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.

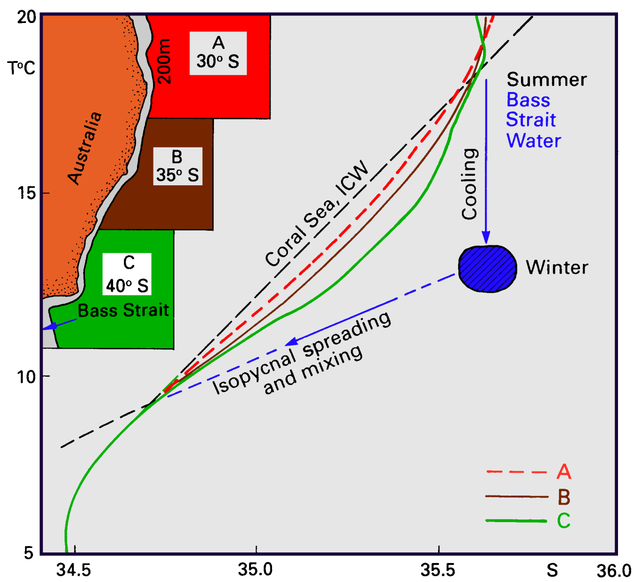

Fig 17.1. Temperature-salinity diagrams for Indian Central Water (ICW) and for Western South Pacific Central Water (WSPCW) in the Coral Sea and for three Tasman Sea areas as identified in the inset. Also shown are the summer and winter T-S characteristics for Bass Strait Water. See text for details.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

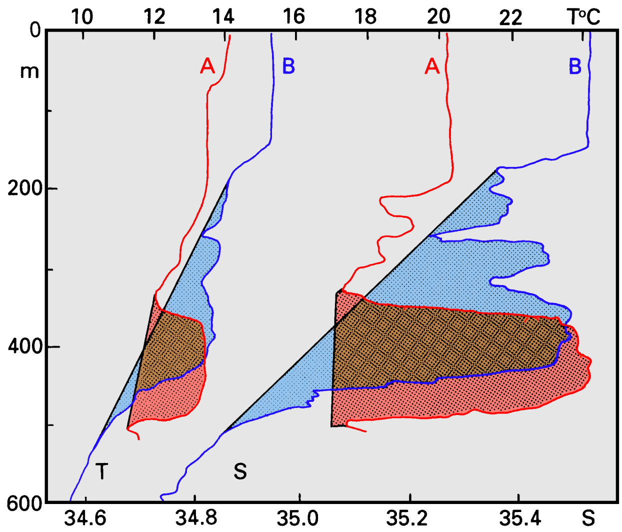

Fig. 17.3. The Bass Strait Water undercurrent in the Tasman Sea as seen in vertical profiles of temperature (T) and salinity (S). Temperature and salinity anomalies produced by the undercurrent are shaded. See Fig. 17.2

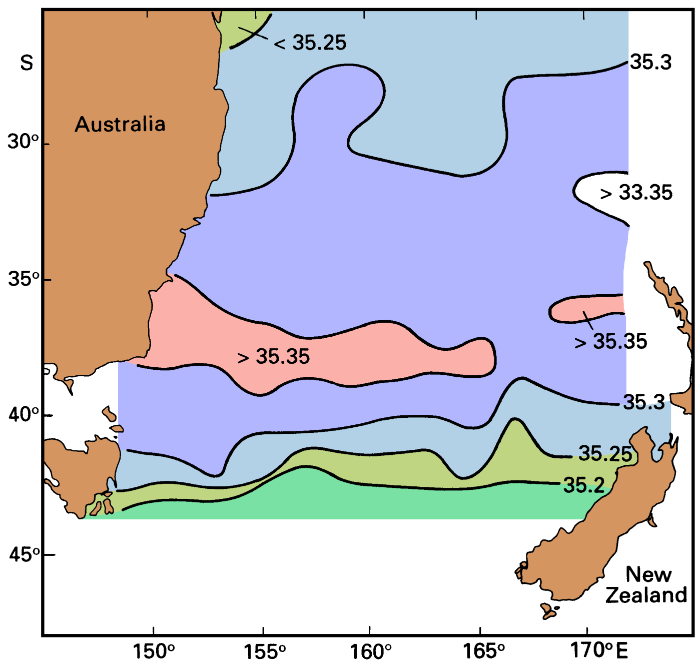

Fig. 17.4. Salinity on the 14.5°C isothermal surface in the Tasman Sea. The depth of this surface varies between less than 100 m in the south to 350 m in the north. High salinity along 37°S indicates Bass Strait Water influence. From Villanoy and Tomczak (1991).

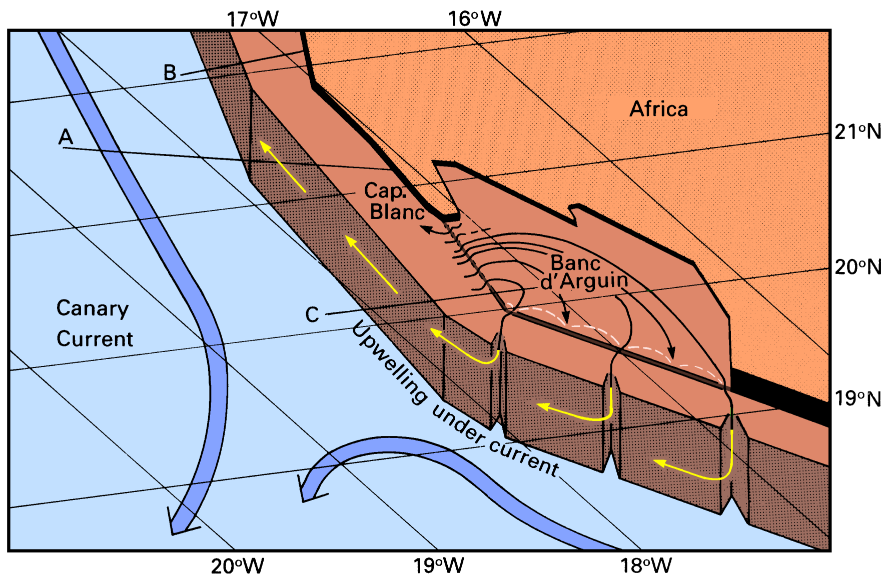

Fig. 17.5. Sketch of the circulation on Arguin Bank. Lines *A*, *B*, and *C* indicate the locations of the sections shown in Figs 17.6 and 17.8. Adapted from Peters (1976).

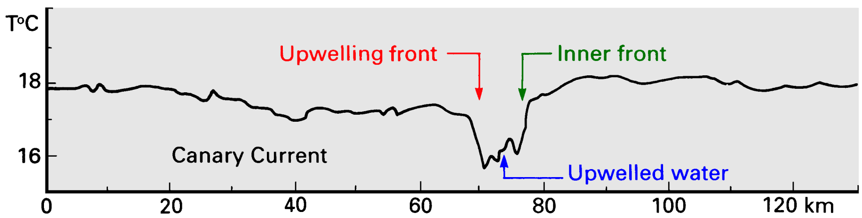

Fig. 17.6 Sea surface temperature along 20°N (section *C* in Fig. 17.5) from the Canary Current to Arguin Bank. The upwelling front, produced by the surfacing of the thermocline, separates the coastal upwelling zone to the right from the region of cold water advection in the Canary Current. The inner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

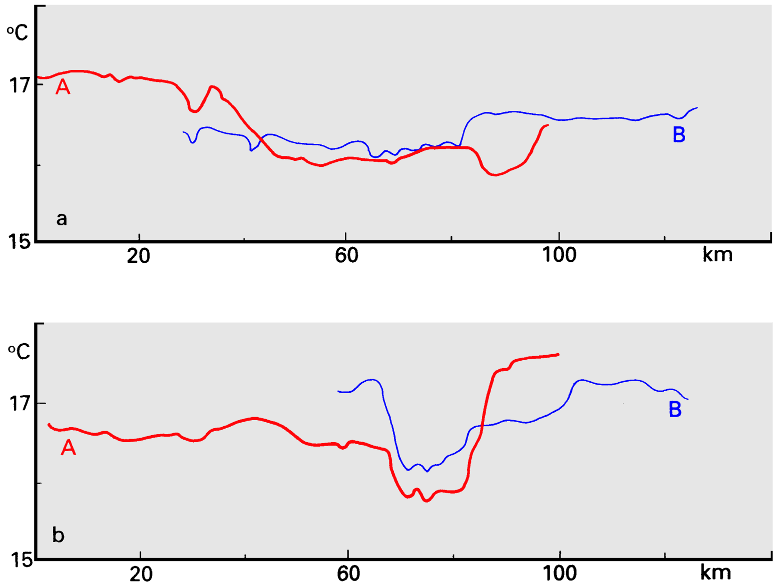

Fig. 17.8. Sea surface temperature along transects *A* and *B* of Fig. 17.5. (a) During weak upwelling, (b) during an event of strong upwelling. Note the absence of fronts and coastal warm water when the upwelling is weak. Note also how the inner front is eroded by mixing as the coastal water moves northward 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

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.

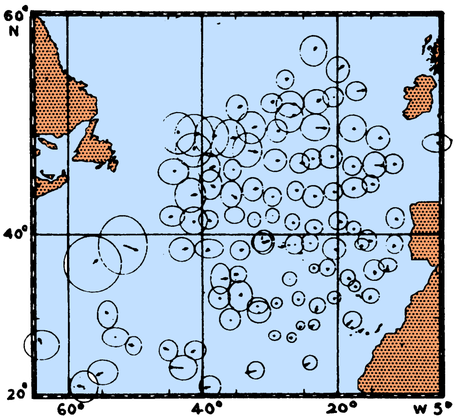

Fig. 17.10. Regional distribution of geostrophic turbulence in the North Atlantic Ocean. The arrows show the mean surface current, the axes of the ellipses give the mean north-south and east-west eddy velocities. Mean eddy velocities increase threefold from Ireland to Newfoundland. In the centre of the 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.