Equatorial Upwelling Events in the Central Pacific[1]

KLAUS WYRTKI AND GERARD ELDIN[2]

Department of Oceanography and Hawaii Institute of Geophysics, University of Hawaii, Honolulu 96822

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ABSTRACT

Observations made during the Hawaii to Tahiti Shuttle Experiment allow the study of the time and space scales of equatorial upwelling. Individual upwelling events can be identified. Each is caused by a burst in the trade winds lasting from 10 to 20 days. Sea level drops, surface temperature decreases, and near-surface isotherms rise by several tens of meters. The vertical velocity of upwelling is at least 3 m per day. Five such upwelling events were observed during the 18-month experiment.

1. Introduction

The existence of equatorial upwelling in the Pacific and the other oceans has been known for a long time and its basic reasons have been discussed by Cromwell (1953) and Knauss (1963). Upwelling manifests itself by a tongue of cool water along the equator stretching from the Galapagos Islands to the date line (Robinson, 1976). The upwelling is caused by the divergence of the surface Ekman drift at the equator under the permanent easterly trade winds. This divergence is balanced by a convergence of geostrophic flow due to the east–west pressure gradient along the equator. Because the vertical distribution of the two meridional flows is different, a strong vertical circulation results, which leads to equatorial upwelling. The average rate of upwelling in the Pacific has been estimated as about 50×10^6 m^3 s^{-1} (Wyrtki, 1981). This steady state may be disturbed by fluctuations of the wind and by the passage of equatorially trapped internal waves. Little observational evidence has so far been collected about the time and space scales of the response of equatorial upwelling to such disturbances. The observations made during the Hawaii to Tahiti Shuttle Experiment (Wyrtki *et al.*, 1981) allow a first description of such upwelling events.

2. Data

Cross-equatorial hydrographic sections were obtained by the Shuttle ship at 150, 153 and 158°W on 15 cruises between February 1979 and June 1980 (Taft and Kovala, 1981). In addition, 35 temperature sections were taken by aircraft (Stroup *et al.*, 1981). Winds were continuously recorded at a cluster of three moorings near 153°W (Knox and Halpern, 1981) and at an automatic station on Jarvis Island at 160°W (Vitousek *et al.*, 1980). The three moorings also measured sea and air temperatures and subsurface currents. Sea level and temperature at a depth of 5 m were recorded at Jarvis Island with a pressure recorder.

3. Equatorial upwelling

Evidence of equatorial upwelling can be found on virtually all temperature sections, but the apparent intensity of upwelling, indicated by the intensity of thermocline uplifting, varies considerably. The average temperature distribution at 150°W (Fig. 1a) shows cool upwelled water of temperature less than 27°C at the equator flanked on both sides by water of temperature more than 28°C. All isotherms above 20°C are uplifted relative to their depth away from the equator. The 26°C isotherm is uplifted by about 30 m. The upwelling area defined by the isotherm doming is about 2° latitude wide. A case of very strong isotherm uplifting at 150°W during cruise 11 in January 1980 is illustrated in Fig. 1b. Water of temperature less than 26°C has reached the sea surface in a thin band, but the entire area between 4°S and 2°N is occupied by water cooler than 27°C. The only case during which no upwelling was evident is seen during cruise 15 along 158°W, when a homothermal layer 70 m deep with temperatures above 28°C stretched from the Southern Hemisphere across the equator to 7°N into the Countercurrent (Fig. 1c).

To evaluate the development of equatorial upwelling in time and to discuss the sequence of events, time series of several pertinent parameters are shown in Fig. 2. The east–west component of wind stress derived from the wind recorders at Jarvis Island and

[1] Hawaii Institute of Geophysics contribution No. 1291.

[2] Visiting scientist from O.R.S.T.O.M. (Office de la Recherche Scientifique et Technique Outre-Mer), France.

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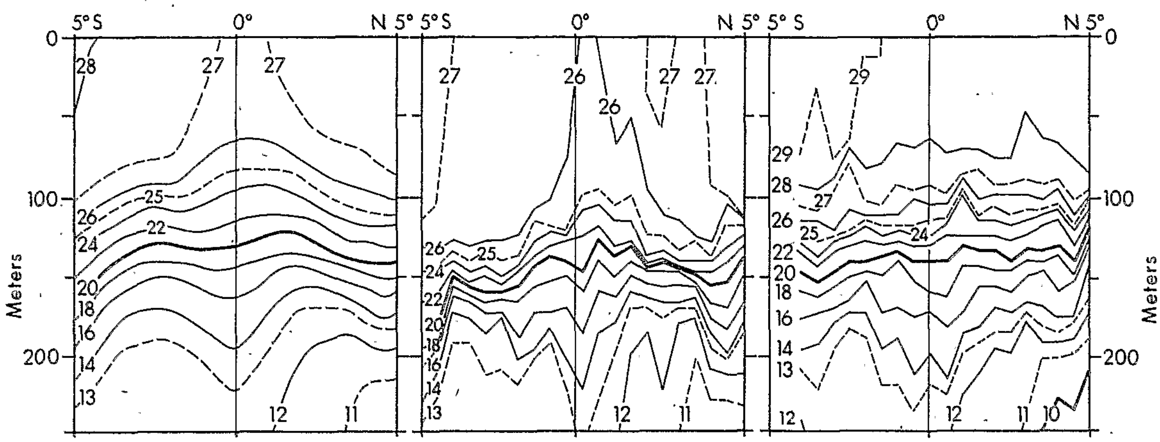

FIG. 1. Cross-equatorial temperature sections obtained during the Shuttle Experiment. (a) Average temperature (°C) at 150°W; (b) temperature (°C) during an intense upwelling event at 150°W observed on 13 January 1980; (c) absence of upwelling at 158°W on 7 June 1980. Temperature profiles are at intervals 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 isotherm slowly descends from an average depth of about 50 m during the beginning of the Shuttle Experiment to about 80 m near its end. No annual signal 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 on 12 July at 153°W on Flight 17 and is also recorded at the buoys; the 26°C isotherm reaches the sea surface. On 19 June the same isotherm is at 65 m and

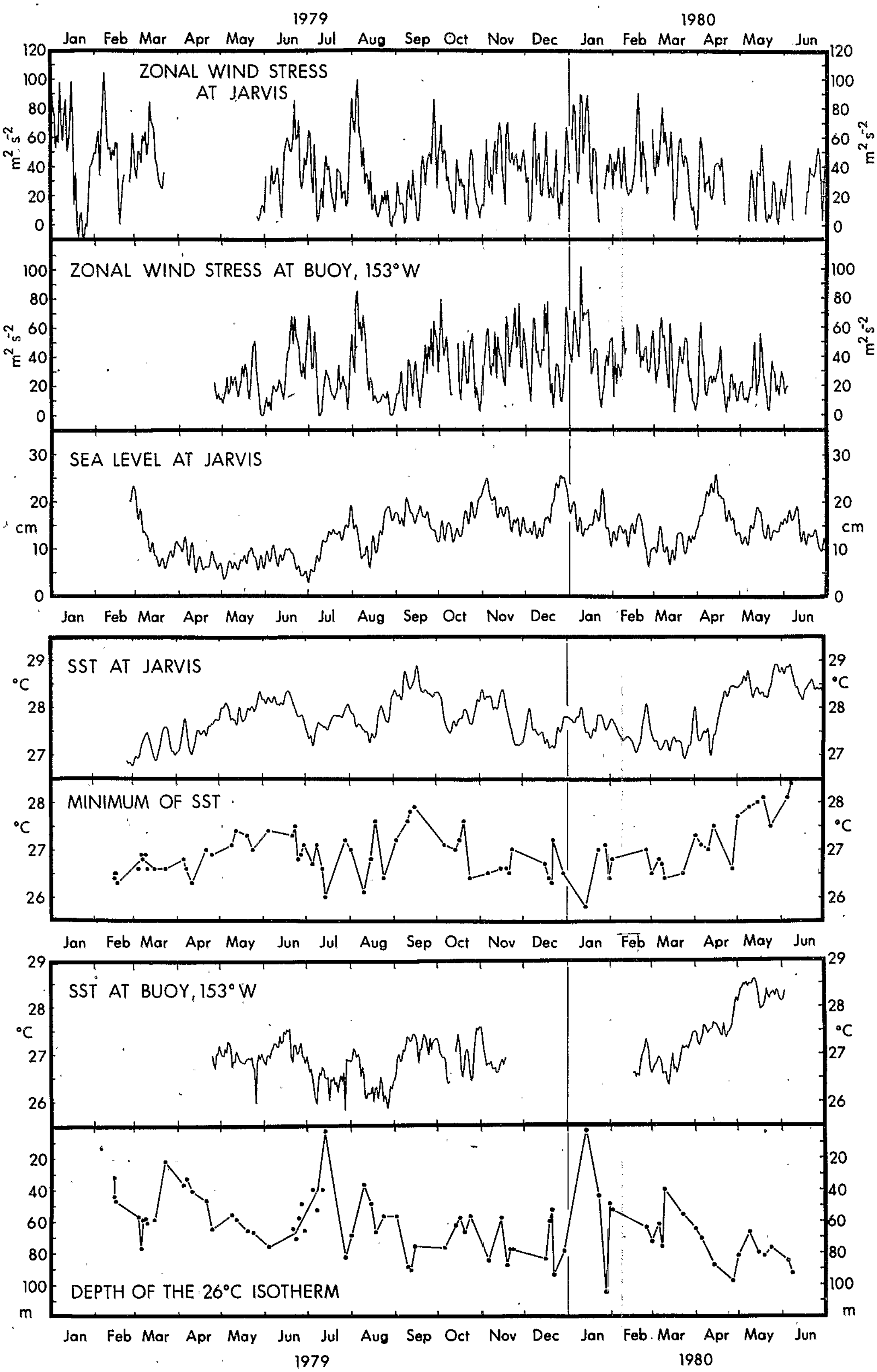

FIG. 2. Time-series of parameters related to equatorial upwelling events including zonal component of wind stress (m^2 s^{-2}) at Jarvis Island (a) and at the buoy cluster at 153°W (b); sea level at Jarvis Island (0°23′S, 160°02′W) (c); sea surface temperature (°C) at Jarvis Island (d); the minimum of 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 surface.

After a period of very weak winds in August, wind speeds increase during September and reach a peak at the end of the month. This peak coincides with a decrease in sea level and is followed by a minimum in sea surface temperature in early October. In response, the 26°C isotherm rises by about 20 m. In October, November and early December wind speeds remain above average but fluctuate strongly; surface temperature decreases slowly, but the depth of the 26°C isotherm increases, probably as part of the seasonal deepening of the undercurrent core. The wind fluctuations during this period seem to be too fast for the ocean structure to respond in an organized and measurable fashion.

During the second part of December winds become weak and sea level rises immediately. The next burst of winds in early January lasts for about two weeks. Sea level drops by about 12 cm, reaching a minimum on 9 January. Surface temperature decreases to less than 26°C at 150°W as observed on leg 11 on 13 January. On 28 December the 26°C isotherm is found at 79 m at the same position, and consequently it rises by 79 m during 16 days with an average speed of about 5 m per day. Unfortunately, no other temperature sections were obtained during this period, and surface temperature sensors at all three buoys had failed. In the second half of January the winds decrease and the 26°C isotherm drops as fast as it had risen, reaching 44 m on 22 January at 153°W and 105 m on 27 January at 158°W.

During the remainder of the Shuttle Experiment winds decrease steadily, surface temperature rises to more than 28°C, and the 26°C isotherm drops to more than 80 m. Sea level remains high, about 10 cm higher than a year earlier. During the last cruise a thick surface layer warmer than 28°C had covered the upwelling at 158°W (Fig. 1c), but also at 153 and 150°W there was little indication of upwelling.

During the early part of the experiment wind observations are unfortunately interrupted. The records at Jarvis show three strong bursts of wind in the first three months of 1979. On 22 March 1979 the 26°C isotherm reaches 22 m at 150°W on cruise 2, which 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 observations in our analysis. Moreover, similar isotherm uplifting was not observed along 153°W on flights 25 and 33, which were only two days after flights 24 and 32, respectively.

5. Discussion and conclusions

During the Shuttle Experiment five discrete upwelling events were observed over a period of 18 months. Each event was caused by a burst in the prevailing easterly trade winds, lasting from ten days to three weeks. These bursts usually followed a period of noticeably weak winds, and wind stress during the bursts was typically twice the long-term mean. For three of the five cases the speed of upwelling could be determined from the rise of the 26°C isotherm as 3–5 m per day. This speed represents a lower limit, because it might have been greater if observations had been more frequent, because vertical mixing might partly offset the vertical advection of isotherms and because the upwelling event is superimposed on the mean climatological upwelling. Each upwelling event is also accompanied by a drop in sea level, indicating that water is being removed from the equator by divergent Ekman flow. The speed of upwelling and the depth of the cooler water before upwelling determine the time after which upwelling becomes evident at the sea surface by the appearance of cooler water; this time is typically 10–20 days. During the Shuttle period strong winds never lasted for more than 20 days and consequently no case of prolonged upwelling could be observed.

Halpern (1980) has estimated the vertical velocity of upwelling from measurements with an array of current-meter moorings at the equator and 110°W. He finds average upwelling of 3 m day^{-1} and large fluctuations. A three-month long time series of vertical velocity also indicates an event scale of 10–20 days.

The end of each upwelling event is determined by the decrease of the winds. This reduces the Ekman divergence and allows water to converge to the equator under the influence of the north–south pressure gradients created by the low equatorial sea level. The 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 and south of the cool tongue (Wyrtki, 1981).

The north–south scale of upwelling on climatological time scales is given by the width of the cool tongue, a value of about 400–500 km. But the event scale appears to be much narrower. During all the short events observed, the cool upwelling water was obvious at only one station. Since stations were 55 km apart, the strongest upwelling was probably concentrated in a band less than 100 km wide. This has probable dynamical reasons still to be explored.

Even less can be said about the east–west scales of upwelling events. The fluctuations of the east–west component of the equatorial trade winds are certainly coherent over distances greater than 10° longitude, even on time scales of 10–20 days (Sadler and Kilonsky, 1981). The good correlation of sea level at Jarvis with the upwelling events at 150°W, 1000 km away, is another indicator for a large zonal space scale of these events. On the other hand, it does not seem accidental that four of the five upwelling events were observed at 150°W, one at 153°W and none at 158°W. This may be partly due to surface temperature at 158°W being about 0.6°C higher and the surface layer above the 26°C isotherm being 14 m thicker than at 150°W. Consequently, upwelling will be less obvious at 158°W and winds will have to last longer than at 150°W to produce a similar effect. Such long bursts of winds apparently did not happen during the Shuttle period.

One might also consider the possible effect of equatorial waves on upwelling. Such waves will certainly be generated in response to a variable wind stress, and in particular in response to a wind patch. These equatorial waves, however, have maximum vertical speeds in the thermocline and not near the sea surface above it.

The conditions on longer time scales might also be considered. The hydrographic structure, sea level and sea surface temperature have a clearly developed annual signal that is related to the annual advance of 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 for any experiment to study the variability of equatorial upwelling.

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