Chapter 17

Aspects of advanced regional oceanography

The discussion of the Atlantic Ocean's adjacent seas completes our tour of the circulation and hydrographic structure of the world ocean. The material presented in sixteen chapters covered a lot of ground and might leave the impression that not much remains to be told. To dispel any doubt in that respect and to demonstrate that the word "Introduction" in the title of this book describes its contents correctly, we devote one chapter to some examples of what can be considered advanced regional oceanography. Compared to the introductory level of this book, advanced regional oceanography is an extension of the material in two respects. It addresses the oceanography of a given region in much greater detail than is possible in a introduction, and it explains it in terms of oceanographic concepts and processes which are beyond the scope of an introductory text. Advanced regional oceanography requires an advanced understanding of ocean dynamics (e.g. an understanding of internal waves or double diffusion). However, in this chapter we will restrict our discussion to examples that can be discussed on the basis of concepts already introduced in earlier chapters.

Example 1: Modification of Central Water in the Tasman Sea

Our first example brings us back to the discussion of Central Water of the South Pacific Ocean. It was pointed out in Chapter 9 that Western South Pacific Central Water (WSPCW) which is found in the Tasman and Coral Seas and just east of New Zealand has properties virtually identical to Indian and South Atlantic Central Water. This was taken as an indication that conditions in the Subtropical Convergence (STC) of the southern hemisphere from where these Central Waters originate do not vary much around the globe. Figure 17.1 compares again the T-S diagrams of Indian Central Water (ICW) with those of WSPCW at various locations in the western Pacific Ocean. It is noted that in the Tasman Sea WSPCW is significantly more saline than ICW and that this difference increases from north to south. In contrast, WSPCW in the Coral Sea has properties very similar to those of ICW. While this indicates that the water found in the thermocline of the Coral Sea originates from the southern STC (i.e. from south of the Tasman Sea), it also shows that it could not have taken the most direct route through the Tasman Sea where the salinity is higher than both in the north and in the south. The regional salinity maximum in the Central Water of the Tasman Sea requires a regional source of high salinity water. This source is located at the western edge and is known as the Bass Strait Water Cascade.

Bass Strait is a shallow sea between Tasmania and mainland Australia connecting the Tasman Sea with the Great Australian Bight. It is about 80 m deep in the centre and has sills slightly deeper than 50 m on either side. During summer its hydrographic properties are the same as those of Tasman Sea surface water, but in winter it cools faster than the Tasman Sea, and its T-S characteristics differ (Figure 17.1). This comes from the fact that cooling of Tasman Sea surface water during autumn and winter deepens the surface mixed layer through convection. As long as the mixed layer is shallower than the depth of Bass

Strait the same process operates in that region as well, and the mixed layer temperature in both regions is the same. When that depth is reached, Bass Strait Water cools faster than the surface water in the Tasman Sea, where continued entrainment of water from below keeps the cooling rate low. By the middle of winter the mixed layer in the Tasman Sea has reached a thickness of 100 m and more and Bass Strait Water is some 5°C colder and therefore denser than Tasman Sea surface water. A density front is established between the two regions and maintained by geostrophic adjustment, i.e. a northward current of Bass Strait Water on the western side (Figure 17.2). When this water approaches the southern coast of the Australian mainland it sinks down along the continental slope and continues as an undercurrent at the depth where it finds water of its own density. The sinking process, known as the Bas Strait Water Cascade, occurs in a well defined region and is apparently associated with a canyon. As is seen from the T-S diagram of Fig 17.1 cascade water is more saline and slightly warmer than Tasman Sea water of the same density and can therefore be identified along its path by its high salinity. As it flows eastward into the Tasman Sea the Coriolis force keeps it close to the Australian shelf, forcing it to turn northward at the southeastern corner.

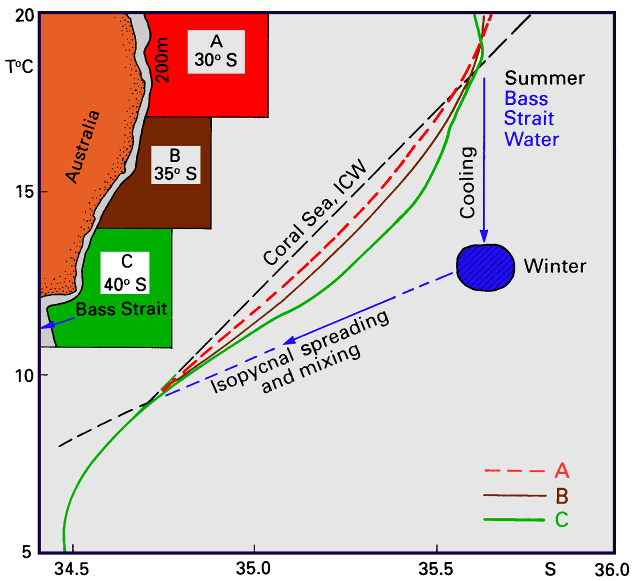

Fig 17.1. Temperature-salinity diagrams for Indian Central Water (ICW) and for Western South Pacific Central Water (WSPCW) in the Coral Sea and for three Tasman Sea areas as identified in the inset. Also shown are the summer and winter T-S characteristics for Bass Strait Water. See text for details. Adapted from Tomczak (1981d).

The warm and saline undercurrent from the Bass Strait Water Cascade (Figure 17.3) proceeds northward until it encounters the East Australian Current, which sweeps it out into the open Tasman Sea. It quickly loses its identity as a water mass, through mixing in eddies and turbulence in the strong current shear associated with the East Australian Current, but its contribution to the properties of the Tasman Sea thermocline is seen in the anomalously high salinities of WSPCW at the temperature characteristic for Bass Strait Water during winter (Figure 17.4).

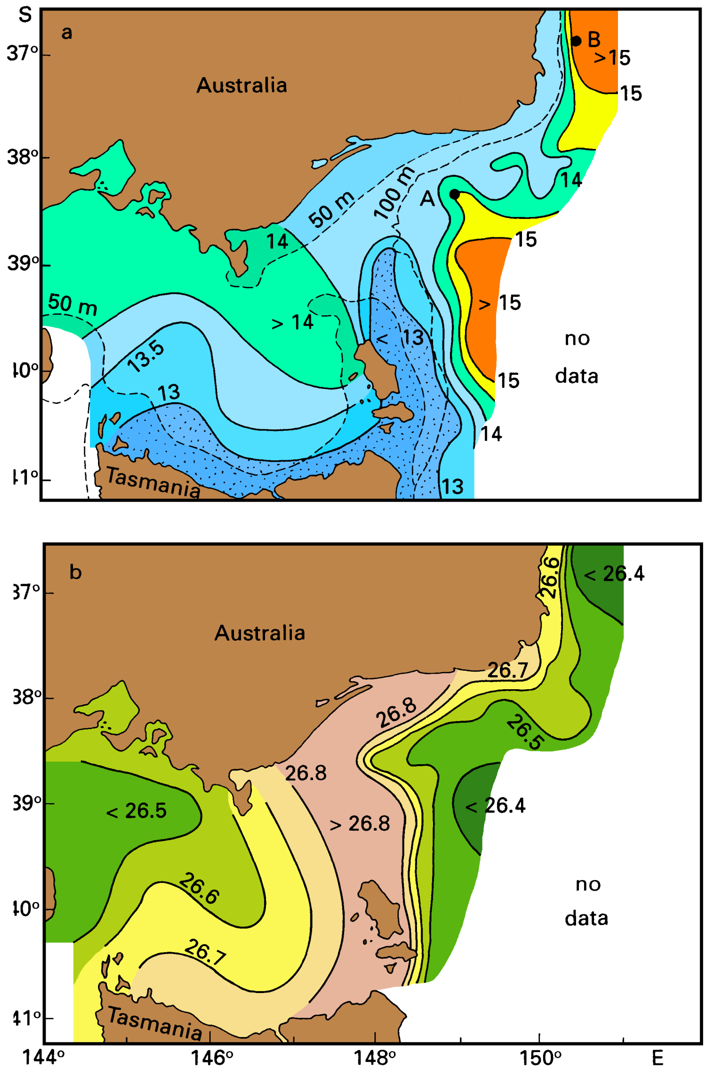

Fig. 17.2. Hydrographic conditions in Bass Strait and in the adjacent Tasman Sea during winter 1981, showing the front between the two regions.

(a) Surface temperature (°C),

(b) surface density (σ_t).

The broken lines are isobaths; the large dots indicate the positions of the stations used in Fig. 17.3. From Tomczak (1985).

The impact of Bass Strait water on WSPCW properties is an example of modification of oceanic water masses by coastal, shelf, and estuarine processes. Similar situations are found in many parts of the world ocean. Another undercurrent of high salinity and temperature has been observed in the Great Australian Bight east of Spencer Gulf. As in the case of Bass Strait, the undercurrent results from an increase of the water density on the shelf (i.e. in Spencer Gulf) above the density of the adjacent oceanic surface water. In this case the density increase is the result of very high evaporation in the arid desert environment of South Australia. It is therefore caused by a salinity increase and requires only a minor drop of sea surface temperature in early autumn to start the sinking motion. The resulting undercurrent, however, shows very similar behaviour and properties. Another example of shelf influence on thermocline waters is described below.

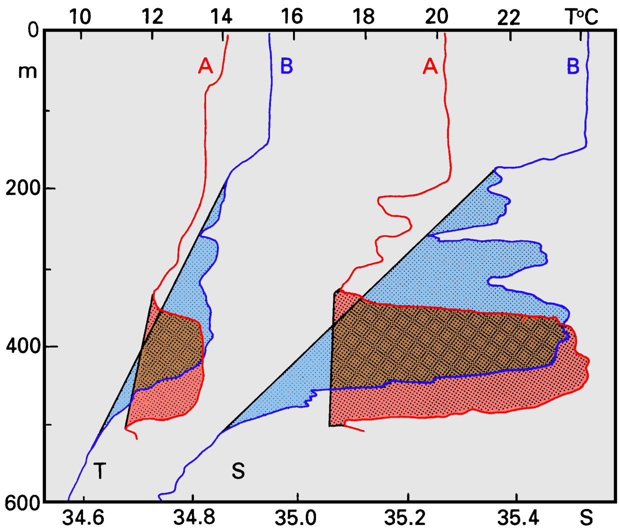

Fig. 17.3. The Bass Strait Water undercurrent in the Tasman Sea as seen in vertical profiles of temperature (T) and salinity (S). Temperature and salinity anomalies produced by the undercurrent are shaded. See Fig. 17.2

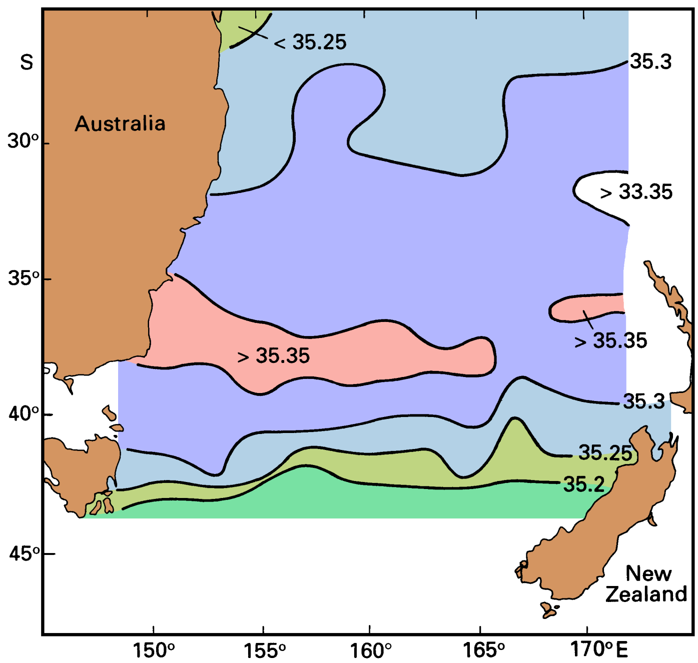

Fig. 17.4. Salinity on the 14.5°C isothermal surface in the Tasman Sea. The depth of this surface varies between less than 100 m in the south to 350 m in the north. High salinity along 37°S indicates Bass Strait Water influence. From Villanoy and Tomczak (1991).

Example 2: Mixing in the Canary Current upwelling region

In our second example we return to the discussion of coastal upwelling in the North Atlantic Ocean. The Canary Current upwelling system is probably the most complex of all coastal upwelling systems, particularly near Cape Blanc (20°S) where the frontal zone between North Atlantic Central Water (NADW) and South Atlantic Central Water (SACW), which was discussed in Chapter 15, reaches the African coast. Mixing between the two water masses, which have very different hydrographic properties, competes with the effect of upwelling, to the degree that it is often difficult to decide whether an observed increase in primary productivity is the result of upwelling, mixing, or advection. Compared with NACW, its southern counterpart SACW has similar temperatures but somewhat lower salinities and much higher nutrient concentrations. An intrusion of SACW into a region originally occupied by NACW can double or treble the nutrient levels in the euphotic zone within hours, bringing phosphate concentrations up from less then 0.5 to 1.5 μg-at/l and raising silicate concentrations from below 5 to 15 μg-at/l. Upwelling of NACW could not lift those nutrients levels above concentrations of 1.0 and 7 μg-at/l, respectively, even if it came from 500 m depth which is very unlikely. It is seen that in the vicinity of the frontal zone horizontal advection of water masses can be at least as important for primary productivity as the upwelling process itself. Because the water masses on either side have different temperatures and salinities but identical densities, the front is density compensated and thus not associated with geostrophic flow but dynamically passive. It is therefore quite common to see parcels of water drift off across the front and continue their movement as intrusions or lenses embedded in different water (A typical example of such a situation was seen in Figure 15.12). As these parcels are lifted into the euphotic zone with the upwelling, the nutrient content in the surface layer is constantly varying depending on the source water masses of the layers. The result is an extremely patchy nutrient distribution in the surface layer and a corresponding patchiness in primary productivity.

The situation is complicated further by the presence of a coastal water mass originating from Arguin Bank, a large expanse of water less than 20 m deep off the coast of Mauritania. Situated in an arid desert climate typical for coastal upwelling regions, this body of water is exposed to strong solar heating and evaporation. The effect on its hydrological properties is determined by the circulation which in turn is driven by loose coupling with the oceanic circulation. The Canary Current veers southwestward at Cape Blanc, leaving the region seaward of Arguin Bank to a northward flowing countercurrent (Figure 17.5). Flow on the bank is therefore anticyclonic; water leaves the bank in the south and is replaced from the north. There is thus little difference between oceanic water properties and those of Arguin Bank Water in the north; but in a bay with average depths around 10 m not much heat is required to warm the water and increase its salinity through evaporation, and by the time the water reaches the southern part of the bank it leaves it with a salinity above 39 and a temperature above 24°C. In the coastal upwelling region, offshore water with an equivalent density is found between 100 - 200 m depth, so Arguin Bank Water sinks to that level where it joins the general northward movement and the upwelling process. This sinking motion occurs against the background of coastal upwelling and is therefore not continuous but concentrated in canyons where it occurs in bursts, alternating with vigorous upwelling events.

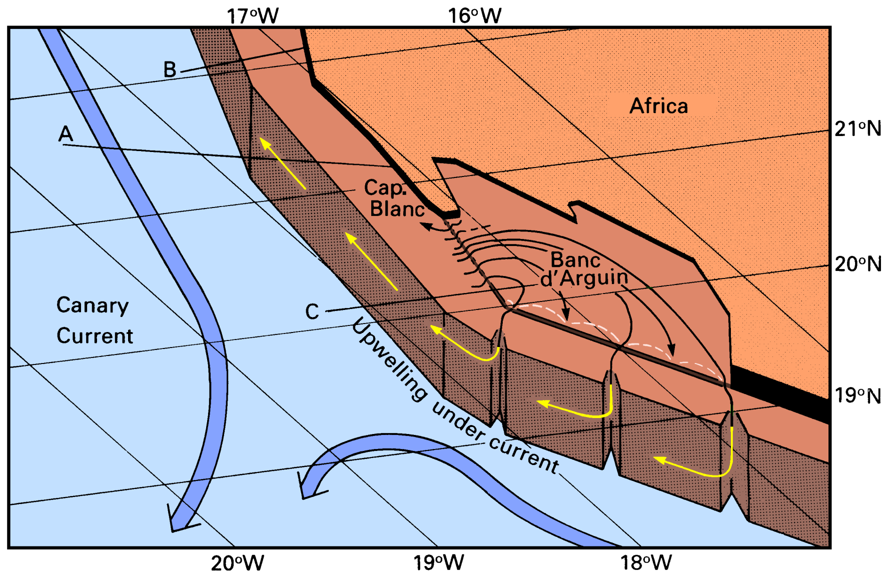

Fig. 17.5. Sketch of the circulation on Arguin Bank. Lines *A*, *B*, and *C* indicate the locations of the sections shown in Figs 17.6 and 17.8. Adapted from Peters (1976).

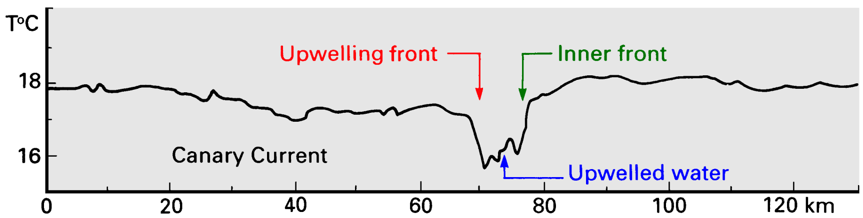

Fig. 17.6 Sea surface temperature along 20°N (section *C* in Fig. 17.5) from the Canary Current to Arguin Bank. The upwelling front, produced by the surfacing of the thermocline, separates the coastal upwelling zone to the right from the region of cold water advection in the Canary Current. The inner front separates Arguin Bank Water from oceanic water. From Tomczak and Miosga (1976).

Similar to the situation encountered with the Bass Strait Water Cascade, the coastal water is retained at the surface by a front situated roughly inside the shelf break. This produces the unique situation that the coastal upwelling zone is bounded by two fronts (Figure 17.6) and appears in the sea surface temperature as a narrow band of cold water between warmer water on either side. The temperature contrast from upwelled water to coastal water is typically 2 - 3°C and occurs over a 10 km distance. This corresponds to a gradient of about 0.2°C km^{-1}. The maximum observed gradient, which occurs only over a narrow strip of 1 - 2 km in the centre of the front, is usually in the range 0.5 - 1.0°C km^{-1} but can reach up to 2°C km^{-1}. This is more than the gradients encountered in the upwelling front but still an order of magnitude smaller than upwelling fronts observed in the Benguela Current upwelling region. Despite the relatively small thermal contrast the front is easily located even visually (Figure 17.7), since Arguin Bank Water is even less transparent than upwelled water, which on behalf of its high productivity is already low in transparency. Sand dust blown across the water from the Sahara desert contributes significantly to the discoloration of Arguin Bank Water.

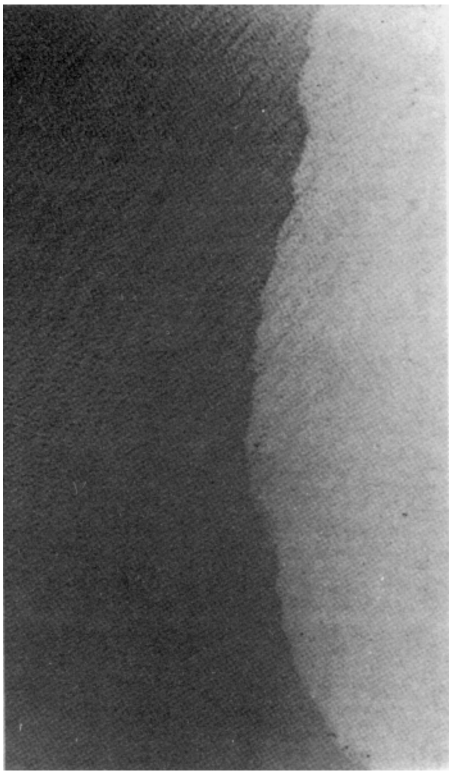

Fig. 17.7. A photograph of the sea surface in the region of the Arguin Bank front. Arguin Bank Water is on the right, the cold band of upwelling water to the left. The north-south extent of the area shown is about 18 km. From Tomczak and Miosga (1976).

North of Cape Blanc the upwelling system follows the dynamics sketched in Figure 8.25 which shows lowest temperatures at its inshore edge. However, during periods of intensified upwelling - so-called upwelling events, which are related to the variability of the synoptic weather systems and thus last for about 5 days - the Arguin Bank front has been observed to extend northward past Cape Blanc, indicating that the mass deficit produced by the increase in Ekman transport towards the sea is balanced mainly by northward advection of Arguin Bank Water (Figure 17.8). The main effect of the increase in wind speed is offshore movement of the upwelled water, leaving the inshore region free to be filled by warm and saline shelf water. During these events the upwelling region north of Cape Blanc shows the same structure usually found further south, i.e. a narrow band of cold upwelled water between the oceanic water of the Canary Current and northward moving water from Arguin Bank.

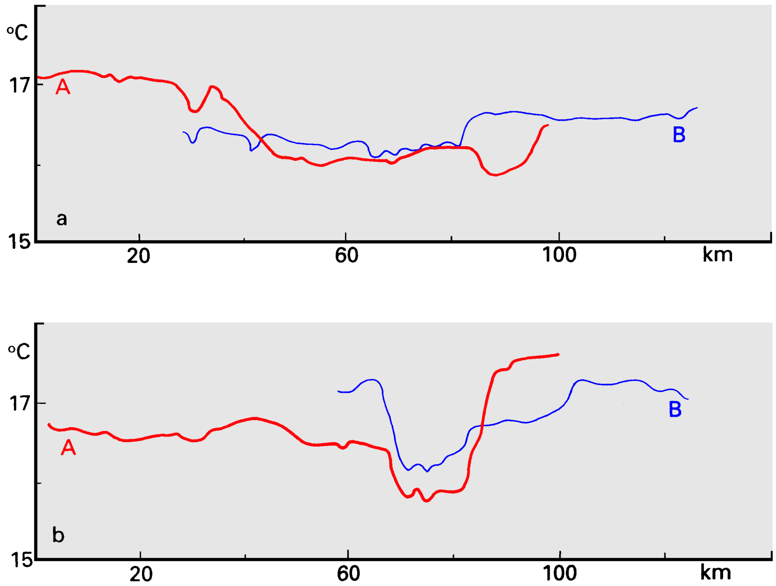

Fig. 17.8. Sea surface temperature along transects *A* and *B* of Fig. 17.5. (a) During weak upwelling, (b) during an event of strong upwelling. Note the absence of fronts and coastal warm water when the upwelling is weak. Note also how the inner front is eroded by mixing as the coastal water moves northward from *A* to *B*. The observations for (a) and (b) were taken ten days apart. From Tomczak (1981e).

Ocean variability and mixing

The two examples discussed above have to suffice as evidence for the important role coastal and shelf processes can play in shaping the details of the oceanic circulation and hydrography. There are many other places along the oceanic rim where similar situations can be found. As already said at the beginning of this chapter, more insight into physical processes is required to fully understand and describe the impact of the vigorous mixing and dynamic interactions between the ocean and its shelf waters. We leave this difficult topic here and conclude the examples of advanced regional oceanography with a brief discussion of ocean variability, i.e. the role of processes occurring on time scales from days to months and space scales from centimeters to hundreds of kilometers, leaving the longer time scales and larger space scales to a detailed discussion in the last three chapters.

Ocean variability occurs in many forms. Not all forms are present everywhere in the world ocean, and identifying their regional distribution is part of advanced regional oceanography. Instability of western boundary currents and associated ring shedding is one form of ocean variability; it is part of the dynamics of these currents and restricted to well defined regions of the world ocean. Other forms (such as the eddies seen in Figure 4.9) are related to the dynamics of the vast geostrophic interior of the ocean and therefore more ubiquitous; but their intensity varies in space and time, and a task of advanced regional oceanography would be to quantify their regional occurrence (an attempt to achieve this is presented in Figure 4.8). Interleaving of water masses (as seen in Figure 15.12) is yet another form of variability; it occurs preferably in the vicinity of fronts but can be found in regions of quite uniform property distributions as well, as we saw in our discussion of the outflow of Mediterranean Water into the Atlantic Ocean and its "meddies" (Figure 15.5).

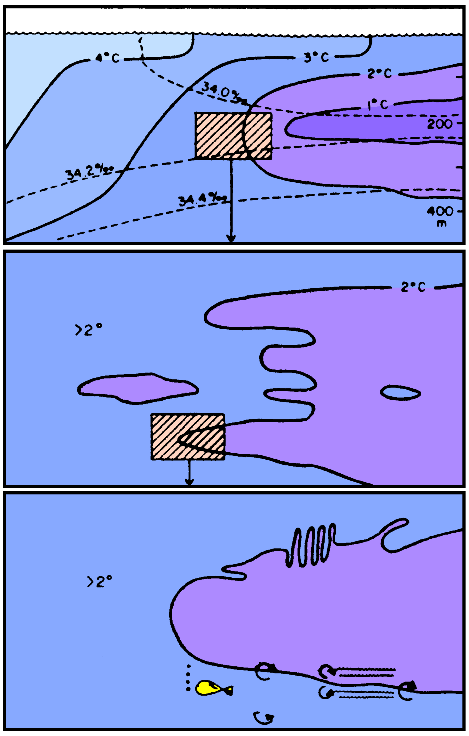

Fig. 17.9. Scales of ocean variability: oceanic fronts and eddies (top), interleaving (middle), double diffusion (bottom). Approximate overall horizontal and vertical sizes for the three sketches are 500 - 1000 km and 500 m (top), 100 - 150 km and 100 m (middle), and 20 km and 25 m (bottom), each panel being an enlargement of the shaded region in the panel above. See text for more explanation. From Joyce (1977).

More detailed investigation reveals a sequence of variability scales spanning several orders of magnitude that tend to co-exist (Figure 17.9). The largest scale is set by features of the mean circulation; Figure 17.9 uses the Antarctic Polar Front as an example (compare the top panel of the figure with Figure 6.8a). The deviation of isotherms and isohalines from horizontal is an indication for geostrophic flow normal to the section. Eddies have similar scales, they differ from oceanic fronts mainly by the fact that their dimensions are comparable in all horizontal directions; in comparison, frontal dimensions are small across the front but large in a direction parallel to the front. The top panel of Figure 17.9 can therefore also be seen as representing half an eddy. The important point to observe here is that at these scales ocean variability is geostrophic, i.e. water movement in its elements is in geostrophic balance. The eddies cause departures from the long-term mean flow, and a time series of currents looks very much like a slowed-down version of turbulence as it is observed for example in a water pipe. The difference is that turbulence in a pipe is too fast to be affected by the Coriolis force. In the ocean, flow variations caused by eddies are always accompanied by geostrophic adjustment of the density field. This type of variability is therefore also called *geostrophic turbulence*. The intensity of geostrophic turbulence varies in space and time. To give an example, Figure 17.10 shows the regional distribution of eddy velocities associated with geostrophic turbulence for the North Atlantic Ocean as deduced from satellite-tracked surface drifters.

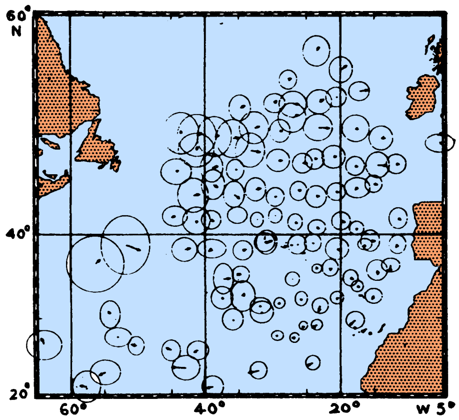

Fig. 17.10. Regional distribution of geostrophic turbulence in the North Atlantic Ocean. The arrows show the mean surface current, the axes of the ellipses give the mean north-south and east-west eddy velocities. Mean eddy velocities increase threefold from Ireland to Newfoundland. In the centre of the subtropical gyre (35°N, 30°W) mean eddy velocities are below 10% of the values reached east of Newfoundland. Note also that with few exceptions, mean eddy velocities are larger than the mean current and therefore cause reversals of the mean flow. From Krauss and Käse (1984).

Around the edges of the eddies are found the *intrusions* (the middle panel of Figure 17.9). They cause deformations of isotherms and isohalines which have to be density-compensated to keep the stratification stable. It follows that on this scale ocean variability is not constrained by geostrophy. It would appear that this should make interleaving the dominant turbulence mechanism in the ocean, since so much less energy is required to disturb the oceanic mean state if there is no need for geostrophic adjustment. However, there is plenty of energy available from the atmosphere at the scales of geostrophic turbulence, while not much energy goes into the generation of turbulence at the interleaving scale. Advanced study would show that other processes, such as the stirring of water above sills or the injection of water from the shelf, are required to trigger significant interleaving. As a consequence, interleaving is less common than geostrophic turbulence and shows a much more uneven regional distribution.

The basis of both geostrophic turbulence and interleaving is that water particles are physically moved from one area of the mean circulation to another. If many such moves take place the result will be the mixing of water properties. Because all water properties, and in particular heat and salt, are exchanged with the turbulent movement of the particles, the rate of mixing is the same for all of them. This is not the case on the third and smallest turbulence scale (the bottom panel of Figure 17.9) which is linked with molecular diffusion processes.

The molecular diffusivity of salt is about two orders of magnitude smaller than the molecular diffusivity of heat; in other words, it is much more difficult to exchange salt on the molecular scale than it is to exchange heat. This gives rise to striking instabilities in the stratification. As an example, consider the situation in the oceanic thermocline.

Temperature decreases with depth, and so does salinity in most parts of the world ocean. On its own, the vertical salinity gradient would result in an unstable density stratification; but this instability is more than compensated by the vertical temperature gradient. Molecular diffusion tends to reduce the temperature and salinity gradients; but the temperature gradient is reduced much faster than the salinity gradient, and if this continues long enough there comes a moment when the temperature gradient is no longer sufficient to compensate for the salinity-induced instability. Convection sets in and occurs in narrow vertical tubes of rising low salinity water between narrow tubes of sinking salty water ("salt fingers"). The process is known as *double diffusion* and has been well documented in the laboratory. Observation in the ocean is much more difficult, and it is generally only possible to infer its existence from its impact on the larger scales. The physics of double diffusion are beyond the scope of this book, so two observations must do. It can be shown that a stratification characterized by a stable salinity but unstable temperature gradient is modified by double diffusion in such a way that the continuous gradients are replaced by a series of layers of uniform temperature and salinity. This type of stratification is sometimes found in the Arctic and Antarctic oceans when fresher melt water overlies warm salty ocean water. The predicted layers have indeed been found under drifting ice islands where the water column is sheltered from atmospheric disturbances and molecular diffusion is the dominant mixing process (Figure 17.11). It is also known that the intensity of double diffusion depends on the ratio of the rate of change of temperature and salinity with depth (the slope of the temperature-salinity (T-S) curve in the T-S diagram). This would indicate that double diffusion plays a more prominent role in the thermocline of the Atlantic and Indian Oceans than in the Pacific Ocean which has a much steeper T-S relationship (less salinity change for the same temperature change). Detailed description of the regional differences has to be left for an advanced textbook.

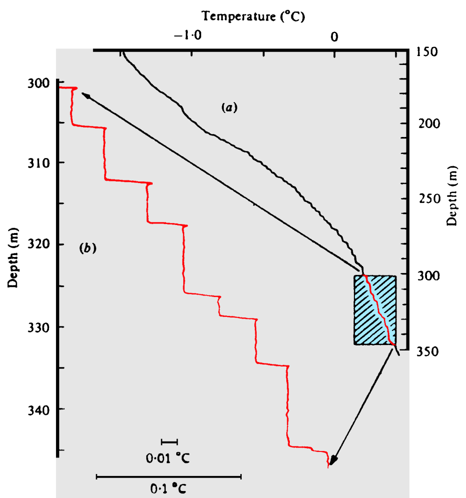

Fig. 17.11. Evidence for double diffusion in the form of homothermal layers observed under an Arctic ice island. From Neal *et al*. (1969).