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Key Points:

- We describe the representation of heat advection in individual assimilation products
- The recharge phase is characterized by upper-ocean cross-equatorial mass transport
- Differences are small but GECCO does not correctly reproduce the meridional divergence

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Heat advection processes leading to El Niño events as depicted by an ensemble of ocean assimilation products

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Abstract The oscillatory nature of El Niño-Southern Oscillation results from an intricate superposition of near-equilibrium balances and out-of-phase disequilibrium processes between the ocean and the atmosphere. The main objective of the present work is to perform an exhaustive spatiotemporal analysis of the upper ocean heat budget in an ensemble of state-of-the-art ocean assimilation products. We put specific emphasis on the ocean heat advection mechanisms, and their representation in individual ensemble members and in the different stages of the ENSO oscillation leading to EN events. Our analyses consistently show that the initial subsurface warming in the western equatorial Pacific is advected to the central Pacific by the equatorial undercurrent, which, together with the equatorward advection associated with anomalies in both the meridional temperature gradient and circulation at the level of the thermocline, explains the heat buildup in the central Pacific during the recharge phase. We also find that the recharge phase is characterized by an increase of meridional tilting of the thermocline, as well as a southward upper-ocean cross-equatorial mass transport resulting from Ekman-induced anomalous vertical motion in the off-equatorial regions. Although differences between data sets are generally small, and anomalies tend to have the same sign, the differences in the magnitude of the meridional term are seen to be key for explaining the different propagation speed of the subsurface warming tendency along the thermocline. The only exception is GECCO, which does not produce the patterns of meridional surface Ekman divergence (subsurface Sverdrup convergence) in the western and central equatorial Pacific.

1. Introduction

El Niño-Southern Oscillation (ENSO) [Ballester *et al.*, 2016] is the dominant source of interannual variability worldwide and one of the most important modes of variability in the tropical Pacific, with far-reaching influences on the whole climate system [Jin, 1997a, 1997b; Meinen and McPhaden, 2000; Wang, 2002; Brown and Fedorov, 2010; Ballester *et al.*, 2011, 2013; Petrova *et al.*, 2016]. The large amplitude of ENSO anomalies in the tropical Pacific is essentially explained by the strong coupling between the Walker circulation, the zonal gradient of sea surface temperature and the longitudinal tilt of the thermocline (i.e., the so-called Bjerknes feedback [Bjerknes, 1969; Wyrtki, 1975]). These interactions are however modulated by out-of-phase negative feedbacks that bound the amplitude and reverse the sign of interannual anomalies. According to the delayed oscillator theory, this reversal is explained by the differential propagation speed of wind-induced oceanic Kelvin and Rossby waves [Battisti, 1988; Schopf and Suarez, 1988]. While eastward-propagating Kelvin waves quickly deepen the warm ocean layer in the eastern Pacific [Wang, 2002], westward Rossby waves travel at lower speeds, and start to shallow the thermocline only after being reflected as Kelvin waves at the western boundary [Fedorov and Brown, 2009].

Among other models that have been proposed, the recharge oscillator emphasizes the time delay between anomalies in longitudinally averaged thermocline depth and eastern Pacific sea surface temperature [Jin, 1997a, 1997b]. In this conceptual framework, a deeper-than-normal thermocline suppresses the active upwelling in the eastern Pacific and favors the growth of an El Niño (EN) event and the weakening of the trade winds, whose curl generates poleward Sverdrup transport that discharges the heat in the upper ocean and reverses the sign of ENSO [Meinen and McPhaden, 2000]. This theory, therefore, hypothesizes that the oscillatory nature of ENSO results from the balance between equatorial zonal winds and the pressure gradient associated with the equatorial thermocline tilt, as well as from the disequilibrium between the mean

basin-wide thermocline depth and the meridional convergence or divergence of Sverdrup transport due to tropical wind stress curl anomalies [Jin, 1997a, 1997b; Singh and Delcroix, 2013].

Zonal and vertical currents are indeed intimately connected through the energy balance, because a significant fraction of the wind power is converted into buoyancy power [Brown and Fedorov, 2010]. This transfer explains how the energy supplied by enhanced trade winds to the westward South Equatorial Current (SEC) in the central Pacific is converted into downward (upward) mass fluxes in the western (eastern) Pacific that distort local ocean isopycnals and deepen (shoal) the thermocline [Brown et al., 2011]. The increased (decreased) thermocline tilting in the equatorial Pacific associated with stronger (weaker) than normal trade winds induces large cold (warm) anomalies in sea surface temperature in the eastern Pacific, which are amplified by the ocean-atmosphere coupling and extended to the central Pacific by means of zonal advection.

The zonal advective, the Ekman pumping and the thermocline feedbacks have been described as the three major dynamical processes contributing to the amplification of temperature anomalies during the onset of ENSO events [Jin and Neelin, 1993]. Thus, assuming a small initial warm perturbation in the equatorial surface, the coupled system rapidly responds by weakening the trade winds and reducing the zonal tilting of the equatorial thermocline [Jin et al., 2006], which in turn generates anomalous eastward geostrophic currents in the central and eastern Pacific [Santoso et al., 2013]. The upper ocean response is characterized by the decrease of the depth of the thermocline and the generation of anomalous zonal currents in the central and eastern Pacific, which together, amplify the initial anomalies and bring the oscillation to a mature phase [Jin and An, 1999]. These mechanisms also play an important role in the dampening and reversal of ENSO conditions when Sverdrup mass divergence starts to discharge the heat content in the equatorial Pacific after the mature phase of EN conditions.

The main objective of the present work is to perform an exhaustive spatiotemporal analysis of the upper ocean heat budget, emphasizing similarities and differences between the individual members of an ensemble of state-of-the-art ocean assimilation products. We specifically focus on the ocean heat advection mechanisms that characterize the stages of the ENSO oscillation leading to EN events. With a similar approach, we have recently shown how the ensemble mean of these ocean assimilation products can provide important insights into mechanisms that contribute to the heat buildup in the western Pacific 18–24 months before EN events [Ballester et al., 2015]. Here we extend these analyses, and we not only describe the similarities and differences between ensemble members, but also generalize them to the whole ENSO cycle and the entire Pacific ocean basin. The main results emerging from this study can hence be used to understand to what extent results derived from these assimilation products can be used to describe the dynamics of ENSO, such as those in Ballester et al. [2015]. Additionally, they can be taken as a reference for validation and assessment of numerical simulations. While still largely disagreeing in some key dynamical processes, because of the large differences in their underlying models, assimilation techniques and assimilated observations [Ray et al., 2015], these products provide the best and most complete spatiotemporal picture of the ocean subsurface available to date. Indeed, all the data sets used in this study correctly capture all the EN and La Niña (LN) events, and they only differ in some of the processes leading to their mature phase, as analyzed and discussed below.

After discussing our methodology (section 2), we use the ensemble of assimilation products to describe the transitions that characterize the swing between phases of the oscillation, from a climatological neutral base state (section 3) to the generation of a subsurface warm buildup in the western Pacific (section 4), the recharged phase in basin-wide equatorial heat content (section 5) and the onset and mature phases of EN (section 6). Discussion and summary are provided in sections 7 and 8, respectively.

2. Methods

The onset of EN events is characterized by an initial subsurface heat buildup in the western Pacific, the subsequent eastward movement of the accumulated warm waters along the equatorial thermocline (i.e., recharge mode in the central Pacific) and the final rapid amplification of temperature anomalies in the eastern Pacific due to the coupled ocean-atmosphere Bjerknes feedback [Ballester et al., 2015]. The present article describes the role of heat advection in each of these three stages of the oscillation before the mature phase of EN events. To this aim, we analyze the different terms of the temperature tendency equation,

which links the potential temperature (θ) tendency to the zonal (U_{adv}), meridional (V_{adv}) and vertical (W_{adv}) heat advection, thermal forcing (Q) and residual terms (R) through:

$$\frac{\partial \theta}{\partial t} = U_{adv} + V_{adv} + W_{adv} + Q + R. \quad (1)$$

We note that certain data assimilation techniques may result in this evolution equation being nonconservative. We do not explicitly compute the thermal forcing as our focus is on the equatorial subsurface below the mixed layer, with climatological depths ranging from 20 m in the eastern Pacific to 70 m in the western Pacific, where the effect of Q is small. The interannual anomalies of the heat advection components are expressed as

$$U'_{adv} = -\bar{u} \frac{\partial \theta'}{\partial x} - u' \frac{\partial \bar{\theta}}{\partial x} - u' \frac{\partial \theta'}{\partial x} + \overline{u' \frac{\partial \theta'}{\partial x}}, \quad (2)$$

$$V'_{adv} = -\bar{v} \frac{\partial \theta'}{\partial y} - v' \frac{\partial \bar{\theta}}{\partial y} - v' \frac{\partial \theta'}{\partial y} + \overline{v' \frac{\partial \theta'}{\partial y}} \quad (3)$$

and

$$W'_{adv} = -\bar{w} \frac{\partial \theta'}{\partial z} - w' \frac{\partial \bar{\theta}}{\partial z} - w' \frac{\partial \theta'}{\partial z} + \overline{w' \frac{\partial \theta'}{\partial z}}, \quad (4)$$

where the overbar and the prime denote the climatological and anomalous components, respectively, and u , v and w the zonal, meridional and vertical current velocities. A 13-term running average (1/24, 1/12, ..., 1/12, ..., 1/12, 1/24) is used to calculate the interannual anomaly component from detrended monthly variables. Other low-frequency filters were tested, such as a recursive Butterworth procedure [Ballester *et al.*, 2011], but similar results were found. Given that the contribution of the nonlinear advection terms (i.e., the last two terms in equations (2–4)) is generally small compared to the other components, they will not be explicitly described in this work, although they are implicitly included in the U'_{adv} , V'_{adv} and W'_{adv} terms throughout this article.

Ocean potential temperature and zonal and meridional current velocities are obtained from five assimilation products: NEMOVAR-COMBINE (model: NEMO v3.0) [Balmaseda *et al.*, 2010], GECCO (MITgcm) [Köhl and Stammer, 2008], SODA2.2.6 (POP2.x) [Carton and Giese, 2008], ORAS4 (NEMO v3.0) [Balmaseda *et al.*, 2013], and ORAS3 (HOPE) [Balmaseda *et al.*, 2008]. Vertical velocity is diagnosed by integrating horizontal divergence down from the surface, with surface values assumed to be equal to the time tendency of sea surface height.

EN events are chosen according to the classification of the Climate Prediction Center: December 1963, 1965, 1968, 1972, 1976, 1982, 1986, 1990, 1997, 2002, 2006 [CPC, 2015]. In those cases when EN conditions are observed in the tropical Pacific for two consecutive boreal winters (i.e., 1968/1969, 1976/1977, 1986/1987, 1990/1991), only the first year is considered for the calculation of the composite anomalies, given that the main objective of the article is the description of the onset of these events. The 1994 event was excluded from the analyses because it was the continuation of a previous warm event starting in 1990, with warm sea surface temperature anomalies persisting in the central and eastern tropical Pacific for almost 6 years [Trenberth and Hoar, 1996]. A similar criterion is used for the selection of LN years in Figure 2: December 1964, 1966, 1970, 1973, 1975, 1984, 1988, 1995, 1998, 2007 [Ballester *et al.*, 2015].

3. Climatological and ENSO Year Features

The main climatological features that characterize the circulation and the thermodynamic structure of the equatorial and off-equatorial Pacific Ocean in the assimilation products is consistent with previous observational studies and is shown in Figure 1. The westward SEC is simulated in the tropical south Pacific, extending from 20S to about 3–4N across the equator, where it is largely driven by the trade winds (Figures 1a and 1b). The circulation in the equatorial Pacific is also characterized by the eastward Equatorial Under Current (EUC), a subsurface current 200–400 km wide transporting 30–40 Sv along the tilted equatorial thermocline [Izumo, 2005, Figure 1b]. The zonal velocity of the EUC is strongest in the central Pacific at about 140W, and

Figure 1. Climatological temperature and ocean currents. Multiproduct average for the common period 1961–2000 of potential temperature ($^{\circ}\text{C}$, (a–f) shading and contours) and (a–c) zonal, (d–f) meridional, and (a–f) vertical ocean velocities (m/s, arrows) averaged in latitude over the south off-equatorial (8–4S), (b) equatorial (2S–2N), and (c) north off-equatorial (4–8N) regions, and averaged in longitude in the (d) western (150–160E), (e) central (160–150W) and (f) eastern (100–90W) Pacific (see the vertical green lines). The temperature contour interval is 1°C , and the thick contour corresponds to the 20°C isotherm.

is primarily driven by the east-west pressure gradient in the equatorial plane, in turn determined by the strength of the easterly zonal wind stress. In the northern off-equatorial Pacific, the assimilation products also consistently reproduce the North Equatorial Counter Current (NECC), whose eastward transport is determined by Sverdrup dynamics and whose spatial structure is constrained by near-equatorial zonal wind stress [Yu *et al.*, 2000, Figure 1c].

The role of the trade winds is also key for explaining the spatial distribution of temperature and vertical currents along the equatorial Pacific. The dynamical forcing associated with the easterly wind stress piles up warm waters to the western Pacific and deepens the local thermocline there (Figures 1b and 1d). In this area, horizontal current convergence in the ocean surface induces weak downwelling motion above the thermocline level, at the narrow westernmost edge of the longitudinal band with strong easterlies (i.e., 150–160E) [Ballester *et al.*, 2015]. In the eastern Pacific, the wind stress forcing shoals the thermocline, with a zonal contrast of about 100–120 m in thermocline depth between the western and eastern parts of the basin (cf. Figures 1b, 1d, and 1f). Ekman-driven equatorial upwelling in the central Pacific and coastal upwelling in the eastern Pacific bring to the surface cold water from below the thermocline level, which explains the equatorial minimum in surface temperatures relative to the off-equatorial bands (Figures 1e and 1f). The rising motion in the central equatorial Pacific is part of the shallow meridional overturning circulation, with surface poleward divergence, symmetric downwelling motion in the off-equatorial bands and meridional convergence at the pycnocline level [Izumo, 2005, Figures 1a–1c, and 1e].

The difference between the average thermodynamic structure in the equatorial and off-equatorial Pacific Ocean during the mature phase of EN and LN events is shown in Figure 2. During LN (EN) events, the strengthening (weakening or even reversal) of the trade winds increases (decreases) the tilting of the thermocline and enhances (reduces) most features of the oceanic circulation, including the SEC, the EUC, the downwelling motion in the western Pacific and the shallow meridional overturning cells in the central Pacific. Two prominent exceptions are found in the assimilation products. First, the eastward NECC is weakened (intensified) during the growing and mature phases of LN (EN) as a result of the decreased (increased) wind stress curl north of the equator [Hsin and Qiu, 2012, Figure 2c]; together with the enhancement (suppression) of the westward SEC, this contributes to the westward (eastward) displacement of the warm pool and the development of ENSO anomalies. Second, the equatorial upwelling in the far eastern Pacific starts

Figure 2. Same as Figure 1, but for the difference of El Niño minus La Niña years. The temperature contour interval is 0.25°C , and the thick contour corresponds to the 0°C isotherm.

to be suppressed (intensified) just a couple of months before the peak of LN (EN) events (Figure 2f), being a primary out-of-phase reversal mechanism for the oscillatory nature of ENSO [Battisti, 1988; Jin, 1997a, 1997b].

4. Growth of the Warm Buildup in the Western and Central Pacific

Figures 3–5 depict the multiproduct average of the composite of EN events for the range of lags corresponding to the generation of the subsurface heat buildup in the western Pacific, between 36 and 25 months before El Niño events. The temperature tendency at these lags determines the anomalies that are observed later, during the peak of the subsurface heat buildup in the western Pacific at lag -21 months. The stippling highlights the inter-product similarities by showing the areas where anomalies have the same sign and magnitude larger than $\pm 0.25^{\circ}\text{C}/\text{yr}$ for all or all but one the members of the ensemble, an approach that we take throughout the manuscript. The inter-product differences corresponding to the vertical, meridional and zonal advection terms are additionally presented in Figures 6 and 7, respectively. Here we show the longitude-depth values along a narrow band in the equatorial Pacific (2S – 2N , Figures 3 and 6, and 7), and two latitude-depth meridional transects representative of the processes that lead to the initial stages of the heat buildup in the warm pool (150 – 160E , Figure 4) and to the east of the dateline (160 – 150W , Figure 5).

Ballester *et al.* [2015] highlighted fundamental differences between the dynamical origin of the subsurface warming in these two regions. On the one hand, in the western Pacific (130 – 170E), anomalous downwelling motion from an upper layer (0 – 75 m) of horizontal convergence to a subsurface layer (75 – 190 m) of horizontal divergence deepens the thermocline and advects heat downward (cf. Figure 3a). Horizontal convergence near the surface is found to be an intricate combination of surface (0 – 60 m) zonal convergence and subsurface (40 – 75 m) meridional convergence, while the horizontal divergence just above the thermocline level is explained by the zonal component and the intensification of the EUC in the central Pacific. On the other hand, a completely different regime prevails in the central Pacific, where surface (0 – 60 m) zonal and meridional divergence and subsurface (60 – 190 m) zonal and meridional convergence generate strong upwelling motion (cf. Figure 3a).

Analysis of the heat budget reveals that most of the spatial structure of surface and subsurface heat anomalies in the equatorial Pacific is explained by zonal and vertical advective processes (cf. shading and contours in Figure 3b). To the west of 170E , vertical advection determines a large fraction of the subsurface warming (Figure 3h). In particular, this contribution is primarily explained by the vertical advection of the climatological

Figure 3. Multiproduct average of temperature tendency and heat advection during the growing phase of the heat buildup leading to El Niño events. Composite anomalies are averaged over the equatorial band (25°N–25°S), and for the range of lags between 36 and 25 months before the major El Niño events. (a) The anomalous zonal and vertical velocity (m/s, arrows), the anomalous potential temperature ($^{\circ}\text{C}$, shading), the climatological 20°C isotherm (black contour) and the anomalous 20°C isotherm (green contour). (b–h) Time tendency of potential temperature ($^{\circ}\text{C/yr}$, contours) and the heat advection terms specified in the plot titles ($^{\circ}\text{C/yr}$, shading). The contour interval is 0.25°C/yr , with solid (dashed) lines depicting positive (negative) anomalies. The dark (light) stippling denotes areas where heat advection anomalies have the same sign and magnitude larger than $\pm 0.25^{\circ}\text{C/yr}$ for all (all but one) the members of the ensemble.

temperature by anomalous currents ($-w'\partial\bar{\theta}/\partial z$, not shown), which results from the combination of (rather weak) downwelling anomalies (Figure 3a) and the strong climatological vertical gradient of temperature (Figure 1b). Meridional cross-sections in the warm pool confirm the dominant role of vertical advection within the tropical band in this region (Figure 4). A tendency toward subsurface warming is present from 125 to 8N, which approximately corresponds to the latitudinal range with anomalous downwelling motion (Figure 4a). Nevertheless, the warming is clearly larger right at the equator near the thermocline, where both the vertical gradient of temperature and the anomaly in downward vertical velocity are largest (Figures 1d and 4a). Figure 4 also confirms the negligible contribution of zonal and meridional heat advection in this region at these very initial stages of the composite of EN events, regardless of the specific latitudes within the tropical band.

To the east of 170E, the subsurface warming is a complex combination of different mechanisms. The largest contribution to subsurface warming is associated with temperature and circulation changes along the equatorial thermocline from the zonal and vertical terms (Figure 3b). For the zonal component, the advection of

Figure 4. Same as Figures 3a and 3e–3h, but for the meridional transect of temperature, meridional and vertical currents and heat advection in the 150–160E sector.

a combination of the three heat advection components (cf. shading and contours in Figure 5b). Notably, while smaller than the zonal and vertical components, the meridional advection plays an important role in determining the tendency of subsurface temperature in the off-equatorial bands (cf. Figures 5c and 5e). In fact, the zonal and vertical advection components tend to cancel each other in the latitude intervals 6–2S and 2–5N at 100 m (Figures 5c, 5d, and 5f), just above the thermocline (Figure 5a). Therefore, the warming tendency in this area is entirely explained by the meridional advection of anomalous heat by climatological currents ($-\bar{v}\partial\theta'/\partial y$, not shown), which advects relatively less cold off-equatorial temperature anomalies (Figure 5a) through the equatorward lower branch of the climatological shallow meridional overturning cells (Figure 1e).

anomalous heat by climatological currents (i.e., $-\bar{u}\partial\theta'/\partial x$) and the advection of climatological temperature by anomalous currents ($-u'\partial\bar{\theta}/\partial x$) have similar contributions, with anomalies reaching up to $+1^\circ\text{C}/\text{yr}$ in the central Pacific near the thermocline level (not shown). Note that nearly opposite anomalies are found for the vertical component (i.e., $-\bar{w}\partial\theta'/\partial z$ and $-w'\partial\bar{\theta}/\partial z$, not shown).

The tendency in subsurface warming in the central Pacific is to a large extent explained by $-(\bar{u}\partial/\partial x + \bar{w}\partial/\partial z)\theta'$ (cf. shading and contours in Figure 3c). This contribution is associated with the negative eastward and upward gradient of subsurface temperature along the equatorial thermocline due to increasing LN-like conditions (Figure 3a), which is advected to the central Pacific by the climatological EUC (Figure 1b). These factors generate positive zonal advection at the level of the thermocline and negative (positive) vertical advection above (below) the thermocline (Figures 3f and 3h), which together explain the warming tendency observed below 100–120 m (Figure 3c). The overall contribution of $-(u'\partial/\partial x + w'\partial/\partial z)\bar{\theta}$ is instead associated with the tilted stratification of the ocean (Figure 1b), which is advected by the intensification of the EUC (Figure 3a). The diapycnal component of these circulation anomalies is not negligible in this case, with eastward anomalies defining areas of warm advection at 160W and upward anomalies generating cold advection at 180 and 140W (Figure 3d).

Unlike what is seen in the warm pool, the tendency in subsurface temperatures in the central Pacific results from

Figure 5. Same as Figures 3a, b, and 3e–3h, but for the meridional transect of temperature, meridional and vertical currents and heat advection in the 160–150W sector.

recharge mode in equatorial heat content are shown in Figures 8 and 9. These figures show averages for lags between 27 and 16 months before EN events, whose tendencies determine the subsequent peak of the recharge phase on average at lag-09 months. This range of lags is characterized by the peak in LN-like conditions (Figure 8a) and a tendency toward warming of the equatorial Pacific around the level of the thermocline (contours in Figure 8). In this section we show, as we did in the previous section, the longitude-

Although the above-described advection processes are generally reproduced by the five assimilation products, there are still substantial differences in their magnitude and spatial extent at this early stage of the oscillation. For example, all products reproduce the subsurface warming due to vertical advection to the west of 170E, but its intensity and extent vary greatly among the data sets: the downward advection is weak in ORAS3, confined to a narrow band in SODA2.2.6, close in magnitude to the composite average in NEMOVAR-COMBINE and ORAS4, and strong and extending eastward to 160W in GECCO (Figure 6). These differences are essentially a reflection of differences in the ocean circulation, because of differing pattern and magnitude of the vertical velocity in each data set (not shown). In turn, the warming due to zonal advection in the central Pacific is directly related to the subsurface warming tendency to the west of 170E, which is zonally advected along the thermocline by the EUC (not shown). Differences are even larger for the meridional component, with no contribution to the subsurface warming in GECCO, around average values in SODA2.2.6 and ORAS3, and a strong positive contribution in NEMOVAR-COMBINE and ORAS4 (Figure 7). The large differences in the magnitude of the meridional term are seen to be key for explaining the different propagation speed of the subsurface warming tendency along the thermocline (cf. contours in Figure 7, find more details in the Discussion section below).

5. Transition to the Recharged Phase

The composites corresponding to the development of the basin-wide

Figure 6. Same as Figure 3h, but for the individual members of the ensemble.

This warm anomaly can be further decomposed in its zonal and vertical contributions, which have large positive values at the level of the thermocline and at depths just above the thermocline, respectively, as a result of the corresponding advection of the subsurface anomalous heat buildup in the western Pacific (Figure 8a). The diapycnal transport is characterized by the intensification of the Ekman-induced upwelling motion in the central Pacific (Figure 8a), which drives cold waters to the surface and explains the negative anomalies in $-w'\partial\bar{\theta}/\partial z$. The zonal component $-u'\partial\bar{\theta}/\partial x$ is positive at the level of the thermocline, due to the weak intensification of the zonal component of the EUC (Figures 8a and 8f). Nevertheless, anomalies in the vertical component dominate, controlling the overall sign of $-(u'\partial/\partial x + w'\partial/\partial z)\bar{\theta}$ (Figure 8e).

depth composite along the equatorial Pacific (2S–2N, Figure 8), as well as a latitude-depth meridional transect in the central Pacific (160–150W, Figure 9). This transect corresponds to the central part of the area of maximum warming rate (i.e., largest temperature tendency anomalies) and meridional mass exchange between the equatorial plane and the off-equatorial bands. Despite some minor residual, the sum of the zonal, meridional and vertical advection terms explains the general basin-wide subsurface warming near the thermocline (cf. shading and contours in Figure 8b).

Similarly to what seen at earlier lags, in this stage of the ENSO cycle the most important contribution to the subsurface warming in the equatorial central Pacific results once again from both zonal and vertical heat advection terms (Figure 8c). While the intensification of the EUC is confined to 160–120W and is weaker than in the previous phase, heat advection anomalies are largely explained by the strong eastward and upward gradient of anomalous temperature that characterizes the peak of the LN-like conditions (Figure 8a). Further decomposition of these terms reveals that $-(\bar{u}\partial/\partial x + \bar{w}\partial/\partial z)\theta'$ (Figure 8d) and $-(u'\partial/\partial x + w'\partial/\partial z)\bar{\theta}$ (Figure 8e) have opposite signs, showing how temperature and circulation anomalies have opposing tendencies in the subsurface temperatures in the central Pacific.

The mean advection of temperature anomalies $[-(\bar{u}\partial/\partial x + \bar{w}\partial/\partial z)\theta']$ is largely positive in the whole central Pacific, from 170E to 120W and between 40 and 180 m (Figure 8d).

Figure 7. Same as Figure 3g, but for the individual members of the ensemble.

The inter-product comparison reveals that very large differences exist between ensemble members at this intermediate phase of the oscillation. For example, the sum of the zonal and vertical advection terms contributes differently to the temperature tendency in each data set: This contribution is positive in 170E–150W at 100–200 m depth in NEMOVAR-COMBINE and ORAS4, weakly positive in 160E–140W at around 100 m in SODA2.2.6, and positive in the whole equatorial Pacific at 20–140 m in ORAS3 and at 80–160 m in GECCO (Figure 10). These differences primarily arise from the vertical component, given that the diapycnal upwelling of cold subsurface waters across the thermocline is subject to large uncertainties (not shown). The meridional term only differs in magnitude and not in sign or spatial extent (Figure 11). Interestingly, the larger the contribution of the meridional advection to the subsurface warming, the larger the warming tendency near the thermocline level

While playing a relatively minor role in the central Pacific (cf. Figures 8c and 8g), meridional heat advection along the equatorial thermocline explains a large fraction of the subsurface warming in the eastern Pacific. Advection of mean temperatures by anomalous currents, $-v'\partial\bar{\theta}/\partial y$, and of anomalous temperatures by mean currents, $-\bar{v}\partial\theta'/\partial y$, both contribute to the overall meridional advection pattern. As shown in Figure 9a, in the central Pacific anomalous currents are characterized by a strengthening of the shallow meridional overturning cells, with anomalous surface Ekman divergence, off-equatorial downwelling and subsurface convergence. Thus, the subsurface convergence in meridional currents advects climatological off-equatorial warmer waters to the equator (Figure 1e). Similarly, temperature anomalies show weak warming in the off-equatorial regions near the level of the thermocline (Figure 9a), which are advected to the equator by the climatological shallow meridional overturning circulation (Figure 1e). When considering the near-equatorial band as a whole (e.g., 10S–10N), it is clear from the latitudinal transect that the combined contribution of the zonal (Figure 9c) and vertical (Figure 9e) components dominate over the meridional term in determining the subsurface warming in the recharge mode (cf. Figures 9b and 9d). Results also show that half of the contribution of the meridional advection is explained by the intensification of the ocean circulation, and the other half by the deepening of the off-equatorial thermocline (Figures 8j, 8j, and 9a).

Figure 8. Multiproduct average of temperature tendency and heat advection during the growing phase of the recharge mode leading to El Niño events. Composite anomalies are averaged over the equatorial band (2S–2N), and for the range of lags between 27 and 16 months before the major El Niño events. (a) The anomalous zonal and vertical velocity (m/s, arrows), the anomalous potential temperature (°C, shading), the climatological 20°C isotherm (black contour) and the anomalous 20°C isotherm (green contour). (b–j) Time tendency of potential temperature (°C/yr, contours) and the heat advection terms specified in the plot titles (°C/yr, shading). The contour interval is 0.25°C/yr, with solid (dashed) lines depicting positive (negative) anomalies. The dark (light) stippling denotes areas where heat advection anomalies have the same sign and magnitude larger than $\pm 0.25^\circ\text{C/yr}$ for all (all but one) the members of the ensemble.

(cf. shading and contours in Figure 11), regardless of the contribution of the zonal and vertical components (Figure 10); this finding highlights the dynamical importance of the meridional term during this phase.

Figure 9. Same as Figure 8a, 8c, and 8f–8h, but for the meridional transect of temperature, meridional and vertical currents and heat advection in the 160–150W sector.

of climatological temperature by anomalous zonal currents (i.e., $-u'\partial\theta/\partial x$) is well known to be largely responsible for the warming tendency in the central and eastern upper ocean (zonal advection feedback [An and Jin, 2001], see Figure 12f as a reference). On the other hand, the advection of climatological temperature by anomalous vertical currents ($-w'\partial\theta/\partial z$) has been described as a fundamental process for the warming in the far eastern Pacific (Ekman pumping feedback [Jin et al., 2006, Figure 12h]).

The meridional transect in the central Pacific shows that the largest heat anomalies are confined to the latitudinal range 5S–5N, between 100 m and the thermocline level (Figure 13a). The vertical velocity anomalies are characterized by strong upwelling north of the equator (6N–9N) and strong downwelling south of it (4S–1S) (Figure 13a). Thus, anomalies in the northern hemisphere tend to restore the thermocline to its

6. Onset of EN Events

Figures 12 and 13 depict composites corresponding to the onset and growing phase of EN events, with averages for lags between 15 and 4 months before EN events, whose tendencies determine the mature phase of EN. The oceanic conditions during this phase are characterized by warm subsurface anomalies along the equatorial thermocline and the weakening (enhancement) of the EUC in the western (eastern) Pacific (Figure 12a). The SEC, the downwelling motion in the warm pool and the coastal upwelling in the far eastern Pacific are also found to be weaker than in the climatology (Figure 12a). These anomalies are known to be associated with the tendency toward warm EN conditions in the central and eastern Pacific, including the beginning stage of weakened trade winds, the flattening of the equatorial thermocline and the development of a subsurface cold buildup in the western Pacific (contours in Figure 12). The role of advective processes is here briefly revisited through the longitude-depth composite along the equatorial Pacific (2S–2N, Figure 12) and the latitude-depth meridional transect in the central Pacific (160–150W, Figure 13).

Similarly to the previous phases, the combination of the three advection terms is in good agreement with the magnitude and spatial structure of the tendency in subsurface temperature (cf. shading and contours in Figure 12e). Note that this correspondence is again primarily explained by the combined contribution of the zonal and vertical advection (Figure 12b). On the one hand, the advection

of climatological temperature by anomalous zonal currents (i.e., $-u'\partial\theta/\partial x$) is well known to be largely responsible for the warming tendency in the central and eastern upper ocean (zonal advection feedback [An and Jin, 2001], see Figure 12f as a reference). On the other hand, the advection of climatological temperature by anomalous vertical currents ($-w'\partial\theta/\partial z$) has been described as a fundamental process for the warming in the far eastern Pacific (Ekman pumping feedback [Jin et al., 2006, Figure 12h]).

The meridional transect in the central Pacific shows that the largest heat anomalies are confined to the latitudinal range 5S–5N, between 100 m and the thermocline level (Figure 13a). The vertical velocity anomalies are characterized by strong upwelling north of the equator (6N–9N) and strong downwelling south of it (4S–1S) (Figure 13a). Thus, anomalies in the northern hemisphere tend to restore the thermocline to its

Figure 10. Same as Figure 8c, but for the individual members of the ensemble.

ency at the thermocline level and as a function of the time lag. This comparison highlights the generally synchronous evolution of the temperature tendency and the advection along the tilted equatorial thermocline (cf. shading and contours in Figure 14c). In the central Pacific (170E–110W), this tendency is to a large extent explained by the advection of the initial subsurface warm buildup in the western Pacific by the climatological EUC (i.e., $-(\bar{u}\partial/\partial x + \bar{w}\partial/\partial z)\theta'$, Figure 14d). Instead, the anomalous downwelling motion of surface climatological warm waters ($-w'\partial\bar{\theta}/\partial z$) is the key process explaining the evolution of the subsurface warming in the western (130–170E) and far eastern (110–90W) Pacific (Figure 14e). Note that the transition between advective processes near 170E is explained by a clear regime shift in the vertical structure of horizontal divergence [Ballester et al., 2015].

climatological depth, while those in the southern hemisphere contribute to the deepening of the thermocline near the equator by intensifying the northernmost edge of the downwelling branch of the southern shallow meridional overturning cell (Figure 1e). Interestingly, inter-hemispheric differences in vertical velocity anomalies increase the meridional tilting of the thermocline and generate southward cross-equatorial mass transport in the upper 50 m of the ocean (Figure 13a).

The meridional heat advection is negative in the central and eastern Pacific near the level of the thermocline (Figure 12g). As a result, this component starts contributing to the weakening of the heat content in the equatorial Pacific subsurface already in the recharge phase (Figure 12a), before the onset of EN and the activation of the Bjerknes feedback. The decomposition of this term shows that $-\bar{v}\partial\theta'/\partial y$ is larger and has opposite sign relative to $-v'\partial\bar{\theta}/\partial y$ (cf. Figures 12c and 12d). Indeed, $-\bar{v}\partial\theta'/\partial y$ ($-v'\partial\bar{\theta}/\partial y$) shows large negative (weak positive) anomalies in the off-equatorial regions (Figures 13c and 13d), near the areas of climatological (anomalous) subsurface equatorward convergence and strong meridional contrast in anomalous (climatological) temperature (Figures 1e and 13a).

7. Discussion

Figure 14 provides an integrated view of the ENSO oscillation by showing the contribution of the processes described throughout the manuscript to the equatorial temperature tendency

Figure 11. Same as Figure 8g, but for the individual members of the ensemble.

phase is characterized by colder (warmer) temperature anomalies at the equator than in the off-equatorial regions (e.g., Figure 13a). Thus, the meridional circulation of the climatological shallow meridional overturning cells warms (cools) the equatorial and off-equatorial thermocline before the mature phase of LN (EN) events ($-\bar{v}\partial\theta'/\partial y$; Figures 14i and 15i). On the other hand, the strengthening (weakening) of the equatorial trade winds during LN (EN) conditions, as well as the associated changes in off-equatorial wind stress curl, induces anomalous subsurface equatorward (poleward) Sverdrup transport of mass [Jin, 1997a, 1997b]. The disequilibrium balance between these processes generates a delayed warming (cooling) at the level of the thermocline ($-\nu'\partial\bar{\theta}/\partial y$; Figures 14j and 15j).

In this regard, the synchronous evolution of LN (EN) conditions and the equatorward warm (cold) advection in the central and eastern Pacific, as illustrated in Figure 14g, is shown to be compatible

The latitudinal heat advection structure is illustrated in the meridional transect in the central Pacific shown in Figure 15. The contribution of the meridional term is negligible right at the equator, where the Ekman-induced upwelling motion dominates, but it rapidly increases poleward, attaining anomalies that are already large at 2S and 2N (Figure 15g). Here we considered a narrow equatorial band (i.e., 2S–2N) in order to isolate the ascending branch of the shallow meridional overturning cells from the descending branches at around 8–3S and 3–8N (Figure 1e). Note that this circulation is associated with a cross-shaped anomalous pattern in both the zonal and vertical advection terms (Figures 15f and 15h), which is not found in their combined contribution (Figure 15c). The meridional advection remains qualitatively unmodified when the latitudinal range considered for the equatorial averages in the longitude-depth plane includes the descending branches of the cells (e.g., 6S–6N, not shown).

The peak in LN (EN) conditions is approximately in phase with the equatorward meridional warm (cold) advection in the central and eastern Pacific (Figure 14g). The phase of the oscillation in which the contribution of meridional advection reaches its peak is however the result of the combination of two different processes with different temporal evolution. On the one hand, the discharge (recharge) phase in basin-wide equatorial heat content leads to LN (EN) events by approximately 9 months [Meinen and McPhaden, 2000]. This

Figure 12. Multiproduct average of temperature tendency and heat advection during the growing phase of El Niño events. Composite anomalies are averaged over the equatorial band (25°S–25°N), and for the range of lags between 15 and 4 months before the major El Niño events. (a) The anomalous zonal and vertical velocity (m/s, arrows), the anomalous potential temperature (°C, shading), the climatological 20°C isotherm (black contour) and the anomalous 20°C isotherm (green contour). (b–h) Time tendency of potential temperature (°C/yr, contours) and the heat advection terms specified in the plot titles (°C/yr, shading). The contour interval is 0.25°C/yr, with solid (dashed) lines depicting positive (negative) anomalies. The dark (light) stippling denotes areas where heat advection anomalies have the same sign and magnitude larger than $\pm 0.25^\circ\text{C/yr}$ for all (all but one) the members of the ensemble.

with the recharge theory formulated by Jin, which is mathematically described by the tilting mode and the recharge-discharge phase. The tilting mode characterizes the quick oceanic response to enhanced (weakened) easterly wind stress in the central tropical Pacific during LN (EN) conditions, which is proportional to the zonal tilting of the equatorial thermocline. The recharge (discharge) phase provides the required memory between opposite phases of the tilting mode. This transition period is characterized by the time tendency toward anomalous equatorward (poleward) Sverdrup convergence (divergence) of mass due to enhanced (weakened) easterly wind stress in the western and central tropical Pacific, and its associated change in the off-equatorial curl, which ultimately tends to deepen the thermocline.

The recharge theory is also found to be compatible with the longitudinal transition in the mechanisms explaining the initial subsurface heat buildup on either side of 170°E, as well as the subsequent eastward propagation along the equatorial thermocline. Near and east of the dateline, anomalous easterly (westerly)

Figure 13. Same as Figure 12, but for the meridional transect of temperature, meridional, and vertical currents and heat advection in the 160–150W sector.

trade winds during LN (EN) events are associated with the tendency toward equatorward (poleward) Sverdrup mass convergence (divergence) and the deepening of the thermocline (e.g., Figure 8j). Near the edge of the warm pool, easterly (westerly) wind stress anomalies and the associated anticyclonic (cyclonic) curl anomalies are weaker, and therefore this delayed effect is smaller (Figure 8j). The oceanic response in this region appears to be more directly controlled by the zonal convergence (divergence) of the zonal wind stress along the equator, which favors anomalous surface ocean horizontal convergence (divergence) and downward (upward) motion during LN (EN) events [Ballester *et al.*, 2015; Figures 3h and 4e]. This process explains the much faster, albeit still somewhat delayed, response of subsurface temperatures in the warm pool (e.g., the zero contour in Figure 14e crosses longitude 160E at lag +03). The present article clarifies, within the context of the recharge oscillator theory, the relative contribution, spatial extent and delayed effect of each of the mechanisms involved in the subsurface buildup in the western and central Pacific, and its eastward propagation.

Finally, it is worth noting how relatively large agreement exists between assimilation products during the three phases of the oscillation. Differences between data sets are generally small, and anomalies tend to be similarly signed (see Figures 6–10, and 11). The only exception is GECCO, which does not produce the

Figure 14. Multiproduct average of temperature tendency and heat advection before and after the peak of El Niño events. Composite anomalies are averaged over the equatorial band (2S–2N) at the depth of the tilted thermocline. (a) The anomalous potential temperature (°C, shading) and its time tendency (°C/yr, contours). The contour interval is 0.50°C/yr, with solid (dashed) lines depicting positive (negative) anomalies. (b–j) Zero contour of the time tendency of potential temperature, as well as the heat advection terms specified in the plot titles (°C/yr, shading). The solid (dashed) horizontal green line shows the phase corresponding to the peak of EN events (LN-like conditions). The stippling denotes anomalies in which heat advection has the same sign and magnitude larger than $\pm 0.25^\circ\text{C}/\text{yr}$ for all the members of the ensemble.

patterns of meridional surface Ekman divergence (subsurface Sverdrup convergence) in the western and central equatorial Pacific observed in the other assimilation data sets, therefore featuring the weakest subsurface heat buildup. The relatively coarser resolution of this product ($1^\circ \times 1^\circ$) might partially explain some of these differences. More importantly, this assimilation data set is obtained by fitting the model simultaneously to all available data over the whole 50 year period, iteratively for 23 iterations by first running the forward model to calculate the model data misfit formulated as a cost function, followed by an adjoint model run to calculate the gradients of this cost function [Köhl and Stammer, 2008]. Given the relatively large computational demands of this approach, the optimization did not fully converge after this number of iterations. Although this solution was found to be good enough to investigate the underlying processes and causes of other phenomena such as the Meridional Overturning Circulation [Köhl and Stammer, 2008], it

Figure 15. Same as Figure 14, but along the latitudinal axis and averaged over the longitudinal range 160–150W.

seems that the model was not brought into a good level of agreement with the data regarding those processes that are fundamental to the generation and growth of EN events.

Mechanisms described in the present work can be used as a reference for the validation of numerical simulations from intermediate and complex coupled climate models. For example, a key feature highlighted here is the asymmetry between the northern and the southern tropical hemispheres during the recharge mode. This phase is characterized by an increase of the meridional tilting of the thermocline and the southward cross-equatorial mass transport in the upper ocean as a result of the anomalous upwelling (downwelling) motion in 6–8N (3–15). Yu and Mechoso [2001] showed that the anomalies in vertical velocity are due to the latitudinal distribution of zonal wind stress anomalies, which induces areas of convergence and divergence of meridional Ekman transport. Nonetheless, the climate model used by Yu and Mechoso [2001] simulated vertical anomalies of equal sign in both hemispheres and no cross-equatorial anomalies between the off-equatorial regions, in disagreement with our results. This highlights how the present work provides a description of dynamical processes that the climate modeling community might use as metrics to test the performance of the ENSO oscillation in state-of-the-art climate models.

8. Summary

The present work describes different processes that control subsurface temperatures and thermocline depth during the generation of El Niño events through a careful analysis of the subsurface heat budget. We emphasize the role that different processes play in the evolution of subsurface warm anomalies during the different stages of the oscillation. Main results include:

1. To the west of 170E, the vertical advection of climatological temperature by anomalous currents, induced by surface horizontal convergence, downwelling motion and subsurface divergence, was shown to explain alone the initial subsurface warming in the equatorial and off-equatorial Pacific during the warm buildup stage, between monthly lags -33 to -21 before the peak of EN events.
2. The role of horizontal advection was found to be confined to the east of 170E, explaining the tendency toward the return to climatological conditions of subsurface temperatures in the central Pacific, both through zonal and vertical advection along the equatorial thermocline and through meridional advection right above this level.
3. These two mechanisms were also shown to explain a large fraction of the subsurface warming associated with the recharge phase in basin-wide heat content. On the one hand, along the meridional axis, the equatorward advection of heat was shown to be explained to the same extent by anomalies in the meridional gradient of subsurface temperature and anomalies in the meridional ocean circulation. On the other hand, along the equatorial plane (i.e., combination of the zonal and vertical components), the anomalous heat accumulated in the western Pacific was seen to be advected to the central Pacific by the climatological currents. This contribution was found to be partially counterbalanced by the advection of climatological temperature by the anomalous currents, which is dominated by anomalous diapycnal upwelling of cold subsurface waters.
4. The large differences in the magnitude of the meridional term were seen to be key for explaining the different propagation speed of the subsurface warming tendency along the thermocline.

All terms in this analysis are inferred from an ensemble of state-of-the-art ocean assimilation products, focusing on those processes that are robustly produced by all the members of the ensemble, as well as those that are differently simulated by a subset of data sets. The combined use of multiple ocean analysis products provides a reference for a three-dimensional description of mechanisms leading to the generation of EN events. Additionally, it allows for a more detailed validation and assessment of mechanisms previously inferred from intermediate and complex coupled climate models, as well as for the determination of the limits in the use of assimilation products for the validation itself.

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