Interannual variability in northeast Pacific circulation

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Received 21 April 2005; revised 16 August 2005; accepted 28 December 2005; published 4 April 2006.

[1] Interannual variability of the circulation in the northeast Pacific Ocean is explored through a joint analysis of expendable bathythermograph (XBT) and expendable conductivity-temperature-depth (XCTD) data, satellite altimetry, and output from a model that was constrained by ocean data. XBT temperature profiles with high spatial resolution are available in the eastern North Pacific along two repeated transects. These ship tracks, along with the coast of North America, define a closed “box” which is used to study the time-mean circulation and its variability on interannual timescales. Geostrophic velocities from XBT data are compared with geostrophic velocities from model output as well as the full model velocity fields. Correlations in variability on interannual timescales between transport in the subpolar gyre and in the subtropical gyre are present in both model output and data. The nature of the variability, and its relation to the changes of the strength of the North Pacific Current (NPC), which supplies the water for both gyres, are explored. Interannual variability in gyre transport is found to be related to both the bifurcation of the NPC, resulting in an anticorrelation in transport between the two gyres, and to variations in NPC strength, resulting in simultaneous changes in the two gyres. The dominant signal is found to be a long-term increase in the NPC, which results in a strengthening of the subtropical gyre. Possible connections with local-scale wind stress changes and with the El Niño/Southern Oscillation phenomenon are also explored.


1. Introduction

[2] The present work investigates the mean state and interannual variability of the circulation of the eastern North Pacific. As shown schematically in Figure 1, the eastward-flowing North Pacific Current (NPC) supplies source water for both the subpolar gyre and the subtropical gyre. The Gulf of Alaska and the California Current region are both important upwelling zones with high productivity. A description of the large-scale circulation of the northeast Pacific region is an important step toward understanding the biologically and economically important elements that are imbedded in it.

[3] Historically, coastal processes and their impacts on economically important fisheries have been the focus of much research in the northeast Pacific [Royer, 1998; Hickey, 1998; Chelton et al., 1982]. For example, the collapse of the California sardine fishery in the 1940s resulted in the establishment of the California Cooperative Oceanic Fisheries Investigations (CalCOFI) program [Bograd and Lynn, 2003]. This project now provides one of the longest continuous oceanographic time series available. Other research in the region focuses on the Pacific Decadal Oscillation (PDO), a large-scale climate phenomenon affecting much of the North Pacific Ocean [Mantua et al., 1997]. One motivation for the study of the PDO was the covariability between salmon catches in Alaska and those in Washington and Oregon.

[4] Most previous in situ studies in this region consist of individual hydrographic transects or short-term surveys which provide synoptic views of the large-scale circulation [Reed, 1984; Musgrave et al., 1992]. These studies provide very little information on long-term variability. The CalCOFI data set described above has long temporal extent, but its spatial coverage is limited. Another long time series of in situ data in this region is a hydrographic line of 13 stations from British Columbia to Ocean Station Papa (located at 50°N, 145°W), known as Line P, which has been occupied regularly since 1959. These data have been used for a variety of studies from short-term observations of El Niño effects [Freeland, 2002] to 25-year analyses of dynamic height variability [Tabata et al., 1986]. Like CalCOFI, the spatial extent of Line P is small relative to the two-gyre system considered here.

[5] In a notable study on long-term variability in the northeast Pacific, Chelton and Davis [1982] examined coastal sea level and observed a coherent rise and fall on interannual timescales along the full North American coast. This led the authors to hypothesize “a quasi-permanent transport of the West Wind Drift in the central North Pacific which bifurcates in the eastern North Pacific either with most of the transport turning northward or most of the transport turning southward” as a possible mode of long-
term variability in this region. This concept is represented schematically in Figure 2a: The input through the North Pacific Current is constant, but the bifurcation of the transport varies on interannual timescales. Other possible modes of variability are also presented. In Figure 2b, variability in the two gyres (and therefore across the XBT lines) is a direct result of variability in the source waters in the NPC. Changes in the volume of source water are split proportionately between the two gyres; the bifurcation does not change. In Figure 2c, in the upper layer, transport into the region differs from transport out of the region. In this case, the upper layer volume changes on interannual time-scales, as net volume is either stored in the box or exported from the region.

In this study we examine the circulation of the North Pacific and the connections between the subtropical and subpolar gyres in the context of these interannual modes of variability. The study is based on a joint analysis of observational data and output from a global ocean circulation model constrained by most of the available ocean observations from 1992 to 2002. Each mode of variability will be considered, independently and in combination with the others, in order to determine its relative contribution to the total variability, and to characterize variability in this region as fully as possible.

In situ data, remotely sensed data, and model output are used in this analysis. Each of these contributes to the overall description of the region but has its limitations. In situ data provide direct measurements of some aspects of the circulation, at specific times and locations, with known uncertainty. The main data set here consists of two repeated tracks of XBT data, a zonal section from San Francisco, California, to Honolulu, Hawaii, and a meridional section from Valdez, Alaska, to Honolulu, Hawaii. These lines are referred to as PX37 and PX38, respectively (Figure 1). Along these lines, spatial resolution is high, but elsewhere, in situ measurements are sparse. Remotely sensed data are also available. These data have good spatial and temporal coverage and resolution, but provide information only at the surface; information about the subsurface structure must be inferred indirectly. A third information source, model output, also has good spatial and temporal coverage, but resolution is lower than data. In an attempt to reduce uncertainties, models have been developed that assimilate data while maintaining dynamical consistency. Here the results from the “Estimating the Climate and Circulation of the Ocean” (ECCO) data assimilation effort [Stamper et al., 2002] will be used; see A. Köhl et al. (Interannual to decadal changes in the ECCO global synthesis, submitted to Journal of Physical Oceanography, 2005) (hereinafter referred to as Köhl et al., submitted manuscript, 2005) for a detailed description of the model, the assimilation approach, and large-scale aspects of interannual variability in the solution. At this point, formal uncertainties of the estimates are unknown and difficult to determine, owing to limitations in computer resources. An important aspect of the present study is therefore to determine the model’s skill in the North Pacific through comparison with observations, in order to subsequently use the model to study components of the ocean circulation in that region that are otherwise difficult to observe. For that purpose, the data and the model output are compared within our study region. They are then used in a combined interpretation to obtain a better representation of the region’s circulation.

The structure of the paper is as follows: In section 2, we describe the data and the model. Section 3 presents methods of comparing the model output with the data. Section 4 describes the mean state of the circulation as well as its variability, as derived from both model and data. Discussion of the results is in section 5, and conclusions are presented in section 6.

2. Data and Model
2.1. XBT and Satellite Data

The primary in situ data source in this analysis is the High Resolution Expendable Bathythermograph (HR_XBT) Program. XBTs are deployed from merchant ships along commercial shipping routes. The data consist of temperature profiles, with a nominal depth of 800 m, having high along-track resolution of 30–50 km in the open ocean and 10 km near boundaries or interesting features, such as the rough topography of the Hawaiian Ridge. The temporal resolution of approximately four cruises per year is only marginally adequate to resolve the seasonal cycle, but should be sufficient to resolve low-frequency variability, with the caveat that temporal aliasing and eddy noise can mask
long-period changes. On each cruise, there are occasional expendable conductivity-temperature-depth (XCTD) casts, which provide salinity information for the calculation of density. Their spatial resolution is lower, approximately one cast per 500 km. Temperature profiles are interpolated onto a uniform grid with resolution of 10 m depth by 0.1° latitude (for meridional sections) or longitude (for zonal sections), over a depth range of 800 m. In the lower latitudes, salinity is determined from a historical T-S relationship, and the resulting density profiles are used to determine geostrophic velocities. This process is described in detail by Gilson et al. [1998]. At higher latitudes, where temperatures are lower, more accurate salinity measurements are needed to maintain the accuracy of the density estimates, because of the nonlinearity of the equation of state. The process is complicated further by a subsurface temperature inversion in the Gulf of Alaska which renders T-S relationships ambiguous [Lagerloef, 1994]. Therefore, north of 38°, a depth-salinity relationship is used to determine salinity in the upper 300 m instead of a T-S relationship. The climatology used is a combination of data from the World Ocean Atlas 2001 and the XCTD drops from our cruises. To determine the accuracy of this process, dynamic height relative to 700 m was calculated for individual XCTD drops and compared with dynamic height from XBT temperatures and salinity from the D-S/T-S relationship. Since XCTDs were not included on a cruise-by-cruise basis, they can be considered independent for the purpose of this comparison. For the 150 XCTDs with data to 700 m or more, the RMS difference was 2.9 cm, only slightly worse than errors at lower latitudes found by Gilson et al. [1998]. Variations in the error were not dependent on either latitude or time.

Data collection began in 1991 on line PX37 and 1993 on line PX38 and continues to the present; data through 2002 are used in this analysis.

[10] Satellite altimetry, which provides high temporal and spatial coverage unavailable from in situ data, is also employed in this project. Sea surface height (SSH) has been measured continuously by TOPEX/Poseidon (T/P) since 1992. Precise geoid information is not yet available, so absolute SSH cannot be determined. However, SSH anomaly is well suited for studies of variability. The present

Figure 2. Possible modes of variability of the northeast Pacific. (a) Amount of incoming water remains constant, but the ratio of transport going north to that going south changes in time; output variability depends on the bifurcation. (b) Volume of water entering the box changes, and transport in each of the two gyres changes proportionately. In this case the north/south ratio is constant and output variability depends only on input variability. (c) Volume of upper layer water leaving the box is not the same as the volume entering the box. In this case, upper layer storage in the box or net outflow occurs.
study is based on a merged SSH product from T/P, ERS-1, ERS-2, and Jason, provided by AVISO [Ducet et al., 2000]. This product is an objective analysis in space and time of all those data sets onto a regular spatial grid with 1/3-degree resolution and a time step of seven days. For our analysis, we created monthly mean fields by averaging the 7-day grids.

2.2. Model

[11] The model output used here is obtained by the ECCO Project by constraining the ECCO model with most of the available basin- and global-scale data sets. The goal of ocean data assimilation is to synthesize in situ and satellite data with the dynamics embedded in ocean models to obtain the best possible dynamically consistent description of the changing ocean. Stammer et al. [2002] provide details of the assimilation approach. The assimilation was performed on a global 1° grid over the period 1992 through 2002 by bringing the ECCO model into consistency with the ocean data using the model’s adjoint. Data constraints included satellite altimetry, surface drifter velocities, and hydrographic information from conductivity-temperature-depth sensors (CTDs), moorings, and floats (Kölh et al., submitted manuscript, 2005). In addition, monthly means of temperature and salinity were constrained by the Levitus 1994 hydrographic fields [Levitus and Boyer, 1994; Levitus et al., 1994]. The model’s initial conditions and daily surface forcing fields were adjusted to bring the model into consistency with most available data sets. The model drift in temperature and salinity over the 11-year period was also constrained at each grid point to avoid numerical drift away from observed changes in hydrographic conditions. This constraint is required to reduce drift resulting from, for example, surface forcing errors. However, it does not eliminate observed changes in, for example, SSH, as discussed in detail by Kölh et al. (submitted manuscript, 2005).

[12] The ECCO model output used here is provided on a global 1° grid as monthly averages over the 11-year period. Temperature, salinity, sea surface height, and all components of the velocity field are included in the output. Because the surface forcing required to bring the model into consistency with the data is estimated during the assimilation procedure, surface wind stress, heat flux and freshwater flux are also part of the solution. See Stammer et al. [2004] for a discussion of the quality of the estimated surface fluxes.

3. Approach

[13] Comparisons between data and model results are an essential step in gaining an understanding of the skill of the model as well as its limitations. However, performing a meaningful comparison is complicated. The inherent differences between what was measured and what the model resolves must be considered. The ocean contains variability from processes taking place on a wide range of spatial and temporal scales, and XBT casts are quasi-synoptic point measurements. In contrast, model fields have 1° spatial resolution; eddies and other mesoscale features are not resolvable on these scales. Temporal variability is further reduced by the use of monthly means. Thus the model fields are significantly smoother than the data in both time and space. In addition, although some drift in temperature or transport might occur in the ocean, constraints reduce the drift in the model. It is important to keep this in mind when evaluating the model’s performance against long timescales in the data (see Kölh et al. (submitted manuscript, 2005) for a detailed discussion of long-term changes in the model).

[14] Velocity is calculated from XBT casts by combining temperature measurements with historical salinity, corrected with XCTDs, to determine density (see Gilson et al. [1998] for details). Horizontal density gradients are used to calculate cross-track geostrophic velocities, as

\[
-\Phi' = \int_0^p \phi dp = \int_0^p \left( \frac{1}{\rho(S, T, p)} - \frac{1}{\rho(35, 0, p)} \right) dp 
\]  

(1)

\[
u_g(p) - \nu_g(p_0) = \left( \frac{1}{f} \right) \left( \frac{\partial}{\partial x} (\Phi'(p) - \Phi'(p_0)) \right) 
\]

(2)

\[
u_g(p) - \nu_g(p_0) = \left( \frac{1}{f} \right) \left( \frac{\partial}{\partial x} (\Phi'(p) - \Phi'(p_0)) \right) 
\]

Here \( \rho \) is the density, \( p \) is the pressure, and \( f \) is the Coriolis parameter. For comparison, model temperature and salinity fields were interpolated onto the same locations as the data. The interpolated model fields were used to calculate geostrophic model velocities, which are directly comparable to the geostrophic velocities calculated from the data. In addition, the model has complete velocity fields that include an ageostrophic component. For each XBT track, the component of model velocity perpendicular to the track was extracted, and compared with the geostrophic velocities calculated from the model and the data. These comparisons reveal both the magnitude of ageostrophic components such as Ekman flow, and the difference between data and model results.

[15] Geostrophic velocities are calculated relative to a “level of known motion” (\( p_B \) in equation (2)) of 800 m, the nominal maximum depth of the XBTs. Velocities were calculated relative to zero flow at 800 m, and also relative to model velocity at 800 m. A comparison between the total horizontal flow into the box and the model’s vertical velocity fields was used to determine which reference velocity resulted in a more physically consistent picture. If upper layer volume is converging in the box, physical consistency requires vertical transport to be downward, out of the box. This would lead to anticorrelation between horizontal and vertical transport. Correlations were higher when the reference velocity was taken from the model fields. Accordingly, all transports shown below are calculated relative to the model’s velocity at 800 m. Figure 3 shows the magnitude and variability of the component of transport resulting from the velocity at 800 m.

[16] Ocean velocities and the associated transport can be decomposed into a geostrophic component (\( T_g \)), an Ekman component (\( T_{Ek} \)), and a residual ageostrophic component (\( T_a \)).

\[
T = T_g + T_{Ek} + T_a 
\]  

(3)
The left-hand side \( T \) can be calculated directly from model velocity fields. However, only the first term on the right-hand side can be determined from the data. Geostrophic volume transport is the flux of volume through an area due to geostrophic velocities.

\[
T_g = \int \frac{\nabla^v_x}{\rho_f} \cdot dA,
\]

The second term on the right-hand side, the Ekman transport \( T_{Ek} \), can be calculated from the ECCO-estimated wind stress according to

\[
T_{Ek} = \frac{\tau^y}{\rho_f} - \frac{\tau^x}{f}.
\]

Here \( \rho \) is seawater density, \( f \) is the Coriolis parameter, and \( \tau^y \) and \( \tau^x \) are the components of wind stress. For our analysis, the wind stress was interpolated to the XBT lines, the Ekman transport was calculated at each location along the line, and the component of this transport normal to the ship track was subsequently extracted. This cross-track Ekman transport \( T_{Ek} \) has a magnitude of \( 1.4 \pm 0.65 \) Sv (1 Sv = 1 Sverdrup = \( 10^6 \) m\(^3\) s\(^{-1}\)) on line PX37 and \( 0.56 \pm 0.68 \) Sv on line PX38. The non-Ekman ageostrophic component of transport \( T_a \) was computed as the residual that remains after the geostrophic \( T_g \) and Ekman \( T_{Ek} \) parts were removed from the total transport \( T \). \( T_a \) is small, with a magnitude of only \( -0.55 \pm 0.36 \) Sv on line PX37 and \( -0.16 \pm 0.39 \) Sv on line PX38.

\[{17} \] In the remainder of this paper, cross-track volume transports are estimated in three ways for each calculation.

\[
T_1 = \int_{-800}^{0} v^m dA,
\]

\[
T_2 = \int_{-800}^{0} \left( v^g + v^m(-800) \right) dA + T_{Ek}.
\]

\[
T_3 = \int_{-800}^{0} \left( v^g + v^m(-800) \right) dA + T_{Ek}.
\]

The first estimate, \( T_1 \), is based on the cross-track component of the model’s full velocity field \( v^m \). The second estimate, \( T_2 \), uses geostrophic velocities from the data \( v^g \), relative to model results at 800 m \( v^m(-800) \), to determine cross-track transport. Ekman transport \( T_{Ek} \), determined from the model’s wind stress, is added to this quantity at each time step. For the final estimate, \( T_3 \), geostrophic velocities calculated from model output \( v^g_m \), relative to model results at 800 m, are used to estimate transport. As with the estimate from data, an Ekman component is added.

\[{18} \] Model output is linearly interpolated in space to the locations of the XBT casts using the month closest to the time of the observations. During the comparisons of results from the model and the data, only those months when XBT data are available are considered. During 1992–2002, 42 sections along line PX37 and 30 sections along line PX38 were recorded. Although the cruises are approximately evenly spaced in time, it is possible that some features of

![Figure 3. Component of transport resulting from using the model velocity at 800 m as a reference velocity instead of zero, across lines (a) PX37 and (b) PX38. In each case, the thin solid line shows the time series of monthly transport estimates. The thick solid line shows the same time series, subsampled at only those months when XBT data are available. The thin and thick dashed lines show the same results for the full and subsampled time series after smoothing using a 12-month running mean.](image-url)
the flow are misrepresented or missed entirely as a result of the low sampling rate. In order to determine the effects of both spatial and temporal aliasing on the transport, the A VISO SSH data were used. Geostrophic surface transport along the lines was determined from A VISO SSH data. Since A VISO SSH data are available for each month, the calculation is made first using all available months, and then using only months when XBT data are available. As shown in Figure 4a, along line PX37, the two A VISO-based estimates match closely, indicating that aliasing from mesoscale features is not a problem. Along PX38, there are significant differences, indicating that aliasing in this region is important (Figure 4b). The implications will be discussed later in this paper.

[19] Since this analysis is focused on interannual variability, it is helpful to accentuate lower frequencies by filtering out higher frequency components. Monthly estimates were produced through linear interpolation of the data and high-frequency variability was subsequently removed with a 12-point running filter. Model output was processed

Figure 4. A VISO-based estimates of surface transport using all data points (circle markers) and only using XBT months (thin dashed lines) are shown for (a) PX37, (b) PX38, (c) AKC, (d) NPC, and (e) NHRC. Also shown are the three estimates of transport anomaly (mean removed), scaled to the same approximate magnitude as A VISO-based transport anomalies. Differences between thin dashed lines and lines with circle markers are solely effects of aliasing.
the same way: it was subsampled to the months when XBT data are available and smoothed with a 12-month filter. The following discussion will be based on those smoothed results.

4. Circulation

Figure 1 shows the mean transport stream function in the top 800 m, as calculated from model output. Transport from the full model velocity field is integrated across the region, relative to zero at the coast, according to

\[ \Psi(x, y) = \int_{\text{coast}}^{z} T_u(x, y, z) \, dz = -\int_{\text{coast}}^{z} T_v(x, y) \, dx. \]  

The stream function is computed over all months when XBT data are available during the eleven-year period covered by the model, 1992–2002. The expected two-gyre structure is evident, with the eastern half of the anticyclonic subtropical gyre covering the basin between latitudes of about 15°N and 45°N, and the cyclonic subpolar gyre mostly contained within the Gulf of Alaska. About 15 Sv are transported by the subtropical gyre, while the subpolar gyre carries about 9 Sv. Because transport from the XBT data can only be computed across ship tracks, the spatial structure depicted in the figure cannot be compared to data directly. Comparisons are limited to cross-track transports.

Figure 5 shows upper-layer (0–800 m) cross-track volume transport as a function of time. Transport across line PX38 (around 9 Sv), transport across line PX37 (around −9 Sv), and total lateral transport into the box (PX37+PX38) are shown. Each quantity is estimated from the full model field, from the data using geostrophy, and from the model using geostrophy (see equations (6)–(8)). Positive values indicate flow into the box (northward flow across PX37 or eastward flow across PX38). The large magnitude of interannual variability in the region is apparent. Net transport across PX38 is positive, dominated by the eastward-flowing NPC, while net transport across PX37 is negative, dominated by the subtropical gyre exporting water from the region. Data-based estimates of the time-mean gyre transport determined from cross-track transports give results similar to those from the model transport stream function in Figure 1: about 18 Sv in the subtropical gyre and about 9 Sv in the subpolar gyre. The net upper layer transport into the box estimated from the full velocity fields has a mean value of 1.03 Sv and a standard deviation of 1.02. The downwelling of −1.00 ± 0.88 Sv calculated from the model’s vertical velocity fields confirms that in the time-mean, volume is balanced in the box. Geostrophic estimates of horizontal transport are significantly larger: 2.39 ± 1.79 Sv in the data, and 1.73 ± 1.09 Sv in the model. This suggests that the full flow fields are required for budget estimates.

Across line PX37, geostrophic upper layer transport in the model closely matches that in the data, with a root-mean square (RMS) difference of 0.88 Sv. The estimate from the full model velocity also agrees well with the data, with an RMS difference of 0.77 Sv. The agreement of these estimates, especially when supported by the evidence from AVISO SSH that aliasing is not an issue on this line (Figure 4a), indicates that the description of the variability here is valid. In contrast, there are substantial differences between model- and data-based transport estimates across PX38. The geostrophic model estimate is slightly closer to the data (RMS difference of 1.91 Sv) than the estimate using the full velocity field (RMS difference of 2.08 Sv). A local minimum in the model estimate in 1994 is not seen in the data, while a peak in the data estimate during 1998 is entirely missing from the model estimates. These disparities result in large differences in total transport. The previous indications of an aliasing problem in this region reinforce the message that the variability will not be well described without further analysis.

In order to examine the differences between the data and the model along line PX38, we divided the line into three segments, distinguished by the direction of flow (see Figure 1), and calculated transport for each section. The northernmost section is the Alaska Current (AKC). This westward-flowing current, north of about 54°N, is the upper arm of the subpolar gyre. South of the AKC, the North Pacific Current (NPC) is the broad section of eastward flow. The lower boundary of the AKC, defined as the point separating westward AKC flow from eastward NPC flow, is constant in time. The flow turns west again south of about 30°N, in the North Hawaiian Ridge Current (NHRC). To find the boundary between the NHRC and the NPC, transport was integrated along the track, starting from Hawaii. The minimum in integrated transport marks the
point where the currents change from flowing westward to flowing eastward, and thus the southern edge of the NPC. This location is highly variable, especially for data estimates of transport. If the boundary between the NPC and the NHRC is not determined correctly, discrepancies could appear in the transport estimates that do not reflect the information in either the data or the model.

[24] Figure 6 shows transport estimates for all three sections. In the AKC, shown in Figure 6a, there is reasonable agreement between model and data results. Both tendency and magnitude agree for most of the time series. The RMS difference between the data and model geostrophic estimates is 0.56 Sv, and instantaneous differences rarely exceed one Sverdrup. This region shows the largest differences between the estimate from the full model field and the model geostrophic estimate, indicating the local importance of friction terms. In this region, the transport anomaly calculated from the AVISO SSH has the same structure as the transport anomalies calculated from the other three estimates (Figure 4c). Also, the difference between the full AVISO SSH time series and the subsampled one was insignificant. This indicates that in this region, aliasing does not change the results obtained from the data.

[25] Differences are larger in the NPC (Figure 6b). The main difference between the estimates is that the magnitude of the data estimate is approximately 4 Sv larger than either model estimate. The slopes are similar throughout the time series, and all three estimates capture a large increase in transport between 1995 and 1999. The standard deviation of the difference between the model geostrophic estimate of transport and the data estimate is 0.81 Sv. The large bias but small standard deviation suggests that there is no fundamental difference between model and data in the structure of variability. Details of the location of the boundaries between the currents could lead to the bias. In this region, aliasing effects, as estimated using AVISO-based surface transports, were again found to be negligible (Figure 4d).

[26] In the southernmost segment of PX38, the NHRC, both a bias and a structural difference are evident (Figure 6c). The data estimate is biased high, by approximately 2 Sv. It is interesting to note that in this case, as in the NPC, the magnitude of the data estimate is larger than the magnitude of the model estimates. However, when the sum of NHRC and NPC transport across PX38 is considered, the magnitudes of data and model estimates are similar (see Figure 5). Unlike the NPC, the slopes of the estimates do not match for the NHRC. The most significant discrepancy occurs in 1998, when the magnitude of transport surges in the model estimate but decreases sharply in the data estimate. We note that it is this discrepancy in the
NHRC estimate that leads to the large difference between model and data estimates of total transport (see Figure 5). In this region, the synoptic nature of in situ data could have a large effect on data-based estimates. Previous observations have demonstrated that “the NHRC appears and disappears on timescales of less than one month” [Bingham, 1998]. As a result, in this region, monthly means may not be adequate for comparisons with synoptic data. In addition, complex small-scale dynamics occur in this region as a result of the rough topography of the Hawaiian ridge that are not resolved by the model’s one-degree grid. As with the NPC, the details of boundary location could lead to problems. In this region, the analysis of the SSH data shows that aliasing is very important in this region. Use of the full time series gives an estimate of transport anomaly that is very different from the subsampled results (Figure 4e). This is a clear indication that aliasing is important in this region. The shape of the surface anomaly transport matches the model estimates of transport more closely than that of the data. Taken together, the conclusion is that the variability in this region is not as well described by the data as it is by the model.

5. Modes of Variability

It is evident that there is strong variability in the upper-layer circulation in this region. In order to explore the variability and possible covariability of the two gyres, we compare their transports (Figure 7). For this comparison, the subtropical gyre is defined as the sum of the transport across PX37 and the transport across the southernmost segment of PX38 (the NHRC). The subpolar gyre consists of the AKC segment of PX38. Since the NPC transport contributes to both gyres, and its bifurcation is undetermined and possibly variable, it is not included in either gyre. In this way, it is possible to distinguish changes in the gyres from changes in the NPC. All estimates in Figure 7 show large variability. In the subpolar gyre, data and model estimates are similar, but in the subtropical gyre there are significant differences. Data estimates of the gyres are uncorrelated with each other, but there are weak positive correlations between the gyres in both the model geostrophic estimate \( r = 0.24 \) and the full model estimate \( r = 0.44 \). This suggests a relationship between the gyres in the model that is fundamentally different than the relationship in the data.

One possible mode of variability, illustrated schematically in Figure 2a, assumes a constant NPC, and attributes the variability of transport in the gyres to changes in the bifurcation of the incoming flow. This would lead to anticorrelation between the two gyres; an increase in southward transport would correspond to a decrease in northward transport. Although the NPC is not constant, its bifurcation can still modulate the gyre transports. In order to isolate the effect of the bifurcation, the transport in each gyre must be normalized by the magnitude of the NPC at each time step.

Figure 8 shows the normalized gyre estimates, or equivalently, the fraction of NPC transport in each gyre. Correlation coefficients of these time series are \(-0.54\) in the data, \(-0.39\) in the full model, and \(-0.34\) in the geostrophic model. While the time series are short, and statistical significance is low, this negative correlation in all cases suggests that a change in bifurcation is occurring. The timescale of the variability is approximately 4 years. The anticorrelation is most evident in late 1997, when a significant El Niño event occurred in the tropical Pacific. A minimum in the normalized subtropical gyre transport, shown in Figure 8, is in agreement with previous studies of transport in the California Current that associate El Niño events with decreased horizontal circulation [Chelton et al., 1982]. Previous studies of El Niño connections with the

![Figure 7](image1.png)

**Figure 7.** Estimates of volume transports in the subpolar gyre (AKC) and the subtropical gyre (PX37+NHRC). The solid lines are the estimates from the data, the dashed lines are the estimates from the model velocity fields, and the dotted lines are the estimates from model geostrophic velocities.

![Figure 8](image2.png)

**Figure 8.** Similar to Figure 7, but showing the gyre transports which have been normalized by the amount of the instantaneous incoming transport (the NPC). Dashed lines represent model estimates from the full velocity fields, solid lines indicate data estimates, and dotted lines indicate model geostrophic estimates. As before, the Ekman transports were added to the geostrophic estimates.
subpolar gyre have been inconclusive [McGowan et al., 1998], but Figure 8 shows a concurrent peak in normalized subpolar transport. While the time series is too short to come to any conclusions about correlations between the 4–7 year El Niño cycle and the changes in bifurcation, it is at least apparent that the idea is worth consideration.

[30] Gyre transports could also covary with the magnitude of the incoming NPC transport (Figure 2b): the transport in each gyre would increase when the incoming transport increased. Figure 9 shows the magnitude of the transports of the NPC (the incoming flow) and of each of the gyres. As in previous plots, the estimates of the subpolar gyre transport agree fairly well. Data estimates for NPC transport and subtropical gyre transport are greater than the model estimates for the same quantities. In the NPC, the difference is a simple bias, but in the subtropical gyre there is substantially more variability in the data estimate than in either model estimate. However, even in the data estimate, a long-term upward trend is evident in both the subtropical gyre and the NPC. Some correlation is apparent between the NPC and the subpolar gyre as well: the main signal in the subpolar gyre is a peak in 1998 which also appears in the NPC. Again, the statistical significance of these correlations is low, the correlations of NPC with gyre transports are shown in Table 1.

[31] The largest signal in this time series is the long-term increase in NPC transport. This robust signal is observed in all three estimates of transport (see Figure 6b), and indicates changes on timescales longer than the 4-year signal in the bifurcation. Previous research has suggested the possibility of decadal variation in this region [e.g., McGowan et al., 1998; Chelton et al., 1982], and the increase in NPC could support these theories. Figure 10 shows the transport stream function, from the model, averaged over the wedge-shaped region shown in Figure 1. SSH from the model output and from the AVISO analysis are also plotted. Interannual variability of more than 3 cm is apparent. The high correlation between steric height and SSH shows that changes are mainly due to density variability rather than mass storage. An exception can be found in 1997, when the SSH anomaly as measured by the model and by the AVISO analysis is almost a centimeter higher than steric height anomaly would suggest, indicating net mass storage in the box. We note that in this

is significantly stronger in the latter period, but the subpolar gyre remains essentially unchanged. The change in stream function is shown in Figure 10c.

[32] For a region in Sverdrup balance, the vertically integrated mass transport can be calculated from the curl of the wind stress as

\[ Hv = \int_{\text{coast}}^{x} \frac{1}{\rho_0 s} \left( \frac{\partial Y_x}{\partial x} - \frac{\partial X_y}{\partial y} \right) \, dx. \]  

(10)

Here \( X_x \) and \( Y_y \) are the zonal and meridional components of wind stress, respectively. Using equation (10), the difference in Sverdrup transport between the two time periods was determined. The result is shown in Figure 10d. This region of the northeast Pacific is narrow enough that these signals could propagate across it in the time period we consider, especially since the wind variability in question occurs in the western portion of the domain, lowering the effective propagation distance of the signal. The similarities between Figures 10c and 10d support the hypothesis that the large-scale changes in the model’s transport stream function are largely due to long-term changes in the local wind-forcing of the east North Pacific.

[33] The modes of variability described so far have been based on the premise that mass is constant in the upper layer of the box, so that the transport out of this layer must be the same as the incoming transport. However, if mass were stored in the upper layer, the variability of outgoing transport would be independent of the incoming volume, as in Figure 2c. This possibility can be explored by considering SSH variability: changes in steric height, which result from density changes; and changes in mass storage in a region, i.e., uncorrelated to any density change. Steric height changes (\( \Delta h \)) are determined as

\[ \Delta h = \frac{1}{g} \int_{-H}^{0} \delta(T, S, p) \, dp. \]  

(11)

Here \( g \) is gravitational acceleration (9.8 m s\(^{-2}\)), \( \delta \) is specific volume anomaly, and \( p \) is pressure. Figure 11 shows a time series of steric height anomalies, from the top-to-bottom model field, averaged over the wedge-shaped region shown in Figure 1. SSH from the model output and from the AVISO analysis are also plotted. Interannual variability of more than 3 cm is apparent. The high correlation between steric height and SSH shows that changes are mainly due to density variability rather than mass storage. An exception can be found in 1997, when the SSH anomaly as measured by the model and by the AVISO analysis is almost a centimeter higher than steric height anomaly would suggest, indicating net mass storage in the box. We note that in this

**Table 1.** Correlations Between Gyre Transport Magnitudes and the Magnitude of the NPC

<table>
<thead>
<tr>
<th></th>
<th>Subtropical Gyre</th>
<th>Subpolar Gyre</th>
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<tbody>
<tr>
<td>Model (full)</td>
<td>0.85</td>
<td>0.71</td>
</tr>
<tr>
<td>Model (geostrophic)</td>
<td>0.81</td>
<td>0.54</td>
</tr>
<tr>
<td>Data</td>
<td>0.58</td>
<td>0.57</td>
</tr>
</tbody>
</table>
region, the model’s density is more strongly dependent on salinity than temperature: The correlation between a time series of halo-steric height and steric height is $\rho_0 \cdot 0.92$ (higher salt leads to higher density which leads to lower steric height), while the correlation between thermo-steric height and steric height is only 0.31. A comparison with in situ data from Line P confirms that the salinity signal in the model has skill and suggests the need for more in situ salinity measurements to study the North Pacific and to constrain models.

Figure 5 shows that the total time-mean horizontal upper-layer transport into the box is not zero. The downwelling transport, calculated from the model fields of vertical velocity, accounts for $-1.00 \pm 0.88$ Sv of transport. When this is added to the horizontal transport, the result is a net inflow of $0.02 \pm 0.61$ Sv. In the model, then, horizontal transport into the box in the upper layer is balanced by vertical transport into a deeper layer and then horizontal deep transport out of the region. As Figure 5 illustrates, the temporal variability of this process is strong. The question of how this variability propagates into the deeper layers of the ocean can be considered using the model’s velocity fields. When the transport is calculated for the full depth of the ocean, following the same process of subsampling to XBT months, interpolating, and then using a 12-point filter, the transport into the box is $0.34 \pm 2.3$ Sv. The large
variability of this quantity is surprising. If, instead of subsampling, all months of model output are used, the transport into the box is \(-0.88 \pm 0.77\) Sv. The reduced variability, signified by low standard deviation, illustrates that the subsample-and-interpolate method used throughout this analysis has the potential to exaggerate some signals and minimize others. This possibility, an effect of aliasing, cannot be corrected. There is still a slight negative bias in the full time series calculation. If the wedge defined by the XBT lines is replaced by a square box, such as that outlined in dashed lines in Figure 1, the need for spatial interpolation of the velocity vectors is eliminated, and the time-mean transport into the box decreases to \(0.1 \pm 0.02\) Sv. These results emphasize that caution must be used when drawing conclusions about the nature of variability from data without including model information, or from model output without validation of data.

6. Conclusions

[35] The objective of this work was to analyze high resolution data sets together with a coarser resolution data assimilating model in order to provide an improved description of the mean and time-varying circulation of the northeast Pacific from 1992 to 2002. Both the model and the data agree in that, on average, in the upper 800 m, about 25 Sv flow eastward in the North Pacific Current between latitudes 30°N and 52°N. This broad eastward current bifurcates west of the North American coast along 47°N. About 8 Sv turn north and follow the coastline of the Gulf of Alaska into the Alaskan Current, while the remaining 17 Sv turn south and then west in the subtropical gyre. The broad gyre circulation reaches as far south as 15°N, where it joins the North Equatorial Current. The flow that recirculates north of Hawaii is referred to as the North Hawaiian Ridge Current. This current has an average magnitude of about 9 Sv, but its standard deviation is 2.5 Sv, or more than 25% of the magnitude of the flow, indicating very high variability.

[36] The largest interannual signal observed during this period, and also simulated by the data-constrained model, was the increase in the transport of the NPC, and the associated intensification of the subtropical gyre. This signal is observed as a rise in NPC and subtropical gyre transport magnitudes through the first half of the time series, as illustrated in Figure 9. Stream functions in Figure 10 demonstrate the structure of the increase in the strength of the subtropical gyre. This evolution is associated with changes in the curl of the wind stress. The difference in 4-year time-mean maps of Sverdrup transport, as calculated from the curl of the wind stress, has the same structure as the difference in the stream function for the same time periods (Figures 10c and 10d). The vertical structure of this change is illustrated by the sections of mean velocity along PX38 prior to and after the change (Figure 12). After smoothing the data over 4 degrees, differences between model and data are still evident. Model output is smoother, and the small-scale structure in the data represents both unresolved structure and noise from temporal aliasing of the in situ measurements. However, both data and model still show intensification of the jet structure at the southern end of the line and an increase in the large-scale incoming transport between about 40°N and 50°N. This signal is an example of the “NPC-dependent variability” illustrated in Figure 2b.

[37] On a shorter timescale, covariability between the gyres indicative of changes in the bifurcation of the incoming current is evident. This is the pattern suggested by Chelton and Davis [1982] as depicted in Figure 2a. To observe this, we compare the outflows of the two gyres. Net transport across PX37 is the part of the subtropical gyre water that does not recirculate north of Hawaii, and hence is the part closest to the California coast described by Chelton and Davis [1982]. The AKC plays the same role in the subpolar gyre. Model and data estimates of these transports are shown in Figure 13. Anticorrelated variability on a 3- to 4-year timescale is evident. The largest increase in subpolar gyre strength occurs at the peak of the 1997–1998 El Niño event. Another way of observing this mode of variability is the fraction of NPC transport, as shown in Figure 8. Again, the largest signal is during 1997–1998. While the eleven years observed here are not enough to find a true correlation on the timescales of ENSO, the similarities indicate a possible teleconnection of the subpolar regions with the tropics.

[38] The third mode of variability illustrated in Figure 2, the storage of upper layer volume in the box, is not evident in this region of the ocean. Model sea surface height, satellite altimetry, and steric height calculated from the model’s temperature and salinity fields are all in good agreement. There is a significant change in steric height, which we find to be mainly a result of changes in salinity rather than temperature. The importance of salinity in the large scale changes in this region underscores the need for more in situ measurements to observe change and help constrain models. As noted by Freeland and Cummins [2005], the Argo project will be instrumental in providing new salinity data in this and other regions.
Overall, the agreement between the model and the data was good with some notable exceptions. To some extent, this is to be expected, since the assimilation approach did incorporate the same data we compare against. However, as discussed by Köhl et al. (submitted manuscript, 2005), the underlying ECCO solution to first order was not constrained by the XBT profile data set we use here, but by the climatological hydrography, the altimetry and other surface data. Taking this into account, our comparison can be seen as a quasi-independent check of the consistency of the ECCO solution with in situ data.

Along line PX37 and in the Alaska Current region, data and model estimates agreed very well both in their mean and time-varying components. However, problems in determining and resolving the boundary between the NPC and NHRC led to differences in mean and time-varying transports between the data and model. When the estimates are summed, the mean bias essentially disappears. The offsets shown in Figures 6a and 6b therefore have to be considered somewhat artificial.

More problematic is the fact that data and model based transport estimates show different temporal variability, particularly in the NHRC. While XBT data have fairly high spatial sampling density, they are available only once every quarter. This temporal sampling is not adequate and will inevitably lead to aliasing of the energetic eddy field. Comparisons of subsampled versus full estimates of geo-
the Ocean (ECCO) funded by the National Oceanographic Partnership Program. The HR XBT data were collected through support from NSF grant OCE00-95248. The participation of many commercial vessels, officers and crew, and SIO shipriders in the HR XBT program is gratefully acknowledged.

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Figure 13. Anticorrelated variability of transport in the Alaska Current and the California Current (the net transport across PX37 is considered to be the California Current in this case). Both data and model estimates show clear anticorrelation, with AKC strengthening when the California current is weak, with timescale of variability of approximately 3 years.

Acknowledgments. Reanalysis surface forcing fields from the National Center for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) are obtained through a computational grant at NCAR. Computational support from the National Partnership for Computational Infrastructure (NPACI) and the National Center for Atmospheric Research (NCAR) is acknowledged. Supported in part through ONR (N00014-99-1-0049), NOAA grant NA17RJ1231 (GODAE, Argo) and by the NASA Ocean Surface Topography Science Working Team through JPL contract 961424. This is a contribution of the Consortium for Estimating the Circulation and Climate