Observations of the Leeuwin Current off Western Australia

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ABSTRACT

Shipboard observations made in May 1982 showed a definite poleward surface flow (the Leeuwin Current) over the West Australian shelf from 22°S to 28°S. The surface current was relatively fresh, warm, low in dissolved oxygen concentration, and high in nutrients. The current flowed against a strong wind. Only a small portion of its flux of 4×10^6 m^3 s^{-1} came from the Northwest Shelf. There was a subsurface equatorward current at a few hundred meters depth which was salty, high in oxygen concentration and low in nutrients. Observations from previous cruises show a surface geopotential gradient that could drive the surface current. It is suggested that winter deepening of the mixed layer may allow the geopotential gradient to overcome the wind stress.

1. Introduction

The eastern boundary currents in subtropical latitudes of the Atlantic and Pacific Oceans include the Peru, Benguela, California and Canary Currents, all of which have a prevailing flow toward the equator with substantial upwelling and concomitant high biological productivity.

The eastern boundary current of the Indian Ocean shows no signs of an equatorward flow nor upwelling (Wooster and Reid, 1962). The thermocline is relatively deep, fish productivity low and the rainfall at Perth (32°S) is four times that of San Diego (32°N). There is, in fact, a southward surface flow in autumn and winter of low-salinity tropical water which flows from at least 29°S, around Cape Leeuwin (34°S) and eastward into the Great Australian Bight (Cresswell and Golding, 1980). Legeckis and Cresswell (1984) give evidence from satellite imagery that this "Leeuwin Current" flows southward from at least Northwest Cape (22°S). The purpose of the present paper is to confirm their findings and to discover the subsurface structure of the Leeuwin Current, which has apparently not been previously determined.

We find that there was indeed a poleward warm current along the edge of the West Australian continental shelf during May 1982, from at least 22 to 28°S. There was a strong salinity front, inside of which the water was relatively warm, fresh and high in nutrients. Some of the water offshore of the front also appeared to be moving poleward. There was an equatorward moving undercurrent below about 150 m depth, carrying relatively salty, low-nutrient, high-oxygen water.

2. Data collection

The bulk of the data reported here was collected between 10 and 16 May 1982 on five hydrographic sections 300 km seaward to shore between 22°S and 27°25'S; Fig. 1b may be used for location. Samples were taken with Niskin bottles in a rosette on a Neil Brown CTD, and analyzed for temperature, salinity and dissolved oxygen, inorganic phosphate, nitrate and silica concentrations using methods described by Major *et al.* (1972). Surface temperature, salinity and water depth were continuously monitored. Surface current was estimated from ship set using satellite and radar navigation, corrected by 3% of the wind velocity (a calibration estimated from trials with the ship hove to in relatively still water). Expendable bathythermographs (Sippican T7) were used approximately every 40 km between hydrographic sections to construct five more temperature sections. Mixed layer depth was defined as the first depth at which the temperature was 0.2°C less than at 2 m depth.

Water-velocity profiles in the vertical were made by suspending an Aanderaa RCM-4 on fixed lengths of taut hydrographic wire for half-hour periods at each of several depths while the ship was drifting. The instrument depth, temperature and relative speed and direction were telemetered acoustically, and decoded on board. The ship set was used to correct the relative velocities to absolute. Water velocities were also estimated from satellite positions of a drogued buoy.

Some hydrographic data also reported here were taken in May 1971, with samples taken by serial cast using Niskin bottles at standard depths and analyzed only for temperature and salinity.

3. Observations

Figure 1a shows the observed surface currents (from ship set and satellite buoy displacement) as well as analyzed surface isotherms. The isotherm analysis has been influenced by a satellite infrared picture made

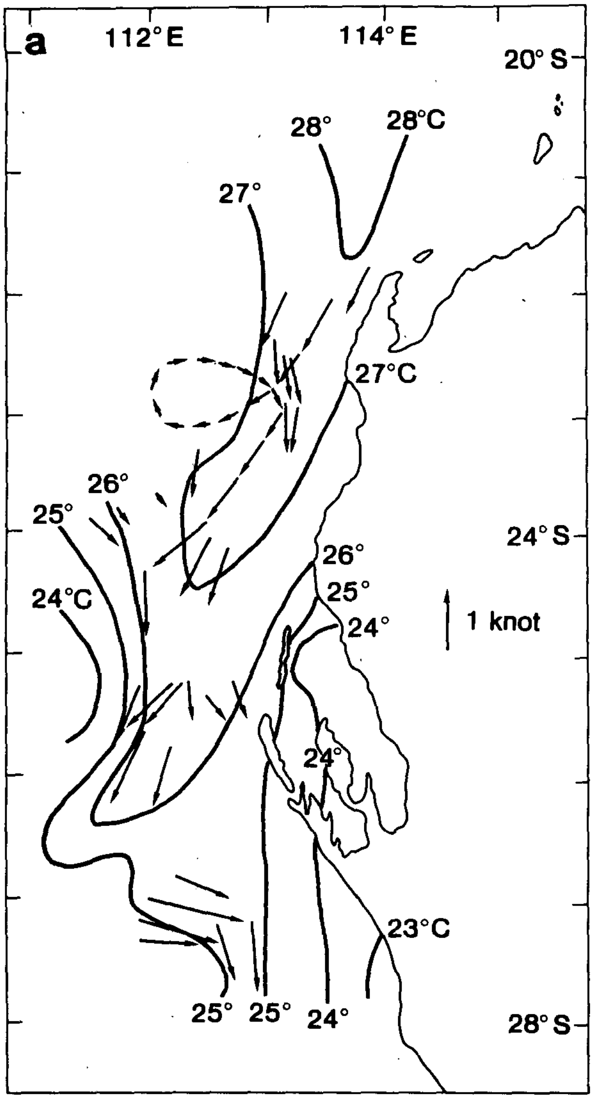

FIG. 1. (a) Surface isotherms (°C) and surface current off Western Australia during May 1982.

by A. F. Pearce (personal communication, 1982) during the cruise. There is an apparent relationship between poleward current in Fig. 1a and the mixed layer depth in Fig. 1b.

The five hydrographic sections were drawn; the section due west from Shark Bay, along 25.4°S, was found to be representative, and is shown here as Fig. 2. Another section is given by Thompson and Cresswell (1983); all data are provided by Thompson (1982). Fig. 2a shows a downward tilt of the isopycnals toward the coast, consistent with a poleward vertical shear of the current. A surface warm core, shown in Fig. 2b, is also consistent with a surface current carrying tropical water poleward, as is the low-salinity core shown in Fig. 2c. One also sees a strong salinity front near the edge of the shelf. There is a strong salinity maximum core at about 200 m depth over the upper slope, in a region of downward tilt of the isopycnals. Both Figs. 2c and 2b, and the data for Fig. 2a, show a flattening of isopleths near 350 m depth.

Figure 2d shows a dissolved oxygen concentration (oxyty) minimum core coincident with the salinity minimum. There is an offshore oxyty maximum, apparently divided into two parts: one near the salinity front and one below the salinity maximum. Fig. 2e shows a dissolved silica maximum core on the shelf close to the salinity minimum core; there is a strong front in silica concentration coincident with the salinity front. Beyond that is a minimum in silica, apparently a core, coincident with the lower oxyty maximum and extending to about 350 m depth. The nitrate concen-

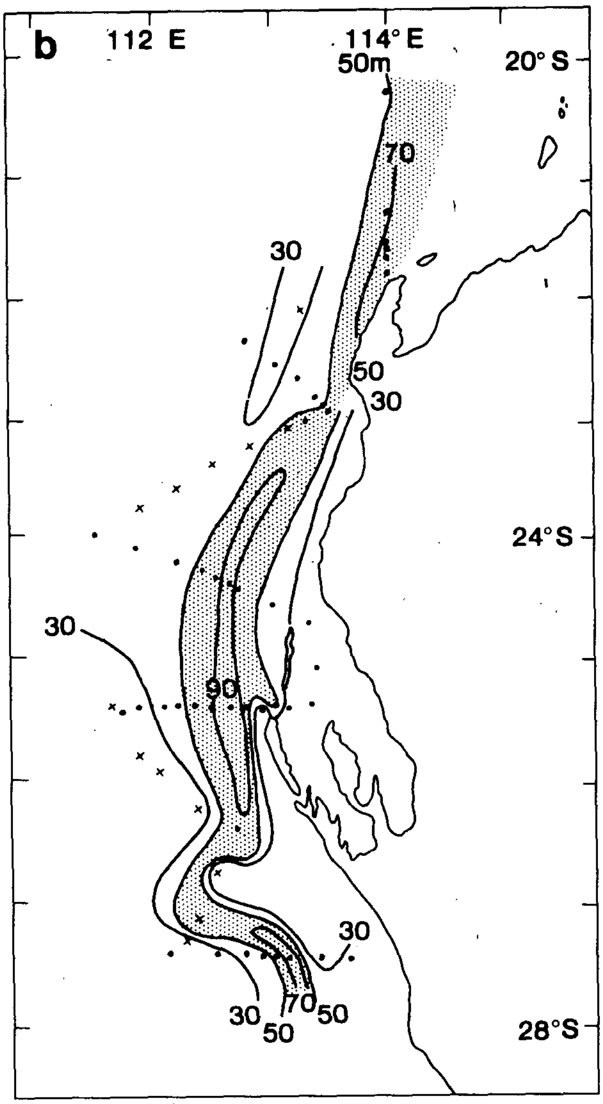

FIG. 1. (b) Mixed layer depths off Western Australia, 10–16 May 1982, with isobaths of 30, 50 and 70 m contoured. The mixed layer depth exceeded 50 m within the stippled area; crosses denote an XBT station, and dots denote a CTD station.

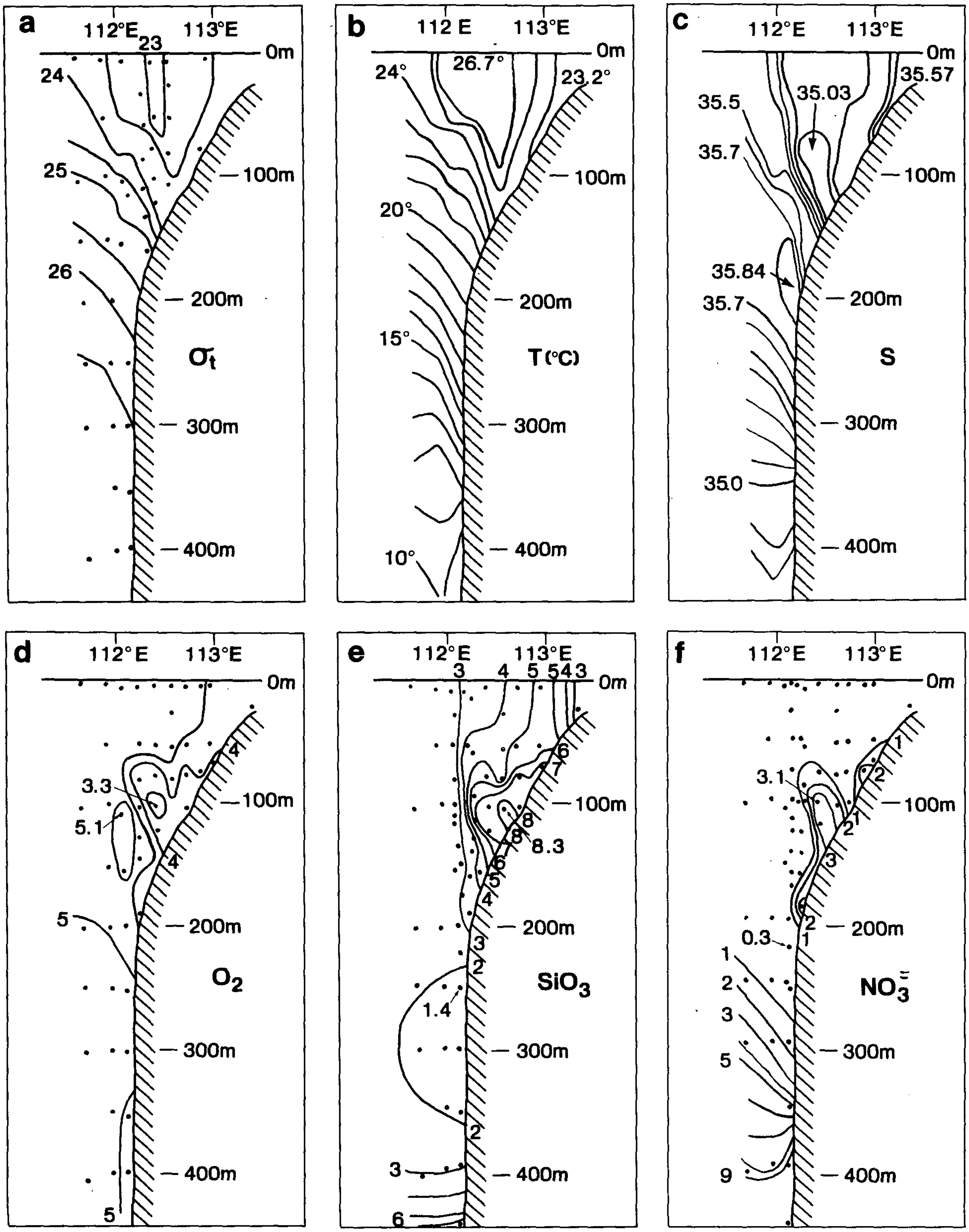

FIG. 2. A hydrographic section along 25°24'S over the continental slope and shelf of Western Australia: (a) isopycnals (σ_t) at 1 kg m^{-3} intervals; (b) isotherms (T) at 1°C intervals; (c) isohalines (S) at 0.1‰ intervals; (d) isopleths of dissolved oxygen in ml l^{-1}; (dots are sampled positions); (e) isopleths of dissolved silica in μmol l^{-1}; (dots are sampled positions); (f) isopleths of nitrate nitrogen in μmol l^{-1}; (dots are sampled positions).

tration in Fig. 2f has a very similar pattern to that of the salinity, but reversed: one finds a nitrate maximum core on the shelf, a strong front at the edge of the shelf and a minimum at ~200 m depth near the 300 m isobath. The isopleths of nitrate concentration slope downward inshore to ~350 m depth, consistent with advection of low-nitrate water between the 200 and 350 m isobaths. The phosphate section (not shown) is very similar to the nitrate section. All three nutrient sections show surface enrichments over the shelf compared to offshore, consistent with mixing or advection of nutrients into the photic zone.

All four other sections show the features described above (Thompson, 1982). The main latitudinal differences were that the slope of the isopycnals extended deeper in the more poleward sections (400 m at 27°25'S versus less than 250 m at 22°S), that the nutrient maxima and salinity minimum weakened poleward and that the salinity maximum detached from the surface and became deeper and weaker equatorward. This latter feature is illustrated in Fig. 3, which shows the salinity core along approximately the 600 m isobath, from hydrographic data collected every two degrees from 20 to 34°S during May 1971. Another difference was that the salinity section into Northwest Cape, near 22°S, showed a salinity maximum on the inner part of the shelf, inside the salinity minimum at the edge of the (narrow) shelf.

In planning the 1982 cruise, we suspected that the Leeuwin Current might be driven by a surface pressure difference due to the warm tropical water stacking up higher than cooler subtropical water and that there might be an equatorward undercurrent driven by a reversed pressure gradient at depth, caused by the deeper thermocline in the midlatitude gyre. Consequently, we plotted the geopotential anomaly versus latitude from the 1971 cruise, for the surface layer (0–300 db) in Fig. 4a and a deeper layer (300–1100 db) in Fig. 4b. There is a definite trend for the surface layer to be about one-half meter thicker at the equatorward end of the Leeuwin Current than at the poleward end. Hamon (1965) found very similar trends in

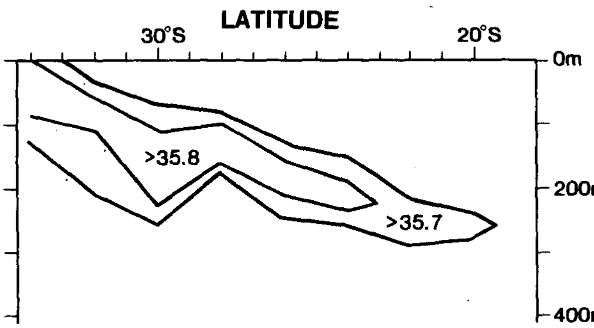

FIG. 3. The 35.8 and 35.7‰ isohalines on a depth versus latitude section constructed from serial Niskin casts made every 2° of latitude along the continental slope of Western Australia.

five cruises along 110°E. There is a less definite trend for the deeper layer (300–1100 db) to be about a tenth of a meter thinner at the equatorward end.

We also measured the actual velocities at depths to 300 m near the 320 m isobaths off Northwest Cape and off Shark Bay. The equatorward component of current for the two sites is plotted in Fig. 5. There was definitely a poleward surface current and an equatorward undercurrent at both sites. Also plotted in Fig. 5a is the geostrophic velocity component computed from Fig. 2a, using 150 m as the level of no motion. With that level of no motion, poleward flux of water computed from Fig. 2a was 4.5×10^6 m^3 s^{-1}. The geostrophic velocity component in Fig. 5b is from the section near 22.5°S given by Thompson and Cresswell (1983).

The wind was observed to blow from the SSE at 20–25 knots over the entire two-week cruise from Broome (18°S) to Fremantle (32°S).

4. Discussion

The surface isotherms in Fig. 1 bear a strong resemblance to Fig. 2 of Legeckis and Cresswell (1984), who interpreted the poleward arch of the isotherms as caused by advection. The directly measured ship sets, buoy sets and currents reported here corroborate their hypothesis; there *is* a warm poleward current along the edge of the West Australian shelf in May.

There is also a poleward geostrophic current according to the density structure of Fig. 2a; one can even see the advection of potential density. Just looking at Fig. 2a (potential density) or Fig. 2b (temperature), one would guess a level of no motion at 350 m or deeper. The salinity in Fig. 2c with its maximum core at 200 m might cause uneasiness, but the isohalines do flatten at 350 m. The oxyty section (Fig. 2d) does not help much. However, the low-silica core of Fig. 2e clearly indicates an equatorward flow, since tropical water in the Indian Ocean is comparatively high in silica (Warren, 1981a). This interpretation is supported by the negative anomalies in nitrate (Fig. 2f) and phosphate. In this case, examination of the nutrients could save one from making an error in the level of no motion which would double the estimate of surface speed (Fig. 5a) and quadruple the estimate of transport. Of course, a salinity core as presented in Fig. 3 would give a strong indication of the direction of flow. In fact, Fig. 3 suggests a level of no motion of 150 m or shallower at 25.4°S, fitting well with Fig. 5.

The indications of a subsurface equatorward flow are in fact strong enough so that Fig. 5 might be better interpreted as a confirmation of the usefulness of suspending an Aanderaa current meter than as a confirmation of the level of no motion at 150 m. We had feared overspeeding of the Savonious rotor due to wave pumping, but this apparently did not occur. The current meter measurements of Fig. 5b (22°S) are thought

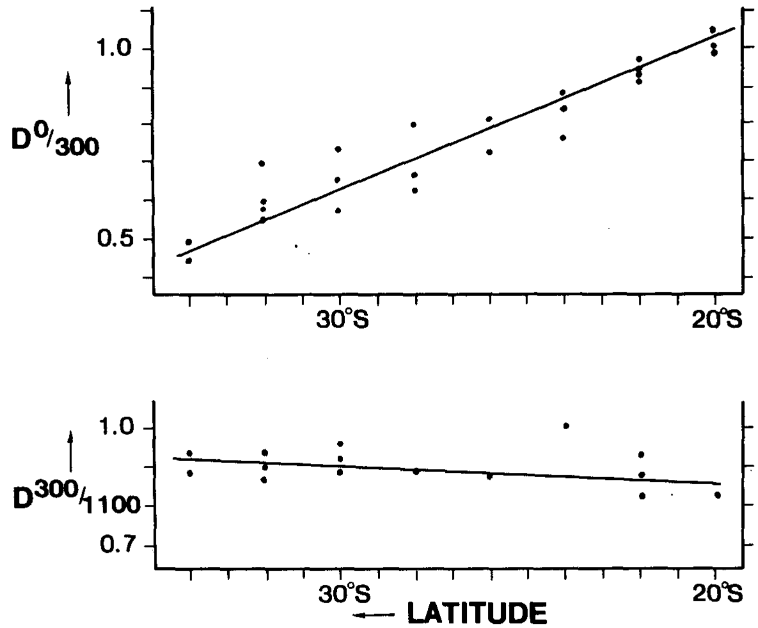

FIG. 4. Geopotential anomaly in $m^2 s^{-2}$ plotted versus latitude from hydrographic stations made in 1971 off Western Australia, with straight-line fits: (a) between the surface and 300 m depth; (b) between 300 and 1100 m depths.

to be reliable, since conditions were unusually propitious: the wind set of the ship exactly canceled the current set, to within the small measurement error, both radar and visual. We were within a few kilometers of high hills on a clear day, and appeared dismayingly stationary for having come to measure a current. The ship lay in the lee of the hills so the waves were very small, despite a 20 knot wind, and rolling was negligible. Conditions were not so propitious at 25.4°S, yet Fig. 5a seems comparable to Fig. 5b.

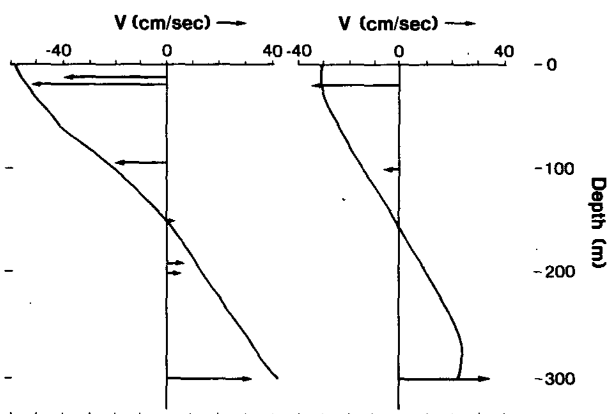

FIG. 5. Equatorward component of velocity in the Leeuwin Current plotted versus depth. The arrows are measurements from a current meter; the curve from geostrophy, using 150 m as a level of no motion: (a) at 25.4°S, the arrows at 15 and 190 m depth were observations made on the 200 m isobath, the arrows at 18, 100, 150, 200 and 300 m were made on the 320 m isobath; (b) near 22°S on the 320 m isobath.

The undercurrent found here is transporting high-salinity, high-oxyty South Indian Central Water (SICW) northward. The water must go somewhere; presumably into the interior of the Indian Ocean. The SICW core of Fig. 3 here closely parallels a similar core at 110°E found by Rochford (1969, his Fig. 13). Since we know (Fig. 5) that there is an equatorward flow of up to 0.4 m s^{-1} carrying of order 5×10^6 m^3 s^{-1} (5 Sv), let's speculate that the equatorward flux of SICW sketched by Rochford does not actually take place along 110°E, but along the continental slope (near 114°E) and spreads westward. An interesting consequence of this speculation then immediately follows from Rochford's Fig. 13, wherein we see the SICW core sliding under the "tropical oxygen minimum" core centered at $\sim$180 m. This implies a rather sharp downward attenuation of the shallow oxyty minimum and would answer Warren's (1981b) puzzle as to why that should happen.

The slope of the line in Fig. 4b is not significantly different from zero. Nonetheless, now that we know there is an equatorward flow on the slope around 300 m depth, it is reasonable to note that the equatorward pressure gradient indicated for 300 m by Fig. 4b would push that current. The slope of geopotential anomaly shown in Fig. 4a is definitely positive and strong enough that the pressure gradient near the surface is poleward, again pushing the observed current. Therefore, the observed system is energetically compatible with a system of convection driven by buoyancy source (heat and rain) in the tropics and sink in the sub-Antarctic. The surface layer is thickened in the tropics and runs downhill to the subtropics, where the added mass sits on the lower layer, which is thus pushed equatorward. This convection process is prevented in mid-ocean by the constraint of rotation. Apparently the presence of the continental slope allows this constraint to be broken, though no theory of how will be given here.

The observations reported here shed some light on why this eastern boundary current is so different from the other four. Thompson and Veronis (1983) suggested that a longshore wind on the Northwest Shelf might push the Leeuwin Current. This suggestion has been falsified—the wind blew across the Northwest Shelf, and nowhere between Broome and Fremantle did the wind stress have a poleward–longshore component. This was also true during the preceding two-week cruise covering the whole shelf from Darwin to past Broome (R. J. Edwards, personal communication, 1982). The wind stress on the Leeuwin Current itself was 0.1 N m^{-2} against the current, which would be enough to remove all of the poleward component of momentum in Fig. 5 within 5 days.

The mixed layer in the Leeuwin Current is of order 50 m thick (Fig. 1b). A wind stress of order 0.1 N m^{-2} spread over 50 m of water provides an acceleration of 2×10^{-6} m s^{-2}. A geopotential gradient of 0.5 m^2 s^{-2} over 1500 km (Fig. 4a) provides an acceleration of 3.3 $\times 10^{-6}$ m s^{-2}. The geopotential gradient can overcome the wind stress mixed over this depth. In fact, it is sufficient to accelerate the surface current poleward everywhere the mixed layer is deeper than 30 m. The mixed layer depth off Peru is typically 10 m, and it is 15 m off Northwest Africa and off California (Wooster and Reid, 1962). With such shallow wind-mixed layers, a similar wind stress would overcome a similar geopotential gradient, causing an equatorward surface current. (One would expect the remaining geopotential to drive a poleward undercurrent below the shallow mixed layer.) Thus the difference in mixed layer depth is intimately connected with the difference in surface current direction and hence temperature.

The mixed layer depth should be a function of season, with greater depth in winter. Rochford (1969) showed that the mixed layer in the eastern Indian Ocean is typically more than 50 m deep in the winter, but only 25 m in summer. Hence, one might explain the seasonality of the Leeuwin Current without requiring any seasonal change in wind stress: in summer, with a thin mixed layer, the wind can overcome the geopotential gradient and push the surface water equatorward; in winter, with a thicker mixed layer, the wind stress is less effective, and the water moves poleward. Cresswell and Golding (1980) observed advection of tropical water to 32°S starting each April (early winter) and ending about October (early summer), consonant with the suggestion.

Unfortunately, cause and effect are hard to separate. One expects a warm poleward current to form a deep mixed layer in winter. One expects a cold upwelling current (as off Peru) in summer to have a thin mixed layer.

It appears valid to conclude from the observations that the mixed layer in the Leeuwin Current was thick enough to allow the observed geopotential gradient to overcome the wind stress and drive a poleward current. A poleward current was observed, both directly and in the distribution of tracers. The poleward flux of water was approximately 4×10^6 m^3 s^{-1}. There was also an equatorward current below 150 m depth, with a flux of order 5×10^6 m^3 s^{-1}.

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