Modeling the Oceanic Response to Westerly Wind Bursts in the Western Equatorial Pacific

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(Manuscript received 4 April 1996, in final form 30 January 1998)

ABSTRACT

A 3D tropical upper-ocean circulation model is employed to study the generation mechanism for subsurface reversing currents forced by strong westerly wind bursts in the western equatorial Pacific. The westerly wind bursts last from a few days to a few weeks and reverse the surface current from westward to eastward, setting up a zonal pressure gradient that generates a westward subsurface current. Although less affected by the local wind, the eastward Equatorial Undercurrent (EUC) decelerates. The authors verified the hypothesis that the subsurface reversal is a local response to westerly wind bursts. However, the model response is very sensitive to the wind fetch. The observed reversal of the subsurface current can only be reproduced by wind bursts with a zonal extent of less than 700 km. This is because the zonal extent determines the timescale on which the pressure gradient is set up by equatorial waves. This timescale must be shorter than the timescales of vertical mixing and downwelling in order for the pressure-driven subsurface westward acceleration to overtake the eastward acceleration due to downward transport of momentum. The results emphasize the importance of resolving spatial variations in simulating the upper ocean response to atmospheric forcing.

The influence of off-equator winds on equatorial currents is also investigated. It is found that the remote effect of an off-equator wind can be larger on subsurface currents than on surface currents. An off-equator westerly (easterly) wind decelerates (accelerates) the EUC but has little effect on the surface equatorial current. When a cyclone with a westerly wind at the equator and an easterly off-equator wind is present, the local response near the surface is dominated by the westerly wind. But the remote effect of the off-equator wind significantly modifies the local wind effect at the depth of the EUC.

1. Introduction

Easterly trade winds prevail over most of the equatorial Pacific, generating a westward South Equatorial Current (SEC) at the surface and an eastward pressure gradient force (PGF) along the equator. The PGF in turn drives a downgradient eastward Equatorial Undercurrent (EUC) in the thermocline. During boreal winter, westerly winds from the Asian monsoon replace the trade winds over the western equatorial Pacific. Superimposed on this seasonal variation are sporadic and energetic westerly wind bursts that reflect different processes, for example, the Madden–Julian oscillation. The first hydrographic measurements under strong westerly winds were made by Hisard et al. (1970) along the 170°E meridian. They reported a vertical profile consisting of three currents: an eastward surface current in the direction of the winds, a preexisting eastward EUC (presumably maintained by the basin-scale PGF), and a westward subsurface reversing current (SSRC) between the two eastward currents. The principal goal of this study is to understand the SSRC generation mechanism through the use of a 3D tropical upper-ocean circulation model.

The SSRC is located below the surface mixed layer in the upper reaches of the main thermocline. The results of Hisard et al. (1970) provide little temporal information on the SSRC and neither the evolution of the wind forcing nor that of the local oceanic response were well resolved. More recent observations (McPhaden et al. 1992) obtained from the TOGA TAO Array mooring at the equator, 165°E, present simultaneous measurements of zonal surface wind, zonal velocity, and tem-
FIG. 1. Daily average of (a) zonal wind, (b) zonal currents at depth, (c) 10 m and (d) 0–300 m temperatures from a mooring located at 0°, 165°E during the period November 1989–January 1990. The contour interval in (c) is 20 cm s⁻¹ with westward flow shaded, and eastward flow larger than 40 cm s⁻¹ hatched (Fig. 4 in McPhaden et al. 1992). (e) Daily averages of the surface zonal winds between 5°S and 8°N along 165°E (Fig. 3 in McPhaden et al. 1992).
associated with vertical shear balances the surface wind stress. However, the surface layer has constant thickness, and it is decoupled from the deep ocean in this simple model. McPhaden et al. (1988) improved the model by coupling the surface layer with the deep ocean through a perturbation displacement at the interface. Their model explains the large vertical shear in the surface layer but it fails to resolve the observed three-layer profile because it does not take the deep ocean into account.

A few studies have focused on energy propagation in the equatorial ocean. Anderson and Gill (1979) found that on a β plane, the inertial wave energy does not remain where it is produced but moves as a group back and forth across the equator. Their studies were followed by Schopf et al. (1980), who investigated the dispersion of low-frequency internal Rossby waves due to changes of deformation radius caused by the β effect. They found that the ray paths for the equatorial Rossby waves are sinusoidal, with their energy channeled initially toward the equator. Motivated by the deep currents, McCreary (1984) studied the response of a linear stratified equatorial ocean to low-frequency equatorial winds. He showed that energy propagates vertically as well as horizontally. All of the above studies describe mechanisms by which the equatorial current system can be affected by remote winds.

Other numerical studies focused on the remotely forced oceanic response to westerly wind bursts. Hindcasts of the 1982–83 ENSO event were performed by Harrison and Craig (1993) using a primitive equation ocean circulation model forced with climatological monthly mean winds. They showed that local zonal wind stress variations could qualitatively explain changes in the upper-ocean zonal flow. They also demonstrated the possibility that westerly wind events in the western equatorial Pacific account for the variability not explained by local forcing in the central equatorial Pacific. Their results, however, did not provide an explanation for temperature changes. Kindle and Phoebus (1995) numerically simulated the Eastern Pacific sea level variations to westerly wind bursts prior to, and during, the 1991–92 El Niño using reduced gravity models. They found that Kelvin waves generated by the westerly winds were responsible for sea level variations in the Eastern Pacific and that these westerly wind events were associated with the development of tropical cyclones in the Western Pacific. Neither of these two studies, however, examined the effect of off-equator winds on the equatorial current system.

The three principal objectives of our study are 1) to examine the upper-ocean response of the western equatorial Pacific to local westerly wind bursts; 2) to investigate the generation mechanism of the SSRC; and 3) to examine the effect of off-equator winds. The rest of this paper is organized as follows. Section 2 is a brief description of the model that will be used to carry out the numerical experiments. In section 3, the response of an initially resting ocean to a subbasin-scale wind burst and to easterly and westerly basin-scale winds is studied and the basic dynamics are established. In section 4, the initial circulation is set by basin-scale easterly winds, and the response of this ocean to westerly wind bursts of different fetch and location is investigated. The summary and discussion are presented in section 5.

2. Model description

a. The model physics

A general circulation model originally developed by Gent and Cane (1989) is employed to investigate the aforementioned objectives. The model ocean is (a) assumed to be hydrostatic, (b) employs the full primitive equations, and (c) includes a fully active heat equation. The model consists of a surface mixed layer and an active upper ocean; the deep model ocean is at rest since it is presumably not important to the timescales we are considering here. A new vertical mixing algorithm is implemented to relax the unrealistic limitation of constant depth of the surface mixed layer in the original implementation. This modification leads to an improved simulation of the large-scale equatorial circulation (Chen et al. 1994). In the new algorithm, the surface mixed layer depth is calculated using the Kraus–Turner bulk mixed layer model. Vertical turbulent mixing in the interior is determined by a dynamic instability criterion based on the Richardson number. Weak background vertical mixing for momentum and heat is set below the surface mixed layer in order to achieve computational stability (Table 1). A more complete discussion of this background mixing can be found in Gent and Cane (1989).

The density is chosen as a function of temperature only, and the state equation is

$$\rho = \rho_0 (1 - \alpha T),$$

where $\rho$ is the density, $\rho_0$ the reference density, $T$ the temperature, and $\alpha$ the coefficient of thermal expansion. This study focuses on the oceanic response on timescales $O(10 \text{days})$, consistent with the observed duration of the westerly wind bursts (Fig. 1). For the sake of

| Domain $(x \times y \times z)$ | $50^\circ \times 40^\circ \times 400 \text{m}$ |
| Grid points $(x \times y \times z)$ | $120 \times 60 \times 13$ |
| Finest horizontal resolution | $\frac{1}{3}^\circ \times \frac{1}{3}^\circ$ |
| Initial vertical resolution | $10 \times 20 \text{m}, 2 \times 50 \text{m}, 100 \text{m}$ |
| Shapiro filter | Order 8 every $\frac{1}{3} \text{day}$ |
| Lorenz N-cycle | $N = 4$ |
| $\Delta t$ | 1 h |
| BINT | $0.5 \times 10^{-4} \text{m s}^{-1}$ |
| CINT | $0.5 \times 10^{-4} \text{m s}^{-1}$ |
simplicity, salinity is not regarded as an active component in determining the density field.

**b. The model geometry**

The model is confined to a rectangular basin, centered at the equator, with a zonal extent of 50° and a meridional extent of 40° (Fig. 2). Extratropical regions are included in order to limit the influence of the boundaries on the equatorial region. The model contains 120 × 60 grid points in the horizontal plane (Table 1), and these are unevenly stretched in both zonal and meridional directions to provide finer resolution along the equatorial waveguide and near the 14.2° meridian (Xc), where the center of a westerly wind burst will be placed.

The model uses a sigma coordinate in the vertical direction and the active upper ocean is divided into 13 layers. Eighth order centered differencing is used in the horizontal and second order in the vertical. An order-8 Shapiro filter is applied every eight time steps for horizontal smoothing (Table 1). To improve the accuracy of time integration, a Lorenz 4-cycle scheme is employed. More detailed justification for using the Shapiro filter and the Lorenz scheme can be found in Gent and Cane (1989). The integration time step is one hour.

**c. Initial and boundary conditions**

The model ocean is initially at rest and the temperature field is horizontally uniform throughout the basin. The bottom of the active portion of the model ocean is initially placed at 400-m depth, and the ocean above it is divided into ten 20-m layers (including the mixed layer), two 50-m layers, and one 100-m layer (Table 1). The initial temperature profile is based on measurements obtained from the Hawaii–Tahiti Shuttle CTD stations between 5°S and 5°N, 150°W and 158°W (Stroup et al. 1981). This central equatorial Pacific profile is chosen because it represents average conditions and because it provides a reasonable initial basin-scale state for subsequent westerly wind burst experiments. No-slip and no-heat-flux boundary conditions are applied at all lateral boundaries. Near the zonal boundaries, a reduced-order Shapiro filter is applied for horizontal smoothing.

**d. Atmospheric forcing**

The ocean is driven by momentum and heat fluxes. There are few wind field datasets available in this region and their resolutions are too low to capture the temporal and spatial structure of westerly wind bursts (Harrison and Giese 1991). We therefore use an idealized wind burst in an attempt to gain insight on the fundamental physics of the oceanic response to westerly wind bursts at the expense of realistic simulations. Analysis of data obtained during COARE IOP will eventually shed light on the detailed structure of westerly wind bursts.

Strong winds above the warm pool are primarily westerly. In addition, idealized experiments have shown that the meridional wind stress has a relatively small impact.
on zonal oceanic currents (Harrison and Craig 1993). Therefore, in all experiments we impose only the zonal component of the wind field. Two types of steady winds are used: basin-scale winds with a duration of one year to represent the basin-scale easterlies and westerlies; and subbasin-scale winds with a much shorter duration, 10 days, to represent the energetic westerly wind bursts (Table 2). The basin-scale winds are uniform meridionally and sinusoidal zonally (a half sine wave) and are expressed as follows:

$$\tau^*_x(x) = \tau_0 \sin(\pi x/L_x),$$

(2)

where $\tau_0$ is the amplitude and $L_x$ is the zonal extent of the model ocean. The subbasin-scale wind bursts are simulated as Gaussian in both zonal and meridional directions; that is,

$$\tau^*(x, y) = \tau_0 \exp \left[ -\frac{(x - X_0)^2}{\Delta x} \right] \exp \left[ -\frac{(y - Y_0)^2}{\Delta y} \right],$$

(3)

where $(X_0, Y_0)$ is the center of the wind burst and $(\Delta x, \Delta y)$ are the Gaussian half-widths in zonal and meridional directions, respectively. In the following experiments, $\Delta y$ is prescribed as the equatorial deformation radius for the first baroclinic mode, which is about 350 km for the present model configuration, and $\Delta x$ is allowed to vary (Fig. 3a). All the winds are switched on and off as a step function. For the prescribed spatial structure and extent, little energy leaks into gravity waves.

The heat flux at the sea surface, $Q$, is parameterized as proportional to the difference between the apparent atmospheric equilibrium temperature ($T_A$) and the sea surface temperature ($T_s$) (Haney 1971)

$$Q = \theta (T_A - T_s),$$

(4)

where $\theta$ is 40 W m$^{-2}$ °C$^{-1}$, and $T_A$ is prescribed as a constant 29°C, which is the annual mean SST in the warm pool as well as the initial SST of the model ocean. This choice of $T_A$ is made to minimize the heat flux at the ocean surface. The annual mean net heating from the ocean surface is estimated to be between 0 and 20 W m$^{-2}$ in the western equatorial Pacific (Gent 1991).

3. Response of an initially resting ocean

In order to illustrate some important aspects of equatorial dynamics and to set a benchmark for subsequent experiments, the behavior of an initially resting ocean to both subbasin-scale and basin-scale winds is studied. First, the oceanic response to a subbasin-scale westerly wind burst (expt 1) is examined, followed by the response to a basin-scale easterly wind (expt 2). Experiment 2 is then contrasted with the oceanic response to a basin-scale westerly wind of identical structure (expt 3) to illustrate the nonlinear behavior of the model. Parameters of wind forcing for these three experiments are listed in Table 2; the model ocean is initially at rest in all three experiments.

a. Basic dynamics

For basic equatorial dynamics, readers are referred to McCreary (1985), who has thoroughly discussed equatorial theory and carefully compared solutions to models of all types. Here we only emphasize some important aspects of equatorial dynamics that will later be used to interpret our model results. We start from the simplest situation.

When a meridionally symmetric and $x$-independent zonal wind stress is switched on over an initially resting ocean in a zonally bounded basin, the oceanic response is zonally uniform in the interior until zonal variations are imposed by Kelvin waves from the western boundary and Rossby waves from the eastern boundary. The

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**Table 2.** Wind forcing parameters for expts 1–3: $\Delta x$ and $\Delta y$ represent Gaussian half-widths in the $x$ and $y$ directions, respectively; $(X_0, Y_0)$ is the central location of the wind; $\tau_b$ is the amplitude of the wind burst in expt 1, and $\tau_E$ is the amplitude of the basin-scale wind in expts 2–3; $X = 14.2^\circ$, and $L_x$ is the width of the model ocean.

<table>
<thead>
<tr>
<th>Expt</th>
<th>Wind type</th>
<th>Zonal structure</th>
<th>Meridional structure</th>
<th>Center $(X_0, Y_0)$</th>
<th>$\tau_b$ or $\tau_E$ (dyn/cm$^2$)</th>
<th>Duration (days)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>wind burst</td>
<td>Gaussian ($\Delta x = 5^\circ$)</td>
<td>Gaussian ($\Delta y = 3^\circ$)</td>
<td>$X_0, 0^\circ$</td>
<td>1.0</td>
<td>10</td>
</tr>
<tr>
<td>2</td>
<td>trade</td>
<td>sinusoidal</td>
<td>uniform</td>
<td>$L_x/2, 0^\circ$</td>
<td>−0.5</td>
<td>365</td>
</tr>
<tr>
<td>3</td>
<td>westerly</td>
<td>sinusoidal</td>
<td>uniform</td>
<td>$L_x/2, 0^\circ$</td>
<td>0.5</td>
<td>365</td>
</tr>
</tbody>
</table>

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![Figure 3](image-url)
meridional velocity at the equator is zero and the zonal momentum equation can be simplified as (Cane 1980)

\[ u_t = -wu_z + v_x u_z, \tag{5} \]

where \( u \) and \( w \) are zonal and vertical velocities, respectively, \( v_x \) is the coefficient of vertical mixing, and subscripts \( t \) and \( z \) represent temporal and vertical derivatives. With a westerly wind, the Ekman convergence induces a downward (negative) \( w \) right at the equator; with \( u \) and \( u_z \) also positive, both terms on the right-hand side of (5) are positive. This means that, below the surface, both vertical advection and turbulent mixing act to transfer the surface’s eastward momentum downward. With an easterly wind stress, the vertical mixing term changes its sign; the advection term, however, remains positive. Advection contributes to reduce the surface westward momentum and it does so by advecting negative shear upward. Therefore, for winds of both directions vertical advection remains positive and gives rise to a positive tendency.

Integrating (5) from the surface to some depth where the surface wind stress \( \tau_z \) disappears, one obtains

\[ \frac{\partial}{\partial t} \int u \, dz = \tau_z + \int (-wu_z) \, dz. \tag{6} \]

In a linear ocean the last term vanishes, and the transport increases with time in the direction of the wind (Yoshida 1959). In a nonlinear ocean, advection increases the downward transport of surface momentum for westernlies but decreases it for easterlies. As a result, a reversing wind changes not only the direction of the current but also its magnitude and position due to nonlinear effects. As discussed by Philander and Pacanowski (1980), the meridional circulation also modifies the linear surface jet, causing the westward surface jet to be shallower and broader and the eastward surface jet to be deeper and narrower. The difference between the current structures under reversing winds indicates the importance of nonlinearities.

b. Subbasin-scale wind bursts

The typical zonal fetch of a westerly wind burst is about 1000 km (Harrison and Giese 1991; Eriksen 1993), so in experiment 1 the zonal half-width of the wind burst is chosen as 5° (Table 2). The zonal velocity and temperature fields on day 10 are shown in Fig. 4. Since the forcing period is only 10 days, the ocean is at the onset of adjustment and is in a transient state. Under the wind burst, a Yoshida jet with the balance of

\[ u_i = \tau_z / \rho_i h \tag{7} \]

has grown in the surface mixed layer of depth \( h \), building up a pressure gradient force (PGF) against the wind stress due to the zonally confined nature of the forcing. Below the surface, this PGF generates a westward current in the upper thermocline between the mixed layer and the main thermocline. Meanwhile, a meridional circulation, with surface convergence and subsurface divergence connected by downwelling at the equator, is being formed. As a result, the surface isotherms are pushed down and subsurface isotherms are brought up due to the meridional circulation. Therefore, in a meridional section, the thermocline is deeper at the equator and shallower near the edges of the wind patch. The surface influence is restricted to the upper thermocline, and isotherms below 200 m undergo little change.

Meanwhile, the first Kelvin wave is emitted from the eastern edge of the patch (Moore and Philander 1977) and propagates eastward at \( c_1 = 2.57 \text{ m s}^{-1} \) (the phase speed of the first baroclinic mode in the model), and a first Rossby wave is emitted from the western edge and propagates westward at about one-third of the Kelvin wave’s speed. These planetary waves will eventually establish Sverdrup balance. However, the wind burst persists for only 10 days, which is too short for the planetary waves to spread the influence of the surface wind across the whole basin.

c. Basin-scale winds

The ocean is forced from rest with basin-scale easterly winds in experiment 2 and with basin-scale westerly winds in experiment 3, for a one-year spinup period. In a basin of similar width, it takes about 150 days for the equatorial currents to reach equilibrium (Philander and Pacanowski 1980). After a year of integration, initialization transients have decayed sufficiently for the purpose of our experiments. The final zonal velocity and temperature fields at the end of the first year are plotted in Fig. 5 for experiment 2 and in Fig. 6 for experiment 3.

At the surface, the easterly wind (expt 2) drives a westward SEC, which has a half-width of the first baroclinic deformation radius (350 km, Fig. 5). The SEC carries surface water westward and results in upwelling near the eastern boundary. Meanwhile, downwelling is induced near the western boundary, and an eastward flowing EUC is generated by the pressure gradient associated with the wind stress. This circulation pattern causes the thermocline to shoal toward the east and to deepen in the west. In the meridional direction, poleward Ekman drift generates divergence at the surface and convergence below, bringing up the surface isotherms and pushing down the subsurface isotherms at the equator. The EUC in the western part of the basin is shallower in the model (Fig. 5) than in the observations (Fig. 1). One reason for this is that the zonal extent of the model ocean is only about one-third as wide as that of the Pacific. McPhaden (1993) examined the dependence of the EUC depth and speed on trade wind fetch analytically and found that the EUC is significantly shallower and weaker in the Atlantic than in the Pacific due to the shorter trade wind fetch.

In response to a westerly wind (expt 3), an eastward surface current and a weak westward undercurrent are generated (Fig. 6). The directions of these currents re-
Fig. 4. Zonal velocity (left) and temperature (right) fields for exp 1 on day 10 (a) at the surface, (b) along the equator, and (c) along meridian Xc. Units are cm s$^{-1}$ and °C, respectively. The thick lines represent 0 cm s$^{-1}$.
verse with the reversing wind. Under westerly winds, Ekman convergence at the surface brings water from midlatitude to equatorial regions, causing a much deeper thermocline than under easterly winds. The surface current is of greater vertical extent than its counterpart in experiment 2; however, it is of greatly diminished magnitude since the momentum input from the wind at the surface is now being distributed to a much thicker layer. Philander and Pacanowski have studied the evolution of equatorial currents using a multilevel primitive equation model by forcing the ocean with reversing winds. When they compared their linear and nonlinear solutions they found that, under westerly winds, the nonlinear mechanisms intensify the linear surface eastward jet and weaken the linear westward undercurrent, and vice versa. The asymmetric characteristics that the model ocean exhibits under opposing winds show that the linear response is significantly modified by vertical and meridional advection. In addition, the undercurrent is located at the top of the thermocline under easterly winds, but in the center of the thermocline under westerly winds. This implies that nonlinearities also change the location of the EUC relative to the thermocline.

4. Response of initially circulating ocean to westerly wind bursts

Wind measurements in the western equatorial Pacific are scarce. Harrison and Giese (1991) classified the westerly wind events based on daily averaged wind records from island stations into four types according to their meridional locations. They found that westerly wind events occurring during boreal winter are predominantly centered either at the equator or near 5°–7° S. The meridional scale of the zonal wind anomaly is approximately 3°–5° latitude, and their duration is typically 10 days. Harrison and Giese conclude that most westerly wind events are connected with tropical cyclones, suggesting mesoscale spatial and temporal variability in the wind field.

Seven additional numerical experiments were performed (Table 3). The first four are designed to investigate the local response to a westerly wind burst centered at the equator and how the response varies with wind fetch. The next three are designed to study the effect that off-equator wind bursts have at the equator. Since easterly trade winds prevail in the equatorial Pacific, the final state of experiment 2 is used as the initial state for all subsequent experiments.

Following spinup with basin-scale easterly winds, experiment 4 is forced with a wind burst identical to that in experiment 1 (10° zonal fetch). The wind-driven eastward surface current connects with the existing EUC and no subsurface westward current is generated between them (not shown here). This kind of behavior is also discussed by Cane (1980), and he attributes it to the strong downward advection of surface momentum. To generate a westward subsurface current, as observed, the wind patch must have a smaller zonal scale.

a. Single wind burst centered at the equator

1) Standard experiment

In experiment 5, the zonal extent of the wind is 2° (Δx = 1° longitude). The amplitude of the wind burst is 1.5 dyn cm⁻², which gives a maximum of 1.0 dyn cm⁻² after accounting for the background trade wind (Fig. 3a). The zonal velocity on day 10 is shown in Fig. 7 (note a change in the field of view from previous figures). At the surface, a local eastward surface jet with an amplitude of 0.4 m s⁻¹, similar to experiment 1, is generated (cf. Fig. 4a to Fig. 7a). It is symmetrical about the equator and has a width of about 4° latitude. A westward subsurface current with a maximum speed of 0.2 m s⁻¹ develops below the surface jet at the depth of 100 m (Fig. 7b). It is deeper near the eastern edge of the wind burst, rising toward the west, and it is connected with the background SEC at the western edge of the wind burst. The meridional convergence under the westerly wind patch is reinforced at the eastern edge by the zonal convergence between the eastward jet and the westward SEC. The subsurface reversing current lies parallel to the shallow isotherms between the base of the mixed layer and the main thermocline (Fig. 7b).

Below the subsurface current lies the basin-scale EUC, which remains at its original position, though with a slightly reduced magnitude of about 30 cm s⁻¹. Thus, we see a three-layer structure under the westerly wind burst: an eastward surface jet, the EUC, and a westward subsurface current in between.

This structure has been observed in many vertical sections along 156°E between December 1992 and March 1993 (Delcroix et al. 1993a), where the magnitude, width, and the position of the subsurface reversing current are all in reasonable agreements with the model results. The three-layer structure is also frequently observed in northern winter along 165°E; however the position of the reversal seems deeper than the model results (Delcroix and Eldin 1995). An interesting feature of the model results is the appearance of two zonal velocity cores (~20 cm s⁻¹) about 1.5° away from the equator on both sides (Fig. 7c). The meridional position of these velocity cores in the model agrees well with observations at 156° and 165°E; however, these velocity cores are at a depth of 60 m, which is shallower than the observations (Delcroix et al. 1993a; Delcroix and Eldin 1995). The disagreement is perhaps due to the shallow thermocline of the initial state.

To look at the impact of the wind burst more directly, the difference in zonal velocity between day 0 and day 10 is plotted in Fig. 8. Eastward propagating Kelvin waves, characterized by a zonal velocity maximum at the equator, are emitted from the eastern edge of the wind patch. Westward propagating Rossby waves, char-
Fig. 5. Zonal velocity and temperature fields for exp 2 on day 365. Legend is as per Fig. 4.
Fig. 6. Zonal velocity and temperature fields for expt. 3 on day 365. Legend is as per Fig. 4.
acterized by symmetric features about the equator, are emitted from the western edge of the patch. The zonal velocity increased by about 1.0 m s$^{-1}$ at the surface near the center of the wind patch and decreased more than 0.2 m s$^{-1}$ in the upper thermocline. There is a symmetric recirculation about the equator east of the wind burst, which is a Rossby wave generated at the eastern boundary by the reflection of the Kelvin wave excited during spinup.

To find out how the current and temperature structures are related, their daily mean profiles on days 1 and 10 under the center of the wind patch are depicted in Fig. 9. Since the wind burst is switched on at the beginning of day 1, profiles of day 1 are already influenced by the surface wind and are different from the initial profiles. However, the influence of the surface wind on day 1 has reached no deeper than 30 m (Figs. 9a and 9b), and one can envision the initial profiles by extrapolating subsurface values upward. The surface mixed layer is initially very deep (150 m) in the western part of the ocean due to the circulation generated by the basinwide trade winds. The velocity profile shows a westward SEC at the surface and an eastward EUC at 130-m depth, with a large vertical shear in the mixed layer (Fig. 9b). Comparing the initial velocity profile with the temperature profile, one finds that the core of the EUC is in the upper thermocline just beneath the surface mixed layer. On day 10, a warmer mixed layer of about 75-m depth has replaced the initial mixed layer. The mixed layer temperature is warmer because of Ekman convergence but the temperature profile in the thermocline remains almost unchanged. The magnitude of the EUC is slightly reduced, and a SSRC is generated just beneath the base of the new mixed layer at a depth of 90 m. These model results resemble observations (Fig. 1) in that the SSRC straddles the mixed layer and the main thermocline but contrast with results from earlier studies that defined the SSRC completely within the mixed layer (Stommel 1960; McPhaden et al. 1988).

The evolution of zonal velocity and temperature at the location of maximum wind is plotted in Fig. 10. The surface jet strengthens and deepens with time, reaching a depth of about 60 m. This depth is about 40 m shallower than the observational results shown in Fig. 1 but it agrees well with the depth of most sections obtained by Delcroix et al. (1993a) at 156°E in December 1992. Meanwhile, the westward current dives below the surface jet and moves downward as time progresses (this is not clearly seen in the observations).

The results of this idealized experiment agree, qualitatively, with the observations (Fig. 1), suggesting that subsurface reversing currents, frequently observed during boreal winter, are likely a local response to westerly wind bursts. A dynamical diagnosis of model results follows.

2) Dynamical Diagnosis

In these experiments, the zonal momentum balance at the equator can be approximated by

$$u_t = -P/ho_0 + \nu_z u_{zz} - \omega u_z,$$  \hspace{1cm} (8)

where horizontal advection and mixing terms have been neglected because they are small compared to other right-hand terms. The right-hand terms in (8) correspond to daily averaged PGF, vertical mixing, and vertical advection, respectively. Waongne (1989) showed that in the nonlinear equatorial Atlantic, the primary balance above the thermocline for an annually averaged state is between these right-hand terms regardless of surface forcing or of the existence of a meridional advection term related to asymmetry.

To investigate the role that each term plays in the local response generation process, the temporal evolution of each right-hand term in (8) is contoured, for the upper 150 m at the center of the wind patch (Fig. 11).

In the numerical experiments, the horizontal advection term of zonal momentum is an order of magnitude smaller than the other terms in (8), and therefore is not shown in Fig. 11. Let us look at the evolution of the PGF first. Within a day, it changes from eastward to westward in the upper 90 m. It continues to increase up to day 4 and then stays almost constant (Fig. 11a).

The vertical mixing term is

$$\int \nu_z u_{zz} dz = \nu_z u_{zz, \text{upper}} - \nu_z u_{zz, \text{lower}}$$

for each numerical layer. The wind stress $\tau_z$ is applied at the surface as $\nu_z u_{zz, \text{surface}}$, so vertical mixing is maximum in the surface mixed layer (Fig. 11b). In this model, $\nu_z$ is not specified; $\nu_z u_{zz}$ is determined by instability criteria. The effective vertical mixing coefficient, $\nu_z$ averaged over the 10 days, is plotted in Fig. 12b. Its value is calculated by dividing the mean vertical mixing by the mean zonal velocity shear. The result has a maximum of 3.5 cm$^2$ s$^{-1}$ at the surface and decreases downward quickly with an e-folding depth of 50 m. In contrast to the pressure gradient force, vertical mixing weakens near the surface during the first few days due to the deepening of the mixed layer. A subsurface maximum grows and moves downward, diffusing eastward
Fig. 7. Zonal velocity and temperature fields for exp 5 on day 10. Legend is as per Fig. 4 but the meridional and zonal extent of the figures is diminished.
surface momentum to deeper waters. The subsurface reversing current remains at depths where the direct influence of the wind is small.

Vertical advection for each numerical layer is calculated as vertical momentum flux, that is, the difference between $uw$ at the upper and lower interfaces. Under the wind patch, horizontal convergence generates strong downwelling at the bottom of the mixed layer, advecting eastward momentum downward to layer 2 (Fig. 11c). The vertical advection term is negative near the surface, and its magnitude increases as Ekman convergence develops on timescales of the local inertial period, which is about 9 days. Below the surface layer, the vertical advection term is positive; farther down it becomes negative again as the vertical shear of zonal velocity reverses.

Based on measurements made in the eastern equatorial Pacific, Dillon et al. (1989) have approximated the pressure gradient profile as Gaussian with a depth scale between 100 and 140 m, and the turbulent momentum flux as an exponential function with a depth scale of 30 to 90 m. Under basin-scale trade winds (expt 2), our model results show depth scales of 180 m for the PGF and 110 m for vertical mixing (Fig. 12c). The model-derived depths are larger than the measurements because the numerical experiments are in the western part of the basin.

The 10-day means of PGF, vertical mixing, and advection are shown in Fig. 12a. Both the PGF and the vertical mixing terms are strongest near the surface and decrease with depth; vertical mixing, however, decreases more rapidly than the PGF. The PGF under the westerly wind burst (expt 5) is three times as large as that under trade winds, while the vertical mixing is six times as large. The PGF and vertical mixing under the wind burst (expt 5) are shallower than under the basin-scale winds (expt 2) because the adjustment is not complete. Observational estimates of the PGF under a westerly

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**Fig. 8.** Zonal velocity change between day 0 and day 10 for expt 5. Legend is as per Fig. 4.
wind burst in the equatorial Pacific compare favorably with our model results (Smyth et al. 1996).

Below the surface there is a region where the current is accelerated in the direction of the PGF because the zonal PGF exceeds the vertical mixing term. Under basin-scale trade winds, this region occupies up to three-quarters of the model ocean except for a wide eastern boundary and a thin western boundary. Wacongne (1989) has examined the dynamical regimes in a steady-state Atlantic simulation and she refers to the region where the PGF dominates as “subregion 2.” Blanke and Delecluse (1993) embedded a turbulence closure scheme in an ocean general circulation model to improve vertical mixing and the simulation of equatorial ocean dynamics. As a consequence of the scheme, the mixed layer is deeper in the western Atlantic than that reported by Wacongne, and this allows a larger influence of the pressure gradient. Wacongne’s so-called subregion 2 is enlarged to cover more than half of the basin west of the longitude of wind reversal, resulting in better agreement with our results.

Fig. 9. Vertical profiles of (a) temperature and (b) zonal velocity on days 1 and 10 in expt 5 under the center of the westerly wind patch \( (X_c, 0^\circ) \). Solid and dashed lines represent profiles on day 1 and day 10, respectively. Horizontal arrows in (a) represent the depth and direction of zonal currents.

Fig. 10. Evolution of (a) zonal velocity and (b) temperature under the center \( (X_c, 0^\circ) \) of the westerly wind patch in expt 5. Contours are in \( \text{cm s}^{-1} \) and \(^\circ\text{C} \), and thick lines represent 0 \( \text{cm s}^{-1} \) and 27\(^\circ\text{C} \) contours.
The vertical advection changes sign with the vertical shear of zonal velocity, showing a maximum at 45 m and a minimum at 100 m. The sum of the three forces represents the overall acceleration during the wind burst, giving rise to an eastward acceleration near the surface and to a westward acceleration below the surface, as shown in Fig. 12a. The PGF is the driving force for the subsurface reversing current, aided by vertical advection.

3) Wind Bursts of Variable Fetch

In the next series of experiments, we investigate changes in the local oceanic response due to variations in the zonal fetch $2\Delta x$ of the wind burst. Figure 13 shows vertical sections of zonal velocity along the equator for experiments 5, 6, and 7. The subsurface reversal is the strongest in experiment 5 when $\Delta x = 1^\circ$. As $\Delta x$ increases, the surface jet deepens and strengthens, and
the SSRC and the EUC move farther downward. Both the SSRC and the EUC decrease in amplitude with increasing $\Delta x$; this result is consistent with Cane’s (1980) analysis. Although a westward subsurface current grows with increasing $\Delta x$ under the eastern edge of the patch, we do not investigate its behavior because it is of smaller zonal scale and it is not a part of the main subsurface current. When $\Delta x = 3^\circ$, the SSRC disappears under the center of the wind patch and an eastward minimum is instead observed between the surface current and the EUC. We see that the zonal extent, $2\Delta x$, of the westerly wind burst is the key factor in determining the generation of the SSRC. For our model parameters, the SSRC is produced only when the zonal scale of the wind burst is smaller than about $6^\circ$ longitude.

To investigate the dynamical cause of this behavior, the evolution of the subsurface PGF at a depth of 100 m at the center of the wind patch ($X_c, 0^\circ$N) is plotted in Fig. 14. The PGF is initially eastward in order to balance the basin-scale easterly winds. As soon as the westerly wind is switched on, the eastward PGF starts to diminish locally and then reverses direction. But the reversal time and the growth rate (represented by the slope of the line) clearly vary with $\Delta x$; the larger $\Delta x$ is, the more delayed the local PGF reversal is and the slower the westward PGF grows.

While the westward PGF is building up, the surface momentum is transferred downward both by vertical advection and by turbulent mixing. As shown in Fig. 12a, turbulent mixing decreases more rapidly with depth than does the PGF so that the net subsurface force is westward. Vertical mixing is determined by the magnitude of the wind stress and it remains unchanged when the zonal fetch of the wind burst varies. The growth rate and magnitude of the PGF, however, are dependent upon the zonal fetch of the wind burst. The fate of the SSRC depends on which of the two competing effects—the PGF or the vertical mixing—dominates. Kelvin and Rossby waves, which help establish the PGF, have less distance to travel under smaller wind patches; therefore the PGF grows more quickly and is of sufficient intensity to exceed the downward momentum transport of eastward momentum and to generate a westward subsurface current. Our model results suggest $6^\circ$ longitude as the threshold zonal fetch. When the wind is smaller than this threshold, the PGF dominates and a westward subsurface current is generated. When the wind is larger than this value, the downward momentum transport by vertical mixing and by advection accelerates the subsurface current eastward before the planetary waves have time to propagate across the forcing region and establish the westward PGF; the velocity profile has a minimum separating the surface jet from the EUC, but there is no reversal.

b. Single wind burst centered off the equator

In order to investigate the equatorial influence of off-equator winds, experiment 8 repeats experiment 5 but with the westerly wind burst centered off the equator at $Y_0 = 6.2^\circ$N. The distance between this location and the equator is approximately twice the first baroclinic deformation radius. Change of zonal velocity field over 10 days due to the off-equator westerly wind is plotted in Fig. 15. It is obvious that the model response is not limited to the area underneath the wind patch; a westward and equatorward propagation of energy at the surface and an eastward and equatorward propagation below the surface can be seen as well.

Wind-induced Rossby waves propagate westward as well as toward the equator, appearing as a northeast-southwest oriented dipole at the surface with a maximum zonal velocity change of 20 cm s$^{-1}$ (Fig. 15a). Below the surface, the zonal velocity is accelerated westward by up to 15 cm s$^{-1}$ (Fig. 15b). The subsurface change peaks at 100-m depth and is roughly at the same zonal location as the center of the wind patch (Figs. 15b and 15c). The resulting zonal velocity changes, both at the surface and below, have similar spatial scale with that of the wind burst. The subsurface response of the
equatorial ocean is westward and opposite to the surface wind; thus it opposes the EUC. Experiment 8 demonstrates that off-equator westerly winds can have significant influence on surface and subsurface equatorial currents.

Experiment 9 is a repeat of experiment 8 but with the wind direction reversed. Meridional sections of zonal velocity from a control run (without any wind bursts) and from experiments 8 and 9 are shown in Figs. 16a–c, respectively. From a comparison of the two experiments with the control run, we conclude that the off-equator westerly wind decelerates the EUC on the order of 10 cm s\(^{-1}\) (Figs. 16a and 16b), while the easterly wind accelerates the EUC on the order of 10 cm s\(^{-1}\) (Figs. 16a and 16c). In both experiments, the net force exerted upon the EUC is in a direction opposite to that of the wind. At the surface, zonal velocity changes at the equator are relatively small. The results imply that
off-equator winds have a larger impact on the EUC than on surface currents at the equator.

In the observational record (Fig. 1c), the EUC decelerates 20 cm s\(^{-1}\) during the first half of the record, accelerates during the second half, and appears to be uncorrelated with the occurrence of local westerly winds (Fig. 1a). A strong westerly wind burst is observed at 5°S, 165°E (Fig. 1e) when the EUC decelerates, and strong and persistent easterly wind events are observed at 5° and 8°N along 165°E when the EUC accelerates. It is plausible that the observed off-equator wind bursts were responsible for the observed changes in the EUC as per experiments 8 and 9.

c. A stationary tropical cyclone

Observational results indicate that westerly wind bursts in the western equatorial Pacific are often associated with tropical cyclones (Harrison and Giese 1991; Keen 1982), which are quasi-stationary during the initial stage of their development (Elsberry 1995). Results from previous sections have shown the important effects of both equatorial and off-equatorial winds on the equatorial currents. To address the question of what happens when the two types of wind coexist, an investigation on

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Fig. 14. Evolution of the PGF at the center of the westerly wind patch in model layer 5 (about 100-m depth) in expts 5, 6 and 7. The solid line represents PGF in expt 5, dashed and dash-dotted lines represent PGF in expts 6 and 7, respectively.

Fig. 15. Change of the zonal velocity field over 10 days due to the off-equator westerly wind in expt 8. Units are in cm s\(^{-1}\).
the equatorial response to a stationary tropical cyclone is carried out in experiment 10.

The model ocean is forced with a stationary tropical cyclone consisting of two zonal winds that are equally strong but of opposing directions: a westerly wind at the equator and an easterly wind centered at 6.2°N. The composite wind pattern is expressed as follows:

\[
\tau^*_w(x, y) = |\tau_w| \exp \left[ -\left( \frac{x - X_w}{\Delta x} \right)^2 \right] \times \left\{ \exp \left[ -\left( \frac{y - Y_w}{\Delta y} \right)^2 \right] - \exp \left[ -\left( \frac{y - Y_e}{\Delta y} \right)^2 \right] \right\},
\]

where \((X_w, Y_w)\) and \((X_e, Y_e)\) are the centers of the westerly and the easterly wind patches, respectively (Fig. 3b).

Figure 17 shows meridional sections of zonal velocity under each of the two branches of the wind, first separately and then together, simulating a cyclone (expts 5, 9, and 10). When there is a westerly wind at the equator, a three-layer structure is generated, as discussed earlier. When there is an easterly off-equator wind (Fig. 17b), the surface current is asymmetric about the equator, with little change of its magnitude at the equator. The EUC is remotely accelerated to over 50 cm s\(^{-1}\), becoming twice as fast as that in experiment 5.

When these two wind bursts are specified simultaneously as an idealized tropical cyclone, the three-layer structure is still present, including the eastward surface jet, the westward SSRC, and the EUC (Fig. 17c). This suggests that the local wind at the equator still dominates the equatorial ocean response. The remote effect of the off-equator easterly can be seen by comparing the ve-
loicity fields in experiments 5 and 10. The comparison shows that the EUC is 10 cm s$^{-1}$ faster when driven by the cyclone than by the westerly wind alone.

5. Summary and discussion

An improved version of the Gent and Cane tropical upper-ocean circulation model (Chen et al. 1994) is forced with basin-scale easterly (expt 2) and westerly winds (expt 3). The oceanic responses to these basin-scale winds are qualitatively different; the depth of the EUC is much shallower under the easterly winds than under the westerly winds. This asymmetry has been attributed to nonlinear vertical advection in previous studies.

We studied the local response of the model ocean, initially forced by basin-scale easterly winds, to westerly wind bursts. The results suggest that the observed three-layer structure in the zonal velocity field (Fig. 1c) is primarily a local response of the equatorial ocean to the strong westerly wind events. The eastward Yoshida jet is driven directly by the wind, while the SSRC is driven by the PGF associated with the wind burst and established by equatorial waves.

Experiments forced with equatorial westerly wind bursts of different fetch (expts 5 to 7) demonstrate the importance of spatial variations in the wind field, for example, tropical cyclones, in determining the local response. At the onset of the westerly wind, the surface current reverses its direction and accelerates eastward. Meanwhile, the eastward PGF built by the trade winds diminishes and a westward PGF develops. This westward PGF in turn balances the wind stress at the surface, accelerates the subsurface fluid westward, and decelerates the EUC originally generated by the basin-scale PGF. Downward momentum mixing plays a competing role in this process.
role against the PGF by accelerating the current eastward. The PGF setup time, which is the time it takes for waves to travel across the wind fetch (usually a few days), determines which of the two effects is dominant. Experiments 5 to 7 show that the SSRC develops only when its zonal fetch is less than 6° of longitude—the larger the fetch, the weaker the SSRC, with the SSRC disappearing completely when the zonal fetch of the wind burst exceeds 6° of longitude. Analysis of the momentum balance suggests that the threshold for zonal wind fetch grows with decreasing magnitude of the vertical advection term.

There are indications that the zonal extent of wind bursts may be at least 15° longitude (Harrison and Giese 1991). However, the winds reported by Harrison and Giese are daily averaged, and therefore the zonal extent and structure cannot be adequately resolved. For example, a wind of 200-km zonal scale translating at 5 m s⁻¹ zonally would appear as 1000 km wide when daily averaged. Thus, better temporal and spatial resolutions are desirable for studies related to temporal changes of wind field.

Westerly wind bursts are characteristic to the western equatorial Pacific where the mixed layer is deep due to the prevailing easterly winds. The SSRC is only generated when the initial EUC is deep enough to be beyond the reach of near-surface vertical mixing. In different regions, the oceanic response would not be the same due to different density and velocity profiles.

Sensitivity to the intensity of wind stress was tested by repeating experiment 1 with wind bursts of different magnitudes (0.5 and 1.5 dyn m⁻², respectively). The resulting velocity fields are shown in Fig. 18. A comparison with the velocity field of Fig. 4 shows that the downward transport of momentum, as well as the magnitude and depth of both surface jet and subsurface current, increase with increasing wind stress. This result agrees with the analytical solution of McCreary (1985), who found that the local response to wind forcing is

![Fig. 18. Zonal velocity field from repetitions of expt 1 with wind stress of 0.5 and 1.5 dyn cm⁻², respectively. Units are cm s⁻¹. The thick lines represents 0 cm s⁻¹.](image)

**Table 4. Sensitivity to the wind stress magnitude for a circulating background:** $U_s$ is the velocity of the surface jet and $U_{SSRC}$ is the velocity of the SSRC.

<table>
<thead>
<tr>
<th>Wind stress (dyn m⁻²)</th>
<th>$U_s$ (m s⁻¹)</th>
<th>$U_{SSRC}$ (m s⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.5</td>
<td>0.32</td>
<td>-0.15</td>
</tr>
<tr>
<td>2.5</td>
<td>0.49</td>
<td>-0.18</td>
</tr>
<tr>
<td>3.5</td>
<td>0.66</td>
<td>-0.29</td>
</tr>
<tr>
<td>4.5</td>
<td>0.75</td>
<td>-0.40</td>
</tr>
<tr>
<td>5.5</td>
<td>0.80</td>
<td>-0.42</td>
</tr>
</tbody>
</table>
directly proportional to the strength of the wind. Sensitivity studies for a circulating background were also performed and the results are listed in Table 4. For a wind patch of given zonal extent, the magnitude and thickness of the surface jet, as well as the magnitude and depth of the SSRC, increase with increasing intensity of the wind stress.

Sensitivity on the strength of vertical mixing has also been examined. When mixing in the bulk mixed layer is reduced, larger shear at the base of the mixed layer is created, and the $R$, mixing scheme generates more mixing to reduce this shear. Therefore, the two mixing schemes compensate each other so as to keep the water column stable. When both mixing schemes are artificially weakened, vertical advection tends to increase due to a larger vertical shear of zonal velocity near the surface, and as a result there is little change in the downward transport. We speculate that if the downward transport of eastward momentum (by vertical mixing and by advection) were weaker, the threshold for the zonal fetch of westerly wind burst would increase.

Current meter measurements from equatorial TOGA TAO buoys at $0^\circ$, $156^\circ$E and $0^\circ$, $165^\circ$E show nonzero correlation between the SSRC at the two locations (M. McPhaden 1995, personal communication), indicating that the SSRC might be generated by winds with zonal fetch greater than $10^\circ$ longitude. If this is the case, then the threshold of zonal wind fetch for generating SSRC is underestimated by the model. This discrepancy could have been caused by the large vertical advection in our model, which exaggerates the downward transport of surface momentum. However, the SSRC observed by McPhaden was rarely associated with an eastward surface jet, implying the absence of westerly wind bursts. In addition, the maximum of the SEC often appears below the surface, suggesting the possible influence of other dynamical processes, for example, inertial oscillation (Smyth et al. 1996). To identify these processes, simultaneous high resolution current meter and wind field measurements are necessary.

Results from experiments forced with a single off-equator wind burst (expts. 8 and 9) indicate that remote wind bursts excite waves that propagate horizontally as well as vertically. The resulting Rossby waves affect the EUC more effectively than the surface current at equator. For the given meridional position in experiments 8 and 9, westerlies weaken the EUC and easterlies strengthen it. The energy beam is found to propagate downward at a slope $-\sigma N_s$ (where $\sigma$ is the forcing frequency and $N_s$ the local Väisälä frequency) in a linear model and under equatorial winds (McCreary 1984). Further theoretical and numerical studies on the response of the equatorial ocean to off-equator winds at different meridional positions are needed to better understand energy propagation.

When a westerly wind burst at the equator coexists with an easterly off-equator wind burst (resembling the zonal wind pattern of a tropical cyclone), an eastward surface current and a westward subsurface current develop locally at the equator. At depth, the remote effect of the easterly wind overcomes the local effect of the westerly wind, accelerating the EUC. Since the mean wind is weak in the western equatorial Pacific, atmospheric convection, which could later develop into cyclones, is of primary importance (Kindle and Phoebeus 1995). The numerical results suggest that off-equator easterly wind bursts were perhaps responsible for the unexplained acceleration of the EUC during the first half of the record.

The salinity effects on density are not considered in this study. However, recent observational results from TOGA COARE have shown that westerly winds are also associated with a series of squalls that consist of high winds and large rainfalls. The freshwater associated with the squalls tends to stay at the surface and form a fresher layer. When this happens, the bottom of the mixed layer is actually at the halocline, which is much shallower than the thermocline, and the temperature remains unchanged across the base of the mixed layer. The existence of the halocline increases the density stratification of the upper ocean and will ultimately weaken downward vertical mixing. The effect of salinity on the vertical mixing is investigated in a companion paper (Zhang and Rothstein 1998, submitted to J. Geophys. Res.). It would be interesting to investigate how an active salinity field modifies the SSRC.

Acknowledgments. We thank Dr. D. Chen for sharing the model code. This work was supported by the U.S. Department of Commerce National Oceanic and Atmospheric Administration through Grants NA46GP0187 and NA16RC0228.

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