UPPER OCEAN UPWELLING, TEMPERATURE, AND ZONAL MOMENTUM
ANALYSES IN THE WESTERN EQUATORIAL PACIFIC

by

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To my wife Jenny, our son Reilly, and our baby on the way.
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Upper Ocean Upwelling, Temperature, and Zonal Momentum Analyses in the Western Equatorial Pacific

Robert William Helber

ABSTRACT

The air-sea interaction thermodynamics of the western equatorial Pacific, the Earth's largest region of warm SST, is a major component of the global climate system. Along the equator, warm pool thermodynamics and momentum are influenced by equatorial ocean visco-inertial boundary layer dynamics that occur within a few degrees of the equator because of the sign reversal of the Coriolis force. Designed to study this system, COARE Enhanced Monitoring Array (EMA) observations of temperature, salinity, velocity, and surface meteorology were centered at 0, 156°E from February 1992 through April 1994. They sampled variability on the equator over larger space/time-scales than the concurrent Intensive Flux Array (IFA) centered at 2°S, 156°E. The EMA data are examined within the context of the larger scale equatorial Pacific and the El Niño conditions that occurred at that time. There is a structural change in the equatorial Pacific near the dateline resulting from the winds that are strong, steady, and easterly in the east and generally weak, punctuated by westerly wind bursts, in the west. East of the dateline the EUC's speed and transport increases downstream, while in the west it tends to be zonally uniform, consistent with the extra-tropical ocean interior water pathways that tend to converge on the equator east of the dateline. At 0°, 156°E in the western Pacific deep, seasonal upwelling (appearing stronger after the peak of the 1991/92 El Niño than during the following weaker El Niño year) occurs within the thermocline in boreal summer with magnitudes as large as upwelling in the eastern Pacific cold tongue. This large upwelling is associated with large downward turbulent heat flux and large turbulent shear stress. While the inferred mixing is quantitatively inconclusive because of unresolved potential errors, it is consistent with the visco-inertial boundary layer concepts from early theory [e.g. Arthur 1960; Robinson 1960; Stommel 1960; and Charney and Spiegel 1971]. These findings suggest that the equatorial thermodynamics differ from those of the IFA. Further process experimentation is necessary to quantify these results.
1.1 Introduction

The global temperature balance requires that the ocean and atmosphere transport energy towards the poles, with the equator at the center of this global energy flux divergence. Incoming short-wave radiation exceeds outgoing long-wave radiation between 35°S and 35°N, indicating a net heat influx to the tropical and subtropical regions [e.g. Bryden and Imawaki 2001]. While both the ocean and the atmosphere may contribute equally to the global temperature balance, each transports energy differently. In the ocean, the energy transport happens on longer time-scales and is more geographically constrained than in the atmosphere. One of the ocean's largest energy influxes occurs along the equator, because of the combined effect of equatorial ocean dynamics and large incoming net surface heat flux. Equatorial dynamics result in large upwelling of relatively cold water and a local maximum in surface heat flux-induced warming at the equator that drops off rapidly within a few degrees of latitude, where upwelling reverses to downwelling. From an ocean circulation viewpoint, cool subthermocline water enters the equatorial region while warmer near-surface water exits and flows poleward, suggesting that non-isentropic processes must occur en route. In contrast, large outward heat flux occurs from the western boundary currents that transport warm water poleward below an increasingly cooler atmosphere. These processes along with water mass formation at high latitudes result in a poleward energy transport within the Pacific ocean and a net northward energy transport from the southern to the northern hemispheres in the Atlantic ocean. Since large ocean heat influx occurs on the equator and large outward ocean heat flux occurs within the western boundary regions, where these regimes meet is central in our understanding of the global temperature budget and the western equatorial Pacific is the subject of this dissertation.

The primary focus is on the equatorial ocean dynamics beneath the western Pacific warm pool. Since the warm pool is the largest warm water pool on Earth, it directly impacts global climate on annual to interannual time-scales [e.g. Weisberg and Wang 1997a; Jin 1997; Clarke and Shu 2000; Wang 2001]. In the eastern Pacific, equatorial ocean dynamics result in a shallow thermocline and a tongue of cold sea surface temperature (SST) that originates both on the equator as well as emanating from the eastern boundary. While the warm SST in the western Pacific covers tens of degrees of latitude and longitude, a narrow strip of cooler temperature connected to the cold
Figure 1. Mean zonally averaged a) zonal velocity, b) temperature, and c) salinity from the NCEP Pacific Ocean Reanalysis. Zonal and temporal averages are from grid locations 170.25°E to 108.75°W and from 1980 to 2000. The contour intervals are $5 \times 10^{-2} \text{ m s}^{-1}$, °C, and 0.1, respectively.

tongue tends to occur along the equator (see Figure 6 in Chapter 2). This suggests that equatorial ocean dynamics influence warm pool SST even in the western Pacific where the thermocline is relatively deep.

1.2 Central and eastern equatorial Pacific structure

Incident net heat flux is taken up along the equator by the cool upwelled equatorial waters, but the poleward energy transport within the ocean is not direct. In the Pacific, warm equatorial surface water generally flows poleward via Ekman transport, but just north of the equator below the Inter-Tropical Convergence Zone (ITCZ), for
example, water is downwelled. At mid-latitudes water is subducted, flowing back towards the equator within the thermocline. Together these circulations comprise the Subtropical Cell (STC) [e.g. McCreary and Lu 1994]. A qualitative understanding of this circulation can be seen in zonally and temporally averaged meridional sections of zonal velocity, temperature, and salinity from National Centers for Environmental Prediction (NCEP) Pacific Ocean Reanalysis products (Figure 1).

The zonal velocity features present in this zonally averaged section of the Pacific ocean from 170°W to 110°W, are the EUC on the equator centered at ~90 m depth, the westward South Equatorial Current (SEC) that straddles the equator at the surface, and the eastward North Equatorial Countercurrent (NECC) centered at ~7°N (Figure 1a). The Tsuchiya jets are not resolved, but the top of the Equatorial Intermediate Current, or the first in a series of reversing equatorial jets below the EUC, is seen. Qualitative justification for the STC appears in the meridional temperature section (Figure 1b). In response to negative wind stress curl, water subducted via Ekman pumping within the mid-latitude subtropical gyres (20° to 40°) flows equatorward within the thermocline along extra-tropical water pathways [e.g. Liu and Philander 2001]. Near the surface, the easterly trade winds result in Ekman divergence at the equator and consequently there is upwelling and a communication between the thermocline water from mid-latitudes and the surface water that flows poleward. The details of how this communication occurs in a narrow band along the equator and how these different water masses coalesce are still unclear. North of the equator the potential vorticity constraint imposed by the NECC, coinciding with the thermocline shape about the NECC core, makes the route to the equator indirect, and thus more water reaches the equator from the south. At the equator, the isotherms above the EUC core bow upward suggesting upwelling, which links the mid-latitude subducted water with the Ekman divergence at the surface. Below the EUC core the isotherms bow downward suggesting downwelling [Knauss 1966; Weisberg and Qiao 2000]. The well-mixed thermostad below the EUC, which has nearly uniform temperature and salinity, is also characteristic of the equator. In addition, high salinity water approaches the equator from the south within the thermocline (Figure 1c).

1.3 Equatorial dynamics

Equatorial dynamics are primarily a result of the horizontal Coriolis force, which vanishes at the equator, changing sign from the northern to the southern hemispheres. As a consequence, an equatorial boundary region exists where relative vorticity, vertical advection, and turbulent eddy viscosity and diffusion are enhanced. This boundary region results in zonal currents such as the equatorial undercurrent (EUC), the Tsuchiya jets, and a succession of deep reversing equatorial jets, in addition to equatorial upwelling and the well-mixed thermostad below the EUC. Following Charney and Spiegel [1971] this boundary region will be termed "visco-inertial," referring to the enhanced eddy viscosity and inertial dynamics at the equator.

Analysis of the dynamics within the equatorial visco-inertial boundary region reveals at least three regimes operating on different scales that coexist and interact. The
regimes that will be discussed are an inertial regime where pressure and friction are neglected, a frictional regime governed by a balance of pressure gradient and friction, and a linear equatorial wave regime. While any one of these regimes may dominate the variability at a given scale, the general nature of equatorial ocean variability results from a combination of all three.

To demonstrate the nature of these regimes consider the zonal momentum

\[
\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} - \beta y v = - \frac{1}{\rho} \frac{\partial P}{\partial x} + A_H \left( \frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial y^2} \right) + A_V \frac{\partial^2 u}{\partial z^2} \tag{1.1}
\]

and the vertical vorticity component

\[
\frac{\partial \zeta}{\partial t} + u \frac{\partial \zeta}{\partial x} + v \frac{\partial \zeta}{\partial y} + w \frac{\partial \zeta}{\partial z} - \zeta \frac{\partial w}{\partial z} + \beta y \frac{\partial w}{\partial z} - \frac{\partial w}{\partial y} \frac{\partial u}{\partial z} + \frac{\partial w}{\partial x} \frac{\partial v}{\partial z} = \frac{1}{\rho^2} \left( \frac{\partial \rho}{\partial y} \frac{\partial P}{\partial x} - \frac{\partial \rho}{\partial x} \frac{\partial P}{\partial y} \right) + A_H \left( \frac{\partial^2 \zeta}{\partial x^2} + \frac{\partial^2 \zeta}{\partial y^2} \right) + A_V \frac{\partial^2 \zeta}{\partial z^2} \tag{1.2}
\]
equations for the equatorial ocean. In this treatment \( u, v, \) and \( w \) are the east, north, and vertical velocity components in the \( x, y, \) and \( z \) directions, respectively, \( \beta \) represents the Coriolis parameter's rate of change with latitude, \( \rho \) is density, \( P \) is pressure, and \( \zeta = \partial v/\partial x - \partial u/\partial y \) is the vertical vorticity component. Continuity, \( \partial u/\partial x + \partial v/\partial y + \partial w/\partial z = 0 \), has been applied in equation 1.2. Friction is parameterized assuming the Reynolds stresses are linearly dependent on spatial gradients of the flow field with constant eddy viscosity coefficients, \( A_H \) and \( A_V \). Near the equator the horizontal Coriolis force is relatively small, but its change with latitude is largest and consequently the Coriolis parameter \( (f) \) is approximated by \( \beta y \). Brackets above and below equations 1.1 and 1.2 indicate the terms involved in the inertial and frictional regimes. The equatorial wave regime includes the linearized inertial terms balanced with pressure gradient force. The meridional momentum equation is similar to (1.1) but is not shown because where the Coriolis force terms balance with the advection terms, the magnitude of variability in the \( x \) and \( y \) directions is identical. This will be denoted using \((u,v) = U(u',v')\) and \((x,z) = L(x',y')\) so that the two horizontal components of velocity and length are the same. Scaling the equations 1.1 and 1.2 with the additional factors \( w = Ww' \), \( z = Hz' \), \( t = Tt' \), \( \rho = \rho_o \rho' \), and \( P = P_o P' \) results in
In cases where the Coriolis force terms are not of identical magnitude with the advection terms, the two horizontal directions can have different scales and the meridional momentum must then be considered. For this description of the equatorial visco-inertial boundary layer equations 1.3 and 1.4 are adequate.

In the following analysis each regime will be considered separately, deriving inherent scales independently of the other regimes. The failure of a single regime to be consistent with equatorial observations is an indication of the need for the coupling of two or more regimes to describe the nature of the equatorial ocean. The piecewise approach is somewhat historical and helps highlight why each regime fails, consequently providing information about how the regimes must couple.

1.3.1 Inertial regime

Early investigation of inertial motions by Whipple [1917] showed the paths of free particles on a smooth rotating globe oscillating about the equator. More recently and more relevant to observations, Fofonoff and Montgomery [1955] demonstrated the essential dynamics of the EUC with an inertial balance of relative and planetary vorticity in the absence of vortex stretching. With respect to equation 1.4, divided by $\beta U$, this is represented by

$$
U \frac{\partial u'}{T \partial t} + \frac{U^2}{L} \left( u \frac{\partial u'}{\partial x} + v \frac{\partial u'}{\partial y} \right) + \frac{WU}{H} w' \frac{\partial u'}{\partial z} = \beta LUy'v' \tag{1.3}
$$

$$
= -\frac{P_0}{\rho_0 L} \frac{1}{\rho} \frac{\partial P'}{\partial x'} + \frac{A_n U}{L^2} \left( \frac{\partial^2 u'}{\partial x^2} + \frac{\partial^2 u'}{\partial y^2} \right) + \frac{A_p U}{H^2} \frac{\partial^2 u'}{\partial z^2}
$$

$$
\frac{U}{LT} \frac{\partial \zeta'}{\partial t'} + \frac{U^2}{L^2} \left( u \frac{\partial \zeta'}{\partial x} + v \frac{\partial \zeta'}{\partial y} \right) + U \beta v' - \frac{\beta WL}{H} y' \frac{\partial w'}{\partial z'}
$$

$$
+ \frac{WU}{HL} \left( \frac{w' \frac{\partial \zeta'}{\partial z} - \zeta' \frac{\partial w'}{\partial z} + \frac{\partial w'}{\partial x} \frac{\partial v'}{\partial z} - \frac{\partial w'}{\partial y} \frac{\partial u'}{\partial z} \right)
$$

$$
= \frac{P_0}{\rho_0 L^2} \frac{1}{\rho^2} \left( \frac{\partial \rho'}{\partial y} \frac{\partial P'}{\partial x} - \frac{\partial \rho'}{\partial x} \frac{\partial P'}{\partial y} \right) + \frac{A_n U}{L^2} \left( \frac{\partial^2 \zeta'}{\partial x^2} + \frac{\partial^2 \zeta'}{\partial y^2} \right) + \frac{A_p U}{LH^2} \frac{\partial^2 \zeta'}{\partial z^2} = 0. \tag{1.4}
$$

In cases where the Coriolis force terms are not of identical magnitude with the advection terms, the two horizontal directions can have different scales and the meridional momentum must then be considered. For this description of the equatorial visco-inertial boundary layer equations 1.3 and 1.4 are adequate.

In the following analysis each regime will be considered separately, deriving inherent scales independently of the other regimes. The failure of a single regime to be consistent with equatorial observations is an indication of the need for the coupling of two or more regimes to describe the nature of the equatorial ocean. The piecewise approach is somewhat historical and helps highlight why each regime fails, consequently providing information about how the regimes must couple.

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$$
\frac{U}{\beta L} \left( u \frac{\partial \zeta'}{\partial x} + v \frac{\partial \zeta'}{\partial y} \right) + v' = 0. \tag{1.5}
$$

Immediately from the balance between relative and planetary advection, we obtain a length-scale

$$
L = \sqrt{\frac{U}{\beta}}. \tag{1.6}
$$
With respect to equation 1.3, 1.6 represents the length scale where advective and local rates of change have the same magnitude as the Coriolis force. Using a zonal velocity speed of $U = 0.8 \, \text{m s}^{-1}$ which is a characteristic speed of the EUC in the central Pacific, a horizontal length scale of $\sim 200 \, \text{km}$ is obtained. With this scaling alone, however, information regarding the relative magnitude of vertical advection is unavailable.

An implicit result of the continuity equation is

$$W = \frac{HU}{L}, \quad (1.7)$$

which is equivalent to the horizontal and vertical advection terms of equation 3 being of the same magnitude $\left( U^2/L = WU/H \right)$. Combining (1.7) and (1.6) so that all three components of advection are the same order as the Coriolis term gives

$$W = H \beta U. \quad (1.8)$$

Figure 2. Vertical velocity scale $W$ plotted as a function of $U$ and $H$ according to equation 1.8. The numbers at the top are $L$ and $U$ associated with the nearby line.
With equation 1.8 we can estimate the magnitudes of $W$ and $H$ that are associated with various values of $U$ and $L$ in the inertial regime.

Figure 2 summarizes equations 1.6 and 1.8. For example, if we choose values from the line representing $U = 0.8 \, \text{m} \, \text{s}^{-1}$ with $W = 5 \times 10^{-5} \, \text{m} \, \text{s}^{-1}$ we get $H = 12 \, \text{m}$. This suggests that for mean EUC speeds and large values of vertical velocity scale, for example, the vertical length scale is short relative to the EUC's vertical height (~100 m). While vertical advection on the equator is more important relative to mid-latitudes, the vertical scale required by equation 1.8 is smaller than would be expected to explain the EUC and equatorial upwelling. Alternatively, a larger vertical scale would imply unrealistically large vertical velocity. In either case the inertial regime alone is incomplete, indicating that pressure force and/or friction broadens the inherent vertical scale and/or limits the magnitude of vertical velocity on the equator.

To further demonstrate the nature of the inertial regime at the equator, consider the conservation of vorticity for an inviscid ocean where zonal and vertical gradients (i.e. stretching and tilting) are neglected relative to meridional gradients. In this case the dimensional equation 1.2 reduces to [Fofonoff and Montgomery, 1955; revisited by Cane, 1980]

$$\nu \frac{\partial}{\partial y} \left( \beta y - \frac{\partial u}{\partial y} \right) = 0. \tag{1.9}$$

Integrating the conserved quantity, $\beta y - \partial u/\partial y = \text{const.}$, from the equator to some latitude $y_0$ gives

$$u(0) \approx u(y_0) + \frac{1}{2} \beta y_0^2. \tag{1.10}$$

Figure 3. Zonal velocity of initially stationary water parcels that approach the equator from latitudes 1, 1.4, 1.8, 2.2, 2.6, and 3° as labeled.
Equation 1.10 requires that a water parcel with zero zonal velocity approaching the equator from 2.6° of latitude would have an eastward speed of nearly 1 m s⁻¹ at the equator. Figure 3 shows the velocity of stationary water parcels that approach the equator from various latitudes. Since symmetry about the equator requires a sharp cusp in zonal velocity at the equator, which is not observed, frictional damping must play a significant role.

The above inertial requirement for a cusp at the equator is exacerbated by the fact that vortex stretching exists because of the meridional thermocline structure. Consider the vertical vorticity component steady state inertial balance after applying equation 1.7

\[ \frac{U^2}{L^2} \left( u' \frac{\partial \zeta'}{\partial x'} + v' \frac{\partial \zeta'}{\partial y'} + w' \frac{\partial \zeta'}{\partial z'} - \zeta' \frac{\partial w'}{\partial z'} + \frac{\partial w'}{\partial x'} \frac{\partial v'}{\partial z'} - \frac{\partial w'}{\partial y'} \frac{\partial u'}{\partial z'} \right) + U \beta \left( v' - y' \frac{\partial w'}{\partial z'} \right) = 0. \]  

(1.11)

Equation 1.11 includes vortex stretching and tilting, and vertical and horizontal advection, which are all the same order as the planetary vorticity terms. This suggests that the horizontal and vertical shears of the equatorial zonal currents are important, and the equatorial inertial regime requires a substantially more complicated vorticity equation than the quasi-geostrophic vorticity applicable at mid-latitudes. In addition, the presence of planetary vorticity stretching results in an even sharper cusp in zonal currents at the equator, in the absence of friction.

![Figure 4](image-url)  

**Figure 4.** A sketch of a characteristic meridional-depth temperature section across the equator (such as in Figure 1b) showing the surface layer and the thermocline layer, which encompasses the EUC. The vertical lines represent vortex tubes that stretch near the surface while being advected poleward because of Ekman divergence, and in the thermocline while being advected equatorward because of geostrophic convergence.
Qualitatively the planetary vorticity terms of (1.11) \((v' - y' \partial w'/\partial z')\) apply to the near-surface equatorial Pacific since poleward of the equator the surface layer deepens (Figure 4; see also Figure 1b). In the surface layer, fluid flows poleward because of Ekman divergence so that in the northern hemisphere \(v' > 0\), while the vorticity column stretches \(y' \partial w'/\partial z' > 0\). In the southern hemisphere the opposite occurs, \(v' < 0\) and \(y' \partial w'/\partial z' < 0\). This creates a balance tendency near the surface in both hemispheres. In contrast, below the surface at depths within the EUC, the thermocline broadens equatorward as water converges because of geostrophy. The broadening of the thermocline layer towards the equator suggests that the planetary vorticity terms do not balance, since \(v' < 0\) while \(y' \partial w'/\partial z' > 0\), as water approaches the equator from the north. Similarly both terms reverse sign as water approaches from the south. As a result, positive relative vorticity must make up for the imbalance in the thermocline, which tends to increase the zonal velocity cusp at the equator. Since a velocity cusp does not occur in nature, friction must be enhanced at the equator within the thermocline to account for the inertial imbalance.

1.3.2 Frictional regime

The following approach used to describe the frictional regime ignores the Coriolis and advection terms. Observations presented in Chapter 5 suggest that two of the dominant terms of the zonal momentum balance are pressure gradient force and vertical friction. To explain the EUC, Arthur [1960] balanced pressure and vertical shear stress

\[
\frac{\partial P}{\partial x} = -\frac{\partial \tau}{\partial z}.
\]

A vertical scale depth \(H\) follows from

\[
H = \frac{\tau}{\partial P} / \partial x,
\]

where \(\tau\) is the surface wind stress. Using values observed in the central equatorial Pacific [Qiao and Weisberg 1997] of \(\tau = 6 \times 10^{-2} \ N \ m^{-2}\) and \(\partial P/\partial x = 6 \times 10^{-4} \ N \ m^{-2}\) we get \(H = 100 \ m\), which is the value obtained by Arthur. With respect to equations 1.3 we have

\[
-\frac{1}{\rho} \frac{\partial P'}{\partial x} + \frac{A_c U \rho_o L}{H^2 \rho_o} \frac{\partial^2 u'}{\partial z^2} = 0.
\]
Figure 5. Frictional regime vertical scale (equation 1.14) versus $U$ using $P_0/\rho_0 L = 3 \times 10^{-7}$ m s$^{-2}$ and values of $A_v$ in units of $10^{-4}$ m$^2$ s$^{-1}$.

We see that for a balance $A_v U \rho_0 L / H^2 P_0 = 1$, where

$$P_0/\rho_0 L = (1/\rho_0) \partial P/\partial x \approx 6 \times 10^{-7} \text{ m s}^{-2} \text{ and } U = 0.8 \text{ m s}^{-1}$$

we get

$$A_v = \frac{H^2 P_0}{U \rho_0 L} = 75 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}, \quad (1.14)$$

which is large and suggests the mechanism for diffusing the velocity cusp at the equator and for maintaining the thermostad beneath the EUC. Figure 5 shows how the vertical scale varies with zonal velocity using an across EUC averaged zonal pressure gradient force of $P_0/\rho_0 L = 3 \times 10^{-7}$ m s$^{-2}$. Stommel [1960] discussed the necessity for large vertical friction but also suggested that non-linear inertial terms are also important. Missing from the isolated frictional regime is the role of the earth's rotation, which is responsible for geostrophic and Ekman convergence/divergence and upwelling on the equator brought about by the coupling of inertial and frictional regimes.
1.3.3 Equatorial waves and the nearly geostrophic regime

Linear equatorial waves play a major role in the adjustment of the equatorial ocean to wind forcing. Consider the linearized shallow water zonal momentum equation, similar to that used by Matsuno [1966] to derive equatorial wave solutions

\[
\frac{\partial u}{\partial t} - \beta y v + g \frac{\partial h}{\partial x} = 0.
\]

The surface height deviation is \( h \) and \( g \) is the acceleration of gravity (alternatively, for a reduced gravity model, \( h \) is the interface depth and \( g \) is the reduced gravity). The local rate of change term indicates that equatorial waves are not entirely geostrophic and suggests a merging of inertial and geostrophic regimes about the equator. Using the same scaling factors as in (1.3) and (1.4) with the additional requirement that the vertical scale \( H \) is applied such that \( h = H h' \) gives

\[
\frac{U u'}{T} \frac{\partial u'}{\partial t'} - \frac{\beta L U y' v'}{L} + \frac{g H h'}{L} \frac{\partial h'}{\partial x'} = 0. \tag{1.15}
\]

To evaluate the terms relative to the Coriolis force term divide (1.15) by \( \beta L U \)

\[
\frac{1}{\beta L T} \frac{\partial u'}{\partial t'} - \frac{y' v'}{\beta U L^2} + \frac{g H h'}{\beta U L^2} \frac{\partial h'}{\partial x'} = 0. \tag{1.16}
\]

For all terms to be order one, characteristic time and length scales are obtained such that

\[
T = \frac{1}{\beta L} \tag{1.17}
\]

and

\[
L = \sqrt{\frac{g H}{\beta U}}. \tag{1.18}
\]

If we relate \( \sqrt{g H} \) to a wave speed and further suppose that this is the characteristic speed for the purpose of scaling, then

\[
L = \frac{c}{\sqrt{\beta}}.
\]
Thus we note that:

(i). \( T \) is the inertial period at latitude \( L \),

(ii). Phase speed is \( c = \sqrt{gH} = \frac{L}{T} \gtrsim U \),

(iii). The Rossby number is \( \frac{c}{\beta L} = 1 \), so equatorial waves are ageostrophic, and

(iv). The Rossby radius of deformation is \( R_d = \frac{\sqrt{gH}}{f} = \frac{c}{\beta} = L \).

Using a characteristic equatorial internal wave speed \( c = 2 \, \text{m} \, \text{s}^{-1} \) and \( \beta = 2.3 \times 10^{-11} \, \text{ms}^{-1} \) a length scale of \( \sim 300 \, \text{km} \) and a corresponding time-scale of \( \sim 2 \) days is obtained. Equatorial waves are then tightly confined to the equator and operate on short time-scales, or perturbations about the mean flow. It is important to note that the linear equatorial wave regime only applies where the water velocity is sufficiently smaller than the wave speeds (ii) so that the non-linear advection terms are negligible.

1.4 Summary and dissertation structure

The goal of this dissertation is to better understand the western equatorial Pacific from observations within the equatorial visco-inertial boundary layer. Alone, each regime of the equatorial boundary layer discussed above is incomplete in some way, since the real equatorial ocean is a combination of all regimes as originally suggested by Robinson [1960], Stommel [1960], Charney [1960], and Charney and Spiegel [1971]. The inertial regime tends to become cuspatte at the equator with an unrealistically short vertical scale and/or large vertical velocity, while the frictional regime provides a mechanism for frictional dissipation at the equator. The equatorial wave regime describes equatorial responses to perturbations about the mean flow, but not how the mean flows are maintained. Only through the coupling of these regimes does the complete nature of the equatorial ocean become clear. The inertial regime has a characteristic horizontal length scale of \( \sim 200 \, \text{km} \) suggesting that equatorial dynamics are tightly confined about the equator. At a distance \( \sim 2^\circ \) from the equator the influence of equatorial dynamics is substantially diminished. The zonal velocity cusp of the inertial regime and the large vertical eddy coefficient of the frictional regime both suggest that turbulent eddy viscosity and diffusion is enhanced at the equator. Consistent scales in the inertial regime maintain that vertical velocity with a magnitude of the order \( 10^{-5} \, \text{m} \, \text{s}^{-1} \) occurs at the equator independent of forcing. This suggests that, while the winds in the western equatorial Pacific are not as steady as in the eastern Pacific, relatively large vertical velocity must still occur.

Because of the warm pool's size and associated internal energy, the western Pacific has a large influence on global climate. Prior to the onset of El Niño events, westerly wind bursts occur over the western equatorial Pacific generating Kelvin wave pulses that transverse the Pacific ocean. On seasonal time-scales, the warm pool shifts north and south with an asymmetry, where the warm pool core tends to reside south of
the equator for most of the year. On the equator, even in the western Pacific, SST tends to be cooler than SST just a few degrees to the north or south. Efforts to clarify the details of how variability associated with the warm pool evolves, may improve our understanding of climate in general.

In Chapter 2 observations of the entire equatorial Pacific ocean are discussed before focusing on the western equatorial Pacific. Whereas Figure 1 provides a useful steady state picture of the eastern and central equatorial Pacific, a similar average of the western equatorial Pacific would be misleading. The seasonal evolutions of the warm pool and the associated gradients have interannual variability as large as the seasonal cycle. In the western equatorial Pacific, the thermocline is relatively deep and the surface winds tend to be weak although they are punctuated by westerly wind bursts. As a result, the EUC is deeper and weaker, and during the summer equatorial upwelling is large (Chapter 3). Because of the deeper thermocline, however, SST is influenced less strongly by equatorial dynamics than in the eastern equatorial Pacific. In contrast to the eastern Pacific, the EUC in the west tends to maintain the same speed downstream until near the dateline. This is because extra-tropical water pathways tend not to converge along the equator in the west. Instead, pathways reaching the equator from the west approach through low-latitude western boundary currents [e.g. Liu and Philander, 2001] and therefore do not contribute to EUC acceleration downstream. In the central and eastern Pacific, extra-tropical water pathways converge on the equator and the EUC's speed and transport increases eastward. While the observed structure of the equatorial variability in the western Pacific is different from the central and eastern Pacific, the underlying dynamics are the same.

Vertical velocity estimated from the divergence of horizontal velocity data recorded in the western equatorial Pacific at 0°, 156°E is analyzed in Chapter 3. This estimate suggests that geostrophic divergence is responsible for relatively deep equatorial upwelling because of a westward-directed pressure gradient force. Westerly wind driven Ekman convergence is too weak to produce downwelling; instead, during the summer, strong upwelling occurs in the western Pacific warm pool that is as large as upwelling in the eastern Pacific cold tongue. Unlike in the east, upwelling in the west is only large within the thermocline, not in the surface mixed layer, and thus has little impact on SST.

The temperature balance in Chapter 4 examines turbulent mixing at 0°, 156°E from a larger scale perspective than that achieved by microstructure estimates. Justified through mixing length arguments, the large-scale turbulent flux estimate suggests that mixing arises because of the large upwelling below the mixed layer and within the EUC. Sub-mixed layer mixing is also suggested by the momentum balance analyses in Chapter 5, but attempting to resolve the shear stress is problematic because the observing array did not fully resolve the pressure gradients.
Chapter 2

Western Equatorial Pacific Upper Ocean Variability near 0°, 156°E from February 1992 through April 1994

2.1 Introduction

The western Pacific warm pool is a subject of interest from several different viewpoints. As a region of maximum sea surface temperature (SST), air-sea coupling results in large moisture convergence and rainfall. Westerly wind bursts (WWBs) that tend to occur in boreal fall-winter, intensify convection and precipitation creating low sea surface salinity [e.g. Henin et al. 1998]. Intensified WWBs during El Niños shift the atmospheric convection, and the ocean’s response to convection, farther east [e.g. Kessler and McPhaden 1995a; Weisberg and Wang 1997b]. Upper ocean circulation exhibits a reversing jet structure driven by WWBs above the eastward equatorial undercurrent (EUC) [e.g. Cronin et al. 2000]. Meridional salinity gradients occur within the EUC because of a high salinity tongue south of the equator at ~170 m and a low salinity tongue north of the equator at ~200 m [e.g. Gouriou and Toole 1993]. Low-latitude western boundary currents as well as other current systems feed the equatorial western Pacific [e.g Tsuchiya et al. 1989; Fine et al. 1994; Bingham and Lukas 1995], and this water potentially transverses the entire Pacific Ocean [e.g. Toggweiler et al. 1991]. Subducted mid-latitude water also reaches the western Pacific via the western boundary [e.g. Liu and Huang 1998]. Below the thermocline, within 3° of the equator, are the origin of basin-wide eastward currents, the EUC and the Tsuchiya jets [e.g. Knauss 1960; Tsuchiya 1975]. Deeper equatorial jets extend down thousands of meters [e.g. Firing et al. 1998]. As a result, the western Pacific ocean dynamics and thermodynamics may directly impact global climate on annual to interannual time-scales [e.g. Weisberg and Wang 1997a; Jin 1997; Clarke and Shu 2000; Wang 2001].

The Tropical Ocean Global Atmosphere – Coupled Ocean Atmosphere Response Experiment (TOGA-COARE) was implemented to address these issues [Webster and Lukas 1992]. During the COARE Intensive Observing Period (IOP), from November 1992 until February 1993, observations by many investigators were made in the tropical western Pacific [Godfrey et al. 1998]. Coincident with the IOP observations was the Enhanced Monitoring Array (EMA), centered on the equator at 156°E and deployed over a longer time period from February 1992 through April 1994. This dissertation and several studies of the upper ocean temperature, salt, and momentum balances, are based on these moored observations [e.g. Cronin and McPhaden 1997, 1998; Cronin et al. 2000; Kennan and Niiler 2002b; Helber and Weisberg 2001].
Figure 6. Mean sea surface temperature in °C averaged over the months of a) July, August, and September of 1992 and 1993, b) over the months of February, March, April, May, November, December, and January starting in February 1992 and ending in January 1994, and c) over the months during the COARE IOP from November 1992 until February 1993. Locations of the Coupled Ocean Atmosphere Response Experiment (COARE) Enhanced Monitoring Array (EMA) moorings used in this analysis are shown in d).
Prior to COARE, Mangum et al. [1990] suggested that horizontal advection is important in the warm pool and that the zonal pressure gradient experiences seasonal reversals. Several COARE studies confirm the importance of advection [e.g. Cronin and McPhaden 1997; Richards and Inall 2000] and the pressure gradient reversals [e.g Cronin et al. 2000; Helber and Weisberg 2001], but more questions have arisen. Various microstructure experiments conducted near 2°S, 156°E report that turbulent mixing quickly diminishes below 100 m [Wijesekera and Gregg 1996; Smyth et al. 1996b]. On the contrary, EMA data suggest that mixing may increase within and below the EUC on the equator (Chapter 4) consistent with the maintenance of a thermostad beneath the equatorial thermocline [Liu and Philander 2001]. These issues remain unanswered because many warm pool dynamical and thermodynamical processes are unresolved by COARE observations (Chapter 5), and microstructure observations conducted on the equator lasted only a week [M. C. Gregg, personal communication 2003].

This chapter offers a large-scale, upper ocean description based on moored EMA velocity, temperature, salinity, and surface wind data (Figure 6) and three equatorial transects (156°E to 170°W) of shipboard ADCP and CTD data. We attempt to set a context for the shorter IOP measurements from the Intensive Flux Array that was centered south of the EMA at 1°45’S, 156°E [e.g. Weller and Anderson 1996; Wijesekera and Gregg 1996; Smyth et al. 1996a,b; Huyer et al. 1997; Feng et al. 1998, 2000]. The western equatorial Pacific is also contrasted with the eastern equatorial Pacific. Relevant western Pacific modeling studies [Rothstein et al. 1998; Zhang and Rothstein 1998; Richardson et al. 1999] as well as Pacific ocean tropical-extratropical thermocline exchange analyses [Liu et al. 1994; McCreary and Lu 1994; Liu and Huang 1998; Johnson and McPhaden 1999] are discussed relative to the observations. The EMA sampling period also included an El Niño that peaked in January 1992 (Figure 7). Our descriptions are qualified relative to this climate state.

Figure 7. The Standardized Southern Oscillation Index (SOI) from the Climate Diagnostics Center, over the COARE EMA time period.
2.2 Data

The EMA (Figure 6) centered at 0°, 156° E was deployed within the pre-existing Tropical Atmosphere Ocean (TAO) array and lasted from February 1992 until April 1994. EMA data used for this dissertation came from four subsurface moorings deployed at 0°, 154° E, 0°, 157° 30´ E, 0° 45´ N, 156° E, and the surface TAO moorings at 2° N, 156° E, 0°, 156° E and 2° S, 156° E, which were equipped with additional instrumentation for COARE [Weisberg et al. 1993, 1994; Kutsuwada and Inaba 1995; Cronin and McPhaden 1997; Iwao et al. 1998]. The data include surface winds, sea surface temperature (SST), and subsurface temperature, salinity, and horizontal velocity. Velocity data are from subsurface moorings with upward looking Acoustic Doppler Current Profilers (ADCPs) except at 0°, 156°E, which was a surface mooring with a downward looking ADCP. All subsurface temperature data are from surface moorings. Vertical velocity is calculated via the continuity equation by vertically integrating zonal and meridional gradients of horizontal velocity. Random measurement errors for horizontal velocity are ~0.02 m s⁻¹, but for horizontal gradients finite difference errors dominate and grow with depth upon vertical integration. While these errors may be large, analysis suggests that they do not exceed the vertical velocity estimate [see Chapter 3 and Helber and Weisberg 2001]. Zonal dynamic height gradients, referenced to 500db, calculated using moorings located at 0°, 154°E and 0°, 156°E, and additional TAO moorings located at 0°, 160.5°E and 0°, 165°E are also used. Errors associated with these estimates are analyzed by Cronin et al. [2000] and are not likely to influence the general observations discussed in this Chapter. As will be shown in Chapter 5 the errors associated with the zonal pressure gradient estimates are substantial when considering the details of the zonal momentum balance. Three EMA mooring deployment cruises also provided Conductivity, Temperature, and Depth (CTD) and Acoustic Doppler Current Profiler (ADCP) transects along the equator from 156°E to 170°W.

To place these EMA data within the context of the larger scale variability, TAO data from across the Pacific ocean and Reynolds SST are also examined. Quantities from the TAO array include the surface winds, 20°C isotherm depth, dynamic height, and SST. The Reynolds SST product helps track the location of the warm pool. Additional information about these data is available in Cronin and McPhaden [1997] and Helber and Weisberg [2001] (see also http://pmel.noaa.gov/tao and http://www.cdc.noaa.gov).

2.3 State of the equatorial Pacific during COARE

The equatorial Pacific ocean is composed of two regimes with the transition near the dateline. Eastward of the dateline zonal winds are generally strong, steady, and easterly*, weakening from late boreal winter into spring [e.g. Yang et al. 1997; Yu and McPhaden 1999a] while the zonal gradient of SST is also weak (Figure 8a and 10). In

* When discussing surface winds the meteorological convention will be adopted. For example, easterly winds refer to winds blowing from the east, while westerly winds are from the west, southerly winds are from the south, etc.
the far eastern Pacific a strong southerly component appears that also weakens into boreal spring. This southerly component decreases toward the west where near the dateline there is a northerly component yearlong (Figures 8b). SST exhibits a large seasonal cycle in the eastern equatorial Pacific, being warmest in boreal spring when the thermocline is shallow and the easterlies are weakest. Before SST is at a maximum, the 20°C isotherm shoals and dynamic height drops in adjustment to varying zonal winds brought about by forced equatorial wave propagation [e.g. Hayes et al. 1991]. A similar seasonal response is found in the equatorial Atlantic ocean [e.g. Weisberg and Tang 1987].

Subsurface variability in the eastern equatorial Pacific is characterized by a sharp thermocline that shoals eastward of the dateline, as seen in the mean temperature profiles of Figure 11. The EUC also shoals eastward while gaining speed and transport (Figure 12) [Johnson et al. 2002]. At 110°W, the 20°C isotherm remains at the base of the surface mixed layer but becomes deeper, relative to the surface layer toward the west. The thermocline tends to be deepest in boreal winter, whereas SST is warmest in the spring. Surface westward currents flow generally above 30 m at 110°W with a relatively inconsistent seasonal variability, but become deeper, stronger, and more seasonal towards the west. At 170°W, westward flow can be as deep as 150 m and is strongest in January, February, or March. Meridional currents across the equatorial Pacific (not shown) are weaker and have shorter variability time-scales than the zonal currents, with the exception of the tropical instability waves [e.g., reviewed by Qiao and Weisberg 1995].
The western equatorial Pacific is substantially different. It is characterized by weaker winds, which are punctuated by westerly wind bursts (WWBs) and upper ocean zonal gradients of temperature, zonal velocity, and pressure, which are more variable and of a smaller scale. The WWBs tend to occur in boreal fall/winter and are generally west of the dateline (Figure 8a), whereas the meridional winds tend to blow towards the summer hemisphere (Figures 8b). The highest SST in the western equatorial Pacific occurs in boreal fall, in contrast with the east where SST is highest in boreal spring (Figure 10). The warm pool generally follows the sun, but in the western Pacific during COARE there is asymmetry in its evolution. West of ~170°W in boreal summer, the warm pool straddles the equator while the sun is farther north and in boreal winter, the warm pool is south of the equator directly below the sun (Figures 6, 15, 16, and 17). This seasonal asymmetry with the warm pool favoring the southern hemisphere contrasts with that in the east where the Inter-Tropical Convergence Zone favors the northern hemisphere. Just prior to the IOP in late October, the warm pool begins its transition from straddling the equator to a position farther toward the southeast. By boreal spring the warm pool is at its eastern most position, while on the equator SST increases toward the east until the dateline (Figure 10). The depth of the 20°C isotherm and surface dynamic height tends to peak near the dateline and either flatten out or drop toward the west (Figure 9). Often there is a shoaling of the surface mixed-layer west of the dateline and, as will be shown below, this corresponds to a westward zonal pressure gradient force.
Figure 10. Longitude-time plot of SST in °C along the equator for the COARE EMA time period. Data are averaged from 2°S to 2°N and over five-consecutive days of sampling.

The surface mixed layer appears deepest near the 170°W mooring and shoals towards the west as the thermocline broadens (Figure 11). Zonal currents in the western equatorial Pacific are characterized by a subsurface westward jet, above the EUC and at times below an eastward surface current ("reversing jets"), that appears in boreal fall and can last until June [Cronin et al. 2000] (Figure 12). During this time of year the EUC is weaker and deeper. In the boreal summer, the water column down to 250 m flows eastward and the EUC is strongest. The speed and depth of the EUC remains the same at 147°E, 156°E, and 165°E, tending not to accelerate or shoal toward the east as it does east of the dateline. Meridional velocity shows slight seasonal variability at 156°E, tending to be weak (~0.15 m s⁻¹) southward above the EUC in July and August (not shown).

Interannual variability is apparent in the equatorial Pacific [e.g. Weisberg and Wang 1997b; Yu and McPhaden 1999b] particularly in the strength of the WWBs and the location of the transition between the two regimes of the equatorial Pacific or the eastern edge of the warm pool [e.g. Picaut et al. 2001]. In January 1992 the WWB extends as far east as 140°W while the positions along the equator of the deepest 20°C isotherm and the highest surface dynamic height are near 150°W. These features do not extend as far to the east in subsequent years of weaker El Niño conditions. The depth of the thermocline in the equatorial Pacific is also deepest in 1992, becoming shallower thereafter (Figure 11).
Figure 11. Subsurface temperature from TAO moorings along the equator in the Pacific ocean located at a) 143°E, b) 156°E, c) 165°E, d) 170°W, e) 140°W, and f) 110°W. Data are in units of °C and are smoothed with a 30 day lowpass filter. The contour interval is 2°C. To the right of each contour is the corresponding depth profile averaged over available data within the standard deviation envelope.
The interannual variation is produced in part by the generation of Kelvin wave pulses. Before the El Niño that peaks in boreal winter 1991/92, two Kelvin wave pulses are generated in response to westerly winds in the western equatorial Pacific (Figure 8), one pulse in September and another in November [Kessler and McPhaden 1995a,b] (Figure 9). The Kelvin wave response to westerlies initiates as a deepening of the thermocline (or the 20°C isotherm) and an increase in dynamic height that propagates eastward across the Pacific ocean, similar to that found during the 1982/83 El Niño [e.g. Tang and Weisberg 1984] (Figure 9). We refer to these as pulses, rather than waves, since they are shaped by a combination of forced Kelvin and Rossby wave responses to both winds and subsequent boundary reflections [e.g., Cane and Sarachik, 1977; Weisberg and Tang 1987], such that the initial deepening is followed by a shoaling, or conversely. Knox and Halpern [1982] first observed this pulse-like behavior attributed to Kelvin waves. During the height of the 1991/92 El Niño two more Kelvin wave pulses are generated in January and February 1992. In addition, a relatively weak Rossby wave pulse generated in January 1992 can be seen in Figure 9a and b propagating more slowly westward from the dateline. In boreal winter 1992/93, similar but weaker equatorial Kelvin wave pulse responses to the weaker westerly winds occur [Boulanger and Menkes 1995; Delcroix et al. 2000].

The equatorial Kelvin wave pulses of 1992/93 are generated in October-November and December, within the COARE domain just prior and during the IOP (11/92 through 2/93). The warmest and deepest surface layer occurs during the 91/92 El Niño, as a result of the Kelvin wave pulses that are generated near the dateline in November and January and arrive at 110°W in December and February, respectively (Figure 11). The westward propagating Rossby wave pulse in 1992 is evident in the sharp thermocline drop that occurs in mid-January at 170°W and in early-February at 165°E. The next deepest surface layer is because of the Kelvin pulse generated by a WWB near 165°E in late December 1992 that arrives at 140°W in late January 1993. The composite end results of such pulse propagation are thermocline slope distributions adjusted to changing wind distributions.
Figure 12. Subsurface zonal velocity from TAO moorings along the equator in the Pacific ocean located at a) 147°E, b) 156°E, c) 165°E, d) 170°W, e) 140°W, and f) 110°W. Data are in units of $10^{-2}$ m s$^{-1}$ and are smoothed with a 30 day lowpass filter. The contour interval is $15\times10^{-2}$ m s$^{-1}$ and the zero contour line is thicker. To the right of each contour is the corresponding depth profile averaged over available data within the standard deviation envelope.
2.4 Relation between wind, SST, and depth averaged zonal velocity

Easterly winds tend to accompany eastward depth averaged (30 to 220 m) zonal velocity, east of 170°W (Figure 13). The largest variance from the mean is in November through March 1991/92 when WWBs occur as far east as 170°W. At that time eastward velocity is decreased at 140°W approximately one month after the WWBs at 170°W. During boreal fall/winter 1992/93 the WWBs are weaker, remain farther west, and the corresponding zonal velocity response is weaker. Throughout the COARE EMA time period, the winds at 156°E and 165°E are more variable yet the depth averaged zonal velocity is still eastward on average, suggesting that non-local forcing through wave propagation has a major influence on zonal velocity in the western equatorial Pacific.

To demonstrate the importance of non-local forcing in the western equatorial Pacific, we parameterize upwelling-induced SST changes in response to windstress and thermocline depth using the following relation:

\[ SST = A \frac{\tau_x}{h_T} + T_0 \]

[Battisti 1988; Kessler and McPhaden 1995b]. Here SST is parameterized as proportional to zonal wind pseudostress \( \tau_x \) \((U^2/\rho \), where \( U \) is zonal wind and \( \rho \) is seawater density\) and inversely proportional to the depth \( h_T \) of an isotherm \( T \). The isotherm \( T \) is chosen to represent the top of the thermocline. The parameters \( A \) and \( T_0 \) are determined through a linear least squares fit. The idea behind this relationship is that Ekman divergence-induced upwelling is proportional to windstress while the ability of upwelling to influence SST is inversely proportional to thermocline depth. This works well in the cold tongue where the seasonal cycle of SST is large, but in the western Pacific where the interannual variation in SST is as large as the seasonal signal this relationship breaks down [C. Wang, personal communication 2002]. Figure 14 shows simulated SST (thin-line) relative to the observed SST (thick-line) at TAO mooring locations across the equatorial Pacific ocean. The isotherms chosen for each longitude are given in Table 1 and represent the top of the thermocline (Figure 11).

<table>
<thead>
<tr>
<th>Longitude</th>
<th>156°E</th>
<th>165°E</th>
<th>170°W</th>
<th>140°W</th>
<th>110°W</th>
</tr>
</thead>
<tbody>
<tr>
<td>Isotherm</td>
<td>27°C</td>
<td>27°C</td>
<td>25°C</td>
<td>23°C</td>
<td>20°C</td>
</tr>
</tbody>
</table>

The seasonal variation of SST is captured at 140°W and 110°W but not the higher frequency variability. Failure of the simulation farther west is consistent with the two regimes of the equatorial Pacific described above. For time-scales shorter than seasonal the departure between the simulated and observed SST is approximately the same across the entire Pacific, suggesting that the processes responsible for the high frequency variability in SST are not a direct response to local windstress and thermocline depth.
Figure 13. Depth averaged (30 to 220 m) zonal velocity (dark line, left-axis) and zonal surface wind (light line, right-axis) at TAO mooring locations on the equator at a) 156°E, b) 165°E, c) 170°W, d) 140°W, and e) 110°W. Units are m s\(^{-1}\) and positive corresponds to eastward current and westerly winds.
Figure 14. Simulated SST (thin-line) relative to the observed SST (dark-line) at TAO mooring locations on the equator at a) 156°E, b) 165°E, c) 170°W, d) 140°W, and e) 110°W.
Figure 15. Latitude-longitude SST maps in the western Pacific ocean averaged for every month of 1992. Units are °C with a contour interval of 0.5°C.
Figure 16.  Latitude-longitude SST maps in the western Pacific ocean averaged for every month of 1993. Units are °C with a contour interval of 0.5°C.
Figure 17. Latitude-longitude SST maps in the western Pacific ocean averaged for January, February, and March of 1994. Units are °C with a contour interval of 0.5°C.

2.5 The western Pacific equatorial transects

The University of South Florida conducted three equatorial transects from 170°W to 156°E during deployment/recovery cruises for moorings at 0°, 170°W and in the COARE EMA. In 1992 both CTD and shipboard ADCP measurements were made starting at ~171°W on 30 January and ending at ~156°E on 9 February (Figure 18). During this transect the salinity maximum deepens westward coinciding with the thermocline and the relatively high zonal and meridional velocity shear region above the EUC. West of ~174°E, the EUC has a southward component while its meridional component is nearly zero towards the east (Figure 18b). Within, or just above the thermocline, the flow tends to be northwestward with the northward flow particularly strong near the 176°E and 179°W. The EUC is centered south of the equator (see Figure 21) and tends to have relatively uniform speed zonally, between 0.4 to 0.6 m s⁻¹, on the equator from 156°E to 170°W, though the EUC’s vertical height decreases towards the east as the fresh surface westward flow deepens with the thermo/pycno-cline (Figure 18a and c-e). The zonal pressure gradient force tends to be westward, penetrating deeper towards the east, consistent with the deepening westward flow near the surface (Figure

29
Figure 18. Shipboard ADCP a) zonal and b) meridional velocity transects along the equator from the R/V Moana Wave starting 31 January 1992 at 171°W and ending on 8 February 1992 at 156°E. Also, c) temperature, d) salinity, e) sigma-t, and f) zonal pressure gradient force from CTD casts made along the same transect starting on 30 January 1992 at 170.5°W and ending on 9 February 1992 at 156°E. Zonal pressure gradient force is calculated from dynamic height references to 330 m. To the right of the CTD data contours are the corresponding zonally averaged depth profiles within the standard deviation envelope in the same units.
Figure 19. Moored ADCP a) zonal and b) meridional velocity profiles from the equator at longitudes 156°E, 165°E, and 170°W averaged from 17 March to 29 March 1993. The profiles are plotted relative to the zero lines at the mooring locations. Also, c) temperature, d) salinity, e) sigma-t, and f) zonal pressure gradient from CTD casts made along the equator from the R/V Wecoma starting on 17 March 1993 at 156°E and ending on 29 March 1993 at 173°W. Zonal pressure gradient force is calculated from dynamic height references to 330 m. To the right of the contours are the corresponding zonally averaged depth profiles within the standard deviation envelope in the same units.
Figure 20. Shipboard ADCP a) zonal and b) meridional velocity transects along the equator from the R/V Moana Wave starting 21 March 1994 at 170°W and ending on 31 March 1994 at 156°E. The units are $10^{-2}$ m s$^{-1}$ and the contour interval is $20*10^{-2}$ m s$^{-1}$. Also, c) Temperature, d) salinity, e) sigma-t, and f) zonal pressure gradient force calculated from dynamic height references to 330 m. To the right of the contours are the corresponding zonally averaged depth profiles within the standard deviation envelope in the same units.
Short periods of eastward zonal pressure gradient force coincide with weakening of the westward flow near the surface or strengthening of the EUC (e.g. near 180°). Zonal pressure gradient data cycles from positive to negative and back in 8 to 12 degrees (900 to 1300 km). This is found in all three transects and is suggestive of tropical instability waves (TIW) [Qiao and Weisberg 1995]. SST has a small eastward gradient with the core of the warm pool farther southeast (Figure 15).

In 1993 CTD casts were made starting at 156°E on 17 March and ending at 173°W on 29 March. Horizontal velocity profiles averaged from 17 March to 29 March are also plotted along the equator from moorings at 156°E, 165°E, and 170°W (Figure 19). At 156°E and 165°E, there is a westward jet and a strong pycnocline enhanced by salinity at the mixed layer base (Figure 19d), which suggests that westward flow brings fresh water from the east. In contrast, at 170°W, the westward jet is absent coinciding with a weak pycnocline since salinity varies little over the upper 150 m. Being strong at 170°W and weak at 165°E, the EUC accelerates potentially since the pressure gradient force is strong and eastward near 170°E and remains eastward from 177°E to at least until 174°W (Figure 19 a and f). Near the EMA, however, the pressure gradient force is westward, consistent with Figure 22d. In this transect, the zonal pressure gradient force fluctuations have a wavelength of 5 to 8 degrees (600 to 900 km) and coincide with the strong pycnocline west of 175°E.

In 1994 both CTD and shipboard ADCP measurements were made starting at ~170°W on 21 March and ending at ~156°E on 31 March (Figure 20). Warm fresh water near 156°E coincides with near-surface northwestward flow and produces an eastward pressure gradient force. A strong eastward zonal pressure gradient force near 167°E coincides with a strong surface northward flow with nearly zero zonal component in contrast to the westward flow on either side. Eastward pressure gradient force near the dateline results in an acceleration of the EUC towards the east from ~0.4 m s⁻¹ near 180° to ~0.8 m s⁻¹ near 170°W. The westward pressure gradient force near 176°W, however, only slightly influences the zonal acceleration, affecting the meridional flow more strongly. At this time the eastern cold tongue reaches farther west also keeping the warm pool core farther west (Figure 17).

These transects were made during a season when the winds are weak and the warm pool is positioned southeast of the EMA (Figures 15, 16, and 17). Compared to the downstream acceleration of the EUC in the eastern Pacific, the EUC's speed is relatively uniform zonally (0.4-0.6 m s⁻¹) west of the dateline. West of ~174°E, the EUC tends to have a southward component while its meridional component is nearly zero towards the east. The thermocline is relatively salty and resides just above the EUC within the high velocity shear region. According to Mangum et al. [1990], historical data suggest that the zonal pressure gradient force of the western equatorial Pacific in boreal winter is westward near the surface, reversing to eastward at the EUC core, and westward again below that. The data here suggest a tendency for the western equatorial Pacific zonal pressure gradient force to change sign with a wavelength consistent with TIW, whereas the zonally averaged sign may change from year to year. Therefore, the interannual
variability is large, and the overall zonally averaged trend from 1992 to 1994 is that the thermo/pycnocline becomes successively deeper, the subsurface salinity maximum weakens, and the pressure gradient switches westward to eastward (see the zonally averaged profiles of Figures 18, 19, and 20).

2.6 The COARE EMA domain

2.6.1 Flow field

The zonal current structure near 0°, 156°E is characterized by a seasonal acceleration/deceleration of the equatorial undercurrent (EUC). This seasonal feature also occurs at the off equator mooring locations (±0.75°) suggesting that this is not merely a meandering of the EUC. In June and July the EUC tends to accelerate (Figure 12b) coincident with the eastward zonal pressure gradient force found below 100 m (Figure 22d). While there may be large finite difference errors associated with horizontal gradients, major features are likely to be robust [see Chapter 3 and Cronin et al. 2000]. After its peak in July through August, the EUC decelerates while the entire water column flows eastward and the top of the thermocline is relatively shallow (Figure 11b). During the EUC deceleration there is an opposing (westward) zonal pressure gradient force (Figure 22d), the warm pool straddles the equator (Figures 6a, 15 and 16), and the meridional velocity tends to be southward (not shown). The meridional structure of the zonal velocity is characterized by the EUC being stronger south of the equator during July, August, and September in 1992 and 1993 (Figure 21a).

In late August or early September, a subsurface westward jet appears between 50 and 160 m while the EUC is weakest and the surface layer is deep or deepening. The westward flow is strongest in boreal winter and persists until June while the warm pool transitions towards the southeast (Figure 6b). The exception is during the 1991/92 El Niño when the westward flow disappears in December 1991 shortly after its onset. In subsequent years the westward flow lasts longer. To demonstrate the persistent nature of the westward jet during 1992, 1993 and 1994, an average of zonal velocity for the months of February, March, April, May, November, December, and January starting in February 1992 through January 1994 is performed and the meridional structure is shown in Figure 21b (Figure 6b uses the same averaging). The westward flow is stronger north of the equator while the EUC is stronger south of the equator.

The COARE IOP is during a time period when the subsurface westward jet is strong (Figure 21c). The surface flow, which is stronger towards the south (Figure 21c) at 156°E, is first eastward (Figure 12b, November and December) and then westward (January and February). Also, the EUC core is south of the equator, the thermocline is relatively deep (Figure 11), and the warm pool is southeast of the COARE domain (Figures 6c, 15, and 16).
Figure 21. Moored ADCP zonal velocity meridional sections averaged over months of July, August, and September of 1992 and 1993, b) over months of February, March, April, May, November, December, and January starting in February 1992 through January 1994, and c) over the COARE IOP from November 1992 through February 1993. Units are 10^{-2} m s^{-1}, the contour interval is 10*10^{-2} m s^{-1}, and the zero contour line is thicker.
Figure 22. Time-depth contours and mean depth profiles of a) zonal divergence, $\frac{\partial u}{\partial x}$, b) meridional divergence, $\frac{\partial v}{\partial y}$, c) vertical velocity, $w$, and d) zonal pressure gradient force, $-\frac{1}{\rho} \frac{\partial P}{\partial x}$. To the right of each contour is the corresponding mean depth profile within the standard deviation envelope. Units for divergence, vertical velocity, and pressure gradient are $10^{-7}$ s$^{-1}$, $10^{-5}$ m s$^{-1}$, and $10^{-7}$ m s$^{-2}$ with contour intervals of $4 \times 10^{-7}$ s$^{-1}$, $4 \times 10^{-5}$ m s$^{-1}$, and $2 \times 10^{-7}$ m s$^{-2}$, respectively. The zero contour line in each panel is thicker.
2.6.2 Divergence and upwelling

Near $0^\circ$, $156^\circ$E zonal velocity ($u$) is convergent ($\partial u/\partial x < 0$) on average during COARE (Figure 22a). Seasonally, the EUC decreases its speed eastward (converges) during July, August, and September, while the warm pool straddles the equator and prior to the onset of the westward jet. This deceleration is because of a westward zonal pressure gradient force (Figure 22d) [Chapter 3; Cronin et al. 2000] that is large at that time and associated with the western edge of the warm pool. Below 120 m the zonal pressure gradient changes sign on average, as is expected to drive the EUC. Also during this season, the meridional velocity ($v$) is most strongly divergent ($\partial v/\partial y > 0$) because of geostrophic divergence (Figure 22b, Chapter 3).

The sum of these two components of divergence results in convergence in vertical velocity ($\partial w/\partial z < 0$, not shown) and upwelling on average (Figure 22c; See Chapter 3 for a discussion of errors associated with the vertical velocity estimate). Divergence during this upwelling season can also be seen in the vector diagrams from 21 July, 1993 to 2 August, 1993 (Figures 23). The whole water column is flowing eastward and zonal convergence can be seen as shorter arrows on the eastern side of the array and longer arrows on the western side of the array (particularly for the 7/21 profile). Meridional divergence occasionally can be seen where the northern arrows point north, while the southern arrows point south. After the upwelling season there is a transition as the warm pool moves southward where divergence changes and downwelling occurs. The onset of the westward jet in October and November coincides with inertial gravity wave energy found in the period band centered at $\sim$20 days [Eriksen et al., 1998; Chapter 4] and weaker divergence entirely changing the nature of the flow field (compare Figures 23 and 24). The arrows in Figure 24 point northwards one day and then 8 to 12 days later they point towards the south. This occurs as the warm pool transitions southeastward.

Before this transition season, during June through mid-October, vertical velocity exhibits the strongest upwelling, the surface layer is shallow, and the warm pool straddles the equator. Once the westward jet develops, during October through May, the vertical velocity is more variable (Figure 22c), as are the horizontal components of velocity (Figure 24). This pattern is found in both years of data, though upwelling is stronger in 1992, consistent with the stronger El Niño of that year.
Figure 23. Velocity vectors from EMA moorings at depths of 40, 60, 80, 100, 120, 140, 160, 180, 200, 220, and 240 m for every four days staring at 21 July 1993 and ending 2 August 1993. Hourly data are smoothed with an 8-day lowpass filter and then sampled every four days. The tails of the arrows are spaced proportionally with the distance between the moorings.
2.7 Discussion and conclusions

There are two regimes of the equatorial Pacific ocean with a transition that is seasonally and interannually modulated about the dateline. Variability east of the dateline is characterized by strong steady easterlies and upper ocean zonal gradients that are also strong, steady, and have length scales on the order of 1000 km. A strong seasonal cycle exists in the thermocline and the SST, and the EUC gains speed and transport as it shoals eastward with the thermocline. In contrast, the western equatorial Pacific is characterized by weak winds punctuated with WWBs, while the upper ocean zonal gradients are more variable and of smaller length scales. Sea Surface Temperature is among the warmest on Earth, and the thermocline is relatively deep and zonally flat or at times sloping up towards the west. The EUC tends to maintain a relatively constant speed and transport at 147°E, 156°E, and 165°E, and there is a seasonal appearance in boreal fall of a subsurface westward jet above the EUC and at times below eastward surface currents. Meridional velocity has a stronger seasonal variability and the seasonal evolution of SST is asymmetric about the equator.

The transition region shifts seasonally and interannually being farther towards the east in winter and anomalously far towards the east during an El Niño. The warmest surface water in February 1992 was near 170°W (Figure 15 and 18c) during the strong El Niño of 1991/92, and shifted westward during the progressively weaker El Niños of
In March 1993 the warmest water was near 170°E (Figure 16 and 19c) while in March 1994 it was near the EMA (156°E, Figure 17 and 20c). In addition, the zonally averaged pressure gradient force along the equator from 156°E to 170°W was westward in February 1992 and eastward in March 1994 (Figures 18-20).

![Figure 25](image)

Figure 25. A cartoon displaying the two seasonal modes of the western equatorial Pacific.
Seasonal variability in the western equatorial Pacific is bi-modal and is linked to an asymmetrical shifting of the warm pool, which is not always directly beneath the Sun. For most of the year the warm pool remains south of the equator shifting to a position straddling the equator in boreal summer (Figures 6, 15, 16, and 17). This asymmetry is opposite to that in the east where the ITCZ tends to remain north of the equator.

In the months of July, August, and September, the warm pool SST core straddles the equator just east of the EMA (Figures 6a). Beginning in May or June the entire water column down to 250 m flows eastward with a small southward component. After its peak strength in July the EUC begins to decelerate because of a westward pressure gradient force, which is potentially associated with the western edge of the warm pool core that straddles the equator east of the EMA. As a result, geostrophic divergence drives strong upwelling below the relatively deep mixed layer. The characteristics of this season are summarized in Figure 25a.

In contrast, during boreal fall through spring, encompassing the COARE IOP, the warm pool SST core is farther towards the southeast (Figure 6b) and the zonal pressure gradient force is more variable (Figure 25b). The EUC is weaker and deeper and there appears a subsurface westward jet, which starts in September and lasts until June. A change in sign of average pressure gradient force, from westward to eastward, below the westward flow, is consistent with the EUC (Figure 22d). This is also coincident with the WWBs and the observed wave energy in a broad period band centered at ~20 days.

Recent modeling studies have investigated the upper ocean response to WWBs producing reasonable agreement with the magnitude, depth, and response time-scale of the resulting subsurface westward jet [Zhang and Rothstein 1998; Richardson et al. 1999]. Simulation of the western equatorial Pacific, however, is complicated by several factors. The lack of strong steady forcing results in a transient region governed by episodic WWBs and non-local forcing. Simulation of SST parameterized using windstress and thermocline depth (section 2.4) suggests that in the west local and non-local wave forcing, variable surface fluxes, and mixing effects govern the western Pacific. An aspect of the EMA observations that differ from recent modeling studies is the influence of westerly wind-driven divergence. Some model results of the western Pacific produce downwelling [e.g. Seager and Murtugudde 1997; Rothstein et al. 1998] because of westerly wind-driven Ekman convergence. The analysis in Chapter 3 suggests that the downwelling influence of the westerly winds is not strong enough to overcome the upwelling influence by geostrophic divergence during boreal summer months (Figure 22). The disparity could be because of the inability of the models to correctly reproduce the pressure gradient force, which changes on shorter zonal length scales than in the eastern Pacific. This may be a consequence of limited salinity data in the western equatorial Pacific, as evident by subsurface pressure gradient inaccuracies (not shown) in the Climate Diagnostics Center Pacific Ocean Reanalysis data.
It is a well-documented feature that the thermocline slopes upward to the east in the central and eastern Pacific, but the shoaling of the thermocline west of 170°W (Figures 11) has not seen much attention. This feature is important because it suggests a pressure gradient adverse to the EUC. Geostrophic streamlines on the $\sigma_i = 24.5$ surface derived by Kessler [1999] suggest that the pressure gradient substantially weakens west of 170°W on the equator. This is consistent with the observed steady speed of the EUC at these longitudes. It is not until east of 170°W that the EUC begins to accelerate where the eastward pressure gradient force is large. The lack of acceleration of the EUC in the west is consistent with tropical-extratropical exchange studies that suggest two general pathways towards the equator [Liu at al. 1994; McCreary and Lu 1994; Liu and Huang 1998; Johnson and McPhaden 1999; Liu and Philander 2001]. One pathway leads water towards the low-latitude western boundary currents that feed the warm pool. Another pathway from the north is described as a more circuitous route to the equator because of the potential vorticity constraint imposed by the trough/ridge system associated with the North Equatorial Current (NEC) and the North Equatorial Countercurrent (NECC). From the south, the absence of a southern counterpart of the NECC allows for a more direct route to the equator. For this reason, more water reaches the equator from the southern hemisphere. The waters that take the interior pathways tend to reach the equator east of the dateline. These mid-latitude ventilated waters that converge on the equator east of the dateline account for the EUC acceleration over the eastern Pacific, while in the western Pacific there is less downstream EUC variability because the feed is primarily from the western boundary.

Another feature of the equatorial oceans mentioned in Chapter 1, that is not well understood, is the thermostad region that occurs below the EUC and consists of near 13°C water that remains weakly stratified and well mixed [e.g. Tsuchiya 1981; Johnson and Moore 1997]. Microstructure experiments tend to be short in duration and do not explain this feature [e.g. Wijesekera and Gregg 1996], whereas mooring observations (while inconclusive because of unresolvable potential error sources) suggest larger mixing with depth (Chapters 4 and 5). Upwelling in the western equatorial Pacific extends at least as deep as 250 m (Figure 22c) suggesting that strong vertical mixing below the EUC is possible, but more complete equatorial process experiments are needed to clarify this issue.

As discussed in Chapter 1, the horizontal length scales of variability in the equatorial visco-inertial boundary layer are short. For this reason, the divergence, zonal pressure gradient, and upwelling estimations discussed are representative of a narrow region within ~200 km of the equator. As the warm pool migrates in its asymmetrical seasonal pattern, these dynamics alter its variability along the equator. The observed magnitudes of pressure and upwelling are consistent with the analysis in Chapter 1, suggesting that turbulent eddy viscosity and diffusion are enhanced at the equator. In the next chapter, vertical velocity will be diagnosed before more detailed temperature and momentum balances are explored in Chapters 4 and 5, respectively.
Chapter 3

Equatorial Upwelling in the Western Pacific Warm Pool

3.1 Introduction

Equatorial upwelling results from the competing influences of Ekman and geostrophic flows [e.g. Wyrtki 1981]. Where the easterly trade winds prevail in the equatorial eastern and central Atlantic and Pacific oceans, Ekman divergence and geostrophic convergence produce upwelling above the equatorial undercurrent (EUC) with downwelling below [e.g. Wyrtki and Kilonsky 1984; Bryden and Brady 1985; Weingartner and Weisberg 1991; Qiao and Weisberg 1997; Weisberg and Qiao 2000]. This is a consequence of the scale mismatch between the near-surface Ekman divergence and the zonal pressure gradient induced geostrophic convergence that penetrates deeper.

In other equatorial regions, where the prevailing easterly trade winds are not present, equatorial upwelling is poorly understood. In this Chapter we describe equatorial upwelling in the western Pacific warm pool, where winds are generally light and where during El Niño phases of El Niño-Southern Oscillation (ENSO) large episodic westerly wind bursts occur. The seasonal occurrence of these strong westerly wind bursts within a relatively calm region result in a pattern of reversing zonal currents that produce contrasting seasons with different vertical structure. A transition occurs when the entire eastward flowing water column, down to at least 260 m, is followed by a season where there is westward flow sandwiched between an eastward surface current and the eastward flowing EUC [Cronin et al. 2000]. Equatorial upwelling in the warm pool is therefore expected to be different from that in the central and eastern Pacific and the Atlantic.

Moored subsurface acoustic Doppler current profilers (ADCP) were deployed as part of the Coupled Ocean Atmosphere Response Experiment (COARE) Enhanced Monitoring Array (EMA) in the western Pacific warm pool centered at 0°, 156°E. The array consisted of five moorings that recorded horizontal velocities down to 260 m. Vertical velocity is obtained using the continuity equation by vertically integrating horizontal divergence calculated via finite difference [e.g. Bubnov 1987; Halpern and Freitag 1987; Halpern et al. 1989; Weingartner and Weisberg 1991; Johnson and Luther 1994; Qiao and Weisberg 1997; and Weisberg and Qiao 2000]. An indirect method of

estimating vertical velocity is required since values of vertical velocity, even in regions of strong equatorial upwelling, are only of the order 10^{-5} \text{ m s}^{-1}. The use of moored ADCP measurements in the present analysis provides time-depth coverage of vertical velocity from February 1992 until April 1994, from the surface to 250 m depth.

3.2 Field program, data, and methods

3.2.1 The COARE Enhanced Monitoring Array

During the COARE Intensive Observing Period (IOP) an extensive observational network was positioned near 2° S, 156° E [Godfrey et al. 1998]. The purpose of the EMA was to set the IOP observations within the context of the larger scale variability. A portion of the EMA (Figure 26) was centered at 0°, 156° E, lasted from August 1991 until April 1994, and was deployed within the pre-existing Tropical Atmosphere Ocean (TAO) array.

Our analysis primarily uses moorings that were located at 0°, 154° E, 0°, 156° E, 0°, 157° 30′ E, 0° 45′ N, 156° E and 0° 45′ S, 156° E. Subsurface horizontal velocity time-series at these moorings were profiled hourly from depths as shallow as 10 m down to 260 m [Weisberg et al. 1993, 1994; Kutsuwada and Inaba 1995; Cronin et al. 2000] (see Figure 26 and Table 2).

Table 2. Temporal and spatial coverage of the Enhanced Monitoring Array Acoustic Doppler Current Profiler data\(^1\). With a sampling increment of 10m, the top and bottom bins are the upper and lower sampling depths after initial data editing procedures, respectively.

<table>
<thead>
<tr>
<th>Location</th>
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<th>Top Bin</th>
<th>Bottom Bin</th>
<th>Sample Rate</th>
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</thead>
<tbody>
<tr>
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<td>6 Nov 92-16 Feb 93</td>
<td>30 m</td>
<td>240 m</td>
<td>Hourly</td>
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<tr>
<td></td>
<td>1 Mar 92-15 Feb 93</td>
<td>30 m</td>
<td>240 m</td>
<td>Daily</td>
</tr>
<tr>
<td>0°,156°E</td>
<td>29 Aug 91-8 Mar 93</td>
<td>10 m(^2)</td>
<td>255 m</td>
<td>Hourly</td>
</tr>
<tr>
<td>0°,156°E</td>
<td>12 Mar 93-27 Apr 94</td>
<td>30 m</td>
<td>260 m</td>
<td>Hourly</td>
</tr>
<tr>
<td>0°,157.5°E</td>
<td>11 Feb 92-4 Apr 94</td>
<td>30 m</td>
<td>260 m</td>
<td>Hourly</td>
</tr>
<tr>
<td>0.75°N,156°E</td>
<td>11 Feb 92-4 Apr 94</td>
<td>30 m</td>
<td>260 m</td>
<td>Hourly</td>
</tr>
<tr>
<td>0.75°S,156°E</td>
<td>11 Feb 92-3 Apr 94</td>
<td>30 m</td>
<td>260 m</td>
<td>Hourly</td>
</tr>
</tbody>
</table>

\(^1\)All moorings were subsurface with upward looking ADCPs, except at 0°, 156°E. See note \(^2\).
\(^2\)The 10 m top bin was available only when the surface moored ADCP was downward looking from 28 August 1991 until 3 March 1993. The rest of the time the ADCP was moored subsurface, upward looking, and the top bin was at 30 m.

Also a part of the EMA, were surface meteorological and subsurface temperature and salinity data that are useful for diagnosing the forcing components responsible for vertical velocity. Cronin and McPhaden [1997] computed wind stress using surface meteorological data from the mooring located at 0°, 156°E, with the use of the COARE v2.5b bulk flux algorithm. In addition, Cronin et al. [2000] derived dynamic height referenced to 500 m using subsurface temperature and salinity data from TAO moorings.
Figure 26. The portion of the COARE EMA relative to Pacific Ocean Reanalysis sea level pressure and surface wind stress monthly averages for March and September 1992. Sea level pressure is in units of cm and wind stress is in dyne cm\(^{-2}\). See Table 2 for data coverage details.

Located at 0°, 154°E, 0°, 156°E, 0°, 160.5°E, and 0°, 165°E. These data are used to analyze the zonal momentum on the equator.

To provide a larger scale representation of the zonal pressure gradient force and wind field, Pacific Ocean Reanalysis [Leetmaa and Ji 1989] fields were obtained from the Climate Diagnostic Center. The surface pressure and wind stress for March and September 1992 in the tropical Pacific is shown in Figure 26.
3.2.2 Horizontal velocity components

Various descriptions of the data used in this analysis are available in Kutsuwada and Inaba [1995], Helber and Weisberg [1998], Cronin et al. [2000], and Chapters 4 and 5. To highlight features of the data relevant to the vertical velocity estimation, contour plots of the horizontal velocity components (denoted $u$ and $v$, corresponding to eastward and northward in the $x$ and $y$ coordinate directions, respectively) are shown in Figure 27. For display purposes, the data are low-pass filtered to exclude time-scales shorter than 25 days.

The two-year long subsurface horizontal velocity observations reveal a seasonal cycle described in Chapter 2, which is also present in vertical velocity. The zonal current structure is characterized by a two to three month season in boreal summer when the entire upper ocean is flowing eastward. Starting in August, during this eastward flow, the EUC begins a deceleration that lasts for several months. In September, while the EUC is decelerating, there is an abrupt onset of a subsurface westward flow between 50 to 150m.
This transition, characterized by the onset of the subsurface westward flow, delineates two dynamic realms that are also apparent in the vertical velocity estimation. All of these features appear at all moorings of the EMA during both years and represent large-scale features relative to the array.

Because of the surface flow and the seasonally occurring subsurface westward flow the $u$ component record-length mean is near zero at 100 m and above. Below 100 m the influence of the EUC on the mean $u$ profile dominates (Figure 27). On average the $v$ component is weakly southward, $< 0.1 \text{ m s}^{-1}$, near the surface and from 90 m to 200 m. Near the average depth of the thermocline top, ~70 m, the mean $v$ component is zero.

3.2.3 Vertical velocity

The divergence of the vertical velocity component (denoted $w$ and positive upward in the $z$ coordinate direction) is obtained from the divergence of horizontal velocity components using the continuity equation,

$$\frac{\partial v}{\partial z} = -\left[ \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right].$$

With the condition that $w$ is zero at the surface, integration in $z$ produces $w$ as a function of depth.

When performing this calculation using data from a mooring array, additional assumptions arise out of necessity. Because finite differences are used for $\partial u/\partial x$ and $\partial v/\partial y$, the control volume implicitly assumed by the continuity equation becomes defined by the mooring spacing. In this situation, the largest zonal and meridional dimensions of the control volume are 389 km and 167 km, respectively, depending on the combination of the moorings used. Since the data are in 10 m depth bins, the vertical dimension of the control volume is 10 m. The horizontal divergence obtained at each sampling depth is then integrated in $z$ using the trapezoidal method. Under the conditions of this method, $w$ is obtained accurately with minimal finite difference error when the curvature in the spatial variability of the horizontal currents is small relative to the size of the array and the depth sampling interval. The relatively small vertical sampling interval combined with the use of the trapezoidal integration method minimizes finite difference error in the vertical. The finite difference error may be large because of the horizontal size of the array, however, and is discussed in section 3.2.4.2.

The five moorings result in three different estimates of both $\partial u/\partial x$ and $\partial v/\partial y$, producing nine different estimates of $w$, which we refer to as phases one through nine. We call them phases because an effect of varying the finite differences, is to shift the estimated phase of a wave that propagates past the array. Figure 28 identifies the phase numbering convention. Since phase five $w$ estimation uses center differences for both $\partial u/\partial x$ and $\partial v/\partial y$, the phase of a passing wave is estimated without shift. Phase six $w$
Figure 28. This figure identifies the numbering convention used to identify the phases of the vertical velocity estimation. Triangles mark the locations of the horizontal velocity moorings. The first row of numbers (1, 2, and 3) represents phases that calculate meridional divergence, $\partial v / \partial y$, using moorings located at 0.75°N, 156°E and 0°, 156°E. The second row uses moorings at 0.75°N, 156°E and 0.75°S, 156°E for meridional divergence, while the third row uses moorings at 0°, 156°E and 0.75°S, 156°E. The first column of numbers (1, 4, and 7) represents phases that calculate zonal divergence, $\partial u / \partial x$, using moorings located at 0°, 156°E and 0°, 154°E. The second column uses moorings at 0°, 157.5°E and 0°, 154°E for zonal divergence, while the third column uses moorings at 0°, 157.5°E and 0°, 156°E.

would estimate the phase of a passing wave shifted zonally in one direction, while phase four would estimate the phase of the same passing wave shifted zonally in the other direction. In addition to being the estimate of $w$ without phase shift, phase five is also the largest scale estimate from the array being identical to the average of all nine estimates.

Figure 29 shows the resulting $w$ from the nine combinations of divergence calculations. Present in all nine estimates is a bi-modal character of $w$. In boreal summer (June, July, and August), there is upwelling in each phase. After the onset of the subsurface westward flow in September, $w$ is more variable and primarily negative.

The average vertical velocity profiles displayed in Figure 30 show a profile that can vary in magnitude from phase to phase, particularly with the deeper values below 175 m, by as much as $5 \times 10^{-5}$ m s$^{-1}$. The vertical velocity profile being positive over most of the upper ocean water column, is consistent with a westerly wind driven geostrophic divergence, but does not account for the convergence expected from westerly wind driven Ekman transport. Above 75 m, $w$ is not substantially different from zero in most
phases. Westerly winds, though not constant, prevail over the averaging time period with a mean of 0.7 m s\(^{-1}\). In the presence of these winds it is surprising not to find downwelling. Upwelling driven by geostrophic divergence apparently cancels Ekman convergence and downwelling.

Figure 29. Time-depth contours of the nine phases of the vertical velocity estimation during the first year of data from 1 March 1992 until 15 February 1993. Units are 10\(^{-5}\) m s\(^{-1}\), contour interval is 5\(\times\)10\(^{-5}\) m s\(^{-1}\), and the number in the upper left hand corner of each panel corresponds to the numbering convention used in the text and shown in Figure 28. All data are filtered to remove oscillations of period 25 days or less.
Figure 30. Mean vertical velocity (thick solid line) versus depth for all nine phases of the vertical velocity estimation. The averaging period is 1 March 1992 until 15 February 1993, the units are $10^{-5}$ m s$^{-1}$, the dashed lines are plus and minus the standard deviation of the mean, and the phase number of each panel correspond to the numbering convention used in the text.
3.2.4 Error analysis

3.2.4.1 Random error

There are two types of random errors inherent in the $w$ estimation. The first type is because of the ADCPs’ error of $\sim 0.02 \text{ m s}^{-1}$ for each hourly ensemble average and the mooring positions error of $\sim 0.02$ degrees. Using standard propagation of error [i.e. Bears 1957; Chatfield 1978], these error sources in $w$ grow with depth, amounting to less than $9 \times 10^{-6} \text{ m s}^{-1}$ at 260 m for the estimate filtered to remove oscillations with a period of 5 days or less. A longer period filter reduces this error since the number of ensembles averaged together increases with increased filter period. This is not a substantial error compared to the systematic errors discussed in the next section.

The second type of random error is because of the fact that geophysical time-series are modulated over the time-scales of interest. Consequently any averaging in the experimental method has an inherent random error even in the event of perfect instruments. This error is the inherent geophysical random error, which depends on the record length relative to the intrinsic bandwidth of the time-series. Since this error is not fully quantifiable, we are left with the caveat that this experiment, if performed at some other time, would produce different results. Where possible, we will identify known deviations by longer time-scale variability such as ENSO.

When calculating averages as in Figure 27, 30, 31, 32, and 38, the standard deviation of the mean is obtained by dividing the standard deviation of the time-series by the square root of the number of degrees of freedom [e.g. Parratt, 1961]. In all cases the number of degrees of freedom, calculated from the integral of the correlation function (integral time-scale), and the standard deviation of the time-series is determined at every depth. These random error determinations are separate from the systematic errors.

3.2.4.2 Systematic errors

The dominant errors in the $w$ estimation are systematic and come from three sources: systematic instrument error, surface extrapolation error, and finite difference error. Systematic ADCP errors are compass bias, which is specified by the manufacturer to be accurate within $2^\circ$, and tilt, which is determined negligible from surface echo amplitude returns. To quantify how large compass error could become in the vertical velocity estimate, we performed a worst-case scenario using phase five $w$ estimation. Horizontal velocity vectors from the four moorings used in phase five $w$ are rotated by $2^\circ$ such that the velocity vector from the $0^\circ$, $157.5^\circ$E mooring is rotated clockwise $2^\circ$ while the velocity vector from the $0^\circ$, $154^\circ$E mooring is rotated counterclockwise $2^\circ$. In this way, the compass error from the two ADCP mooring data sets used in $\partial u/\partial x$ would add (Figure 31b). The same alternating rotation error method is used for $\partial v/\partial y$. A similar analysis was performed on an estimate of $w$ from horizontal ADCP velocity data at $0^\circ$. 
Figure 31. Time-depth contours of a) vertical velocity phase five, b) compass calibration error, c) surface extrapolation error, d) combined propagation of extrapolation and compass error, and e) vertical velocity phase variance. Vertical velocity phase standard deviation is computed as the standard deviation of the nine vertical velocity estimates at each depth and time. Propagation of error is used because each error is independent of the other. Units are $10^{-5}$ m s$^{-1}$ and the contour interval for a) and e) is $5 \times 10^{-5}$ m s$^{-1}$, while the contour interval for b), c), and d) is $10^{-5}$ m s$^{-1}$. To provide error estimates on the time-scales of interest the filtering in these data is a 120 hr (5 days) low-pass filter.

140°W in the Pacific Ocean [Weisberg and Qiao, 2000] during the Tropical Instability Wave Experiment (TIWE). Since compass error grows with depth, the largest possible error was estimated at the TIWE array to be nearly two-thirds of the vertical velocity value at 250 m. The present analysis reveals smaller compass error for this experiment, presumably because of the weaker zonal currents in the warm pool. The largest compass
error is found when the zonal velocity is greatest from April until September 1992 coinciding with the time period when the horizontal velocity components are strongest, most divergent, and produce upwelling.

Surface extrapolation error results from side lobe reflection which, for the upward looking ADCPs of this experiment, make velocity measurements above 30 m unavailable. Because, in the calculation of \( w \), the horizontal divergence is connected with the boundary condition at the surface \( (w=0) \), these near-surface velocity values are estimated by extrapolation. Since the center mooring, located at 0°, 156° E, had a downward-looking ADCP during the EMA-IOP, velocity data as shallow as 10 m are available (Table 2). The near-surface measurements from this mooring are used to estimate the magnitude of the extrapolation error. This error is smallest when the velocity profiles are extrapolated as a constant from the shallowest bin (see Table 2) to the surface.

Surface extrapolation error is taken as the square root of the difference of the squares of measured (when available) and constant extrapolated near-surface values. When measured values are not available, the error is taken as the square root of the difference of the squares of constant and constant shear extrapolated near-surface values. The extrapolation error is independent of depth below 30 m and the resulting error in the phase five \( w \) estimation is depicted in Figure 31c.

Error in \( w \), because of the extrapolated near-surface values, is largest when the horizontal velocity components near the surface (<30 m) have large vertical shear that is not accounted for by the constant extrapolation. During the months of June and October 1992, the \( u \) component velocity near the surface at 0°, 156°E, is strong westward and has large vertical shear. This coincides with periods of large surface extrapolation error (Figure 31c). Extrapolation error seems independent of the vertical structure of the deeper currents brought about by the subsurface westward flow.

Finite difference error occurs because the data are recorded at finite intervals in space and time requiring that the model equations be discretized as described in section 3.2.3. Consequently, the spatial curvature of velocity measurements creates a bias in the \( w \) estimation. To get an idea as to the magnitude of the effect of the curvature in horizontal velocity, nine \( w \) phases were plotted in Figure 29. Gradients in horizontal velocity components are not homogeneous in space; otherwise, the \( w \) phases would be identical. Unfortunately, the curvature in the horizontal currents that would produce spatial variance in \( w \), is also what causes finite difference error. Consequently, part of the variance between the nine phases is true \( w \) asymmetry, and the remainder is because of errors, including finite difference error. It is not possible to identify how much of the variance is erroneous. Figure 31e displays the standard deviation of the nine phases at each depth and time. It is reassuring to note that the total standard deviation is not much larger than the magnitude of \( w \), suggesting that the error is smaller than \( w \). Moreover, as will be seen in the following sections, physically identifiable \( w \) responses are apparent, indicating that the estimate is robust.
3.3 Results and analysis

3.3.1 First year, February 1992 through February 1993

3.3.1.1 Observed divergence

On average, over the first year of the EMA and above 125 m, the $u$ component velocity is convergent, while between 125 m and 210 m, it is divergent (Figure 32a). Below 210 m, zonal flow is again convergent. The zero crossings of $\partial u/\partial x$ correspond to the depth of the subsurface westward flow (~125 m) and the EUC (~210 m). A similar vertical structure is also found in the zonal pressure gradient (ZPG) force (Figure 32d). The ZPG force, averaged over available data, has the strongest magnitude near the surface, and is directed westward $(- (1/\rho) \partial P/\partial x < 0)$. Mean ZPG force is directed eastward over the approximate depth range where $\partial u/\partial x$ is positive. These profiles of average $\partial u/\partial x$ and $-(1/\rho) \partial P/\partial x$ are consistent with each other. Near the surface where the ZPG force tends to decelerate the zonal flow downstream, $u$ is convergent. Conversely, at the depth where the ZPG force tends to accelerate the zonal flow downstream, between approximately 110 and 220 m, $u$ is divergent.

The $v$ component velocity is divergent, on average, over the upper 250 m (Figure 32b) with a maximum at around 100 m. From geostrophic arguments, a negative ZPG force on the equator would result in meridional divergence by producing poleward geostrophic flow just off the equator. Near the surface, the mean $\partial v/\partial y$ profile is divergent and the ZPG force is negative, but at the depth where the ZPG force becomes positive, $\partial v/\partial y$ is still divergent suggesting that ageostrophic forcing is important. Moreover, above 100 m, $\partial v/\partial y$ becomes smaller in opposition to increased negative ZPG force, suggesting that westerly wind driven Ekman convergence tends to oppose the geostrophic divergence.

Temporally, when $\partial u/\partial x$ is largely negative (positive), $\partial v/\partial y$ is largely positive (negative). A good example is from June through August 1992, when $\partial u/\partial x$ is negative and $\partial v/\partial y$ is positive (Figure 32a and 32b). This coincides with the strong upwelling and negative ZPG force prior to the onset of the subsurface westward flow and during the deceleration of the EUC (Figures 27, 31, and 32d). During July, August, and early September, $\partial v/\partial y$ is positive and the ZPG force is negative. Starting in September, the ZPG force is more variable (with more positive patches) coincident with a more variable $\partial v/\partial y$ (with more negative patches). The measure of how much $\partial u/\partial x$ and $\partial v/\partial y$ fail to cancel is of course given by $\partial w/\partial z$ (Figure 32c). During June through August 1992, when horizontal divergence is large, $\partial w/\partial z$ is also large and primarily negative which results in positive $w$ on average.
Figure 32. The three terms of the continuity equation, from divergences corresponding to phase five wind estimation and zonal pressure gradient, are shown in time-depth contours. The terms are a) zonal divergence, $\partial u / \partial x$, b) meridional divergence, $\partial v / \partial y$, c) vertical divergence, $\partial w / \partial z$, and d) zonal pressure gradient calculated via dynamic height from moorings located at $0^\circ$, $160.5^\circ$E and $0^\circ$, $154^\circ$E. The contour interval is $5 \times 10^{-7}$ s$^{-1}$ for divergence and $2 \times 10^{-7}$ m s$^{-2}$ for pressure gradient. Divergence and pressure data are filtered to remove oscillations of period less than 5 days.
3.3.1.2 Modes of variability

Using time domain Empirical Orthogonal Function (EOF) computations, we analyze the modes of variability inherent in the phase five $w$ estimate. Two EOFs are computed, one for each horizontal velocity component ($u$ and $v$) from all five moorings and from 30 m to 240 m depth. The means are removed and data are filtered to remove oscillations with a period shorter than 8 days. Velocity components are separated in order to exploit the inherent symmetries of equatorial ocean dynamics, instead of attempting to interpret the phase information obtained by using velocities in complex form ($u + iv$).

Divergence and vertical velocity calculated from reconstructed data summed over many modes are examined. The first four modes, representing 51%, 29%, 8%, and 2% of the total variance in $u$ and 28%, 21%, 10%, and 8% of the total variance in $v$, capture most of the variability. Phase five $w$, calculated directly from the filtered, de-meaned $u$ and $v$, and $w$, calculated from the sum of modes 1 through 4 $u$ and $v$, are shown in Figure 33 a and b, respectively.

To identify how the variability is distributed, mode 1 and the sum of modes 1 through 4 are used to compute the contribution of $\partial u/\partial x$ and $\partial v/\partial y$ to the $w$ estimate. Using only the $u$ component EOF output, $w$ is computed from $\partial u/\partial x$ for mode 1 and the sum of modes 1 and 2, 1 through 3, and 1 through 4 (Figure 33 c, e, g, and i) and similarly for the $v$ component EOF output (Figure 33 d, f, h, and j). Vertical and horizontal lines of EOF mode 1 (Figure 33 c and d) result from the zero lag nature of a time domain EOF. Summing modes provides phase information and hence structures become more general (Figure 33 e-j). Figure 33 i and j demonstrate that the contributions to the divergence by $\partial u/\partial x$ and $\partial v/\partial y$, independently calculated, tend to oppose one another as expected.

An additional EOF analysis is performed to identify the percentage of variance coherent for each velocity component in the horizontal plane. This is done by computing the EOFs for $u$ or $v$ separately at each depth from 30 m to 240 m. A total of 22 EOF runs per component result in a depth average of 85% and 62% of total averaged variance for $u$ and $v$, respectively. Summing over the first four modes shows a depth average of 99% and 97% of the $u$ and $v$ variances are accounted for, respectively. This suggests that at each depth, the array resolved the horizontal wavelength of the variability.
Figure 33. Time domain EOF analysis using $u$ and $v$ components separately and excluding the surface extrapolation. Time-depth contours of a) de-meaned $w$, b) $w$ via sum of modes 1 through 4 $u$ and $v$, c) $w$ via mode 1 $\partial u/\partial x$, d) $w$ via mode 1 $\partial v/\partial y$, e) $w$ via sum of modes 1 & 2 $\partial u/\partial x$, f) $w$ via sum of modes 1 & 2 $\partial v/\partial y$, g) $w$ via sum of modes 1 through 3 $\partial u/\partial x$, h) $w$ via sum of modes 1 through 3 $\partial v/\partial y$, i) $w$ via sum of modes 1 through 4 $\partial u/\partial x$, and j) $w$ via sum of modes 1 through 4 $\partial v/\partial y$. The contour interval for is $5\times10^{-5}$ m s$^{-1}$ with negative values shaded.
Figure 34. Percent of total variance from single depth, single velocity component EOFs, for every sample depth from 30 m to 240 m. The time period is from 1 March, 1992 through 15 February, 1993. Solid lines represent $u$ velocity and dashed-dot lines represent $v$ velocity, with first mode in the heaviest lines and modes two and three in successively thinner lines, respectively.
3.3.2 Both years, February 1992 through April 1994

3.3.2.1 Zonal momentum balance and $w$

The zonal momentum equation evaluated with *in situ* zonal pressure gradient, zonal windstress, and the three-dimensional velocity data, provides an independent consistency check for $w$. Cronin et al. [2000], in terms of layered zonal momentum balances using COARE EMA data and the present estimate of $w$, showed that although the lowest order dynamics are in response to wind stress and zonal pressure gradients, the non-linear terms can be large. Taking another brief look at this balance we use the zonal momentum equation integrated with depth to $z=\text{-}240$ m

$$\int_{\text{-}240}^{0} \left[ \frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} \right] dz = -\frac{1}{\rho} \int_{\text{-}240}^{0} \frac{\partial P}{\partial x} dz + \frac{1}{\rho} \left( \tau_{x,0} - \tau_{x,\text{-}240} \right).$$

3.1

In this treatment $P$ is pressure, $\rho$ is density, and $\tau_{x}$ is the Reynolds stress where $\tau_{x,0}$ is equal to the surface zonal windstress and $\tau_{x,\text{-}240}$ is the vertical turbulent stress at 240 m obtained as a residual. The results shown in Figure 35 indicate that vertical advection, which was calculated using the phase five $w$ estimate, agrees in magnitude and temporal variability with pressure and surface windstress. The seasonality in vertical advection is similar to that of the other terms, showing larger magnitude in the boreal fall after the local deceleration of the EUC. The residual, turbulent stress at 240 m, also shows a seasonality indicating heightened mixing during boreal summer. More discussion of the zonal momentum balance is in Chapter 5.

3.3.2.2 Components of $\partial v/\partial y$

Evaluation of equation 3.1 reveals that the zonal momentum balance on the equator is three-dimensional, highly frictional, and pressure driven. The magnitude of the acceleration terms suggests that the earth’s rotation will make a negligible contribution in the vicinity of the equator. Nevertheless, geostrophic and Ekman divergences resulting from the Coriolis force just off the equator will contribute to the dynamics within the inertial boundary layer on the equator. To obtain a quantitative measure of the influence of Ekman and geostrophic dynamics, we use the zonal momentum equation to estimate the divergence of $v$ across the equator. Taking a finite difference of the zonal momentum equation evaluated at a distance $y$, north and south of the equator, we obtain

$$\frac{\Delta v}{\Delta y} \equiv \frac{1}{yf} \frac{Du}{Dt} + \frac{1}{yf\rho} \frac{\partial P}{\partial x} - \frac{1}{yf\rho} \frac{\partial}{\partial z} \tau_{y}.$$ 3.2

where $f$ denotes the Coriolis parameter and we have assumed that the material derivative, $Du/Dt$, zonal pressure gradient, $\partial P/\partial x$, and vertical divergence of zonal stress, $\partial \tau_{y}/\partial z$,
are uniform across the equator from $-y$ to $y$. The first term on the right represents the acceleration resulting in meridional divergence, whereas the second and third terms represent geostrophic and Ekman contributions to meridional divergence. Though acceleration and pressure gradient are not likely to be uniform across the equator, equation 3.2 identifies the major factors governing meridional divergence.

Figure 35. All terms of the zonal momentum balance integrated to 240 m. The terms are indicated to the left of each panel where the terms in [ ] correspond to the dashed line. Units are $10^5$ m$^2$ s$^{-2}$. 

60
Coherence analysis suggests that $\partial v/\partial y$ at time-scales greater than two months, derive from the acceleration and the zonal pressure gradient terms more so than the zonal windstress (Figure 36). This is qualified by the limited resolution at low frequency, since the coherence peak width is smaller than the analysis bandwidth. The implied importance of the acceleration term, suggests the region is time dependent and non-linear, and the implied importance of the pressure gradient term suggests that geostrophy contributes to equatorial upwelling. Far field winds, however, are not accounted for in this analysis and may contribute to divergence via equatorial wave-propagation and may be part of the acceleration term.

In attempt to determine the latitude at which geostrophic dynamics are important to the equatorial divergence, we perform a linear least squares regression between integrated zonal pressure gradient force and integrated meridional divergence. The obtained slope is an estimator of $1/y_f$, which multiplies the pressure gradient force term in equation 3.2. The resulting coefficient corresponds to a value of $y$ at $\pm 1.5^\circ$ from
the equator. Using this value of $y$, we calculate the second term on the right-hand-side of equation 3.2 as an estimate of the geostrophic $\partial v / \partial y$. Integrating with depth, we obtain an estimate of $w$ resulting solely from geostrophic meridional divergence. For comparison, we also integrate the observed $\partial v / \partial y$ to obtain an estimate of $w$ solely from observed meridional divergence. The results show a reasonable agreement between these two calculations indicating that geostrophic dynamics are an important part the observed $w$ (Figure 37). In particular, note the agreement during the summer season upwelling as discussed in Chapter 2. A caveat of this estimation is that the pressure gradient is not uniform across the equator, as is evident from the surface pressure fields from the Pacific reanalysis data. The pressure gradient in the western Pacific is opposite in direction to that in the central and eastern Pacific, and is asymmetric near the equator (Figure 26).

3.3.2.3 Meridional asymmetry

Because of the lack of data at the mooring located at 0°, 154°E after February 1992, there are only three possible $w$ phases for the second year of EMA data. The variance of these three estimates during the second year, is larger than the first, but can be accounted for by local forcing. In August 1993, there was a strong northward meridional wind across the equator (Figure 38a, dashed line). The result from surface Ekman divergence was downwelling north of the equator and upwelling to the south (Figure 38b and 38d) - a response to local meridional winds first suggested by Cromwell (1953). Similar patterns of asymmetric $w$ can be seen at other times during the record as well.
Figure 38. Local winds, and phases three, six, and nine of vertical velocity estimation, for the time period from 10 February 1992 until 4 April 1994 are shown. Panel a) is the zonal (solid) and meridional (dashed) wind in units of m s\(^{-1}\). The units of vertical velocity b) phase three, c) phase six, and d) phase nine are 10\(^{-5}\) m s\(^{-1}\). The panels to the right of the time-depth contours are the corresponding mean depth profile with the associated standard deviation of the mean. Data are filtered to remove oscillations of period 25 days or less.
For example, consider the opposite asymmetric response of $w$ to southward wind events in May 1992, April 1993, and February and March 1994. Also, the northward wind in October and November 1992 resulted in asymmetric $w$. There are events that produce unexpected $w$ responses suggesting other dynamics are important (for example, June 1992).

By using depth-averaged $w$ and calculating its coherence with meridional wind stress, we obtain more quantitative support for the observed meridional asymmetry in $w$ (Figure 39). Phase three $w$ is potentially coherent with meridional windstress at low frequencies and is nearly $\pi$ radians out of phase. Phase nine $w$ may also be coherent with meridional windstress at low frequencies, but is nearly in phase while phase six $w$ is not coherent with windstress. This clear relation between meridional windstress and asymmetries in $w$ occurs in a region where the local zonal winds are not coherent with $w$. The same caveat of limited low frequency resolution applies as in Figure 36.

![Figure 39](image-url)

Figure 39. Coherence of $w$ phases three, six, and nine with meridional winds for the longest period of available winds from 22 March, 1993 until 27 September, 1993. The bandwidth for the calculation is 0.002 cph resulting in 10 degrees of freedom and the significance level is 0.46 indicated by the solid line in the upper panels. The lower panels represent the phase in units of $\pi$ corresponding to the coherence squared directly above.
3.4 Discussion and Conclusions

The major finding in this paper is that large equatorial upwelling occurs in the western Pacific warm pool in the presence of westerly winds. The two-year mean $w$ estimation on the equator at $156^\circ$E is near zero from the surface down to 60 m, and then grows increasingly positive down to the deepest data at 260 m, reaching a value in excess of $3 \times 10^{-5}$ m s$^{-1}$. Local zonal winds are primarily westerly during this time period, but are not coherent with the $w$ estimation, and downwelling is not found near the surface. Instead, geostrophic divergence counteracts the effect of Ekman convergence resulting in negligible vertical motion near the surface. This is also found by Lagerloef (2000, personal communication) using satellite altimetry and wind data. Using techniques developed by Lagerloef et al., [1999], estimates of both Ekman and geostrophic components of equatorial upwelling tended to cancel, resulting in small vertical motion near the surface in the warm pool.

In contrast, the $w$ profile estimated in the central Pacific ($0^\circ$, $140^\circ$W) by Weisberg and Qiao [2000], exhibits upwelling above the EUC core and downwelling below. The difference is that at $0^\circ$, $140^\circ$W, the ZPG force is eastward and the easterly trade winds blow consistently. Both Ekman and geostrophic dynamics play an important role. Recent model results also differ from the present analysis by producing downwelling in the western Pacific [Rothstein et al. 1998]. This disparity may stem from incomplete model physics as manifested in turbulence parameterization.

Random errors in the $w$ estimation are estimated to be less than $10^{-5}$ m s$^{-1}$, while systematic errors (finite difference error, systematic instrument error, and surface extrapolation error) may be larger. Although the magnitude of all possible errors cannot be fully obtained, results suggest that the $w$ estimate is larger than the net accumulation of errors.

Upwelling is found to be largest in the boreal summer months when the entire upper water column is flowing eastward after the EUC intensity peaks and begins to decelerate against an adverse ZPG force (Figures 27, 31a, and 32). The zonal momentum balance indicates that turbulent stress is also larger when upwelling is prominent during the boreal summer. Vertical velocity becomes variable with extended periods of downwelling after the onset of subsurface westward flow in September. Interannual variability is also observed with larger upwelling during the stronger El Niño in 1992, as compared with 1993.

Agreement of the geostrophic $w$, with $w$ computed from observed $\partial v/\partial y$, suggests that the seasonal cycle in upwelling is consistent with the large-scale evolution of the pressure field. The latitude at which the geostrophic term of equation 3.2 is most closely linearly related with observed $\partial v/\partial y$, is also consistent with the inertial boundary layer length scale $(u/\beta)^{1/2}$. Using $u=0.4$ m s$^{-1}$ (Figure 27) and $\beta=2 \times 10^{-11}$ s$^{-1}$ m$^{-1}$, we find the boundary scales to about $1.3^\circ$ from the equator, which is close the value of $1.5^\circ$ found in section 3.3.2.2.
Increased variance between $w$ phases is observed in 1993 relative to 1992 and is attributed to local meridional winds. In August 1993, a strong northward wind event produced downwelling north of the equator and upwelling to the south. Smaller asymmetries in $w$ are also found in response to meridional winds throughout the two-year record. ZPG variations control the low frequency variations in $w$, while local meridional winds have a direct effect. Local zonal winds are not coherent with $w$, but large-scale winds result in the pressure gradients observed locally.

In a fully three-dimensional flow, as found on the equator, vertical advection may be an important contributor to the balances of mass, heat, and momentum. Since the role of vertical advection in these balances is tied to mixing [e.g. Weisberg and Qiao, 2000], improved observations of equatorial upwelling are necessary for advancing the parameterizations of mixing in coupled ocean-atmosphere models. This has been recognized for the equatorial cold tongue. Here we show similar importance for the warm pool, where seasonal and interannual variations of the zonal pressure gradient give rise to large, relatively deep upwelling. More specifically designed arrays of velocity profiles, coupled with temperature, salinity, and surface flux measurements can provide an effective means of diagnosing vertical velocity and the advective contributions to the balances of mass, heat, and momentum.
Chapter 4

Upper Ocean Thermodynamics on the Equator in the Western Pacific Warm Pool

4.1 Introduction

The western Pacific warm pool is the Earth’s largest mass of open ocean warm water. Sea surface temperature (SST) in the warm pool exceeds 29°C, and its location varies seasonally from straddling the equator in boreal summer to being centered south of the equator in winter/spring (Figure 40). Horizontal gradients are small, ~10^{-3} °C km^{-1}, but the thermocline structure and the surface mixed layer depth at one location can vary considerably because of the shifting location of the warmest water. For example, in mid-October 1992 at 0°, 156°E, the 28.5 °C isotherm was ~20 m deep (see Figure 42d), but by early November this isotherm dropped to ~80 m as the warm pool shifted towards the southeast. An increase of 3 °C over 60 m of water in less than a month corresponds to a local internal energy change requiring an average net heat flux of ~300 W m^{-2}.

It is important to understanding how warm pool internal energy and SST vary, because of their impact on climate. Small fluctuations in SST can have a significant influence on the global climate system [e.g. Palmer and Mansfield 1984]. For this reason, the warm pool is closely linked to climate variability such as the El Niño - Southern Oscillation (ENSO) [e.g. Meyers et al. 1986; Jin 1997; Weisberg and Wang 1997a; Clarke and Shu 2000]. Increasing knowledge about the warm pool dynamics and thermodynamics may lead to an improved understanding of ENSO.

A study by McPhaden and Hayes [1991] found that surface heat fluxes are the primary factor in surface layer internal energy and SST variability on the equator at 165°E, but many factors have a role. For example, Mangum et al. [1990] suggest that horizontal advection is important between 128°E and 170°E where zonal winds may be responsible for deep seasonal zonal pressure gradient reversals. Several studies found episodic horizontal advection events in the warm pool [Richards et al. 1995; Cronin and McPhaden 1997; Richards and Inall 2000], and Ralph et al. [1997] suggest that advection is important on average. In addition, advection may be the dominant factor in warm pool salinity changes [Smyth et al. 1996b], which can result in a salt stratified “barrier” layer [Lukas and Lindstrom 1991] that separates the mixed layer from the top of the thermocline and inhibits turbulent fluxes [Godfrey and Lindstrom 1989].

The Tropical Ocean Global Atmosphere (TOGA) Coupled Ocean - Atmosphere Response Experiment (COARE) provided further upper ocean thermodynamic balance analyses in the western Pacific warm pool. These used data from the Intensive Flux
Array (IFA), centered near 2° S, 156° E during the Intensive Observing Period (IOP), from November 1992 through February 1993 [Godfrey et al. 1998]. Several of these analyses estimated turbulent heat flux from microstructure measurements made during two cruises of the R/V *Moana Wave* [Smyth et al. 1996b; Wijesekera and Gregg 1996; Feng et al. 1998; Feng et al. 2000]. Smyth et al. [1996b] estimated a 24-day cruise-averaged (20 December until 12 January) turbulent heat flux into the thermocline of 17 ± 10 W m⁻². Under calmer conditions during the previous cruise, with measurements made near the same location, Wijesekera and Gregg [1996] estimated a 22-day cruise-averaged (11 November until 3 December) turbulent heat flux to be less than 10 W m⁻². A three-dimensional upper ocean balance by Feng et al. [2000] claimed closure to within 10 W m⁻². Their estimate of turbulent heat flux from cruise 2, for example, is 11 W m⁻² downward at 50 m. While consistent with each other, the location and limited duration of these analyses leaves open questions about variability on the equator and on time-scales longer than a few weeks. Estimates from the IFA may not represent equatorial inertial boundary layer variability since the array was located at ~2°S (see Chapter 1). Longer time-scale features of the warm pool’s equatorial seasonal evolution were not observed by the IFA because the sampling occurred in November and December 1992, while the warm pool resided south of the equator. The seasonal variability mode that occurs in July, August, and September, therefore, was not sampled (Chapter 2).

The COARE Enhanced Monitoring Array (EMA) was designed to place the off-equatorial IOP observations within the context of both these larger scales and the different dynamical balances that occur directly on the equator. Overlapping the IOP, the EMA deployment lasted from February 1992 until April 1994, and it was centered on the equator at 156°E. The EMA sampled the warm pool as it transitioned from straddling the equator to being positioned farther towards the southeast (Chapter 2). Sampling on the equator also included the variability unique to the equatorial inertial boundary layer where the sign change in the horizontal Coriolis force creates a region of increased relative vorticity and divergence.

One question of interest is whether or not turbulent mixing is enhanced within the equatorial visco-inertial boundary layer where the currents, their shears, and the vertical velocity are all strong. Using microstructure measurements, Crawford and Osborn [1979] first suggested enhanced mixing on the equator above the Equatorial Undercurrent (EUC) core while subsequent investigations have produced varied results [e.g. Moum et al. 1986; Peters et al. 1989]. Hebert et al. [1991] found that mixing was enhanced on the equator, though temporal variability tends to mask spatial variability of the microstructure measurements. Recently, Gregg et al. [2003] found that dissipation from internal wave breaking is substantially reduced at the equator relative to mid-latitudes, for low shear regions well below (between 400 m and 890 m depth) the EUC. Moored observations on the equator estimate the eddy coefficient to be at the high end of the range estimated from microstructure [Qiao and Weisberg 1997; Weisberg and Qiao 2000]. Using a one-dimensional mixed layer model to simulate the diurnal cycle in the
July, August, and September, 1992

Jan, Feb, March, April, May, Nov, and Dec, 1992

Figure 40. Average Reynolds climatological SST for a) July, August and September and b) January, February, March, April, May, November, and December 1992, and a close up view (c) of the Coupled Ocean Atmosphere Response Experiment (COARE) Enhanced Monitoring Array (EMA) mooring locations used in this analysis.

central equatorial Pacific, Schudlich and Price [1992] found that turbulent mixing penetrates farther below the diurnal mixed layer on the equator than at mid-latitudes. This result, they hypothesized, is because of the strong, persistent vertical shear in the
upper 100 m associated with the westward South Equatorial Current (SEC) overlaying the eastward EUC. In the warm pool, however, Clayson and Kantha [1999] suggested that turbulent mixing does not penetrate below the mixed layer because the EUC is too deep and weak. Anderson et al. [1996] also suggest that in the warm pool, the mixed layer is shallow enough to avoid entrainment cooling. In contrast, COARE moored observations show time-dependent vertical structure in the equatorial zonal currents that suggests elevated mixing [Cronin et al. 2000; Helber and Weisberg 1998].

The analysis in this Chapter estimates a turbulent mixing coefficient on the equator as a residual of the temperature balance equation using COARE EMA data. The data are available as daily averages and the spacing between the moorings is on the order of 100 km. Since this is a balance analysis, the resulting estimate derived from the residual represents mixing on all space and time-scales smaller than those resolved by the array. The influence of equatorial visco-inertial boundary layer dynamics, coupled with the seasonally reversing jet structure on the equator [Cronin et al. 2000] and the seasonal movements of the warm pool, may have important consequences on our estimates. For these reasons, our estimates of turbulent mixing may be substantially different from those of the IFA.

We add to the surface layer analysis of Cronin and McPhaden [1997, hereafter referred to as CM97], by including a vertical velocity estimate and using a level model framework to analyze the data over the upper 250 m of the water column. The inclusion of a vertical velocity estimate is important because temperature balance analyses from mooring data have traditionally been limited without a direct estimate of vertical advection [Bond and McPhaden 1995; Cronin and McPhaden 1998; Wang and McPhaden 1999]. A level model approach is chosen because vertical advection enters into the calculation without the need to estimate the entrainment terms of a layer analysis, which introduce additional finite difference and layer depth estimation errors.

4.2 Data

4.2.1 The Enhanced Monitoring Array

The EMA (Figure 40) centered at 0°, 156° E was deployed within the pre-existing Tropical Atmosphere Ocean (TAO) array and lasted from February 1992 until April 1994. In addition to three TAO moorings at 2° N, 156° E, 0°, 156° E and 2° S, 156° E the EMA had four moorings at 0°, 154° E, 0°, 157° 30´ E, 0° 45´ N, 156° E, and 0° 45´ S, 156° E. Subsurface horizontal acoustic Doppler current profiler (ADCP) velocities were recorded at 0°, 154° E, 0°, 156° E, 0°, 157° 30´ E, 0° 45´ N, 156° E, 0° 45´ S, 156° E, 2°N, 156°E, and 2°S, 156°E [Weisberg et al. 1993, 1994; Kutsuwada and Inaba 1995; Iwao et al. 1998; Cronin et al. 2000] and temperature profiles were recorded at 0°, 154° E, 0°, 156° E, 0°, 157° 30´ E, 2° N, 156° E, and 2° S, 156° E (Figure 40). The central 0°, 156° E mooring recorded surface meteorological data, consisting of solar radiation, air temperature, relative humidity, rain, and wind [CM97].
Coincident with the COARE IOP, from 15 September until 24 December 1992 (hereafter referred to as the EMA-IOP), the EMA had more extensive temperature coverage making three-dimensional thermodynamic calculations possible (see Table 3 [CM97]). All data are available in daily averages and the spatial coverage of subsurface temperature and velocity data are listed in Tables 3 and 4, respectively. Since the vertical resolution for temperature was irregular and relatively sparse compared to the velocity measurements by ADCP, the temperature data are linearly interpolated to match the ADCP depth sampling. Salinity measurements [Cronin and McPhaden 1998] were sparser than the temperature measurements and are not used.

Table 3. Temporal and spatial coverage of the Enhanced Monitoring Array temperature data.

<table>
<thead>
<tr>
<th>Location</th>
<th>Dates</th>
<th>Sampled Depths (m)¹</th>
</tr>
</thead>
<tbody>
<tr>
<td>0°,154°E</td>
<td>6 May 92-2 June 93</td>
<td>1,25,50,75,100,125,150, 200,250,300,500</td>
</tr>
<tr>
<td>0°,156°E</td>
<td>29 Aug 91-12 Dec 94</td>
<td>1,3,5,10,11,20,30,33,50,51,57,75,83, 100,101,107,125,132,150,157,175,183,200,201, 207,225,232,250,282,300,382,400,482,500²</td>
</tr>
<tr>
<td>0°,157.5°E</td>
<td>16 Sep 92-16 Dec 93</td>
<td>1,25,50,75,100,125,150, 200,250,300,500</td>
</tr>
<tr>
<td>2N°,156°E</td>
<td>19 Sep 92-2 Feb 93</td>
<td>1,25,50,75,100,125,150, 200,250,300,500</td>
</tr>
<tr>
<td>2S°,156°E</td>
<td>13 Sep 92-19 Nov 93</td>
<td>1,25,50,75,100,125,150, 200,250,300,500</td>
</tr>
</tbody>
</table>

¹Prior to use in the analysis data were interpolated to 10 m increments to match with the depth sampling of the ADCP velocity data.
²Not all depths were sampled simultaneously. For example, from 15 September until 24 December the depth sampling was 1, 5, 10, 33, 57, 83, 107, 132, 157, 183, 207, 232, 382, 482.

Table 4. Temporal and spatial coverage of the Enhanced Monitoring Array Acoustic Doppler Current Profiler data. With a sampling increment of 10m, the top and bottom bins are the upper and lower sampling depths after initial data editing procedures, respectively.

<table>
<thead>
<tr>
<th>Location</th>
<th>Dates</th>
<th>Top Bin</th>
<th>Bottom Bin</th>
</tr>
</thead>
<tbody>
<tr>
<td>0°,154°E</td>
<td>6 Nov 92-16 Feb 93</td>
<td>30 m</td>
<td>240 m</td>
</tr>
<tr>
<td></td>
<td>1 Mar 92-15 Feb 93</td>
<td>30 m</td>
<td>240 m</td>
</tr>
<tr>
<td>0°,156°E</td>
<td>29 Aug 91-8 Mar 93</td>
<td>10 m¹</td>
<td>255 m</td>
</tr>
<tr>
<td>0°,156°E</td>
<td>12 Mar 93-27 Apr 94</td>
<td>30 m</td>
<td>260 m</td>
</tr>
<tr>
<td>0°,157.5°E</td>
<td>11 Feb 92-1 Apr 94</td>
<td>30 m</td>
<td>260 m</td>
</tr>
<tr>
<td>.75°N,156°E</td>
<td>11 Feb 92-4 Apr 94</td>
<td>30 m</td>
<td>260 m</td>
</tr>
<tr>
<td>.75°S,156°E</td>
<td>11 Feb 92-3 Apr 94</td>
<td>30 m</td>
<td>260 m</td>
</tr>
<tr>
<td>2°N,156°E</td>
<td>6 Nov. 92-15 Feb. 93</td>
<td>30 m</td>
<td>160 m</td>
</tr>
<tr>
<td>2°S,156°E</td>
<td>6 Nov. 92-17 Feb. 93</td>
<td>30 m</td>
<td>210 m</td>
</tr>
</tbody>
</table>

¹The 10 m top bin was available only when the ADCP was downward looking from 28 August 1991 until 3 March 1993. The rest of the time the ADCP was upward looking and the top bin was at 30 m. At all other locations the ADCPs were upward looking.
4.2.2 Vertical Velocity

Vertical velocity ($w$) is estimated (Figures 41c and 42c) via the continuity equation by integrating the divergence of the horizontal velocity components ($u$, $v$) (Chapter 3). This estimation assumes that seawater is incompressible, the ocean surface is rigid, the velocity profile above 30 m may be estimated by extrapolation, and the velocity measurements represent the average velocity into the control volume.

Incompressibility and rigid lid assumptions are the most reliable. To extrapolate the horizontal velocity to the surface, where there are no data, the velocity profile is assumed to be constant from the shallowest data bin up to the surface. The resulting error is slightly larger than the random instrument errors, but they are still small relative to the vertical velocity estimate (see Appendix A).

The assumption that the velocity measurements represent the average velocity into the control volume is a consequence of finite differences. The array size determines the control volume, and the resulting finite difference errors are potentially the largest source of error in the $w$ estimation. These errors, proportional to the spatial curvature in the horizontal velocity components, are not readily estimated. Similar $w$ estimates by Weisberg and Qiao [2000] from the Tropical Instability Wave Experiment were consistent with ancillary analyses, although in the central Pacific, the horizontal velocity components were more uniform than in the warm pool during COARE [e.g. Huyer et al. 1997; Eldin et al. 1994]. Nevertheless, analysis of the present estimate suggests that the errors do not overwhelm the $w$ estimate (Chapter 3).

4.2.3 Subsurface Data

Contour plots of the velocity components ($u$, $v$, and $w$), temperature ($T$), and partial derivatives of $T$ with respect to time ($\partial T/\partial t$) and depth ($\partial T/\partial z$) at $0^\circ$, $156^\circ$E are shown in Figure 41. The coordinates ($x$, $y$, $z$) and velocity components ($u$, $v$, $w$) are positive in the eastward, northward, and upward directions, respectively. For display purposes, the data in Figure 41 are smoothed with a 30-day low-pass Butterworth filter.

Subsurface data reveal seasonal patterns in velocity and temperature that represent the large-scale variability at the EMA moorings during a moderate El Niño. In the boreal summer while the warm pool straddles the equator, the water column down to at least 250 m flows eastward (Figure 41, see also Chapters 2 and 3). During this season the surface isothermal layer is relatively shallow, meridional velocity tends to be southward, and vertical velocity exhibits the strongest upwelling especially as the EUC decelerates. During the rest of the year, starting in September, a mid-layer westward flow occurs between the mixed layer and the EUC ["reversing jet", Cronin et al. 2000]. The onset of the westward jet occurs as the warm pool begins to transition southward before the surface layer deepens (Figure 41). When the westward jet is well developed, the warm
Figure 41. Three-dimensional velocity and temperature data from the EMA mooring located at 0, 156°E are displayed in time-depth contour plots. Depth is plotted as positive downward in all figures whereas the $z$ coordinate used in the analysis equations (Section 4.3) is positive upward. Data are smoothed to remove periods of oscillation of less than 30-days. Figure shows a) $u$, b) $v$, c) $w$, d) $T$, e) $\partial T/\partial t$, and f) $\partial T/\partial z$. Horizontal velocity units are $10^{-2}$ m s$^{-1}$ with a contour interval of 15 $10^{-2}$ m s$^{-1}$ and units for $w$ are $10^{-5}$ m s$^{-1}$ with a contour interval of $5 10^{-5}$ m s$^{-1}$. Units for $T$, $\partial T/\partial t$, and $\partial T/\partial z$, are °C, $10^{-7}$ °C s$^{-1}$, $10^{-2}$ °C m$^{-1}$ with contour intervals of °C, $5 10^{-7}$ °C s$^{-1}$, 3 $10^{-2}$ °C m$^{-1}$, respectively. To the right of each contour plot is the corresponding record length average profile within the standard deviation envelope in the same units. The dark line through each contour represents the depth of the 28.5°C isotherm.
Figure 42. Three-dimensional velocity and temperature data from the EMA-Intensive Observational Period (IOP) (15 September 1992 through 24 December 1992) located at 0°, 156°E are displayed in time-depth (positive downward) contour plots. Data are smoothed (except for in d) to remove periods of oscillation of less than 8-days. Figure shows a) $u$, b) $v$, c) $w$, d) $T$, e) $\partial T/\partial t$, and f) $\partial T/\partial z$. Horizontal velocity units are $10^{-2}$ m s$^{-1}$ with a contour interval of $15 \times 10^{-2}$ m s$^{-1}$ and units for $w$ are $10^{-5}$ m s$^{-1}$ with a contour interval of $5 \times 10^{-5}$ m s$^{-1}$. Units for $T$, $\partial T/\partial t$, and $\partial T/\partial z$, are °C, $10^{6}$ °C s$^{-1}$, $10^{2}$ °C m$^{-1}$ with contour intervals of °C, $2 \times 10^{6}$ °C s$^{-1}$, $2 \times 10^{2}$ °C m$^{-1}$, respectively. The bold line in each panel represents the 28.5°C isotherm. To the right of each contour plot is the corresponding record length average profile within the standard deviation envelope in the same units.
pool is shifted southeastward, the surface layer is deep, the EUC is weak, and the vertical velocity is weaker and more variable (see Chapter 2).

The EMA-IOP is during the season when the warm pool transitions from straddling the equator to its position farther southeast. The data for this period are shown in Figure 42, where, for display purposes, all the data except temperature (Figure 42d) are smoothed using an 8-day low-pass Butterworth filter. The daily temperature data are used to show the variability independent of filtering. The depth of the 28.5 °C isotherm deepens from ~40 m to ~80 m coincident with the succession of westerly wind bursts (WWBs) beginning in October (Figure 43) and the southward shift of the warm pool. Prior to these WWBs, the surface flow is westward and the surface layer is shallow. Subsequently, the surface layer deepens, and the surface flow is eastward.

Comparing the meridional velocity with the local rate of temperature change (Figure 42b and e) reveals a frequency-doubling phenomenon. Throughout the EMA-IOP, $v$ exhibits an approximate 20-day periodicity, whereas $\partial T/\partial t$ exhibits an approximate 10-day periodicity. Frequency doubling for $\partial T/\partial t$ relative to $v$ can be accounted for by meridional advection of a temperature field that is symmetric about the equator. It may also be accounted for by vertical advection because frequency doubling is also seen in $w$. The time and depth ranges of the most coherent packet of these oscillations spans October through early November and within the thermocline. On the equator they exhibit upward phase propagation and downward energy flux as evident in the contours of $\partial T/\partial t$, $w$, and $v$. Similar oscillations in $\partial T/\partial t$ are observed at all moorings and are stronger (weaker) south (north) of the equator. This is consistent with the suggestion by Eriksen et al. [1998] that these oscillations are generated south of the equator by WWBs.

4.2.4 Surface Flux Data

The surface meteorological data collected at 0°, 156°E are processed with the COARE bulk flux algorithm [Fairall et al. 1996a,b] to produce the surface latent and sensible heat fluxes. Net longwave radiation is estimated using the Clark et al. [1974] algorithm [CM97]. Following the Weller and Anderson [1996] convention, positive net surface heat flux, $Q_0$ (units of W m$^{-2}$), corresponds to ocean heating. As in CM97

$$Q_0 = (1 - \alpha) Q_{sw} + Q_{lw} + Q_{lat} + Q_{sen} + Q_{rf}$$

(4.1)

where albedo, $\alpha=0.055$, $Q_{sw}$ is the shortwave radiation, $Q_{lw}$ is the net longwave radiation, $Q_{lat}$ is the latent heat flux, $Q_{sen}$ is the sensible heat flux resulting from the difference between air temperature and SST, and $Q_{rf}$ is the sensible heat flux resulting from rain. All the components of the net heat flux are displayed in Figure 43 along with the wind stress components.
Figure 43. Surface flux output from the COARE bulk flux algorithm input with surface meteorological data from the buoy located at 0°, 156°E. The quantities are a) zonal wind stress, \( \tau_x \), b) meridional wind stress, \( \tau_y \), c) latent heat flux, \( Q_{\text{lat}} \), d) sensible heat flux, \( Q_{\text{sen}} \), e) shortwave radiation, \( Q_{\text{sw}} \), f) longwave radiation, \( Q_{\text{lw}} \), g) sensible heat flux from rain, \( Q_{\text{rf}} \), and h) net surface flux, \( Q_0 \). Positive heat flux is downward and the smoothed solid lines for \( Q_{\text{sw}} \) and \( Q_0 \) represent 8-day low-pass filtered data.
In response to the WWBs of October and November 1992 the surface cooled because of increased latent and sensible heat fluxes and reduced solar radiation. The sensible heat flux from rain made episodic contributions. The southward transition of the warm pool in boreal fall coincides with the WWBs and surface cooling at 0°, 156°E.

4.3 Methods

We diagnose the temperature balance according to

$$\frac{\partial T}{\partial t} + \mathbf{u} \cdot \nabla T = \frac{1}{\rho_0 c_p} \frac{\partial q}{\partial z},$$

where $T$ is temperature, $\mathbf{u} = (u, v, w)$ is the velocity vector, $q$ is the vertical heat flux (positive downward) by non-advective processes, and $\rho_0 c_p$ is the specific heat capacity per unit volume, where $\rho_0$ is the reference density ($\rho_0 c_p = 4.088 \times 10^6$ J °C$^{-1}$ m$^{-3}$).

When evaluating equation 4.2 with in situ data, finite sampling and instrument inaccuracies produce inherent limitations. The daily average limits the time-scale for the resolvable fluctuations, while the spatial resolution results in finite difference errors, since the moorings are separated by 83 to 390 km. Additional differencing errors arise because gradients of different quantities are calculated over different distances. For example, in the meridional direction, temperature was recorded at 2°N and 2°S, whereas velocity was recorded at 0.75°N and 0.75°S.

To clarify the effects of limited resolution, temperature, velocity, and heat flux are partitioned into two terms: $T = \tilde{T} + T'$, $\mathbf{u} = \tilde{\mathbf{u}} + \mathbf{u}'$, and $q = \tilde{q} + q'$. The unadorned variables represent the ‘actual’ values that we are attempting to estimate, the tilde ($\sim$) quantities represent the averaged measured values and the primed (’) quantities represent the unknown higher frequency perturbations and also include random instrument and geophysical errors. To clarify finite difference errors, gradients are partitioned into two terms denoted $\partial / \partial x = \partial / \partial \tilde{x} + \partial / \partial x'$ (similarly for $t$, $y$, and $z$ gradients). The tilde ($\sim$) terms represent the gradients calculated from the sample locations and times, while the primed term represents error in the gradient calculation because of curvature in the velocity or temperature field. (Note: Captions and labels of Figures 41, 42, and 43 use unadorned variables and gradients to represent the averaged measured quantities and gradients. The remaining Figure captions and labels use the notation as defined here to emphasize the unresolved scales.)
Taking the daily average of $T$ we get $\langle T' \rangle = \langle \tilde{T} + T' \rangle = \tilde{T} + \langle T' \rangle$. Applying this average and the partitioned variables and gradients to equation 4.2 and combining and renaming some terms results in

$$\frac{\partial \tilde{T}}{\partial t} + u \frac{\partial \tilde{T}}{\partial x} + v \frac{\partial \tilde{T}}{\partial y} + w \frac{\partial \tilde{T}}{\partial z} + \langle u' \frac{\partial T'}{\partial x} \rangle + \langle v' \frac{\partial T'}{\partial y} \rangle + \langle w' \frac{\partial T'}{\partial z} \rangle = \frac{1}{\rho \sigma c_p} \frac{\partial q}{\partial z} + R + FD.$$  

(4.3)

The terms that are linear in primed quantities, such as $\langle \partial T'/\partial t \rangle$, $\langle \partial q'/\partial z \rangle$, $\langle \tilde{u} \partial T'/\partial x \rangle$, etc., are combined and renamed $R$. These are likely to be small since high frequency fluctuations tend to cancel and are of the order of random instrument and geophysical errors. Terms including finite difference error, such as $\langle \partial \tilde{T}/\partial t \rangle$, $\langle \tilde{u} \partial \tilde{T}/\partial x \rangle$, $\langle u' \partial T'/\partial x \rangle$, etc., are combined and renamed $FD$. The magnitude of the $FD$ may be large when the advective length scale is small relative to the scale of the array (see Appendix A). Products of primed terms are retained because they result in Reynolds flux divergences that are unresolved by the experiment. Daily average is suitable for this purpose because there is low co-variability between temperature and velocity at periods around a few days. Tests were done using other averages and the results are not substantially affected. This notation is used so that the first four terms and some of the flux divergence terms of equation 4.3 are identical in concept to those terms calculated from in situ data.

Since the $w$ estimate is derived from application of the continuity equation, it is assumed that $\partial u'/\partial x + \partial v'/\partial y + \partial w'/\partial z = 0$ and the primed advection terms become the Reynolds stress divergences

$$\langle u' \frac{\partial T'}{\partial x} \rangle + \langle v' \frac{\partial T'}{\partial y} \rangle + \langle w' \frac{\partial T'}{\partial z} \rangle = \frac{\partial}{\partial x} \langle u'T' \rangle + \frac{\partial}{\partial y} \langle v'T' \rangle + \frac{\partial}{\partial z} \langle w'T' \rangle.$$  

Integrating equation 4.3 yields

$$\rho \sigma c_p \int_{z_0}^{0} \left[ \frac{\partial \tilde{T}}{\partial t} + u \frac{\partial \tilde{T}}{\partial x} + v \frac{\partial \tilde{T}}{\partial y} + w \frac{\partial \tilde{T}}{\partial z} + \frac{\partial}{\partial x} \langle u'T' \rangle + \frac{\partial}{\partial y} \langle v'T' \rangle + \frac{\partial}{\partial z} \langle w'T' \rangle \right] dz + \rho \sigma c_p \langle w'T' \rangle \bigg|_{z_0}^{0}$$  

$$= \tilde{q}_{z_0}^{0} + \int_{z_0}^{0} [R + FD] dz + FI,$$  

(4.4)

where $z_0$ is a depth below the surface ($z_0 < 0$), $FI$ represents finite integration error, and the units are W m$^{-2}$. Finite integration error is similar in principle to finite difference error because of the finite sampling interval with depth.
The vertically integrated non-advective vertical flux divergence is the radiative part of the surface heat flux,

\[ \tilde{q}_0^0 = (1-\alpha)\tilde{Q}_{sw} + \tilde{Q}_{lw} - \tilde{Q}_{pen}, \quad (4.5) \]

where \( \tilde{Q}_{pen} \) is the portion of the incoming short-wave radiation that penetrates the water column to depth \( z_0 \) evaluated as a double exponential [CM97; Siegel, et al., 1995]. Absent from equation 4.5 are terms representing turbulent diffusion since these processes result in Reynolds stresses. Consequently, the remaining surface heat fluxes are part of the integrated primed vertical heat flux divergence such that

\[ -\rho_o c_p \langle w' T' \rangle_{z_0}^0 = \tilde{Q}_{lat} + \tilde{Q}_{sen} + \tilde{Q}_{ef} - \tilde{Q}_{turb}, \quad (4.6) \]

where \( \tilde{Q}_{turb} \) is the vertical turbulent heat flux at depth \( z_0 \). Equation 4.6 indicates that latent and sensible heat fluxes are diffusive processes unlike those of equation 4.5. Combining equations 4.4, 4.5, and 4.6, with the use of 4.1, we obtain

\[ \rho_o c_p \int_{z_0}^0 \left[ \frac{\partial \tilde{T}}{\partial t} + \tilde{u} \frac{\partial \tilde{T}}{\partial x} + \tilde{v} \frac{\partial \tilde{T}}{\partial y} + \tilde{w} \frac{\partial \tilde{T}}{\partial z} \right] d\tilde{z} = \tilde{Q}_0 - \tilde{Q}_{pen} - \tilde{Q}_{turb} \]

\[ -\rho_o c_p \int_{z_0}^0 \left[ \frac{\partial}{\partial x} \langle u' T' \rangle + \frac{\partial}{\partial y} \langle v' T' \rangle \right] d\tilde{z} + \int_{z_0}^0 [R + FD] d\tilde{z} + FI. \quad (4.7) \]

With the available in situ data, each term of equation 4.7 can be estimated except the vertical turbulent heat flux, \( \tilde{Q}_{turb} \), the integrated horizontal (turbulent) heat flux divergences and errors. The combination of these terms comprises the residual:

\[ \text{Residual} = -\tilde{Q}_{turb} - \rho_o c_p \int_{z_0}^0 \left[ \frac{\partial}{\partial x} \langle u' T' \rangle + \frac{\partial}{\partial y} \langle v' T' \rangle \right] d\tilde{z} + \int_{z_0}^0 [R + FD] d\tilde{z} + FI. \]

The daily average defines the upper limit of the temporal scale of the unresolved processes (Reynolds fluxes), but the array size defines the upper limit of the spatial scale. Since the residual is from an integrated balance, all of the errors are also integrated and thus tend to grow with depth. For this reason, the deeper values are more error prone. The sign convention of the residual is such that positive corresponds to a heat flux into the control volume via upward vertical turbulent heat flux and/or a net heat influx from the integrated horizontal turbulent fluxes.
4.4 Results

4.4.1 Material derivative, \( d\tilde{T}/d\tilde{t} \)

The daily estimates of the measured (tilde) terms on the left side of equation 4.3 (after smoothing with an 8-day low-pass Butterworth filter) are shown in Figure 44. Data in subsequent figures employ the same smoothing for display purposes. For an isentropic flow field, the sum of the advection terms cancel with the local rate of change, 
\[ \partial \tilde{T}/\partial \tilde{t} + \mathbf{u} \cdot \nabla \tilde{T} = 0, \]
and this tendency is often found. During October and November, there is a sequence of \( \sim \)10 day oscillations where negative (positive) \( \tilde{w}\partial \tilde{T}/\partial \tilde{z} \) opposes positive (negative) \( \partial \tilde{T}/\partial \tilde{t} \). This balance is a robust aspect of the analysis since \( \tilde{w} \) governs the variability of \( \tilde{w}\partial \tilde{T}/\partial \tilde{z} \) and is derived from the divergence of horizontal velocity, independent of temperature.

Horizontal advection, however, lacks the oscillations found in \( \partial \tilde{T}/\partial \tilde{t} \) and is substantially smaller. This is troublesome because the v component exhibits the \( \sim \)20 day oscillations (Figure 42b) suggesting that a frequency doubling should occur with a tendency of \( \tilde{v}\partial \tilde{T}/\partial \tilde{y} \) to offset the \( \sim \)10 day oscillations found in \( \partial \tilde{T}/\partial \tilde{t} \). This may be a consequence of FD and/or because of the scale mismatch between \( \partial \tilde{T}/\partial \tilde{y} \) and \( \partial \tilde{v}/\partial \tilde{y} \).

To test the latter possibility, we employed the v component and temperature data from 2°N and 2°S using a centered difference, flux divergence scheme
\[ \tilde{v} \partial \tilde{T}/\partial \tilde{y} = \partial \tilde{v}\tilde{T}/\partial \tilde{y} - \tilde{T} \partial \tilde{v}/\partial \tilde{y}. \]  
(4.8)  

Here \( \tilde{v}\partial \tilde{T}/\partial \tilde{y} \) is calculated from the right-hand-side of equation 8 using velocity and temperature from the moorings at 2°N, 0°, and 2°S at 156°E. The velocity data at 2°N and 2°S were limited in time and depth beginning in mid-November and above 160 m (Table 4). During this time, a frequency-doubling tendency with the correct phase is observed, but the magnitude is insufficient to offset \( \partial \tilde{T}/\partial \tilde{t} \) (not shown). This result may be because the mooring spacing is too large or the data are not available when the most coherent oscillations in \( \partial \tilde{T}/\partial \tilde{t} \) occurred. Alternatively, meridional advection may not be of prime importance since the \( \partial \tilde{T}/\partial \tilde{t} \) oscillations are largest in the thermocline whereas the v oscillations are largest above the thermocline.

The longer records from the EMA have also been analyzed, but data gaps prevent the terms of \( d\tilde{T}/d\tilde{t} \) from overlapping for a longer period than the EMA-IOP. Available data indicates that during the season of upwelling in the boreal summer, there is the potential for an even greater imbalance between local rate of temperature change and vertical advection. Since horizontal advection over the longer records still makes only minor contributions to the temperature balance, large mixing on the equator is implied.
Figure 44. Time-depth contours of material derivative of temperature (measured \[ \cdot \] terms of equation 4.3) are plotted in units of \( 10^{-6} \, ^\circ\text{C} \, \text{s}^{-1} \) with a contour interval of \( 2 \times 10^{-6} \, ^\circ\text{C} \, \text{s}^{-1} \) throughout. The terms are a) \( \partial T/\partial t \), b) \( \bar{u} \partial T/\partial x \), c) \( \bar{v} \partial T/\partial y \), d) \( \bar{w} \partial T/\partial z \), and e) \( dT/dt \). To the right of each contour plot is the corresponding record length average profile within the standard deviation envelope in the same units. The dark line through each contour represents the depth of the 28.5\(^\circ\text{C} \) isotherm.
The EMA-IOP is a relatively quiet time in the cycle of vertical advection (and potentially mixing) at 0°, 156°E. During the boreal summer, when the warm pool straddles the equator, vertical advection is stronger.

4.4.2 Vertically integrated balance

In Figure 45 the mean depth profile of the tilde terms and the residual of equation 4.7 are displayed within their standard deviation and standard deviation of the mean envelopes. The standard deviation below 170 m decreases slightly for local rate of change, but for vertical advection, and consequently the residual, variance continues to grow with depth. Horizontal advection has a small variance. An estimate of the random error associated with these mean profiles is the standard deviation of the mean, which is equivalent to the standard deviation divided by the square root of the number of degrees of freedom. For the residual, the number of degrees of freedom is 45, calculated by estimating the equivalent bandwidth of the raw daily time-series. The number of degrees of freedom is equal to twice the equivalent bandwidth times the record length. The result is represented by the dotted line on Figure 45, and it suggests that the mean residual is significantly different from zero with respect to random errors.

Figure 45. Mean profiles of the depth-integrated analysis (equation 4.7). The quantities are a) \( \int \frac{dT}{dt} dz \), b) \( \int [u \frac{\partial \tilde{T}}{\partial x} + v \frac{\partial \tilde{T}}{\partial y}] dz \), c) \( \int w \frac{\partial \tilde{T}}{\partial z} dz \), and d) the residual in units of \( 10^2 \text{ W m}^{-2} \). The dash-dot and dotted lines correspond to the standard deviation and the standard deviation of the mean, respectively. Standard deviation of the mean was computed using 45 degrees of freedom as determined from an equivalent bandwidth estimation.
A physical basis for these results should entail a co-variability between the terms of analysis that are summed and integrated to produce the residual. These are shown as a function of time and depth in Figure 46. Near the base of the surface (mixed) layer, delineated by the 28.5°C isotherm, and below, warming (\( \int \partial \tilde{T} / \partial t \ dz > 0 \)) generally corresponds with a deepening thermocline and a small residual. Conversely, cooling (\( \int \partial \tilde{T} / \partial t \ dz < 0 \)) generally corresponds with a shoaling thermocline and a large negative residual. During the period from 23 September to 7 October, the near-surface warming is accounted for by surface heat flux, but below the 28.5°C isotherm we see a cooling followed by a warming. The cooling portion has a large negative residual, whereas the warming portion has a smaller residual (Figure 46 e and f). For the cooling, \( \int w \partial \tilde{T} / \partial z \ dz \) tends to be positive and in opposition to the negative (warming) influence of horizontal advection, leaving \( \int \partial \tilde{T} / \partial t \ dz \) largely unchecked. For the warming, both the vertical and horizontal advection terms tend to reinforce accounting for the smaller residual (centered on 1 October). Similar arguments apply to the entire time-series (compare \( \partial \tilde{T} / \partial t \) with the residual at the 28.5°C isotherm), particularly when the thermocline is shallow, and in most cases the cooling events have the largest residuals. In general, advective effects largely account for periods of mixed layer and upper thermocline deepening and warming, whereas periods of mixed layer and upper thermocline shoaling and cooling are suggestive of increased vertical mixing.

The period of maximum rate of thermocline deepening occurs over approximately one week bracketing October 21\textsuperscript{st}. Near the surface, \( \int \partial \tilde{T} / \partial t \ dz < 0 \) since \( \tilde{Q}_0 < 0 \). Below 30 m and across the thermocline, however, \( \int \partial \tilde{T} / \partial t \ dz > 0 \) because of ocean dynamical effects. The contribution by horizontal advection changes sign during this time, whereas the vertical advection term dominates and accounts for the relatively small residual from 20-23 October. Since this period of largest downward vertical advection and largest local temperature increase coincides with relatively small residual, these results cannot simply be dismissed as error.
Figure 46. The results from the depth integrated analysis down to 150 m. The quantities are a) $\tilde{Q}_s$ (positive heat flux is downward), b) $\int \partial \tilde{T}/\partial t \, d\tilde{z}$, c) $\int [\tilde{u} \partial \tilde{T}/\partial x + \tilde{v} \partial \tilde{T}/\partial y] \, d\tilde{z}$, d) $\int \tilde{w} \partial \tilde{T}/\partial \tilde{z} \, d\tilde{z}$, e) the residual, and f) the residual sampled at the 28.5 °C isotherm depth. Units are W m$^{-2}$ for the surface heat flux and the residual sampled at the 28.5 °C isotherm. For b through e the units and contour interval are 10² W m$^{-2}$ and 2*10² W m$^{-2}$, respectively. To the right of each contour plot is the corresponding record length average profile within the standard deviation of the mean (45 degrees of freedom) envelope in the same units. The dark line through each contour represents the depth of the 28.5°C isotherm.
The patterns described above are quantified by the correlation between the Figure 46 terms: \( \int \frac{\partial \hat{T}}{\partial t} \, dz \) and \(-\int \hat{w} \frac{\partial \hat{T}}{\partial z} \, dz\). Maximum correlation occurs at zero lag and is significant (at the 95% level) between depths of 60 m and 110 m (Figure 47). Significant correlation is an important aspect of the analysis because the variability in the vertical temperature advection term comes primarily from \( w \), which is independent of temperature. While a correlation does not support the magnitude of the variability, it at least suggests that the sign and timing of the fluctuations are consistent.

![Graph showing correlation coefficient between two terms](image)

Figure 47. Zero lag correlation coefficient between \( \int \frac{\partial \hat{T}}{\partial t} \, dz \) and \(-\int \hat{w} \frac{\partial \hat{T}}{\partial z} \, dz\) at each depth. The negative sign indicates that these terms balance each other and the 95% significance level, represented by the vertical line, is 0.532.

The largest downward (negative) and upward (positive) residuals violate the above general trend and degrade the correlation. During the third week in October there is a large downward residual that is due in part to a horizontal advection event that bolsters the vertical advection and local cooling to produce a large negative residual. Another large downward vertical advection event (warming) at the end of November also coincides with a local cooling. Vertical advection is unable to account for the large local warming during the last week in October, resulting in a large upward turbulent heat flux.

Errors tend to accumulate with depth when integrating downward, consequently the variance gets larger (Figure 45) and the correlation between local temperature rate of change and vertical advection is not significant (Figure 47). The variability below 150 m (not shown), however, still reveals considerable balance between \( \int \frac{\partial \hat{T}}{\partial t} \, dz \) and \( \int \hat{w} \frac{\partial \hat{T}}{\partial z} \, dz \), but the trend of small residual associated with local warming is not found.
This is because variance in $\int \partial T/\partial t \, dz$ tends to decrease with depth below ~170 m, whereas $\int \bar{w} \partial T/\partial z \, dz$ does not (Figure 45), resulting in a larger residual. Also, there is a phase shift between $\int \partial T/\partial t \, dz$ and $\int \bar{w} \partial T/\partial z \, dz$ that at times increases with depth, preventing a better balance. The phase shift is not consistent, and therefore, a strong lagged correlation does not occur. This contributes to the residual's unrealistically large variance below 150 m, though the mean remains within a reasonable magnitude.

Figure 48. The record length average temperature mixing coefficient, $K_{mix}$, plotted versus depth in units of $10^{-4}$ m$^2$ s$^{-1}$. The heavy solid-line represents the estimate averaged over the EMA-IOP (15 September - 24 December) while the dashed-dot line represents the estimate averaged over the first half of the EMA-IOP (15 September - 31 October) and the dashed line over the second half (1 November - 24 December).
Using the residual we define a temperature mixing coefficient given by

\[ K_{mix} = \frac{-\text{Residual}}{\rho c_p \bar{\partial T/\partial z}}, \]  
(4.9)

where the residual and \( \partial \bar{T}/\partial z \) are averaged over the EMA-IOP (Figure 48). The form of equation 4.9 is based on mixing-length theory for vertical turbulent heat flux

\[ (K_T \approx Q_{turb}/\rho c_p \bar{\partial T/\partial z}), \]

where \( K_{mix} \) is analogous with a vertical eddy temperature diffusivity coefficient \( K_T \), since \( Q_{turb} \) may be the dominant term of the residual. The negative sign in the numerator appears because of the sign convention of the residual, where positive corresponds to upward vertical turbulent heat flux and/or horizontal heat influx (see section 4.3). The results show that \( K_{mix} \) grows with depth. Since the residual contains horizontal turbulent fluxes, in addition to vertical turbulent heat flux at depth \( -z \), this suggests that mixing \textit{in general} may be important below the surface layer and within the time varying reversing jets and the EUC.

There are two distinct regimes during the EMA-IOP. The first regime, from 15 September until 31 October 1992, is when the depth of the 28.5°C isotherm transitions from relatively shallow (~40 m) to relatively deep (~70 m), which is associated with the shifting of the warm pool southeastward. The second regime, from 1 November until 24 December 1992, is when the 28.5°C isotherm remains relatively deep (~70 m). During the transition period, the average residual evaluated at the depth of the 28.5°C isotherm is -125 W m\(^{-2}\) (downward and/or horizontally outward heat flux). During the period of deep mixed layer, the average residual is -90 W m\(^{-2}\). The decrease in magnitude of the residual is reflected in the estimates of \( K_{mix} \) that are larger above 120 m during the transition period.

In light of the correlations, outliers notwithstanding, a systematic finding for this EMA-IOP upper ocean (above ~150 m) equatorial analysis is that downwelling (upwelling) events tend to have residuals of relatively small (large) magnitudes. Moreover, upwelling residuals tend to be negative, consistent with an increase in potential energy by turbulent mixing, whereas downwelling residuals are sometimes positive. Such a positive residual may result from unresolved processes such as horizontal interleaving of water masses, whereupon turbulent mixing can account for residuals of either sign depending upon the relative temperature of the water injected equatorward. The systematic differences in the upwelling and downwelling events for this EMA-IOP, rectifies the residual time-series, resulting in the mean negative value (implying a downward and/or horizontally outward turbulent heat flux) at the 28.5°C isotherm.
4.5 Comparison with previous work

Microstructure measurements from the central equatorial Pacific reported by Gregg et al. [1985] show the same order of magnitude \( (10^4 \text{ m}^2 \text{ s}^{-1}) \) of vertical turbulent heat transport coefficient, \( K_h \), as the mixing coefficient, \( K_{\text{mix}} \), defined in the previous section. The processes represented by \( K_h \) and \( K_{\text{mix}} \), however, are not necessarily the same. The coefficients are similar in that they both represent turbulent diffusion of temperature, but differ in that \( K_h \) is from microstructure and \( K_{\text{mix}} \) is from a macro-scale balance analysis that may contain contributions from horizontal exchange processes and accumulated errors.

The vertical turbulent heat flux estimated near the surface by Gregg et al. [1985] is order \( 10^2 \text{ W m}^{-2} \), and is similar in magnitude to the averaged residual of this analysis. Using their values of \( K_h \) at depth also implies vertical turbulent heat flux with a similar magnitude. These numbers are also comparable in magnitude to those estimated below the EUC core by Weisberg and Qiao [2000] using moored measurements. This suggests that the magnitude of the present averaged residual represents reasonable average values for turbulent heat fluxes.

In the Pacific warm pool, the depth of the EUC is deeper (~200 m) and consequently the estimated structure of \( K_h \) is different. Near the surface, our estimate \( (K_{\text{mix}}) \) is similar to the eddy diffusivity \( (K_p) \) of Wijesekera and Gregg [1996, their Figure 1g], but our estimate increases while theirs decreases below 70 m. The difference is more than 2 orders of magnitude below 125 m (Figure 48). The corresponding turbulent flux estimates also differ. Wijesekera and Gregg [1996] estimated a vertical turbulent heat flux of less than 10 W m\(^{-2}\) at the base of the mixed layer and Smyth et al. [1996b] estimated an average turbulent heat flux of 18 W m\(^{-1}\) into the thermocline. At 50 m depth, Feng et al. [2000] estimated peak vertical turbulent heat flux of less than 100 W m\(^{-2}\) and segment averages as large as 12 W m\(^{-2}\) downward. Our average residual at the depth of the 28.5°C isotherm is -125 W m\(^{-2}\) during 15 September through 31 October and -90 W m\(^{-2}\) during 1 November through 24 December, 1992. The maximum peak value is -500 W m\(^{-2}\) in mid-October.

The larger values of our mixing estimates, relative to microstructure estimates, may be the result of several factors. Our estimate is made directly on the equator at the center of the visco-inertial boundary layer, whereas the Wijesekera and Gregg, Smyth et al., and Feng et al. analyses were off the equator near 1.75°S, 156°E, outside the boundary layer. This alone does not explain the discrepancy because additional microstructure measurements made on the equator during COARE, suggest similar results to that of Wijesekera and Gregg [1996] below 150 m [M. C. Gregg, personal communication, 2003]. Near the surface and at a subsurface peak near 125 m depth, dissipation estimates are elevated relative to 1.75°S, suggesting that dissipation within the equatorial inertial boundary layer is indeed enhanced. The values below 75 m, however,
are still substantially smaller than the present results. This may be because the nature of
the present balance methods are inherently different from microstructure methods in that
our estimates contain large-scale processes and potential horizontal flux contributions.
This is important because several aspects of turbulent mixing are still in question. For
example, the importance of non-local mixing effects, the effects of the interaction
between diapycnal mixing and lateral gradients, and the turbulent energy contribution
directly from eddies is currently not understood, i.e. the transition from macro-scale
stirring to micro-scale mixing [Muller and Garret 2002]. Thus, the nature of the fully
three-dimensional exchange processes on the equator remains an important topic to be
clarified.

4.6 Discussion

This chapter adds to the CM97 western equatorial Pacific EMA temperature
balance by using a level model framework and including a vertical velocity estimate.
Owing to differences in the local dynamics/thermodynamics and methods, the EMA and
IFA estimates are inherently different. The IFA was centered at an off-equatorial
location outside of the equatorial inertial boundary layer (1.75°S, 156°E), with a
sampling duration of less than a month and during a time when the warm pool was
centered south of the equator. The EMA-IOP was farther north relative to the warm pool
center, and most importantly, it was in the equatorial inertial boundary layer where
ageostrophic dynamics result in increased currents, vertical shears, relative vorticity, and
divergence. Additionally, the scales measured by the EMA balance analyses include
large-scale processes not captured by the microstructure measurements of the IFA.

The EMA-IOP, 15 September until 24 December 1992, occurs when the warm
pool transitions from straddling the equator to a position farther southeast. At 0°, 156°E
this results in a transition from a shallow surface layer with westward flow, to a deeper
surface layer with eastward flow. There is a subsurface westward jet throughout this time
that strengthens in the last half of the EMA-IOP after WWB onset. In boreal summer,
prior to the EMA-IOP, the warm pool straddles the equator, the upper ocean flow is
eastward, and upwelling is strongest. The warm pool begins its transition towards the
southeast in October after the onset of a subsurface westward jet followed by variable
vertical advection (see Chapter 2). The present experiment is within this seasonal
transition regime. The IOP begins in November after the transition.

Potential error sources are represented explicitly in the analysis equations in an
attempt to clarify their origin. Vertical turbulent heat flux may be the dominant term in
the residual, but the potential for non-zero integrated horizontal turbulent heat flux and
finite difference errors cannot be disregarded. There is potential for large finite
difference errors and/or experimental bias because of high frequency processes such as
solitary internal waves observed in the COARE domain in January and February 1993
[Pinkel 2000]. This high frequency variability may also exist as actual horizontal turbulent temperature fluxes. For this reason, we use the residual to estimate a temperature mixing coefficient, \( K_{\text{mix}} \). Estimates of \( K_{\text{mix}} \) grow with depth below 150 m (Figure 48) because of \( \partial \tilde{T} / \partial z \) (Figure 42f), since the residual is relatively constant with depth (Figure 45). While the magnitude of the residual's variance seems unrealistically large, the mean profiles are significantly different from zero with respect to random errors, and there is a significant correlation between the independently acquired terms \( \int \partial \tilde{T} / \partial t \, dz \) and \( - \int \bar{w} \partial \tilde{T} / \partial z \, dz \). Accurate estimates of finite difference errors (the largest likely source) and the nature of the turbulent horizontal temperature flux contribution cannot be made. Therefore, we are left with plausible but inconclusive evidence of elevated turbulent mixing on the equator and across the thermocline.

Above 150 m we find a tendency for local cooling to occur with surface layer depth shoaling, upwelling, and large negative residual, and local warming to occur with surface layer deepening, downwelling, and small residual. While these tendencies do not always occur, they are consistent enough to rectify the residual, resulting in a mean negative value (implying a downward turbulent heat flux) of reasonable magnitude. Below 150 m these tendencies are less consistent.

The largest and most rapid thermocline deepening occurs in mid-October. The residual for this event is small and coincides with the largest estimated vertical advection. We surmise that the surface layer deepening is primarily by advection and that the residual is not merely a reflection of vertical advection error. The thermal eddy diffusivity coefficient profiles estimated from the residuals above 120 m, are larger during the higher vertical shear regime prior to the thermocline transition than (Figure 48), and they are in reasonable agreement with the previous equatorial estimates of Gregg, et al. [1985].

Mixing length theory [e.g., Tennekes and Lumley, 1972] provides a framework in which to comment on the consistency between the observed scales and the estimated eddy diffusivities. The vertical eddy temperature diffusivity scales with \( w \) and vertical length scale \( l \), such that \( K_T \sim wl \). The application of this theory is reasonable, provided \( w \) is correlated with temperature and \( T'/l \sim \partial T/\partial z \), where \( T' \) is the temperature perturbation. Figure 47 confirms the first requirement, and from Figure 42 we have \( T' \sim 3 \, ^\circ C \) and \( l \sim 50 \, m \), suggesting that \( \partial T/\partial z \sim 0.06 \, ^\circ C \, m^{-1} \), which is the magnitude observed (Figure 42f). Therefore, using an approximate \( w \) magnitude of \( 2 \times 10^{-5} \, m \, s^{-1} \), we estimate \( K_T \sim 10 \times 10^{-4} \, m^2 \, s^{-1} \), demonstrating a consistency with observed scales and the analogy with \( K_{\text{mix}} \) in equation 4.9. These scales, as evident in the vertical structure of the zonal and meridional velocity components, result from layering and interleaving of water masses. Water masses entering the equatorial inertial boundary layer are deformed as
relative vorticity and divergence increase, altering the macro-scale and potentially facilitating mixing. For example, without mixing the EUC would be a cuspate feature. How mixing occurs, both horizontally and vertically, and how homogeneous these processes are in both time and space all remain important questions to be answered. The nature of the equatorial inertial boundary layer coupled with the large observed time-scales of variability suggest that mixing is both large and non-homogeneous. Just how to approach this problem through a combination of techniques remains an illusive question, and one of importance to ocean modeling in relation to climate.
Chapter 5

Large Turbulent Shear Stress in the Equatorial Western Pacific

5.1 Introduction

Using EMA observations and similar methods to those used for the temperature balance in Chapter 4, zonal momentum balances have also been performed. The time period available for three-dimensional zonal momentum balance analyses, however, is longer since velocity measurements are more extensive than temperature. Cronin et al. [2000] analyzed the zonal momentum balance from March 1992 through March 1994. Using the same data, but over a shorter time period (May 1992 through December 1992) when the most reliable zonal dynamic height gradients were available, Kennan and Niiler [1998, 2002a,b] also analyzed the zonal momentum balance.

These momentum analyses concern the portion of the upper ocean within the center of the western equatorial Pacific warm pool during an extended El Niño that peaked in January 1992. Surface winds were westerly on average with intense westerly wind bursts. Subsurface zonal velocity was characterized by a deep EUC (relative to other areas of the equatorial Pacific) and a seasonally occurring (September through May) subsurface westward flow above the EUC centered near 120 m [“reversing jets,” Cronin et al. 2000]. Vertical velocity exhibited a similar seasonal pattern with strong upwelling in boreal summer, and downwelling, on average, in boreal fall and winter (Figure 49, see also Chapters 2 and 3).

Cronin et al. [2000; hereafter CMW] found that across a surface layer (bounded by the 28°C isotherm), the local zonal acceleration balanced with the zonal surface wind stress and below the surface layer it balanced with the zonal pressure gradient force, both to lowest order. Using the same data, but casting the momentum analysis in a different light, Kennan and Niiler [1998, 2002a,b; hereafter KN] concluded that local accelerations of the zonal currents were primarily driven by the vertical shear stress divergence at all depths. While both analyses provide a valid point of view regarding the momentum balance, there are inherent limitations because of experimental constraints. In the CMW analysis the lowest order balance was complicated by the residual, which suggested large vertical shear stress divergence of competing magnitude with the zonal pressure gradient force and the local acceleration. The KN analysis imposed an additional assumption that the vertical shear stress is zero at the EUC core. In this note we identify the essential differences between these analyses, calculate a zonal shear stress profile, and suggest that the experiment did not fully resolve the physics.

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Figure 49. Eight-day low-pass filtered a) zonal wind and subsurface b) zonal ($u$) and c) vertical ($w$) velocity from 10 May through 20 December 1992. Wind is in units of m s$^{-1}$, $u$ is in units of 10$^{-2}$ m s$^{-1}$, and $w$ is in units of 10$^{-5}$ m s$^{-1}$. The contour interval for $u$ is 15*10$^{-2}$ m s$^{-1}$ and for $w$ is 5*10$^{-5}$ m s$^{-1}$. To the right of each contour is the corresponding mean depth profile within the standard deviation envelope. The cross (X) on the right hand axis of panel a) designates the mean value of the zonal wind (1.09 m s$^{-1}$).

5.2 Methods Contrasted

The difference between the two analysis methods was the treatment of zonal pressure gradient. Moored subsurface salinity and temperature observations were used to produce dynamic height at locations 0°, 154°E, 0°, 156°E, and 0°, 160.5°E. CMW referenced the dynamic height to 500 db, while KN referenced it to the surface. Using the reasonable assumption that the pressure gradient at 500 db was small, CMW used the 500 db referenced zonal dynamic height gradient as the total zonal pressure gradient. In contrast, KN assumed the surface referenced zonal dynamic height gradient to be the
baroclinic pressure gradient and obtained the barotropic pressure gradient using the zonal momentum balance. By setting the turbulent shear stress to be identically zero at the EUC core, the barotropic pressure gradient was obtained as the residual of the zonal momentum equation evaluated with \textit{in-situ} data. It was then compared with TOPEX/POSEIDON sea surface height data for a short time period of overlapping data in October, November, and December 1992. Reasonable agreement was achieved.

The approach of KN may have improved the absolute pressure gradient magnitude and its temporal variability, but the depth dependence comes from the baroclinic portion of the pressure gradient, which was the same in both methods. For this reason, pressure gradient profiles from these methods differed only by a time-dependent magnitude offset.

5.3 Zonal Momentum Balance

Following a similar approach to that of the temperature balance in Chapter 4, we diagnose the equation

\[
\frac{\partial \tilde{u}}{\partial \tilde{t}} + u \frac{\partial \tilde{u}}{\partial \tilde{x}} + v \frac{\partial \tilde{u}}{\partial \tilde{y}} + w \frac{\partial \tilde{u}}{\partial \tilde{z}} + \frac{1}{\rho_0} \left( \frac{\partial \tau_{xx}}{\partial \tilde{x}} + \frac{\partial \tau_{xy}}{\partial \tilde{y}} + \frac{\partial \tau_{xz}}{\partial \tilde{z}} \right) = - \frac{1}{\rho_0} \frac{\partial \tilde{P}}{\partial \tilde{x}} + R + FD,
\]

where \( u, v, \) and \( w \) have been partitioned into averaged measured values (denoted with \( \sim \)) and unresolved higher frequency perturbations (denoted with \( \prime \)). The partial derivatives relative to \( \sim \) variables, represent finite differences associated with the moorings spacing (see section 4.3). The unresolved perturbations are contained in the Reynolds stresses, such that

\[
\tau_{xx} = \rho_0 \langle u'u' \rangle, \quad \tau_{xy} = \rho_0 \langle v'u' \rangle, \quad \text{and} \quad \tau_{xz} = \rho_0 \langle w'u' \rangle.
\]

Errors consist primarily of random (R) and finite difference (FD), while there are other less important systematic errors such as compass error discussed in Chapter 3.

Using subsurface velocity from the TOGA-COARE divergence array, the Eulerian material derivative terms \( \left( \dot{D}u/\dot{D}t = \partial u/\partial t + \overline{u} \partial u/\partial x + \overline{v} \partial u/\partial y + \overline{w} \partial u/\partial z \right) \) are calculated for the momentum analyses (Figure 50b). One of the largest terms, next to local acceleration (Figure 50a), is vertical advection (not shown) and is possibly the most error prone aspect of \( \dot{D}u/\dot{D}t \). The error derives from the estimate of vertical velocity, which is difficult to measure because of its small magnitude (order \( 10^{-5} \, \text{m s}^{-1} \)). Nevertheless, a similar array in the past did provide physically reasonable vertical velocity estimates in the central equatorial Pacific ocean [Weisberg and Qiao, 2000].
Figure 50. Eight-day low-pass filtered a) local acceleration, b) material derivative, and zonal pressure gradient force from the pair of moorings at c) 0°, 156°E and 0°, 154°E, d) 0°, 160.5°E and 0°, 154°E, and e) 0°, 160.5°E and 0°, 156°E. The units are $10^{-7} \text{ m s}^{-2}$ and the contour interval is $4 \times 10^{-7} \text{ m s}^{-2}$. To the right of each contour is the corresponding mean depth profile within the standard deviation envelope.
Error analysis of the vertical velocity used in the present momentum balance revealed that while the error cannot be completely determined, the calculation is supported by ancillary evidence (Chapter 3).

Another potential source of error in the zonal momentum balance is the pressure gradient estimate, the quality of which is limited by the availability of salinity data (used to calculate density), vertical temperature and salinity resolution, and the mooring spacing. For this Chapter we chose a time period from 10 May until 20 December 1992, when temperature and salinity were measured at 0°, 154°E, 0°, 156°E, and 0°, 160.5°E. Salinity coverage is a problem because dynamic height is referenced to 500 db, but salinity is available only down to 75 m at 0°, 154°E and down to 30 m at 0°, 160.5°E. At 0°, 156°E salinity is available down to 200 m [Cronin and McPhaden, 1998]. The vertical structure of density could be improperly sampled because of coarse vertical resolution and the lack of salinity below these depths. This combined with the zonal spacing of the moorings (2° to 6.5°), that do not match up with the velocity measurements zonal spacing (between moorings at 0°, 154°E and 0°, 157.5°E), make higher mode variability less likely to be resolved.

In order to calculate density at deeper depths, salinity is estimated using temperature-salinity relationships derived from CTD and climatological data. The result provides three possible estimates of zonal pressure gradient using different combinations of moorings (Figure 50c, d, and e). Examining the three estimates indicates that the pressure gradient is substantially different depending on which moorings are used for the calculation. Particularly, during September through December, when the subsurface westward flow is present, the pressure gradient seems to have the largest zonal variability.

The residual of the zonal momentum balance, or the terms that cannot be calculated directly, consists of shear stress and errors

\[
\text{Residual} = \frac{1}{\rho_0} \left( \frac{\partial \tau_{xx}}{\partial x} + \frac{\partial \tau_{xy}}{\partial y} + \frac{\partial \tau_{xz}}{\partial z} \right) + R + FD.
\]

Depending on which pressure gradient is used, the depth profile of the residual does not change much but the magnitude offset changes substantially (Figure 51).
Figure 51. Eight-day low-pass filtered vertical divergence of zonal turbulent stress calculated using pressure gradient from the pair of moorings at a) $0^\circ$, 156°E and $0^\circ$, 154°E, b) $0^\circ$, 160.5°E and $0^\circ$, 154°E, and c) $0^\circ$, 160.5°E and $0^\circ$, 156°E. The units are $10^{-7} \text{ m s}^{-2}$ and the contour interval is $4 \times 10^{-7} \text{ m s}^{-2}$. To the right of each contour is the corresponding mean depth profile within the standard deviation envelope.

Depth integrating the vertical shear stress divergence from the surface to depth $z$, using the surface wind stress boundary condition, produces

\[
\int_{-z}^{0} \frac{\partial \tau_{x_0}}{\partial z} \, dz = \tau_{x_0} \bigg|_{z} - \tau_{x_0} \bigg|_{-z}
\]

\[
= \rho_0 \int_{-z}^{0} \text{(Residual)} \, dz - \int_{-z}^{0} \left( \frac{\partial \tau_{xx}}{\partial \bar{x}} + \frac{\partial \tau_{xy}}{\partial \bar{y}} \right) \, dz - \rho_0 (R + F D),
\]

or
Figure 52. Mean zonal velocity and turbulent shear stress depth profile averages during 1992 over the time periods from (a and b, respectively) 10 May through 20 December 1992, (c and d) 10 May through 31 August 1992, and (e and f) 1 September through 20 December 1992. Zonal turbulent stress profile was calculated using pressure gradient from the pair of moorings at 0°, 156°E and 0°, 154°E (dashed line), 0°, 160.5°E and 0°, 154°E (solid line), and 0°, 160.5°E and 0°, 156°E (dotted line). Units are $10^{-5} m^2 s^{-2}$.
\[
\tau_{xz\mid_{-z}} = \tau_{xz\mid_{0}} - \rho_0 \int_{-z}^{0} (\text{Residual}) \, dz + \int_{-z}^{0} \left( \frac{\partial \tau_{xz}}{\partial x} + \frac{\partial \tau_{xy}}{\partial y} \right) \, dz + \rho_0 \left( R + FD \right). \tag{5.2}
\]

The first two terms on the right-hand-side of 5.2 can be estimated from the data, and the result provides a rough estimate of vertical turbulent shear stress as it varies with depth. The value of the estimate is dependent on the degree to which horizontal shear stress and errors are negligible (Figure 52b).

An important aspect of these results is that the vertical profile of the vertical shear stress estimate only crosses zero once and only for two of the profiles. This is in disagreement with the Reynolds analogy that vertical shear stress is proportional to the mean shear, \( \tau_{xz} \approx \rho_0 A \frac{\partial U}{\partial z} \), the basis for the KN assumption that the vertical shear stress is zero at the EUC core. Two of the pressure gradient profiles result in shear stress that does cross zero but not at the correct depths, and none of the profiles have the correct vertical structure. The Reynolds analogy would require shear stress to cross three times, corresponding to the three depths where the mean \( u \) profile has zero vertical shear (Figure 52 a and b). Because the stress profiles have large variation because of the different pressure gradient estimates used, the experiment’s ability to fully resolve pressure gradient is suspect. Horizontal shear stresses and/or, more likely, errors resulting from observational resolution are substantial.

Since the three-dimensional velocity field changes character in September, we make two additional mean shear stress estimates averaged over time periods each of different, but more uniform regimes. During the period from 10 May until 31 August, while the EUC is broad and strong and vertical velocity is upward (Figure 49 b and c), the shear stress still does not cross zero where the Reynolds analogy would suggest (Figure 52 c and d). During the period from 1 September until 20 December, characterized by the subsurface reversing jets, shear stress has less vertical structure when the zonal velocity has more vertical structure (Figure 52 e and f).

Since the effect of a pressure gradient magnitude offset (the essential difference between CMW and KN) is to rotate the shear stress profile about the wind stress boundary condition (through vertical integration), it would not help simply to adjust the barotropic pressure gradient. It is the baroclinic portion of the pressure gradient that fails to provide the depth-dependence required for the Reynolds analogy to work.

5.4 Conclusions

The residuals of the CMW and the KN zonal momentum balances suggest large vertical divergence in zonal shear stress. CMW choose to focus on the apparent lowest order balance of pressure, wind stress, and local acceleration. KN in contrast, note the relatively large turbulent shear stress and pose some arguments for why that might be so. For example, KN suggest that the negative eddy viscosity implied by the residual shear stress profile may occur through wave rectification processes. Based on the arguments
presented in Chapter 1, we suggest that large turbulent shear stress is an inherent consequence of the equatorial visco-inertial boundary layer.

The lowest order balances of the local zonal acceleration with the zonal wind stress over the surface layer and with the zonal pressure gradient force are supported (CMW). In late September and early October 1992 three easterly wind pulses coincide with local decelerations (Figure 49a, 50a). Zonal velocity from the surface to around 160 m decelerates from late August to early September coincident with a westward zonal pressure gradient force (Figure 50a, c, d, e). These balance tendencies are illustrated by CMW over a two-year time period, even when the zonal pressure gradient is even less well resolved. Because of poor vertical resolution of temperature and salinity data, limited availability of salinity data at depth, and the pressure estimate’s zonal spacing that do not match up with the velocity measurements zonal spacing, higher vertical mode and consequently higher order balances are less likely to be resolved.

The essential difference between the methods of CMW and KN is the treatment of the barotropic pressure gradient. CMW assume that the pressure gradient is small at 500 db and KN assume that the shear stress is zero at the EUC core. Whichever method provides the best barotropic pressure gradient is of secondary importance because the problematic vertical structure leading to inconsistent Reynolds analogy is dependent on the baroclinic portion of the pressure gradient.

A similar experiment in the central equatorial Pacific revealed a shear stress profile that crossed zero within 10 m of the EUC core without adjustment [Qiao and Weisberg, 1997]. In that region the vertical structure was considerably different, being a region under the influence of strong steady easterly trade winds. In the western equatorial Pacific the winds are not steady and there is the added complication of strong, reversing subsurface jets and asymmetries brought about by local topography. Consequently, it is not surprising that simplified models used to interpret coarsely sampled data do not produce easily interpreted shear stress profiles.

The apparent inconsistency between the estimated vertical shear stress profile and the Reynolds analogy is likely the result of under-sampling in time and space resulting in large finite difference errors. Figure 49 illustrates the temporal under-sampling, revealing two regimes with a transition in-between. The first is from May through August when vertical velocity is upward and the whole upper ocean is flowing eastward. The second is from September through December when there is heightened vertical and temporal variability and the existence of reversing subsurface jets. Figure 50 (c, d, and e) illustrates the spatial under-sampling primarily of the zonal pressure gradient, which is potentially the largest source of error in the balance. The baroclinic portion of the pressure gradient fails to produce the depth dependence required by the Reynolds analogy. While the lowest order results of the momentum balance are physically realistic, attempting to resolve the vertical stress divergence proves to be problematic. Future moored experiments must improve spatial and temporal resolution of temperature and salinity to resolve this issue.
Chapter 6

Summary and Conclusions

Observations of upper ocean variability and the temperature and momentum balances during February 1992 through April 1994 near 0, 156°E in the western equatorial Pacific are described. The inherent scales of the EMA observations represent variability with horizontal scales of 100-200 km, vertical scales of 10-200 m, and time-scales that range from weekly to interannual. These scales are large relative to the concurrent observations obtained south of the equator (~2°S) within the COARE IFA during the IOP. Observations were made about the equator in order to sample the visco-inertial boundary layer that exists within ~200 km of the equator. For this reason, the EMA data places the IOP/IFA within a larger spatial and temporal context. With this goal in mind, the western Pacific is described relative to the entire equatorial Pacific and the moderate El Nino conditions present during the EMA time period. Analyses are also performed that diagnose equatorial upwelling and temperature and momentum balances.

Chapter 1 sets the stage for the equatorial observations by reviewing equatorial dynamics via scaled momentum and vorticity equations. The analysis shows (consistent with historical development), how the equatorial region is a visco-inertial boundary layer with a horizontal scale of ~200 km [e.g. Fofonoff and Montgomery 1955; Arthur 1960; Charney 1960; Robinson 1960; Stommel 1960; Charney and Speigel 1971]. Frictional dissipation, within the equatorial boundary layer, must be elevated in order to smooth out a zonal velocity cusp that would otherwise form. Since the IFA was located at ~2°S, 156°E, it sampled a dynamical regime inherently different from that of the EMA. In addition, the western Pacific is where both equatorial and western boundary regimes coalesce. These ocean dynamics, coupled with tropical atmosphere dynamics, result in the largest warm water pool on Earth impacting global climate on time-scales from intra-seasonal to inerannual and longer.

The western equatorial Pacific exhibits episodic westerly wind bursts and far field forcing that results in interannual variability that is as large as the seasonal cycle (Chapter 2). In contrast, the eastern equatorial Pacific has steady easterly trade winds, which create steady large-scale gradients. Another difference between the eastern and western equatorial Pacific is that subducted extra-tropical water pathways toward the equator tend to reach the western Pacific only via low-latitude western boundary currents, whereas interior water pathways converge along the equator east of the dateline. As a result, the
EUC accelerates eastward of the dateline, while in the west, the EUC maintains a relatively constant speed and transport. Equatorial transects of zonal pressure gradient west of 170°W show depth reversals consistent with the reversing jet structure and zonal reversals suggestive of tropical instability waves.

Vertical velocity estimated from horizontal velocity divergence is analyzed in Chapter 3. Results suggest that deep seasonal upwelling within the thermocline in the boreal summer is driven by geostrophic divergence and is as large as upwelling in the eastern Pacific cold tongue. Ekman convergence from westerly winds is evidently less important. Since the thermocline is deeper in the west, however, the impact of upwelling on SST is small, relative to that in the eastern Pacific. This result is contrary to some GCMs that tend to produce downwelling in the western Pacific. Large vertical velocity is also suggestive of large vertical mixing consistent with the equatorial visco-inertial boundary layer discussed in Chapter 1.

The temperature balance analysis (Chapter 4) suggests a trend where large upwelling is associated with local cooling and large downward turbulent heat flux, whereas downwelling is associated with local warming and relatively small turbulent heat flux. While some of the results are inconclusive because of indeterminable potential errors and do not extend below the EUC, large vertical turbulent heat flux is consistent with thermostad maintenance. Mixing length arguments support the results and add plausibility to enhanced mixing on the equator. The scales discussed are larger than those associated with microstructure measurements made south of the equator (~°2S) within the IFA (and also for a short time on the equator) and cannot be compared directly.

In Chapter 5, the momentum balance analysis suggests that while the lowest order balances are achieved, there is an inconsistency between the estimated shear stress profile and a Reynolds analogy. The fundamental limitation to these momentum balance analyses appears to be the failure of the EMA to resolve the variations inherent in the baroclinic portion of the pressure gradient. Analogous limitations also occur in the temperature balance analysis. Nevertheless, the large residual of the momentum balance suggest that large turbulent shear stress exists, consistent with the need to dissipate the equatorial zonal current cusp that would otherwise form.

The requirement for enhanced mixing within the equatorial visco-inertial boundary layer is highlighted in this dissertation. Since definitive error estimation is not possible, scaling arguments based on dynamical equations are used to support some of the results. Therefore, plausible but not definitive conclusions are made about equatorial mixing. As discussed in Chapter 4, there is a disparity in experimental results concerning mixing on the equator and many questions remain to be answered about mixing in general [e.g. Muller and Garrett 2002]. New process experiments are necessary to address these issues. Advances in understanding can be achieved with current technology provided that experiments are conducted with a sufficient set of measurements. Provisions to estimate three-dimensional subsurface and surface
temperature, salinity, momentum, and turbulent fluxes are essential. The minimum requirements are subsurface temperature, salinity, and velocity measurements at all points of a flux divergence array and a full suite of surface meteorological measurements. Resolving the horizontal divergences of the temperature and salinity fluxes, and of the velocity is essential to improve our understanding of the upper ocean temperature, mass and momentum balances. Sampling complications may be reduced by making use of equatorial ocean symmetry properties away from complicated land boundaries, as existed in the COARE region. Moreover, for specific studies of the equatorial cold tongues, which provide the basis for the SST gradients of the Tropics that govern the coupling of the ocean and Atmosphere there, it is advisable to position the experiment array where the primary factors that influence divergence are largest. Since these factors are the zonal component of wind stress and the zonal pressure gradient loading to Ekman divergence and geostrophic convergence, respectively, the array should be positioned about the equator where these are largest. At the same time and location, other temperature, salinity, momentum, and turbulent flux experiments using different methods should also be deployed in order to broaden the scales and processes that are observed.

Clarifying the mixing requirements of the equatorial visco-inertial boundary layer is important for understanding and modeling global climate. Since tropical SST is a critical element of the coupled ocean-atmosphere system it is important for models to simulate it properly. Yet, existing coupled ocean-atmosphere models all tend to perform poorly for SST over the equatorial cold tongue regions [e.g. Mechoso et al. 1995; Latif et al. 2001; Davey et al. 2002]. Improving prediction depends on improving turbulence parameterizations, which depend on our understanding of the functional dependence of the mixing processes. This requires new process experiments, the need for which is highlighted in this dissertation.
References


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Appendices
Appendix A: Error Analysis

A.1. Advective time-scale

Peak values in the residual particularly below 150 m are unrealistically large and put the validity of the high frequency variability in question. Figure 53 shows that at most depths water transverses the array in ~10 days, suggesting that the low frequency processes discussed are likely resolved.

A.2. Surface extrapolation error

Because of side lobe reflection associated with upward looking ADCP’s, velocity measurements above 30 m are unavailable. Since the center mooring, located at 0°, 156°E, had a downward looking ADCP during the EMA-IOP, near-surface velocity data are available at that location (Table 4). The near-surface measurements from the center mooring are used to determine the type of extrapolation that works best and to estimate the magnitude of the extrapolation error. The error is calculated as the root mean square difference between the actual near-surface data at 0°, 156°E and the values interpolated from the 30 m bin. It was found that the minimum error was achieved by extrapolating the velocity values as a constant from the shallowest bin to the surface at the moorings where the near-surface values were unavailable. Extrapolation error is most important in terms that require horizontal gradients of velocity since data are used from one or more of the moorings where near-surface values are not available. Consequently, this error most substantially affects the vertical advection term because of the $w$ estimation. Figure 55 shows the combination of random instrument and surface extrapolation error, both of which are relatively small. Errors resulting from temporal and spatial resolution are potentially larger.

A.3. Random instrument error

The magnitude of random errors of averages is estimated using the standard deviation of the mean, which is the standard deviation divided by the number of degrees of freedom (see Figures 45 and 46). For the daily values only the random instrument error can be estimated. Temperature and velocity measurements have small random error, ~0.01-0.09° C and ~2 cm s$^{-1}$, respectively. Using propagation of error, Figures 54 shows the relative magnitude of the random instrument error associated with horizontal advection terms. Figure 55 shows the relative magnitude of the random instrument and surface velocity extrapolation error associated with vertical advection and the total derivative. We at least can neglect these errors since they are small relative to our estimate.
Figure 53. Progressive vector diagrams of horizontal velocity from the mooring located at 0°, 156°E and depths 20, 40, 60, 80, 100, 120, 140, 160, 180, 200, 220, and 240 m are plotted against latitude and longitude. The circles designate the mooring locations and plus signs mark the location every ten days starting at 16 September 1992 and ending 24 December 1992. The start of the progression is at the location 0°, 156° E.
A.4. Finite difference and other unaccounted for errors

Both finite difference and random geophysical errors cannot be definitively estimated and are left as unknowns. Finite difference errors result from spatial curvature of the differenced quantity, and the terms including $FD$ and $FI$ in equations 4.3 and 4.7 are included to acknowledge their existence. They are potentially the source of largest error, particularly because of the $w$ estimate, since it is critically dependent on horizontal divergence estimates. Random geophysical errors could be largely because of the length of the data sets relative to the time-scale of the signals we are attempting to resolve. Instead of speculating the magnitude of these errors, ancillary evidence has been cited to help justify the results.
Figure 54. The time-depth contours and associated random error of zonal (a and b) and meridional (c and d) advection, respectively. Units are $10^{-7}$ °C s$^{-1}$ with a contour interval of $2 \times 10^{-7}$ °C s$^{-1}$ throughout. The magnitude of the random error is very small since the temperature measurement error ranges from 0.01 to 0.09 °C. Finite difference error is not estimated and may be larger.
Figure 55. The time-depth contours and associated random instrument and surface velocity extrapolation error of vertical advection (a and b) and material derivative (c and d), respectively. Units are $10^{-6}$ °C s$^{-1}$ with a contour interval of $2 \times 10^{-6}$ °C s$^{-1}$ throughout. The magnitude of the random error is very small since the temperature measurement error at 0°, 156°E is 0.01°C. Error results primarily from surface velocity extrapolation. Finite difference error is not estimated and may be larger.
About the Author

Robert William Helber was born a U.S. citizen on 9 June 1967 in West Palm Beach, Florida. He earned a Bachelor of Science in Physics with a minor in Mathematics at the Virginia Commonwealth University (VCU), in 1992. Continuing at VCU and specializing in General Relativity, he earned a master's degree in physics in August 1994. That same month he entered the Marine Science Ph.D. program at USF as a graduate research assistant with Dr. Robert Weisberg as his major professor. During his time at USF he has participated in many research cruises as a USF scientific diver, assisting in numerous surface and subsurface mooring deployments. Robert has expertise in the analysis of time-series and large oceanographic data sets. Upon graduation he will begin a postdoctoral position under Dr. Weisberg to perform diagnostic studies of the equatorial Atlantic cold tongue with funding from NOAA CLIVAR Atlantic.