

## Seasonal mixed layer heat budget of the tropical Atlantic Ocean

Gregory R. Foltz, Semyon A. Grodsky, and James A. Carton

Department of Meteorology, University of Maryland, College Park, Maryland, USA

Michael J. McPhaden

Pacific Marine Environmental Laboratory, NOAA, Seattle, Washington, USA

Received 2 August 2002; revised 16 January 2003; accepted 5 March 2003; published 14 May 2003.

[1] This paper addresses the atmospheric and oceanic causes of the seasonal cycle of sea surface temperature (SST) in the tropical Atlantic on the basis of direct observations. Data sets include up to 4 years (September 1997 to February 2002) of measurements from moored buoys of the Pilot Research Array in the Tropical Atlantic (PIRATA), near-surface drifting buoys, and a blended satellite in situ SST product. We analyze the mixed layer heat balance at eight PIRATA mooring locations and find that the seasonal cycles of latent heat loss and absorbed shortwave radiation are responsible for seasonal SST variability in the northwest basin (8°N–15°N along 38°W). Along the equator (10°W–35°W), contributions from latent heat loss are diminished, while horizontal temperature advection and vertical entrainment contribute significantly. Zonal temperature advection is especially important during boreal summer near the western edge of the cold tongue, while horizontal eddy temperature advection, which most likely results from tropical instability waves, opposes temperature advection by the mean flow. The dominant balance in the southeast (6°S–10°S along 10°W) is similar to that in the northwest, with both latent heat loss and absorbed solar radiation playing important roles. *INDEX TERMS:* 4231

Oceanography: General: Equatorial oceanography; 4247 Oceanography: General: Marine meteorology; 4504 Oceanography: Physical: Air/sea interactions (0312); 4512 Oceanography: Physical: Currents; *KEYWORDS:* tropical Atlantic, heat balance, mixed layer

**Citation:** Foltz, G. R., S. A. Grodsky, J. A. Carton, and M. J. McPhaden, Seasonal mixed layer heat budget of the tropical Atlantic Ocean, *J. Geophys. Res.*, 108(C5), 3146, doi:10.1029/2002JC001584, 2003.

### 1. Introduction

[2] The seasonal cycle of shortwave radiation at the top of the atmosphere is primarily semiannual in the tropics, with maxima in boreal spring and fall. On the other hand, variability in the tropical Atlantic ocean-atmosphere system is dominated by the annual harmonic. The difference reflects seasonal changes in the radiative properties of the atmosphere and the dynamics and thermodynamics of the ocean. In this study we use direct observations, primarily from recently deployed moorings of Pilot Research Array in the Tropical Atlantic (PIRATA) [Servain *et al.*, 1998], to examine the causes of the seasonal cycle of sea surface temperature (SST) in response to seasonally varying surface heating and winds.

[3] Net surface heat flux is a combination of latent and sensible heat loss, shortwave radiation absorption, and net longwave emission. Sensible heat loss is insignificant ( $<10 \text{ W m}^{-2}$ ) due to small air-sea temperature differences, while net emission of longwave radiation is a relatively constant  $\sim 50 \text{ W m}^{-2}$  [da Silva *et al.*, 1994]. Seasonal variations in latent heat loss and downwelling surface shortwave radiation are more significant. Both are influenced by the latitudinal movement of the narrow band of clouds associ-

ated with the Intertropical Convergence Zone (ITCZ), and shortwave radiation is additionally influenced by changes in the solar zenith angle. In the northern tropics, latent heat loss is lowest during boreal summer and fall, when winds are weak and relative humidity is high ( $>85\%$ ). Latent heat loss rises during boreal winter and spring when the ITCZ is close to the equator, low-level humidity is lower, and the northeast trade winds are stronger. In the eastern equatorial zone, latent heat loss has only weak seasonal variations, as low-level relative humidity and wind speed are fairly steady throughout the year [da Silva *et al.*, 1994].

[4] In contrast to the shortwave radiation at the top of the atmosphere, surface shortwave radiation has a significant annual harmonic at most locations. North of 5°N, surface solar radiation reaches a maximum in boreal spring, when the ITCZ is near its southernmost position and the solar zenith angle is high. Between the equator and 5°N, there is a strong semiannual component, with maxima in boreal spring and fall, while on the equator the annual harmonic is again significant, with increasing amplitude toward the west reaching maximum in boreal fall. The reduced amplitude in the east is due to the appearance of reflective stratus clouds in boreal fall over the cool waters of the eastern basin [Klein and Hartmann, 1993; Philander *et al.*, 1996].

[5] Like the tropical atmosphere, the tropical ocean also has a strong annual harmonic. A westward shift of warm

(>27°C) SST in the latitude band 5°N–15°N occurs in boreal summer, along with the development of a tongue of cool 23°C SST along and just south of the equator east of 30°W. This shift occurs concurrently with the annual growth and eastward expansion of the Atlantic warm pool west of 50°W [Wang and Enfield, 2001]. In the west, seasonal changes in SST are weak in the equatorial zone, while north of 8°N a strong annual harmonic appears with a maximum in boreal fall. South of the equator, SST reaches its maximum in boreal spring with an annual harmonic that increases in amplitude eastward to up to 3°C near the African coast [Reynolds and Smith, 1994].

[6] The surface wind field is dominated by the northeast trade winds to the north of the ITCZ and the southeast trade winds to the south, with weakened winds in between [da Silva et al., 1994]. To the east a monsoonal circulation develops in boreal summer, causing the northeast trades to reverse direction. Zonal currents develop in response to these changing winds and to the resulting Ekman divergence. On and south of the equator the westward South Equatorial Current is strongest in boreal summer, with speeds of 55 cm s<sup>-1</sup> in the central basin [Richardson and Reverdin, 1987]. Close to the latitude of the ITCZ (5°N–10°N) the eastward North Equatorial Countercurrent is strong during boreal summer and fall with speeds of 35 cm s<sup>-1</sup>. Between these two major current systems, and on the northern edge of the cold tongue, lie the strong meridional fluctuations of tropical instability waves, which we anticipate are important in transporting heat into the cold tongue in the eastern and central basins [Weisberg and Weingartner, 1988].

[7] A number of observational [e.g., Wyrski, 1981; Enfield, 1986; Hayes et al., 1991; Wang and McPhaden, 1999; Swenson and Hansen, 1999] and modeling [e.g., Koberle and Philander, 1994; Kessler et al., 1998] studies have addressed the causes of the seasonal cycle of SST in the eastern equatorial Pacific. Observational studies in particular have had difficulty closing the heat budget. Hayes et al. [1991] used mooring observations in the eastern equatorial Pacific, together with surface meteorology, to calculate contributions to the mixed layer heat budget during 1986–1988. They found discrepancies between their forcing and actual heat storage that in some cases exceeded 100 W m<sup>-2</sup> and attributed these discrepancies to errors in parameterizations of mixed layer depth, entrainment, and meridional eddy heat divergence.

[8] Wang and McPhaden [1999] used a combination of 15 years of daily mooring observations and climatological surface meteorology to estimate terms in the equatorial Pacific mixed layer heat budget. They estimated a mean seasonal cycle and found significant contributions from net surface heat flux and horizontal advection. They found discrepancies as high as 120 W m<sup>-2</sup> in the east and attributed these differences to a combination of two missing terms: vertical entrainment and vertical diffusion. Consistent with these results, Swenson and Hansen [1999] used a combination of drifting buoys and vertical temperature profiles to investigate the heat budget of the cold tongue. They found that strong seasonal cycles of entrainment and horizontal advection account for a large fraction of seasonal SST variability.

[9] In the equatorial Atlantic, Hastenrath [1977] and Merle [1980] used climatologies of surface heat flux and

heat storage to deduce that horizontal and vertical temperature advection are necessary to balance the annual mean net surface heat flux. Molinari et al. [1985] used atmospheric and oceanic measurements from the First Global Atmospheric Research Program's (GARP) Global Experiment (FGGE) in 1979 to evaluate explicitly the effects of surface energy fluxes and zonal temperature advection on the seasonal cycle of mixed layer temperature. Their results revealed important contributions from zonal advection between 3°N and 9°N. They calculated the sum of vertical and meridional advection/diffusion as a residual and found this term to be important within 3° of the equator. Unfortunately, none of the above studies was able to calculate explicitly the effects of meridional advection or vertical entrainment/diffusion on the mixed layer heat balance.

[10] Weingartner and Weisberg [1991a, 1991b] examined the seasonal heat budget based on 1 year of observations from an equatorial mooring at 28°W (midbasin). They concluded that upwelling creates the cold SST tongue in boreal spring, while SST increases in boreal summer as the result of tropical instability waves. In late summer and fall, advection terms are small and compensating, and diffusion at the base of the mixed layer balances net surface heat gain. In boreal winter, SST increases in response to a net surface heat flux concentrated by the shallower mixed layer.

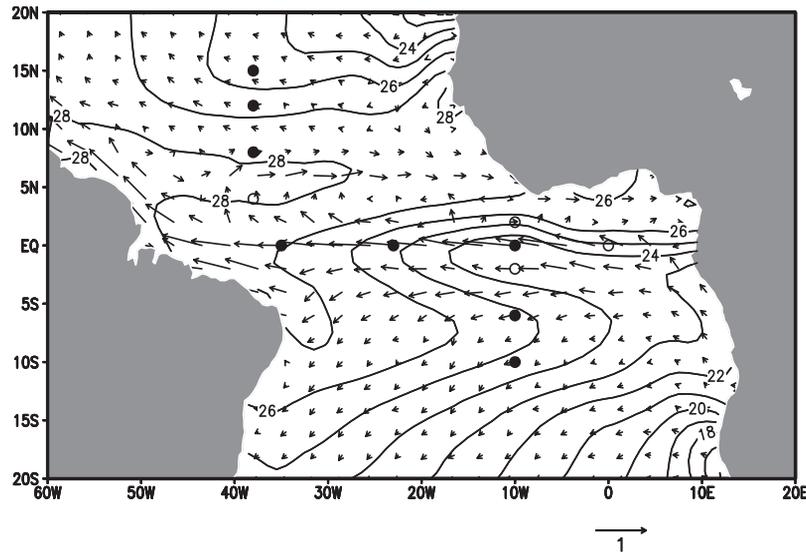
[11] Modeling studies have also stressed the importance of ocean dynamics in the equatorial Atlantic heat budget. Carton and Zhou [1997] concluded that zonal mass divergence causes cooling in the equatorial region east of 20°W, while meridional Ekman divergence plays an important role in the west. Their results indicate that solar heating is most important south of 5°S and north of 10°N, while latent heat loss is dominant in the western basin between 8°N and 12°N. DeWitt and Schneider [1999] also found that advection plays a crucial role in the equatorial region, while the importance of latent heat loss increases toward the subtropics.

[12] The studies mentioned in this section reveal that surface fluxes as well as horizontal and vertical temperature advection play a major role in shaping the seasonal cycle of SST in the equatorial Atlantic. However, the limited duration (~1 year) and spatial coverage of previous in situ observational programs has hindered efforts to quantify these contributions. In this study we use a variety of in situ and satellite measurements, with extended duration and spatial coverage, to calculate explicitly all contributions (with the exception of vertical turbulent diffusion) to the mixed layer heat balance to obtain a quantitative picture of the tropical Atlantic seasonal mixed layer heat balance.

## 2. Data and Methods

[13] The mixed layer heat budget represents a balance of several terms [Stevenson and Niiler, 1983; Moisan and Niiler, 1998]

$$h \frac{\partial T}{\partial t} + h(\mathbf{v} \cdot \nabla T + \overline{\mathbf{v}' \cdot \nabla T'}) + (T - T_{-h})w_e + \nabla \cdot \int_{-h}^0 \hat{\mathbf{v}} \hat{T} dz = \frac{q_0 - q_{-h}}{\rho c_p}. \quad (1)$$



**Figure 1.** Locations of the Pilot Research Array in the Tropical Atlantic (PIRATA) moored buoys (solid and open circles). Solid circles indicate buoys with data records generally at least 2 years in length. Background curves and arrows are climatological July surface temperature [Reynolds and Smith, 1994] and near-surface velocity [Grotsky and Carton, 2001], respectively. Reference velocity arrow is  $1 \text{ m s}^{-1}$ .

The terms represent, from left to right, local storage, horizontal advection (separated into mean and eddy terms), entrainment, vertical temperature/velocity covariance, and the combination of net atmospheric heating and vertical turbulent diffusion at the base of the mixed layer. Here  $h$  is the depth of the mixed layer,  $T$  and  $\mathbf{v}$  are temperature and velocity, respectively, vertically averaged from the surface to a depth of  $-h$ ,  $\bar{T}'$  and  $\bar{\mathbf{v}}'$  are deviations from the time means (the overbar represents a time mean),  $\hat{T}$  and  $\hat{\mathbf{v}}$  represent deviations from the vertical average,  $q_0$  is net surface heat flux, and  $q_{-h}$  represents the sum of heat flux due to penetrative shortwave radiation and turbulent mixing at the base of the mixed layer. Swenson and Hansen [1999] estimate that the vertical temperature/velocity covariance term in equation (1) is  $<10\%$  as large as other terms, and we therefore proceed to neglect this term. Entrainment velocity may be rewritten as  $w_e = \partial h / \partial t + \nabla \cdot h \mathbf{v}$  following Stevenson and Niiler [1983, see their equations (2) and (3)], and is associated with a mass flux that crosses an isopycnal surface.

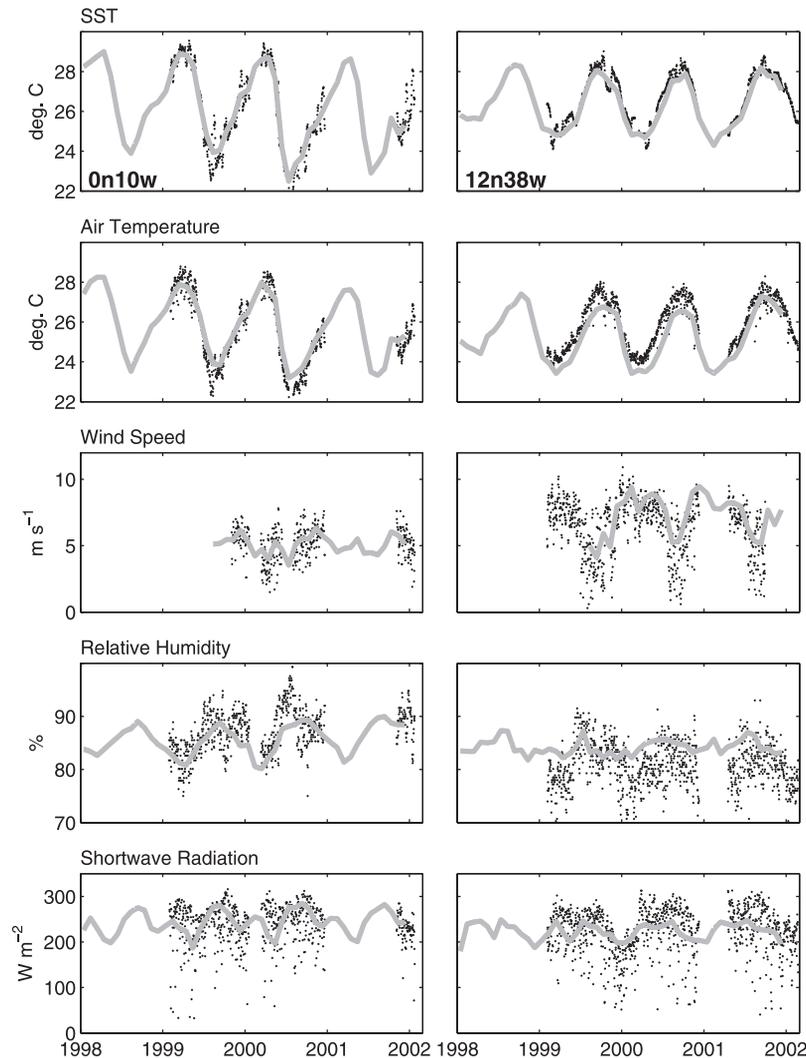
[14] Estimation of the terms in equation (1) requires knowledge of vertically averaged mixed layer horizontal velocity. Unfortunately, vertical profiles of velocity are not available at the PIRATA mooring locations. We therefore use alternative methods (described later in this section) to estimate mean and eddy advection terms in equation (1).

[15] The PIRATA mooring array consists of 12 buoys (see Figure 1). We focus on eight with record lengths exceeding 2 years. Deployed in 1997 to study ocean-atmosphere interactions, these Next Generation Autonomous Temperature Line Acquisition System (ATLAS) buoys measure subsurface temperature at 11 depths between 1 and 500 m with 20-m spacing in the upper 140 m. Air temperature and relative humidity are measured at a height of 3 m above sea level, while shortwave radiation and wind velocity are measured at 3.5 and 4 m, respectively. The sampling interval is 10 min for all variables except short-

wave radiation, which is sampled at 2-min intervals. The instrument accuracies are water temperature within  $\pm 0.01^\circ\text{C}$ , wind speed  $\pm 0.3 \text{ m s}^{-1}$  or 3% (whichever is greater), air temperature  $\pm 0.2^\circ\text{C}$ , and relative humidity  $\pm 3\%$  [Freitag et al., 1994, 1999, 2001; Lake et al., 2003]. Here we use both 10-min and daily averaged data, which are transmitted in near real time via satellite by Service Argos, Inc.

[16] The mooring on the equator at  $10^\circ\text{W}$  has the shortest data record for most variables, yet a clear seasonal cycle is still discernable for all surface variables with the possible exception of wind speed (see Figure 2). The data record at  $12^\circ\text{N}$  is one of the longest, and it reveals a strong seasonal cycle for all surface variables. Net surface shortwave radiation is available directly from the PIRATA moorings assuming an albedo of 6%. We form a daily-mean seasonal cycle by averaging all available PIRATA measurements on a given day from each year. We then create a monthly mean seasonal cycle from these data. On the basis of the analysis of Medovaya et al. [2002] we anticipate that buoy tilting (caused by winds and currents) and aerosol and salt buildup on the sensors may lead to shortwave radiation measurement errors as high as  $20 \text{ W m}^{-2}$ . The problem of aerosol buildup is caused by westward advection of dust from the Sahara desert and is likely most severe at  $12^\circ$  and  $15^\circ\text{N}$  along  $38^\circ\text{W}$  owing to lack of rainfall which would otherwise periodically wash the radiometer dome.

[17] The amount of shortwave radiation absorbed in the mixed layer depends on the depth of the mixed layer and the optical transparency of the water. We have considered two models for penetrative shortwave radiation at the base of the mixed layer. One uses an empirical formula based on chlorophyll  $a$  concentration,  $Q_{\text{pen}} = 0.47 Q_{\text{surf}} e^{-(0.027 + 0.0518 \text{CHL}^{0.428})h}$ , where  $Q_{\text{surf}}$  is surface shortwave radiation, CHL is the chlorophyll  $a$  concentration ( $\text{mg m}^{-3}$ ) from the Sea-viewing Wide Field-of-view Sensor (SeaWiFS), and  $h$  is the depth of the mixed layer [Morel, 1988]. The second model follows Wang and

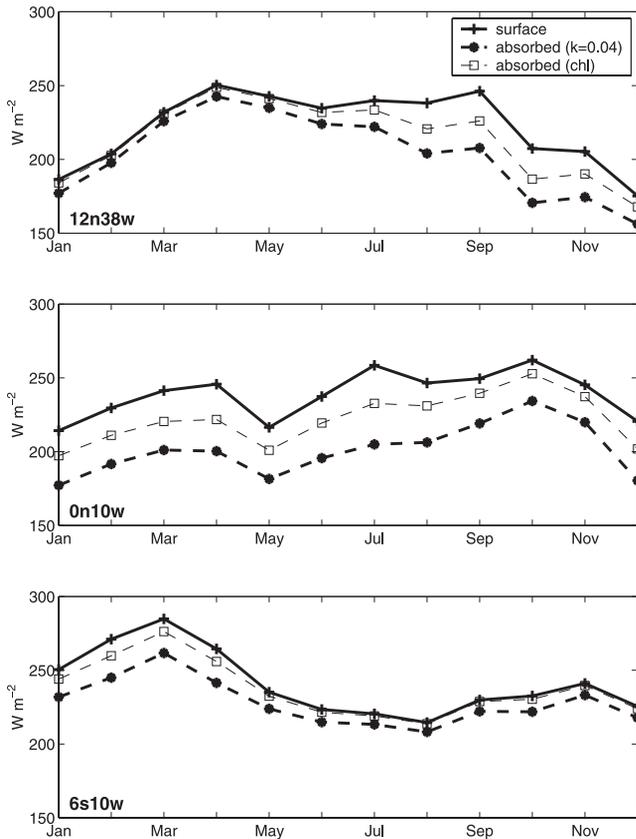


**Figure 2.** PIRATA surface atmospheric and oceanic measurements at (right) 12°N, 38°W and (left) 0°N, 10°W during 1998–2002. Shaded curves represent *Reynolds and Smith's* [1994] sea surface temperature, National Centers for Environmental Prediction/National Center for Atmospheric Research reanalysis [*Kalnay et al.*, 1996] air temperature, relative humidity, surface shortwave radiation (all at a height of 2 m), and QuikSCAT near-surface wind speed.

*McPhaden* [1999] in assuming exponential decay of surface radiation with a constant 25-m  $e$ -folding depth. The seasonal cycle of penetrative solar radiation is strong at 12°N (Figure 3) owing to variations in mixed layer depth (varying from 21 m in October to 67 m in March and calculated according to the method described later in this section) but is weak along the equator at 10°W, where seasonal changes in mixed layer depth are relatively small (from 17 m in July to 34 m in October). Comparison of the two penetrative radiation models reveals that the chlorophyll-dependent equation always predicts more absorption. The bias is greatest along the equator at 10°W, where chlorophyll concentration is high. However, since the bias is nearly constant throughout the year at all locations, both models result in similar seasonal cycles of absorbed solar radiation. For simplicity, and to avoid uncertainties inherent in the empirical formula, we will use the constant 25-m  $e$ -folding depth model to estimate penetrative radiation.

[18] Latent heat flux depends on surface humidity, wind speed, air temperature, and SST. Here we rely on a bulk parameterization,  $Q_e = \rho_a L_e C_e W (q_s - q)$ , where  $Q_e$  is the latent heat flux,  $\rho_a$  is air density,  $L_e$  is the latent heat of vaporization,  $C_e$  is the transfer coefficient,  $W$  is wind speed,  $q$  is the water vapor mixing ratio, and  $q_s = 0.98 q_{\text{sat}}(T_s)$  is the interfacial water vapor mixing ratio, which is assumed to be proportional to the saturation water vapor mixing ratio (the factor of 0.98 accounts for salinity effects). Tests of this algorithm, developed from the Coupled Ocean-Atmosphere Response Experiment (COARE) in the tropical west Pacific [*Fairall et al.*, 1996], have revealed a bias of  $1.5 \text{ W m}^{-2}$  (COARE estimates are lower) and a RMS scatter of  $<20\%$  [*Fairall et al.*, 1996].

[19] The COARE algorithm includes a model to estimate the effects of a diurnal warm layer and cool skin temperatures on latent heat flux. To determine the importance of these effects, we compared 10-min latent heat flux estimates calculated without taking into account cool skin and warm



**Figure 3.** Climatological shortwave radiation at the surface ( $Q_{\text{surf}}$ ) and absorbed in the mixed layer of depth  $h$  ( $Q_{\text{abs}} = Q_{\text{surf}}(1 - 0.47e^{-kh})$ ), calculated with a constant  $e$ -folding depth of 25 m ( $k = 0.04$ ) and a depth that depends on chlorophyll  $a$  concentration ( $k = k(\text{CHL})$ ), at three PIRATA locations.

layer effects to estimates that include both effects. We find that the cool skin effect dominates at all mooring locations, leading to a reduction in latent heat flux of  $4\text{--}8\text{ W m}^{-2}$  (mean bias over the length of each data record), with very little seasonal dependence. The effect is greatest at  $8^\circ$  and  $12^\circ\text{N}$  along  $38^\circ\text{W}$ , where it can reduce latent heat flux by as much as  $10\text{ W m}^{-2}$  on a monthly basis. The diurnal warm layer increases latent heat flux by  $<0.5\text{ W m}^{-2}$  at all locations. Unfortunately, gaps in the PIRATA shortwave radiation records (solar radiation is required for calculation of both warm layer and cool skin effects) have led us to neglect both warm layer and cool skin effects in our analysis.

[20] It is also possible for short-term ( $<1$  day) fluctuations of humidity and wind speed to affect monthly latent heat flux estimates. We compared estimates made from 10-min and monthly measurements (all using 10-min, scalar-averaged wind speed) and found mean biases of  $<3\text{ W m}^{-2}$  at all locations. These results are similar to those of *Esbensen and McPhaden* [1996], which indicate that short-term correlations of scalar-averaged wind speed and humidity have very little impact on latent heat loss in the equatorial Pacific. However, we find seasonal variations to be more significant along  $38^\circ\text{W}$ , where 10-min estimates are lower by  $\sim 10\text{ W m}^{-2}$  during April and August–September at  $8^\circ\text{N}$  and during

August at  $12^\circ\text{N}$ , with very little bias during the remainder of the year (generally  $<2\text{ W m}^{-2}$  on a monthly basis). We therefore use 10-min measurements of air temperature, SST, wind speed, and relative humidity to estimate latent heat flux.

[21] We obtained monthly climatological net longwave radiation from the *da Silva et al.* [1994] surface marine atlas. This parameter has an annual mean of close to  $50\text{ W m}^{-2}$  at all locations and varies less than  $\pm 10\text{ W m}^{-2}$  on a seasonal basis.

[22] Mixed layer depth estimates are affected by the development of shallow diurnal mixed layers. To avoid averaging these shallow effects into our estimates of mixed layer depth, we use hourly SST, averaged between 0500 and 0700 local time, together with daily PIRATA subsurface temperature. We use linear interpolation to calculate the mixed layer depth as the depth at which temperature is  $0.5^\circ\text{C}$  below SST [*Hayes et al.*, 1991]. We then average these daily estimates to form a monthly mean cycle. This definition of mixed layer depth has the advantage that vertically averaged mixed layer temperature is very close to SST but has the disadvantage that vertical mixing at the base of the mixed layer may not be negligible. Errors in our mixed layer depth estimates result mainly from the 20-m vertical resolution of PIRATA subsurface temperature data. We estimate these errors by considering an idealized vertical temperature profile that is homogeneous within the mixed layer and linearly decreasing below. If we assume that the vertical gradient of temperature in the seasonal thermocline is  $\sim 0.1^\circ\text{C m}^{-1}$  (typical of the tropical Atlantic [*Levitus and Boyer*, 1994]), we find that our daily mixed layer depth estimates are too low by 0–5 m.

[23] It is possible for surface freshwater fluxes to create salinity stratification within a deeper, nearly isothermal mixed layer [e.g., *Ando and McPhaden*, 1997]. In such cases, mixed layer depths based on temperature overestimate the true mixed layer depth. To assess whether this effect is important in the tropical Atlantic, we have compared mixed layer depth estimates based on vertical profiles of potential density ( $0.125\sigma_\theta$  criterion) to those with potential temperature ( $0.5^\circ\text{C}$  criterion) at each mooring location using the climatological ( $1^\circ \times 1^\circ \times 1\text{-month}$  resolution) data sets of *Monterey and Levitus* [1997]. We find that, in the annual mean, mixed layer depth estimates based on temperature exceed those based on density by  $<6\text{ m}$  at each location. These differences are similar in magnitude, but opposite in sign, to the uncertainties associated with our estimates based on linear interpolation of vertical temperature profiles (discussed in this section). We therefore ignore both in our analysis.

[24] To calculate horizontal temperature advection and vertical velocity, we first estimate the seasonal cycle of near-surface horizontal velocity following the procedure of *Grodsky and Carton* [2001]. This procedure combines more than 100 years of historical ship drifts, TOPEX/Poseidon sea level (1992–2001), ERS 1/2 surface winds (1992–2001), and velocity from hundreds of drifting buoys, deployed during 1997–2001 and drogued at a central depth of 15 m, to produce estimates with a  $2^\circ$  latitude  $\times$   $3^\circ$  longitude  $\times$  1-month resolution. The procedure uses optimal interpolation, with a first-guess field consisting of long-term mean ship drift and drifter velocity, and applies

corrections based on drifter/ship drift velocity and sea level. The analysis relies on simple assumptions such as geostrophy and Ekman balances off the equator and equatorial dynamics close to the equator. These velocity estimates are multiplied by 20-year (1982–2001) climatological monthly SST gradients [Reynolds and Smith, 1994] in order to estimate monthly horizontal mixed layer heat advection. We also use divergence of these velocity estimates, as well as estimates of the time derivative of mixed layer depth based on PIRATA subsurface temperature, to calculate  $w_e$ . Horizontal gradients of mixed layer depth are estimated from a monthly climatology based on bathythermograph temperature profiles [White, 1995].

[25] We anticipate that meridional velocity in the mixed layer is primarily the result of Ekman drift (since the meridional component of geostrophic velocity is weak) that decreases with increasing depth. Our velocity estimates were calculated mainly from ship drifts, measured a few meters below the surface, and drifting buoys, with a 7-m drogue centered at a depth of 15 m. We thus expect that these values overestimate vertically averaged mixed layer meridional velocity under most circumstances, since the depth of the mixed layer remains  $>15$  m at all locations we consider. We therefore apply a correction that assumes a linear decrease in meridional velocity from the observed value at 15 m to 0 at  $-h$ . No correction is applied for  $h < 15$  m since we cannot accurately estimate surface velocity needed for interpolation from 15 m to the surface. This correction leads to annual mean mixed layer heat advection adjustments (with respect to values obtained from constant vertical profiles of meridional velocity) of  $<12$   $\text{W m}^{-2}$  at all locations. On a monthly basis, adjustments are  $<20$   $\text{W m}^{-2}$  at all locations, with the exception of  $8^\circ\text{N}$ ,  $38^\circ\text{W}$ , where it is more than  $30$   $\text{W m}^{-2}$  during February as the result of a deep ( $\sim 70$  m) mixed layer. Since we cannot estimate the vertical distribution of zonal velocity, we have not applied a correction to zonal velocity estimates.

[26] It is known that tropical instability waves significantly heat the equatorial Atlantic mixed layer during boreal summer and fall through horizontal Reynolds heat fluxes. Since the typical period of tropical instability waves is  $<1$  month, our climatological monthly heat advection estimates do not resolve them. Because in situ velocity measurements from the PIRATA buoys are not yet available, we are unable to estimate horizontal eddy advection directly. We therefore resort to the method of Baturin and Niiler [1997] and Swenson and Hansen [1999] to first calculate horizontal heat advection as the difference between the total and local time derivatives of mixed layer temperature

$$h(\mathbf{v} \cdot \nabla T) = h \left( \frac{dT}{dt} - \frac{\partial T}{\partial t} \right). \quad (2)$$

The total time derivative is estimated from the SST measurements made by the (quasi-Lagrangian) drifting buoys, while the local derivative is estimated from climatological SST. We then subtract monthly climatological heat advection estimates (discussed previously) from equation (2) in order to estimate eddy heat advection. Estimates on the equator are uncertain owing to poor spatial and temporal drifter coverage and the drifters' tendency to diverge from the equator.

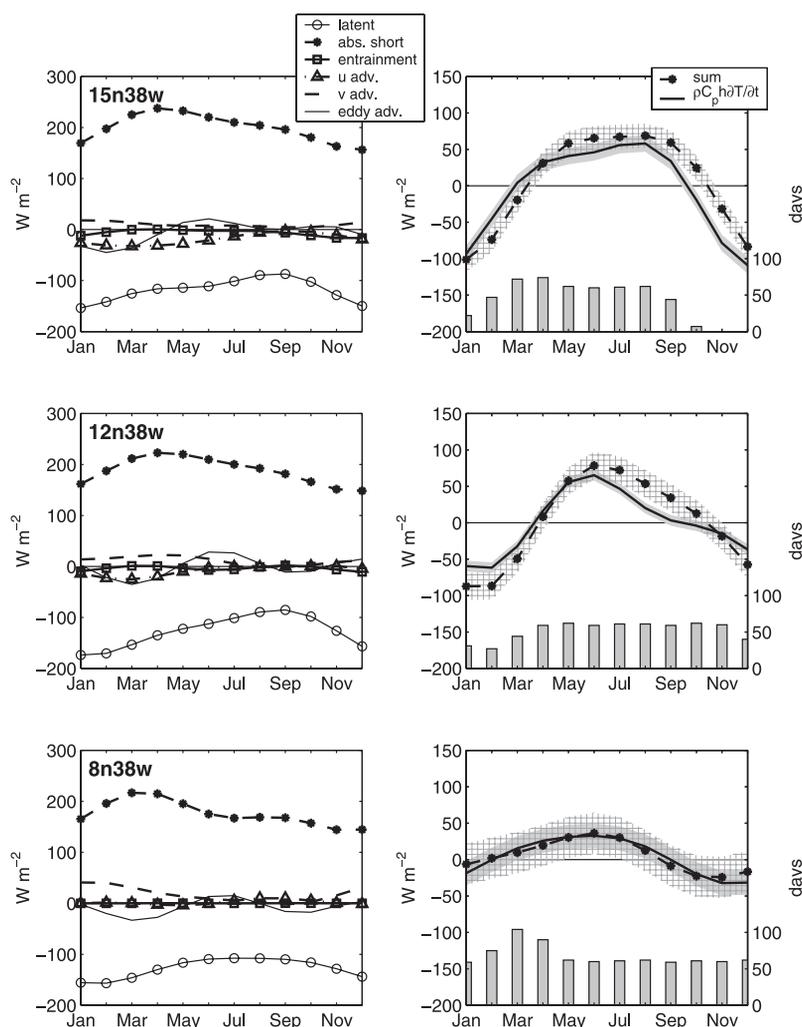
[27] Since we are interested primarily in the seasonal cycle, we eliminate high-frequency variability by fitting the monthly averaged data to annual and semiannual harmonics using least squares (Fourier) analysis. We use the standard deviation from each harmonic fit as an estimate of the uncertainty associated with each term in equation (1). These error estimates account for high-frequency variability (a period  $<6$  months) that our data cannot accurately reproduce. We also anticipate errors resulting from the combination of missing PIRATA data and interannual variability. In particular, it is possible for climatologies of different PIRATA variables to incorporate data from different time periods (see Figure 2). For this reason we display the number of daily PIRATA measurements that go into each climatological monthly estimate at each location. Monthly estimates of each term in equation (1) use a maximum of  $\sim 120$  individual daily measurements (since most PIRATA moorings have been operational for  $\sim 4$  years). Low counts ( $\ll 120$ ) indicate high uncertainty.

### 3. Results

[28] In this section we examine the balance of terms in equation (1) along meridional sections in the west ( $38^\circ\text{W}$ ) and east ( $10^\circ\text{W}$ ) and a zonal section along the equator (see Figure 1 for buoy locations). Examination of the mixed layer heat balance along  $38^\circ\text{W}$  (Figure 4) shows, as expected, that away from the equator surface, heat flux balances local storage. Both absorbed shortwave radiation and latent heat loss terms have strong seasonal cycles at all mooring locations along  $38^\circ\text{W}$  (see Table 1), and both are dominated by the annual harmonic. Absorbed solar radiation is highest in boreal spring, when the ITCZ is displaced to the south and the mixed layer is deep. Since surface shortwave radiation and mixed layer depth are very nearly in phase at  $8^\circ\text{N}$  and  $12^\circ\text{N}$  along  $38^\circ\text{W}$ , the amplitude of absorbed shortwave radiation at these locations is even larger than that of surface shortwave (see Figure 3).

[29] The seasonal cycle of latent heat loss along  $38^\circ\text{W}$  lags absorbed solar radiation by  $\sim 5$  months. Latent heat loss is at a maximum during boreal winter when the northeast trade winds are strong and relative humidity is low, while absorbed solar radiation is near its minimum. This phase relationship results in a strong seasonal cycle of local heat storage at  $15^\circ\text{N}$  and  $12^\circ\text{N}$ , with maximum storage in boreal summer. At  $8^\circ\text{N}$  the intense northeast trade winds of mid-winter induce meridional advection of heat that acts to warm the mixed layer and reduce the seasonal changes in SST. In comparison to the equatorial locations (see Figure 5), eddy advection is less significant at all locations along  $38^\circ\text{W}$ . This is consistent with the results of Hansen and Paul [1984] in the Pacific, which show that eddy heat transport is significant only within  $\pm 5^\circ$  of the equator.

[30] Our estimates of heat storage resulting from the sum of terms generally agree with the observed seasonal cycle of heat storage at all locations along  $38^\circ\text{W}$  (to within  $\pm \sim 50$   $\text{W m}^{-2}$ ). The agreement is least good at  $15^\circ\text{N}$  in fall, when the number of observations is very low. We attribute the excess heating during boreal fall and excess cooling during spring at this location to inaccurate representations of latent heat loss due to a lack of PIRATA surface meteorological data. The discrepancies at  $15^\circ\text{N}$ ,  $38^\circ\text{W}$  are reduced when Com-



**Figure 4.** (left) Individual contributions to the heat balance (equation (1)) in the form of latent heat flux, absorbed shortwave radiation, entrainment, mean zonal and meridional heat advection, and eddy heat advection (defined as total advection minus climatological monthly advection). Plots in Figure 4 (left) show least squares fits of mean plus annual and semiannual harmonics to monthly data. (right) Sum of the terms in Figure 4 (left) (plus longwave and sensible, which have been omitted from Figure 4 (left)) and the actual mixed layer heat storage rate. Shading and cross-hatching in Figure 4 (right) indicate error estimates based on standard deviations of monthly data from least squares harmonic fits. Bars in Figure 4 (right) indicate the number of days in each month for which all PIRATA-based terms in Figure 4 (left) are available (maximum of  $\sim 120$  days for each month, corresponding to  $\sim 4$  years of data: September 1997 to February 2002).

prehensive Ocean-Atmosphere Data Set (COADS) climatological latent heat flux [da Silva *et al.*, 1994] is substituted for PIRATA-based estimates.

[31] Along the equator, absorbed shortwave radiation and zonal advection have significant seasonal variations (Figure 5 and Table 2). The seasonal cycle of absorbed shortwave is enhanced by the fact that mixed layer depth and surface solar radiation vary in phase (the mixed layer is deepest in fall when skies are clear). Seasonal variations of absorbed shortwave radiation are larger in the west owing to larger variations of both mixed layer depth and surface shortwave radiation. Horizontal heat advection has strong seasonal variations along the equator (except at  $10^\circ\text{W}$ ) because of the seasonal variations in currents and stronger horizontal temperature gradients. In contrast, the seasonal cycle of

latent heat loss plays a lesser role. The extremely low values of latent heat loss at  $0^\circ$ ,  $10^\circ\text{W}$  during boreal summer are the result of very high relative humidity (88–90%) and modest wind speed ( $\sim 5 \text{ m s}^{-1}$ ). The high values of humidity are possibly the result of a very short data record at this location (see Figure 2). Most of the monthly averaged latent heat values during boreal summer contain data from only the year 2000, and it appears that relative humidity was anomalously high during that year (presumably associated with anomalously low SST). Our subsequent discussion identifies this factor as one of several potential causes of the large discrepancy between the sum of terms and heat storage rate at this location during boreal summer.

[32] Examination of the mixed layer heat budget at  $0^\circ$ ,  $35^\circ\text{W}$  reveals strong seasonal variations in zonal advection,

**Table 1.** Amplitude and Phase of Annual and Semiannual Cosine Harmonics (With Respect to 1 January) and Annual Mean of Terms in the Mixed Layer Heat Balance

	Annual Mean, W m <sup>-2</sup>	Annual Amplitude, W m <sup>-2</sup>	Annual Phase, months	Semiannual Amplitude, W m <sup>-2</sup>	Semiannual Phase, months
<i>15°N, 38°W</i>					
Latent	-120	30	8	10	3
Surface shortwave	210	40	6	20	3
Absorbed shortwave <sup>a</sup>	200	40	5	10	3
Entrainment	-10	10	5	0	3
Zonal advection	-20	20	9	0	3
Meridional advection	10	10	2	0	1
Eddy advection	-10	20	8	20	5
Heat storage	-10	80	6	30	3
Mixed layer depth <sup>b</sup>	60	20	2	10	1
<i>12°N, 38°W</i>					
Latent <sup>a</sup>	-130	40	8	10	4
Surface shortwave	210	30	6	10	3
Absorbed shortwave	190	30	5	10	3
Entrainment	0	0	7	10	3
Zonal advection	-10	10	9	0	6
Meridional advection	10	10	3	0	5
Eddy advection	0	10	8	20	0
Heat storage	0	50	7	20	5
Mixed layer depth <sup>b</sup>	40	20	4	0	2
<i>8°N, 38°W</i>					
Latent	-130	30	8	10	5
Surface shortwave	200	20	4	20	3
Absorbed shortwave <sup>a</sup>	180	30	4	20	3
Entrainment	0	0	7	0	4
Zonal advection	0	0	9	0	2
Meridional advection	20	20	2	10	1
Eddy advection	-10	10	8	20	0
Heat storage	0	30	6	0	2
Mixed layer depth <sup>b</sup>	40	20	4	0	3

<sup>a</sup>Explains the largest amount of variance at this location.

<sup>b</sup>Annual mean, annual amplitude, and semiannual amplitude values for mixed layer depth are given in meters.

with a period of maximum cooling in late boreal summer and weaker values during winter and spring. The phase and large amplitude of this term are explained by the fact that the zonal surface currents and temperature gradient vary in phase and are strongest during boreal summer when the equatorial cold tongue and the westward South Equatorial Current are well developed (see Figure 1). Warming associated with horizontal eddy heat advection is also greatest during boreal summer and fall, when tropical instability waves are present in the western basin. The strength of the equatorial cold tongue and westward surface currents decreases throughout boreal fall and winter, resulting in decreased cooling by mean and eddy advection during this period.

[33] Absorbed shortwave radiation also has significant seasonal variability at 0°, 35°W. The pronounced maximum in boreal fall coincides with clear skies and a deep mixed layer, while the minimum in spring results from cloudiness associated with the ITCZ and a shallow mixed layer. Despite a deep mixed layer in boreal fall, entrainment reaches a maximum during this period. Upward velocity at the base of the mixed layer at this time is supported by a decrease in the intensity of the westward South Equatorial Current in the eastern and central basin (resulting in an increase in zonal mass divergence) and by a strengthening of the easterly trade winds (resulting in an increase in

meridional mass divergence). As expected, latent heat loss undergoes weak seasonal changes at 35°W associated with small variations in near-surface relative humidity.

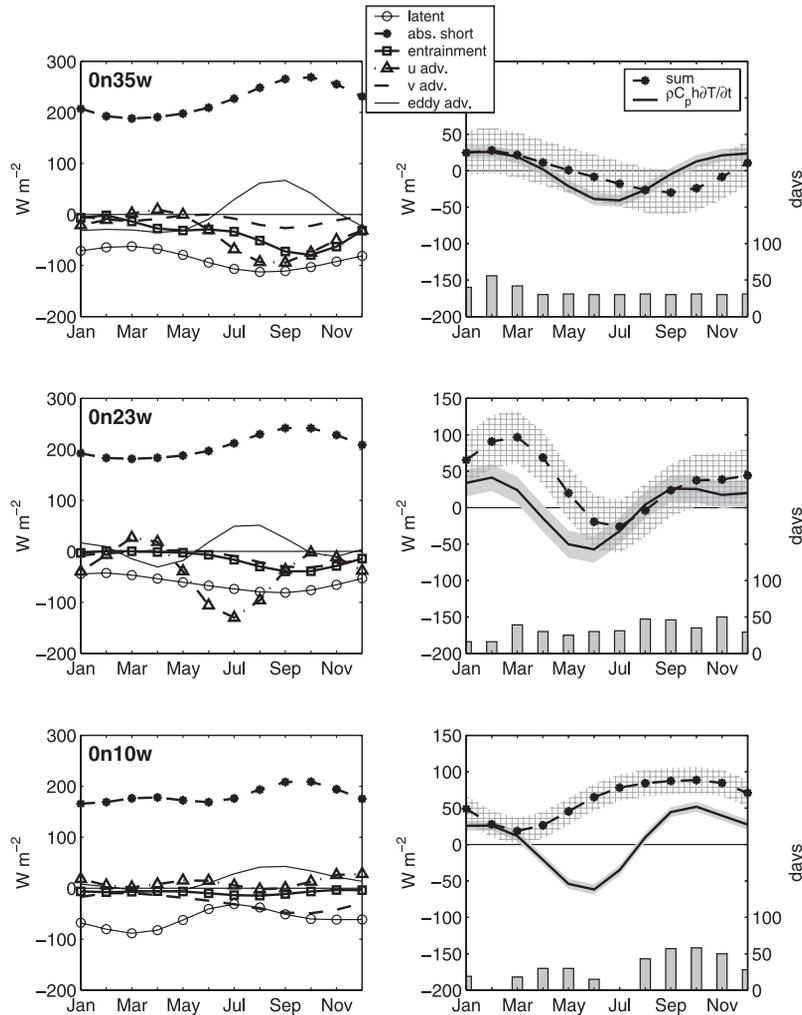
[34] The mixed layer heat balance at 0°, 23°W shares many similarities with the one at 35°W. The maximum in zonal heat advection at 0°, 23°W occurs in boreal summer (about a month earlier than at 35°W) and is stronger than the corresponding maximum at 35°W. This increased amplitude is explained by stronger westward surface currents and sharper zonal temperature gradients at 23°W owing to its position nearly midway between the center of the cold tongue to the east and warm SST to the west (see Figure 1). Meridional heat advection cools the mixed layer at 23°W during the second half of the year owing to northward currents in the presence of a strong northward temperature gradient on the northern edge of the cold tongue.

[35] Absorbed shortwave radiation at 23°W varies with nearly the same phase as at 35°W, but the annual mean and seasonal variations are smaller at 23°W owing to weaker annual mean and seasonal variations in mixed layer depth. Entrainment at 23°W is also less significant than at 35°W, although its seasonal cycle has nearly the same phase and is supported by similar mechanisms (meridional mass divergence resulting from easterly wind stress and zonal mass divergence resulting from a decrease in intensity of the South Equatorial Current in the eastern basin).

[36] In contrast to the conditions at 0°, 23°W and 0°, 35°W, seasonal variations of zonal heat advection are insignificant along the equator at 10°W. This location is near the center of the seasonal cold tongue (see Figure 1) so that westward surface currents and temperature gradients are out of phase. The westward South Equatorial Current is strongest in boreal summer, when the cold tongue is well developed in the eastern basin and the zonal temperature gradient at 10°W is small. The temperature gradient becomes increasingly negative throughout boreal summer and is strongest in October, when the cold tongue is weak and centered east of 10°W. However, by this time, zonal surface currents are weak near 10°W, and zonal heat advection remains small. In contrast, meridional temperature gradients are strong during the second half of the year at 10°W when meridional velocity is  $\sim 10$  cm s<sup>-1</sup>. As a result, meridional heat advection cools the equatorial mixed layer significantly during boreal summer and fall.

[37] Seasonal changes in shortwave radiation have a significant influence on mixed layer heat content at 10°W. In contrast to the other equatorial locations, absorbed shortwave radiation at 10°W has a significant semiannual harmonic, with a weak maximum in boreal spring and a stronger maximum in fall. The maximum in boreal fall results from simultaneous maxima in mixed layer depth and surface shortwave radiation and would be greater if not for the presence of reflective low-level clouds over the cool surface waters in early boreal fall. The weak maximum in boreal spring occurs when the mixed layer is shallow, but surface solar radiation is relatively strong since the ITCZ fails to reach all the way to the equator in the eastern basin.

[38] In general, the agreement between the sum of forcing terms in equation (1) and the actual local heat storage rate is better off the equator along 38°W than along the equator (Figures 4 and 5). Along the equator, agreement is best at



**Figure 5.** As in Figure 4, but for locations along the equator.

$35^{\circ}W$ , where zonal advection, entrainment, and latent heat loss tend to balance absorbed solar radiation and eddy heat advection. The agreement is worst at  $10^{\circ}W$ , where the number of daily measurements is small for most months. The greatest residual in the heat balance at  $10^{\circ}W$  occurs during June and July, when the sum of terms predicts warming of  $\sim 50 W m^{-2}$  while the actual mixed layer heat content decreases by  $\sim 50 W m^{-2}$ . Explanations for this discrepancy include underestimates of entrainment, vertical turbulent diffusion, and latent heat loss. We first consider entrainment.

[39] Our results along the equator suggest that entrainment is most important (in terms of both annual mean and seasonal variations) in the west. We find that entrainment cooling is more significant at  $35^{\circ}W$  than in the central and eastern basin, where the mixed layer is significantly shallower. We anticipate that during boreal summer the easterly component of wind stress causes strong meridional divergence of mixed layer velocity  $\pm 2^{\circ}$  from the equator. It is likely that our mixed layer velocity estimates (which have a  $2^{\circ}$  meridional resolution) do not adequately resolve this process, resulting in entrainment estimates that are too low during boreal summer. Thus one possible explanation for the missing source of cooling at  $10^{\circ}W$  is that we have

underestimated meridional velocity-divergence-induced entrainment. However, an average entrainment velocity of  $\sim 5 \times 10^{-3} cm s^{-1}$  is required to explain the  $\sim 100 W m^{-2}$  discrepancy during boreal summer at  $10^{\circ}W$ . This is nearly an order of magnitude larger than the maximum upwelling rate calculated by *Weingartner and Weisberg* [1991a] ( $0.6 \times 10^{-3} cm s^{-1}$ ), averaged over eight months at  $0^{\circ}, 28^{\circ}W$ , and it suggests that entrainment alone cannot account for the additional cooling at  $0^{\circ}, 10^{\circ}W$ .

[40] It is also possible that seasonal variations of vertical turbulent diffusion at the base of the mixed layer may alter mixed layer heat content at  $0^{\circ}, 10^{\circ}W$ . We have completely neglected this term since we do not have estimates of the turbulent exchange coefficient. *Hayes et al.* [1991] calculated this term explicitly in the eastern equatorial Pacific and found seasonal variations of up to  $150 W m^{-2}$  that were associated with changes in the vertical profiles of temperature and horizontal velocity.

[41] An additional factor accounting for the discrepancy at  $10^{\circ}W$  involves our estimates of latent heat loss. As discussed previously, data from only 1 year (2000) were used in the monthly latent heat flux estimates at this location. During this year, relative humidity was anomalously high, leading to estimates of latent heat that are  $50 W$

**Table 2.** As in Table 1, but for Locations Along the Equator

	Annual Mean, $W m^{-2}$	Annual Amplitude, $W m^{-2}$	Annual Phase, months	Semiannual Amplitude, $W m^{-2}$	Semiannual Phase, months
<i>0°</i> , <i>35°W</i>					
Latent	-90	30	3	0	4
Surface shortwave	230	30	10	10	4
Absorbed shortwave	220	40	10	10	4
Entrainment	-40	30	3	20	1
Zonal advection	-40	50	3	10	5
Meridional advection	-10	10	3	10	0
Eddy advection <sup>a</sup>	0	50	9	20	3
Heat storage	0	30	1	10	4
Mixed layer depth <sup>b</sup>	60	20	10	0	2
<i>0°</i> , <i>23°W</i>					
Latent	-60	20	2	0	1
Surface shortwave	240	20	9	10	4
Absorbed shortwave	210	30	9	10	4
Entrainment	-20	20	3	10	1
Zonal advection <sup>a</sup>	-40	50	1	50	4
Meridional advection	-10	20	4	0	0
Eddy advection	10	20	8	30	1
Heat storage	0	40	0	20	3
Mixed layer depth <sup>b</sup>	30	10	10	0	1
<i>0°N</i> , <i>10°W</i>					
Latent <sup>a</sup>	-60	20	8	10	1
Surface shortwave	220	10	8	10	3
Absorbed shortwave	180	20	10	10	4
Entrainment	-10	0	1	0	5
Zonal advection	10	10	0	10	6
Meridional advection	-30	20	3	0	1
Eddy advection	20	20	9	10	2
Heat storage	10	50	11	20	3
Mixed layer depth <sup>b</sup>	20	10	11	0	4

<sup>a</sup>Explains the largest amount of variance at this location.

<sup>b</sup>Annual mean, annual amplitude, and semiannual amplitude values for mixed layer depth are given in meters.

$m^{-2}$  lower than climatological estimates [da Silva et al., 1994]. Such additional cooling could partially explain the discrepancy at  $10^{\circ}W$ .

[42] Our results at  $0^{\circ}$ ,  $35^{\circ}W$  and  $0^{\circ}$ ,  $23^{\circ}W$  along the equator generally agree with the results of Weingartner and Weisberg [1991b], who analyzed 1 year of upper ocean heat content data on the equator at  $28^{\circ}W$ . They found a balance between meridional eddy advection and vertical and zonal mean flow advection at 10 m. Their results indicate a period of cooling during mid-April through mid-May associated with upwelling-induced entrainment, followed by a period of warming mid-May through mid-July associated with enhanced meridional heat advection from tropical instability waves and cooling due to mean westward advection. At  $23^{\circ}W$  and  $35^{\circ}W$  we also find these terms to be important. During the remainder of the year, Weingartner and Weisberg show that ocean dynamics do not contribute significantly to changes in SST and conclude that vertical diffusion must balance the net surface heat flux. In contrast, we find that entrainment is most important during the second half of the year. Interestingly, Carton and Zhou [1997] found that entrainment associated with meridional velocity divergence is an important source of cooling at these locations, while we find that zonal divergence is most important. The  $25 W m^{-2}$  excess cooling during boreal fall at  $0^{\circ}$ ,  $35^{\circ}W$  (Figure 5) is likely due to an overestimate of zonal temperature advection, which likely overestimates the vertical scale of the South Equatorial Current.

[43] In contrast to the conditions discussed above, south of the equator at  $10^{\circ}W$  the mixed layer depth and surface shortwave radiation vary nearly out of phase. As a result, the amplitude of the seasonal cycle (annual plus semiannual harmonics) of absorbed shortwave radiation at these locations is  $\sim 15 W m^{-2}$  lower than that of surface shortwave. In contrast, latent heat loss has strong seasonal variations at  $6^{\circ}S$  and  $10^{\circ}S$ . Maximum latent heat loss occurs in boreal summer, when the southeast trade winds are strong and relative humidity is low (see Figure 6 and Table 3).

[44] At these locations, near-surface currents are from the northeast throughout the year. Zonal temperature advection is strongest during boreal summer, when the South Equatorial Current advects cool water westward. At  $6^{\circ}S$ , weak meridional advection provides cooling during boreal summer, when the equatorial cold tongue is well developed to the north, and heating during the remainder of the year, when the meridional temperature gradient is reversed. In contrast, at  $10^{\circ}S$  the meridional temperature gradient is northward throughout the year, so meridional advection provides a year-round source of heat.

#### 4. Summary

[45] This paper examines the mixed layer heat budget in the tropical Atlantic, on the basis of measurements from eight PIRATA moorings and a variety of other in situ and satellite sources, in an attempt to explain the strong seasonal cycle of SST. We have followed the formalism of Stevenson and Niiler [1983], which equates seasonal changes in mixed layer heat content to various atmospheric and oceanic forcing mechanisms. This study is similar in spirit to several previous studies in the equatorial Pacific [Hayes et al., 1991; Swenson and Hansen, 1999; Wang and McPhaden, 1999] which use in situ near-surface atmospheric and subsurface oceanographic measurements to estimate directly as many terms as possible in the mixed layer heat budget. Our main results are as follows:

[46] 1. In the western (along  $38^{\circ}W$ ) and eastern (along  $10^{\circ}W$ ) tropical Atlantic, changes in mixed layer heat content are balanced primarily by changes in net surface heat flux (latent heat loss and solar heat gain). As the equator is approached, contributions from horizontal heat advection become increasingly important.

[47] 2. Along the equator in the western basin ( $35^{\circ}W$ ) the seasonal cycles of zonal heat advection (resulting from the seasonally varying flow of the South Equatorial Current), eddy heat advection (presumably associated with tropical instability waves), entrainment (caused by zonal mass divergence), and net surface heat flux all contribute significantly to seasonal SST variability.

[48] 3. The mixed layer heat balance in the central equatorial Atlantic ( $23^{\circ}W$ ) is similar to that in the west ( $35^{\circ}W$ ), with the exception that seasonal variations of latent heat loss and entrainment are significantly smaller and zonal advection is stronger at  $23^{\circ}W$ .

[49] 4. In the eastern equatorial Atlantic ( $10^{\circ}W$ ), cooling from meridional advection and warming from eddy advection tend to balance, so seasonal changes in SST tend to reflect seasonal variations in absorbed shortwave radiation. Entrainment and unresolved vertical diffusion may explain anomalous cooling, as discussed below.

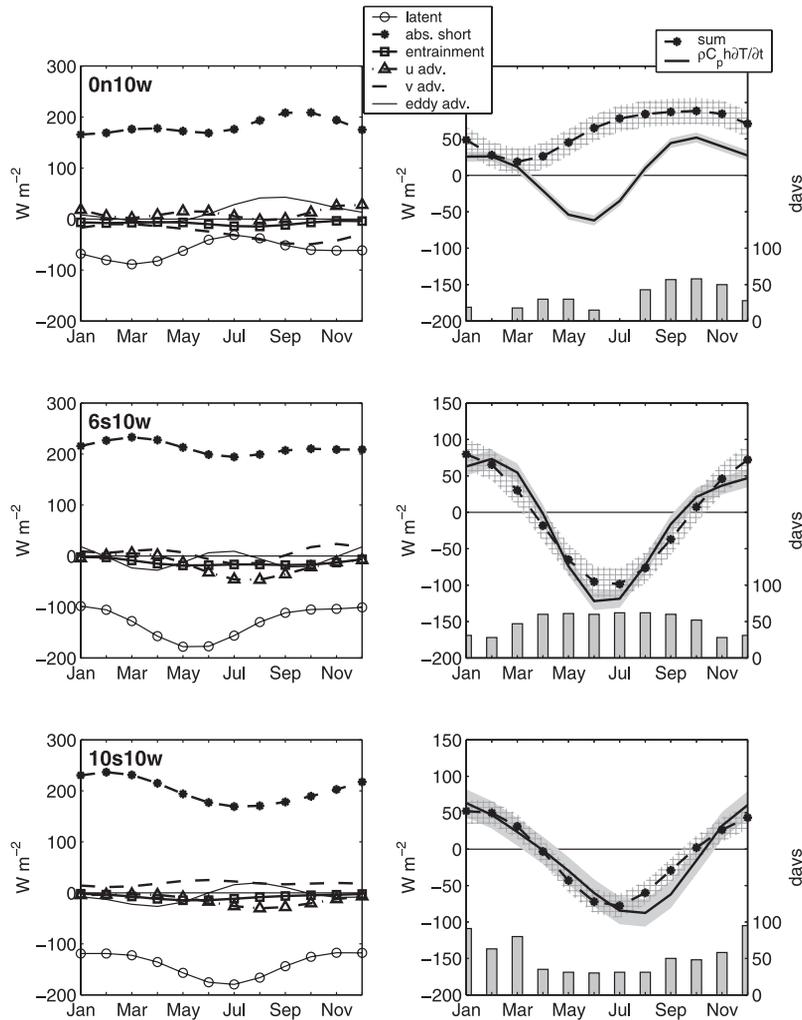


Figure 6. As in Figure 4, but for locations along  $10^{\circ}\text{W}$ .

[50] Our observation-based results generally agree with the modeling results of *Carton and Zhou* [1997]. Their results show that solar heating is most important south of  $5^{\circ}\text{S}$  and north of  $10^{\circ}\text{N}$ . We also find that absorbed solar radiation is important in this region, but that latent heat loss also plays an important role, especially in the eastern basin south of  $5^{\circ}\text{S}$ . Along the equator they show that ocean dynamics have a large influence on seasonal SST variability. They find that boreal summer cooling is the result of zonal divergence of mass east of  $20^{\circ}\text{W}$  and meridional divergence to the west of  $30^{\circ}\text{W}$ . We find that both meridional and zonal divergence are important in the west ( $0^{\circ}$ ,  $35^{\circ}\text{W}$ ), while meridional divergence (we suspect) is important in the east ( $0^{\circ}$ ,  $10^{\circ}\text{W}$ ).

[51] Our results are also similar to those of the observation-based analysis of *Weingartner and Weisberg* [1991a]. They show that heating from tropical instability waves and cooling from zonal advection oppose each other during boreal summer at  $28^{\circ}\text{W}$  on the equator. Our results along the equator at  $23^{\circ}\text{W}$  and  $35^{\circ}\text{W}$  also show strong contributions from zonal and eddy advection during this time period. However, they indicate that upwelling between 10 and 75 m is most intense during boreal spring, while our results show

that entrainment is important at  $23^{\circ}\text{W}$  and  $35^{\circ}\text{W}$  only during boreal fall.

[52] The largest errors in our analysis include a missing source of cooling (up to  $100\text{ W m}^{-2}$ ) during May–July at  $0^{\circ}$ ,  $10^{\circ}\text{W}$  and during boreal spring at  $0^{\circ}$ ,  $23^{\circ}\text{W}$ . Vertical entrainment is the most likely explanation for the additional cooling needed during May–July at  $0^{\circ}$ ,  $10^{\circ}\text{W}$ . We expect that meridional divergence of mass within the mixed layer induces upwelling at the base of the mixed layer during May–July, when the southeast trade winds are strong on the equator. Unfortunately, the meridional resolution of our mixed layer depth and meridional velocity estimates leads to an underestimation of this process. We also suspect that vertical diffusion at the base of the mixed layer contributes significant cooling at this location.

[53] We believe that the discrepancy along the equator at  $23^{\circ}\text{W}$  can also be explained by inaccurate representation of entrainment. At  $28^{\circ}\text{W}$ , *Weingartner and Weisberg* [1991a] find that upwelling within the upper 75 m is strongest during boreal spring and weak during boreal fall. Although their analysis was limited to only 1 year, their estimates of meridional velocity divergence (calculated with a resolution of  $0.5^{\circ}$ ), and hence upwelling, are likely more accurate than

**Table 3.** As in Table 1, but for Locations Along 10°W

	Annual Mean, W m <sup>-2</sup>	Annual Amplitude, W m <sup>-2</sup>	Annual Phase, months	Semiannual Amplitude, W m <sup>-2</sup>	Semiannual Phase, months
<i>0°, 10°W</i>					
Latent <sup>a</sup>	-60	20	8	10	1
Surface shortwave	220	10	8	10	3
Absorbed shortwave	180	20	10	10	4
Entrainment	-10	0	1	0	5
Zonal advection	10	10	0	10	6
Meridional advection	-30	20	3	0	1
Eddy advection	20	20	9	10	2
Heat storage	10	50	11	20	3
Mixed layer depth <sup>b</sup>	20	10	11	0	4
<i>6°S, 10°W</i>					
Latent <sup>a</sup>	-130	40	0	10	2
Shortwave	230	20	2	10	3
Absorbed shortwave	210	10	2	10	3
Entrainment	-10	10	1	0	1
Zonal advection	-20	20	2	10	4
Meridional advection	10	10	1	10	5
Eddy advection	-10	10	0	20	1
Heat storage	-10	90	1	30	3
Mixed layer depth <sup>b</sup>	50	10	9	10	6
<i>10°S, 10°W</i>					
Latent <sup>a</sup>	-140	30	1	10	4
Shortwave	210	40	2	10	3
Absorbed shortwave	200	30	2	0	3
Entrainment	-10	10	0	0	2
Zonal advection	-10	10	2	0	5
Meridional advection	20	0	7	0	6
Eddy advection	-10	20	9	10	1
Heat storage	-10	70	1	10	5
Mixed layer depth	50	20	9	0	1

<sup>a</sup>Explains the largest amount of variance at this location.

<sup>b</sup>Annual mean, annual amplitude, and semiannual amplitude values for mixed layer depth are given in meters.

ours. We also find that the sum of forcing terms and heat storage rate differ by a substantial amount ( $\sim 50 \text{ W m}^{-2}$ ) during boreal fall at 15°N, 38°W. We attribute these differences to errors in our estimates of latent heat loss, which are highly suspect owing to a lack of data. Concurrent in situ estimates of subsurface velocity and temperature, combined with horizontal gradients of velocity, temperature, and mixed layer depth, are required for a complete analysis of all terms affecting the equatorial mixed layer heat balance.

[54] Despite these limitations and the relatively short mooring records, we have been able to show that the seasonal mixed layer heat balance in the tropical Atlantic is quite complex, with many terms contributing at most locations. To complete our understanding at seasonal periods, additional work is needed to quantify the roles of entrainment and vertical turbulent diffusion in the equatorial heat balance. Already the results provide a base from which interannual and decadal variability, which are both linked to the annual cycle, can be addressed.

[55] **Acknowledgments.** The authors thank Jennifer Donaldson for her significant contributions during the early stages of this study. We also acknowledge two anonymous reviewers who provided valuable critiques of the originally submitted manuscript. This work was supported by NOAA's office of Oceanic and Atmospheric Research and Office of Global Programs. The authors also gratefully acknowledge the support provided by the National Science Foundation (OCE9812404). We are grateful to the Drifter DAC of the GOOS Center at NOAA/AOML for providing the drifter data set. QuikSCAT wind has been obtained from the NASA/NOAA-sponsored

system Seaflux at JPL through the courtesy of W. Timothy Liu and Wenqing Tang.

## References

- Ando, K., and M. J. McPhaden, Variability of surface layer hydrography in the tropical Pacific Ocean, *J. Geophys. Res.*, *102*, 23,063–23,078, 1997.
- Baturin, N. G., and P. P. Niiler, Effects of instability waves in the mixed layer of the equatorial Pacific, *J. Geophys. Res.*, *102*, 27,771–27,793, 1997.
- Carton, J. A., and Z. X. Zhou, Annual cycle of sea surface temperature in the tropical Atlantic Ocean, *J. Geophys. Res.*, *102*, 27,813–27,824, 1997.
- da Silva, A., A. C. Young-Molling, and S. Levitus, *Atlas of Surface Marine Data 1994*, vol. 1, *Algorithms and Procedures*, NOAA Atlas NESDIS, vol. 6, U.S. Dep. of Commer., Washington, D. C., 1994.
- DeWitt, D. G., and E. K. Schneider, The processes determining the annual cycle of equatorial sea surface temperature: A coupled general circulation model perspective, *Mon. Weather Rev.*, *127*, 381–395, 1999.
- Enfield, D. B., Zonal and seasonal variations of the near-surface heat balance of the equatorial Pacific Ocean, *J. Phys. Oceanogr.*, *16*, 1038–1054, 1986.
- Esbensen, S. K., and M. J. McPhaden, Enhancement of tropical ocean evaporation and sensible heat flux by atmospheric mesoscale systems, *J. Clim.*, *9*, 2307–2325, 1996.
- Fairall, C. W., E. F. Bradley, D. P. Rogers, J. B. Edson, and G. S. Young, Bulk parameterization of air-sea fluxes for Tropical Ocean-Global Atmosphere Coupled-Ocean Atmosphere Response Experiment, *J. Geophys. Res.*, *101*, 3747–3764, 1996.
- Freitag, H. P., Y. Feng, L. J. Mangum, M. P. McPhaden, J. Neander, and L. D. Stratton, Calibration procedures and instrumental accuracy estimates of TAO temperature, relative humidity and radiation measurements, NOAA Tech. Memo. ERL PMEL-104, 32 pp., Pac. Mar. Environ. Lab., National Oceanic and Atmospheric Administration, Seattle, Wash., 1994.
- Freitag, H. P., M. E. McCarty, C. Nosse, R. Lukas, M. J. McPhaden, and M. F. Cronin, COARE Seacat data: Calibrations and quality control procedures, NOAA Tech. Memo. ERL PMEL-115, 89 pp., Pac. Mar. Environ. Lab., National Oceanic and Atmospheric Administration, Seattle, Wash., 1999.
- Freitag, H. P., M. O'Haleck, G. C. Thomas, and M. J. McPhaden, Calibration procedures and instrumental accuracies for ATLAS wind measurements, NOAA Tech. Memo. OAR PMEL-119, 20 pp., Pac. Mar. Environ. Lab., National Oceanic and Atmospheric Administration, Seattle, Wash., 2001.
- Grodsky, S. A., and J. A. Carton, Intense surface currents in the tropical Pacific during 1996–1998, *J. Geophys. Res.*, *106*, 16,673–16,684, 2001.
- Hansen, D. V., and C. A. Paul, Genesis and effects of long waves in the equatorial Pacific, *J. Geophys. Res.*, *89*, 10,431–10,440, 1984.
- Hastenrath, S., Hemispheric asymmetry of oceanic heat budget in the equatorial Atlantic and eastern Pacific, *Tellus*, *29*, 523–529, 1977.
- Hayes, S. P., P. Chang, and M. J. McPhaden, Variability of the sea surface temperature in the eastern equatorial Pacific during 1986–1988, *J. Geophys. Res.*, *96*, 10,553–10,566, 1991.
- Kalnay, E., et al., The NCEP/NCAR 40-year reanalysis project, *Bull. Am. Meteorol. Soc.*, *77*, 437–471, 1996.
- Kessler, W. S., L. M. Rothstein, and D. Chen, The annual cycle of SST in the eastern tropical Pacific, diagnosed in an ocean GCM, *J. Clim.*, *11*, 777–799, 1998.
- Klein, S. A., and D. L. Hartmann, The seasonal cycle of low stratiform clouds, *J. Clim.*, *6*, 1587–1606, 1993.
- Koberle, C., and S. G. H. Philander, On the processes that control seasonal variations of sea surface temperatures in the tropical Pacific Ocean, *Tellus, Ser. A*, *46*, 481–496, 1994.
- Lake, B. J., S. M. Noor, H. P. Freitag, and M. J. McPhaden, Calibration procedures and instrumental accuracy estimates for ATLAS air temperature and relative humidity measurements, NOAA Tech. Memo., Pac. Mar. Environ. Lab., National Oceanic and Atmospheric Administration, Seattle, Wash., in press, 2003.
- Levitus, S., and T. P. Boyer, *World Ocean Atlas 1994*, vol. 4, *Temperature*, NOAA Atlas NESDIS, vol. 4, U.S. Dep. of Commer., Washington, D. C., 1994.
- Medovaya, M., D. E. Waliser, R. A. Weller, and M. J. McPhaden, Assessing ocean buoy shortwave observations using clear-sky model calculations, *J. Geophys. Res.*, *107*(C2), 3014, doi:10.1029/2000JC000558, 2002.
- Merle, J., Seasonal heat budget in the equatorial Atlantic Ocean, *J. Phys. Oceanogr.*, *10*, 464–469, 1980.
- Moisan, J. R., and P. P. Niiler, The seasonal heat budget of the North Pacific: Net heat flux and heat storage rates (1950–1990), *J. Phys. Oceanogr.*, *28*, 401–421, 1998.
- Molinari, R. L., J. F. Festa, and E. Marmolejo, Evolution of sea surface temperature in the tropical Atlantic Ocean during FGGE, 1979, 2. Ocean

- nographic fields and heat balance of the mixed layer, *J. Mar. Res.*, *43*, 67–81, 1985.
- Monterey, G. I., and S. Levitus, *Seasonal Variability of Mixed Layer Depth for the World Ocean, NOAA Atlas NESDIS*, vol. 14, 5 pp., U.S. Dep. of Commer., Washington, D. C., 1997.
- Morel, A., Optical modeling of the upper ocean in relation to its biogenous matter content (case 1 waters), *J. Geophys. Res.*, *93*, 10,749–10,768, 1988.
- Philander, S. G. H., D. Gu, D. Halpern, G. Lambert, N. C. Lau, T. Li, and R. C. Pacanowski, Why the ITCZ is mostly north of the equator, *J. Clim.*, *9*, 2958–2972, 1996.
- Reynolds, R. W., and T. M. Smith, Improved global sea surface temperature analyses using optimum interpolation, *J. Clim.*, *7*, 929–948, 1994.
- Richardson, P. L., and G. Reverdin, Seasonal cycle of velocity in the Atlantic North Equatorial Countercurrent as measured by surface drifters, current meters, and ship drifts, *J. Geophys. Res.*, *92*, 3691–3708, 1987.
- Servain, J., A. J. Busalacchi, M. J. McPhaden, A. D. Moura, G. Reverdin, M. Vianna, and S. E. Zebiak, A Pilot Research Moored Array in the Tropical Atlantic (PIRATA), *Bull. Am. Meteorol. Soc.*, *79*, 2019–2031, 1998.
- Stevenson, J. W., and P. P. Niiler, Upper ocean heat budget during the Hawaii-to-Tahiti shuttle experiment, *J. Phys. Oceanogr.*, *13*, 1894–1907, 1983.
- Swenson, M. S., and D. V. Hansen, Tropical Pacific ocean mixed layer heat budget: The Pacific cold tongue, *J. Phys. Oceanogr.*, *29*, 69–81, 1999.
- Wang, C. Z., and D. B. Enfield, The tropical Western Hemisphere warm pool, *Geophys. Res. Lett.*, *28*, 1635–1638, 2001.
- Wang, W. M., and M. J. McPhaden, The surface-layer heat balance in the equatorial Pacific Ocean. Part I: Mean seasonal cycle, *J. Phys. Oceanogr.*, *29*, 1812–1831, 1999.
- Weingartner, T. J., and R. H. Weisberg, On the annual cycle of equatorial upwelling in the central Atlantic Ocean, *J. Phys. Oceanogr.*, *21*, 68–82, 1991a.
- Weingartner, T. J., and R. H. Weisberg, A description of the annual cycle in sea surface temperature and upper ocean heat in the equatorial Atlantic, *J. Phys. Oceanogr.*, *21*, 83–96, 1991b.
- Weisberg, R. H., and T. J. Weingartner, Instability waves in the equatorial Atlantic Ocean, *J. Phys. Oceanogr.*, *18*, 1641–1657, 1988.
- White, W. B., Design of a global observing system for gyre-scale upper ocean temperature variability, *Prog. Oceanogr.*, *36*, 169–217, 1995.
- Wyrtki, K., An estimate of equatorial upwelling in the Pacific, *J. Phys. Oceanogr.*, *11*, 1205–1214, 1981.

---

J. A. Carton, G. R. Foltz, and S. A. Grodsky, Department of Meteorology, Computer and Space Sciences Building, University of Maryland, College Park, MD 20742, USA. (carton@atmos.umd.edu; gregoryf@atmos.umd.edu; senya@atmos.umd.edu)

M. J. McPhaden, Pacific Marine Environmental Laboratory, NOAA, 7600 Sand Point Way NE, Seattle, WA 98115, USA. (mcphaden@pmel.noaa.gov)