where the density of the oceanic water matches its own; it then spreads on this density surface, where it can be traced through the ocean basin by its high salinity. (A prominent example is the Eurafrican Mediterranean Sea; its impact on the salinity of the Atlantic Ocean has already been noted in the discussion of Figure 2.5.) Because the salinity of the oceanic water is increased as it passes through the mediterranean sea, this type of mediterranean sea is also known as a concentration basin.

If, on the other hand, precipitation over the mediterranean sea exceeds evaporation, the freshwater gain drives an outflow into the ocean basin through the upper layer. It also decreases the surface density, and the resulting density difference at the sill causes inflow of oceanic water through the lower layer and additional compensating outflow in the upper layer. A strong pycnocline is established, and renewal of the deeper waters is strongly inhibited. Inflow of oceanic water is usually the only renewal process of any significance, and if the connection across the sill is narrow or the deep water volume large, deep water renewal is not always sufficient to prevent the depletion of oxygen in the deep basins. In these cases the sea is devoid of life (apart from sulfur-reducing bacteria) below the pycnocline. This type of mediterranean sea is also known as a dilution basin.

Mediterranean seas are a special class of marginal seas, which are defined as those parts of the World Ocean that are separated from the major deep ocean basins by topographic features such as islands or bay-like coastline configurations. Examples of marginal seas are some of the major shelf regions, e.g. the North Sea or the East China Sea, and topographically semi-enclosed ocean regions, e.g. the Tasman Sea or the Bay of Bengal. While the circulation and stratification in these marginal seas may be strongly modified by thermohaline or tidal forcing it is still dominated by the wind. Mediterranean seas are the only marginal seas where thermohaline forcing dominates. (For readers familiar with estuarine dynamics, mediterranean seas can be defined as those marginal seas which display a circulation of the estuarine type.)

Bottom topography

A look at the topography of the Arctic Seas (Figure 7.2) clearly establishes their mediterranean character. The major connection with the three oceans is to the Atlantic Ocean where a 1700 km wide opening exists along a large oceanic sill running from Greenland across to Iceland, the Faroe Islands and Scotland. Approximate sill depths are 600 m in Denmark Strait (between Greenland and Iceland), 400 m between Iceland and the Faroe Islands, and 800 m in the Faroe Bank Channel (between the Faroe Islands and Scotland). Minor openings to the Atlantic Ocean exist through the Canadian Archipelago, mainly through Nares Strait and Smith Sound with a sill depth of less than 250 m and Barrow Strait and Lancaster Sound with about 130 m sill depth. The connection with the Pacific Ocean through Bering Strait is only 45 m deep and 85 km wide and of little consequence for the Arctic circulation (It is important for the global freshwater balance; see Chapter 18).

Within the confines of the Arctic Mediterranean Sea are the Greenland, Iceland, and Norwegian Seas, and the Arctic or North Polar Sea proper which includes the various regions of the large Siberian shelf area, i.e. (beginning at Bering Strait and moving westward) the Chukchi, East Siberian, Laptev, Kara, Barents, and White Seas, and the Lincoln and Beaufort Seas on the Greenland-Canadian-Alaskan shelf. The Greenland,

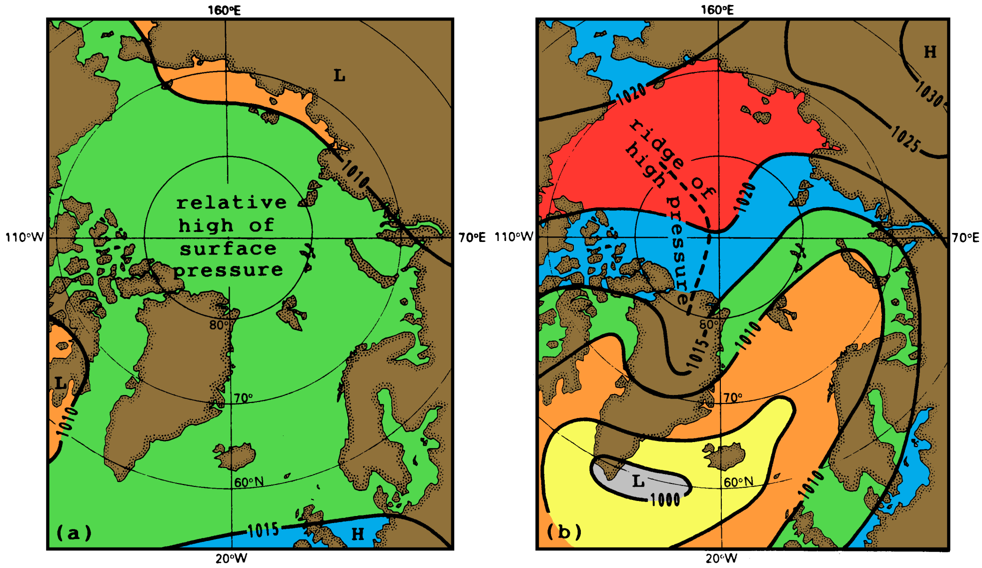

Fig. 7.3. Air pressure (hPa) at sea level over the Arctic Mediterranean Sea. (a) July mean, (b) January mean, both for the period 1950 - 1980. Data from University of East Anglia (1992).

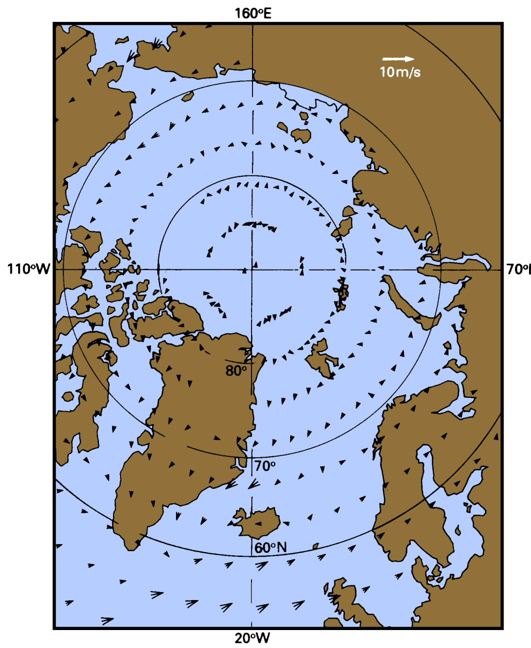

Fig. 7.4. Surface winds over the Arctic Mediterranean Sea.

(a) annual mean,

(b, page 87) July mean,

(c, page 87) January mean. See Figure 1.2 for data sources.

The Arctic Mediterranean Sea has a total volume of $17 \cdot 10^6$ km^3, which is 1.3% of the World Ocean. If the seas south of Fram Strait are excluded, the Arctic Sea proper covers an area of about $12 \cdot 10^6$ km^2 and contains $13 \cdot 10^6$ km^3 of water. This represents about 3% of the World Ocean area but only 1% of its volume. The reason is found in the large expanse of shelf area. Along the American coast the shelf is only 50 - 90 km wide, but on the Siberian side its width exceeds 800 km in most places. The shelf is also rather shallow, 20 - 60 m in the Chukchi Sea, and probably similar in the East Siberian Sea, 10 - 40 m in the Laptew Sea, an average depth of 100 m in the Kara Sea, and 100 - 350 m in the Barents Sea. It represents nearly 70% of the surface area of the Arctic Sea. Numerous large rivers empty into the Arctic shelf seas, reducing their salinity. These shallow shelf areas therefore greatly influence surface water conditions in the Arctic Mediterranean Sea.

The wind regime

Again, Figures 1.2 - 1.4 contain the relevant information but projection on polar coordinates gives the better representation. Figures 7.3 - 7.4 show air pressure and surface winds over the Arctic region. The comments made about the validity of wind stress estimates over the Southern Ocean are even more relevant here, since the Arctic ocean includes ocean basins at very high latitudes where wind observations are extremely sparse. Nevertheless, while the winds of Figure 7.4 may not give the magnitudes of all wind stresses right they give some indication of their directions.

High pressure in the vicinity of the north pole determines the wind system over the Arctic Mediterranean Sea through the year. It is more prominent during winter when it takes the form of a ridge from the Canada Basin towards northern Greenland. Pressure gradients are reduced during summer, but the pressure near the pole is still higher than over the continents. Most of the Arctic seas are therefore under the influence of the Polar Easterlies and display anticyclonic (west-ward) surface circulation, in contrast to the Southern Ocean where the effect of the Polar Easterlies is only noticeable in a weak westward current along the Antarctic continent. Winds are much stronger in winter, and the annual mean resembles the January distribution (Figure 7.4). Over the Greenland and Norwegian Seas the wind system is dominated by the Icelandic atmospheric low, which generates cyclonic water movement.

The estimated surface circulation, shown in Figure 7.5, is derived from drift tracks of research stations on ice islands and ships, trapped in the ice either accidentally or deliberately. The most famous of these crossings of the polar seas was the drift of the research vessel Fram, built by the Norwegian explorer Fritjof Nansen to withstand the pressure of the ice, during 1893 - 1896. Earlier, the American vessel Jeanette had been caught in the ice of the Chukchi Sea near Bering Strait in November 1879 and crushed by the ice in June 1881; wreckage from that vessel was recovered in 1884 on the southwest coast of Greenland. Other early observations include the drifts of the Maud from Bering Strait to the New Siberian Islands in 1918 - 1925 and of the Sedov which took $2^1/_2$ years to drift from the New Siberian Islands to Spitsbergen during 1937 - 1940.

These and the numerous drift tracks from ice stations (such as NP1, NP4 and T3 in Figure 7.5) of the last forty years establish a picture of mean westward circulation around a centre of motion close to the centre of the atmospheric high, with outflow from the Arctic Mediterranean Sea along the coast of Greenland and inflow along the Norwegian coast. The

on the eastern side. The inflowing water has its origin in the temperate and subtropical gyres of the North Atlantic Ocean. Its low density is due to the high temperatures in the Gulf Stream extension. Current speeds in the East Greenland and Norwegian Currents are usually in the vicinity of 0.2 m s^{-1} but can reach 0.5 m s^{-1} on occasions. Data from current meter moorings deployed in the East Greenland Current for one year (Foldvik *et al.*, 1988) show a decrease of average current speed with depth to about 0.05 m s^{-1} at 600 m; the passage of eddies can increase these values more than threefold or reverse the flow. Generation of strong currents in the polar anticyclonic gyre by the wind is inhibited by permanent ice coverage; average speeds are close to 0.02 m s^{-1} (2 km per day). Some indication of eastward flow produced by the West Wind Drift is seen on the Siberian shelf. However, measurements in these regions are sparse, and the currents are likely to be influenced by coastline topography and river inflow; so most of the estimates of flow on the shelf are hypothetical. Eastward flow on the Alaskan shelf is well documented as a wind-driven extension of the inflow through Bering Strait.

Precipitation and ice

Compared to conditions for data collection in the Southern Ocean, collection of rainfall and snowfall data on the stable platforms of drifting ice stations is no serious challenge. Reasonable information on rainfall and snowfall is therefore available. Precipitation is low in the region of the Polar Easterlies but significant in the subpolar regions which are dominated by the West Wind Drift and its associated large variability and high storm frequency. Since most of the precipitation occurs over ice which does not melt until it is exported from the Arctic region, local snowfall does not play a major role in the oceanic mass budget. The major contribution comes from precipitation over Siberia and the resulting river run-off, estimated in total as 0.2 Sv. With evaporation over ice being comparatively low, the Arctic Mediterranean Sea is a dilution basin, i.e. its outflow is fresher on average than its inflow.

Precipitation over Greenland feeds the glaciers which produce the several thousand icebergs annually found in the East Greenland, West Greenland, and Labrador Currents. A few of them - and in winter quite a few, between 50 and 100 - reach the area south of the Newfoundland Bank and enter the main shipping route between North America and Europe. When the cruise liner Titanic hit one of these on her maiden voyage in 1912 and sank, taking 1490 lives, the International Ice Patrol Service was established. Between March and July it monitors and reports the positions of all icebergs that may become a danger to transatlantic shipping.

A remarkable feature of the ice distribution is the extreme southward extent of the ice-affected region along the American continent and the extreme northward extent of the region which is permanently ice-free along the Norwegian coast (Figure 7.6). Nowhere else in the world ocean can ports at 70° latitude be reached by sea during the entire year, as is the case with the Norwegian cities of Tromsö and Hammarfest; and nowhere else do icebergs reach 40° (the latitude of southern Italy) as they do south of the Newfoundland Bank. This is of course the result of the temperature difference between the outgoing and incoming water which is so marked that it can even be seen in severely space and time averaged data (for example in Figure 2.5). The average temperature of the Norwegian Current is 6 - 8°C; the average temperature of the East Greenland Current is below -1°C. This temperature

the sill. The second modification concerns the renewal of the water in the deep basins. The water below the sill depth is several degrees colder than the water entering the Arctic Seas from the Atlantic Ocean and is thus denser. The inflowing water therefore does not sink but spreads through an intermediate layer, and a hydrographic station from anywhere in the Arctic Mediterranean Sea (the Norwegian Sea being the only exception) displays a layering of three water masses (Figure 7.7). Finally, as a third modification it will be seen that outflow of water from the Arctic Seas is not restricted to the surface layer.

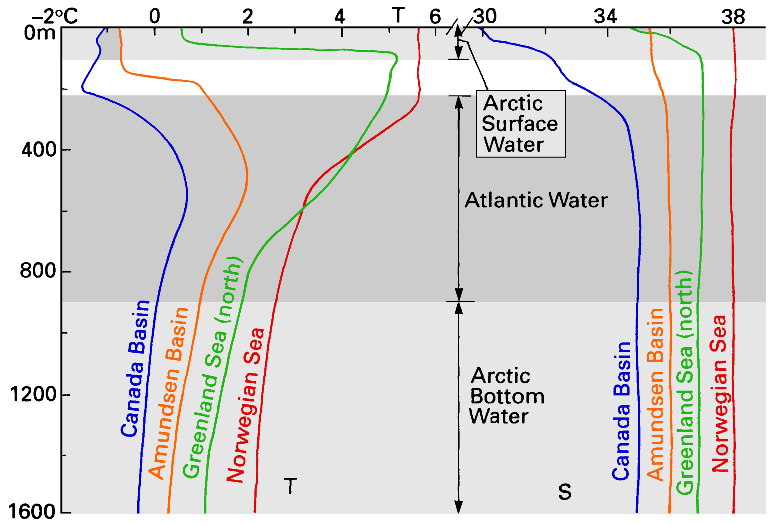

Fig 7.7. Temperature (T °C) and salinity (S) profiles in the Arctic Mediterranean Sea. The scale is correct for the Canada Basin; other profiles are offset by 1 unit from each other. Arctic Surface Water is not present in the Norwegian Sea, where inflow of Atlantic Water extends to the surface. From Coachman and Aagaard (1974) and Osborne *et al*. (1991).

We begin the discussion of the water masses by looking at *Arctic Bottom Water*. For a long time it was believed that its formation region is in the Greenland or Norwegian Seas. It is now known that the formation process involves the interplay of two sources, Greenland Sea Deep Water and water from the Arctic shelf regions. *Greenland Sea Deep Water* is formed during winter in the central Greenland Sea, where the cooling of surface water causes intense vertical convection. Sinking of water to the bottom occurs in events, clearly related to the passage of storm systems; the events last less than a week and are limited to regions a few kilometers across. In each event, individual cooling cycles occur on even smaller time and space scales. At the beginning of each cycle the surface layer is quite fresh (see Figure 7.7), and concentration of salt is required to initiate sinking. This is achieved by ice formation. Eventually the density increase is sufficient to overcome the barrier posed by the warm but saline water below. Sinking sets in and is compensated by upwelling of the warm water which melts the ice, bringing the cycle to an end. The

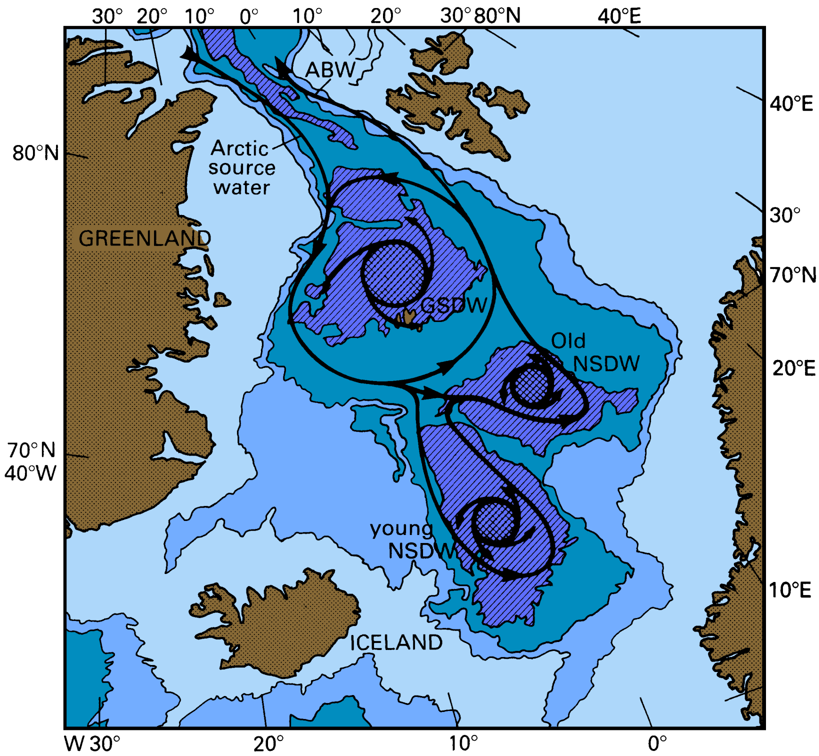

Fig. 7.9. Diagram of Arctic Bottom Water formation showing the circulation at and below 2000 m depth. Approximate 1000 m, 2000 m, and 3000 m contours are shown, with basins deeper than 3000 m shaded. "Arctic source water" is the water from the Amundsen and Nansen Basins which contain contributions from the Arctic shelf. GSDW: Greenland Sea Deep Water, NSDW: Norwegian Sea Deep Water. Deep convection occurs in the cross-hatched areas. ABW: Arctic Bottom Water, the product of the mixing process. After Smethie *et al*. (1988)

Figure 7.9 is a schematic representation of Arctic Bottom Water formation. The Norwegian Sea, which apparently does not cool enough to experience deep winter convection, nevertheless plays an important role in the formation process as a mixing basin where Arctic Bottom Water obtains its final characteristics. Greenland Sea Deep Water, the densest component, is confined to the centre of a cyclonic gyre in the centre of the Greenland Sea. Its temperature is consistently below -1.1°C, lower than the temperature of *Norwegian Sea Deep Water* (-0.95°C), and the lowest of all bottom water in the Arctic Mediterranean Sea. It entrains from its flanks water from the depths of the Amundsen and Nansen Basins which contains the contribution from the Arctic shelf. A salinity maximum in the East Greenland Current at 1500 m depth is interpreted as evidence for the presence of the shelf contribution; Smethie *et al*. (1988) estimate the amount of Arctic Bottom Water produced on the shelf at up to 0.1 Sv and the sinking of surface water at 0.5 Sv. Further downstream, about 1 Sv of Greenland Sea Deep Water enters the southern Norwegian Sea,

Bottom Water formation but sufficiently high to bring the water into contact with Atlantic Water has been estimated at 2.5 Sv (Aagaard *et al*., 1981), not much less than the flow of Atlantic Water into the Arctic Sea. With its very low temperature (near the freezing point) the sub-surface layer of sunken shelf water acts as a heat shield for the surface layer: By entraining Atlantic Water in the canyons and reducing its temperature sufficiently, the water of the sub-surface layer prevents the Atlantic Water from melting the ice layer above.

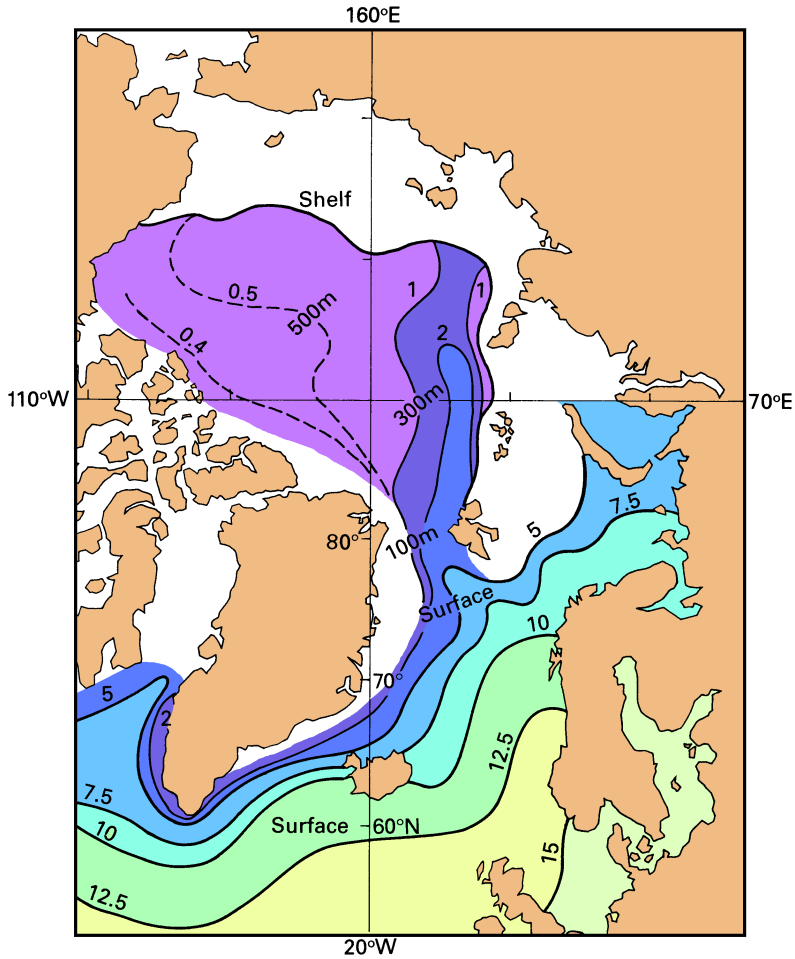

Fig. 7.10. Temperature (°C) at the depth of the temperature maximum caused by inflow of Atlantic Water during summer. Approxi-mate depths are indicated.

A remarkable feature of the sub-surface layer is that it contains the swiftest currents of the Arctic Seas. They are usually of the order of 0.3 - 0.6 m s^{-1}, last only for up to a fortnight, and appear to be linked with the movement of subsurface eddies, or lenses. The lenses are an indication for the intensity of the mixing; they are a few tens of kilometers in

be quite large, and knowledge of long-term mean transports in the major contributors to the mass budget is still unsatisfactory.

The Arctic Mediterranean Sea receives water from the Norwegian Current, through Bering Strait, and from river run-off. On average, the same amount of water leaves the Arctic Mediterranean Sea through the East Greenland Current, the Canadian Archipelago, as meltwater and ice, and in the overflow of Arctic Bottom Water discussed below. The Norwegian Current transports about 10 Sv of Atlantic Water northward. Some 4 Sv leave the Norwegian Sea towards the Atlantic Ocean, as outflow of Arctic Bottom Water across the Greenland-Iceland-Scotland Ridge (see below). Of the remaining 5 - 6 Sv, the West Spitsbergen Current carries 3 - 5 Sv into the Amundsen and Nansen Basins, while 1 Sv flows through the Barents Sea and enters the Arctic region between Franz Josef Land and Novaya Semlya. Transport through Bering Strait is well documented (Coachman and Aagaard, 1988) as ranging between 0.6 Sv in winter and 1.1 Sv in summer, while river run-off is estimated at 0.2 Sv. Outflow in the East Greenland Current is estimated at about 3 - 5 Sv, which includes 0.1 - 0.2 Sv (4000 - 5000 km^3 per year) of meltwater and ice. Transport through the Canadian Archipelago into the Atlantic Ocean via Baffin Bay is estimated at 1 - 2 Sv.

A summary of the mass budget is presented in Figure 7.12. Obviously the figures are adjusted to give zero total balance. However, the various contributions differ by an order of magnitude, and errors in the estimates for the major components exceed most of the minor contributions to the budget. All figures, particularly those for the major components, have to be seen as representing our best knowledge to date but are not final.

An unresolved feature of the budget is the volume flow in the recirculation of the Norwegian and Greenland Seas. Recent estimates based on tracer measurements give figures for the recirculation of Arctic Bottom Water in the region as low as 1 Sv, while direct current measurements, which include the recirculation of Atlantic Water (Coachman and Aagaard, 1974), lead to figures as high as 25 Sv. The huge differences are of little consequence for estimates of the mass exchange between the Arctic Mediterranean Sea and the Atlantic Ocean but are crucial for a correct assessment of the mixing processes in the seas south of Fram Strait.

The Arctic region plays a major role in the world's climate, and changes in heat content of the Arctic Mediterranean Sea are likely to influence future climate trends in most parts of the world. Establishing an Arctic heat budget is therefore an important task. Among its prerequisites is a correct mass budget, since heat advection by currents is responsible for the net heat gain of the region. The above estimates of volume flow require some refinement for that purpose; for example, the 5 Sv transported in the East Greenland Current consist of about 1 - 2 Sv of Arctic Surface Water and 3 - 4 Sv of Atlantic Water. Using this kind of split and the average temperature and salinity of both water masses, the heat and freshwater transport of the East Greenland Current can be established. It is found (Aagaard and Greisman, 1975) that on average, the Arctic Sea gains $108 \cdot 10^{12}$ W (most of it through the import of warm Atlantic Water in the West Spitsbergen Current but up to 30% through the export of ice and another 15% through the export of Arctic Surface Water in the East Greenland Current, through the Canadian Archipelago, and through Bering Strait). This heat flux must compensate for the heat loss experienced at the surface. The total salt budget is of course balanced; but the split into components gives an idea of the degree of mixing experienced in the Arctic Sea: The Atlantic Water which enters the region with the West Spitsbergen Current imports about $250 \cdot 10^6$ kg of salt per year; the Atlantic Water which

entraining some 3 Sv of water from the side. This flow continues along the Mid-Atlantic Ridge until it encounters the Gibbs Fracture Zone, a break in the Ridge deep enough to allow passage of water below 3000 m depth. The western overflow increases its transport by entraining at least another 3 Sv of water along its path. It follows the continental slope around southern Greenland where it is joined by the other overflow component. This component is slightly warmer and more saline (1.8 - 3.0°C and 34.98 - 35.03 salinity) than the western component (0.0 - 2.0°C and 34.88 - 34.93), a result of the higher temperatures and salinities of the water entrained in the east.

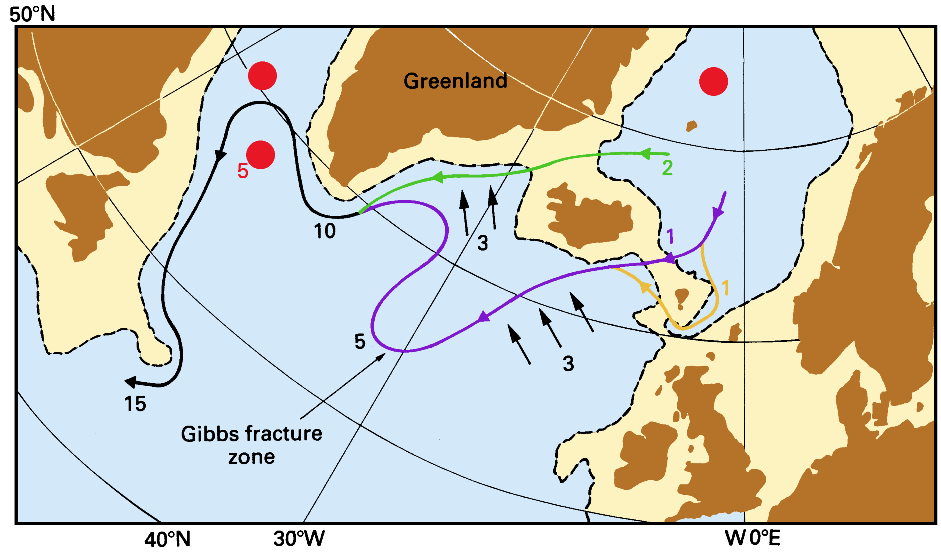

Fig. 7.13. The path of Arctic Bottom Water overflow. Numbers indicate volume transport in Sv. Deep convection is indicated by dots. The broken line is the 1000 m contour.

Observations of tritium concentrations in the overflow have shown (Peterson and Rooth, 1976) that the source of the overflow is not water from the bottom of the Norwegian Sea but from depths close to 1000 m. This is of course to be expected since the water has to pass over shallow sills and does that only intermittently when water residing behind the sill is lifted up by a hundred meters or so. The triggering mechanism for this uplift are atmospheric disturbances. Intense storm systems locally generate cyclonic winds. The corresponding Ekman transports are set up within a matter of hours, and they point outward from the centre of the low pressure region, causing upwelling. This is accompanied by a depression of the sea surface which, by our Rule 1a of Chapter 3, is reflected in a rise of the thermocline. The Greenland-Iceland-Faroe-Scotland Ridge is in the West Wind region and therefore sees plenty of storm systems passing by. Each storm lifts the water behind the sill above the sill depth and produces an overflow event. The combined effect of all these events is a flow of 4 Sv of Arctic Bottom Water into the eastern Atlantic Ocean.

Like the Greenland Sea, the Labrador Sea is a region of intense surface cooling and deep winter convection. On arriving in the Labrador Sea from the southern (Atlantic) entrance,

and there is no doubt that it describes a valid recirculation path for North Atlantic Deep Water. Whether it is the major path is not certain, and the transport estimates involved have yet to be firmly tested. According to Figure 7.15 the rate of North Atlantic Deep Water formation should be equal to the rate of water loss of the Agulhas Current to the Atlantic Ocean which, as we shall see in the discussion of the Indian Ocean, carries most of its water back to the east. It also assumes that all North Atlantic Deep Water upwells in the Southern Ocean and no fraction of it enters the Atlantic Ocean again from the Pacific Ocean after one or more complete cycles in the Circumpolar Current, to recirculate northward. However, in Chapter 9 it will be seen that North Atlantic Deep Water can be identified in the abyssal layers of the Pacific Ocean. How much of it passes through Drake Passage into the Atlantic Ocean is unknown at present. There is also evidence that large amounts of NADW make their way through Drake Passage in the form of Antarctic Intermediate Water (Rintoul, 1991).

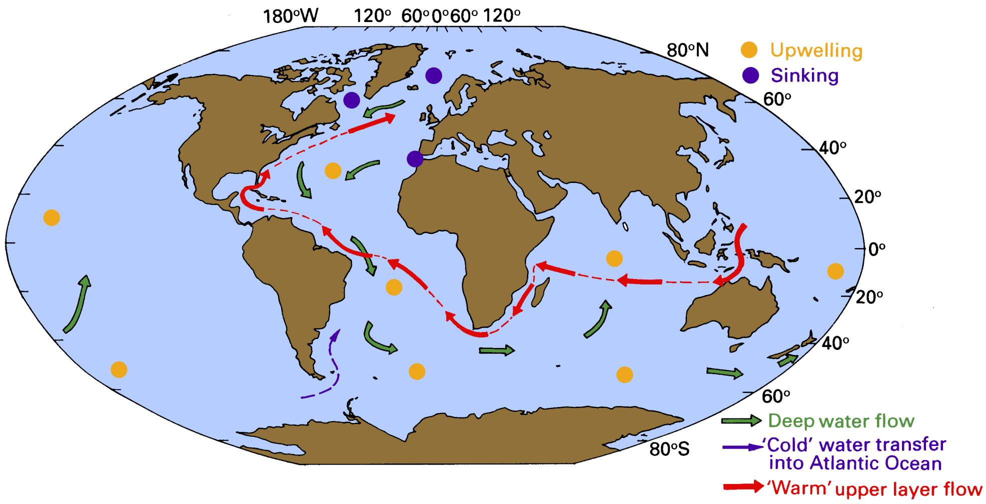

Fig. 7.15. The recirculation path of North Atlantic Deep Water (NADW) through the world ocean. Open arrows indicate flow of NADW and Antarctic Intermediate Water produced in the western south Atlantic Ocean from upwelling of NADW at the Antarctic Divergence. Dots and crosses indicate movement into and out of the thermocline. Full arrows indicate flow of Central Water. Broken arrows indicate deep flow of NADW which is considered insignificant in Gordon's model. Adapted from Gordon (1986a).

The second aspect concerns changes of temperature and salinity along the recirculation. From Figure 7.16 it can be seen that temperature along the path mainly reflects warming during the upwelling and cooling during the sinking. Salinity, on the other hand, undergoes repeated changes, indicating various episodes of mixing with surrounding waters, dilution from rainfall, and evaporation. Occasionally such episodes occur in well defined regions and influence the water properties to such a degree that the water can be identified as a new water mass. As an example, the low salinity water which leaves the Indonesian Archipelago has become known as Australasian Mediterranean Water. From the point of view of the