Chapter 15

Hydrology of the Atlantic Ocean

The hydrology of the Atlantic Ocean basins is deeply affected by the formation and recirculation of North Atlantic Deep Water, which was discussed in Chapter 7. The injection of surface water into the deeper layers is responsible for the high oxygen content of the Atlantic Ocean. It is also intricately linked with high surface salinities, as will become evident in Chapter 20. When compared with other ocean basins, the basins of the Atlantic Ocean are therefore characterized by relatively high values of salinity and dissolved oxygen.

Precipitation, evaporation, and river runoff

Precipitation over the Atlantic Ocean varies between 10 cm per year in the subtropics, with minima near St. Helena and the Cape Verde Islands, and more than 200 cm per year in the tropics. The region of highest rainfall follows the Intertropical Convergence Zone (ITCZ) in a narrow band along 5°N. A second band of high rainfall, with values of 100 - 150 cm per year, follows the path of the storm systems in the Westerlies of the North Atlantic Ocean from Florida (28 - 38°N) to Ireland, Scotland, and Norway (50 - 70°N). In contrast to the situation in the Pacific Ocean, no significant decrease in annual mean precipitation is observed from west to east; however, rainfall is not uniform across the band through the year. Most of the rain near Florida falls during summer, whereas closer to Europe it rains mainly in winter. A third band of high rainfall with similar precipitation values is associated with the Westerlies of the South Atlantic Ocean and extends along 45 - 55°S.

The precipitation-evaporation balance (*P-E*; Figure 1.7) reflects the rainfall distribution closely, since over most of the region evaporation varies much less than precipitation. The influence of the ITCZ is seen as a region of positive *P-E* values north of the equator. In the vicinity of South America the region extends to 30°S and along the coast of Panama, a result of extreme annual mean rainfall conditions over land. A similar southward extension is found near the African continent. The band of high rainfall in the northern hemisphere Westerlies is also evident as a region of positive *P-E* balance; its counterpart in the southern hemisphere does not come out very well due to lack of data.

Compared to the Pacific and Indian Oceans, the total downward freshwater flux (i.e. the *P - E* balance averaged over the ocean area) is obviously smaller in the Atlantic than in the other two oceans. Maximum *P - E* values are considerably less, and the areas with less than 100 cm per year cover a proportionately much larger area. The effect on the sea surface salinity is somewhat alleviated by the fact that the land drainage area of the Atlantic Ocean is much larger; it includes nearly all of the American continent, Europe, large parts of Africa, and northern Asia (Siberia). Many of the world's largest rivers - including the Amazon, Orinoco, Mississippi, St.Lawrence, Rhine, Niger, and Congo Rivers - empty into the Atlantic Ocean, others - the Nile, Ob, Jenisej, Lena, and Kolyma Rivers - into its mediterranean seas. In these adjacent seas river runoff plays an important role in the salinity balance and consequently influences their circulation. Overall, however, the contribution from rivers to the fresh water flux of the Atlantic Ocean cannot compensate for the low level of rainfall over the sea surface.

Sea surface temperature and salinity

As noted earlier, the map of sea surface salinity (SSS; Figure 2.5b) resembles the *P-E* distribution (Figure 1.7) outside the polar and subpolar regions. Poleward of the Westerlies, the SSS values decrease further, despite the decrease in rainfall and *P-E* values, as a result of freshwater supply from glaciers and icebergs. In the north Atlantic Ocean this effect is concentrated in the west and linked with advection by the East and West Greenland Currents and the Labrador Current. This produces a sharp salinity increase across the boundary between the Labrador Current and the Gulf Stream (the Polar Front). A similar effect is seen in the southern hemisphere along the boundary between the Malvinas and Brazil Currents. The low SSS values along South Africa and Namibia, on the other hand, are the result of Indian Ocean water extrusions from the Agulhas Current, which were discussed in Chapters 11 and 12.

In the subtropics, the water with high salinities flows westward with the North and South Equatorial Currents. Continuous evaporation along its way increases the surface salinity further. The SSS maxima are therefore shifted westward relative to the *P-E* maxima. In the southern hemisphere this process continues into the South American shelf, where the Brazil Current advects the high salinity water southward. In the northern hemisphere the region of high surface salinity does not reach the American shelf because the North Brazil, Guyana, and Antilles Currents carry tropical water of low salinity across the equator into the northern hemisphere. This water is needed to feed the Gulf Stream and its extensions and is therefore not available to dilute the waters to the east. (This contrasts with the situation in the Pacific Ocean, where the low salinity water of the western region is recirculated in the Equatorial Countercurrent.) The highest sea surface salinities of the world ocean are therefore found in the region of the Canary and North Equatorial Currents.

The distribution of sea surface temperature (SST; Figure 2.5a) shows similarities with the Pacific Ocean, particularly in the southern hemisphere where surface temperatures are much the same in the central parts of both oceans. Advection by the Brazil Current and upwelling along the Namibian coast are responsible for the marked SST differences between west and east in the subtropics. The thermal equator is at about 5°N and coincides with the Doldrums or ITCZ. In the west it extends northward into the Gulf of Mexico. The region of weak and variable winds is limited to the narrow band of the ITCZ; there is no analogue to the large region of extremely light winds found in the region north of Papua New Guinea. This is probably the reason why maximum SST values in the Atlantic Ocean are 2°C lower than in the Pacific Ocean. The contouring interval of Figure 2.5a shows the highest temperatures as above 26.0°C; actual annual mean SST values are in fact above 27.0°C over most of the region.

The major feature of the SST distribution is the marked departure of the isotherms from a zonal distribution and the associated crowding along the Polar Front in the northern hemisphere. The temperature difference between the east and west coasts north of 40°N and its consequences for the local climate have often been noted. In fact, the SST difference between the shelves off northern Japan and Oregon is only marginally smaller than the SST difference between the shelves off Newfoundland and France (about 6°C and 8°C, respectively). But in the Pacific Ocean this difference develops over more than twice the zonal distance available in the Atlantic Ocean, and the isotherms cross the latitude circles at a much smaller angle. The departure from zonal isotherm orientation in the North Atlantic

Ocean is enhanced by water exchange with the Arctic Mediterranean Sea; the 5°C isotherm (and the 35 isohaline) angles across the ocean basin from 45°N near Newfoundland to 72°N off Spitsbergen.

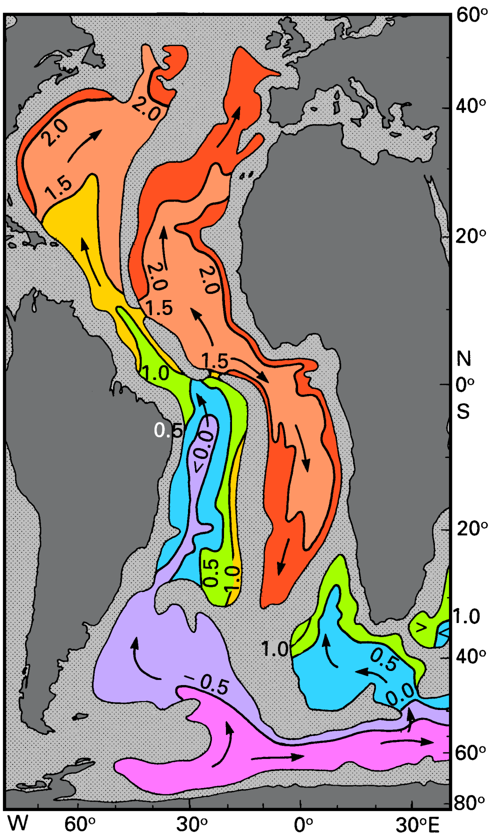

Fig. 15.1. Potential temperature below 4000 m depth and inferred movement of Antarctic Bottom Water. Adapted from Wüst (1936).

Abyssal water masses

When compared with the other two oceans, the abyssal layers of the Atlantic Ocean display a hydrographic structure full of texture and variety. This results mainly from water exchange with mediterranean basins, particularly the Arctic and Eurafrican Mediterranean Seas.

Below 4000 m depth, all Atlantic Ocean basins are occupied by *Antarctic Bottom Water* (ABW). This water mass spreads northward from the Circumpolar Current and penetrates the basins east and west of the Mid-Atlantic Ridge. On the eastern side its progress comes to a halt at the Walvis Ridge; but on the western side it penetrates well into the northern hemisphere past 50°N. A map of potential temperature below 4000 m depth (Figure 15.1) shows a gradual temperature increase from the Southern Ocean to the Labrador Basin through mixing with the overlying waters. It also indicates how Antarctic Bottom Water enters the eastern basins north of the Walvis Ridge from the equator by passing through the Romanche Fracture Zone (Figure 8.2). As a result, potential temperature increases slowly both northward and southward from the equator in the eastern basins, and potential temperatures north and south of the Walvis Ridge differ by more than 1°C.

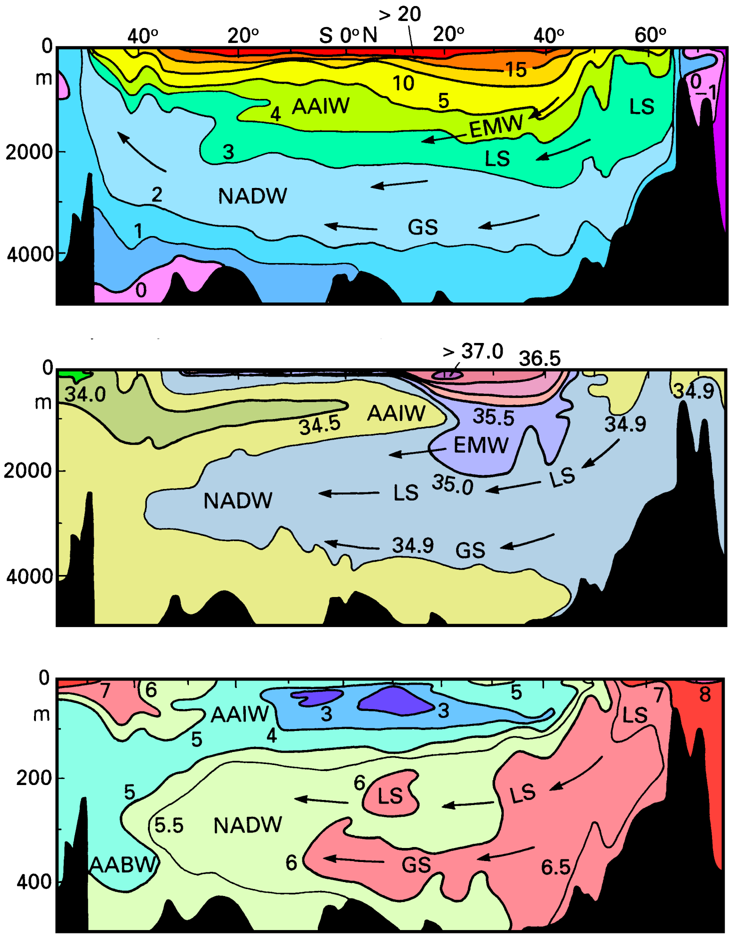

Fig. 15.2. A section through the western basins of the Atlantic Ocean. (a) Potential temperature (°C), (b) salinity, (c) oxygen (ml/l). See Fig. 15.7 for position of section. AABW: Antarctic Bottom Water, AAIW: Antarctic Intermediate Water, NADW: North Atlantic Deep Water originating from the Labrador Sea (LS) or the Greenland Sea (GS) or containing a contribution of Eurafrican Mediterranean Water (EMW). Adapted from Bainbridge (1980).

The occurrence of *Arctic Bottom Water* (ABW) in undiluted form is restricted to the immediate vicinity of the Greenland-Iceland-Scotland Ridge. As already discussed in Chapter 7 its main impact is its contribution to the formation of *North Atlantic Deep*

Water (NADW) which fills the depth range between 1000 m and 4000 m. In vertical sections (Figure 15.2) it is seen as a layer of relatively high salinity (above 34.9) and oxygen (above 5.5 ml/l) extending southward from the Labrador Sea to the Antarctic Divergence. More detailed inspection reveals two oxygen maxima in the subtropics, at 2000 - 3000 m and 3500 - 4000 m depth, indicating the existence of two distinct Deep Water varieties. The upper maximum can be traced back to the surface near 55°N and reflects the spreading of NADW formed by mixing Arctic Bottom Water with the product of deep winter convection in the Labrador Sea. The lower maximum has its origin in the Greenland-Iceland-Scotland overflow region and indicates that some Deep Water is formed before the Arctic Bottom Water reaches the Labrador Sea, through mixing of overflow water with the surrounding waters. East of the Mid-Atlantic Ridge this is the only mechanism for the formation of Deep Water, and this NADW variety, variably known as lower or eastern NADW, is therefore particularly prominent in the eastern basins. Deep Water of Labrador Sea origin, referred to as middle or western NADW, is less dense than the eastern variety, and the two varieties remain vertically layered along their southward paths.

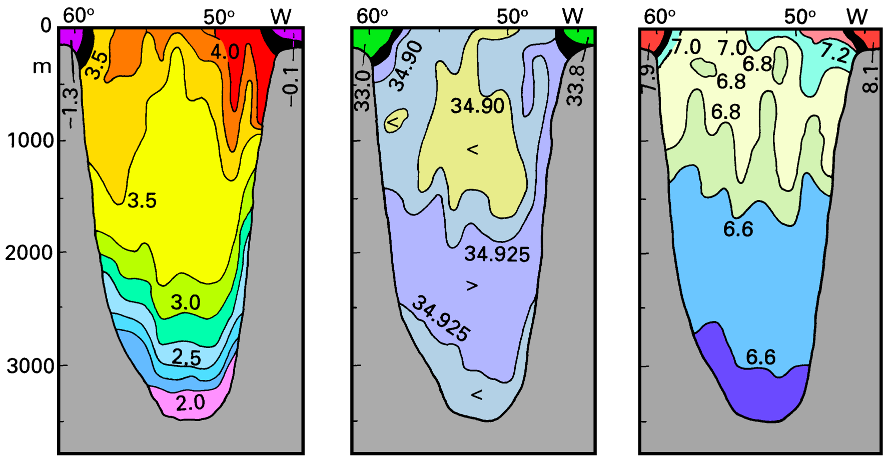

Fig. 15.3. A section through the southern Labrador Sea along approximately 60°N. (a) temperature (°C), (b) salinity, (c) oxygen (ml/l). Data from Osborne *et al*. (1991).

The long-term stability of NADW properties depends on the degree of atmospheric and oceanic variability during its formation period. A section through the Labrador Sea (Figure 15.3) shows a huge volume of nearly homogeneous water, with temperatures of 3.0 - 3.6°C and salinities of 34.86 - 34.96 and consistently high oxygen content, surrounded by strong cyclonic circulation. This *Labrador Sea Water* is the product of deep convection during the winter months. Observations show that deep winter convection is not an annual event; Clarke and Gascard (1983) report the formation of 10^5 km^3 of water with 2.9°C and 34.84 salinity in 1976 but virtually no formation of new water in 1978. Present estimates are that convection occurs in 6 out of 10 years. This produces significant variation in the properties of Labrador Sea Water; for the years 1937, 1966, and 1976

Clarke and Gascard report respective T-S combinations of 3.17°C and 34.88, 3.4°C and 34.9, and 2.9°C and 34.84. However, not all of this variability is passed on to the NADW since mixing of Labrador Sea Water with Arctic Bottom Water does not occur during a single passage through the Labrador Sea. Variations in the rate of production of Labrador Sea Water are therefore averaged over the number of loops the Arctic Bottom Water performs around the area. A rough calculation based on the volume transports and velocities of the last chapter gives 2 - 3 loops, performed over 12 - 18 months. This suggests that interannual variations of Labrador Sea Water properties are transmitted into NADW at about half the original magnitude.

A third variety of North Atlantic Deep Water sometimes found in the literature as upper NADW is really NADW from the Labrador Sea (middle or western NADW) with traces of *Eurafrican Mediterranean Water* (EMW). This water mass leaves the Strait of Gibraltar with a temperature of about 13.5°C and a salinity of 37.8; but within less than 250 km its temperature and salinity are reduced by mixing to 11 - 12°C and 36.0 - 36.2. Starting from these characteristics EMW spreads isopycnally across the ocean, mixing gradually with the Deep Water above and below. Relative to NADW of the same density it has anomalously high salinities and temperatures. Figure 15.4 shows it as a salinity and temperature maximum at 1000 m depth near the upper distribution limit of NADW. The Mediterranean Water is carried northward along the Portuguese shelf under the influence of the Coriolis force and mixes into the subtropical gyre circulation, eventually spreading southward and westward. The core of the salinity and temperature anomaly sinks as the water spreads, and at the 2000 m level (Figures 2.5e and f) traces of Mediterranean Water can be seen all across the North Atlantic Ocean, with some high salinity water crossing the equator in the west and proceeding southward.

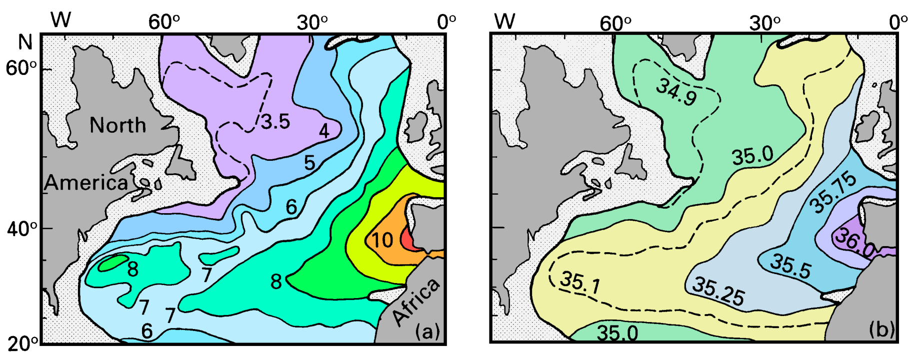

Fig. 15.4. Temperature (°C) (a) and salinity (b) in the North Atlantic Ocean at 1000 m depth.

Although the influence of the Mediterranean Water is strong enough to put its mark on the long-term mean distribution of oceanic properties, it is wrong to imagine the spreading of EMW as a process of smooth isopycnal movement with equally smooth diapycnal diffusion. It has to be remembered that the eastern basins of all oceans are characterized by slow mean motion but high eddy activity. Mediterranean Water is therefore injected into the

NADW in the form of subsurface eddies, rotating lenses which contain a high proportion of Mediterranean Water in its core. The rotation shelters EMW from the surrounding NADW; it prevents mixing and keeps the lens together over large distances. Lenses of Mediterranean Water, often referred to as "meddies" (Figure 15.5), have been found as far afield as in the Sargasso Sea. Direct observations in eddies from the Canary Current region showed rotational velocities of 0.2 m s^{-1} in general southward movement of about 0.05 m s^{-1}. The salinity and temperature anomalies found in the long-term average distribution have to be seen as the result of a process in which many such meddies travel through the upper NADW range at any particular moment in time, slowly releasing their load of extra salt and heat into the surrounding Deep Water.

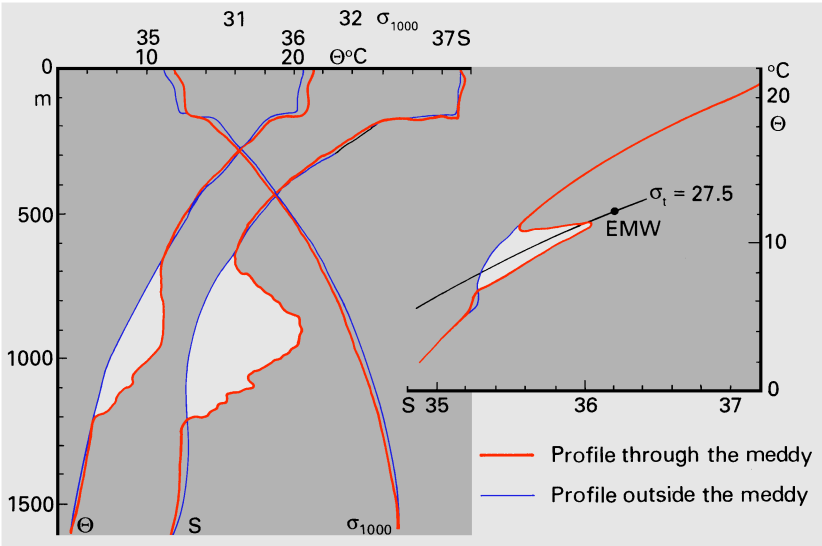

Fig. 15.5. An example of a "meddy", a rotating lens of Mediterranean Water found some 2500 km south-west of the Strait of Gibraltar near 26°N, 29°W. Note that the density profiles inside and outside are nearly identical. (σ_t gives density at atmospheric pressure, σ_{1000} at a pressure equivalent to 1000 m depth.) Adapted from Armi and Stommel (1983).

Along the western boundary of the ocean the mean flow of Deep and Bottom Water becomes stronger than eddy-related movement and can therefore be seen in hydrographic sections. Figure 15.6 shows the Deep Water as a salinity maximum at 2000 - 3000 m and a temperature maximum at 1400 - 2000 m, concentrated against the South American shelf. Both features are nearly 1000 km wide, probably wider than the associated boundary current as a result of mixing. Intensification of Antarctic Bottom Water flow is indicated by the shape of the isotherms and isohalines below 4000 m; upward slope towards the coast is consistent with a northward "thermal wind" increasing in speed with depth (Rule 2a of Chapter 3). A similar intensification occurs on the western side of the Cape Basin along the Walvis Ridge. Because the basin is closed in the north below the 3000 m level, the flow follows the depth contours in cyclonic motion, and the Bottom Water leaves the basin on the eastern side towards the Indian Ocean (Nelson, 1989). The current is swift enough to remove sediment along the base of the continental rise and produce a band of exposed rock face at 5000 m depth.

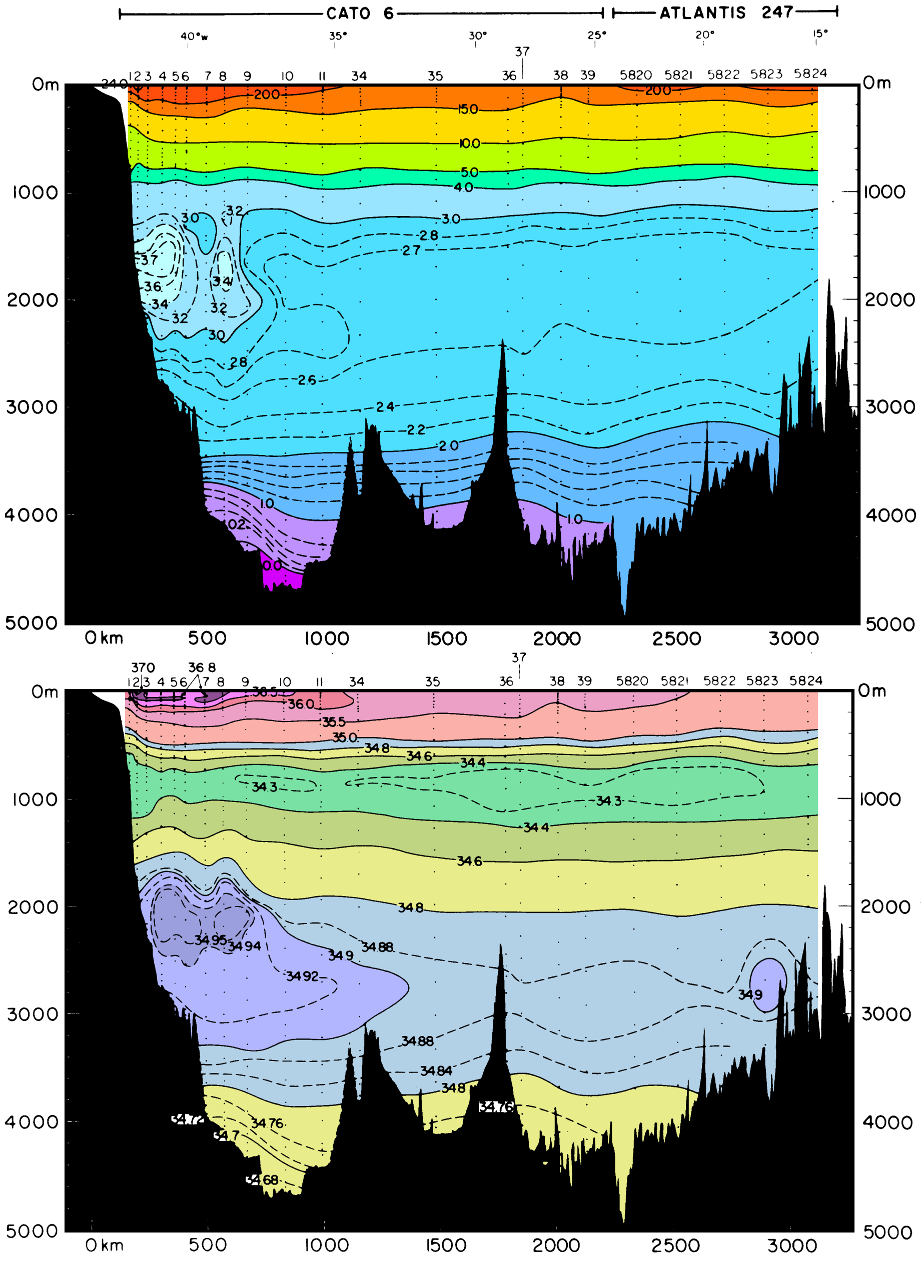

Fig. 15.6. A section through the western Atlantic Ocean along 30°S. (a) Potential temperature (°C), (b) salinity. From Warren (1981a).

Antarctic Circumpolar Water has the same density as North Atlantic Deep Water but is colder and fresher (Figure 6.13). In the absence of NADW it would take its place in the Atlantic Ocean; however, the southward advance of the Deep Water reduces its influence. Detailed analysis (Reid, 1989) shows northward propagation of some Circumpolar Water both below and above the Deep Water.

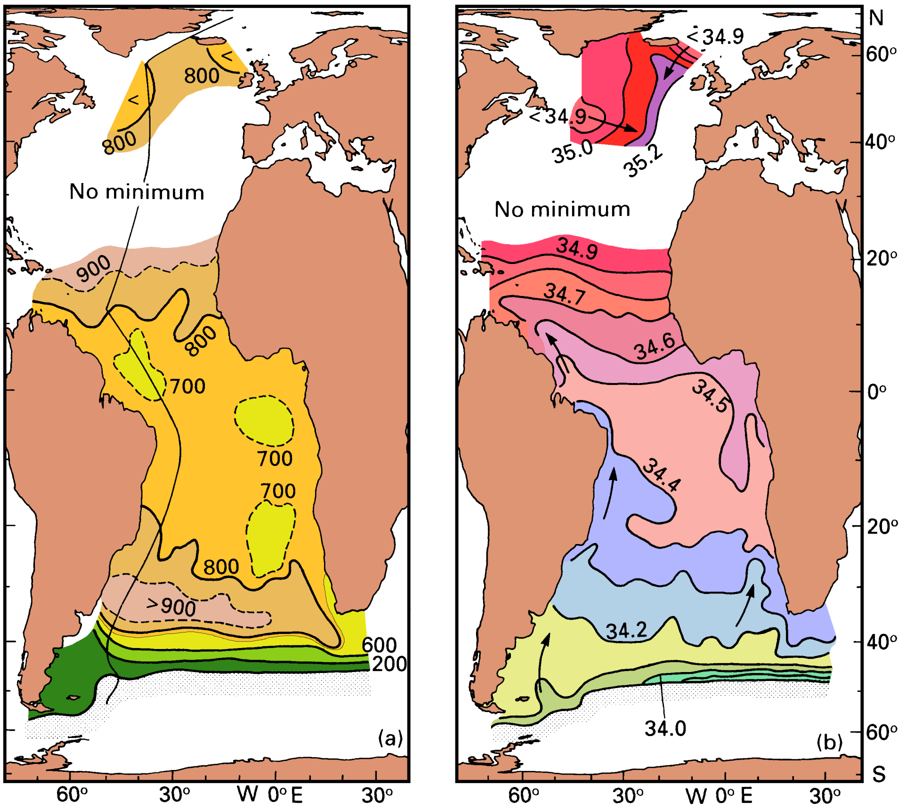

Fig. 15.7. Maps of the salinity minimum produced by the Intermediate Waters. (a) Depth of the minimum, (b) salinity at the depth of the minimum. The location of the section shown in Fig. 15.2 is also indicated in (a). After Wüst (1936) and Dietrich *et al*. (1980).

Above the Deep and Circumpolar Waters is the Intermediate Water, characterized as in the other oceans by its low salinity. Figure 15.7 gives the depth of the salinity minimum produced by the spreading of this water mass and the salinity at that depth. The outstanding feature is the pronounced lack of symmetry relative to the equator. The dominant water mass is the *Antarctic Intermediate Water* (AAIW). Formed mostly in the eastern south Pacific and entering into the Atlantic Ocean through Drake Passage and with the Malvinas Current, it spreads isopycnally into the northern hemisphere. Concentration of its flow along the western boundary is indicated by the northward extension of the isohalines with the Guyana and Antilles Currents and in Figure 15.6 by the widening of the isohalines around the salinity minimum at 1000 m depth. In the eastern basins its movement is masked by eddies, particularly Agulhas Current eddies propagating northward (Chapter 11). Observations along the south African continental rise near 1000 m depth (Nelson, 1989) indicate that AAIW participates in the cyclonic motion of abyssal water masses in the Cape Basin, even though the Walvis Ridge does not pose a barrier to flow at AAIW level.

As discussed in Chapter 6, formation of AAIW in the Atlantic Ocean occurs through water mass conversion in the Southern Ocean with limited direct atmospheric contact. The only region where winter convection contributes to AAIW formation is in the Scotia Sea. Most of the AAIW enters from a formation region in the eastern Pacific Ocean (England *et al.*, in press). As a consequence, Atlantic AAIW differs little from AAIW in the other two oceans. Close to the formation region it has a temperature near 2.2°C and a salinity of about 33.8. Mixing with water from above and below erodes the salinity minimum; by the time AAIW reaches the Subtropical Convergence it has properties closer to 3°C in temperature and about 34.3 in salinity. The gradual weakening and eventual disappearance of the minimum towards north can also be seen in the T-S diagrams of Figure 15.9.

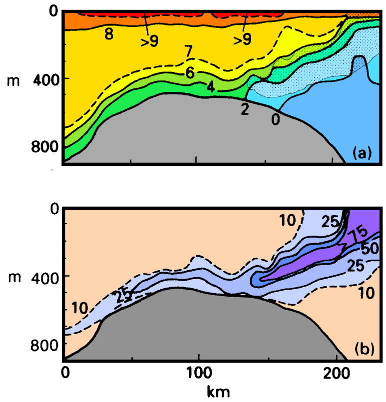

Fig. 15.8. A section across the Iceland-Scotland Ridge 150 km northwest of the Faroe Islands showing the spreading of Arctic Intermediate Water.
(a) Temperature (°C), with the region of low salinity (<34.9) shaded;
(b) presence of Arctic Intermediate Water (percentage of volume).
Adapted from Meincke (1978).

The occurrence of *Arctic Intermediate Water* (AIW) is restricted to two small regions in the north, a western variety formed in the southern Labrador Sea at temperatures near 3°C and 34.5 salinity, and an eastern variety which originates in the Iceland Sea at temperatures below 2°C and near 34.6 salinity. Both are subducted in locations along the Polar Front of the north Atlantic Ocean, in the west at the boundary between the Gulf Stream and the Labrador Current and in the east along a frontal region between the North Atlantic and East Iceland Currents. Their influence on the hydrography is limited by their proximity to the formation regions for NADW, which absorbs their low salinities over short distances. The eastern variety in particular cannot be recognized much beyond the sill where it sinks; its salinity minimum does not extend past 60°N (Figure 15.7). The influence of the western variety is felt most strongly in the North Atlantic Current, but its salinity minimum can be traced into the Bay of Biscay and southward to 40°N.

Because of the rapid absorption of Arctic Intermediate Water into the Deep and Bottom Water complex the existence of Intermediate Water in the north Atlantic Ocean is often ignored. A hydrographic section across the Iceland-Scotland Ridge (Figure 15.8), however, gives clear evidence that this water fits our definition of a water mass as a body of water with a common formation history: The Arctic Bottom Water, with salinities near 35.0 and temperatures below 3°C, is retained behind the sill and enters the Atlantic Ocean episodically, while the Intermediate Water sinks from the surface and is continuously subducted.

Water masses of the thermocline and surface layer

Two well-defined water masses occupy the Atlantic thermocline. Both are characterized by nearly straight T-S relationships. A south to north succession of T-S curves (Figure 15.9) shows a sudden shift to higher salinities some 1500 km north of the equator, indicating different hydrographic properties north and south of about 15°N.

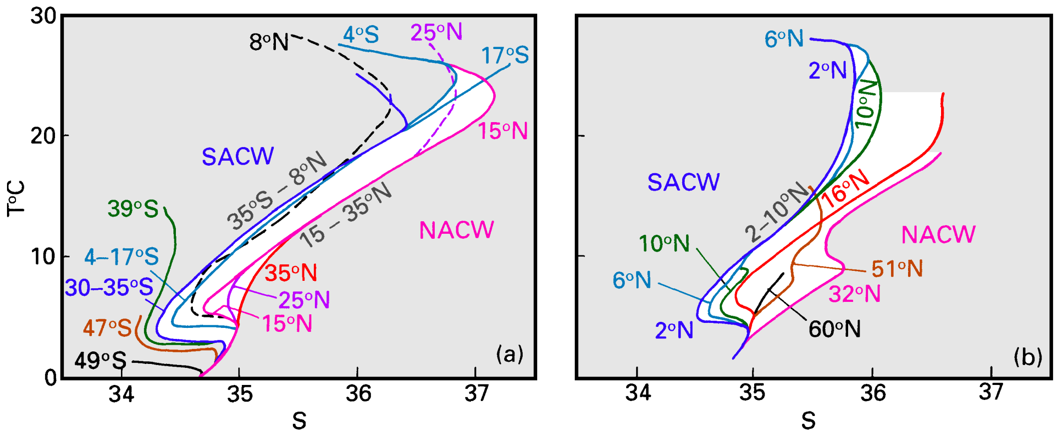

Fig. 15.9. T-S diagrams for stations along two meridional sections. (a) Western basins, (b) eastern basin (northern hemisphere). Note the northward weakening of the AAIW salinity minimum, the deep salinity maximum produced by the inflow of Eurafrican Mediterranean Water (most prominently at 32°N in the east), and the sudden transition from SACW to NACW south of 15°N. Data from Osborne *et al*. (1991).

South Atlantic Central Water (SACW), the water mass south of 15°N, shows rather uniform properties throughout its range. Its T-S curve is well described by a straight line between the T-S points 5°C, 34.3 and 20°C, 36.0 and is virtually the same as the T-S curves of Indian and Western South Pacific Central Water. This reflects the common formation history of all Central Waters in the southern hemisphere, which are subducted in the Subtropical Convergence (STC). Although the STC is well defined and continuous across the south Atlantic Ocean, detailed comparison between the T-S relationship along a meridional track across the STC with the T-S curve of SACW (in the manner described with Figure 5.4) reveals that the T-S properties of SACW in the tropics are closer to those typical for the Subtropical Convergence of the western Indian Ocean near 60 - 70°E, than those found along the STC in the Atlantic Ocean (Sprintall and Tomczak, in press). This indicates that much of the SACW is not subducted at the Atlantic portion of the STC but is in fact Indian Central Water (ICW) brought into the Atlantic Ocean by Agulhas Current eddies (see Chapter 11), in agreement with the ideas of North Atlantic Deep Water recirculation discussed in Chapter 7. Mixing in the eddy separation region and possibly in the Agulhas Current itself does not change the T-S characteristics of the inflowing ICW but redistributes the contributions of the water types which make up the T-S curve, enhancing in particular the volume of water near 13°C. This water type, also known as *13° Water* (Tsuchiya, 1986), thus turns into a variety of Subtropical Mode Water; the associated thermostad can be traced from Namibia to the coast of Brazil near 10°S, along the North Brazil Current and into the eastward flowing components of the equatorial current system. It is worth noting that, unlike other Subtropical Mode Water varieties, 13° Water is not formed in contact with the atmosphere.

Some Central Water formation does occur in the western south Atlantic Ocean, in the confluence zone of the Brazil and Malvinas Currents (Gordon, 1981). It is responsible for a high salinity variety of SACW (36 salinity is reached at 17°C instead of 20°C; see Figure 15.9). This SACW variety is recirculated within the southern subtropical gyre and therefore restricted to the western south Pacific Ocean.

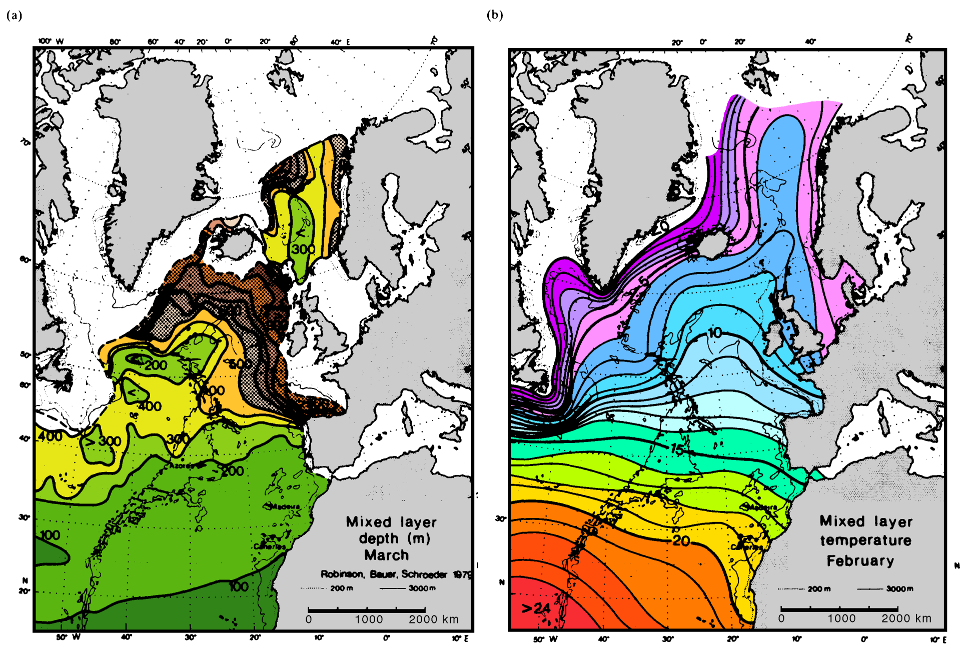

Fig. 15.10. Winter convection in the north Atlantic Ocean. (a) Mean mixed layer depth (m) in March, depths in excess of 600 m are shaded; (b) mean mixed layer temperature (°C) in February. Adapted from Robinson *et al.* (1979).

North Atlantic Central Water (NACW) can again be characterized by a nearly straight line in the T-S diagram, with some variation of the T-S relationship within the water mass. Typically, the T-S curve connects the T-S point 7°C, 35.0 with the points 18°C, 36.7 in the east and 20°C, 36.7 in the west. The regional differences stem from property variations in the formation region. Although curl(τ/f) is negative over most of the north Atlantic Ocean (Figure 4.3), a subtropical convergence as a region of more or less uniform subduction of Central Water from the surface cannot be identified. This is apparently because the subtropical gyre extends further north in the eastern north Atlantic than in the north Pacific Ocean, allowing the surface waters of the North Atlantic Current to cool much more during their northeastward passage than those of the North Pacific Current. Winter convection in the north Atlantic Ocean therefore reaches much deeper. Figure 15.10 shows that at the end of winter it reaches the bottom of the thermocline along the European shelf as far south as northern Spain. This affects all temperatures below 12°C. As a consequence, NACW enters the thermocline at those temperatures by a process of horizontal "injection" rather than isopycnal subduction, and its properties are influenced by water masses from the Arctic Mediterranean and Labrador Seas which participate in the convection. The deepest convection occurs in the northeast, and the corresponding vertical transfer of salt from the surface layer reaches its greatest depth there, raising the salt content of the eastern NACW variety.

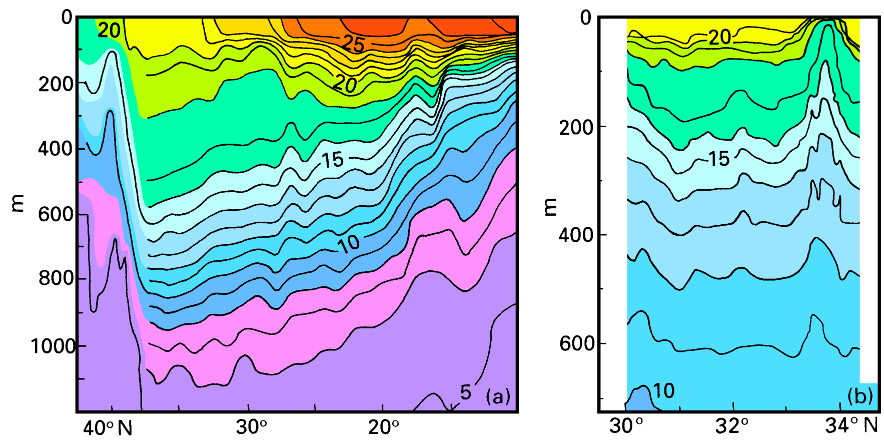

Fig. 15.11. Temperature sections indicating the presence of Subtropical Mode Water. (a) Through the Sargasso Sea along 50°W, (b) near Madeira along approximately 18°W. Note the increased isotherm spacing (thermostad) in the 17 - 18°C range. Adapted from McCartney (1982) and Siedler *et al*. (1987).

Above 12°C NACW is formed through the usual process, i.e. surface subduction of winter water. Again, this process does not occur uniformly across the region but involves Mode Water formation. Large volumes of Central Water are formed every winter in the Sargasso Sea at temperatures around 18°C. They appear in vertical temperature profiles as a permanent thermostad at 250 - 400 m depth (Figure 15.11) and represent a variety of

Subtropical Mode Water known as the *18° Water*. A third variety, the *Madeira Mode Water*, is formed north of Madeira and indicated by a summer thermostad at 70 - 150 m depth (Figure 15.11). Both mode waters together contribute more than half the volume of NACW.

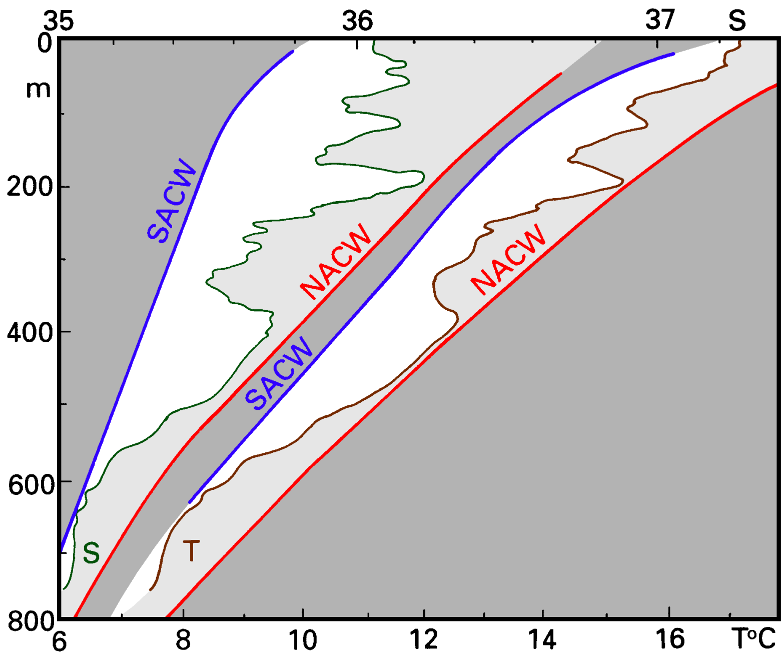

Fig. 15.12. (Left) Temperature (°C) and salinity as functions of depth at a station in the eastern part of the water mass boundary between SACW and NACW, showing strong interleaving. Reference curves for undisturbed SACW and NACW conditions are indicated. The station is located some 25 km west of the position marked in Fig. 14.17. From Tomczak and Hughes (1980).

The transition from South to North Atlantic Central Water occurs as a front along approximately 15°N which extends from below the mixed layer to the bottom of the thermocline. In the east it bends northward past 20°N, following the southern limit of the Canary Current and sharpened by the confluence with the circulation around the Guinea Dome. In general, SACW penetrates northward underneath NACW (see Chapter 14 for details), giving the front a downward slope from south to north. With SACW and NACW occupying the same density range, the front is density-compensated, i.e. the effect of the temperature change across the front is compensated by the effect of the salinity change and the front is not noticed in the density field. Parcels of water from either side of the front can therefore be moved easily across the front on isopycnal surfaces. The resulting multitude of intrusions, filaments, and lenses (Figure 15.12) makes the structure of the front quite complicated. Further west the front loses its identity, as mixing between SACW and

NACW in the North Equatorial Current erodes the horizontal gradients. Eventually the mixture is carried north in the Guyana and Antilles Currents to complete the route of SACW from the Agulhas Current eddies to the formation region of North Atlantic Deep Water.

Because the separation zone between both Central Waters is located more than 1500 km north of the equator and the SACW/NACW mixture produced in the west is not returned into the equatorial current system but transported northward, there is no opportunity to form a special equatorial water mass in the manner seen in the Pacific Ocean. The opposing eastward and westward equatorial flows leave, however, their mark in the hydrographic properties of SACW. Observations from the region of the Guinea Dome show a small but well-defined salinity decrease, from south to north, across the boundary between the North Equatorial Countercurrent and the North Equatorial Current near 10°N (Figure 15.13). The front between the two SACW varieties slopes downward toward the north, the low salinity variety moving eastward above westward movement of the high salinity variety.

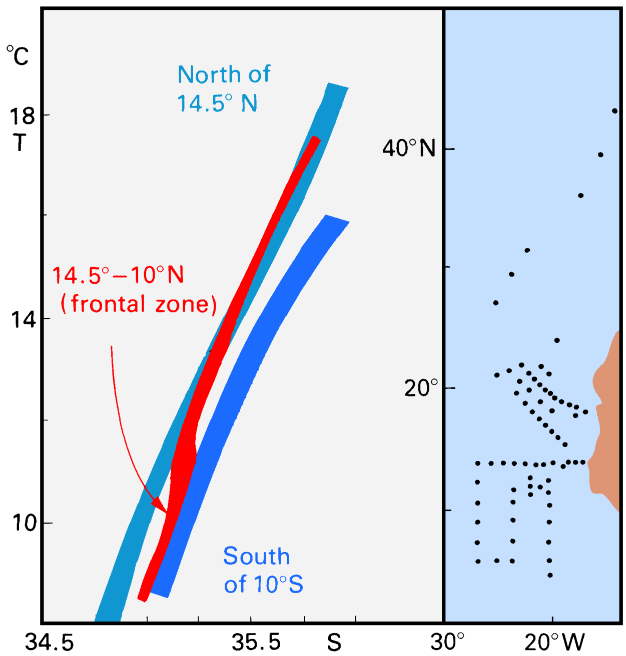

Fig. 15.13. (Right) T-S relationships in the South Atlantic Central Water of the eastern tropical Atlantic Ocean, showing two SACW varieties separated by a frontal zone between the North Equatorial Current and the Equatorial Countercurrent. The stations used for the construction of the T-S diagram are shown on the right.

Main aspects of the hydrographic structure above the permanent thermocline were already discussed in an earlier section of this chapter. A major feature of the tropical Atlantic Ocean is the existence of a barrier layer in the region of the Guyana Current (Figure 15.14). This region is characterized by net water loss to the atmosphere (the major region where rainfall exceeds evaporation being east of 40°W; Figure 1.7). Local freshening of the surface layer, the mechanism that produces the barrier layer in the Pacific Ocean, can therefore not be responsible here. It appears that the high salinities found at the surface in the subtropics (Figure 2.5b) are subducted towards the equator at the upper end of the temperature/salinity range of the Central Water. This creates a salinity maximum above the Central Water in the tropics, which is then advected westwards towards the equator into regions of uniform temperature in the equatorial current system. The result is salinity stratification in the isothermal surface layer. Figure 15.14 shows active subduction (indicated by negative barrier layer thickness; see Chapter 5 for an explanation of the mechanism) for August - October south of 12°S, coupled with the formation of a barrier layer to the north. During February - April the same process occurs in the vicinity of 20°N. The two sources alternate in renewing the barrier layer structure in the west, where the barrier layer is found during all seasons. In this region an accurate heat and mass budget of the surface layer cannot be achieved without taking advection into account.

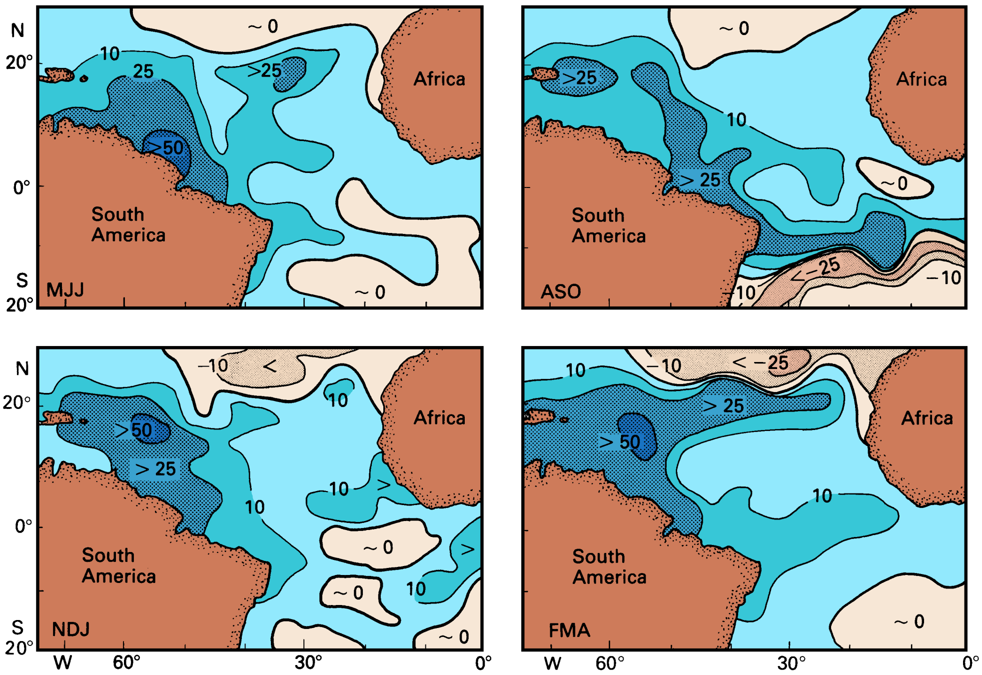

Fig. 15.14. Seasonal mean barrier layer thickness (m) in the tropical Atlantic Ocean. (a) May - July, (b) August - October, (c) November - January, (d) February - April. The barrier layer is located below the mixed layer (see Fig. 5.6 for mixed layer depth). Contours are given for layer thickness of 50 m, 25 m, 10 m, 0 m, -10 m, and -25 m. Subduction regions, indicated by values less than -10 m, are lightly shaded, regions with a barrier layer thickness >25 m are shown with dark shading. Adapted from Sprintall and Tomczak (1992).

A description of the hydrographic conditions in the shelf regions of the Atlantic Ocean is beyond the scope of this book, but one region deserves mention. The large volume of water between the Gulf Stream and the continental shelf is isolated from direct contact with the oceanic water masses of its depth range by the western boundary current. Its properties are formed through a complex process of interaction between water on the shelf, from the Labrador Current, and from the Gulf Stream. Water on the shelf has very low salinity (below 34, a result of freshwater inflow from the St. Lawrence River). The Labrador Current also carries low salinity water. Mixing of the various components produces a water mass known as Slope Water, which extends over the upper 1000 m along the north American continental rise north of Cape Hatteras (35°N) and is characterized by a nearly linear T-S relationship similar to that of NACW but with much lower salinity. This water is frequently trapped in cyclonic Gulf Stream Rings and transported across the Gulf Stream into the Sargasso Sea, as seen in the example of Figure 15.15. As a result, variations of hydrographic properties in the permanent thermocline of the Atlantic Ocean are largest in the northern Sargasso Sea.

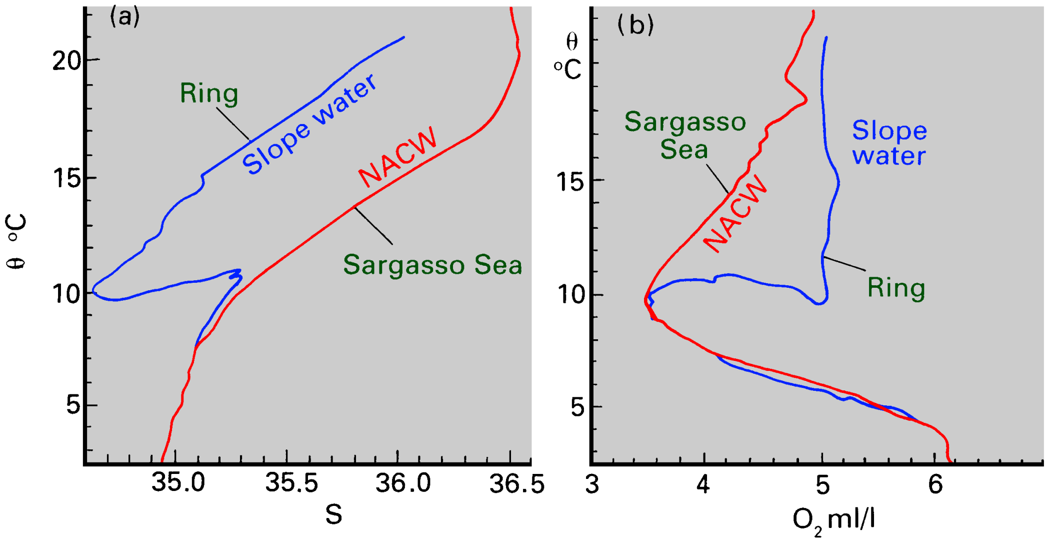

Fig. 15.15. An example of Slope Water advection into the Sargasso Sea in a Gulf Stream ring. (a) Θ-S diagram, (b) Θ–oxygen diagram, for a station outside ("Sargasso Sea") and inside ("Ring") a cyclonic Gulf Stream ring. Higher oxygen concentration in the Slope Water indicates more recent contact with the atmosphere. From Richardson (1983b).