mixing may also have contributed to the development of the water mass. The spread of the *T-S* curves which appear in fig. 183 can be accounted for by assuming that the sinking of surface water took place in different areas where waters of the same density but of slightly different temperatures and salinities occurred, or took place in different years. Some of the differences are regional and can be used with advantage in detailed study of the mixing of water masses.

The thickness of the North Atlantic Central Water is closely related to the character of the currents in the North Atlantic Ocean. Within a region of a strong current the thickness will be great on the right-hand side of the current and small on the left-hand side, according to the general rule that in the Northern Hemisphere light water is present on the right-hand side of a current.

THE CURRENTS OF THE NORTH ATLANTIC OCEAN. The system of currents in the North Atlantic (chart VII) is dominated by the North Equatorial Current to the south and the Gulf Stream system to the north. The North Equatorial Current flows from east to west in the trade-wind region and is fed by the southeasterly currents off the west coast of North Africa. Corresponding to flow from the northwest, water of relatively high density and low temperature is found off the African coast, as is evident from the charts of surface temperatures (charts II and III). The temperature close to the coast is also lowered by upwelling from moderate depths due to the action of prevailing northwesterly winds, but this upwelling does not exercise influence as widespread as does the corresponding upwelling off the coasts of southwest Africa or, particularly, as does that off the west coasts of North and South America. Details as to the upwelling off northwest Africa are still lacking, but the results of the work of the *Meteor* in these waters in 1937 (Defant, 1937a) can be expected to give much interesting information.

The wide North Equatorial Current runs, as already stated, from east to west, but does not follow an absolutely straight course. Schumacher (1940) points out that north of 15°N, on most of the monthly charts of the surface currents of the North Atlantic Ocean which he has prepared by means of ships' observations, the North Equatorial Current bends to the north (to the right) when approaching the mid-Atlantic Ridge and to the south (to the left) after having passed the ridge. Schumacher also draws attention to the fact that on Defant's chart (1936b) of "average currents in the upper layers as derived from the inclination of the discontinuity layer in the troposphere," these bends are much more conspicuous and that they appear reduced at the surface, probably owing to the superimposed pure wind drift. The maximum deflection to the north occurs directly above the mid-Atlantic Ridge and it seems therefore that the bends are due to the crossing of the ridge, although the latter lies more than 3000 m below the sea surface. The character of the deflection is not in agreement with Ekman's theory of the influence of the bottom configuration, but it is in agreement with the type of deflection which would be expected if a deep relative current flows over a ridge (p. 466). The observations suggest, therefore, that in most months of the year the east-west flow of the North Equatorial Current north of 15°N comprises water masses between the surface and a depth of more than 3000 m, and that the influence of the ridge upon the distribution of density extends through all these layers, but the question requires further examination.

In the western part of the Atlantic Ocean the North Equatorial Current joins the branch of the South Equatorial Current which has crossed the Equator and which, according to fig. 183, p. 669, carries characteristically different water masses. Mixing takes place between these water masses and the corresponding North Atlantic water, and the waters in the Caribbean Sea are therefore intermediate in character (table 77, p. 639). Thus, the part of the North Equatorial Current which continues into the Caribbean Sea carries water which is mixed with water of South Atlantic origin, whereas the northern branch of the North Equatorial Current which flows along the northern side of the Great Antilles as the Antilles Current carries water which is identical with that of the Sargasso Sea.

The North Equatorial Current terminates in the current through the Yucatan Channel and the Antilles Current. The continuation of these currents represents the beginning of the Gulf Stream system which dominates the circulation of a great part of the North Atlantic Ocean. Following the nomenclature of Iselin (1936), the term "Gulf Stream System" is used to include the whole northward and eastward flow beginning at the Straits of Florida and including the various branches and whirls found in the eastern North Atlantic that can be traced back to the region south of the Newfoundland Banks. This system can be subdivided into the three following parts:

1. The Florida Current. The northward-moving water from the Straits of Florida to a point off Cape Hatteras where the current ceases to follow the continental slope. The Florida Current can be traced directly back to the Yucatan Channel, as explained on p. 642, because the greater part of the water flowing through this strait continues on the shortest route to the Straits of Florida and a small amount only sweeps into the Gulf of Mexico, later to join the Florida Current. After having passed the Straits of Florida the current is reinforced by the Antilles Current, but the name "Florida Current" is retained as far as to Cape Hatteras.

2. The Gulf Stream. The mid-sector of the system, from the region where the current first leaves the continental slope off Cape Hatteras to

the region to the east of the Grand Banks in about long. 45°W where the stream begins to fork. This application of the name "Gulf Stream" represents a restriction of the popular term, but such a restriction is necessary in order to introduce clear definitions.

3. The North Atlantic Current. The name is used as a general term covering all the easterly and northerly currents of the North Atlantic from the region to the east of the Grand Banks where the Gulf Stream divides. The branches of the North Atlantic Current are often masked by shallow and variable wind-drift surface movements, which have become commonly known as the North Atlantic Drift.

The terminal branches of the Gulf Stream System are not all well known, but among the major ones are the Irminger Current, which flows towards the west to the south of Iceland, and the Norwegian Current, which enters the Norwegian Sea across the Wyville Thomson Ridge and can be traced ultimately into the Polar Sea (p. 661). Other more irregular branches which turn to the south have been examined by Helland-Hansen and Nansen (1926), who find that they terminate in great whirls off the European coast.

THE FLORIDA CURRENT. The energy of the Florida Current appears to be derived directly from the difference in sea level between the Gulf of Mexico and the adjacent Atlantic coast, the observed difference between Cedar Keys and St. Augustine being 19 cm. (p. 642). Assuming that this hydrostatic head accounts for all of the energy, and assuming frictionless flow, Montgomery (1938) finds that the velocity through the Straits of Florida should be 193 cm/sec, which is somewhat higher than the average velocity at the center of the current. The difference in level is probably maintained by the trade winds, and the energy of the Florida Current is therefore derived from the circulation of the atmosphere.

Within the current flowing through the Straits of Florida, the distribution of density must adjust itself in the usual manner; that is, the lighter water must be found on the right-hand side of the current and the denser water on the left-hand side, and the sea surface, instead of coinciding with a level surface, must accordingly rise towards the right-hand side of the current. In the case of the Florida Current this rise amounts to about 45 cm, sea level at the coast of Cuba being about 45 cm higher than at the American mainland (Dietrich, 1936).

The character of the currents in the Straits of Florida was established by the outstanding measurements made in the years 1885 to 1889, by the United States Coast and Geodetic Survey from the survey vessel *Blake*, commanded by J. E. Pillsbury. Pillsbury's observations of currents, carried out from a vessel anchored in deep water in a swift stream, are among the classical data in physical oceanography, not so much because they give complete information as to the average currents, but mainly because they made possible a convincing demonstration of the correctness of the later methods used for computing relative currents (Wüst, 1924). The upper right-hand graph in fig. 184 shows the observed average velocity distribution in a section through the narrowest part of the Straits of Florida between Fowey Rocks, south of Miami, Florida, and Gun Cay, south of Bimini Islands, as plotted by Wüst from Pills-

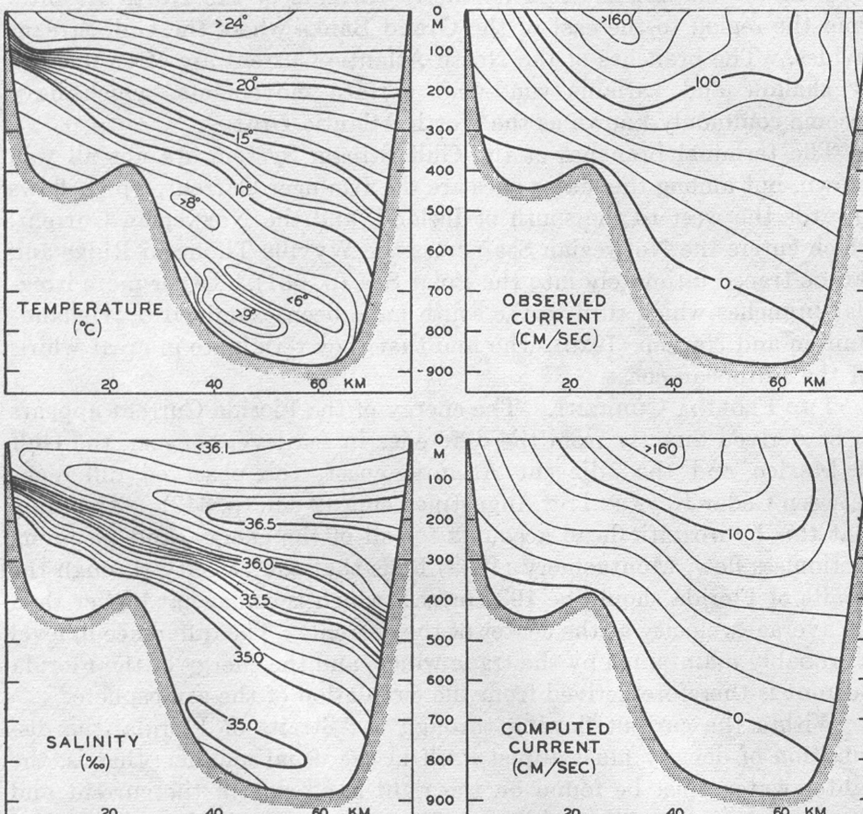

Fig. 184. *Left:* Observed temperatures and salinities in the Straits of Florida. *Right:* Velocities of the current through the Straits according to direct measurements and according to computations based on the distributions of temperature and salinity (after Wüst).

bury's data. To the left are shown the corresponding distributions of temperature and salinity as represented by Wüst on the basis of temperature measurements made by Bartlett from on board the *Blake* in 1878 and published by Agassiz in 1888, and on the basis of temperature and salinity observations from on board the Coast and Geodetic Survey vessel *Bache* in 1914. Wüst has, by means of these data, computed the velocity distribution shown in the lower right-hand graph in fig. 184, which is in remarkable agreement with the observed one. In order to arrive at absolute values of the velocity, Wüst had to assume a known velocity at some depth; on the basis of the distribution of temperature and salinity he assumed an inclined surface of no motion at some distance from the bottom, as shown by the curves marked 0. This curve nearly coincides with the curve of zero velocity as derived from Pillsbury's measurements. A complete correspondence is thus found between observed and computed currents; this single example, therefore, has greatly contributed to increasing confidence in the correctness of computed relative currents in general.

On the basis of measurements and computations Wüst finds that the average transport of water through the Straits of Florida is 26 million m^3/sec. The transport probably shows an annual variation and may differ from year to year, but so far little is known about such fluctuations (Montgomery, 1938, 1941, Iselin, 1940).

In its further course the Florida Current closely follows the continental slope, flowing most swiftly directly along the slope. The shallow coastal waters to the left of the Florida Current remain more or less at rest; often the transition from these waters to the blue waters of the Florida Current is so abrupt that the border of the Florida Current can be seen as a line stretching from horizon to horizon. After emerging from the Straits of Florida, the current is soon joined by the Antilles Current which, according to Wüst (1924), carries about 12 million m^3/sec. Owing to the moderate depth to the bottom, the current remains relatively shallow, not more than about 800 m deep, and carries no water colder than 6.5° until it leaves the Blake Plateau in about lat. 33°N. According to Iselin (1936) the current increases steadily in volume by absorption of Sargasso Sea water, and as it leaves the Blake Plateau both the depth and the volume suddenly increase as a result of the joining in of water of a temperature considerably below 8° which comes from the southwestern Sargasso Sea.

THE GULF STREAM. The middle portion of the Gulf Stream System, for which the name Gulf Stream is retained, continues as a well-defined and relatively narrow current which, in contrast to the Florida Current, flows at some distance beyond the continental shelf. To the right of the current is the Sargasso Sea water, as previously, but to the left are now found two water masses, the coastal water which covers the shallow shelf areas and the slope water which, at temperatures between 4° and 10°, is very similar to the Gulf Stream water (Iselin, 1936) but at higher temperatures is of lower salinity. Within the upper layers of the slope water great seasonal variations in temperature and salinity occur and, in addition, eddies of Gulf Stream water occasionally intrude.

The surface velocities of the Gulf Stream are very high, the computed values reaching, in lat. 36°N, long. 73°W, more than 120 cm/sec (Iselin, 1936) and in lat. 38°N, long. 69°W, 140 cm/sec (Seiwell, 1939). On the assumption of no motion at a depth of 2000 m where the isosteres are nearly horizontal, the volume transport of the Gulf Stream off Chesapeake Bay is between 74 and 93 million m^3/sec (Dietrich, 1937a, Iselin, 1936 and 1940), and off Woods Hole it is about 72 million m^3/sec (Seiwell, 1939). If these figures are correct, they indicate that between 38 and 57 million m^3/sec of Sargasso Sea water and deep water have been added to the Florida-Gulf Stream after the Antilles Current, carrying 12 million m^3/sec, joined the flow of 26 million m^3/sec through the Straits of Florida. Similarly, between 34 and 53 million m^3/sec would have to be discharged towards the south from the Gulf Stream between Chesapeake Bay and long. 45°W, off the "tail" of the Grand Banks where, according to Soule (1939) the transport of thc Gulf Stream is somewhat less than 40 million m^3/sec. These conclusions are not supported by observations between the line Chesapeake Bay-Bermuda and the Bahamas, or between Bermuda and long. 45°W. The available data indicate that the inflow north of the Antilles Current does not exceed 15 to 20 million m^3/sec and between Bermuda and long. 45°W the southward flow of Gulf Stream water does not exceed 15 million m^3/sec. The computed transport can, however, be interpreted differently.

The dynamics of the Florida Current and the Gulf Stream, particularly the downstream increase in volume as far as Cape Hatteras, is not clearly understood. Rossby (1936) has compared the Florida Current and its continuation, the Gulf Stream, to a wake stream which emerges from the Straits of Florida, and has examined the effect on such a stream of stresses due to lateral mixing. In a wake stream in homogeneous water the momentum transport remains constant whereas the volume (mass) transport increases downstream, the increase being due to inflow from the sides. Expanding the theory to a stratified medium, Rossby finds that a "compensation current" in the direction of flow must develop on the right-hand side of the wake stream, whereas to the left a countercurrent in the opposite direction must appear, and this picture agrees in general with the pattern of the Gulf Stream and its surroundings. Another important aspect of the theory is that owing to the lateral stresses a transverse circulation should develop, water being absorbed from the oceanic areas to the right of the current and discharged into the countercurrent to the left. Such a mechanism would account for the presence of eddies of Gulf Stream water in the slope current, but Defant (1937b) and Ekman (1939) have warned against immediate acceptance of the theory because several of the necessary assumptions appear not to be fulfilled. Regardless of whether the theory is confirmed or disproven, it has been greatly stimulating, particularly because of its emphasis on the importance of lateral mixing. In this connection the possibility may be mentioned that the developments of the countercurrent and the transverse circulation may not necessarily be associated with a wake stream, but may be related to lateral mixing which takes place regardless of the character of the current, or to cooling or heating when the current flows in a north-south direction. These possibilities should be examined because countercurrents and transverse circulations appear to develop wherever a current flows parallel to a coast (Kuroshio, Peru Current, California Current).

A satisfactory theory of the Gulf Stream must not only account for the increase in volume transport in the direction of flow and the fact that this increase takes place without evidence of strong inflow from the southeast, but it also must account for another important feature which

TABLE 81

AVERAGE SEA LEVEL ALONG THE NORTH AMERICAN EAST COAST REFERRED TO SEA LEVEL AT THE COAST OF FLORIDA-GEORGIA

Locality	Sea level (cm) Avers, 1927	Sea level (cm) Rappleye, 1932	Mean	Distance along coast (km)	Slope
St. Augustine, Fla. / Fernandina, Fla. / Brunswick, Ga.	0	0	0	0	
					6×10^{-8}
Norfolk, Va.	4	7	6	1000	
					35×10^{-8}
Cape May, N.J. / Atlantic City, N.J. / Fort Hamilton, N.J.	16	24	20	1400	
					13×10^{-8}
Boston, Mass. / Portland, Me.	25	30	28	2000	
					12×10^{-8}
Halifax, Nova Scotia	—	35	35	2600	

has been given considerable attention (Dietrich, 1937b) without having been explained satisfactorily (Ekman, 1939). Precise leveling along the American east coast shows that the mean sea level increases towards the north from St. Augustine, Florida, to Halifax, Nova Scotia, the most conspicuous increase taking place directly north of Cape Hatteras. Table 81 summarizes the results of the precise leveling, according to Avers (1927) and Rappleye (1932). In the table the sea level along the coast has been referred to that on the coast of Florida and southern Georgia, values from stations less than 200 km apart having been combined as averages. The distances along the coast from Florida–Georgia are entered and also the values of the slope of the sea surface. These values are of the same order of magnitude as those derived from data in the Caribbean Sea where Parr (1937) computed a slope of the sea surface of 17×10^{-8} and Sverdrup (1939b) a slope of 12×10^{-8}.

If this upward slope were due to the distribution of mass in the ocean, the average density of the Gulf Stream water off the continental shelf would have to decrease in the direction of flow and the Gulf Stream would have to flow uphill. Dietrich (1937b), however, has shown that along the continental slope the distribution of density does not indicate any rise of the sea surface if the topography of the surface is referred to the oxygen minimum layer, and the same is true if the topography of the sea surface is referred to the 2000-decibar surface. These conclusions are not altered by taking into consideration the small effect of differences in atmospheric pressure. A discrepancy therefore exists between the results of precise leveling and the results of what Dietrich calls oceanographic leveling. It is not surprising that such discrepancies appear because, as explained on p. 407, oceanographic observations can give information only as to the topography of the sea surface *relative* to some selected surface in the ocean and information as to the *absolute* topography of the sea surface must be derived from precise leveling along the coasts.

The question now arises whether it is possible to reconcile the different observations and at the same time arrive at transport values which are compatible with the distribution of density. The results of precise leveling are so accurate that the upward slope of the sea surface north of Cape Hatteras must be taken as established. It is also established that this slope of the sea surface is not compensated for by the distribution of mass, as is the case in the Caribbean Sea, for which reason it follows that the water must actually be piled up against the coast. The lack of compensation is understood if one considers that the piling up takes place in the shallow waters along the coast and does not extend to any distance beyond the continental slope. However, a transition must exist from the region of uncompensated piling up of mass to the region where compensation can take place, and it appears reasonable to assume that the transition takes place along the edge of the continental shelf. There the sea level sinks to the value indicated by the oceanographic data and, consequently, a current to the south runs along the continental slope, following approximately the absolute contour lines of the surface. A current to the south must also flow over the shallow portion of the shelf where it flows downhill and where the balance of forces is maintained by the effect of friction.

The transports can be calculated by considering that, owing to the piling up of water along the coast, the actual profiles of the isobaric surfaces slope downwards from the coast towards the outer part of the slope water and then rise because of the distribution of density. If this consideration is correct, the computed transport of Gulf Stream water must come out too high if based on the oceanographic data alone because these give no indication of the southward transport of the slope water along the continent.

This reasoning can be illustrated by certain computations based on the *Atlantis* section of April 17–23, 1932, using the stations 1231–1225 which have been used extensively by Iselin and Dietrich. It is assumed here that over the shelf the piling up of water leads to the sea level on the coast being 10 cm higher, that the same piling up effect is present at the border of the continental shelf, and that no effect is found beyond a distance of 40 km from the border. Furthermore, it is assumed that beyond that distance the 2000-m level is a level of no motion. On the basis of such assumptions one arrives at the following figures:

Transport to the southwest of slope water inside of station 1229	19 million m^3/sec.
Transport to the northeast of slope and Gulf Stream water outside of station 1229	74 million m^3/sec.
Net transport to the northeast	55 million m^3/sec.

According to this computation the net transport of the Gulf Stream is of 55 against 74 million m^3/sec if the southward motion of the slope water is disregarded. A net transport of about 55 million m^3/sec is in agreement with the previous conclusion (p. 676) that not more than 15 to 20 million m^3/sec circulate in the gyral to the right of the Gulf Stream; and if the reasoning is correct, another gyral is present on the left-hand side within which approximately 20 million m^3/sec circulate. The slope water on the coastal side of the Gulf Stream is essentially of the same character as the Gulf Stream water, but has a slightly lower salinity owing to admixture with coastal water. The southward motion directly off the continental shelf is not associated with high surface velocities because one has to deal with a "slope current," and on the assumptions made the velocity of the current is about 10 cm/sec.

Figure 185 gives a schematic picture of the conditions which have been described. The lines provided with arrows represent transport lines, each line corresponding to a transport of about 10 million m^3/sec. The two gyrals on both sides of the Gulf Stream are shown and, in addition, an eddy is indicated between the slope-water gyral and the Gulf Stream where the latter is at a greater distance from the coast. Such an eddy, rotating clockwise, has repeatedly been observed to the south of Nova Scotia (Iselin, 1936), but the clockwise rotation has been difficult to explain because the eddy was considered an offshoot of the Gulf Stream. The direction of rotation is better understood, on the other hand, if the eddy is considered as a part of the slope-water gyral. The inset diagram shows a profile of the sea surface along the line *A-B*. The line marked 1 represents the profile which is obtained from oceanographic observations and the line marked 2 is the profile which is assumed to exist on the basis of these observations and the results of precise leveling.

It remains to account for the piling up of the water masses on the continental shelf. It can be suggested that this piling up is maintained by the prevailing winds over a large area of the North Atlantic Ocean. Southwesterly winds over the northern parts of the North Atlantic may maintain a higher sea level along the northern borders of the ocean and, consequently, a slope of the sea surface along the eastern and western borders.

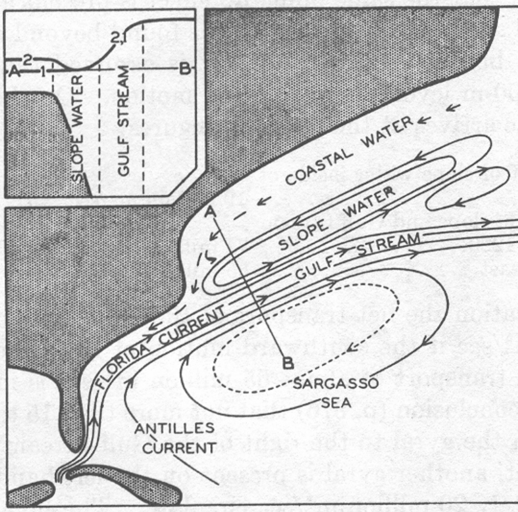

Fig. 185. Schematic representation of the character of the Gulf Stream, taking results of precise leveling into account. *Inset:* Profiles of the sea surface along the line *A-B*. Profile 1 derived from oceanographic data only; Profile 2, from these data and the results of precise leveling.

The North Atlantic Current. The North Atlantic Current represents the continuation of the Gulf Stream after it leaves the region to the east of the "tail" of the Grand Banks. Beyond this region the Gulf Stream loses its characteristics as a well-defined current and divides into branches that are often separated by countercurrents or eddies. Some of the branches turn south but others continue towards the east across the mid-Atlantic Ridge, being flanked on the northern side by waters of the Labrador Current that have been mixed with Gulf Stream water.

The contrast between the Gulf Stream and the North Atlantic Current to the north of the Azores is illustrated in fig. 186, which shows two temperature profiles on the same scale. To the left in the figure are shown the isotherms in a vertical section from Chesapeake Bay towards Bermuda according to observations at the *Atlantis* stations 1231–1226. The Gulf Stream is here concentrated within the narrow band in which the isotherms slope steeply downward toward the right. The section to the right in the figure runs north-northwest from the Azores to lat. 48°N and is based on the observations on board the *Altair* during the International Gulf Stream Expedition in 1938 (Defant and Helland-Hansen, 1939). In this section the isotherms generally slope downward toward the south, indicating a flow towards the east, but the slope is not uniform and countercurrents or eddies are present between the east-flowing branches of the current. In the section is shown the location of the *Altair* cone, discovered during the expedition, which probably represents a submarine volcano rising from depths of about 3500 m to 980 m (the section does not cross the cone at its highest point). This cone appears to exercise an appreciable influence on the hydrographic conditions. The effect on the temperature distribution is seen in fig. 186 and in the salinity distribution a similar disturbance appears even more conspicuously. The resulting distribution of density indicates a counter-clockwise eddy above the cone, extending to a depth of at least 1500 m. The presence of such an eddy, which appears clearly both at the surface (Neumann, 1940) and at a depth of about 100 m (Wüst, 1940), is quite in agreement with the considerations as to the effect of bottom topography upon relative currents which was set forth on p. 466.

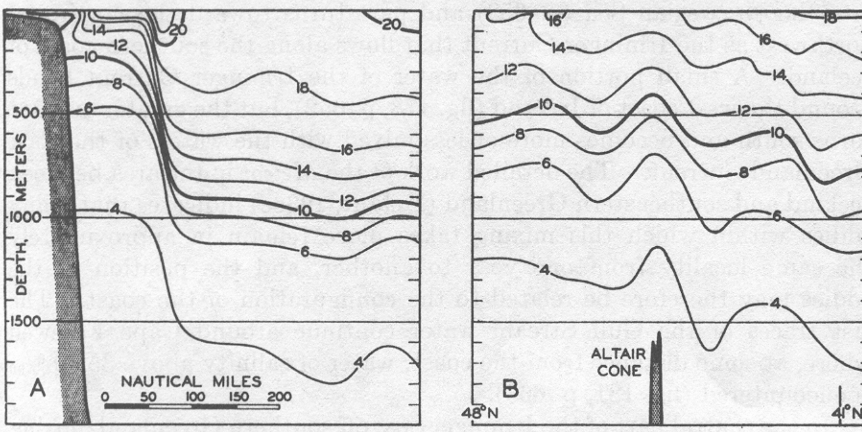

Fig. 186. Temperature profiles across the Gulf Stream off Chesapeake Bay and across the North Atlantic Current to the north of the Azores.

The intensive work conducted by the International Gulf Stream Expedition of 1938 (Defant and Helland-Hansen, 1939) clearly shows the complicated details of the oceanographic conditions. Between June 1 and 22, 1938, the German vessel *Altair* and the Norwegian vessel *Armauer Hansen* occupied 159 stations in an area of less than 100,000 square miles, and at one station the *Altair* anchored and made hourly observations of temperature, salinity, and currents at a number of depths between the surface and 800 m for a period of 90 hours. The dense network of stations showed even greater irregularities than one might expect. At 600 m, for example, the temperature varied between approximately 7° and 13°, and differences up to 5° were observed at distances less than 40 miles. Some of the observed features may be due to the influence of the bottom topography and others may be related to traveling disturbances. Regardless of how the features are interpreted, they do show that caution has to be exercised when drawing conclusions from a few scattered observations because such scattered data may not be representative. They also show that an intensive mixing takes place in mid-ocean.

A series of stations which were occupied in 1931 by the *Atlantis* along the mid-Atlantic Ridge in long. 30°W demonstrates that in spite of irregularities one can distinguish between two major branches of the North Atlantic Current. The northern branch flows between lat. 50° and 52°N, at the boundary between the water of the Gulf Stream System and the Subarctic Water, and carries water which represents Gulf Stream water mixed with waters of the Labrador Current. The other branch flows approximately in lat. 45°N and carries undiluted Gulf Stream water. The northern branch continues mainly towards the east-northeast and divides up. Part of the water flows across the Wyville Thomson Ridge into the Norwegian Sea (p. 653) and part turns toward the north and northwest as the Irminger Current that flows along the southern coast of Iceland. A small portion of the water of the Irminger Current bends around the west coast of Iceland (fig. 178, p. 652), but the greater amount turns south and becomes more or less mixed with the waters of the East Greenland Current. The detailed work of the *Meteor* in the area between Iceland and southeastern Greenland (Defant, 1936c) indicates that many eddies within which this mixing takes place remain in approximately the same locality from one year to another, and the position of the eddies may therefore be related to the configuration of the coast. The last traces of the Gulf Stream water continue around Cape Farewell where, at some distance from the coast, water of salinity above 35.00 ‰ is encountered (fig. 181, p. 663).

In the central part of the Irminger Sea, off southern Greenland, mixing between the North Atlantic water and the Labrador Sea water leads to the formation of Subarctic Water of uniform salinity close to 34.95 ‰ and a temperature which from a depth of a few hundred meters and downwards is nearly 3°C. When the surface layers are cooled in winter to temperatures below 3°, vertical convection currents develop, reaching from the surface to the bottom and leading to renewal of the deep water. Table 82 contains data from two *Meteor* stations which were occupied in early spring and at which nearly uniform water was present. The oxygen observations have been included in order to show the high oxygen content of the deep water. The intense mixing that takes place in winter in this area and in the Labrador Sea (p. 664) must have important bearing on the productivity of the subarctic waters.

The branch of the North Atlantic Current that crosses the mid-Atlantic Ridge in approximately lat. 45°N turns to the right and continues as an irregular flow toward the south between the Azores and Spain, sending whirls into the Bay of Biscay. No distinct currents exist in this region, as is evident from the discussions of the conditions in the eastern North Atlantic by Helland-Hansen and Nansen (1926), but a diffuse transport of water toward the south takes place. Some of this water enters the Mediterranean as a surface current and flows out again across the sill in the Strait of Gibraltar as water of very high salinity which spreads at intermediate depths, and influences conditions of the greater part of the North and South Atlantic Oceans (p. 670). The larger amount of the upper water masses continues toward the south and finally joins the North Equatorial Current.

TABLE 82

HYDROGRAPHIC CONDITIONS IN THE IRMINGER SEA IN EARY SPRING

Depth (m)	*Meteor* 121, March 9, 1935 Lat. 56°37′N, Long. 44°54.5′W					*Meteor* 79, March 30, 1933 Lat. 59°38′N, Long. 40°42.5′W				
	Temp. (°C)	S (°/₀₀)	σ_t	O_2 (ml/L)	O_2 (%)	Temp. (°C)	S (°/₀₀)	σ_t	O_2 (ml/L)	O_2 (%)
0......	2.82	34.87	27.80	7.24	96	4.07	34.96	27.76	6.92	95
50......	3.01	34.90	27.81	7.26	97	4.07	34.97	27.77	6.99	96
100......	3.09	34.92	27.82	7.24	97	4.06	34.96	27.77	6.81	94
200......	3.17	34.93	27.82	6.70	90	3.98	34.97	27.77	6.84	94
800......	3.26	34.96	27.83	6.98	94	3.76	34.95	27.77	6.60	90
1000......	3.20	34.95	27.83	6.96	93	3.33	34.89	27.77	6.64	89
1500......	3.17	34.93	27.82	6.99	94	3.28	34.94	27.82	6.39	86
2000......	3.23	34.93	27.82	——	—	2.84	34.96	27.88	6.37	87

TRANSPORT. On the basis of the above discussion and numerous computations of transports, fig. 187 has been prepared, giving a schematic picture of the volume transport of the currents which carry North Atlantic Central Water or Subarctic Water. No lines of equal transport are entered but the different branches of the current system are indicated and the approximate volume transport in millions of cubic meters per second is stated. The presentation has been derived by computing the transport of water of temperature higher than 7° between a number of selected stations north of lat. 20°N, and adjusting the figures in order to take the continuity of the system into account. To the northwest of a line which can be drawn roughly between the Straits of Florida and the English Channel it has been assumed that the 2000-decibar surface could be taken as a surface of no motion, but to the southeast of that line it has been assumed in general that the 7° isothermal surface was a surface of no motion. This treatment was necessary because the condition of continuity has to be satisfied and because a reversal of the direction of flow appears to take place below the 7° isothermal surface over large parts of the southeastern area. It is not claimed that the picture in fig. 187 is accurate in details, but it is believed that it gives an approximately correct idea of the circulation of the upper water within the greater part of the Atlantic Ocean.

The selected surface of no motion has several features in common with the corresponding surface which Defant (1941) arrived at by using the third method discussed on p. 457, without taking the equation of continuity into account. Defant's results as to the prevailing currents differ somewhat, however, from ours, particularly along the north coasts of the Antilles where he finds currents toward the southeast dominating, whereas according to our analysis the currents are directed toward the northwest in agreement with other generally accepted results.

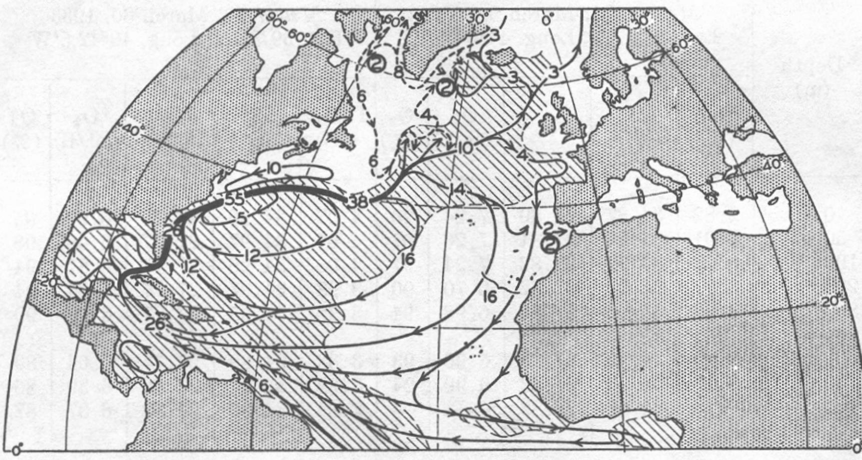

Fig. 187. Transport of Central Water and Subarctic Water in the Atlantic Ocean. The lines with arrows indicate the direction of the transport, and the inserted numbers indicate the transported volumes in millions of cubic meters per second. Full-drawn lines show warm currents, dashed lines show cold currents. Areas of positive temperature anomaly are shaded.

The representation in fig. 187 shows the Florida Current and the Gulf Stream as the only well-defined currents of the Atlantic Ocean. It also shows that the greater amount of the waters of the Gulf Stream turns south before reaching the Azores and circulates around the Sargasso Sea.

In the figure the areas are shaded in which the average surface temperature is higher than the general average for that latitude. The shaded portions therefore represent the areas with positive temperature anomalies as referred to the average temperatures of the Atlantic Ocean, and the unshaded portions represent the areas of negative temperature anomalies. As should be expected, water which is transported from lower to higher latitudes is relatively warm, whereas water which is transported from higher to lower latitudes is relatively cold. If the velocity distribution within the different branches of the currents were known it would be possible to compute the net amounts of heat which are carried by the ocean currents across parallels of latitude and also to compute the amounts of heat given off or taken up within different areas, but so far no such calculations have been attempted.

Another characteristic of the system of currents is brought out by a comparison between the transports and the temperatures off Chesapeake Bay and to the north of the Azores (fig. 186). The Gulf Stream off Chesapeake Bay transports large volumes of water of temperatures above 16° but the North Atlantic Current to the north of the Azores carries only small amounts of water as warm as 16°. This apparent reduction in temperature must be due to the fact that the warmer waters of the

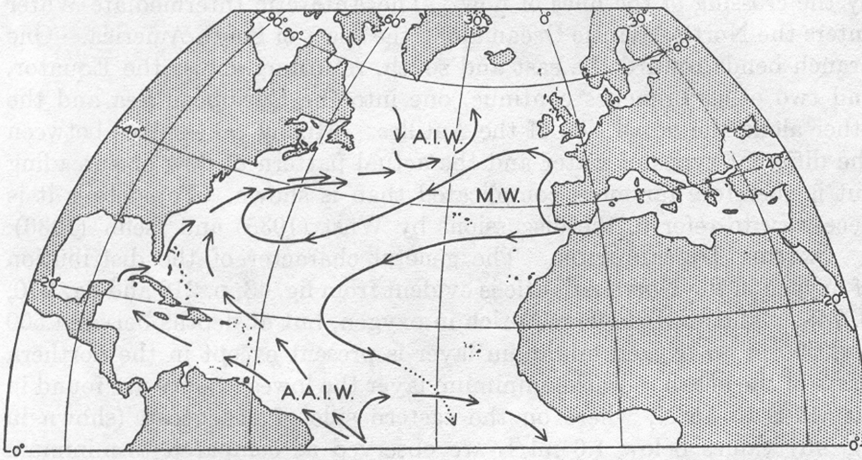

Fig. 188. Approximate directions of flow of the intermediate water masses of the North Atlantic. A.I.W., Arctic Intermediate Water; M.W., Mediterranean Water; A.A.I.W., Antarctic Intermediate Water.

upper layers have been carried south before reaching the Azores, whereas the somewhat colder waters at greater depths have continued toward the east. If such is the case, the law of the parallel solenoids is not fulfilled, owing possibly to cooling in the direction of flow (Parr, 1936).

As previously mentioned (p. 629), 6 million m³/sec of South Atlantic Upper Water enter the North Atlantic Ocean along the coast of South America. Correspondingly, 6 million m³/sec of North Atlantic water sink in different localities and return to the South Atlantic as a deep-water flow. The three regions where sinking takes place and the amounts of water sinking from the surface are shown by circles and inserted numbers that represent rounded-off values. About 2 million m³/sec sink outside the Strait of Gibraltar and about the same amount sinks in the Labrador Sea. Both of these values are based on fairly accurate data (pp. 647 and 666), wherefore it follows that a similar amount sinks in the third region within which bottom and deep water is being renewed, namely the region to the southeast of southern Greenland.

The direction of flow of the different types of intermediate waters of lower temperature does not always coincide with the direction of flow of the upper water. On the basis of the character of the water and the results of dynamic calculations, fig. 188 has been prepared, giving a tentative picture of the spreading of the Arctic Intermediate Water, the Mediterranean and the Antarctic Intermediate Water. The Arctic Intermediate Water is present in the northern region only. The Mediterranean water partly bends north before turning west and partly spreads directly toward the west. Some of this water turns south and continues across the Equator below the Antarctic Intermediate Water, as indicated by the crossing of the lines of flow. The Antarctic Intermediate Water enters the North Atlantic Ocean along the coast of South America. One branch bends toward the east and south, returning across the Equator, and two other branches continue, one into the Caribbean Sea and the other along the north side of the Antilles. Mixing takes place between the different types of water and the actual pattern of flow or spreading out is therefore far more complicated than is shown. For details it is necessary to refer to the discussions by Wüst (1935) and Iselin (1936).

Oxygen Distribution. The general character of the distribution of oxygen in the North Atlantic is evident from fig. 43, p. 210 and fig. 210, p. 748. The upper layers are rich in oxygen, but at depths between 500 and 900 m an oxygen minimum layer is present except in the northern parts of the ocean. In the minimum layer the lowest values are found in lat. 15°S to 15°N, where on the eastern side of the ocean (shown in fig. 43) values below 1.0 ml/L are observed as compared to minimum values of about 3.0 ml/L on the western side (fig. 210). To the north of 30°N the contrast between the two sides of the ocean is reversed and the lowest values are found on the western side, although they do not drop much below 4.0 ml/L.

Below the oxygen-minimum layer the oxygen content increases rapidly with depth; the deep water of the North Atlantic contains very large amounts of oxygen which decrease somewhat, however, from north to south. In the region where the North Atlantic Deep Water is formed, the oxygen content is at all depths higher than 6.5 ml/L (see table 82, p. 683) and from that area the oxygen content gradually decreases to 5.5 ml/L at the Equator and less than 5 ml/L in 45°S. In the same distance the percentage saturation decreases from about 85 to 75 or 70. The southern part of the Weddell Sea is the only other oceanic area in which water at depths below 2000 m contains as much oxygen as does the deep water of the North Atlantic Ocean.

Adjacent Seas of the Indian Ocean

The Red Sea. Before discussing the water masses and currents of the Indian Ocean it is of advantage to deal with the one important adjacent sea of the Indian Ocean, the Red Sea, the waters of which exercise an influence similar to that of the Mediterranean water in the Atlantic, but less widespread.

Extending between lat. 12°N and 30°N, the Red Sea fills a long and narrow basin which at the northern end is closed except for the communication through the Suez Canal and at the southern end is separated from the Gulf of Aden by a shallow sill. The total length of the Red Sea is nearly 1800 km and the width 270 km. Outside the shallow and reef-bound coastal waters the general depth is about 700 m, but the bottom is very irregular, and apparently isolated depressions exceeding 2000 m occur in several places. The sill at the southern entrance of the Red Sea lies 140 km inside of the narrow Strait of Bab el Mandeb off Hanish Island, where the greatest depth is only about 100 m. The fairly complete sonic soundings of the John Murray Expedition have failed to reveal any deeper channel (Thompson, 1939b).

The Red Sea is located in a region which is characterized by such an arid climate that evaporation from the water surface greatly exceeds the small precipitation. There is no runoff because no rivers enter the Red Sea. Along the entire length the prevailing winds blow consistently from the north-northwest during half of the year, from May to September, but during the other half of the year, October to April, the north-northwest winds reach only as far south as lat. 22° or 21°N, and south of 20°N the wind direction is reversed, south-southeast winds dominating.

The climatic conditions and the prevailing winds determine the character of the waters in the Red Sea and the exchange of water between the Red Sea and the Gulf of Aden. Owing to the excessive evaporation the surface salinity of the water in the northern part of the Red Sea reaches values between 40 $^0/_{00}$ and 41 $^0/_{00}$. In summer the temperature of the surface water is very high, mostly exceeding 30°, but in winter the temperature is decreased, particularly in the northern end where the average temperature in February is as low as 18°. The lowering of the temperature in winter, together with the intense evaporation in that season, leads to the formation of deep water that fills the entire basin of the Red Sea below the sill depth and has a salinity between 40.5 $^0/_{00}$ and 41.00 $^0/_{00}$ and a temperature between 21.5° and 22°. The formation of this deep water is further facilitated by the character of the currents, which is related to the character of the prevailing winds. According to Barlow (1934), who has examined 6100 direct observations of currents, averaging them by months for the areas 12° to 20°N and 20° to 28°N, the surface current flows toward the north-northwest (into the Red Sea) from November to March and to the south-southeast (out of the Red Sea) from June to September, with transition stages in April, May, and October. During November to March, when the current flows in, the velocity is less in the northern half of the sea where the flow is directed against the wind than it is in the southern half. For this reason a convergence develops which facilitates the formation of the deep water. Superimposed upon the longitudinal flow are cross currents which probably are due to irregular eddies. Thompson (1939a) points out that the prevailing winds must bring about a transverse circulation which, in summer when the north-northwest winds prevail, must lead to upwelling along the Arabian coast and piling up of surface water along the African coast and, in winter, to a transverse circulation in the opposite direction to the south of lat. 20°N; and he finds this concept confirmed by the available observations. Owing to the prevailing currents in the longitudinal direction a piling up of water must take place in winter at the northern end of the Red Sea and in summer at the southern end, because there the Strait of Bab el Mandeb is too narrow and shallow to permit a free outflow.

The oxygen content of the deep water is very low in spite of sinking of surface water in winter. According to Thompson, the oxygen content shows an annual variation, values higher than 2.0 ml/L being observed at the end of the winter, whereas at the end of the summer most values were lower than 1 ml/L. Thompson attributes the low oxygen values and the annual variation to a very rapid consumption of oxygen, which at depths between 350 and 600 m appears to amount to about 2 ml/L/year. This rate of oxygen consumption is by far the highest which has been found at such depths, but appears reasonable in view of the very high water temperature (about 22°C). The observations of the John Murray Expedition in September and in April–May both indicate the existence of a region of minimum oxygen content in the central portion of the Red Sea at depths between 300 and 500 m. Thompson attributes this minimum to a vertical rotational movement, which in winter is related to the sinking of surface water in the northern part of the Red Sea and rising of deep water inside of the sill, but which, in summer, in part reverses owing to the piling up of water inside of the sill.

The exchange of water between the Red Sea and the adjacent parts of the ocean takes place through the Suez Canal and through the Strait of Bab el Mandeb. The exchange through the Suez Canal is of no importance to the water and salt budget of the Red Sea but shows some interesting details (Wüst, 1934). Wüst points out that any flow of water through the Suez Canal is greatly complicated by the fact that the canal passes through the Bitter Lakes, the bottoms of which consist of layers of salt which are gradually being dissolved, thus increasing the salinity of the waters in the canal to a concentration above that of the Red Sea or Mediterranean Sea waters. In October–December the salinity at the surface of the canal above the Great Bitter Lake is as high as 50.00 ‰ and at the bottom it is above 55.00 ‰. The flow through the Suez Canal is mainly determined by three factors: (1) the difference in sea level between the Red Sea and the Mediterranean Sea, (2) the prevailing local winds, and (3) the great salinity of the canal waters due to the solution of the salt layers of the Bitter Lakes. The sea level is higher at Suez on the Red Sea than at Port Said on the Mediterranean, except in July–September, and because the difference in sea level dominates, the surface flow is directed from the Red Sea to the Mediterranean in all seasons except in July–September, when it is reversed. The highly saline bottom water flows from the Bitter Lakes towards the Mediterranean in all seasons, and from July to December an outflow of this water towards the Red Sea also takes place. The volumes of water and the amount of salt which are transported through the Suez Canal are, however, too small to be significant as far as conditions in the Red Sea or the Mediterranean are concerned.

The exchange of water between the Red Sea and the Gulf of Aden is subject, according to Vercelli (1925) and Thompson (1939b), to a distinct annual variation which is related to the change in the direction of the prevailing winds in winter and summer. In winter, when south-southeast winds blow in through the Strait of Bab el Mandeb, the surface layers are carried from the Gulf of Aden into the Red Sea, and at greater depths highly saline Red Sea water flows out across the sill. In summer, when north-northwest winds prevail, the surface flow is directed out of the Red Sea and at some intermediate depths water from the Gulf of Aden flows in, having a lower salinity and a lower temperature than the outflowing surface water. At still greater depths highly saline Red Sea water appears to flow out over the sill, but it is probable that this outflow is much reduced as compared to the outflow in winter. On the basis of direct measurements of currents at anchor stations, Vercelli found that in winter the average inflow amounts to approximately 0.58 million m^3/sec, whereas the outflow of Red Sea water amounts to approximately 0.48 million m^3/sec. No measurements are available for summer.

Owing to the complicated character of the water exchange in summer, the average salinity of the in- and outflowing water is not known and, furthermore, the excess evaporation from the Red Sea is not well determined, for which reasons a computation of the exchange of water cannot be based on a consideration of the salt balance. Vercelli even thinks it possible that more salt is carried in than out and that the salinity of the Red Sea is increasing. He also points out that owing to the rapid change in salinity through the Strait of Bab el Mandeb, the outflowing tidal currents carry water of higher salinity than the inflowing, such that tidal currents assist in the transport of salt out of the Red Sea. He concludes that in winter the net inflow of 0.1 million m^3/sec is nearly twice as great as the net loss of water by evaporation, for which reason the water level in the Red Sea must rise during winter. According to Vercelli the average annual evaporation excess from the Red Sea amounts to about

3.5 m, the evaporation being considerably greater in summer than in winter; but the latter conclusion, which is based on consideration of pan measurements at shore stations, appears doubtful because the annual variation of evaporation over an extensive sea surface is entirely different from that over land.

The final result of this discussion is not very conclusive. As far as the conditions in the Indian Ocean are concerned, the greatest interest is attached to the amount of Red Sea water that flows out over the sill and spreads at an intermediate depth in a manner similar to the spreading of the Mediterranean Sea water in the Atlantic Ocean. According to Vercelli the amount that flows out in winter is, as already stated, 0.48 million m^3/sec, but in summer the outflow must be considerably smaller. The average annual amount is therefore probably between 0.3 and 0.4 million m^3/sec, that is, approximately one sixth of the amount which flows out through the Strait of Gibraltar. This conclusion is in agreement with the fact that, as presently will be shown, the Red Sea water is of less importance in the Indian Ocean than is the Mediterranean water in the Atlantic Ocean.

The Persian Gulf is so shallow that any exchange of water between it and the adjacent Gulf of Oman is of small significance. The average depth of the Persian Gulf is only 25 m and the maximum depth is about 90 m. It appears to be filled by water of a nearly uniform salinity of about 38.00 $^0/_{00}$ and some exchange must take place with the waters of the Gulf of Oman, but the character of this exchange has not been examined.

The Indian Ocean

THE WATER MASSES OF THE INDIAN OCEAN. Oceanographically, the southern limit of the Indian Ocean can be placed in the region of the Subtropical Convergence, according to which definition the Indian Ocean extends to approximately lat. 40°S. The ocean is closed toward the north and all the water masses in the upper layers are therefore such as are characteristic of the middle latitudes and the equatorial regions. No subpolar water mass enters the Indian Ocean. A considerable number of oceanographic stations have recently been occupied in the equatorial areas of the Indian Ocean by the *Dana*, *Snellius*, and John Murray Expeditions, but as yet most of the data from the two latter are not available. The *Discovery* expeditions have occupied many stations along the African east coast and off South Africa, and a few to the south and southwest of Australia, but no accurate observations are available from the entire central and southern part of the Indian Ocean except a few from the *Planet* Expedition in 1906. Any discussion of the types of water in the Indian Ocean must therefore be of a preliminary character and the classification given here will have to be subject to future corrections.

Within the surface waters the temperature increases rapidly toward the north from the Subtropical Convergence and, in the equatorial regions, it is uniformly high during the greater part of the year, between 25° and 29°. In August lower temperatures, down to 22°, are found along the southeast coast of Arabia and the east coast of Africa as far south as to the Equator, in consequence of the upwelling under the influence of the prevailing southwest monsoon. In February lower temperatures are similarly found in the Gulf of Oman and the Bay of Bengal in consequence of the effect of the northwest monsoon. The salinity of the surface waters shows, to the west of Australia, the common subtropical maximum, and in the equatorial and northern regions is subject to considerable annual variations which are related to the changing monsoons and the annual variation in the precipitation (see chart VI).

At subsurface depths three larger water masses can be shown to exist, the Indian Ocean Central Water and the Indian Ocean Equatorial Water, both of which extend to moderate depths, and the deep water, present below a depth of roughly 2000 m. Transition types are found and, furthermore, two types of water spread at mid-depths, the Antarctic Intermediate Water and the Red Sea Water.

In the inset map in fig. 189 are shown the approximate regions in which the Central Water mass and the Equatorial Water mass are found. Within the area of the Central Water mass seven widely separated stations have been selected, as indicated in the figure. The temperature-salinity relation at these stations is shown in fig. 189 (left), from which it is evident that nearly all the observed temperatures and salinities fall on a straight line between the points $T = 8°$, S = 34.60 ‰, and $T = 15°$, S = 35.50 ‰. A remarkable agreement exists between such widely separated stations as *Discovery* 427 off Port Elizabeth in South Africa and B.A.E. 75 (Howard, 1940) to the southwest of Australia, and between the latter station and *Dana* station 3938 to the northwest of Madagascar. The deeper values from *Dana* station 3938 deviate, however, showing higher salinities that indicate presence of Red Sea water. Traces of this water are also found at *Dana* station 3960 to the south of Madagascar. The southern limit of the Central Water mass coincides with the approximate location of the Subtropical Convergence. In the region of the convergence the surface temperature and salinity vary rapidly with latitude, and the horizontal T-S relation in that region in certain seasons agrees very well with the vertical T-S relation between depths of approximately 100 m and 800 m within the Central Water mass. It is therefore probable that this water mass has been formed by sinking at the Subtropical Convergence. The northern limit of the Central Water mass cannot be determined, owing to lack of observations.

The Equatorial Water of the Indian Ocean is not so well-defined as the Central Water, but at most stations to the north of the Equator a considerable similarity exists in the T-S relationship. Figure 189 (right) also shows the corresponding temperature and salinity values at six stations close to, or north of, the Equator and at two stations to the south of the Equator. At most of the northern stations the corresponding temperatures and salinities fall nearly on a straight line between the points $T = 4°$, $S = 34.90$ ‰, and $T = 17°$, $S = 35.25$ ‰. In the figure is also entered the T-S relation of the Central Water, and it appears that at the southern stations water occurs which is intermediate in char-

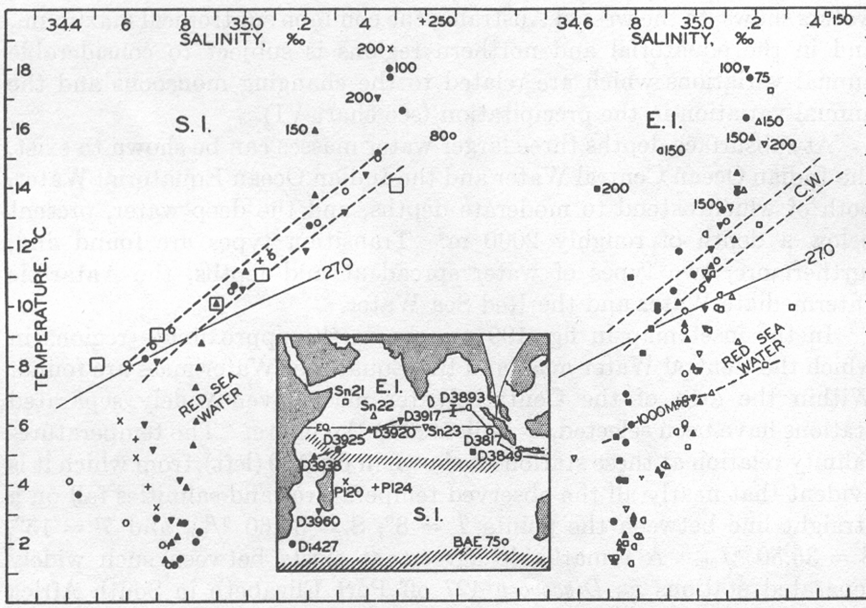

Fig. 189. Temperature-salinity relations in the Indian Ocean. The depths of the shallowest values are indicated. Large squares represent winter surface values in the region of the subtropical convergence. Inset map shows locations of stations used and boundaries of water masses. *Abbreviations:* Sn, *Snellius;* D, *Dana;* Di, *Discovery;* P, *Planet;* BAE, *B.A.N.Z. Expedition;* E.I., Equatorial Indian Ocean; S.I.. South Indian Ocean.

acter between the Equatorial Water and the Central Water. This water of intermediate character is particularly conspicuous at the two most southern stations, *Dana* 3849 and *Dana* 3925, at which similar T-S relations exist below 10°. At *Dana* station 3849, located to the southwest of Sumatra, the low-salinity water of temperature higher than 10° is probably of Pacific origin, and if a more detailed study were to be made it would be necessary to subdivide the regions. At the two northwestern stations, *Snellius* 21 and 22 (van Riel, 1932a), Red Sea water is evidently present, as seen from the conspicuously high salinities at a depth of 750 m. The rough division used here corresponds to some extent to the divisions introduced by Lotte Möller (1929), but several of her subdivisions have been combined.

The spreading of the Red Sea water in the Indian Ocean is probably similar to that of the Mediterranean water in the Atlantic Ocean, but owing to lack of data it is impossible to carry out a detailed analysis comparable to that of the Mediterranean component made by Wüst (1935). Repeated observations at the same stations in the Gulf of Aden have given very different salinity values, a condition which strongly indicates that the outflow is intermittent. A seasonal variation in the outflow appears to be well established, but great year-to-year differences may also occur and such fluctuations complicate the conditions at greater distances. The best idea of the spreading of the Red Sea water is obtained

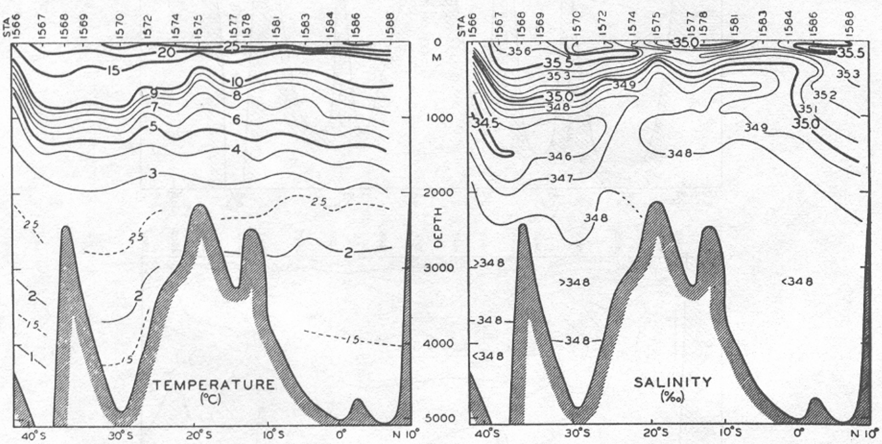

Fig. 190. Distribution of temperature and salinity in a vertical section along the east coast of Africa. Location of section shown on inset map in fig. 192. (After Clowes and Deacon.)

from a *Discovery* section parallel to the African east coast, the location of which is shown in the inset map, fig. 192 (p. 697). The distributions of temperature, salinity, and oxygen in this section have been discussed by Clowes and Deacon (1935), from whose paper fig. 190 and fig. 192 have been reproduced. Figure 190 shows the Red Sea water penetrating as a tongue of water of high salinity, which sinks from a depth of about 500 m at station 1588 in lat. 8°N (see fig. 192) to a depth of about 1250 m south of lat. 20°S between southern Madagascar and the mainland. This water has a low oxygen content, as is evident from fig. 192. Clowes and Deacon draw attention to the fact that the oxygen distribution indicates a further penetration of the Red Sea water to the south, such that the last traces of this water may possibly be present in lat. 40°S.

The *Discovery* section off the African east coast is the only longitudinal section that can be constructed in the Indian Ocean by means of redinall data. Other sections have been constructed by Lotte Möller (1929) by means of older data, but these are not included here because the original data appear to be subject to systematic errors.

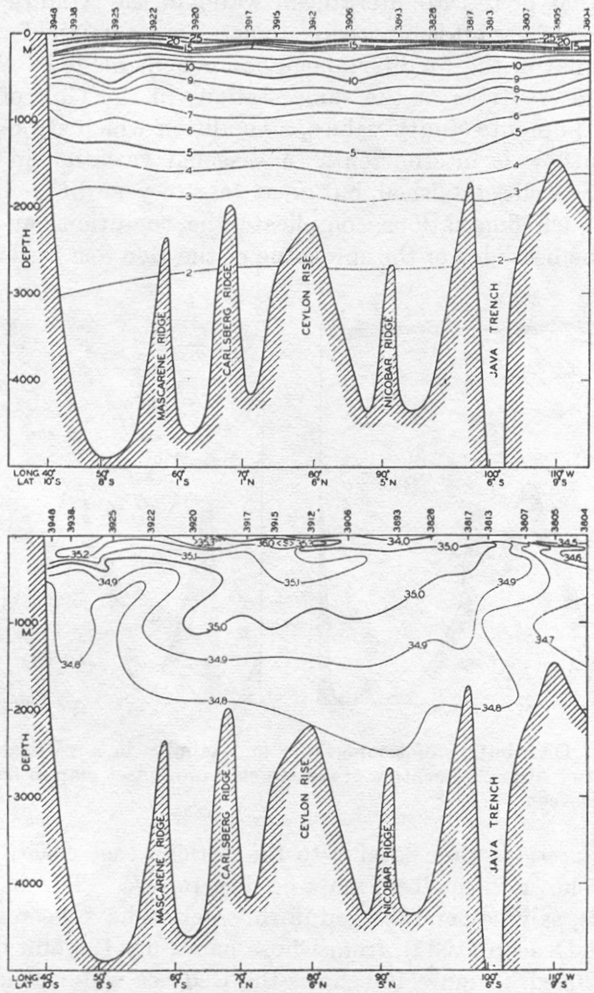

Fig. 191. Distribution of temperature and salinity in a vertical nearly east-west section in the equatorial region of the Indian Ocean. Location of section shown by line I on inset map in fig. 189.

An approximately east-west section in the equatorial regions has been constructed by means of *Dana* observations, using stations along line I in fig. 189. The section follows the coasts of Java and Sumatra, crosses the Bay of Bengal, and runs in a south-southeasterly direction from Ceylon to Cape Delgado on the coast of east Africa in lat. 10°S, lying to the north of the Equator between longitudes 66°E and 97°E. Figure 191 shows the distribution of temperature and salinity in this section. The bottom topography is roughly indicated, mainly on the basis of the soundings made at the different stations. From the temperature section it is seen that in the tropical part of the Indian Ocean one finds a stratification similar to the one encountered in corresponding latitudes of the Atlantic Ocean. Below a surface layer of high and relatively uniform temperatures, having a thickness of 75 to 100 m, the temperature decreases abruptly, the decrease often exceeding 10° in 50 m. Below this layer of rapid decrease the temperature drops more slowly toward the deep water, which has a temperature between 3° and 1.2°. The abrupt drop of temperature below the surface layer appears to be particularly conspicuous to the north of the Equator.

The surface salinity is below 35.00 $^0/_{00}$ except to the west of long. 70°E, but a subsurface salinity maximum is present over a large part of the section, the maximum being found near the upper limit of the sharp thermocline, as was the case in the Atlantic Ocean. In the Indian Ocean this high-salinity surface water appears to come from the Arabian Sea.

Below the salinity maximum the salinity gradually decreases with increasing depth, until the nearly constant value for the deep water of about 34.76 $^0/_{00}$ is reached between depths of 2500 and 3000 m; but to the right in the section, low-salinity water is present that probably comes from the Pacific Ocean where water of similar characteristics is found (p. 707). The Red Sea water is not very conspicuous. To the extreme left it is shown by the downward bend of the 34.8 isohaline and by the greater distance between the 8° and 7° isotherms.

In the middle part of the section the equatorial water is present, the bulk of which has a temperature below 10° and a salinity less than 35.10 $^0/_{00}$. This water probably contains some admixture of Red Sea water but mainly it is formed and maintained by slow processes of mixing between the high-salinity water from the Arabian Sea and the deep water; the origin of the latter will be discussed when dealing with the deep-water circulation of the oceans.

The Currents of the Indian Ocean. In the southern part of the Indian Ocean a great anticyclonic system of currents appears to prevail, comparable to the corresponding systems of the South and North Atlantic Ocean except that it is subjected to greater annual variations. Between South Africa and Australia the current is directed in general from west to east. In the southern summer the current bends north before reaching the Australian Continent and is joined by a current which flows from the Pacific to the Indian Ocean to the south of Australia. In winter the current appears to reach to Australia and in part to continue towards the Pacific along the Australian south coast. To the north of 20°S the South Equatorial Current flows from east to west, reaching its greatest velocity during the southern winter when the southwest monsoon over the northern part of the ocean represents a direct continuation of the southeast trade winds on the southern side of the Equator. In this season the current is reinforced by water from the Pacific Ocean, which enters the Indian Ocean to the north of Australia, but in the southern summer the flow to the north of Australia is reversed.

In both seasons of the year part of the South Equatorial Current turns south along the east coast of Africa, feeding the strong Agulhas Stream. To the south of lat. 30°S the Agulhas Stream is a well-defined and narrow current which extends to a distance from the coast of less than 100 km. As is to be expected in a flow towards the south in the Southern Hemisphere, the coldest water is found inshore and the sea surface rises when departing from the coast. Off Port Elizabeth the rise amounts to about 29 cm in a distance of about 110 km (Dietrich, 1936). To the south of South Africa the greater volume of the waters of the Agulhas Stream bends sharply to the south and then toward the east, thus returning to the Indian Ocean by joining the flow from South Africa toward Australia across the southern part of that ocean, but a small portion of the Agulhas Stream water appears to continue into the Atlantic Ocean (Dietrich, 1935). Owing to the reversal of the direction of the main current to the south of Africa, numerous eddies develop, resulting in a highly complicated system of surface currents which probably is subjected to considerable variations during the year and variations from one year to another.

To the north of lat. 10°S the surface currents of the Indian Ocean, which are probably nearly identical with the currents above the tropical discontinuity surface, vary greatly from winter to summer owing to the different character of the prevailing winds. During February and March when the northwest monsoon prevails, the North Equatorial Current is well developed and an Equatorial Countercurrent is present with its axis in approximately 7°S. Along the African east coast between the Gulf of Aden and lat. 5°S the current is directed towards the south. In August–September when the southwest monsoon blows, the North Equatorial Current disappears and is replaced by the Monsoon Current, which flows from west to east. Along the coast of east Africa the current is directed north from lat. 10°S, water of the Equatorial Current crosses the Equator, and considerable upwelling takes place off the Somali coast. The Equatorial Countercurrent does not appear to be present in this season.

At present nothing is known as to the motion of the water below the tropical discontinuity in the northern part of the Indian Ocean. The character of the water, together with the very low oxygen values which are found to the north of the Equator, indicates that no strong currents exist and that only a sluggish flow takes place.

In the southern part of the Indian Ocean the Antarctic Intermediate Water probably flows north, but the flow must be less well-defined than the corresponding flow in the South Atlantic Ocean, because in the Indian Ocean the Antarctic Intermediate Water loses its typical characteristics on a shorter distance from the Antarctic Convergence (Sverdrup, 1940). The probable flow of the deep water will be discussed later on.

The data from the Indian Ocean are too scanty to permit many quantitative calculations as to the amount of water carried by the different branches of the current. The only reliable figure which is available is found in Dietrich's study of the Agulhas Stream, which transports a

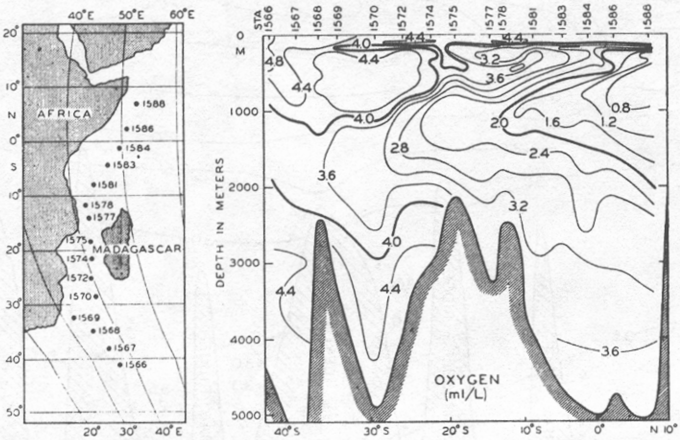

Fig. 192. Distribution of oxygen (ml/L) in a vertical section along the east coast of Africa. Location of section shown on map. (After Clowes and Deacon.)

little more than 20 million m³/sec. A transport map similar to the one for the North Atlantic Ocean cannot be constructed.

Oxygen Distribution. The distribution of oxygen in the Indian Ocean is not known in detail, but some of the characteristic features can be seen from the *Discovery* and *Dana* sections in figs. 192 and 193. According to the *Discovery* section along the African east coast the oxygen content of the Indian Ocean decreases from the south toward the Equator. The Red Sea water is characterized by low oxygen content, but the content increases in the direction in which the Red Sea water spreads, probably due to mixing with the over- and underlying water masses. The Antarctic Intermediate Water is relatively high in oxygen and the same statement applies to the deep water, in which values ranging from 4.4 to 3.6 ml/L are observed. To the north of 20°S an intermediate minimum of oxygen is present at a depth of approximately 200 m. According to the *Dana* section the oxygen content of the tropical waters decreases abruptly within the thermocline to an intermediate minimum, which at all stations to the north of the Equator is found at a depth of about 150 m. The lowest values were observed at *Dana* station 3912, where at 150 m the oxygen content was 0.21 ml/L. Below the minimum there is a secondary maximum between 300 and 400 m, which is present at all stations except the most easterly, at which water of Pacific origin was found. The low oxygen content of the Red Sea water is indicated at

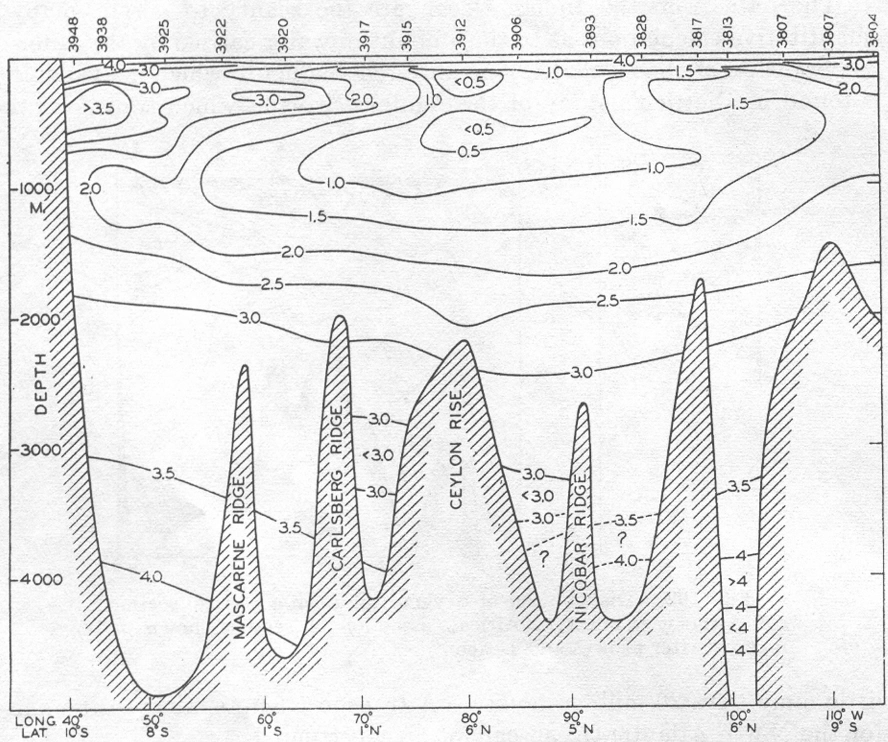

Fig. 193. Distribution of oxygen (ml/L) in a vertical nearly east-west section in the equatorial region of the Indian Ocean. Location of section shown by line I on inset map in fig. 189.

the left in fig. 193 by the form of the 2.0-ml/L curve. The deep water has evidently an oxygen content which is lower than that of the deep water further south, values below 3.0 ml/L being found at depths greater than 3000 m at several stations.

The South Pacific Ocean

Water Masses of the South Pacific Ocean. Above the deep water the water masses of the Pacific Ocean are of more complicated character than those of the other oceans. Subantarctic Water is of small significance in the Atlantic and Indian Oceans, but the South American

Continent, owing to its far southward extent, deflects large quantities of Subantarctic Water to the north along the west coast of South America, such that in the Pacific Ocean this water exercises an influence which extends beyond the Equator. Similarly, Subarctic Water masses are present in large quantities in the North Pacific, where they are carried toward the east and toward the south along the coast of North America as far as to lat. 25°N. Another equally important reason for the difference between the Atlantic and the Pacific Ocean is that in the Pacific

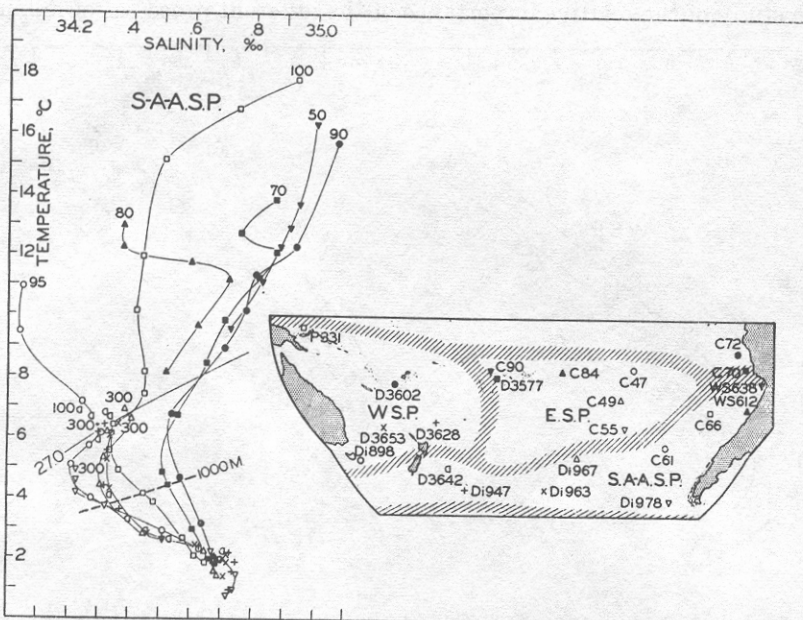

Fig. 194. Temperature-salinity relations within the Subantarctic Water of the South Pacific. The depths of the shallowest values are indicated. Locations of stations and boundaries of water masses are shown on the inset map. *Abbreviations:* D, *Dana;* C, *Carnegie,* Di, *Discovery,* W.S., *William Scoresby,* SA-A.S.P., Subantarctic South Pacific; W.S.P. Western South Pacific; E.S.P., Eastern South Pacific.

the circulation of the enormous water masses of that ocean is more sluggish and an intense mixing of different water masses with development of a uniform body of water over the entire ocean does not take place, as it does in the Atlantic Ocean.

Turning first to the South Pacific Ocean, one finds to the south of lat. 40°S the Subantarctic Water mass which, in the upper layers, is characterized by a salinity between 34.20 ‰ and 34.40 ‰ and a temperature between 4° and 8°. This is evident from the *T-S* curves in fig. 194, according to which the water at *Dana* station 3642 off New Zealand was of the same character as the water at the *Discovery* station 967 half way between New Zealand and South America. Part of this

Subantarctic Water bends to the north along the coast of South America, and in course of time the temperature of the surface layers is raised by heating and the salinity is increased owing to evaporation. Consequently, the *T-S* curves bend more and more to the right, as shown in the figure, and simultaneously, the salinity minimum of the Antarctic Intermediate Water decreases in intensity. Lateral and vertical mixing are of importance to the change in the temperature-salinity relationship indicated by the curves in fig. 194, but the available data are insufficient for a study of the relative importance of the different processes of mixing.

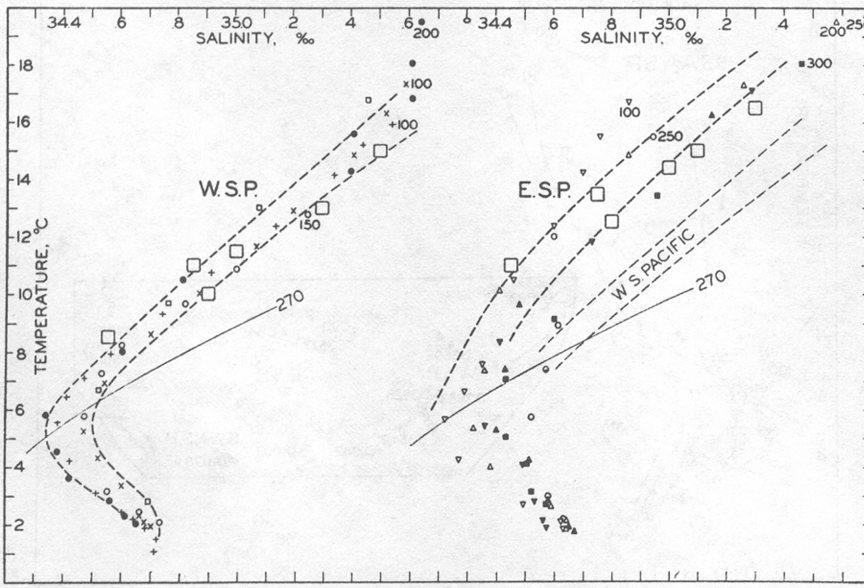

Fig. 195. Temperature-salinity relations in the western and eastern South Pacific. The depths of the shallowest values are shown. Squares represent winter surface values in the region of the Subtropical Convergence. Locations of stations used and boundaries of water masses shown on inset map in fig. 194. W.S.P., Western South Pacific; E.S.P., Eastern South Pacific.

In the western part of the South Pacific Ocean a water mass is encountered that is similar to the Central Water of the Indian Ocean. This is evident from fig. 195, in which the corresponding temperature and salinity values at a number of *Dana* stations are entered, together with those from *Discovery* station 898 off Tasmania and *Planet* station 331 in lat. 2°S and long. 152°E. A comparison with fig. 188, p. 685, shows that the Indian Ocean Central Water and the water in the western Pacific are practically identical.

The large squares in the figure show the corresponding values of surface temperatures and salinities in August in lat. 35° to 45°S, long. 150°E to 160°W (according to Schott's charts, 1935). Again a region is found on the poleward side of the water mass within which the horizontal *T-S* relation agrees with the vertical *T-S* relation of the water mass.

In the eastern South Pacific the few available data indicate the existence of another water mass which has, between temperatures of 10° and 18°, a salinity nearly 0.50 $^0/_{00}$ lower than that of the western body of water. The definition of this water mass is mainly based on the observations at a number of *Carnegie* stations, four of which are shown in fig. 195. At *Dana* station 3580, located in 19°S, 163°W, the water is intermediate in character between the eastern and western masses, and the boundary region between these water masses can therefore be placed approximately along the meridian of 165°W. The squares in fig. 195 (right) show corresponding values of surface temperature and salinity in August in lat. 35° to 40°S and long. 150° to 120° W.

In the South Pacific Ocean we therefore encounter three distinctly different upper water masses: the Subantarctic Water mass, which is more or less uniform to the south of 40°S and which changes its characteristics as it moves north along the coast of South America, and the western and the eastern South Pacific Central Water masses, which are separated from each other by a region of transition in about 165°W. The eastern Central Water mass does not extent north of 10°S but the western may extend nearly to the Equator, as indicated by the observations at *Planet* station 331. The existence of two characteristically different central water masses indicates that in the South Pacific Ocean the circulation is split up in two large cells, the location of which appears to be related to the prevailing winds. In the southern winter the atmospheric pressure shows two distinct areas of high pressure over the Pacific Ocean, one eastern with its center in approximately lat. 28°S and long. 100°W, and one western which extends partly over the Pacific Ocean with its center over eastern Australia. The surface currents (chart VII), which only in part reflect the circulation in the deeper layers, and in part show the wind drift of surface water, indicate in most seasons the existence of a region of weak and irregular currents in about 160°W which is the approximate region where the transition between the two central water masses takes place.

Below the upper water masses Antarctic Intermediate Water is present, in the east probably to within 10° to 15° from the Equator, and in the west to the Equator. In the South Pacific no salinity maximum is found below the intermediate water, but the salinity increases toward the bottom or remains constant below a depth of 2500 or 3000 m.

Currents of the South Pacific Ocean. The only major current of the South Pacific Ocean which has been examined to some extent is the Peru Current. Following the nomenclature proposed by Gunther (1936), the name Peru Current will be applied to the entire current between the South American Continent and the region of transition towards the eastern South Pacific Central region; the part of the current which is close to the coast will be called the Peru Coastal Current, whereas the part which is found at greater distances will be called the Peru Oceanic Current.

As is evident from the character of the waters (fig. 194), the origin of the Peru Current has to be sought in the subantarctic region, part of the Subantarctic Water which flows towards the east across the Pacific Ocean being deflected towards the north when approaching the American Continent. The total volume of water in the current does not appear to be very great. On the basis of a few *Discovery* stations it is found that the transport lies somewhere between 10 and 15 million m^3/sec, and this figure includes transport of the upper water layers and of Antarctic Intermediate Water. The western limit of the current appears to be diffuse and cannot be well established on the basis of the available data, but it is probable that the current extends to about 900 km from the coast in 35°S. The northern limits, according to Schott (1931), can be placed a little south of the Equator where the flow turns toward the west.

The current being wide and the transport small, the velocities are quite small. This must be true particularly in the case of the Peru Oceanic Current which, however, is little known. The Peru Coastal Current is better known, partly because numerous observations of surface currents and surface temperatures are available off the coasts of Peru and Ecuador between the Equator and 10°S, and partly because of the extensive examination of the coastal waters which was conducted by the R.R.S. *William Scoresby* in 1931 and which has been discussed in detail by Gunther (1936). Within the Peru Coastal Current upwelling represents a very conspicuous feature. The upwelling is caused by the southerly and south-southeast winds which prevail along the coasts of Chile and Peru and carry the warm and light surface waters away from the coast, resulting in cold water being drawn from moderate depths toward the surface. On the basis of numerous sections close to the coast, Gunther concludes that the upwelling water comes from depths between 40 and 360 m, the average depth being 133 m. The upwelling therefore represents only an overturning of the upper layers and no water from greater depths is ever drawn to the surface.

The process of upwelling will naturally influence the current parallel to the coast because the distribution of density is altered, as explained on p. 500. The *William Scoresby* observations clearly show that the upwelling is an intermittent process, greatly influenced by local winds, and that reversal of the wind direction frequently leads to subsidence, that is, to re-establishment of the stratification characteristic of the undisturbed conditions.

According to Schott and to Gunther, the most active upwelling occurs in certain regions separated by regions in which the upwelling is less intense. Both authors recognize four such regions between lat. 3°S and 33°S, but they do not agree on the extent of the different regions, probably because the regions are not absolutely fixed or because the locations ascribed to the different regions may depend upon the available data. Gunther has particularly examined the two northern regions where the most intense upwelling occurs in 5°S and 15°S respectively, and has shown that the surface temperatures in the winter of 1931 (June to August) indicate the existence of two tongues of warm water which approach the coast to the south of the regions of intense upwelling. The upwelled water, on the other hand, leaves the coast as tongues of cold water, and consequently the distribution of surface temperatures shows alternate tongues of warm and cold water. Schott's analysis and other observations indicate that the locations of these tongues do not vary much from one year to another, and the tongues must therefore be either permanent or recurrent. Gunther interprets these tongues as demonstrating the existence of swirls off the coast, assuming that within one branch of the swirl upwelled water moves out and within another branch oceanic water moves in toward the coast.

In early winter, April to June, the shoreward-directed branch of the northern swirl is well developed and carries water of high temperature in toward the coast in latitudes 9° to 12°S. It may even appear as an inshore warm current which brings great destruction to the animal life of the coastal waters. In the discussion of the California Current it will be shown that during the period of upwelling this current is similarly characterized by a series of swirls on the coastal side of the current, whereas from November to February, when there is practically no upwelling, a warm countercurrent flows to the north along the coast.

Close to the coasts of Peru and Chile a subsurface countercurrent flows south, as is evident from the section prepared by Gunther, showing the distribution of salinity between the surface and a depth of 400 m at a distance of approximately 180 km from the coast and between lat. 3° and 36°S. The subsurface current appears to originate at the equatorial end of the section, where it is present at a depth of less than 100 m, but sinks gradually when progressing to the south, being found at nearly 300 m in lat. 36°S. The water of this countercurrent is of the type which has been called Pacific Equatorial Water (p. 706), as is evident from the *T-S* curves in fig. 194, in which the data from two stations within the countercurrent, stations WS 638 and WS 612, have been plotted. To the north of lat. 25°S, where excessive evaporation increases the salinity, the salinity of the subsurface countercurrent is lower than that of the surface water, whereas to the south of about 25°S the salinity of the countercurrent is higher than that of the surface waters. Consequently, the upwelling brings relatively low-salinity water to the surface off the coast of Peru, but relatively high-salinity water to the surface off the coast of Chile. The subsurface countercurrent has its equivalent off the coast of California (p. 725); there, however, it does not begin at the Equator but in approximately 25°N.

At the northern boundary of the Peru Coastal Current certain characteristic seasonal changes take place. During the northern summer the Peru Coastal Current extends just beyond the Equator where it converges with the Equatorial Countercurrent, the waters of which in summer mainly turn towards the north. In winter this countercurrent is displaced further to the south and part of the warm but low-salinity water of the Countercurrent turns south along the coast of Ecuador, crossing the Equator before converging with the Peru Coastal Current. The warm south-flowing current along the coast is known as El Niño and is a regular phenomenon in February and March, but the southern limit mostly lies only a few degrees to the south of the Equator. Occasionally, major disturbances occur which appear to be related to changes in the atmospheric circulation (Schott, 1931). In disturbed years, such as in 1891 and in 1925, El Niño extends far south along the coast of Peru, reaching occasionally past Callao in 12°S. According to Schott, the duration of El Niño periods in 1925 was as follows:

Locality	Latitude	Dates	Duration, days
Off Lobitos	4°20′S	Jan. 20 to April 6	76
Off Puerto Chicana	7°40′S	Jan. 30 to April 2	63
Off Callao	12°20′S	March 12 to March 27	15
Off Pisco	13°40′S	March 16 to March 24	8

These figures show that the warm surface waters of the equatorial area slowly penetrated to the south, but withdrew much more rapidly, because the time interval between the appearance of the warm water off Lobitos and off Pisco was 44 days, whereas the time interval between the disappearance of the warm water at the two localities was only 13 days. The surface temperature of the water in March, 1925, was nearly 7° above the average, as is evident from the following compilation:

Locality	Average temperature in March (°C)	Temperature in March, 1925 (°C)
Lobitos	22.2	27.3
Puerto Chicana	20.3	26.9
Callao	19.5	24.8
Pisco	19.0	22.1

Details as to the surface salinity are not available, but normally the surface salinity between 5°S and 15°S is above 35.00 $^0/_{00}$, whereas the waters of El Niño have a salinity between 33.00 and 34.00 $^0/_{00}$.

The extreme development of El Niño leads to disastrous catastrophes of both oceanographic and meteorological character. The decrease of the temperature of El Niño toward the south indicates that the waters are mixed with the ordinary cold coastal waters, and during this mixing process the organisms in the coastal current, from plankton to fish, are destroyed on a wholesale scale. Dead fish later cover the beaches, where they decompose and befoul both the air and the coastal waters. So much hydrogen sulphide may be released that the paint of ships is blackened, a phenomenon known as the "Callao painter." More serious is the loss of food to the guano birds, many of which die of disease or starvation or leave their nests, so that the young perish, bringing enormous losses to the guano industry. The meteorological phenomena which accompany El Niño are no less severe. Concurrent with a shift in the currents a shift of the tropical rain belt to the south takes place. In March, 1925, the precipitation at Trujillo in 8°S amounted to 395 mm, as compared to an average precipitation in March of the eight preceding years of only 4.4 mm. In 1941 the rainfall was again excessive, but no details are available. These terrific downpours naturally cause damaging floods and erosion. In the 140 years from 1791 to 1931 twelve years were characterized by excessive rainfall at Piura in lat. 5°S and twenty-one years by moderate rainfall which was, however, greatly in excess of the average. During the remaining nearly one hundred years the rainfall was close to nil. A greater development of El Niño is therefore not an uncommon phenomenon, but the catastrophic developments appear to occur on an average of once in twelve years. The records reveal no periodicity because the interval between two disastrous years varies from one year to thirty-four years.

El Niño is not the only current that brings warm water to the coast of Peru with subsequent destruction of the organisms near the coast. High temperatures off the coast appear to be an annual occurrence in the months of April to June, in about lat. 9° to 12°S, that is, at and to the north of Callao. These high temperatures are due, as pointed out by Gunther, to the greater development of the warm branch of the northern swirl; the water that approaches the coast is in this case offshore oceanic surface water of high temperature and relatively high salinity (p. 703). The disastrous effect on the marine organisms is very much milder than that of El Niño, but otherwise similar in character. It may lead to the killing of plankton and fish and to the migration of guano birds, but is ordinarily observed mainly by reason of a change in the color of the coastal water and development of hydrogen sulphide. Locally, these changes are known by the name of *aguaje*, also used synonymously with

"Callao painter." The approach of the oceanic water toward the coast is not accompanied by any disastrous meteorological conditions.

On leaving the coast the waters of the Peru Current join the waters of the South Equatorial Current, which flows all the way across the Pacific towards the west but is known only as far as surface conditions are concerned. The subsurface data are inadequate for computation of velocities and transports, and it is therefore not possible to give any numerical values. Some features of this current will, however, be dealt with when discussing the currents of the equatorial regions of the Pacific, and it will be shown that the cold water along the Equator does not represent a continuation of the Peru Coastal Current but is due to a divergence along the Equator within the South Equatorial Current.

The other currents of the South Pacific Ocean are even less known, but from the character of the water masses it appears that two current systems exist, the nature of which may be revealed by future exploration. One big gyral appears to be present in the eastern South Pacific; in the western South Pacific annual variations are so great that in many regions the direction of flow becomes reversed, as is the case off the east coast of Australia. No chart of the transport by the current can be prepared.

The Equatorial Region of the Pacific Ocean

WATER MASSES OF THE EQUATORIAL PACIFIC. In the equatorial region of the Pacific and below the tropical discontinuity layer, one finds an Equatorial Water mass of a remarkably uniform character which extends over the entire Pacific Ocean from east to west, as is evident from fig. 196, in which the temperature-salinity values at nine stations have been plotted. This Equatorial Water mass has its greatest north-south extension along the American coast, where it is present between latitudes 18°S and 20°N. Towards the west it appears to become narrower although, as will be shown, Equatorial water is occasionally found near the Hawaiian Islands. For points still further west detailed information is lacking, but at *Dana* station 3750, close to the Equator in long. 135°E, characteristic Equatorial Water was encountered, as is evident from the figure.

The Equatorial Water is probably formed off the coast of South America by gradual transformation of the Subantarctic Water, as suggested by the *T-S* curves in fig. 194, and spreads to the north and to the west. In the extreme western part of the Pacific the *Dana* data indicate that some of this water enters the North Pacific Ocean.

The Pacific Equatorial Water mass is characterized by a nearly straight *T-S* correlation between $T = 15°$, $S = 35.15$ ‰, and $T = 8°$, $S = 34.6$ ‰. At a depth of about 800 m where the temperature is about 5.5°, a salinity minimum exists in which the minimum values lie between 34.50 ‰ and 34.58 ‰. Below a depth of 1000 m the *T-S* curve is again a nearly straight line, the temperature decreasing and the salinity increasing toward the bottom, where the corresponding values are 1.3° and 34.70 $^0/_{00}$.

The water masses near the surface are separated from the Equatorial Water mass by a layer of transition within which the temperature decreases and the density increases so rapidly with depth that in many localities the layer has the character of a discontinuity surface. In

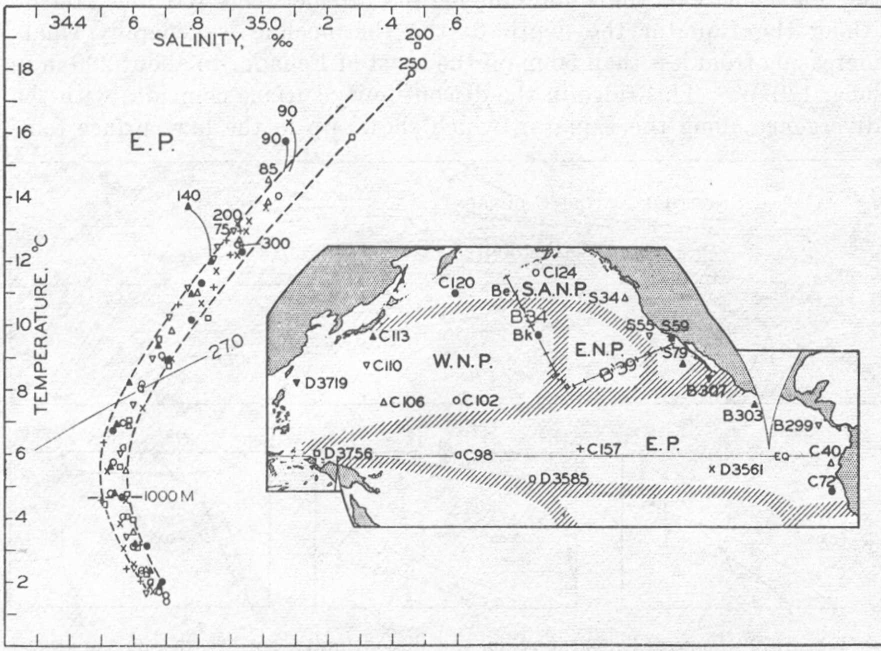

Fig. 196. Temperature-salinity relations in the Equatorial Pacific. The depths of the shallowest values are shown. Locations of stations and boundaries of water masses are shown in the inset map. *Abbreviations:* C, *Carnegie,* B, *Bushnell,* S, *E. W. Scripps,* D, *Dana,* B34, *Bushnell,* 1934; B39, *Bushnell,* 1939; E.P., Equatorial Pacific; W.N.P., Western North Pacific; E.N.P., Eastern North Pacific; S.A.N.P., Subarctic North Pacific.

or directly above the discontinuity layer a salinity maximum is present over large areas. Within the surface layer the temperature is nearly constant to a depth that varies from as little as 10 or 15 m up to nearly 100 m. These conditions are quite similar to those in the tropical regions of the Atlantic and Indian Oceans, but the scarcity of data makes it impossible to study the upper layers in the equatorial part of the Pacific in such detail as can be done in the Atlantic Ocean. A few general features can, however, be pointed out. In fig. 197, which covers the tropical region of the Pacific Ocean, is shown the approximate depth to the thermocline by means of curves drawn at intervals of 50 m. The contours are drawn as smooth curves because so few data are available that no details can be incorporated. The figure brings out features which are similar to those in the Atlantic Ocean (fig. 171, p. 633) and are even more marked. The depth to the thermocline is very small on the eastern side of the Pacific Ocean, where a sharp decrease in temperature is encountered at depths mostly less than 50 m, but toward the west the depth to the thermocline generally increases and on the western side of the Pacific Ocean the minimum depths are between 150 and 200 m. Along the Equator the depth to the thermocline is generally small, increasing from less than 50 m off the coast of Ecuador to about 200 m in long. 170°W. This ridge in the discontinuity surface coincides with the divergence along the Equator, which shows up in the low surface tem-

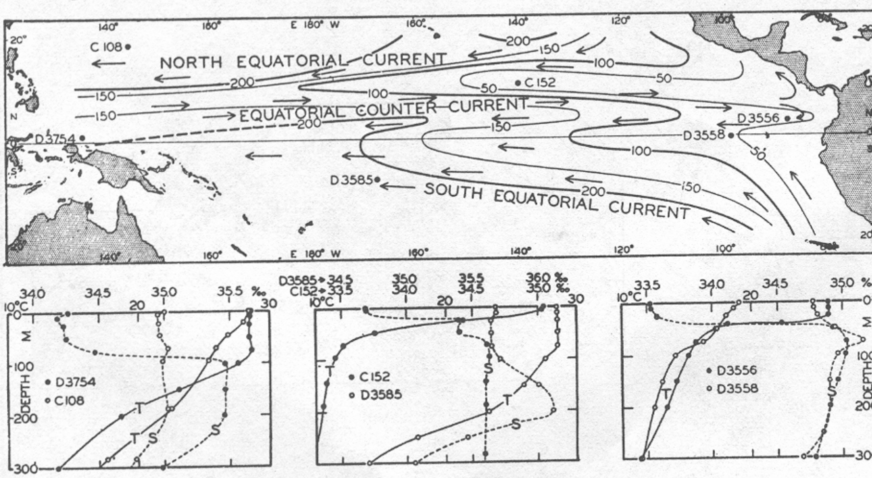

Fig. 197. *Upper:* Topography of the discontinuity surface in the equatorial region of the Pacific and corresponding currents. *Lower:* Vertical temperature and salinity curves at six stations, the locations of which are shown in the upper figure. C, *Carnegie,* D, *Dana.*

peratures extending from the coast of South America and halfway across the Pacific Ocean (charts II and III). Another ridge nearly parallels the Equator in about lat. 8°N; this is associated with the Equatorial Countercurrent. In the lower part of the figure are shown the vertical temperature and salinity distributions at six stations. These curves demonstrate the increase of the depth to the thermocline toward the west, the presence of the surface layer of uniform temperature, and the presence of the salinity maximum near the thermocline. This maximum is particularly conspicuous in the South Pacific and appears to be less developed in the North Pacific Ocean.

Currents of the Equatorial Pacific. The chart of the depth to the thermocline immediately gives an idea of the currents in the upper layers in the tropical region. If the motion of the water below the discontinuity layer is small, the current in the upper layers must be related to the slope of the discontinuity layer in such a manner that in the Northern Hemisphere the discontinuity layer sinks to the right of an observer looking in the direction of the current and, in the Southern Hemisphere, sinks to the left (p. 445). In fig. 197 arrows have been entered on the basis of this rule, showing the North and South Equatorial Currents flowing toward the west and between them the Equatorial Countercurrent flowing toward the east. The South Equatorial Current is present on both sides of the Equator and extends to about 5°N but the North Equatorial Current remains in the Northern Hemisphere. Off South America the flow is directed more or less parallel to the coast line, turning gradually west when approaching the Equator.

The Equatorial Countercurrent is remarkably well developed in the Pacific Ocean where, according to charts by Puls (1895), it is present at all seasons of the year, lying always in the Northern Hemisphere but further away from the Equator in the northern summer. In this season the velocities of the current also appear to be higher, reaching values up to 2 knots at the surface. The structure of the water masses was first demonstrated by the *Carnegie* section in approximately long. 140°W, which was obtained in October, 1929. Figure 198A and 198B show the distribution of temperature and salinity in this section between the surface and 300 m. The figure brings out the great variation in the depth to the thermocline in a north-south direction, and the presence of the surface layer of uniform temperature and the tongues of maximum salinity. In figure 198C is shown the velocity distribution, computed on the assumption of no motion at the 700-decibar surface. These computations are uncertain near the Equator, but the resulting picture is remarkably consistent. In this case the Equatorial Countercurrent lies between 3°N and 10°N, as indicated by the letters stating the direction in which the currents flow.

Montgomery and Palmén (1940) have shown that the Equatorial Countercurrent in the Pacific Ocean, as well as that in the Atlantic Ocean, is maintained by the piling up of the light surface water against the western boundary of the ocean. They have made use of one *Dana* station, 3558, in long. 99°07′W, and of three *Dana* stations, 3775, 3756, 3767, in long. 134°44′E, all near the Equator. A comparison between the dynamic heights of the sea surface relative to the 1000-decibar surface shows that in long. 135°E the surface lies 62.6 dyn. cm higher than in long. 99°W, but the difference decreases with increasing depth and the 300-decibar surface is parallel to the 1000-decibar surface. If dynamic centimeters are replaced by centimeters one obtains the result that along the Equator the slope of the sea surface is 4.5×10^{-8}. This slope of the surface and the slopes of the isobaric surfaces above the 300-decibar surface can be maintained by an east wind of velocity 4 m/sec,

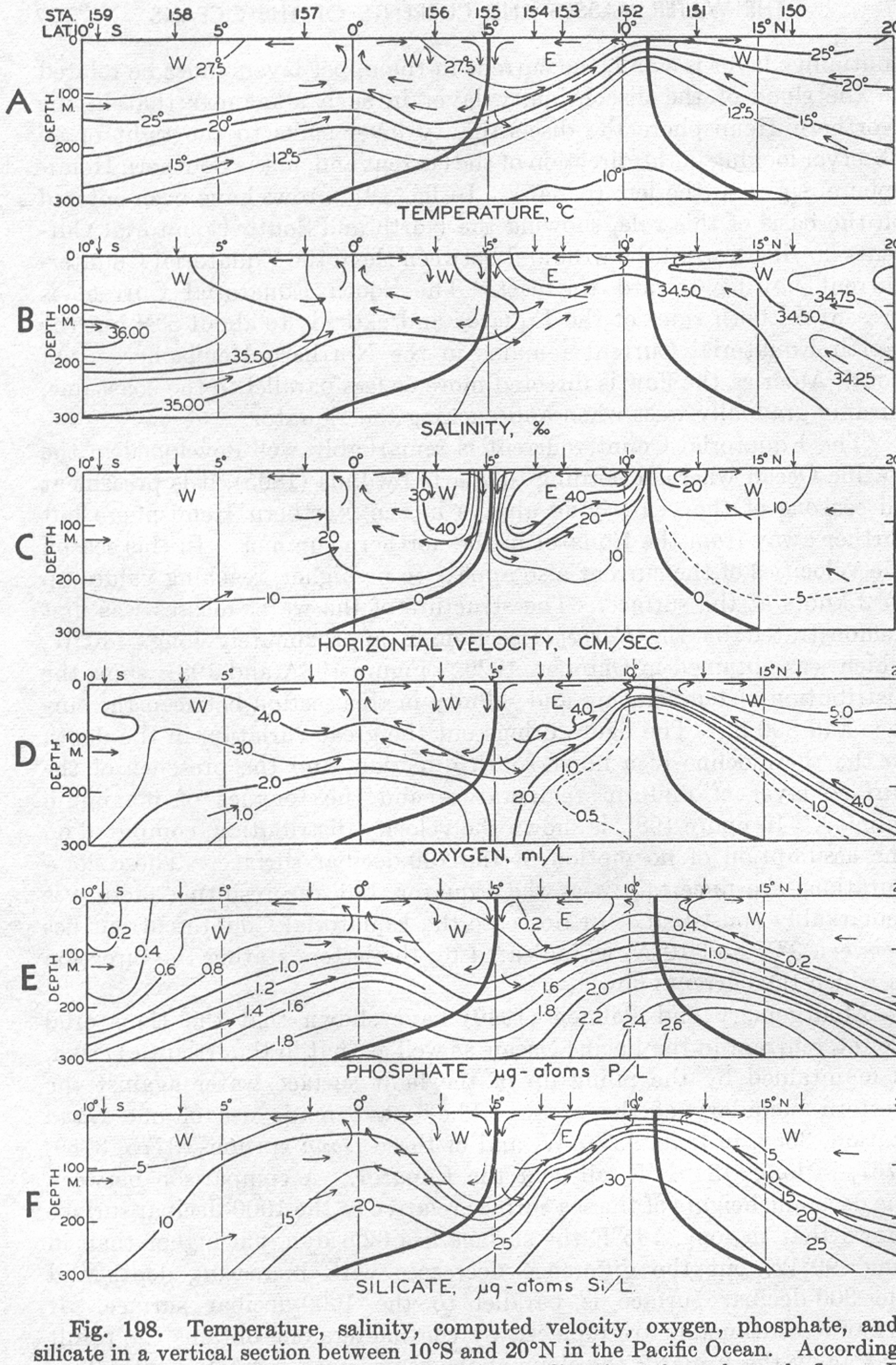

Fig. 198. Temperature, salinity, computed velocity, oxygen, phosphate, and silicate in a vertical section between 10°S and 20°N in the Pacific Ocean. According to the *Carnegie* observations. Arrows indicate direction of north-south flow. *E* indicates flow to the east; *W*, flow to the west.

which is lower than the observed velocity of the trades. The stress of wind, therefore, more than accounts for the observed conditions (p. 488). Owing to the slope of the sea surface the Equatorial Countercurrent flows downhill in the calm belt between the trade winds, and the distribution of mass adjusts itself to the presence of the current. The maximum velocity at the surface, as computed from the *Carnegie* section (fig. 198C) is a little over 50 cm/sec or about 1 knot, in good agreement with values reported from ships.

The *Carnegie* section gives a transport to the east by the Equatorial Countercurrent of approximately 25 million m^3/sec; the volume transport of this countercurrent in the Pacific is therefore comparable to that of the Florida Current. The surface observations seem to indicate that the transport is somewhat less in the western part of the ocean and that water is drawn into the current as it crosses the Pacific Ocean.

Within the Equatorial Countercurrent and between that current and the Equator, a distinct transverse circulation is superimposed upon the flow toward the east or the west. The character of this transverse circulation is evident, particularly from the distribution of salinity, oxygen, phosphate, and silicate (figs. 198B, D, E, and F). The arrows shown in the figures have been derived from the distribution of oxygen and have been transferred to the other representations, where they fit equally well with the course of the isolines. Within the Equatorial Countercurrent descending motion takes place at the southern boundary and ascending motion at the northern boundary, and between the Equator and the countercurrent descending motion takes place at the boundary of the countercurrent and ascending motion at the Equator. Thus, two cells appear with divergence at the northern limit of the countercurrent and at the Equator, and with a convergence at the southern boundary of the countercurrent. This system, which appears so clearly in the *Carnegie* section, is quite similar to the one which Defant (1936b) has derived for the Equatorial Countercurrent in the Atlantic Ocean on the basis of theoretical considerations (fig. 172, p. 635).

Owing to the convergence and the divergences the water does not flow due east or due west; but a spiral motion is superimposed upon the major current, and within the countercurrent this motion carries water from the northern to the southern boundary at the surface and carries water in the opposite direction at depths between 50 and 200 m. Within the Equatorial Current the surface water moves from the Equator toward the countercurrent, but at depths between 100 and 150 m the water moves in the opposite direction. To the north of the Equatorial Countercurrent and to the south of the Equator, subsurface water moves toward the divergences at the surface, and this water must originate from the regions of the Tropical Convergences which lie outside the section under consideration. According to Defant's estimate the maximum north-south component of velocity is not more than one fifth of the east-west component.

The divergences at the Equator and at the northern boundary of the countercurrent must be expected to be regions of high productivity, because the ascending motion brings water which is rich in plant nutrients into the euphotic zone. The plankton collections made on board the *Carnegie* when crossing the Equator confirm this expectation (see p. 788). The region in the vicinity and to the west of the Galapagos Islands is known to be abundant in marine life, but observations are not detailed enough to show whether two maxima of abundance of marine life occur, related to the two divergences. From the above discussion it is evident that the equatorial divergence, with all its consequences, is not related to the proximity of land, and that the conditions met with in the Galapagos area and to the west do not simply represent a continuation of the conditions off the coast of Peru, as had previously been assumed.

The character of the Equatorial Countercurrent is complicated both at the origin of the current between New Guinea and the Philippines and at the termination against the American coast. Schott (1939) has shown that large seasonal changes take place to the north of New Guinea. From June to August the South Equatorial Current follows the north coast of this Island and converges sharply with the North Equatorial Current in about lat. 5°N, where the countercurrent begins. From December to February, part of the North Equatorial Current bends completely around off the southern islands of the Philippines, sending one branch toward the southeast along the north coast of New Guinea, and another branch, the countercurrent, toward the east.

Similarly, great seasonal variations occur in the Central American region, as is evident from the data presented on the United States Hydrographic Pilot Charts, but in this region the picture is complicated by numerous eddies, the locations of which appear to vary from one year to another. The only persistent features are that the Equatorial Countercurrent is in most months well developed between lat. 5°N and 6°N and that the greater volume of water transported by the Countercurrent is deflected to the north and northwest, where a strong current prevails off the coast of Central America. Another branch, weaker and much more irregular, turns to the south. In the Gulf of Panama large seasonal changes occur which are associated with the change in the prevailing wind direction (Fleming, 1941).

The North Pacific Ocean

Water Masses of the North Pacific Ocean. In the North Pacific Ocean one encounters a Subarctic Water mass which is characterized in about lat. 50°N by an average temperature between 2° and 4° and by a salinity which at the surface may be as low as 32.00 ‰ but increases to approximately 34.00 $^0/_{00}$ at a depth of a few hundred meters and below that depth increases slowly to about 34.65 $^0/_{00}$ at the bottom. In general this water mass is carried toward the east. When reaching the American west coast it is deflected toward the south, where it enters a region of different climatic conditions. Here the temperature of the upper layers is raised by heating and the salinity is increased by excess evaporation and mixing, so that the *T-S* curves gradually swing toward the right, as shown in fig. 199, right.

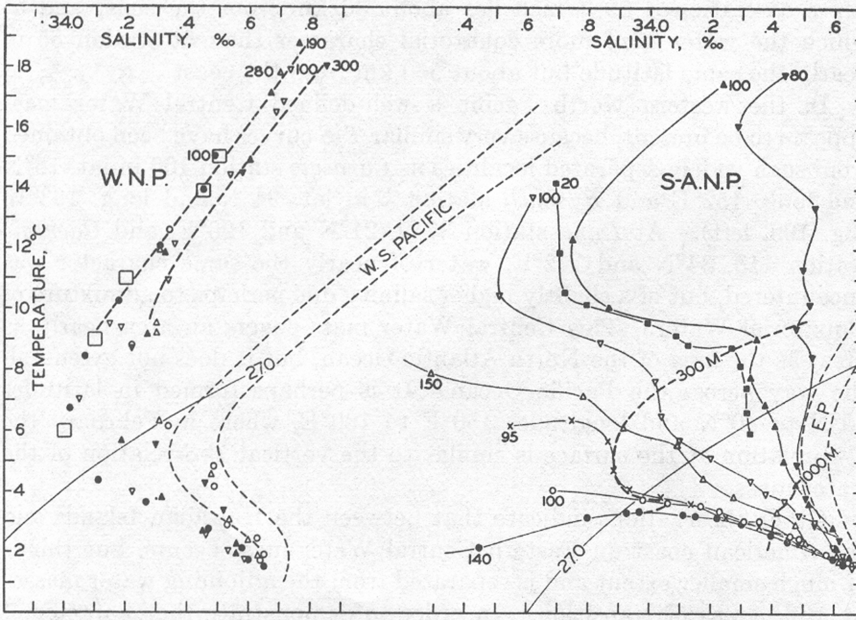

Fig. 199. Temperature-salinity relations in the North Pacific Ocean. The areas of the Subarctic and the Western North Pacific waters and the locations of stations are shown in fig. 196. The depths of the shallowest values are entered, and to the left the large squares represent winter surface values near the Subtropical Convergence. W.N.P., Western North Pacific; S-A.N.P., Subarctic North Pacific.

The term Subarctic Water should be applied strictly only to the water to the north of 45°N, but for the sake of convenience the name will be used for the entire North Pacific Subarctic Water mass, the character of which is illustrated to the right in fig. 199. In about lat. 23°N, off the south coast of Lower California, the Subarctic Water converges with the Equatorial Water. At Station B 307 in lat. 22°22′N, nearly pure Equatorial Water was present; at Station S 79 in lat. 25°22′N the water below a depth of 300 m, where a temperature of 9° and a salinity of 34.35 $^0/_{00}$ was observed, approached the equatorial type. The bend in the *T-S* curve at that station suggests that an intrusion of Equatorial Water takes place below the depth of 300 m where the bend occurs.

Below 300 m the stations between lat. 22°N and 45°N show *T-S* curves intermediate in character between those characteristic of the true Subarctic Water of the North Pacific and the Equatorial Water; this fact suggests that the water at these stations is formed by lateral mixing between those two large and well-defined water masses. Such mixing is facilitated by a northward penetration along the coast of water that contains a large component of Equatorial Water (Sverdrup and Fleming, 1941, Tibby, 1941). This penetration is shown in fig. 199 by the *T-S* curve at station S 59, which lies about 30 km from the coast and at which the water is of more equatorial character than at station 55 in nearly the same latitude but about 550 km from the coast.

In the western North Pacific a well-defined Central Water mass appears to be present, because very similar *T-S* curves have been obtained from such widely separated localities as *Carnegie* station 106 in lat. 16°N and long. 152°E and *Bushnell* station *k* in lat. 24°N and long. 163°W (fig. 199, left). At *Dana* station 3719, 21°N and 126°E, and *Carnegie* station 113, 34°N and 142°E, water of nearly the same character was encountered, but of a slightly higher salinity due perhaps to admixture of Equatorial Water. This Central Water mass covers an area nearly as great as the area of the North Atlantic Ocean, but it does not extend all the way across the Pacific Ocean. It is perhaps formed in latitudes 30°N to 40°N and longitudes 150°E to 160°E, where in February the *T-S* relation at the surface is similar to the vertical *T-S* relation of the water mass.

Other observations indicate that between the Hawaiian Islands and the American coast an Eastern Central Water mass occurs, but this is of much smaller extent and is separated from the adjoining water masses by wide areas of transition. In order to demonstrate the character of this water mass, the *T-S* curves have been plotted from a series of stations between San Diego and the Hawaiian Islands occupied by the *Bushnell* in 1939, and from a series of stations between Dutch Harbor, Aleutian Islands, and the Hawaiian Islands occupied by the *Bushnell* in 1934 (fig. 200).

At the two stations closest to the coast of California, Subarctic Water was present of the same character as that shown by the curves in fig. 199. Between stations 309 and 310 a very marked transition took place; at station 309 the water of a temperature of 15° had a salinity of 33.55 ‰, whereas at station 310 the corresponding salinity was 34.40 ‰. In the central area between the coast and the Hawaiian Islands the *T-S* curves show considerable similarity to each other, but the slope of the curves is much greater than the slope of the corresponding curves which characterize the Western Central Water mass. In order to demonstrate this feature the *T-S* relations of the latter water mass have been indicated in the figure. Close to the Hawaiian Islands station 316 shows considerable admixture of Equatorial Water, particularly at temperatures lower than 10°. This *T-S* curve is quite similar to the one which was obtained at *Carnegie* station 150 to the east-southeast of the Hawaiian Islands in lat. 16°N, long. 137°W. It appears, therefore, that the northern boundary of Equatorial Water must occasionally nearly reach the Hawaiian Islands and that the region around the islands is a boundary region within which water masses of very different character may be encountered.

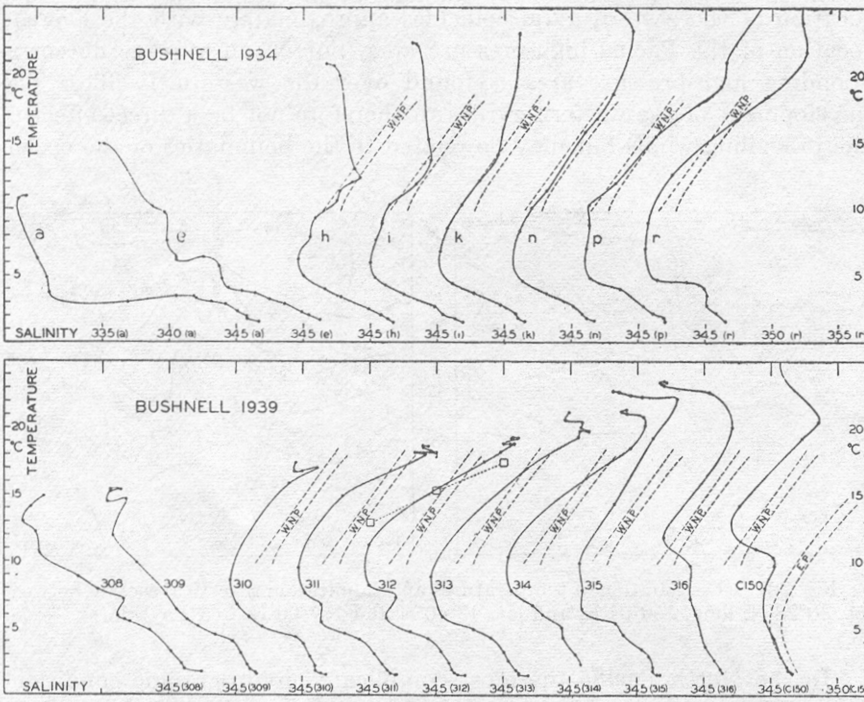

Fig. 200. *Upper: T-S* curves at a number of *Bushnell* stations of 1934 between the Aleutian and the Hawaiian Islands. *Lower: T-S* curves at a number of *Bushnell* stations of 1939 between San Diego, California, and the Hawaiian Islands, and at station *Carnegie* 150 to the southeast of the Hawaiian Islands. The *T-S* relations of the Western North Pacific and the Equatorial Pacific are shown. Large squares represent winter surface values in about lat. 40°N, long. 140°W. The stations used are located along B34 and B39 in fig. 196.

Proceeding south from the Aleutian Islands, the northern stations show Subarctic Water, but between stations *e* and *h* a transition takes place. At station *e* Subarctic Water is found, but at station *h* Western Central Water is present, as is evident from the lines marked WNP in the figure. This same water mass is present at stations *i*, *k*, and *n*, but between stations *n* and *p* a new transition takes place, water more similar to the Eastern Central Water being present at stations *p* and *r*.

From this description it is evident that in the North Pacific (as in the South Pacific, see p. 701) we encounter three typically different water masses which are separated by regions of transition: the Subarctic Water mass which penetrates toward the south along the west coast of North America and two Central Water masses of which the western occupies the larger area. The existence of these two Central Water masses indicates that in the North Pacific the circulation is also characterized by the presence of two separate gyrals of which the western is the larger. The location of the eastern gyral coincides approximately with the average location of the Pacific high-pressure area, but on an average no corresponding high-pressure area is found over the western Pacific. The development of the western gyral can therefore not be a direct effect of the prevailing winds but may be related to the boundaries of the ocean.

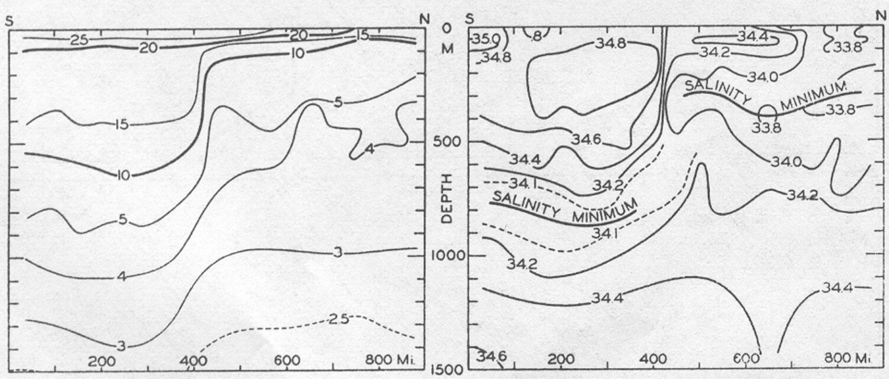

Fig. 201. Distribution of temperatures and salinities in a vertical section between lat. 30°28′N, long. 150°04′E, and lat. 42°40′N and 157°00′E. (After Uda.)

In the North Pacific the most complicated hydrographic conditions are encountered in the waters to the northeast of Japan, where the Subarctic Water carried south by the cold Oyashio meets the warm Central Water carried northeast by the Kuroshio. The Kuroshio follows closely along the coast of Japan as far as lat. 35°, where it turns east, sending branches off toward the northeast, where intensive mixing with the waters of the Oyashio takes place. Numerous eddies are present and the different branches of the current are often separated by sharp lines of demarcation, which are visible as "tide rips" on the sea surface (Uda, 1938a).

The sharp transition from one water mass to another is illustrated in fig. 201, showing temperatures and salinities in a vertical section between the points lat. 30°28′N, long. 150°04′E and lat. 42°40′N, long. 157°0′E, about 1000 km from the coast. The waters of the Kuroshio are present to the south of lat. 36°N, where an abrupt transition takes place to waters which, above 200 m, represent Kuroshio water greatly diluted by

Subarctic Water. Still further north, in lat. 41°N, is another transition to the typical Subarctic Water mass.

Both to the north and to the south of lat. 36°N an Intermediate Water is present, characterized by a salinity minimum. Below the Kuroshio water the salinity minimum is found at a depth of about 800 m, and the lowest salinity values are between 34.00 $^0/_{00}$ and 34.1 $^0/_{00}$, whereas below the mixed water to the north of lat. 36°N the salinity minimum is found at a depth of 300 m, the lowest salinities being less than 33.8 $^0/_{00}$. The oxygen content of the Northern Intermediate Water is considerably higher than that of the Southern.

Intermediate Water is present below the Central Water masses all over the North Pacific Ocean. On the basis of the *Carnegie* data, the

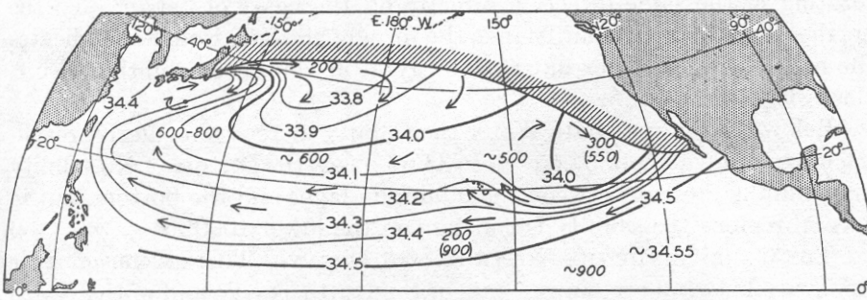

Fig. 202. Salinity in the layer of minimum salinity of the North Pacific. The direction of flow is indicated by arrows. The approximate depths (in meters) of the minimum values are shown. Where two minima are present, one of the depths is entered in parenthesis.

Bushnell data, and the numerous observations on Japanese expeditions in the western part of the Pacific, the somewhat schematic picture in fig. 202 has been prepared, showing the value of the salinity at the minimum layer. Smooth curves have been drawn in the western area in order to bring out the main features, omitting the numerous details which would confuse the presentation. Within the Subarctic Water no salinity minimum is present; the curves have therefore been carried only to the boundary region between the Subarctic Water and the Central Water masses. The depth of the layer of minimum salinity is indicated by a few numerical values. In lat. 10°N two depths have been entered because there the layer of salinity minimum appears to divide, one branch rising toward the surface at the northern limit of the Equatorial Countercurrent, and one branch sinking and approaching the minimum layer of the Equatorial Water mass (see fig. 198). Similarly two depths are entered in lat. 28°N, long. 130°W, because several stations in that region show a double minimum (see fig. 200).

In order to understand the distribution of salinity within the minimum layer it is necessary to discuss the probable flow of the water, to which we shall return when dealing with the currents. The high salinity values along the western boundary of the ocean from the Philippine Islands to Japan show that the Intermediate Water moves toward the north in that region. The salinity minimum here is found at depths between 600 and 800 m; when encountering the Subarctic Water the greater part of the Intermediate Water probably turns around at that depth, forming a big eddy off the coast of Japan. The Intermediate Water of lower salinity, which to the north of the convergence between the Kuroshio and the Oyashio is found at depths of 300 m or less, does not flow directly south but flows toward the east, sending branches off to the south. A large clockwise gyral consequently appears to be present in the western Pacific approximately between long. 160°W and the coast of Asia, with a whirl rotating in the same direction directly off the coast of Japan. To the northeast of the Hawaiian Islands the form of the 34.0 isohaline indicates the existence of another but smaller gyral in which the rotation also is clockwise.

Below the Intermediate Water the salinity increases regularly, reaching maximum values of 34.65 ‰ to 34.67 ‰ at the bottom. No salinity maximum is present between the minimum layer and the bottom, but in several regions the salinity is constant in the lower 1000 m.

THE CURRENTS OF THE NORTH PACIFIC OCEAN. There is considerable similarity between certain currents of the North Pacific and of the North Atlantic Oceans, but there are also striking differences due mainly to the occurrence of large quantities of Subarctic Water in the North Pacific, as contrasted to the small amounts of that type in the North Atlantic. The Subarctic Water of the Pacific and the currents that carry Subarctic Water are present in the northern and eastern areas and similarity to the North Atlantic is therefore found in the southern and western part of the ocean, where the North Equatorial Currents correspond to each other and where the Kuroshio corresponds to the Florida Current and the Gulf Stream.

The North Equatorial Current of the Pacific Ocean runs from east to west, increasing in volume transport because new water masses join the current from the north. The very beginning of the North Equatorial Current is found where the waters of the Equatorial Countercurrent turn to the north off Central America. To these water masses are later added the waters of the California Current, which have attained a relatively higher temperature and salinity owing to heating and evaporation, and which have been mixed with waters of the tropical region. Between the American coast and the Hawaiian Islands, Eastern North Pacific Water is added to the equatorial flow, and to the west of the Hawaiian Islands a considerable addition of Western North Pacific Water appears to take place. In about long. 160°W the volume transport of the North Equatorial Current above a depth of 1000 m is, according to *Carnegie* observations, about 45 million m^3/sec, and this value is approximately equal to the maximum transport of the corresponding current in the Atlantic Ocean. No details are known as to the character of the current, and only a few isolated computations of velocities can be made. These indicate that one has to deal with a broad and relatively deep current within which the velocities are mostly less than 20 cm/sec. A vertical transverse circulation is present, with ascending motion at the northern boundary of the Equatorial Countercurrent and with descending motion at the Tropical Convergence. The waters therefore move in a spiral-like fashion and the motion is probably highly complicated owing to the great distance between the regions of major divergence and convergence and to the presence of local divergences and convergences.

It is probable that, before reaching the western boundary of the ocean, the Equatorial Current begins to branch off, mainly to the north but partly to the south, feeding the countercurrent. To the east of the Philippine Islands a definite division of the current takes place, one branch turning south along the coast of the island of Mindanao and the larger branch turning north, following closely the east side of the northern Philippine Islands and the island of Formosa. The intensity of the flow to the south varies with the season (p. 712) and so probably does the flow to the north.

After having passed the island of Formosa, the warm waters continue to the northeast between the submarine ridge on which the Riukiu Islands lie and the shallow areas of the China Sea. On reaching lat. 30°N the current bends to the east and then to the northeast, following closely the coast of Japan as far as lat. 35°N. The name Kuroshio is particularly applied to the current between Formosa and lat. 35°N, but in agreement with the nomenclature used when dealing with the currents of the Atlantic, and following Wüst (1936), we shall apply the name "Kuroshio System" to all branches of the current system and shall introduce the following subdivisions:

1. The Kuroshio. The current running northeast from Formosa to Riukiu and then close to the coast of Japan as far as lat. 35°N.

2. The Kuroshio Extension. The warm current which represents the direct continuation of the current and flows nearly due east, probably in two branches, and can be traced distinctly to about long. 160°E.

3. The North Pacific Current. The further continuation of the Kuroshio Extension which flows toward the east, sending branches to the south and reaching probably as far as long. 150°W.

To these three major divisions of the Kuroshio System can also be added the *Tsushima Current*, the warm current that branches off on the left-hand side of the Kuroshio and enters the Japan Sea following the western coast of Japan to the north, and the *Kuroshio Countercurrent*, part of a large eddy on the right-hand side of the Kuroshio.

THE KUROSHIO. The great similarity between the Kuroshio and the Florida Current has been pointed out by Wüst (1936). Wüst compares the Kuroshio between the Riukiu Islands and the continental shelf to the current through the Caribbean Sea and the flow of the Kuroshio across the Riukiu Ridge in lat. 30°N to the flow through the Straits of Florida. This part of the Kuroshio, as well as the current to the south of Shiono-misaki in lat. 33°N, long. 135°W, has been studied by Japanese oceanographers, who have conducted current measurements and have occupied a large number of oceanographic stations. The results have been discussed mainly by Wüst (1936) and Koenuma (1939). According to Koenuma, the current between Formosa and the southern of the Riukiu Islands reaches to a depth of about 700 m and the maximum velocity near the surface is 89 cm/sec. These figures are based on computations, but are in good agreement with direct measurements at one station in the passage. The transport amounts to about 20 million m^3/sec. Similar conditions are encountered further north, between the northern of the Riukiu Islands and the continental shelf; here, however, a weak countercurrent appears to be present on the left-hand side of the main flow. The maximum velocities in this profile are somewhat above 80 cm/sec and the computed transport is 23 million m^3/sec.

The profile to the south of Shiono-misaki has been discussed both by Wüst and Koenuma, who do not agree in their interpretation of the measurements, particularly as to the depth of the level of no motion. Wüst assumed that the level of no motion coincides approximately with the 10° isothermal surface, which nearly represents the upper boundary of the Intermediate Water; but this assumption leads to computed velocities which are considerably below the observed ones. Koenuma, on the other hand, assumes that close to the coast the observed velocity of the Intermediate Water, 16 cm/sec to the northeast, is correct, and furthermore assumes that at greater distances from the coast the Intermediate Water flows toward the southwest at a velocity of about 5 cm/sec. These assumptions are based on a number of considerations that appear to be valid. In this manner Koenuma arrives at a computed velocity distribution in good agreement with observations. The Kuroshio runs here very close to the coast with maximum velocities up to 160–180 cm/sec (up to 3½ knots), but extends only to a distance of 140 km from the coast. At greater distances a countercurrent flows to the southwest within which maximum velocities up to 20 cm/sec are encountered, in agreement with ships' observations. The character of the Kuroshio is here closely related to that of the Florida Current to the south of Cape Hatteras. The transport of the Kuroshio between the coast and its outer limit is greatly increased, but the countercurrent on the right-hand side carries large amounts of water in the opposite direction. It is probable that a considerable annual variation takes place in the transport, but so far little information is available on the subject.

The temperatures of the Kuroshio water are comparable to those of the Florida Current but the salinities are much lower, the maximum salinity in the Kuroshio being slightly less than 35.00 $^{0}/_{00}$, whereas the maximum salinity in the Florida Current is about 36.50 $^{0}/_{00}$. This difference reflects the general lower salinity of the Pacific as compared to the Atlantic. The temperature of the Kuroshio is subject to a large annual variation (p. 131) related to excessive cooling in winter by cold offshore winds. Off Shiono-misaki the annual range at the surface amounts to nearly 9° and at 100 m the range is still nearly 4.5°.

THE KUROSHIO EXTENSION. In lat. 35°N, where the Kuroshio leaves the coast of Japan, it divides into two branches; one major branch turns due east and retains its character as a well-defined flow as far as approximately long. 160°E, and one continues toward the northeast as far as lat. 40°N where it bends toward the east. The major branch is evident on charts showing the anomaly of the surface temperature or the difference between air and surface temperatures in winter. According to Schott (1935) a tongue within which this difference is greater than 4°C extends east toward long. 170°E, that is, beyond the eastern limit of the well-defined flow. Between long. 155° and 160°E considerable water masses turn toward the south and southwest, forming part of the Kuroshio countercurrent which runs at a distance of approximately 650 km from the coast as the eastern branch of a large whirl on the right-hand side of the Kuroshio.

According to the vertical sections of temperature and salinity in fig. 201, the Kuroshio Extension is, in long. 153°E, still a narrow current, but continuing toward the east a section in long. 162°E (Uda, 1935) shows that the current has been broken up into a number of branches separated by eddies and countercurrents. The change corresponds to the one which, in the Atlantic Ocean, takes place between the regions to the south of the Grand Banks and to the north of the Azores (see fig. 186, p. 681).

The northern branch of the Kuroshio Extension becomes rapidly mixed with the cold waters of the Oyashio that flow south close to the northeastern coasts of Japan, reaching nearly to 35°N. The extensive work of Japanese oceanographers (Uda, 1935, 1938b) shows that along the boundary between the Kuroshio Extension and the Oyashio numerous eddies develop, within which a thorough mixing of the water masses takes place. From the sections in fig. 201 it is evident that to the north of lat. 36°N the Kuroshio waters, which can be traced to lat. 41°N, have been greatly diluted by the Oyashio water. These processes of mixing gradually form the water mass that is present in the northwestern

Pacific Ocean, which has been called the Subarctic Water of the North Pacific (fig. 199, p. 713). These sections bring out another feature mentioned in the discussion of the Intermediate Water. To the north of lat. 36°N Intermediate Water of a salinity as low as 33.8 ‰ is found at a depth of 300 m or less. This water, which contains a considerable amount of oxygen, is probably formed in winter at the convergence between the Kuroshio Extension and the Oyashio and sinks from the surface in a manner similar to the sinking of the Antarctic Intermediate Water. From the region of sinking this Intermediate Water flows toward the east and spreads over the greater part of the North Pacific (fig. 202). To the south of lat. 36°N Intermediate Water of salinity between 34.0 ‰ and 34.1 ‰ is present at a depth of about 800 m, but this water represents the most northern extension of the Intermediate Water that flows north along the coast of Japan and turns around as part of the big whirl on the right-hand side of the Kuroshio.

THE NORTH PACIFIC CURRENT. Under this name is understood the general eastward flow of warm water to the east of long. 160°E. Details of this flow are not known, but from the *Bushnell* section of 1934 (fig. 200) it is evident that Kuroshio water crosses long. 170°W, because at a number of *Bushnell* stations the *T-S* curves are similar to those of the Kuroshio Water. The greater part of this water appears to turn around toward the south before reaching 150°W, and only a small portion continues and flows south between the Hawaiian Islands and the west coast of North America after having been mixed with waters of different origin. The main part of the North Pacific Current does not, therefore, extend across the Pacific Ocean but turns back toward the west in the longitude of the Hawaiian Islands.

THE ALEUTIAN (SUBARCTIC) CURRENT. To the north of the North Pacific Current one finds a marked transition to an entirely different type of water, the Subarctic Water, which also flows toward the east. According to the *Bushnell* observations the transport of Subarctic Water between the Aleutian Islands and lat. 42°N amounts to about 15 million m^3/sec above the 2000-decibar surface. This water mass must have been formed by mixing of Kuroshio and Oyashio water, the temperature of the mixture having been reduced by cooling and the salinity of the upper layer decreased by excessive precipitation. One branch of the Aleutian Current turns north and enters the Bering Sea, following along the northern side of the Aleutian Islands and circling the Bering Sea counterclockwise. In the Bering Sea these waters are further cooled, and flowing south they reach the northern islands of Japan as the cold Oyashio. The amount of water in this gyral is not known, but inflow and outflow from the Bering Sea must be nearly equal, wherefore it follows that the 15 million m^3/sec of Subarctic Water which continues east on the southern side of the Aleutian Islands are supplied by the Kuroshio, the waters of which have been completely transformed by mixing and external influences.

Before reaching the American coast the Aleutian Current divides, sending one branch toward the north into the Gulf of Alaska and another branch toward the south along the west coast of the United States. The former branch is part of the counterclockwise gyral in the Gulf of Alaska. It enters the gulf along the American west coast, and since it comes from the south it has the character of a warm current in spite of the fact that it carries Subarctic Water. It therefore exercises an influence on the climatic conditions similar, on a small scale, to that which the North Atlantic and the Norwegian Currents exercise on the climate of northwestern Europe. The major branch, which turns toward the south along the west coast of the United States, is known as the California Current, but before dealing with this it is desirable to discuss the warm-water currents between the Hawaiian Islands and the American West Coast.

THE EASTERN GYRAL IN THE NORTH PACIFIC OCEAN. The existence of an eastern gyral is recognized mainly by the character of the water masses and by the results of computations based on observations between the Hawaiian Islands and the coast of California and between the Hawaiian Islands and the Aleutian Islands. The study of water masses (fig. 200, p. 715) brought out that a distinctly different water mass was present in the region to the east of the Hawaiian Islands, a mass in part formed at the boundary between the warmer waters and the Subarctic Water between long. 130° and 150°W. Computation of currents and transport, the results of which are shown schematically in fig. 205, p. 727, leads to the conclusion that a clockwise rotating gyral is present in the eastern North Pacific with its center to the northeast of the Hawaiian Islands. It is probable that the location of this gyral changes with the seasons and shifts from year to year, so that occasionally the gyral may lie entirely to the northeast of the Hawaiian Islands, whereas in other circumstances the Hawaiian Islands may lie inside the gyral. If the gyral is displaced considerably to the north, Equatorial Water may reach as far north as to the Hawaiian Islands, as was observed in 1939 at *Bushnell* station 316 (see fig. 200, p. 715). At the surface the gyral is masked by wind-driven currents.

The existence of this gyral is confirmed by the values of the salinity at the salinity maximum (fig. 202). The high values which are found to the northeast of the Hawaiian Islands must be associated with a flow to the northeast, and the presence of two salinity minima in about lat. 28°N and long. 130°W must be related to the circulation. The upper minimum at which the salinity is about 33.95 ‰ is probably formed where the Central Water mass spreads over the Subarctic Water; the lower minimum at which the salinity is about 34.10 ‰ represents the direct continuation of the minimum in the northeast-flowing branch of the gyral. In the southern branch of the gyral the minima merge to a single one at about 500 m.

THE CALIFORNIA CURRENT. The California Current represents, as already explained, the continuation of the Aleutian Current of the North Pacific. The name is applied to the southward flow between lat. 48° and 23°N, where the Subarctic Water converges with the Equatorial (fig. 199, p. 713). The outer limit of the California Current is represented by the boundary region between the Subarctic Water and the Eastern North Pacific Central Water, and lies in lat. 32°N at a distance of approximately 700 km from the coast, according to the observations by the *Carnegie* in 1929, the *Louisville* in 1936, and the *Bushnell* in 1939. The total volume transport of the California Current above the 1500-decibar surface is probably not more than about 10 million m³/sec, and in view of the great width of the current no high velocities are encountered except within local eddies. As a whole, the current represents a wide body of water which moves sluggishly toward the southeast.

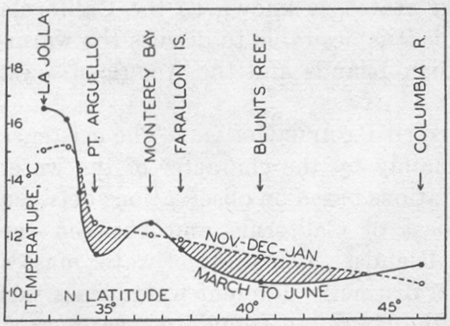

Fig. 203. Surface temperatures along the coast of California in March to June and in November to January. In regions of intense upwelling, the average surface temperature is lower in March to June than in December to January.

In spring and early summer the California Current is a counterpart to the Peru Current, several characteristic features of the two currents being strikingly similar. During these months north-northwest winds prevail off the coast of California, giving rise to upwelling that mostly begins in March and continues more or less uninterruptedly until July. Records of surface temperatures show that on the coast the lowest temperatures regularly occur in certain localities separated by regions with higher surface temperatures. This is demonstrated by fig. 203, in which are plotted the surface temperatures between lat. 32°N and 45°N along the American west coast in March to June and in November, December, and January. In the regions of intense upwelling the spring temperatures are lower than the winter temperatures, but in regions of less intense upwelling they are higher.

Recent work of the Scripps Institution of Oceanography shows that from the areas of intense upwelling tongues of water of low temperature extend in a southerly direction away from the coast; these tongues are separated from each other by tongues of higher temperature extending in toward the coast. Within the tongues of higher temperature the flow is directed to the north, whereas within the tongues of low temperature the flow is directed to the south. Swirls thus appear on the coastal side of the current, similar to those which Gunther has demonstrated within the Peru Current.

To the north of lat. 30°N the two most conspicuous centers of upwelling are located in lat. 35°N and 41°N. The two swirls associated with these centers of upwelling and the alternating movements away from and toward the coast are shown in fig. 204, which is based on observations in May to July, 1939. The coast to the south of 34°N is under the influence of water which returns after having traveled a long distance as surface water, and, correspondingly, the surface temperatures are much higher there than those encountered where the water was recently drawn to the surface (fig. 203). A third region of intense upwelling is found perhaps in about latitude 24°N, on the coast of Lower California (Thorade, 1909).

The upwelling water rises from moderate depths only, probably less than 200 m (Sverdrup and Fleming, 1941), and the phenomenon therefore represents only an overturning of the upper layers such as is the case in other regions which have been examined. According to McEwen (1934) the rate of upwelling is about 20 m a month and on an average the process is therefore a very slow one. It is, however, of the greatest importance to the productivity of the waters off the coast, because the rising water brings nutrient salts into the euphotic zone. The importance of upwelling to the distribution of phytoplankton (p. 785) has been discussed by Sverdrup and Allen (1939).

During the entire season of upwelling a countercurrent that contains considerable quantities of Equatorial Water flows close to the coast at depths below 200 m (Sverdrup and Fleming, 1941). This subsurface countercurrent appears to be analogous to the subsurface countercurrent off the coast of Peru, the existence of which was shown by Gunther (p. 703). In spring and early summer the currents off the coast of California are therefore nearly a mirror image of those off the coast of Peru, but the similarity is found in this season only because the character of the prevailing winds off California changes in summer, whereas off Peru the winds blow from nearly the same direction throughout the year.

Toward the end of the summer the upwelling gradually ceases and the more or less regular pattern of currents flowing away from and toward the coast breaks down into a number of irregular eddies, some of which carry coastal waters far out into the ocean (Johnson, 1939). Other eddies carry oceanic waters in toward the coast, particularly in the regions between the centers of upwelling, as shown by Skogsberg (1936) in his discussion of the waters of Monterey Bay (p. 131).

In the fall upwelling ceases, and in the surface layers a countercurrent develops, the Davidson Current which in November, December, and

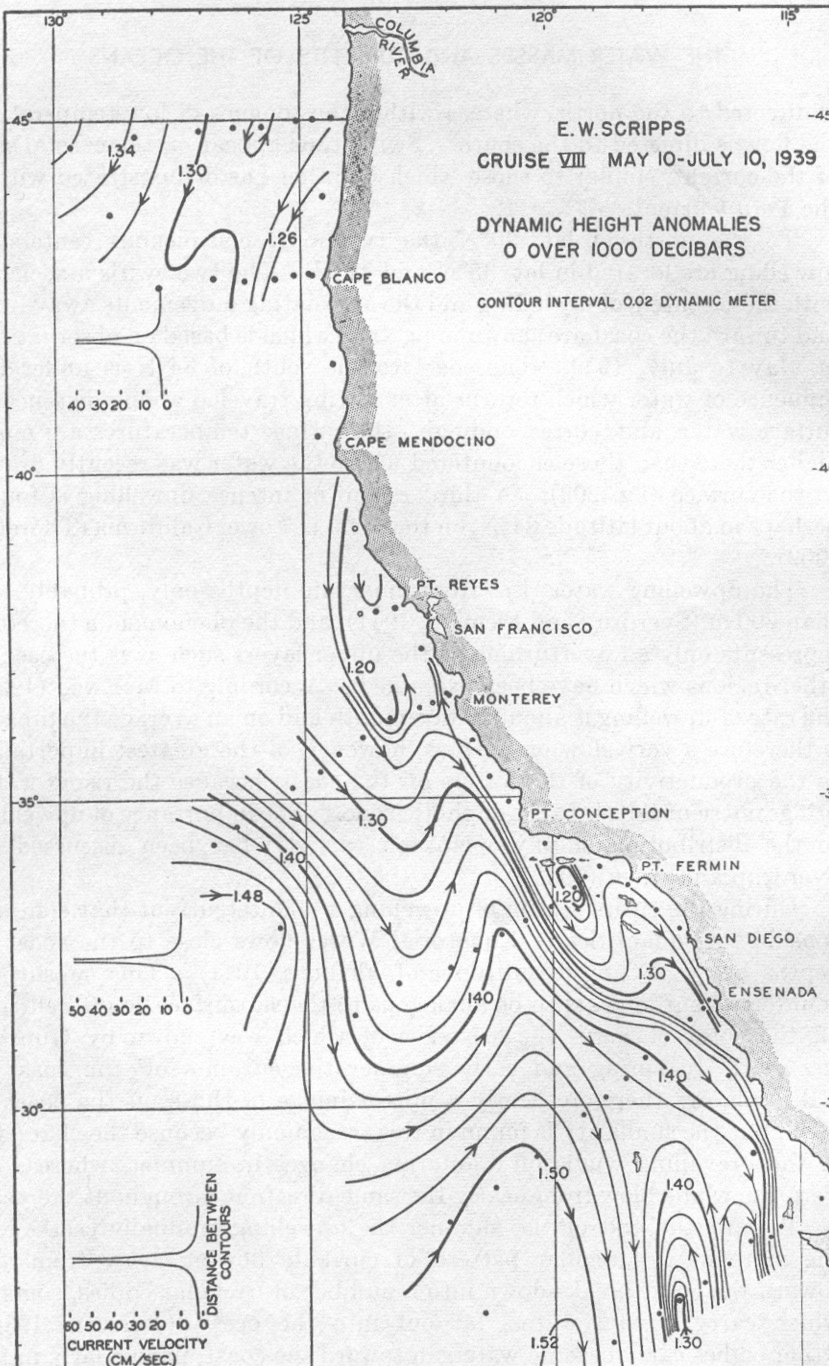

Fig. 204. Geopotential topography of the sea surface relative to the 1000-decibar surface off the American west coast in May to July, 1939, showing flow alternating away from and towards the coast.

January runs north along the coast to at least lat. 48°N. In this season the subsurface countercurrent still exists and the main difference between the seasons without and with upwelling is therefore that in the former a countercurrent is present at all depths on the coastal side of the California Current, whereas when upwelling takes place the countercurrent has disappeared in the surface layer where, instead, a number of long-stretched swirls have developed. This difference suggests that in the absence of prevailing winds that cause upwelling, a countercurrent would appear on the coastal side, as is the case in other localities (see p. 677). In the presence of upwelling the overturn of the surface layers destroys the countercurrent above a depth of 200 m and leads to an entirely different pattern of flow.

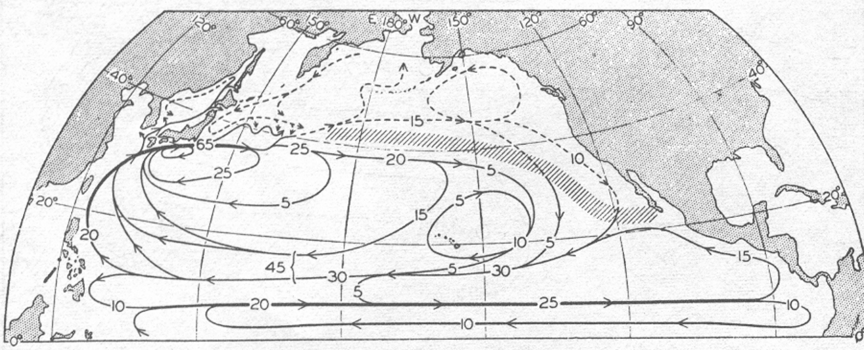

Fig. 205. Transport chart of the North Pacific. The lines with arrows indicate the approximate direction of the transport above 1500 meters, and the inserted numbers indicate the transported volumes in millions of cubic meters per second. Dashed lines show cold currents; full-drawn lines show warm currents.

Transport. On the basis of the values of the volume transport of the different branches of the current system that have been mentioned and others that have been computed, the schematic picture in fig. 205 has been prepared. The lines with arrows give the approximate direction of the transport and the numbers give the volume transport in millions of m^3/sec. In this case the transport numbers include the motion of the Upper and the Intermediate Water because the Intermediate Water of the Pacific appears to flow in general in the same direction as the Upper Water (fig. 202). The lines showing the direction of transport are full-drawn where the upper water masses are warm and dashed where they are cold.

The figure brings out that in the North Pacific Ocean the Equatorial Countercurrent and the Kuroshio are the two outstanding, well-defined currents. Over the greater part of the Pacific Ocean weak or changing currents are present, and the transport numbers that are entered refer therefore to broad cross sections. The figure intends only to bring out the major features which have been discussed, representing the first attempt at a synthesis of the available information as to the circulation in the North Pacific Ocean.

The Oxygen Distribution in the Pacific Ocean. In the South Pacific Ocean no oxygen observations are available from the central and eastern parts except for a few at two *Carnegie* stations in about lat. 33°S and long. 110°W. On board the *William Scoresby* oxygen determinations were made in the waters of the Peru Coastal Current, but so far the data have not been published. From the western South Pacific and from the equatorial part of the South Pacific oxygen values are available at a number of *Dana* and *Carnegie* stations.

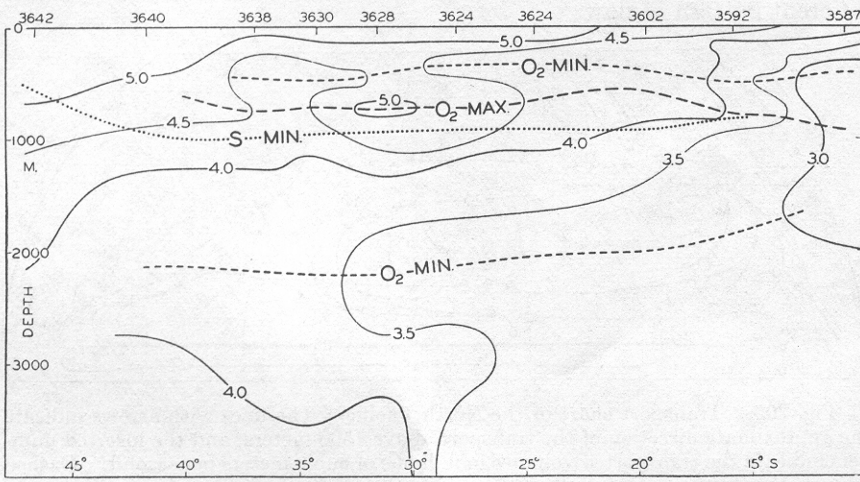

Fig. 206. Oxygen distribution (ml/L) in a vertical section along the 180° meridian between latitudes 48°S and 10°S. (According to observations of the *Dana*.)

Figure 206 shows the distribution of oxygen in a vertical section which nearly follows the meridian of 180° between latitudes 48°S and 10°S. Near the Antarctic the oxygen content of the water is generally high, decreasing more or less regularly from the surface to a depth of about 2000 m, where the content is a little below 4 ml/L. At greater depth a small increase takes place, such that below 2500 m the content is greater than 4 ml/L. To the north of 40°S an oxygen minimum is present at about 400 m, but below this minimum an intermediate maximum is found between 600 and 800 m. This intermediate maximum lies a few hundred meters above the core of the Antarctic Intermediate Water, which is indicated by the line marked S-min. At about 2000 m a second layer of minimum oxygen is found and below this the oxygen content slowly increases toward the bottom. In a horizontal direction an abrupt decrease in the oxygen content takes place at about 15°S. At 11°S the oxygen content is below 3 ml/L between 300 m and 2000 m and a minimum value of 2.22 ml/L is found at 400 m, but the content of the deep water is as high as further to the south.

According to observations at two *Carnegie* stations in about lat. 33°S and long. 110°W (see Schott, 1935), the oxygen distribution is there quite similar to that found off New Zealand; but from analogy with conditions within the Benguela Current it appears highly probable that the oxygen content of the subsurface waters is low within the Peru Current, particularly within the Peru Coastal Current. At *Dana* stations 3559, 3560, and 3561, located to the south of the Equator between long. 104° and 117°W, minimum oxygen values as low as 0.09 ml/L were observed at 400 m. This low-oxygen water probably comes from the coast off Peru, indicating that there a minimum oxygen layer is present within which the oxygen content is even lower than that found off the coast of South Africa. Such conditions should be expected because of the great productivity of the Peru coastal waters and, consequently, the great consumption of oxygen at subsurface depths by decomposition of sinking remnants of organisms.

In the North Pacific the oxygen content of the subsurface waters is as a rule lower than in the South Pacific, but a similar contrast exists between west and east, the oxygen content in the eastern areas being lower than in the western. The lowest oxygen values are consistently found 400 or 500 m below the salinity-minimum layer where such a layer is present, and in the Subarctic Waters at a depth of 600 or 800 m. Minimum values higher than 2 ml/L are found only in the extreme western part to the west of long. 160°E and between the Equator and 36°N. Over the greater part of the North Pacific the minimum values are below 1 ml/L, and off the west coasts of Central America and Mexico the content is nearly nil. Below the minimum layer the oxygen content increases toward the bottom, at which values somewhat above 3 ml/L are found in the western part and somewhat below 3 ml/L in the eastern part.

The most striking feature of the oxygen distribution in the North Pacific is the presence of a very large body of water within which the oxygen content is exceedingly small. This body is found off the American coast between lat. 28°N and the Equator; it extends toward the west like a wedge, gradually becoming more and more narrow, and reaches at least past long. 140°W. Close to the coast the layer of oxygen content less than 0.25 ml/L is found between the depths of 200 to 300 m and 1200 m, and according to *Carnegie* observations it is found in long. 140°W between 100 and 600 m. It is probable, as already stated, that another water mass of equally low oxygen content extends toward the west from the coast of Peru, but the two water masses appear to be separated by a region of slightly higher oxygen content.

The character of the oxygen distribution off the American west coast is illustrated in fig. 207, showing a section which runs parallel to the coast at a distance of a few hundred miles and extends from lat. 45°N to the Equator. Between 20°N and 10°N is a region within which the oxygen content is so low that no trace of oxygen can be determined by the Winkler method, but hydrogen sulphide is not present. This low-oxygen water is also found in the Gulf of California (see table 83), where in some localities, owing to upwelling, the oxygen content is practically nil at a depth of less than 100 m.

A systematic study of the types of organisms which may be found in this oxygen-poor water has not yet been made. A few scattered observations indicate that organisms are found even where the oxygen content is

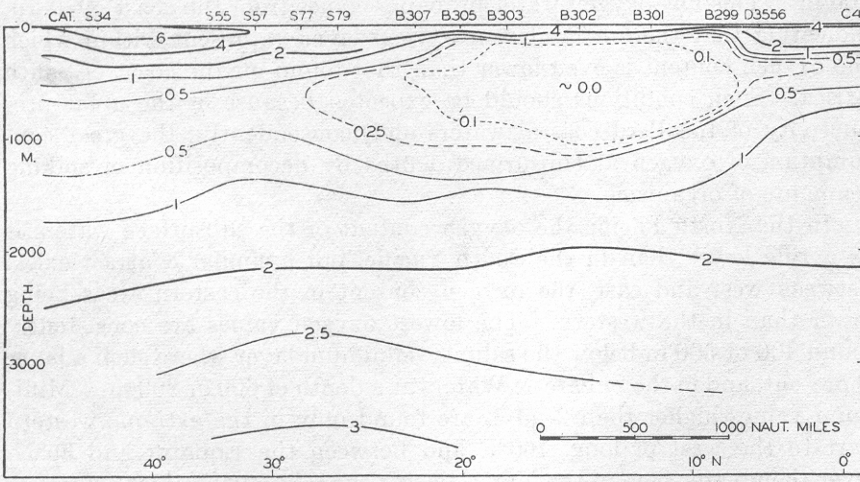

Fig. 207. Oxygen distribution (ml/L) in a vertical section along the west coast of North and Central America at a distance of a few hundred miles from the coast (mainly based on observations of *E.W. Scripps* and *Bushnell*).

as low as 0.1 ml/L; the further examination of the distribution of such organisms and their adaptation to the extreme conditions represents one of the fascinating biological problems of the Pacific Ocean.

The Adjacent Seas of the North Pacific Ocean

None of the adjacent seas of the North Pacific Ocean exercise any appreciable influence on the water masses of the ocean and it has therefore not been necessary to discuss them until now.

Gulf of California. The Gulf of California lies in a climatic region which is comparable to that in which the Red Sea is located, but hydrographically the Gulf of California is entirely different from the Red Sea, the reason being that the Gulf of California is not separated from the adjacent ocean by a submarine ridge. To 28°30′N the bottom of the Gulf is trough-shaped, with the greatest depth at the opening to the south. The soundings of the *E. W. Scripps* in 1939 and 1940 have shown, however, that the bottom is very irregular with several deeper basins, although at the entrance free communication with the waters of the Pacific exists above depths of nearly 3,000 m. A ridge running nearly north-south in long. 112°40′ separates the inner portion of the Gulf from the outer, the sill depth of the ridge being about 200 m. Inside of the ridge lies a deep trench between the peninsula of Lower California and the island of Angel de la Guardia.

The character of the waters in the Gulf is shown in table 83, containing temperature, salinity, and oxygen at three stations: station B 305 in

TABLE 83

TEMPERATURE, SALINITY, AND OXYGEN IN AND NEAR THE GULF OF CALIFORNIA

(*Bushnell* Station off the entrance to the Gulf and *E. W. Scripps* Stations in the Gulf)

Depth (m)	*Bushnell* 305 March 24, 1939 20°00′N, 108°16′W			*E. W. Scripps* VII-27 March 4, 1939 26°21′N, 110°46′W			*E. W. Scripps* VII-53 March 19, 1939 28°46.5′N, 113°08′W		
	Temp. (°C)	S (‰)	O_2 (ml/L)	Temp. (°C)	S (‰)	O_2 (ml/L)	Temp. (°C)	S (‰)	O_2 (ml/L)
0	23.78	34.88	5.06	15.95	35.22	5.37	15.80	35.12	5.43
25	22.00	35.02	5.20	15.45	.16	4.87	14.06	.07	4.90
50	17.22	34.70	1.45	15.34	.15	4.90	13.81	.08	4.27
100	13.15	.78	0.17	13.44	34.82	0.41	13.11	.01	3.13
200	11.50	.75	0.16	12.00	.78	0.25	12.50	34.95	2.32
400	8.71	.55	0.10	9.60	.65	0.11	11.83	.84	1.63
600	6.69	.52	0.07	7.26	.52	0.07	11.50	.83	1.58
800	5.45	.52	0.09	5.54	.52	0.12	11.34	.80	1.53
1000	4.70	.52	0.20	4.46	.52	0.26	11.18	.79	0.96
1500	3.10	.58	1.01	2.89	.61	0.84	(10.98	.76	0.62)[a]
2000	2.21	.63	1.84	2.52	.60	1.26			
2500	1.85	.66	2.35	(2.56	.62	1.28)[b]			
3000	1.82	.65	2.46						

[a] At 1270 m. [b] At 2400 m.

20°00′N off the entrance to the Gulf, station EWS VII-27 in the middle portion of the Gulf, and station EWS VII-53 in the trench south of Angel de la Guardia. In March, when the stations were occupied, the surface temperatures in the Gulf were lower than those off the entrance, the difference being due to cooling by northwesterly winds and to local upwelling. The salinity, on the other hand, was higher in the Gulf owing to excessive evaporation. Between depths of 100 and 1500 m the waters in the middle portion of the Gulf were nearly identical with those off the entrance. Some differences were found at 2000 m and 2400 m because station EWS VII-27 was located in one of the basins, the sill depth of which was probably about 1500 m. In the trench to the north of the transverse ridge, basin conditions existed from a depth of about 200 m and to the bottom. The bottom water, which had a temperature of about 11° and a salinity of 34.76 ‰, was probably formed by mixing of water flowing in across the sill with water sinking from the surface during periods of excessive cooling and evaporation. The admixture of surface water is indicated by the higher oxygen content as compared to the oxygen content of the water in the middle portion of the Gulf.

During the cruise of the *E. W. Scripps* in 1939 a number of stations were occupied in the shallow northern portion of the Gulf. The most northern station was located at a distance of about 70 km from the mouth of the Colorado River where the depth to the bottom was 60 m. In this locality the influence of fresh water from the Colorado River was not perceptible, the surface salinity being 35.31 ‰, whereas the highest surface salinity, 35.50 ‰, was observed at the neighboring station only about 35 km to the southeast.

No well-defined exchange of water between the Gulf and the adjacent parts of the Pacific Ocean could be established. It is probable that such an exchange takes place by irregular currents in the upper layers and by a slow inflow of deep water and outflow of surface water.

Bering Sea. In the deeper portions of the Bering Sea the water masses are similar to those in the subarctic region of the North Pacific, as is evident from table 84, showing temperatures, salinities, and oxygen values at stations on the south and the north sides of the Aleutian Islands and at a third station about 240 km to the north of the Aleutian Islands. Below a depth of about 200 m the water masses at these three stations were practically identical. At 200 m the temperature shows a minimum at all stations but, according to the more detailed original observations, the lowest temperatures were found somewhat above 200 m. At the two northern stations an oxygen maximum appeared at the depth of the temperature minimum and at the southern station the oxygen content was high. This situation suggests that the depth of the intermediate temperature minimum represents the depth to which convection currents reach in winter. Such a layer of minimum temperature has been established at a number of stations in the subarctic region and lower temperature values have been observed further toward the west, where winter cooling is more intense.

The shallow shelf areas in the eastern and northern part of the Bering Sea are covered by water of a considerably lowered salinity owing to dilution by runoff from the large Alaskan rivers. The influence of the waters of the Yukon River is particularly conspicuous. In summer warm water of low salinity is found along the northeastern coast of the Bering Sea and this water continues through Bering Strait into the Polar

Sea, where it can be traced for some distance along the coast of northwest Alaska.

Some of the water that flows into the Bering Sea around the Aleutian Islands contributes in summer to the current which flows through Bering Strait (see p. 655), but the major amount of water turns around and flows south along the coast of Kamchatka, reaching the northern of the Japanese Islands as the cold Oyashio (Barnes and Thompson, 1938).

TABLE 84

TEMPERATURE, SALINITY, AND OXYGEN AT ONE STATION TO THE SOUTH OF THE ALEUTIAN ISLANDS AND AT TWO STATIONS IN THE BERING SEA

(*Bushnell* MSS; Barnes and Thompson, 1938)

Depth (m)	Station B-a, August 18, 1934, 50°30′N, 175°16′W			Station G-8, June 13, 1933, 52°37′N, 177°20′W			Station C-107, August 21, 1934, 55°04′N, 168°49′W		
	Temp. (°C)	S (‰)	O_2 (ml/L)	Temp. (°C)	S (‰)	O_2 (ml/L)	Temp. (°C)	S (‰)	O_2 (ml/L)
0	10.91	32.92	6.24	6.57	33.28	8.85	9.77	32.50	6.33
25	10.50	.87	6.40	4.13	.33	6.82	5.52	.94	5.36
50	6.00	33.04	7.22	3.86	.31	5.65	4.26	33.12	4.90
100	3.48	.12	7.10	3.54	.37	5.07	3.77	.31	4.32
200	3.11	.75	2.60	3.25	.46	5.40	3.20	.46	4.36
400	3.42	34.13	0.63	3.43	.87	2.08	3.44	.86	2.21
600	3.22	.22	0.46	3.31	34.09	1.10	3.27	34.13	0.90
800	2.98	.34	0.44	3.11	.23	0.62	3.04	.25	0.63
1000	2.74	.44	0.58	2.89	.34	0.65	2.79	.38	0.52
1500	2.17	.52	0.96	2.31	.49	0.74			
2000	1.88	.58	1.64	1.91	.58	1.19			
2500	1.72	.59	2.20	1.70	.63	1.67			
3000	1.61	.64	2.58	1.61	.65	1.81			

OKHOTSK SEA. The few data available from the Okhotsk Sea (Krümmel, 1911) show that in winter excessive cooling of the waters takes place such that even in summer the temperature at a depth of 100 to 200 m is as low as −1.4°. The salinity of the water is also low, being about 33.1 ‰. It is probable that the cold intermediate layer of the Okhotsk Sea flows out between the Kurile Islands and contributes to the maintenance of the layer of intermediate temperature minimum in the northwestern part of the Pacific and in the Bering Sea; but there is no evidence that the salinity of the water in the Okhotsk Sea in winter is increased so much by freezing that bottom water is formed, nor does any such formation take place in Bering Sea. One finds, therefore, no region in the North Pacific in which processes go on similar to those which lead to the formation of bottom water in the Antarctic (p. 611), or similar to those which produce bottom water in the North Atlantic (p. 664).

JAPAN SEA. The Japan Sea is a basin in which the greatest depth is about 3700 m. The deepest sill is found in the Tsushima Strait between Korea and Japan, where the maximum depth is about 150 m. In the Japan Sea (Suda and Hidaka, 1932, Uda, 1934) above 400 m there exists a striking contrast between the waters along the west coast of Japan and those along the east coast of Korea. A branch of the Kuroshio, the Tsushima Current, flows into the Japan Sea and carries water of high temperature and high salinity toward the north. Branches of the current flow out through the straits between the northern Japanese islands and part of the water continues along the west side of Sakhalin Island, turns around, and flows south after having been cooled and diluted. The contrast between the waters on the east and the west sides of the Japan Sea is shown by the data in table 85. Schott (1935) points out that the cold water along the mainland side cannot come from the Okhotsk Sea through the narrow strait between the Asiatic coast and Sakhalin Island, because the strait at its narrowest is only 6.7 km wide and 12 m deep. The cold water must therefore have been formed in the Japan Sea by excessive cooling in winter and must have been diluted by river water. The water below a depth of about 400 m is of a temperature slightly above 0° and a salinity a little above 34.0 ‰.

TABLE 85

TEMPERATURES AND SALINITIES IN THE JAPAN SEA

(*Syunpu Maru*)

Depth (m)	East side, August, 1930, 41°N, 140°E; sounding, 887 m		West side, July, 1930, 41°N, 132°E; sounding, 3300 m	
	Temp. (°C)	S (‰)	Temp. (°C)	S (‰)
0	27.00	(32.68)	19.30	33.73
25	22.04	34.14	5.35	.98
50	17.73	.36	2.58	34.00
100	12.45	.47	1.23	.00
150	9.30	.30	0.74	.01
200	6.54	.16	0.50	.02
400	0.91	.04	0.24	.04
600	0.19	.03	0.18	.07
800	0.16	.02	0.13	.11
1000			0.16	.11
1500			0.15	.06

On a small scale the Japan Sea is comparable to the Arctic Mediterranean, into which flows a branch of the North Atlantic Current carrying warm water of high salinity which is cooled and diluted so that the outflowing current carries cold water of low salinity. The contrast between the eastern and western sides of the Japan Sea corresponds to the contrast between the eastern and western sides of the Norwegian Sea and also to the contrast between the eastern and western sides of the Labrador Sea. The main difference is that no great outflow of cold water takes place from the Japan Sea; the cold water on the western side is mainly part of an eddy.

Yellow Sea and East China Sea. In both of these the surface salinity is greatly reduced by runoff from rivers, and temperature and salinity alike are subjected to great annual variations. The waters are shallow and the processes that take place have small bearing on conditions at greater distances from the coast.

South China Sea. In the South China Sea, between the Philippine Islands and the Asiatic mainland, a basin is found within which the greatest depths exceed 4600 m and which is in communication with the adjacent part of the Pacific Ocean through the passage between the Philippine Islands and Formosa, where the sill depth is between 2500 and 3000 m. Table 86 contains observations of temperature, salinity, oxygen, and computed potential temperatures at a *Dana* station located

Table 86

TEMPERATURE, POTENTIAL TEMPERATURE, SALINITY, AND OXYGEN AT STATIONS IN THE SOUTH CHINA SEA AND IN THE SULU SEA

(*Dana*)

Depth (m)	*Dana* 3714, May 20, 1929, 15°22′N, 115°20′E; sounding, 4240 m				*Dana* 3685, April 4, 1929, 7°22′N, 121°16′N; sounding, 4825 m			
	Temp. (°C)	θ (°C)	S (‰)	O_2 (ml/L)	Temp. (°C)	θ (°C)	S ‰	O_2 (ml/L)
0	29.54	29.54	33.73		27.08	27.08	34.08	
50	24.07	24.06	34.04	4.57	24.37	24.36	.16	3.67
100	18.66	18.64	.52	2.33	20.93	20.91	.31	2.17
200	14.46	14.43	.60	2.44	14.56	14.53	.51	1.74
400	9.94	9.85	.48	2.23	11.47	11.41	.51	.59
1000	4.38	4.30	.56	1.85	10.11	9.98	.50	.43
1200	3.58	3.49	.60	.83	.10	.95	.49	.42
2000	2.53	2.38	.58	.89	.14	.88	.51	.33
3000	.38	.13	.63	2.47	.28	.86	.51	.40
4000	.44	.08	.63	.50	.42	.84	.49	.48
4750					.56	.85	.50	.46

in the central portion of the South China Sea. In the upper layers the salinity is lower than it is further east, owing to admixture of river water, but at depths between 200 and 3000 m the water is of the same character

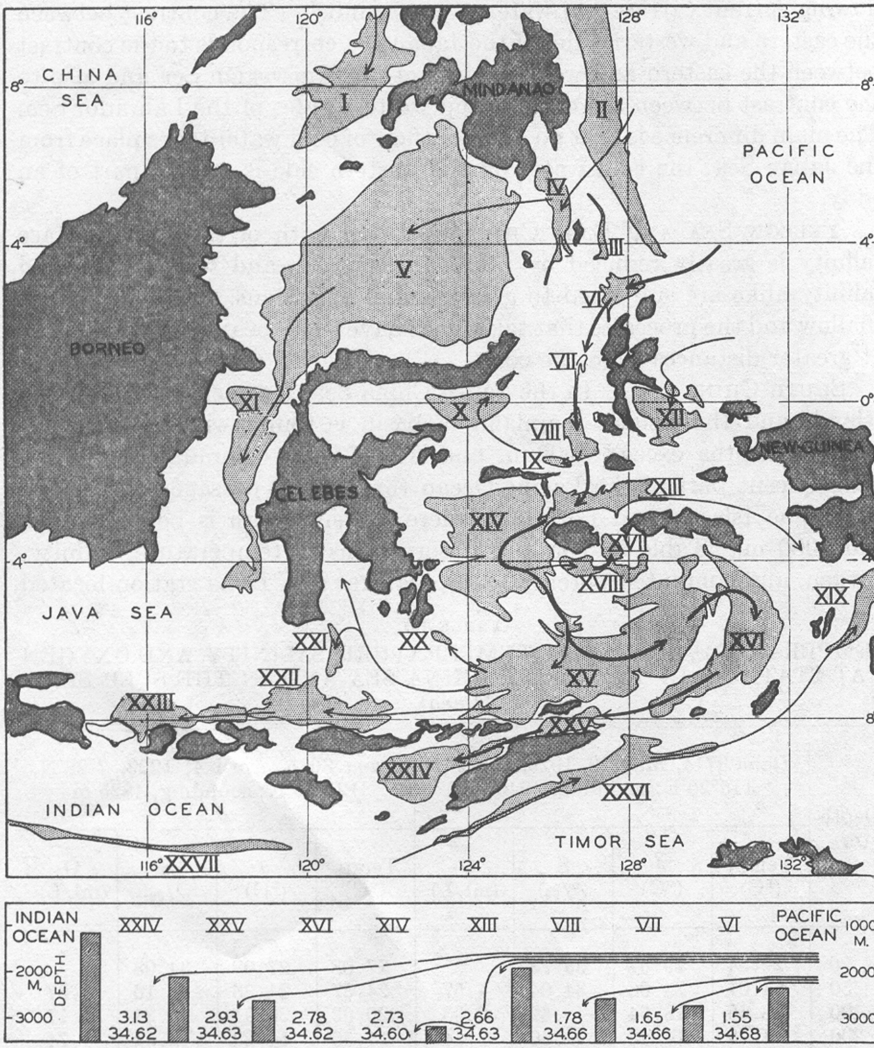

Fig. 208. *Upper:* Basins of the East Indian Archipelago, and direction from which the basin waters are renewed. *Lower:* Direction of renewal and potential temperature and salinity of the basin waters along the heavy line in the upper figure (according to van Riel).

as that of the adjacent parts of the West Pacific, whereas below 3000 m basin conditions exist, the temperature increases toward the bottom, and the salinity remains constant. The increase in the temperature is not sufficiently great to cause instability because the potential temperature decreases slightly. The oxygen content is nearly constant below 3000 m, whereas in the open ocean it increases below that depth.

THE WATERS OF THE EAST INDIAN ARCHIPELAGO. Detailed examination of the waters of the East Indian Archipelago has been conducted on the *Snellius* expedition, and the following summary is based mainly on van Riel's discussions (1932b, 1934, 1938). In fig. 208 are shown the numerous basins which are found in this region of highly complicated bottom topography, also the direction from which the renewal of water in these basins takes place. The names of the basins and their probable sill depths and maximum depths are listed in table 87. The flow of the surface water, according to van Riel (1932b), is also approximately in the direction of the arrows. The high-salinity water, which in the western Pacific is found at depths between 100 and 200 m, flows into the seas between the East Indian Islands, where the thickness of the high-salinity layer decreases in the direction of flow and the maximum salinities are reduced. Between Borneo and Celebes the layer of maximum salinity disappears in about lat. 2°S, but to the west of New Guinea it can be followed to about 8°S. In the southwestern portion of the areas under consideration, then, a region exists without an intermediate salinity maximum and with low-salinity water in the upper layers.

From the arrows in fig. 208 it is evident that the water in the deep basin in the Sulu Sea is renewed from the north by inflow of water from the South China Sea. The sill depth between the South China Sea and the Sulu Basin is probably about 400 m, and the water passing the sill has a potential temperature of about 9.9° and a salinity of about 34.50 ‰. These conditions are illustrated by the observations at *Dana* station 3685 in the Sulu Sea (table 86), from which it is seen that in the Sulu Sea the temperature increases with depth below 1200 m but the potential temperature decreases slightly except in the lower 1000 m. The oxygen content is somewhat lower than that of the South China Sea, and remains practically constant between 500 and 5000 m.

In all of the other basins shown in fig. 208 except the Timor Trench and the Sunda Trench, renewal of the deep water in the basins takes place from the Pacific Ocean. The potential temperature and the salinity of the water in the different basins is directly related to the manner of renewal and to the type of communication that exists. At the bottom of the figure is shown schematically how the renewal takes place in the series of basins joined by the heavy line in the upper part of the figure. The potential temperatures and the salinities at or directly below the sill depths are entered.

In all basins examined on the *Snellius* Expedition, a layer of minimum temperature was found a little below the sill depth across which renewal takes place. Below the layer of minimum temperature a small increase toward the bottom was observed, but this increase was in all instances somewhat less than the adiabatic one, so that the potential temperature decreased toward the bottom. The salinity of the water in the basins appears to be so constant that the greatest observed differences, amount-

Table 87

BASINS AND TRENCHES IN THE EAST INDIAN ARCHIPELAGO

(According to van Riel, 1934)

Number*	Name	Sill depth (m)	Maximum depth (m)	Observed minimum temperature (°C)	Observed minimum temperature Depth (m)	Salinity of deep water (°/oo)
I	Sulu basin	400	5580	10.08	1225	34.49
II	Mindanao trench		10500	1.56	3490	.63
III	Talaud trough	3130	3450			
IV	Sangihe trough	2050	3820	2.40	2550	.64
V	Celebes basin	1400	6220	3.58	2475	.56
VI	Morotai basin	2340	3890	1.81	2490	.65
VII	Ternate trough	2710	3450	1.85	2761	.67
VIII	Batjan basin	2550	4810	2.06	2970	.66
IX	Mangole basin	2710	3510			
X	Gorontalo basin	2700	4180	2.20	2740	.63
XI	Makassar trough	2300	2540	3.59	2133	.58
XII	Halmahera basin	700	2039	7.76	1839	.60
XIII	Boeroe basin	1880	5319	3.02	3240	.61
XIV	Northern Banda basin	3130	5800	3.04	2990	.62
XV	Southern Banda basin	3130	5400	3.06	2720	.60
XVI	Weber deep	3130	7440	3.07	2990	.61
XVII	Manipa basin	3100	4360	3.10	3185	.60
XVIII	Ambalaoe basin	3130	5330	3.08	3235	.61
XIX	Aroe basin	1480	3680	3.90	2240	.65
XX	Boetoeng trough	3130	4180			
XXI	Salajar trough	1350	3370	3.86	1750	.60
XXII	Flores basin	2450	5130	3.22	2480	.61
XXIII	Bali basin		1590	3.58	1488	.61
XXIV	Sawoe basin	2100	3470	3.39	2360	.61
XXV	Wetar basin	2400	3460	3.16	2500	.61
XXVI	Timor trench	1940	3310	2.67	2254	.71
XXVII	Sunda trench		7140	1.18	4230	.71

* Numbers refer to fig. 208.

ing to 0.02 °/oo, lie inside the experimental error of the determinations. Stable stratification prevailed, then, and no evidence was found of instability which might be caused by heating from the interior of the earth. In most basins the bottom water contained appreciable amounts of oxygen, but in some small basins of shallow sill depth the bottom water contained no oxygen but considerable quantities of hydrogen sulphide.

The water in the Timor and Sunda Trenches originates from the Indian Ocean, as is evident from the high salinity of the water, 34.71 ‰ in contrast to values between 34.60 ‰ and 34.66 ‰ in all the other basins of the Archipelago.

The classical example on adiabatic increase of the temperature toward the bottom is found in the Mindanao Trench (the Philippines Trench), on the east side of Mindanao, in which a depth in excess of 10,000 m has been recorded. From the somewhat uncertain observations of the *Planet* in 1907–1908, Schott (1914) concluded that the adiabatic temperature increased toward the bottom and that the stratification was unstable, but Wüst (1929) showed that the observations could be interpreted differently and that indifferent equilibrium probably exists. His

TABLE 88

TEMPERATURE, POTENTIAL TEMPERATURE, AND SALINITY IN THE MINDANAO TRENCH

(*Snellius* station 262, May 15–16, 1930, 9°40′N, 126°51′E; sounding, 10,068 m)

Depth (m)	Temp. (°C)	θ (°C)	S (‰)
2,470	1.82	1.65	34.64
2,970	1.66	1.44	.66
3,470	1.58	1.31	.67
3,970	1.59	1.26	.67
4,450	1.64	1.25	.67
5,450	1.78	1.26	.67
6,450	1.92	1.25	.67
7,450	2.08	1.24	.68
8,450	2.23	1.22	.69
10,035	2.48	1.16	.67

conclusion has been confirmed by observations on the *Snellius* Expedition, according to which the potential temperature actually decreases slightly with depth, whereas the observed differences in salinity are within the limits of the experimental errors. An extract from the *Snellius* observations in the Mindanao Trench is given in table 88. Thus, stable stratification appears to exist even in the deepest troughs, but a state of indifferent equilibrium is closely approached.

The Water Masses of the Oceans: A Summary

In figs. 209A and 209B are shown the characters of the water masses that have been discussed, their regions of formation, and their distribution. The chart in fig. 209A and the *T-S* curves in fig. 209B should

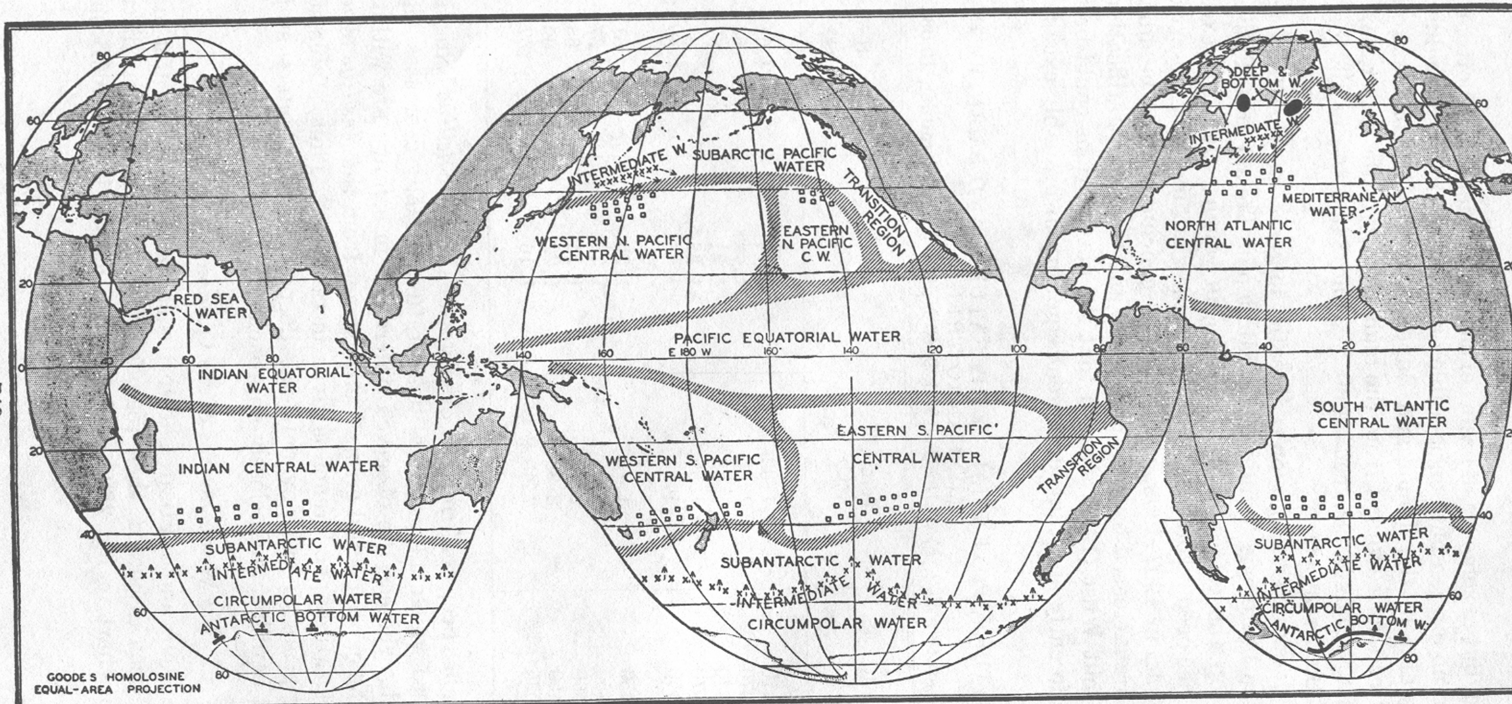

Fig. 209A. Approximate boundaries of the upper water masses of the ocean. Squares indicate the regions in which the central water masses are formed; crosses indicate the lines along which the antarctic and arctic intermediate waters sink.

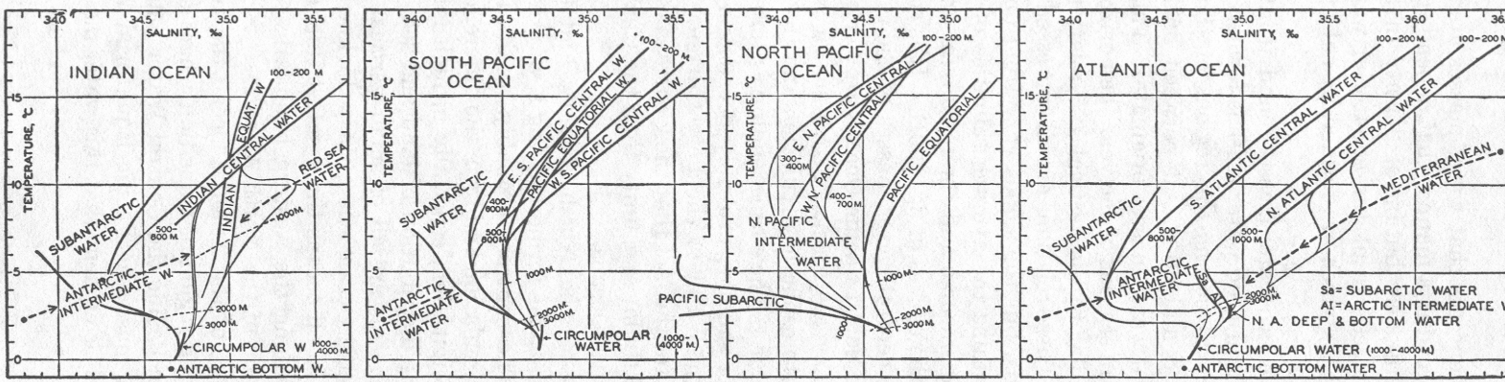

Fig. 209B. Temperature-salinity relations of the principal water masses of the oceans.

together illustrate the concepts that water masses are formed at the sea surface and sink and spread in a manner that depends upon their density in relation to the general distribution of density in the oceans This applies to all water masses except the equatorial water masses of the Indian and Pacific Oceans, which are formed by subsurface processes of mixing.

With the exception of two water masses found at intermediate depths and to which we shall return, no water mass is formed at the sea surface in low latitudes, but in all oceans the regions of the subtropical convergences (between 35°S and 40°S and between 35°N and 40°N) are regions where the central water masses originate. This concept is based on the fact that in certain seasons of the year the horizontal T-S relations in these regions are similar to the vertical T-S relations of the different central water masses. These relations are all expressed on a T-S diagram by nearly straight bands, but the slopes vary from one ocean region to another. Table 89 contains the average salinities of the central water masses at different temperatures and the maximum deviations from the averages, according to the curves in fig. 209B. It is seen from the table and the figure that the central water masses of the South Atlantic, the Indian, and the western South Pacific Oceans are very similar, as should be expected, because they are formed in regions in which the external influences, that is, the atmospheric circulation and the processes of heating and cooling, are similar. The corresponding water mass of the eastern South Pacific is of lower salinity, probably because of admixture of the low-salinity Subantarctic Water of the Peru Current. Such admixture may also be responsible for the fact that the Central Water of the western South Pacific has a slightly lower salinity than the Central Waters of the Indian and South Atlantic Oceans.

The Central Waters of the North Atlantic and the North Pacific Oceans are quite different, the former having a very high and the latter a very low salinity. The contrast probably results from the different character of the ocean circulation and from the differences in the amounts of evaporation and precipitation, especially in high latitudes, which are intimately related to the distribution of land and sea.

The central water masses are all of small vertical extension, particularly in the North Pacific Ocean where their thickness over large areas is only 200 to 300 m. In all oceans the greatest thickness of the central water masses is found along the western boundaries; it reaches 900 m in the Sargasso Sea region of the North Atlantic.

In the equatorial part of the Atlantic Ocean the two Central Water masses are separated by a region of transition where the T-S relation is intermediate, but in the Pacific the Central Water masses are separated by a well-defined water mass, the Pacific Equatorial Water. From the T-S curves in fig. 209B it is evident that this water mass is formed in the

TABLE 89

SALINITIES OF THE CENTRAL WATER MASSES OF THE OCEANS IN PARTS PER THOUSAND AT STATED TEMPERATURES

Temperature (°C)	8°	10°	12°	14°	16°
South Atlantic	34.64 ± 0.08	34.86 ± 0.08	35.11 ± 0.08	35.37 ± 0.09	35.64 ± 0.10
Indian Ocean	34.65 ± 0.07	34.89 ± 0.06	35.13 ± 0.07	35.37 ± 0.08	35.62 ± 0.09
Western South Pacific	34.58 ± 0.07	34.80 ± 0.06	35.04 ± 0.08	35.30 ± 0.08	35.55 ± 0.09
Eastern South Pacific		34.54 ± 0.07	34.70 ± 0.08	34.88 ± 0.08	35.08 ± 0.08
North Atlantic	35.12 ± 0.09	35.37 ± 0.09	35.63 ± 0.09	35.88 ± 0.09	36.12 ± 0.09
Western North Pacific		34.24 ± 0.07	34.38 ± 0.06	34.52 ± 0.06	34.67 ± 0.07
Eastern North Pacific			34.11 ± 0.09	34.32 ± 0.08	34.62 ± 0.08

South Pacific because it is similar to the water masses of that ocean, but it has a higher salinity than any of the water masses of the North Pacific. In the northern part of the Indian Ocean a corresponding Equatorial Water mass is present, which at a temperature of 15° or higher is of the same salinity as that of the Pacific, but at lower temperatures it is of a higher salinity. The higher salinities indicate admixture of Red Sea water, but in general the manner in which the water mass is formed is not clear.

The central and equatorial water masses are covered by a surface layer 100 to 200 m thick, within which the temperature and the salinity of the water vary greatly from one locality to another, depending upon the character of the currents and the exchange with the atmosphere, and within which great seasonal variations occur in middle latitudes. A discussion of the surface layer is not included in this summary. The surface layer, the central water masses, and the upper portions of the equatorial water masses together form the oceanic troposphere (p. 141).

The Subantarctic Water occurs between the central water masses of the southern oceans and the Antarctic Convergence. This water has nearly the same character all around the earth, and is therefore considered as belonging to the waters of the Antarctic Ocean (p. 606). The Subantarctic Water is of low salinity and is probably formed by mixing and vertical circulation in the region between the Subtropical and the Antarctic Convergences. In the North Atlantic the corresponding Subarctic Water is found in a small region only and is of relatively high salinity, but in the North Pacific it is of wide extension and of low salinity. The Subarctic Water must be formed by processes which differ from those that maintain the Subantarctic Water. In the southern oceans the Antarctic Convergence represents a continuous and well-defined southern boundary of the Subantarctic Water, but in the northern oceans the corresponding Arctic Convergence is found in the western parts of the oceans only, and in large areas there exists no marked *northern* boundary of Subarctic Waters. This contrast between south and north must be related to the differences in the distribution of land and sea and is reflected in the character of the waters. The Subarctic Waters are similar to the corresponding Arctic Intermediate Waters, but the Subantarctic Water is distinctly different in character from the Antarctic Intermediate Water.

Below the central water masses the intermediate waters are found in all oceans. The Antarctic Intermediate Water is the most widespread. This water, in contrast to the central waters, sinks along a well-defined *line*, and the water which leaves the surface is not a water mass but a *water type*, which, all around the Antarctic Continent, is characterized by a salinity of 33.8 ‰ and a temperature of 2.2°. After sinking, the water spreads to the north, mainly between the σ_t surfaces $\sigma_t = 27.2$ and $\sigma_t = 27.4$, and mixes with the over- and underlying waters. In this manner a *water mass* is formed, characterized by a salinity minimum which, with increasing distance from the Antarctic Convergence, becomes less and less pronounced. In the Atlantic Ocean, in which an equatorial water mass is lacking, the salinity minimum of the Antarctic Intermediate Water extends across the Equator and can be traced to about 20°N, but in the Indian and South Pacific Oceans the Antarctic Intermediate Water reaches only to about 10°S. In the Pacific Ocean a salinity minimum in the Equatorial Water can be interpreted as showing the last traces of the Intermediate Water.

In the North Atlantic the corresponding Arctic Intermediate Water is formed to the east of the Grand Banks of Newfoundland, but probably in small quantities because it appears only in a limited area of the northwest Atlantic. In the North Pacific Ocean the Arctic Intermediate Water, on the other hand, is present between lat. 20°N and 43°N, except off the American west coast where the Subarctic Water flows south. This Intermediate Water is probably formed mainly to the northeast of Japan, but it is added to off the American west coast, where at a depth of 500 to 600 m Subarctic Water spreads below the intermediate water that originates further to the west. Correspondingly, two salinity minima are found in that region (fig. 200, p. 715, and fig. 202, p. 717).

Two other intermediate water masses are of importance, namely those formed in the Atlantic and the Indian Oceans by addition of Mediterranean and Red Sea water, respectively. The Mediterranean water that flows out along the bottom of the Strait of Gibraltar has a salinity of 38.1 $^0/_{00}$ and a temperature of 13.0°, but it is rapidly mixed with surrounding Atlantic water and spreads mainly between the σ_t surfaces $\sigma_t = 27.6$ and $\sigma_t = 27.8$, that is, below the Antarctic Intermediate Water. It can be traced over wide areas by an intermediate salinity maximum. The spreading of the Red Sea water is not so well-defined, but over large parts of the equatorial and western regions of the Indian Ocean the Red Sea water is recognized by a salinity maximum at a σ_t value of about 27.4.

Below the intermediate water the deep ocean basins are filled by deep and bottom water, the maximum densities of which vary from $\sigma_t = 27.90$ in the North Atlantic to 27.75 in the North Pacific. These water masses are formed in high northerly latitudes in the Atlantic Ocean and in high southerly latitudes close to the Antarctic Continent in the Weddell Sea area, and to the south of the Indian Ocean. The spreading of these water masses will be dealt with in the following discussion of the deep-water circulation of the oceans.

The Deep-water Circulation of the Oceans

In the preceding sections reference has frequently been made to the deep and bottom waters of the different oceans, the character of which is illustrated in figs. 161, 168, 183, 189, 195, 196, and 199.

TABLE 90

TEMPERATURE, SALINITY, AND OXYGEN CONTENT BELOW 2000 METERS AT SELECTED STATIONS

(Met = *Meteor*, Atl = *Atlantis*, Da = *Dana*, B.A.E. = B.A.N.Z. Antarctic Research Exped., Di = *Discovery*, B = *Bushnell*, EWS = *E. W. Scripps*)

Atlantic Ocean, and Atlantic Antarctic Ocean

Depth (m)	Met 127 March 12, 1935 50°27.5′N, 40°14.5′W			Atl 1223 April 19, 1932 33°19′N, 68°18′W			Met 279 March 17, 1927 19°16′N, 27°27′W		
	°C	S‰	O_2	°C	S‰	O_2	°C	S‰	O_2
2000	3.32	34.92	6.30	3.62	34.97	6.08	3.54	34.98	5.07
2500	.22	.93	.26	.37	.97	.04	.11	.96	.30
3000	2.97	.93	.17	2.95	.96	5.99	2.76	.94	.27
3500	.63	.95	.28	.61	.94	6.03	.49	.92	.32
4000	.38	.95	.34	.45	.92	.06	.39	.89	.42
5000	—	—	—	.54	.90	5.88	—	—	—

Depth (m)	Met 86 Dec. 4, 1925 32°49′S, 40°01′W			Met 135 March 7, 1926 39°46′S, 22°12′E			Met 129 Feb. 22–23, 1926 58°53′S, 4°54′E		
2000	2.97	34.77	4.71	2.68	34.76	4.70	−0.26	34.67	3.37
2500	3.10	.90	5.53	.54	.82	.99	−0.36	.67	.52
3000	2.86	.92	.65	.32	.81	5.14	−0.42	.66	.59
3500	.15	.89	.46	1.93	.80	.04	−0.51	.65	.67
4000	0.77	.69	4.88	.42	.78	4.97	−0.55	.64	.79

Indian Ocean, and Indian Antarctic Ocean

Depth (m)	Da 3917 Dec. 5, 1929 1°45′N, 71°05′E			B.A.E. 75 March 19, 1930 36°41′S, 114°55′E			Di 858 April 24, 1932 60°10′S, 63°55′E		
	°C	S‰	O_2	°C	S‰	O_2	°C	S‰	O_2
2000	2.68	34.79	—	2.70	34.60	—	0.90	34.72	4.41
2500	.09	.76	3.22	.28	.68	—	.63	.70	.51
3000	1.84	.79	2.78	1.93	.72	—	.34	.68	.48
3500	.66	.75	3.17	.59	.73	—	.11	.68	.66
4000	.71	.74	.61	.25	.74	—	−0.09	.67	.77

Pacific Ocean, and Antarctic Pacific Ocean

Depth (m)	B Aug. 18, 1934 50°30′N, 175°16′W			EWS VIII-77 July 3, 1939 28°02′N, 122°08′W			Da 3745 July 8, 1929 3°18′N, 129°02′E		
	°C	S‰	O_2	°C	S‰	O_2	°C	S‰	O_2
2000	1.88	34.58	1.64	2.13	34.61	1.89	2.24	34.67	2.56
2500	.72	.59	2.20	1.83	.63	2.44	1.86	.69	.82
3000	.61	.64	.58	.65	.64	.72	.65	.69	3.15
3500	.50	.68	3.00	.55	.67	3.00	.62	.70	.26
4000	—	—	—	—	—	—	.57	.70	.27

Depth (m)	Da 3561 Sept. 24, 1928 4°20′S, 116°46′W			Da 3628 Dec. 15, 1938 31°25′S, 176°25′W			Di 950 Sept. 7, 1932 59°05′S, 163°46′W		
	°C	S‰	O_2	°C	S‰	O_2	°C	S‰	O_2
2000	2.30	34.63	2.53	2.42	34.60	3.32	1.70	34.73	4.27
2500	.01	.64	.75	.16	.64	.25	.34	.73	.33
3000	1.84	.65	.86	1.89	.68	.75	.09	.72	.37
3500	.70	.66	.96	.49	.72	4.27	0.91	.71	.20
4000	.64	.67	3.00	.22	.72	.52	.87	.70	.06
5000	—	—	—	.02	.71	.55	—	—	—

The manner in which the deep and bottom water is formed has been discussed (p. 138), and certain statements as to the deep-water circulation have been made, but so far no general review of the deep-water circulation has been presented. Table 90 has been prepared in order to facilitate such a review. It contains temperatures, salinities and oxygen values of the deep and bottom water at fifteen selected stations, six in the Atlantic, three in the Indian and six in the Pacific Ocean, including the adjacent parts of the Antarctic Ocean. The content of this table will not be dealt with separately but must be examined as the discussion proceeds.

In order to understand the deep-water circulation, one has to bear in mind that deep and bottom waters represent water the density of which became greatly increased when the water was in contact with the atmosphere, and that this water, by sinking and subsequent spreading, fills all deeper portions of the oceans. The most conspicuous formation of water of high density takes place in the subarctic and in the antarctic regions of the Atlantic Ocean. The deep and bottom water in all oceans is derived mainly from these two sources, but is to some extent modified by addition of high-salinity water flowing out across the sills of basins in lower latitudes, particularly from the Mediterranean and the Red Sea.

In the North Atlantic Ocean, North Atlantic Deep and Bottom Waters flow to the south, the flow being reinforced, and the upper deep water being modified, by the high-salinity water flowing out through the Strait of Gibraltar. The newly formed deep and bottom water has a high oxygen content which decreases in the direction of flow. Figure 210 and the values given in table 90 demonstrate the character of these waters and show particularly the increase in salinity at moderate depth caused by addition of Mediterranean water. Antarctic Bottom Water flows in the opposite direction, from south to north, and has been traced beyond the Equator to lat. 35°N (Wüst, 1935). The spreading to the north of the Antarctic Bottom Water is illustrated in fig. 211, showing the potential temperatures below a depth of 4000 m. Owing to admixture of this water the salinity of the bottom water of the North Atlantic decreases toward the south.

The North Atlantic Deep Water crosses the Equator and continues toward the south above the Antarctic Bottom Water. On the other hand, it sinks below the Antarctic Intermediate Water and therefore, in the South Atlantic Ocean, it becomes sandwiched between the Antarctic Intermediate Water and the Antarctic Bottom Water, both of which are of lower salinity. In a vertical section the deep water of the South Atlantic Ocean is therefore characterized by a salinity maximum, but, owing to mixing with the overlying and the underlying water, the absolute value of the salinity at the maximum decreases toward the south.

The bottom water of antarctic origin is colder than the deep water and to the south of about latitude 20°S the Antarctic Intermediate Water is also colder. In a vertical section the deep water therefore shows a

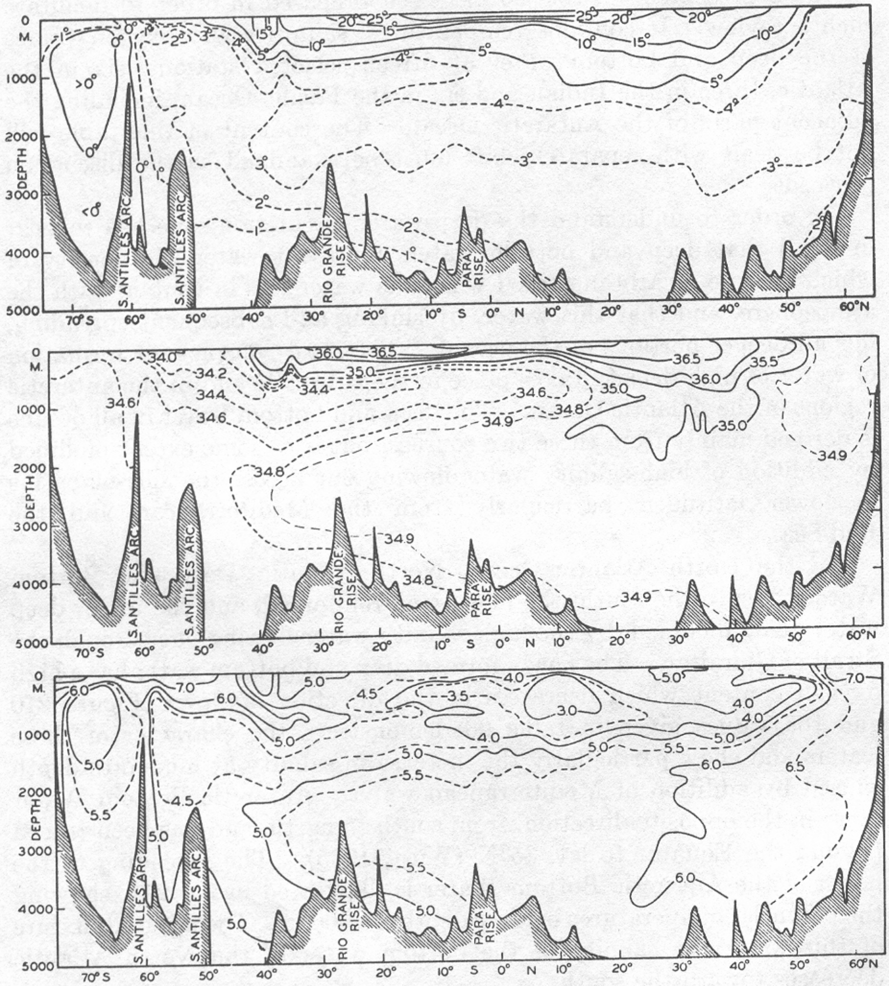

Fig. 210. Vertical sections showing distributions of temperature, salinity, and oxygen in the Western Atlantic Ocean (after Wüst).

temperature maximum and also a decreasing temperature to the south, owing to admixture from above and from below.

In the South Atlantic Ocean a large amount of water of antarctic origin, bottom water or intermediate water, returns to the Antarctic after having been mixed with the south-moving deep water. According to the computations which were discussed on pp. 465 and 629, the transport across the Equator of North Atlantic Deep Water amounts to

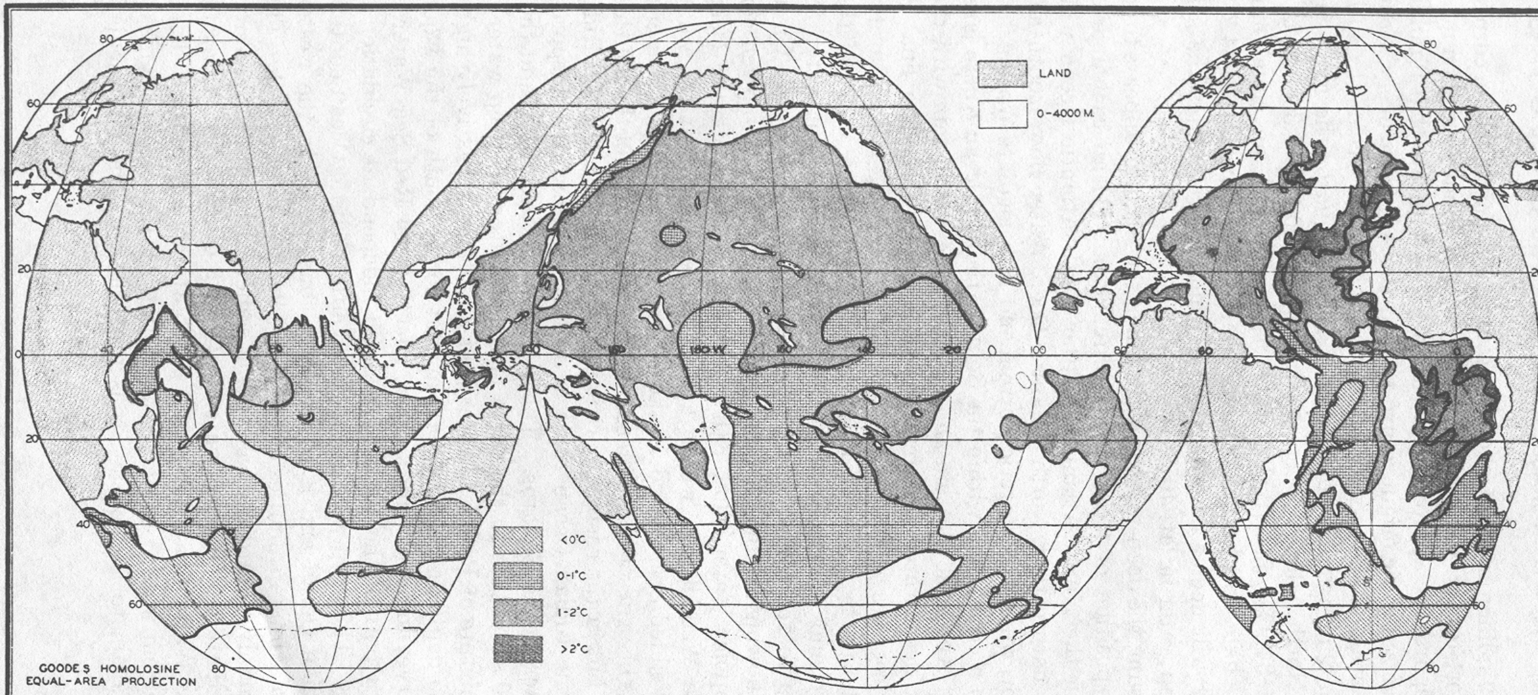

Fig. 211. Potential temperature at depths exceeding 4000 meters (after Wüst).

about 9 million m^3/sec, whereas between 20° and 30°S the corresponding transport toward the south of deep water is about 18 million m^3/sec. If these figures are approximately correct they indicate that 9 million m^3 of water of antarctic origin return toward the Antarctic every second. An examination of the salinity in the South Atlantic Ocean at depths below 1600 m confirms this conclusion. At the Equator the average salinity of the North Atlantic Deep Water between 2000 and 3500 m is approximately 34.93 ‰. If this water is mixed with an equal amount of intermediate and bottom water of salinity approximately 34.7 ‰, the salinity will decrease to 34.81 ‰, which approximates the salinity of the deep water in lat. 40°S.

In sum, the deep-water circulation of the Atlantic appears to represent a superposition of two types of circulation: (1) an exchange of water between the North Atlantic and the South Atlantic Ocean which is of such a nature that North Atlantic Deep Water flows south across the Equator, whereas Antarctic Bottom and Intermediate Waters flow north; and (2) a circulation within the South Atlantic Ocean where large quantities of Antarctic Bottom and Intermediate Water mix with the south-flowing deep water and return to the Antarctic. The final result of these processes is that the deep water reaching the Antarctic Ocean from the north is diluted as compared to the deep water of the North Atlantic and is of a lower temperature. This is the water that contributes to the formation of the large body of Antarctic Circumpolar Water flowing around the Antarctic Continent. The oxygen content of Circumpolar Water is lower than that of the North Atlantic Deep Water and the Weddell Sea water, and decreases somewhat toward the east from Weddell Sea toward Drake Passage (see p. 621). The circulation that has been described is present in the western part of the South Atlantic Ocean, but in the eastern part it is impeded by the Walfish Ridge.

In the Indian Ocean there is no large southward transport of deep water across the Equator. The *T-S* diagram in fig. 189 and the data in table 90 show that to the north of the Equator the deep water contains an admixture of Red Sea Water that maintains a relatively high salinity down to depths exceeding 3000 m, but to the south of the Equator the *T-S* curves indicate only a slight effect of the Red Sea Water. In the southern part of the Indian Ocean an independent circulation must be present. The deep water from the South Atlantic Ocean continues into the Indian Ocean and is particularly conspicuous in the western part, where maximum salinities of 34.80 ‰ have been observed. This water flows mainly toward the east, being somewhat diluted by admixtures of intermediate and bottom waters. On the other hand, Antarctic Intermediate Water flows north and the bottom temperatures demonstrate that bottom water also moves north (fig. 211); these water masses must return again to the south. It is probable that the intermediate water and the bottom water mix with the deep water and that the return flow takes place within the latter. A slight admixture of deep water from the region to the north of the Equator, which is compensated for by bottom water penetrating across the Equator, appears to maintain the salinity of the Indian Ocean Deep Water at a higher level than would be the case if no addition of saline water took place. Thus the influence of the Red Sea can probably be traced to the Antarctic, but by a mechanism somewhat different from the one which has been suggested by Lotte Möller (1929, 1933) and which has been discussed particularly by Thomsen (1933, 1935).

From the Indian Ocean the Antarctic Circumpolar Water with its components of Atlantic and Indian Ocean origin enters the Pacific Ocean. The *Discovery* and *Dana* observations in the Tasman Sea between Australia and New Zealand, and in the Pacific to the east of New Zealand, show that the salinity of the deep and bottom water has been reduced so much that the maximum values lie between 34.72 ‰ and 34.74 ‰. These maximum salinities are found at depths between 2500 and 4000 m, the salinity of the water close to the bottom being slightly lower. From the region where the deep water enters the Pacific Ocean the salinity decreases both toward the north and toward the east. The *Discovery* data indicate that below the Antarctic Convergence a core of water of salinity higher than 34.72 ‰ is found, which represents water of the Circumpolar Current; but to the north of this region of maximum salinity, values below 34.70 ‰ prevail, increasing uniformly toward the bottom. The *Carnegie* and the *Dana* data similarly show that north of 40°S the highest salinities are found near the bottom. The structure of the water masses of the Pacific differs completely, therefore, from that found in the other oceans, where the highest salinities are encountered in the deep water and not in the bottom water.

This feature can be explained if one assumes that in the South Pacific Ocean there also exists a circulation which is similar to that of the South Atlantic and Indian Oceans, namely, that intermediate and bottom water flow to the north and that a flow of deep water to the south takes place. This north-south circulation is superimposed upon a general flow from west to east. The Pacific Deep Water is, therefore, of Atlantic and Indian origin but has become so much diluted by admixture of intermediate and bottom water that the salinity maximum has disappeared. These conclusions as to the character of the deep-water circulation of the South Pacific are in agreement with the concept of Deacon (1937a), who has shown that the deep water of the Pacific moves toward the south and rises within the Antarctic region in a similar manner to that of the deep water of the Atlantic and Indian Oceans.

The more or less closed systems of the deep-water circulation in the Southern Hemisphere between the Antarctic Ocean and the Equator are related to the fact that near the Equator all ocean currents tend to flow in east-west directions and that transport of water across the Equator takes place only when it is required to maintain the same sea level in both hemispheres. Such is the case in the Atlantic Ocean, where deep water is formed in the Northern Hemisphere. The water that sinks in high latitudes or flows out of the Mediterranean Sea spreads at great depths, continues across the Equator, and is replaced by horizontal flow from the South Atlantic Ocean. In the Indian Ocean a similar mechanism operates, but on a very small scale, because sinking takes place only in the Red Sea and the amounts of deep water formed there are small and exercise a significant influence only upon conditions to the north of the Equator. In the Pacific Ocean no mechanism exists that will give rise to a considerable exchange of water across the Equator, because conditions are such that no deep water is formed in the North Pacific.

TABLE 91

EXTREME AND AVERAGE VALUES OF SALINITIES AT STATIONS IN THE WESTERN PACIFIC BETWEEN LONG. 152°W AND 130°E, LAT. 30°S AND 5°N, AND AT STATIONS IN THE EASTERN PACIFIC BETWEEN LONG. 90°W AND 130°W, LAT. 5°S AND 30°N, WITH CORRESPONDING AVERAGE VALUES OF TEMPERATURE AND OXYGEN CONTENT

(From data of *Dana*, *Bushnell*, and *E. W. Scripps*)

Depth (m)	Western Pacific					Eastern Pacific				
	S, min. (‰)	S, max. (‰)	S (‰)	Temp. (°C)	O_2 (ml/L)	S, min. (‰)	S, max. (‰)	S (‰)	Temp. (°C)	O_2 (ml/L)
2500	34.64	34.71	34.674	1.98	3.20	34.62	34.69	34.655	1.85	2.36
3000	.68	.71	.690	1.78	3.37	.64	.69	.669	1.71	2.56
3500	.69	.72	.698	1.64	3.48	.65	.70	.677	1.56	2.94

On the other hand, continuity must exist between the South and the North Pacific, for which reason the depths of the North Pacific Ocean are filled by water of the same character as that found in the northern portion of the South Pacific (fig. 212). The most accurate data available indicate that a very small exchange of deep water takes place between the two hemispheres and that a sluggish motion to the north may occur on the western side of the Pacific Ocean, whereas a sluggish motion to the south may occur on the eastern side. A comparison of salinities at the depths of 2500, 3000, and 3500 m in the eastern and western parts of the North Pacific is found in table 91, from which it is seen that the salinity in the eastern part is about 0.02 ‰ lower than that in the western. This difference lies at the limit of the accuracy of the observations, but in the table the ranges of the observed values have also been entered and these ranges are not completely overlapping, for which reason the difference probably is significant. It can be explained if the deep water of the

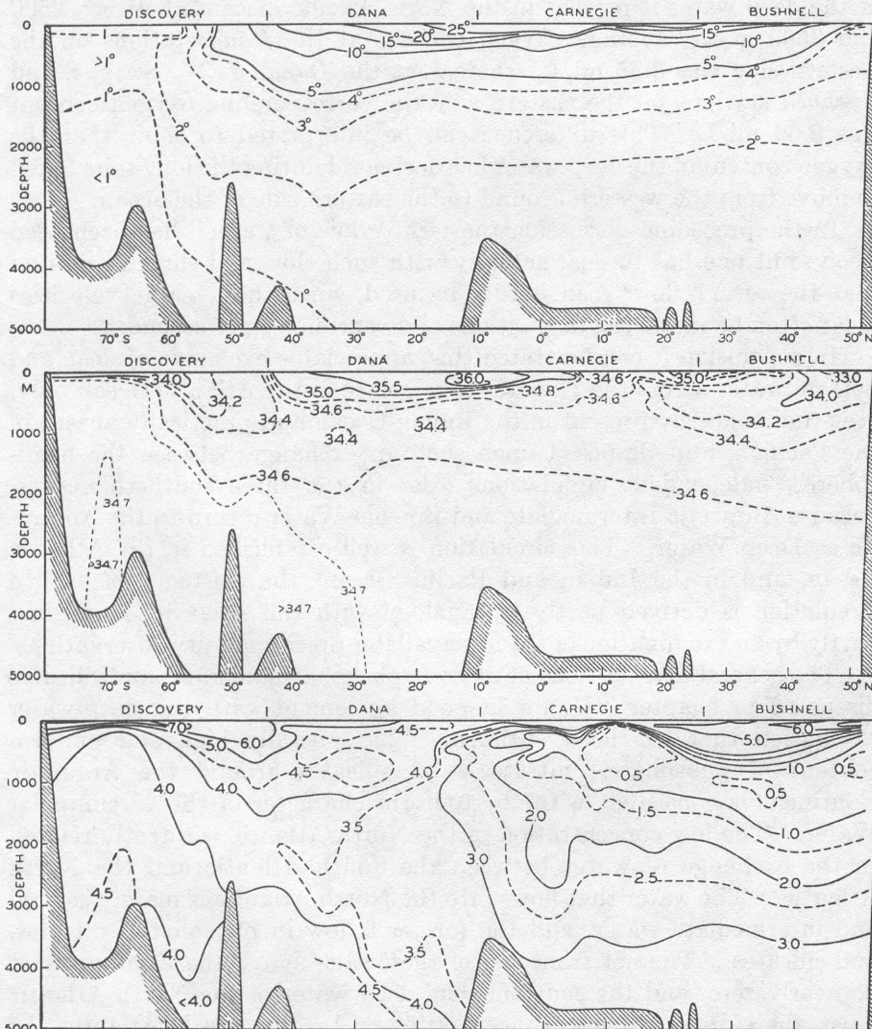

Fig. 212. Vertical sections showing distribution of temperature, salinity and oxygen in the Pacific Ocean, approximately along the meridian of 170°W.

North Pacific circulates slowly in a clockwise direction and if a slight renewal takes place on the western side, while on the eastern side small amounts are transported to the Southern Hemisphere. The lower salinity of the deep water on the eastern side would then be due to vertical admixture of low-salinity water where the deep water crosses the Pacific Ocean. The slightly lower temperatures on the eastern side indicate that this admixture takes place in high latitudes. The oxygen content of the water substantiates the conclusion that a slow clockwise circulation of the deep water is present in the North Pacific, because between 2500 and 3500 m the average oxygen content at the *Dana* stations on the western side was 3.35 ml/L, whereas at the *Dana*, *E. W. Scripps*, and *Bushnell* stations on the eastern side the corresponding oxygen content was 2.62 ml/L. This difference can be interpreted to show that the oxygen content of the deep water has decreased during the long time taken to move from the western around to the eastern side of the ocean.

In the preceding discussion the term "flow of water" has been used freely, but one has to deal actually with such slow and sluggish motion that the term "flow" can hardly be used, since the average velocities must often be measured in fractions of a centimeter per second.

In conclusion it can be stated that appreciable exchange of deep and bottom water across the Equator takes place in the Atlantic Ocean only. It is rudimentarily present in the Indian Ocean and practically absent in the Pacific. Superimposed upon such an exchange between the hemispheres, independent circulations exist in the three southern oceans, because Antarctic Intermediate and Bottom Water return to the Antarctic as Deep Water. This circulation is well established in the Atlantic Ocean, and in the Indian and Pacific Oceans the existence of such a circulation is derived partly by analogy with the Atlantic Ocean and partly by an examination of the few available precise salinity observations.

The general distributions of oxygen, phosphates, nitrates and silicates discussed in chapter VII are in good agreement with the deep-water circulation that has been outlined. The generally high and uniform content of phosphates, nitrates, and silicates around the Antarctic Continent is consistent with the uniform character of the Circumpolar Water. The low concentration in the North Atlantic is directly related to the exchange of water between the South Atlantic and the North Atlantic. The water that flows into the North Atlantic is mainly central and intermediate water, and the former is low in phosphates, nitrates, and silicates. The net transport of these salts across the Equator must be nearly zero, and the south-flowing deep water of the North Atlantic must therefore have low concentrations. In the South Atlantic the deep water becomes rapidly mixed with Antarctic Intermediate Water and Antarctic Bottom Water, both of which contain so much phosphate, nitrate, and silicate that the contents in the deep water increase as it flows south. The North Pacific Deep Water, on the other hand, contains great amounts of phosphates, nitrates and silicates, in agreement with the concepts that a small exchange of deep water takes place between the South and North Pacific and that in the North Pacific the circulation of the deep water is so slow that there the concentrations of these salts are increased by further decomposition of organic matter. In the Indian Ocean conditions appear to be more or less similar to those in the South Pacific. Some of the major features of the distributions, therefore, are related to the deep-water circulation, but others are intimately related to the biological economy of the sea as explained in chapter VII.

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