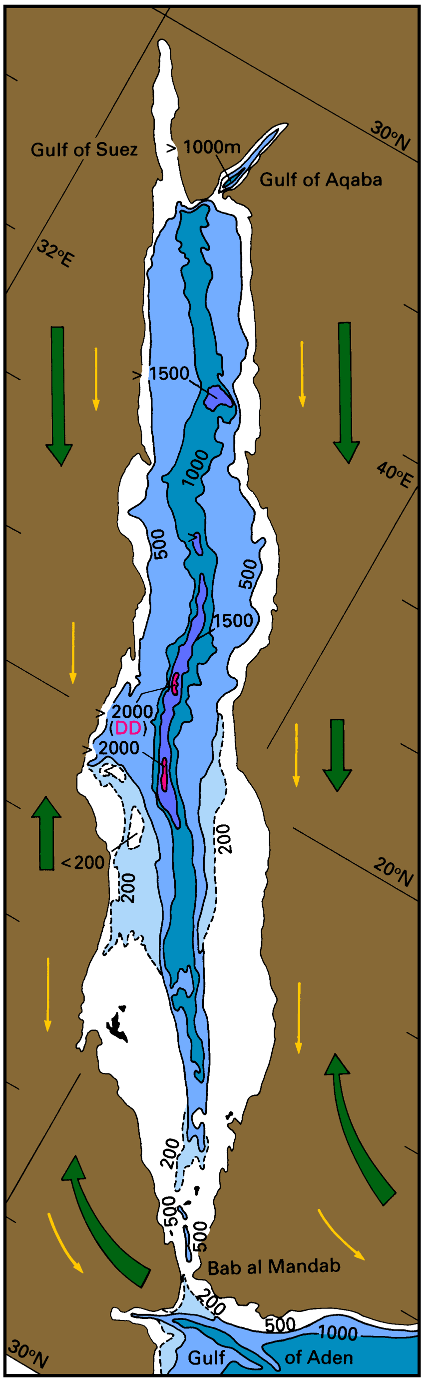

during summer), are a period of active convection. Which of the two regions is responsible for the formation of deep water has been a matter of debate for many years. Recent observations of bomb radiocarbon indicate that both contribute, but in different ways. The winter water from the Gulf of Suez is very dense; it falls down the continental slope and fills the depths of the Red Sea below 1000 m. When the water starts its descent it has a temperature below 18°C and a salinity above 42; but mixing quickly modifies this, and the water soon ends up with the standard Red Sea deep water properties of 21.5°C and 40.6 salinity. Winter water from the northern Red Sea is slightly less dense; it slides down on the appropriate isopycnal surface and spreads below the surface mixed layer. The resulting circulation is indicated in Figure 13.2. A lower circulation cell with slow upward movement, gradual loss of oxygen, and a northward return flow at its upper limit is capped by southward movement of water from the second source. This is reflected in an oxygen minimum below the thermocline, indicating that the oldest water is at the top of the

Fig. 13. Topography of the Red Sea. *DD* marks the Discovery Deep, the depression where the first observations of hot brines were made. Arrows indicate prevalent wind directions for summer (full arrows) and winter (open arrows). Depths are in m.

lower circulation cell. It is worth noting that even at the oxygen minimum, Red Sea Water has a higher oxygen content than the very old Indian Central Water found at the same depth in the Indian Ocean thermocline; consequently, east of Bab el Mandeb Red Sea Water manifests itself through an oxygen maximum. Oxygen values in the 150 - 200 m thick surface layer of the Red Sea correspond to saturation values, which at these high temperatures are relatively low (less than 4 ml/l in summer).

Estimates for the residence time of deep water are just as controversial as identification of its sources and vary from a few years to two centuries. Recent radiocarbon data indicate a residence time slightly less than 40 years. This is the best available estimate at present.

The outflow of Red Sea water into the Arabian Sea is clearly visible in Figure 13.2, most prominently in salinity. It is seen that the dense water flows down the continental slope to a depth of 1500 m and more; but the bulk of the water is modified by mixing soon after passing Bab el Mandeb and spreads between 500 - 1000 m with a temperature of 13 - 14°C and a salinity of 36.5 or less. The magnitude of the outflow at Bab el Mandeb is quite small. Measurements during the summer of 1982 indicated only 0.25 - 0.3 Sv; similar observations over a few months in 1965 showed intermittent outflow of about 0.5 Sv depending on the direction of the wind. Nevertheless, the concentration of salt in the outflow is sufficient to guarantee that Red Sea Water can be traced in the Indian Ocean thermocline even into the southern hemisphere, as we saw in the last chapter.

Seasonal variations in the outflow are related to the monsoons, which determine the surface circulation in the Red Sea in general. Winds over the Red Sea form part of the general monsoon system of the Indian Ocean but are modified by the influence of the land, which establishes a belt of low air pressure from Asia towards northern Africa during summer and a centre of high pressure over northern Africa during winter. The resulting pressure gradients against the region of constant air pressure in the tropics produce northwesterly winds throughout the year north of 20°S. South of that latitude winds are from the northwest during summer but reverse to southeasterly during winter. Throughout the entire region winds are generally stronger in winter than in summer. The resulting flow is northward in winter, south of 20°S under the direct forcing of the wind which supports surface inflow through Bab el Mandeb, further north driven by the sinking of surface water even though winds north of 20°S are predominantly from the northwest. In summer the wind opposes surface inflow, but during most of the time its strength is insufficient to suppress the surface inflow necessary to maintain the water budget. Occasionally water movement at the surface is southward and inflow occurs intermittently at intermediate depths. It is estimated that the water stays in the upper layer circulation for about 6 years before sinking.

A remarkable feature of the Red Sea is the extremely high water temperature and salinity found in various depressions of the sea floor (Figure 13.3). This is the result of geothermal heating through vents in the ocean crust, which brings minerals contained in the crust and in the sediment into solution. The resulting brine is dense enough to remain at the ocean floor even at very high temperatures. Values close to 58°C have been recorded, together with "salinities" in excess of 300. The salinity readings were obtained by diluting brine samples until their salinity came into the range of normal CTD instruments. Such determinations are invariably incorrect since the universal rule that the relative composition of sea salt remains the same throughout the world ocean does not hold in brines brought up from fissures in the earth's crust - the content in metal ions is much higher than in the

because the residence time is much shorter. Since Indian Central Water near the Strait of Hormuz is older and its oxygen content lower than at Bab el Mandeb, the oxygen maximum produced by the outflowing water is even more marked. Persian Gulf Water also tends to have somewhat lower density than Red Sea Water (on account of its higher temperature) and therefore tends to stay above the main thermocline rather than penetrating it. Nevertheless, at some distance from the Strait of Hormuz it is often difficult to separate traces of Red Sea and Persian Gulf Water, and the two water masses are often regarded as one.

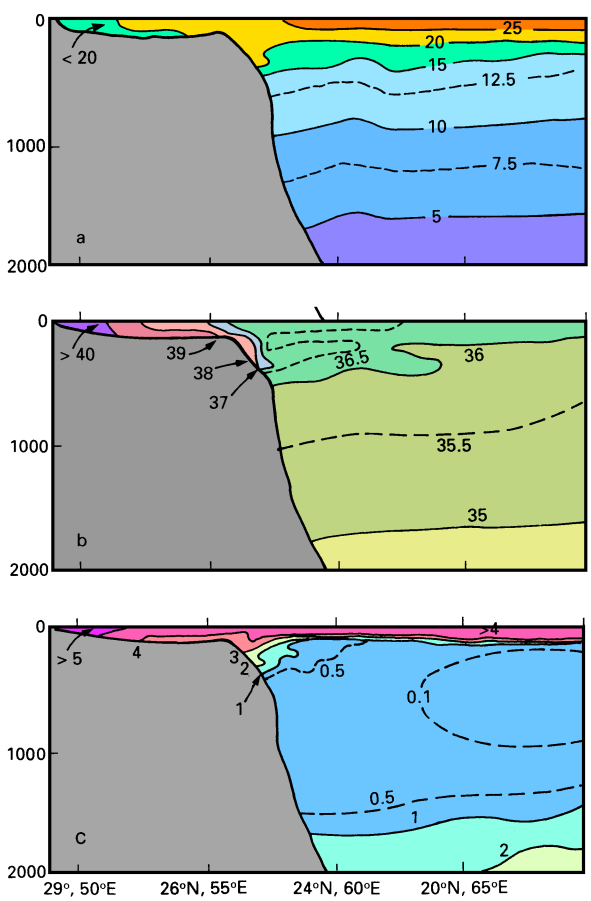

Fig. 13.4. Hydrographic section along the axis of the Persian Gulf, winter conditions

(a) Potential temperature (°C)

(b) salinity,

(c) oxygen (ml/l).

The Australasian Mediterranean Sea and the Indonesian throughflow

Of all the regional seas of the world ocean, the Australasian Mediterranean Sea displays without doubt the most complicated topography. It consists of a series of very deep basins with very limited interconnections, each basin being characterized by its own variety of bottom water. The exact number of deep basins found within its borders is difficult to define; most nautical charts usually recognize at least eight basins under their own name

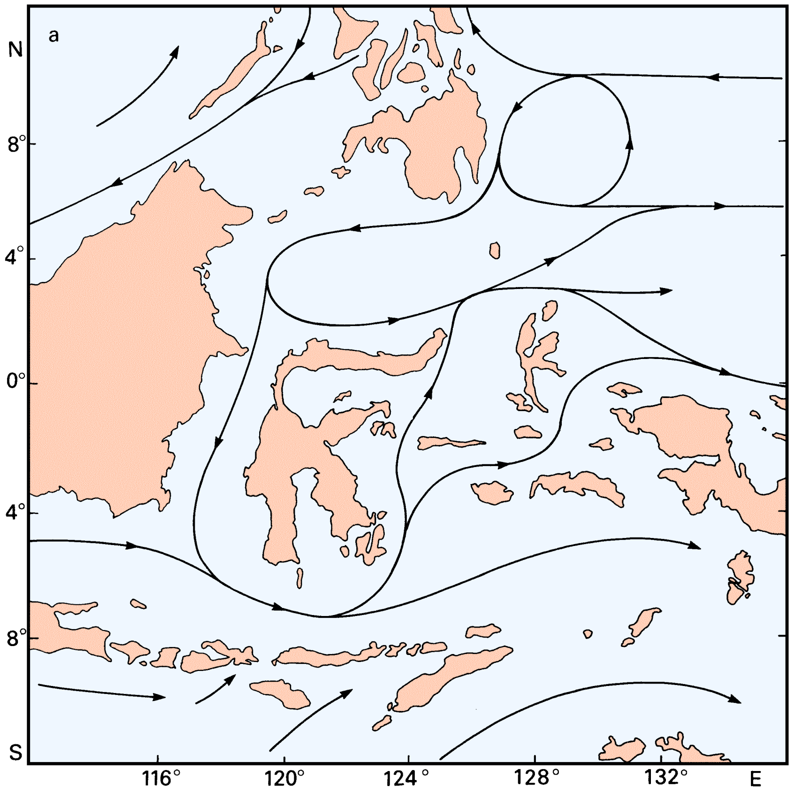

Fig. 13.6. Surface currents in the Australasian Mediterranean Sea. (a) In February (north monsoon, minimum throughflow), (b, page 223) in August (south monsoon, maximum throughflow).

Java Seas; it is shallow in the west but over 2000 m deep in the east where it is connected without obstruction to the Sulawesi Sea in the north.

The climate of the Australasian Mediterranean Sea is characterized by monsoonal winds and high rainfall. Winds blow from the south, curving across the equator with a westward component in the south and an eastward component in the north, during May - September and in nearly exactly the opposite direction during November - March (Figure 1.2). Rain occurs at all times of the year and exceeds 400 cm in the annual mean near the junction of the Intertropical and South Pacific Convergences. As in the western equatorial Pacific Ocean, it is released from strong localized convection cells that are only a few kilometers in diameter, reach high into the upper atmosphere, and are surrounded by cloud-free regions of sinking air. As a consequence, solar heat input over the Australasian Mediterranean Sea is high (Figure 1.5) despite relatively high cloud-cover (Figure 8.5). Evaporative heat loss is high on account of high sea surface temperatures; but on balance the ocean receives more heat than is lost to the atmosphere. There is also more freshwater gain than evaporation, and the *P-E* balance (Figure 1.7) is strongly positive.

Direct current measurements in the Australasian Mediterranean Sea are available for only a few locations and are often of short duration. The net transport is believed to be westward at all times, from the Pacific to the Indian Ocean. It occurs as a western boundary current (i.e. with highest velocities along Mindanao and Kalimantan) and is made up of two components, the surface current driven by the monsoons and the deeper reaching interoceanic throughflow. Although the wind-driven flow opposes the throughflow during May - September and follows it during November - March, the total westward transport reaches a maximum in August and goes through a minimum in February. The reason for this apparent contradiction is seen when the circulation of the Australasian Mediterranean Sea is considered in conjunction with that of the Indian and Pacific Oceans. During November - March the Equatorial Countercurrent of the Indian Ocean is fully developed, supplying water to the region where the outflow from the Australasian Mediterranean Sea occurs and raising the sea level. As a result the pressure gradient from the Pacific into the Indian Ocean is small, and the throughflow is at its minimum. During May - September the Equatorial Countercurrent in the Indian Ocean is replaced by the South Equatorial Current which expands northward under the south monsoon, drawing water away from the eastern Indian Ocean. This lowers the sea level in comparison to the Pacific Ocean and produces maximum throughflow.

First estimates for the throughflow based on geostrophic calculations from a very limited data base gave annual mean values of 2 Sv or less. More recent studies indicate that the throughflow maximum should be in the range 12 - 20 Sv, while the minimum is estimated at 2 - 5 Sv. None of these estimates are derived from direct current observations. Some are the result of numerical models of the world ocean circulation with fairly coarse resolution and are derived as balances between total Pacific and Indian Ocean transports. Others are based on Sverdrup dynamics and calculate the transport from hydrographic observations. The current is concentrated in the upper layers and decays markedly with depth, with little transport occurring below 500 m; this makes the assumption of a depth of no motion below 1000 m reasonably acceptable. Recent observations from current meters moored in the west Flores Sea and in Lombok Strait for the period January 1985 - March 1986 showed consistent flow from the Pacific to the Indian Ocean of 0.9 m s^{-1} and more during August. During October - March the flow was interrupted by frequent reversals of 10 - 20 day duration, but when it set southwestward it still attained 0.6 m s^{-1}. These and similar observations indicate that the fairly large transports inferred from numerical models and Sverdrup calculations might well be realistic. Annual mean transport through Lombok Strait works out at about 1.7 Sv, with virtually no flow during November - January, 1 Sv in February - June, and maximum transport of 4 Sv in August (Murray and Arief, 1988). Given that Lombok Strait is one of the minor passages, these figures suggest fairly large total throughflow. A recent attempt to estimate the flow through all channels into the Indian Ocean from geostrophic calculations gave a total transport of 24 Sv, again pointing towards a large throughflow .

Surface currents can reverse seasonally despite continuous net westward throughflow; this is known for Lifamatola Strait, the passage from the Molucca Sea to the Buru Basin which leads into the North Banda Sea, where the current sets northward during August but southward during February. A sketch of the surface circulation constructed to the best of available knowledge is given in Figure 13.6.

Currents below 500 m depth are even less well surveyed than upper layer currents. Estimates from geostrophic calculations indicate concentration of the flow in the upper few

Figure 13.7 gives a hydrographic section from the Pacific Ocean north of Halmahera through the Molucca, Banda and Sawu Seas into the Indian Ocean south of Timor, obtained during one of the rare expeditions into the region. Unfortunately no salinity data were obtained on the Pacific side, so the character of the Australasian Mediterranean Sea as a dilution basin between the two oceans does not come out as clearly as it could. However, salinity in the Banda Basin is seen to vary by less than 0.06 over the entire water column, and lowest salinities are found in the Sawu Sea. The effect of freshwater input at the surface comes out more clearly in a comparison of T-S diagrams along the path of the throughflow (Figure 13.8) which shows that the vertical salinity gradient of the Pacific Central Water virtually disappears during the passage through the Indonesian seas. Mixing with Indian Central Water restores the gradient and brings the T-S diagram nearly back to its original form. This water mass conversion affects the upper 1000 m of the water column, despite the fact that adding freshwater at the surface increases the stability. Turbulent mixing must therefore occur over a large depth range and must be able to overcome the strong density gradient. Indications for strong mixing at great depth can be seen at sills such as Lifamatola Strait. Figure 13.7 shows that the water that fills the Buru and Banda basins is drawn from about 1500 m depth some 500 m above the sill depth, indicating that strong bottom currents of probably tidal character are able to mix the water over some hundreds of meters. Support for this conclusion comes from the oxygen data (Figure 13.9) which indicate an oxygen maximum above the sill as a result of downward mixing of water from above.

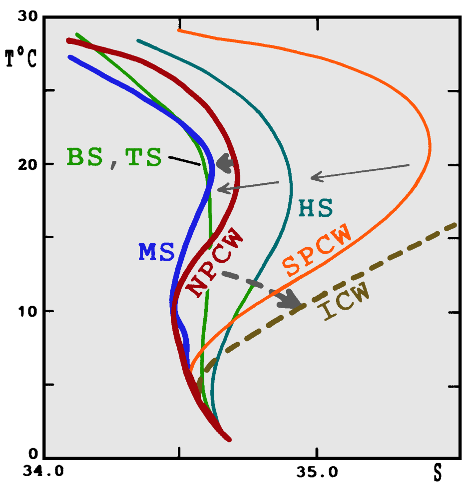

Fig. 13.8. Temperature-salinity diagrams along the path of the Indonesian through-flow, showing the transformation of Pacific Central into Australasian Mediterranean Water (demonstrating the character of the Australasian Mediterranean Sea as a dilution basin) and subsequently into Indian Central Water. South Pacific Central Water (SPCW) passes through the Halmahera Sea (HS) into the South Banda (BS) and Timor Seas (TS). North Pacific Central Water (NPCW) passes through Makassar Strait (MS) to the Timor Sea (TS). Both are then converted into Indian Central Water (ICW). Adapted from Ffield and Gordon (1992).

The mixing process in the upper 1000 m of the Australasian Mediterranean Sea is unique since it achieves nearly complete homogenization of the salinity field without destroying

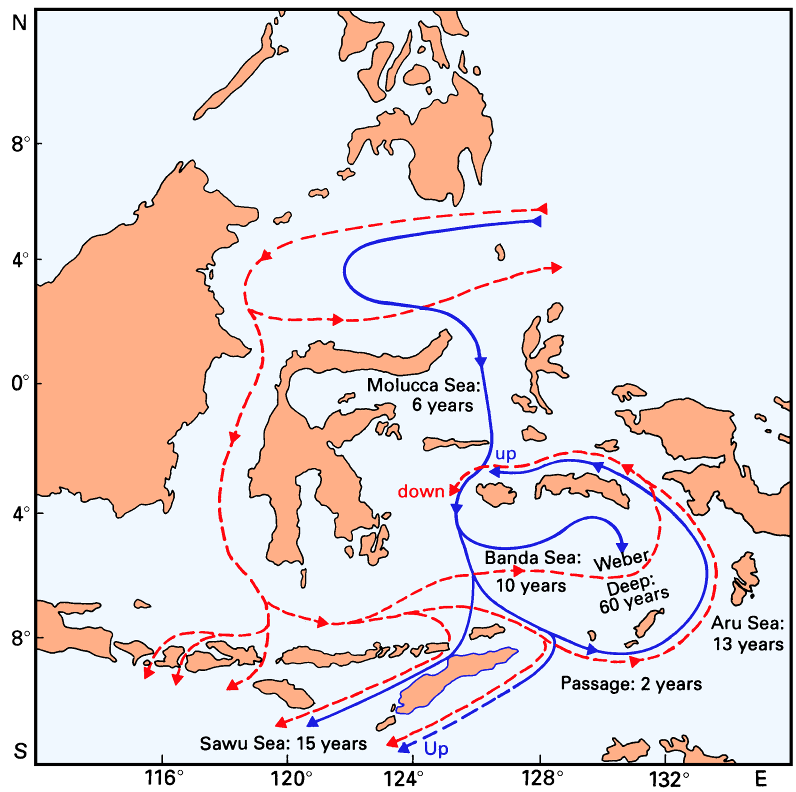

Fig. 13.10. Sketch of intermediate (broken lines) and deep water movement (full lines). Estimated transit times of bottom water are also indicated. Adapted from van Bennekom (1988).