Interannual variability in North Pacific heat and freshwater budgets

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ABSTRACT

Transports of volume, heat, and freshwater in the North Pacific Ocean from 1992 to 2004 are analyzed using a long-term high resolution expendable bathythermography (XBT) dataset and output from a data-assimilating model. Estimates of geostrophic transport from the data are compared with the model transport to close the volume budget north of the trans-Pacific XBT track. Advective transport from both model output and data are combined with surface fluxes to determine budgets of heat and freshwater in the closed region. The northward heat transport across the XBT track is estimated to be 0.74 ± 0.1 pW, and has variability of almost 0.5 pW on 3–4 year time scales, while freshwater transport is estimated to be −0.1 ± 0.06 Sv. The balance between northward advective heat transport and surface heat flux gives a time-varying estimate of heat storage that compares well with observations. A similar balance is found between model estimates of advective freshwater transport and surface freshwater flux. Despite a scarcity of observations and uncertainties in all components, this analysis results in nearly closed budgets of volume, heat, and freshwater. Mean estimates of advective transport of both heat and freshwater agree with previous estimates. An analysis of each component of the heat budget with latitude indicates that a relative lack of time-variability of the surface component is consistent throughout the North Pacific. The dominant advective component is driven by changes in the wind stress curl field. For both heat and freshwater storage, strong signals occur concurrently at all latitudes. This behavior could indicate that these signals are controlled by large-scale dynamics, rather than small-scale disturbances from which signals would need to propagate to be widely felt. The analysis demonstrates the value of bringing models and data together, resulting in budgets that are consistent with observations, yet provide a comprehensive look at the variability of North Pacific heat and freshwater storage that would be unavailable from data alone.

1. Introduction

The transport of heat and freshwater by the large-scale ocean circulation is a critical component in the global climate system. Of the 5 pW of heat carried northward out of the tropics by the ocean/atmosphere system, the ocean circulation is responsible for about 2 pW (Trenberth and Caron, 2001), and interannual variability is significant (Roemmich et al., 2001). A better understanding of the magnitude and frequency of variability in the ocean is essential in determining its role in the climate and ecosystems of the planet. Even synoptic estimates of heat and freshwater transport are difficult to obtain, and the details of variability on longer time scales remain to be explored. As data accumulate, and global and regional-scale models are developed and refined, estimates of mean transports are improving and variability on a range of temporal and spatial scales is being quantified.

The heat transport in the Pacific Ocean has been the subject of several previous investigations. The magnitude of this component of the global heat budget has been estimated both directly from hydrographic transects and indirectly from basin integrals of surface fluxes. Early hydrographic estimates (Bryan, 1962) left even the sign of this transport in dispute. More recent work has determined that heat is transported poleward, from the equator to higher latitudes where heat is radiated to space. Recent direct estimates of heat transport into the North Pacific from hydrography, using ship tracks on or near 24°N have converged between 0.5 and 1.0 pW, with uncertainties ranging between 0.1 and 0.3 pW (Bryden et al., 1991; Macdonald, 1998; Roemmich et al., 2001; Ganachaud and Wunsch, 2003; Talley, 2003; Uehara et al., 2008). Estimates from models show a similar mean value, and indicate that year-to-year changes can reach 0.4 to 0.5 pW (Kawai et al., 2008). Calculating heat transport from top of the atmosphere radiation minus atmospheric heat transport, as in Trenberth and Caron (2001), gives similar results. Integrated air-sea heat flux provides another method for estimating heat transport, using the concept that heat radiated to space at higher latitudes must have been transported poleward from lower latitudes.
latitudes. Estimates made using this method vary broadly depending on the flux product used. For example, for the time period 1992–2001, the ERA-40 product (Uppala et al., 2005) implies 0.31 pW heat transport across 20°N, while the NCEP reanalysis (Kalnay et al., 1996) implies 0.65 pW. These various results show general agreement, but a full understanding of the variability had not yet been achieved.

The oceanic freshwater budget, including storage, transport, and air-sea flux (precipitation minus evaporation, $P−E$) is an important diagnostic for the hydrological cycle. However, the spatial structure of $P−E$ is more complex than that of air-sea heat flux, and relatively few large-scale estimates of freshwater transport have been made. Historically, salinity data are much sparser than temperature data, so there are few estimates of freshwater storage. Indirect estimates of freshwater convergence from integration of air-sea freshwater fluxes are available, but uncertainties in precipitation, evaporation, and runoff introduce large error bars into such estimates. Ganachaud and Wunsch (2003) use hydrographic sections from the World Ocean Circulation Experiment (WOCE) and various surface climatologies to calculate freshwater convergence of $0.14 \pm 0.26 \times 10^8$ kg s$^{-1}$ in the North Pacific. By integrating air-sea fluxes using Bering Strait transport as a reference point, Wijffels et al. (1992) estimated that all freshwater transport in the Pacific Ocean is northward. More recently, Uehara et al. (2008) estimated freshwater transport using a time series of XBT transects, and found a freshwater convergence of $0.08 \pm 0.07$ Sv in the western Pacific and a freshwater divergence of $−0.26 \pm 0.11$ Sv in the eastern half of the North Pacific basin. A comprehensive understanding of time-varying freshwater transport is still an important goal.

In this study, a long time series of high-resolution expendable bathythermograph (XBT) transects and a regional ocean state estimate from a data-assimilating numerical model are used to estimate the time-mean and variability of heat and freshwater transport in the North Pacific between 1992 and 2004. The XBT transect, at approximately 24°N, defines a southern boundary to a closed region for which budgets are calculated (Fig. 1). Temperature and salinity are used to find geostrophic velocities which lead to data-based estimates of advective heat and freshwater transport. These are compared with similar quantities from the model estimate. The model solution also includes heat and freshwater air-sea flux fields. The sum of the advective and surface flux components gives an estimate of the time-varying storage in the region north of the XBT transect. Estimates of heat content from broadscale XBT data provide a consistency check for the heat storage component. The model results are analyzed further to examine the latitudinal variation of heat and freshwater transport, as well as how these properties vary on interannual time scales.

2. Model and data

2.1. Model

This analysis uses output from a data-assimilating model developed by the Estimating the Climate and Circulation of the Ocean (ECCO) Consortium (see Stammer et al., 2002a). This model has been developed to provide estimates of time-varying ocean circulation by constraining the MIT general circulation model with ocean data using the adjoint method. It runs forward to estimate the time-varying ocean state, and then calculates a cost function from the misfit between the resulting state estimate and the constraining data. The adjoint of the model is then used to obtain the gradient of this cost function with respect to the control variables, taken here to be the initial temperature and salinity fields, and the time-varying surface forcing fields. A standard optimization algorithm uses these gradients to adjust these control variables to bring the model into closer agreement with the data while maintaining model dynamical consistency. The process is repeated iteratively to minimize the cost function. This approach to data assimilation is explained in further detail in Stammer et al. (2002b).

To obtain the state estimate used in this analysis, the ECCO model was used in a regional setting in the Pacific Ocean north of 26°S. The estimate was performed for the time period from 1992–2004. Constraining data include hydrography, satellite altimetry, sea surface temperature and salinity, profile data from floats, drifter-derived velocities, and winds from scatterometry. A complete description of datasets included in the assimilation is found in Lu et al. (2002). Although the XBT data are included as a constraint, it is important to compare the assimilation results with this and other input data sets to test the consistency of prior assumptions. The bin-averaging of data onto the model grid prior to assimilation, the large amount of other data assimilated, and the dynamically consistent assimilation approach ensure that the comparison retains some validity. The state estimate used in this analysis results from assimilating all subsurface data with increased weights relative to the surface data, to increase their relative influence on the solution, as described in detail by Douglass et al. (2009). The model has horizontal resolution of 1° in latitude and longitude and 50 vertical levels. The level thicknesses increase from 5 m at the surface to 500 m at depth. Model output consists of daily sea surface height (SSH) fields, monthly means of temperature, salinity, and zonal, meridional, and vertical velocity, and adjustments to the initial conditions and surface forcing as determined by the assimilation. Initial temperature and salinity fields are taken from the Levitus 1994 climatology (Levitus and Boyer, 1994; Levitus et al., 1994), and the NCEP reanalyses of heat flux, freshwater flux, and zonal and meridional wind stress are used for initial surface forcing (Kalnay et al., 1996). The model domain is bounded by land to the north and to the east, but the southern and western boundaries are open to exchange with the South Pacific, and with the Indian Ocean through the Indonesian Throughflow. Temperature, salinity, and zonal and meridional velocity obtained from a global ECCO state estimate (Wunsch and Heimbach, 2007) are prescribed on the boundaries.
2.2. XBT data

Estimates of geostrophic velocity and transport are based on temperature profiles from XBT measurements along ship track PX37/10/44 (Fig. 1). Collection of temperature profiles from a cargo ship along this track from San Francisco, CA, to Honolulu, HI, to Guam, to Hong Kong began in 1991 and continues to the present. Cruises take place approximately every three months. Between 1992 and 2004, the period of this study, there were 49 cruises. These measurements have high along-track spatial resolution, ranging from 50 km in the open ocean to 10 km near shallow topography or interesting features, for a total of more than 300 profiles per transect. The probes measure temperature to a nominal depth of 800 m, with vertical spacing of 10 m. In addition, occasional expendable conductivity-temperature-depth (XCTD) probes and, in recent years, profiles from Argo floats provide salinity data. These sparse salinity measurements (up to 500 km spacing, with a maximum of 12 profiles per transect) are combined with historical measurements to determine the salinity field at the time of each cruise. From temperature and salinity, density and geostrophic velocity relative to 800 m are calculated as described in Gilson et al. (1998).

3. Volume transport

The first step of this analysis is a comparison of the model and data estimates of volume transport, which must be balanced to determine a meaningful estimate of property fluxes.

This comparison demonstrates similarities and differences between the model and the data, and illuminates some of the difficulties encountered when comparing them.

For comparison with the model, simulated temperature and salinity are interpolated spatially to the locations of XBT casts and geostrophic transport is calculated relative to 800 m. It is clear (Fig. 2A) that while there are discrepancies between the total geostrophic transport estimated by the model and that estimated by XBTs, the main features of variability, such as the local maxima in 1995 and 1998, and local minima in 1996 and 1999, are replicated well. The correlation between the model and data-based estimates of geostrophic transport across the transect is 0.65 (here and throughout the analysis, correlations are significant at a 95% confidence level).

Also shown is the transport as calculated from the simulated velocity fields in the top 800 m, with Ekman transport subtracted. The model geostrophic transport is nearly identical to the model transport from the velocity fields. This shows consistency within the model, and confirms that the geostrophic and Ekman components of transport together capture all of the significant variability demonstrated by the full velocity field.

Another issue that must be taken into account when comparing model and data estimates is temporal aliasing. The model output consists of monthly means, which are compared to snapshots from cruises every three months. To quantify the possible effects of aliasing, the model output was subsampled at the times when cruises occurred. Both the subsampled model output and the data were linearly interpolated to obtain monthly estimates, and then smoothed over 12 months to remove annual

![Fig. 2](image-url)
and higher frequency variability. A visual comparison between the subsampled and interpolated version of the model estimate and the full time series shows the impact of this aliasing. Smoothed versions of both the full and subsampled time series are shown in Figs. 2A, for both the model transport estimated from the velocity fields and for that estimated from geostrophy. It is evident that both estimates preserve the main character and variability of the transport, but the differences in transport can reach a Sverdrup (1 Sverdrup = 1 Sv = 10^6 m^3 s^-1) or more.

Geostrophic velocities are calculated relative to a reference velocity at a level of known motion. In this case, the reference level used is 800 m, the nominal depth of the XBT casts. Rather than assuming zero velocity at 800 m, the model provides an estimate of velocity at the reference level. The magnitude of transport resulting from applying this velocity throughout the top 800 m is −0.14 Sv with a standard deviation of 1.62 Sv (Fig. 2B). This reference velocity has large horizontal variability as well (Fig. 3). It is evident from the large positive velocity at the western boundary that the Kuroshio extends deeper than 800 m. This reference velocity is small along the remainder of the track, mainly negative in the eastern and western segments and close to zero in the central segment. Discontinuities at 145°E and 155°W indicate ports of call. Transports in the west section appear to be relatively stable, with standard deviation reduced to 0.55 Sv.

A final issue to be considered with the comparisons made here concerns the track used for the XBT cruises. In 1999, the endpoint of this track changed from Taiwan to Hong Kong. The change can be seen in Fig. 1, which shows all cruise tracks throughout the time series. To fully enclose the region north of the transect, and to ensure consistency between the cruises, the transport through the Taiwan Strait must be considered. Using the model, the transport through the Taiwan Strait is found to be consistently northward, with a mean of 2.2 Sv and a strong seasonal cycle leading to standard deviation of 1.2 Sv. This is similar to, although slightly higher than, prior research indicating that mean annual transport through the Taiwan Strait is highly variable but averages about 2.0 Sv (Wang et al., 2003; Wu and Hsin, 2005; Jan et al., 2006). This transport occurs at warm temperatures and therefore will contribute to heat transport. The model estimate of northward temperature transport is 0.12 ± 0.18 pW, which agrees relatively well with an observational estimate of 0.13 to 0.24 pW (Jan et al., 2006). The model estimate of 84.5 ± 37.1 kg/s of salt transport compares reasonably well with previous measurements of 53.33 to 81.74 kg/s (Jan et al., 2006). Because model estimates are generally consistent with previous measurements, these estimates are added to the XBT-derived volume, heat, and freshwater estimates for cruises terminating in Taiwan.

The sum of geostrophic transport from observations, relative to the model's velocity at 800 m, and Ekman transport calculated from the model's wind fields, provide a data-based estimate of transport in the top 800 m of 0.92 ± 2.05 Sv (Fig. 2D). (Here and throughout this paper, uncertainties are provided as the standard deviation of the time series after smoothing over one year). The full time series of transport from the model has a mean of −2.1 ± 1.3 Sv. When that time series is subsampled to the times when XBT data are available prior to smoothing, the resulting estimate is −2.5 ± 1.5 Sv. All three estimates have similar features such as a peak in 1996, another in 1998, and a slow increase from 2000 through 2005, and standard deviations of all three estimates indicate that interannual variability is larger than a Sverdrup.

To diagnose some of the differences between model and data estimates of total transport, the track is decomposed geographically. The west (Hong Kong to Guam), central (Guam to Honolulu), and east (Honolulu to San Francisco) sections are separated by the ports of call along the cruise track. Transports in the west section (Fig. 4A) are −6.1 ± 3.95 Sv and −6.3 ± 5.45 Sv for the model and data, respectively. While the means are similar, the model smooths each of the peaks of variability, leading to its smaller standard deviation. Transport estimates in the central section (Fig. 4B) are 14.0 ± 4.15 Sv and 17.4 ± 5.75 Sv for model and data, respectively. This section shows the highest magnitude of transport as well as the highest variability. The data-based estimate is distinctly higher for much of the time series. In the east section, the model estimate of transport is −10.0 ± 1.1 Sv and the data estimate is −10.2 ± 2.2 Sv (Fig. 4C). The magnitude is still large, but variability is smaller, as are model-data differences. Differences between model and data are shown explicitly in Fig. 4D. The differences in the west section, which includes the Kuroshio, are the largest, reaching magnitudes of almost 10 Sv. Differences in the central section are also significant. Further, it is evident that the model-data differences in these two sections are antecorrelated (r = −0.75).

The mean streamfunction (Fig. 1) shows that transport in the central segment arises from westward-flowing North Equatorial Current (NEC), while the west section includes the NEC and the Kuroshio. Fig. 4 shows that the model has lower mean across-track transport in the central section. This indicates that the core of the NEC is about 1° lower in latitude in the model, with a corresponding fraction of the NEC transport flowing westward south of Guam instead of north of it. Thus, the data estimate of northward transport in the central section is higher than the model estimate. In the west section, large southward transport is apparent as the NEC flows westward across the path of the

![Fig. 3. Cross track reference velocities from the model results at 800 m. Velocities are plotted for each time step in gray, with the mean overlaid (thick black line). Discontinuities at 145°E and 155°W indicate ports of call.](image-url)
XBt section. This is balanced by the northward transport in the Kuroshio. Prior research has shown that a higher bifurcation latitude of the NEC leads to a proportionately smaller northward transport in the Kuroshio (Qu and Lukas, 1996). Accordingly, in the west section, the model's smaller and farther south NEC transport (which appears as southward transport across this section) is balanced by a larger Kuroshio transport, while the larger and farther north NEC in the data is balanced by a smaller Kuroshio transport. Thus, a small southward offset in the southern boundary of the subtropical gyre explains the anticorrelated variability of the central and western sections. While the mean difference between the model and data estimates of across-track transport affects estimates of mean heat transport, discussed below, the variability is similar in model and data.

4. North Pacific heat budget

The goal here is to develop a full heat budget for the North Pacific, beginning with the region north of the XBT transect. Positive values indicate heat into the closed region, either through advective transport at the boundaries or through air-sea flux of heat at the ocean surface. The sum of the advective and surface flux components is equal to the heat storage in the region, which can also be calculated as the time-rate of change of the heat content. Each component is discussed below.

4.1. Advecove component

For both the Ekman and geostrophic components of the volume transport budget, corresponding temperature transports are calculated by integrating the product of velocity and temperature. The model and data estimates of geostrophic temperature transport across transect PX37/10/44 are shown in Fig. 5A. The correlation between the model and data estimates is 0.64. The most significant feature in this time series is the peak in 1998, which is most likely associated with the El Niño event occurring in 1997–1998. Another peak in 1995 is larger in the XBT estimate than in the model, but is present in both. Both model and data also show an increase in geostrophic transport from 1999 to 2000. For the Ekman component (Fig. 5B), the sea surface temperature is used to approximate the temperature of the transport. In the case of the model, sea surface temperature is the temperature in the shallowest grid box, centered at 2.5 m depth, while in the data the shallowest XBT measurement (the zero measurement) is used. This approximation was shown by Wijffels et al. (1994) to introduce negligible error into heat transport estimates. Fig. 5B shows only the model estimate, because SST is so similar between model and data that the two estimates are indistinguishable. The Ekman temperature transport variability is highly correlated with the Ekman transport variability, indicating that volume changes are dominant in this component.

Mass (or in the case of this model, volume) must be balanced to determine heat transport. Although the net southward geostrophic transport in the upper 800 m is largely canceled by
Large volumes of the warmest water transport in the North Pacific consists of a shallow overturning model estimate. The structure of total meridional temperature volume conservation in the data, and the full velocity fields for the transports are used as in the heat budget, that is, the “best transport occurs, for the model and the data. The same heat warm flow is compensated with a geostrophic return flow in the northward Ekman component, there is a small residual (Fig. 2D). In the model, where transport in the top 800 m has a mean of $-2.1 \pm 1.3$ Sv, compensation for this unbalanced transport occurs below 800 m. In the data, where the mean transport is smaller and more variable with $0.74 \pm 2.3$ Sv, compensation might either occur in the top 800 m due to a reference velocity adjustment, or deeper. In terms of heat transport, there are two extremes for how this data-based volume imbalance could be compensated. The first is a barotropic compensation. This depth-independent velocity would have a mean temperature equal to the mean top-to-bottom along-track temperature of the ocean, calculated using the model to be 2.9 C. The other extreme, hereafter the baroclinic balance after Roemmich et al. (2001), is achieved by adjusting the 800 m reference velocity in the warmest portion of the XBT transect, near the Kuroshio. There, the mean 0–800 m temperature is 14.3 C. After balancing the volume transport, the heat transport is then the sum of the temperature transports from the geostrophic, Ekman, and balance components. We take the mean of the barotropic and baroclinic cases as the “best” estimate (Fig. 6), and the spread between them as one measure of the uncertainty of this component. The resulting total advective heat transport is estimated to be $0.66 \pm 0.11$ pW from model output and $0.81 \pm 0.11$ pW from XBT data.

A final perspective on advective heat transport is obtained by examining the temperature structure of the mean volume transport. Fig. 7a shows the temperature classes in which transport occurs, for the model and the data. The same heat transports are used as in the heat budget, that is, the “best estimate” of geostrophic plus Ekman plus the adjustment for volume conservation in the data, and the full velocity fields for the model estimate. The structure of total meridional temperature transport in the North Pacific consists of a shallow overturning cell (Roemmich et al., 2001). Large volumes of the warmest water are transported north, both as Ekman transport in the mixed layer, and at the western boundary within the Kuroshio. This warm flow is compensated with a geostrophic return flow in the thermocline, at a broad range of cooler temperatures (Fig. 8). The warmest classes of water also have the largest standard deviation, indicating a higher level of variability. This is another indication of the influence of the highly variable Ekman flow, which occurs in the warmest surface waters, as opposed to the cooler and more stable geostrophic return flow.

4.2. Surface flux

The net surface heat flux, integrated over the region north of the XBT track, is always negative, indicating that heat is lost from the ocean to the atmosphere in that area (Fig. 6). This is consistent with poleward advective heat transport in the ocean. Both the model (0.45 ± 0.08 pW) and NCEP (0.62 ± 0.10 pW) estimates of surface heat flux show distinctly less variability than the geostrophic and Ekman terms. This indicates that variability in heat storage north of this XBT transect results from the imbalance of a strongly variable advective heat transport with a relatively stable surface heat flux.

4.3. Heat storage

In any closed volume, heat storage results from the residual between advective heat transport into the volume and surface net heat flux out of the volume. Storage can be calculated from the model results as the sum of these terms, or as the time-rate of change of heat content in the region. The change in heat content anomaly determined from a combination of broadscale XBT data and satellite altimetric height (Willis et al., 2004, hereafter WRC) is used to calculate heat storage for comparison with the model estimate. This heat content product has been corrected for a systematic bias due to XBT fall-rate errors as outlined in Wijffels.
et al. (2008). As shown in Fig. 6, the model estimate of storage is biased high relative to the WRC estimate. The storage from the model estimate is \(0.21 \pm 0.17\) pW, indicating a mean increase in heat content. Storage from the WRC product, on the other hand, has a mean of \(0.00 \pm 0.13\) pW. This difference in mean heat transport indicates a drift in model temperature. Similar numerical drifts have been noted in other ECCO applications (Köhl et al., 2007), and the drift in this application is discussed in Douglass et al. (2009). While there is some evidence of an increase in temperature in this region from 1992 to 2004 (Levitus et al., 2005), the drift observed here is more likely a numerical effect. Despite this, as was the case with volume transport, the variability in the two estimates of heat storage is similar, with a correlation of \(r = 0.68\).

4.4. Total heat budget

The terms of the heat budget for the region north of PX37/10/44 are shown in Fig. 6. Advecive heat transport is northward, with a magnitude of \(0.66 \pm 0.11\) pW in the model and \(0.81 \pm 0.11\) pW in the along-track XBT data. Net surface heat flux into the atmosphere north of the transect is estimated to be \(0.45 \pm 0.08\) pW in the model, which is smaller than the NCEP estimate of \(0.62 \pm 0.10\) pW. These lead to an estimate of \(0.21 \pm 0.17\) pW of heat stored in the model. This is compared with storage from the WRC product, which is found to be \(0.00 \pm 0.13\) pW through direct analysis. Taken as a whole, the heat budget analysis demonstrates the large interannual variability in heat transport in the North Pacific. Variability of air-sea heat flux is secondary to that of advective heat transport, which can change by as much as \(0.5\) pW over a 2–3 year span. These changes in advective heat transport, while integrated air-sea heat flux stays relatively constant, lead to large variability in heat storage north of the transect. Both the observation- and model-based estimates show changes with a range of almost \(0.7\) pW.

The meridional structure of mean heat transport can be estimated from the model. This calculation also places the current analysis in the context of other recent estimates of North Pacific heat transport, from hydrographic transects and from integrated surface flux (Fig. 9). The difference between the advective transport from the model and the integrated model air-sea flux indicates mean storage in the North Pacific, which could be explained by temperature drift in the model. The estimate from the model’s advective field is similar in structure to other estimates, particularly the estimate made by Trenberth and Caron (2001) based on the ECMWF reanalysis with constraints from top-of-the-atmosphere radiation data. The integrated model surface flux is somewhat lower than most of the other results, with the exception of the unadjusted NOC (Josey et al., 1998) and the Woods Hole (Yu et al., 2008) surface heat flux fields, which have substantially lower net heat loss by the ocean north of the XBT transect.

A comprehensive view of the heat budget of the North Pacific comes from comparing the spatial and temporal variability of all terms concurrently. Fig. 10 shows temporal and spatial variability of northward heat transport, integrated surface flux, and heat storage estimated from the model (Figs. 10A, B, and C.
respectively), and heat storage estimated from the WRC product (Fig. 10D). At all latitudes, the time-variability of the integrated surface heat flux is much smaller than that of the advective heat transport. This shows that conclusions drawn from the heat budget north of PX37/10/44 (Fig. 6) hold throughout the North Pacific. The boundary between the subtropical and subpolar gyres is evident in the advective heat transport near 35 N. Changes are coherent from the equator through 35 N, indicating the strong connection between equatorial and subtropical variability. In this region, advective heat transport has significant variability, as does heat storage. Conversely, in the subpolar gyre, advective heat transport is much less variable, and the changes that do occur are uncorrelated with the changes that are evident in the subtropical and equatorial regions.

The comparison between the two estimates of heat storage is instructive. The strongest feature in both figures, an increase on the order of 1 pW, coincides with the El Niño event in 1998. This event causes an increase in advective heat transport (Fig. 10A) while the integrated surface heat flux (Fig. 10B) remains unchanged. This increase reaches to nearly 20 N almost immediately, and disappears abruptly, rather than propagating as time progresses. Smaller equatorial and subtropical features show this trait as well. Equatorial signals in heat transport extend from the equator through the subtropical gyre to as far north as 35 N, but not beyond, reinforcing the concept of the separation of the subtropical and subpolar gyres with respect to heat transport, which is echoed in the integrated heat storage estimate (Fig. 10C). While this border between the gyres has been found in other estimates of heat transport, such as those shown in Fig. 9, it is possible that it is made more distinct in the model synthesis due to the absence of eddy noise. The WRC estimate (Fig. 10D) has similar character, but the transition between the gyres is less distinct, and all features are of lower magnitude than those in the model estimate. Both estimates indicate that heat storage in the equatorial region and the subtropical gyre can change by as much as 0.5 pW over a time span as short as two years.

5. Freshwater budget

To estimate a freshwater budget, the region north of PX37/10/44 is considered first, and then the analysis is expanded to the full North Pacific. As with heat, the advective and surface flux components are discussed separately, and then combined to give a full, time-varying budget of freshwater in this region. Throughout the analysis, positive values indicate freshwater into the ocean.

5.1. Advective component

Determining the advective freshwater transport is less straightforward than calculating its heat-related counterpart. Because the variability in salinity is small compared to its mean, salt transport is highly correlated with volume transport. To isolate the salinity signal, an equivalent freshwater flux is calculated. To do so, we assume that salt is conserved in any enclosed volume:

\[ \rho \ast V_{in} \ast S_{in} = \rho \ast (V_{out} - FW) \ast S_{out} \]  

(1)

In this equation, \( V_{in} \) is incoming volume transport, \( S_{in} \) and \( S_{out} \) are incoming and outgoing salinity, respectively, and \( \rho \) is the density of seawater. Solving Eq. (1) for \( FW \), we obtain the amount of freshwater needed to account for the difference between salinity of incoming transport and salinity of outgoing.
In the integral form of these relations used here to calculate freshwater transport, $S_{in}$ and $S_{out}$ are transport-weighted averages.

As was the case with temperature, the salinity data from the HR-XBT dataset are limited to the top 800 m. Salinity is determined from a combination of historical data, XCTD casts, and profiles from Argo floats where available. For more information on determination of salinity, see Gilson et al. (1998). As with temperature, most variability occurs in the top 800 m. A comparison between the model estimates of full depth equivalent freshwater transport, and the equivalent freshwater transport in the top 800 m, is shown in Fig. 11. Differences are small relative to the magnitude of the signal and its variability, indicating that the upper 800 m is adequate for analysis of freshwater transport variability on the interannual time scales and basin-wide spatial scales considered here.

Model and data-based estimates of equivalent freshwater transport are shown in Fig. 12A. The model estimate of freshwater transport is $-0.14 \pm 0.06$ Sv, while the data estimate is

\[ FW = V_{in} \frac{S_{out} - S_{in}}{S_{out}} \]  

(2)

In the integral form of these relations used here to calculate freshwater transport, $S_{in}$ and $S_{out}$ are transport-weighted averages.

Fig. 10. Latitude-varying heat budget of the North Pacific. (A) Northward heat transport. (B) Integrated air-sea heat flux into the North Pacific. (C) The change in model heat content. (D) The change in heat content from the WRC product. Units are pW, and contour intervals on all plots are 0.2 pW. Thick black line is the zero contour and shaded areas indicate negative values.

Fig. 11. The thick solid line shows the full depth equivalent freshwater transport, while the thin solid line shows the top 800 m equivalent freshwater transport. The dashed line shows the difference.
both mean and variability of these time series is encouraging. This feature is much more distinct in the data. The similarity in increased exported freshwater between 1998 and 1999, although minimum in the data several years later. Both show a feature of decreases initially, and then increases in the latter part of the time export freshwater from the region. The magnitude of this export correlation of 0.58. Both indicate that the effect of advection is to increase surface freshwater flux (precipitation, evaporation, and runoff) is estimated to be 0.07 ± 0.08 Sv, which is slightly lower than the NCEP estimate of 0.09 ± 0.04 Sv (Fig. 12B). The changes that take place in surface freshwater flux over the course of the time series are gradual, an increase through about 1999 followed by a decrease through the remainder of the time series. Unlike surface heat flux, the magnitude and time scale of variability of integrated surface freshwater flux is similar to that of advective freshwater transport and integrated freshwater storage.

5.3. Freshwater storage

The model estimates mean freshwater storage (advective plus surface components) to be −0.07 ± 0.10 Sv (Fig. 12C). There are

--0.11 ± 0.06 Sv. The two estimates have similar variability, with a correlation of 0.58. Both indicate that the effect of advection is to export freshwater from the region. The magnitude of this export decreases initially, and then increases in the latter part of the time series, with minimum in the model estimate in 1995 and a minimum in the data several years later. Both show a feature of increased exported freshwater between 1998 and 1999, although this feature is much more distinct in the data. The similarity in both mean and variability of these time series is encouraging.

Further understanding of the advective freshwater transport comes from examining the evolution of the salinity profiles in time. Along-track salinity is averaged spatially and temporally to obtain a mean profile. At each time step, this mean profile is subtracted from a spatially averaged profile, to obtain a time series that shows the evolution of salinity anomaly as a function of depth and time. Results from model and data are shown in Fig. 13. Anomalies of up to 0.1 psu occur in the model and the data. Both the model and the data have lower than average salinity in the top 200 m from 1995 to 1998, followed by a distinct shift to above-average salinity in the top 200 m until 2002. Anomalies are larger and extend deeper in the model than in the data, which is probably due to the finer-scale resolution of the data. An interesting aspect of these features, evident in both the model and the data, is that they originate at the surface, and then propagate to depths of 200 m or more over time. The same phenomenon of surface salinity anomalies being subducted has been observed at Station Aloha just north of Hawaii, and was attributed to the subduction and southward advection of winter rainfall anomalies (Lukas, 2001; Stammer et al., 2008).

As with temperature, the mean structure of the transport can be understood through a breakdown into salt classes (Fig. 7B). Overall, there is southward transport in the saltiest and freshest classes, with northward transport in the mid-range salinity classes. High standard deviations indicate significant variability in the transport with salinity higher than 34 psu, but are still less than the magnitude of the transport in most locations, leaving no doubt about the direction of the transport. The data indicate strong northward transport at around 34.5 psu, while model components have considerably smaller magnitude. Compensation occurs in the highest salinity classes, where the data again show larger magnitude than the model, but in the southward direction. Transport with salinity below 34.0 psu, on the other hand, is consistently southward. As a percentage of transport, the variance is approximately the same as in the higher salinity classes. In the North Pacific, at the latitude of the XBT track, salinity this low is found almost exclusively in the California Current, the eastern boundary current of the subtropical gyre. Computing the transport in the geographical sections (not shown) confirms that this low-salinity transport is exclusive to the east section of the XBT transect. Standard deviation smaller than the magnitude of the transport indicates the consistency of this current in transport and salinity on interannual time scales, in agreement with previous research (Schneider et al., 2005).

5.2. Surface flux

The model estimate of surface freshwater flux (precipitation, evaporation, and runoff) is estimated to be 0.07 ± 0.08 Sv, which is slightly lower than the NCEP estimate of 0.09 ± 0.04 Sv (Fig. 12B). The changes that take place in surface freshwater flux over the course of the time series are gradual, an increase through about 1999 followed by a decrease through the remainder of the time series. Unlike surface heat flux, the magnitude and time scale of variability of integrated surface freshwater flux is similar to that of advective freshwater transport and integrated freshwater storage.

The model estimates mean freshwater storage (advective plus surface components) to be −0.07 ± 0.10 Sv (Fig. 12C). There are
not enough salinity data to estimate salt content directly. As a function of latitude, model-estimated mean meridional freshwater transport (Fig. 14) is compared with the integrated surface freshwater flux from the NOC surface flux dataset (Josey et al., 1998), and with previous estimates of freshwater transport from hydrographic surveys. Unlike heat flux, the surface freshwater fluxes are not integrated relative to zero at the northern boundary. Using Wijffels et al. (1992) estimate of 0.8 Sv at 32.5 psu transport through the northern boundary, along with the model estimate of mean salinity of 34.5 at the southern boundary, Eq. (2) leads to an equivalent freshwater transport of 0.04 Sv out at the northern boundary. This is in agreement with Talley (2008)”s estimate of 0.07 ± 0.02 Sv of transport through the Bering Strait, so 0.04 Sv is used as a fixed reference point for the integration of surface freshwater flux.

The time evolution of the meridional equivalent freshwater transport at each latitude is shown in Fig. 15A. In comparison with the advective heat transport shown in Fig. 10A, the advective freshwater transport is relatively stable throughout the time series. The heat transport showed an oscillation, on a time scale of approximately 3 years, between the equator and 15 N, in direction of heat transport. In contrast, fresh water is transported northward between about 10N and 20N, and southward at higher latitudes. Again, Wijffels et al. (1992) concluded from an integration of surface freshwater flux southward from the Bering Strait that freshwater transport in the Pacific should be northward everywhere, in disagreement with these results. However, in this model, the Bering Strait is closed to transport. These results have the same magnitude of convergence and divergence as Wijffels et al. (1992), and future model analyses should correct this deficiency.

The integrated surface flux varies with latitude, but like the advective component it is consistent in both direction and magnitude for most of the time series, with precipitation dominating near the equator and in the Northern Pacific and evaporation dominating in the subtropical gyre (Fig. 15). As in the comparison between estimates, the integration uses 0.04 Sv of exported freshwater through the Bering Strait as its endpoint. Although the closure of the Bering Strait in the model is not ideal, prior research indicates that while freshwater is transported through the Pacific to the Bering Strait, this transport does not affect convergence within the region (Talley, 2008). The most significant signal in integrated surface flux is a decrease in $P - E + R$ reaching south to the equator during the El Niño event in 1998. The main features of integrated freshwater storage include a decrease in storage of freshwater between 20 N and 30 N in the first 2–3 years of the time series and again during El Niño and an increase in approximately 2002 (Fig. 15). Overall, in freshwater, opposing signals cancel out to ensure the conservation of salt throughout the region.

6. Forcing

The North Pacific Ocean has a highly variable heat budget, with fluctuations of up to half a petawatt in storage on a time scale of 3–4 years (Fig. 6). This variability in storage arises mainly from changes in the temperature of advective heat transport. Broken down geographically, most transport in the central section of the XBT track is in the NEC, while transport in the west section includes the NEC and the Kuroshio. Changes in the latitude of the NEC bifurcation result in corresponding changes in the transport across these two segments of the track. Such shifts have been associated with El Niño events (Qiu and Lukas, 1996). When the NEC transport is at its largest (during El Niño events), bifurcation occurs at a higher latitude. This leads to a smaller transport volume in the Kuroshio, and a lower net northward transport west of Guam.

Here, volume transport is correlated with temperature transport. In the high El Niño phase, the increased westward NEC transport in the central section consists of warmer water. In the western section, the higher NEC bifurcation latitude and corresponding smaller Kuroshio transport lead to a smaller northward temperature transport. These combine to create a strong correlation between the Southern Oscillation Index (SOI) and the temperature transport in the central and western sections of the transect (Fig. 16). Correlations are visually apparent and statistically significant. For the model estimate, correlation between the temperature transport in the west section and SOI is 0.86, while temperature transport in the central section is anticorrelated with the SOI at −0.86. In the data estimate, correlations with the SOI are 0.76 for temperature transport in the west section and −0.77 for temperature transport in the central section.

These two segments are both strongly affected by El Niño, but in a canceling sense. There is no clear correlation between the SOI and the total net heat transport at the latitude of the PX37/10/44 transect, but the conclusion that the SOI strongly affects North Pacific heat transport is supported by the latitude-varying heat budget as shown in Fig. 10. There are increases in northward heat transport corresponding with El Niño, but the maximum latitude is no more than 25 N, indicating the extent of the direct effects. Although it does not extend throughout the North Pacific, and in the case of freshwater transport, it does not even reach the latitude of the XBT transect, El Niño is associated with the most significant signals in freshwater and heat storage alike.

According to Sverdrup dynamics, the meridional transport across a zonal section should be proportional to the integrated wind stress curl across that section. Fig. 17 shows the integrated wind stress curl for zonal sections from the eastern edge of the basin extending to Honolulu and Guam. Correlation with estimates of meridional transport for the same segments from
the model velocity fields is evident. From Guam (latitude 13°N) to the eastern boundary, this correlation is 0.69, and in the eastern section (latitude 22°N) the correlation is 0.78. These results imply that wind stress curl is a dominant factor in the dynamics of the North Pacific, as suggested by previous research (e.g. Deser et al., 1999; Qiu, 2003).

7. Conclusions

This analysis uses 13 years of in situ data and output from a data-assimilating model to quantify the mean and variability of heat and freshwater transport in the North Pacific Ocean. By combining the advective transport and air-sea fluxes into a closed region, budgets have been calculated which characterize the nature and magnitude of interannual variability of large-scale temperature and salinity fields, and their relationship to variability in circulation. Once the level of agreement between the model and data estimates of the budget for the closed region north of the XBT transect was established, a full North Pacific view of the time and latitude varying budget was developed for both heat and freshwater.

This analysis provides a nearly closed budget of volume transport in the top 800 m for the North Pacific, for both mean and time variability. The mean northward heat transport is 0.66 ± 0.11 pW in the model and 0.80 ± 0.11 pW in the XBT data. These agree with previous estimates of northward heat transport in the North Pacific. The surface component of heat flux has also been quantified. Variability in the advective component dominates the total storage variability in the heat budget. Analysis of each component of the heat budget with latitude indicates that this relative lack of variability in the surface component of heat flux is consistent throughout the North Pacific. The advective component varies strongly through the equatorial region and the subtropical gyre, but within the model the subpolar gyre does not appear to exchange heat with lower latitudes. In all regions, the advective component dominates. At the latitudes of Guam (13°N) and Honolulu (21°N), close to the XBT transect, Sverdrup dynamics indicate that changes in the wind stress curl field are the controlling factor in volume transport, and thus heat transport, throughout the North Pacific.

A budget for freshwater was calculated as well. Results indicate a time-mean advective transport of freshwater across the XBT track of −0.14 ± 0.06 Sv from the model estimate and

![Fig. 15.](image-url)
rather than small-scale disturbances from which signals would need to propagate to be widely felt.

The agreement of the model and the data is an encouraging sign that model-based syntheses have improved in quality, and that more can be learned from using models and data together than from using either alone. In particular, it provides some validation that the expanded latitude and time-varying budgets shown in this analysis provide meaningful representations of large-scale budgets of heat and freshwater transports in the North Pacific over the last decade. For the climate-related study provided here, the dynamical consistency of the assimilation approach is essential, as it is for many other climate applications of ocean state estimation. Much emphasis is needed on improving ongoing efforts, by increasing the model resolution, but also by enhancing the control space, e.g., by including ocean mixing parameters (see Stammer, 2005). Another remaining problem is the drift in model temperature, which to some extent is not supported by data. If the mean storage terms should, in fact, be smaller, then we are left with some ambiguity in whether storage bias is balanced by bias in air-sea flux or ocean advection terms.

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