CHAPTER 8

Observed Mean State of the Oceans

As we have seen before the oceans are a very important component of the climatic system. Indeed, with their high heat capacity the oceans store large amounts of solar energy that can be released later in the form of sensible and latent heat into the atmosphere. The oceans are also an important vehicle to transport energy from low to high latitudes, thereby reducing the north–south gradient of temperature. Through these processes the oceans play a crucial role as moderators of the earth's climate. Furthermore, they are the main source of the water that falls as precipitation over the continents, supplying the runoff for the major rivers in the world. Through air–sea interaction the oceans are also important in the formation and modification of certain types of air masses.

The major surface currents in the world ocean are shown schematically in Fig. 8.1. They are thought to be instrumental in making the climate more amenable in certain regions of the globe which otherwise might be inhospitable. For example, the relatively mild climate of northwestern Europe compared with the climate of other regions in the same latitude belt may be associated with the advection of warmer water in the Gulf Stream system.

It is clear that the study of the climate of the oceans is not only important *per se* but also crucial for understanding the climate in the atmosphere.

8.1 MEAN TEMPERATURE STRUCTURE OF THE OCEANS

8.1.1 Global distribution of the temperature

The horizontal temperature distribution at the ocean surface is shown in Fig. 8.2(a). As expected, the highest values are found over the tropical regions with maxima over the western equatorial Pacific and the Indian Oceans, whereas the strongest north–south gradients are found in mid to high latitudes, being most pronounced in the Southern Hemisphere.

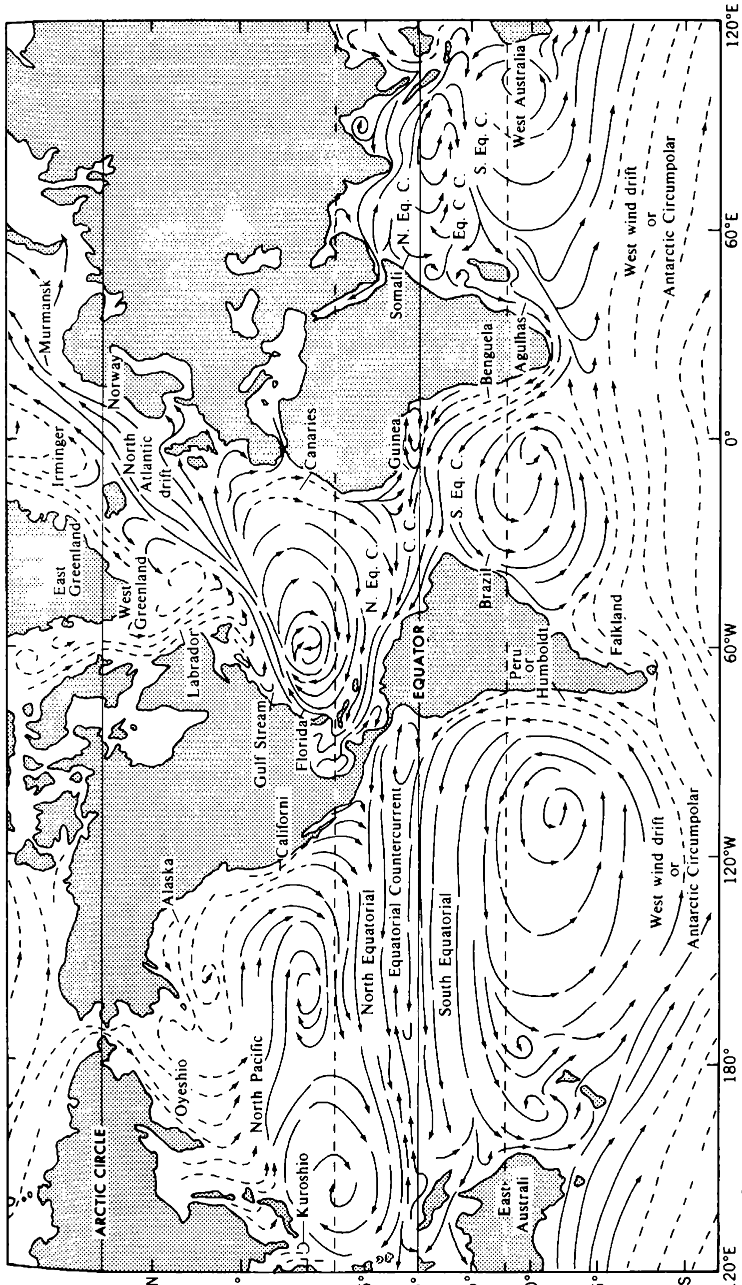


FIGURE 8.1. A map of the major surface currents in the world ocean during northern winter (from Tolmazin, 1985).

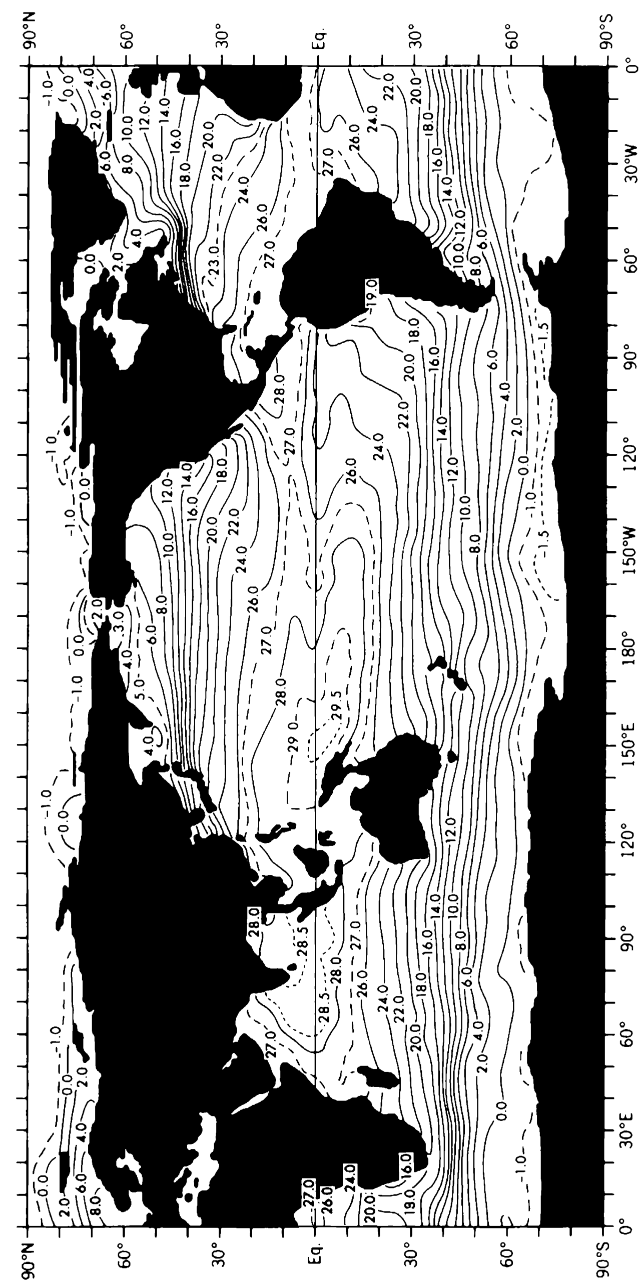
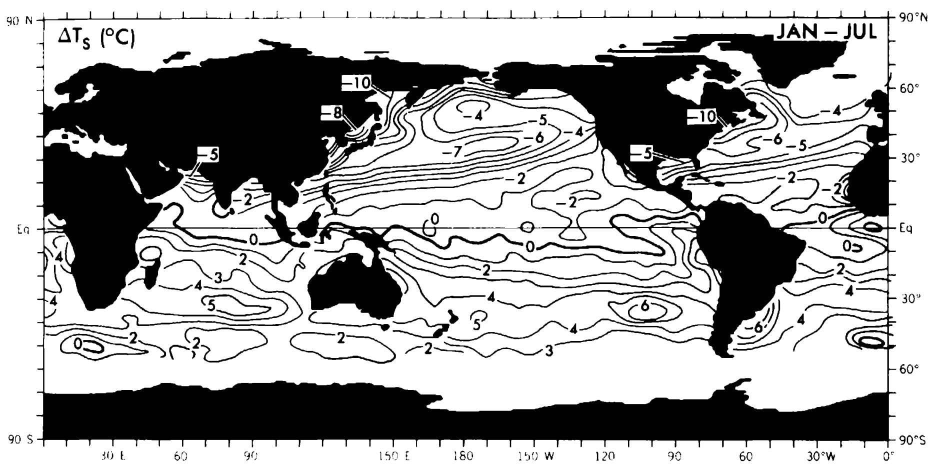


FIGURE 8.2a



b

FIGURE 8.2. Global distributions of the sea surface temperature for annual-mean conditions (a) from Levitus (1982) and the January–July difference of the sea surface temperature (b) in °C.

To first order, the sea surface isotherms are zonal in character but with some distortion due to the influence of the continents. This influence manifests itself in the warm poleward currents (e.g., Gulf Stream, Brazil, and Kuroshio Currents) and the cold equatorward current systems (e.g., the Labrador, Canary Islands, Benguela, California, and Peru currents). Thus comparing both sides of the oceans, we find that in the subtropics the waters on the west sides tend to be warmer than those on the east sides, whereas in high latitudes the opposite tends to occur. This is evident in the North Atlantic Ocean where the Norwegian Sea is anomalously warm due to the poleward transport of heat in the Gulf Stream system.

The low temperatures in the cold equatorward currents on the eastern side of the subtropical gyres are reinforced by upwelling of even colder, nutrient-rich water from great depths. The main regions in which this coastal upwelling occurs are clearly depicted in Fig. 8.2 by the equatorward dipping of the isotherms, namely the Canary current (10–40 °N), the California current (25–40 °N), the Benguela current (10–30 °S), the Peru current (5–45 °S), and (during the southwest summer monsoon) the Somali current regions (0–15 °N). The upwelling in these regions is associated with local wind regimes and is most intense when there is a strong wind blowing equatorward (poleward in case of the Somali current) parallel to the coast, leading to a large Ekman transport away from the coast and directed to the right of the wind in the Northern and to the left in the Southern Hemisphere [see Fig. 8.3 and Eq. (3.24a)]. These wind regimes are connected with the position of the large subtropical atmospheric anticyclones which move poleward in summer and equatorward in winter.

To obtain an order of magnitude estimate of the vertical velocity at the bottom of the upwelling layer, we will assume that w_E is uniform in a strip of width L_x along the coast which is assumed to be oriented in the north–south direction (see Fig. 8.3). Then the amount of mass transported across an imaginary vertical "wall" of length L_y at a distance L_x from the coast [see Eq. (3.24a)] is such that