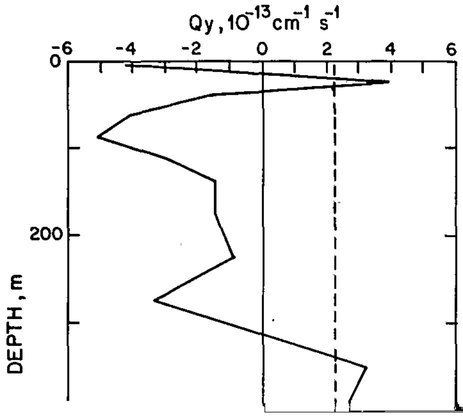
300 m. Recall that it was at this depth, at the Cluster C site (15°N), that a critical layer was observed. Remarkably, $Q_y \rightarrow 0$ at the Cluster C latitude. Indeed, close examination of the $Q_y(z)$ profile (Figure 2) used in Keffer (1982a) to calculate shear modes, shows that the sign reversal happens at 300 m. This comes from a completely independent data set using NODC Nansen bottle data.



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

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

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curves σ_o and σ_p. This suggests that about half of the remaining variability for this model is accounted for by changes in SST between the equator and 30°N.

These conclusions are further supported by a more comprehensive study by Manabe

extratropical interactions, *etc.*) also contribute to the interannual variability of time averages, and boundary forcings alone cannot explain the total observed variance.

It is reasonable to conclude that although for short and medium range the instantaneous

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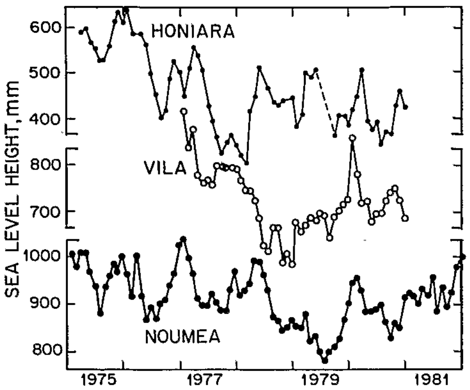

FIGURE 2 *(Henin)*
Monthly mean sea level values recorded at Honiara (9°15.6'S, 159°34.2'E), Vila (17°26.4'S, 168°11.4'E) and Noumea (22°10.8'S, 166°15.6'E).

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 eastward 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 at Honiara and Noumea. These 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.

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Christian Henin
O.R.S.T.O.M.
B.P. A5
Noumea Cedex
New Caledonia

Drifting Buoy Clusters in the Atlantic Equatorial Undercurrent

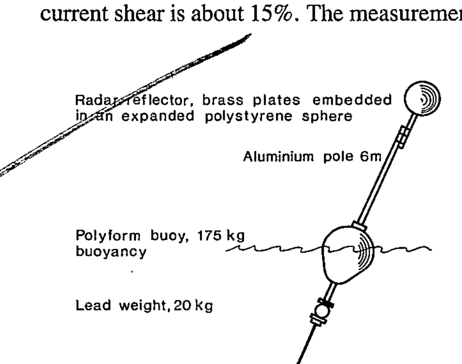
The large horizontal shear of the jet-like Equatorial Undercurrent (EUC) could be the source of high horizontal mixing rates. Contrary to this concept is the occurrence of the narrow (*e.g.*, 200 km wide) high salinity core across the equatorial Atlantic (Katz *et al.*, 1980). Thus, two questions are raised: are high mixing rates compensated by the horizontal advection 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 dif-

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.

During the CIPREA experiment, simultaneous current measurements were carried out from the drifting ship to determine the vertical shear of the current. The current shear in the vertical direction induces an error in the

because of the drag on the buoy and on the wire. The error calculated from the observed current shear is about 15%. The measurements



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Dr. David Halpern
JISAO
University of Washington, AK–40
Seattle, WA 98195 U.S.A.

The **Tropical Ocean-Atmosphere Newsletter (TO-AN)** is published bimonthly by the University of Washington's Joint Institute for the Study of the Atmosphere and Ocean (JISAO), with support from NOAA's Equatorial Pacific Ocean Climate Studies (EPOCS) Program, to provide timely dissemination of information about tropical atmospheric and oceanic research studies. Uncopyrighted articles and diagrams may be reprinted with appropriate citation. Material for publication, comments, and requests for subscriptions may be sent to Dr. David Halpern, JISAO, University of Washington, AK–40, Seattle, WA 98195, USA.

Meetings

AGU Fall Meeting, 7-15 December 1982, San Francisco

Sessions of interest are (a) the SEAREX program, (b) upper ocean dynamics and (c) equatorial dynamics. Contact American Geophysical Union, 2000 Florida Avenue N.W., Washington, DC 20009 (tel: 202-462-9603).

Second Conference on Climate Variations, 10-14 January 1983, New Orleans

Sessions of interest are (a) climate modeling and prediction, (b) climate variations, (c) the Southern Oscillation and (d) tropical sea surface temperature variations. Contact American Meteorological Society, 45 Beacon Street, Boston, MA 02108 (tel: 617-227-2425).

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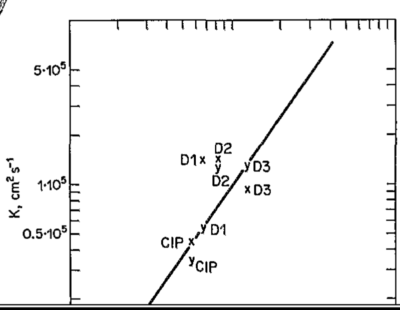
agree within this range.

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 spreading 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 Ebbesmeyer (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 curred 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 Ebbesmeyer (1976). The computed coefficients (Figure 2) agree with those given by Okubo (1971).



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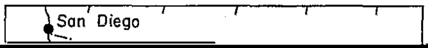

there significant seasonal differences in the oc-

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