A Discussion of Flow Pathways in the Central and Eastern Equatorial Pacific

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(Manuscript received 19 September 2005, in final form 20 August 2006)

ABSTRACT

An eddy-permitting global ocean model is used to interpret kinematics within the central and eastern equatorial Pacific Ocean, from 160°E to the coast of America. Because of high levels of variability in this region, observational studies of meridional flow are contradictory, in particular as to whether the net flow is northward or southward. Unlike most oceanographic datasets, model output can be analyzed at high temporal and spatial resolution, providing clues to real ocean behavior. In the model, a net southward flow occurs across the equator east of 160°W, at most density layers throughout the year. In the central Pacific, from 160°E to 160°W, the net flow is northward but varies with season and occurs primarily in the mixed layer. This is a key region for the flow of Equatorial Undercurrent water into the Northern Hemisphere. The three-dimensional flow is very complex and seasonally dependent. It is vital that these flows are analyzed in an isopycnal framework, or else the pathways are very misleading. In the first half of the year, evidence is found of meridional tropical cells on either side of the equator out to 5°N. These cells appear to exist without any need for diapycnal downwelling. In the second half of the year, when tropical instability waves are active, the cells are overlaid by a strong surface southward flow that appears to be a bolus-type transport. This transport is not apparent unless the flow is calculated in the aforementioned manner.

1. Introduction

While interannual variation of equatorial currents has been studied intensively in the Pacific Ocean, we still lack a full understanding of the annual mean and seasonal cycle of the Pacific Ocean, to which El Niño–Southern Oscillation (ENSO) is phase locked. One particular shortfall occurs in the meridional component of flow on and near the equator. It is well known that there are vigorous time-mean zonal flows along the equatorial Pacific, away from boundaries. We are primarily concerned here with the Equatorial Undercurrent (EUC) and South Equatorial Current (SEC), as they account for most of the depth-integrated flow.

There is also intense time-mean upwelling, to feed the poleward Ekman flows on either side of the equator. However, meridional flow near the equator is neither well measured observationally, nor well understood theoretically. Our aim is to explore the kinematics of near-equatorial Pacific flow in an eddy-permitting global ocean model.

Understanding meridional flow is a complex and inherently nonlinear problem, made more so because of the change of sign and strong gradient of planetary vorticity across the equator, creating a vorticity barrier there. Because of this barrier, no steady flow across the equator can occur in the absence of friction and diffusion. Further nonlinearity is introduced by the high variability tropical instability waves (TIWs), which propagate on the northern side of the equator. Their large amplitude makes them difficult to analyze and they can introduce a bolus-type transport (Gent et al. 1995) to the balance thereby making the dynamics harder to comprehend.

Because of this high variability in the equatorial Pacific, data must be averaged to remove the noise and
hence spatial and/or temporal resolution is lost. For this reason we use model output, which provides high-resolution information, to explore pathways for cross-hemispheric flow. Since friction (both vertical and lateral) is thought to be important in allowing time-mean cross-equatorial flow, we chose a model with what is believed to be one of the more realistic friction schemes presently available—the nonlinear Smagorinsky biharmonic scheme as implemented by Griffies and Hallberg (2000). Using model output also allows us to demonstrate the importance of considering meridional flow in a density and time-varying framework, which will show to be essential in resolving the contribution of TIWs and the resulting bolus transport, in the second half of the year.

We compare our model results with a wide variety of previous papers, some of which rely on model output, others on pure observations, and which together provide a whole range of qualitative insights into equatorial Pacific mean flow. These papers are due to Kessler et al. (2003); inverse models from Wijffels (1993) and Sloyan et al. (2003); observations from Johnson et al. (2001), Weisberg and Qiao (2000), and Poulain (1993); and other modeling studies from Liu et al. (1994) and Fukumori et al. (2004). A comprehensive review of our present understanding of meridional flow can be found in Brown (2005); here we quote results as relevant. We outline the flow pathways in our model compared to earlier (sometimes conflicting) ideas. We make no claim for the “rightness” or otherwise of our model-based picture, but it does provide a detailed, dynamically self-consistent picture that may be used (with caution) as a guide to interpreting ocean observations.

Wijffels (1993) discovered an upper-ocean pathway for cross-equatorial flow using an inverse model, and a simple schematic of this pathway is reproduced in Fig. 1. Water was found to flow from the Southern Hemisphere into the western boundary currents, which feed the EUC along the dark gray line from the SEC to the EUC in the figure. The water then upwelled along the EUC, becoming less dense. From the surface waters it then traveled southward, warming further and advecting westward in the SEC (light gray line). In the western half of the basin it was able to then cross into the Northern Hemisphere—both in the western boundary current and the ocean interior—eventually feeding into the Indonesian Throughflow via the Mindanao current. We will confirm this pathway in the model output and explore it in more detail.

Sloyan et al. (2003) also studied the circulation patterns of the equatorial Pacific, with an inverse model. At 2°N in the eastern Pacific, their results were more consistent with Sverdrup theory than Wijffels, finding northward flow. However at 2°S, and even 8°S, they found northward flow, which contradicts Sverdrup predictions and Wijffels’ observations.

Small meridional tropical cells (TCs) are thought to sit either side of the equator out to 5°N and S, as described by Wyrtki and Kilonsky (1984) and more recently Lu et al. (1998), Kessler et al. (1998), and Johnson et al. (2001). Liu et al. (1994) conducted investigations of flow pathways in these TCs by tracking particle paths in an idealized model. They found that the particles flowed along the EUC at the equator and upwelled in the eastern part of the ocean. They then follow a simple “corkscrew”-like pattern—upwelling at the equator, drifting poleward in the Ekman flux in both hemispheres (out to 5°), westward in the SEC, and then downwelling and converging back to the equator in the geostrophic flow. They rejoin the EUC where they continued their eastward and upward flow. The cycle continued, with a net westward displacement each time, until they reached the western boundary or broke away in the interior. We will show that in our model these TCs do exist, but must be measured carefully considering the time-varying position of the isopycnals.

Observations of the cell-like structure were found by Johnson et al. (2001), with an extensive study of ADCPs over a large region of the equatorial Pacific from 170°W to 95°W. They found a net flow of 10 Sv (1 Sv = 10⁶ m³ s⁻¹) northward across the equator, which flowed in from the Southern Hemisphere in the lower arm of the tropical cells, then upwelled and continued northward in the upper arm of the Northern Hemisphere cell. The error bars on these measurements were quite large however. Other observational studies include Weisberg and Qiao (2000) and Poulain (1993) with somewhat conflicting results. Poulain (1993) also shows that much of the equatorial upwelling occurs within less than half a degree from the equator, suggesting a need for highly resolved data.

In depth coordinates, the presence of meridional tropical cells is clear in observations and models. Hazleger et al. (2001), however, argue that calculations along time-varying isopycnals show that these cells appear to be an artifact of the coordinate system. First they showed that converting flow to a density coordinate system resulted in much weaker TCs. This is because isopycnals are slanted in the EUC region and so integrations at constant z levels create an often spurious impression that upwelling or downwelling exists across isopycnal surfaces. They also showed that incorporating the high-frequency mass fluxes allows TCs to be almost completely compensated by eddy-induced
overturning. This occurs because of correlations between velocity and isopycnal thickness variations, also known as bolus transport (Gent et al. 1995; McIntosh and McDougall 1996; McWilliams and Danabasoglu 2002).

Hazeleger et al. (2001) also support their argument with the fact that the downwelled water in the assumed TCs is far too warm to match the subsurface equatorial waters, and a mechanism for this cooling is not known (McCreary and Lu 1994). For this reason, where possible our discussions will also take place in a density layer coordinate system, and we will estimate instantaneous transport between isopycnals.

To understand the flow pathways in the equatorial region, we study an eddy-permitting global ocean model run with full variability included, on density layers. Section 2 discusses the model used in deriving our data and the methods used to obtain flow estimates. Section 3 presents our results for the annual mean, first on depth integral and then by density layer. We will show in section 4 that the annual mean is somewhat deceptive in highlighting cross-equatorial flow pathways by presenting the flow pathways induced by seasonal variability. In this section we will also consider the role of TIWs in cross-hemispheric flow through bolus transport near the surface.

2. Model output and processing

a. Model details

The model is a global version of the Australian Community Ocean Model (ACOM3; see Schiller 2004 and Schiller and Godfrey 2003 for further details) and is based on Geophysical Fluid Dynamics Laboratory (GFDL) Modular Ocean Model 3 (MOM3) code (Pacanowski and Griffies 2000), forced with European Remote Sensing Satellite (ERS) scatterometer winds. The

![Diagram of ocean flow pathways](image-url)
resolution is $\frac{1}{6}^\circ$ latitudinally and $\frac{1}{2}^\circ$ in longitude, allowing tropical instability waves to be fairly well resolved. The model is run from July 1992 to October 2000, after an 8-yr spinup, based on seasonal climatology derived from daily/original forcing fields. There are 36 vertical levels of which the top 10 are 10 m thick and increase thereon with a thickness of 225 m by the 1000-m depth.

A nonlinear Smagorinsky biharmonic mixing scheme (Griffies and Hallberg 2000) was used for horizontal viscosity. In the vertical, a hybrid Chen mixing scheme (Schiller et al. 1998; Power et al. 1995) is applied with a Niiler–Kraus-like bulk mixed layer near the surface, and gradient Richardson number mixing below.

Surface fluxes, apart from incoming shortwave radiation, are calculated by coupling the OGCM to an atmospheric boundary layer model as described by Kleeman and Power (1995) and Schiller and Godfrey (2003). The ABLM consists of a single-layer model atmosphere (boundary layer as well as a portion of the cloud layer) that is in contact with the surface. Wind fields at the 850-hPa level are assumed to be representative of the circulation in the atmospheric boundary layer and are prescribed [National Centers for Environmental Prediction (NCEP)].

Within $15^\circ$N/S in the Indian and western Pacific Oceans, we use a well-defined regression of solar shortwave radiation on outgoing longwave radiation (OLR; Shinoda et al. 1998) to obtain the former.

The upward sensible and latent eddy heat fluxes and net upward longwave radiation are parameterized with traditional bulk formulas (see Kleeman and Power 1995 for details).

Freshwater fluxes were calculated from the simulated evaporation (latent heat) together with precipitation from the Carbon Dioxide Information Analysis Center (CDIAC) Microwave Sounding Unit (MSU) precipitation dataset.

b. Flow calculations

As found by Hazeleger et al. (2001), it is important to study meridional flow on spatially and time-varying density layers. The first issue is that flow in the equatorial Pacific is mostly along density layers, which slope upward in the east. Erroneous upwelling and downwelling can therefore occur if averaging is done at constant $z$ levels. Second high-frequency eddy motions induce a mass transport, much like a bolus transport, due to correlations between velocity and density coordinates. Therefore if flow is calculated between mean positions of density layers, errors are introduced. To address this in our model output, each time the model output is collected (every three days), the depth of the density layers is calculated, and then the corresponding meridional and zonal velocities are integrated vertically between them.

The depth-integrated flow is taken to the 1026.5 kg m$^{-3}$ isopycnal, which corresponds to approximately 300 to 350 m (see Fig. 2a). It was found that this isopycnal captured most of the flow and avoided the problems this particular model has with bottom boundary layers. For ease in understanding the flows, and to aid comparison to previous work, the flow is calculated within boxes bounded by 130°E, 160°E, 140°W, 120°W, 100°W, and the American coastline, and 10°S, 4°S, 2°S, 0°, 2°N, 4°N, and 10°N. Having model output means that ideally we could have boxes the size of the model grid. This is not helpful as the picture is then too complex to understand and the variability is quite high. We have chosen these particular layers and boxes to capture the flows we are trying to describe without making the picture too complex.

In the vertical, the isopycnal layers are chosen as follows: from the surface to 1022.5 kg m$^{-3}$ (light layer), to 1023 kg m$^{-3}$ (mixed layer 1), to 1023.5 kg m$^{-3}$ (mixed layer 2), 1024.5 kg m$^{-3}$ (upper thermocline), 1026 kg m$^{-3}$ (lower thermocline), and the remaining water to 1026.5 kg m$^{-3}$ (deep layer). The layers called “light layer,” “mixed layer 1,” and “mixed layer 2” are not present everywhere in the Pacific, and generally outcrop in the Southern Hemisphere. Their outcropping positions are highly variable in time. Mean depths of the isopycnals used in the analysis along the equator (Figs. 2a–c) slope upward to the east, where some of them outcrop.

At the base of each density layer an approximate calculation is made for annual mean cross-isopycnal flow. This is done by considering the horizontal transport balance even though the depth of the isopycnal is not constant. The rate of change of the vertical position of each isopycnal, time-averaged over the whole period, is negligible compared to the divergence of the horizontal flow above the isopycnal. The flow through the base of the light layer is calculated first, and it is assumed that there is no flow into the surface of the ocean, that is, precipitation and evaporation are negligible. Subsequent deeper layers are then calculated. Our assumption of negligible evaporation and precipitation does introduce errors in downwelling and upwelling, but they are minor.

3. Annual mean flow pathways

a. Depth integral

Both model and theoretical time-mean streamfunctions show that depth-integrated flow is eastward along
the equator, with a southward tendency after about 150°W (Figs. 3b,d). Here the theoretical streamfunction is calculated with the same ERS scatterometer winds used in the model, so the theoretical streamfunction is a linear approximation to the full model results. Note that in the calculation of Sverdrup flow (Figs. 3a,c) it is important to have the high horizontal resolution of the ERS wind stress to capture the detail of the flow in the eastern Pacific (Kessler et al. 2003). Differences between the model and theoretical streamfunctions, particularly the strength of the currents, largely represent the contribution of nonlinearity and friction in the system (as noted by Kessler et al. 2003). There may also be some discrepancy due to Sverdrup transport at depth; however, we believe this is minimal.

The model does a reasonable job of capturing the observed zonal currents (Fig. 4). The narrow equatorial jet—referred to below as the “EUC,” for convenience, even though the EUC is overlaid by some westward-flowing part of the SEC—is of comparable strength.
though it is too strong in the west in the model. The EUC is sandwiched between northern and southern branches of the “SEC,” which are weaker than observed. To the north lies the North Equatorial Counter Current (“NECC”), which is also weaker than observed across the basin.

For later comparisons, the depth-integrated flows from the model have been quantified according to an array of boxes set up between 10°N and 10°S (Fig. 5). We note that the flow through the boxes may not appear to obey continuity. This is due partly to rounding errors and also a weak diapycnal flow across the 1026.5 kg m\(^{-3}\) isopycnal at the base of each box.

The EUC reaches a maximum at around 120°W, driven by the zonal pressure gradient associated with the density gradients in Fig. 2a. The increase in the EUC is fed partly by recirculating water in the SECs, as shown by Wyrtki (1967). Upon reaching the Galapagos Islands (around 92°W) the flow bifurcates to the north and south. Eastward flow is still present east of the Galapagos Islands (Lukas 1986), however the majority of the flow is poleward, as expected from the change in sign of the zonal pressure gradient (Fig. 2a). Note also that by 160°W the southern half of the EUC is significantly stronger than the northern portion, and this discrepancy continues to grow as the EUC travels eastward.

Partitioning the Pacific along the equator leaves the North Pacific as a closed system, bar the Indonesian Throughflow and the Bering Strait. Across this boundary there is a net flow of 7.2 Sv northward, due to a net northward flow of 15.9 Sv in the western boundary currents and a net return of 8.7 Sv east of 160°W. As the Bering Strait is around 0.6 Sv (Aagaard et al. 1985), this leaves most of the flow to exit via the Indonesian Throughflow or through the 1026.5 kg m\(^{-3}\) isopycnal. The exact path that water takes to reach the Northern Hemisphere cannot be understood from this two-dimensional picture. In the next section we will study the flow in the vertical, by density layer, to address the conflicting views of where and how water gets into the Northern Hemisphere.
b. Density layers

We now look at the flow profiles along instantaneous density layers as described in section 2b. The long-term mean position of these layers (Fig. 2) clearly demonstrates that had we chosen to look at the flow at a particular depth, it would cross over many different density layers.

The same horizontal boxes are used as in the depth-integrated study (though note the color scale for the arrows is different). Diapycnal flows are now included with dots representing diapycnal upwelling and crosses representing diapycnal downwelling. While in the previous section we found a net southward cross-equatorial flow east of 160°W, other authors have found northward flow at some depths in this region. We wish to find the major flow pathways and establish if northward flow does occur in our model at any particular density levels. It turns out that flow behavior is qualitatively different in the top three layers and in the bottom three. It is convenient to discuss these two groups separately.

1) The lowest three layers

The deep layer (Fig. 6f) shows flow between 1026 and 1026.5 kg m⁻³. There are only very weak diapycnal flows through the bottom of these boxes (all less than 1 Sv). Apart from the western boundary current there is very little cross-equatorial flow. Flow at this depth would encounter the strong potential vorticity barrier at the equator with little in the way of friction, away from the boundaries, to allow it to change hemispheres. There is a pathway for the eastward-traveling equatorial water, already in the Northern Hemisphere, to turn and flow northward past 2°N between 160°W and the
coast of America, in agreement with Sloyan et al.’s (2003) results.

In the “lower thermocline layer” (Fig. 6e), the diapycnal flow symbols (dots and crosses) represent water in this layer flowing from or into the “deep layer” below. Again this diapycnal flow is quite weak and none are greater than 1 Sv. The largest horizontal flows are dominated by the western boundary currents and the EUC along the equator. The bulk of the depth-integrated EUC appears to be accounted for in this layer. The EUC also shows some tendency for the southern half to be stronger than the northern, but not to the extent of the depth integral discussed previously.

East of 140°W, water is lost via upwelling to the layer above (as indicated by the “dots” in Fig. 6d) being the upper thermocline. In the easternmost box, there is strong poleward flow in both hemispheres of similar magnitude to the initial strength of the EUC, of around 5 Sv.

In the model at least, a large part of the EUC originates near the west and leaves near the east. The increase in the EUC in the central Pacific is due to meridional convergence, rather than upwelling. In the Southern Hemisphere this is fed both by a recirculation of EUC water from the east and from an inflow of water across 10°S. In the Northern Hemisphere, the water that feeds the EUC can be separated into two sections. East of 140°W, the eastern outflow of the EUC seems to be feeding back into itself (as it did in the Southern Hemisphere). West of 140°W the EUC is fed from water that originated in the west of the basin and from the layer below.

In the “upper thermocline layer” (Fig. 6d) the strongest flows are again western boundary currents and the eastward flow between 2°N and 2°S, though they are much weaker than in the deep thermocline layer below. Again the strength of the EUC appears to be stronger in the Southern Hemisphere than the north, east of 140°W. In the lower thermocline layer this occurred partly due to a southern cross-equatorial flow; in this layer it is more attributable to upwelling. The intense upwelling from lower thermocline to upper in the far east also helps to feed southward flows past 4°S. Poleward of 2°N and 2°S, zonal flows are quite similar in sign and magnitude to those in the lower thermocline layer.

The upper thermocline layer (Fig. 6d) is close to the surface in the eastern part of the basin and so can be
Fig. 6. Same as in Fig. 5; however, flow is now separated in the vertical into six density layers: (a) light layer ([1022.5–1022.5 kg m$^{-3}$]), (b) mixed layer 1 (1022.5–1023 kg m$^{-3}$), (c) mixed layer 2 (1023–1023.5 kg m$^{-3}$), (d) upper thermocline (1023.5–1024.5 kg m$^{-3}$), (e) lower thermocline (1024.5–1026 kg m$^{-3}$), and (f) deep layer (1026–1026.5 kg m$^{-3}$). The top three layers outcrop at the surface, which is also shown in the figure. Diapycnal flow is represented in this picture with dots indicating upwelling from the layer below and crosses indicating downwelling into the layer below. Strength of the flow is indicated by the number and shading of the arrows (a different scale has been used than in Fig. 5).
considered part of the mixed layer, with Ekman flow away from the equator (Johnson et al. 2001). In the west, the upper thermocline layer subducts beneath the three upper layers, leaving the region of Ekman influence. Meridional flow becomes purely geostrophic, and therefore flow is equatorward in the western half, as required by the pressure fields associated with Fig. 2a.

2) The upper three layers

The remaining three layers (Figs. 6a–c) are sections within the mixed layer, showing different flow patterns compared to the previous deeper three. The zonal flow at the equator is now due mostly to the westward SEC and so the recirculation cells that were evidenced in the lower layers and feeding back in the EUC no longer exist. The outcropping positions of these layers (Fig. 2c) also vary in time and so the layers do not always exist everywhere in the basin at all times. These layers are more easily subjected to the wind forcing, turbulent mixing, and Ekman flow, which is also partly dependent on when they are exposed to the surface.

Right on the equator, flow is generally southward east of 160°W and northward to the west. Away from the equator, flow is westward in the SEC, and particularly strong in the Southern Hemisphere, entraining water from the upper thermocline as it travels.

Traditionally surface flow within a few degrees of the equator is thought to be poleward due to Ekman dynamics. This holds in the Southern Hemisphere and in mixed layer 2. The light layer and mixed layer 1, in Figs. 6a,b, introduce a new feature. They show similar flow patterns to mixed layer 2, with upwelling and northward cross-equatorial flow in the west. The difference is that the average position of the outcrop of these two layers lies north of the equator, in the central and eastern Pacific—that is, water in these density classes is only intermittently present to the south of the average outcrop line. Over these regions, the time-mean cross-equatorial flow vectors—and those at 2°N, in Fig. 6b—are southward, that is, equatorward, contrary to what we expect of Ekman fluxes. Diapycnal downwelling occurs as this southward flow crosses the equator. Later we will show a connection between TIWs and this light southward flow.

3) On tropical cells

The model results are consistent with large subtropical cells (STCs) sitting either side of the equator with strong Ekman flows extending beyond 10°N at the surface replenished by equatorward flows beneath. The return flow is a combination of the strong western boundary currents and interior flow to the equator. Within these large cells are tropical cells that are smaller and extend to ±5°.

Traditional descriptions of TCs include poleward surface Ekman flow, downwelling, and then equatorward convergence. In z coordinates our model also has this meridional flow structure (Fig. 8a) and downwelling (not shown). We note here that in the upper thermocline layer and mixed layer 2 (Figs. 6c,d), there is no evidence of diapycnal downwelling, that is, only one arrow tail (cross), as opposed to many arrow heads (dots), which represent diapycnal upwelling. So how do TCs exist?

There are a number of ways that a TC structure can be formed. Water in the upper and lower thermocline flows eastward in the EUC. As it flows eastward, it is also flowing “upward” due to the slope of the thermocline, with no need to change density layer, though much diapycnal upwelling does occur in the region (Brady and Bryden 1987). When it reaches the east, it travels poleward. The region of poleward flow across 2°S and 2°N extends to 160°W in the upper thermocline and 140°W in the lower. This water can then either 1) leave the region past 10°N or S, 2) upwell into the layer above, or 3) flow westward and recirculate back into the EUC. If the water is flowing westward along an isopycnal, it is also traveling “downward,” giving the impression of downwelling in z coordinates.

Within this framework it is possible for a water parcel to continue this recycling pathway a number of times within its own density layer before upwelling into the layer above. It is still possible to see the TCs predicted by Liu et al. (1994), except that now we have shown that TCs can exist without any change in density. This is also consistent with Lu et al.’s (1998) criticisms of TCs not having a mechanism for downwelling. This TC behavior is not the only flow that occurs in the equatorial region. As discussed above, there are the large STCs that dominate the flow in and out of the region. The finer details of the flow are also not shown here; we have only demonstrated possible pathways that a water parcel could travel. It appears however that a parcel could enter the equatorial region either via the western boundary currents or the interior flow and circulate many times in the TC structure with the option of becoming lighter, though this is not necessary.

4) Flow into the Northern Hemisphere

Despite the large inflow to the equatorial region of high density water from the Southern Hemisphere [both from the western boundary current (WBC) and interior], there does not seem to be a direct pathway for the required amount of this Southern Hemisphere flow to continue north across the equator and beyond 10°N.
The majority of the WBC water turns east and flows into the EUC. Flow out of the region past 10°N is mostly in the lightest layer, so there must be a pathway whereby the denser WBC water in the Southern Hemisphere is converted to very light density water.

However, the broad flow pattern that we describe in Figs. 6–8 is very similar horizontally and in density space to that of Wijffels (1993; see discussion in the introduction and Fig. 1). A difference to Wijffels’ results is that we find diapycnal upwelling occurring along much of the circuitous path of water parcels from the southeastern tropical Pacific to the northeastern tropical Pacific. First, water parcels experience diapycnal upwelling between 2°S and 8°S right across the basin as they travel westward, and cross the equator northward in the west. Water parcels that then stay north of the equator experience further diapycnal upwelling. This occurs once these parcels reach the East Pacific, as they shallow and flow north across 10°N. In Wijffels’ diagram, the upwelling occurs primarily near the start of this path, at the eastern end of the EUC in the Southern Hemisphere. Our study has also allowed us to investigate the pathway in more detail, which shows that in the process of moving west, a water parcel may undergo
Fig. 8. Meridional slices of meridional flow averaged over 160° to 100° W, shown from 6° S to 6° N for every third month beginning in January. (left) Flow on depth coordinates (m s⁻¹). Mean position of density contours is overlaid showing every 0.5 kg m⁻³ for the density layers shown in Fig. 2. (right) Flow on density coordinates. Flow is depth integrated between every instantaneous 0.5 kg m⁻³ contour interval at each time interval. Flow is in meters squared per second per density layer, shown for every 0.5 kg m⁻³. The thick black line shows the average value of the surface density for that particular month.
many recirculations into the EUC via the TCs discussed above.

4. Seasonal cycle flow pathways

We now explore the seasonal cycle of meridional flow over the same regions. We begin with the seasonal, depth-integrated meridional flow to show that cross-equatorial flow has a strong temporal component. We will then show that west of 140°W most of this variability occurs in the mixed layer. Already there has been research into features of the equatorial seasonal cycle such as the springtime surge (STS) of the EUC (e.g., Yu and McPhaden 1999) and TIWs (e.g., Baturin and Niiler 1997), however to our knowledge nobody has done this for cross-equatorial flows specifically.

a. Basinwide flow

Seasonal depth-integrated flow is quite noisy in this model, particularly east of 140°W (Fig. 7a). Integrating the flow from the east to get a “net” volume flux shows that this noise in the east is inconsequential (Fig. 7b). In the west, however, there are significant seasonal differences.

In April, a depth-integrated southward cross-equatorial flow appears at around 140°–130°W. This southward signature then propagates westward, reaching 160°E by September. At other times, the flow west of 140°W is northward across the equator.

From November to April the net flow is northward across the equator (Fig. 7b). Then from April to November, the net flow becomes quite strongly southward. The timing of this southward flow shows some correlation to the annual Rossby wave that crosses the Pacific at 5°N (Yu and McPhaden 1999).

A further demarcation is made at the 1023.5 kg m⁻³ isopycnal to separate the flow into a “top layer” from the surface to 1023.5 kg m⁻³ [i.e., down to the base of mixed layer 2, (Figs. 7c,d) and a “bottom layer” from 1023.5 to 1026.5 kg m⁻³ (Figs. 7e,f)]. It is clear that the seasonal variability of the cross-equatorial flow is mostly accounted for by flow within the top layer. In the top layer east of 140°W, the flow is generally southward all year. West of 140°W the southward flow propagates westward from April as seen in the depth-integrated flow. Note also that this layer is relatively thin east of 140°W and therefore cannot be responsible for much of the flow. The lower layer must account for most of the spatial variability, however on zonal integration much of this detail is lost.

b. Contribution of tropical instability waves

Typical TC structure, as found in previous studies, can be seen in the model by averaging flow from 160° to 100°W (Fig. 8, left). The same flow constructed along isopycnals and between instantaneous density layers shows a very different picture (Fig. 8, right). In the first half of the year we find that a TC-like structure still exists (poleward flow in the lighter layers and equatorward beneath). In the second half of the year the TC-like structure is no longer clear. Instead a strong southward flow develops at the surface, originating from very light water in the Northern Hemisphere of less than 1023 kg m⁻³.

This viewpoint however still has limitations. The figures only show two-dimensional meridional flows in the north/south direction. The sources feeding these flows can either be from zonal flow (i.e., the EUC weakening and strengthening) or from upwelling. So it is not possible to determine if this light southward flow in the second half of the year is continuing southward and becoming denser or feeding the SEC at the surface.

We do not construct a meridional streamfunction, as done by Hazeleger et al. (2001), as this incorporates basinwide information, including strong western boundary currents, while the features we are interested in, TCs, only exist from about 160° to 100°W. Hazeleger et al. also only looked at the annual mean from which they concluded that the TCs do not exist. Here our monthly analysis shows that they do exist in the first half of the year. In the second half they are masked by a surface southward flow.

Bolus transport may explain some of this lighter surface flow at the surface of the ocean in the second half of the year. An example of surface meridional flow is shown in Fig. 9 for a snapshot at 24 September 1993 in the model. Tropical instability waves sit just north of the equator in addition to other waves along and south of the equator. The overlaid contours of surface density, which show perturbations from 1023 kg m⁻³, have a strong correlation to surface meridional velocity. This figure suggests that when southward flow occurs between 2° and 5°N, it transports lighter water southward while northward velocities advect denser waters. Of course this is a highly nonlinear region with a lot of wave activity, and so southward flow can be caused by many different features.

Closer inspection shows that in the second half of the year when TIWs are present, there is a strong correlation between surface flow and the surface density (Fig. 10). In the first half of each year there is no clear connection between the two. The correlation, for example, over the second half of 1995 is 0.48. We label this bolus transport because of the distinct correlation between density and velocity. To further justify the flow as bolus transport via its true definition involving the depth of
the isopycnal, Fig. 10b shows the same flow but now for the period June 1995 to March 1996, and the thickness of the isopycnal, which in this case is the distance from the surface to the depth of the 1023.5 kg m$^{-3}$ isopycnal, is overlaid. Clearly when the flow is northward, there is a tendency for the depth of the 1023 kg m$^{-3}$ isopycnal to be zero, that is, all the southward-flowing water has density less than 1023 kg m$^{-3}$. When the flow is southward, the depth of the 1023 kg m$^{-3}$ isopycnal is non-zero. So sporadic “fingers” of lighter water spill from the north toward the equator, and it is the time average of these that is seen in the profiles of Fig. 8, in the lightest layers in July to January.

5. Discussion and conclusions

Observations of meridional equatorial flow in the eastern and central Pacific are limited, and where they do exist, they are often contradictory. For this reason we have used results from a global ocean model to try to get an indication of how water might flow in this region.

Our analysis of the flow is limited by studying it in large boxes over large density layers, so there is most likely further detail still to be found. Increasing the number of boxes however also introduces a higher level of complexity when trying to interpret the results. While the long-term mean and seasonal pathways are fairly well understood, Fukumori et al. (2004) have already shown, with a particle tracking model, that high variability in this region is important in allowing flow to “short-circuit” these mean flows and seasonal pathways. The advantage of our analysis over a particle-tracking approach (a true Lagrangian analysis) is that we can determine the instantaneous density of the flows.

Nevertheless, we have found that within our scale limitations, over the region 140°W to the coast of America, cross-equatorial flow is generally southward over all density layers and throughout the seasonal cycle. We did not find a northward flow across 2°S as shown in Sloyan et al. (2003) and are inclined to believe it does not exist, in particular because their error bars are so large for meridional flows—not large enough,
however, to incorporate all of our findings. Our results for annual mean southward flow are supported by Sverdrup theory, which is a key indicator of flow patterns in the ocean. Sloyan’s study used an inverse model that relied on fairly simplistic conditions for meridional velocity that we feel are not justified. Johnson et al. (2001) also found a net northward flow in the top 300 m of the ocean from their observations, again with large error bars. Generally observations would be considered more reliable than model output. In the case of the equatorial Pacific however the flow is so noisy and the time series limited as to introduce many errors and bring the observations into question. We, as well as Hazeleger et al. (2001), have demonstrated the importance of measuring the flow on isopycnals, which was not done in Johnson et al. (2001) and may have introduced further errors into their results.

One striking discrepancy between our model and earlier work is the surface flow in the east. Theory (e.g., Philander and Delecluse 1983) and observations (e.g., Kessler et al. 1998) suggest that the northward wind in the east should lead to a northward cross-equatorial surface flow. ACOM3 however has a southward flow. We are unsure as to why the model behaves this way; it is possibly due to an overly strong meridional pressure gradient in the north acting against the wind-forced flow, which is examined further in Brown et al. (2007).

By partitioning the flow into boxes and by density layer we were able to explore, in detail, many of the

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**Fig. 10.** (a) Surface meridional flow at 3°N and 120°W for the period June 1994 to July 1996 (m s⁻¹; filled line), and surface density (kg m⁻³), at the same position, where it is different from 1023 kg m⁻³ (dotted line). (b) Surface meridional flow (cm s⁻¹) from June 1995 to March 1996. Dotted line is the thickness of the isopycnal, which in this case is the distance from the surface to the depth (m) of the 1023 kg m⁻³ isopycnal.
complex flows in the equatorial Pacific. One pathway of interest is how water flows from the Southern Hemisphere to the north. We found evidence in the model to support the observed pathway found by Wijffels (1993; Fig. 1). Wijffels found that water upwelled along its travels through the EUC, surfaced, and then traveled southward and westward before crossing into the Northern Hemisphere in the mixed layer of the western to central Pacific. The resolution of our model output has allowed us to study her conclusions in more detail and we find further circulations that water parcels may follow between the EUC and the northward surface flow. On annual mean, we found upwelling into lighter density layers occurring not only in the EUC but also as water traveled westward in the SEC(south) and SEC(north). The change in density of the water most probably occurred at the surface in the westward-flowing SEC, where the water was exposed to surface heating. Our analysis on density layers also revealed that the water parcels did not necessarily remain outside the EUC once they left it in the east. Some of the water that leaves the EUC is able to recirculate through the SEC and rejoin the EUC farther west. We found possible pathways whereby parcels could continue this recirculation until either upwelling into the layer above or leaving the equatorial region, in a modified version of the tropical cell (TC) concept as discussed by Liu et al. (1994). This pathway is ultimately linked with TCs, discussed below.

After these recirculations as a water parcel becomes less dense, it is able to travel northward between 160°E and 160°W in the mixed layer aided presumably by the wind stress to alter its potential vorticity, as predicted by Wijffels in the mechanism discussed above. On seasonal cycle, however, this flow is strongly variable with northward flow in March and southward in September. This flow variability is an important addition to Wijffels findings as it limits the opportunities for the northward cross-equatorial flow.

To simplify understanding we have included a schematic of the flow pathways in Fig. 11. We make no claim that this schematic fully describes the flow in the equatorial Pacific; we use it merely as a tool to help explain some of the pathways we are interested in. Four density layers are shown with the previously defined light layer and mixed layer 1 combined in Fig. 11 (top); below that is mixed layer 2, followed by the upper thermocline and the lower thermocline. The deep layer discussed previously has been omitted as it has minimal interaction with the layers above. The red and blue arrows represent the zonal flows from the EUC and SEC, respectively. The black and purple arrows both show flow within a density layer. Note that in space, each layer shoals to the east. Because of this slope of the density layers, as water flows within the layer it has two components, horizontal and vertical. This means that water flowing westward also has a downward component and vice versa. The light brown arrows represent diapycnal upwelling and downwelling. Note that we have generally only found upwelling in these density layers, apart from the top layer where there is diapycnal downwelling from the lightest layers to mixed layer 2. In subsequent layers the light brown arrows represent upwelling from that layer to the one above. The region over which the upwelling occurs is shaded in light brown.

In the upper and lower thermocline, the schematic (Fig. 11) shows the recirculation pathway in the layers, that is, water enters from the west and flows to the east, where it turns and flows back to the west. Some continues westward in the SEC while the remaining water recirculates back into the EUC. In each layer there is also upwelling in the east, though over a wider region in the upper thermocline. Water can therefore either continue recirculating within the same density layer or “jump up” to the layer above. Very little water leaves our domain of interest (10°S to 10°N) at these densities.

In the top two diagrams of Fig. 11, which represent the flow in the mixed layer, the flow is somewhat different as the Ekman flow now dominates over the

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Fig. 11. Schematic of flow pathways in the equatorial Pacific Ocean for approximately 5°S to 5°N and from 130°E to the South American coast. This diagram is given as an aid for explaining the flow pathways found in the earlier, more complex Fig. 5. Major flow pathways are indicated for (top) the light layer and mixed layer 1 representing flow from the surface to the 1023 kg m⁻³ isopycnal, (second panel) mixed layer 2 for flow between the 1023 and 1023.5 kg m⁻³ isopycnals, (third panel) the upper thermocline, 1023.5–1024.5 kg m⁻³, and (bottom) the lower thermocline, 1024.5–1026 kg m⁻³. Red and blue represent zonal flows of the EUC and SEC. Black and purple arrows are both flow within each density layer. Light brown arrows represent diapycnal upwelling and downwelling over the light brown shaded regions. The small surfacing of mixed layer 2 (shown in Fig. 6c) is ignored and blended together in the top layers in the bite taken out of the top panel. The light brown downward arrows in the top panel represent mixing of water carried southeast of the “bite” by eddy processes (seasonal and intraseasonal). The light brown shading shifts from the far east Pacific in the upper and lower thermocline, to the west and North Pacific in mixed layer 2, because the top surface of mixed layer 2 is the ocean surface in the southeast part of the region. This diagram is purely a schematic of some of the possible flows evidenced in Fig. 6.
equatorward geostrophic flow. In mixed layer 2 water enters either from upwelling from the layer below or from the west in the lighter part of the EUC. At the eastern boundary it turns and flows westward, however, there is no recirculation back into the EUC, as the Ekman flow is now stronger than the equatorward geostrophic component. Instead it continues westward in the SECs. Right at the equator there is southward flow east of about 160°W and northward flow to the west.

The lightest layers (Fig. 11, top) outcrop in the south and east. Across this boundary a southward flow appears, which we relate to a seasonal bolus transport.

Our research also allowed us to explore these smaller TCs. Earlier work on the cells focused on a downwelling component that existed at around ±5° from the equator. Our analysis of the annual mean flow on density layers shows that very little diapycnal downwelling occurs. As water flows westward in the SEC, there are two routes it can take. First if it flows adiabatically, it must travel “downward” with the sloping isopycnal. In depth coordinates this would appear as a downwelling even though the density of the water need not change. Alternatively the water parcel could be warmed or undergo mixing so that its density is lowered sufficiently for it to stay at the surface as it advects westward into a lower density regime. The result is “diapycnal upwelling” into the layer above, as can be seen in Fig. 6 for the top few layers (this forms part of the upwelling in the Wijffels recirculation method discussed above). Of the water parcels that subduct beneath the surface (away from the region of Ekman influence), they are forced equatorward by geostrophy where they return to the eastward-flowing EUC. So in the model the TCs still exist but in a more three-dimensional context, much like the cells described by Liu et al. (1994). This pathway also addresses the problem that Lu et al. (1998) had with TCs not having a mechanism for cooling and sinking. We note that Brady and Bryden (1987) estimated diapycnal upwelling in the region of 0.75°N–0.75°S, 150°–110°W to be strongest at a depth of about 100 m at 1 × 10⁻⁵ m s⁻¹. This value equates to less than a Sverdrup in our 2° × 20° boxes, which is even weaker than our estimates.

Hazeleger et al. (2001) showed that TCs do not, or barely, exist on an annual mean, Lagrangian-type, density framework. We studied meridional velocity in a similar framework but for the seasonal cycle (Fig. 8), which revealed that the cells do exist at some times of the year. In our study we did not construct a meridional streamfunction like Hazeleger, which would account for these differences. A streamfunction necessarily incorporates the strong western boundary currents and gives a total basin perspective of the flow. This does not allow any conclusion about what is happening in the eastern part of the basin alone.

In the first half of the year, the meridional flow does suggest a poleward flow in the lightest layers with a return equatorward flow beneath. As discussed above, the flow need not have a diapycnal downwelling to connect the two flows. In the second half of the year, a southward flow appears from 5°N through 5°S, overlaying what may still be some remnant of TCs. Hazeleger et al. (2001) suggested that the disappearance of TCs in this framework was due to eddies, and indeed in the second half of the year TIWs appear and may be the mechanism that counteracts the TC flow.

The southward surface flow in the second half of the year may be some type of bolus transport related to the TIWs. We showed a strong correlation between the southward flow and low density water that appeared to be related to the structure of the TIWs, causing the southward flow mentioned above (Fig. 9 and Fig. 10).

Ideally this study needs to be repeated for model output that runs over a longer time period. We were limited to eight years of output, many of which were El Niño years in which TIW activity is reduced. We also did not take evaporation and precipitation into account. While we do not believe that these will have a dramatic impact, they should be investigated.

We have been able to show that the meridional flow, at least within our model, is very complex, and caution should be taken when averaging. It is very important that the flow is studied in an isopycnal framework with the flow calculated on instantaneous density layers, or else the results can be misleading. We hope that eventually an observational system that is able to instantaneously measure density and velocity so that the effect of the highly complex and three-dimensional TIWs and their resulting bolus transports can be accounted for may be implemented in the equatorial Pacific. This describes the kinematic component to our work. Our accompanying papers attempt to understand the dynamics of the flow, that is, they examine why the meridional flow behaves as it does.

Acknowledgments. We thank Drs. William Kessler and Trevor McDougall for their advice throughout this project. Enormous thanks are given to Dr. Russ Fiedler for computing assistance. Figure 1 was provided by Dr. Susan Wijffels, and data for Fig. 4a were provided by Dr. Greg Johnson.

REFERENCES


