Surface Layer Heat Balance in the Eastern Equatorial Pacific Ocean on Interannual Time Scales: Influence of Local versus Remote Wind Forcing*

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ABSTRACT
The authors use a new and novel heat balance formalism for the upper 50 m of the Niño-3 region (5°N–5°S, 90°W–150°W) to investigate the oceanographic processes underlying interannual sea surface temperature (SST) variations in the eastern equatorial Pacific. The focus is on a better understanding of the relationship between local and remote atmospheric forcing in generating SST anomalies associated with El Niño–Southern Oscillation (ENSO) events. The heat balance analysis indicates that heat advection across 50-m depth and across 150°W are the important oceanic mechanisms responsible for temperature variations with the former being dominant. On the other hand, net surface heat flux adjusted for penetrative radiation damps SST. Jointly, these processes can explain most of interannual variations in temperature tendency averaged over the Niño-3 region. Decomposition of vertical advection across the bottom indicates that the mean seasonal advection of anomalous temperature (the so-called thermocline feedback) dominates and is highly correlated with 20°C isotherm depth variations, which are mainly forced by remote winds in the western and central equatorial Pacific. Temperature advection by anomalous vertical velocity (the “Ekman feedback”), which is highly correlated with local zonal wind stress variations, is smaller with an amplitude of about 40% on average of remotely forced vertical heat advection. These results support those of recent empirical and modeling studies in which local atmospheric forcing, while not dominant, significantly affects ENSO SST variations in the eastern equatorial Pacific.

1. Introduction

Bjerknes (1969) pioneered ENSO theory by proposing a positive air–sea feedback operating during El Niño development. He argued that “A decrease of the equatorial easterlies weakens the equatorial upwelling, thereby the eastern equatorial Pacific becomes warmer and supplies heat also to the atmosphere above it. This lessens the east-west temperature contrast within the Walker Circulation and makes that circulation slow down.” He suggested that main oceanic process involved in the positive air–sea feedback was the variation of equatorial upwelling associated with variations in the easterly trade winds. He hypothesized that it was the local winds in the eastern Pacific that were important, though since then we have learned that it is remote wind forcing from the central and western Pacific that is dominant in the evolution of El Niño and La Niña (e.g., Wyrtki 1975; Neelin et al. 1998). Nonetheless, interest has grown recently in the possible role of local wind forcing in the eastern Pacific because for some ENSO events, it is a nonnegligible contributor to the development of ENSO SST anomalies. Since Bjerknes (1969), many studies have been carried out to identify physical processes underlying SST variations in the equatorial Pacific Ocean. Some of these were observationally based empirical studies (e.g., Wyrtki 1981; Stevenson and Niiler 1983; Bryden and Brady 1985; Hayes et al. 1991a; Wang and McPhaden 1999, 2000, 2001,
hereafter WM), while others were modeling efforts using various models and configurations (e.g., Philander 1987; Seager et al. 1988; Chen et al. 1994; Kessler et al. 1998; Borovikov et al. 2001; Vialard et al. 2001; Lee et al. 2004; Kim et al. 2007). Although general consistency can be identified among different studies, there are still points of disagreement. This is partially because of the specific heat budget formalisms used, the spatial scales on which the processes are examined, and the differences in model forcing fields and model parameterizations of unresolved physical processes. Moreover, because of limitations in available observations, many observation-based heat balance analyses cannot achieve closure of heat budget without making assumptions that are difficult to validate (e.g., Qiu 2002; Wang and McPhaden 1999). For ocean general circulation models (OGCMs) that do not have an explicit mixed layer model embedded in them, it is not a very easy task to close the heat budget either (e.g., Vialard et al. 2001; Kim et al. 2006). Below we provide a brief review of specific issues relevant to our study.

Vialard et al. (2001) developed a local mixed layer heat balance formalism for their OGCM to study oceanic mechanisms underlying SST variations during the 1997/98 El Niño event. The mixed layer depth (MLD) in their study is defined diagnostically and thus varies both spatially and temporally. By spatial averaging of local balance over the Niño-3 region, they find subsurface processes (including both entrainment and vertical turbulent diffusion) are the essential driving mechanism for temperature tendency, while surface heat flux exerts damping effect. Wang and McPhaden (2000, 2001) used a local mixed layer heat balance formalism to study interannual heat balance variability, mainly based on Tropical Atmosphere Ocean (TAO) buoy observation at four fixed locations (i.e., 165°E, and 170°, 140°, and 110°W). In Wang and McPhaden’s analyses, vertical heat flux out of the base of the mixed layer (including vertical turbulent diffusion and entrainment) was derived as a residual \( Q_m \), which also contained unresolved processes, plus computational and observational errors. Using 140° and 110°W as two representative locations for the eastern equatorial Pacific, Wang and McPhaden found that vertical heat flux out of the base of the mixed layer and adjusted net surface heat flux were the two largest terms. Both vertical heat flux and zonal heat advection were positively correlated with temperature tendency, indicating their role as driving mechanisms. Both adjusted net surface heat flux and meridional heat advection [mainly due to the high-frequency part associated with tropical instability waves (TIWs)] exert damping effects on temperature change. Lee et al. (2004) developed a novel integrative heat balance formalism for the upper 50 m of the Niño-3 region. Their analysis from 1997 to 2000 indicated both vertical heat advection across 50-m depth and zonal heat advection across 150°W are important, with the former being dominant. Meridional advection was generally smaller than zonal advection. Kim et al. (2007) developed a similar formalism following Lee et al. (2004) to analyze an ocean data assimilation product for the period of 1993–2003, but with the MLD determined diagnostically based on density criteria other than constant at 50 m. It was found that the mixed layer temperature tendency over the Niño-3 domain could be explained mainly by subsurface processes (including both entrainment and vertical turbulent diffusion), horizontal advection, and surface heat flux, with the former two processes as driving mechanisms and the latter as an opposing mechanism. Zonal advection was primarily caused by large-scale processes (rather than high-frequency TIW activity). Zonal advection across the west section at 150°W contributed more than that across the east section at 90°W. Vertical heat advection and vertical turbulent mixing at the base of the mixed layer contributed almost equally to the temperature tendency over the domain.

On the basis of historical empirical and modeling studies, including those mentioned earlier, it is generally agreed that there are two essential physical processes for interannual SST variations in the equatorial Pacific: zonal advection of mean temperature gradient by anomalous current (the so-called zonal advective feedback) and vertical advection by mean upwelling of anomalous subsurface temperature (the so-called thermocline feedback; e.g., Jin and An 1999: An and Jin 2001). The zonal advective feedback mainly operates in the central Pacific, depending on the anomalous zonal current variations and mean negative zonal SST gradient along the equator. The zonal advective feedback is also central to the advective–reflective oscillator paradigm (Picaut et al. 1997). For the eastern equatorial Pacific, thermocline displacements in the presence of mean upwelling alter the temperature of the water upwelled to the surface (Zebiak and Cane 1987; Battisti and Hirst 1989). This thermocline feedback results primarily from remotely wind-forced thermocline displacements associated with equatorial wave dynamics and is a key element of most ENSO theories. These conclusions were consistent with a recent theoretical study by Jin et al. (2006). Within the simple recharge oscillator framework, they proposed a “Bjerknes (BJ) stability index” for ENSO evolution. Using this index they identified thermocline and zonal advection as the main driving mechanisms for ENSO evolution, and thermal damping is the principal source of decay. They also described a secondary “Ekman” feedback due to local wind forcing in the eastern Pacific as a driving mechanism for ENSO.
The effects of local wind forcing in the eastern Pacific on the evolution of ENSO has been noted in some previous modeling and observational studies (e.g., McPhaden 1999; Vintzileos et al. 2005; Vecchi and Harrison 2006). Recent empirical analysis by Zhang and McPhaden (2006, hereafter ZM06) for example pointed out that local wind variations have non-negligible effects on interannual SST variations in the eastern equatorial Pacific. Through regression analysis, they found that a zonal wind stress ($\tau_x$) anomaly of 0.01 N m$^{-2}$ leads to approximately a 1°C SST anomaly over the Niño-3 region ($5^\circ$N–$5^\circ$S, 90$^\circ$W–150$^\circ$W). Wind stress anomalies on the order of 0.01 N m$^{-2}$ occurred in the eastern equatorial Pacific during the 1982/83 and 1997/98 El Niños, accounting for about one-third of the maximum SST anomaly during these events. ZM06’s strictly empirical analysis was later verified in follow-up numerical OGCM sensitivity tests (Zhang and McPhaden 2008, hereafter ZM08). Their analysis also indicated that in the Niño-3 region, a $\tau_x$ anomaly of 0.01 N m$^{-2}$ leads to about 1°C SST anomaly and that air–sea heat fluxes tend to damp interannual SST anomalies generated by other physical processes at a rate of about 40 W m$^{-2}$ (°C)$^{-1}$.

In this study, we use the OGCM described in ZM08 to examine the surface layer heat balance for interannual SST variations in the eastern equatorial Pacific. The purpose is to apply the new and novel methodology of Lee et al. (2004) and Kim et al. (2007) to identify the relative importance of local and remote wind forcing on the processes that affect SST variability in the eastern equatorial Pacific. This work expands on previous model heat balance studies in the eastern Pacific by more quantitatively distinguishing between Ekman and thermocline feedbacks. This paper is organized as follows. In section 2, data sources and processing techniques are introduced. The OGCM and its configurations are briefly introduced in section 3. Then its performance is validated against available observations in section 4. The mixed layer heat budget formalism is discussed in section 5. Then we examine the effects of various physical processes on the mixed layer temperature balance for the Niño-3 region as a whole. Lastly, a discussion and conclusions are presented in sections 6 and 7, respectively.

2. Datasets

In this section we introduce datasets used in this study. Some of these datasets are used to force the model, while others are used to validate the model. The 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) provides products with global coverage, available from September 1957 to August 2002 (available online at http://data-portal.ecmwf.int/data/d/era40_daily/). Momentum, heat, and freshwater flux forcing for the OGCM simulation are derived from the original 6-hourly ERA-40 reanalysis dataset.

The SST used in this study is derived from blended in situ and satellite analyses, which are available every week from November 1981 to present at 1° latitude × 1° longitude resolution (Reynolds et al. 2002). SST anomalies are computed relative to a monthly climatology for a 1971–2000 base period constructed following the method of Reynolds and Smith (1995) (more information can be found online at http://www.cpc.noaa.gov/products/predictions/30day/SSTs/explanation.html). For the model run, a reconstructed monthly SST dataset (Smith and Reynolds 2003) with 2° latitude × 2° longitude resolution is used in the heat flux correction from January 1979 to November 1981. Hereafter, this mixed SST dataset for the period from January 1979 to present is referred as the Reynolds SST dataset.

Monthly-mean potential temperature and salinity data are derived from the World Ocean Atlas 1998 (WOA98; NOAA 2009). The horizontal resolution of this dataset is 1° latitude × 1° longitude. There are 33 standard vertical levels covering the depth range from 0 to 5500 m with fine resolution in the upper ocean. This dataset is used in the initialization of the OGCM and also the relaxation of temperature and salinity in sponge layers during the OGCM runs.

A subsurface temperature dataset is derived from an ocean subsurface analysis system (Smith 1995a,b) by the Centre for Australian Weather and Climate Research (CAWCR). This dataset has 14 levels in the upper 500 m with 1° latitude × 2° longitude horizontal resolution and monthly temporal resolution starting from 1980. It combines XBT, Argo, and mooring data by optimal interpolation. The TAO array (Hayes et al. 1991b; McPhaden et al. 1998) provides most of the subsurface temperature information for the CAWCR dataset for the equatorial band between 5°S and 5°N during this period (Smith and Meyers 1996). The CAWCR subsurface temperature is used to derive the 20°C isotherm depth (hereafter $Z_{20}$) by linear interpolation in the vertical direction. Anomalies of $Z_{20}$ and subsurface temperature were calculated relative to a monthly climatology with a base period from 1980 to 2000. These data were used for model validation.

The Ocean Surface Current Analyses–Real time (OSCAR) Project (available online at http://www.oscar.noaa.gov/) provides gridded surface current derived from satellite altimeter and scatterometer measurements (Lagerloef et al. 1999). The 1° latitude × 1° longitude gridded dataset is available every 5 days since 1993. These analyzed currents are compared with model current simulations in section 4.
3. Model configuration and simulation

The OGCM and its setup are described in detail by ZM08 (see their section 2), so only key attributes are discussed here. The OGCM is a modified version of the Princeton Ocean Model (POM), which was developed by Princeton University (Mellor 2004). POM is a free-surface general circulation model, with a special topography following sigma vertical coordinate. The horizontal grid uses curvilinear orthogonal coordinates and an “Arakawa C” differencing scheme. An implicit differencing scheme is used for the vertical direction, which eliminates time constraints and permits fine vertical resolution. The model domain covers the tropical and subtropical Pacific basin (50°N–40°S, 120°–70°W) with constant 1° zonal resolution. Meridional resolution is $\frac{1}{3}^\circ$ for the tropical band between 10°N and 10°S, then linearly increases poleward from 10°N (10°S) up to the model boundaries, where the meridional resolution is about 2°. Sponge layers are applied to the region within 10° of the northern and southern boundaries to reduce the effects of the artificial boundaries. Within these sponge layers, horizontal viscosity and diffusion are artificially increased, and temperature and salinity are relaxed to seasonal climatologies from the WOA98 dataset (section 2). For the present study, the ocean is assumed to have a flat bottom of 4000-m depth. There are 29 sigma layers in the vertical, with 10 uniform layers in the upper 100 m, 9 layers in the upper 100–400 m, and additional 10 layers below that down to 4000 m. For the inner domain—that is, outside of the sponge layers—the coefficients of horizontal eddy viscosity and diffusivity are both set to 1500 m$^2$ s$^{-1}$. Vertical mixing coefficients are derived from a Richardson number-dependent formula (Pacanowski and Philander 1981). The surface forcing terms consist of momentum, heat, and freshwater fluxes derived from the 6-h ERA-40 reanalysis dataset. A correction term is also added to the prescribed net heat flux, which is a common practice for multiyear forced ocean general circulation modeling (e.g., Vintzileos et al. 2005; Kim et al. 2007; ZM08). The correction term relaxes model SST to observed weekly Reynolds SST with relaxation coefficient derived from a 3-yr climatology of ECMWF analyses (Barnier et al. 1995), which varies spatially and temporally between about $-30$ and $-50$ W m$^{-2}$ °C$^{-1}$ for the tropical Pacific. Surface freshwater flux—that is, evaporation minus precipitation ($E - P$)—provides surface forcing for the salinity equation. Because of uncertainties in $E - P$, a weak relaxation with a time scale of 90 days is added to prevent salinity from drifting too far from the WOA98 seasonal climatology for the several upper layers.

After a spinup run from rest with monthly climatological forcing for 10 yr, the OGCM is integrated from January 1979 to August 2002. The OGCM produces output every 2.5 days and thus can resolve high-frequency signals such as TIWs. To ensure an exact balance, heat budget terms (section 5) are also saved every 2.5 days. Monthly means, monthly climatological seasonal cycles, and interannual anomalies are derived from the model output. Monthly anomalies are smoothed twice with a 5-month running-mean filter to emphasize interannual variations. For comparison, monthly anomalies from validation datasets are derived with reference to their own climatologies and are then smoothed in time twice with a 5-month running-mean filter.

4. Evaluation of OGCM performance

Before we can use the OGCM as a tool to diagnose surface layer heat balance, it is necessary to first validate the model performance by comparing model output to available observational datasets. ZM08 have compared the Niño-3 SST simulated by the model with that from the Reynolds SST dataset, mean vertical profiles of model zonal velocity at three representative locations (170°, 140°, and 110°W) on the equator with TAO buoy measurements, and model sea surface height with merged Ocean Topography Experiment (TOPEX)/Poseidon and Jason-I altimeter measurements (see ZM08, their section 3.1). Here we show additional comparisons to support the reliability of the OGCM.

The classical equatorial surface current system is well defined in the OGCM simulation (Fig. 1a). From north to south, there are the westward North Equatorial Current (NEC), eastward North Equatorial Countercurrent (NECC) and westward South Equatorial Current (SEC). The SEC is clearly separated into a northern and a southern branch with a minimum on the equator. Annual mean surface currents at 15-m depth from the OGCM (Fig. 1a) compare favorably with those from the OSCAR dataset (Fig. 1b), which is derived from satellite altimeter and scatterometer data (section 2). Nonetheless, differences can also be found (Fig. 1c). In the western equatorial Pacific, the OGCM produces a weaker NECC. In the eastern equatorial Pacific, a stronger SEC can be found on the equator and just south of the equator, while a weaker SEC can be found just north of the equator (2°–5°N). Some of these differences may result from errors in the OSCAR analysis and therefore represent an upper bound on model error.

The OGCM also simulates the seasonal variations of surface zonal currents along the equator well. The westward propagation of zonal surface currents agrees with that from the OSCAR dataset in both magnitude and phase (not shown). This westward propagation is
consistent with TAO and ship drift data as described in Yu and McPhaden (1999), who explained the propagation by invoking wind-forced equatorial wave dynamics.

The vertical distribution of zonal velocity along the equator from the OGCM simulation (Fig. 2a) is compared with that from observations (Johnson et al. 2002, hereafter J02; Fig. 2b). The OGCM reproduces the main vertical structure of zonal velocity along the equator: a westward SEC at the surface and an eastward equatorial undercurrent (EUC) between 50 and 300 m, with an eastward-shoaling EUC core. However, there are several obvious differences between the observations and the OGCM simulation. For example, the EUC is weaker than in the observations (Fig. 2c). Also, the vertical range of the EUC gradually increases westward (Fig. 2a), but it is almost constant in J02’s analysis (Fig. 2b). Some of these differences could be due to uncertainties in the observational analysis; alternatively, decadal variations in the EUC might account for some of the differences since the OGCM simulation is from 1979 to 2002, while J02 is mainly based on measurements from the 1990s. However, it is more likely that the differences primarily reflect shortcomings in the model.

ZM06 showed that interannual SST variations in the central and eastern equatorial Pacific can be represented by a linear combination $\tau^x$ and $Z_{20}$ [see ZM06, their Eq. (5)]. Thus, the interannual variations of $\tau^x$, $Z_{20}$, and SST are compared to observations here. The same wind product is used in ZM06’s empirical analysis and in forcing the OGCM (Fig. 3a). High $\tau^x$ standard deviation can be found between 160$^\circ$E and 160$^\circ$W in the central and western tropical Pacific. The standard deviation of interannual $Z_{20}$ from the model (Fig. 3b) is similar to that derived from the CAWCR monthly subsurface temperature dataset (Fig. 3c). There are three centers of high standard deviation: one in the eastern equatorial Pacific along the equator and two off the equator in the western Pacific along the western boundary. The OGCM produces a relatively high standard deviation of $Z_{20}$ extending poleward from the equator along the eastern boundary, which is related to poleward propagation of Kelvin waves originating at the equator. This feature is
absent in the CAWCR dataset, which may be related to the relatively poor data coverage along the coast. The standard deviation of SST from the OGCM (Fig. 3d) is very similar to that from the Reynolds SST dataset (Fig. 3e), as would be expected from our relaxation constraint. High standard deviations exist east of the date line along the equator and between 0° and 10°S off the South American coast. The same multiple regression model developed by ZM06 was also applied to the model output, producing similar regression slopes as in the ZM06’s empirical analysis (not shown).

Both $Z_{20}$ and SST from the OGCM simulation show weaker interannual variability than those derived from observations. This could be associated with a somewhat diffuse thermocline produced by our OGCM—a common weakness for many OGCMs that can lead to weaker interannual variability (Meehl et al. 2001). A lack of an Indonesian Throughflow in our model may partially contribute to the relatively weak thermocline variability, since it has been found that blockage of the Indonesian straits reduces model-simulated seasonal-to-interannual thermocline depth variations in the central to eastern equatorial Pacific (Lee et al. 2002).

The focus of the current study is the interannual surface layer heat balance in the eastern equatorial Pacific. However, higher-frequency variability cannot be ignored, because it may also exert influence on lower-frequency surface layer temperature variations through nonlinear interactions. For example, the dominant source of intraseasonal variability is TIWs, which may significantly affect the local mixed layer heat budget (e.g., Philander et al. 1986; Menkes et al. 2006). TIWs were first discovered by Legeckis (1977) from infrared satellite images and then later verified by in situ data (e.g., Halpern et al. 1988) and numerical models (e.g., Philander et al. 1986; Masina and Philander 1999). They have wavelengths of about 1000 km and periods of about 20–35 days. TIWs gain energy from both kinetic and potential energy of the mean flow (Philander et al. 1986).

A snapshot from the OGCM simulation shows instantaneous temperature and horizontal current vectors at 15 m over the eastern equatorial Pacific (Fig. 4a). The TIWs are most pronounced in the north flank of the cold tongue, with wave lengths of about 1000 km, consistent with observations (e.g., Legeckis 1977). Another snapshot at the same time from the OGCM simulation shows...
instantaneous vertical velocity and horizontal current vector (Fig. 4b). Downwelling with clockwise horizontal circulation takes place over cold crests and upwelling happens over warm troughs, which reduces the potential energy of the mean flow and feeds TIWs (Philander et al. 1986).

The longitudinal range for TIWs in the model simulation is from about 170°E to 120°W. This range is shifted slightly westward compared to the observations, which show TIWs in the region from 160°E to 100°W (Contreras 2002). This westward shift is consistent with the westward shift of the cold tongue in the model (not shown), which is a common problem for OGCMs (e.g., Stockdale et al. 1998).

In summary, the validation of the model in this section and in ZM08 supports its usage to diagnose the physical processes underlying surface layer heat balance in the equatorial Pacific.

5. Heat budget of the Niño-3 region

We are interested in the large-scale heat balance of the mixed layer averaged over the eastern Pacific rather than at discrete locations. Thus, we focus on the Niño-3 region and fix the bottom boundary at a constant depth of 50 m (Fig. 5). This depth is close to the mean MLD for the eastern equatorial Pacific and is commonly used in intermediate models (Zebiak and Cane 1987; Battisti 1988). For the ease of description, the six surfaces defining the box are called section W for the section along 150°W, section E along 90°W, section N along 5°N, section S along 5°S, section B for the bottom section at 50-m depth, and section U at sea surface. For the surface section U, net surface heat flux accounts for the heat exchange between ocean and atmosphere. For all the other sections, there are two physical processes for heat exchange, that is, advection and diffusion. It is straightforward and unambiguous to calculate diffusion across one section, because both temperature gradient and diffusion coefficients can easily be obtained from model output. This is not the case for heat advection though. Traditionally, heat advection (which is also sometimes referred to as advective heat transport) across section A is calculated as
\[
\rho C_p \int_A (\bm{V} \cdot \bm{n}) T \, dS,
\]

where \( \rho C_p \) is the volumetric heat capacity of seawater, \((\bm{V} \cdot \bm{n})\) is the normal velocity, and \(T\) is the temperature at the section. However, the physical meaning behind this expression is ambiguous, since the volume transport \( \int_A (\bm{V} \cdot \bm{n}) \, dS \) across that section is not usually zero.

Actually, Eq. (1) calculates “temperature transport” rather than heat transport, which moreover depends on the definition of temperature scale. For example, the Celsius and Kelvin scales produce different transport values even if the flow condition is exactly the same (Hall and Bryden 1982).

To explicitly evaluate the effect of heat transport across each section on the temperature change of the

FIG. 4. A snapshot of TIW activity: (a) horizontal current (vectors) and temperature (shades and contours) at 15-m depth and (b) vertical velocity (shades and contours) and horizontal current (vectors) at 15-m depth.
Here, we calculate the heat transport using a modified scheme following Lee et al. (2004),

$$\rho C_p \int_A (\mathbf{V} \cdot \mathbf{n})(T - T_{\text{ave}}) dS,$$

where $T_{\text{ave}}$ is the volume-averaged temperature of the study domain, which is time-dependent. By using the average temperature as a reference, this modified scheme does not require zero net mass flux at each section and does not depend on the reference of zero temperature (Fig. 5).

The physical implications of the two schemes represented by (1) and (2) are quite different. For example, if we assume a uniform inflow condition at a section $A$ (see case I in Table 1), then the traditional scheme (1) produces positive values, which indicate a warming effect due to the inflow. In contrast, the calculation by the modified scheme (2) depends on the relationship between the temperature at this section $T$ and the domain-averaged temperature $T_{\text{ave}}$, that is, warming (cooling) if the former is warmer (cooler) than the latter. It has clear physical meaning, that is, inflow of warmer (cooler) water warms up (cools down) the domain. Similarly, for the outflow condition (case II in Table 1), the traditional scheme indicates cooling, while the modified scheme indicates warming (cooling) if $T$ is cooler (warmer) than $T_{\text{ave}}$. In the other example illustrated in Table 1 (case III), when $T$ is equal to $T_{\text{ave}}$, the calculation by the traditional scheme depends on the flow condition and is usually not zero, while it is zero in the modified scheme regardless of the flow conditions. Physically, this means that moving the water of the same temperature into or out of the domain does not change the domain's average temperature.

As we shall see later, utilizing this modified scheme provides new insights into the processes controlling Niño-3 SSTs on ENSO time scales.

The heat balance over the upper 50 m of the Niño-3 region can be obtained by the integration of the temperature equation over the whole domain (Lee et al. 2004),

$$Q_t = Q_{\text{advB}} + Q_{\text{advW}} + Q_{\text{advE}} + Q_{\text{advS}} + Q_{\text{advN}} + Q_{\text{diffB}} + Q_{\text{diffW}} + Q_{\text{diffE}} + Q_{\text{diffS}} + Q_{\text{surf}},$$

with this modified scheme. For instance, the heat advection across the bottom section $B$ ($Q_{\text{advB}}$) can be expressed as

$$\int_B w_{50m} (T_{50m} - T_{\text{ave}}) dx dy = \int_B w_{50m} \delta T dx dy,$$

where $w_{50m}$ is vertical velocity (temperature) at the depth of 50 m and $\delta T = T_{50m} - T_{\text{ave}}$, is the vertical temperature difference. As we shall see later, utilizing this modified scheme provides new insights into the processes controlling Niño-3 SSTs on ENSO time scales.

With an interannual standard deviation of 13.4 W m$^{-2}$, $Q_{\text{surf}}$ is uncorrelated with the temperature tendency term (Fig. 6a; Table 2), while heat advection through $Q_{\text{advB}}$ and $Q_{\text{advN}}$ are negligible most of the time (Fig. 6b).

Term $Q_{\text{advB}}$ dominates on interannual time scales with an interannual standard deviation of 9.9 W m$^{-2}$ (Fig. 6a; Table 2); $Q_{\text{advB}}$ generally warms (cools) the Niño-3 region during El Niño (La Niña) events, as evident from the positive correlation with the temperature tendency term (Fig. 6a; Table 2). The second largest term is $Q_{\text{advW}}$, with an interannual standard deviation of 4.7 W m$^{-2}$. It is also positively correlated with the tendency term but with a variance that is about half of $Q_{\text{advB}}$ (Fig. 6a). Term $Q_{\text{advN}}$ also contributes weakly, while heat advection through $Q_{\text{advE}}$ and $Q_{\text{advS}}$ is negligible most of the time (Fig. 6b).

With an interannual standard deviation of 13.4 W m$^{-2}$, $Q_{\text{surf}}$ is uncorrelated with the temperature tendency term at zero lag (Table 2) but significantly anticorrelated with SST ($-0.81$, exceeding 90% significance; Fig. 6b).

1 For the ease of comparison, hereafter heat budget components across all surfaces are normalized by the surface area of the Niño-3 region and are thus expressed in equivalent of average surface heat flux with units of watts per meters squared.
indicating that the adjusted net heat flux acts to damp SST variability on interannual time scales. The combination of the two largest heat advection terms $Q_{\text{advB}}$ and $Q_{\text{advW}}$ with $Q_{\text{surf}}$ accounts for most of the temperature variations over the box (Fig. 6c; Table 2). Lee et al. (2004) drew a similar conclusion based on their numerical simulation though only for the period of 1997–2000.

To further investigate the two largest advection terms $Q_{\text{advB}}$ and $Q_{\text{advW}}$, we decompose them using a separation of variables technique. For example, $Q_{\text{advB}}$ in Eq. (3) can be further decomposed as

$$Q_{\text{advB}} = \int_B \int_B \int_B \left( \overline{w_{50m}} \frac{\partial T}{\partial y} + \overline{w'_{50m}} \frac{\partial T'}{\partial y} \right) \, dx \, dy$$

where overbars denote monthly climatological mean seasonal cycles and primes denote anomalies from these cycles. Equation (5) gives four components of $Q_{\text{advB}}$ that is, 1) mean seasonal vertical advection of mean seasonal temperature, 2) mean seasonal advection of anomalous temperature, 3) anomalous advection of mean seasonal temperature, and 4) anomalous advection of anomalous temperature.

Averaging Eq. (5) over the mean seasonal cycle reveals that the first component dominates the annual mean state of $Q_{\text{advB}}$ with a value of $-49.5 \ \text{W m}^{-2}$, while the fourth nonlinear component contributes weak warming $5.4 \ \text{W m}^{-2}$. Contributions from the second and third components are negligible by definition. Consequently, the annual mean of $Q_{\text{advB}}$ is about $-43.6 \ \text{W m}^{-2}$ (refer to Table 2 for mean values of heat budget terms). Using a box model ($5^\circ \text{S} - 5^\circ \text{N}, 170^\circ \text{E} - 100^\circ \text{W}; 0-50 \ \text{m}$), Wyrtki (1981) obtained a similar magnitude (about $-53 \ \text{W m}^{-2}$) of mean heat advection at 50-m depth by inferring an upwelling rate of 50 Sv (1 Sv = $10^6 \ \text{m}^3 \ \text{s}^{-1}$) and a mean temperature difference of 3°C between upwelled water and the water leaving in the Ekman layer.

Averaged over the mean seasonal cycle, vertical diffusion $Q_{\text{diff}}$ contributes cooling of $-11.5 \ \text{W m}^{-2}$, which is about one-quarter of $Q_{\text{advB}}$, while the interannual standard deviation of $Q_{\text{diff}}$ is about one-sixth of $Q_{\text{advB}}$ (Table 2). The relative small magnitude of vertical diffusion for both the mean and interannual time scales is mostly related to our choice of fixed MLD. If the heat balance were formulated relative to a time-dependent MLD rather than a fixed depth (refer to the appendix), then vertical heat diffusion would contribute as much as the vertical heat advection and entrainment on interannual time scales (not shown).

Consistent with the thermocline feedback mentioned in the introduction, the mean seasonal vertical advection of anomalous temperature dominates interannual anomalies of $Q_{\text{advB}}$ (Fig. 7a; Table 3). High correlation (0.91, exceeding 90% significance) can be found between this component and $Z_{20}$ anomalies averaged over the Niño-3 region (Fig. 7b; Table 3). The fluctuations of $Z_{20}$ cause differences between the volume-averaged temperature and temperature at the bottom of the Niño-3 box at 50 m. A deeper thermocline is usually associated with a weaker temperature difference, which leads to an anomalous warming tendency. Consequently, we designate this component as “$Q_{\text{T}}$” to indicate its close connection with the thermocline feedback. Also consistent with ZM06’s empirical analysis, anomalous vertical advection of mean seasonal temperature is positively correlated (0.78, exceeding 90% significance) with $\tau'$ averaged over the Niño-3 region (Fig. 7c). A relaxation of the trade winds (positive $\tau'$ anomalies) leads to a reduction in local equatorial upwelling, which causes an anomalous warming tendency. This is the oceanic process originally proposed by Bjerknes (1969) as the cause of warming during El Niño, though we now know it is the

<table>
<thead>
<tr>
<th>Case</th>
<th>Conditions</th>
<th>$\rho C_p \int A (\mathbf{V} \cdot \mathbf{n}) T , dS$ [Eq. (1)]</th>
<th>$\rho C_p \int A (\mathbf{V} \cdot \mathbf{n}) (T - T_{\text{ave}}) , dS$ [Eq. (2)]</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>Inflow</td>
<td>“Warming”</td>
<td>Warming if $T &gt; T_{\text{ave}}$</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Cooling if $T &lt; T_{\text{ave}}$</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td>Warming if $T &lt; T_{\text{ave}}$</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>$0$</td>
</tr>
<tr>
<td>II</td>
<td>Outflow</td>
<td>“Cooling”</td>
<td></td>
</tr>
<tr>
<td>III</td>
<td>$T = T_{\text{ave}}$</td>
<td>“Warming” for inflow</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>“Cooling” for outflow</td>
</tr>
</tbody>
</table>

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<td>Cooling if $T &lt; T_{\text{ave}}$</td>
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<td>Cooling if $T &gt; T_{\text{ave}}$</td>
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<td></td>
<td>Warming if $T &lt; T_{\text{ave}}$</td>
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<tr>
<td></td>
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<td>$0$</td>
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<td></td>
</tr>
<tr>
<td>III</td>
<td>$T = T_{\text{ave}}$</td>
<td>“Warming” for inflow</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>“Cooling” for outflow</td>
</tr>
</tbody>
</table>

where overbars denote mean state of $Q_{\text{advB}}$ with a value of $-49.5 \ \text{W m}^{-2}$, while the fourth nonlinear component contributes weak warming $5.4 \ \text{W m}^{-2}$. Contributions from the second and third components are negligible by definition. Consequently, the annual mean of $Q_{\text{advB}}$ is about $-43.6 \ \text{W m}^{-2}$ (refer to Table 2 for mean values of heat budget terms). Using a box model ($5^\circ \text{S} - 5^\circ \text{N}, 170^\circ \text{E} - 100^\circ \text{W}; 0-50 \ \text{m}$), Wyrtki (1981) obtained a similar magnitude (about $-53 \ \text{W m}^{-2}$) of mean heat advection at 50-m depth by inferring an upwelling rate of 50 Sv (1 Sv = $10^6 \ \text{m}^3 \ \text{s}^{-1}$) and a mean temperature difference of 3°C between upwelled water and the water leaving in the Ekman layer.

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$$Q_{\text{advB}} = \int_B \int_B \int_B \overline{w_{50m}} \frac{\partial T}{\partial y} \, dx \, dy$$

$$= \int_B \int_B \int_B \overline{w_{50m}} \frac{\partial T}{\partial y} + \overline{w'_{50m}} \frac{\partial T'}{\partial y} + \overline{w'_{50m}} \delta T + \overline{w'_{50m}} \delta T' \, dx \, dy,$$ (5)
remote forcing that is more important. Jin et al. (2006) proposed the name “Ekman feedback” for the SST changes linked to local wind changes in the eastern Pacific. Similarly, we designate this component as “$Q_{E K}$” to indicate its close connection with local wind variations through the Ekman feedback. Comparing the interannual standard deviations of $Q_{E K}$ and $Q_{T C}$, the magnitude of the local Ekman feedback is on average about 40% of the remote thermocline feedback in terms of vertical heat advection in the Niño-3 region (Table 3). The $Q_{E K}$ is not significantly correlated with either $Q_{a d v B}$ or $Q_{I}$ because it is not dominant in either. The Ekman feedback mechanism does, however, account for significant SST variability that is not associated with the thermocline feedback, consistent with ZM06’s empirical analysis.

The nonlinear vertical advection term $Q_{N L}$ is strongly anticorrelated with the Niño-3 SST (correlation coefficient $-0.71$, exceeding 90% significance level; Fig. 7d),
also weakly anticorrelated with both total $Q_{advB}$ and tendency $Q_t$ (Fig. 7a; Table 3). For example, there is anomalous cooling related to this term during most of 1997 and then anomalous warming in most of 1998. Thus, $Q_{NL}$ acts as a damping term (Fig. 7d) to reduce interannual temperature anomalies.

Lee et al. (2004) also reached similar conclusions regarding heat advection across the bottom section based on their heat budget analysis over the Niño-3 region. However, they did not decompose total vertical advection into its components and thus did not discuss the relative magnitude of remote versus local processes in the Niño-3 SST balance.

The same decomposition technique shows that anomalous zonal advection of mean seasonal temperature is the dominant component of $Q_{advW}$ (Fig. 8; Table 4). This indicates that the difference between volume-averaged temperature and temperature at section W does not vary significantly on interannual time scales. Instead, fluctuations of zonal transport across the section W are responsible for heat transport variations through this section. Positive transport leads to more advective warming since the mean temperature difference is usually positive (i.e., section W is usually warmer than the average temperature of the Niño-3 box). This is in good agreement with previous observational results (McPhaden and Picaut 1990; Picaut et al. 1996), the advective feedback mechanism described by Picaut et al. (1997), and the OGCM simulation by Lee et al. (2004).

6. Discussion

a. Comparison of ZM06’s empirical analysis

In ZM06’s empirical analysis (see their section 4), a multiple linear regression model was developed based on a simplified mixed layer temperature equation. Two vertical advection processes (mean vertical advection of anomalous temperature gradient and anomalous vertical advection of mean temperature gradient, respectively) and one thermal damping term are considered. Based on dynamical arguments and observational evidence, the mean vertical advection of anomalous temperature gradient was found to be associated with $Z_{20}$ variations, while the anomalous vertical advection of mean temperature gradient was associated with local $\tau^s$ variations.

Our heat budget analysis in the Niño-3 box indicates that vertical heat advection across the bottom interface at 50-m depth is indeed the dominant driving mechanism for temperature change over this domain. The separation of variables technique further verifies that the mean seasonal vertical advection of anomalous temperature dominates interannual anomalies in vertical heat advection. This component is highly correlated with the Niño-3 $Z_{20}$ anomalies. The anomalous vertical advection of mean seasonal temperature is found to be second largest component and its effects cannot be neglected, especially during strong ENSO events. These findings indicate those assumptions behind the simplified surface layer temperature equation are quite reasonable and valid. It may be the reason why such a simple regression model can account for SST variations with only $Z_{20}$ and $\tau^s$ as inputs.

b. TIWs

As mentioned in section 4, as a major source of intraseasonal variability, TIWs may also influence the local mixed layer heat budget (e.g., Philander et al. 1986; Menkes et al. 2006). The strength of TIWs is related to the shear of the zonal equatorial currents, with horizontal shear between westward SEC and eastward NECC and between SEC and EUC both playing important roles. TIWs are most active during boreal fall and winter seasons and disappear during boreal spring and early summer (e.g., Kessler et al. 1998). In the regions on the equator and just north of the equator in the eastern Pacific, TIWs may contribute an equatorward heat flux of O(100 W m$^{-2}$) during active periods (Philander et al. 1986). On interannual time scales, TIWs are almost absent during El Niño events and are stronger than normal.

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TABLE 2. Mean and interannual standard deviation of heat balance terms normalized by the surface area in the Niño-3 region. Also shown is the correlation coefficient of interannual anomalies for each heat balance term with $Q_t$. Estimation uncertainties of mean values and 90% significance levels based on Emery and Thomson (2001) are also shown in parentheses. Correlations in bold are significant at the 90% level.

<table>
<thead>
<tr>
<th>Heat balance term</th>
<th>Mean (W m$^{-2}$)</th>
<th>Interannual standard deviation (W m$^{-2}$)</th>
<th>Correlation with $Q_t$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tendency ($Q_t$)</td>
<td>0.1 ± 2.7</td>
<td>12.2</td>
<td>1.0 (0.19)</td>
</tr>
<tr>
<td>Heat transport across W section ($Q_{advW}$)</td>
<td>−10.3 ± 0.8</td>
<td>4.7</td>
<td>0.47 (0.18)</td>
</tr>
<tr>
<td>Heat transport across E section ($Q_{advE}$)</td>
<td>−1.1 ± 0.5</td>
<td>1.8</td>
<td>0.68 (0.27)</td>
</tr>
<tr>
<td>Heat transport across N section ($Q_{advN}$)</td>
<td>−2.3 ± 0.6</td>
<td>1.9</td>
<td>0.34 (0.22)</td>
</tr>
<tr>
<td>Heat transport across S section ($Q_{advS}$)</td>
<td>−5.7 ± 0.5</td>
<td>3.0</td>
<td>−0.15 (0.14)</td>
</tr>
<tr>
<td>Heat transport across B section ($Q_{advB}$)</td>
<td>−43.6 ± 1.5</td>
<td>9.9</td>
<td>0.74 (0.28)</td>
</tr>
<tr>
<td>Adjusted net surface heat flux ($Q_{surf}$)</td>
<td>73.8 ± 2.3</td>
<td>13.4</td>
<td>0.02 (0.10)</td>
</tr>
<tr>
<td>Heat diffusion across B section ($Q_{diffB}$)</td>
<td>−11.5 ± 0.4</td>
<td>1.6</td>
<td>0.39 (0.22)</td>
</tr>
</tbody>
</table>
FIG. 7. (a) Interannual anomalies of $Q_{advB}$ (W m$^{-2}$) and its components [refer to Eq. (5)]. (b) Comparison of mean vertical advection of anomalous temperature ($Q_{TC}$) with $Z_{20}$ anomalies averaged over the Niño-3 region. (c) Comparison of anomalous vertical advection of mean temperature ($Q_{EK}$) with $\tau^*$ anomalies averaged over the Niño-3 region. (d) Comparison of $Q_{NL}$ with Niño-3 SST.
during La Niña events (e.g., Contreras 2002). Thus, on and just north of the equator in the eastern Pacific, TIW activity can cause anomalous cooling (warming) during El Niño (La Niña) events, damping interannual temperature variations (e.g., Wang and McPhaden 2001). Our OGCM successfully captures this damping effect (Fig. 9).

The effects of TIW activity on the local heat balance are quite sensitive to the region or location of interest. Since the TIWs affect the heat balance mainly by equatorward heat transport, there are counterbalancing effects in neighboring regions to the north and south; that is, anomalous warming between the equator and 2°N is associated with anomalous cooling north of 4°N during La Niña events, and the reverse generally happens during El Niño events. Nonetheless, eddies associated with TIWs mainly act to redistribute heat within the Niño-3 region and thus do significantly affect the total heat balance of the whole domain (e.g., Kim et al. 2007). Consequently, TIWs do not register strongly in our large-scale Niño-3 heat balance analysis.

c. Nonlinear vertical advection term

A significant portion of variability in the nonlinear vertical advection term is intraseasonal, mainly associated with TIW activity. In addition to redistributing heat horizontally as mentioned earlier, TIWs can also affect vertical heat redistribution. As pointed out by Philander et al. (1986) and verified by our OGCM simulation (Fig. 4), the spatial structures of TIWs are organized in the following way: horizontally convergent flow with downwelling in the cold crests and divergent flow with upwelling in the warm troughs. Consequently, the mean effect of TIW-related vertical heat advection is to warm the upper layer averaged over the TIW-active region. During El Niño (La Niña), weaker (stronger) TIW-related vertical heat advection can cause anomalous cooling (warming) in the upper layer, thus working as an extra damping effect for upper-layer temperature variations. In our current study over the Niño-3 region, we found that the nonlinear vertical advection is weakly anticorrelated with temperature tendency but strongly anticorrelated with temperature anomalies (section 5; Table 3). This is in agreement with the finding by Kim et al. (2007), though we did not separate out vertical advection associated with TIWs explicitly as they did.

Nonetheless, Jin et al. (2003) and An and Jin (2004) found that nonlinear vertical advection can contribute to warming during both El Niño and La Niña events based on their local pointwise mixed layer heat balance analysis on the equator. Thus, nonlinear vertical advection

\[
Q_{\text{adv}} = \rho c_p \int_{W} \int_{0}^{B} \tilde{T} dz dy
\]

\[
Q_{\text{EC}} = \rho c_p \int_{W} \int_{0}^{B} \tilde{\nabla} \cdot \tilde{T} dz dy
\]

\[
Q_{\text{NL}} = \rho c_p \int_{W} \int_{0}^{B} \tilde{\nabla} \cdot \tilde{\nabla} \tilde{T} dz dy
\]

Table 3. Various components of vertical heat transport through section B, \(Q_{\text{advB}}\), their interannual standard deviations, correlations with \(Q_{\text{advB}}\), and \(Q_t\) (see section 5). The 90% significance levels based on Emery and Thomson (2001) are shown in parentheses. Correlations in bold are significant at the 90% level.

<table>
<thead>
<tr>
<th>Component</th>
<th>Interannual standard deviation (W m(^{-2}))</th>
<th>Correlation coefficient with (Q_{\text{advB}})</th>
<th>Correlation coefficient with (Q_t)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(Q_{\text{advB}})</td>
<td>9.9</td>
<td>1.0 (0.20)</td>
<td>0.74 (0.28)</td>
</tr>
<tr>
<td>(Q_{\text{EC}})</td>
<td>13.1</td>
<td>0.88 (0.35)</td>
<td>0.62 (0.25)</td>
</tr>
<tr>
<td>(Q_{\text{NL}})</td>
<td>5.8</td>
<td>0.19 (0.23)</td>
<td>0.10 (0.24)</td>
</tr>
</tbody>
</table>

FIG. 8. Term \(Q_{\text{advW}}\) (along 150°W). The expression for \(Q_{\text{advW}} = \rho c_p \int_{W} \int_{0}^{B} \tilde{\nabla} \cdot \tilde{T} dz dy\) is similar to the vertical heat advection expressed in Eq. (5), where \(u(T)\) is zonal velocity (temperature) at section W and \(\delta T_x = T - T_{\text{ave}}\) is zonal temperature difference.
does not damp upper-ocean temperature variations as in our study. They showed that during the development of El Niños, warming from the nonlinear term in the eastern equatorial Pacific is due to the combination of deeper-than-normal $Z_{20}$ and stronger-than-normal upwelling associated with increased trade winds. The reverse generally happens during La Niñas, which also causes warming. Thus, the nonlinear vertical advection term can lead to the asymmetry of ENSO cycles (e.g., maximum positive SST anomalies during El Niño events are larger than maximum negative SST anomalies during La Niña events). To resolve the discrepancy between their results and ours about the role of nonlinear vertical advection on surface ocean temperature variations, we recalculated vertical advection at the bottom of the mixed layer along the equator using a local pointwise mixed layer temperature balance analysis \[ Eq. (A3); \] Fig. 10]. Following An and Jin (2004), the MLD is assumed to be constant as 50 m for this recalculation. Under this assumption, the entrainment velocity is equal to the vertical velocity at 50-m depth, and vertical heat advection is all due to entrainment \[ Eq. (A1) \]. Noticeable effects from the nonlinear term (Fig. 10d) only show up during the strong 1982/83 and 1997/98 El Niño events. Weak anomalous cooling can be found in the eastern equatorial Pacific before El Niño peaks and strong anomalous warming afterward. Thus, this nonlinear vertical advection term based on our local temperature balance analysis still indicates an overall damping effect, consistent with what we found from our integrative heat budget analysis of the Niño-3 region.

The discrepancy could be due to differences in data sources and processing. Jin et al. (2003) and An and Jin (2004) used the monthly National Centers for Environmental Prediction (NCEP) ocean data assimilation product and the Simple Ocean Data Assimilation (SODA) product with vertical velocity derived by continuity. Consequently, TIW effects may not be as well captured as in our study, where vertical velocity is directly simulated and available every 2.5 days. Another possible source of discrepancy is that their analyses are limited to the equator, where TIW effects are a little weaker. Our current study over the Niño-3 region includes the most active zone of TIWs.

7. Conclusions

To describe the effects of various physical processes on interannual variations of the mixed layer temperature in the eastern equatorial Pacific, we carry out a heat budget calculation based on a novel formalism considering the Niño-3 region as a whole. Heat advection

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**Table 4.** As in Table 3, but for zonal heat transport through section W along 150°W $Q_{advW}$.

<table>
<thead>
<tr>
<th>$Q_{advW}$ and its components</th>
<th>Interannual standard deviation (W m$^{-2}$)</th>
<th>Correlation coefficient with $Q_{advW}$</th>
<th>Correlation coefficient with $Q_t$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$Q_{advW} = \rho C_p \int_{0}^{\infty} u \delta T_y \ dy \ dz$</td>
<td>4.7</td>
<td>1.0 (0.22)</td>
<td>0.47 (0.18)</td>
</tr>
<tr>
<td>$\rho C_p \int_{0}^{\infty} u \delta T_y \ dy \ dz$</td>
<td>1.5</td>
<td>0.49 (0.31)</td>
<td>0.03 (0.22)</td>
</tr>
<tr>
<td>$\rho C_p \int_{0}^{\infty} u \delta T_y \ dy \ dz$</td>
<td>4.3</td>
<td>0.94 (0.34)</td>
<td>0.55 (0.20)</td>
</tr>
<tr>
<td>$\rho C_p \int_{0}^{\infty} u \delta T_y \ dy \ dz$</td>
<td>0.9</td>
<td>-0.19 (0.23)</td>
<td>-0.34 (0.24)</td>
</tr>
</tbody>
</table>

**Fig. 9.** Comparison of the Niño-3 SST (°C) and $Q_{eddy}$ [W m$^{-2}$, see Eq. (A2)] averaged over the same latitudinal range (0°–1°N) but different longitudinal ranges.
across the bottom (mainly related to the mean seasonal vertical advection of anomalous temperature) and across the western section along 150°W (mainly related to the anomalous zonal advection of mean seasonal temperature) are important for temperature changes over the Niño-3 box with a constant depth of 50 m, with the heat advection across the bottom section being dominant. Net surface heat flux tends to work against these processes. Jointly, surface heat flux and heat advection across both the bottom and western sections explain most of the variations in temperature tendency averaged over the Niño-3 box (Fig. 6). Decomposition of vertical advection across the bottom reveals physical processes underlying each component. The mean seasonal advection of anomalous temperature—that is, the thermocline feedback $Q_{TC}$—is the dominant component and is highly correlated with $Z_{20}$ variations (Fig. 7b). Our previous study also found that interannual Niño-3 $Z_{20}$ variations were forced mainly by remote wind forcing in the western and central equatorial Pacific (ZM06); this implies that $Q_{TC}$ can be generally regarded as remotely forced. The anomalous vertical advection of mean seasonal temperature, $Q_{EK}$, is associated with local $r^3$ variations (Fig. 7c) and is in magnitude about 40% of remotely forced $Q_{TC}$. This relationship is consistent with the conclusion based on the ZM06’s empirical analyses (cf. Fig. 7 with Fig. 6 of ZM06). Nonlinear vertical heat advection $Q_{NL}$ is highly anticorrelated with Niño-3 SST and acts as a thermal damping term (Fig. 7d). The conclusions derived from the current study are generally in agreement with both observation-based analyses and other numerical simulations (e.g., Wang and McPhaden 2000, 2001; Vialard et al. 2001; Lee et al. 2004; Kim et al. 2007).

The earlier-mentioned conclusions from the OGCM simulation generally validate the physical assumption of the simplified SST equation employed in our previous empirical analyses [see Eq. (1) of ZM06], which serves as the starting point of regression analyses. For the eastern equatorial Pacific, the thermocline feedback is the main mechanism responsible for interannual variations of mixed layer temperature, with zonal advective feedback and anomalous vertical advection related to local wind variations (the Ekman feedback) also playing important secondary roles. Net surface heat flux tends to damp interannual variations produced by other physical processes.

ZM06, ZM08, and the current study point out the important role of local atmospheric forcing (especially

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**FIG. 10.** (a) Interannual anomalies of vertical temperature advection $-\rho C_p w_{50m} \delta T$ averaged over 0.5°S–0.5°N (W m$^{-2}$), where $\delta T = T - T_{cl}$ is vertical temperature difference (refer to the appendix for details). (b) As in (a), but for vertical advection of anomalous temperature by mean vertical velocity $-\rho C_p \bar{w}_{50m} \delta T'$ (where overbar represents climatological mean seasonal cycle and prime denotes anomalies from these cycle). (c) As in (b), but for vertical advection of the mean temperature by anomalous vertical velocity $-\rho C_p w'_{50m} \delta T'$. (d) As in (b), but for vertical advection of anomalous temperature by anomalous vertical velocity $-\rho C_p w'_{50m} \delta T'$. For this calculation, the MLD is assumed to be constant as 50 m (see section 6). Contour interval is 10 W m$^{-2}$.
wind stress) on interannual SST variations in the eastern equatorial Pacific. All these studies are from an oceanic point of view in which the ocean responds, but does not actively feed back, to the atmospheric. However, extensive air–sea interactions are involved in ENSO dynamics, including local air–sea interaction in the eastern tropical Pacific (e.g., Vecchi and Harrison 2006; ZM06). It is also possible that local atmospheric variations (especially wind stress) are not solely decided by the air–sea interaction processes within the Pacific basin. For example, Wang (2006) found coupled patterns between the eastern tropical Pacific, tropical Atlantic, and the Central American continent. As a result, influences from outside of the Pacific basin may also affect the evolution of ENSO events through modifying local atmospheric variations in the eastern equatorial Pacific.

Recent studies point out the differences between “conventional” El Niños and “date line” (Larkin and Harrison 2005) or Modoki El Niños (Ashok et al. 2007). The difference is related to the location of maximum SST anomalies during the peak of El Niño events: eastern Pacific for “conventional” events and central Pacific for “date line” or Modoki El Niño events. It is also found that different teleconnection patterns exist in response to these two different types of El Niños (Kumar et al. 2006; Weng et al. 2007). It is an open question as to what underlying physical processes (oceanic and atmospheric) account for these differences. We note that since the largest differences between the two kinds of El Niño events is in the eastern equatorial Pacific, the methodology developed in this study can be extended to address this question from the perspective of mixed layer dynamical and thermodynamical processes.

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APPENDIX

Local Mixed Layer Heat and Temperature Budgets

Many historical studies of the mixed layer heat balance use a local formalism, so here we also briefly show results using this formalism to validate our OGCM performance. The inclusion of complete thermodynamics in the OGCM allows the heat and temperature budgets of the mixed layer to be diagnosed in detail for comparison with those historical studies. The temperature equation can be integrated locally over the simulated mixed layer to calculate the heat budget (Moisan and Niiler 1998):

$$\rho C_p h \frac{\partial T}{\partial t} = -\rho C_p h \mathbf{v}_a \cdot \nabla T_a - \rho C_p \mathbf{v} \cdot \left( \int_{-h}^{0} \mathbf{v} T \, dz \right)$$

$$- \rho C_p \int_{-h}^{0} D(T) \, dz - \rho C_p W_\nu(T_a - T_{-h})$$

$$- \rho C_p \left( k \frac{\partial T}{\partial z} \right)_{-h} + (Q^\phi + Q^\text{pen}). \quad (A1)$$

The variables are defined as follows: \( \rho C_p \) is the volumetric heat capacity of seawater; \( h \) is the MLD; \( \mathbf{v}_a(T_a) \) is the vertically averaged horizontal velocity vector (temperature) over the MLD; \( \mathbf{v}(T) \) is the deviation from the vertically averaged velocity (temperature); \( D(T) \) is the horizontal diffusion; \( W_\nu \) is entrainment velocity; \( Q^\phi \) is the net surface heat flux, including the heat flux correction term; and \( Q^\text{pen} \) is penetrative shortwave radiation at the base of the mixed layer. The left-hand side of the equation is the mixed layer heat tendency term \( \dot{Q^\text{mld}} \). The right-hand side consists of horizontal advection \( (\overline{Q_a}, \overline{Q_t}) \), vertical temperature and velocity covariance \( Q^\text{cov} \), horizontal diffusion \( Q^\text{hdiff} \), entrainment \( Q^\text{ent} \), vertical turbulent diffusion at the base of the mixed layer \( Q^\text{vdif} \), net surface heat flux \( Q^\phi \) and penetrative shortwave radiation at the base of the mixed layer \( Q^\text{pen} \).

The horizontal advection term can be further separated into low-frequency and high-frequency (i.e., eddy) components—for example, \( \overline{Q_a} = -\rho C_p h \overline{(\mathbf{v}_a \cdot \nabla) T_a + (\mathbf{v}_a \cdot \nabla) T} \), where overbars denote monthly means and primes denote deviations from monthly means. The separation of low-frequency and high-frequency components is somehow arbitrary, but the analysis is not very sensitive to the exact definition of the low-frequency and high-frequency components as long as the cutoff period is about one month. Vialard et al. (2001) used a similar separation period of 30 days. The eddy components of both zonal and meridional heat advection are further grouped with horizontal diffusion \( Q^\text{hdiff} \) and vertical temperature and velocity covariance \( Q^\text{cov} \) to form an “eddy” term \( Q^\text{eddy} \).

After averaging to monthly means, the original equation can be rewritten as (overbars representing monthly means are omitted for simplicity)

$$Q^\text{mld} = Q^\text{ulp} + Q^\text{vlp} + Q^\text{cddy} + Q^\text{ent} + Q^\text{vdif} + (Q^\phi + Q^\text{pen}), \quad (A2)$$

where \( Q^\text{ulp}(Q^\text{vlp}) \) is low-frequency zonal (meridional) heat advection. The last two terms \( Q^\phi \) and \( Q^\text{pen} \) are combined into one “adjusted net surface heat flux” \( Q^\text{oadj} \) following Wang and McPhaden (1999), which represents the net surface heat flux absorbed by the mixed layer.
Divided by $\rho C_p h$, the mixed layer heat budget Eq. (A2) can be converted into the mixed layer temperature budget equation

$$T^\text{mld}_t = T^\text{ulp}_t + T^\text{vlp}_t + T^\text{eddy}_t + T^\text{ent}_t + T^\text{vdif}_t + T^\text{0adj}_t. \quad (A3)$$

From left to right, the local mixed layer temperature budget terms are temperature tendency $T_t$, low-frequency zonal advection $T^\text{ulp}_t$, meridional advection $T^\text{vlp}_t$, eddy flux term $T^\text{eddy}_t$, entrainment $T^\text{ent}_t$, vertical diffusion $T^\text{vdif}_t$, and adjusted net surface heat flux $T^\text{0adj}_t$.

Compared with the mixed layer heat balance, the mixed layer temperature balance provides a similar but distinct perspective. As indicated by Wang and McPhaden (1999), one obvious difference is that mass balance (or equivalently depth) effects on heat balance [Eq. (A2)] are eliminated from the tendency and advection terms in the mixed layer temperature budget (i.e., $T^\text{mld}_t$, $T^\text{ulp}_t$, and $T^\text{vlp}_t$); thus, it is easy to make a site-by-site comparison.

WM applied a similar local formalism to calculate the mixed layer temperature balance in the equatorial Pacific Ocean on both seasonal and interannual time scales. Their analyses were mainly based on observational datasets, which thus serve as a benchmark to evaluate output from this OGCM. Here we choose one representative location ($0^\circ$, $140^\circ W$) to show the mixed layer temperature balance on interannual time scales (Fig. A1). Following

![Fig. A1. Interannual anomalies of the mixed layer temperature budget terms (°C month$^{-1}$) at location $0^\circ$, $140^\circ W$ [refer to Eq. (A3)]: (a) $T^\text{0adj}_t$, (b) $T^\text{ulp}_t$, (c) $T^\text{vlp}_t$, (d) $T^\text{eddy}_t$, and (e) heat flux due to exchange with $T^\text{sub}_t$. The tendency $T^\text{mld}_t$ is plotted in each panel in dash lines. Note that the vertical axes in (a) and (c) are different than those in (b)–(d).](image-url)
WM, the MLD is defined with a temperature criterion given by the depth where the temperature is lower than the SST by 0.5°C (Hayes et al. 1991a). We also group $T_{\text{ent}}$ and $T_{\text{vdp}}$ as one term-temperature change because of the heat exchange with the subsurface ocean beneath the mixed layer, that is, $T_{\text{sub}} = T_{\text{ent}} + T_{\text{vdp}}$. Note that this term was calculated as a residual by WM.

The local mixed layer temperature balance analysis indicates that $T_{\text{osad}}$ and the exchange with the subsurface ocean $T_{\text{sub}}$ are the two dominant terms with the largest interannual variations, and they tend to oppose each other. Both $T_{\text{sub}}$ and $T_{\text{ulp}}$ contribute positively to the mixed layer temperature change, while $T_{\text{osad}}$ and $T_{\text{eddy}}$ work together to damp SST anomalies. The effects of $T_{\text{vdp}}$ are almost negligible, with the exception of periods of warming following the strong 1982/83 and 1997/98 El Niño events. The interannual mixed layer temperature balance at $0^\circ$, 140°W simulated by the OGCM is in good agreement with the observation-based analysis by Wang and McPhaden (2000, 2001).

This local mixed layer heat balance formalism can also be applied with a fixed MLD as in section 5c, where we examine nonlinear terms for comparison with previously published work.

REFERENCES


