Pedlosky (1982) described a two-layer model in which a critical layer and a weak potential vorticity gradient are present in the lower layer. The mean shear is slightly supercritical. In the limit of low frequency, slightly dissipative waves, an instability develops, grows, and then feeds back to the mean flow and potential vorticity gradients, resulting in a finite-amplitude state where the potential vorticity gradient in the lower layer (i.e., the critical layer) has been homogenized. Although this is a simple model, more physical arguments, such as Bretherton's (1966), would also suggest that potential vorticity gradients within a critical layer may be especially susceptible to erosion and eventual homogenization due to the large particle excursions in the layer.

Cluster C is located far downstream in a 2500 km long current, and most likely will be observing the finite-amplitude state of any developing waves and resulting potential vorticity gradient. Indeed, no significant down-gradient heat fluxes were observed (Keffer, 1982b).

Two additional questions suggest themselves. First, from Figure 1, it can be seen that the isopycnal \( q_0 = 26.8 \) that is suspected of containing a critical layer at the Cluster C latitude and where \( Q_y \to 0 \), is the same isopycnal that has undergone extensive homogenization within the subtropical gyre (latitudes 20°-36°) due to processes described by Rhines and Young (1982). This may be coincidence or it may be due to the interactions between the two processes.

Second, although the condition \( Q_y \to 0 \) at \( z_c \) removes the requirement for critical layer instability, it is unclear what it implies for baroclinic instability in general.

References

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The 1978 Occurrence of High Sea Surface Salinity in the Eastern Coral Sea

Long-term changes in sea surface salinity in the Coral Sea, including high salinity values occurring in the periods 1957-58 and 1972-73, have been described by Donguy and Henin (1975). These phenomena were related to the El Nino events along the western coast of South America (Donguy and Henin, 1981). During these periods the Intertropical Convergence Zone (ITCZ) was on the equator and the South Equatorial Current (SECC) was particularly noticeable. A zonal ridge is commonly observed near 15°-18°S, inducing an eastward flow on the south side (the South Equatorial Countercurrent, STCC), a westward flow on the north side (the South Equatorial Current, SEC) and the eastward flowing South Equatorial Countercurrent (SECC) which passes through the Solomon Archipelago.

Scarcity of hydrological casts in the Coral Sea prompted the development of data collection using XBT measurements from merchant ships. In this way, data have been collected routinely along shipping lanes between New Caledonia and Japan, New Guinea, Vanuatu and Fiji (Meyers and Donguy, 1980). However, the XBT program only became operational in 1979, so no data are available to infer the current system during 1978 when the high salinities were observed.

The relationship between monthly mean values of dynamic height and sea level in tropical areas (Wyrski, 1980) was exploited to monitor the current system variations in the Coral Sea. Sea level data from three stations were used. Honiara (Solomon Islands) lies in the dynamic trough between the SECC and the
Erosion of Potential Vorticity Gradients by Critical Layers in the Atlantic North Equatorial Current

Bretherton (1966) showed that the presence of a critical layer, which occurs at depth \( z_c \), where \( U(z_c) = c_p \), where \( U(z) \) is the zonal mean flow profile and \( c_p \) is a disturbance phase speed, implies the instability of the flow. At depth \( z_c \), the particle speed equals the phase speed and a given particle is always exposed to the same phase of a wave cycle. Particles which are initially moving north will continue to do so, and conversely, southward moving particles continue to move south. These excursions imply a large north-south excursions of potential vorticity unless the potential vorticity gradient \( \nabla \psi_p \) vanishes at \( z_c \). If \( \nabla \psi_p \neq 0 \), the resulting flux of potential vorticity can only be balanced by growth of the instability.

Using the POLYMODE Array III Cluster C data set, Keffer (1982a) found four independent pieces of evidence for the existence of a critical layer at 300 m depth within the Atlantic North Equatorial Current. (1) The 3.5 cm s\(^{-1}\) westward cross-correlation phase velocity corresponds to the 300 m flow velocity. (2) The primary temperature balance at 300 m is \( T' + UT'_z = 0 \), where \( T' \) is the temperature perturbation measured at the current meter. (3) The moored temperature measurements indicated a maximum eddy potential energy at 300 m. (4) Historical Nansen bottle data from the National Oceanographic Data Center (NODC) indicated a maximum eddy potential energy at 300 m.

Given the existence of a critical layer and the importance of the north-south eddy flux of potential vorticity, it becomes important to ask: What is the mean potential vorticity gradient at \( z_c \)?

Figure 1 is a contour plot of potential vorticity along the GEOSECS cruise track (~50°W) in the western Atlantic from McDowell et al. (1982). Potential vorticity was evaluated from

\[
Q = -\frac{\bar{\rho} \frac{\partial \bar{\theta}}{\partial z}}{\rho \frac{\partial \bar{z}}{\partial z}}
\]

where \( \bar{\rho} \frac{\partial \bar{\theta}}{\partial z} \) is the vertical adiabatic density gradient, \( \rho \) is the density and \( f \) is the Coriolis parameter. In Figure 1, \( Q \) is contoured as a function of surface referenced density anomaly \( \sigma \) and latitude. Such a prescription for \( Q \) is consistent for large scale slow motions where relative vorticity and horizontal components of vorticity are small.

Also shown in the North Equatorial Current area is a curve representing the density at

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**References**


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**FIGURE 1 (Keffer)**

Potential vorticity \((10^{-11} \text{ cm}^{-1} \text{ s}^{-1})\) along 50°W computed from GEOSECS data and contoured as a function of density \((\sigma)\) and latitude. The stippled region is where \( Q < 0 \) and these latitudes will be a likely site of baroclinic instability. Just below this region is a heavy solid line that marks the 300 m isodepth. Note that at 15°N, \( Q \to 0 \) at this depth. This may be due to critical layer homogenization. The dashed line is the winter outcrop.
Monthly mean sea level values recorded at Honiara (9°15.6'S,159°34.2'E), Vila (17°26.4'S,168°11.4'E) and Noumea (22°10.8'S, 166°15.6'E).

SEC. Vila (Vanuatu) is located either on the ridge between SEC and STCC or on the north side of the eastward STCC. Noumea (New Caledonia) has always been observed to be in the STCC.

Because of the uncertain location of the ridge between 15°S and 18°S, the sea level difference is significant only between Vila and Noumea, which are always in the south westward flux. Available time series of mean sea levels for the 1975-81 period are shown in Figure 2. Using harmonic analysis, Wyrtki (1980) showed annual and semi-annual sea level variations are masked by a southward travelling low sea level disturbance which was recorded in late 1977 at Honiara, in late 1978 at Vila and in mid 1979 at Noumea. In order to eliminate the variations of periods shorter than one year, a twelve month running mean was used before computing cross-correlation coefficients between Honiara and Vila, Vila and Noumea, and Honiara and Noumea. The largest correlation coefficients were obtained at lags of 11 months for Honiara-Vila, 5 for Vila-Noumea and 16 for Honiara-Noumea, indicating a southward movement of the low sea level disturbance of about 4 cm s⁻¹.

The geostrophic flow between stations is related to the difference of dynamic heights at these locations, so the relationship between dynamic height and monthly mean sea level allows us to use the difference of monthly mean sea level between two islands to monitor the magnitude of the geostrophic transport. Thus, the southward travelling low sea level disturbance observed in the Coral Sea produces a change in the current strength.

The sea level difference between Vila and Noumea (Figure 1), which is representative of the magnitude of the eastward flowing STCC, shows a very well marked minimum in 1978, which coincides with the sea surface salinity maximum near 21°S-22°S. Normally the eastward STCC carries low salinity water and moves at an average speed of 25 cm s⁻¹. However, in 1978 the rate of flow was considerably weaker, about 5 cm s⁻¹. It therefore seems that the higher salinities observed are a consequence of the smaller volume of low salinity waters being carried to the Coral Sea. The origin of the southward moving low sea level disturbance has not yet been elucidated.

### References


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## Drifting Buoy Clusters in the Atlantic Equatorial Undercurrent

The large horizontal shear of the jet-like Equatorial Undercurrent (EUC) could be the source of high horizontal mixing rates. Contrary to this concept is the occurrence of the narrow (e.g., 200 km wide) high salinity core across the equatorial Atlantic (Katz et al., 1980). Thus, two questions are raised: Are high mixing rates compensated by the horizontal advective tendencies of meridional circulation cells and is horizontal mixing small enough to be neglected compared to the loss due to vertical mixing?

Four experiments with drifting buoy clusters were carried out in the equatorial Atlantic (Table 1) to measure horizontal turbulent diffusion rates. The 21 m² drogues (Figure 1) were located within the EUC. Ship’s radar was used to track the surface buoys. Because of the relatively long (up to 6 hours) hiatus between the satellite fixes, one buoy is arbitrarily chosen from the cluster to be a reference buoy and the other buoys are positioned relative to the reference buoy every 30 minutes. The track of the reference buoy is determined relative to the ship, which itself is positioned by satellite navigation.

**TABLE 1 (Fahrbach)**

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Number of buoys</th>
<th>Start time</th>
<th>Duration</th>
<th>Mean position</th>
</tr>
</thead>
<tbody>
<tr>
<td>CIPREA</td>
<td>6</td>
<td>1700 GMT</td>
<td>1.83 d</td>
<td>00°36'S</td>
</tr>
<tr>
<td></td>
<td></td>
<td>15 AUG 78</td>
<td></td>
<td>03°43'W</td>
</tr>
<tr>
<td>D1</td>
<td>9</td>
<td>1030 GMT</td>
<td>2.08 d</td>
<td>00°30'N</td>
</tr>
<tr>
<td></td>
<td></td>
<td>14 FEB 79</td>
<td></td>
<td>21°25'W</td>
</tr>
<tr>
<td>D2</td>
<td>8</td>
<td>1730 GMT</td>
<td>1.88 d</td>
<td>00°16'W</td>
</tr>
<tr>
<td></td>
<td></td>
<td>18 MAY 79</td>
<td></td>
<td>20°59'W</td>
</tr>
<tr>
<td>D3</td>
<td>5</td>
<td>1800 GMT</td>
<td>1.60 d</td>
<td>00°01'S</td>
</tr>
<tr>
<td></td>
<td></td>
<td>14 JUN 79</td>
<td></td>
<td>21°38'W</td>
</tr>
</tbody>
</table>

Because of the drag on the buoy and on the wire. The error calculated from the observed current shear is about 15%. The measurements were performed to measure horizontal turbulent diffusion rates in the equatorial Atlantic.
agree within this range.

The distribution of the buoy positions within a cluster is described in a Cartesian coordinate system with the origin at the center of gravity of the cluster. During the CIPREA experiment, the spread of the drifters seemed rather isotropic, whereas during D2, the spreading was dominated by strong horizontal north-south shear. These observations indicate the difficulty in separating influences due to space scales greater and smaller than the array size. Molinari and Kirwan (1975) and Okubo and Ebbsmeyer (1976) assumed a horizontally linear varying mean current to which a random turbulent velocity is added. Using the CIPREA and D2 measurements, the equations for the mean velocity components and their first horizontal derivatives are solved by minimizing the turbulent velocity term.

Experiment D1 was carried out between current meter moorings at the equator and at 1°N. The horizontal derivative calculated from the drifters and the moored current meters agreed within the errors. This gives some confidence to the methods. The divergence of the mean field does not differ significantly from zero. This corresponds with the temperature observations that no significant cooling occurred during the experimental period. If we assume that the standard deviation of the drifter coordinates is equivalent to a mixing length and that the turbulent velocity is equal to the turbulence intensity, then the horizontal mixing coefficients can be calculated from the method described by Okubo and Ebbsmeyer (1976). The computed coefficients (Figure 2) agree with those given by Okubo (1971).

References


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Energetic Propagating Sea Level Events Along the Pacific Coast of Mexico

Along low latitude ocean boundaries (including the equator), baroclinic long waves are a preferred mode of oceanic response to transient wind forcing. Poleward propagating disturbances with frequencies of 0.1-0.5 cpd and properties similar to those of internal Kelvin waves have been observed equatorward of 15°S along the Peru coast (Smith, 1978). The source of forcing for these waves—along the coast or in the equatorial waveguide—is not known and is one of the questions prompting current research. As such waves propagate poleward along an eastern boundary, the internal Rossby radius of deformation decreases with increasing latitude. When it is comparable to the offshore scale of the continental margin, the waves are theoretically expected to be more of a hybrid form, retaining some of the characteristics of internal Kelvin waves, while becoming more like topographically supported continental shelf waves. For typical topographic scales this transition should occur in the 10-20 degree latitude range (see, e.g., Allen and Romea, 1980), though Brink (1982) finds that even at lower latitudes the waves are hybrids. Christensen and de la Paz (1982) have observed propagating sea level disturbances along the continental Pacific coast of Mexico between Salina Cruz (16°N) and Guaymas (28°N), and conclude that they have internal Kelvin wave characteristics (see locations in Figure 2). Their results are based primarily on a correlation analysis of year-long sea level records for 1971. From the lagged correlations they find poleward propagation speeds of 190-380 km day⁻¹ (100 km day⁻¹ = 1.16 m s⁻¹). They call attention to the existence of solitary, energetic events of elevation (10-30 cm) whose alongshore coherence and propagation are evident from time series plots. Their work leaves unanswered the questions of how and where these large amplitude disturbances are generated, and how their propagation characteristics agree or disagree with coastal trapped wave theory.

We have examined the 1971 data and find that the energetic events in that year were confined to the summer season (May-October), suggesting they are generated by eastern North Pacific tropical storms that normally occur in the same season. Figure 1 shows the sea level time series for the 1971 tropical storm season, spaced vertically in proportion to station separation. Single, coherent events aligned along straight off-vertical lines can be seen, indicating poleward propagation speeds of the order of 250 km day⁻¹. To document the atmospheric forcing of the events, we examined the eastern North Pacific storm tracks for the 1971 storm season. Compared with other years, 1971 was remarkable for the relatively large number of tropical storms and hurricanes that passed within 600 km of the Mexican coast. Nine tropical storms (most of them of hurricane intensity) coincided temporally and spatially with corresponding sea level events in the Acapulco-Manzanillo region; the tracks for these are shown in Figure 2. All storm centers passed within 500-600 km of the coast. The tracks are only plotted where tropical storm or hurricane force conditions were observed; hence, near-coastal intense storm activity was mainly limited to the region between Guatemal and Manzanillo.

To see if the forcing by tropical storms is further corroborated by coastal wind records, we examine the maximum daily wind speeds reported by Mexican airports at Salina Cruz, Acapulco, Mazatlan and Guaymas. We found that wind speeds significantly above the usual daily maximum had occurred at Acapulco and/ or Salina Cruz, coinciding with the passage of storms and sea level events. The solid circles in Figure 1 show where both nearby tropical storms were reported, and coastal winds were stronger than usual. In contrast, during the periods of sea level event passage at Mazatlan and Guaymas, no significant local wind increases could be found. These results suggest that the energetic sea level events were generated...
Instabilities on the Equatorial Beta Plane

The equatorial ocean and the tropical atmosphere are regions of strong latitudinal current and wind shear, which results in instabilities at low latitudes. Kuo (1978) and Philander (1978) have studied the barotropic and baroclinic instability of zonal equatorial currents, while Dunkerton (1981) has discussed the inertial instability on the equatorial beta plane when the zonal wave number is zero.

This brief note will show that the equatorial Kelvin wave and gravity waves in horizontal shear are unstable. Rossby waves can also be unstable, but will not be studied here. The mean zonal flow U contains a linear shear in the north-south direction, i.e., U(y) = Sy, where S is constant and y is the latitude. A linear shear is always barotropically stable according to the "Rayleigh-Kuo" criterion, thus "filtering out" unstable Rossby waves from our calculations. This linear shear makes it possible to separate variables and formulate our model, which consists of the linearized shallow water equations for an inviscid, stratified fluid on the equatorial beta plane. In nondimensional form, the model equations are (Boyd, 1980):

\[ i
\]

\[ i \nu - (y - T(y))\nu + ik\phi = 0 \]

\[ y\nu + i\nu + \phi = 0 \]

\[ iku + \nu + ik\nu \phi = 0 \]

where k is the zonal wavenumber; \( \nu \) is the Doppler shifted frequency \( (U(y) - c) \); k is the phase speed of the waves; subscript y denotes differentiation with respect to latitude y; \( u \) and \( \nu \) are the zonal and meridional velocities, respectively; \( \phi \) is the height; and \( T(y) = dU/dy = S \).

This model is too simplified to allow a direct comparison with the viscous and diabatic atmosphere or ocean, where the dynamical and physical processes are very complicated. However, in terms of mathematical analysis, the model is sufficient to provide a basic understanding of the mechanisms that trigger instabilities in the equatorial region.

In an effort to reduce the calculations that arise in computing the complex eigenvalues, \( c, \alpha \), of equation (1), different approximations were introduced to gain some insight into the different modes of instability. Note that a phase speed with a positive imaginary part, \( c_{im} \), characterizes an unstable wave that grows as \( \exp(\alpha t) \), where t is time.

The first approximation is the "long wave approximation" (Boyd and Christidis, 1982), which filters out all gravity waves and eliminates the wavenumber \( k \) as an explicit parameter in our calculations. A shooting method was used for the computation of the

References


