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ABSTRACT

The physical oceanography of the Tasman and Coral Seas is reviewed with an emphasis on the deep currents. There are many uncertainties in the deep circulation pattern. The available data are used to develop an idealised circulation to estimate the likely path taken by water flowing from a depth of 5000 m in the Tasman Sea. The model suggests that the water would finally reach the surface layers south of the Antarctic Convergence with a median delay of 600 years.

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1. INTRODUCTION

The oceans form a vast network of water masses with a complex worldwide circulation pattern. Consequently, pollution added to one coastline or body of water will ultimately spread throughout all of the oceans. In spite of the wealth of knowledge of the mean surface currents, little is known of the worldwide circulation pattern. Very little also is known of the intermediate scale circulation or of the deep currents.

This report reviews the knowledge of deep ocean currents of the south-west Pacific in the context of the worldwide circulation pattern. In Section 2, the basic physical oceanography necessary to understand currents and mixing processes is discussed. In Section 3, the topography and circulation of the Tasman and Coral Seas are reviewed and the circulation is then related to the worldwide circulation patterns in Section 4. Finally in Section 5, the accrued information is used to estimate the likely transport of polluted water into or out of the deep Tasman and Coral Seas.

This study was prompted by the suggestions in some countries for radioactive waste disposal into the deep oceans. Ocean current transport and dispersion is only one aspect of the problem. Other aspects include the effects of biological transport and concentration, sediment concentration, methods of release and the complex near-shore currents, but these will not be discussed in this report.

2. PHYSICAL OCEANOGRAPHY

2.1 Seawater

The oceans are horizontally stratified with the densest water at the bottom and the least dense water at the surface. Although the density range is small, it is sufficient to restrict greatly the vertical movement of ocean water. The density of seawater depends on three variables - the temperature, the salinity and the pressure.

Salinity is essentially the mass of the total dissolved salts in the seawater; generally it is between 33 and 37 g kg^{-1}. The temperature of ocean water ranges from -2 to +30°C but the median temperature is only 2.1°C [Montgomery 1958].

Seawater is slightly compressible (Table 1) and this causes adiabatic heating as a random sample of seawater moves to greater depths. The temperature of seawater is expressed either as the *in situ* temperature or as a potential temperature; the latter is the temperature attained by the sample if it were raised adiabatically to the sea surface. The potential density of the sample is derived in a similar manner.

TABLE 1

DENSITY AS A FUNCTION OF DEPTH FOR SEAWATER OF 0°C *in situ* TEMPERATURE AND 35 g kg^{-1} SALINITY

(After Sverdrup *et al.* 1942)

Depth (m)	0	2000	4000	6000
Density (kg m^{-3})	1028.1	1037.5	1046.4	1055.0

The world oceans are really very uniform; 90 per cent of the ocean waters have potential densities between 1026.56 and 1027.92 kg m^{-3} [Montgomery 1958]. Because of this small range, the potential density is often expressed in terms of σ_t:

$$\sigma_t = \rho_{S,T,O} - 1000 ,$$

where $\rho_{S,T,O}$ is the density in kg m^{-3} at the salinity S, the temperature T, and a pressure of one atmosphere.

The surface layers of the oceans, down to a depth of about 100 m, are well mixed by wave action. Below this, in mid and tropical latitudes, the temperature decreases with depth to produce a region of great stability (the thermocline) which extends down to about 1000 m. This stability produces a strong stratification in the surface ocean currents. Below the thermocline, the abyssal region shows only small variations in properties. The thermocline is limited to the upper 1000 m by the slow upward transport of abyssal waters. In the Antarctic region, there is no thermocline and, in some places, the circumpolar current extends from the surface to the ocean bottom at 3000 to 5000 m [Neumann 1968].

The concept of water masses is often used to trace the circulation of ocean water. A water mass is characterised by its salinity, temperature and other properties that are maintained as the water flows from its point of origin. The temperature and salinity are conserved by the mixing of two different water masses. Other properties such as dissolved oxygen, oxygen-18 and phosphate are also used to provide a better identification of a water mass.

2.2 Ocean Currents

All ocean currents derive their energy from the sea-air interface. Most of the surface currents are driven by the prevailing winds, whereas the deep currents flow in response to density differences created at the sea-air interface by evaporation, ice formation and atmospheric cooling. Some deep currents also flow in response to pressure differences created by the wind driven currents.

The deflecting force of the Earth's rotation (the Coriolis force) prevents a current flowing in the direction of the pressure gradient. For a frictionless current to flow horizontally without a change in velocity, the horizontal pressure gradient force must be equal and opposite to the Coriolis force [Neumann 1968]. Such a current is called a geostrophic current. Since the Coriolis force is always normal to the direction of motion, the geostrophic current can only flow at right angles to the pressure gradient and hence maintains the gradient. The pressure gradient and the geostrophic current must be coexistent, *i.e.* one cannot exist without the other.

A standard method for determining deep ocean currents relies on measuring the pressure gradient and then calculating the coexisting geostrophic current. The pressure gradient cannot be determined directly because the depth cannot be measured with sufficient accuracy. However, the density profiles can be determined from measurements of salinity and temperature, and the pressure caused by the weight and overlying water can be calculated. The pressure field must be referenced to a level of no motion because the contribution of a sloping sea surface cannot be measured with sufficient accuracy. The ocean is assumed to be horizontally stratified with the currents flowing in the horizontal plane. The geostrophic current is the flow necessary to produce a Coriolis force that will balance the calculated pressure differences. By measuring the properties in several locations, a contour diagram of the current can be built up.

The geostrophic method of current determination suffers from several problems. Oceanic water movements can accelerate, and frictional and vertical movements can also play a significant role. In water movements that reach to the bottom, friction must be included in the balance of forces. Ideally, the pressure field should be determined simultaneously at all locations, but the usual method is for one ship to make all the measurements hence different measurements may be made several days apart. Such results are subject to periodic or non-periodic disturbances that affect the temperature or salinity. The uncertainty introduced in transforming the relative pressure field into an absolute pressure field using a level of no motion is especially important for deep currents. The deep currents are often so slow that it is hard to identify a suitable level of no motion for use as a reference. Notwithstanding all these difficulties, the geostrophic approximation is the principal source of data on the volume transport of deep currents.

Brown *et al.* [1975] tried to test the geostrophic approximation in the North Atlantic Ocean as part of the Mid-Ocean Dynamics Experiment (MODE) but failed. They succeeded in directly measuring the geostrophic bottom pressures but found that these were related primarily to large scale fluctuations whereas the observed bottom currents were related to intermediate size eddies (about 200 km diameter). Currents derived from the geostrophic approximations have greater validity for large scale ocean circulation than isolated current measurements. Direct measurements of the deep currents are made difficult by the periodic and non-periodic variations in flow.

The equations governing geostrophic flow show that a non-accelerating, frictionless current must move along surfaces of constant potential density (isopycnic surfaces). This requirement also follows if the water circulation pattern is stationary. The surfaces of constant σ_t are good approximations to the constant entropy surfaces and water is assumed to move over the surface without change of energy. The properties on a constant σ_t surface can be analysed for information about the water circulation: this is called isentropic analysis. Movement over the isopycnic surface should proceed without change of temperature or salinity. If changes occur, they must be the result of mixing.

Stommel & Arons [1960a,b] developed an abyssal circulation model based on geostrophic flow in idealised ocean basins. The geostrophically balanced flow is induced by the upwelling of deep water to the upper layers and this results in a flow in the interior of the ocean basins that is directed towards the poles. To satisfy continuity, a narrow frictional or inertial current, which is not geostrophically balanced, was postulated on the western boundary. The source of this deep boundary current is sinking water at high latitudes. The existence of the deep western boundary currents has been observed in all the oceans. In the South Pacific, the current flows along the east coast of New Zealand rather than Australia, because the Tasman Sea is closed off to the north at depths greater than 3000 m [Warren 1973].

The Stommel & Arons model applies to an idealised ocean. The real ocean floor tends to consist of a series of basins or trenches of various depths connected by saddle points or sills. The densest seawater will sink to the bottom of a basin. The bottom water will be heated slowly by turbulent conduction from overlying warmer water and also by a small contribution from the Earth's interior. New cold water entering the basin will flow under the older water giving it a small vertical velocity. Ultimately, the water will rise above a sill and flow into the next basin.

2.3 Mixing Processes

The ocean water is in continual movement with sizes ranging from the major ocean currents to centimetre size eddies. The ocean currents produce eddies of the order 200 km diameter. These eddies produce a cascade of eddies of ever decreasing size with the transfer of kinetic energy towards the smallest eddies. Finally, eddies of a centimetre size dissipate the energy into heat through the viscosity of water [Okubo 1971].

The movement of ocean water acts to spread and bodily transport any patch of tracer or pollutant material. The wandering of the patch is attributed to currents and eddies which are larger than the patch, whereas the spreading is due to eddies which are smaller or comparable in size to the patch. As the patch becomes larger, it is subject to spreading by larger eddies which contain more energy than the smaller eddies.

In oceanic mixing the transport of the tracer is known as 'advection' and the spreading is known as 'diffusion' or 'turbulent diffusion'. In fact, nearly all significant spreading in the oceans is produced by the combination of many small advection processes, whereas molecular diffusion is negligible. The term 'diffusion' tends to be used interchangeably with the term 'mixing' whenever the mixing process is being distinguished from the bulk transport.

Observations of diffusing dye patches in the upper mixed layer of the oceans have shown the horizontal variance of the distribution to be proportional to $(\text{time})^{2.3}$ [Okubo 1971]. This contrasts with the normal Fickian diffusion where the variance is directly proportional to the elapsed time.

Nevertheless, mixing processes in the ocean are usually analysed in terms of a standard diffusion-advection equation, but with a varying eddy-diffusivity, K, defined by

$$K = -J_\theta / \underline{\nabla}\theta \quad ,$$

where J_θ is the averaged flux of property θ, and $\underline{\nabla}\theta$ is the gradient of the property. The eddy-diffusivity usually exceeds the molecular diffusivity by several orders of magnitude even at very early times.

The horizontal eddy-diffusivity in the surface layers typically ranges from 0.03 $m^2\ s^{-1}$ after one hour spreading to 4000 $m^2\ s^{-1}$ after 100 days' spreading [Pritchard *et al.* 1971]. Vertical diffusion, in contrast, is controlled primarily by small motions characteristic of stable stratified water and remains relatively constant. Below the main thermocline, the vertical eddy-diffusivity is typically about $10^{-4}\ m^2\ s^{-1}$. Within the thermocline, values down to $6 \times 10^{-7}\ m^2\ s^{-1}$ have been reported for short-term tracer experiments [Okubo 1971]. This contrasts with estimates of $10^{-4}\ m^2\ s^{-1}$ in the thermocline obtained from analysis of oceanic variables such as temperature. Okubo suggests that the difference occurs because the vertical distribution of the oceanic variables represents a long-term average with horizontal transport processes making a major contribution.

For a constant eddy-diffusivity, the standard deviation, σ, of the contaminant distribution in one dimension would be

$$\sigma = (2Kt)^{\frac{1}{2}} \quad ,$$

where K is the eddy-diffusivity and t is the elapsed time.

A velocity shear between different depths can produce horizontal mixing even without random horizontal water movements. The theory of shear diffusion is reviewed by Okubo [1971] and Pritchard *et al.* [1971].

Very little is known about the mixing processes and turbulence in the deep ocean water. The velocity shear produced by geostrophic currents is too small for instabilities to occur [Okubo 1971]. However, since at abyssal depths the density stratification is weak, shears from internal waves or bottom topography might produce instabilities.

3. TASMAN AND CORAL SEAS

3.1 Topography

The Tasman and Coral Seas are the part of the South-East Pacific Ocean which is bounded by Australia, New Guinea, the Solomon Islands and New Zealand (Figure 1). The division into the Coral Sea and the Tasman Sea is not well defined but the area north of about 25°S is considered to be the Coral Sea. The Coral Sea comes under the influence of the Central Pacific, whereas the Tasman Sea is largely under the influence of the Antarctic Ocean. The general oceanography of the Tasman and Coral Seas has been reviewed by Rotschi & Lemasson [1967].

The complex bottom topography of the region is shown by the contours marked on Figure 1. The Tasman Sea contains the East Australian Basin in the west and the Norfolk Trough in the east. The East Australian Basin has an average depth of 4600 m and a maximum of 5900 m. To the south, the East Australian Basin is limited by the Indo-Antarctic Ridge at a depth of about 3500 m. To the north, the East Australian Basin connects with the Coral Sea Basin over a sill at a depth less than 3000 m.

The Coral Sea comprises three basins - the Solomon, Coral Sea and New Hebrides. The Solomon Basin connects with the Central Pacific across a sill that lies at a depth of between 3000 and 4000 m. The Solomon Basin has several deep trenches, the deepest (9100 m) being the New Britain Deep. The Coral Sea Basin is an abyssal plain lying at a depth of between 4400 and 4800 m. The basin is connected to the Solomon Basin by a sill at a depth of 3800 m. The New Hebrides Basin has a maximum depth of 7700 m and connects to the South Fiji Basin by a sill at a maximum depth of 5900 m.

3.2 Circulation

The ocean waters of the Coral and Tasman Seas show the strong stratification typical of tropical and subtropical oceans. The different water masses spread in quasi-horizontal layers one above the other. Wyrtki [1962] identified seven major water masses in the region; each water mass was characterised by extreme values of salinity and oxygen. The identification is not exhaustive as the water masses can be further subdivided, or different identifying characteristics can be used.

The present discussion will consider the water features and circulations at representative depths. It must be remembered that the circulation is not strictly horizontal, so the depth of a given feature can vary at different locations.

3.2.1 Surface

The surface circulation depends on meteorological conditions and only the East Australian Current and the Trade Wind Drift are permanent features. The East Australian Current is a strong, narrow, southward flow near the edge of the continental shelf between 27°S and 32°S. Further south, it turns to the east or north and produces a series of southward moving eddies which carry the water as far as 40°S [Boland & Hamon 1970, Hamon 1965, Hamon 1968, Highley 1967]. The system is complex and variable (Figure 2). The current is caused by the accumulation in the Coral Sea of water which escapes towards the south. The Trade Wind Drift brings surface water from the Central Pacific into the Coral Sea with a flow rate which depends on the season [Rotschi & Lemasson 1967].

The volume transports of the East Australian Current are quite variable and transports in the range of 12 to 57 x 10^6 m^3 s^{-1} have been measured. The total width of the current is as much as 150 km with the stronger flowing core (>0.5 m s^{-1}) being 40 to 80 km wide. The velocity of the current decreases to half the surface value at a depth of 250 m; below this depth, it gradually decreases to zero. Direct current measurements using neutrally buoyant floats show that the current can reach depths of 2300 m [Boland & Hamon 1970].

At the southern limit, the East Australian Current meets the Antarctic Circumpolar Current and the sinking water forms the Subtropical Convergence that extends from Tasmania to New Zealand [Stanton 1973]. Another important convergence, the Tropical Convergence, is formed north of New Zealand at the southern limit of the Trade Winds, where the wind direction changes from south-east to south-west.

The East Australian Current originates in the Coral Sea at south of 15°S in summer and south of 20°S in winter [Scully-Power 1973]. North of 20°S in winter, water from the south equatorial current enters the region from the east and flows north into the Solomon Sea.

3.2.2 Sub-surface to 500 m

The water at depths of less than 500 m originates in the convergences found in the tropical and sub-tropical oceans. Water masses can travel large distances and still maintain their identity. Ultimately, the identity will be lost when the water type mixes with adjacent layers or when the water reaches the surface at a point of divergence in the surface currents. There are considerable difficulties in trying to identify different water masses and their origins. The circulation is seasonal and might change from one year to another.

The circulation of the upper 500 m has been investigated by water mass core analysis [Wyrtki 1962], by isentropic analysis on various σ_t surfaces [Rochford 1960b, 1968, 1969] and by geostrophic flow analysis [Scully-Power 1973]. Only the main circulation pattern will be described here.

The principal structural feature in the upper 600 m is a salinity maximum that occurs between the surface and 250 m depth. This water originates east of the Coral Sea in the sub-tropical South Pacific Ocean and flows south in the East Australian Current (Figure 3a).

North of 35°S there is an oxygen minimum between 150 m and 500 m depth which is probably formed locally and implies a long residence time. This water probably forms the boundary between the South Pacific Water and the underlying Antarctic Intermediate Water. South of 30°S there is an oxygen maximum at about 300 m, which is characteristic of water originating at the surface south of the Tasman Sea and flowing north. Figure 3b shows the circulation at the upper oxygen minimum and Figure 3c shows the circulation at the oxygen maximum.

3.2.3 Antarctic Intermediate Water

The Antarctic Intermediate Water is formed at the sea surface near the Antarctic Convergence. Because the precipitation exceeds evaporation in this region, the water type is characterised by a low salinity, a high oxygen content and a temperature in the range 4 to 6°C.

Wyrtki [1962] investigated the flow of the Antarctic Intermediate Water by water mass analysis, and Rochford [1960a] and Johnson [1973] have carried out an isentropic analysis on the 27.20 σ_t and 27.10 σ_t surfaces, respectively. In the Tasman Sea the core of the Antarctic Intermediate Water, defined as the salinity minimum, is found at a depth of 1000 m, but it rises to 700 m in the north Coral Sea. An oxygen maximum is found at 50 to 200 m above the salinity minimum. Wyrtki [1962] suggests that the two layers do not coincide because of a greater oxygen loss by mixing with underlying water.

The Antarctic Intermediate Water enters the Tasman Sea mainly from the east between New Zealand and Fiji and this flow continues north into the Coral Sea. There is a flow from the south but it is poorly developed, possibly because of the effect of the southward East Australian Current. A zone of mixing exists in the centre of the Tasman Sea between 30°S and 40°S (Figure 3d). Johnson [1973] found a small clockwise circulation pattern in the central Tasman Sea.

3.2.4 Oxygen minimum at 1200 to 2500 m

The whole layer between 1200 and 2500 m is of a low oxygen content, with a gradual increase in oxygen from north to south. A region of lower oxygen content is found in the East Australian Basin between 30 and 40°S. The salinity-temperature characteristics of the oxygen-minimum water show it to be a mixture of Antarctic Intermediate Water and deep water, both of which flow northwards and have higher oxygen content [Rochford 1960c, Wyrtki 1962]. The oxygen-minimum water flows slowly south and returns water to the Antarctic Ocean (Figure 3e).

The oxygen minimum must be formed in the depths by oxygen consuming processes; this indicates a considerable residence time. The formation of the minimum probably occurs over all of the Central Pacific. Rochford [1960c] suggested that the phosphate rich waters at a depth of 2000 m in the South Tasman Sea were probably of Indian Ocean origin.

3.2.5 Deep and bottom water (deeper than 2500 m)

Between 2500 and 3000 m there is a weak salinity maximum with a potential temperature of about 1.7°C. Wyrtki [1962] suggested that this water is probably the last trace of the North Atlantic Deep Water.

The bottom water of the East Australian Basin is characterised by minimum potential temperatures and by salinities slightly lower than those of the deep water. This water originates during the formation of pack ice around the Antarctic continent in winter.

At depths below 3000 m, the waters of the Coral Sea are higher in temperature and lower in oxygen than waters at the same depths in the Tasman Sea. This is indicative of the sill depth of about 3000 m connecting the Coral and Tasman Seas [Rochford 1960c]. Wyrtki [1961] derived the circulation of the bottom waters from measurements of potential temperature and salinity. The deep water at about 3000 m in the Tasman Sea overflows the two sills into the Coral Sea and flows to the bottom of the Coral Sea Basin. This water becomes slightly heated and diluted by mixing with the overlying water and ascends as more dense water flows into the basin. The ascending water enters the Solomon Basin through a channel at a sill depth of about 3400 m. Part of this water forms the bottom water in the Bougainville-New Britain Trench.

The deep and bottom water of the Central Pacific Basin extends south between the Solomon and New Hebrides Islands and between the New Hebrides and Fiji. In the vicinity of the San Cristobal Trench, it mixes with the higher salinity water escaping from the Solomon and Coral Sea Basins and flows into the New Hebrides Basin. From the New Hebrides Basin, water flows at the 3000 m level into the Norfolk Trough and fills it to its southern end. This circulation pattern is shown in Figure 3f.

The vertical distribution of temperature in a basin below the inlet and outlet sills is determined by the balance between the turbulent conduction of heat from higher levels and the ascending motion of water. Wyrtki [1961] analysed the heat balance in the Coral Sea Basin. Assuming a vertical eddy diffusivity of 10^{-4} m^2 s^{-1}, Wyrtki obtained a vertical velocity of 1.5×10^{-7} m s^{-1} (4.7 m $year^{-1}$) and a rate of inflow of 4.5×10^4 m^3 s^{-1}. This is a very small flow compared with ocean currents;

it could be supplied by a current of only 0.02 m s^{-1} flowing over a sill 20 km wide with a vertical extension of 100 m.

Boland & Hamon [1970] have made direct current measurements at 3000 m in the Tasman Sea using neutrally buoyant floats. They observed currents in the range 2 to 7 cm s^{-1} but there was no clear correlation with the East Australian Current. Laird & Ryan [1969] measured the bottom current in the Tasman Sea at four locations but only for periods of less than 1.2 hours. At one location, the current varied from 1 to 4 cm s^{-1} over 30 minutes. The results of the MODE experiment in the North Atlantic [Brown *et al.*, 1975] suggest that the current measurements in the Tasman Sea would be related to intermediate-scale eddies and not to the large-scale volume transport of deep ocean water.

The SCORPIO Expedition measured hydrographical sections across the Pacific from Australia to South America at 43°S and 28°S [Warren 1973]. In the Tasman Sea, the cold bottom water was found to warm from a potential temperature of θ = 0.62°C at 43°S to θ = 0.80°C at 28°S. Warren estimated that this warming process corresponded to a northward flow rate of 6×10^{-2} cm s^{-1}. However, when he tried to explain the other measurements he was forced to conclude that, 'The deep water of the Northern Tasman Sea is perplexing: neither salinities nor phosphates nor geostrophic velocities fit into a simple interpretation'. He suggested that steady state conditions may not apply; this could raise some doubts about some of the detailed circulation patterns derived from data taken over several years.

4. WORLDWIDE CIRCULATION

4.1 General Circulation

The three major oceans are linked by the Circumpolar Current which flows around Antarctica almost unobstructed by land masses. The Circumpolar Current is driven by the prevailing west winds and transports water from one ocean to the next. The surface speeds are only in the range 0.15 to 0.20 m s^{-1}, but the volume transport is greater than for any other current system. The lack of a thermocline allows the current to reach depths of 3000 to 5000 m and to be significantly affected by the bottom topography. Estimates of the volume transport vary between 114×10^6 and 165×10^6 m^3s^{-1} [Gordon 1975, Neumann 1968].

There is a continuous exchange of water between the Circumpolar Current and the major oceans (Figure 4). The Antarctic Surface Water is mixed with Sub-antarctic Water at the Antarctic Convergence and sinks to form the

Antarctic Intermediate Water that flows north into all the oceans at a depth of about 1000 m. The Antarctic Bottom Water, which is formed mainly in the Weddell Sea during freezing of the pack ice, has a very high density and sinks down the continental shelf and spreads eastwards and northwards on the ocean bottom. Between the northward flowing intermediate and bottom waters there is a southward flowing deep water at depths of between 1500 and 2500 m. The Circumpolar Current is superimposed on this circulation and transports to the east all water lying within about 7° latitude of the Antarctic Convergence.

Deep water originates in the North Atlantic as well as around the Antarctic continent. The North Atlantic Deep Water flows south and becomes somewhat diluted with Antarctic Intermediate and Bottom Water before reaching the Circumpolar Current. The Circumpolar Current consists of a mixture of both the North Atlantic Deep Water and the Antarctic Deep Water.

The potential temperature of the bottom water is lowest in the Weddell Sea; it slowly increases as the bottom water moves along its path and mixes with overlying water (Figure 5). The bottom water is transported eastwards by the Circumpolar Current and spreads northwards into the major oceans. In the South Pacific (Figure 6), the bottom temperatures are indicative of the western boundary current in the South Pacific Basin.

Prominent tongues of Circumpolar Water extend into the Tasman Sea at the 3 km depth [Callahan 1972]. These are probably due to a northward deflection of the Circumpolar Deep Water by the Macquarie Ridge south of New Zealand. However, the Tasman Sea is closed off to the north at depths greater than 3 km, preventing the formation of a deep current to supply the North Pacific Deep Water. Instead, the principal deep current in the Pacific flows north, off the east coast of New Zealand, at a depth of 3 to 4 km. Observations made by the SCORPIO Expedition [Warren 1973] have enabled the geostrophic current distribution to be determined (Figure 7). The northward volume transport in the current is about 19×10^6 m^3 s^{-1}. This current flows through the narrow channel linking the North and South Pacific Deep Waters (Figure 6).

In the Atlantic, Indian and South Pacific Oceans, the salinity increases as the bottom water flows north because of mixing with the overlying water which is more saline. However, in the North Pacific the overlying water is less saline and the salinity of the bottom water decreases along the flow path [Lynn & Reid 1968].

4.2 Circulation and Mixing Models

Estimates of the time scale of the worldwide circulation patterns can be derived from the distribution of tracer materials (such as oxygen, carbon-14, strontium-90) that either decay or are consumed. Oxygen is gradually consumed by biological processes and is only replenished when the water returns to the surface layers. The rate of utilisation of oxygen would be expected to vary as a function of location and the amount of detritus reaching abyssal depths. Several abyssal circulation models using vastly simplified oceans have been proposed to interpret tracer measurements. The models consist either of constant depth rectangular oceans or of a network of well-mixed 'boxes'.

4.2.1 Advection lateral-mixing models

Advection lateral-mixing models use a constant depth ocean with the abyssal circulation proposed by Stommel & Arons [1960a,b] (Section 2.2). The tracers are assumed to be spread by large-scale advection and mixed by a horizontal eddy-diffusion. The abyssal circulation comprises narrow western boundary currents carrying the water towards the equator and geostrophically balanced flows within the ocean basins which carry the water towards the poles. Over the whole ocean, water is slowly returned to the surface layers with a small uniform vertical velocity.

An advection lateral-mixing model was applied to the oxygen distribution in the North Atlantic by Arons & Stommel [1967]. From this, they derived 600 to 700 $m^2\ s^{-1}$ for the lateral mixing coefficients and 2 to 2.5 $cm^3\ m^{-3}\ year^{-1}$ for the mean oxygen consumption rate. (The saturation value for oxygen dissolved in seawater at 0°C is 8000 $cm^3\ m^{-3}$, *i.e.* 8000 cm^3 of gas at NTP dissolves in 1 m^3 of seawater at 0°C.)

Kuo & Veronis [1970, 1973] have used an advection lateral-mixing model to interpret the worldwide oxygen distribution at a depth of 4 km. Figure 8 shows the observational data they are trying to fit and Figure 9 shows their idealised ocean model with a constant depth of 5 km. The sources of deep water in the North and South Atlantic Oceans were assumed equal and the total source strength was made equal to the total uniform upwelling over the entire oceans. The best fit to the observed oxygen distribution is shown in Figure 9b and the optimum values of the parameters were: horizontal eddy-diffusivity $K = 600\ m^2\ s^{-1}$, upwelling velocity $w_o = 1.5 \times 10^{-7}\ m\ s^{-1}$; oxygen consumption $\nu = 2\ cm^3\ m^{-3}\ year^{-1}$, and abyssal recirculation of the Antarctic Circumpolar Current $R = 35 \times 10^6\ m^3\ s^{-1}$. Actually, the model only depends on the ratios w_o/k, ν/k and R/k, and it was necessary to accept an estimate of ν based on observation to define the absolute parameters.

Models such as this obviously have severe limitations; one can object to the use of constant values of K, w_o and ν, to the use of a mixing coefficient or to the lack of any realistic topographical description of the oceans. Nevertheless, this simple model does reproduce the gross features of the abyssal oxygen distribution.

4.2.2 Box models

Box models divide the oceans into a small number of well-mixed boxes or reservoirs. Data on radioactive tracers are then used to determine the exchange rates between the different boxes and the mean residence times within the boxes. The selection of the boxes is obviously somewhat arbitrary and the number of boxes is usually kept small to limit the number of unknown exchange rates. As a recent example, Broecker & Li [1970] use only three boxes comprising the North Atlantic Deep Water, the Pacific and Indian Deep Waters and the warm surface water in their analysis of carbon-14 and inorganic carbon data.

Pritchard *et al.* [1971] evaluated many published box models to determine the best estimates of the residence times of water molecules in various subdivisions of the real ocean. They concluded that the residence time of a water molecule in the deep Pacific and Indian Oceans is about 1000 to 1300 years, whereas it is only about 10 years in the surface waters. The mean vertical velocities at the interface between the deep water and the water above it are estimated to be from 0.3 to 1.4 x 10^{-7} m s^{-1} (1 to 4 m $year^{-1}$).

Box models are useful for interpreting limited amounts of data, but great care is necessary when using the results because of the extreme simplicity of the model.

4.3 Vertical Velocities

The existence of the thermocline and the stratification of deep ocean waters suggests that the deep water has a relatively uniform upwelling velocity. Many methods have been used to determine the vertical velocity of deep water and the similarity of the results gives confidence in the uniformity of upwelling. This does not apply to the well-mixed waters less than 100 m deep, which have localised divergences caused by currents and winds.

Table 2 summarises some estimates of the mean upwelling velocity from a variety of methods. The Coral Sea heat balance, the box model and the advection lateral-mixing methods have all been discussed in previous

Sections.

Knauss [1962] estimated the flow rate of Pacific Bottom Water from carbon-14 data. He derived a mean vertical velocity by assuming that all of the northward flow was lost to the surface layers by a uniform upward movement of water.

Warren [1973] determined the geostrophic flow between New Zealand and South America across a section at 28°S. The net transport at depths greater than 2 km was 15×10^6 m^3 s^{-1}. This volume would support a mean upwelling of 3.9 m $year^{-1}$ over the whole of the Pacific north of 28°S.

The weapons tests in the Pacific have added a large quantity of carbon-14 since 1959. Pritchard *et al.* [1971] have analysed the vertical distribution in 1959, 1965 and 1966 to determine the vertical transport and hence the vertical velocity and diffusivity. The thermocline data gave a mean vertical velocity of 2.1×10^{-7} m s^{-1} and a vertical diffusivity, K_z, of 1.2 cm^2 s^{-1} .

About 2.1×10^{19} g of pack ice is formed in the Antarctic each year. Munk [1966] estimated that this results in the formation of 9×10^{20} g of deep and bottom water annually. The sinking of this quantity of water would produce a mean upwelling velocity in all the oceans of 3 m $year^{-1}$ at the 2000 m depth. More deep and bottom water would be produced in the Arctic.

TABLE 2

ESTIMATES OF THE UPWELLING VELOCITY OF DEEP WATER

Method	Upwelling Velocity		
	cm s^{-1}	m $year^{-1}$	Reference
Heat balance in Coral Sea	1.5×10^{-5}	4.7	Wyrtki [1961]
Carbon-14 in Pacific Bottom Water	2.5×10^{-5}	7.5	Knauss [1962]
Box models	$0.3\text{-}1.4 \times 10^{-5}$	0.9 - 4.4	Pritchard *et al.* [1971]
Weapons test carbon-14	2.1×10^{-5}	6.6	Pritchard *et al.* [1971]
Geostrophic flow into North Pacific	1.25×10^{-5}	3.9	Warren [1973]
Thermocline analysis	4×10^{-5}	12.6	Veronis [1969]
Advection lateral-mixing	1.5×10^{-5}	4.7	Kuo & Veronis [1973]
Vertical carbon-14 distribution	1.4×10^{-5}	4.4	Munk [1966]

5. A DEEP CIRCULATION MODEL FOR THE TASMAN AND CORAL SEAS

There are many uncertainties about the flow of deep and abyssal waters and the worldwide circulation patterns. However, it is important to estimate the likely flow and travel times even on the present inadequate knowledge of the ocean. To this end, a very simplified model will be developed for the deep circulation in the Coral and Tasman Seas and for the effects of mixing.

5.1 Circulation

A uniform upwelling velocity will be assumed for all the water below the thermocline. The data discussed in Section 4.3 gave a mean upwelling velocity of 6 m year^{-1}.

The deep water flows slowly north in the East Australian Basin with a mean velocity of the order of 6×10^{-4} m s^{-1} (20 km year^{-1}) [Warren 1973, Knauss 1962]. At the north end of the East Australian Basin, the bottom water must rise over the 3 km depth sill before flowing to the 5 km floor of the Coral Sea Basin (Figure 10). No account will be taken of the additional time spent upwelling in subsequent basins because this effect is included in some of the estimates of the mean vertical velocity. Bottom water from 5 km depth will take about 330 years to reach 3 km. Between 3 km and 2.5 km the water will be flowing north to fill the Coral Sea basins. At the uniform upwelling rate, the water will take 80 years for the 500 m rise. At this stage the upwelling water will be rising all over the Tasman and Coral Seas. Very little bottom water from the Tasman Sea flows out into the Central Pacific Basin; this water is supplied by the deep current to the east of New Zealand.

Above 2.5 km depth the oxygen-minimum water flows south. The velocity of this water is too small to measure, *i.e.* it is less than about 5×10^{-3} m s^{-1}. It is likely that all deep water entering the Tasman and Coral Seas flows back south as this water mass. The oxygen-minimum water is present from 1.5 km to 2.5 km depth, *i.e.* a depth range of only 1 km compared with a depth range of 3 km for the deep water. However, the horizontal extent of the ocean at 2 km depth is perhaps double that available for the deep water. Hence if the deep water is assumed to move north with a velocity of 6×10^{-4} m s^{-1}, then the oxygen-minimum water should move south with a velocity of about 1×10^{-3} m s^{-1} (32 km year^{-1}).

At a velocity of 1×10^{-3} m s^{-1}, water would take about 150 years to travel from the Coral Sea to the Antarctic Convergence.

The water flowing south would still be subject to upwelling and the time taken for water to rise from 2.5 km to 1.5 km would be about 170 years.

Overlying the oxygen-minimum water is the Antarctic Intermediate Water which flows mainly northwards with its core at a depth of about 1000 m. The maximum velocity of the Antarctic Intermediate Water in the region is about 5×10^{-2} m s^{-1}, *i.e.* 1600 km $year^{-1}$ [Scully-Power 1973]. The mean velocity is probably considerably less than this and a value of 10^{-2} m s^{-1} is a reasonable estimate. In the Tasman Sea between 30 and 40°S, there is very little net flow of the Antarctic Intermediate Water. Close to the east coast of Australia, the Antarctic Intermediate Water comes under the influence of the East Australian Current and flows south. The Antarctic Intermediate Water ultimately becomes mixed with the surface water or the oxygen-minimum water.

Upwelling oxygen-minimum water will mix with the Antarctic Intermediate Water. In the time taken for water to rise from 1500 m to 500 m at the uniform upwelling velocity, the Antarctic Intermediate Water could travel a distance greater than the extent of the Pacific Ocean. Hence it is likely that a contaminant in the Antarctic Intermediate Water would be spread throughout the Pacific.

The surface layers are well mixed by the strong currents and the complex circulation patterns.

5.2 Mixing

Superimposed on the circulation patterns are the diffusion or mixing processes. The vertical diffusivity in deep water below the thermocline is about 10^{-4} m^2 s^{-1} [Okubo 1971]. Using the formula given in Section 2.3, the standard deviation of the distribution of a contaminant after 100 years would be about 800 m, and after 1000 years it would be 2.5 km. These distances are of the same order as the distance travelled by a molecule having the mean upwelling velocity.

The magnitude of horizontal diffusion in deep water is very uncertain. There have been virtually no direct measurements, the only estimates being derived from the advection lateral-mixing models, which use vastly simplified circulation patterns. Using such a model, Kuo & Veronis [1973] obtained a lateral eddy-diffusivity for deep water at 600 m^2 s^{-1}. This is equal to the surface coefficient for spreading after 30 days and a factor of 50 less than the surface coefficient after 365 days' spreading.

After 100 years of spreading with a lateral eddy-diffusivity of 600 m^2 s^{-1}, the lateral standard deviation of the patch would be 2000 km, which corresponds to 18 degrees of latitude. Just as the vertical spreading was comparable to the vertical distance travelled, the horizontal diffusion is of the same order as the horizontal spreading. It is not clear if this is typical of oceanic mixing processes which are driven by large-scale circulation or if it is a consequence of the over-simplified advection lateral-mixing model.

5.3 Time-scales and Destinations

The empirical model developed above has been used to calculate the distribution of times when water released in the bottom of the Tasman Sea would reach the surface. The mixing processes have been assumed to produce the Gaussian distribution characteristic of a constant eddy-diffusivity.

Water rises from 5000 to 2500 m with a velocity of 6 m year^{-1} while undergoing mixing with a vertical eddy-diffusivity of 10^{-4} m^2 s^{-1}. No correction has been made for the proximity of the ocean floor. The calculated rate of arrival of the released water at the 2500 depth is shown on Figure 11. The median time of arrival is 415 years.

The water is assumed to reach the 2500 m depth uniformly over the Tasman and Coral Seas north of 20°S. The southward moving oxygen-minimum water then carries the released water south at 1×10^{-3} m s^{-1} while mixing with a horizontal eddy-diffusivity of 600 m^2 s^{-1}. The calculated rate of arrival at the Antarctic Convergence at 57°S is also shown on Figure 11. The median time for the water to travel from depths of the Tasman Sea to the Antarctic Convergence is 600 years. Ten per cent of the water will have reached the convergence within 300 years and 90 per cent will have arrived by 1120 years.

The water surfacing south of the Antarctic Convergence will travel eastwards with the Circumpolar Current and finally go north again with the bottom water or the Antarctic Intermediate Water.

There are many possible errors in a model like this; the velocities and diffusivities are poorly known, the ocean upwelling and currents are not uniform, and the circulation is likely to be complex. Nevertheless, the model does provide an estimate of the time-scales of the circulation and the range of arrival times. It has been assumed that the abyssal water in the Tasman Sea moves slowly northward. One major uncertainty is a possible deep circulation, coupled perhaps to the Circumpolar Current, that could cause an exchange between Circumpolar Current Water and deep Tasman Sea Water.

6. CONCLUSION

The observational data on the deep circulation in the Tasman and Coral Seas is barely adequate to make a prediction of the destinations and travel times of water released in the deep Tasman Sea. Water released in this sea probably flows into the Coral Sea and finally reaches the ocean surface in the Antarctic Circumpolar Current south of the Antarctic Convergence. Estimates of the flow velocities and eddy-diffusivity have been used in the circulation model to determine the time scale of the bulk transport. The median time for water to travel from the deep Tasman Sea to the surface near Australia is 600 years but the eddy-diffusion processes cause a quartile deviation of 200 years. After reaching the surface layers perhaps 50 per cent of the water will be returned to the abyssal depths by the formation of deep and bottom water around the Antarctic Continent.

The inaccuracies in the model are principally due to uncertainties in the circulation pattern and, to a lesser extent, to errors in the numerical data. Nevertheless, the arrival times are very unlikely to be in error by more than a factor of two. The calculated median travel time is consistent with the estimated mean residence time of 1000 years for the Pacific Deep Water.

Although this report has concentrated on the Tasman and Coral Seas, the times obtained for deep water from 5000 m to travel to the surface layers are likely to be typical of most oceans. Transport by ocean currents has been shown to be sufficiently slow for radioactive materials with half-lives shorter than 30 years to have decayed to low levels by the time they reach the surface. After 20 half-lives, the quantity of a radioactive element has been reduced by a factor of one million. The uptake of radionuclides by the sediments would further reduce the concentration. With few exceptions, all chemical species introduced into the ocean are ultimately removed from the seawater and deposited on the ocean bottom. Much biological transport is downwards as excretory products and dead organisms sink to lower levels. However, little is known of the magnitude of the upward biological transport of radionuclides from abyssal depths, and this process could be as important as that of the bulk transport by ocean currents for some elements.

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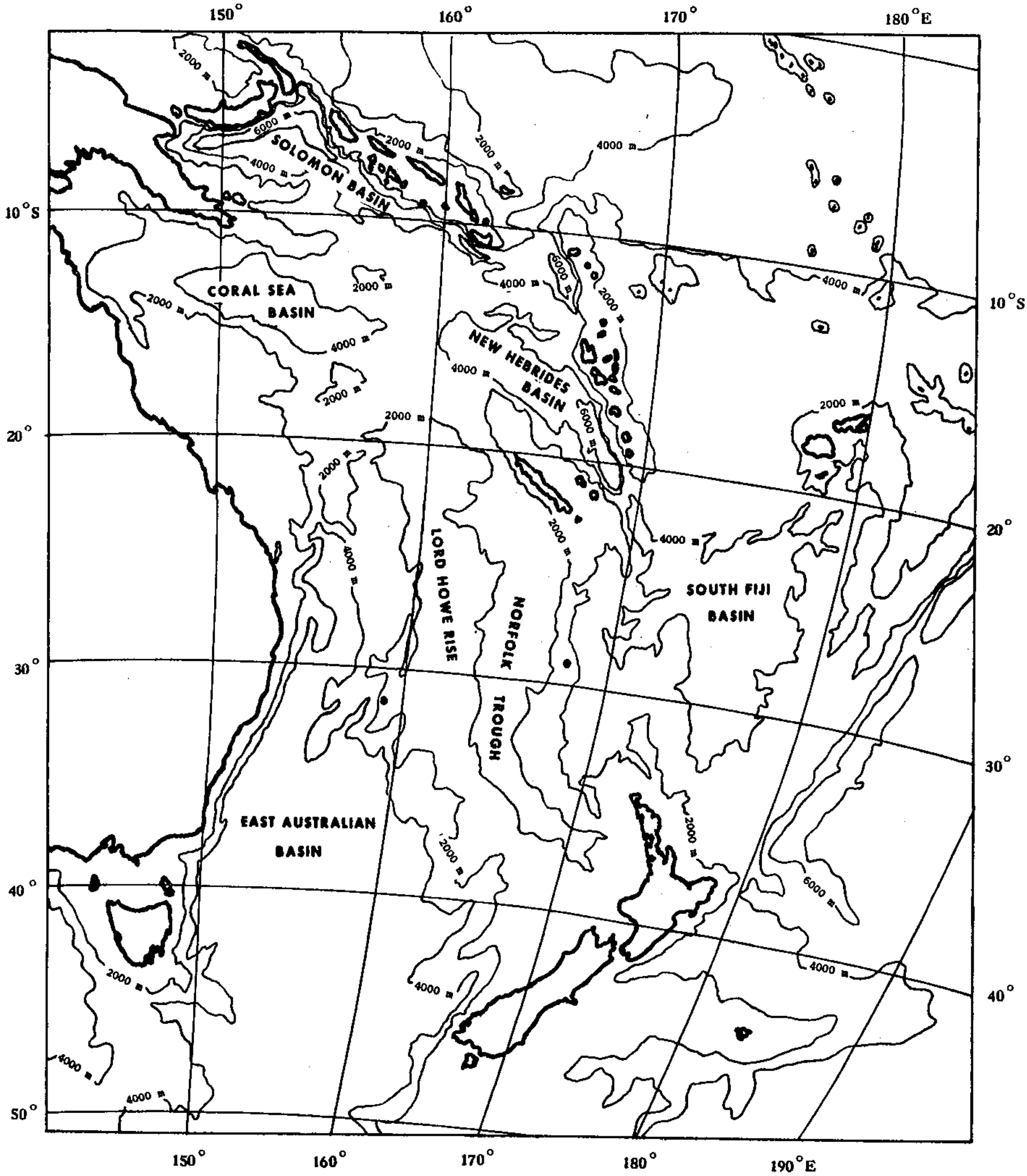


Figure 1 The bottom topography of the Tasman and Coral Seas.

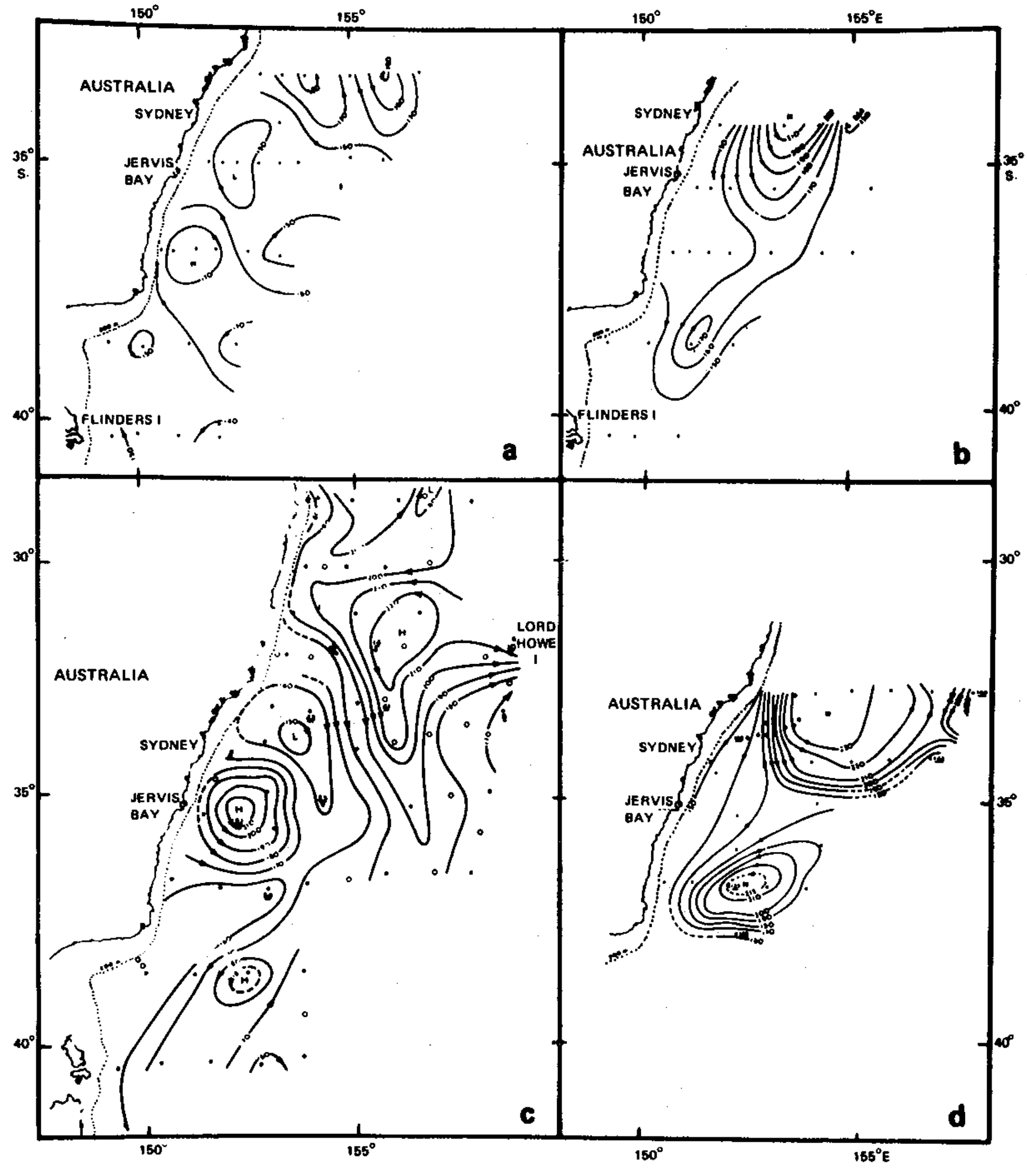


Figure 2 The dynamic surface circulation showing the variability of the East Australian Current. (a) September 1963; (b) November 1963; (c) January-February 1974; (d) March 1964 (Source: Hamon 1965).

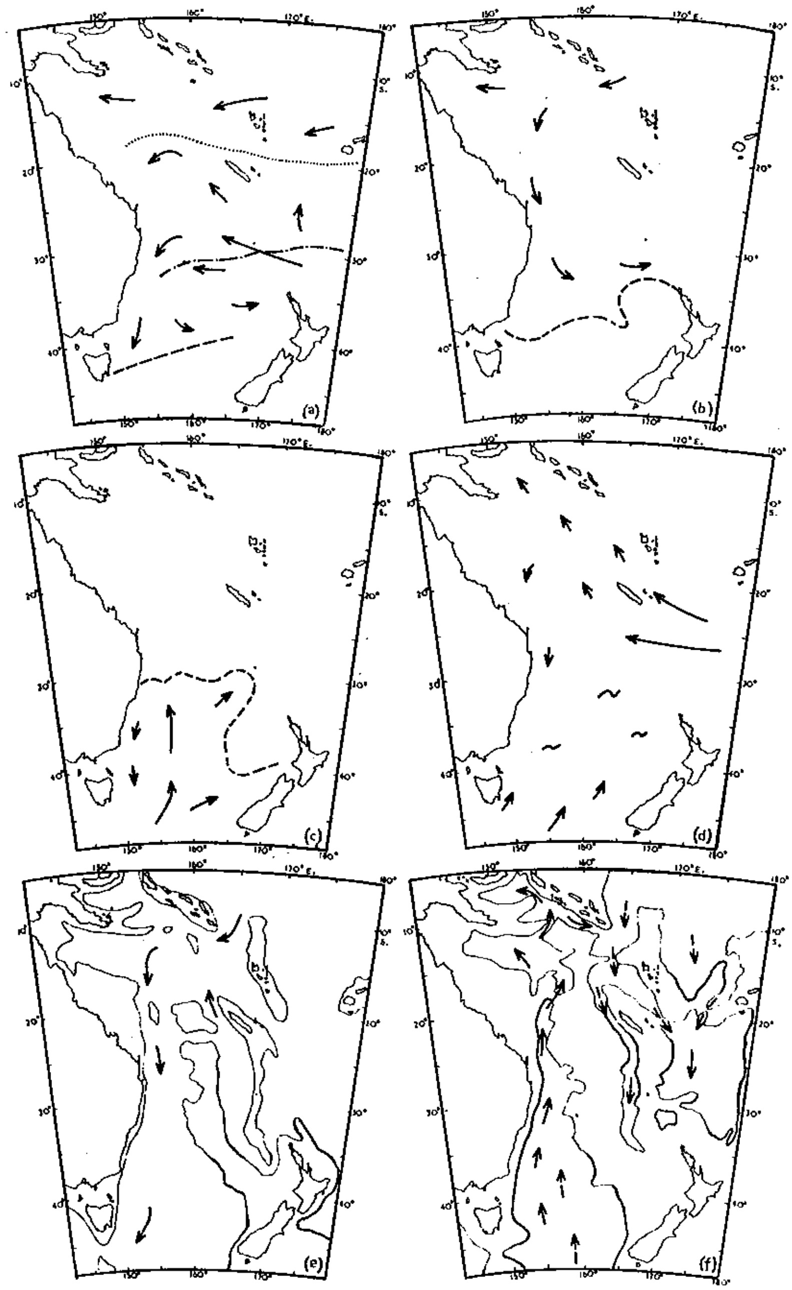


Figure 3 The spreading of different water masses as determined by Wyrtki [1962]. (a) 125-250 m; (b) upper oxygen minimum 150-500 m; (c) oxygen maximum 150-400 m; (d) Antarctic Intermediate Water 650-1100 m; (e) deep oxygen minimum 1000-2500 m, 200 m contour shown; (f) deep and bottom water, 3000 m contour shown.

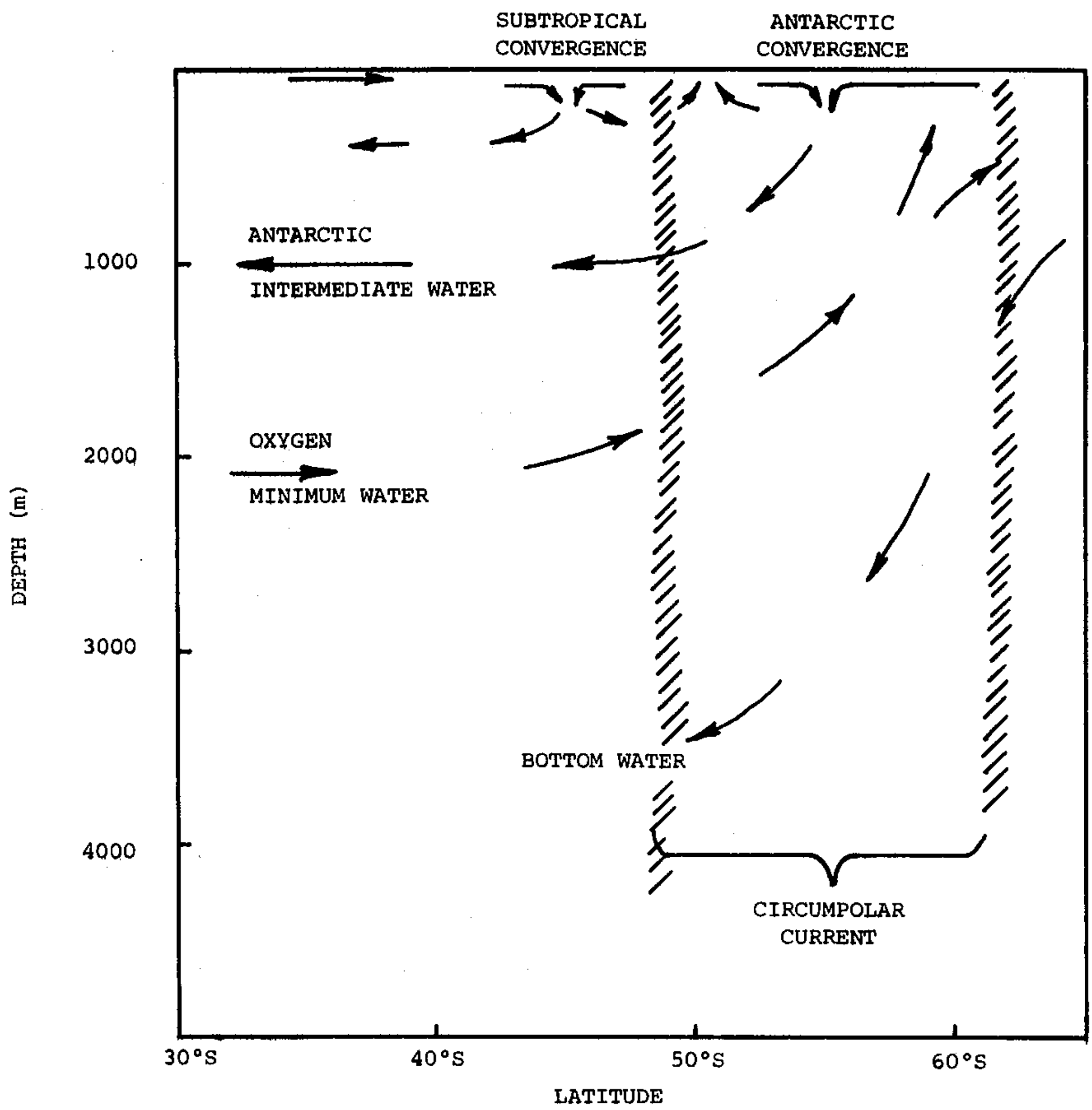


Figure 4 The circulation in a vertical section through the circumpolar current.

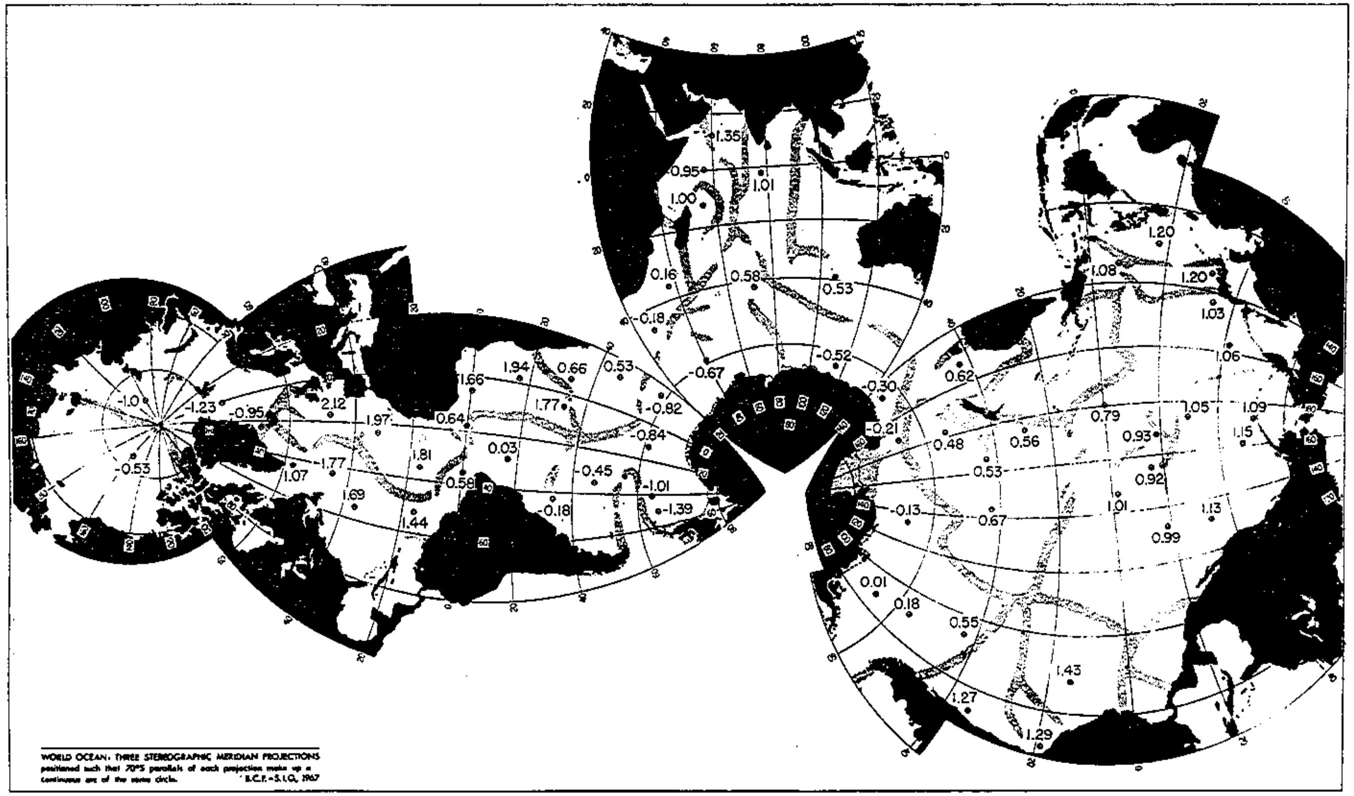


Figure 5 Potential temperatures at abyssal depths in the world oceans (Source: Lynn & Reid 1968).

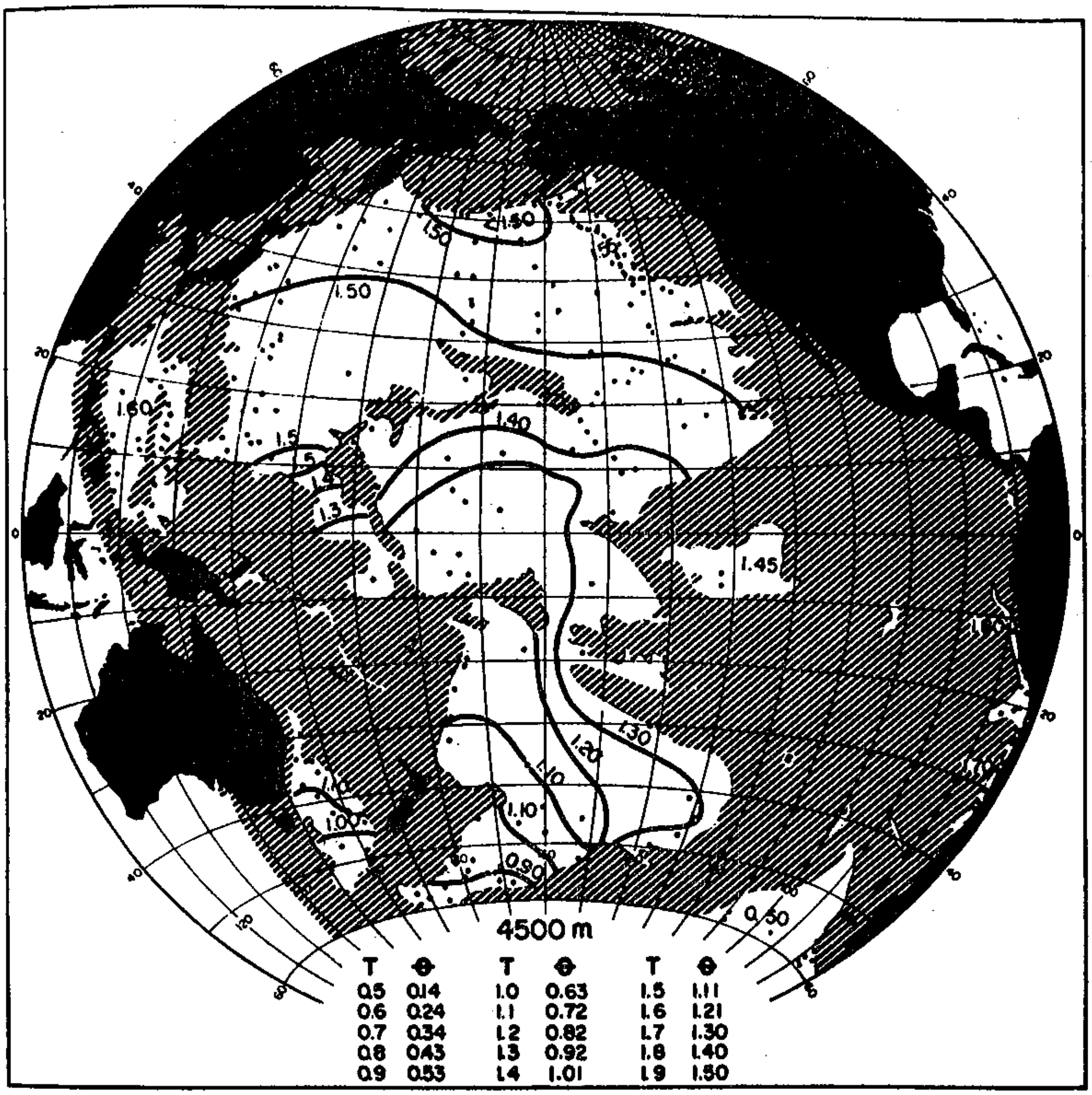


Figure 6 *In situ* temperatures at 4500 m in the Pacific Ocean. The cross-hatching shows areas of the Pacific where the depth is less than 4500 m. The corresponding potential temperatures, θ, are shown in the table (Source: Knauss 1962).

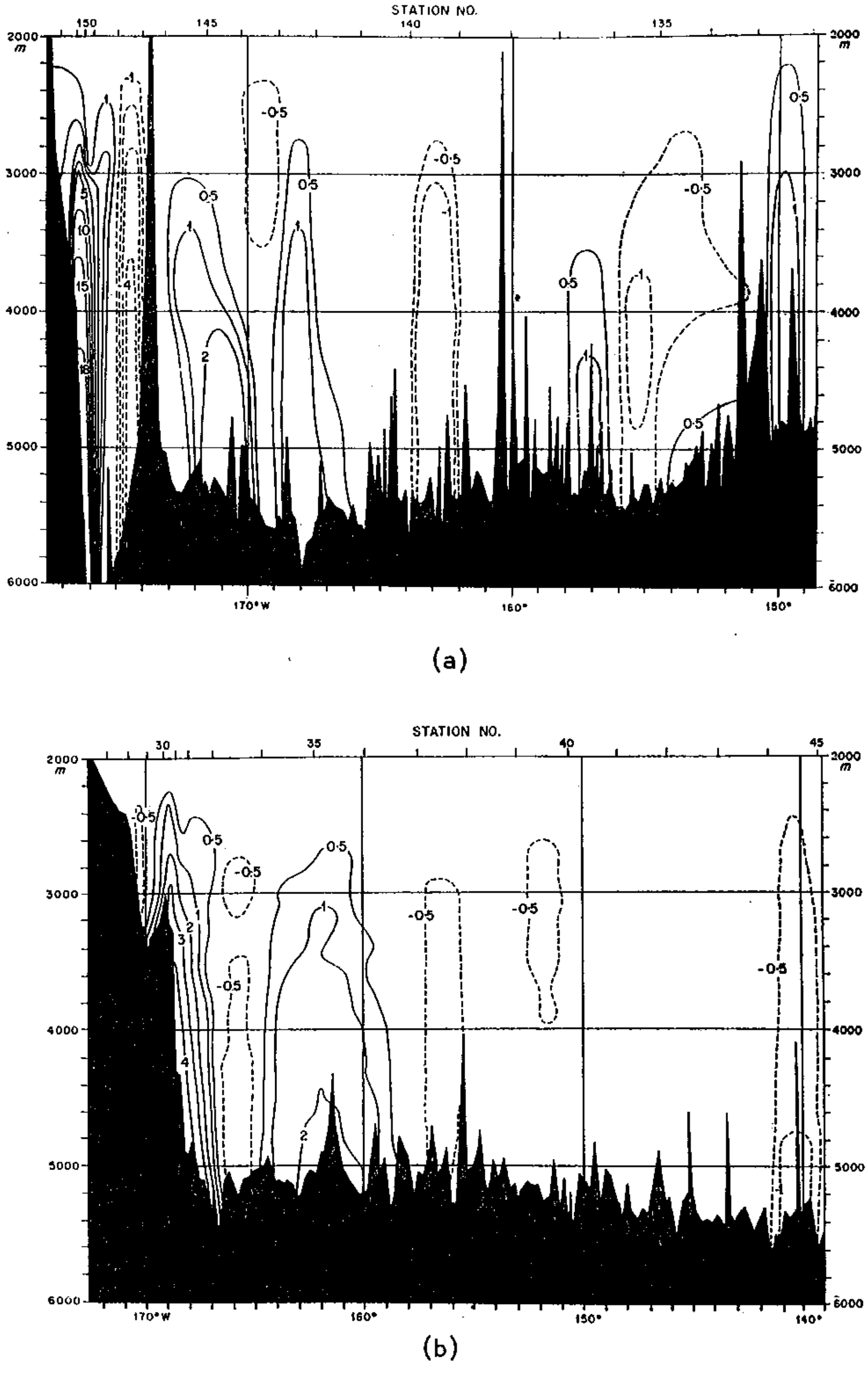
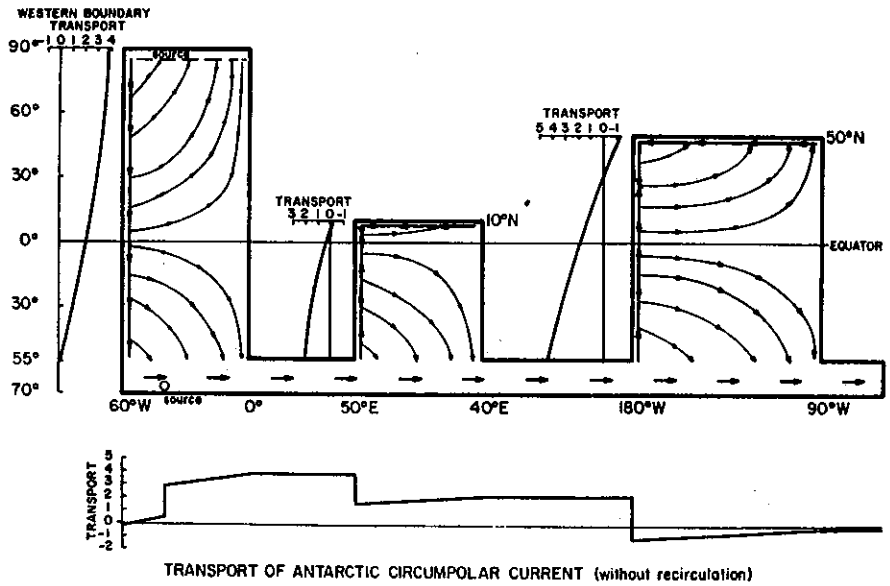
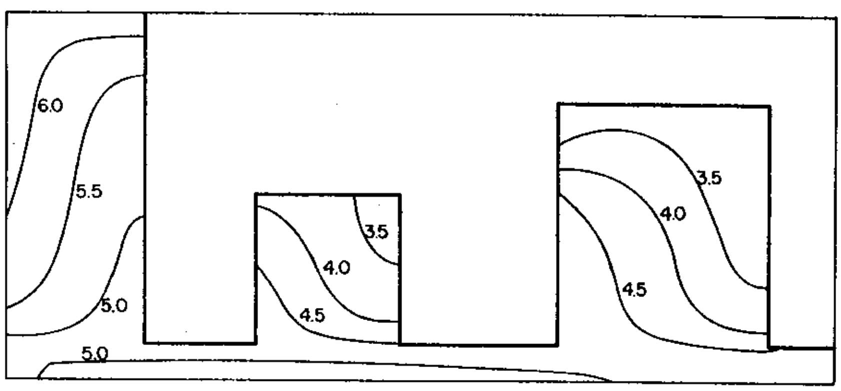


Figure 7 The deep western boundary current of the South Pacific flowing to the east of New Zealand. Contours are labelled in cm s^{-1}; solid lines indicate northward flow, dashed lines indicate southward flow (a) latitude 28°S; (b) latitude 43°S (Source: Warren 1973).

Figure 8 The distribution of dissolved oxygen in $m\ell\ \ell^{-1}$ at 4 km depth (Source: Kuo & Veronis 1973).



(a) Idealised model of world ocean;



(b) best fit to the observed oxygen distribution.

Figure 9 The calculation of the dissolved oxygen distribution in an idealised ocean model by Kuo & Veronis [1973]. The calculated distribution should be compared with the observed data shown in Figure 8.

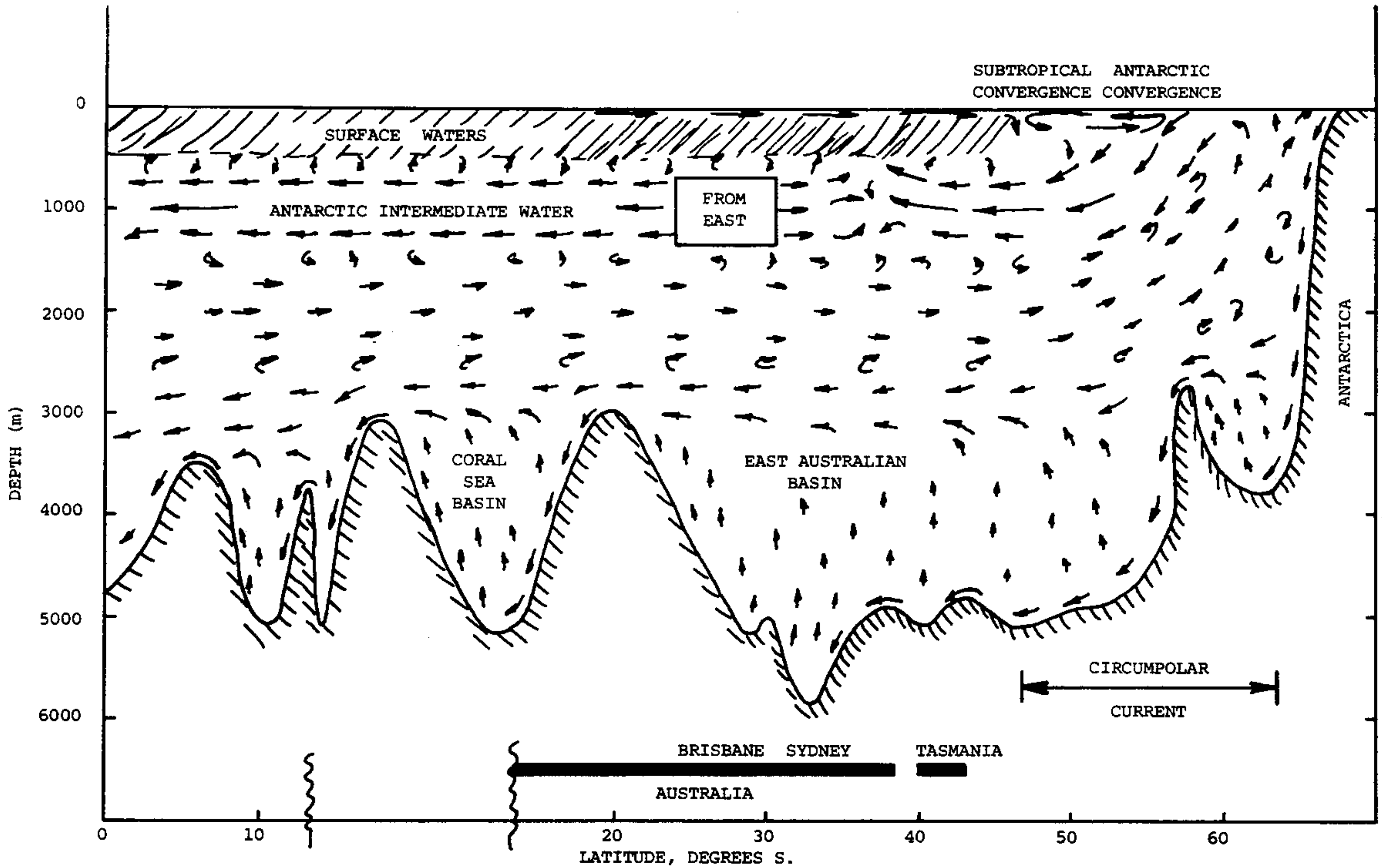


Figure 10 The circulation model from Antarctica to the Central Pacific through the Tasman and Coral Seas. The Antarctic Intermediate Water enters the Tasman Sea from an easterly direction north of New Zealand at the approximate position of 'from East' label.

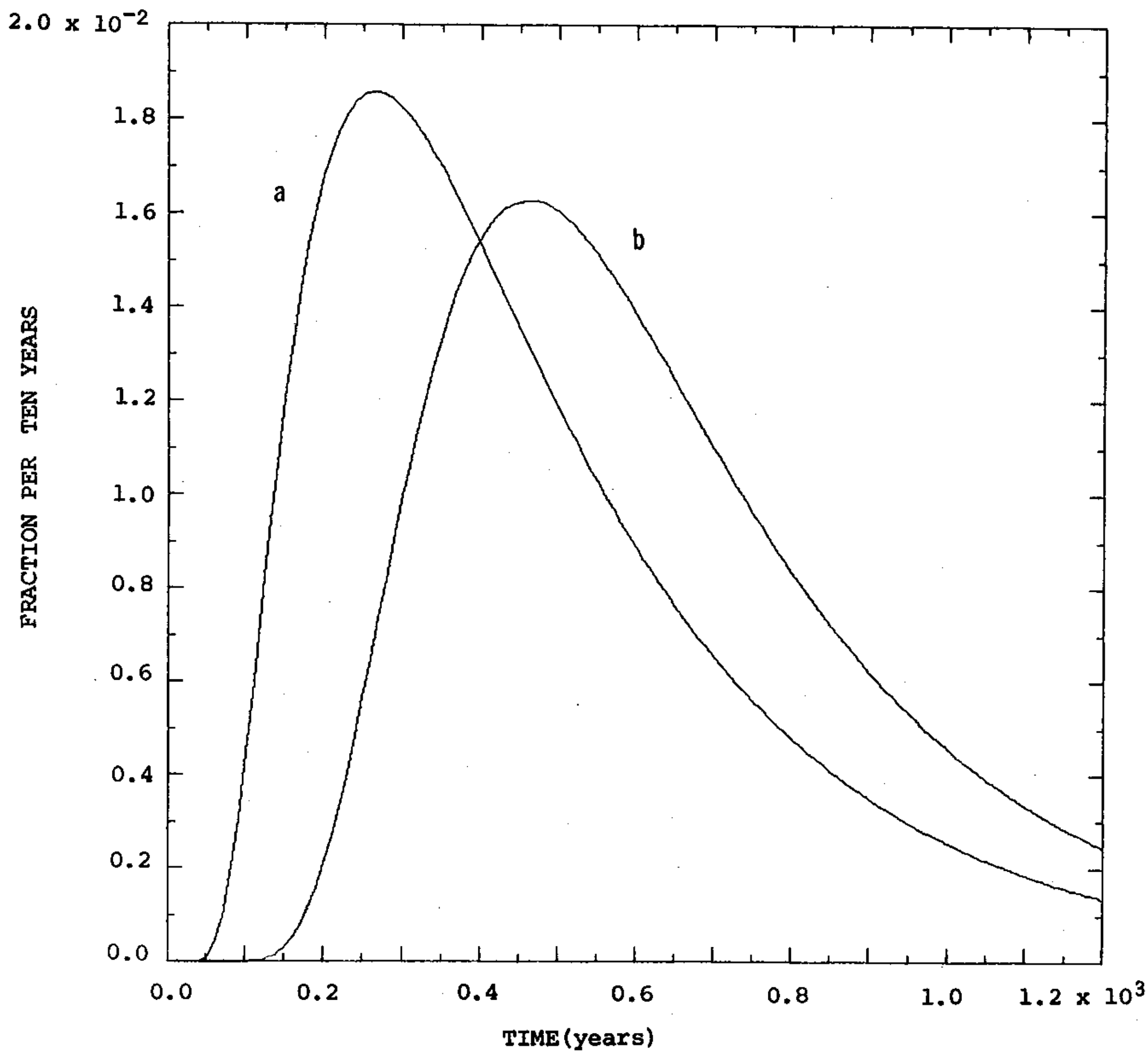


Figure 11 The calculated times for water released at 5000 m in the Tasman Sea to (a) reach 2500 m depth and (b) reach the surface layers south of the Antarctic Convergence, on the basis of the model.