Chapter 5

Water mass formation, subduction, and the oceanic heat budget

In the first four chapters we developed the concept of Ekman pumping, Rossby wave propagation, and the Sverdrup circulation as the steady-state balance between these two processes. We then showed how the *depth-integrated* flow in most ocean regions is well described by this concept. However, this is evidently not the full story of the oceanic circulation. The oceans carry heat from the tropics to polar latitudes, and they carry cold water from the poles towards the equator. The details of these transport processes, which are restricted to certain depth ranges, are hidden in the Sverdrup circulation. They have to be resolved if we want to understand the ocean's role in climate variability and climate change.

One thing we do know is that the ocean carries about as much heat towards the poles as the atmosphere does; but since its time scales are so much larger the ocean has a large capacity to act as a damping mechanism for rapid fluctuations in our climate. Conversely, much of the long-term variability of the climate may be related to the ocean as it slowly releases heat stored from earlier rapid climate changes. Topographic details play a large part in determining patterns of ocean heat transport. This is the point of interaction between regional oceanography and geophysical fluid dynamics. It is essential for theoretical studies to know to what degree regional oceanic features have to be part of the modelling process.

Generally speaking, most of the heat and salt exchange with the atmosphere occurs at the ocean surface in a layer which during most of the year is less than 150 m deep. Once a water parcel is removed from the surface layer its temperature and salinity do not change until it rises back up to the surface again, usually many years later. Evidence for this can be found in the temperature and salinity maps of Figures 2.5c and 2.5f for 2000 m depth. The temperature everywhere is less than 4.2°C, even in the tropics; such cold waters can only have acquired their low temperatures in the polar regions and moved into the tropics without much mixing with the warmer waters above. Since independent evidence obtained, for example, from ^{14}C dating sets the residence times of these deeper waters at hundreds of years, this lack of mixing is remarkable. Yet mixing is not entirely absent: Note that the highest temperatures at 2000 m occur just west of the Strait of Gibraltar, where they coincide with very high salinities. The pattern suggests that a "water mass" of relatively warm, salty water enters the Atlantic Ocean from the Mediterranean Sea and moves westward across the ocean, mixing as it goes.

This example illustrates a common feature found throughout the ocean. Water masses with well-defined temperature and salinity characteristics are created by surface processes in specific locations, which then sink and mix slowly with other water masses as they move along. Since these movements are so slow, it is usually unrealistic (and because of the presence of eddy motion impractical in any case) to measure them directly. It is easier to deduce these movements and the strength of mixing from the distribution of the water properties themselves.

The analysis of water mass movements and mixing can assist in understanding the deep oceanic circulation where the estimation of steric height relative to an assumed depth of no motion finds its limits. Steric height estimation provides valuable insights into the workings of the vigorously moving top kilometer of the ocean where accurate knowledge of the depth of no motion is not critical. It cannot resolve the slow flows at great depth. The best hope to achieve a complete description of the circulation in the ocean at all depths is to

Common names of known water masses usually relate to their major area of residence. Unfortunately, this can give rise to ambiguity since the same name may be used for a well defined water mass or simply for water found in a certain region. To avoid this confusion we adopt the convention that water masses are always identified by capitals. For example, "Bottom Water" can stand for Antarctic, Arctic, or other Bottom Water but always refers to a water mass, while water found at the bottom of an oceanic region may be referred to as "bottom water" without implying that it is a known and well defined water mass. Likewise, we use the term "intermediate water" occasionally for intrusions of water at intermediate depth; in contrast, "Intermediate Water" is used to indicate well-defined water masses.

It is important to note that exclusive occupation of an oceanic region by a single water mass occurs only in the formation regions. As the water masses spread across the ocean they mix, and several water masses are usually present at an oceanic location. However, water masses occupy a measurable volume, which is the sum of the volumes occupied by all its elements regardless of their present whereabouts. It is possible to determine the percentage contribution of all water masses to a given water sample, because the water mass elements retain their properties, in particular their potential temperature and their salinity, when leaving the formation region. Water masses can therefore be identified by plotting temperature against salinity in a so-called T-S diagram. An elementary description of T-S diagrams and their use can be found, for example, in Dietrich *et al.* (1980). Figure 5.1 is an example of a T-S diagram from a tropical ocean region and shows how observational data can be used to identify water masses from their T-S combinations. It is seen that the properties of Central Water in the Coral Sea correspond closely to those in its formation region, indicating that little mixing with other water masses occurred along its way. In contrast, the intermediate and deep water masses are not present with their original T-S values; their properties are modified by mixing with water above and below, and their presence is indicated by salinity or temperature extrema in the vicinity of the T-S combinations found in their formation regions.

From Figure 5.1 it is seen that T-S relationships alone are insufficient to describe a water mass. Particularly in the upper ocean water masses undergo property changes in response to atmospheric conditions, as indicated in Figure 5.1 by the increase in the standard deviation as the surface is approached. For a full description of a water mass it is necessary to include information about the degree of spatial and long-term variability during its formation, as expressed through its standard deviation. Some water masses, such as the Intermediate Water in Figure 5.1, require only a single T-S combination (T-S point) and a standard deviation; others, such as the Central Water, require a set of T-S combinations, or a T-S relationship, together with a standard deviation envelope. Since data sets to determine standard deviations require observations over several years and are not always available, most books identify water masses by one or more T-S points without standard deviations. Points in the T-S diagram are called *water types*, and water mass definition points are known as *source water types*. In reality, very little - if any - of the water belonging to a water mass has exactly the T-S properties of the corresponding source water types. But most T-S values are very close, within the (often unknown) standard deviation. Numerical T-S values for water masses given in later chapters have to be understood in this way. In general, the standard deviation is very small for abyssal water masses but increases rapidly as the upper layers are approached.

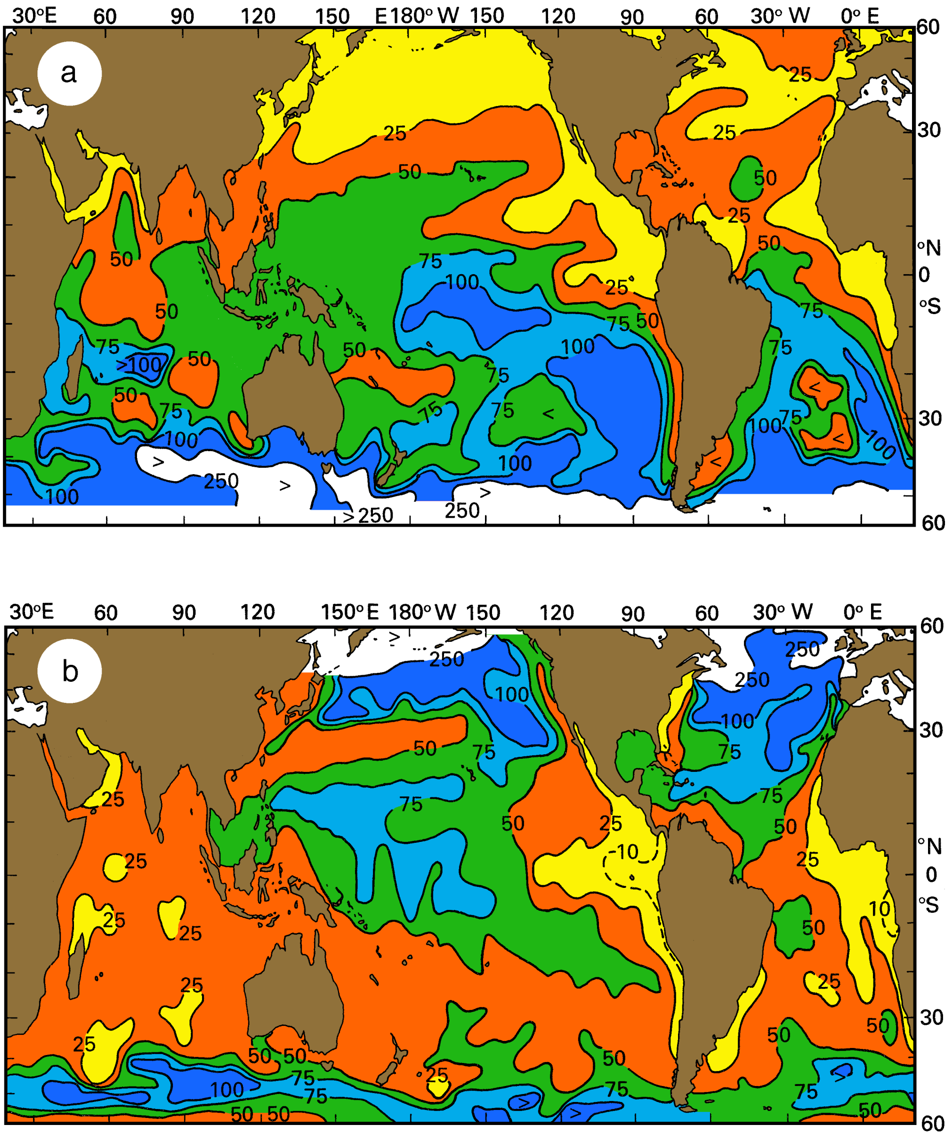

Fig. 5.2. Mean depth of the surface isothermal layer (m). (a) August - October, (b) February - April. Contouring levels are 10 m (dashed line), 25 m, 50 m, 75 m, 100 m, 250 m. Adapted from Sprintall and Tomczak (1990).

Subduction

What maintains the permanent thermocline and prevents its erosion from mixing with the waters below and above? The principal factors are the combination of water mass forma-

central north Atlantic Ocean (now known as North Atlantic Central Water) and water from the shelf area to the west. It is now used to identify thermocline water masses in all three oceans.

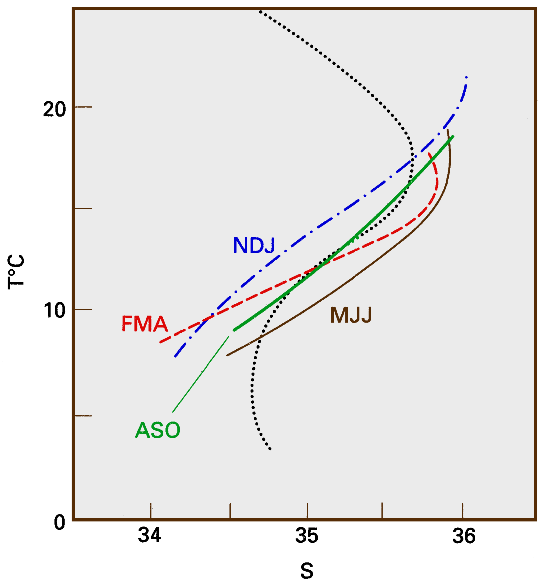

Fig. 5.4. A comparison of T-S diagrams across the Subtropical Convergence (STC), along 102.5°E between 30°S and 45°S, and across the permanent thermocline at 20°S. T-S diagrams across the STC are shown for August - October (ASO), November - January (NDJ), February - April (FMA), May - July (MJJ). The remaining curve is the vertical T-S diagram at 20°S. Adapted from Sprintall and Tomczak (1993).

Studies of T-S diagrams and of property distributions on isopycnal surfaces led to the conclusion that mixing across isopycnal surfaces is generally much weaker than mixing on isopycnal surfaces. The principle does not apply universally; western boundary currents, the Equatorial Undercurrent, and frontal regions are among the regions where mixing across density surfaces, or *diapycnal mixing*, contributes significantly to the exchange of properties. But in general, and over vast ocean regions, it is safe to neglect diapycnal mixing as a first guess for the oceanic circulation.

Figure 5.5 summarizes the discussion of thermoclines and subduction. Intermediate Water, which spreads just below the permanent thermocline, is also produced by subduction. Although the driving agent is not Ekman pumping but mixing and convection in the region between two strong currents the mechanism is the same, movement along isopycnal surfaces towards the equator.

The barrier layer

Traditionally, it has been assumed that the depth over which the temperature is uniform can be used as an indication of the depth of the mixed layer, i.e. the layer affected by surface mixing processes. This assumption probably developed more out of necessity than physical argument, since many upper ocean observations lack information on salinity and some way had to be found to determine the depth of the surface mixed layer from temperature information alone. The better data base of today allows us to check that assumption, thereby gaining a better understanding of the processes of water mass formation. Figure 5.6

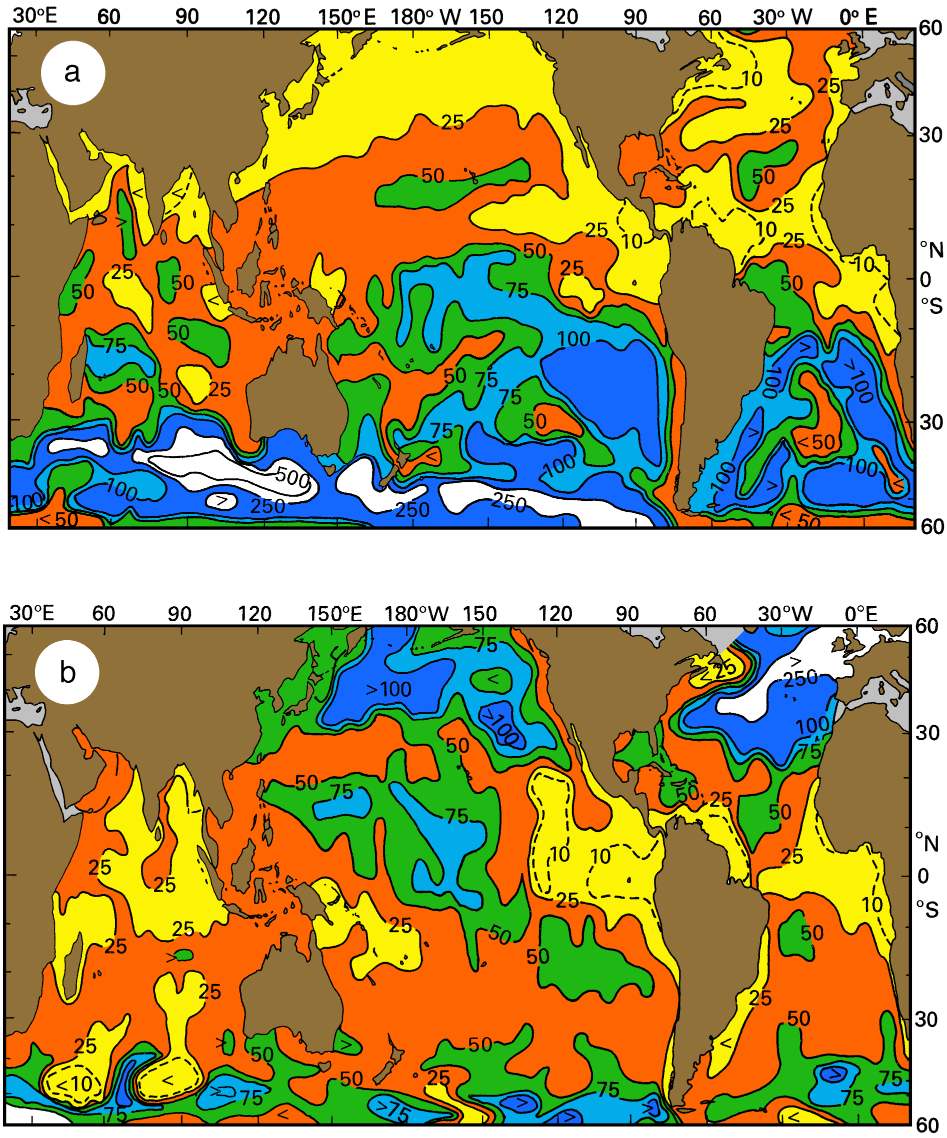

Fig. 5.6. Mean depth of the surface isopycnal layer (m). (a) August - October, (b) February - April. Contouring levels are 10 m (dashed), 25 m, 50 m, 75 m, 100 m, 250 m, 500 m. Adapted from Sprintall and Tomczak (1990).

The layer between the halocline and the thermocline is now referred to as the *barrier layer*, because of its effect on the mixed layer heat budget. The mixed layer receives large amounts of heat from solar radiation. In a steady state situation this source of heat has to be balanced by one or more sinks. In the absence of a barrier layer, i.e. where the surface mixed layer

There is therefore no heat flux through the bottom of the mixed layer, and other sinks have to come into play to prevent a permanent rise of mixed layer temperature. The dynamics of the three regions which show a barrier layer through most of the year, the western Pacific, the equatorial Atlantic, and the Bay of Bengal in the Indian Ocean, differ distinctly from each other; they will be discussed in more detail in the chapters which deal with the individual oceans.

Another region of positive difference in isothermal vs isohaline layer depth is in the polar latitudes where a thermocline does not exist: The surface water, constantly cooled by the atmosphere, is so dense that it sinks virtually to the bottom of the ocean (the refinements of this process will be discussed in Chapters 6 and 7), leaving the water column isothermal to great depth. The salinity, on the other hand, is seasonally influenced by freshwater release from the ice shelf and from icebergs. The resulting reduction in density can inhibit the sinking of water, and a halocline can develop. This process, seasonal development of a pycnocline as a result of freshwater flux from the ice regions, is indicated by the large differences in Figure 5.7 south of 50°S and during February to April near Newfoundland and Labrador and south of Alaska and the Aleutian Islands.

The third region of non-zero differences in Figure 5.7 is located in the subtropics and displays negative values, indicating that the first density change below the surface is found at a depth *greater* than the isothermal layer thickness. This can only occur if the density change produced by the change in temperature across the thermocline is compensated by an appropriate salinity change. To explain this feature we note from Figure 2.5 that in the subtropics where the water of the permanent thermocline is subducted, sea surface temperature and salinity both decrease rapidly towards the poles. The T-S diagram which describes the meridional variation of temperature and salinity across the subduction zone thus nearly follows an isopycnal (compare Figure 5.3). As a consequence, the isothermal layer thickness, say z_1 at station C, is evaluated correctly by the 0.5°C criterion of Figure 5.2. The equivalent density criterion, however, is not exceeded until the greater depth z_2 is reached, because of the compensating salinity effect. It is seen that the difference in layer thickness shown in Figure 5.7 for the subtropics does not, as in the case of the tropics, result from different isothermal and isohaline layer thickness. We might even regard it as an artifact produced by the subduction process. On the other hand, the discussion in Chapters 9, 12, and 15 will show that the negative differences in the subtropics are a reliable indicator for subduction of thermocline water.

The picture that emerges particularly in the subtropical gyres is a set of independent flows on isopycnal surfaces which, when depth-integrated, collectively satisfy the Sverdrup relationship. In the following chapters, we shall use the Sverdrup circulation as a guide for the discussion and fill in the depth-dependence by looking at water mass properties where necessary. One instance where this need will arise is whenever we want to evaluate oceanic transports of heat or salt, because both involve integrals over products of velocity and temperature or salinity which require knowledge of the distribution of currents with depth. To see this, consider the example of warm water moving poleward with a velocity of 0.4 m s^{-1} in the upper 1000 m and cold water moving equatorward with 0.1 m s^{-1} between 1000 m and 5000 m. The net transport of heat is obviously poleward, but the net mass transport is zero. Clearly, the vertically integrated flow patterns of Chapter 4 are inadequate for estimating heat or salt transports.