Wind-Driven Ageostrophic Transport in the North Equatorial Countercurrent of the Eastern Pacific at 95°W

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ABSTRACT

Observations of horizontal velocity from two shipboard acoustic Doppler current profilers (ADCPs), as well as wind, temperature, and salinity observations from a cruise during June–July 2001, are used to compute a simplified mean meridional momentum balance of the North Equatorial Countercurrent (NECC) at 95°W. The terms that are retained in the momentum balance and derived using the measurements are the Coriolis and pressure gradient forces, and the vertical divergence of the turbulent stress. All terms were vertically integrated over the surface turbulent layer. The $K$-profile parameterization (KPP) prescribed Richardson number ($Ri$) is used to determine the depth of the turbulent boundary layer $h$ at which the turbulent stress and its gradient vanish. At the time of the cruise, surface drifters and altimeter data show the flow structure of the NECC was complicated by the presence of tropical instability waves to the south and a strong Costa Rica Dome to the north. Nonetheless, a consistent, simplified momentum balance for the surface layer was achieved from the time mean of 19 days of repeat transects along 95°W with a 0.5° latitude resolution. The best agreement between the ageostrophic transport determined from the near-surface cruise measurements and the wind-derived Ekman transport was obtained for an $Ri$ of 0.23 ± 0.05. The corresponding $h$ ranges from ~55 m at 4°N to ~30 m within the NECC core (4.5°–6°N) and shoaling to just 15 m at 7°N. In general, the mean ageostrophic and Ekman transports decreased from south to north along the 95°W transect, although within the core of the NECC both transports were relatively strong and steady. This study underscores the importance of the southerly wind-driven eastward Ekman transport in the turbulent boundary layer before the NECC becomes fully developed later in the year through indirect forcing from the wind stress curl.

1. Introduction

In the eastern tropical Pacific Ocean, positive wind stress curl associated with the intertropical convergence zone (ITCZ) drives the eastward flowing North Equatorial Countercurrent (NECC; Sverdrup 1947). As such, the seasonal strength and path of the NECC is strongly tied to the seasonal migration of the upwelling curl of the ITCZ. The NECC is relatively weak in May, it then strengthens during July–October as the ITCZ moves north, and is strongest in November when the ITCZ is most northerly and the curl consequently increases the thermocline slope across the NECC. On average, the NECC shifts poleward as it flows from west to east across the tropical Pacific (Donguy and Meyers 1996; Johnson et al. 2002), bounded by westward flows to the north in the North Equatorial Current (NEC) and to the south by the South Equatorial Current (SEC). In the eastern tropical Pacific, the mean velocity field from over 27 years of surface-drogued drifters shows strong surface flow of ~30 cm s$^{-1}$ in the NECC between 4.5° and 7.5°N (Fig. 1a). The near-zonal path of the NECC is interrupted in the far eastern Pacific when the ITCZ interacts with the wind jets that blow through mountain

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gaps in the Central American cordillera (Kessler 2002, 2006). These mountain-gap winds produce strong zonal gradients of wind stress curl that can change the thermocline slope of the upper ocean, acting to weaken and strengthen the ridge–trough system that supports the NEC and the NECC, and set up the cyclonic flow around the Costa Rica Dome (≈9°N, 87°W in Fig. 1a).

The presence of tropical instability waves (TIWs) and their associated westward-propagating anticyclonic eddies (Kennan and Flament 2000) further complicates the flow of the NECC in the eastern tropical Pacific. The TIWs are characterized by cusp-shaped wave patterns of SST along the equatorial cold tongue and typically occur during July–November when the cold tongue is most developed. This pattern results in well-defined, large cross-equatorial fronts that are clearly visible in remotely sensed images of SST and ocean color. On average, the TIWs have periods of 20–40 days, wavelengths of 1000–2000 km, and westward phase speeds of ≈0.5 m s\(^{-1}\) (Qiao and Weisberg 1995; Contreras 2002). As such, the presence of the eddies and the meridional excursions of the TIWs directly act to perturb the flow within and along the southern border of the NECC. Both the TIWs, and their eddies, are thought to be generated by a barotropic instability associated with the strong velocity shear between the westward-flowing SEC and the eastward-flowing NECC and/or the Equatorial Undercurrent (EUC; Philander 1978; Contreras 2002; Willett et al. 2006). In this paper, we will show that flow within the NECC in the eastern Pacific Ocean was directly influenced by the presence of TIWs and a strong Costa Rica Dome during the time of a field campaign in June–July 2001.

Attempts in the past to observe a geostrophic balance of the NECC in the eastern Pacific have largely failed (Knauss 1961; Johnson et al. 1988), with hydrographic observations typically yielding geostrophic flows of \(O(10 \text{ cm s}^{-1})\) (Montgomery and Stroup 1962; Tsuchiya 1974). The observed discrepancy from a geostrophic balance may result from the inadequate spatial resolution of the hydrography, a dominantly ageostrophic NECC flow, or a combination of both. Even with a 50-km resolution synoptic survey along 110°W, Leetma and Molinari (1984) found that geostrophic velocities and the acoustic Doppler current profiler (ADCP) directly measured surface velocities differed by up to 20 cm s\(^{-1}\). The top-most bin resolved by their ADCP measurements was at ≈30-m depth, and they were possibly missing much of the surface-trapped ageostrophic or Ekman contributions.

![Fig. 1. Mean surface velocity (m s\(^{-1}\)) from 6-h drifting buoy data in the eastern tropical Pacific Ocean averaged over 1979–2006 for (a) the entire period and (b) June–July. The scale velocity vector is found in the upper-right-hand corner. Drifter velocities were interpolated by kriging to 6-hourly locations, and then vector averaged in 0.5° latitude by longitude bins. The location of the cruise track along 95°W is shown by the solid line.](image-url)
To date, simultaneous direct observations of the vertical structure of mixed layer currents and density gradients in the eastern tropical Pacific have been unavailable to investigate the near-surface momentum balance. Of critical importance to adequately resolve the momentum terms is that the in situ observations must resolve the fine scales of the property gradients in the uppermost layer. In this study we investigate a simplified vertically integrated meridional momentum balance of the near-surface turbulent boundary layer using direct observations from a 19-day repeat north–south transect at 95°W across the NECC in the eastern tropical Pacific during June–July 2001. In the simplified balance, the only momentum terms of importance retained are the Coriolis and pressure gradient forces, and the vertical divergence of the turbulent stress. A time-mean of each of these momentum balance terms is explicitly calculated along 95°W using the available current, density, and wind measurements. The velocity measurements were made from two ADCP systems that highly resolved the currents from the very near-surface layer at 6- to 250-m depth. The shallow velocity measurements provide fundamental and unprecedented means for adequately resolving the Ekman contribution to the near-surface momentum balance in the eastern Pacific.

A major challenge to closing the momentum balance in the ocean surface layer is the determination of the no-stress level, which is commonly considered to be the depth where boundary layer mixing resulting from turbulence is suppressed by dissipation and a stabilizing buoyancy gradient. The turbulence is driven primarily by the surface wind stress and buoyancy fluxes. Various algorithms have been proposed for a vertical mixing scheme that adequately describes the oceanic boundary layer physics and determines the boundary layer depth (see Large et al. 1994 for a complete discussion of these schemes). In fact, Large et al. (1994) developed their own parameterization of the oceanic boundary layer mixing, referred to as the “K-profile parameterization” (KPP). KPP directly predicts the no-stress level in the boundary layer to be the minimum depth at which a bulk Richardson number (Ri), a measure of the relative importance of stabilizing stratification to destabilizing shear, is equal to a critical value (Ri∗). Large et al. (1994) used various modeling and upper-ocean observational studies to suggest that the value of Ri∗ should lie between 0.25 and 0.50, with the range of values largely reflecting the chosen vertical resolution. These values are consistent with laboratory experiments that show shear stratified flow becomes unstable when the Ri decreases to a value below a Ri∗ of 0.25 (Turner 1973).

There have also been attempts to close the momentum budget without appealing to a mixing model. Stommel (1960) tried to apply the Ekman (1905) balance equations across the equator, integrating in both the meridional and vertical, using EUC as a no-stress boundary condition. In the western equatorial Pacific, Kennan and Nüüs (2003) also used the fact that the turbulent dissipation disappears at the EUC core to close the momentum flux budget on the equator. However, these approaches require prior knowledge of the depth of the no-stress level, such as in the EUC core.

In this paper, we take an alternative approach that indirectly diagnoses the no-stress level from the simplified momentum balance of the near-surface turbulence layer using the KPP prescribed Ri. Essentially, we ask the question: At what value of Ri∗ can we best achieve closure of the momentum balance within the near-surface turbulence layer? An independent measurement of the wind-forced Ekman transport using ship-based wind data is compared to the ageostrophic transport estimated from the cruise density and velocity data. We will show that agreement between these independent estimates is best achieved for Ri∗ of 0.2–0.3, corresponding to no-stress depth levels from ~50 at 4° to ~15 m at 7°N.

The outline of this paper is as follows: 1) The datasets and the simplified meridional momentum balance are described in section 2; 2) in section 3 we describe the large-scale horizontal circulation observed in the eastern tropical Pacific during the June–July 2001 cruise; 3) in section 4 we estimate the time-mean vertically averaged near-surface meridional momentum balance along 95°W using the observed shipboard data, and the latitudinal variation and balance of the momentum budget terms for each of the diagnosed no-stress levels that correspond to the KPP prescribed Ri are also discussed; and 4) the conclusions follow in section 5.

2. Data and methods
   a. Data

The primary hydrographic datasets used in this analysis were collected as part of a field campaign in the eastern Pacific Ocean aboard the R/V Roger Revelle from 13 June to 28 July 2001. Repeat meridional transects were made along 105°, 100°, 95°, 90°, and 85°W, crossing the anticipated axis of the NECC at each longitude between ~3° and ~8°N. Measurements along each transect included standard shipboard meteorological observations, the release of surface drifters, and shipboard underway ADCP and SeaSoar CTD transects. In this manuscript, we focus on the repeat transects along 95°W.

1) Shipboard ADCP

Velocity measurements of the surface layer were made using two underway shipboard ADCP systems. The R/V
Roger Revelle is normally equipped with a hull-mounted 150-kHz narrowband ADCP manufactured by RD Instruments. This unit was configured to sample at $1 \text{ Hz}$ and averaged every 300 s using an 8-m vertical bin and an 8-m pulse length. The shallowest bin of currents resolved by this 150-kHz ADCP was centered at 17 m. To capture the expected strong surface shear of the NECC above the working range of the 150-kHz ADCP, we also installed a 600-kHz Workhorse ADCP in a second well in the hull of the Revelle for the June–July 2001 cruise. This higher-frequency ADCP sampled 1-m bins beginning at a 6-m depth. Although the depth range of the 600-kHz ADCP was nominally 60 m, consistently good data returns were obtained to only a 45-m depth.

Over the depth range of 17–45 m for which both ADCPs sampled, the two datasets were compared by resampling the 600-kHz data, using a triangular filter, to the 8-m depths comparable to the sampling characteristics of the 150-kHz ADCP. The resulting data differed in the mean by less than 1 cm s\(^{-1}\). Possible ship hull effects on the shallowest measured flow were also investigated by comparing the measured currents from both ADCPs in the direction of, and normal to, the velocity of the ship. While no significant correlation was found between the observed velocity and the ship speed, there was an effect on the observed vertical shear. At depths above 17 m (corresponding to the shallowest bin of the 150-kHz ADCP), there was a weak correlation between negative shear (i.e., drag by the ship) and ship velocity (Fig. 2). Although the correlation was less than 0.08, the resulting shear bias exceeded $-4 \times 10^{-4}$ s\(^{-1}\) in the shallowest bin (6 m) of the 600-kHz ADCP for flow that is parallel to the ship heading (Fig. 2a). A small effect of only $-1 \times 10^{-4}$ s\(^{-1}\) was also found in the shallowest bin (17 m) of the 150-kHz ADCP. The shear bias was much smaller for flow that is normal to the ship heading (Fig. 2b). Since the ship heading was primarily north–south along 95°W, this suggests that the bias is less significant for the normal or cross-flow component (i.e., the NECC), which is the focus of our study. Nonetheless, the trend toward a ship drag on the flow is clear, and we wish to avoid any bias from ship heading. Hence, a subjective decision was made to correct only the five shallowest bins of the 600-kHz data, although the correction is small and accounts for only about 1% of the variance (the correlation between shear and ship speed was <0.1). Conversely, such a procedure is not common, so the deeper data, including the 150-kHz data, were not adjusted. The corrected shears were then integrated from the shallowest good bin (6 m) to the surface and then transformed back into the earth coordinate velocity components.
For both ADCP systems, absolute currents were calculated by subtracting the ship velocity over the ground, determined using global positioning system (GPS) satellite position fixes, from the relative current measured by the ADCP. Single-fix accuracy for the dual frequency P(Y) code (military or precision) GPS receiver on the Revelle is typically ~5 m (Chereskin et al. 1997). Heading corrections were made using an Ashtech GPS attitude sensing array (King and Cooper 1993). Errors in the GPS position, or because of instrumental errors in the absolute velocity computed from the ensemble-averaged data, are expected to be of $O(1 \text{ cm s}^{-1})$ (Chereskin et al. 1997; Chereskin and Harding 1993). Both the 150- and 600-kHz ADCP data were processed using the University of Hawaii data processing routines (CODAS3). Postprocess manual-editing of the data also excluded excessively noisy data, particularly at the bottom of the velocity profiles. Velocity artifacts from the ship turning at the end of each transect are known to introduce larger errors from the GPS position, and hence the ADCP velocities here were also eliminated. The ADCP data were then ensemble averaged over five minutes (equivalent to ~0.02° in space). Velocity profiles were formed from ~6- to 250-m depth by interpolating to 4-m depth bins using a piecewise spline.

For this study, we use the repeat ADCP sections along 95°W to determine the oceanic vertical shear for the parameterization of the Ri and in the momentum balance of the near-surface layer in the eastern tropical Pacific. In both these calculations, missing data in the surface layer were assumed to be equal to the temperature and salinity data at the shallowest depth (typically 10 m) of each CTD profile.

2) UNDERWAY SEAsoAR CTD

Underway measurements of temperature, conductivity (salinity), fluorometry, and pressure were obtained by towing a SeaSoar profiler consisting of a modified Chelsea SeaSoar body and hydraulic power pack. The SeaSoar was equipped with a Sea-Bird SBE911plus CTD with redundant temperature and conductivity sensors, a Wet Labs fluorometer, and attitude sensors for pitch, roll, propeller speed, and wing angle. The SeaSoar was typically flown along a sawtooth pattern, maintaining a uniform cycling period of approximately six minutes between 10 and 250 m. The data were collected at a frequency of 24 kHz. The SeaSoar data were processed following Rudnick and Martin (2002), who uses various procedures to minimize salinity spiking (e.g., Lueck and Picklo 1990). The processed temperature and salinity data were binned with a vertical resolution of 4 m and a temporal resolution of 10 minutes, which is equivalent to an alongtrack distance resolution of 0.05°.

The repeat SeaSoar CTD sections along 95°W (12 transects concurrent with the ADCP measurements) are used to determine the oceanic buoyancy in the parameterization of the Ri and the geostrophic contribution to the momentum balance of the near-surface layer in the eastern tropical Pacific. In both of these calculations, the missing data in the surface layer were assumed to be equal to the temperature and salinity data at the shallowest depth (typically 10 m) of each CTD profile.

3) WIND STRESS

During the cruise on the R/V Roger Revelle, underway ocean winds were measured using an R. M. Young anemometer mounted at 10 m above the reference water line on the science mast at the bow of the ship. The wind speed and direction are recorded every 30 s. The shipboard wind measurements along 95°W were converted to stress using the Coupled Ocean–Atmosphere Response Experiment, version 3.0 (COARE3.0) bulk formula (Fairall et al. 2003) and averaged in 0.5° latitude bands between 4° and 8°N.

These wind stress data will be used to compute the wind-driven Ekman transport in the simplified momentum balance. Errors in the wind stress data can be ascribed to bias in the shipboard bow-mounted anemometer measurements when the winds were from the stern; and errors in the bulk formulation of the wind stress as result of differences in air–sea conditions, fetch, wave field, etc. It is difficult to quantify any of these errors. Nonetheless, as we will show, the Ekman transport estimated using the shipboard winds is found to be consistent with ageostrophic transport estimated from the shipboard ADCP and CTD data. Ekman transports calculated from a “relative” wind referenced to the surface currents from the ADCP measurements (e.g., Fairall et al. 2003) were within the error bars of those estimated from the “unreferenced” shipboard wind observations. Hence, for brevity in the following, only the latter estimates of Ekman transport will be discussed.

4) ANCILLARY DATA

A number of other publicly available datasets are used primarily to provide context for the large-scale conditions in the eastern Pacific during the time of the June–July 2001 cruise described in section 3. Surface trajectories were obtained from more than 100 drifting
buoys deployed during the field campaign. The drifters were surface velocity program (SVP) buoys, which track the water motion at 15-m depth and typically have an accuracy of 1 cm s\(^{-1}\) in 10 m s\(^{-1}\) winds (Niiler et al. 1995; Niiler 2001). The raw Argos position data from the drifters were processed and interpolated by kriging to 6-hourly locations. Wind-slip corrections, generally fewer than 3 cm s\(^{-1}\), were also applied following Niiler et al. (1995). Estimates of sea surface height (SSH) anomalies were obtained from the French satellite processing facility for archiving, validation, and interpretation of Satellite Oceanographic data (AVISO) and are a merged product of all the available satellite data. We used the processed delayed mode (DT-REF) 7-day fields on a 1/3° Mercator grid. Ocean surface winds from a space and time blend of Quick Scatterometer (QuikSCAT)-KU2000 observations and the National Centers for Environmental Prediction (NCEP) analyses (Chin et al. 1998) were used to describe the larger-scale conditions during the field campaign. The 6-hourly surface wind data, on a 0.5° × 0.5° grid, were converted to stress using the COARE3.0 bulk formulas (Fairall et al. 2003).

b. A simplified meridional momentum budget

Our goal is to determine an appropriate momentum balance that describes the ageostrophic wind-driven transport in the near-surface turbulent boundary layer across the zonally flowing NECC at 95°W in the eastern tropical Pacific. Thus, in the momentum budget, our focus is on the meridional momentum equation

\[
\frac{dv}{dt} + fu = -\frac{\partial p}{\partial y} + \frac{\partial \tau^y}{\partial z} + \nu \nabla^2 v, \tag{1}
\]

where the acceleration (local acceleration plus the advective terms) and the Coriolis force are on the left-hand side and the pressure gradient force, the vertical divergence of the turbulent stress, and the viscosity terms are on the right-hand side. We will assume that acceleration and other higher-order terms, such as the turbulent diffusion of horizontal momentum and viscosity, to be negligible. This is a reasonable first approach, and in fact we computed the northward acceleration term from the SVP drifter data (Fig. 1) and found it to be much smaller than the Coriolis force retained in (1).

Thus, in our simplified linear momentum balance of the ageostrophic near-surface transport, the terms of importance in the meridional momentum Eq. (1) are the Coriolis force, the pressure gradient force, and the vertical divergence of the turbulent stress,

\[
f u = -\frac{\partial p}{\partial y} + \frac{\partial \tau^y}{\partial z}, \tag{2}
\]

where \(f\) is the Coriolis parameter, \(u\) is zonal velocity, \(p\) is pressure, and \(\tau^y\) is the meridional wind stress with a value of \(\tau^y_{\text{surf}}\) at the surface.

The ageostrophic transports are expected to be surface trapped. Hence, for our simplified meridional momentum balance, we vertically integrate Eq. (2) to the depth \((h)\) at which the turbulent stress vanishes, and apply the Boussinesq assumption,

\[
f \int_{-h}^{0} u \, dz = -\frac{1}{\rho_o} \int_{-h}^{0} \frac{\partial p}{\partial y} \, dz + \frac{\tau^y_{\text{surf}}}{\rho_o}, \tag{3}
\]

where \(\rho_o\) is the background water density. Below the near-surface turbulent boundary layer \((h)\), we assume a hydrostatic and geostrophic balance such that

\[
f u_G = f u_{\text{ADCP}}(-h) + g \int_{-h}^{z} \frac{\partial \sigma}{\partial y} \, dz',
\]

where \(\sigma = (\rho - \rho_o)/\rho_o\), \(\rho\) is the depth-dependent water density, and \(h)\) is a depth or layer where the flow is assumed to be in geostrophic balance, or analogously, that the ageostrophic flow has vanished. In this case we assume a layer depth for \(h)\) between 200 and 250 m, which is the maximum depth of our ADCP and hydrographic data, where the flow is assumed to be geostrophic. Thus, \(u_G\) is the geostrophic component of the zonal velocity estimated from the SeaSoar hydrographic data that is referenced to the ADCP observed zonal velocity \(u_{\text{ADCP}}\) averaged between 200- and 250-m depth. The difference between the ADCP velocity (which measures both the ageostrophic and geostrophic components) and the geostrophic velocity from the hydrographic data will be the ageostrophic velocity (see also Chereskin and Roemmich 1991). Therefore, our simplified vertically integrated meridional momentum Eq. (2) becomes

\[
\int_{-h}^{0} [u_{\text{ADCP}} - u_{\text{ADCP}}(-h)] \, dz
- \int_{-h}^{0} [u_G - u_G(-h)] \, dz = \frac{\tau^y_{\text{surf}}}{\rho_o f}, \tag{4}
\]

where the left-hand side of (4) represents the ageostrophic transport within the turbulent boundary surface layer and the right-hand side is the locally wind-driven transport within the boundary layer, which is the classical Ekman transport.

To close the near-surface linear momentum balance of Eq. (4), we need to determine the depth at which the turbulent stress vanishes. The depth of the near-surface
turbulent boundary layer is dependent on the surface forcing and its effect on the ocean buoyancy and shear profiles. The KPP prescribed 

$$\text{Ri}(h, y, t) \approx \frac{1}{h} \int_{-h}^{0} \left( \frac{\partial \sigma}{\partial z} \right)^{2} + \left( \frac{\partial v}{\partial z} \right)^{2} \, dz - \frac{1}{h} \int_{-h}^{0} \frac{N^2 \, dz}{\int_{-h}^{0} S^2 \, dz}.$$  

(5)

parameterizes the relative importance of the stabilizing stratification (defined by the buoyancy frequency $N$) to the destabilizing velocity shear ($S$). As suggested by Large et al. (1994), we compute the depth-integrated values for both the buoyancy frequency $N$ and the velocity shear $S$ over the turbulent surface layer. In this way, the determined Richardson number is not dependent on, and less sensitive to, the vertical resolution of the data. Thus, Eq. (5) offers a potential scheme for determining the boundary layer depth $h$. The depth is equal to the shallowest depth at which the Ri defined by Eq. (5) equals the $\text{Ri}_\text{c}$ (Large et al. 1994). Below $h$, the flow is essentially geostrophic. Large et al. (1994) suggest that in the near-surface layer, the turbulent stress should vanish for a critical Ri between 0.2 and 0.5. Our approach then is to determine the boundary layer $h$ from Eq. (5) using the shipboard buoyancy and velocity profiles for a similar range of specified Ri. An alternative approach would be to compute $h$, at which the ageostrophic transport is equal to the Ekman transport from Eq. (4), and then compute the Ri from Eq. (5) for that water column. In practice, we found there is little difference between the two approaches and hence, for brevity, we only present the first methodology.

3. Conditions in the tropical eastern Pacific Ocean, June–July 2001

In general, during June–July the ITCZ shifts northward. During June–July 2001, the meridional wind stress between 3$^\circ$ and 9$^\circ$N along 95$^\circ$W shows a wide latitudinal variation in the strength of the winds (Fig. 3). For the most part, the winds were primarily southerly and stronger at lower latitudes, becoming more northerly at northern latitudes. This is consistent with Mitchell et al. (1989), who noted that winds in the eastern tropical Pacific are characterized by a strong southerly component that turns eastward after crossing the equator and converges with the northeast trade winds near 10$^\circ$N in the ITCZ. During the period of the cruise along 95$^\circ$W (black box in Fig. 3), there is strong variability in the meridional wind stress. The northerlies extended as far south as 5.5$^\circ$N during 7–8 July 2001, only to reverse from 9 to 12 July 2001 to the strongest southerlies observed over the whole time series.
The historical drifter data for the long-term average June–July period show an eastward flow between 4°–8°N and a westward flow between the equator and 4°N (Fig. 1b). These zonal flows appear to make up the northern and southern limbs of an anticyclonic gyre centered at 4°N, 95°W, which is not strongly apparent in the annual surface velocity (Fig. 1a). The June–July period also coincides with the annual western expansion and intensification of the Costa Rica Dome (Kessler 2006), which is centered at 9°N, 90°W (Fig. 1b).

The drifter data and SSH anomalies during the field campaign show the expected strong zonal flows but equally strong meridional flows (Fig. 4). The low sea level in the Costa Rica Dome is well developed and extends from 105°–90°W to 7°–10°N. In fact the strongest eastward flows in the drifters are clearly associated with the southern limb of this cyclonic circulation (Fig. 4), marked by the negative sea level gradient between 5° and 8°N during the field campaign (Fig. 5). South of 5°N, the sea level and its gradient (Fig. 5) are broken up by a series of mesoscale features, and no coherent zonal flow can be deduced (Fig. 4). Drifters deployed along 95°W, and earlier along 105°W, are entrained equatorward, moving anticyclonically about the anomalous SSH highs (Fig. 4). These anticyclonic motions are clearly associated with TIW eddies, which were also evident in satellite SST images, and SeaSoar fluorometer and temperature sections taken along 95°W during the cruise (not shown). The implied circulation during the cruise derived from the sea level maps and drifter trajectories (Figs. 4, 5) confirm the historical drifter patterns during June–July (Fig. 1b): east of 95°W the eastward extension of the anticyclonic circulation is closer to the equator, whereas west of 95°W the flow appears to be broken up by a number of mesoscale TIW features. The only zonally contiguous flow structure in the drifter data that can be labeled the NECC is the strong eastward flow associated with the southern limb of the Costa Rica Dome.

The vertical section of the measured (ADCP) mean zonal current at 95°W shows strong eastward flow of more than 60 cm s⁻¹ within the upper 50 m between 5° and 7.3°N (Fig. 6a). Below this upper layer, the eastward velocities steadily decrease with depth and narrow in latitudinal range. At 200-m depth, the eastward flow is ~15 cm s⁻¹ and confined between 5.4° and 6.1°N. Figure 6a suggests that most of the eastward flow that might be attributed to the NECC has been resolved by the ADCP surveys in both depth and latitude range. Mean densities from the SeaSoar CTD surveys at 95°W were used for calculating the geostrophic component of the flow (Fig. 6b), referenced to the averaged ADCP zonal velocity between 200- and 250-m depth. Unlike the measured ADCP zonal velocity, between 5.6° and 6.5°N the zonal geostrophic velocity has a subsurface maximum of ~55 cm s⁻¹ at ~30–50-m depth. Deeper in the water column, the pattern and strength of the geostrophic flow is similar to the ADCP measured total flow.
Most ageostrophic flow [the difference between the ADCP total flow (Fig. 6a) and the geostrophic flow derived from the SeaSoar data (Fig. 6b)] at 95°W occurs within the upper 50 m (Fig. 6c). The ageostrophic flow has virtually vanished below 100-m depth. The strongest eastward (30 cm s$^{-1}$) ageostrophic flow is found between 4.5° and 5.2°N, although eastward ageostrophic shear is present in the upper 30 m as far north as 6°N. Detection of this strong very shallow flow is only possible because of the highly resolved velocity measurements in this layer by the high-frequency ADCP that measured the very near-surface flows and with higher vertical resolution than afforded by the low-frequency ADCP. Clearly, much of the shallow ageostrophic flow would not have been resolved if, as is often the case for shipboard ADCP measurements, only the low frequency, lower vertical resolution velocity data were available. Relatively strong westward ageostrophic flow occurs at the southern end of the transect equatorward of 4.5°N. This latitude coincides with the absence of the sea level gradient as measured by the altimeter (Fig. 5) and was also strongly influenced by the presence of the TIWs during the time of the cruise (Fig. 4).

4. Time mean, simplified momentum balance along 95°W

To evaluate the Richardson number [Eq. (5)] at the time of the cruise along 95°W, the buoyancy frequency is calculated from the mean density profiles of the SeaSoar CTD data (Fig. 7a), and the mean vertical shear is calculated from the horizontal velocity as measured by the ADCP (Fig. 7b). Figure 7c shows the shallowest depth at which the $R_{i}$, defined by Eq. (5), equals the $R_{i, c}$ of 0.2–0.7 (at 0.1 intervals). This $h$, therefore, coincides with the depth of the turbulent boundary layer for each of these choices of $R_{i, c}$.

For $R_{i, c}$ of 0.2 and 0.3, the patterns are similar, with the boundary layer depths shoaling northward from 4° to 6.5°N (Fig. 7c). This pattern is related to the elevated values in the buoyancy frequency as the thermocline shoals northward along the 95°W transect (Fig. 7a). In the shear profiles (Fig. 7b), the shoaling pattern is still evident, but it is not as gradual or distinct. This may imply that there is relatively little momentum transport below this relatively shallow layer. In the tropics and equatorial region, small-scale nighttime convective turbulence and internal waves are known to penetrate into the stratified top of the main thermocline (Dillon et al. 1989; Wijesekera and Dillon 1991). However, it is not clear whether these processes carry any horizontal momentum deeper into the ocean (Moum et al. 1992; Johnson and Luther 1994). Boundary layer depths of ~55 m for $R_{i, c} = 0.2$ occur south of 5°N because of the influence of the stronger shear at this latitude and depth range (Fig. 7b). Depths of ~30–40 m for $R_{i, c} = 0.2$ are found in the core of the NECC between 5.2 and 6.5°N (Fig. 6). Here, the buoyancy frequency is relatively low,

![Fig. 5. Time series of the SSH anomaly (cm; colors) and its meridional gradient (10$^{-5}$ cm; contours) during 23 Jun–28 Jul 2001 at 95°W. The green box encloses the 19-day period and the latitudinal track (green line) of the cruise observations undertaken at 95°W.](image)
and thus the depth of the boundary layer is largely determined by the depth variation of the shear profile. North of 6.5°N, transect maxima in both buoyancy (Fig. 7a) and shear (Fig. 7b) are found and the boundary layer depth for \( \text{Ri}_c = 0.2 \) and 0.3 outcrop at the surface north of 7°N. For increasing values of \( \text{Ri}_c \) (Fig. 7c), the depth of the boundary layer increases by more than a third (\( \sim 20 \) m) because of the influence of larger buoyancy and smaller shear (Figs. 7a,b) over these depth ranges. South of 5°N, an increase in shear coincides with elevated values of the buoyancy frequency at depth. These high values of the buoyancy frequency at approximately 70–80-m depth are likely related to the cold, relatively deep surface layers associated with the TIWs. The TIWs result in the abrupt change in the boundary layer depth near 5°N for all \( \text{Ri}_c \). Between 5.5° and 6.5°N for higher values of \( \text{Ri}_c \), the boundary layer depth is determined by the buoyancy profile.

The Ekman transport

The simplified momentum balance within the surface layer can now be diagnosed following Eq. (4), from a time mean of the meridional density and velocity transects collected along 95°W during the cruise and the turbulent boundary layer depths defined by the prescribed KPP Richardson numbers. The zonal ageostrophic transport in the near-surface turbulent layer from the left-hand side of Eq. (4) is obtained by differencing the geostrophic transport from the total transport (Fig. 6c). For the depth integration limit, we used the local value of \( h \) as determined for the range of KPP \( \text{Ri}_c \).
shown in Fig. 7c. The mean ageostrophic transport per unit distance \( (m^2 \text{s}^{-1}) \) estimated from the alongtrack cruise data is shown in Fig. 8 for each 0.5° latitude band along 95°W. The Ekman transport on the right-hand side of Eq. (4) is estimated from the meridional wind stress from the shipboard winds (Fig. 8). For clarity, in Fig. 8 we only show results of the ageostrophic calculation for the turbulent boundary layer depths determined for \( \text{Ri}_c \) of 0.2 and 0.3, as these give the most comparable values to wind-estimated Ekman transports. The 95% confidence intervals for both the ageostrophic and Ekman calculations (Fig. 8) were estimated using a standard Student’s \( t \) test, with the degrees of freedom for each 0.5° latitude band given in Table 1.

The Ekman transport estimated from the shipboard wind stress decreases with increasing latitude from \( \sim 2.6 \) at 4° to \( \sim 0.4 \) \( m^2 \text{s}^{-1} \) at 7.5°N. In the core of the NECC, from 4.5° to 6°N, the Ekman transport is fairly steady at \( \sim 2.0 \) \( m^2 \text{s}^{-1} \) and largely falls between the two ageostrophic transport estimates (Fig. 8). In this latitude band, the winds along 95°W are primarily southerly (Fig. 3) and drive eastward Ekman mass transport in the surface boundary layer during the June–July 2001 field campaign (Fig. 8). North of 6°N, the winds are weaker and more variable (Fig. 3), the turbulent boundary layer depth is very shallow (Fig. 7c), and the Ekman transport reduced (Fig. 8). The flow here is primarily geostrophic (Fig. 6b) and related to the relatively strong flow in the southern limb of the Costa Rica Dome. South of 4.5°N, the southerly winds (Fig. 3) would also drive eastward wind-driven flow, although the ageostrophic flow derived from the cruise data in this region is westward (Fig. 6c). This may indicate cyclostrophic effects associated with the westward propagating TIWs and their anticyclonic eddies (Fig. 4). The strong variability of the ageostrophic transport at 4°N reflects the transient presence of the TIWs. For higher critical Richardson numbers at 4°N, the deeper \( h \) (Fig. 7c) results in transect maximum transports ranging from 6.8 \( m^2 \text{s}^{-1} \) for \( \text{Ri}_c = 0.4 \) to 8.6 \( m^2 \text{s}^{-1} \) for \( \text{Ri}_c = 0.6 \) and also exhibits strong variability (not shown).

It is worth pointing out that the Ekman transport derived from the wind field has no information on the vertical structure of the flow and its relation to the mixed layer. Thus, we could have expected significant differences between the depth-integrated ageostrophic transport and the Ekman transport estimates, even when no particular adjustment of the critical \( \text{Ri} \) is made. In the main, the Ekman transport estimated from the shipboard wind stress shows quite good agreement with the ageostrophic transport estimated for \( \text{Ri}_c \) of 0.2 and 0.3 (Fig. 8). A mean value of \( \text{Ri}_c = 0.23 \pm 0.05 \) was determined to give the best agreement by minimizing the

![Fig. 7. Variation of (a) buoyancy frequency (s^{-1}) and (b) vertical shear (s^{-1}) with depth and latitude along 95°W. Notice the different scale between (a) and (b). (c) The depth below which the KPP prescribed \( \text{Ri} \) of 0.2–0.7 are equaled or exceeded vs latitude along 95°W. A 3-point moving average filter has been applied.](image-url)
root-mean-square differences between the ageostrophic and Ekman transport estimates at each latitude (Fig. 8, the error bar is the standard deviation of this estimate). Within the core of the NECC, the ageostrophic transport estimate calculated using the mean value of $R_i = 0.23$, agrees within the error bars of the Ekman transport (Fig. 8). This good agreement suggests that, at least to first approximation, the simplified momentum balance described by Eq. (2) provides a relatively good model describing the flow in the NECC during the 19-day cruise period at 95°W. The differences could be attributed to the neglect of other terms in the simplified balance. Processes such as internal wave variability and cyclostrophic effects associated with the TIWs south of 4.5°S are not accounted for in our simplified model.

5. Conclusions

In this paper, we used a suite of observations collected during a field campaign in June–July 2001 of the eastward-flowing NECC in the eastern tropical Pacific to investigate and quantify the time-mean, vertically averaged, near-surface momentum balance subject to simplifications and classical assumptions. The presence of TIWs, and their anticyclonic eddies, along with the strongly developed westward-elongated Costa Rica Dome, clearly complicated the surface circulation pattern over relatively short timescales in relation to our repeat survey periods along 95°W. The drifter trajectories show strong eastward surface flow associated with the southern limb of the cyclonic gyre associated with the Costa Rica Dome. In the annual mean, Kessler (2002) suggested that the flow around the Costa Rica Dome is not just an eastward extension of the NECC but instead is fed largely from

<table>
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<th>Latitude (°N)</th>
<th>4.0°</th>
<th>4.5°</th>
<th>5.0°</th>
<th>5.5°</th>
<th>6.0°</th>
<th>6.5°</th>
<th>7.0°</th>
<th>7.5°</th>
<th>8.0°</th>
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<td>55</td>
<td>55</td>
<td>98</td>
<td>105</td>
<td>99</td>
<td>63</td>
<td>51</td>
<td>22</td>
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the north. However, the drifter velocities and trajectories clearly suggest that, at least during this June–July time frame, the flow on the southern side of the Costa Rica Dome is indeed fed from the NECC.

Despite the complex circulation patterns evident in individual transects, by averaging over the repeat sections taken along 95°W over the 19-day period, a NECC can be identified that is relatively strong in the upper 50 m between 4.5° and 6°N, decreases in speed, and narrows in latitude with depth (Fig. 6a). Subtracting the geostrophic velocity estimate from the ADCP velocities, we find that most of the ageostrophic flow and vertical shear is confined to the upper 50 m between 4.5° and 6°N (Fig. 6c). Detecting this surface wind-driven flow was only possible because of the repeat high vertical resolution velocity observations obtained in the rarely measured near-surface layer. These direct velocity measurements in the upper 50 m were crucial in allowing us to determine the vertical distribution of the ageostrophic zonal currents used to estimate the vertically integrated ageostrophic transport with confidence. The ageostrophic transport was largely explained by the surface wind-driven flow that is trapped within the turbulent boundary layer. Our results confirm the importance of resolving the finescale gradients of the near-surface layer to successfully separate the wind-driven from the geostrophic components of the flow.

The depth of the boundary layer where the turbulent stress vanishes was determined using the KPP prescribed Richardson number—the ratio of buoyancy production to the destabilizing shear [Eq. (5)]. The corresponding depth of the turbulent boundary layer ranges from ~55 at 4°N to ~30 m within the NECC core and shoaling to the near-surface at 7°N (Fig. 7c). Again, resolving these relatively shallow boundary layer depths was only possible because of the shallower velocity measurements from the high-frequency ADCP. For the simplified surface momentum balance, we found the best agreement of the ageostrophic flow with an independent estimate of the Ekman transport from the shipboard winds was obtained for $R_i < 0.23 \pm 0.05$. This is in relatively good agreement with other modeling, and experimental and upper-ocean observational studies that suggest $R_i$, should lie between 0.25 and 0.5 (Large et al. 1994). The mean Ekman transport decreased from south to north along the 95°W transect during the June–July 2001 cruise; however, within the core of the NECC, the Ekman transport was relatively strong and steady. This study underscores the importance of the southerly wind-driven eastward Ekman transport in the relatively shallow turbulent surface layer before the NECC becomes fully established later in the year by the wind stress curl.

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REFERENCES


