Equatorial Upwelling Events in the Central Pacific[1]

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ABSTRACT

Observations made during the Hawaii to Tahiti Shuttle Experiment allow the study of the time and space

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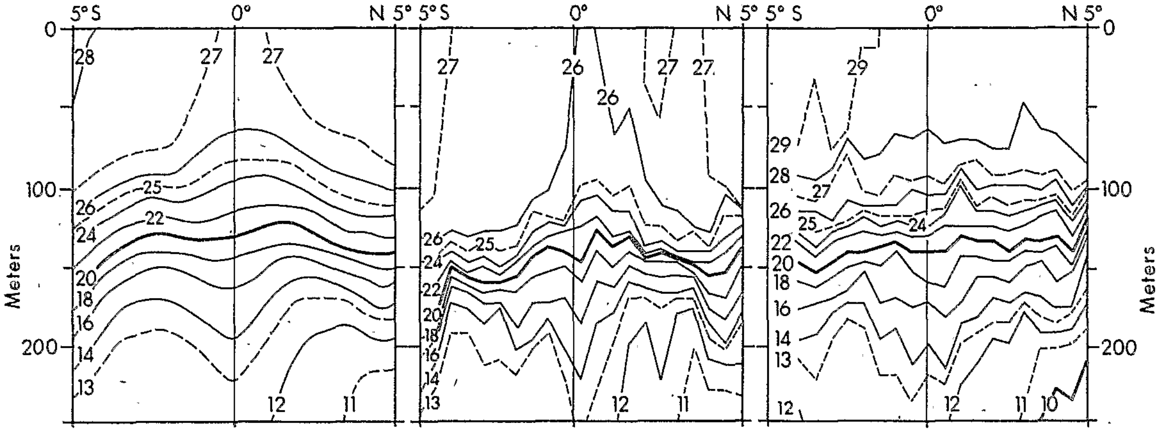

FIG. 1. Cross-equatorial temperature sections obtained during the Shuttle Experiment. (a) Average temperature (°C) at 150°W; (b) temperature (°C) during an intense upwelling event at 150°W observed on 13 January 1980; (c) absence of upwelling at 158°W on 7 June 1980. Temperature profiles are at intervals of one-half degree of latitude.

at the three moorings demonstrate that fluctuations in the wind field are coherent between the two sites and that the same events are obvious in both records. Wind direction is almost always to the west, turning north and south only if wind speed is less than 1 m s^{-1}. The average wind speed during the 18 months of record is only 5.2 m s^{-1} and the maximum is 10 m s^{-1}, which is practically the same as the climatological average. There is no apparent annual signal in the wind records and high-frequency variability dominates the time series. Several bursts of strong wind, lasting 10–20 days can be noted.

The sea-level record is less noisy, and shows a weak annual signal, with sea level lower in April, May and June and higher from October to January, in agreement with data from other equatorial stations. Superimposed are several peaks and troughs lasting a few weeks and having a magnitude of about 10 cm. Surface temperature records from Jarvis and from the buoys show a weak annual cycle with temperatures lower in February and March and higher from June to November. Superimposed are equally strong irregular variations. To characterize upwelling, we have plotted the minimum of sea surface temperature observed within 1° latitude from the equator and the shallowest depth of the 26°C isotherm. Fluctuations of minimum temperature agree very well with temperature at Jarvis, but minimum temperatures are almost 1°C lower, which is partly due to choosing the minimum of temperature for each section, and partly due to the mean east–west temperature gradient.

The parameter most relevant to upwelling is probably the depth of the 26°C isotherm, which fluctuates from its mean position at 65 m upward to the surface and downward to 100 m. Near the equator a well-defined mixed layer is usually not present. The 26°C is obvious in its depth, in contrast to the behavior of deeper isotherms. The 20°C isotherm, situated in the center of the thermocline near the core of the undercurrent, and the 14°C isotherm follow the vertical displacement of the undercurrent. The depths of these isotherms clearly exhibit an annual cycle (Lukas and Firing, 1981), which Meyers (1979) attributes to the annual forcing by the wind stress.

4. Upwelling events

An upwelling event might be caused by an increase of the wind, lasting long enough to cause the ascent of cooler subsurface waters to the sea surface. Such an event should be discernible by isotherms rising above their previous depth. An increase in the wind stress will cause a stronger Ekman transport away from the equator, which should result in a drop of sea level at the equator, and upwelling of subsurface water, which in turn will uplift subsurface isotherms and decrease the sea surface temperature. The speed of such uplifting might give a first estimate of the vertical velocity associated with such upwelling events. If the events are rapid and short, the effects of vertical mixing are probably small compared to those of vertical advection, and entrainment may not be an important factor.

During the Shuttle Experiment several upwelling events can be identified (Fig. 2). From April to June 1979 winds are weak, sea surface temperature rises slowly, and the depth of the 26°C isotherm increases slowly to about 70 m in early June. In late June a three-week period of increased winds follows, causing a drop of sea level at Jarvis in the last days of June and a minimum of sea surface temperature a few days later. The minimum surface temperature is observed

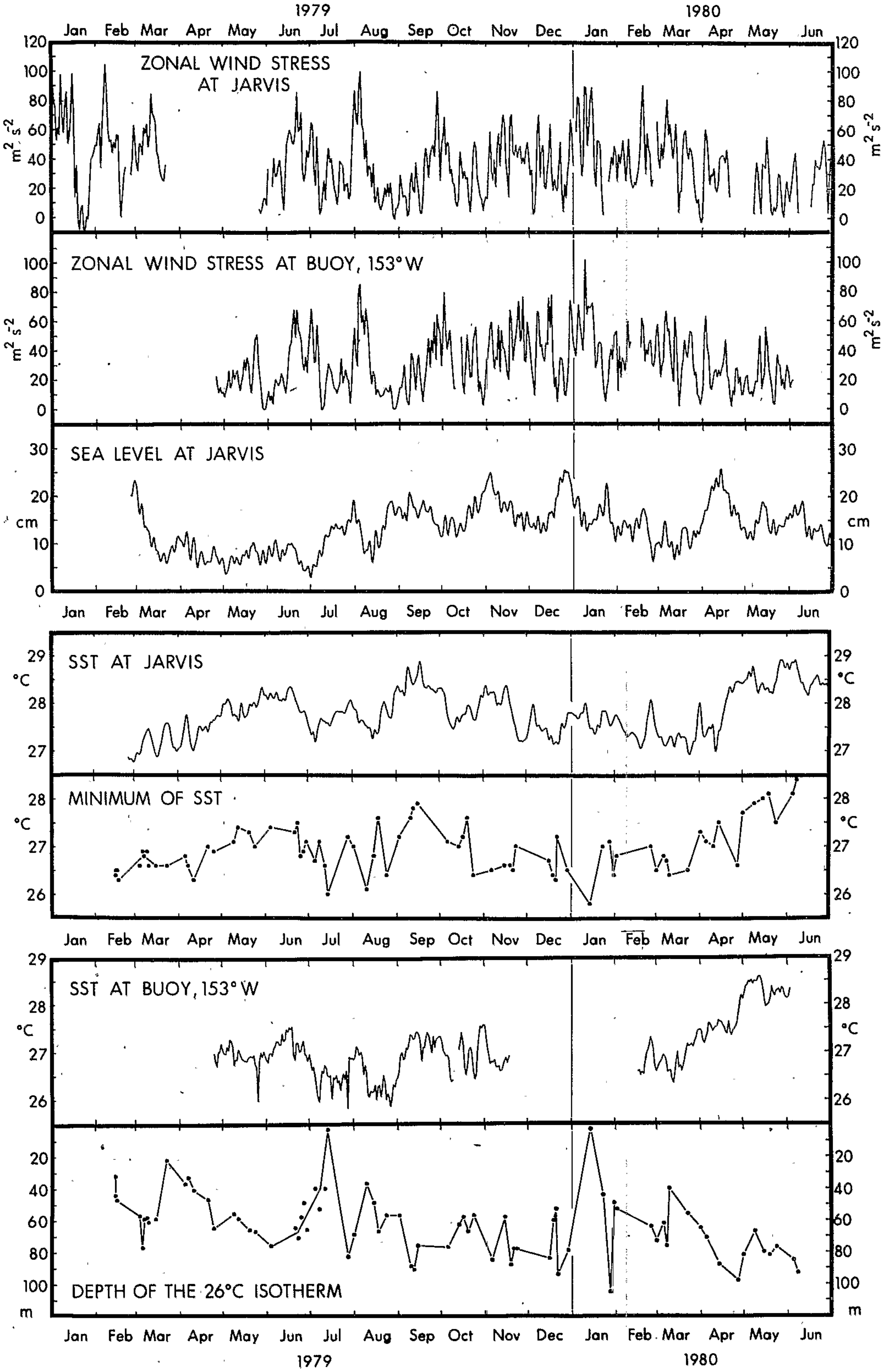

FIG. 2. Time-series of parameters related to equatorial upwelling events including zonal component of wind stress ($m^2\ s^{-2}$) at Jarvis Island (a) and at the buoy cluster at 153°W (b); sea level at Jarvis Island (0°23′S, 160°02′W) (c); sea surface temperature (°C) at Jarvis Island (d); the minimum of sea-surface temperature (°C) observed on each temperature section (e); sea surface temperature (°C) at the buoy cluster at 153°W (f); and highest position of the 26°C isotherm near the equator (g).

on 23 June at 67 m. During this event it rises by 67 m in 20 days with an average speed of about 3 m per day. Immediately after the upwelling event the 26°C isotherm drops to 83 m at 158°W and to 70 m at 153°W, and sea level rises again.

In early August another burst of the winds lasts for about 10 days. Sea level immediately drops, reaching a minimum on 10 August. The 26°C isotherm rises to 37 m as observed on 9 August at 150°W on cruise 6, producing a surface temperature minimum, which is also seen in the records of the buoys at 153°W and at Jarvis. The observed isotherm rise by about 30 m during 10 days again gives an upwelling speed of 3 m per day. Because the wind burst was of shorter duration, the 26°C isotherm did not reach the sea [illegible]

is probably a result of the wind burst in the first half of March. Sea level at Jarvis also drops sharply. The shallow position of the 26°C isotherm during the first cruise is probably due to the wind burst in February, but supporting data are insufficient to analyze other aspects of this event.

The thermal structure observed during flight 24 along 150°W shows the 26°C isotherm at only 9 m at 1.5°N indicating thermocline uplifting (Stroup *et al.*, 1981). Similarly, during flight 32 along 150°W the 26°C isotherm was observed at 15 m at 0.5°N. As deeper isotherms are uplifted by about the same amount, we believe that these particular AXBT drops were failures and should have been deleted in the data processing. Therefore, we have not used these obser- [illegible] cool water previously drawn to the surface drops back to the top of the thermocline and usually overshoots its previous equilibrium depth, indicating a wave-like response. Such a response is especially obvious after the events in July 1979 and January 1980 (Fig. 2). As upwelling ceases, sea level rises.

Upwelling water during these events comes definitely from the uppermost portions of the thermocline since little or no upward movement of water only 2°C cooler can be detected. This situation appears reasonable, as water in the very weakly stratified layer above the thermocline can be more easily advected vertically than the heavier water of the thermocline. The upwelling events seem to be restricted to the weakly stratified surface layer and they turn over water with a temperature difference of less than 2°C. For the mean climatological situation the temperature of the source water for upwelling was determined to be 3–4°C cooler than the water north

the cool tongue from the east into the region and to the ocean-wide changes in the wind field (Meyers, 1979). The variability of the wind is much stronger on time scales of a month or less than at the annual period. Consequently, the observed upwelling events must be considered as a local response to winds of limited duration.

While this study reveals the existence of equatorial upwelling events and describes their general characteristics, much remains to be learned. The events are apparently so fast that the spacing of hydrographic sections in time was the main limiting factor for studying the events in greater detail. Moreover, only five events occurred in 1½ years. Any more systematic study will require the use of clusters of moorings with continuously recording instruments over a long period. The large east–west scale of the phenomenon together with the need to sample strong north–south gradients will require a substantial experimental effort

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