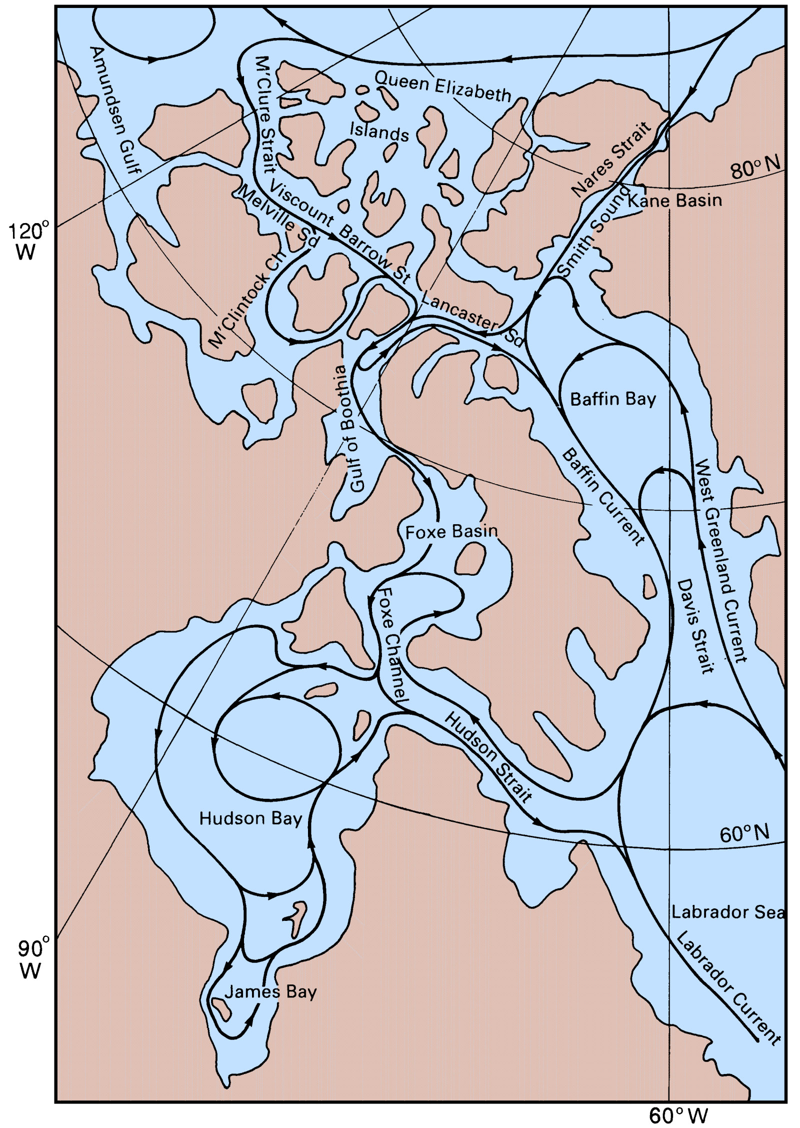

Fig. 16.1. Geography and circulation of the Northwest Passage and connected seas.

The hydrography of *Baffin Bay* and *Northwest Passage* shows a distinct layering of water masses. Unfortunately a generally accepted nomenclature of water masses in the region does not exist, and some names found in the literature can be quite misleading. Two conspicuous features of the vertical temperature distribution in Baffin Bay are a temperature maximum at about 500 m depth below a minimum in the range 50 - 200 m (Figure 16.2). The maximum is the result of Atlantic water inflow with the West

the character of Baffin Bay as a small mediterranean dilution basin and correspondingly slow renewal of its Bottom Water. The details of the formation process are not entirely clear. Irregular discharge of Arctic water through Smith Sound is believed to contribute. There is also evidence that formation of cold, saline water in shallow regions of Smith Sound and subsequent sinking along the continental slope, similar to the formation process of Arctic Bottom Water (Chapter 7), plays a major role (Bourke *et al*. 1989).

The hydrography of *Hudson Bay* is determined by its small water depth of on average only 250 m and by seasonally varying river discharge, which gives it the character of a large estuary. Since estuarine dynamics are not discussed in this book, a few remarks have to suffice. Foxe Basin and Hudson Bay are completely ice-covered during several months. The ice starts to break up from James Bay in June, and Hudson Bay and Foxe Basin are clear of ice by mid-August. Large amounts of freshwater are poured into the bay during this period. As a consequence, temperature and salinity undergo large seasonal changes with a period of strong stratification during late spring and summer. This affects the circulation as well, so Figure 16.1 can only show the water movement as it prevails during two thirds of the year. Lack of ice movement during spring breakup indicates that there is little if any water movement before the stratification is established by the influx of meltwater. The effect of the runoff is felt most strongly towards October when flow across the entire northern bay is northward.

Ice coverage in other parts of Northwest Passage is similarly heavy, but satellite observations show that up to 10% of the region between Lancaster Sound, Smith Sound, and Greenland are ice-free at any time even through winter. The polynya are produced by offshore movement of ice in response to the prevailing wind. The dynamics are essentially those of coastal upwelling, but the situation is unique because in the upper few hundred meters temperature increases with depth. Upwelling thus brings warm Atlantic Polar Water to the surface and keeps the surface layer ice-free.

The Eurafrican Mediterranean Sea

The mediterranean basin between Africa, Europe, and Asia has always been regarded the prototype of a concentration basin, to the extent that it is mostly referred to not as the Eurafrican but simply *the* Mediterranean Sea. It consists of a series of deep basins mostly well connected with each other, the major exception being the Black Sea which has very limited communication with the other subdivisions (Figure 16.3). The mean depth of the Mediterranean Sea is near 1500 m; maximum depths in the various basins are between 2500 m and 5100 m (in a narrow trench off southwestern Greece). A sill between Sicily and Tunisia with a maximum depth near 400 m divides the region into the western and eastern Mediterranean Sea. Maximum depths in the two subdivisions are about 3400 m in the west and about 4200 m in the east (if the deep trench is excluded). The second connection between the western and eastern basins, the narrow Strait of Messina between Sicily and mainland Italy, has a sill depth of only 120 m and is of no significance for the general circulation; its reputation as a treacherous passage for ships stems from its strong tidal currents of 2 - 3 m s^{-1} ascribed to the two monsters Scylla and Charybdis in Homer's masterful account of the adventures of the ancient Greek navigator Odysseus.

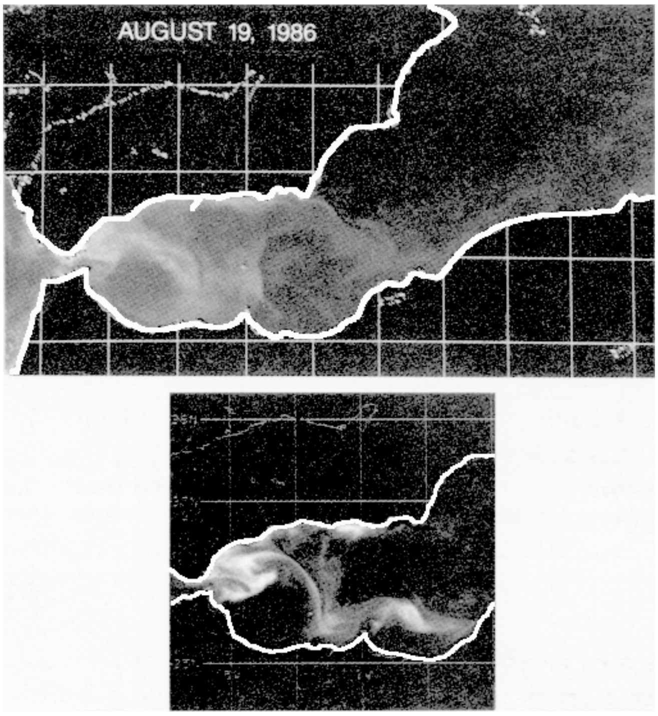

Fig. 16.4. Satellite thermal image of the Alboran Sea showing the inflow of cool Atlantic Water (white is cold, dark is warm). (a) With two eddies and the Almeria - Oran Front at 1° - 2°W, (b) with a single eddy and the front at 4°W. The two-eddy situation is far more frequent. The transition period from one to two eddies is about one month. From Tintore *et al.* (1988)

Note that the total water exchange through Gibraltar Strait is about thirty times what is required to replace the water lost by evaporation. This is because flow between mediterranean seas and the open ocean is mainly driven by the density difference between the water masses on either side of the strait (which is, of course, the result of the freshwater balance in the mediterranean sea). Inflow velocities in excess of 1 m s^{-1} in combination with a rapidly shoaling bottom in a constricted passage result in a situation where normal ocean dynamics give way to hydraulic control of the flow; in other words, in the vicinity of the strait the flow axis and the depth of the interface between the layers are not controlled by geostrophy and deflection by the Coriolis force but by the same processes which govern the flow of water over a weir. The inflowing Atlantic water initially continues eastward as a free jet and breaks into one or two large eddies of 150 km diameter before the Coriolis force can deflect it to continue along the African coast (Figure 16.4). The changeover from the Spanish to the Algerian coast occurs in a narrow current associated with a front, known as the Almeria - Oran Front, which separates the relatively fresh Atlantic water from the salty Mediterranean water (Figure 16.5). The Atlantic inflow then continues as the Algerian Current (Figure 16.6), which for at least 300 km maintains the character of a narrow jet of less than 30 km width with average velocities of 0.4 m s^{-1}, maximum velocities of

20 - 30 m depth. Being formed outside the Mediterranean Sea and identifiable throughout the basin, it can rightly be given the status of a water mass and is usually referred to as *Atlantic Water*.

The character of the Mediterranean Sea as a concentration basin requires that Atlantic Water is converted into denser water that eventually leaves the sea over the sill of the Strait of Gibraltar. This conversion process involves deep vertical convection during winter. It does not act uniformly in the entire mediterranean basin but occurs in three small regions which more than the remainder of the basin are affected by the cold northeasterly winter winds. Very cold air from Siberia is channeled through the valleys of the Alps and descends in bursts of strong winds known as mistral on the Ligurian Sea and the northern Balearic Basin; this is the region of origin for *Mediterranean Deep Water* (MDW), often also called Western Mediterranean Deep Water. Similarly cold winter winds descend on the region between Rhodes and Cyprus and on the northern and central Adriatic Sea and are responsible for the formation of *Levantine Intermediate Water* (LIW). Both are characterized by high salinities, but the details of the formation mechanisms for the two water masses of the lower layer are very different and require some discussion.

The region where MDW is formed is generally under the influence of a cyclonic (anti-clockwise) wind system. This results in a cyclonic ocean circulation with Ekman transports directed outwards and upwelling in the centre. As a consequence the LIW, which is found at about 400 - 500 m depth to the south of the region, here rises to 150 - 200 m, increasing the salinity of the upper 200 m and reducing the salinity contrast over the water column. The result is a geographically well defined region of reduced vertical stability. Winter cooling reduces the surface temperatures throughout the Mediterranean Sea and reduces the stability further; in the Ligurian Sea and the northern Balearic Basin it leaves the upper 200 m of the water column with only marginal stability. When a mistral event reaches the region, rapid additional cooling produces instability and vigorous sinking of the surface water. The sinking occurs in funnels not larger than a few tens of kilometers in diameter and is accompanied by a compensating rise of water from great depth on all sides (Figure 16.7). The water can sink some 800 m within a matter of hours and reach the 2500 m level within days (the maximum depth in the region is near 2900 m). Short and violent as these episodes of MDW formation are, the result of one such episode of a few days' duration supplies enough water to feed the lower layer outflow through the Strait of Gibraltar for several weeks. Newly formed MDW is characterized by a potential temperature of 12.6 - 12.7°C, a salinity of 38.4, and an oxygen content of 4.6 ml/l, much warmer, saltier, and better oxygenated than the North Atlantic Deep Water found at the same depth in the Atlantic Ocean. The residence time of MDW in the Ligurian Sea has been estimated at 11 ± 2 years.

Similar outbreaks of strong cold winter winds occur in the Adriatic Sea, where they are known as bora. However, the stability of the water column is also affected by river runoff which keeps the surface density low. Currents through the Strait of Otranto are therefore as expected for a dilution basin: outflow of low salinity water at the surface and inflow of high salinity Levantine Intermediate Water at its usual depth of 200 - 500 m. The modification to this simple scheme, in the form of an additional outflow layer below the inflow, comes as a result of the shallowness of the Adriatic Sea north of 42.5°N (less than 200 m decreasing to less than 100 m north of 43.5°N). This allows the water in the northern and central regions to cool very fast during a bora event and to attain a density higher than the

the thermocline of several hundred meters on the African side, and MDW is able to leave the Mediterranean Sea (Figure 16.10) at an estimated rate of 0.2 Sv. Once over the sill, MDW and LIW are not much longer recognizable as separate water masses; they sink and spread as *(Eurafrican) Mediterranean Water* as described in Chapter 15.

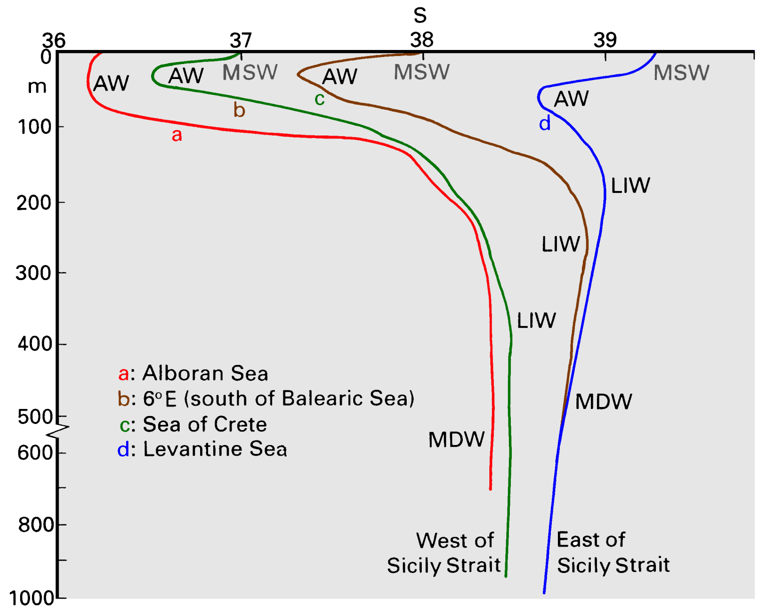

Fig. 16.8. Salinity against depth in the eastern and western Mediterranean Sea. MSW: Mediterranean surface water, AW: Atlantic Water, LIW: Levantine Intermediate Water, MDW: Mediterranean Deep Water. Note the difference in the salinity of MDW east and west of Sicily Strait; evaporation is higher in the eastern Mediterranean Sea, and the sill between Sicily and Tunisia prevents horizontal mixing between the basins.

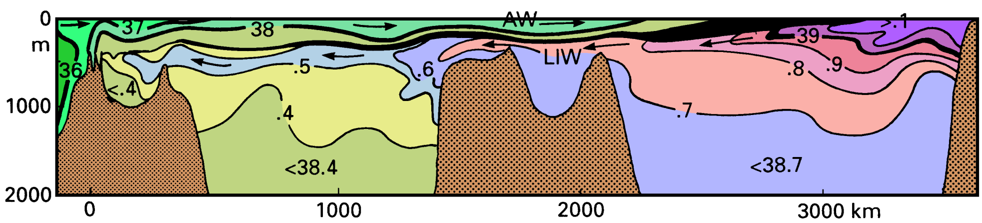

Fig. 16.9. Salinity section along the axis of the Eurafrican Mediterranean Sea. The black region indicates the salinity range 38.4 - 38.5 in the west and 38.4 - 39.1 in the east. Arrows indicate water mass movement; AW: Atlantic Water, LIW: Levantine Intermediate Water. After Wüst (1961).

537,000 km^3), the Black Sea is connected with the world ocean by a narrow passage with three subregions. Bosphorus Strait is on average 60 m deep, 31 km long and at its narrowest point only 760 m wide and has a sill depth near 35 m. In the Marmara Sea the connection broadens to some 75 km width with depths in excess of 1000 m, but further passage to the Eurafrican Mediterranean Sea is again constricted by the Dardanelles, a more than 100 km long narrow waterway between Europe and Asia. Depths in the Black Sea exceed 2000 m throughout (Figure 16.3). The shallow Sea of Azov connects to the Black Sea through Kerch Strait (sill depth 5 m) in the north; it makes up 9% of the area but only 0.5% of the volume of the Black Sea.

Large freshwater input from the Danube, Dniester, Dnieper, Severskiy Donets, and Don rivers produces a positive freshwater balance and gives the Black Sea its character as a dilution basin. Luigi Marsigli argued in 1681 already and verified with laboratory experiments that underneath the well known surface flow that carries low-salinity water from the Black Sea through the Bosphorus, Marmara Sea, and Dardanelles, there should be a flow of salty Mediterranean water in the opposite direction, produced by the salinity (and thus density) difference between the Mediterranean and Black Seas. Modern observations confirm his ideas and give the following freshwater budget, which is believed to have existed since Bosphorus Strait opened 9000 years ago:

input (km^3/year)		output (km^3/year)	
from the Mediterranean Sea	120	to the Mediterranean Sea	260
precipitation	140	evaporation	350
river runoff	350		
total	610	total	610

If this is compared to the total volume of the Black Sea it becomes obvious that water renewal in the basin is extremely slow. This is also evident in the distribution of hydrographic properties (Figure 16.11) which show that the inflow is insufficient to keep the salinity at normal oceanic values and all oxygen is depleted below 200 m depth, resulting in the formation of hydrogen sulfide which makes the Black Sea uninhabitable below 200 m.

The apparent uniformity of T-S properties below 200 m gave rise to the idea that the inflowing Mediterranean Water sinks to the bottom and that water renewal at depth is achieved by very slow upwelling. CTD data obtained over the last 30 years (Tolmazin, 1985b) indicate that the structure is not as uniform as previously believed. They show a ribbon, about 7 - 8 m thick, of Mediterranean water above the shelf floor defying the Coriolis force by bending northward for the first 50 km or so after entering from Bosphorus Strait, apparently under hydraulic control from the shallow sills. The velocities in the inflowing water are quite modest, usually less than 0.1 m s^{-1}; however, given the weak density difference between inflow and outflow, they are sufficient to generate hydraulic control. There is evidence to suggest that the inflow is not continuous but occasionally blocked by weather systems which depress the interface in Bosphorus Strait below the sill depth. Eventually the inflow merges with the general anti-clockwise circulation of the western Black Sea, and the Mediterranean water progresses along the Turkish shelf. All along its path it is exposed to intense mixing through bottom-induced turbulence, which causes parcels of water to separate and float away into the interior. As a result, lenses of warm saline oxygenated water are found at various depths in an environment usually devoid

to 33 and lower. Even when distributed over the upper 150 - 200 m, the associated density decrease is sufficient to de-couple the deep circulation from that at the surface. The water in the deep basins therefore enters from the southern Sargasso Sea, the Venezuela and Columbia Basins being filled from the Jungfern Passage and the Cayman and Yucatan Basins from the Windward Passage (Figure 16.13c). A section following the path through one of the passages (Figure 16.14) shows the remarkable uniformity of potential temperatures below the sill depths and indicates how the water is drawn from a narrow layer at the depth of the sills. Direct observations show the inflow as being confined to a layer less than 200 m thick, as well as being highly variable, often modulated by strong tidal currents, and sometimes suppressed for extensive periods. The large scale distribution of potential temperature averages over many inflow episodes and gives a good indication of the water renewal. The variability of the inflow makes estimation of deep water residence times a difficult task. Numbers found in the literature range from 55 to 800 years.

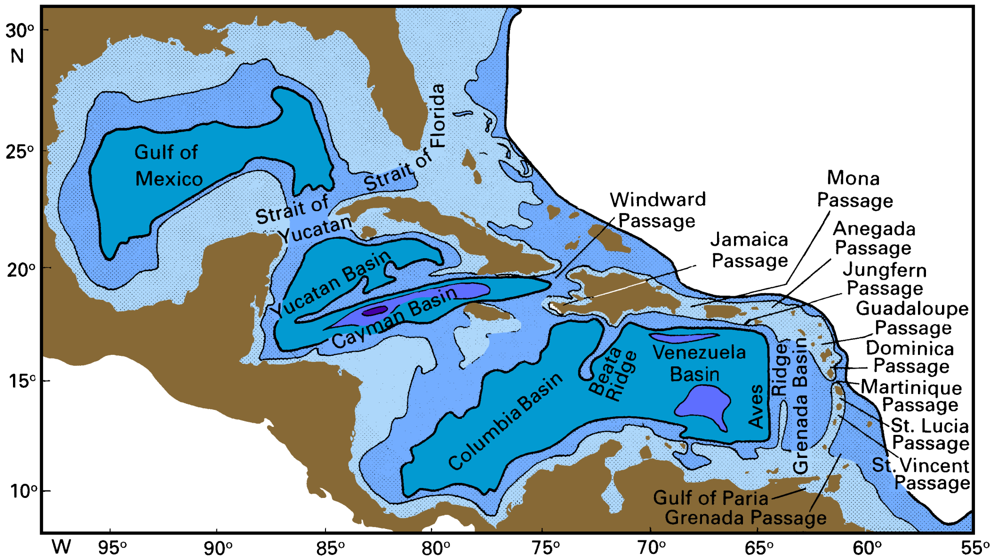

Fig. 16.12. Topography, subdivisions, and major passages of the American Mediterranean Sea. The 1000, 3000, and 5000 m isobaths are shown, and regions less than 3000 m deep are shaded.

Movement above the sill depths of the passages is dominated by throughflow from the Antilles to Yucatan Strait and into the Gulf of Mexico. The details of the circulation are determined by the topography and the location of the source. North Atlantic Deep Water is advected from the north and consequently enters mainly through Windward Passage, with additional inflow through Jungfern Passage (Figure 16.13b). Antarctic Intermediate Water is advected from the southern hemisphere and therefore enters the Caribbean Sea nearly exclusively through the eastern passages (Figure 16.13a). Central Water is advected with the North Equatorial Current from the east and finds its way into the Caribbean Sea through both the eastern and northern passages (Figure 16.15).

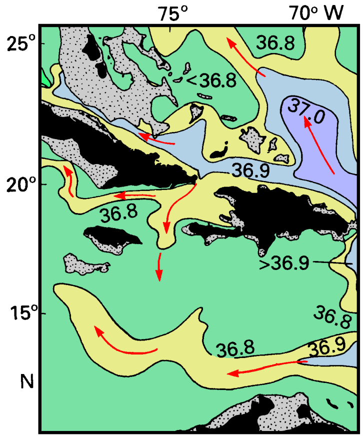

Fig. 16.15. Salinity at the salinity maxi-mum at the top of the Central Water (approximately 150 - 200 m depth). Depths less than 200 m are shaded. After Wüst (1964).

The throughflow through the American Mediterranean Sea is part of the system of western boundary currents of the north Atlantic Ocean and therefore associated with large transports. The total transport through the Caribbean Sea (close to 30 Sv) is well known from detailed measurements in the Strait of Florida, through which all water from Yucatan Strait must leave. How much of this flow enters through the passages of the Lesser Antilles and how much through Windward Passage is less well established. Early geostrophic calculations put the transport through the eastern passages at 26 Sv, leaving 4 Sv for Windward Passage. More recent estimates based on a combination of direct current measurements and circulation models reduce the role of the eastern passages significantly, allocating 15 Sv to the Grenada, St. Vincent, and St. Lucia Passages and 5 Sv to the remaining passages in the Lesser Antilles, leaving 10 Sv for Windward Passage (Kinder *et al*., 1985).

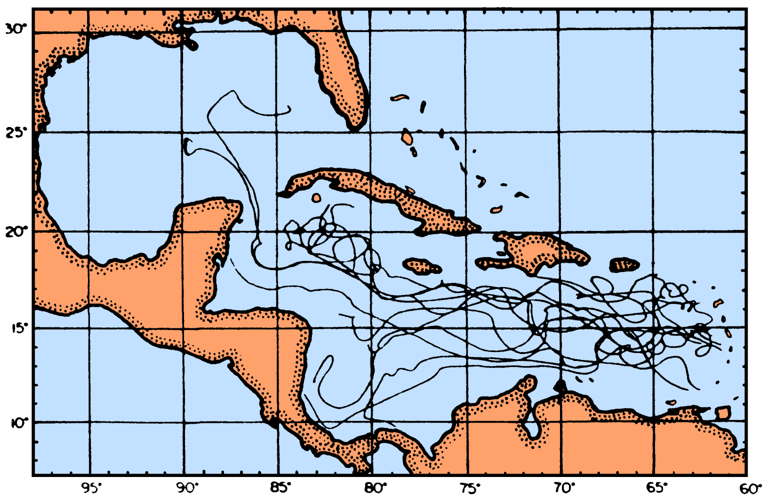

Fig. 16.16. Tracks of 19 satellite-tracked buoys for the period October 1975 - June 1976. From Kinder *et al*. (1985).

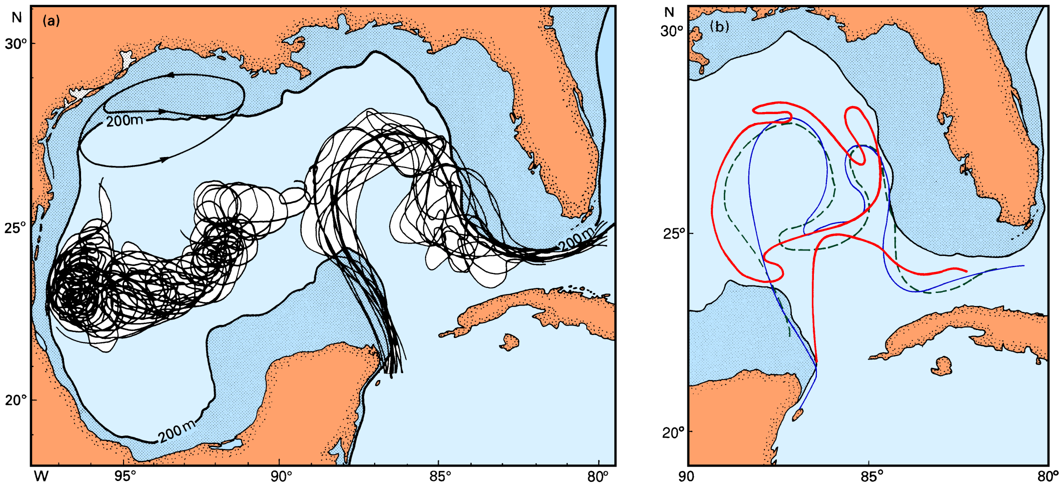

Fig. 16.18. Circulation in the Gulf of Mexico. (a) Mean position of the Loop Current during 1980 - 1984 (heavy line) and positions inferred from satellite observations of sea surface temperature, tracks of satellite tracked drifting buoys indicating eddy movement in the west, and schematic circulation on the northwestern shelf; (b) observed positions of the Loop Current just before (thin and dotted lines) and after eddy shedding (heavy lines). Areas shallower than 200 m are shaded.

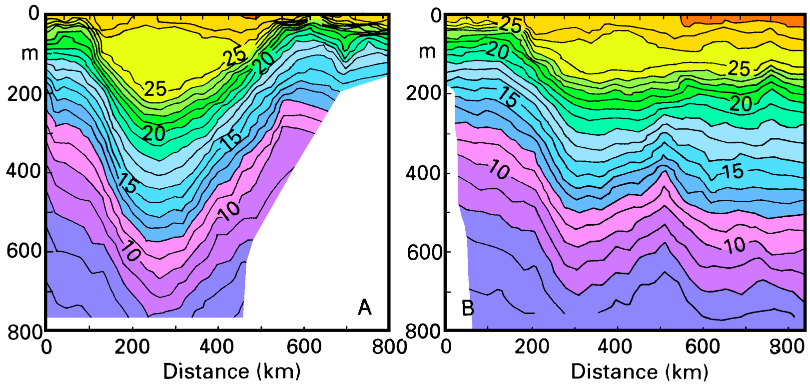

Fig. 16.19. Temperature along sections *A* and *B* shown in Fig. 16.20. Yucatan Strait is on the right. From Lewis and Kirwan (1987).

Occasionally the northern part of the Loop Current separates into a ring and the main current does not reach beyond 25°N for some time (Figure 16.18b). Eddy separation occurs on average every 11 months but can vary between 6 and 17 months (Vukovich, 1988). The eddies or rings are of the anticyclonic, warm core type described in detail in the discussion of the East Australian Current in Chapter 8. Figure 16.19 shows the isothermal core and

The westward passage of the eddies is accompanied by large variations in sea level (as explained with Figures 2.7 and 3.2 - 3.4), and a map of the long-term mean variability of sea level is a useful indicator of eddy movement. Figure 16.20 indicates mean sea level variability of 0.3 m in the eddy formation region and variability levels near 0.2 m along the major eddy drift path. This amounts to about 60% of the variability level in the centre of the Gulf Stream eddy region and indicates the high level of kinetic energy that is dispersed in the western Gulf of Mexico.

The Irish Sea, the North Sea, and the Baltic Sea

The three seas in the last geographical subdivision are all part of the European continental shelf and therefore mostly shallow, the only notable exception being the Norwegian trench in the North Sea. The Irish and North Seas both have long open connections with the Atlantic Ocean proper and are dominated by strong tidal currents and frequent strong winds; they are therefore similar in character. In contrast, communication between the Baltic Sea and the Atlantic Ocean is only indirect (through the North Sea) and severely restricted, which makes the Baltic Sea the fourth mediterranean sea of the Atlantic Ocean. Our discussion of the basics of ocean dynamics in Chapters 1 - 5 did not include the modifications that occur in coastal, shelf, and estuarine areas. With these limitations in our understanding of the processes responsible for water movement and renewal in shallow seas we have to restrict our description of the circulation and hydrology on the European shelf to a few general remarks.

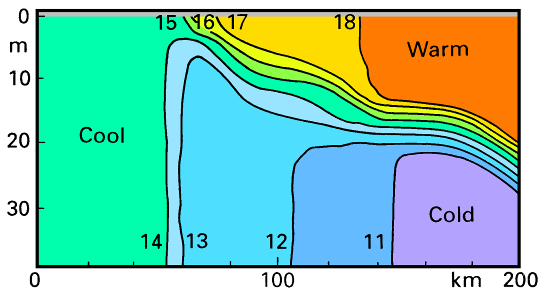

Fig. 16.21. Sketch of a temperature section through a shallow sea front, based on observations from the eastern entry to the English Channel. Note that the temperature on the shallow side (on the left) equals the temperature in the core of the thermocline.

An important element of the dynamics of shallow seas with a good connection to the deep ocean is tidal movement. As the tide enters from the ocean the tidal current increases in magnitude as the water depth decreases. The increase in current speed is not restricted to the surface layer but occurs at all depths. The associated turbulence acts like a giant stirring mechanism trying to break down the stratification. There is therefore a competition between solar heat input at the surface, which acts to stabilize the water column, and mixing from the tides, which attempts to homogenize the water column. In deep water tidal currents are weak and the water is stratified. In shallow water the currents are strong and the water is well-mixed. The transition between the mixed region and the stratified region occurs in a frontal zone known as a shallow sea front. Figure 16.21 gives a sketch of such a front.

upward entrainment of salty water. This is associated with inflow towards the river mouths in the lower layer and outflow near the surface. The situation is typical for estuarine circulation systems, and the reader is referred to textbooks on estuaries and coastal regions for further detail.

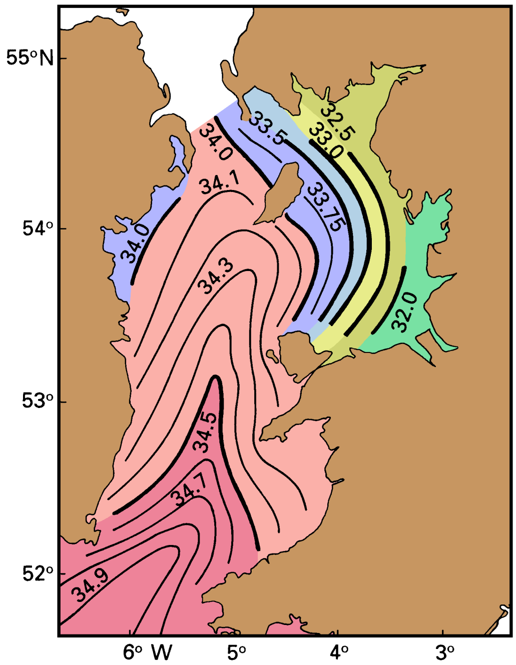

Fig. 16.23. Mean surface salinity in the Irish Sea.

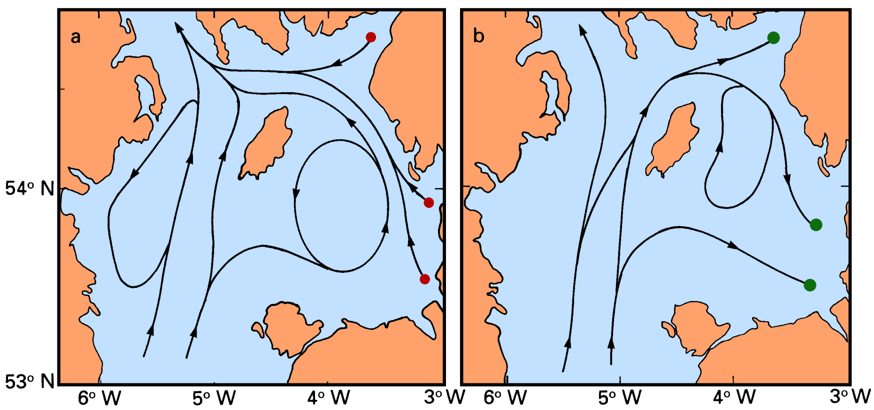

Fig. 16.24. Mean circulation in the northern Irish Sea. (a) Near-surface, (b) near-bottom. Dots indicate entrainment from the near-bottom into the near-surface layer.

The mean circulation can be seen in the salinity distribution which changes little over the year. Figure 16.26 compares the salinity near the sea surface with the distribution near the bottom. It is seen that the salinity difference between the two surfaces is small over most of the North Sea; but significant differences are observed in the north and in particular the northeast. They are related to the water movement and indicate the presence of a two-layer circulation system in the Skagerrak and the Norwegian trench. Low-salinity water from the Baltic Sea enters the North Sea at the surface and joins the northern branch of the anti-clockwise circulation which derives most of its water from the central North Sea. The inflow is intermittent and controlled by the local wind. High salinity water moves eastward from the open Atlantic Ocean along the southern slope of the Norwegian trench and returns along its northern side; some of it fills the depression in the east. Mean currents in the central and southern North Sea do not vary much with depth.

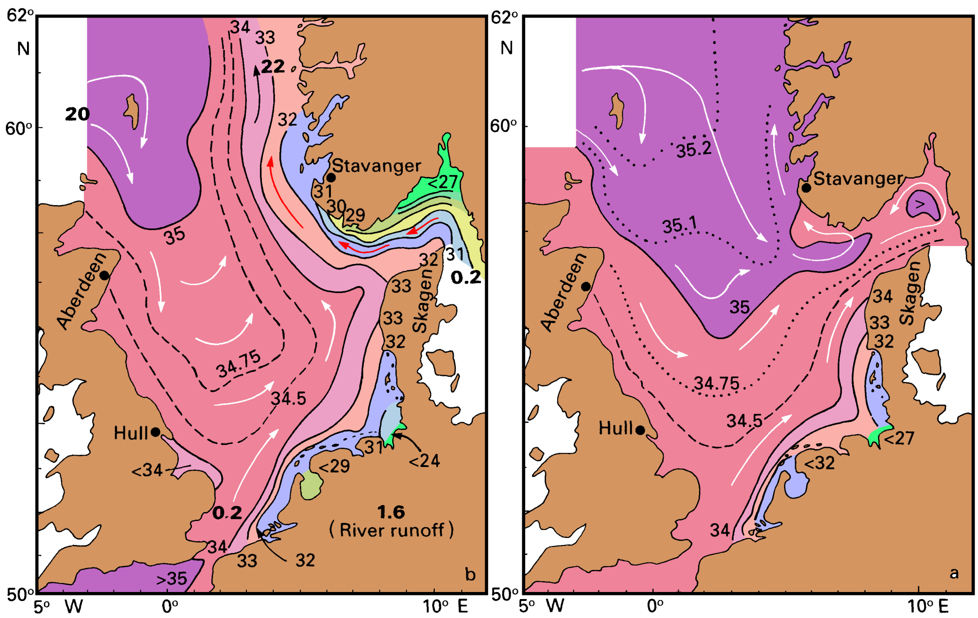

Fig. 16.26. Mean salinity and circulation in the North Sea for June. (a) At 7.5 m depth, (b) above the bottom. Bold numbers give average transport in 1000 km^3 per year (1000 km^3 per year = 0.03 Sv). Adapted from Goedecke *et al.* (1967).

The temperature distribution shows more seasonal variation and regional structure. Sea surface temperatures in the central North Sea vary between 2 - 4°C in winter and 18 - 20°C in summer, the total range increasing monotonically from northwest to southeast. Close to the Dutch and German coast the range increases to -1 - 22°C, bringing occasional ice formation. The vertical structure of the temperature field (Figure 16.27) shows the imprint of shallow sea front dynamics (discussed above with the Irish Sea), a clear indication for strong tidal currents - tides dominate the flow field and sea level in the North Sea at any particular time -, although these fronts have not received much attention in the North Sea.

exchange of water between the Baltic and North Seas carries transports similar to those between the Black and Mediterranean Seas, the amount of basin water requiring renewal through inflow from the open ocean is much less, and de-oxygenation in the deep basins is not permanent. The average salinity (Figure 16.29), on the other hand, is even lower than in the Black Sea, since the river runoff constitutes 2% of the Baltic Sea volume.

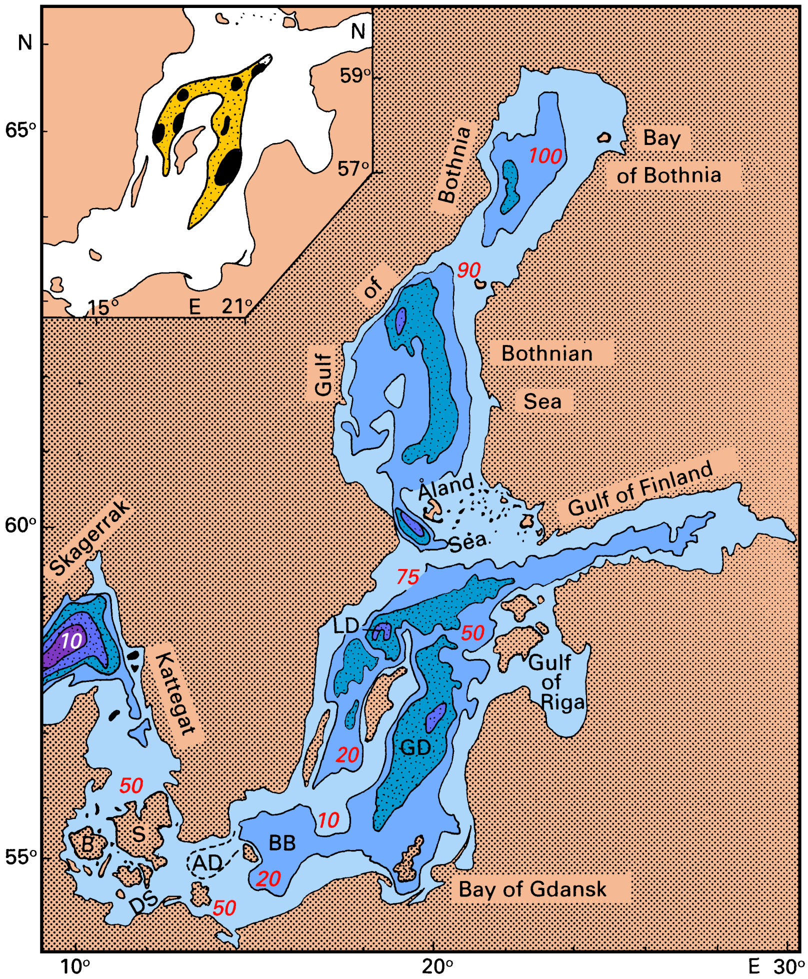

Fig. 16.28. Topography of the Baltic Sea. Depths > 100 m are stippled, with contours drawn for 50, 100, 300, and 500 m depth. AD: Arkona Deep, BB: Bornholm Basin, BS: Belt Sea, DS: Darß Sill, GD: Gotland Deep, LD: Landsort Deep. Numbers in italics give the probability of total ice coverage (*100%* = total ice cover every year). The inset shows the minimum (black) and maxi–mum (stippled) extent of anoxic conditions in the central basins for the period 1979 - 1988; adapted from Nehring (1990).

strongly to small changes induced by human activity. Present plans to connect Denmark and Sweden by a system of bridges or tunnels can have a major impact on deep water renewal if the sill depths or cross-sections of the Belt Sea passages are changed.

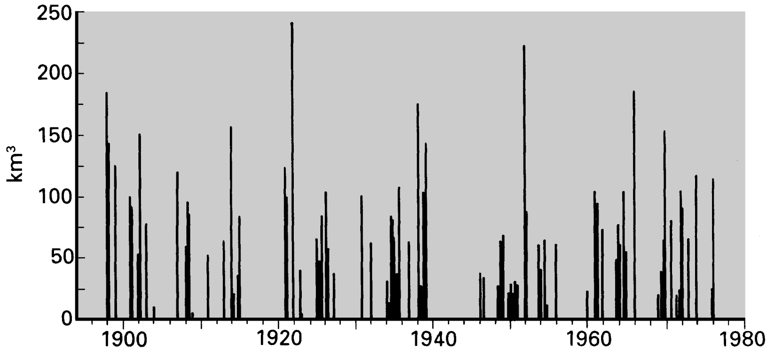

Fig. 16.30. Volumes of high salinity inflow into the Baltic Sea. The gaps during 1914 - 1918 and 1939 - 1945 are from lack of observations. From Matthäus and Franck (1990).

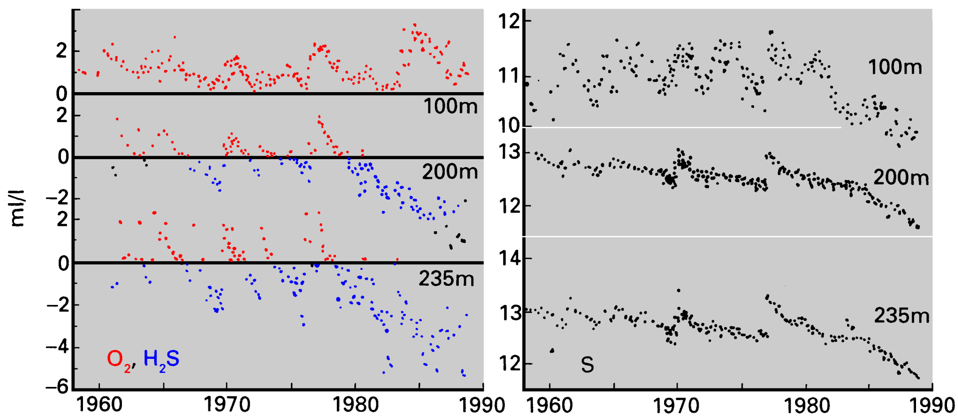

Fig. 16.31. Oxygen O_2 (ml/l), hydrogen sulfide H_2S (ml/l, plotted as negative oxygen), and salinity S at a station in the Gotland Deep. Strong inflow events occurred in 1965, 1970, 1973, and 1976. Note that after each inflow event salinity and oxygen are higher at 235 m than at 200 m, indicating that renewal takes place near the bottom and progresses upward by turbulent diffusion. After Nehring and Matthäus (1990).

*A major inflow event, comparable in volume to earlier major events, occurred during the winter of 1992/1993. A full assessment of its impact on water renewal in the Baltic deep basins was not possible at the time of printing.