Chapter 16

Adjacent seas of the Atlantic Ocean

The hydrography of the Atlantic Ocean is strongly influenced by its adjacent seas, to the degree that a discussion of this ocean would be incomplete without a description of the hydrographic conditions of its adjacent seas. This is particularly true for the Arctic Mediterranean Sea which plays a crucial role in the formation of Deep Water not only for the Atlantic but for the entire world ocean. Its importance for the global oceanic circulation justified treatment of this adjacent sea in a separate chapter (see Chapter 7).

The remaining adjacent seas can be divided on geographical arguments into four groups. The first group contains the waters connected to the Atlantic Ocean proper through the Labrador Sea and consists of the Davis Strait, Baffin Bay, the Northwest Passage, and Hudson Bay. The second group is located between Europe, Africa, and Asia and contains the Eurafrican Mediterranean Sea (which includes the Black Sea). The third group is found near the junction of North and South America and contains the American Mediterranean Sea with its subdivisions the Caribbean Sea and the Gulf of Mexico. The shallow European seas make up the fourth and last group, which contains the Irish and North Seas and the Baltic Sea with its approaches.

Davis Strait, Baffin Bay, the Northwest Passage, and Hudson Bay

The passages between the Arctic Mediterranean and the Labrador Seas, variably known as the Northwest Passage, the North Water, or the Canadian Arctic Archipelago, consist of a maze of islands, channels, straits, and basins of widely different character. Baffin Bay is a deep basin with maximum depths in excess of 2300 m. It is separated from the Labrador Sea by Davis Strait, which has a sill depth of less than 600 m. The two major connections to the Arctic basins are through Nares Strait, the Kane Basin and Smith Sound to the north and through M'Clure Strait, Viscount Melville Sound, Barrow Strait and Lancaster Sound to the west (Figure 16.1). A third passage bifurcates from Lancaster Sound towards south and connects with the Labrador Sea through Prince Regent Inlet, the Gulf of Boothia, the Foxe Basin and Foxe Channel, and Hudson Strait. Sill depths in all three passages are quite shallow; in the northern passage the sill depth is just over 200 m in Smith Sound, while in the western passages it is less than 150 m in Barrow Strait and less than 100 m in the Foxe Basin. The various deep passages between the Queen Elizabeth Islands are blocked to the south by similarly shallow water and do not play a significant role in the water exchange between the Arctic and Labrador Seas.

The circulation in the region is determined by the West Greenland Current and the throughflow from the Arctic Mediterranean Sea. Both influences combine to drive the Baffin Current, a southward flow of about 0.2 - 0.4 m s^{-1} along the western side of Baffin Bay which supplies water to the Labrador Current (Figure 16.1). Flow through the passages is generally southward and westward, but eastward countercurrents are found on the northern sides of Lancaster Sound and Hudson Strait. The countercurrents appear to be more variable in time than the eastward flow along the southern coasts but of comparable strength (0.3 - 0.5 m s^{-1}).

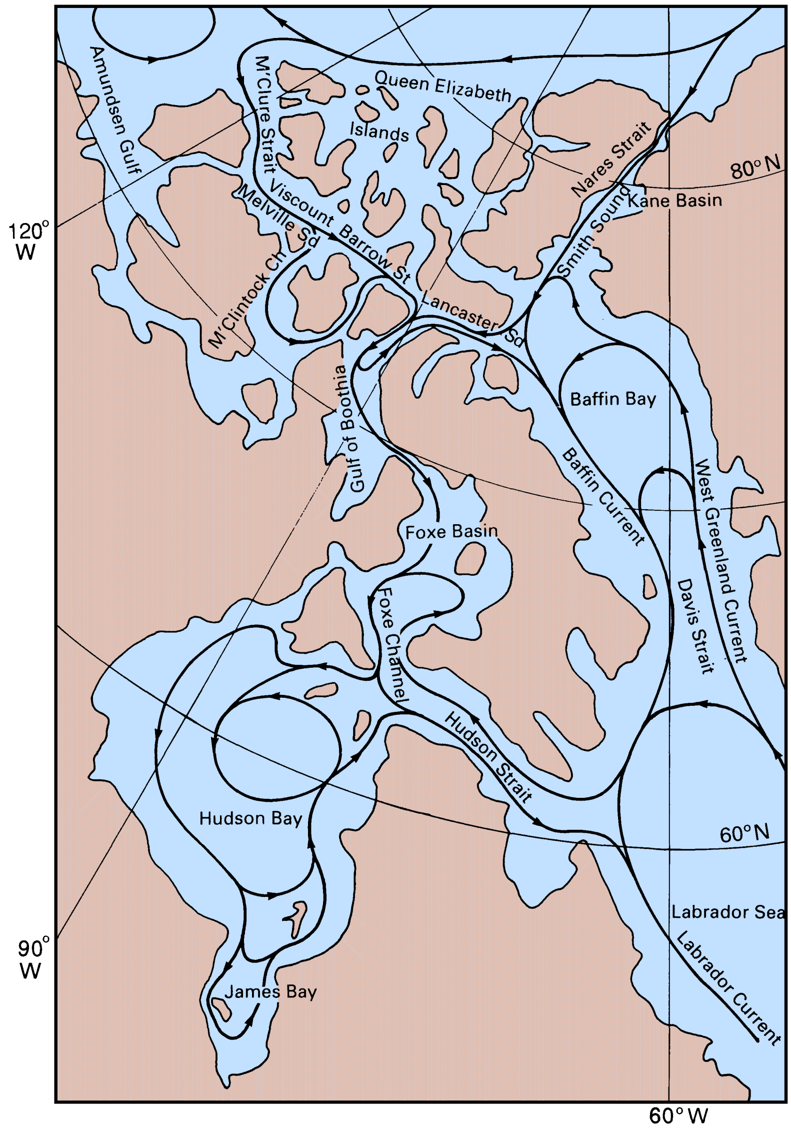


Fig. 16.1. Geography and circulation of the Northwest Passage and connected seas.

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

Greenland Current. This water is sometimes called Atlantic Intermediate Water on account of its position in the water column and the fact that it is brought into the region from the south. However, its origin is in the East Greenland Current and thus ultimately in the Arctic Mediterranean Sea, and there is no relation with what is normally called Intermediate Water (water subducted near the Polar Front and characterized by a salinity minimum). Another, more appropriate name for this water is Polar Atlantic Water.

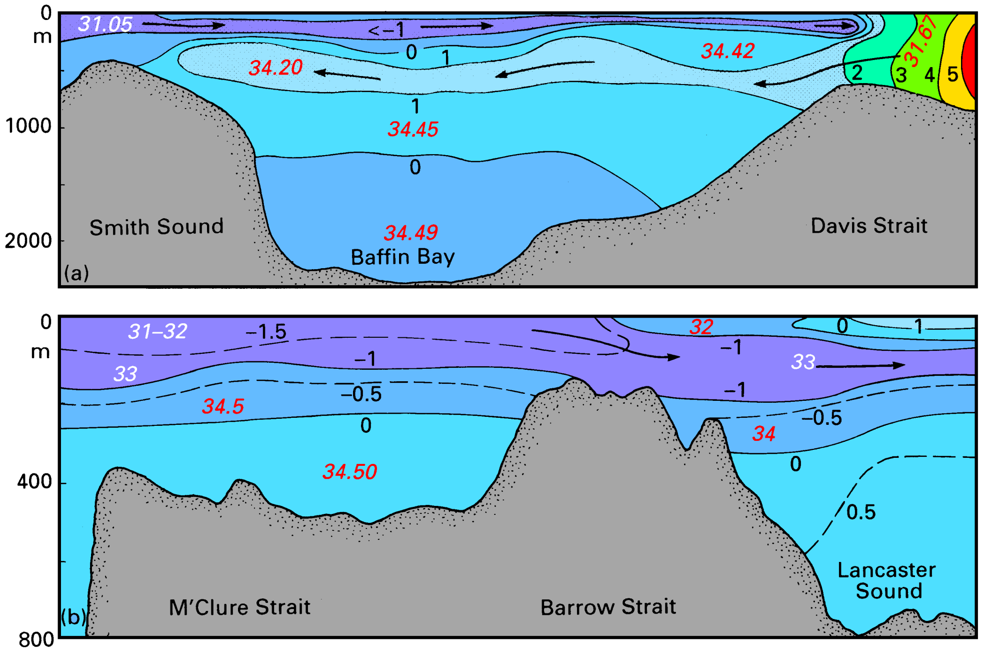


Fig. 16.2. Temperature (°C) in two sections through the Northwest Passage and Baffin Bay. (a) From Smith Sound to Davis Strait, (b) from M'Clure Strait to Lancaster Sound. Arrows indicate water movement. Typical salinities are indicated in *italics*. Adapted from Collin (1966).

The low temperature water above the Polar Atlantic Water is also of Arctic origin but advected from the north. It is drawn from the sub-surface layer of Arctic Surface Water (see Chapter 7) and modified by some injection of brine from the surface during winter freezing. Its salinity is therefore somewhat higher than the salinity of Arctic Surface Water at the same temperatures. This water is sometimes referred to as Arctic Intermediate Water but again not related to the Intermediate Water of the temperate zone.

Above the Arctic Intermediate Water is a thin surface layer of not more than 50 m depth where water properties change with the seasons. Summer temperatures vary between -0.1 - 5.0°C, and salinities are in the range 30.0 - 35.5.

Below the Atlantic Polar Water temperature decreases and salinity increases slowly until both become virtually constant in the Baffin Bay Bottom Water below 1800 m depth (-0.4°C and 34.49 salinity). The low oxygen content (3.6 ml/l) of this water testifies for 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.

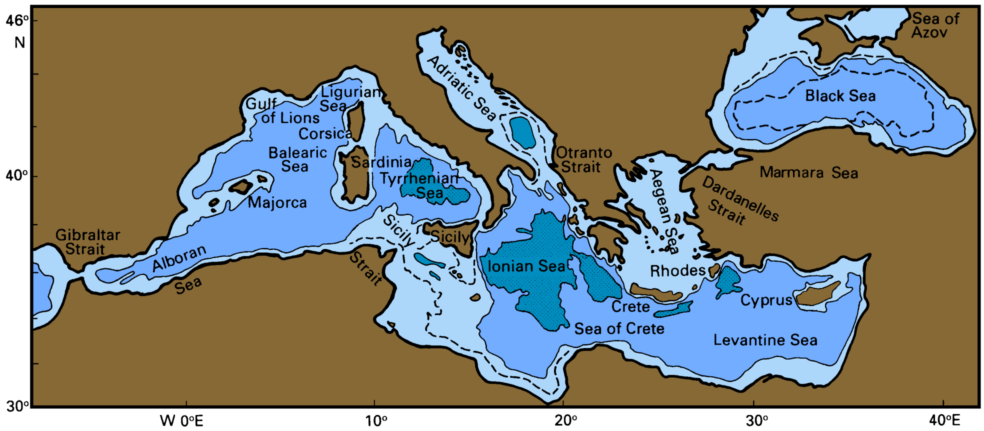


Fig. 16.3. Topography and subdivisions of the Eurafrican Mediterranean Sea. The 1000 m contour is shown, and regions deeper than 3000 m are shaded. The 200 m contour is shown as a broken line where it departs significantly from the 1000 m contour. In addition, the 2000 m contour is shown as a broken line in the Black Sea.

Communication with the Atlantic Ocean proper is through the Strait of Gibraltar which is 22 km wide and has a sill depth of 320 m. This poses a severe limitation on the water exchange and in combination with the atmospheric conditions creates distinctive hydrographic conditions. During most of the year winds over the Mediterranean Sea are from the northwest and carry warm dry air, causing large evaporation. During winter the winds are often northeasterly, bringing dry but cold air into contact with the sea.

Over most of the Mediterranean Sea annual evaporation exceeds rainfall and river runoff by about 1 m, so on average the Mediterranean Sea is a concentration basin. The two exceptions are the Black and the Adriatic Seas which receive large amounts of freshwater from the Danube and Po rivers and therefore are dilution basins. Water exchange between the Atlantic Ocean proper, the western and eastern Mediterranean Sea, and the Black Sea, which follows the principles discussed with Figure 7.1, therefore consists of inflow of Atlantic water through the Strait of Gibraltar in an upper layer, outflow of dense Mediterranean water below, inflow of relatively fresh Black Sea water through the Dardanelles in an upper layer, and outflow of Mediterranean water below (exchange with the Adriatic Sea is similar but not associated with strong currents because of the lack of topographic restriction between the Adriatic and Ionian Seas). An annual mean budget gives the following transports (where the actual numbers have been adjusted to give zero balance and do not imply accuracy to three digits):

inflow (Sv)		outflow (Sv)	
from the Atlantic	1.107	to the Atlantic	1.041
from the Black Sea	0.006		
precipitation	0.027	evaporation	0.111
river runoff	0.011		
total	1.152	total	1.152

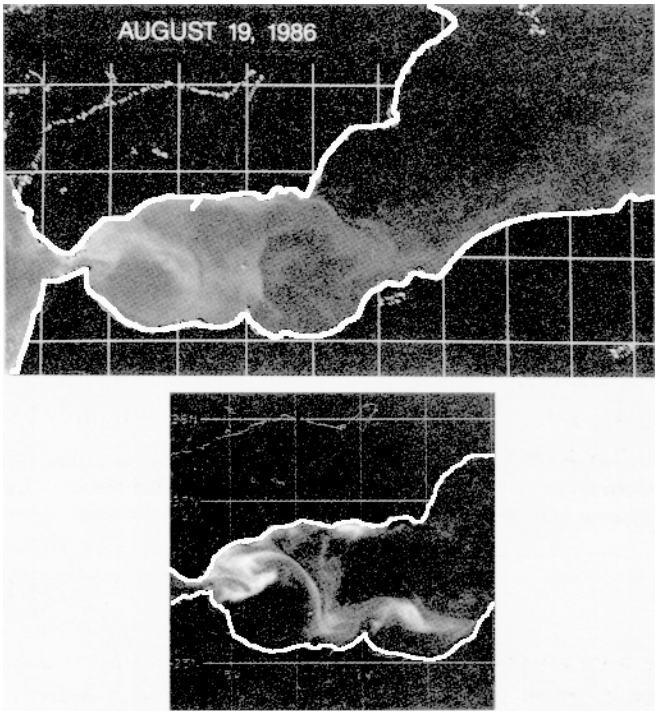


Fig. 16.4. Satellite thermal image of the Alboran Sea showing the inflow of cool Atlantic Water (white is cold, dark is warm). (a) With two eddies and the Almeria - Oran Front at 1° - 2°W, (b) with a single eddy and the front at 4°W. The two-eddy situation is far more frequent. The transition period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

0.8 m s^{-1}, and a total transport of about 0.5 Sv. Having advanced past 4°E the flow becomes more diffuse. In the eastern Mediterranean basins it is dominated by eddies, some large and stationary as indicated in Figure 16.6, others of only 40 - 60 km diameter but reaching to great depth. Observations have shown them to extend to at least 1000 m, with velocities exceeding 0.2 m s^{-1} above the 300 m level.

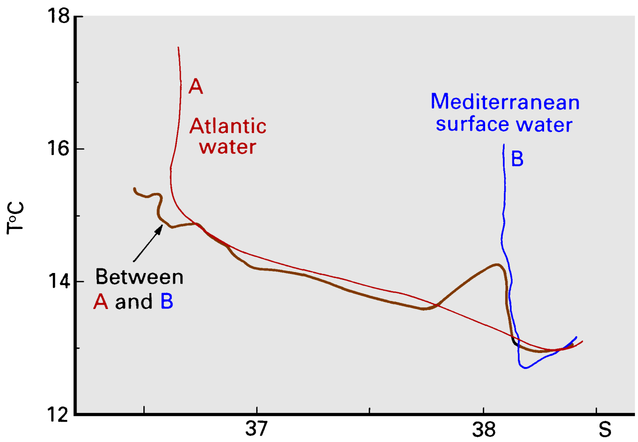


Fig. 16.5. Temperature – salinity diagrams for two stations on either side of the Almeria - Oran front (locations *A* and *B* in Fig. 16.6) and for a station in-between. Note the layering and associated temperature inversion in the front. After Arnone *et al*. (1990).

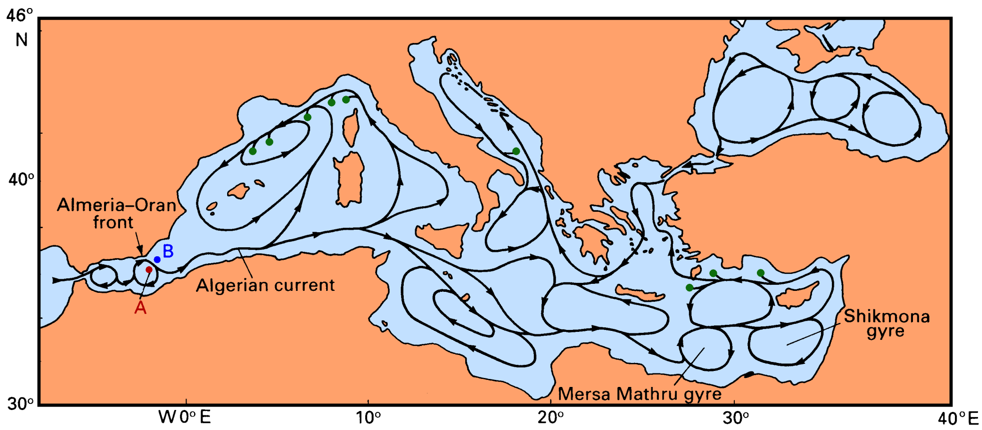


Fig. 16.6. Upper ocean currents in the Eurafrican Mediterranean Sea. Locations marked *A* and *B* in the Alboran Sea refer to the data shown in Fig. 16.5. Dots indicate areas of winter convection.

On its way east the Atlantic water encounters saltier but warmer and consequently less dense Mediterranean water. Outside the Algerian Current it therefore continues submerged under a shallow surface layer of high salinity and can be followed as a salinity minimum at

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 density of LIW in the south. The dense surface water flows southward on the Italian shelf, bypassing the deep southern basin until it encounters a series of canyons near 41.5°N. It falls down the canyons, mixing vigorously with LIW on the way, and leaves Otranto Strait as an outflow below the LIW inflow. Its characteristic properties are 13°C and 38.6 salinity, slightly less than the salinity of 38.7 found in the LIW above (which is warmer - 14°C - and thus less dense). The slight reduction in salinity indicates a contribution from the Po river.

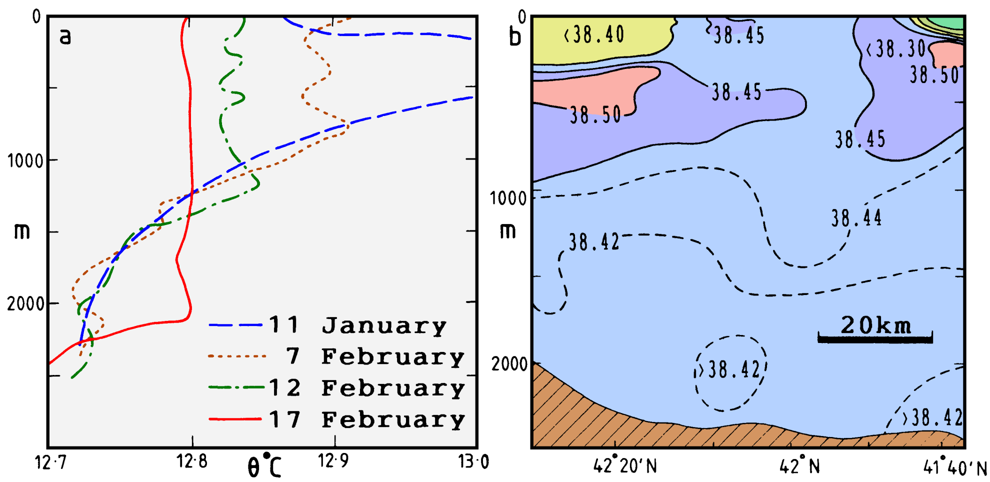


Fig. 16.7. Observations of Mediterranean Deep Water formation in the Balearic Sea. (a) Potential temperature (°C) over a one-month period in 1969 near 42°N, 5°E showing the deepening of the mixed layer to 2200 m, (b) a salinity section along 5°20'E showing a column of newly mixed water near 42°N. Adapted from Sankey (1973).

The cold but relatively fresh bottom water from the Adriatic Sea does not maintain its identity very long. It turns eastward and enters the Levantine Basin where it encounters water freshly formed in the region between Rhodes and Cyprus. The two sources mix, and together they form the Levantine Intermediate Water. The Levantine surface source is much warmer but significantly more saline (15°C and 39.1 salinity). The resulting mixture gives a potential temperature of 13.3°C, a salinity of 38.67, and a very high oxygen content of 5.0 ml/l as the characteristics of LIW.

LIW is saltier but warmer than MDW and has a slightly lower density. As a consequence, a hydrographic station in the central parts of the Mediterranean Sea shows a layering of four water masses (Figure 16.8). LIW is easily identifiable as a salinity maximum underneath the Atlantic Water minimum. A salinity section along the axis of all major basins (Figure 16.9) shows its movement towards the Strait of Gibraltar. In general, MDW is not found above 600 m depth, so the bulk of the outflow through Gibraltar Strait must be provided by LIW. However, in the vicinity of the sill hydraulic control causes an uplift of 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.

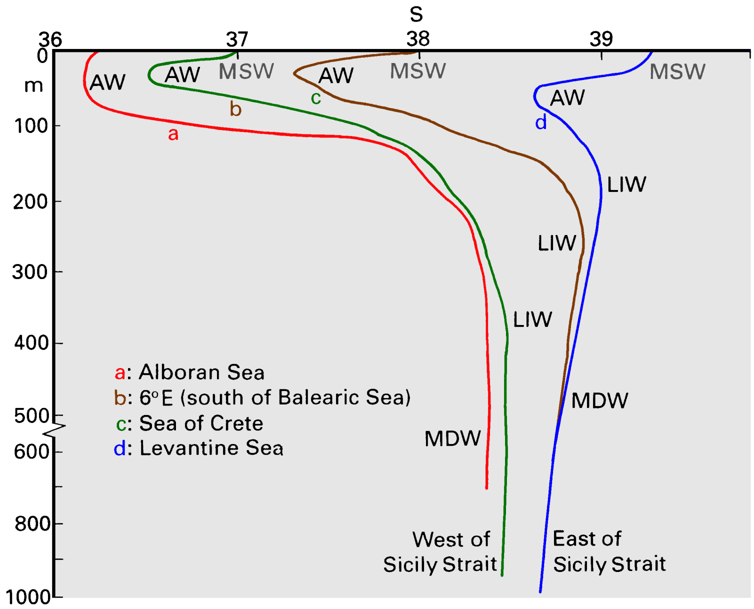


Fig. 16.8. Salinity against depth in the eastern and western Mediterranean Sea. MSW: Mediterranean surface water, AW: Atlantic Water, LIW: Levantine Intermediate Water, MDW: Mediterranean Deep Water. Note the difference in the salinity of MDW east and west of Sicily Strait; evaporation is higher in the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

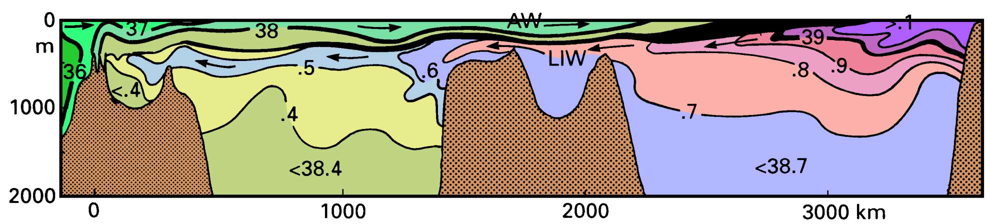
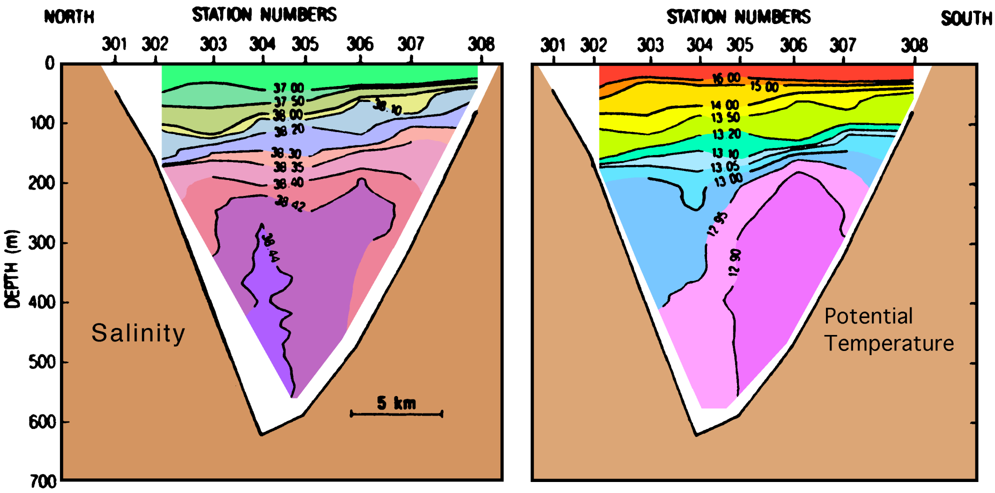
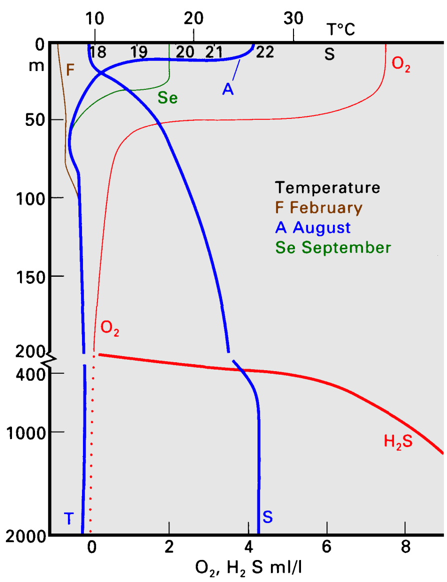


Fig. 16.9. Salinity section along the axis of the Eurafrican Mediterranean Sea. The black region indicates the salinity range 38.4 - 38.5 in the west and 38.4 - 39.1 in the east. Arrows indicate water mass movement; AW: Atlantic Water, LIW: Levantine Intermediate Water. After Wüst (1961).



16.10. Outflow of Levantine Intermediate Water (LIW) and Mediterranean Deep Water (MDW) through Gibraltar Strait as seen in salinity and potential temperature (°C). The section is some 15 km west of the narrowest point. MDW is identified by Θ < 12.90°C, LIW by a salinity > 38.44. From Kinder and Parrilla (1987).



16.11. Vertical profiles of temperature (°C), salinity, oxygen (O_2, ml/l), and hydrogen sulfide (H_2S, ml/l) typical for the Black Sea. From Tolmazin (1985a).

The Black Sea has not been mentioned throughout this discussion since its impact on the Mediterranean Sea is very small and it is more appropriate to describe it as a separate mediterranean basin. The world's largest inland water basin (area 461,000 km^2, volume

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 of oxygen. This is reflected in age estimates derived from radiocarbon measurements, which for the water in the 300 - 2160 m depth range vary from 600 to 2200 years and do not show a systematic variation with depth. Nevertheless, all water below 200 m depth must ultimately be renewed from Bosphorus Strait.

Away from the entry point for Mediterranean water the circulation in the entire water column is dominated by anti-clockwise motion along the continental slope with three anti-clockwise gyres of about identical size filling the western, central, and eastern basin (Figure 16.6). The western gyre brings water from the northwest shelf region into contact with the open Black Sea. In winter the water on the shelf is colder than that in the open sea (through a process explained for Bass Strait water in Chapter 17) and sinks as it leaves the region of shallow water depth. By spreading through the Black Sea at intermediate depth it produces the temperature minimum regularly observed near 75 m (Figure 16.11). A recent series of research cruises (Murray, 1991) will allow a much more detailed description of the circulation and water masses in the Black Sea.

The American Mediterranean Sea

The topography of the Caribbean Sea and the Gulf of Mexico (Figure 16.12) shows a succession of five basins, separated by sills of less than 2000 m depth and set apart from the main Atlantic basins by an island-studded enclosure less than 1000 m deep but containing several passages with sill depths of 740 - 2200 m. This alone identifies the region as a mediterranean sea, similar in structure to the Australasian Mediterranean Sea, a dilution basin in the tropics (see Chapter 13). The similarity extends to the fact that both seas have more than one connection with the main ocean basins and are therefore dominated in their upper layers by throughflow. The difference is that the Australasian Mediterranean Sea is located in a region of large freshwater gain and is a true dilution basin, while over the Caribbean Sea and the Gulf of Mexico evaporation exceeds precipitation by over 1 m per year, as much as in the case of the Eurafrican Mediterranean Sea and too much to be balanced by freshwater input from rivers; so the American Mediterranean Sea should really be a concentration basin. This is indeed correct: The annual mean salinity, averaged over the upper 200 meters, increases from 36.09 at the inflow through the Lesser Antilles to 36.19 in Yucatan Strait and further to 36.39 in the Strait of Florida (Etter *et al.*, 1987). However, for north Atlantic standards these salinities are quite low (Figure 2.5b); they do not reach the salinity values below the surface layer (compare Figure 16.15). The density increase associated with the concentration process is therefore insufficient to cause deep vertical convection. As a result, deep water renewal does not occur through the sinking of surface water (as observed in other concentration basins) but follows the pattern typical for dilution basins, i.e. sporadic inflow of oceanic water from outside. This makes the American Mediterranean Sea rather unique among all adjacent seas of the world ocean.

The reason for the low salinities in the surface layer is advection of Amazon River water with the Guayana and Caribbean Currents. Estimates based on radium measurements indicate that on average, 15 - 20% of the surface water that enters the Caribbean Sea is derived from the brackish waters of the Orinoco and Amazon River estuaries (Moore *et al.*, 1986). The influence of river runoff is strongly seasonal, with strongest flow occurring between May and November; surface salinities in the eastern Caribbean Sea can then drop 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.

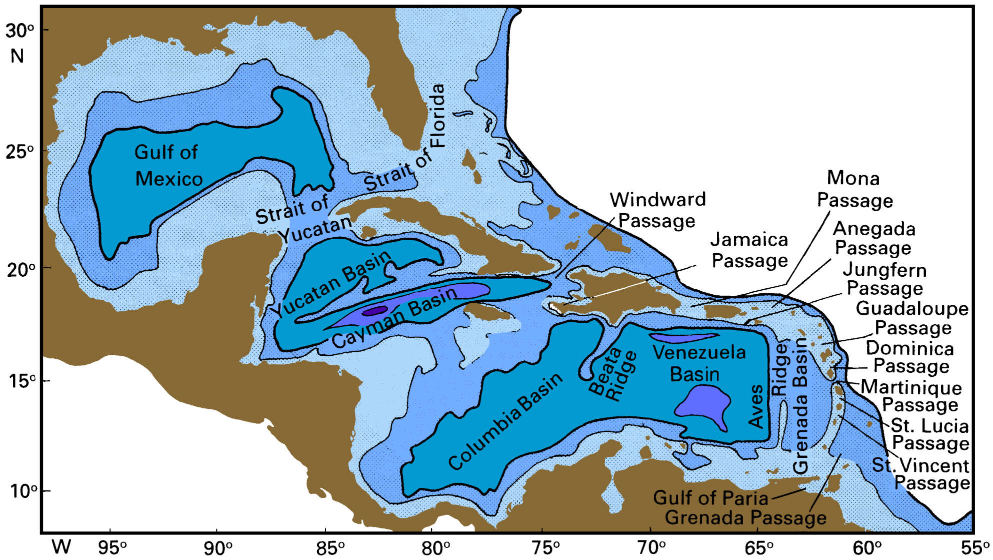


Fig. 16.12. Topography, subdivisions, and major passages of the American Mediterranean Sea. The 1000, 3000, and 5000 m isobaths are shown, and regions less than 3000 m deep are shaded.

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).

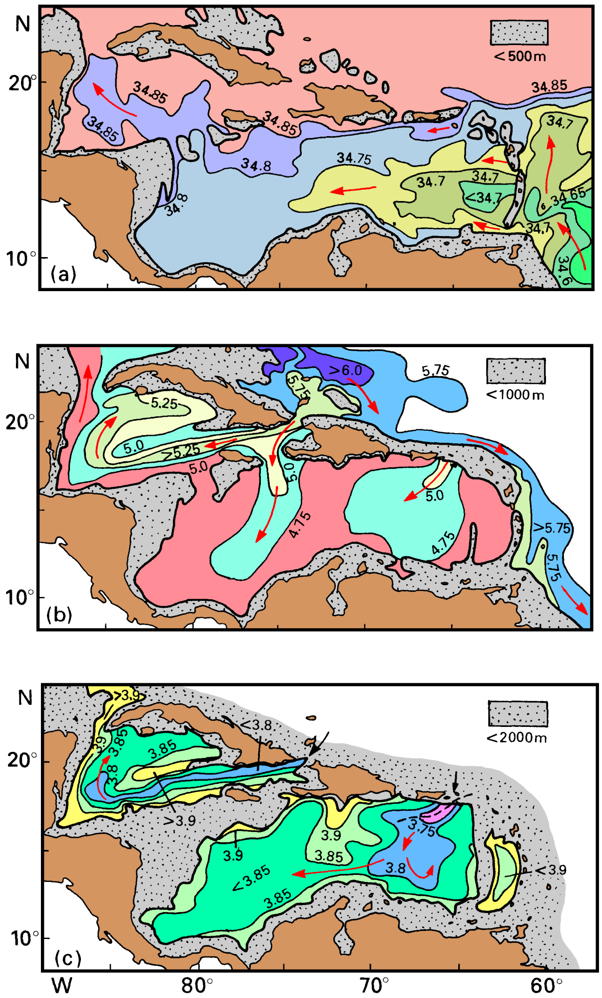


Fig. 16.13 Intermediate and deep water renewal in the basins of the Caribbean Sea. (a) Salinity at the level of the salinity minimum near 700 - 850 m, indicating the path of Antarctic Intermediate Water; (b) oxygen (ml/l) at the level of the oxygen minimum near 2000 m, indicating movement in the upper range of North Atlantic Deep Water; (c) bottom potential temperature (°C), indicating renewal paths for the water in the deep basins. The broken line indicates the location of the section of Fig. 16.14. After Wüst (1963). (The data for (b) were obtained during 1954 - 1958. Data obtained during 1932 - 1937 show oxygen levels higher by 0.3 ml/l in the east and 0.7 ml/l in the west. The indicated flow pattern is the same in both cases.)

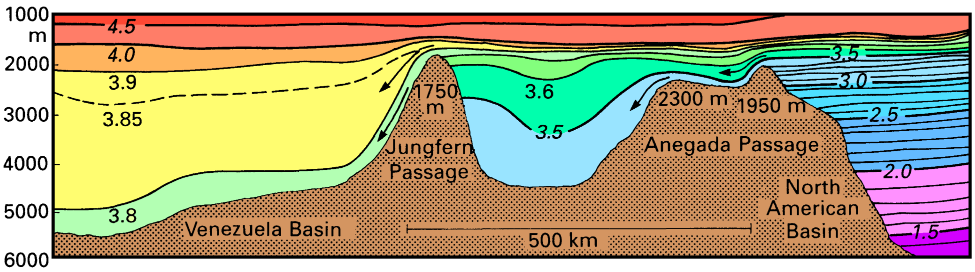


Fig. 16.14. Potential temperature (°C) along the section indicated by the dotted line in Fig. 16.13c. After Wüst (1963).

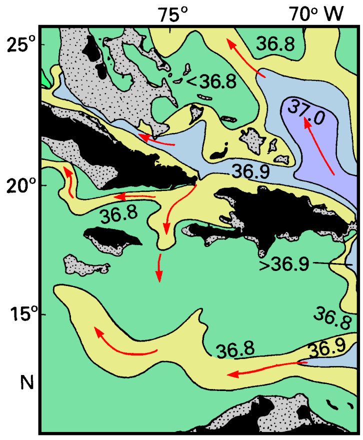


Fig. 16.15. Salinity at the salinity maximum at the top of the Central Water (approximately 150 - 200 m depth). Depths less than 200 m are shaded. After Wüst (1964).

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).

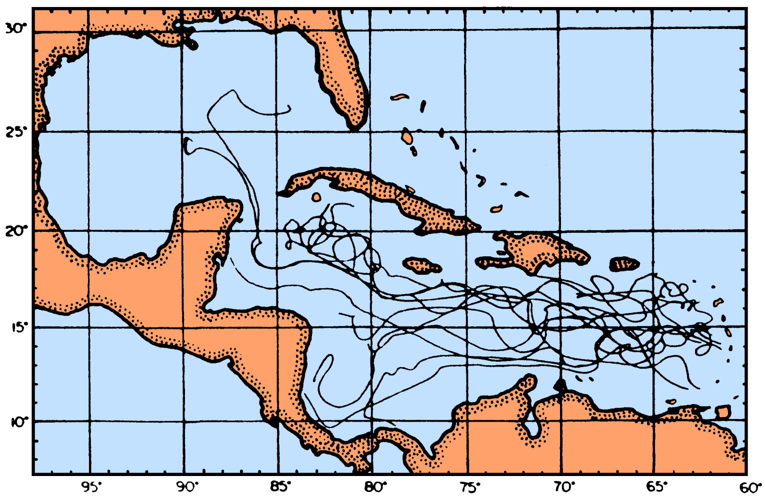


Fig. 16.16. Tracks of 19 satellite-tracked buoys for the period October 1975 - June 1976. From Kinder *et al*. (1985).

The surface flow through the Caribbean Sea has been documented by drifting buoys tracked by satellites. The tracks (Figure 16.16) show the Caribbean Current with speeds around 0.2 m s^{-1} in the Grenada Basin, 0.5 m s^{-1} in the Venezuela, Columbia, and Cayman Basins, and 0.8 m s^{-1} near Yucatan Strait. These velocities are lower than in other western boundary currents and particularly in the western Caribbean Sea lower than the velocities associated with the eddies produced by the current. This makes the currents highly variable and causes occasional flow reversals from westward to eastward in the Yucatan Basin. Systematic eastward flow embedded in generally westward movement is found in the Grenada and possibly also in the Venezuela Basin. Observations above the Aves Ridge (Figure 16.17) show a banded flow structure with eastward flow of variable strength near 13°N. This flow eventually leaves the Caribbean Sea and enters the Atlantic Ocean as the Caribbean Countercurrent (Figure 14.2). Near 16°N the current appears to be mostly eastward; but occasionally the flow turns westward with minimum speeds near 16°N.

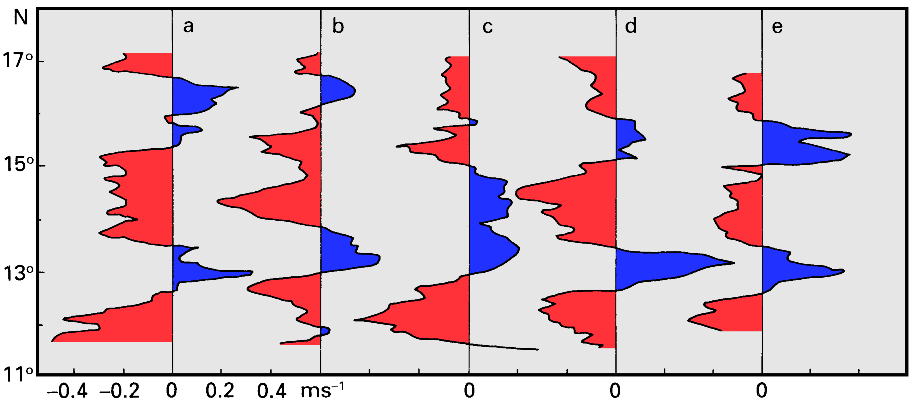


Fig. 16.17. East-west component of velocity (m s^{-1}, positive is eastward), averaged over 0 - 200 m from measurements with an acoustic Doppler current meter, as a function of latitude along 63.55°W in (a) August 1985, (b) January, (c) March, (d) July, (e) October 1986. The velocity scale is correct for (a), other curves are shifted as indicated by the zero velocity line. Adapted from Smith and Morrison (1989).

The continuation of the Caribbean Current through the Gulf of Mexico is known as the Loop Current. This is a true western boundary current which separates from the continental shelf north of Yucatan Strait. It is therefore characterized by instability of its path and periodic eddy shedding. Figure 16.18 gives a summary of the circulation features associated with the Loop Current. The main path followed by the current penetrates the Gulf to about 27°N. The speed of the current in Yucatan Strait usually exceeds 1 m s^{-1} at the surface but falls off with depth, reaching 0.4 m s^{-1} at about 1000 m depth. A southward undercurrent is often found in the last 200 m above the sill depth of 1895 m. It is highly variable, can sustain velocities of 0.05 m s^{-1} over several months and 0.15 m s^{-1} in bursts, but a mean over three years gave a net southward flow of just under 0.02 m s^{-1}.

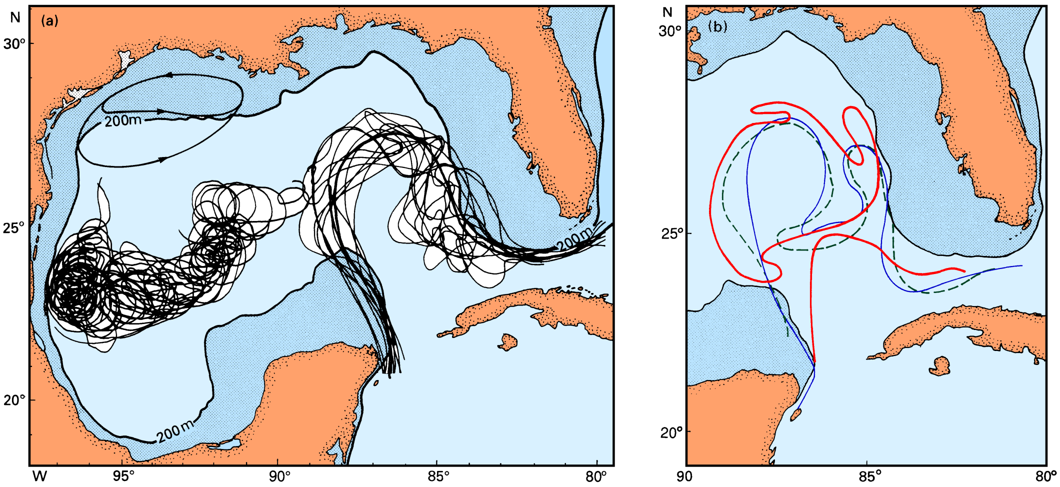


Fig. 16.18. Circulation in the Gulf of Mexico. (a) Mean position of the Loop Current during 1980 - 1984 (heavy line) and positions inferred from satellite observations of sea surface temperature, tracks of satellite tracked drifting buoys indicating eddy movement in the west, and schematic circulation 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.

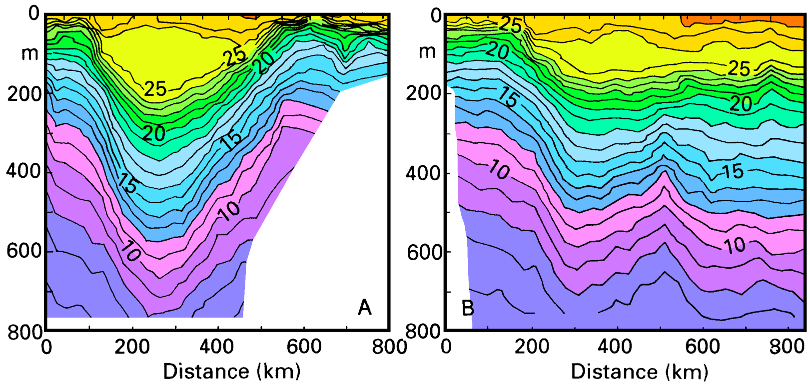


Fig. 16.19. Temperature along sections *A* and *B* shown in Fig. 16.20. Yucatan Strait is on the right. From Lewis and Kirwan (1987).

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 depression of the isotherms for one such eddy as it was crossed by two sections along tracks only about 100 km apart. Note the difference in the depth of the 15 - 25°C isotherms across Yucatan Strait (the southern end of the sections); the associated thermocline slope indicates the Loop Current as it leaves the Strait between the two sections. The steep slope of the isotherms further north in section B indicates where the Loop Current crosses the section to continue towards Florida Strait.

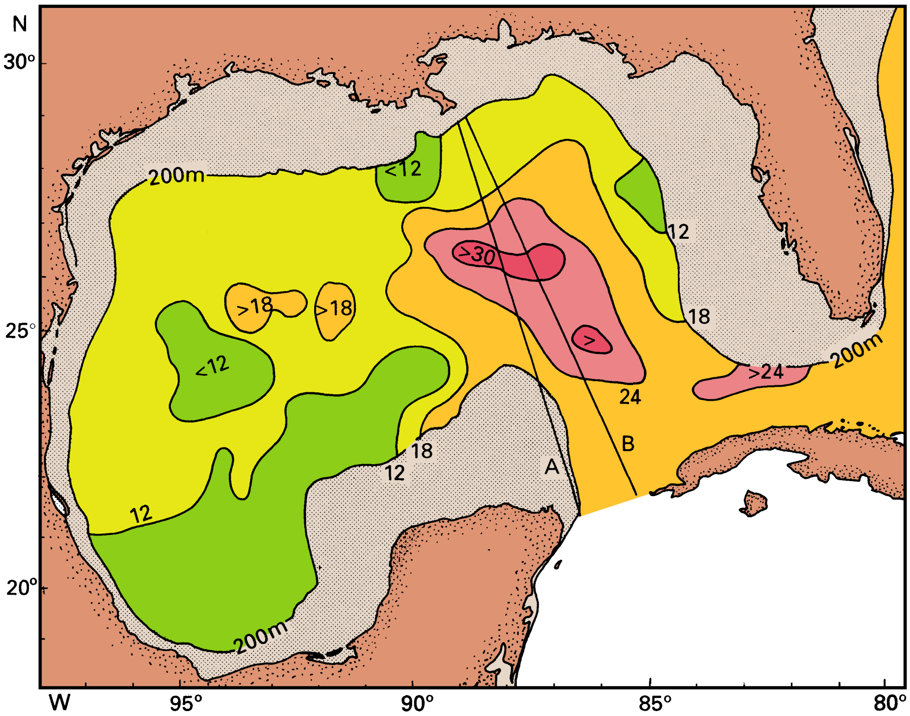


Fig. 16.20. Long term mean variability of sea level (cm). After Maul and Hermann (1985). The lines indicate the positions of sections *A* and *B* of Fig. 16.19.

Once formed the eddies drift away from the Loop Current in a general southwestward direction at 3 - 4 km per day. Like other eddies of western boundary currents they have typical diameters of 200 - 300 km and surface speeds of 1 - 2 m s^{-1} depending on age. When they reach the western continental shelf they still induce shelf currents as high as 0.7 m s^{-1}. Figure 16.18 shows typical eddy paths as reflected in the movement of drifting buoys. It is obvious that the direction of water movement in the western Gulf of Mexico at any particular moment is determined by the eddy field. Given their rate of formation and mean drift speed, between one and three eddies are usually present at any one time. The net circulation, determined geostrophically from smoothed data, indicates anti-cyclonic (clockwise) movement of about 5 Sv around the Gulf and a smaller cyclonic feature in the north with a transport of 8 Sv and linked with the circulation on the northern shelf (Figure 16.18a).

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.

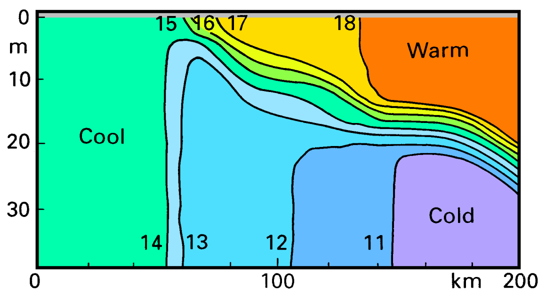


Fig. 16.21. Sketch of a temperature section through a shallow sea front, based on observations from the eastern entry to the English Channel. Note that the temperature on the shallow side (on the left) equals the temperature in the core of the thermocline.

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. Shallow sea fronts occur at well defined locations which are determined by a combination of water depth h and tidal current$u$. Energy arguments show that the parameter which measures the competition between thermal stabilization and tidal stirring is h/u^3.

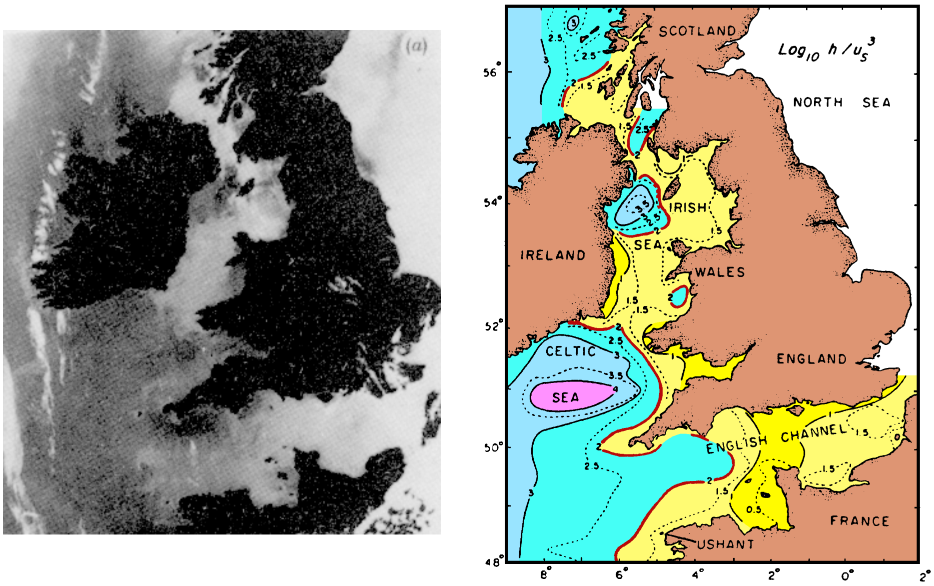


Fig. 16.22. Shallow sea fronts in the Irish Sea. (a) Sea surface temperature derived from satellite observations during May 1980; dark is warm, gray and white is cold (some small scale white features are clouds). The dark regions are stratified and therefore show warm surface temperature, while the gray regions are mixed. Fronts are found along the edges of the gray regions. (b) Contours of $\log_{10}\ h/u^3$; the contour closest to the position of the fronts is highlighted. Adapted from Simpson (1981).

The topography of the *Irish Sea* resembles that of a channel with gentle slopes on either side. Maximum depths along the axis are near 110 - 140 m in the south and exceed 250 m between northern Ireland and Scotland. East of 4.5°W the depth rarely exceeds 50 m. If this topography is combined with the magnitude of the tidal current, the resulting distribution of h/u^3 gives the contours shown in Figure 16.22. By comparing the contours with the sea surface temperature during May 1980 it is seen that most of the Irish Sea is well-mixed throughout the year and separated from the stratified regions by shallow sea fronts in the north and south.

Mean flow through the Irish Sea is weak and difficult to measure directly, due to the dominance of strong tidal flow. It can be deduced from hydrographic properties, which indicate northward movement with inflow from the Celtic Sea and outflow to the Hebrides (Figure 16.23). The circulation in the eastern Irish Sea is modified by freshwater input from several rivers which lowers the sea surface salinity (Figure 16.24) and produces 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.

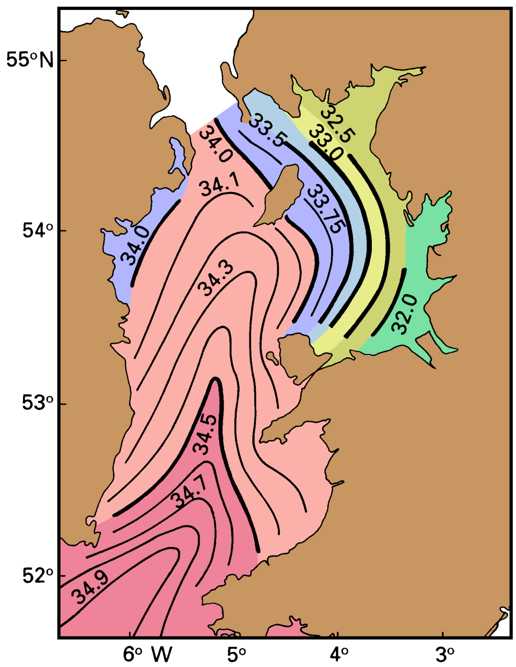


Fig. 16.23. Mean surface salinity in the Irish Sea

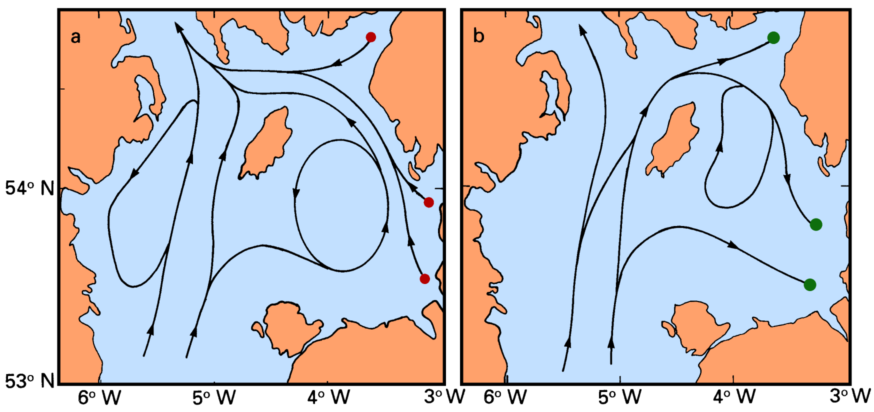


Fig. 16.24. Mean circulation in the northern Irish Sea. (a) Near-surface, (b) near-bottom. Dots indicate entrainment from the near-bottom into the near-surface layer.

The introduction of nuclear power stations resulted in contamination of the European shelf with radioactive cesium nuclides from two nuclear fuel reprocessing plants. Sellafield (formerly Windscale), the larger of the two, is located on the coast of the northern Irish Sea. The spreading of ^{137}Cs confirms the concept of mean northward flow and indicates the passage of Irish Sea water into the North Sea within 3 - 4 years (Figure 16.25). Some ^{137}Cs spreads into the Celtic Sea to the south by tidal dispersion and during occasional wind-driven reversals of the mean flow. The same process disperses some ^{137}Cs from La Hague, the second reprocessing plant which is located on the French coast of the English Channel, westward allowing it to enter the southern Irish Sea with the mean flow.

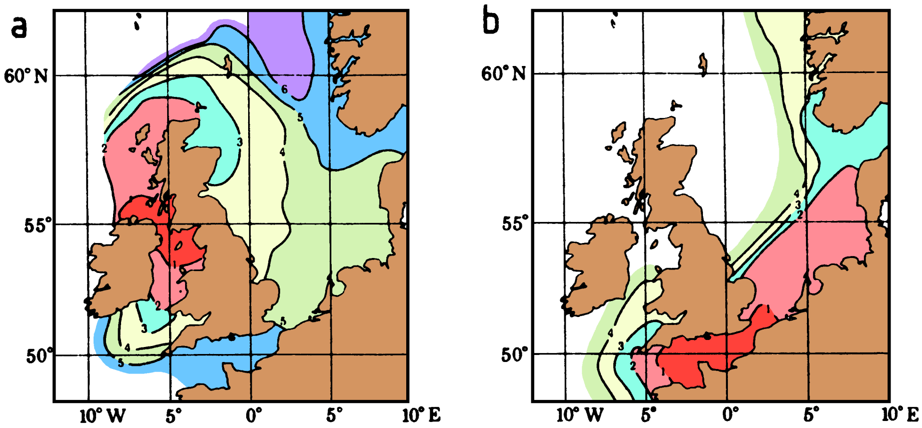


Fig. 16.25. Spreading of ^{137}Cs from Sellafield (a) and La Hague (b), expressed as average time in years required for a water particle to travel to the positions indicated by the contours. The distribution is the result of dispersion and advection with the mean flow, the latter being indicated by the asymmetry of the contours with respect to the release point. From Zimmerman (1984).

The inferences made from the distribution of ^{137}Cs can be extended to give information on the mean water movement of the *North Sea* as well. Effluent from La Hague spreads mainly eastward, indicating a net flow of water through the English Channel from the Bay of Biscay into the North Sea. The North Sea itself displays anti-clockwise circulation, evidenced by the movement of Sellafield effluent along the British coast and the spreading of La Hague effluent along the Scandinavian coast.

The North Sea is generally shallow, with depths around 120 - 150 m in the north decreasing to 100 m between Aberdeen and Stavanger, to 50 m between Hull and Skagen, and to 30 - 40 m further to the southeast. An exception is the Doggerbank, a shallow region that stretches from 2°E to 6°E just south of the 50 m isobath. At 54.3°N, 2°E it rises to 13 m and is feared by fishermen for its dangerous waves which can break in severe weather and have caused the loss of more than one vessel. The other exception is the Norwegian trench which stretches along the Norwegian coastline with depths around 300 m and a width of 35 - 80 km at the 250 m depth contour; between 7°E and 9°E it ends in a depression with a maximum depth of 700 m.

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.

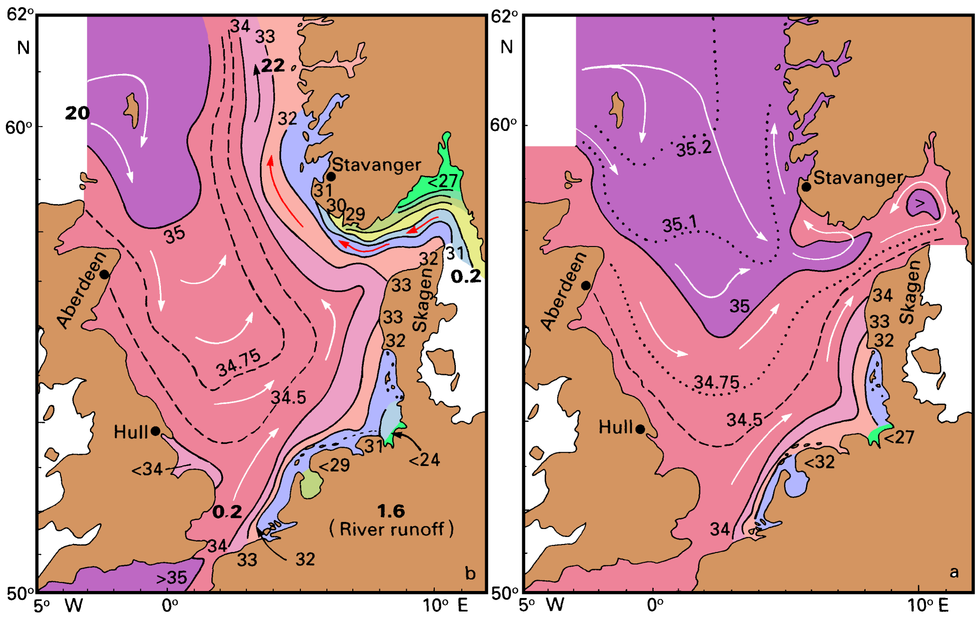


Fig. 16.26. Mean salinity and circulation in the North Sea for June. (a) At 7.5 m depth, (b) above the bottom. Bold numbers give average transport in 1000 km^3 per year (1000 km^3 per year = 0.03 Sv). Adapted from Goedecke *et al.* (1967).

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. Observations of currents near the sea floor show that they are associated with geostrophic flow along the fronts of up to 0.15 m s^{-1}. The development of a summer thermocline is restricted to the region deeper than 50 m and east of 0°W. Regions shallower than this depth and the waters along England and Scotland remain unstratified throughout the year due to strong tidal currents. A strong seasonal thermocline develops in the eastern and southern parts of the Norwegian trench where the salinity stratification produces high vertical stability of the water column and inhibits vertical mixing. The entire North Sea is well mixed vertically during winter; horizontal temperature differences result from the larger seasonal range in shallow coastal water.

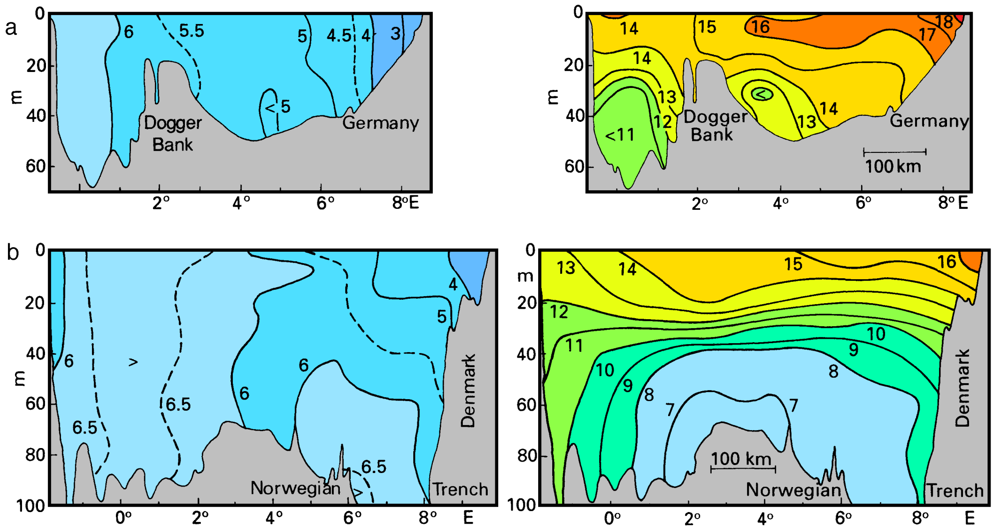


Fig. 16.27. Mean temperature (°C) for February (left) and August (right) based on data for the period 1902 - 1954. (a) Across the Doggerbank along 54.5°N, (b) from Aberdeen to Skagen along 57.5°N. Note the similarity of the isotherm distribution near the Doggerbank and the Scottish coast with the sketch of Fig. 16.21 during summer. The sharpness of the fronts is greatly reduced by averaging over 52 years. From Tomczak and Goedecke (1964).

The *Baltic Sea*, a mediterranean dilution basin, is in many aspects similar to the Black Sea. It has very restricted exchange with the open ocean and a significant freshwater surplus; as a result its salinity is also well below normal oceanic values, and oxygen is regularly depleted in the deep basins and replaced by hydrogen sulfide. Its area is comparable (350,000 km^2), and so is its freshwater surplus (between 300 km^3 per year in February and 770 km^3 per year in May, annual mean 470 km^3 per year). Its system of passages to the North Sea and the open ocean - Belt Sea, Kattegat, and Skagerrak - is also similarly shallow and complicated (Figure 16.28; the Kattegat has a mean depth of 23 m, and the sill depth in the Belt Sea is 18 m). The all important difference lies in the volume of the two mediterranean basins: The mean depth of the Baltic Sea is about 60 m, giving it a total volume of about 20,000 km^3 or 4% of the volume of the Black Sea. Thus, while the 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.

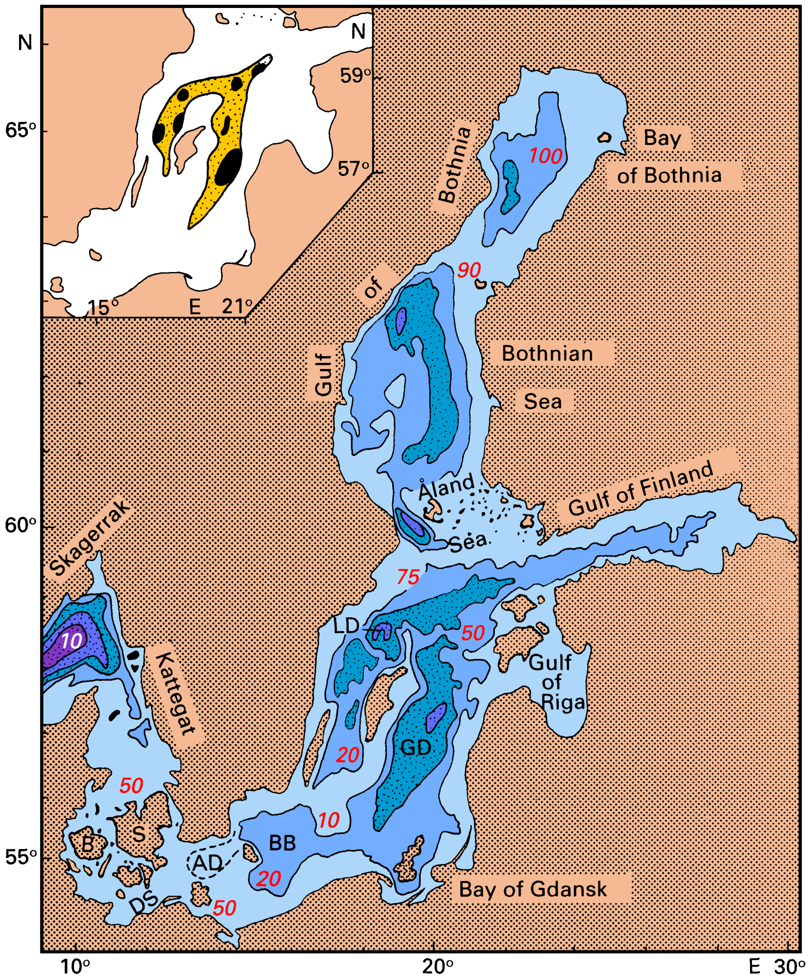


Fig. 16.28. Topography of the Baltic Sea. Depths > 100 m are stippled, with contours drawn for 50, 100, 300, and 500 m depth. AD: Arkona Deep, BB: Bornholm Basin, BS: Belt Sea, DS: Darß Sill, GD: Gotland Deep, LD: Landsort Deep. Numbers in italics give the probability of total ice coverage (*100%* = 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).

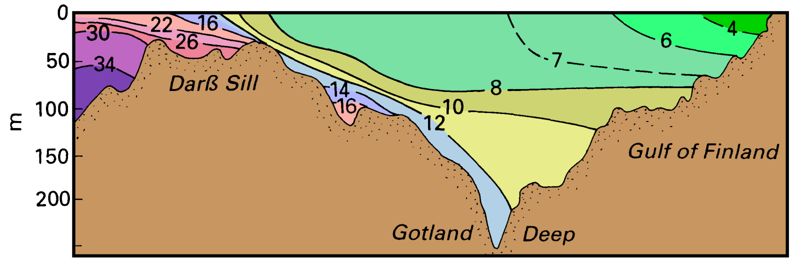


Fig. 16.29. Salinity in a section along the axis of the Baltic Sea. Note the change of contouring interval across the Darß Sill.

The strong salinity stratification provides the water column with ample stability even in winter, and convective mixing during the cooling period is restricted to the upper 70 m or so. This results in a large seasonal temperature variation at the surface and in combination with the low surface salinity leads to regular ice formation (Figure 16.28). Below 70 m depth the temperature is rather stable and quickly approaches the 5.5°C found throughout the year in the deep basins. The thermocline at 70 m is also the upper limit for the occurrence of hydrogen sulfide during extended periods of stagnation; during most years hydrogen sulfide does not occur above 120 - 150 m depth.

Water renewal in the Baltic basins is a highly intermittent process. Individual inflow events last between ten days and three months (average duration 32 days) and are associated with a rise of sea level in the Baltic Sea of some 0.6 m. They are restricted to the period August - April; most of them occur in December. Mean inflow velocities are 0.1 m s^{-1} near the Darß Sill and 0.08 m s^{-1} on approaching the Bornholm Basin. On average, the amount of water transported into the Baltic Sea is 200 ± 100 km^3, which is 35 - 100% of the volume of the Belt Sea. The amount of high salinity water brought in by each event is, however, considerably less, because at the beginning of the inflow the lower layer of the Belt Sea contains a mixture of low salinity Baltic and high salinity North Sea water and has to be emptied before less diluted high salinity water can flow in. Figure 16.30 shows the volumes of high salinity water transported by the 90 major events which were identified for the period 1897 - 1976. It is seen that the events occur in clusters, on average every three years, and alternate with periods of 1 - 4 years without major inflow. The same variability has been found in the meridional wind component over the North Atlantic Ocean (Börngen *et al.*, 1990). This suggests that inflow into the Baltic Sea is controlled by interannual variability of the global climate.

Systematic measurements in the deep basins indicate that long term climate change is also likely to affect the hydrography of the Baltic Sea. Data collected since 1958 (Figure 16.31) indicate that several major renewal events reached the Gotland Deep; however, it is also clear from the data that over the observation period deep water renewal is not frequent enough to prevent a general increase of hydrogen sulfide concentration and that no major inflow event occurred after 1976*. This is accompanied by progressive freshening, a result of diffusion through the halocline. Whether this process will continue is impossible to predict at present. What is clear from the data is that the hydrographic state of the Baltic Sea is finely tuned and susceptible to changes in the world climate. It also reacts 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.

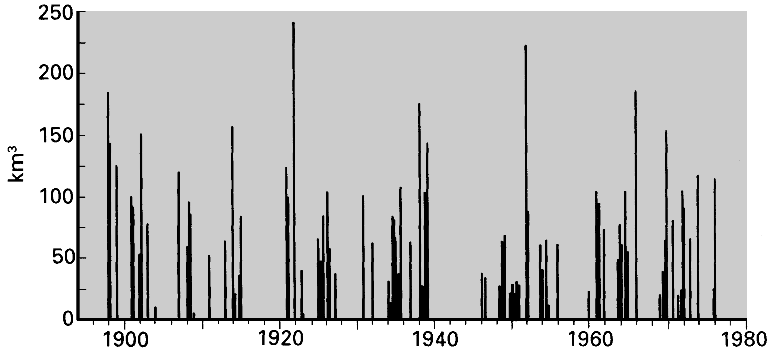


Fig. 16.30. Volumes of high salinity inflow into the Baltic Sea. The gaps during 1914 - 1918 and 1939 - 1945 are from lack of observations. From Matthäus and Franck (1990).

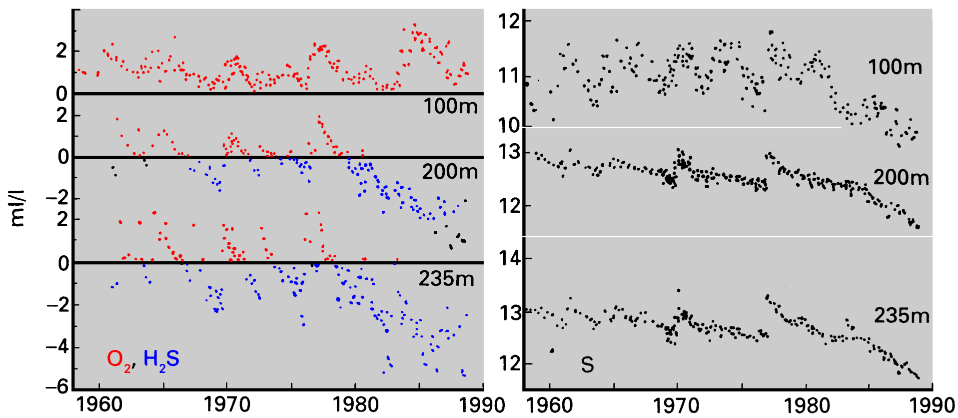


Fig. 16.31. Oxygen O_2 (ml/l), hydrogen sulfide H_2S (ml/l, plotted as negative oxygen), and salinity *S* at a station in the Gotland Deep. Strong inflow events occurred in 1965, 1970, 1973, and 1976. Note that after each inflow event salinity and oxygen are higher at 235 m than at 200 m, indicating 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.