Tropical Pacific near surface currents estimated from altimeter, wind and drifter data

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Abstract

Tropical surface currents are estimated from satellite-derived surface topography and wind stress using a physically based statistical model calibrated by 15 m drogue drifters. The model assumes a surface layer dominated by steady geostrophic and Ekman dynamics. Geostrophy varies smoothly from a beta-plane formulation at the equator to an f-plane formulation in mid latitude, with the transition occurring at about 2-3º latitude. The transition is treated with a Gaussian weight function having a meridional decay scale that is found to be approximately the Rossby radius (~2.2º latitude). The two-parameter Ekman model represents drifter motion relative to wind stress, with down wind flow along the equator and turning with latitude. Velocities computed from satellite data are evaluated statistically against drifter velocities and equatorial current moorings. Examples of the geostrophic and Ekman flow fields in the western Pacific during a westerly wind burst in late December 1992 depict a strong eastward flow and equatorial convergence. A comparison between December 1996 and June 1997 illustrates the basin-wide reversal of equatorial surface flow during the onset of the 1997 El Niño.

1. Introduction

The objective of this study is to estimate the tropical Pacific surface circulation using satellite-derived sea level and wind stress fields and evaluate the results. Our present understanding of the mean surface circulation and climatology is derived from observations of ship drifts, drogued drifting buoy velocities and a small number of current meter moorings (Reverdin et al., 1994; Frankignoul et al, 1996). The scarcity of these data nevertheless requires broad scale averaging and permits only coarse synoptic mapping to address intraseasonal to interannual circulation patterns. Satellite observations provide much higher spatio-temporal views of the ocean on ~100-200 km spatial scales and on time scales of ~2-5 d for scatterometer winds (Liu et al., 1998) and ~10-30 d for altimetric sea level (Fu et al., 1994). Renditions of high resolution velocity fields can be produced through numerical OGCM simulations that incorporate satellite and *in situ* observations as boundary conditions and/or through data assimilation. Several investigating groups are working on such problems, where the techniques are highly specialized and computationally intensive. Our approach is to apply relatively simple lowest order dynamics to estimate velocity directly from satellite-derived variables. In the process uncover certain physical insights into these dynamics. Finally, we describe detailed circulation features that such calculations allow at the spatial and temporal resolution of satellite data.

For practical purposes, we define surface velocity as the motion of a standard WOCE-TOGA 15 m drogue drifter (described below), and then design the analysis to provide a best fit to the available *in situ* drifter data (Fig. 1). Geostrophic and Ekman components are assumed to account for the lowest order dynamics of the surface velocity, and can be obtained independently from surface height and wind stress data. It is well understood that the standard f-plane geostrophic balance, where the velocity is proportional to the height gradient divided by the earth's rotation parameter f, is the lowest order balance for quasi-steady circulation at higher latitudes (Pedlosky, 1979). Geostrophy requires special attention near the equator where $f \rightarrow 0$. The flow at the equator becomes independent of the sea level gradient, which may exist on the equator from ageostrophic balances (Joyce, 1988). Many authors have shown that in close proximity to the equator, a β-plane geostrophic approximation involving the second derivative of surface height provides excellent agreement with observed velocities in the equatorial undercurrent (Lukas and Firing, 1984; Picaut et al., 1989;). The practical problem in analyzing a gridded height gradient field from altimetry, as with a set of dynamic height differences from a cross-equator hydrographic transect, is deciding how to transition the velocity calculation from one geostrophic approximation to the other. Picaut and Tournier (1991) applied a small pressure

correction term to eliminate the meridional slope at the equator, with a meridional trapping scale similar to one we obtain below. Other investigators have used the f-plane calculation to within about 1˚ latitude and then averaged these at the grid points immediately adjacent to the equator north and south (Kessler and Taft, 1987; Johnson et al., 1988). This gives the equivalent to β-plane estimate at the equator. Others applied the β-plane exclusively within a few degrees latitude (Moum et al., 1987; Cornuelle et al., 1993). One must still consider the how valid either approach is at a given latitude. Clearly, there must be some transition between valid and not-valid, and this transition is given shape and form here. It is addressed with a blended analysis approach using weight functions (Appendix A) which are scaled with a regression analysis between height gradients and drifter velocities. The results will be applicable to near-equatorial geostrophic calculations in general, such as with densely spaced dynamic height sections.

The Ekman motion of the drifters is considered with a two-parameter model of the purely wind-driven response. The parameters are a depth scale and a drag coefficient. We then derive the surface velocity as the sum of geostrophic and Ekman components. Results computed from satellite data are evaluated statistically against drifter velocities and equatorial current moorings. Surface velocity maps are presented to illustrate the applications of these data. The first set depicts the surface field during the strong westerly wind burst in the western Pacific warm pool observed during TOGA COARE in late December 1992. The other contrasts the basin-wide surface flow during the recent ENSO cycle between November 1996 cold phase conditions and June 1997 warm phase conditions.

2. Data and Processing

2.1 Surface height field(s):

TOPEX/POSEIDON altimeter sea level anomalies from along-track data (Fu et al., 1994) were interpolated by objective analysis to a 1˚x1˚ grid in the domain 25˚N—25˚S and 90˚E—290˚E(70˚W), centered on the half-degree, with a temporal sample interval of 36/yr (~10 d). Additional details of the interpolation scheme are provided in Appendix B. The time period covers October 1992 - September 1998. The 6-year mean was subtracted to remove any residual marine geoid errors and replaced with a mean dynamic height surface. This was obtained from mean climatological gridded 1˚x1˚ degree temperature and salinity fields of Levitus et al. (1994a&b), which were used to compute

surface dynamic height relative to 1000 dbar (1 dbar = 10^4 Pa) using standard methods. The topography gradient was then computed for each ~10-day map for subsequent analyses and computation of geostrophic current.

2.2 Wind stress fields

We chose the variational analysis of Special Sensor Microwave Imager (SSMI) winds (Atlas et al, 1996) as a proxy for satellite scatterometer winds. These data provided a gridded field suitable for the initial development and testing of our approach. In comparison studies using various model and satellite wind fields to drive a tropical Pacific OGCM, these winds generated the most accurate surface height field when compared to TOPEX/Poseidon data (E. Hackert, personal communication), and these data were available continuously over the 6-year period of this altimeter analysis. 5-day average surface winds were converted to wind stress with the Large and Pond (1981) drag formula and filtered with a 20-day low pass filter and interpolated to the same time-space grid as the surface height analyses.

2.3 Drifter velocities

The standard WOCE-TOGA drifters consist of a 1 m diameter “holey sock” drogue suspended from 10 to 20 m below the surface (average depth 15 m). Extensive calibration tests demonstrate that this drogue design has minimal slippage relative to the current (Niiler et al., 1995). Comparisons with geostrophic velocities were shown by Yu et al. (1996). Bi (1995) estimated drifter decorrelation time scales in the tropical Pacific on the order of 5 to 20 days for zonal and meridional components respectively. We computed drifter velocities from trajectory positions during the period October 1992 - December 1994 (Fig. 1) by computing displacements over 5 d intervals as representative of 5 d averages. Only those trajectory histories when the buoy and drogues remained attached were used. The latitude domain 25˚N to 25˚S was divided into 1 degree latitude bands centered on the half-degree. Drifter velocities were sorted by position into these latitude bins regardless of longitude and time. The surface height gradient and wind stress fields were interpolated linearly to match the time and location for each drifter velocity within the respective latitude bins. These sets formed the basis of the regression analyses carried out within each latitude interval as described in Sect. 3 below.

2.4 Current moorings

Direct observations of current from equatorial current moorings at 10 m depth were obtained from NOAA/PMEL (0, 110˚W; 0, 140˚W; 0, 165˚E)and from R. Weisberg, USF (0, 170˚W). Daily mean currents were calculated, passed through a 21-day running mean filter to provide some smoothing of instability wave variability (Halpern et al., 1988). A 6-month-long current record during TOGA COARE (Webster and Lukas, 1992) from 19 m at 2˚S, 156˚E was obtained from R. Weller (WHOI) and smoothed similarly. All current data were interpolated to the height analysis 36/year time base for direct comparison with the satellite-derived currents discussed in Sect.4.

3. Formulation

3.1 Equations of motion

As described above, the surface height and wind stress fields are to be used to estimate the steady geostrophic and Ekman surface velocity components representing the motion of standard 15-m drogue drifters. Errors will arise from simple steady-state dynamics that omit the terms for local acceleration, momentum advection and lateral viscosity. Local acceleration is likely to be most influential near the equator in the presence of wind-driven jets (Yoshida, 1959) and equatorially trapped waves. Influence of momentum advection and lateral viscosity on the momentum budget has been documented by Johnson and Luther (1994).

Drifter velocity is assumed to represent the average motion in a surface layer of scaling thickness h. The linear steady momentum balance is expressed using conventional notation as:

$$-fhv = -gh\frac{\partial\zeta}{\partial x} + \frac{\tau^x}{\rho} - ru_e \qquad (1)$$

$$fhu = -gh\frac{\partial\zeta}{\partial y} + \frac{\tau^y}{\rho} - rv_e \qquad (2)$$

where r is a linear drag coefficient that represents the vertical viscosity terms as a body force on the Ekman components u_e and v_e. Multiplying (2) by i and adding (1) gives a complex vector form:

$$ifhU + rU_e = -ghZ + \tau \qquad (3)$$

where $\boldsymbol{U}=u+iv$, $\boldsymbol{U}_e=u_e+iv_e$, $\boldsymbol{Z}=\partial\zeta/\partial x + i\partial\zeta/\partial y$ and $\tau=(\tau^x+i\tau^y)/\rho$ is the kinematic stress computed with ρ=1025 kg m^{-3}. The flow is described in terms of **Z** and τ by expressing **U** as the sum of separate geostrophic ($\mathbf{U}_g$) and Ekman ($\mathbf{U}_e$) components:

$$ifU_g = -gZ \tag{4}$$

$$(ifh+r)U_e = \tau \tag{5}$$

3.2 Geostrophic velocity near the Equator: $\mathbf{U}_b$ denotes the geostrophic velocity that can be computed with a β-plane approximation (f=βy) using the derivative of (4)

$$\beta U_b + \beta y\frac{\partial}{\partial y}U_b = ig\frac{\partial Z}{\partial y} \tag{6}$$

The second term is generally neglected very near the equator (y≈0). This term is retained in the method presented here (Appendix A) which provides a fitted solution to the differential equation (6).

At some distance away from the equator the conventional f-plane geostrophic calculation is used, denoted as $\mathbf{U}_f$

$$U_f = \frac{ig}{f}Z \tag{7}$$

A smooth transition in the computed geostrophic current is achieved between the equator and higher latitudes by applying a pair of weight functions, W_b and W_f, and expressing U_g as the weighted sum:

$$U_g = W_b U_b + W_f U_f \tag{8}$$

where W_b=1 and W_f=0 at the equator, and $W_b \rightarrow 0$ and $W_f \rightarrow 1$ as latitude increases. The derivations of $\mathbf{U}_b$ and weights are described in Appendix A. W_b and W_f are shown to be approximated by Gaussian functions with the meridional decay scale as a free parameter. The scale length is found by analyzing the drifter velocities relative to the local topography gradient and wind stress. The following sections describe these analyses and show how the findings relate to the weight function scale analysis given in the Appendix.

3.3 Analysis of geostrophic drifter motion

In the next two sections we evaluate geostrophic and Ekman drifter motion by means of a complex multiple linear regression of the equation

$$U_{drifter} \approx a_1 Z + a_2 \tau \qquad (9)$$

The empirical regression coefficients a_1 and a_2 were computed in each 1° latitude band with the interpolated data described in Sect. 2.3. The complex linear regression produces real and imaginary parts for each coefficient, with the real parts representing along gradient and along wind motion, and the imaginary parts representing cross gradient and cross wind motion, respectively.

Before describing these results, however, the ensemble drifter velocities examined at latitudes greater that 5°, where the f-plane geostrophic approximation is expected to prevail. Fig. 2 shows a scatter plot of zonal drifter velocities versus $-(g/f)\partial\zeta/\partial y$ for all latitudes between 5° and 25° N and S, after accounting for the Ekman terms. The relationship appears linear, but the slope exceeds unity by a factor of about 1.4. We attribute this to a number of possible factors. 1) The lagrangian drifter motion effectively integrates over all length and time scales, whereas the smoothing applied to the interpolation of the altimeter data truncates the wave number-frequency spectrum. There could be significant energy in the gradient field that contributes to the drifter velocity but is not resolved in our analysis. A similar situation where drifter velocity is greater than the geostrophic velocity computed from the analyzed fields is evident in the Gulf of Mexico (P. Niiler, personal communication). 2) The wind fields are much smoother that the height gradient fields (see Sect. 5), leaving a significant unresolved wind-driven component. If this is correlated with the geostrophic motion, it will tend to be

accounted for by the height gradient in the multiple regression analyses.

This condition influenced the regression results, where the a_1 regression coefficients were on average biased high by about this 1.4 factor at these latitudes, and assumed to be influenced the lower latitudes similarly. In Appendix A, it was necessary to take this into account as another free parameter, α, in order to obtain a stable estimate of the appropriate decay scale for the weight functions. The best fit was consistently found with $\alpha \approx 1.4$. This factor is considered only for the purpose of fitting to the weight function and is not taken into account subsequently when computing geostrophic currents from the height gradient fields.

From (7), (8) and (A.1) we have

$$\boldsymbol{U}_g = W_b \boldsymbol{U}_b + a_1 \boldsymbol{Z} \tag{10}$$

with:

$$a_1 = (1 - W_b)\left(\frac{ig}{f}\right) = W_f\left(\frac{ig}{f}\right) \tag{11}$$

which gives a relationship between the weight functions and the a_1 regression coefficient. This expression was used in Appendix A to estimate the appropriate decay scale for W_b using the empirical a_1 coefficients described below.

The real and imaginary parts of a_1 regression coefficients from (9) are shown as small dots (real part) and open circles (imaginary part) in Fig. 3. As shown above, a_1 represents a measure of the contribution of **Z** to the flow. The real parts are generally near zero, consistent with the absence of any real part in (11). The imaginary part of the coefficients increase in magnitude toward the equator until a peak at about 2-3° latitude in each hemisphere, then rapidly decrease to zero and reverse sign in the opposite hemisphere. The r.h.s. of (11) is our model for the imaginary part and is shown as the smooth curve. The form of the model agrees well with the regression coefficients, particularly within the critical latitude range near the equator where the both show a peak amplitude and then decrease equatorward where the function reverses sign. The maxima of the empirical coefficients were

conspicuously larger than the model in the northern hemisphere, partly associated with the amplitude factor described above. The magnitudes were more in agreement in the southern hemisphere. The regression analysis indicates that the application of the weight function in (11) leads to a dynamically consistent description of the flow in relation to the surface height gradient. The peak inflection points in a_1 give the latitude where the influence of the f-plane term is maximum. These occur at a distance from the equator of about the equivalent of the meridional decay scale of the weight function, θ_s=2.2°, as derived in Appendix A. The inflection latitude in the model is governed by θ_s, such that a larger scale would move the inflection poleward and reduce the peak amplitude, and vice versa. It is noted that the equatorial Rossby radius $R_o \approx 2.25°$ latitude (Gill, 1985), so it is apparent that $\theta_s \approx R_o$. The β-plane equatorial solution is applicable within one Rossby radius of the equator and the transition to the f-plane solution occurs near this latitude.

3.4 Ekman motion of drifters

The expression for the combined equatorial geostrophic and Ekman flow $\mathbf{U} = \mathbf{U}_g + \mathbf{U}_e$ is written using (5) and (10)

$$U = W_b U_b + a_1 Z + a_2 \tau \qquad (12)$$

with:

$$a_2 = \frac{1}{ifh + r} = \frac{r - ifh}{r^2 + f^2 h^2} \qquad (13)$$

The coefficient a_2 has units of inverse velocity to scale to the kinematic wind stress (Eq. 3), whereas a_2/ρ will have units $ms^{-1}Pa^{-1}$ to scale to the dynamic wind stress. The real and the imaginary parts represent the velocity components parallel and perpendicular to the wind stress respectively. The imaginary part vanishes and real(a_2)→1/r as f→0 such that the equatorial Ekman flow is directed down wind and the amplitude relative to the wind stress is determined by the inverse of the drag coefficient. The coefficients a_2 were estimated empirically by multiple linear regression of **(9)** (results for a_1 given above). The complex number $r+ifh$ is the inverse of a_2. The real part and imaginary parts (r and fh respectively) are shown as derived from the a_2 regressions coefficients in Fig. 4a. Both these terms

show an increase in scatter and error with latitude. Accordingly, only those values whose error variance was less than the median error were used to derive parameters r and h in order to obtain stable estimates. The horizontal solid line represents the mean value r= 2.15(±.3)x10^{-4} ms^{-1} of this set, and this is most representative of the values close to the equator. The product *fh* shows a clear trend with latitude from which *h* is found by estimating the slope by linear regression with f. The result is h=32.5 ±1.2 m when the fit is again restricted to latitudes where the error is less than the median error.

The real and imaginary parts of a_2 are plotted in Fig 4b with the model (solid curves) given by (13) using constant values of r= 2.15x10^{-4} ms^{-1} and h=32.5 m. There is generally very good agreement between model and empirical results. The imaginary part has a very similar form to a_1 with peak magnitudes near 2°-3° latitude zero crossing at the equator. The regression coefficients are greater than the model at latitudes under 5°, while the magnitudes match well at the higher latitudes. The real part of the model increases monotonically to a peak at the equator given by 1/r. The real regression coefficients are generally smaller than the model at latitudes under 5°, while the magnitudes match well at the higher latitudes. The dashed curves show the result if r is increased by an arbitrary factor of 1.25 to illustrate the model sensitivity. The influence is restricted to near the equator, where the peak amplitudes are reduced. This improves the apparent agreement with the real coefficients and degrades the agreement with the imaginary coefficients. Decreasing r would have the opposite effect of improving the imaginary fit and degrading the real fit. The conclusion is that the derived values for r and h given represent the best approximation for a_2.

Van Meurs and Niiler (1997) proposed a two parameter Ekman model $\mathbf{U_e}=Be^{i\phi}\tau$ which is equivalent to the present formula with $B=1/(r^2 + f^2h^2)$ and ϕ=arctan(fh/r)=arctan(−imaginary(a_2)/real(a_2)) representing an amplitude and turning angle relative to the wind respectively. They also derived a complex-valued mixing depth parameter and calculated a mean magnitude of 33 ±25 m in the Pacific between 45˚N and 50˚N latitude using drifters drogued at 15 m. Ralph and Niiler (1999) estimated h≈26 m ±3 m by differencing mean drifter motion from geostrophic fields in a study extending from tropical to midlatitudes. The similarity of these results indicates that the depth scale parameter may be quite stable over a wide range of latitude and ocean conditions.

The coefficients B and ϕ are plotted from our results in Fig. 5, with B scaled to m s^{-1} Pa^{-1}. The

velocity per unit wind stress is an order of magnitude greater at the equator than at 25˚ latitude, showing the amplified equatorial response. The turning angle indicates a rotation toward 90˚ from the wind with increasing latitude under the assumption of constant r and h (solid curve). The estimates of ϕ from the regression coefficients suggest that the turning is less than the model at higher latitudes and may remain at about $\phi \approx 60^{\circ}$-70°. The average turning angle ϕ found by Van Meurs and Niiler (1997) was 56˚ to the right of the wind for data between 45˚N and 50˚N. Ralph and Niiler (1998) obtained 55˚ ±5˚ averaged over the tropical north and south Pacific basins. Similar angles for 15 m drogues were obtained by Krauss (1993) in the sub polar North Atlantic. In contrast, Schudlich and Price (1998) observed 15 m depth currents at nearly right angles to the wind in current meter measurements of the Ekman layer from the western Atlantic near 35˚N. A constant ϕ at higher latitudes in our model requires r or h have some variation with latitude. There is some suggestion of a trend in the distribution of r in Fig. 4, although the scatter increases at latitudes >10˚, reducing the reliability of the gradient estimate. This issue must be resolved at latitudes beyond our domain (25˚S - 25˚N) and will be the subject of future work. For the present analysis, the assumption r and h are constant is considered to be a reasonable first approximation in the tropical band.

In the context of climate models, Zebiak and Cane (1987) use an equivalent $r=r_sH_1$ (their notation), where their values of r_s=2 day^{-1} and H_1=50 m yield a comparative value of $r=2.9\times10^{-4}$ ms^{-1}. However, use of the thickness scale h in place of H_1 would reduce the equivalent r to a value closer to that derived here. The scaling thickness parameter h is more representative of the averaging depth for the momentum input from the wind to balance the motion of a 15 m drogue drifter than to an explicit mixed layer depth. Nevertheless, the mixed layer depths are often observed on this scale due to freshwater buoyancy stratification, particularly in the warm pool (Lukas and Lindstrom, 1991; Anderson et al., 1996).

3.5 Computing surface velocity

With the gridded surface height gradient and wind stress maps (Sect. 2), the surface velocity was computed from (12) based on the smooth models of a_1, a_2 and W_b presented above and in Appendix A. The β-plane geostrophic term U_b was computed within 5˚ of the equator with the polynomial described in Appendix A and wass ignored at higher latitudes. U_g and U_e were computed separately so that their relative contribution can be evaluated. Flow field examples are given in Sect. 5.

4. Comparison with in situ measurements

4.1 Drifter velocities

A statistical comparison between the satellite-derived and drifter velocities was analyzed with U_g and U_e fields interpolated to drifter samples within 1˚ latitude bins as described in Sect 2.3. Comparison statistics were computed from the ensemble of co-located samples within each bin and are thus presented as a function of latitude. Fig. 6 presents the zonal average components for zonal (u) and meridional (v) velocity components. Comparisons are shown for U_g, U_e and their sum. Caution is advised in interpreting these as basin-wide zonal average because the statistics are biased by the drifter sampling, which is weighted toward the western Pacific warm pool region (Fig. 1).

Zonal velocity (u): An eastward jet-like feature at the equator is prominent and can be attributed to the drifter distribution and the frequency of westerly wind events in the warm pool. Aside from this, the prevailing features of the zonal equatorial current system (SEC, NECC, NEC) are represented. Most of the amplitude is in the geostrophic component. The Ekman contribution is strongest near the equator, 10ºN and near 5ºS, and is otherwise much less significant. The agreement between model and drifter data is very good south of the equator, as well as northward of about 12˚N latitude. Discrepancies as large as ~0.1 ms^{-1} appear in the equatorial jet and the NECC centered near 5-6˚N.

Meridional velocity (v): This is seen to be dominated by the wind-driven flow, with only minor contributions from geostrophy, in direct contrast to the zonal velocity above. The geostrophic current is of the appropriate sign and magnitude that the sum of v_g and v_e is in excellent agreement at all latitudes except an outlier at the southernmost bin. The meridional velocity magnitude peaks between latitudes 5˚-10˚ latitude in both hemispheres, and the equatorial divergence is clearly evident.

Differences and correlations: Fig. 7 depicts standard deviations of differences between drifter and computed u and v components (mean difference removed). The standard errors tend to remain within 0.1-0.15 ms^{-1} for latitudes >10˚. This background error may be associated with interpolation uncertainty, drifter motions that cannot be resolved by the spatial and temporal scales in the analysis,

errors in the height and wind data, and the inevitable discrepancies that arise from such different kinds of measurements. The errors near the equator grow to about 0.3 ms^{-1} for u and about 0.2 ms^{-1} for v. This additional error is attributed to the elevated variability along the equator, the lack of local acceleration terms in our model, and the smoothness of the height fields which can lead to underestimated U_g near the equator. Correlations are larger for the u than for the v components, primarily due to the geostrophic term, varying considerably with latitude. Coefficients range from 0.4-0.9 for u and 0-0.7 for v. The minimum correlation for v is at the equator where the mean is also zero. The weaker correlation for v is in contrast to the much better agreement of the mean (Fig. 6). It is evident that the variations in the meridional drifter velocity are associated with processes not as well resolved by our analysis as are the zonal fluctuations. The slight negative correlation of the geostrophic v term at the equator is physically relevant. It represents the convergent tendency in response to the zonal pressure gradient that is counter to the divergent drifter motion.

4.2 Current meter moorings

The current meter data processing is described in Sect. 2.4 above. The surface velocity maps were linearly interpolated to each mooring location to generate time series for comparison, shown in Fig. 8. Our zonal velocities were significantly biased westward by about ~0.3 ms^{-1} at the two moorings in the eastern basin (110˚W and 140˚W). Much of the bias is comparable in magnitude to the Ekman component, which is predominantly westward in the trade wind zone. Many low frequency variations were well matched, yet significant disparities degrade the correlations to the 0.2 to 0.5 range, with the smallest correlation at 110W. Difference standard deviations were of similar magnitude to the bias (~0.3 ms^{-1}). The mean differences in meridional components were small (0.—0.1 ms^{-1}). Meridional variations were very noisy, largely due to instability waves that are not completely resolved by our gridded analysis, and by high-frequency inertia-gravity waves which are aliased. The meridional velocity correlations are thus near zero, consistent with the drifter correlation at the equator described above. Agreements were better in the western basin at the two TAO moorings (170˚W and 165˚E) and the COARE IMET mooring (2˚S, 156˚E). Zonal mean differences are small (<.07 ms^{-1}) and biased slightly eastward in our data, the correlations much higher (0.6-0.7) and the difference standard deviation reduced (0.17—0.25 ms^{-1}). Meridional comparisons are similar to the eastern moorings with small biases and weak correlations. TAO zonal currents were compared with a shorter duration record of equatorial currents derived from TOPEX/Poseidon results by Menkes et al. (1996) with smaller differences found, particularly in the eastern Pacific. In contrast to the present study, their calculations

employed considerably more filtering with a 70-day Hanning window and 4.5˚ meridional window applied. Their results were also unbiased in that only anomalies were considered. Picaut et al, 1989 compared geostrophic currents derived from adjacent moored thermistor arrays with moored current meters on the equator at 110W and 165E. Their results are consistent with ours at 110W where they found a mean westward bias of about 0.25 ms^{-1} at the surface. They found a much larger westward bias ~0.7 ms^{-1} at 165E and 50 m depth, in contrast to our slightly eastward bias at the surface.

The principal discrepancy in the present comparison with TAO moorings is that our analysis does not generate equivalent eastward velocity along the equator in the eastern basin, introducing the mean westward bias. The prevailing U_e is westward in the trade wind zone, so the bias may be that U_g is not sufficiently eastward to compensate. The three-year mean in our analysis is governed by the geostrophic estimate from the mean Levitus height field (Sect. 2.1). The first consideration is that the smoothness of the Levitus field flattens the surface height curvature at the equator upon which the geostrophic current depends. However, this would have the opposite bias we observe because the mean westward flow occurs with a surface height trough, and our westward bias implies a deeper rather than flatter trough.

Secondly, the Levitus field is not likely to be representative of the true climatic mean of this period. The eastern tropical Pacific is known to have experienced anomalously warm conditions during this time, so that the true mean surface height conditions may have been more conducive to eastward flow than indicated in the historical Levitus field. We examined recent ocean model reanalysis data (RA-6) from NOAA/NCEP to determine if such differences are apparent. (Acero-Shertzer et al., (1997) compare an earlier version, RA-3, with drifters in the Tropical Pacific, concluding that velocity discrepancies, greatest in the western Pacific, were caused primarily by neglecting salinity in the data assimilation.) We found that our equatorial geostrophic flow computed with dynamic heights from RA-6 data averaged over 1993-1995 indicated increased westward zonal velocity bias relative to Levitus in the eastern Pacific, and added a significant eastward bias in the western Pacific as well (not shown). These widen the discrepancies with TAO currents in both regions. The conclusion is either the RA-6 errors remain large and offer no answer to this question, or the bias lies in other aspects of our analysis.

Thirdly, the mean westward surface velocity is minimum at the equator relative to the adjacent

latitudes (<2˚) based on Doppler current profiles (Johnson and Luther, 1994) and geostrophic calculations (e.g. Cornuelle et al., 1993). Consequently, interpolation between grid points on either side of the equator will show westward bias at the midpoint. Mean zonal velocity presented by Johnson and Luther (1994) indicate this effect could account for perhaps up to 30% of the bias we observe.

Lastly, these comparisons are made at locations of considerable vertical shear between the surface and EUC core. The simple Ekman model drag formulation may not account for the associated vertical shear stress and thus the computed U_e is likely to be biased westward. Geostrophic vertical gradients >1.5 ms^{-1} were observed in the top ~50m at 110˚W (Hayes, 1982), indicating that 0.3 ms^{-1} shear may commonly separate the surface U_g from that observed by a current meter at 10 m depth. This alone can account for the observed bias considering that our computation applies the surface height gradient and thus yields a surface U_g only. By comparison, the shear is small at the western mooring sites, accounting for the smaller biases found. The EUC is often asymmetric about the equator in the Eastern Pacific, displaced a full degree south in Hayes' (1982) data for example. Such meandering will influence the moored current meter observation much more than our smoothed and interpolated analysis, and might explain the lower correlations we see in the eastern two mooring sites. Under such circumstances, and considering the great difference in the method of observation, the agreement between these currents may be as good as can be expected.

5. Case Studies

5.1 Westerly wind burst:

Fig. 9 illustrates the surface currents during the strong westerly wind burst which began in late December 1992 during the TOGA COARE field experiment (Weller and Anderson, 1996). These maps represent the ~10-day sample interval beginning 21 December 1992. The Ekman, geostrophic and combined flow patterns are shown at the 1˚ grid resolution. A strong eastward wind-driven Ekman jet and equatorial convergence was well-developed along the equator between 140˚E to 180˚E. This brought about strong zonal convergence between eastward and westward Ekman flow near the date line. A band of eastward geostrophic flow was developed in the NECC extending east of the date line, accompanied by a band of eastward flow along the equator between about 150˚E to 180˚E. The

sum of the two terms indicates considerable structure to the circulation pattern highlighted by the equatorial currents.

5.2 The 1997-98 El Niño onset

Monthly mean current maps for December 1996 and June 1997 illustrate dramatic changes in the surface circulation across the entire Pacific Ocean associated with the development of ENSO. The surface flow in December 1996 represents the conditions during the mild cold event in the tropical Pacific in 1996 that preceded the 1997 El Niño. Negative sea surface temperature (SST) and easterly wind anomalies were present eastward of the date line (Climate Diagnostics Bulletin, 1996). The Ekman flow field (Fig. 10a) shows strong westward and divergent flow across the central basin. The response to a westerly wind burst is apparent along the equator between 140˚E and 170˚E, similar to the event observed during COARE (Fig. 9). This event likely contributed to the onset of the subsequent warm event the following spring (McPhaden, 1999). The December geostrophic currents were westward along the equator in the eastern and central basin, and the NECC was well developed across the entire basin. Zonal current anomalies (not shown) were negative (westward) relative to a six-year December mean, although to a lesser degree than the November anomaly one month prior. These conditions are in stark contrast to those that had evolved just six months later in June 1997 as the El Niño was in full progress. Warm SST anomalies had already exceed 2˚C in the east, and wind anomalies were westerly across the basin (Climate Diagnostics Bulletin, 1997). Ekman currents were weaker, and the trade wind zone had shifted to the eastern third of the basin (Fig. 10b). Geostrophic currents formed a band of strong eastward flow across the basin along the equator and merging with the NECC to the north. Speeds exceeded ~0.5 ms^{-1} in the core, with anomalies exceeding these magnitudes, given that the climatic condition for June is westward flow. This pattern became apparent in March and April, and persisted through the remainder of 1997 (not shown). One can assume that a considerable amount of warm water was advected eastward during this episode. Detailed analyses of the genesis and evolution of the 1997 El Niño currents and thermal advection, and of the warm pool circulation during TOGA COARE, are the subjects of separate studies with these data. The examples given here are intended to provide the reader with an understanding of both the spatial resolution and variability evident in these fields.

6. Summary and Conclusions

This study has introduced a straightforward method to compute steady state surface geostrophic and Ekman currents from satellite derived surface height and vector wind data. The regional focus has been the tropical Pacific, and accordingly it was essential to attain a continuity of flow estimates across the equator where f=0.

In the geostrophic calculation, we introduce a weighted blended analysis combining equatorial and off-equatorial geostrophic estimates; an approach suitable for analyzing conventional *in situ* dynamic height sections as well. In the process of developing the appropriate weight functions, we also describe the relation between the velocity and sea level gradient in the transition between hemispheres. This was verified with a regression analysis between height gradients and drifter velocities. We considered how close to the equator the [f-plane] geostrophic assumption is valid, and likewise, how far from the equator the equatorial β-plane approximation is valid. The answer lies at approximately the distance off the equator of about one equatorial Rossby radius (R_o). This is the decay scale of the Gaussian weight function for the β-plane solution. Secondly, the coefficient a_1 giving the velocity in proportion to the height gradient is maximum at this latitude, indicating the maximum influence of the f-plane solution. It then decreases toward the equator, crossing zero at f=0 so the current and height gradient are independent at the equator, and reverses sign in the opposite hemisphere.

Ekman currents were derived to represent the motion of a 15 m drogue drifter relative to the wind stress. This avoids the complication of a theoretically infinite depth Ekman layer at the equator because we are concerned with the motion at only one level. A two-parameter model defines the motion at 15 m relative to the wind at any latitude, with the flow being directly down wind at the equator. The parameters were determined by regression analysis with drifters. The depth scale parameter *h* appears to be a constant 32.5 m within the latitude range for this study (±25˚). The drag coefficient parameter *r* is also treated as a constant in this range. However, the analysis indicated a weak variation *r* with latitude that implies that the angle between the 15 m drogue drifter velocity and the wind stress becomes steady at about 60-70˚, consistent with other results from higher latitudes reported in the literature. Consequently, caution is advised in extending this Ekman model to higher latitudes with the constant *r* used here.

An extensive comparison between derived currents and both drifter and current mooring *in situ* data are presented. Zonally averaged co-located drifter and interpolated derived currents were in excellent

agreement for the meridional current, which was almost entirely wind-driven. Mean zonal currents were dominated by the geostrophic term, and agreed very well south of the equator. Biases of ~0.05 to 0.1 ms^{-1} were evident in the core of the major zonal currents in the northern hemisphere, where the magnitudes were underestimated in our analysis. The standard deviation errors were about 0.1 ms^{-1} at the higher latitudes of the study, and increased equatorward to ~0.3 ms^{-1}. This error variance is attributed to the local accelerations influencing drifter momentum that are not included in our steady linear formulation. The current mooring comparisons on the equator (10 m depth zonal component) were in much better agreement in the western part of the basin (165˚E and 170˚W), where correlations were ~0.7 and the mean biases were <0.1 ms^{-1}. Similar results were obtained at 2˚S, 156˚E at 19 m depth. Mean biases of about 0.3 to 0.4 ms^{-1} and correlations 0.2 to 0.5 were obtained at 110˚W and 140˚W. Some of this difference can be attributed to interpolation error. However, the strong mean shear above the EUC core produces a velocity gradient on the scale of 0.3 ms^{-1} between the surface and 10 m at these longitudes. We attribute the mean velocity differences primarily to this cause. EUC meandering and instability waves introduce variability to the mooring velocities that is not resolved in our gridded analysis and may account for the weaker correlations observed at these sites.

Two case studies are presented for illustration. The first portrays the surface flow in the west Pacific warm pool during a major westerly wind burst observed during the TOGA COARE experiment. Strong convergence and eastward velocity were evident in the Ekman field, and an eastward geostrophic flow is seen developing. In the second example we contrast the conditions prior to and during the large-amplitude El Niño on 1997. The surface zonal equatorial current was eastward at about 0.5 ms^{-1} across most of the basin in June 1997 as the El Niño was rapidly developing and SST anomalies had exceed 2˚C. This is in complete contrast to six months earlier when the flow was opposite and trade winds were strong in December 1996 as the previous cool SST anomaly phase was coming to an end. These velocity fields can now be used to monitor the monthly circulation patterns with a continuous flow of satellite observations. Further analyses will address intraseasonal to interannual variability, heat transport and other dynamics.

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Appendix A: *Near-equator geostrophic currents and weights*

This appendix describes the β-plane and f-plane geostrophic computations from the height gradients near the equator and the development of the associated weight functions based on the respective error variances. $\mathbf{U}_g$ is estimated as a weighted sum of two geostrophic expressions, $\mathbf{U}_b$ and $\mathbf{U}_f$, given by (6)-(8). The error of $\mathbf{U}_b$ will be smallest at the equator, while $\mathbf{U}_f$ has an error that grows toward the equator as $f \to 0$. The error variances for $\mathbf{U}_b$ and $\mathbf{U}_f$ are given by σ_b^2 and σ_f^2, respectively. We then prescribe the weight functions W_b and W_f to vary inversely with the meridional structure of the respective error variances such that the minimum variance is given to the weighted sum (8).

$$W_b \propto \sigma_b^{-2} \; ; \quad W_f \propto \sigma_f^{-2} \; ; \quad W_b + W_f = 1 \tag{A.1a}$$

W_b and W_f are then expressed by assuming that the ratio of weights equals the inverse of the error variance ratio, which yields:

$$W_b = \frac{\sigma_f^2}{\sigma_f^2 + \sigma_b^2} \quad ; \quad W_f = \frac{\sigma_b^2}{\sigma_f^2 + \sigma_b^2} \tag{A.1b}$$

Before deriving expressions for the weights it is convenient to introduce a unit latitude scale $\theta = y/L$ (with L=111 km) whereby θ becomes latitude in degrees. It will be seen below that this scaling cancels out in the variance ratios and does not influence W_b and W_f. $\mathbf{U_f}$ is then:

$$\mathbf{U_f} = \frac{ig}{\beta L}\frac{1}{\theta}\mathbf{Z} \tag{A.2}$$

The error variance σ_f^2 is obtained by simple error propagation as:

$$\sigma_f^2 = \left(\frac{g}{\beta L}\right)^2 (\theta)^{-2}\sigma_z^2 \tag{A.3}$$

where σ_z^2 is the error variance of **Z**.

The estimate of $\mathbf{U}_b$ and its error variance σ_b^2 begins by writing $\mathbf{U}_b$ in terms of the undifferentiated geostrophic equation (4),

$$\beta y \boldsymbol{U}_b = ig\boldsymbol{Z} \tag{A.4}$$

and making a polynomial fit by defining

$$\boldsymbol{U}_b = \boldsymbol{U}_0 + \boldsymbol{U}_1\theta + \boldsymbol{U}_2\theta^2 + \ldots \tag{A.5}$$

and

$$\boldsymbol{Z} = \boldsymbol{Z}_0 + \boldsymbol{Z}_1\theta + \boldsymbol{Z}_2\theta^2 + \boldsymbol{Z}_3\theta^3 + \ldots \tag{A.6}$$

We write the $\mathbf{U}_n$ in terms of the $\mathbf{Z}_n$ by combining (A.4)-(A.6) and matching terms of the same power of θ to obtain

$$\boldsymbol{U}_n = \frac{ig}{\beta L}\boldsymbol{Z}_{n+1} \tag{A.7}$$

This definition of the $\mathbf{U}_n$ ensures that the velocity field computed from the height gradient terms ($\mathbf{Z}_{n+1}$) satisfies the differentiated form (6), which is written:

$$\boldsymbol{U}_b + \theta\frac{\partial \boldsymbol{U}_b}{\partial\theta} = \frac{ig}{\beta L}\frac{\partial \boldsymbol{Z}}{\partial\theta} \tag{A.8}$$

In practice, U_b is computed within 5° of the equator from (A.5) and (A.7), where the coefficients Z_n are obtained with (A.6) by polynomial fit along meridians in each height gradient map. From (A.7), $\mathbf{U}_b$ depends only on the n>0 terms in $\mathbf{Z}$ and is independent of $\mathbf{Z}_0$. Thus, in fitting the $\mathbf{Z}_n$ we allow for the possibility that there is an ageostrophically balanced height gradient at the equator. The $\mathbf{Z}_n$ were obtained by a polynomial regression to third order (A.6) between latitudes 5°N and 5°S, from which U_b was computed with (A.5) and (A.7). A polynomial to third order in $\mathbf{Z}$, thus second order in $\mathbf{U}$ (A.7), is the lowest order model that allows a velocity maximum or minimum at the equator, a natural feature of the equatorial current system we judged was necessary to preserve. The meridional structure and magnitude of U_b will of course depend on the order chosen, but the development below indicates that there are no fundamental changes to the weight functions if the order is chosen differently.

Using error propagation again, the error variance σ_b^2 is defined from (A.5):

$$\sigma_b^2 = \sigma_{U_0}^2 + 2\theta\rho_{01}\sigma_{U_0}\sigma_{U_1} + \theta^2(\sigma_{U_1}^2 + 2\rho_{02}\sigma_{U_0}\sigma_{U_2}) + 2\theta^3\rho_{12}\sigma_{U_1}\sigma_{U_2} + \theta^4\sigma_{U_2}^2 \tag{A.9}$$

where the ρ_{nm} are the correlations between the nth and mth terms in the polynomial defining $\mathbf{U_b}$. From (A.7), the terms on the r.h.s. of (A.9) can be rewritten as,

$$\sigma^2_{U_n} = \left(\frac{g}{\beta L} \right)^2 \sigma^2_{n+1} \qquad \text{(A.10)}$$

where σ_{n+1}^2 is the error variance of $\mathbf{Z}_{n+1}$, and the ρ_{nm} can similarly be obtained from the correlations between $\mathbf{Z}_{n+1}$ and $\mathbf{Z}_{m+1}$.

The weight functions (A.1b) can be obtained from (A.3), (A.9) and (A.10) and all scaling terms cancel. However, the error variances σ_z^2 and σ_n^2 and the correlations ρ_{nm} must still be specified, which is not trivial to do. It is not apparent how to obtain these rigorously without invoking various assumptions which must be substantiated in some way. Our alternative approach is to introduce an error variance model in terms of a parameter that can later be tested with data. To do this, we first note that σ_b^2 is proportional to a meridional function $F^2(\theta)$ and the variance of the residuals from the polynomial fit. Defining this residual variance to be σ_r^2, we can write

$$\sigma^2_b = \sigma^2_r \left(\frac{g}{\beta L} \right)^2 (\theta)^{-2} F^2(\theta) \qquad \text{(A.11)}$$

where

$$F^2(\theta) = \theta^2 \left[F_0 + F_1\theta + F_2\theta^2 + F_3\theta^3 + F_4\theta^4 \right] \qquad \text{(A.12)}$$

The coefficients F_n are obtained from (A.9) after substituting from (A.10) and re-writing the correlations between the $\mathbf{U_n}$ in terms of the correlations between the $\mathbf{Z_n}$. The variances and correlations of the $\mathbf{Z_n}$ are easily obtained from the polynomial fits used to obtain the $\mathbf{Z_n}$. Only ρ_{02} appears to be non-zero, so that the coefficients of odd powers in (A.9) and (A.11) vanish and the functions remain symmetric about the equator. We will not go into further detail on this, however, as the only point that we wish to establish is that the *form* of (A.11) is appropriate as a function of latitude, and that given σ_r^2 we are now able to compute σ_b^2 .

Substituting the model (A.11) with (A.3) into (A.1b), we have

$$W_b = \frac{1}{1 + F^2(\theta)/R^2} \quad ; \quad W_f = \frac{F^2(\theta)/R^2}{1 + F^2(\theta)/R^2} \tag{A.13}$$

where R^2 is the error variance ratio σ_z^2/σ_r^2. This is the one free parameter in our model, and is quantified by analyzing surface drifter data (below). Note that both $F^2(\theta)$ and R^2 (implicitly through σ_r^2) depend on the the specific latitudes and order of the polynomial used in the fit. If the model order is chosen differently, then the form of the weight functions (A.13) can change as $F^2(\theta)$ changes, but the weights do not necessarily have to change. This is because for a different model order, the value chosen for R^2 must also be re-considered. In fact, we find below that the weight functions are relatively insensitive to the choice of the model order, and to the details of how the correlations are specified (Fig. A.1). We interpret this as evidence that our weight function is robust.

Next, we turned to the analysis of the surface drifter data (see Section 3.3, main text) in order to find the best value of R for a given model order and correlation (ρ_{02}). We are also concerned with the behavior of the weight functions if the model order and correlation structure are changed. In Section 3.3, it is shown how the weight functions influence the ratio between the height gradient and the surface geostrophic current through coefficient a_1 in (10) and (11). This allows us to use the independent drifter regression data to fix the free parameter in our model, R, assuming that the order of the polynomial has been chosen.

In doing these fits we took into account the factor α, the ratio between the drifter velocities and the geostrophic estimates that is discussed in Section 3.3. The procedure simultaneously determined the values of R and α that gave the minimum variance of the difference between the a_1 regression coefficients and $\alpha W_f ig/f$, where $\alpha \approx 1.4$ was consistently found in all cases. We estimated R in this manner for model orders N=1,2, using our best estimates of the correlations in some cases and simply setting the correlations to zero in other cases, and then computed the weight functions that resulted. Example cases are shown on Fig. A.1. The result is that regardless of the model order, and whether the correlations are accounted for or not, the weight functions are a similar quasi-Gaussian shape. They retain a similar length scale by requiring the model functions to approximate the independent drifter a_1 regression data. The N=2 curve that includes the ρ_{02} correlation maintains a Gaussian-like profile with some higher order structure. This is an artifact of the negative correlation of the odd-

numbered Z_n coefficients in the error propagation formula. Such correlation is inevitable when fitting noisy data to a polynomial, and accordingly, no physical significance is given to the apparent structure of W_b in this case. We interpret these overall results as indicating that the W_b must be Gaussian in nature in order to satisfy the error variance criteria (A.1) and thus effectively minimize the variance of the final velocity field that we compute as a weighted average of the $\mathbf{U_f}$ and $\mathbf{U_b}$ approximations.

It is more straightforward to define a universal weight function as a Gaussian:

$$W_b = \exp(-(\theta/\theta_s)^2) \quad ; \quad W_f = 1 - W_b \qquad \text{(A.14)}$$

and select the length scale, θ_s , to match the drifter regression data as above. The best-fit value of θ_s is 2.2°, and the resulting weight function is also shown on Figure A.1. The Gaussian weight function is consistent with the results for any model order, is insensitive to the choice of the correlation structure, and are much easier for other researchers to incorporate into their own work. Thus we adopted this function for the calculations in the main text. We emphasize, however, that this choice of a Gaussian weight function is not arbitrary, but is chosen as an appropriate shape to represent the results of the error analysis.

Appendix B: *Altimeter surface height gridding procedure*

TOPEX/Poseidon sea surface height data along the ground tracks were preprocessed (1) to remove estimates of tide model errors by removing fitted harmonics at the tidal alias frequencies, and (2) to remove long wavelength, high frequency signals that were interpreted as orbit errors or errors in the environmental corrections (e.g., ionospheric or water vapor corrections to the travel time). The data were slightly low-passed along track and subsampled to a 0.25 degree spacing, and the roughly 10-day samples at each location along the ground track were slightly low-passed and temporally interpolated to a set of standard times spaced at exactly 1/36 yr.

These heights were then used to create a 0.25 X 0.25 degree spatially gridded field at each time by creating a weighted average of two estimates of the height field at the desired grid locations. The first estimate was simply the nearest data point. The second estimate was again the nearest data point, but shifted in time to allow for propagation appropriate to the spatial lag between the data point and the desired grid point location. The weights were chosen based on estimates of the proportion of the signals contained in propagating, as opposed to stationary, signals, and are thus a complicated function of spatial location. These weights and the optimal propagation speeds at each location were determined by a separate calculation with TOPEX/Poseidon data. The final grid was computed by low-pass filtering the spatial field at each time step using a bivariate Gaussian set of weights. The length scale for the Gaussian weights was made proportional to the local Rossby radius and the resulting height field retains higher wavenumber variability at high latitudes. The length scale, however, was never allowed to be smaller than 0.5 degrees; this choice was based on the fact that the basic TOPEX/Poseidon spacing is relatively coarse.

A more detailed discussion of the gridding procedure is beyond the scope of this paper. For the present purposes we simply note that the performance of this grid was checked against other possibilities and performed as well or better than the other choices, as measured by the ability to produce reasonable velocity fields. It has proved superior to Optimal Interpolation for a variety of choices of correlation functions. In areas where the Gaussian scale is comparable to other gridded products, such as one provided by the Center for Space Research at the University of Texas, results are similar.

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Figure Captions

Fig. 1: Geographic outline (dashed box) of the analysis region. Drifter trajectories as shown during the period October 1992 - December 1994 were used for computing drifter velocities in the analysis.

Fig. 2: Scatter diagram of the zonal drifter velocities, corrected for Ekman motion, and the geostrophic velocity derived from the interpolated meridional height gradient (abscissa) for latitudes greater than 5°. Drifter velocities are apparently larger than can be explained by the height gradient, which is attributed to the smoothing in the altimeter gridding analyses and interpolation procedures.

Fig. 3: The real and imaginary parts of the regression coefficient a_1 are the solid and open circles, respectively, with error bars (1σ). The heavy curve shows the model of a_1 computed with (11).

Fig. 4: Top Panel: Asterisks give the drag coefficient *r* with error bars estimated from the regression coefficients a_2. The solid horizontal line is the average value $r \approx 2.15 \times 10^{-4}$ ms^{-1} of the values with small errors (see text). This value is most suitable near the equator where the magnitude of *r* is most critical. Open circles give the coefficient *fh* estimated from the regression coefficients a_2 from which *h* is computed as the slope. The solid line indicates the slope with *h*=32.5 m derived by linear regression with f (see text). Bottom Panel: The model using $r=2.15 \times 10^{-4}$ ms^{-1} and *h*=32.5 m (solid line) and a_2 regression coefficient real part (asterisks) and imaginary part (open circles) for the down wind and cross wind velocity component respectively. The dashed line shows the model with r increased by a factor of 1.25 to illustrate the sensitivity to r.

Fig. 5: Top Panel: The amplitude function B of drifter speed per unit dynamic wind stress (Pa) for the model (solid curve) and derived from a_2 regression coefficients with error bars (symbols). Bottom Panel: Same as above except showing the drifter velocity vector angle to the right of the wind stress vector.

Fig. 6: Zonally averaged comparisons between drifter (light curves with open circles) and interpolated satellite-derived velocity components (dark curves), with horizontal bars showing standard error of the mean. Left panels are zonal velocity u and right panels are meridional velocity v, with subscripts g and e denoting geostrophic and Ekman terms respectively.

Fig. 7: Left Panels: Standard deviation differences between drifter and satellite-derived velocity terms (see Fig. 6) for u (circles) and v (triangles) components. Right Panels: Same as left panels except showing correlation coefficients.

Fig. 8a: TAO current mooring 10 m velocity comparisons at the equator and 110˚W. The light curve is mooring data with 21-day running mean and interpolated to 36/year time step of the satellite data. The dark curve is the satellite-derived current spatially interpolated to the mooring site. The dashed curve is the geostrophic term only. Top panels show zonal and bottom panels meridional components respectively.

Fig. 8b: Same as 8a except showing 140˚W mooring data.

Fig. 8c: Same as 8a except showing 170˚W mooring data.

Fig. 8d: Same as 8a except showing 165˚E mooring data.

Fig. 8e: Same as 8a except showing 2˚S, 156˚E IMET mooring data at 19 m depth.

Fig. 9a: The Ekman flow field during the 20-day composite 21 December 1992 to 10 January 1993 during a strong westerly wind burst that occurred during the TOGA COARE experiment. The COARE Intensive Flux Array (IFA) location is shown by the square at 2˚S, 156˚E.

Fig. 9b: Same as 9a except showing the geostrophic term.

Fig. 9c: Same as 9a except showing the sum of Ekman and geostrophic terms.

Fig. 10a: The Ekman term, geostrophic term and their sum (top, middle and bottom panels respectively) for the monthly composite during December 1996, depicting conditions prior to the onset of the 1997 El Niño. Data were sub sampled to a 2˚x5˚ degree grid for these plots.

Fig. 10b: Same as 10a except showing June 1997 after the El Niño onset in the spring of 1997. The eastward geostrophic flow on the equator persisted through December 1997 (not shown).

Appendix

Fig. A.1: Four examples of the weight function W_b for cases discussed in Appendix A. N signifies the order of the polynomial model for U_b in Eq. (A.5).