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UNIVERSITY OF CALIFORNIA, SAN DIEGO

**Large-scale ocean circulation, dynamics, and air-sea exchanges:
Argo observations of the mean and time-varying ocean**

A dissertation submitted in partial satisfaction of the
requirements for the degree
Doctor of Philosophy

in

Oceanography

by

Donata Giglio

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2014

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The dissertation of Donata Giglio is approved, and it is acceptable in quality and form for publication on microfilm and electronically:

Chair

University of California, San Diego

2014

DEDICATION

To my grandparents Elvira, Emilio, Assunta and Sesto.

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ABSTRACT OF THE DISSERTATION

**Large-scale ocean circulation, dynamics, and air-sea exchanges:
Argo observations of the mean and time-varying ocean**

by

Donata Giglio

Doctor of Philosophy in Oceanography

University of California, San Diego, 2014

Dean Roemmich, Chair

The large-scale ocean circulation, dynamics, and air-sea exchanges are investigated, based on observational datasets including Argo and satellite altimetry, and viewed in the framework of modern theoretical ideas. Initially, the wind-driven interannual variability of the subtropical North Pacific is described. Using the extensive Argo dataset, it is seen that the North Pacific gyre varies in the strength of its interior circulation and in its spatial orientation, on interannual timescales. Also, satellite altimetry shows variations in sea surface height that are consistent with Argo steric height changes, and enabled a temporal extension of the 2004-2011 Argo study back to 1993. The next part of the thesis is focused on annual steric variability, including diabatic changes in the surface layer due to air-sea buoyancy

fluxes and adiabatic changes due to wind-forced advection, which are dominant in the subsurface ocean. The annual signal in subsurface (200-2000 db) steric height zonally averaged over a season shows good agreement with the wind-forced vertical advection contribution, both in the global ocean and in different basins. This study, the first on global scale, also shows that the annual vertical advection extends deep into the water column. Next, the global pattern of climatological monthly heat and freshwater fluxes at the ocean surface is estimated using Argo temperature and salinity profile data for the period 2004 to 2013. The air-sea flux estimates from Argo are described in global maps and basin-wide integrals, in comparison to atmospheric reanalysis data and to air-sea flux products based on observations. This ocean-based estimate of surface fluxes is consistent with property variations in the subsurface ocean and indicates greater amplitude for the climatological monthly heat flux values in the subtropics compared to other products. Similarly, the combination of Argo freshwater flux and reanalysis evaporation, suggests greater amplitude for climatological monthly precipitation in the tropics. Finally, this thesis describes the mean field of the North Pacific subtropical gyre using Argo T/S and trajectory data. The gyre is deeper than 1975 db and than the densest ventilated isopycnal, and it is deepest in the northwest subtropics. In the east, its axis tilts northeast with increasing pressure.

Chapter 1

Introduction

Large scale ocean circulation plays a key role in regulating the Earth's climate and the productivity of ecosystems. Even ocean gyres, that are regions of relatively low biological productivity, give a significant total contribution to the global productivity due to their large size [McClain *et al.*, 2004]. Studying the physical processes that govern large scale circulation, is crucial to understand not only nutrient distribution, habitat definition and related fisheries' yields [Sydeman *et al.*, 2011; Shiozaki and Chen, 2013; Johnson *et al.*, 2013], but also migration patterns and mobility of pollutants and marine debris [Day *et al.*, 1985].

The dynamics of large scale ocean circulation, is strongly related to the air-sea exchange of momentum, heat and freshwater fluxes at the ocean-atmosphere interface. Also, the exchange of heat between the ocean and the atmosphere is a key aspect of the solar energy redistribution in the climate system, since a significant portion of the heat redistribution occurs in the ocean [Houghton *et al.*, 1996; Czaja and Marshall, 2006]. Predicting how the circulation is evolving in a changing climate is essential in order to predict ice-melting at the poles, regional climate and in general the state of the global ocean-atmosphere system in the coming decades.

1.1 Observing the ocean with Argo floats.

The Argo array of profiling floats provides measurements of temperature, salinity and pressure with unprecedented resolution on a global scale and no bias towards summer months [Roemmich *et al.*, 2009]. In the time period between January 2004 and June 2014, more than a million profiles were stored with a median number of profiles per 1-degree ocean box of 24. The global distribution of profiles is fairly homogenous in the two hemispheres (with a slight bias towards the Northern Hemisphere), but the density per 1-degree box rapidly decreases poleward of $\sim 60^\circ$ latitude (Fig.1.1, top panel). Also, Argo observations are not available close to the coast and in marginal seas and this limits Argo-based studies of global heat and freshwater exchanges between the atmosphere and the ocean. Still, Argo offers a unique view of the global ocean three dimensional field in the upper 2000 db and equatorward of $\sim 60^\circ$ latitude and it is expanding to marginal seas and higher latitudes. Argo profiles are measured every ~ 10 days and global maps of temperature and salinity can be drawn monthly, but shorter timescales can also be studied in specific regions. Most importantly, Argo does not have a bias towards summer months. This is an issue with historical data, especially in the Southern Ocean. Fig.1.2 shows how the density distribution of Argo profiles is similar in (b) winter and (a) summer also south of $\sim 20^\circ\text{S}$, with more than 100,000 profiles in the region in both seasons.

The density of profiles in the World Ocean Database that reach at least the 900 db level (i.e. ocean station data, CTDs and moored buoys) is shown for comparison with Argo in Fig.1.1 (bottom). The distribution is clearly biased towards the Northern Hemisphere and along ship tracks, and the median number of profiles per 1-degree box is 2, even for a much longer time record (January 1950 to March 2014) than Argo. Also, Fig.1.2c and Fig.1.2d show how the WOD is biased towards summer months, with a total number of profiles south of $\sim 20^\circ\text{S}$ that reduces of more than half in winter (Fig.1.2d).

Argo is an unprecedented dataset of ocean subsurface data, and, as it extends in time, it will also play a key role in understanding how the coupled ocean-atmosphere system is evolving in a changing climate.

1.2 Studying the mean, annual cycle, and inter-annual variability of large-scale ocean circulation using Argo.

The large-scale ocean circulation, dynamics, and air-sea exchanges are investigated here, based on observational datasets including Argo and satellite altimetry, and viewed in the framework of modern theoretical ideas. Combining Argo T/S profiles and altimetry data provides a view of interannual variability on a basin-scale, with satellite observations enabling temporal extensions of the Argo results back to 1993.

Also, 10 years of global Argo data (since 2004) provide 10 realizations of the annual cycle and yield robust results both in the upper ocean and at greater depths, where the signal has a smaller amplitude. In the upper water column, it is particularly interesting to analyze the heat and freshwater budgets and estimate air-sea fluxes of these quantities. The ocean acts as a low-pass filter for high-frequency atmospheric variability, and an ocean based estimate of heat and freshwater fluxes at the ocean-atmosphere interface is now possible thanks to the accuracy and vertical resolution of Argo measurements.

Finally, the mean field of absolute horizontal velocity from a combination of Argo T/S profiles and trajectory data is well-determined, especially in the heavily-sampled North Pacific. The main source of noise in the trajectory data originates from mesoscale variability. Still, the large quantity of data available in some regions allows a robust estimate of the absolute velocity field at 1000 db.

1.3 Overview of the results.

The wind-driven interannual variability of the subtropical North Pacific is described in Chapter 2 of this thesis. It is shown that the interior circulation of the North Pacific gyre varies in strength and spatial orientation on interannual timescales. Also, variations in sea surface height derived from satellite altimetry are consistent with Argo steric height changes. Hence, this enabled an extension

of the 2004-2011 Argo study back to 1993. The observed gyre-scale circulation variability is found to be consistent with regional anomalies in wind-forcing through a time-dependent Sverdrup balance.

The next part of the thesis is focused on annual steric variability, including diabatic changes in the surface layer due to air-sea buoyancy fluxes and adiabatic changes due to wind-forced advection, which are dominant in the subsurface ocean. The annual signal in subsurface (200-2000 db) steric height zonally averaged over a season ($\Delta\eta'$) shows good agreement with the wind-forced vertical advection contribution ($\Delta\eta'_w$) both in the global ocean and in different basins (Chapter 3). This study is the first on a global scale and also shows that the annual vertical advection extends deep into the water column. The contribution of horizontal advection to $\Delta\eta'$ is significant in some regions and consistent with differences between $\Delta\eta'$ and $\Delta\eta'_w$.

In Chapter 4 of this thesis, the Argo temperature and salinity profile data (from 2004 to 2013) are used to estimate the global pattern of climatological monthly heat and freshwater fluxes at the ocean-atmosphere interface. Changes in temperature or salinity are computed in a volume of water above an isopycnal that is below the mixed layer and not subject to mixed-layer entrainment. Horizontal advection components from geostrophic velocity and from Ekman transport, based on wind stress, are also included. The climatological monthly heat or freshwater flux at the ocean surface is estimated as the sum of advective and time tendency contributions. This estimate of air-sea flux from Argo is described in global maps and, also, as basin-wide integrals, and it is compared to atmospheric reanalysis data and to air-sea flux products based on observations. This ocean-based estimate of surface fluxes is unique in that it is consistent with property variations in the subsurface ocean. It indicates greater amplitude for the climatological monthly heat flux values in the subtropics compared to other products. Similarly, the combination of Argo freshwater flux and reanalysis evaporation, suggests greater amplitude for climatological monthly precipitation in the tropics.

The final chapter of my thesis work (Chapter 5) describes the mean state of the North Pacific subtropical gyre, investigating the vertical structure of the

gyre and the meridional transport. Argo trajectory data are used to estimate the absolute velocity at 1000 db and are combined with temperature and salinity profiles to reconstruct the 3d velocity field in the upper 2000 db. The shape of the gyre defined here is that of a bowl that shifts northward as it deepens and is deepest in the northwest of the domain, where the Kuroshio Extension's (south) recirculation region is located. The gyre axis in the eastern portion of the gyre (east of $\sim 180^\circ\text{E}$) tilts northeast with increasing pressure, while does not change as much in the western region. Still, for the deepest isopycnals observed by Argo, a northeast tilt is noticeable in the west too. The gyre circulation is also present on isopycnals that do not ventilate to the north, indicating that the atmospheric forcing does not directly drive the flow in the deeper portion of the gyre. Finally, referencing the flow to the Argo trajectory data (rather than to the 1975 db isobar), yields a stronger transport both in the zonal and meridional direction.

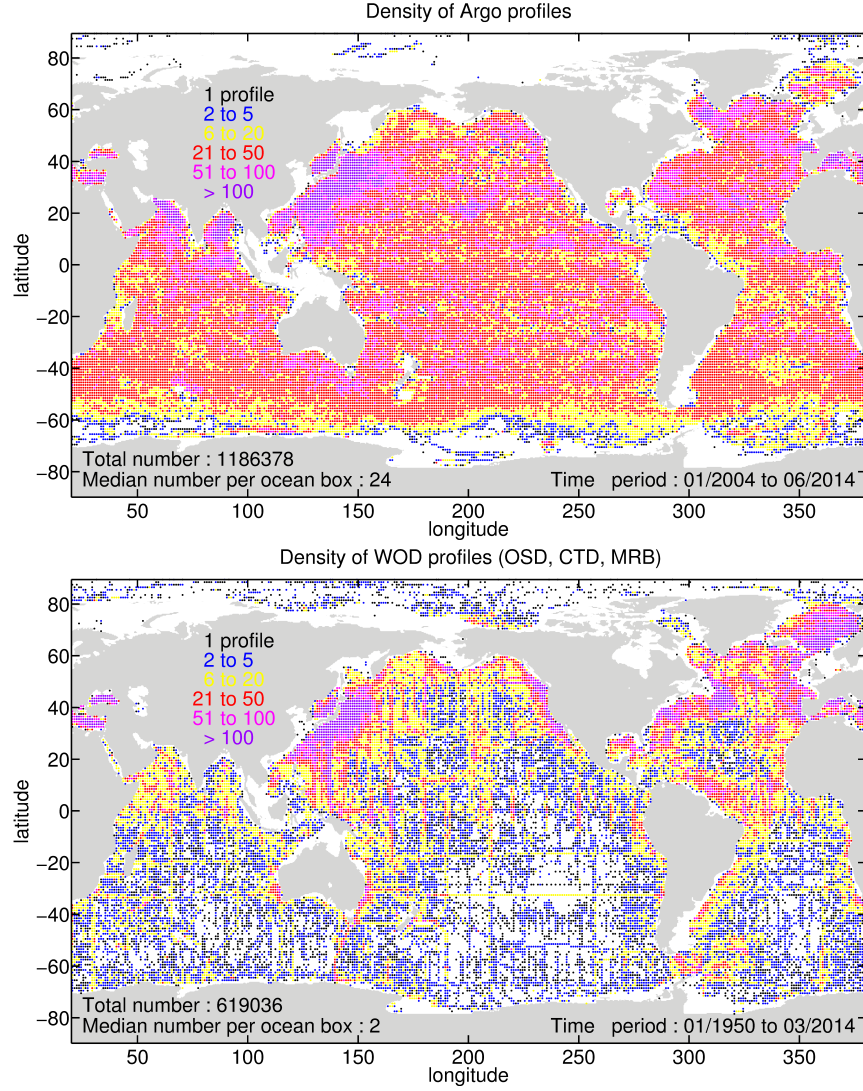


Figure 1.1: Number of profiles in 1×1 degree boxes. The upper panel is for Argo profiles since January 2004 (until June 2014). The lower panel is for Ocean Station Data (OSD), CTD casts and moored buoys (MRB) stored in the World Ocean Database (WOD) that reach at least the 900 db level, for the time period from January 1950 to March 2014.

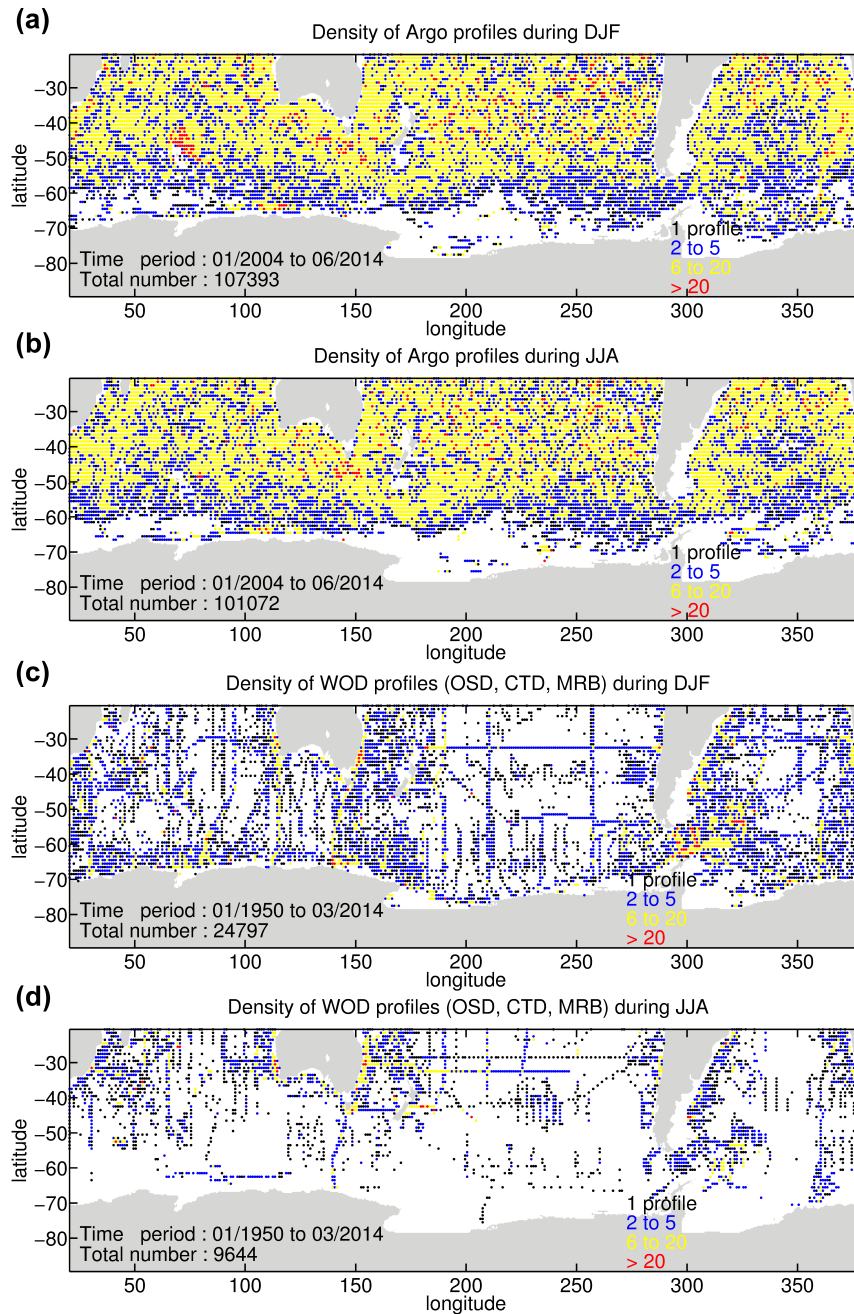


Figure 1.2: Density of profiles as in Fig.1.1, but for (a, c) the Southern Hemisphere summer (December to February, DJF) and (b, d) winter (June to August, JJA). (a, b) are for Argo profiles since January 2004 (until June 2014), (c, d) are for OSD, CTD and MRB profiles stored in the World Ocean Database (WOD) and that reach at least the 900 db level, during the time period from January 1950 to March 2014.

Chapter 2

Wind-driven Variability of the Subtropical North Pacific Ocean

Abstract. The Argo array provides a unique dataset to explore variability of the subsurface ocean interior. This study considers the subtropical North Pacific Ocean during the period from 2004 to 2011, when Argo coverage has been relatively complete in time and space. Two distinct patterns of Argo dynamic height transport function ($\widehat{DH}$) are observed: in 2004 – 2005, the gyre is stronger, and in 2008 – 2009 it is weaker. The orientation of the subtropical gyre also shifts over time: the predominantly zonal major axis shifts to a more northwest-southeast orientation in 2004 – 2005 and to a more southwest-northeast orientation in 2008 – 2009.

The limited temporal range of the Argo observations does not allow analysis of the correlation of ocean transport and wind forcing in the basin for the multi-year time scale (6–8 year period) typical of the dominant gyre patterns. The meridional geostrophic transport anomaly between 180°E and 150°E is computed both from Argo data (0 – 2000 db) and from the Sverdrup relation (using reanalysis winds): similarities are observed in a latitude-time plane, consistent with local forcing playing an important role. A forcing contribution from the eastern subtropics will also reach the region of interest, but on a longer timescale, and it is not analyzed in this study.

Compared with the 8-year Argo record, the longer 19-year time-series of satellite altimetry shows a similar but somewhat modified pattern of variability. A longer Argo record will be needed to observe the decadal-scale fluctuations, to separate interannual and decadal signals, and to ensure statistical confidence in relating the wind-forcing and the oceanic response.

2.1 Introduction

Wind-driven fluctuations in the circulation and temperature structure of the oceanic subtropical gyres have impacts on marine ecosystems and potentially on climate through ocean-atmosphere feedbacks in the western boundary current regions [*Barsugli and Battisti*, 1998; *Kelly et al.*, 2010; *Kwon et al.*, 2010; *Tanimoto et al.*, 2011]. Because of the long time-scales of oceanic response, changes in large-scale ocean circulation are sensitive diagnostics of changing patterns in the noisier atmospheric forcing.

The Argo array, which began in 2000, is now providing the first continuous subsurface observations of gyre-scale variability. Though the time span of Argo is still short, it is now sufficient to begin to see patterns of significant interannual variability: this work describes the period 2004 – 2011, when Argo coverage has been consistently good in time and space.

Previous regional studies in the North Pacific have mostly been limited to altimetry and other surface ocean data, and have described interannual variability in the Kuroshio Extension (KE) system, the North Equatorial Current (NEC) and the North Pacific Current (NPC). The dominant signals in dynamic height of the Subtropical North Pacific are the interannual path and strength variations of the Kuroshio system [*Qiu and Joyce*, 1992; *Qiu*, 2000, 2003; *Qiu and Chen*, 2005, 2010b]. In particular, *Qiu and Chen* [2010b] described the interannual oscillation of the Kuroshio Extension system during 1992 – 2008 in terms of a stable state and an unstable state. A stable dynamical state appeared from late 1992 to mid 1995 and from early 2002 to late 2005; an unstable state from mid 1995 to late 2001 and from early 2006 to the end of the timeseries. During the unstable state the

eddy kinetic energy level is enhanced, and the path meanders substantially. At the same time, the KE jet has a reduced eastward transport and a more southerly flow path. These patterns reverse when the KE system shifts to a stable state. The basin-scale wind forcing has been shown to play a key role in determining decadal KE variability through linear Rossby wave dynamics (see also *Miller et al.* [1998]; *Deser et al.* [1999]; *Lysne and Deser* [2002]; *Qiu* [2003]; *Douglass et al.* [2010]; *Zhang et al.* [2011]). However, Sverdrup transport change, based on wind stress forcing, underestimates the observed KE transport [*Qiu*, 2002b]. This is consistent with the notion that the volume transport change in the Kuroshio Extension is not necessarily controlled by temporal changes in the upstream transport [*Chao*, 1984; *Yoon and Yasuda*, 1987; *Yamagata and Umatani*, 1989; *Sekine*, 1990; *Akitomo, K. et al.*, 1991; *Akitomo et al.*, 1997], but it is also related to changes in the inertial recirculation gyre [*Qiu and Miao*, 2000; *Qiu*, 2002b; *Qiu and Chen*, 2010b]. A recent analysis of altimetric and hydrographic data for the Kuroshio in the East China Sea [*Andres et al.*, 2011] finds that large-scale wind-stress curl forces both a barotropic and a baroclinic mode that reach the region of interest via different waveguides. The barotropic response manifests itself at zero lag, propagating to the western boundary from the east and travelling southward through a coastal waveguide. This suggests that, in general, the absolute transport at a given latitude may differ from the Sverdrup balance along that line of latitude: waveguides can steer the barotropic mode across latitude lines.

Other studies have analyzed transport and location variability of the North Equatorial current in relation to the basin wind-stress curl [*Qiu and Joyce*, 1992; *Qiu and Chen*, 2010a]. In particular, *Qiu and Chen* [2010a] used 17 years of satellite altimeter sea surface height (SSH) data and estimated the bifurcation location of the NEC along the Philippine coast to study its interannual-to-decadal variability. The bifurcation latitude was found between 10° and 15° N, with northern locations observed in late 1992, 1997 – 98, and 2003 – 04 and southern locations in 1999 – 2000 and 2008 – 09. On interannual and longer time scales, variability was generally related to the Niño-3.4 index, with a positive index corresponding to a northern bifurcation and a negative index indicating a southern bifurcation.

However, the exact location of bifurcation was determined by wind forcing in the $12^{\circ} - 14^{\circ}\text{N}$ and $140^{\circ} - 170^{\circ}\text{E}$ band, that includes variability not fully representable by the Niño-3.4 index. When the NEC bifurcates at the northern latitude, its surface transport tends to increase, the Kuroshio and Mindanao Current tend to intensify, and the SSH values tend to decrease in regions surrounding the Philippines and the Indonesian archipelagos. The reverse is true when the NEC bifurcation shifts to the south.

Interannual variability of the North Pacific Current (NPC) is documented by *Qiu* [2002a]; *Douglass et al.* [2006]; *Cummins and Freeland* [2007]. *Qiu* [2002a] describes a steady strengthening over the period from late 1992 to 1998, as observed in altimetric data from the TOPEX/Poseidon satellite mission. Baroclinic Rossby wave dynamics could explain the interannual variability in SSH, which dropped on the northern side of the NPC causing the strengthening: the wind-forced signal propagated slowly from the eastern boundary. Intensification of the subtropical gyre was also associated with an increase in the NPC transport in this period [*Douglass et al.*, 2006]. *Cummins and Freeland* [2007] used dynamic height from Argo, numerical simulations and satellite altimetry data to quantify the relative strength of two orthogonal modes of the NPC variability, the breathing and bifurcation modes [*Freeland*, 2006]. Results suggest that variability of the NPC is associated with large-scale differential Ekman pumping rather than Rossby wave propagation or changes in the Sverdrup-balanced transport.

The goal of the present work is to integrate and expand on these earlier regionally-based studies by taking a basin-wide perspective on variability of the gyre structure and circulation. Section 2.2 describes the data used. Section 2.3 presents the ocean variability as observed in Argo. Section 2.4 relates this variability to wind forcing provided by ECMWF. Section 2.5 presents a summary and conclusions.

2.2 Data

2.2.1 Argo observations

Argo is a broad-scale global array of temperature/salinity profiling floats, initiated in 2000 [Roemmich *et al.*, 2009]. The number of Argo float profiles in the region of interest ($20^{\circ} - 40^{\circ}N$, $120^{\circ}E - 100^{\circ}W$) rapidly increased in time from 2004 to 2006, to about 12,000 per year, and the array has been maintained since then (Fig. 2.1a). An example of measurement locations is shown in Fig. 2.1b for the year 2009. The western portion of the domain is more highly resolved than the interior and the east, but the overall distribution is sufficient to capture basin-scale variability year by year.

After quality control and adjustment of pressure bias, the raw data were gridded monthly on a $1^{\circ} \times 1^{\circ}$ grid. The gridding of Argo temperature and salinity data was done by objective mapping, with latitude-dependant decorrelation scales [Roemmich and Gilson, 2009]. In order to test whether substantial variance is lost by the objective mapping in lightly sampled regions, we also carried out an alternate gridding procedure consisting of bin-averaging of data in overlapping large-scale regions of 555 km radius. Results based on this procedure were little different from the objective mapping case, and are therefore not discussed further.

2.2.2 ECMWF wind momentum flux

Zonal and meridional wind momentum flux data used in this analysis are available online from the European Centre for Medium-Range Weather Forecasts (ECMWF) website. They are synoptic monthly means from the ECMWF Re-Analysis ERA-Interim for the period 1989 – 2011. Dee *et al.* [2011] describe the ERA-Interim data. Brunke *et al.* [2011] found that for wind stress and latent heat fluxes, ERA-Interim performed better than most other products.

2.2.3 AVISO altimetry data

Ssalto/Duacs weekly gridded sea level anomalies are distributed by the AVISO program to study long term as well as seasonal variability [Ducet *et al.*, 2000]. The dataset used here is the reference version on a $1/3^\circ \times 1/3^\circ$ Mercator grid. It is based on two satellites, and it was chosen since it is homogeneous and stable in time over the 1993 – 2010 record. These data were smoothed to a $1^\circ \times 1^\circ$ monthly grid in order to have the same resolution as Argo.

2.3 Ocean variability

2.3.1 Argo data

This study describes interannual variability of the Subtropical North Pacific Ocean ($20^\circ - 40^\circ N$, $120^\circ E - 100^\circ W$) from 2004 through 2011. The ocean variability is analyzed using a transport function $\widehat{DH}$, and anomalies in $\widehat{DH}$ are computed removing the temporal mean and the seasonal cycle. We define $\widehat{DH}$ as the integral of dynamic height DH (a transport function on pressure surfaces) over all the Argo pressure levels, minus the spatial mean at every point in time,

$$\widehat{DH}(x, y, t) = \int_{p=2000db}^{p=0db} DH(x, y, p, t) dp - \left\langle \int_{p=2000db}^{p=0db} DH(x, y, p, t) dp \right\rangle_{x,y}. \quad (2.1)$$

In Eq.2.1, DH is the dynamic height at pressure p referenced to the pressure $p_{ref} = 2000$ db, i.e. $DH(x, y, p, t) = \int_{p_{ref}}^p \delta dp$; in this formulation, $\delta(x, y, p, t) = \frac{1}{\rho} - \frac{1}{\rho^*}$ is the specific volume anomaly, ρ is the density and ρ^* is the density corresponding to $S = 35$ psu and $T = 0^\circ C$ at pressure p . $\widehat{DH}$ was spatially smoothed in space using a Parzen filter of 7 points in the zonal direction and 5 points in the meridional direction. Since the data are gridded at 1° resolution, this procedure filters out features with scales less than about 500 km.

The spatial mean in Eq.2.1 (second term on the right) is an area-weighted average, and it is removed to focus on the circulation rather than on basin-wide water column expansion. $\widehat{DH}$ derivatives in x and y are, in fact, related by geostrophy to zonal and meridional transport, respectively per unit latitude and longitude. For

instance, according to geostrophic dynamics, an anticyclonic gyre in the Northern Hemisphere corresponds to positive $\widehat{DH}$ values (a dome) in the center and negative values (a depression) at the periphery. This is the case for the (temporal) average field of $\widehat{DH}$ in Fig. 2.2a (positive values are red, negative are blue), and it corresponds to the North Pacific Subtropical Gyre.

For each month, the time varying transport function anomalies $\widehat{DH}$ are projected onto large-scale patterns that characterize the North Pacific Ocean, here using the formalism of a least-squares fit. The least-squares fit is of the form:

$$\widehat{DH}(t) = \mathbf{G}\mathbf{m}, \quad (2.2)$$

where $\widehat{DH}(t)$ $[N \times 1]$ is the vector of transport function anomalies at every point in the spatial domain for time t . Since the $\widehat{DH}$ data are smoothed with an 11-month running mean, the first and last 5 months of the time series are not considered. The matrix $\mathbf{G}$ is $[N \times 2]$, where the first column contains the spatial pattern to be fit, and the second column is constant and it is used to remove any residual constant offset. A standard least-squares procedure is used to minimize the misfit between $\widehat{DH}$ and the model $\mathbf{G}\mathbf{m}$, so that $\mathbf{m}(t) = (\mathbf{G}^T\mathbf{G})^{-1}\mathbf{G}^T\widehat{DH}(t)$. The time series of fitted coefficients $\mathbf{m}_1$ provides a measure of the variability of the system, while $\mathbf{m}_2$ is a constant offset. Since the weighted average spatial mean was removed from the data at every point in time, $\mathbf{m}_2$ is close to zero.

The red line in Fig. 2.3a shows $\mathbf{m}_1$ for the time-varying projection of the anomalies onto the time-mean transport function field (Fig. 2.2a). This tests the extent to which the variability is a simple strengthening or weakening of the mean field. The blue line, also, shows $\mathbf{m}_1$, but now for the projection onto the mode 1 empirical orthogonal function shown in Fig. 2.2b [EOF, see *North, 1984; Dommenget and Latif, 2002*]. Since $\mathbf{m}_2$ is close to zero, this blue line corresponds, also, to the temporal mode of EOF 1. The coefficients ($\mathbf{m}_1$) in Fig. 2.3a, corresponding to the time mean and EOF 1 of the transport function, have very similar time series, with both indicating ocean variability of opposite sign between the earlier and later years of the record. Here the sign convention is that positive coefficients correspond to positive anomalies of the fitted field. Thus positive values of the projection onto the mean field (red line in Fig. 2.3a) indicate a strengthening

of the mean anticyclonic gyre, and negative anomalies imply a weaker gyre. For the EOF 1 (blue line), the projections have similar interpretations, but the region of stronger and weaker gyres in each period is mainly confined to the west of the Hawaiian-Emperor seamount chain (see Fig. 2.2b and, for details of the bathymetry, Fig. 2.2f).

For each month in the time series, the fraction of spatial variance explained by the least-squares fit is computed as

$$\sigma_i(t) = 1 - \frac{\left(\widehat{\mathbf{DH}}(t) - \mathbf{G}\mathbf{m}(t)\right)^T \left(\widehat{\mathbf{DH}}(t) - \mathbf{G}\mathbf{m}(t)\right)}{\widehat{\mathbf{DH}}^T(t)\widehat{\mathbf{DH}}(t)}, \quad (2.3)$$

Because the constant m_2 is near zero, σ_i essentially represents the correlation coefficient of $\widehat{\mathbf{DH}}$ and the spatial pattern in the first column of $\mathbf{G}$. As shown in Fig. 2.3b, in both cases, $\sigma_i(t)$ peaks in 2004 – 2005 and in 2008 – 2009, but EOF 1 explains more variance: this is expected since the EOF is defined to explain as much variance as possible in the first mode. The two peaks in Fig. 2.3b indicate when the strong and weak gyre patterns are most evident.

Maps of the transport function anomaly averaged over each year (2004 – 2011) are shown in Fig. 2.4. Contours are for the mean field, and lines are dotted if values are negative (the zero contour is dashed). The spatial anomaly distribution in 2004 is very similar to EOF 1 in Fig. 2.2b and it evolves to a field of opposite sign in 2008. This progression indicates again a stronger gyre (anticyclonic anomalous circulation with a pattern similar to that of the mean field) in 2004 – 2005 and a weaker gyre (cyclonic anomalous circulation) in 2008 – 2009. The anomalous circulation region is mainly confined to the west of the Hawaiian-Emperor seamount chain (see shallow (dark) bathymetry around $180^\circ E$ in the map background) as seen previously. The anomaly map in 2011 is similar but opposite in sign compared to the map in 2007: it may anticipate a stronger gyre period starting in 2012. The complete time-series of area weighted averages of $\widehat{\mathbf{DH}}$ anomaly are shown for two characteristic regions (Fig. 2.5). The first box extends from $140^\circ E$ to $160^\circ E$ and from $20^\circ N$ to $35^\circ N$ (Fig. 2.5a), while the second box is from $160^\circ W$ to $140^\circ W$ and from 30° to $40^\circ N$ (Fig. 2.5b). The time series is shown as a thick gray line in the figure, and it aligns well with the 11-month running

mean (thick black line). In the western box, values are significantly positive in 2004 – 2005 and negative in 2008 – 2009 (Fig. 2.5a). The sign is opposite in each period for the eastern box in Fig. 2.5b. In both areas, the standard error is computed for the average in the box, and this is the measure used to estimate significance (see gray thin lines in the two panels).

Annual means of the pressure anomaly of the 1026.4 kg/m³ isopycnal are also consistent with the transport function anomaly maps in Fig. 2.4 (not shown). They have the same sign, and this indicates that an anomalous depression of the isopycnal corresponds to an anomalous dome in the transport function. A similar behavior is expected since dynamic height and isopycnal pressure approximately mirror each other. The average pressure on the 1026.4 kg/m³ isopycnal is about 400 db, and the choice of analyzing this surface is made since it is in the subtropical gyre thermocline, and it has similar behavior to neighboring isopycnals. Also, large-scale flows tend to follow isopycnals; hence, a density surface is appropriate when describing dynamics. Pressure on the 1026.4 kg/m³ isopycnal in this analysis is smoothed in space in the same way that $\widehat{DH}$ was smoothed, and the spatial mean (area weighted average) is removed at every point in time. The spatial distribution of the anomaly in time is consistent with two different displacements of the gyre during the Argo period. Fig. 2.6b,c show the meridional dependence of pressure anomaly (cyan and purple lines) and of the mean field (green line) averaged in zonal bands. The temporal anomalies are also averaged in time over the periods 2004 – 2005 (cyan line, period 1) and 2008 – 2009 (purple line, period 2), and the error bars in the figure represent the standard error for this computation. In the western part of the basin (Fig. 2.6b), the gyre mean shape (green line) shifts poleward during period 1 (cyan line anomaly) and equatorward during period 2 (purple line anomaly). Filled circles indicate the location of maximum pressure at different times: in 2004 – 2005, it is north of the average position, in 2008 – 2009 it is south. Further east (Fig. 2.6c), the opposite occurs in each period. The gyre has an elliptical ‘bowl’ shape, with the major axis of the ellipse oriented roughly in the zonal direction. Isopycnals are deepest in the western part of the domain around 30 degrees latitude [*Rhines and Young*, 1982; *Luyten et al.*, 1983]. Fig. 2.6b,c

suggests that the gyre’s major axis shifts to a more northwest-southeast orientation in 2004 – 2005 and to a more southwest-northeast orientation in 2008 – 2009. The movement is around a vertical axis located in the western part of the basin ($\sim 170^\circ\text{E}$), and it is small in the sense that the gyre is displaced by a small fraction of its size, but nevertheless significant. Fig. 2.6a provides a schematic. Two pressure contours are drawn for the two periods. In period 1 (cyan line) the gyre is shifted poleward in the west and equatorward in the east relative to period 2 (purple line). This is consistent with the rotation described above.

Fig. 2.6b also highlights steeper isopycnal slopes in the western portion of the basin during period 1, implying a stronger zonal geostrophic flow. In contrast, the anomalous isopycnal slope in period 2 is opposite in sign indicating weaker flow. Further east, pressure anomalies are less effective in modifying the meridional isopycnal slope, and they mainly shift the gyre ‘bowl’ shape (Fig. 2.6c).

2.3.2 Satellite altimetry data

Satellite altimetric sea surface height (SSH) anomalies provide a dataset to explore gyre-scale variability beginning in 1993 and extending through the Argo era. In this analysis, the spatial mean is removed at every point in time, and a spatial smoothing is applied as done for the transport function $\widehat{DH}$. Resulting values can be thought of as approximating anomalies relative to the mean Argo sea surface dynamic height referenced to 2000 db in Fig. 2.2c. A least-squares fit is performed on this altimetric SSH anomaly field, and results can be seen in Fig. 2.3c ($\mathbf{m}_1$) and in Fig. 2.3d (fraction of variance explained). SSH data are fitted to EOF 1 of SSH, as computed for the Argo period (green line) and for the complete SSH timeseries (magenta line). The magenta timeseries in panel c corresponds, also, to the temporal mode of EOF 1 of SSH anomaly for the period 1993 – 2010. The spatial pattern of SSH EOF 1, based only on data during the Argo period, 2004 – 2010, (Fig. 2.2d) differs somewhat from the EOF 1 of the transport function (Fig. 2.2b), and this is in part related to the shallower thermocline in the eastern subtropical Pacific. $\widehat{DH}$ is integrated in pressure, and it corresponds approximately to SSH multiplied by an effective depth. Scaling the

altimetric SSH EOF 1 during Argo by pressure on the 1026.4 kg/m^3 isopycnal, for instance, reduces the huge eastern anomaly values in Fig. 2.2d but not the western ones (not shown). The coefficient $\mathbf{m}_1$ in Fig. 2.3c shows interannual variability of opposite sign in the earlier and later Argo years as described for the transport function $\widehat{DH}$ (panel (a) of the same figure). These oscillations are also observed in the years prior to Argo. The fraction of variance explained (Fig. 2.3d) peaks in each regime but the SSH EOF 1 spatial pattern for the Argo period (green line) explains less variance in the period prior to Argo. This suggests that, in the longer ~ 20 -year period covered by altimetric SSH, either the interannual variability does not have the same larger-scale spatial pattern as during the recent Argo era or a lower frequency pattern of variability is causing the differences in the maps in Fig. 2.2d,e.

Fig. 2.7 is a schematic of the gyre movement using the 260 cm contour of AVISO SSH anomaly data added to the mean Argo sea surface dynamic height referenced to 2000 db. As shown in Fig. 2.6a, in 2004 – 2005 (cyan line) the gyre is poleward in the west and equatorward in the east relative to 2008 – 2009 (purple line). The pink line shows the same contour in 1999 – 2000: in this period, the gyre orientation is similar to 2008 – 2009.

2.4 The atmospheric forcing

Changes in large-scale ocean circulation are sensitive diagnostics of low frequency variability in the noisier atmospheric forcing. The baroclinic component of the oceanic response to such forcing has long time-scales. Analysis of variability in ECMWF wind momentum flux is presented here to investigate its possible relationship to the ocean variability described in Section 2.3.

Zonal and meridional wind momentum flux anomalies induce anomalous heat fluxes into the ocean. Also, momentum and vorticity are transferred to the ocean through anomalous Ekman upwelling velocity [*Qiu and Joyce, 1992; Miller et al., 1998; Deser et al., 1999; Lysne and Deser, 2002; Qiu, 2003; Douglass et al.,*

2010]. The Ekman upwelling velocity (w_{Ek}) is computed as

$$w_{Ek} = \nabla \times \left(\frac{\boldsymbol{\tau}}{\rho_0 f} \right), \quad (2.4)$$

where ρ_0 is the characteristic density of the region, f is the Coriolis parameter and $\boldsymbol{\tau}$ is the wind stress vector. Fig. 2.8 shows maps of the yearly averaged Ekman upwelling anomaly (2004–2011). In the subtropical Pacific ocean west of 180° , the vertical velocity is mostly negative (anomalous Ekman downwelling) in 2004–2006. It is mostly positive (anomalous Ekman upwelling), in 2007–2009. If Sverdrup dynamics are applied to the local anomaly field, the spatial distribution west of 180° indicates a stronger gyre there in the early Argo years and weaker later. In 2011, the anomaly in the region seems to shift to negative again.

Geostrophic meridional transport is computed twice for the 2004–2011 period: first, integrating the Argo geostrophic velocity in the first 2000 db between 150°E and the eastern longitude x_E (the label for this transport will be V_G); second, integrating the Sverdrup relation in the same zonal range (the label for this transport will be V_S):

$$V_G(y, t) = \int_{x_E}^{150^\circ E} \int_{2000}^0 v_G(x, y, p, t) dp dx \quad (2.5)$$

and

$$V_S(y, t) = \int_{x_E}^{150^\circ E} \frac{f_0}{\beta} w_{Ek}(x, y, t) dx, \quad (2.6)$$

where v_G is Argo meridional geostrophic velocity, f_0 is the Coriolis parameter, $\beta = \frac{df}{dy}$ and w_{Ek} is the Ekman upwelling velocity (Eq.2.4) computed from ECMWF momentum fluxes. Note that V_S is the Sverdrup transport ($\frac{1}{\rho_0} \mathbf{k} \cdot \nabla \times \boldsymbol{\tau}$) minus the Ekman transport ($-\frac{\beta}{\rho_0 f_0} \tau^x$), and it is representative of the entire water column, while V_G includes only the first 2000 db. Fig. 2.9 shows V_G (a,b,c) and V_S (d,e,f) anomaly time series versus latitude for different values of x_E ; the anomaly is filtered with an 11-month running mean. If x_E is west of the Hawaiian-Emperor seamount chain (see shallow topography around 180°E in Fig. 2.2f), V_G and V_S anomalies have similar distributions (Fig. 2.9a,d and b,e), but V_G is lagged by about 1 year compared to V_S . The reason for this lag is not clear, and it could be related

to the spatial distribution of the forcing in the zonal range of interest as well as the remote forcing component. The similarity in the patterns is consistent with Sverdrup-like dynamics applied to the local anomaly field of Ekman upwelling velocity in Fig. 2.8: when the wind curl forces Ekman pumping in the region, the water column moves equatorward; when Ekman suction is forced, the water column moves poleward.

The remote forcing by wind-stress curl in the eastern part of the basin in previous years is not analyzed in this study, due to the limited time range available for the Argo data. The gyre-scale oscillation described in this study has a temporal scale of about 6 – 8 years (see Fig. 2.3a). This length is very similar to that of the Argo record, and it does not allow us to assess the significance of correlation between V_G and V_S in different regions of the subtropics. Also, the V_G anomaly distribution seems not to change even when x_E is moved east of 180°E (Fig. 2.9c). Rather, anomalous V_G is mainly due to meridional velocities in the western region. A separate analysis of the transport between 190°E and the eastern longitude x_E shows, in fact, that the anomaly distribution there differs from those in Fig. 2.9a,b,c (not shown).

The anomalies in Fig. 2.9a,b and d,e are of opposite sign in the earlier and later Argo years. This is generally in agreement with results in Section 2.3 for the region west of the Hawaiian-Emperor seamount chain: stronger gyre (negative meridional transport anomaly) in 2004 – 2005 and weaker (positive anomaly) in 2008 – 2009.

Fig. 2.10 shows the meridional geostrophic transport anomaly (Sv) computed from the Sverdrup relation between 150°E and 180°E in 1989 – 2011 (Fig. 2.9e zooms in on a part of this plot). As seen in the AVISO SSH fields, the pattern of Sverdrup transport variability prior to 2004 does not show strong similarities to the recent Argo era (Fig. 2.9e).

2.5 Summary and conclusions

The present work exploits the broadscale spatial coverage and time-series nature of Argo’s subsurface ocean dataset. Argo provides a unique dataset to explore variability of the subsurface ocean interior during the period 2004 – 2011. AVISO altimetry provides a complementary dataset for the Argo period and is also used to extend the analysis farther into the past, along with the ECMWF wind-stress to map atmospheric forcing. Two distinct patterns of the dynamic height transport function $\widehat{DH}$ are observed during 2004 – 2011. One in 2004 – 2005 indicates a stronger gyre, while the other in 2008 – 2009 shows a weaker gyre. This variability is consistent with results of *Qiu and Chen* [2010a] for the strength of the Kuroshio Current and the North Equatorial Current and of *Qiu and Chen* [2010b] for the strength of the Kuroshio Extension jet.

Argo observations are available for only 8 years; hence, it is not possible to study correlation of ocean transport and wind forcing for the multi-year time scale (6 – 8 year period) characteristic of the dominant gyre patterns. The meridional geostrophic transport anomaly between 180°E and 150°E is computed both from Argo data (from the surface to 2000 db) and from the Sverdrup relation: similarities are observed in a latitude-time plane (Section 2.4) and this is consistent with local forcing playing a significant role. Yearly maps of Ekman upwelling confirm that, west of the dateline, the sign of the anomaly indicates (by Sverdrup dynamics) a stronger gyre in 2004 – 2005 and a weaker gyre in 2008 – 2009. A forcing contribution from the eastern subtropics will also reach the region of interest, but on a longer timescale, and is not analyzed in this study.

In addition to the variations in amplitude, a displacement of the subtropical gyre is observed during 2004 – 2011. As described by *Lysne and Deser* [2002], a comparison of the spatial pattern of interannual variance to the mean temperature structure shows that regions of maximum variability are located where the strongest north-south gradients in the mean field are found. The latitudinal boundaries of the wind-driven gyre also correspond to such regions; hence the variance maxima are indicative of north-south shifts in the gyre boundaries. The movement observed in Argo implies that the zonally oriented major axis of the gyre

tilts northwest-southeast in 2004 – 2005 and southwest-northeast in 2008 – 2009. Such variability is consistent with results by *Qiu and Chen* [2010a] for the location of the North Equatorial Current bifurcation and by *Qiu and Chen* [2010b] for the state of the Kuroshio Extension System. The movement can also be recreated with altimetric data and, in 1999 – 2000, the gyre tilts slightly southwest-northeast as in 2008 – 2009.

In general though, in the longer period covered by altimetric SSH and ECMWF wind momentum fluxes, a somewhat different dominant pattern of interannual variability is seen. This is to be expected, given the increasing variance in ocean variability on longer timescales (i.e. the red spectrum), and as illustrated, for example, using SSH for the Argo period and for the longer record in *Roemmich and Gilson* [2009] (their Figure 3.9a,b). In the 20-year SSH timeseries, a lower frequency pattern not seen in the 8-year Argo record modulates the shorter term variability. A longer Argo record is obviously needed to observe the decadal-scale fluctuations, to separate interannual and decadal signals, and for statistical confidence in relating the wind-forcing and the oceanic response. Nonetheless, we have shown here that the 8-year Argo timeseries is long enough in duration and sufficiently extensive in coverage to begin describing interannual variability of the wind-driven subtropical gyre.

Acknowledgments. The Argo data used here were collected and are made freely available by the International Argo Program and by the national programs that contribute to it. The authors’ participation in the Argo Program was supported by U.S. Argo through NOAA Grant NA17RJ1231 (SIOJIMO). The statements, findings, conclusions, and recommendations herein are those of the authors and do not necessarily reflect the views of the National Oceanic and Atmospheric Administration or the Department of Commerce. The efforts of many international partners in planning and implementing the Argo array are gratefully acknowledged. ECMWF ERA-Interim data used in this study have been obtained from the ECMWF Data Server. The AVISO altimeter products were produced by the CLS Space Oceanography Division as part of the Environment and Climate EU ENACT project (EVK2-CT2001-00117) and with support from CNES. Valuable

suggestions for this work were provided by Bruce Cornuelle and by two anonymous reviewers. The objectively mapped Argo data were provided by John Gilson.

Chapter 2, in full, is a reproduction of the material as it appears in *D. Giglio, D. Roemmich, and S.T. Gille* [2012], *J. Phys. Oceanogr.*, 42, 2089-2100. The dissertation author was the primary investigator and author of this work.

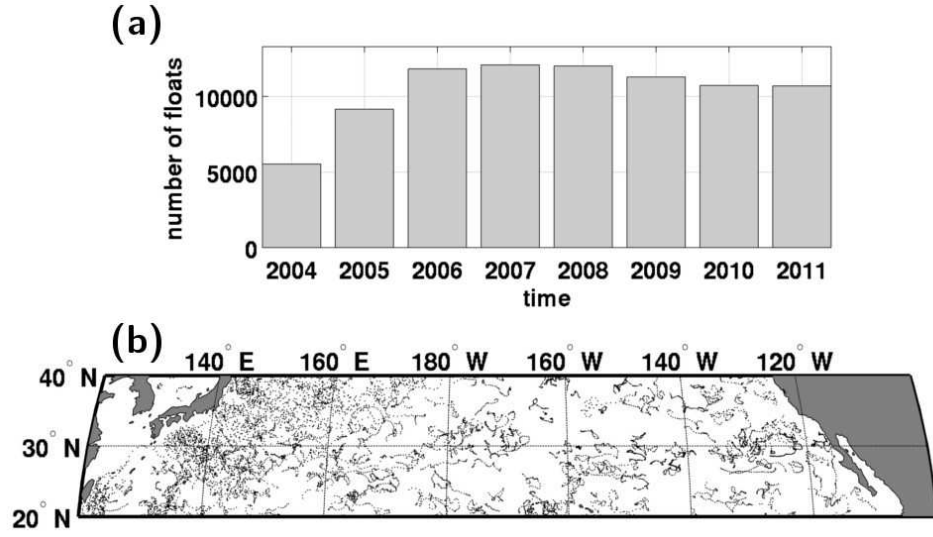


Figure 2.1: (a) Number of Argo profiles in time in the Subtropical North Pacific Ocean ($20^{\circ} - 40^{\circ}N$, $120^{\circ}E - 100^{\circ}W$). (b) Distribution of Argo profiles in the same region in 2009.

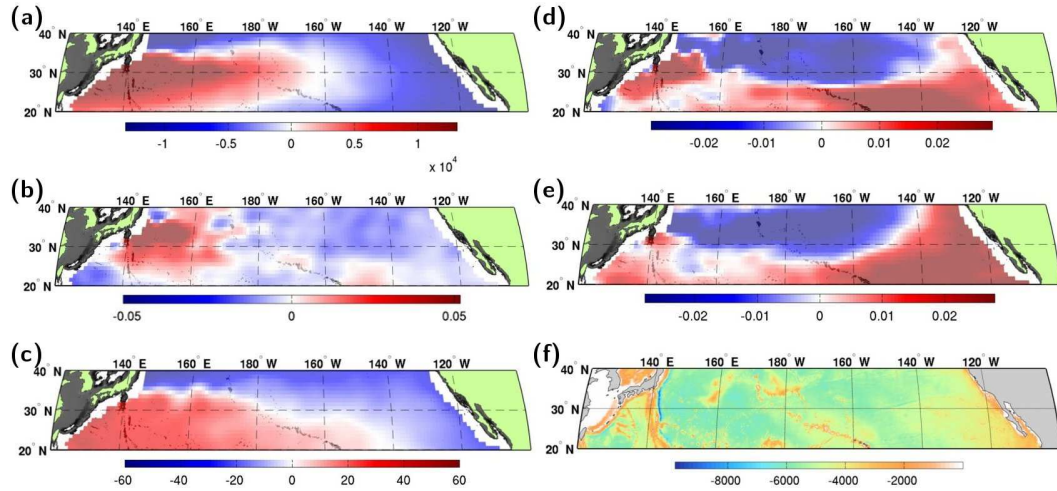


Figure 2.2: (a) Transport function mean field ($\text{cm}^2 \text{s}^{-2} \text{db}$). (b) Transport function EOF 1 spatial pattern. (c) Argo surface dynamic height (temporal) mean field during 2004 – 2010, after removing the spatial mean (cm). (d) AVISO SSH EOF 1 spatial pattern during 2004 – 2010. (e) AVISO SSH EOF 1 spatial pattern during 1993 – 2010. In panels (a-e), the map is in color and the bathymetry is in the background. Bathymetry is shown only when shallower than 2000 m using a gray (deep) to black (shallow) colormap. See (f) for a detailed bathymetry map.

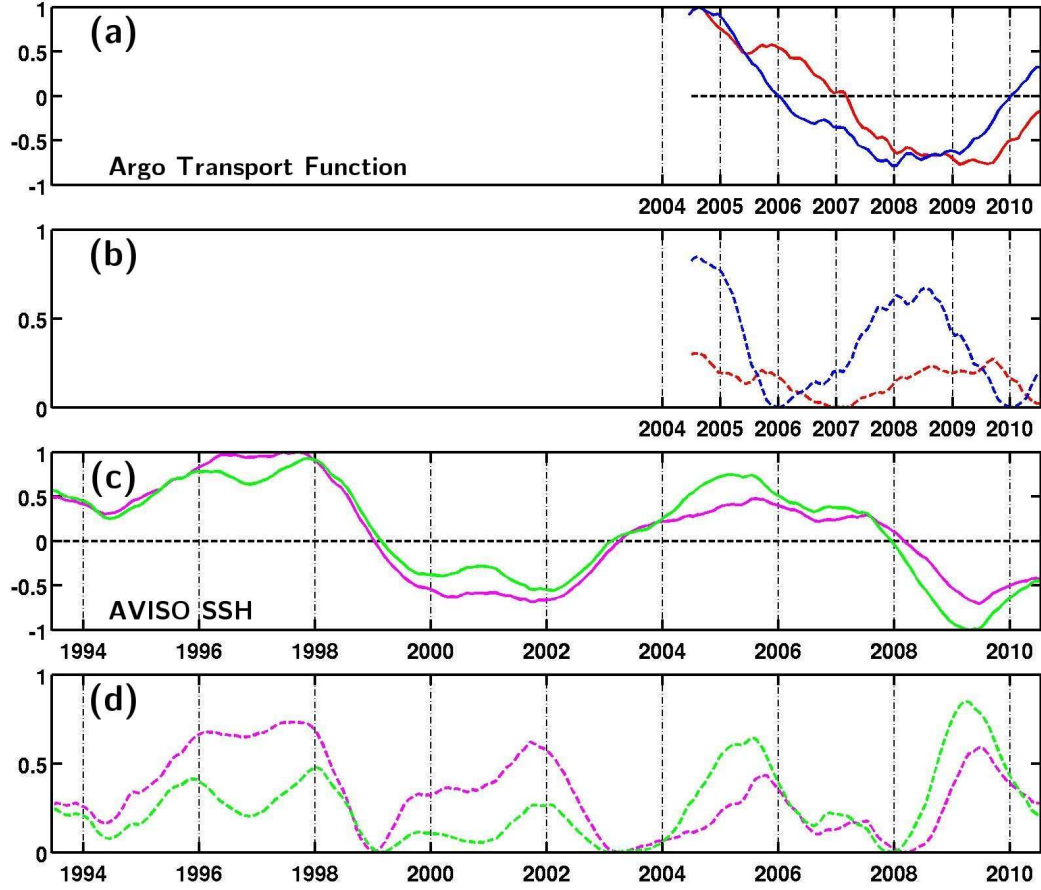


Figure 2.3: (a) Red line: least-squares fit coefficient $m(t)$, when fitting the transport function anomaly to its (temporal) average spatial pattern (Fig. 2.2a). Blue line: temporal mode of the transport function EOF 1 (see Fig. 2.2b for the spatial pattern). Each curve is normalized to its maximum value. (b) Fraction of spatial variance $\sigma_i(t)$ explained by the two timeseries in (a) at every time step. (c) Green line: least-squares fit coefficient $m(t)$, when fitting the AVISO SSH anomaly to its EOF 1 spatial pattern in 2004 – 2010 (Fig. 2.2d). Magenta line: temporal mode of the AVISO SSH EOF 1 in 1993 – 2010 (see Fig. 2.2e for the spatial pattern). Each curve is normalized to its maximum value. (d) Fraction of spatial variance $\sigma_i(t)$ explained by the two timeseries in (c) at every time step.

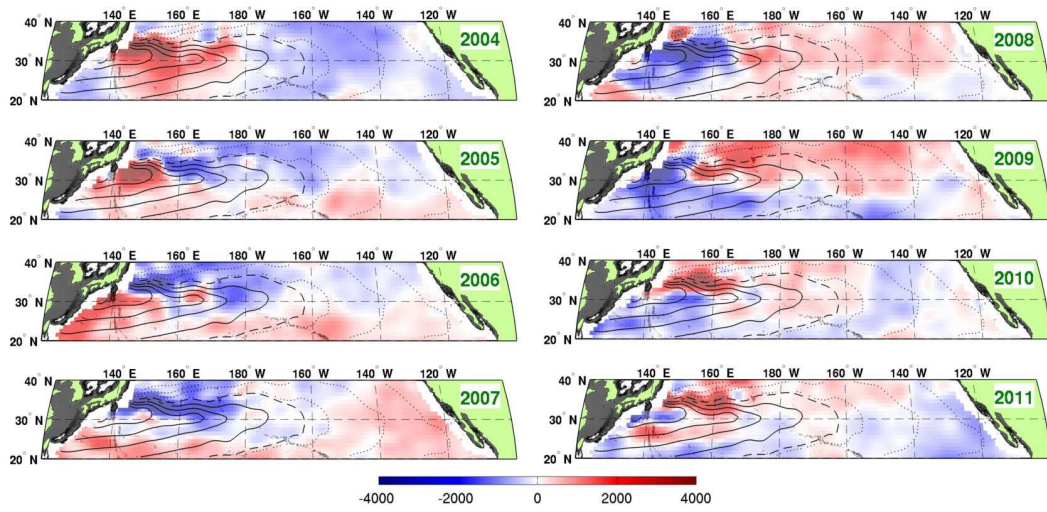


Figure 2.4: Transport function ($\text{cm}^2 \text{s}^{-2} \text{db}$) for each year considered in this study. Colors indicate anomalies relative to the 2004 – 2011 temporal mean: the anomalies in each panel are annually averaged. Contours represent the mean field: lines are dotted if values are negative; the zero contour is dashed. Bathymetry is shown as in Fig. 2.2.

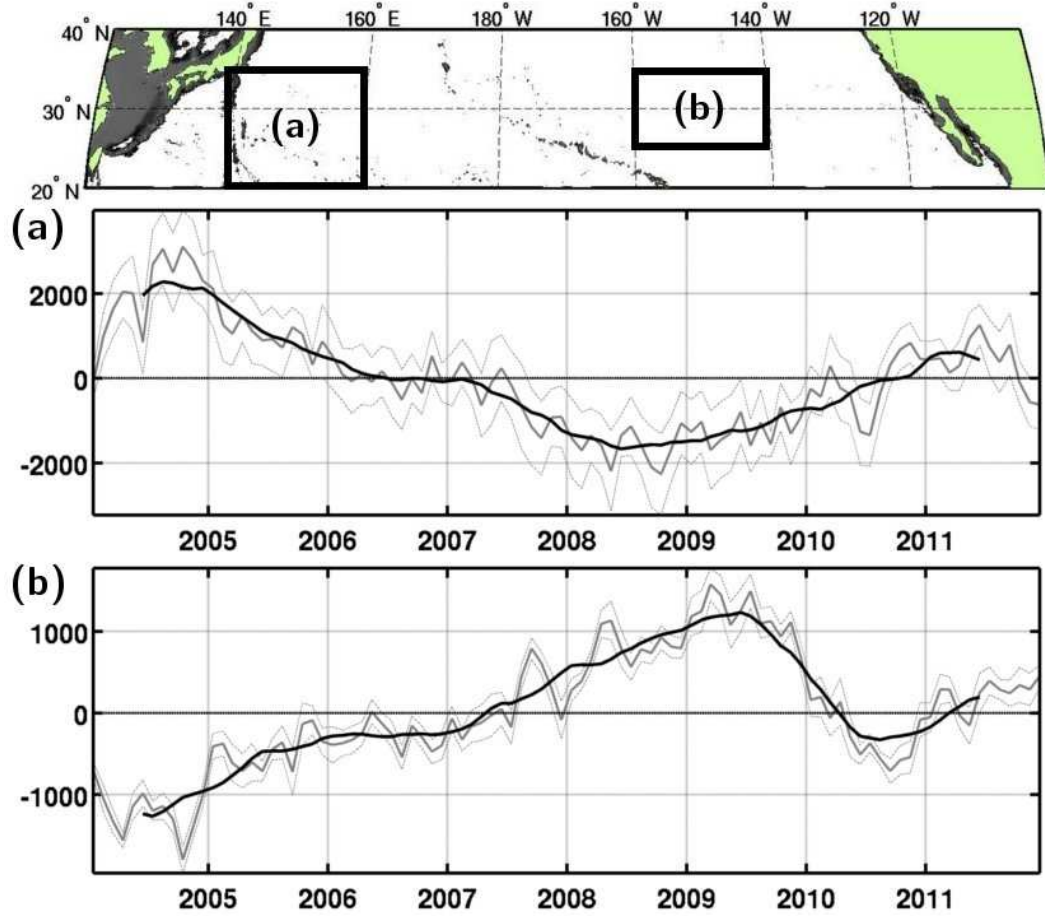


Figure 2.5: (a,b) Transport function anomaly ($\text{cm}^2 \text{s}^{-2} \text{db}$) averaged in the longitude-latitude boxes indicated in the top panel. Thick gray line: anomaly timeseries. Thin gray lines: error bar. Thick black line: 11-month running mean of the anomaly. (a) average in the box at $140^\circ - 160^\circ\text{E}$, $20^\circ - 35^\circ\text{N}$. (b) average in the box at $160^\circ - 140^\circ\text{W}$, $25^\circ - 35^\circ\text{N}$.

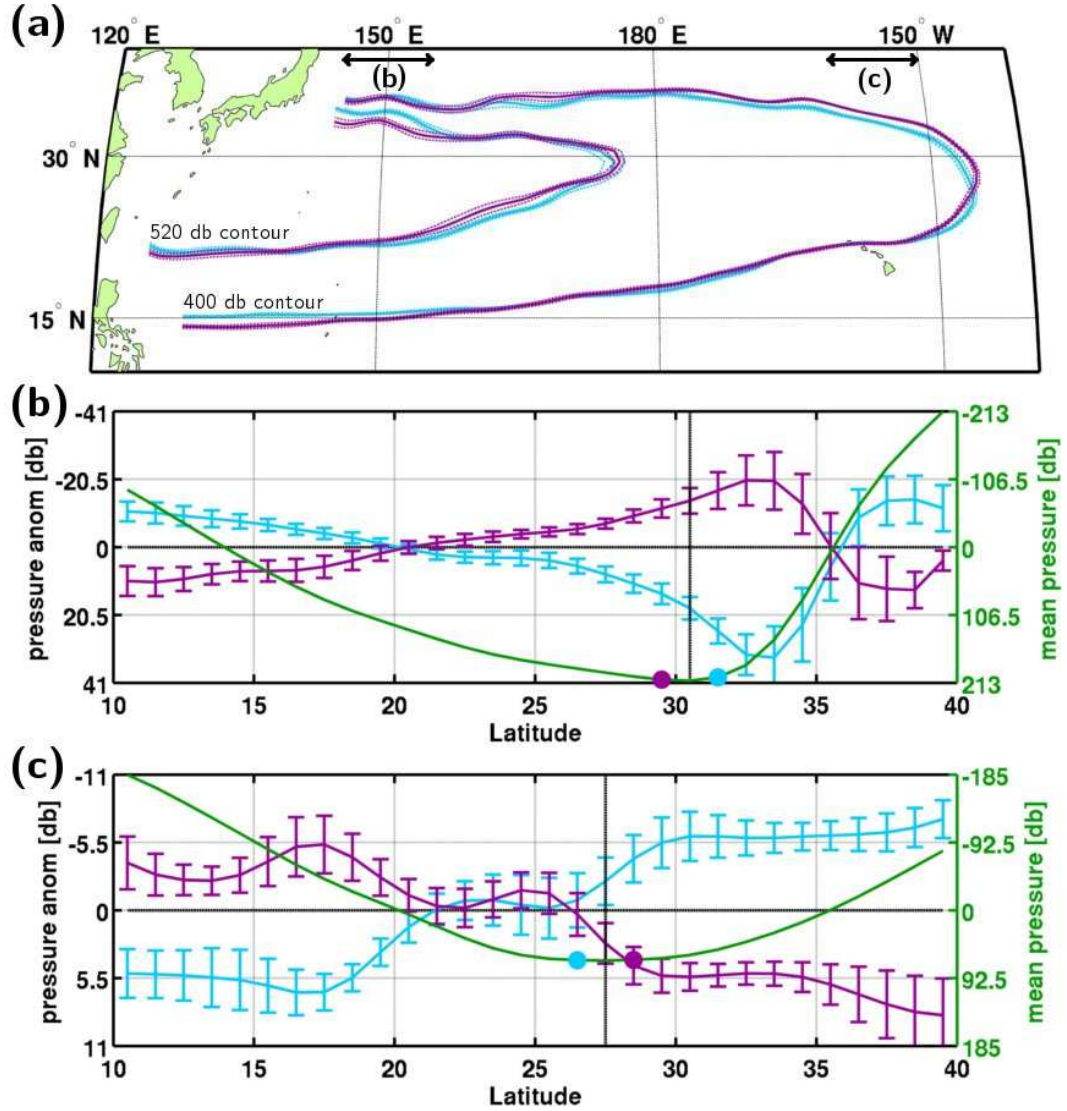


Figure 2.6: (a) Schematic of the gyre movement, showing the 400 db and 520 db pressure contour on the $\sigma_p = 26.4 \text{ kg/m}^3$ isopycnal surface. Cyan line: position in 2004 – 2005 (error bars are dotted). Purple line: same, but in 2008 – 2009. (b) Green line: temporal mean field of pressure (db) on the isopycnal $\sigma_p = 26.4 \text{ kg/m}^3$, averaged over 150 – 160°E; the spatial mean is removed at every point in time. Cyan and purple lines: anomaly of the same variable averaged in 2004 – 2005 (cyan line) and in 2008 – 2009 (purple line). Error bars are also shown. The vertical black line indicates the mean location of maximum pressure; circles show the location of the same in 2004 – 2005 (cyan circle) and in 2008 – 2009 (purple circle). (c) same as (b), but the zonal average is over 160 – 150°W.

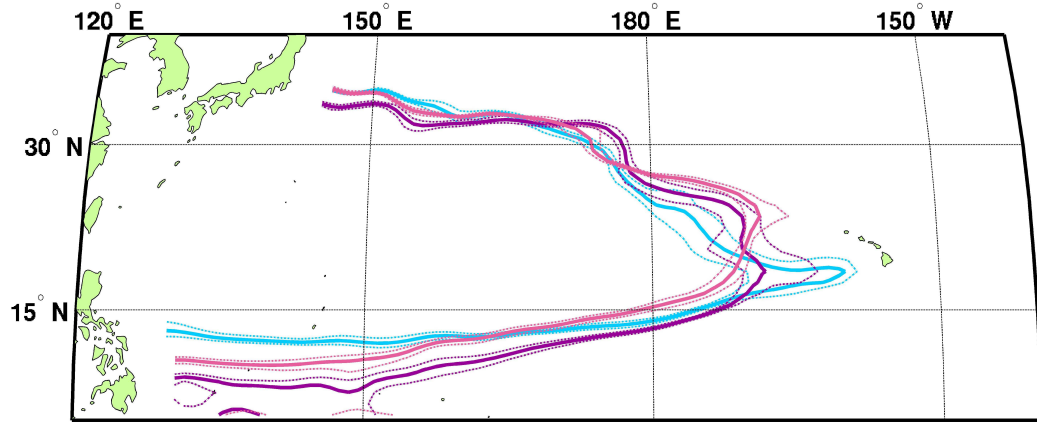


Figure 2.7: (a) Schematic of the gyre movement in different years, showing the 260 cm SSH contour from AVISO data. Pink line: position in 1999 – 2000 (error bars are dotted). Cyan line: same, but in 2004 – 2005. Purple line: same, but in 2008 – 2009.

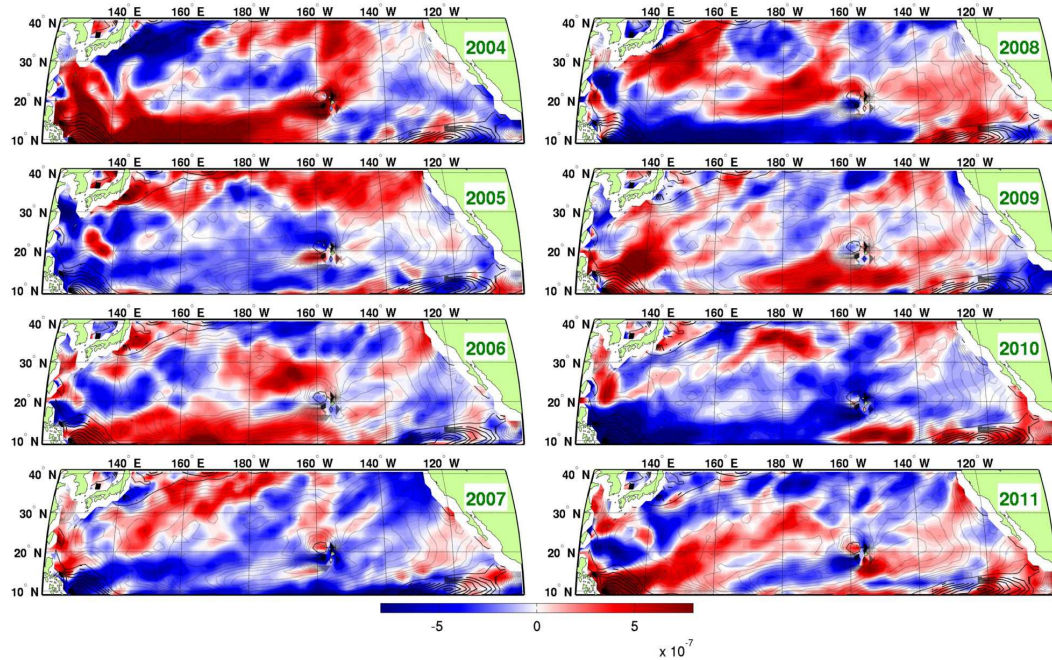


Figure 2.8: ECMWF Ekman upwelling (m/s) for each year considered in this study. Colors indicate anomalies relative to the 2004 – 2011 temporal mean: the anomalies in each panel are annually averaged. Thin contours represent the time-averaged mean field: lines are gray if values are negative, black if positive.

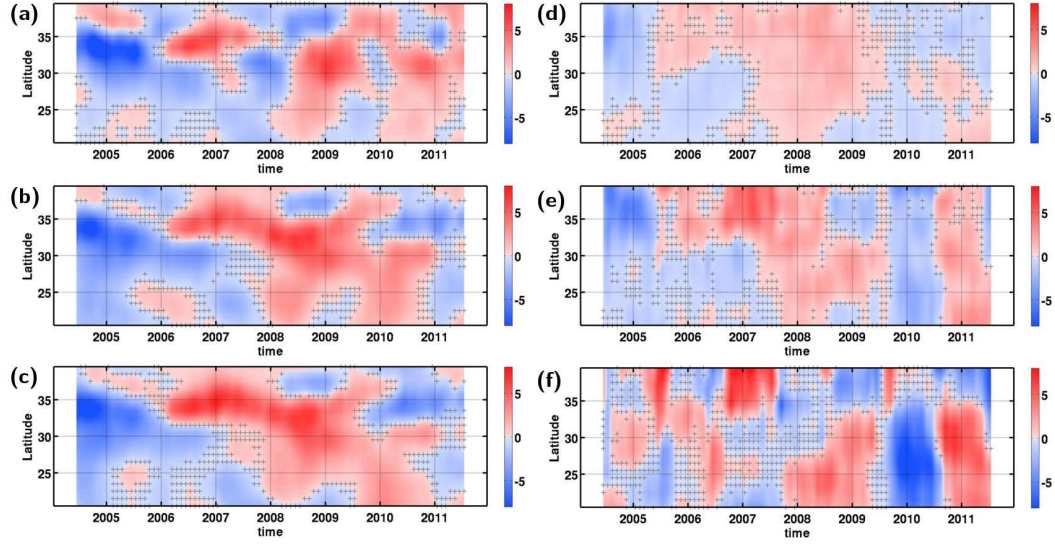


Figure 2.9: Meridional geostrophic transport anomaly (Sv) from Argo (a,b,c) and from the Sverdrup relation using ECMWF Ekman upwelling (d,e,f): 11-month running mean timeseries versus latitude. Transport is computed between $150^{\circ}E$ and different longitudes to the east: (a,d) $150 - 160^{\circ}E$; (b,e) $150 - 180^{\circ}E$; (c,f) $150^{\circ}E - 140^{\circ}W$. Gray + symbols indicate points where the transport anomaly is not statistically different from zero.

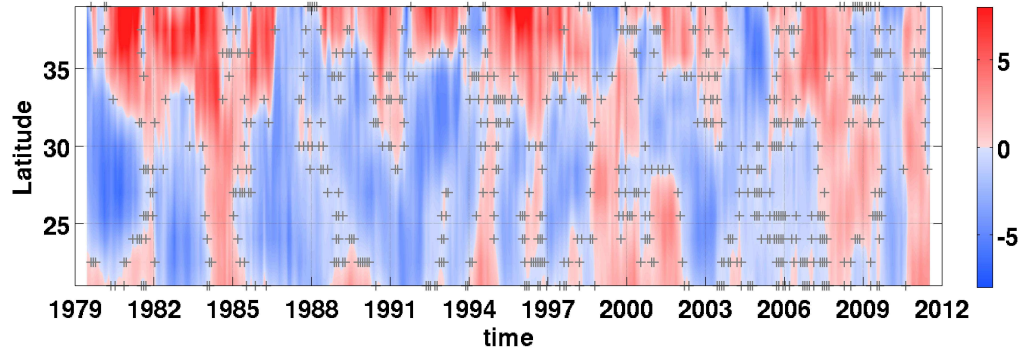


Figure 2.10: Meridional geostrophic transport anomaly (Sv) computed from the Sverdrup relation and ECMWF Ekman upwelling between $150^{\circ}E$ and $180^{\circ}E$: 11-month running mean timeseries versus latitude in 1979 – 2011. Gray + symbols indicate points where the transport anomaly is not statistically different from zero.

Chapter 3

Understanding the annual cycle in global steric height

Abstract. Steric variability in the ocean includes diabatic changes in the surface layer due to air-sea buoyancy fluxes and adiabatic changes due to advection, which are dominant in the subsurface ocean. Here, the annual signal in subsurface steric height (η' below 200 db) is computed on a global scale using temperature and salinity profiles from Argo floats. The zonal average of $\Delta\eta'$ over a season (e.g. $\eta'_{March} - \eta'_{December}$) is compared to the wind-forced vertical advection contribution ($\Delta\eta'_w$) both in the global ocean and in different basins. The results show agreement that extends beyond the tropics. The estimate of $\Delta\eta'_w$ is based on the Ekman pumping and assumes that the seasonal vertical velocity is constant over the depth range of interest. This assumption is consistent with annual isopycnal displacements inferred from Argo profiles. The contribution of horizontal advection to $\Delta\eta'$ is significant in some regions and consistent with differences between $\Delta\eta'$ and $\Delta\eta'_w$.

3.1 Introduction

Observations of the annual cycle in global subsurface temperature and salinity were very limited prior to the implementation of the Argo Program [Roemmich and Gilson, 2009]. Before the array was initially deployed (2000 – 2007), very few salinity measurements were available on a planetary scale, the data had a strong bias toward the Northern Hemisphere and summer months, and the typical maximum depth of temperature observations was 750 m. Argo provides unprecedented resolution and near-global coverage, with observations as deep as 2000 db and the first systematic sampling of subsurface salinity. Fig.3.A1 in the appendix illustrates, for instance, the extreme sparseness of Southern Hemisphere mid-ocean winter data in the historical archive and shows how the spatial and temporal sampling biases have been greatly reduced in Argo. The new data improve our understanding of seasonal variability in the ocean, which is dominated by air-sea exchanges of heat and freshwater in the surface layer, but which is also driven adiabatically by dynamics and in some regions extends deep into the water column. Here the focus is the subsurface ocean and in particular the relationship between the steric height seasonal cycle and wind-forced vertical advection [Gill and Niller, 1973]. A goal is better separation of steric variability into its diabatic and adiabatic components.

Gill and Niller [1973] present the theory of large-scale seasonal variations of sea-level. The steric height component is described in terms of an upper and a lower baroclinic term, to consider separately the effect of steric changes above and below 200 m. They perform a dimensional analysis of the equations for momentum and mass conservation and show that vertical advection of the mean density field by seasonal Ekman pumping velocity balances, at first order, the large-scale tendency of seasonal steric height below 200 m $\left(\frac{\partial \eta'}{\partial t}\right)$. Their assumption regarding the seasonal vertical velocity w' is a linear decay of the Ekman pumping velocity w'_{Ek} (applied at $z = -200$ m) to zero at the bottom ($z = -H$), i.e. $w' = w'_{Ek}(1 + z/H)$. Gill and Niller [1973] expect η' to be fairly small, about 10% of the total steric height signal, but point out that a test of the theory would require measurements of this deeper steric variability, to compare with the estimate from Ekman pumping

velocity. The energy in the background non-seasonal continuum tends to yield noisy observations of η' , but with Argo's temporal and spatial resolution on a planetary scale, the comparison can now be achieved both globally and in different basins. This is the goal of the present study.

Previous work shows that seasonal variations of the meridional overturning circulation extend as deep as 1000 m and that they are related to zonal wind stress and to the east-west slope of sea level [Kanzow *et al.*, 2010; Liu *et al.*, 2011]. Also, subsurface seasonal signals are observed propagating along the equator and in the tropics [Brandt and Eden, 2005; Hosoda *et al.*, 2006; Bunge *et al.*, 2008; Johnson, 2011], sometimes to the full depth of Argo observations in the region [Hosoda *et al.*, 2006; Johnson, 2011]. In the present analysis, zonal averages of the data are considered, and propagation is not discussed.

This study is based on observations and it describes how seasonal steric height in the subsurface ocean relates to wind-driven vertical advection. The comparison is carried out following the theory of Gill and Niller [1973] both on a global scale and in different ocean basins and the focus is on extra-equatorial regions, where the Ekman dynamics hold and the seasonal vertical velocity (w') can be estimated from Ekman pumping. Section 3.2 presents the data. Section 3.3 describes the results, including also a discussion of the depth dependence of w' and of the role of horizontal advection. Section 3.4 presents a summary and conclusions.

3.2 Data

3.2.1 Argo data

Argo profiles provide temperature, salinity, and pressure data on a global scale ($60^\circ S$ - $60^\circ N$) and as deep as 2000 db [Roemmich *et al.*, 2009]. In this analysis, the raw Argo temperature and salinity measurements during 2005 – 2012 were gridded monthly on a $1^\circ \times 1^\circ$ grid, after quality control and adjustment of pressure bias. The gridding was done by objective mapping, with latitude-dependent decorrelation scales [Roemmich and Gilson, 2009]. In order to test

how this procedure may affect the seasonal signal in lightly sampled regions, an alternate gridding method was applied to each month of the 12-month climatology (i.e. the January map combines all the January data during 2005 – 2012). Bin-averaged data in overlapping large-scale regions of 555 km radius, yield results that are little different from the objective mapping case, and are therefore not discussed further.

3.2.2 ECMWF data

Zonal and meridional wind momentum flux data used in this analysis are synoptic monthly means from the European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis, ERA-Interim, for the period 2005 – 2012. *Dee et al.* [2011] describe the ERA-Interim data. *Brunke et al.* [2011] found that for wind stress and latent heat fluxes, ERA-Interim performed better than most other products.

3.3 Results

Consistent with the theory of *Gill and Niller* [1973], the large-scale seasonal tendency of steric height $\left(\frac{\partial \eta'}{\partial t}\right)$, from Argo, is expected to balance, at first order, vertical advection (estimated from ERA-Interim Ekman pumping) of the Argo mean potential density field, i.e.

$$\frac{\partial \eta'}{\partial t}(y, p, t) = \frac{1}{x_2 - x_1} \int_{x_1}^{x_2} \left(- \int_{p_{ref}}^p \frac{1}{\bar{\rho}} \frac{\partial \bar{\rho}}{\partial p} w' dp \right) dx. \quad (3.1)$$

In Eq.3.1, η' is a basin-scale zonal average of seasonal steric height between longitudes x_1 and x_2 , hence it depends only on latitude, pressure and time. η' is defined as $\eta' = \frac{1}{x_2 - x_1} \int_{x_1}^{x_2} \left(-\frac{1}{g} \int_{p_{ref}}^p \frac{\rho'}{\bar{\rho}^2} dp \right) dx$, where g is the gravitational acceleration, $p_{ref} = 2000$ db is the reference pressure, ρ is Argo potential density and the upper limit on p is 200 db. Here, the prime symbol indicates that values are mean seasonal deviations from the time average over the complete timeseries (2005 – 2012), hence $\rho' \ll \bar{\rho}$ (where the bar indicates the time mean). The right

hand side of Eq.3.1 is taken to be a basin-scale zonal average and w' is the seasonal vertical velocity based on ERA-Interim Ekman pumping (see Section 3.3.1). The integral of this term over a season, provides an estimate of the vertical advection contribution to η' during that season ($\Delta\eta'_w$ in the following) and can be compared to how η' changes from the beginning to the end of the season ($\Delta\eta'$).

Steric height is chosen for the comparison between the atmospheric forcing and the oceanic response since it is an integral quantity. Section 3.3.1 presents the assumption made for the depth-dependence of w' . Section 3.3.2 shows observations of the balance in Eq.3.1, comparing Argo $\Delta\eta'$ and $\Delta\eta'_w$ from Ekman pumping. Section 3.3.3 is a discussion of the role of horizontal advection.

3.3.1 Depth dependence of the seasonal vertical velocity

$$w'$$

The estimate of $\Delta\eta'_w$ in Eq.3.1 requires an assumption regarding the vertical profile of w' . In *Gill and Niller* [1973], this assumption is a linear decay from seasonal Ekman pumping (w'_{Ek}) at 200 m to zero at the ocean bottom. In the current study, instead, the depth range of interest is 200 – 2000 db, where Argo observations are available, and w' is considered constant in pressure ($w'(p) = w'_{Ek}$ below 200 db), as suggested by Argo isopycnal displacement. Argo isopycnal displacement provides an estimate of the seasonal vertical velocity, i.e. $w'_\rho = -\frac{\partial \rho'}{\partial t} \left(\frac{\partial \bar{\rho}}{\partial z} \right)^{-1}$. This assumes that all of the seasonal density signal is caused by vertical displacement of isopycnals and it does not account for other mechanisms including surface buoyancy fluxes, horizontal advection and mixing. Surface buoyancy fluxes play a role only in the upper ocean, but horizontal advection and mixing may be relevant also at depth. Despite these limitations, w'_ρ still provides useful information on the depth dependence of w' where horizontal advection and mixing are small, i.e. those regions where w'_ρ is consistent with w'_{Ek} . A comparison between w'_ρ and w'_{Ek} during the Northern Hemisphere winter can be seen in Fig.3.1 (December to February average). w'_ρ and w'_{Ek} are zonally averaged in the Pacific Ocean (panels a,b) and in the Indian Ocean (c,d) and $w'_\rho(700)$ is consistent with w'_{Ek} (panels a,c).

Also, $w'_\rho(1300)$ is not very different from $w'_\rho(700)$ (panels a,c) and, in general, w'_ρ has a large vertical scale (panels b,d in Fig.3.1 and Section 3.A.2). This is consistent with the constant vertical profile assumed for w' in the present study. Similar results are found during the Northern Hemisphere summer and in the Atlantic (see Fig.3.A3 -3.A5 in the appendix). While w'_ρ clearly has a large vertical scale, the Gill and Niller assumption that w' linearly decreases with depth from the Ekman pumping velocity w'_{Ek} (applied at $z = -200$ m) to zero at the bottom provides similar and equally consistent results in the depth range of interest. The same is expected for an alternate vertical structure made from the lowest baroclinic modes.

3.3.2 Argo seasonal steric height and wind-forced vertical advection

Fig.3.2 shows a comparison between the global zonal average of Argo $\Delta\eta'$ and the vertical advection contribution alone ($\Delta\eta'_w$ from Ekman pumping), during the Northern Hemisphere winter and summer. $\Delta\eta'$ is computed directly taking the difference between the beginning and the end of the season (i.e. η' on 1st March minus η' on 1st December and η' on 1st September minus η' on 1st June), while $\Delta\eta'_w$ is the time integral of the right-hand side of Eq.3.1 from December to the end of February and from June to the end of August. Also, black shaded values are smaller than one standard error, where the standard error computation is based on the standard deviation of $\Delta\eta'$ and $\Delta\eta'_w$ over the 8 years (2005-2012) of Argo having adequate global coverage (i.e. 8 independent realizations for each season). As described in *Gill and Niller* [1973], Fig.3.2 shows that the wind-forced vertical advection contribution alone (panels a and c) has similar spatial patterns to the total $\Delta\eta'$ from Argo (b and d). The agreement is significant and extends beyond the tropics for both seasons, although Argo $\Delta\eta'$ has more structure than $\Delta\eta'_w$. This may indicate the presence of strong currents, where seasonal horizontal advection could play a role and alter the balance in Eq.3.1. Differences between $\Delta\eta'$ and $\Delta\eta'_w$ may, also, result from keeping the reference level at 2000 db. Variability below the Argo depth range could be estimated using the quasigeostrophic formalism,

projecting the Argo density field onto the interior barotropic and first baroclinic modes, similarly to [Wang *et al.*, 2013], but this is beyond the scope of the present analysis. Also, while seasonal mixing is observed in the subsurface ocean both in the tropics and in the extra-tropics [Wu *et al.*, 2011; Whalen *et al.*, 2012], it is not addressed here. Finally, in the tropics, deep temperature variability below the thermocline is forced by seasonal Rossby waves [Hosoda *et al.*, 2006; Johnson, 2011], but the present study describes zonal averages of $\Delta\eta'$, globally and in each ocean basin, and propagation is not discussed. The ocean observations (i.e. $\Delta\eta'$) do include the effect of waves, but $\Delta\eta'_w$ does not account for propagation of the wind-induced signals. Since propagation is mostly zonal, the basin-wide zonal integration minimizes the fraction of signal that propagates out of the region during the season, hence the error for not accounting for propagation in $\Delta\eta'_w$.

While Fig.3.2 is global, a similar result is found in zonal averages in the Pacific Ocean (Fig.3.3, (a) and (b)), in the Indian Ocean (Fig.3.3, (c) and (d)) and in the Atlantic Ocean (see Fig.3.A6 in the appendix), for both winter and summer seasons. Northern Hemisphere summer anomalies are of opposite sign to winter and are not shown (see Fig.3.A7 in the appendix). In each basin, the comparison between $\Delta\eta'$ and the vertical advection component deteriorates towards the poles (generally north of 50° latitude), consistent with the increase of unresolved mesoscale features at higher latitudes.

3.3.3 Horizontal advection contribution to $\Delta\eta'$

The zonal and meridional advection contributions to $\Delta\eta'$ (in the following, $\Delta\eta'_u$ and $\Delta\eta'_v$) are computed during the Northern Hemisphere winter, from Argo geostrophic velocity, i.e. $\Delta\eta'_u = \int_{\text{Dec}}^{\text{Feb}} \left(\frac{1}{g} \int_{p_{ref}}^p \frac{1}{\bar{\rho}^2} u'_g \frac{\partial \bar{\rho}}{\partial x} dp \right) dt$ and $\Delta\eta'_v = \int_{\text{Dec}}^{\text{Feb}} \left(\frac{1}{g} \int_{p_{ref}}^p \frac{1}{\bar{\rho}^2} v'_g \frac{\partial \bar{\rho}}{\partial y} dp \right) dt$. u'_g and v'_g are seasonal geostrophic velocity from Argo (i.e. referenced to 2000 db), g is the gravitational acceleration, $p_{ref} = 2000$ db is the reference pressure and ρ is Argo potential density, as in Eq.3.1.

Fig.3.4 shows a comparison of $\Delta\eta'_u$ (in cyan), $\Delta\eta'_v$ (in blue), $\Delta\eta'_w$ (in green) and $\Delta\eta'$ (in red), spatially averaged in two different regions: 44 – 40°S in the Indian Ocean (panel a) and 34 – 30°S in the Atlantic Ocean (panel b). Dashed

lines indicate the seasonal value plus and minus one standard error. The horizontal advection contribution is small and the error bars are relatively large, consistent with the increase of unresolved mesoscale features in the extra-tropics, but the difference between the observed $\Delta\eta'$ from Argo (in red) and the vertical advection component $\Delta\eta'_w$ (in green) is consistent with $\Delta\eta'_v$ (panel a, in blue) and $\Delta\eta'_u$ (panel b, in cyan). In general, caution is needed in asserting the relative importance of horizontal and vertical advection, since results depend on the geographical area of averaging. Unfortunately, where strong currents are present (i.e. where horizontal advection is most important), there are also mesoscale eddies that are unresolved by Argo sampling, but a longer data record will allow a more local analysis at least in some regions. Also, the effect of wave adjustment and propagation is present both in the Argo steric height and in the estimate from observed horizontal advection and has a major role in some regions [*Qiu and Lukas*, 1996; *Chen and Qiu*, 2004; *Hsin and Qiu*, 2012], but is not discussed here.

Adding the surface absolute velocity anomaly from AVISO altimetric data [*Ducet et al.*, 2000] to the Argo geostrophic velocity relative to the surface (i.e. using an absolute velocity anomaly) does not yield statistically different results in the two regions of interest and it makes the estimate noisier due to mesoscale features that are present in AVISO but less so in the smoother Argo gridding.

3.4 Summary and conclusions

The unprecedented global coverage and accuracy of Argo temperature and salinity profiles make it possible to observe the seasonal steric height deviation (η') on global and basin scales during 2005 – 2012. Steric variability in the ocean includes diabatic changes in the surface layer due to air-sea buoyancy fluxes and adiabatic contributions from advection, which is most important below the surface layers. In the subsurface ocean, Argo $\Delta\eta'$ over a season agrees with vertical advection of the mean potential density field by Ekman pumping ($\Delta\eta'_w$), consistent with the theory of *Gill and Niller* [1973], and the agreement extends beyond the tropics. The current analysis compares zonal averages both in the global ocean

and in ocean basins, and describes their latitude-dependence, that is a primary factor in annual cycle signals and noise. Neither the role of propagation nor the role of mixing are discussed. Horizontal advection gives a significant contribution in some regions and is consistent with the differences between $\Delta\eta'$ from Argo and $\Delta\eta'_w$.

The depth dependence of the seasonal vertical velocity (w') is also discussed, as the expression for $\Delta\eta'_w$ is based on the assumption that $w'(p) = w'_{Ek}$, that is, w' constant in pressure over the depth range of interest. While the estimate of w' from Argo isopycnal displacement (w'_ρ) is consistent with w'_{Ek} and shows a large vertical scale, the Gill and Niller assumption that w' linearly decreases with depth from the Ekman pumping velocity w'_{Ek} (applied at $z = -200$ m) to zero at the bottom provides similar results in the depth range of interest.

As *Gill and Niller* [1973] point out, although the dynamic component of the seasonal baroclinic signal is small at depth, it is useful to test the theory using observations. Moreover, this test might be regarded as a measure of Argo's ability, based on 8 years of global coarse resolution profiles, to estimate the annual cycle. Lastly, in some regions, the heat redistribution that corresponds to the dynamically-driven Argo steric height variability in the subsurface ocean amounts to 10 – 15% of the seasonal cycle heat gain in the upper ocean (not shown) and should not be neglected when trying to close the heat balance in the surface layer. With more Argo profiles, it will be possible to extend the analysis in this study beyond large scale zonal averages and to assess the role of propagation in seasonal steric height on a global scale. With the current dataset, averages over smaller areas yield results that are significant only in some regions.

Acknowledgments. The Argo data used here were collected and are made freely available by the International Argo Program and by the national programs that contribute to it. The authors' participation in the Argo Program was supported by U.S. Argo through NOAA Grant NA10OAR4320156 (SIO CIMEC). The statements, findings, conclusions, and recommendations herein are those of the authors and do not necessarily reflect the views of the National Oceanic and Atmospheric Administration or the Department of Commerce. The efforts of many

international partners in planning and implementing the Argo array are gratefully acknowledged. ECMWF ERA-Interim data used in this study have been obtained from the ECMWF Data Server. The AVISO altimeter products were produced by the CLS Space Oceanography Division as part of the Environment and Climate EU ENACT project (EVK2-CT2001-00117) and with support from CNES. Valuable suggestions for this work were provided by two anonymous reviewers. The objectively mapped Argo data were provided by John Gilson.

Chapter 3, in full, is a reproduction of the material as it appears in *D. Giglio, D. Roemmich, and B. Cornuelle* [2013], *Geophys. Res. Lett.*, *40*, 4349–4354. The dissertation author was the primary investigator and author of this work.

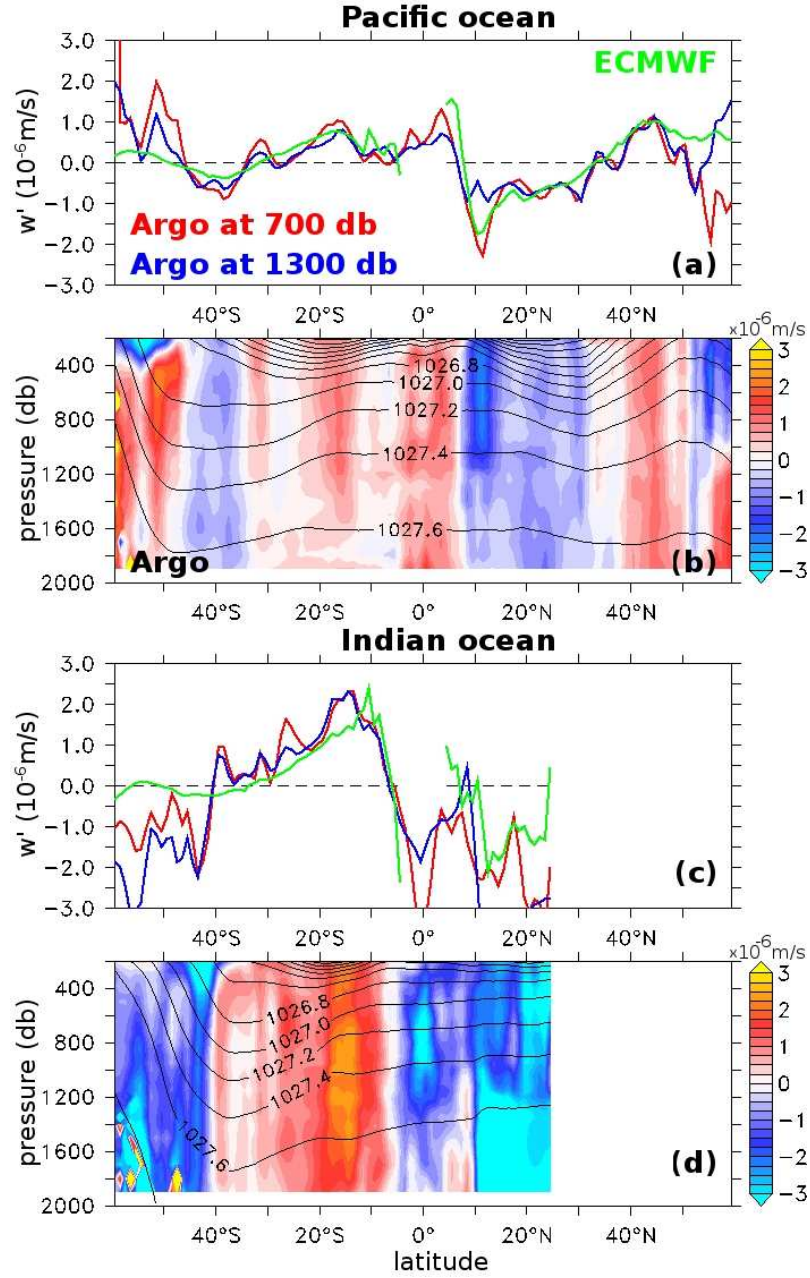


Figure 3.1: Seasonal vertical velocity from Argo isopycnal displacement (w'_ρ): zonal average in the Pacific Ocean (a,b) and in the Indian Ocean (c,d) during the Northern Hemisphere winter. In (a,c), the green line is based on ERA-Interim wind stress Ekman pumping (w'_{Ek}); the red and the blue lines are w'_ρ at 700 db (red line) and 1300 db (blue line). In (b,d), the color is w'_ρ from Argo isopycnal displacement; the black contours are the temporal mean field of Argo potential density, zonally averaged in the same regions.

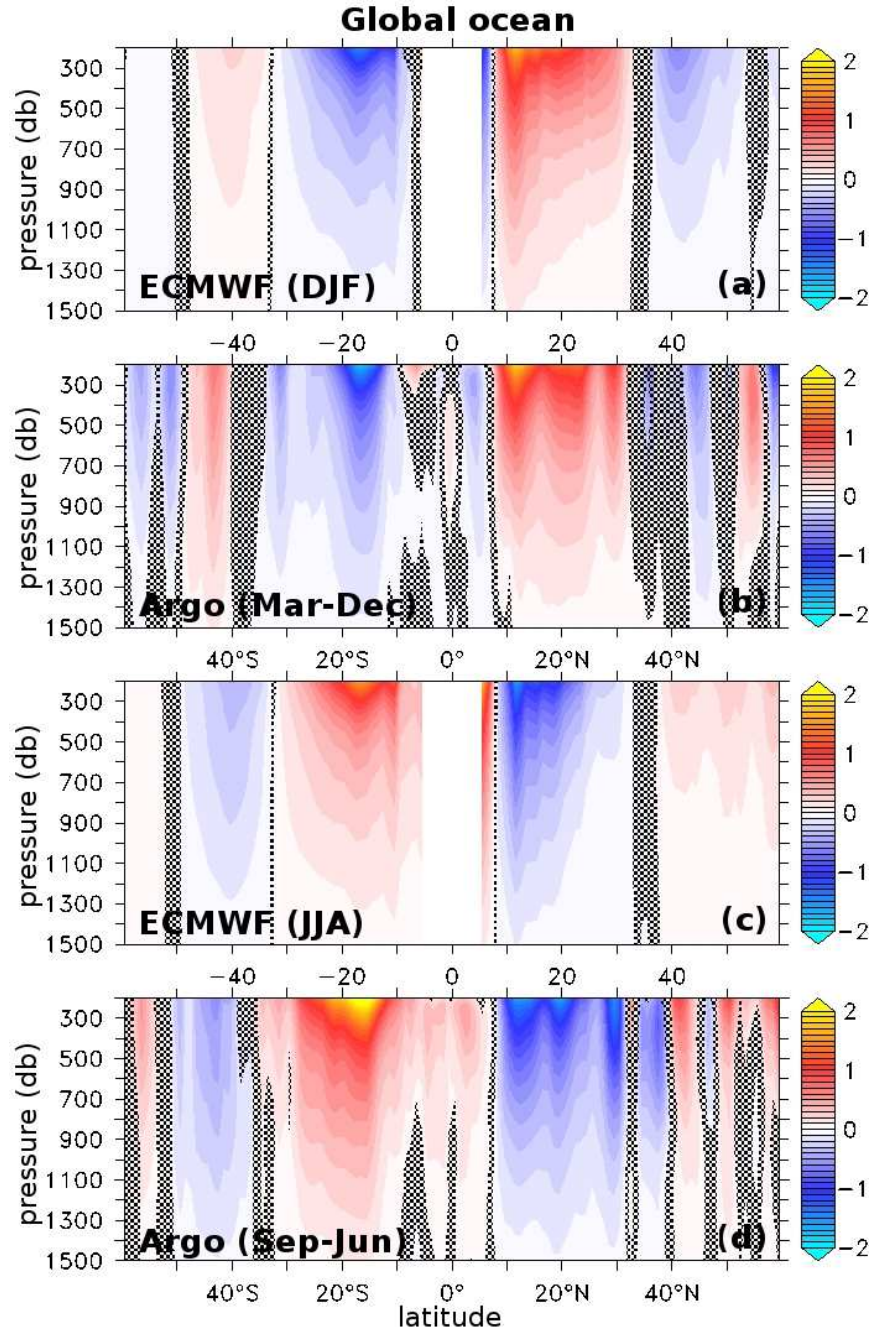


Figure 3.2: Seasonal steric height change during the Northern Hemisphere winter (panels a,b) and summer (panels c,d): zonal average in the global ocean. Panels (a) and (c) show the contribution of vertical advection ($\Delta\eta'_w$). Panels (b) and (d) show $\Delta\eta'$ from Argo. Changes smaller than one standard error are shaded black.

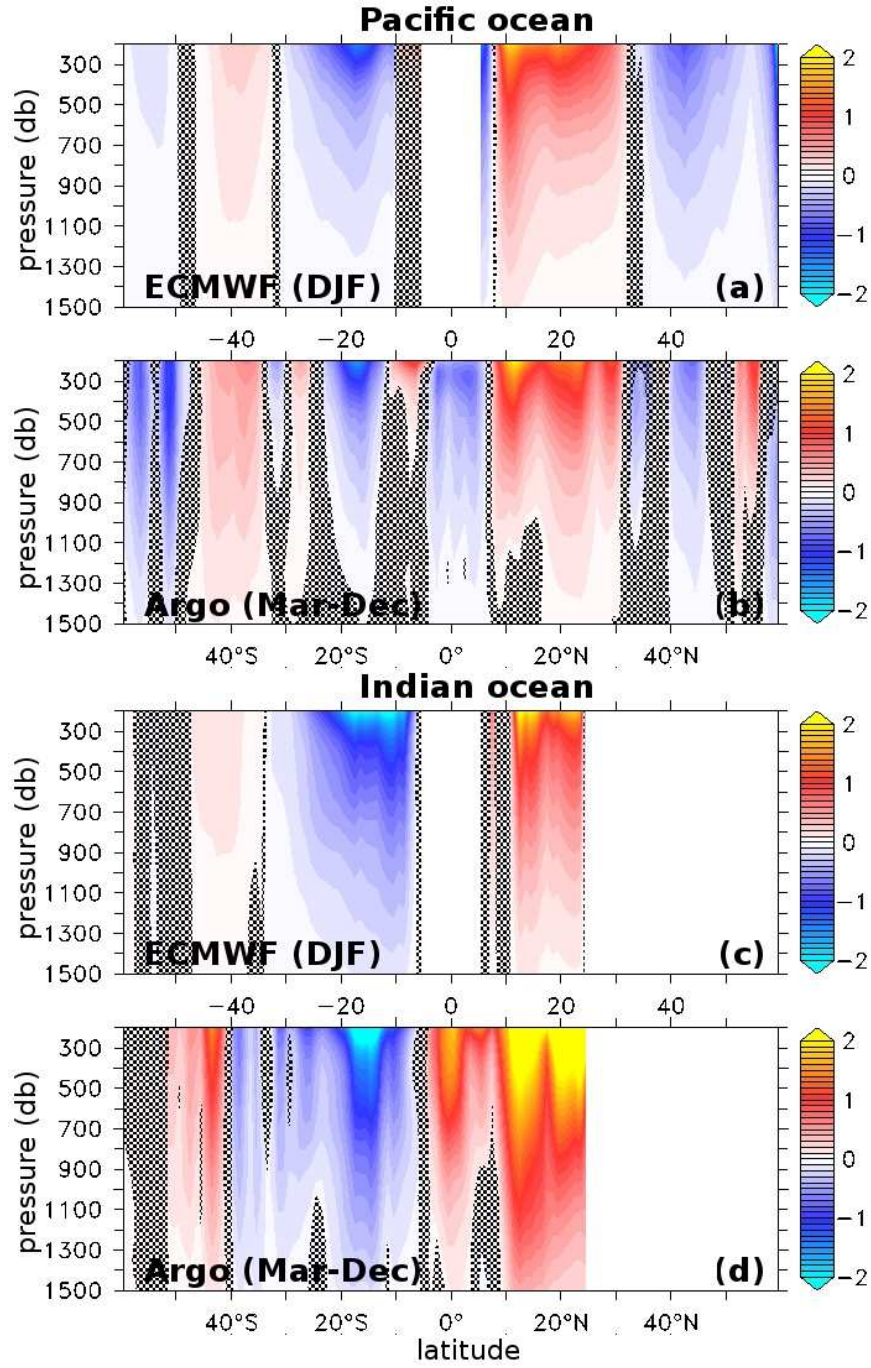


Figure 3.3: Seasonal steric height change during the Northern Hemisphere winter: zonal average in the Pacific ocean (a,b) and in the Indian ocean (c,d). Panels (a) and (c) show the contribution of vertical advection ($\Delta\eta'_w$). Panels (b) and (d) show $\Delta\eta'$ from Argo. Changes smaller than one standard error are shaded black.

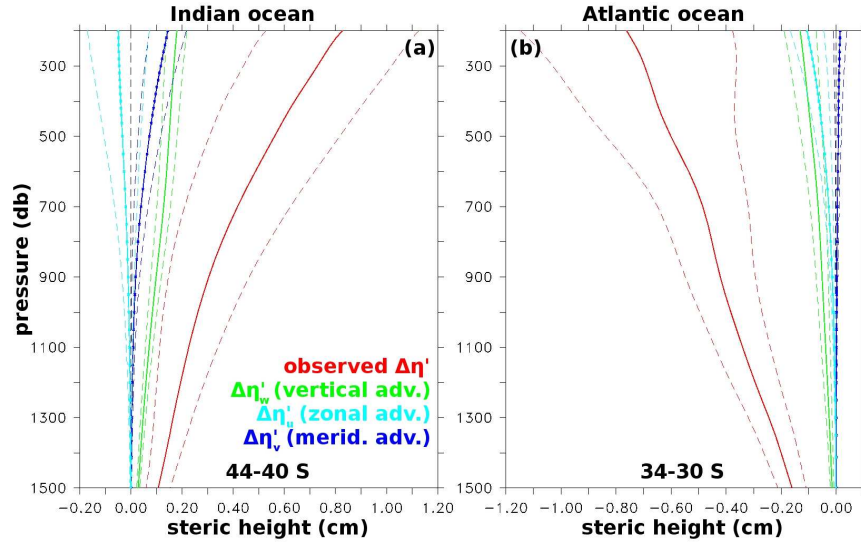


Figure 3.4: Seasonal steric height change during the Northern Hemisphere winter: area weighted average along $20.5 - 145.5^\circ\text{E}$, $44 - 40^\circ\text{S}$ in the Indian Ocean (panel a) and $49.5^\circ\text{W} - 15.5^\circ\text{E}$, $34 - 30^\circ\text{S}$ in the Atlantic Ocean (panel b). The solid red line is Argo $\Delta\eta'$; the advective contributions are in green (vertical advection, $\Delta\eta'_w$), cyan (zonal, $\Delta\eta'_u$) and blue (meridional, $\Delta\eta'_v$). Dashed lines indicate error bars.

3.A Appendix

This appendix includes additional figures (Fig.3.A1 - Fig.3.A7). Further details about these figures can be found in dedicated subsections below or in the main text of this chapter.

3.A.1 Data coverage: the historical dataset and Argo

The historical dataset not only contains many fewer T/S profiles than the million profiles obtained by Argo, it is strongly biased toward the Northern Hemisphere, toward the continents, and toward summer. Fig.3.A1 illustrates the extreme sparseness of historical Southern Hemisphere mid-ocean winter data and shows how the spatial and temporal sampling biases have been greatly reduced in Argo. In particular, the upper panel of Fig.3.A1 shows the number of CTD casts and station data (per one-degree box) for temperature and salinity at least to 1000 m, during the Southern Hemisphere winter (JJA) in 1950 – 2000, while the lower panel is similar but for Argo floats profiles and only in 2008 – 2010. Argo has provided more than an order of magnitude increase in Southern Ocean winter data compared with the historical archive. Even with this large improvement in Argo relative to historical sampling, the mid-latitude oceans are very noisy, and despite the long zonal averaging in our analysis, results are still of marginal significance at some latitudes.

3.A.2 Additional comments to Section 3.3.1

Fig.3.A2 quantifies the ratio of w'_ρ at 700 db to w'_ρ at 1300 db, where w'_ρ is estimated from isopycnal displacement and zonally averaged in the global ocean during the Northern Hemisphere winter. The estimation of w' from isopycnal displacement may not work well where meridional gradients of potential density are high, such as the latitudes of the separated western boundary currents at about 38° north and south. However, the ratio displayed in Fig.3.A2 still provides some insight regarding the vertical scale of w' , and some justification for

our choice of a non-decaying vertical structure (ratio=1). In the tropics, the ratio $w'_\rho(700)/w'_\rho(1300)$ is generally close to 1, indicating that w' does not change much over that depth range. Around 10°N, though, w'_ρ decreases below 1200 db and the ratio is greater than 1. As discussed in the manuscript, the steric height results in this study describe the top 2000 db of the ocean in an integrated sense and are not affected greatly by whether the vertical structure chosen for w' is constant or linearly decaying, as in Gill and Niiler (1973).

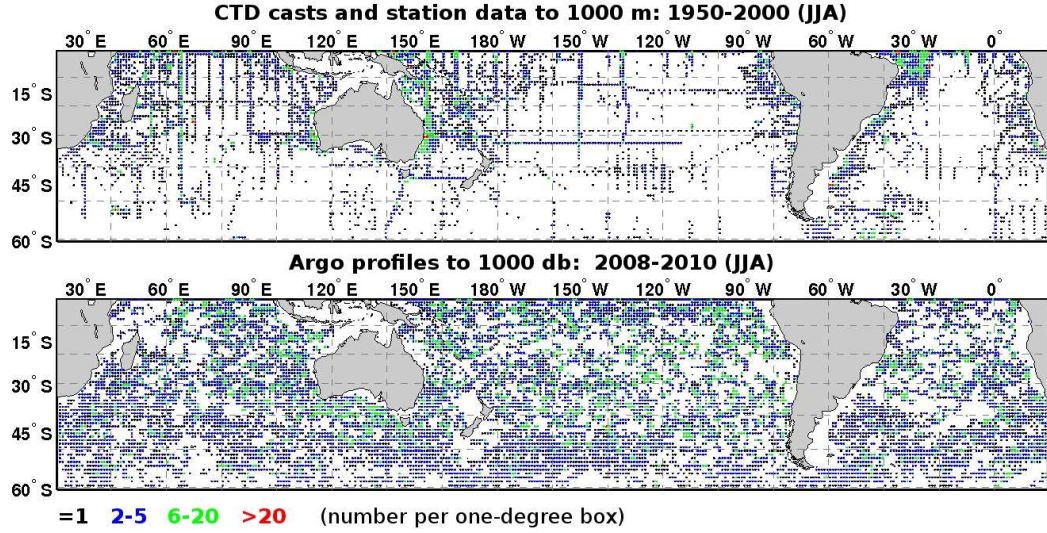


Figure 3.A1: Upper panel: CTD casts and station data (per one-degree box) for temperature and salinity at least to 1000 m, during the Southern Hemisphere winter (JJA) in 1950 – 2000. Lower panels: same as the upper panel but for Argo floats profiles and only in 2008 – 2010.

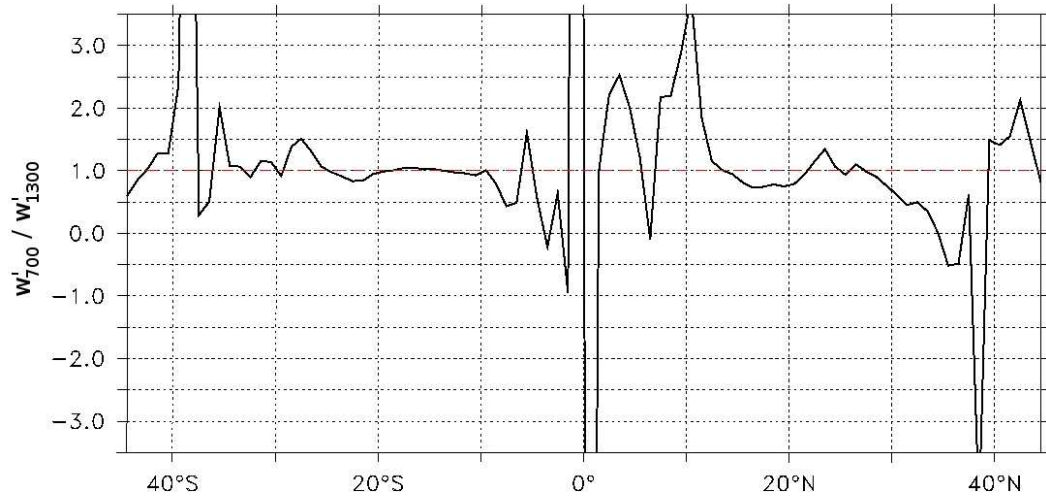


Figure 3.A2: Ratio between the seasonal vertical velocity from Argo isopycnal displacement (w'_p) at 700 db and at 1300 db. w'_p is zonally averaged in the global ocean during the Northern Hemisphere winter.

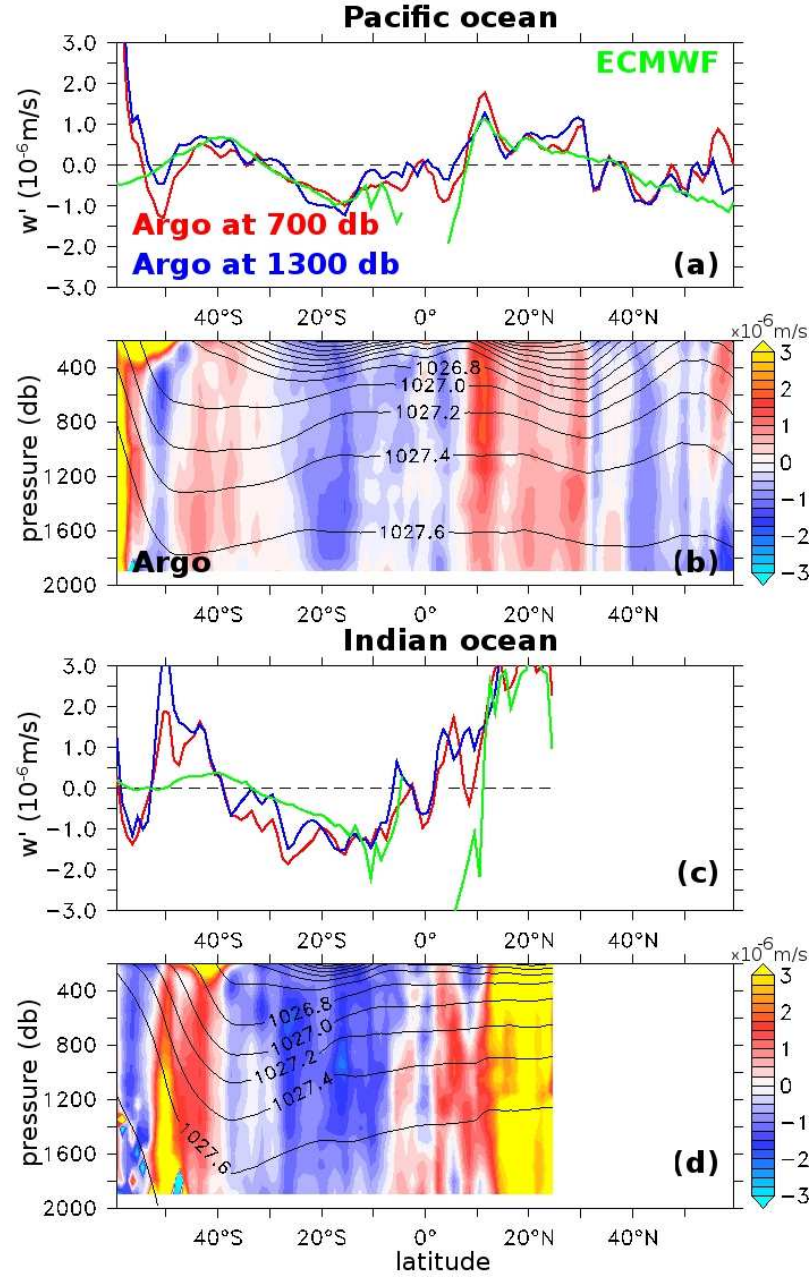


Figure 3.A3: Seasonal vertical velocity from Argo isopycnal displacement (w'_ρ): zonal average in the Pacific Ocean (a,b) and in the Indian Ocean (c,d) during the Northern Hemisphere summer. In (a,c), the green line is based on ERA-Interim wind stress Ekman pumping (w'_{Ek}); the red and the blue lines are w'_ρ at 700 db (red line) and 1300 db (blue line). In (b,d), the color is w'_ρ from Argo isopycnal displacement; the black contours are the temporal mean field of Argo potential density, zonally averaged in the same regions.

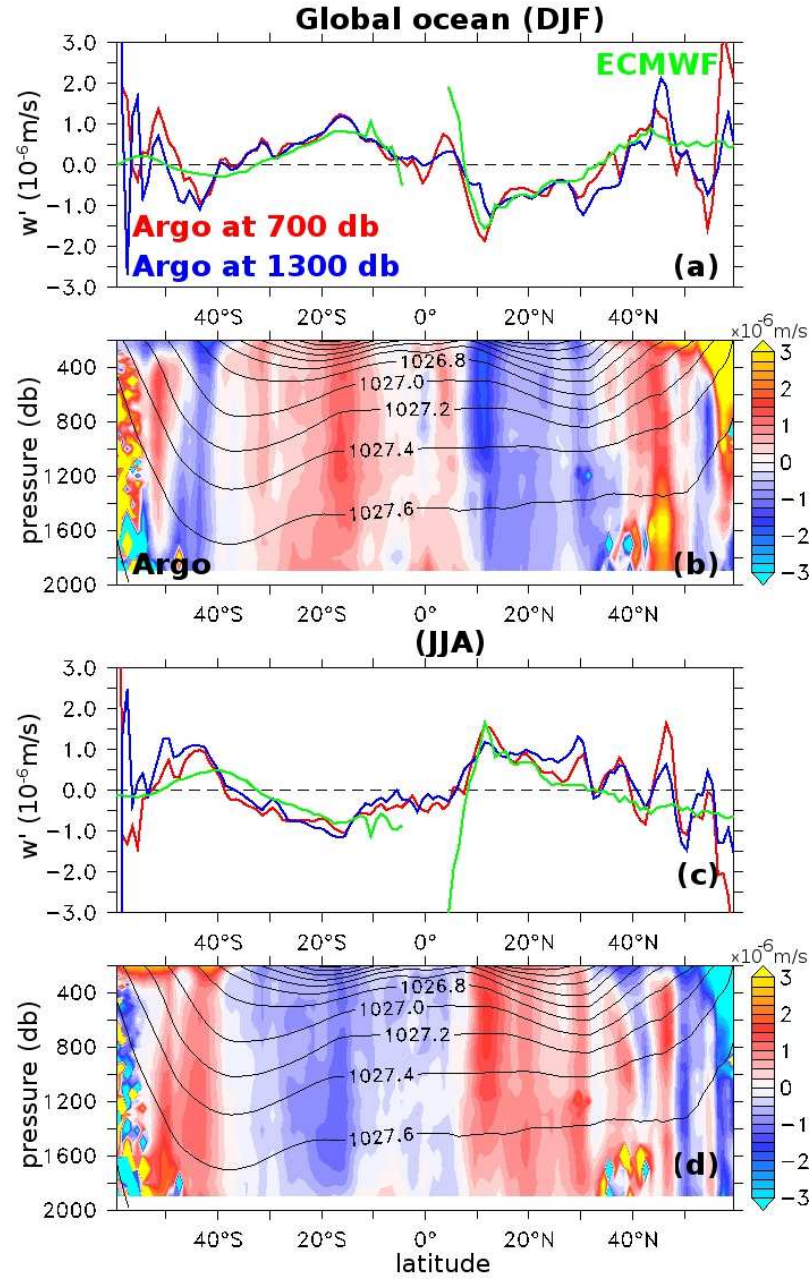


Figure 3.A4: Seasonal vertical velocity from Argo isopycnal displacement (w'_ρ): zonal average in the global ocean during the Northern Hemisphere winter (a,b) and summer (c,d). In (a,c), the green line is based on ERA-Interim wind stress Ekman pumping (w'_{Ek}); the red and the blue lines are w'_ρ at 700 db (red line) and 1300 db (blue line). In (b,d), the color is w'_ρ from Argo isopycnal displacement; the black contours are the temporal mean field of Argo potential density, zonally averaged in the same regions.

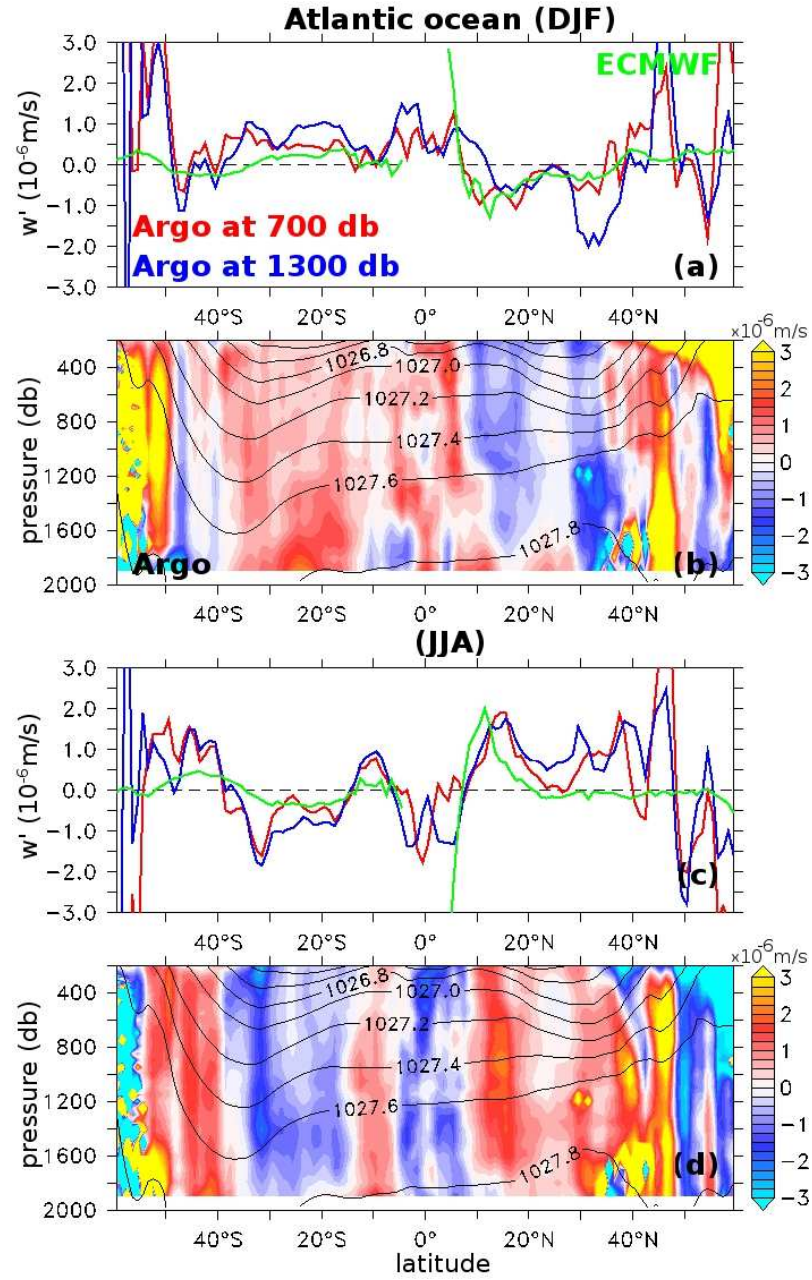


Figure 3.A5: Seasonal vertical velocity from Argo isopycnal displacement (w'_ρ): zonal average in the Atlantic ocean during the Northern Hemisphere winter (a,b) and summer (c,d). In (a,c), the green line is based on ERA-Interim wind stress Ekman pumping (w'_{Ek}); the red and the blue lines are w'_ρ at 700 db (red line) and 1300 db (blue line). In (b,d), the color is w'_ρ from Argo isopycnal displacement; the black contours are the temporal mean field of Argo potential density, zonally averaged in the same regions.

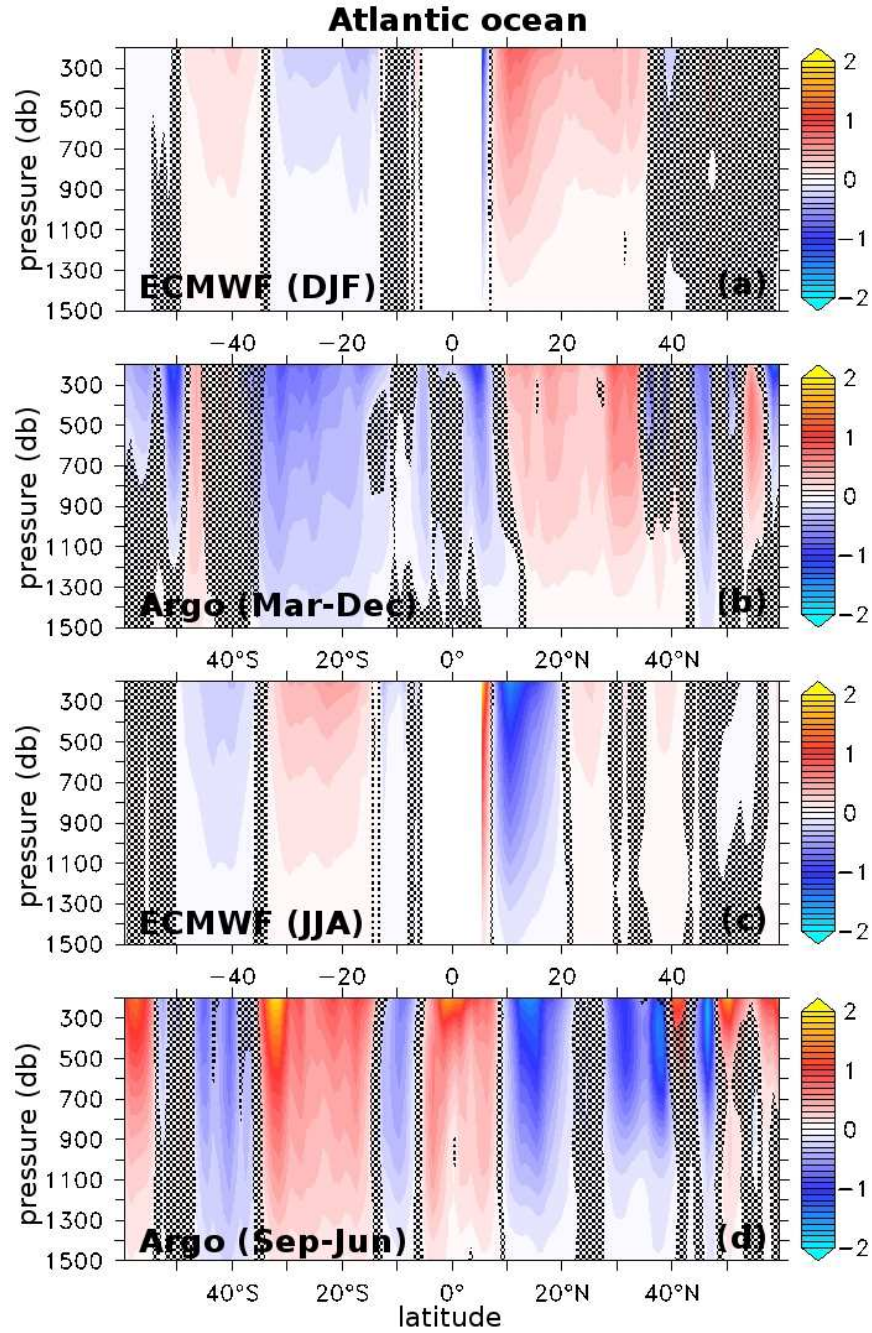


Figure 3.A6: Seasonal steric height change during the Northern Hemisphere winter (panels a,b) and summer (panels c,d): zonal average in the Atlantic ocean. Panels (a) and (c) show the contribution of vertical advection ($\Delta\eta'_w$). Panels (b) and (d) show $\Delta\eta'$ from Argo. Changes smaller than one standard error are shaded black.

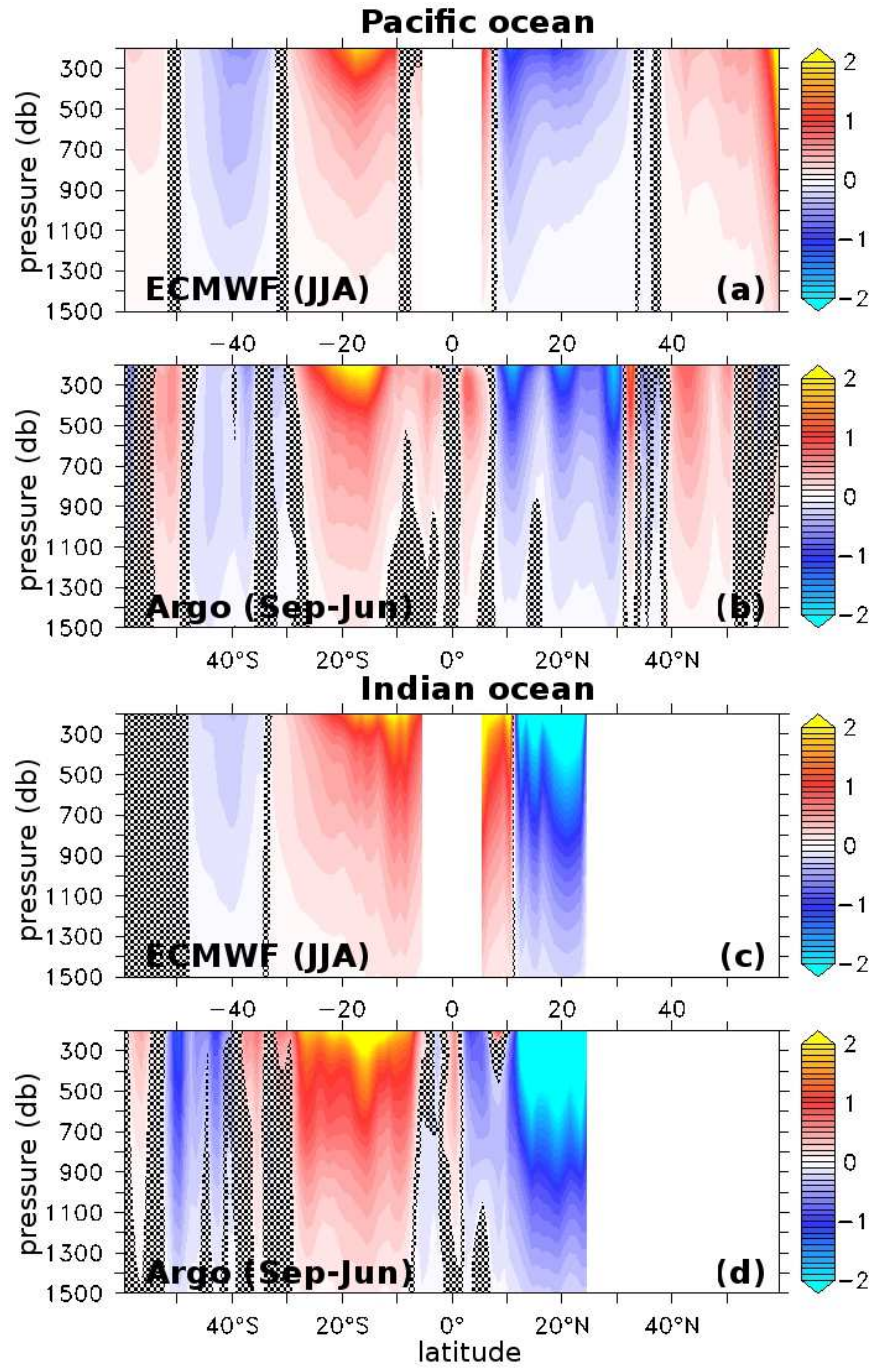


Figure 3.A7: Seasonal steric height change during the Northern Hemisphere summer: zonal average in the Pacific ocean (a,b) and in the Indian ocean (c,d). Panels (a) and (c) show the contribution of vertical advection ($\Delta\eta'_w$). Panels (b) and (d) show $\Delta\eta'$ from Argo. Changes smaller than one standard error are shaded black.

Chapter 4

Climatological monthly heat and freshwater flux estimates on a global scale from Argo

Abstract. The global pattern of climatological monthly heat and freshwater fluxes at the ocean surface is estimated using Argo temperature and salinity profile data for the period 2004 to 2013. Temperature or salinity changes are calculated in a volume of water above an isopycnal that is below the mixed-layer and not subject to mixed-layer entrainment. Horizontal advection components from geostrophic velocity and from Ekman transport, based on wind stress, are also included. The climatological monthly heat or freshwater flux at the ocean surface is estimated as the sum of advective and time tendency contributions. The air-sea flux estimates from Argo are described in global maps and basin-wide integrals, in comparison to atmospheric reanalysis data and to air-sea flux products based on observations. This ocean-based estimate of surface fluxes is consistent with property variations in the subsurface ocean and indicates greater amplitude for the climatological monthly heat flux values in the subtropics compared to other products. Similarly, the combination of Argo freshwater flux and reanalysis evaporation, suggests greater amplitude for climatological monthly precipitation in the

tropics.

4.1 Introduction

Fluxes of heat and freshwater across the ocean-atmosphere interface play a major role in redistributing solar energy around the globe and are a key component of the coupled system. Traditional estimates of these fluxes rely on measurements of basic state variables near and at the ocean surface, to be fed to a bulk formula or to a conversion scheme. Results may not be consistent with property variations in the subsurface ocean due to both deficiencies in the bulk formula and shortcomings in ship-based and satellite observations, especially for evaporation and precipitation [*Schulz, 2003*]. Ship-based measurements are mostly obtained along shipping lines, which are not sufficient in the Tropics and all Southern Hemisphere oceans and introduce a bias since vessels try to avoid bad weather. Also, the data quality varies from weather ships and research quality buoys to voluntary observing ships. Satellite measurements may be limited by cloud coverage of the satellite scene and/or satellite algorithms to analyze the data. The advent of Argo temperature and salinity profiles on a global scale introduces the possibility of estimating heat and freshwater exchange across the ocean-atmosphere interface from an ocean-based perspective, as surface fluxes are the residual of the heat or salt budget in a water column when the ocean dynamics and tendency terms are accounted. The goal of the present analysis is to apply this ocean-based method and produce a global estimate of annual heat and freshwater flux from Argo.

Previous studies have considered heat and salt budgets on regional or basin scales. *Moisan and Niiler [1998]* shows the heat budget for the North Pacific Ocean using a combination of ocean and atmospheric observations, and compares the heat storage rate (HSR) in the ocean to the net heat flux (NHF) at the interface with the atmosphere. The HSR estimates are the integral of temperature profiles down to a chosen isotherm (regarded as the local winter ventilation isotherm) rather than to a constant depth, to remove most of the variance caused by vertical motion of the thermocline (by mesoscale eddies or Rossby waves). In fact, while the seasonal

climatology of HSR by integrating the temperature profiles down to 300 m compares poorly to the seasonal climatology of the NHF estimates, integrating to a constant isotherm results in seasonal HSR climatologies that are very similar to the seasonal climatologies of the NHF. The greatest difference between the seasonal climatologies of the NHF and HSR occurs in the western boundary current region. *Moisan and Niiler* [1998] attributes this increased difference primarily to the large meanders associated with the Kuroshio and Kuroshio Extension, which follow preferential patterns [*Mizuno and White*, 1983] and thus cause error in the estimate of the HSR. Also, the seasonal variability in the horizontal transport of heat is mentioned as a potential source of difference between the seasonal climatologies of the NHF and HSR in this and other regions of the Pacific.

White et al. [2005] also compare the HSR to NHF, over the Pacific Ocean (20°S - 60°N), using World Ocean Circulation Experiment (WOCE) reanalysis products from 1993 to 1999. They quantify the uncertainties in closing the seasonal heat budget including Ekman and geostrophic horizontal advection terms and dissipation from cross-isopycnal mixing at the top of the main pycnocline. Results show that the Ekman contribution dominates horizontal heat advection, and residual dissipation is significant only in the Kuroshio-Oyashio Current Extension.

On similar lines, *Wells and Josey* [2009] presents an upper ocean (0–300 m) heat budget for the North Atlantic (20° - 60°N) using Argo profiling floats (1999 – 2005) and the NCEP/NCAR and NOC surface flux datasets. The budget terms are assessed in 10°×10° boxes and closure is obtained within the error estimates for some regions (particularly the eastern subtropical Atlantic), but not for those boxes that include the Gulf Stream. Also, the analysis of budget closure reveals biases in the NCEP product compared to NOC.

Ren and Riser [2009] perform, instead, the budget for mixed-layer salinity, using Argo for a region in the northeast Pacific ocean. In this study area, the mixed-layer salinity has a strong annual cycle, driven by seasonality in precipitation, evaporation, Ekman advection, and entrainment. The imbalance in closing the budget is attributed to the representation of the entrainment.

Another mixed-layer salinity budget from Argo profiles is shown by *Dong et al.* [2009], who combined Argo with remotely sensed data and re-analysis products in the Southern Ocean (65° - 35° S). The domain-averaged terms of oceanic advection, diffusion, entrainment, and air-sea freshwater flux are largely consistent with the seasonal evolution of mixed-layer salinity, which increases from March to October and decreases from November to February. This seasonal cycle is largely attributed to oceanic advection (mostly Ekman) and entrainment, with air-sea freshwater flux playing only a minimal role. Substantial imbalances in the salinity budget do exist locally, particularly for regions with strong eddy kinetic energy and sparse in situ measurements, and the authors comment that the attempt to close the budget would benefit from improved freshwater flux and surface salinity fields and a better representation of the mixed-layer depth.

Ren et al. [2011] also studied the mixed-layer salinity budget in the Southern Ocean (62° - 45° S) using Argo profiles. In this analysis, the role of sea ice is quantified, and results show a seasonal cycle driven by freshwater flux at the ocean-atmosphere interface, Ekman advection, entrainment, and sea ice. The geostrophic advection and diffusion components are small compared to other terms in the large-scale average, but their contribution can be important locally. Difficulties in closing the budget arise due to errors in surface freshwater flux and entrainment estimates.

A different approach to estimating annual heat exchange at the ocean-atmosphere interface is described by *Fasullo and Trenberth* [2008a]. The net upward surface flux is derived as the residual of the top of the atmosphere (TOA) flux and atmospheric energy budgets, using satellite observations and re-analysis products. Results are compared with direct calculations of ocean heat content and its tendency from ocean temperature datasets (WOA, JMA, and GODAS). The ocean estimates are found to be unrealistically large, with biggest problems over the Southern Ocean, where the scarcity of observations suggests shortcomings in the ocean datasets [*Fasullo and Trenberth*, 2008b].

This chapter presents a global estimate of climatological monthly heat and freshwater flux from Argo. This ocean-based estimate of surface fluxes can improve our understanding of errors in the large-scale patterns inferred from sparse direct

observations or remotely sensed data, first by identifying regional inconsistencies between air-sea flux estimates and the temperature or salinity annual cycles. Errors in heat and freshwater storage are due largely to sampling of temperature and salinity fields. With 10 years of Argo coverage, the annual storage terms are already determined with unprecedented accuracy, and will continue to improve as the Argo timeseries is extended. Corrections to the storage term due to advection are small except in a few locations. This ensures that the systematic errors are small, and that they diminish in regional and global averages (i.e. the net advection is zero for a global average while the net storage is not). Section 4.2 describes the observations and the model output used for the analysis. Section 4.3 describes the method and possible sources of error in the final estimate. Section 4.4 presents the climatological monthly heat (section 4.4.1) and freshwater (section 4.4.2) flux from Argo and compares them to other products. A summary and conclusions are in section 4.5.

4.2 Data

Argo observations.

Argo profiles provide temperature, salinity, and pressure data on a global scale (64.5°S - 64.5°N, except for continental shelves, marginal seas, and ice-covered regions) and as deep as 2000 db [Roemmich *et al.*, 2009]. In this analysis, the raw Argo temperature and salinity measurements during 2004 – 2013 were gridded monthly on a $1^\circ \times 1^\circ$ grid, after quality control and adjustment of pressure bias. The gridding was done by objective mapping, with latitude-dependent decorrelation scales [Roemmich and Gilson, 2009]. In order to test how this procedure may affect the seasonal signal in lightly sampled regions, an alternate gridding method was applied to each month of the 12-month climatology (i.e. the January map combines all the January data during 2004 – 2013). Bin-averaged data in large-scale regions of 555 km radius, yield results that are little different from the objective mapping case, and are therefore not discussed further.

The lack of Argo coverage in coastal, ice-covered and marginal sea regions needs to be considered when extending the present results to the global ocean. In particular, coastal regions are important for the global freshwater flux estimate and marginal seas and ice-covered regions affect both global freshwater and heat flux climatological monthly values from Argo.

ECMWF ERA-Interim Re-Analysis data.

Synoptic monthly means of heat, freshwater and zonal/meridional wind momentum flux from the European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis, ERA-Interim (ERA-Interim), are used in this analysis for the period 2004 – 2013. *Dee et al.* [2011] describe the ERA-Interim data. *Brunke et al.* [2011] found that for wind stress and latent heat fluxes, ERA-Interim performed better than most other products.

OA heat flux and evaporation.

The OA heat flux used here for the period 2003 – 2009 is provided by the WHOI OAFlux project v3 (<http://oafux.whoi.edu>), along with monthly evaporation (for the period 2004 – 2013). The OA product is the synthesis of satellite observations and re-analysis data by objective analysis, to reduce error in each input data source and to produce an estimate that has the minimum error variance [*Yu and Weller*, 2007].

MERRA precipitation.

MERRA, the Modern Era Retrospective-analysis for Research and Application, is an atmospheric observation reanalysis that uses the Goddard Earth Observing System Data Assimilation System Version 5 (GEOS-5) and is performed by NASA’s Global Modeling and Assimilation Office at Goddard Space Flight Center. This project places satellite observations, including those of the NASA EOS suite, in a climate context, for improved estimates of the hydrological cycle

over a broad range of weather and climate time scales. Detailed information about the MERRA project can be found at <http://gmao.gsfc.nasa.gov/research/merra/>. Monthly MERRA precipitation for the period 2004 – 2013 is combined here with OA evaporation to estimate monthly freshwater flux (tagged in the following as MERRA+OA).

SOSE.

The Southern Ocean state estimate (SOSE) for the years 2008–2010 (iteration 60) is used here for heat and freshwater fluxes. The ocean state is estimated with a general circulation model run and convergence to the state estimate is achieved by systematically adjusting the atmospheric driving and initial conditions using an adjoint model [Mazloff *et al.*, 2010]. A cost function compares the model state to in situ observations (Argo float profiles, CTD synoptic sections, seal-mounted SEaOS instrument profiles, and XBTs), altimetric observations [Envisat, Geosat, Jason, Ocean Topography Experiment (TOPEX)/Poseidon)], and other datasets (e.g., sea surface temperatures inferred from infrared and microwave radiometers). In contrast to other data assimilation approaches, no nudging terms are introduced in the dynamical equations, and therefore the state estimate represents a physical solution that satisfies the discretized equations of motion of the model. Mazloff *et al.* [2010] describe in detail the assimilation procedure and the observations used.

4.3 Method.

The Argo estimate of annual heat and freshwater flux at the ocean-atmosphere interface (ξ_T and ξ_S) is based on the heat and freshwater budget in the volume of water above an isopycnal, during the period 2004 – 2013. The budget includes advective (zonal and meridional) and time tendency contributions, and it does not include mixing,

$$\xi_{flux}(x, y, t) = \gamma \left(h \frac{\partial \psi_a}{\partial t} + h \mathbf{V}_a \cdot \nabla \psi_a \right), \quad (4.1)$$

as Eq.1 in *Moisan and Niiler* [1998]. The seasonal horizontal mixing contribution is expected to be small in large-scale averaged fields and poleward of 5° latitude [*Moisan and Niiler*, 1998; *Dong et al.*, 2009; *Ren and Riser*, 2009; *Ren et al.*, 2011; *Piecuch and Ponte*, 2012; *Da-Allada et al.*, 2013]. In Eq.4.1, ξ_{flux} is the estimate of annual heat (or freshwater) flux, ψ is potential temperature (or salinity) from Argo and $\mathbf{V}$ is the horizontal velocity vector (geostrophic component from Argo, referenced to 2000 db, and Ekman component from ECMWF ERA-Interim wind stress). Also, $\gamma = \rho_0 c_p$ for the heat flux estimate and $\gamma = \frac{1}{s_0}$ for the freshwater flux estimate, where ρ_0 (kg/m^3) is surface potential density, $c_p = 4000 \text{ J/K/kg}$ is specific heat and S_0 (PSU) is surface salinity. The subscripts in ψ_a and $\mathbf{V}_a$ indicate that both variables are averaged between the sea surface and the isopycnal depth $h(x, y, t)$. Depth h corresponds to an isopycnal 0.1 kg/m^3 denser than the maximum surface density during the Argo period; hence it is always below the mixed-layer and not affected by mixed-layer entrainment. The time average pressure on such isopycnal is shown in Fig.4.1. Surface flux results from Argo are smoothed in time using a Parzen filter and a 5-month window and the same filtering is applied to the ECMWF, OA, MERRA+OA and SOSE surface fluxes used for comparison in this analysis.

The error on ξ_{flux} (i.e. ξ_{err}) accounts for two contributions: how the flux estimate changes for each month in different years and how it depends on the isopycnal chosen for the average of ψ_a and $\mathbf{V}_a$. The first component (ξ_{err_1}) is not only related to interannual variability, but also to meanders and mesoscale features not resolved by Argo. It is based on the standard deviation of ξ_{flux} for each month during the years of interest (e.g. for the January value, $\xi_{Jan_{err_1}} = \frac{\sigma_{Jan}}{\sqrt{n}}$, where σ_{Jan} is the standard deviation for the January surface flux estimate, and n is the number of years). The second component of the error (ξ_{err_2}) is computed for each month of the climatology from the standard deviation of the flux estimate, letting the isopycnal h vary between one that is $.03 \text{ kg/m}^3$ denser than the maximum surface density and other ones that are $0.33 \times h$, $0.66 \times h$, $1 \times h$ deeper than h in Eq.4.1 (e.g. $\bar{h}_{+.66 \times h}(x, y) = \bar{h}(x, y) + .66 \times \bar{h}(x, y)$), with $n = 4$. Also, when considering large-scale averages or integrals of the surface flux, a third component

is added to the error, i.e. the standard error on a spatial average. In the estimate of this component, only 1 of every 5 points (of the $1^\circ \times 1^\circ$ degree grid) is considered independent.

Fig.4.2 shows components 1 and 2 of the error on the surface flux estimate from Argo during the Northern Hemisphere winter (DJF average, in color), along with the mean Argo dynamic height (dyn-cm, $p_{ref} = 2000$ db) in DJF (black contours).

Overall, ξ_{err_1} (error related to how the DJF average varies in different years) is the largest component for both ξ_T (panel a) and ξ_S (panel b) and it peaks where energetic currents dominate the dynamics. Here, the Argo coverage may not be sufficient to filter out the noise related to mesoscale features and meandering. This problem occurs, for instance, along the ACC, especially in regions of complex bathymetry. In general, where Argo results show big localized differences with other products and/or are not consistent with the large spatial scale of annual atmospheric variability, they are considered unreliable and are not discussed in this manuscript.

As for ξ_{err_2} (error related to the isopycnal choice when computing the surface flux), this component shows larger values at low latitudes with a maximum for the heat flux estimate in the eastern Pacific/Indian where the thermocline shoals (panels c). ξ_{err_2} is much smaller than ξ_{err_1} where the water column is less stratified, since the estimate of ξ_{err_2} is based on how the result changes when performing the budget in a volume with different vertical extent.

4.4 Heat and freshwater flux: climatological monthly values from Argo.

Climatological monthly values for the Argo heat and freshwater flux (ξ_T and ξ_S from Eq.4.1) are described in this section, in comparison to atmospheric re-analysis data and to air-sea flux products based on observations. Positive values of ξ_T and ξ_S indicate flux of heat and freshwater into the ocean. Also, both Argo and

SOSE freshwater fluxes include a possible contribution from ice melting, while the ECMWF ERA-Interim and MERRA+OA products only represent precipitation minus evaporation.

4.4.1 Heat flux.

Fig.4.3 shows the 12-month climatology of heat flux integrated globally over the Argo domain (64.5°S – 64.5°N , with corresponding area of integration of $2.64 \cdot 10^{14} \text{ m}^2$) for Argo, ECMWF ERA-Interim (ERA-Interim) and OA. All the products show the same phasing, with correlation close to 1 for each pair. The OA estimate is consistent with Argo also in amplitude, while ERA-Interim has a significantly larger amplitude from November to January. The analysis of Northern and Southern Hemisphere separately (NH and SH, see panels a and b in Fig.4.A3), reveals that both ERA-Interim and OA monthly values are smaller than Argo in the NH, while the opposite occurs in December and January for ERA-Interim integrated in the SH (OA values are instead smaller than Argo during the SH winter and comparable during summer). The ERA-Interim heat flux globally integrated over the ocean (90°S – 90°N , with corresponding area of integration of $3.63 \cdot 10^{14} \text{ m}^2$) is also shown in Fig.4.3, along with the integral over the ocean regions between 64.5°S – 64.5°N (with area of integration of $3.37 \cdot 10^{14} \text{ m}^2$, that also includes regions where Argo observations are not available). A comparison between the green line and the dashed cyan one in Fig.4.3 provides some insight on how much signal is missing in the limited Argo domain compared to the global ocean (the equatorial band, the marginal seas and sea ice regions are expected to play a main role). The difference reaches almost 1 PW both in June and during the Northern Hemisphere winter. Also, the similarity between the solid and dashed cyan lines indicates how those regions between 64.5°S – 64.5°N not observed by Argo are the main contribution to the missing global signal in the Argo domain. In particular, the influence of missing coastal regions and marginal seas between 64.5°S – 64.5°N , is larger than that of the global zonal band equatorward of 3.5 degree latitude (not shown).

The estimate of global climatological monthly heat flux over the ocean as the residual of the top of the atmosphere (TOA) flux and atmospheric energy

budgets, updated from *Fasullo and Trenberth* [2008a] (FT2008), is indicated in Fig.4.3 as magenta lines (John Fasullo, personal communication). This update to the FT2008 estimate is for the period 2003-2012. In this version, ERA Interim and CERES EBAF are used and changes in EBAF are likely to play a main role in differentiating the current product from the 2008 estimate (John Fasullo, personal communication). The Argo result is noticeably similar to the updated FT2008 estimate (magenta solid line), which uses a different approach starting from satellite observations and re-analysis products. Finally, the difference between the solid and dashed magenta lines is consistent with ERAI (cyan lines), but the amplitude of the climatological monthly values is smaller.

Panel (a) in Fig.4.4 shows the Argo heat flux monthly climatology integrated over the global Argo domain (zonally) in 5 degree latitude bands (PW). The large-scale pattern is similar to the annual heat storage from Argo shown in Fig.4.6 in *Roemmich and Gilson* [2009], consistent with the small role of climatological monthly advection in a global zonal average representation.

The subtropics are the region with both larger monthly values in the Argo estimate and larger differences from ERAI and OA (Fig.4.4b-c, as well as panels d and g in Fig.4.A3). Argo and ERAI are, instead, consistent in the tropics, with OA showing a small bias at some latitudes in this zonal band. Finally, both ERAI and OA show a small bias at higher latitudes (panel b and c, as well as panels e and h in Fig.4.A3).

A regional analysis of the Argo estimate in different ocean basins (Fig.4.A1 shows the area corresponding to each basin) reveals that the integrals in the North Pacific and both in the North and South Atlantic Ocean show greater amplitude of the climatological heat flux monthly values compared to ERAI and OA, indicating a similar bias for both the ERAI and OA products (Fig.4.5, panels a, c and d). The estimates are, instead, consistent across products for the integral over the South Pacific and South Indian Ocean (panel b and e). In the latitude band between 64.5°S – 30.5°S , all the products considered for comparison (ERAI, OA, SOSE) show a similar bias during the Southern Hemisphere winter, with smaller amplitude than Argo climatological monthly values (panel f). Although some

differences in amplitude are seen, all the products show the same phasing, with correlation close to 1 for each pair as in Fig.4.3.

Finally, Fig.4.6 shows the spatial distribution of the Argo heat flux estimate. Contours (in black) are for the Northern Hemisphere winter heat flux (DJF average) from Argo (a), ERAI (b), OA (c) and SOSE (d). Color shading represents, in panel (a), Argo heat flux during the season (same as the black contours) and in panel (b-d), the difference between Argo and each of the other products. Hatched regions indicate where the value is smaller than the error on the Argo estimate ($\xi_{T_{err}}$). ξ_T is smoothed in space (using a Parzen filter of length 7 in latitude and 15 in longitude) to reduce the noise from mesoscale features not resolved in Argo and to focus on the large-scale pattern of atmospheric annual variability. The map of Argo heat flux in Fig.4.6a still shows smaller scale features south of 30°S, especially in the Indian and west Pacific basins, with corresponding localized differences between Argo and the other products in those regions. Some of these smaller scale features are due to unresolved meandering and mesoscale variability along the ACC and are not discussed further. Fig.4.6b-c shows the large spatial scale of the amplitude difference between Argo DJF values in the subtropics and both ERAI and OA. As already seen in Fig.4.4b-c (as well as in panels d and g in Fig.4.A3) both ERAI and OA have a smaller amplitude than the Argo estimate in this zonal band. The Argo signal is, on the other hand, weaker in the western tropical South Pacific, the tropical South Indian ocean and south of 50°S. The large-scale spatial distribution of Argo heat flux (black contours in panel a) is particularly similar to ERAI (rather than OA) in the South Atlantic ocean (north of 20°S) and in the eastern South Pacific tropics. Also, the DJF signal in Argo is almost everywhere stronger than in SOSE (panel d).

Argo heat flux monthly climatology during the Northern Hemisphere summer has a similar spatial pattern compared to DJF but opposite in sign, and the way it relates to the other products is also similar in the two seasons (see JJA in Fig.4.A4).

4.4.2 Freshwater flux.

Fig.4.7 shows the 12-month climatology of freshwater flux integrated globally over the Argo domain (64.5°S – 64.5°N , with corresponding area of integration as for the heat flux) for Argo, ERAI and MERRA+OA. MERRA+OA is mostly consistent with Argo both in phase (correlation coefficient $r = 0.8$) and amplitude (except for the January value), while ERAI has smaller amplitude and shows very low correlation with Argo ($r = 0.22$). The analysis of Northern and Southern Hemisphere separately (NH and SH, see panels a and b in Fig.4.A4), reveals that MERRA+OA monthly values have greater amplitude than Argo in both hemispheres, while ERAI is mostly consistent with Argo in the NH and shows stronger signal in the SH. The ERAI freshwater flux globally integrated over the ocean (90°S – 90°N) is also shown in Fig.4.7, along with the integral over the ocean regions between 64.5°S – 64.5°N (with area of integration that also includes regions where Argo observations are not available). A comparison between the green line and the dashed cyan line in Fig.4.7 provides some insight on how much signal is missing in the limited Argo domain compared to the global ocean. For the freshwater climatological monthly values, the equatorial band, marginal seas and sea ice regions are expected to play a main role along with the coastal band (due to river runoff). Also, the similarity between the solid and dashed cyan lines indicates how those regions between 64.5°S – 64.5°N not observed by Argo are the main contribution to the missing global signal in the Argo domain, as for the heat flux. For the freshwater flux case, the global zonal band equatorward of 3.5 degree latitude plays a role comparable to that of the missing coastal regions and marginal seas between 64.5°S – 64.5°N (not shown).

Panel (a) in Fig.4.8 shows the Argo freshwater flux monthly climatology integrated over the global Argo domain (zonally) in 5 degree latitude bands (Sv). The tropics are the region with larger monthly values in the Argo estimate and, in the Southern Hemisphere, they also show larger differences from ERAI and MERRA+OA (Fig.4.8b-c, as well as panel f in Fig.4.A6). The apparent bias in ERAI and MERRA+OA freshwater flux monthly climatology in the tropics, is mostly related to precipitation, consistent with the difficulty in observing a phe-

nomena that is highly variable in space and time. This can be seen in Fig.4.9, where the estimate of climatological monthly precipitation from combining Argo freshwater flux and ERAI or OA evaporation does not vary much with the evaporation product used (i.e. the two dashed lines are consistent one with the other). Both ERAI and MERRA precipitation have much smaller amplitude than Argo (green and blue lines), with only MERRA's phasing consistent with Argo. In Fig.4.9 the Argo freshwater flux is computed from the tendency term alone, hence the equatorial band is included; also, the Argo domain is extended into coastal areas and marginal seas using values from the closest neighboring regions. This assures that the estimate is only missing the contribution of horizontal advection at the Northern and Southern boundaries of the zonal band of interest.

The global integral of freshwater flux in the Northern Hemisphere subtropics is consistent between the Argo estimate and ERAI, while the MERRA+OA product has greater amplitude (Fig.4.8b-c, as well as panel d in Fig.4.A6). In contrast, the apparent bias is large for both ERAI and MERRA+OA in the Southern Hemisphere subtropics, with low correlation between Argo and the other products (Fig.4.8b-c, as well as panel g in Fig.4.A6).

A regional analysis of the Argo freshwater estimate in different ocean basins (Fig.4.A1) reveals that the global integrals in the South Pacific and in the North Atlantic Oceans are consistent between the Argo estimate and ERAI, while the MERRA+OA product is significantly different (Fig.4.10, panels b and c). The MERRA+OA estimate is consistent with Argo in the South Indian Ocean, while ERAI shows a larger amplitude (Fig.4.10e). In the South Atlantic Ocean and in the latitude band 64.5° - 30.5° S, different products provide very different estimates of the freshwater monthly climatology (Fig.4.10, panels d and f). In the Southern Ocean region, the Argo freshening during the Southern Hemisphere summer is in agreement with input of freshwater into the ocean towards the end of the year [Dong *et al.*, 2009; Ren *et al.*, 2011]. Dong *et al.* [2009] also suggests a role of the ice melting near Antarctica, with subsequent northward Ekman transport. As mentioned at the beginning of this section, ice melting is not included in ERAI and MERRA+OA, but it is in SOSE freshwater flux, hence the large difference between

Argo and SOSE in Fig.4.10f is probably not related to ice but mostly derives from differences in the upper ocean salinity field of the two products. Fig.4.11 shows that the 12-month climatology of salinity in the upper 100 m of the ocean has opposite phase between Argo and SOSE (black lines in the left panel). This difference in phase results mostly from a difference in amplitude of the seasonal signal between the two products, with Argo generally fresher than SOSE in the first half of the year and saltier in the second half (right panel in Fig.4.11). In the latitude band $40^{\circ} - 30^{\circ}\text{S}$, the salinity anomaly is positive in the first half of the year both in Argo and SOSE, but with amplitude larger in SOSE than Argo. The opposite occurs in the band $60^{\circ} - 45^{\circ}\text{S}$ both in phase and amplitude. Only the transition region between the subtropics and higher latitudes (i.e. the latitude band $45^{\circ} - 40^{\circ}\text{S}$) has a monthly climatology of upper ocean salinity that is out of phase between Argo and SOSE and this could be related to representation error of the two products where sharp gradients of salinity are present (e.g. the Agulhas Current). Still, a higher resolution map of Argo annual salinity ($.5^{\circ} \times .5^{\circ}$ grid) shows agreement with the Argo results here. Finally, in the band $42.5^{\circ} - 64.5^{\circ}\text{N}$, the MERRA+OA estimate of climatological monthly freshwater flux is consistent with Argo, while the amplitude is smaller in ERAI (panel e in Fig.4.A6).

Fig.4.12 shows the spatial distribution of annual freshwater flux in the Argo estimate. Contours (in black) are for the Northern Hemisphere winter freshwater flux (DJF average) from Argo (a), ERAI (b), MERRA+OA (c) and SOSE (d). Color shading represents, in panel (a), Argo freshwater flux during the season (same as the black contours), in panel (b-d) the difference between Argo and each of the other products. Hatched regions indicate where the amplitude is smaller than the error on the Argo estimate (ξ_{err}). As for ξ_T , even with spatial smoothing, the map of Argo freshwater flux (Fig.4.12a) shows smaller scale features south of 30°S . As with ξ_T , some of these features are due to unresolved meandering and mesoscale variability along the ACC and are not discussed further.

Fig.4.12c shows the large spatial scale of the amplitude difference between Argo DJF values and MERRA+OA in the Southern Hemisphere tropics, in the subtropical North Pacific, and in the North Atlantic in general. In the Southern

Hemisphere tropics, ERAI shows an overall bias similar to MERRA+OA, except for the Indian and South-West Pacific ocean. The large-scale spatial distribution of Argo freshwater flux (black contours in panel a) is particularly similar to MERRA+OA (rather than ERAI) in some regions of the Pacific Ocean (Northern Hemisphere tropics and south of 30°S). Also, the DJF signal in Argo is almost everywhere fresher than in SOSE (panel d).

Argo freshwater flux anomaly during the Northern Hemisphere summer has a similar spatial pattern compared to DJF (but opposite in sign) and the way it relates to the other products is also similar in the two seasons (see JJA in Fig.4.A7).

4.5 Summary and conclusions.

Heat and freshwater exchanges between the atmosphere and the ocean directly impact sea level and ocean circulation, and play a key role in redistributing solar energy in the climate system. In order to describe and analyze the coupled ocean-atmosphere system, consistency is essential between the estimation of air-sea fluxes of heat and freshwater and that of heat and freshwater storage in the oceans based on temperature and salinity. The indirect nature of most flux estimates and the limited coverage of the supporting data leaves large-scale biases in available datasets [*Schulz*, 2003; *Bosilovich et al.*, 2008].

Estimates of climatological monthly heat and freshwater surface fluxes based on Argo temperature and salinity profiles are consistent with subsurface water properties since they rely primarily on storage terms. For regional to global scale estimation, advective corrections to the heat and freshwater storage terms are small and diminish in relative importance with increasing area (Fig.4.A2). Also, for these large-scale averages, seasonal changes in temperature and salinity are directly interpretable as storage, with little systematic error. The Argo Program has provided over 1.2 million temperature and salinity profiles in the global ocean over the last decade, with little bias in spatial or seasonal distributions. The Argo dataset is of unprecedented accuracy already for estimation of heat and freshwater storage, with an added 120,000 profiles per year continuing to decrease

the estimation errors.

The global integral of climatological monthly heat flux from Argo is limited to the Argo domain, and is missing the contributions of marginal seas, coasts and high latitude oceans. Nevertheless it is consistent with other products and representative, within 1 PW, of the global value. It is notable that even with these spatial limitations, our global estimate agrees well with the update to the *Fasullo and Trenberth* [2008a] estimate, that is based on the top of the atmosphere flux and atmospheric energy budgets from satellite observations and re-analysis products (Fig.4.3).

The global zonal integral of Argo climatological monthly heat flux shows greater amplitude than other products in the subtropics (especially in WBC regions) and in the SOSE region (64.5 S - 30.5 S, during the Southern Hemisphere winter), indicating a similar bias for both ERAI and OA compared to Argo (Fig.4.4 and Fig.4.A3). Similar differences are seen in the North Pacific and in both the North and South Atlantic ocean (Fig.4.5).

The freshwater monthly climatology from Argo can be very informative due to the disagreement (sometimes even in sign) among other products. The global integral in the Argo domain between 64.5 S - 64.5 N is consistent between the Argo estimate and the MERRA+OA but differs significantly from the ERAI estimate of precipitation minus evaporation (Fig.4.7). Also, the combination of Argo freshwater flux with ERAI or OA evaporation shows greater amplitude for climatological monthly precipitation in the tropics than either ERAI or MERRA precipitation (although MERRA's phasing is consistent with Argo). In this region, ERAI and OA evaporation are similar to one another (Fig.4.9).

The integrals of freshwater monthly climatology in the NH subtropics and in the North Atlantic Ocean are consistent between the Argo estimate and ERAI, with an apparent bias in the MERRA+OA product (Fig.4.8 and Fig.4.10). In the South Indian Ocean, instead, the integral from MERRA+OA is consistent with Argo, while the ERAI product shows greater amplitude.

Large differences among products are in the SOSE region (64.5 – 30.5°S), where Argo and SOSE have nearly the same amplitude of the seasonal signal but

are opposite in phase (Fig.4.10f). Both Argo and SOSE include the ice-melting contribution to freshwater and the disagreement between them is consistent with differences in the upper ocean salinity fields between the two products (Fig.4.11).

In this study, the SOSE output spans only 3 years and this could affect the comparison when describing differences between Argo and SOSE. Still, it is interesting to include SOSE in the present analysis since SOSE surface fluxes are adjusted for some level of consistency with observations of the subsurface ocean. A more detailed analysis and comparison to Argo, of the different terms in the SOSE heat and freshwater upper ocean budget, is needed but is beyond the scope of the present work.

For the monthly climatology described in this work, the random coverage errors due to inadequate sampling of eddies, fronts, and other small-scale features is the largest source of error. These coverage errors decrease in regional and global averages and they will continue to decrease over time with a long-term sustained Argo Program. Eventually it may be possible to show how seasonal variability evolves in a changing climate, along with the corresponding impacts on sea level and ocean circulation.

Finally, the Argo estimate of surface fluxes would benefit from filling the gaps in the Argo domain, and indeed the Argo Program is extending sampling into the marginal seas and ice-covered oceans, and is enhancing spatial resolution in western boundary regions. The results from ERAI show that missing regions make up most of the difference between the heat flux integral in the Argo domain and in the global ocean (Fig.4.3). For freshwater, the coastal contribution due to river runoff is crucial too, and the land-ocean exchanges have large impact on salinity there. To make up for the limitations on Argo spatial coverage, supplementary datasets are useful, including satellite measurements of sea surface height, sea surface temperature, and sea surface salinity. Estimates based on correlation of heat and freshwater content in poorly sampled regions with values in the adjacent open ocean observed by Argo floats can extend the estimation to the global domain, as illustrated in Fig.4.9 for climatological monthly tropical precipitation from Argo. Extending the spatial extent of the Argo domain would also improve comparisons

of ocean based freshwater flux with the seasonal signal in ocean mass from GRACE.

The present analysis has focused on the Argo-sampled domain, but extensions of both the sampled regions by Argo and the use of supplementary datasets and correlation analysis will extend these estimates to the global ocean.

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Chapter 4, in full, is a reproduction of the material as it appears in *D. Giglio and D. Roemmich* [2014], *J. Geoph. Res.*, doi: 10.1002/2014JC010083. The dissertation author was the primary investigator and author of this work.

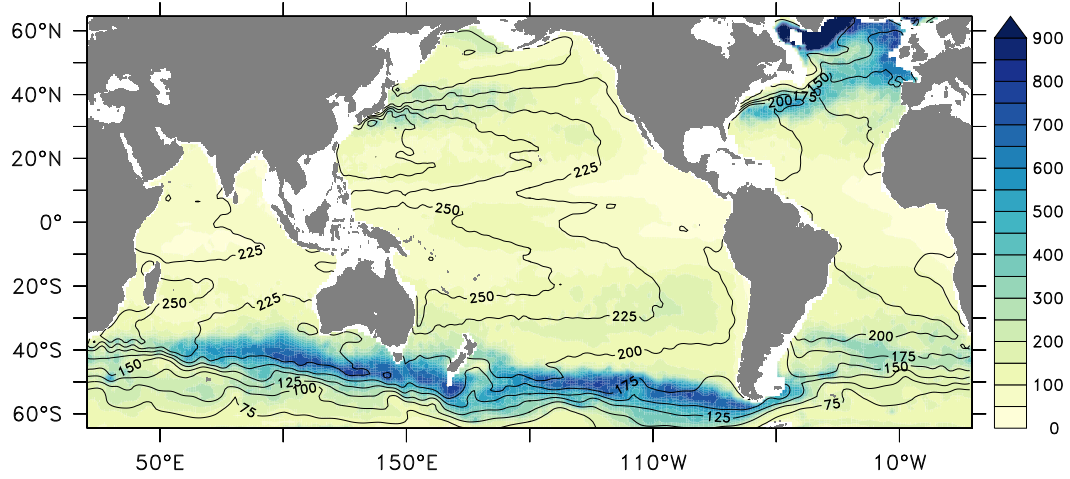


Figure 4.1: Time average pressure (db, from Argo), during 2004 – 2013, on the isopycnal used in Eq.4.1 (in color), i.e. an isopycnal 0.1 kg/m^3 denser than the maximum surface density during the same period. Black contours indicate the mean Argo dynamic height (dyn-cm, $p_{ref} = 2000 \text{ db}$).

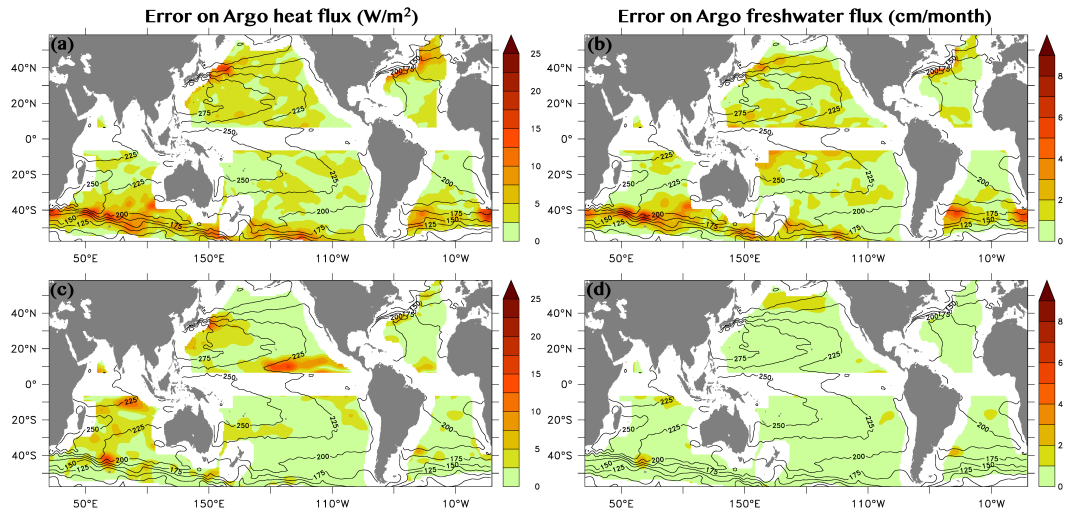


Figure 4.2: Maps of the two components of the error for ξ_T (panels a and c) and ξ_S (panels b and d) in DJF. The error component in panels a and b is related to how the DJF average varies in different years (error term 1, ξ_{err1}). The error component in panels c and d is related to the isopycnal choice when computing the surface flux (error term 2, ξ_{err2}). Black contours indicate the mean Argo dynamic height (dyn-cm, $p_{ref} = 2000 \text{ db}$).

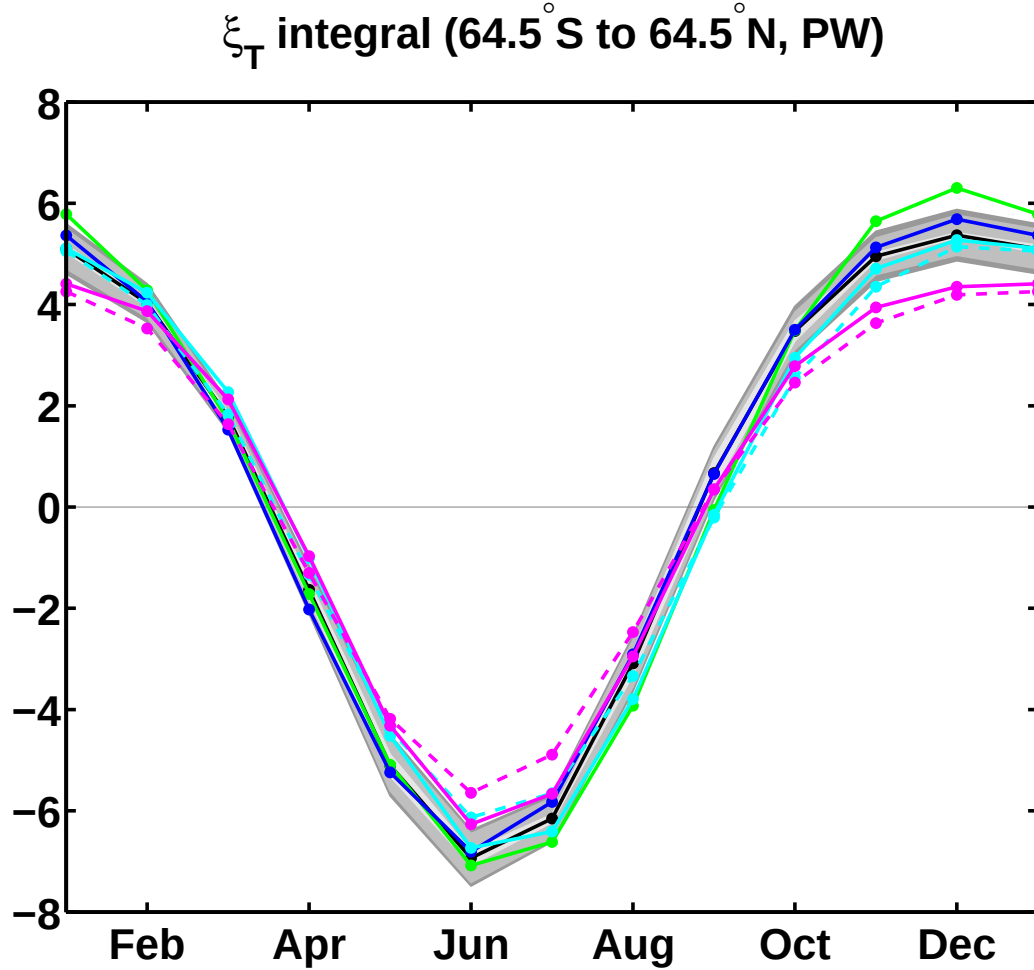


Figure 4.3: Heat flux monthly climatology: integral over the ocean (PW) for Argo (black line, integral over the Argo domain, i.e. no coastal regions nor marginal seas within 64.5°S–64.5°N and except equatorward of 2.5°), ERAI (green line for the same region as the Argo black line, cyan dashed line for the global ocean, cyan solid line for the global ocean within 64.5°S–64.5°N), OA (blue line, for the same region as the Argo black line) and for the updated estimate from *Fasullo and Trenberth* [2008a] (magenta lines: solid, for the global ocean within 64.5°S–64.5°N; dashed, for the global ocean). The errorbars on the Argo estimate are shaded gray: dark gray is the total error ($\pm\xi_{T_{err}}$), light gray is $\pm(\xi_{T_{err2}} + \xi_{T_{err3}})$ (i.e. the error component related to the isopycnal choice in computing Eq.4.1 and the error on the spatial integral) and white is $\pm\xi_{T_{err2}}$.

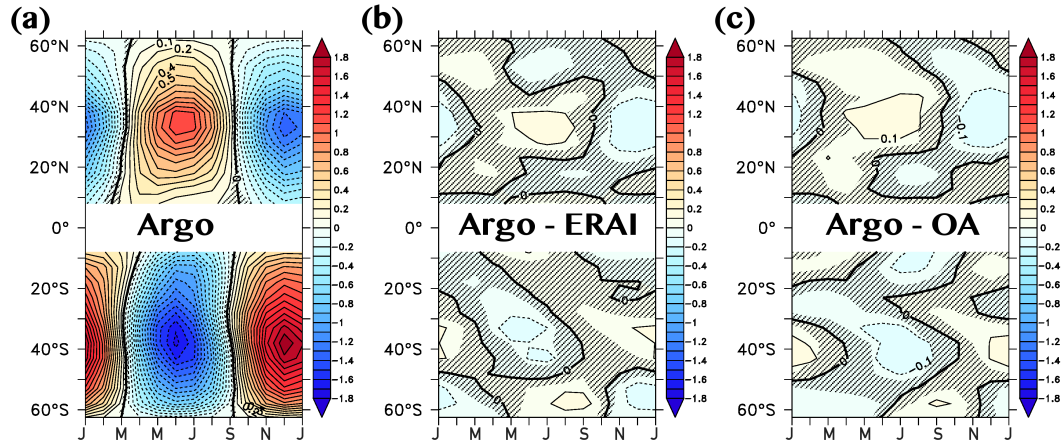


Figure 4.4: (a) Argo heat flux monthly climatology: integral over the global Argo domain (zonally) in 5 degree latitude bands (PW). (b-c) difference between Argo and ERAI/OA. In (a-c), values smaller than $\pm \xi_{T_{err}}$ are hatched black.

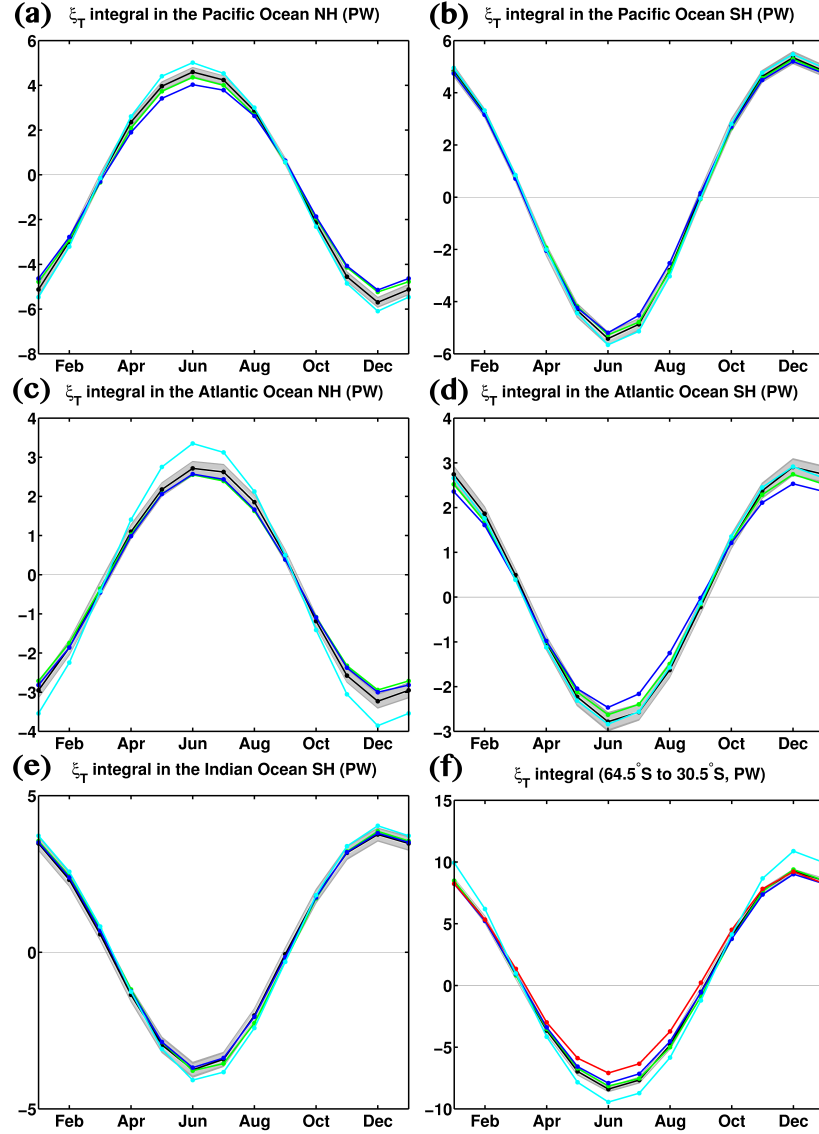


Figure 4.5: Heat flux monthly climatology: integral (PW) in different ocean basins (a-b, North/South Pacific Ocean; c-d, North/South Atlantic; e, South Indian) and in a region of the SOSE domain (64.5°S – 30.5°S , panel f). Black line: Argo, i.e. no coastal regions nor marginal seas, with corresponding areas of integration of $6.41 \cdot 10^{13}$, $6.67 \cdot 10^{13} \text{ m}^2$ for the NH, SH Pacific Ocean, $2.90 \cdot 10^{13}$, $2.98 \cdot 10^{13} \text{ m}^2$ for the NH, SH Atlantic Ocean, $4.36 \cdot 10^{13} \text{ m}^2$ for the SH Indian Ocean, $8.18 \cdot 10^{13} \text{ m}^2$ for the region 64.5°S – 30.5°S . Green/blue/red line: ERAI/OA/SOSE for the same region as the Argo black line. Cyan line: ERAI for the whole ocean region in the basin of interest, with corresponding areas of integration of $7.84 \cdot 10^{13}$, $7.62 \cdot 10^{13} \text{ m}^2$ for the NH, SH Pacific Ocean, $4.03 \cdot 10^{13}$, $3.37 \cdot 10^{13} \text{ m}^2$ for the NH, SH Atlantic Ocean, $4.80 \cdot 10^{13} \text{ m}^2$ for the SH Indian Ocean, $9.57 \cdot 10^{13} \text{ m}^2$ for the region 64.5°S – 30.5°S . The errorbars on the Argo estimate are shaded as in Fig.4.3.

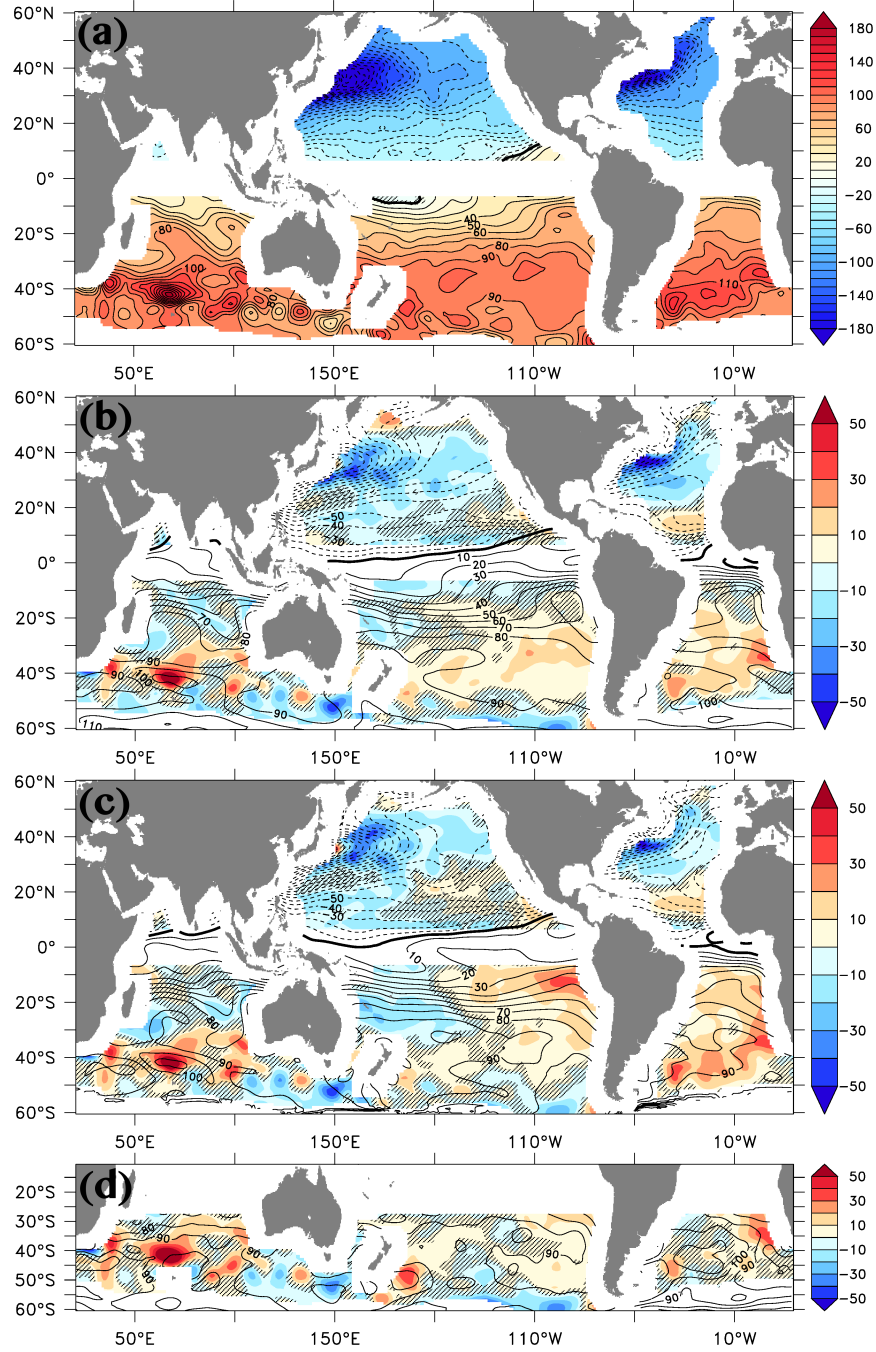


Figure 4.6: Heat flux (Wm^{-2}) during the Northern Hemisphere winter (DJF average, black contours) from (a) Argo, (b) ERAI, (c) OA, (d) SOSE. The contour interval is 10 Wm^{-2} , negative values are dashed and the thick line is zero. In (a), the color follows the black contours. In (b-d), the color is the difference between Argo and each of the other products. Hatched regions indicate values that are smaller than $\xi_{T_{err}}$. Positive values are shaded red (i.e. red indicates that the ocean becomes warmer).

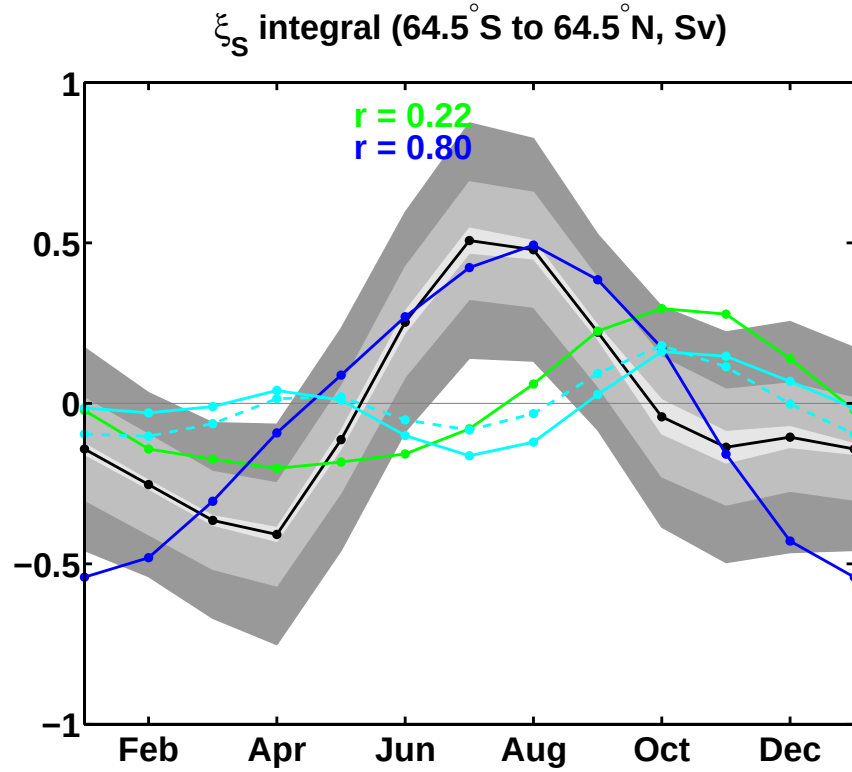


Figure 4.7: Freshwater flux monthly climatology: integral over the ocean (Sv) for Argo (black line, integral over the Argo domain, i.e. no coastal regions nor marginal seas within 64.5°S–64.5°N and except equatorward of 2.5°), ERAI (green line for the same region as the Argo black line, cyan dashed line for the global ocean, cyan solid line for the global ocean within 64.5°S–64.5°N) and for MERRA+OA (blue line, for the same region as the Argo black line). The correlation coefficient (r) between the black and green/blue line is indicated in the legend (at least 95% confidence level). The errorbars on the Argo estimate are shaded as in Fig.4.3.

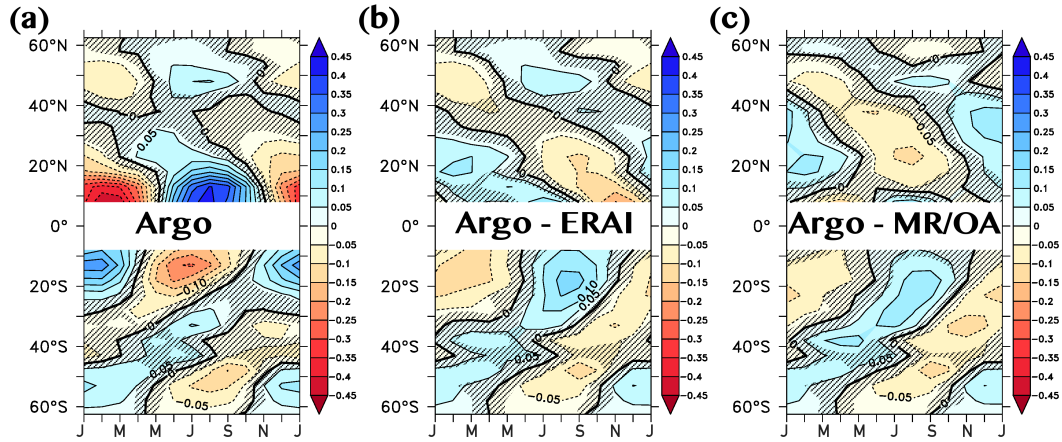


Figure 4.8: (a) Argo freshwater flux monthly climatology: integral over the global Argo domain (zonally) in 5 degree latitude bands (Sv). (b-c) difference between Argo and ERAI/MERRA+OA. In (a-c), values smaller than $\pm\xi_{err}$ are hatched black.

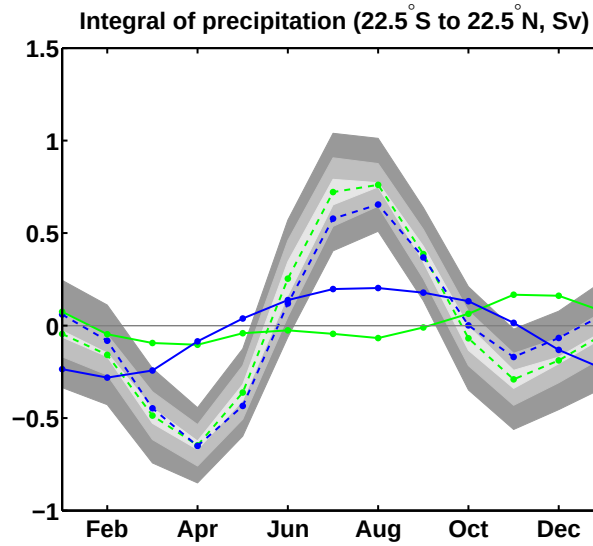


Figure 4.9: Monthly climatology of precipitation: integral over the tropical ocean (22.5°S – 22.5°N , Sv), with corresponding area of integration of $1.51 \cdot 10^{14} \text{ m}^2$. Green/blue solid line: ERAI/MERRA. Green/blue dashed line: estimate from combining Argo freshwater flux (only tendency term used, hence the equatorial band is included) and ERAI/OA evaporation. The Argo domain is extended into coastal regions and marginal seas using values from the closest neighboring regions. The errorbars on precipitation from Argo freshwater flux are shaded for the green dashed line as in Fig.4.3.

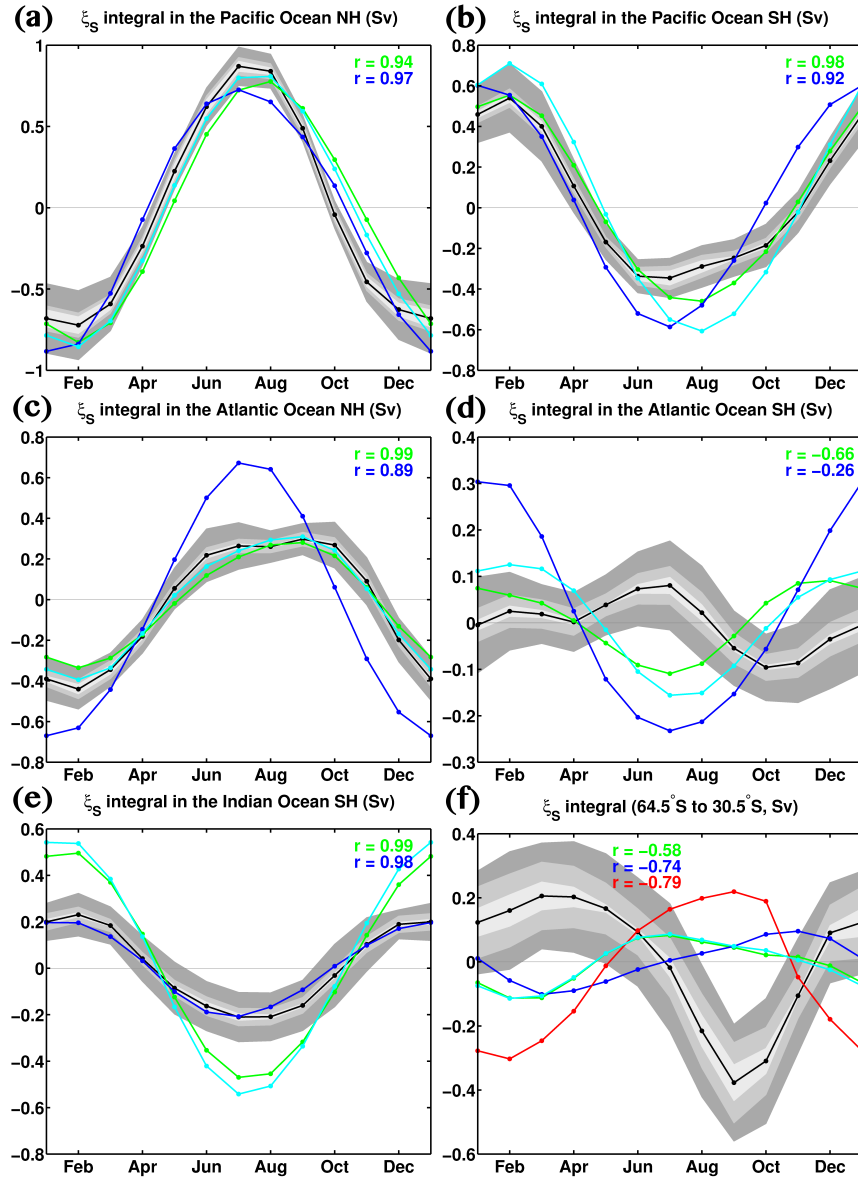


Figure 4.10: Freshwater monthly climatology: integral (Sv) in different ocean basins (a-b, North/South Pacific Ocean; c-d, North/South Atlantic; e, South Indian) and in a region of the SOSE domain (64.5°S–30.5°S, panel d). Black line: Argo, i.e. no coastal regions nor marginal seas. Green/blue/red line: ERAI/MERRA+OA/SOSE for the same region as the Argo black line. Cyan line: ERAI for the whole ocean region in the basin of interest. The area of integration for the different basins is the same as in Fig.4.5. The correlation coefficient (r) between the black and green/blue line is indicated in the legend (at least 95% confidence level). The errorbars on the Argo estimate are shaded as in Fig.4.3.

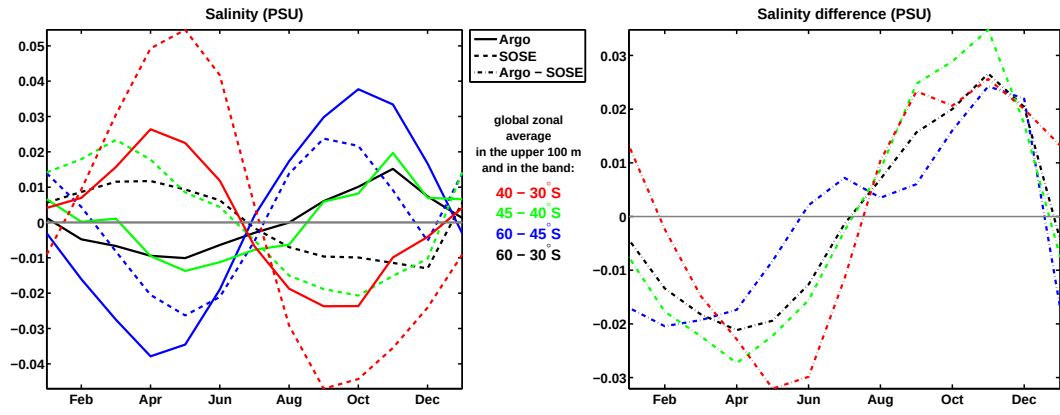


Figure 4.11: Salinity monthly climatology (PSU) in Argo and SOSE (left panel, solid and dash-dotted lines respectively), and difference between the two (right panel): global zonal average in the upper ocean (≤ 100 m) and in the latitude bands $40^\circ - 30^\circ\text{S}$ (red), $45^\circ - 40^\circ\text{S}$ (green), $60^\circ - 45^\circ\text{S}$ (blue), $60^\circ - 30^\circ\text{S}$ (black).

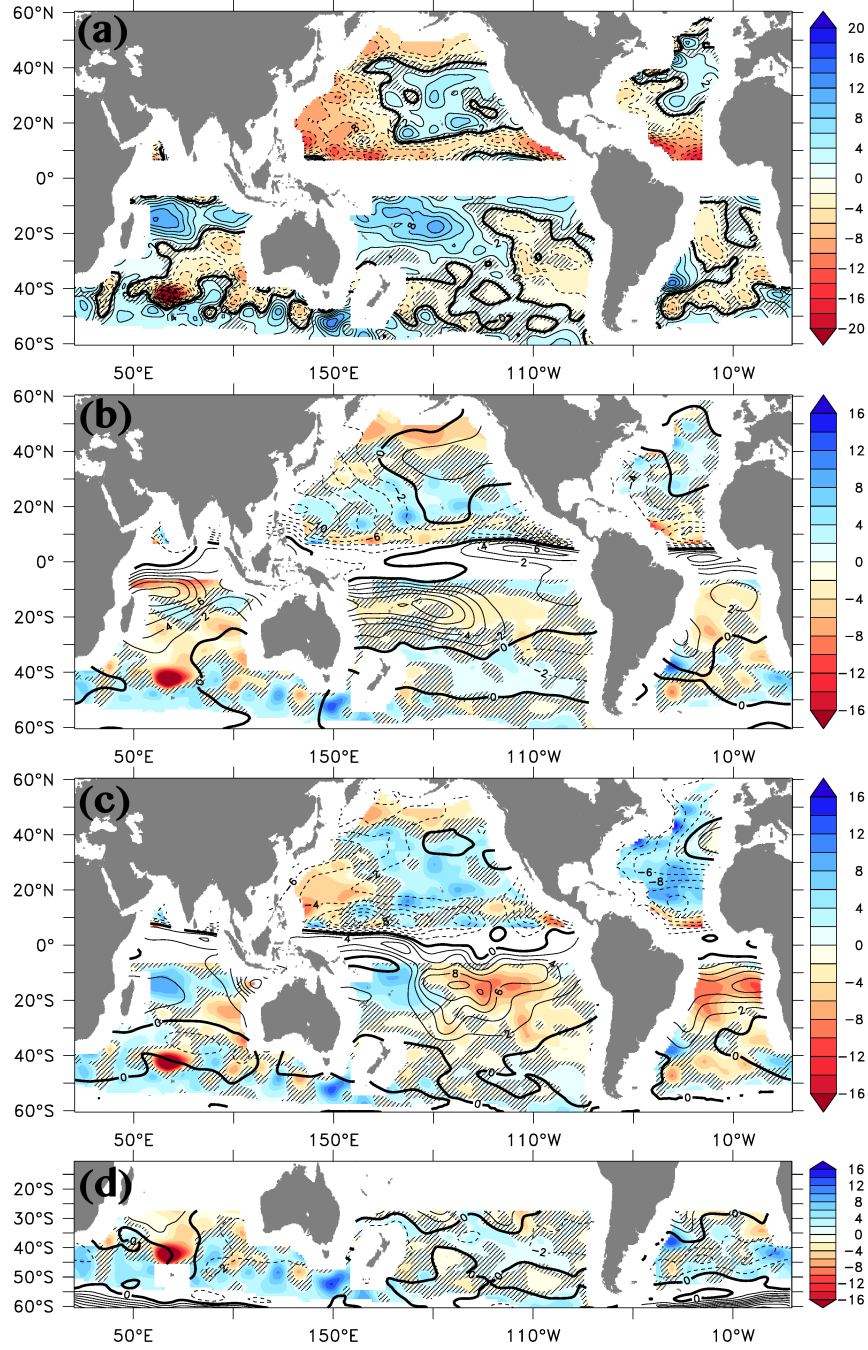


Figure 4.12: Freshwater flux (cm/month) during the Northern Hemisphere winter (DJF average, black contours) from (a) Argo, (b) ERAI, (c) MERRA+OA, (d) SOSE. The contour interval is 2 cm/month, negative values are dashed and the thick line is zero. In (a), the color follows the black contours. In (b-d), the color is the difference between Argo and each of the other products. Hatched regions indicate values that are smaller than $\xi_{S_{err}}$. Positive values are shaded blue (i.e. blue indicates that the ocean becomes fresher).

4.A Appendix

Fig.4.A2 - Fig.4.A7 are an appendix to this chapter. Further details about the figures can be found in the main text of the chapter.

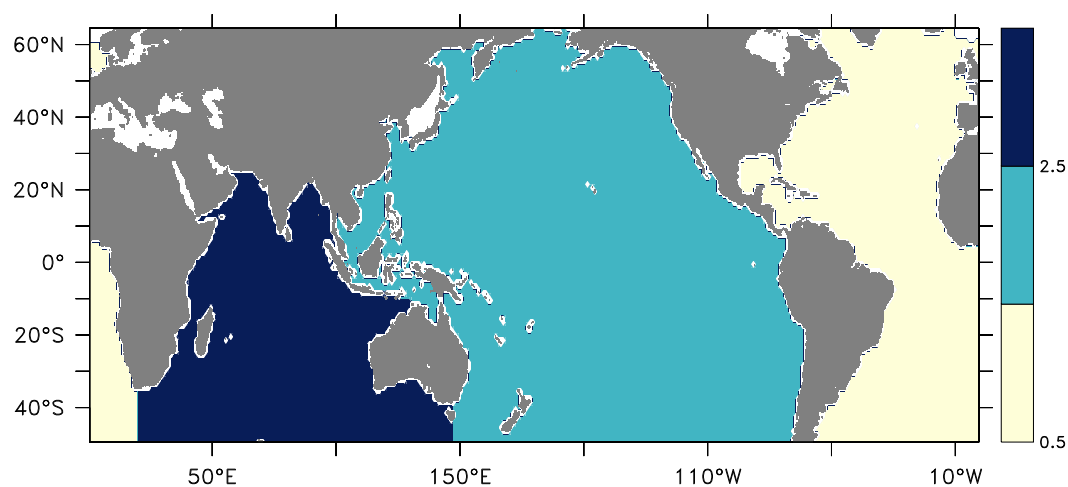


Figure 4.A1: NOAA mask of ocean basins (<http://iridl.ldeo.columbia.edu/SOURCES/.NOAA/.NODC/.WOA09/.Masks/.basin/>). Numbers from 1 to 3 indicate the Atlantic, Pacific and Indian Ocean.

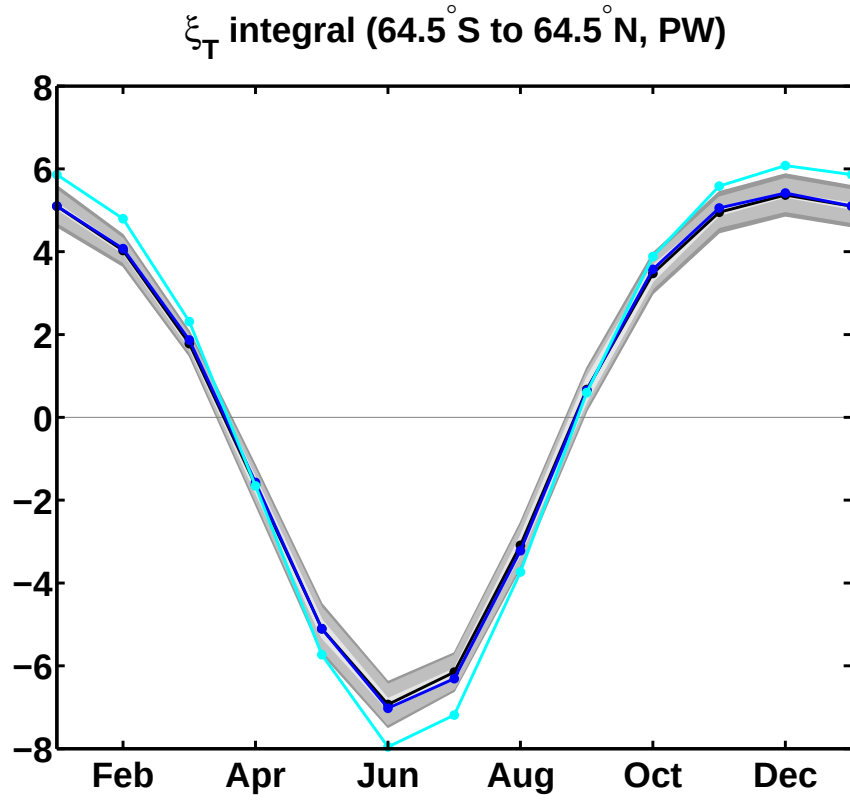


Figure 4.A2: Argo heat flux monthly climatology: integral over the ocean (PW) for the estimate from advection and tendency term (black line), and for the tendency term alone (blue and cyan lines). The black and the blue line are integrals over the Argo domain, i.e. no coastal regions nor marginal seas within 64.5°S–64.5°N, and except equatorward of 2.5° degree latitude. The cyan line includes also the equatorial band.

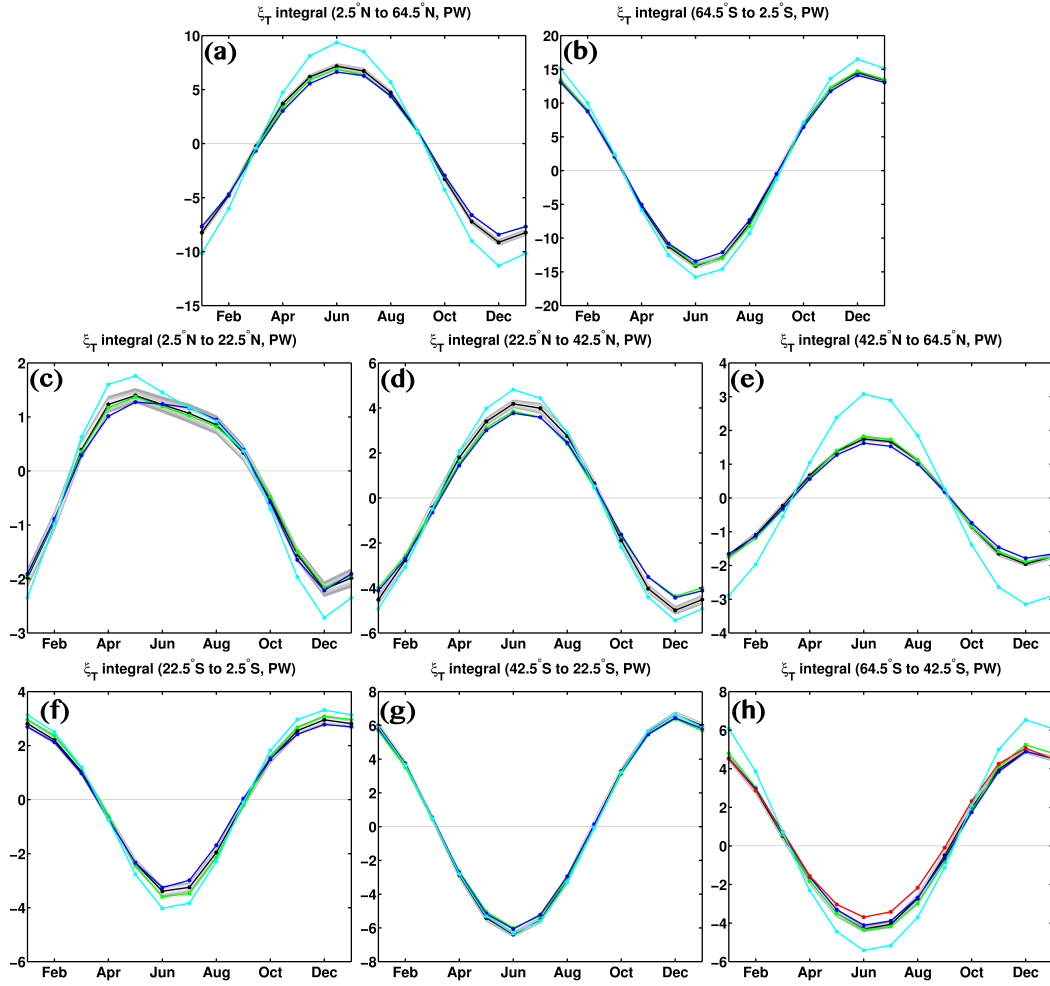


Figure 4.A3: Heat flux climatology: integral over the ocean (PW) in the Northern and Southern Hemisphere (NH and SH, panels a and b), NH tropics/subtropics/higher latitudes (panels c, d, e), SH tropics/subtropics/higher latitudes (panels f, g, h). Black line: Argo, i.e. no coastal regions nor marginal seas, with corresponding areas of integration of $1.00 \cdot 10^{14}$, $1.64 \cdot 10^{14}$ m² for the NH, SH; $5.05 \cdot 10^{13}$, $3.53 \cdot 10^{13}$, $1.44 \cdot 10^{13}$ m² for the NH tropics, subtropics, higher latitudes; $5.85 \cdot 10^{13}$, $6.02 \cdot 10^{13}$, $4.53 \cdot 10^{13}$ m² for the SH tropics, subtropics, higher latitudes. Green/blue line: ERAI/OA for the same region as the Argo black line. Cyan line: ERAI for the whole ocean region in the latitude band of interest, with corresponding areas of integration of $1.32 \cdot 10^{14}$, $1.88 \cdot 10^{14}$ m² for the NH, SH; $6.42 \cdot 10^{13}$, $4.36 \cdot 10^{13}$, $2.43 \cdot 10^{13}$ m² for the NH tropics, subtropics, higher latitudes; $6.71 \cdot 10^{13}$, $6.35 \cdot 10^{13}$, $5.69 \cdot 10^{13}$ m² for the SH tropics, subtropics, higher latitudes. The errorbars on the Argo estimate are shaded as in Fig.4.3.

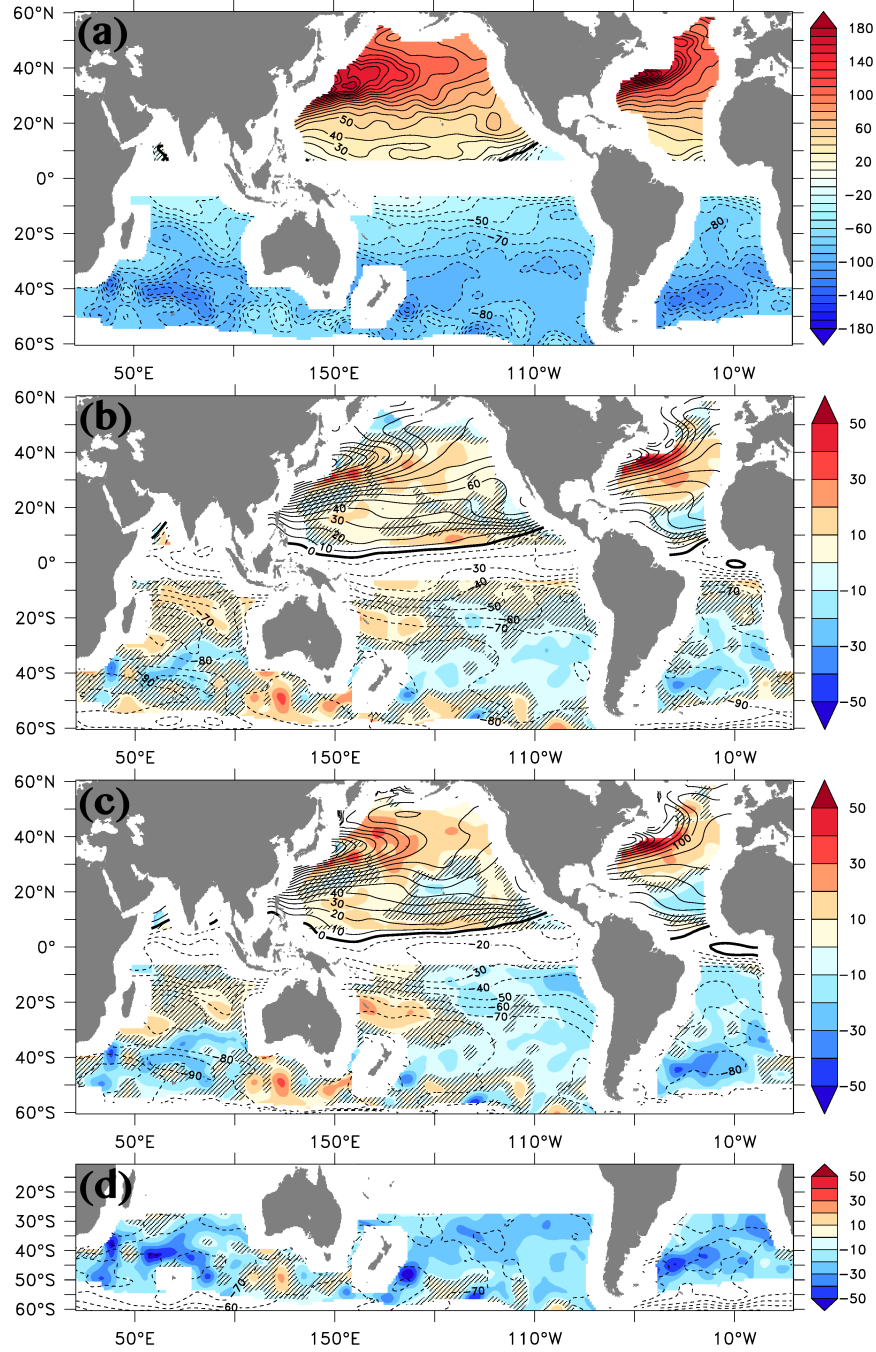


Figure 4.A4: Heat flux (Wm^{-2}) during the Northern Hemisphere summer (JJA average, black contours) from (a) Argo, (b) ERAI, (c) OA, (d) SOSE. The contour interval is 10 Wm^{-2} , negative values are dashed and the thick line is zero. In (a), the color follows the black contours. In (b-d), the color is the difference between Argo and each of the other products. Hatched regions indicate values that are smaller than $\xi_{T_{err}}$. Positive values are shaded red (i.e. red indicates that the ocean becomes warmer).

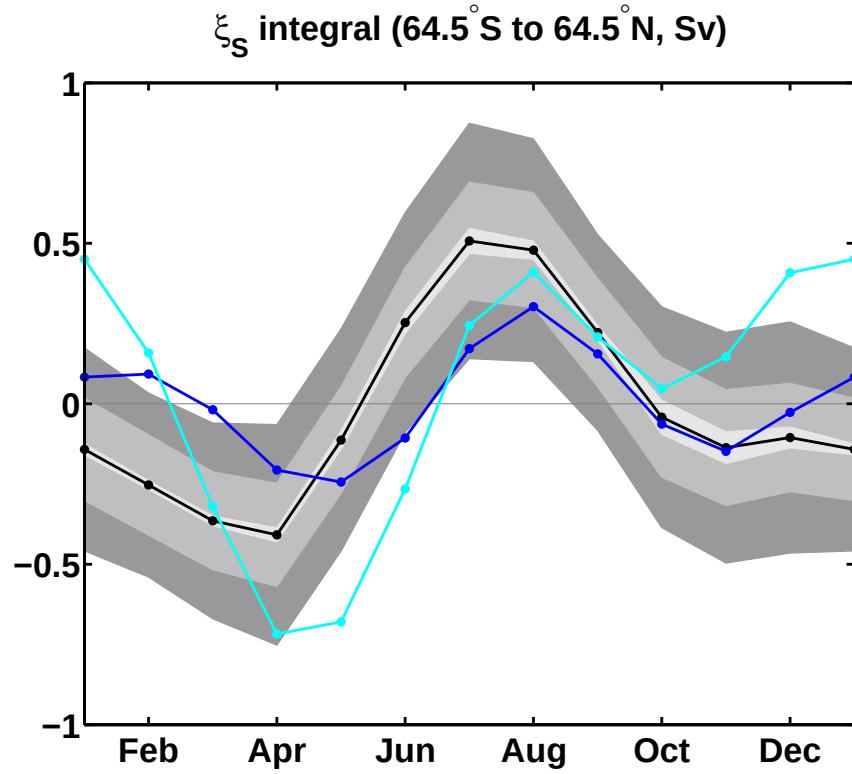


Figure 4.A5: Argo freshwater flux monthly climatology: integral over the ocean (Sv) for the estimate from advection and tendency term (black line), and for the tendency term alone (blue and cyan lines). The black and the blue line are integrals over the Argo domain, i.e. no coastal regions nor marginal seas within 64.5°S–64.5°N, and except equatorward of 2.5° degree latitude. The cyan line includes also the equatorial band.

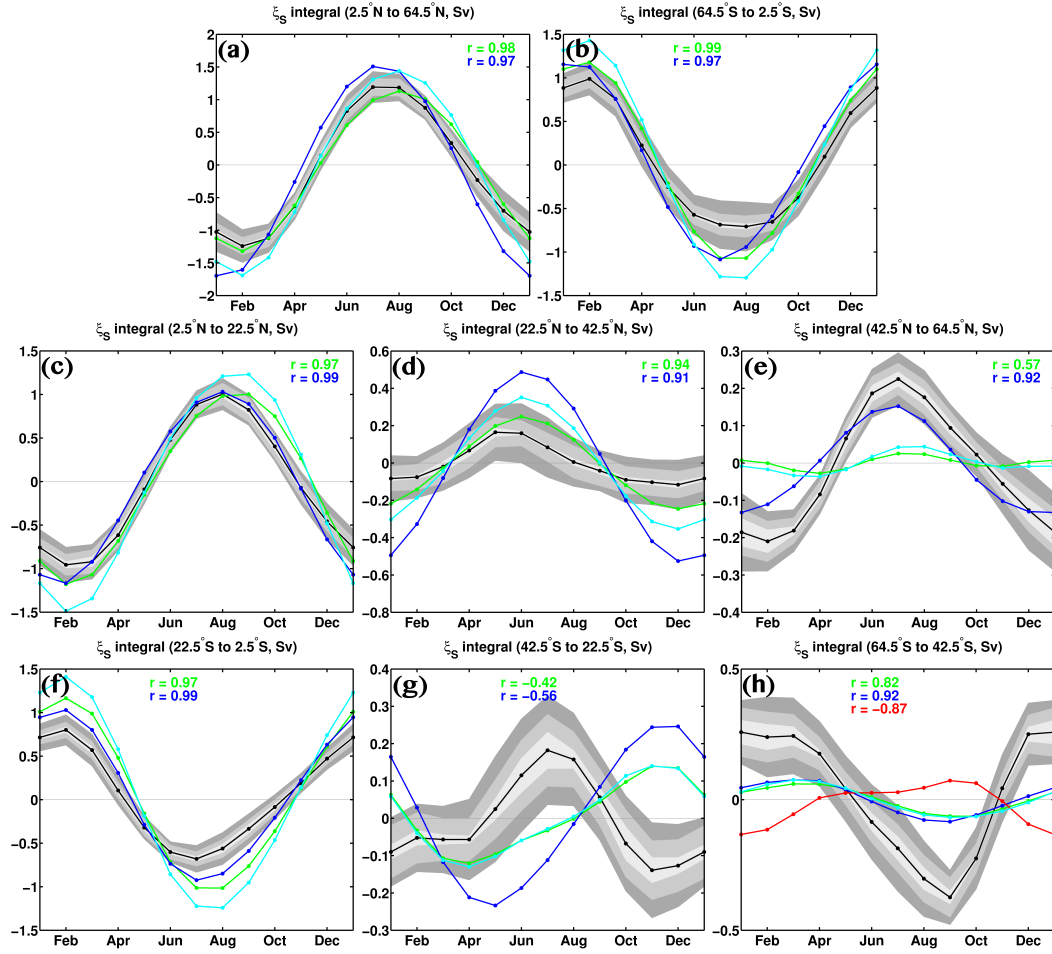


Figure 4.A6: Freshwater flux climatology: integral over the ocean (Sv) in the Northern and Southern Hemisphere (NH and SH, panels a and b), NH tropics/subtropics/higher latitudes (panels c, d, e), SH tropics/subtropics/higher latitudes (panels f, g, h). Black line: Argo, i.e. no coastal regions nor marginal seas. Green/blue line: ERAI/MERRA+OA for the same region as the Argo black line. Cyan line: ERAI for the whole ocean region in the latitude band of interest. The correlation coefficient (r) between the black and green/blue line is indicated in the legend (at least 95% confidence level). The errorbars on the Argo estimate are shaded as in Fig.4.3.

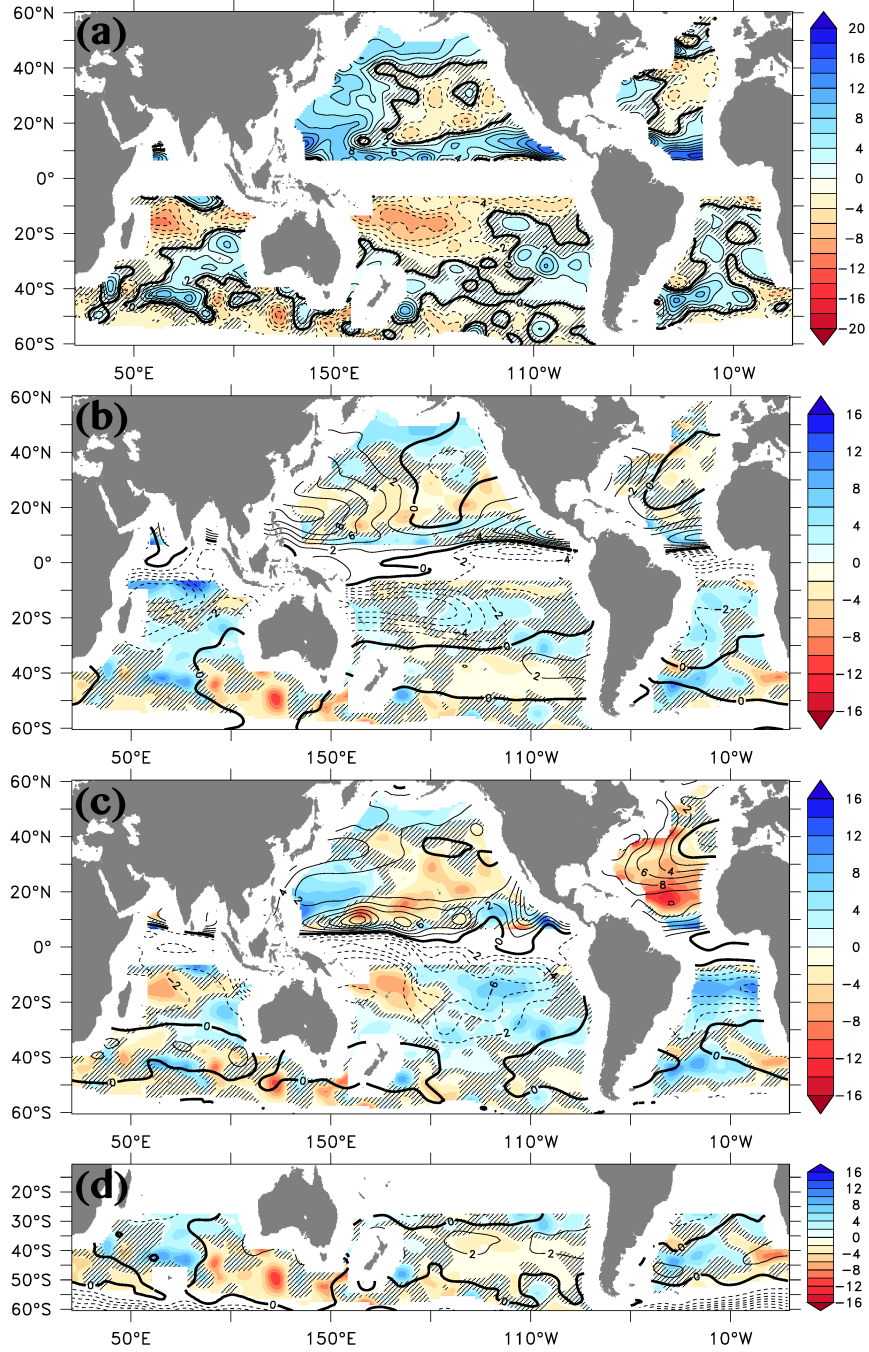


Figure 4.A7: Freshwater flux (Sv) during the Northern Hemisphere summer (JJA average, black contours) from (a) Argo, (b) ERAI, (c) MERRA+OA, (d) SOSE. The contour interval is 2 cm/month, negative values are dashed and the thick line is zero. In (a), the color follows the black contours. In (b-d), the color is the difference between Argo and each of the other products. Hatched regions indicate values that are smaller than $\xi_{S_{err}}$. Positive values are shaded blue (i.e. blue indicates that the ocean becomes fresher).

Chapter 5

The mean field of the North Pacific Subtropical gyre from Argo T/S profiles and trajectory data

Abstract. The three dimensional field of the North Pacific Subtropical gyre is described using Argo T/S profiles and trajectory data. The shape of the gyre defined here is that of a bowl that shifts northward as it deepens and is deepest in the northwest of the domain, where the Kuroshio Extension's (south) recirculation region is located. The gyre axis in the eastern portion of the gyre (east of $\sim 180^\circ\text{E}$) tilts northeast with increasing pressure, while does not change as much in the western region. Still, for the deepest isopycnals observed by Argo, a northeast tilt is noticeable in the west too.

The gyre circulation is present also on isopycnals that do not ventilate to the north, indicating that the atmospheric forcing does not directly drive the flow in the deeper portion of the gyre. Finally, referencing the flow to the Argo trajectory data (rather than to the 1975 db isobar), yields a stronger transport both in the zonal and meridional direction.

5.1 Introduction

Subtropical gyres are a key component of the Earth’s system, for their role in regulating both the Earth’s climate and the productivity of ecosystems. While gyres are regions of relatively low biological productivity, they have a large size and this makes their total contribution significant [McClain *et al.*, 2004]. Studying the physical processes that govern their circulation, is crucial to understanding property distributions (including biology) within the gyre. Also, in the Pacific and Indian Oceans, the bulk of the heat transport from the equator to the poles is associated with wind-driven gyres confined to the thermocline [Ferrari and Ferreira, 2011], and heat redistribution is a key aspect of the climate system.

In classical theories of gyre circulation, (i.e. assuming steady, near-laminar flow and planetary geostrophy), the upper thermocline (below the surface mixed layer) is dominated by advection and ventilated [Luyten *et al.*, 1983]. Below this (i.e. in the unventilated region), the flow is, instead, driven by momentum transfer from layers above [Rhines and Young, 1982]. The basin wide vertically integrated transport of the gyre is thought of in the context of Sverdrup dynamics [Sverdrup, 1947], with the wind driving the circulation by Ekman upwelling/downwelling.

Reid [1997] shows the circulation of the Pacific ocean on different vertical levels from hydrographic data and also describes the subtropical gyres in the Northern and Southern Hemispheres. Other studies have described these subtropical gyres in relation to existing theories [Keffey, 1985; Talley, 1985, 1988; Huang and Qiu, 1994], with the conclusion that the idealized dynamics applies to some regions. Qu [2002] used a climatology of temperature and salinity (World Ocean Atlas 1998) to investigate large-scale aspects of the North Pacific subtropical gyre 3D field. Results showed that, in the central and eastern parts of the basin, the axis of the subtropical gyre, defined as the meridional maximum of dynamic height, tends to move poleward from about 25°N near the surface to about 40°N within the upper intermediate layers. In the western region, instead, the axis was seen at about 30°N and almost unchanged with depth. Also, the vertical distribution was characterized by a northward shift of the bifurcation latitude of the North Equatorial Current at increasing depth and a barotropic nature of the confluence

point between the Kuroshio and Oyashio.

The advent of Argo profiles as well as trajectory data offers a unique opportunity to look at a 3D view of the ocean with unprecedented resolution on a global scale. Absolute geostrophic velocities from Argo have been used to perform a global analysis of the Sverdrup balance in the world ocean [Gray and Riser, 2014], showing a good agreement between observations and theory over the subtropics (away from the boundaries and from regions of strong eddy activity [Yuan *et al.*, 2014]). These results are consistent with analyses (predating Argo) at specific latitudes in the subtropical North Pacific, e.g. [Roemmich *et al.*, 1994; Aoki and Kutsuwada, 2008].

The present study describes the mean flow in the subtropical North Pacific and the 3D shape of the gyre-like circulation observed using Argo T/S profiles and trajectory data. The gyre circulation is defined here as a recirculating flow that closes in the subtropics, i.e. characterized by streamlines that move northward along the coast of Japan, enter the interior of the basin (from the northwest) and circulate all the way around proceeding northward once they approach land in the southwest of the domain (coming from the east). This definition is based exclusively on the streamlines of the flow and leaves out the eastern subtropical regions where the wind-driven flow does not close in the subtropics. Section 5.2 describes the data used for this analysis, and Section 5.3 the method. Results are presented in Section 5.4 and the summary and conclusions in Section 5.5.

5.2 Data

Argo profiles provide temperature, salinity, and pressure data on a global scale (64.5°S - 64.5°N, except for continental shelves, marginal seas, and ice-covered regions) and as deep as 2000 db [Roemmich *et al.*, 2009]. In this analysis, the raw Argo temperature and salinity measurements during 2004 – 2013 were gridded monthly on a $1^\circ \times 1^\circ$ grid, after quality control and adjustment of pressure bias. The gridding was done by objective mapping, with latitude-dependent decorrelation scales [Roemmich and Gilson, 2009].

The Argo trajectory data were quality controlled and used for the estimate of zonal and meridional velocity at the parking pressure (of the float) by the Argo team at SIO (*Park et al.* [2005]; *Scanderbeg et al.* [2014]; Megan Scanderbeg, personal communication). All the velocity data in the pressure range 975 – 1025 db were then objectively mapped as described in Section 5.3, to estimate Dynamic Height at 1000 db.

5.3 Objective Mapping of dynamic height from trajectory data.

Zonal and meridional velocities from Argo trajectory data were binned in 1-degree bins to create a dataset with a reduced noise level. All the quality controlled data available for the time period 2004-2013 and in the the pressure range 975 and 1025 db were included. Fig.5.1 shows the bins, with location assigned at the average latitude and longitude of the observations considered for each bin. Several trajectory data are present in each bin and only one region in the north-east of the domain lacks observations compared to the rest (Fig.5.1a). Also, for both zonal and meridional velocity, the highest variability in the bins corresponds to the Kuroshio Extension, as expected due to meandering and mesoscale features associated with the current (Fig.5.1b,c). Even with the large number of data averaged in each bin, the standard error associated with some of the bins has a bigger amplitude than the average velocity in the bin.

The binned zonal and meridional velocity were used to objectively map the mean field of dynamic height on the 1000 db isobar.

5.3.1 Anisotropic Objective Mapping of Streamfunctions

The general formulation

Objective mapping provides a formal methodology to map irregularly spaced observations onto desired coordinates, while minimizing the error [*Brether-*

ton *et al.*, 1976]. This method can also be applied to horizontal velocity data (u, v) to map dynamic height (η) . *Gille* [2003] derives the general case for such a procedure, i.e. mapping the streamfunction ψ from u, v data when the covariance function $\langle \psi(x, y) \psi(x + r, y + s) \rangle = F(\rho)$ is anisotropic (but symmetric), with $\rho = \rho(r, s)$. Both the covariance functions relating ψ to u or v and the velocity covariance functions can be derived from $F(\rho)$, using the relation between ψ and u, v , i.e. $u = -\frac{\partial \psi}{\partial y}$, $v = \frac{\partial \psi}{\partial x}$, as described in *Gille* [2003] (Eq. A1-A5). The estimate $\hat{\psi}$ of the true streamfunction ψ is then

$$\hat{\psi} = (\mathbf{A}^{-1} \mathbf{P})^T \phi, \quad (5.1)$$

where $\phi = [u_1 \ u_2 \ \cdots \ u_N \ v_1 \ v_2 \ \cdots \ v_N]$ is a column vector of velocity data $[2N \times 1]$, containing all u measurements and then all v measurements; $\mathbf{A} = \tilde{\mathbf{A}} (\mathbf{I} + \epsilon \mathbf{I})$ has dimensions $[2N \times 2N]$ and is the result of including noise in the covariance matrix of ϕ (i.e. $\tilde{\mathbf{A}} = \langle \phi \phi^T \rangle$), with ϵ being the noise to signal ratio (the addition of $\epsilon = \frac{N}{S}$ to the diagonal of $\tilde{\mathbf{A}}$ makes the resulting matrix $\mathbf{A}$ diagonally dominant and therefore easier to invert); $\mathbf{P} = \langle \phi \psi \rangle$ contains the covariances of ψ with ϕ and has dimensions $[2N \times M]$, with M being the number of locations where ψ should be estimated.

The squared error in ψ is

$$\frac{(\hat{\psi} - \psi)^2}{\psi^2} = 1 - \frac{\mathbf{P}^T \mathbf{A}^{-1} \mathbf{P}}{\text{diag}(\mathbf{S})}, \quad (5.2)$$

where $\text{diag}(\mathbf{S})$, $[M \times 1]$, is the diagonal of the covariance matrix of ψ , i.e. $\mathbf{S} = \langle \psi \psi \rangle$.

Application to the subtropical North Pacific

In this study, the dynamic height (η) mean field is estimated at 1000 db starting from Argo trajectory data (u_{obs}, v_{obs}) . By geostrophy, $fu = -\frac{\partial \eta}{\partial y}$, $fv = \frac{\partial \eta}{\partial x}$, where $f = f(y)$ is the Coriolis parameter varying with latitude. If u, v are redefined to be $u = fu'_{obs}$ and $v = fv'_{obs}$, with u_{obs}, v_{obs} (m/s) observed velocities at the pressure level of interest p^* , then, the formalism described in *Gille* [2003] and

summarized above can be applied to estimate η at p^* (now $p^* = 1000$ db). The estimate of the signal covariance for velocity from the assumed covariance for η does not enforce non-divergence of the velocity field [Lavender *et al.*, 2005]. Also, since the estimated covariance functions are between velocity and streamfunction anomalies, a first guess of the mean velocities should be removed from u_{obs}, v_{obs} (i.e. $u'_{obs} = u_{obs} - u_m, v'_{obs} = v_{obs} - v_m$, with u_m, v_m being a first guess of the mean field) and the dynamic height field corresponding to u_m, v_m (i.e. η_m) should be added to the output of the objective mapping to create an estimate for the mean field of η (i.e. $\hat{\eta} = \eta'_{obj.map} + \eta_m$, m^2/s , with the hat symbol indicating that it is an estimate). Here, η_m, u_m, v_m are computed from the gridded Argo T/S and referenced to the 1975 db isobar.

In this study, the covariance function $\langle \psi(x, y) \psi(x+r, y+s) \rangle = F(\rho)$ has the form $F(\rho) = C^2 \exp(-\rho^2)$, with $\rho^2 = \frac{r^2}{L_x^2} + \frac{s^2}{L_y^2}$, hence Eq.A11-A13 in Gille [2003] apply here too. For this form of $F(\rho)$, Eq.A14-A18 in Gille [2003] become:

$$\langle \psi(x, y) u(x+r, y+s) \rangle = -\frac{s}{L_y^2 \rho} \frac{dF}{d\rho} = 2 \frac{C^2 s}{L_y^2} \exp(-\rho^2) \quad (5.3)$$

$$\langle \psi(x, y) v(x+r, y+s) \rangle = \frac{r}{L_x^2 \rho} \frac{dF}{d\rho} = -2 \frac{C^2 r}{L_x^2} \exp(-\rho^2) \quad (5.4)$$

$$\begin{aligned} \langle u(x, y) u(x+r, y+s) \rangle &= \frac{1}{L_y^2} \left[-\left(1 - \frac{s^2}{L_y^2 \rho^2}\right) \frac{1}{\rho} \frac{dF}{d\rho} - \frac{s^2}{\rho^2 L_y^2} \frac{d^2 F}{d\rho^2} \right] \\ &= 2 \frac{C^2}{L_y^2} \exp(-\rho^2) \left[1 - \frac{2s^2}{L_y^2} \right] \end{aligned} \quad (5.5)$$

$$\begin{aligned} \langle v(x, y) v(x+r, y+s) \rangle &= \frac{1}{L_x^2} \left[-\left(1 - \frac{r^2}{L_x^2 \rho^2}\right) \frac{1}{\rho} \frac{dF}{d\rho} - \frac{r^2}{\rho^2 L_x^2} \frac{d^2 F}{d\rho^2} \right] \\ &= 2 \frac{C^2}{L_x^2} \exp(-\rho^2) \left[1 - \frac{2r^2}{L_x^2} \right] \end{aligned} \quad (5.6)$$

$$\begin{aligned} \langle u(x, y) v(x+r, y+s) \rangle &= \langle v(x, y) u(x+r, y+s) \rangle = \frac{rs}{\rho^2 L_x^2 L_y^2} \left(-\frac{1}{\rho} \frac{dF}{d\rho} + \frac{d^2 F}{d\rho^2} \right) \\ &= \frac{4rsC^2}{L_x^2 L_y^2} \exp(-\rho^2) \end{aligned} \quad (5.7)$$

Eq.5.3-5.7 describe how to fill in the matrices $\tilde{\mathbf{A}}$ and $\mathbf{P}$ for the objective mapping ($C^2 = 1$, $L_x = 400\text{km}$ and $L_y = 200\text{ km}$ in this study, with L_x and L_y consistent with the gridded Argo T/S), while $\mathbf{A} = \tilde{\mathbf{A}}(\mathbf{I} + \epsilon\mathbf{I})$ with $\epsilon = 2$. The result of the mapping shows a pattern that is not very sensitive to the choice of ϵ , only the amplitude changes (Fig.5.3). A larger ϵ corresponds to a larger noise (computed a posteriori), since it yields a smaller correction on the first guess from the mapping (left panels in Fig.5.4). Still, the range spanned by N and S even changing ϵ from 0.15 to 4 is not large, especially for N. The ratio between N and S is the actual ϵ of the output map, i.e. $\epsilon_m = \frac{\sum (u_{obs} - \hat{u})^2}{\sum (u_m - \hat{u})^2}$ (with $\hat{u}$ from the output map), and can be compared to the ϵ chosen a priori for the mapping. The a priori $\epsilon = 2$ used here, is consistent with the level of noise expected in trajectory data (from mesoscale variability) and with the small amplitude of the expected correction to the choice of first guess for u_m, v_m (i.e. $u'_{obj.map}, v'_{obj.map}$, the signal). Also, the extent to which ϵ and ϵ_m are consistent (i.e. the extent to which ϵ/ϵ_m gets close to 1) does not improve much by increasing ϵ beyond a value of 2 (right panels in Fig.5.4).

Once $\hat{\eta}, \hat{u}, \hat{v}$ at 1000 db were estimated using trajectory data, the difference with η_m, u_m, v_m was used to apply a correction on η_m, u_m, v_m at other pressure levels and account for a velocity different from zero at 1975 db. Hereforth, the hat symbol is dropped and η, u, v are all intended as estimates of the circulation referenced to the trajectory data (i.e. with a level of known motion at 1000 db based on Argo trajectory data).

5.4 Results.

Fig.5.2a shows (in color) the objectively mapped correction to the first guess of dynamic height (mean field) at 1000 db, i.e. the correction to η_m (with η_m shown in contours). Such a correction increases the gradients of η in the north-west of the domain (Fig.5.2b), indicating a stronger circulation there compared to referencing the flow to the 1975 db isobar (i.e. compared to η_m in Fig.5.2c). This intensification is particularly strong in the Kuroshio Extension's (south) recirculation gyre region,

i.e. just south of the current and between $140 - 160^\circ\text{E}$. In general, there is a noticeable difference between the pattern of dynamic height at 1000 db from Argo (Fig.5.2b) and the steric height at the same level described by *Reid* [1997] (Fig.5e). The same is true for the 1500 and 2000 db levels (not shown). As expected, results here are in better agreement with maps from Argo described by *Gray and Riser* [2014] (see their Fig.4) and the differences are likely related to the details of the mapping procedure and spatial smoothing, and to a shorter time record (December 2004 to November 2010) compared to the current study (2004 – 2013). Finally, the best agreement is observed with the map of dynamic height at 1000 db from Argo shown by *Ollitrault and Colin de Verdière* [2014] (their Fig.16).

The stronger zonal flow compared to the first guess u_m (Fig.5.5d) is shown in Fig.5.5b: the difference between the two (i.e. the velocity corresponding to the objectively mapped correction to η_m) is positive in the Kuroshio Extension region and negative for the return flow (Fig.5.5c). Other features of the zonal flow are also intensified including the objectively mapped correction (e.g. the subtropical countercurrents [*Aoki et al.*, 2002; *Kobashi and Xie*, 2011]), but the amplitude of the resulting total flow in Fig.5.5b is generally smaller compared to the binned observations (Fig.5.5a).

5.4.1 Zonal and meridional transport.

Fig.5.6 shows the zonal transport per degree latitude in the upper 1900 db of the water column ($150 - 170^\circ\text{E}$ average). The eastward transport at 34°N is stronger if the flow is referenced to the Argo trajectory data rather than to the 1975 db isobar. The opposite occurs for the westward flow between 21° and 29°N . Error bars represent the standard error and were computed for each latitude (y_j) from the variance of trajectory data in 3-degree latitude zonal bands centered at y_j .

For the meridional transport between $135 - 260^\circ\text{E}$ (Fig.5.7), the errorbars on the estimate that includes trajectory data are large, since the meridional velocities are quite small compared to mesoscale variability. Still, there is indication of stronger transport at ~ 16 and 31°N for the flow referenced to the Argo trajectory

data.

5.4.2 Geostrophic streamfunction on isopycnal surfaces: the shape of the gyre.

Fig.5.8-5.11 show maps of the geostrophic streamfunction on isopycnal surfaces [McDougall and Klocker, 2010] when including the trajectory data (case 1, in black) and when referencing the flow to the 1975 db isobar (case 2, in gray). The main feature of the circulation is the North Pacific subtropical gyre. On shallow isopycnals, the difference between case 1 and 2 is very small (not shown), but on denser isopycnals it becomes noticeable (Fig.5.8-5.11). Case 1 is characterized by a stronger flow (larger gradients for the streamfunction) and a southern edge of the streamlines that is displaced northward relative to case 2 (Fig.5.8-5.10). The flow is stronger for case 1, also on the deepest isopycnal observed by Argo ($\sigma_\theta = 27.6$ kg/m³, Fig.5.11), indicating a potentially deeper gyre than in case 2. Here, the circulation includes the Kuroshio Extension's (south) recirculation gyre [Qiu *et al.*, 2008; Jayne *et al.*, 2009] and it is not possible to distinguish between this and the traditional wind-driven gyre. Also, on this isopycnal, the circulation is zonally wider in case 1 compared to 2. Finally, the isopycnal surfaces in Fig.5.8-5.11 do not ventilate northward, indicating that the atmospheric forcing is not driving directly the flow at these depths (e.g. by ventilation).

Fig.5.12a maps the maximum potential density surface where a gyre circulation is observed in Argo (with the corresponding pressure in Fig.5.12b). The horizontal boundary of the gyre-like circulation is shown in Fig.5.12c for different potential density levels and was selected, for each isopycnal, based on which streamline goes all around from the northwest to the southwest region of the domain and proceeds northward once there (i.e. the recirculating flow that closes in the subtropics, as described also at the end of Section 5.1). Fig.5.12 is for the flow referenced to 1975 db since the objective mapping of the trajectory data involves derivatives and is not reliable where data are not available on one side of the grid point (i.e. the southwest region of the domain where the flow is close to the

coast). Data remain sparse very close to the coast and in marginal seas. Hence, an assumption is needed about the fact that the streamlines that turn northward once in proximity of the southwest boundary, actually join the Kuroshio (i.e. do not leave the gyre and enter into the South China Sea). This is not expected to change substantially the picture drawn here. The 3D shape of the gyre circulation in Fig.5.12c is that of a bowl that shifts northward as it deepens and is deepest in the northwest of the domain, where the Kuroshio Extension's (south) recirculation gyre is located. The southern edge of the gyre shifts northward as the depth increases, along with the bifurcation of the North Equatorial Current [Qu, 2002]. Also, the northeast edge of the gyre-like circulation does not have a counterpart at shallower levels, since the shallower streamlines there do not proceed northward once they reach the southwest of the domain.

The gyre's axis on different isopycnals is defined (for each longitude) at the latitude where the geostrophic streamfunction is maximum and is shown as a piecewise constant linear interpolation in Fig.5.13. Consistent with the results of Qu [2002], the location of the axis in the eastern portion of the gyre (east of $\sim 180^\circ\text{E}$) tilts northeast with increasing pressure but does not change as much in the western region (where it lays around 30°N). A northeast tilt is noticeable, here, even for the deepest isopycnals observed by Argo.

Finally, the meridional transport within the gyre circulation defined here is not expected to be consistent with an estimate from Sverdrup dynamics, since the Sverdrup balance applies in a vertically integrated framework and for the transport integrated from the eastern boundary. Here, a part of the water column is omitted and not all of the zonal extent of the basin is included.

5.5 Summary and conclusions.

The unprecedented resolution and coverage of the Argo array was used to describe the 3D structure of the North Pacific subtropical gyre. Referencing the flow to the Argo trajectory data (rather than to the 1975 db isobar), yields a stronger transport both in the zonal and meridional direction (Fig.5.6). Also, the

gyre circulation is present on isopycnals that do not ventilate to the north (Fig.5.8-5.11), indicating that the atmospheric forcing does not directly drive the flow in the deeper portion of the gyre. Finally, the 3D shape of the gyre circulation defined here for the subtropical North Pacific is that of a bowl that shifts northward as it deepens and is deepest in the northwest of the domain (Fig.5.12c), where the Kuroshio Extension's (south) recirculation region is located. The gyre axis in the eastern portion of the gyre (east of $\sim 180^\circ\text{E}$) tilts northeast with increasing pressure, but does not change as much in the western region. These results are consistent with *Qu* [2002], who used a climatology of temperature and salinity from the World Ocean Atlas to investigate large-scale aspects of the North Pacific subtropical gyre 3D field. Still, the present analysis from Argo, reveals a slight northeast tilt also for the deepest observed isopycnals in the western portion of the gyre.

With the advent of deep Argo it will be possible to complete the description of the deepest region of the gyre flow, which is missing in this analysis. Here, the circulation is still evident on the deepest isopycnals resolved by Argo, even when referencing the velocity to the 1975 db isobar (Fig.5.11). The stronger circulation (also on this level) for the case that uses trajectory data is consistent with previous studies that show how the Kuroshio Extension's (south) recirculation gyre is deeper than 2000 db [*Jayne et al.*, 2009].

Argo observations are a unique tool to gain a 3D view of the ocean and, in this study, they have allowed us to define the 3D boundary of the gyre with unprecedented resolution and accuracy, based on closed streamlines. With a longer Argo record it will be possible to describe the interannual variability of the gyre and investigate how the transport both within and outside its boundary changes in time. Such variability has important implications for the redistribution of heat and freshwater in the Earth's system, for air-sea exchanges and ultimately for how the climate will evolve.

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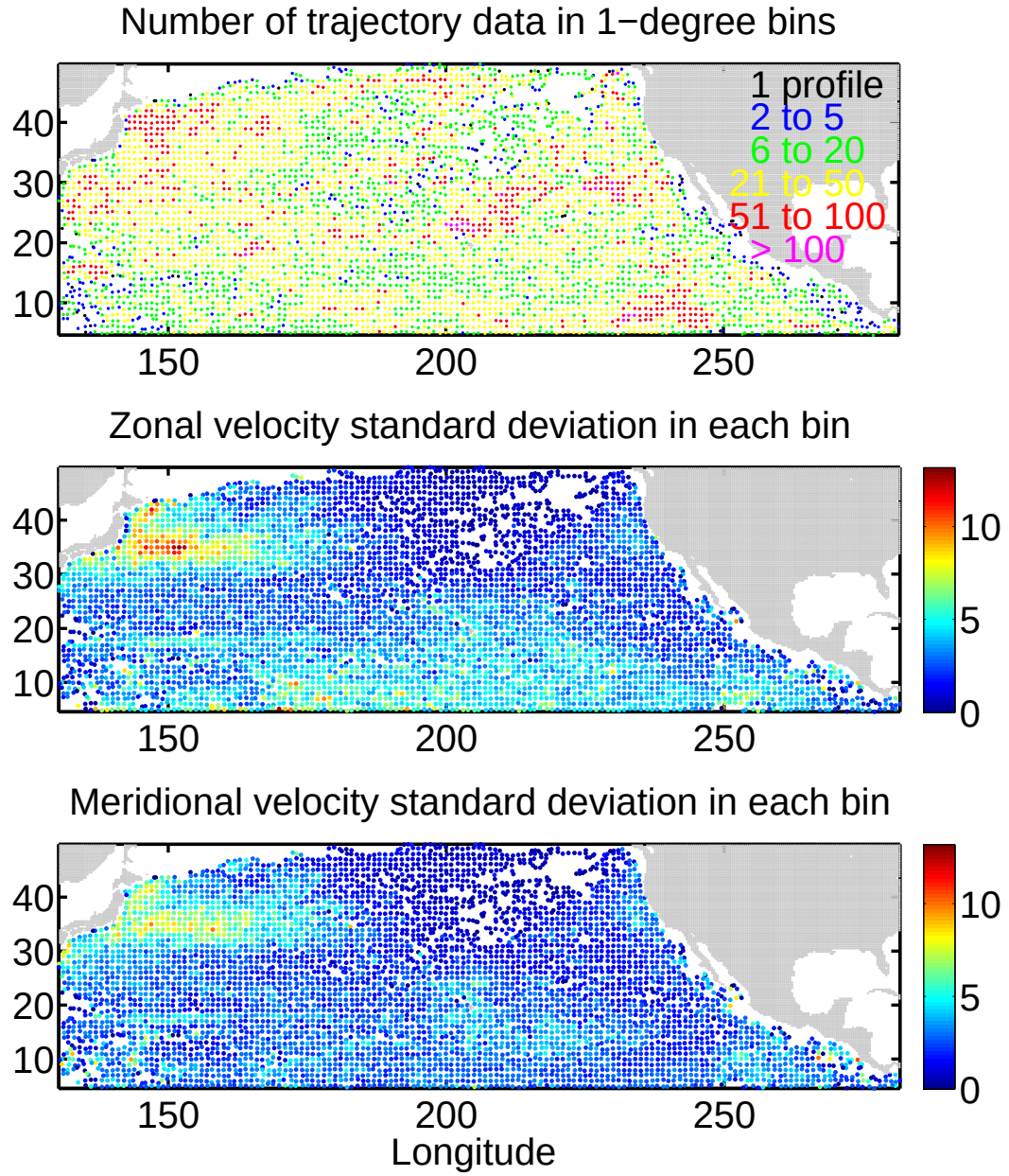


Figure 5.1: Binning of the raw trajectory data in 1-degree bins: (a) number of data in each bin and (b-c) standard deviation for (b) zonal and (c) meridional velocity.

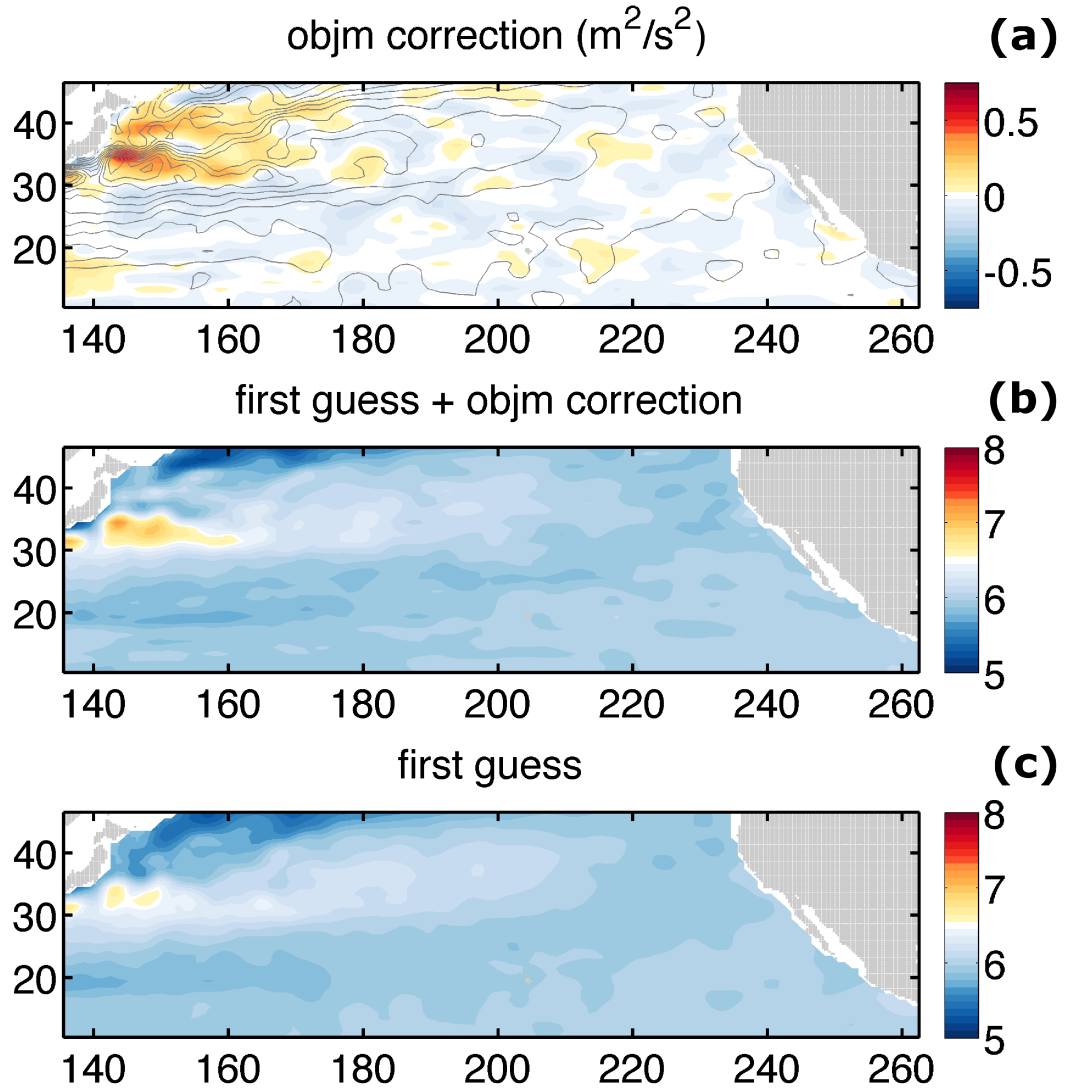


Figure 5.2: Mean dynamic height (m²/s²) from Argo at 1000 db: (b) referenced to the mapped trajectory data, (c) referenced to the pressure level $p_{ref} = 1975$ db (i.e using only mapped T/S profiles) and (a) the difference between the two (in color, with contours that are the same variable as in panel c).

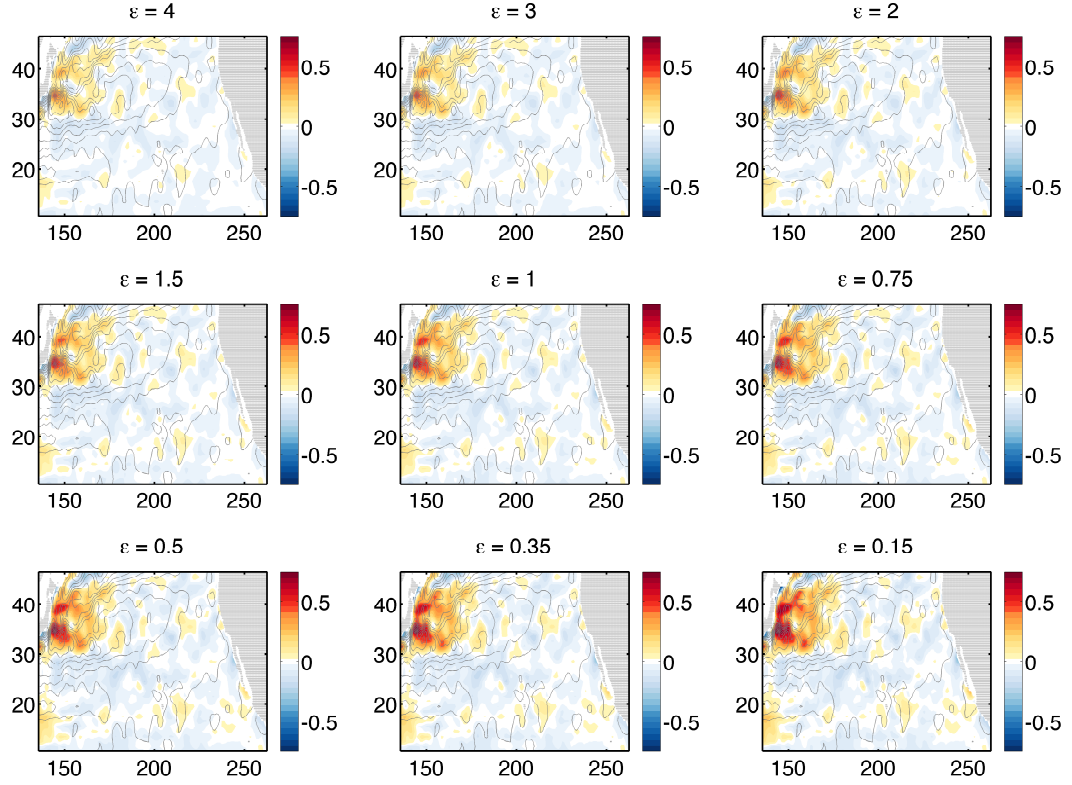


Figure 5.3: Mean dynamic height correction (m^2/s^2) from Argo at 1000 db: difference (in color) between using trajectory data and referencing to the pressure level $p_{ref} = 1975$ db (similar to Fig.5.2a). Contours are the same as in Fig.5.2a. Different panels are for different values of ϵ in the objective mapping.

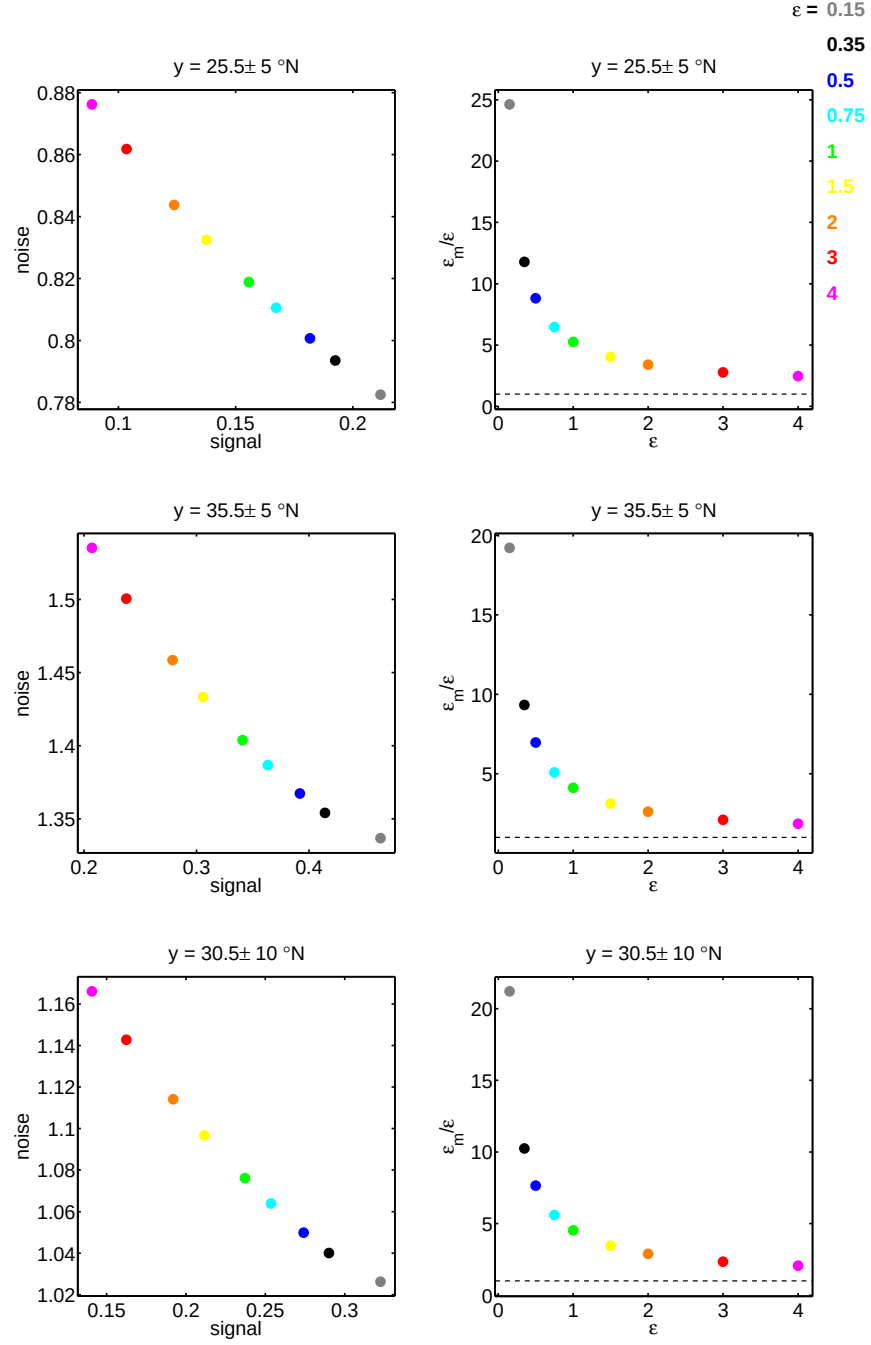


Figure 5.4: Right panels: ratio between ϵ_m (i.e. ϵ computed after the mapping) and ϵ (i.e. used in the mapping). Left panels: signal versus noise computed from the mapped zonal velocity (i.e. N and S used to compute ϵ_m), when using different ϵ (in color) for the mapping. The latitude band of interest for different panels is indicated in the title.

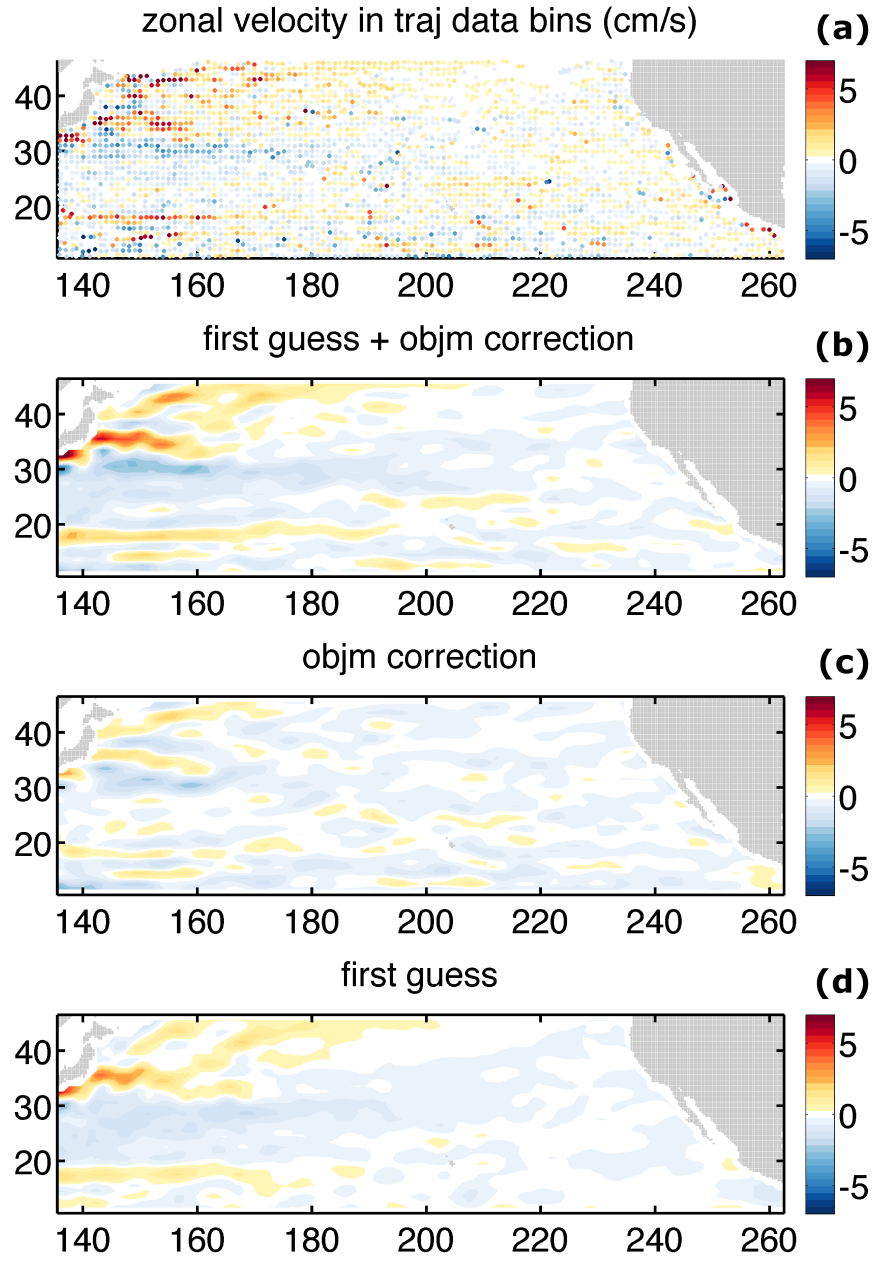


Figure 5.5: Zonal velocity from Argo (cm/s): (a) binned trajectory data, (b) mapped (i.e. referenced to the trajectory data), (d) geostrophic flow referenced to $p_{ref} = 1975$ db, (c) difference between panel b and d.

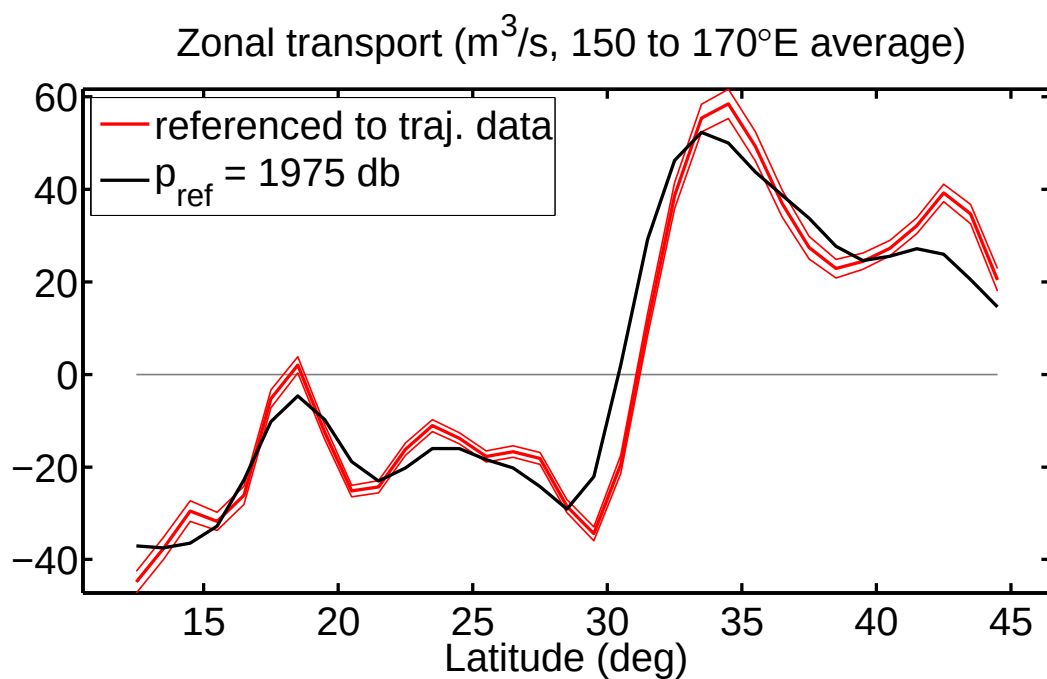


Figure 5.6: Zonal transport per degree latitude (m^3/s) in the upper 1900 db (150 – 170°E average): referenced to the trajectory data (red line, with errorbars) and referenced to $p_{\text{ref}} = 1975 \text{ db}$ (black line).

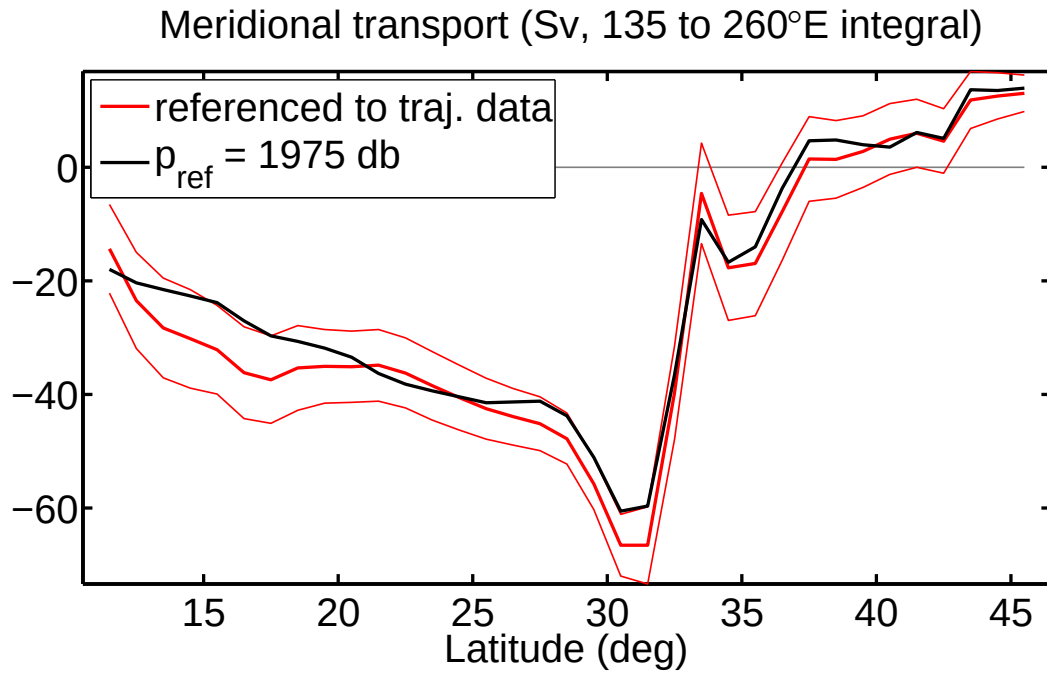


Figure 5.7: Meridional transport (Sv) in the upper 1900 db and between 135 – 260°E: referenced to the trajectory data (red line, with errorbars) and referenced to $p_{ref} = 1975$ db (black line).

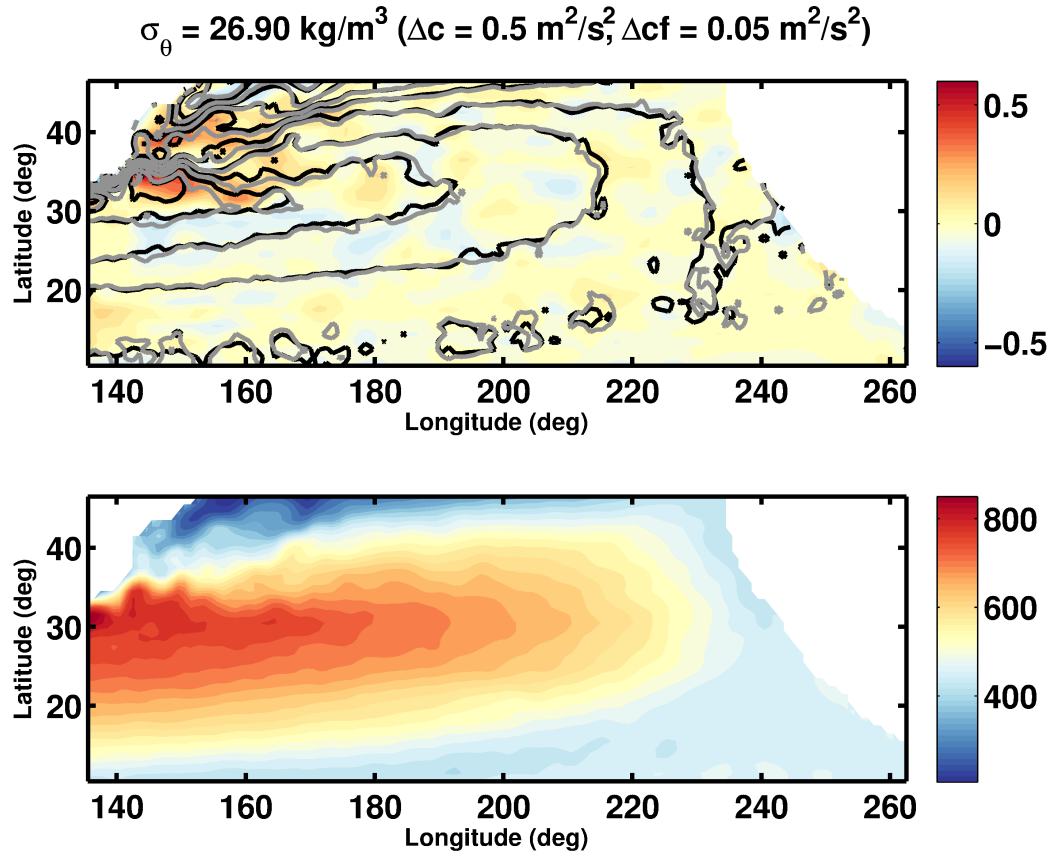


Figure 5.8: Top panel: geostrophic streamfunction on the isopycnal surface $\sigma_\theta = 26.9 \text{ kg/m}^3$ referenced to $p_{ref} = 1975 \text{ db}$ (gray line) and to the trajectory data (black line). The color is the same as in Fig.5.2a. The contour interval is $.5 \text{ m}^2/\text{s}^2$ for the lines and $.05 \text{ m}^2/\text{s}^2$ for the color. Bottom panel: pressure (db) on the isopycnal surface σ_θ (with a contour interval of 25 db).

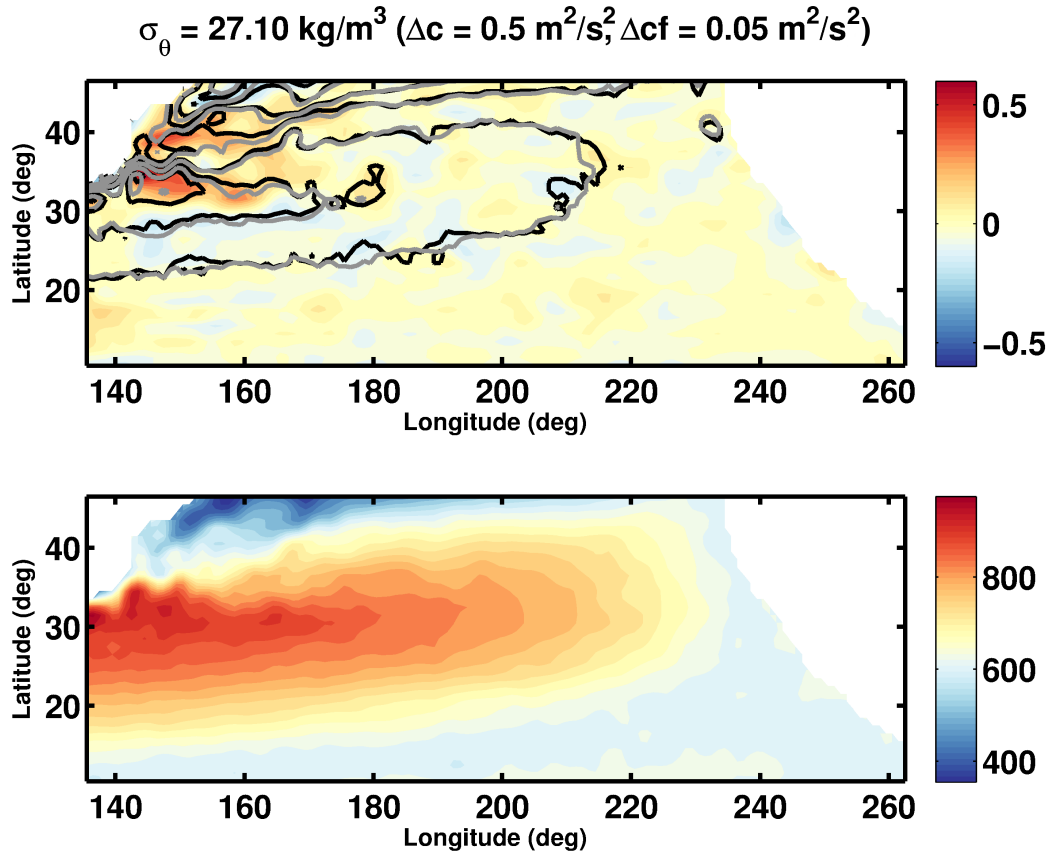


Figure 5.9: Top panel: geostrophic streamfunction on the isopycnal surface $\sigma_\theta = 27.1 \text{ kg/m}^3$ referenced to $p_{ref} = 1975 \text{ db}$ (gray line) and to the trajectory data (black line). The color is the same as in Fig.5.2a. The contour interval is $.5 \text{ m}^2/\text{s}^2$ for the lines and $.05 \text{ m}^2/\text{s}^2$ for the color. Bottom panel: pressure (db) on the isopycnal surface σ_θ (with a contour interval of 25 db).

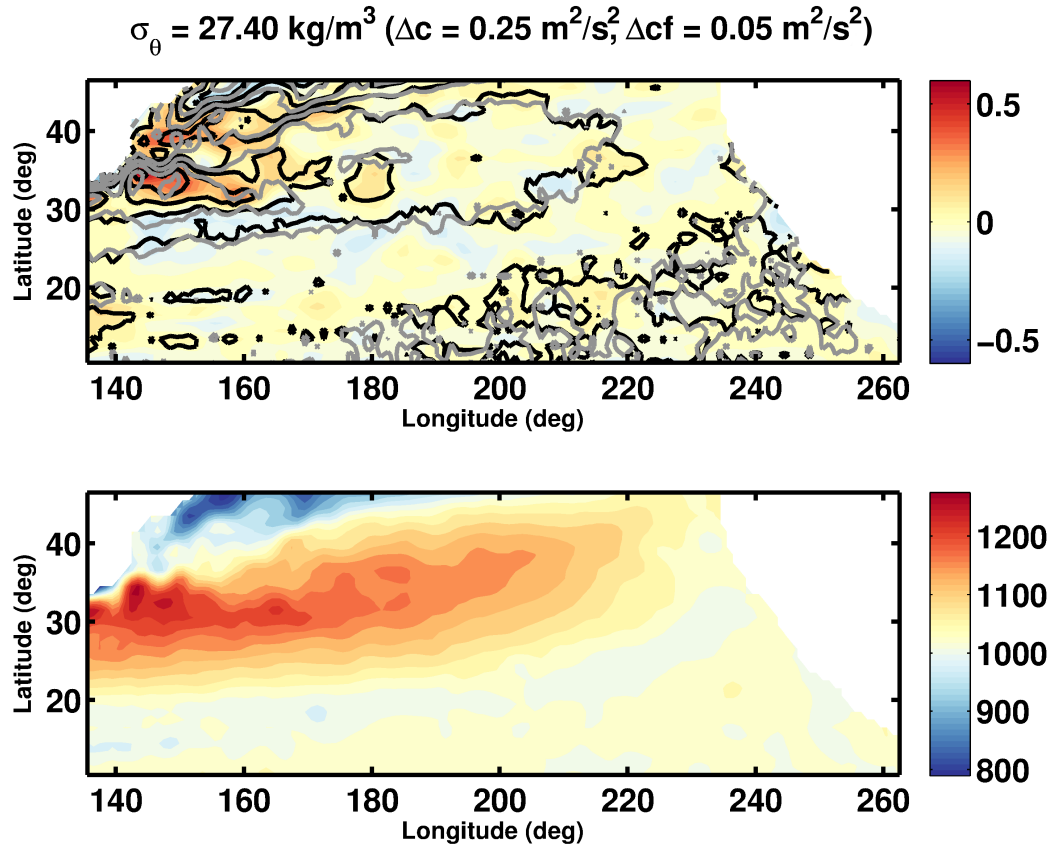


Figure 5.10: Top panel: geostrophic streamfunction on the isopycnal surface $\sigma_{\theta} = 27.4 \text{ kg/m}^3$ referenced to $p_{ref} = 1975 \text{ db}$ (gray line) and to the trajectory data (black line). The color is the same as in Fig.5.2a. The contour interval is .25 m^2/s^2 for the lines and .05 m^2/s^2 for the color. Bottom panel: pressure (db) on the isopycnal surface σ_{θ} (with a contour interval of 25 db).

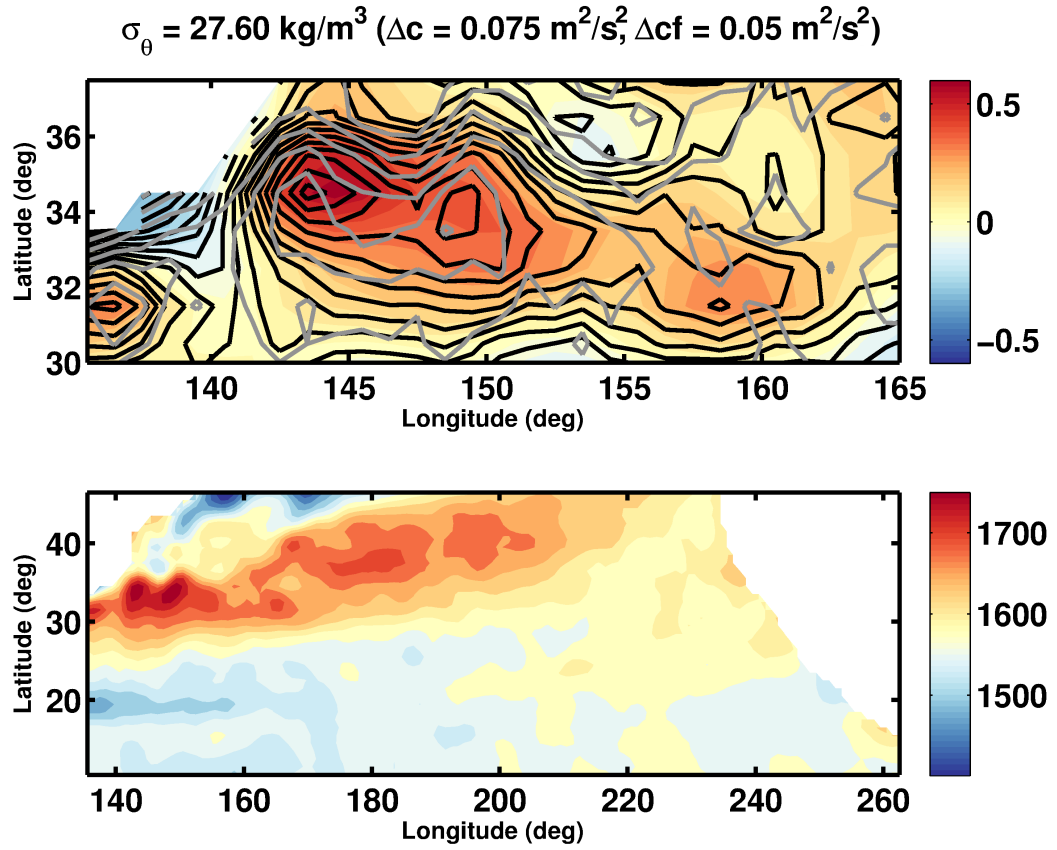


Figure 5.11: Top panel: geostrophic streamfunction on the isopycnal surface $\sigma_\theta = 27.6 \text{ kg/m}^3$ referenced to $p_{ref} = 1975 \text{ db}$ (gray line) and to the trajectory data (black line). The color is the same as in Fig.5.2a. The contour interval is $.075 \text{ m}^2/\text{s}^2$ for the lines and $.05 \text{ m}^2/\text{s}^2$ for the color. Bottom panel: pressure (db) on the isopycnal surface σ_θ (with a contour interval of 25 db).

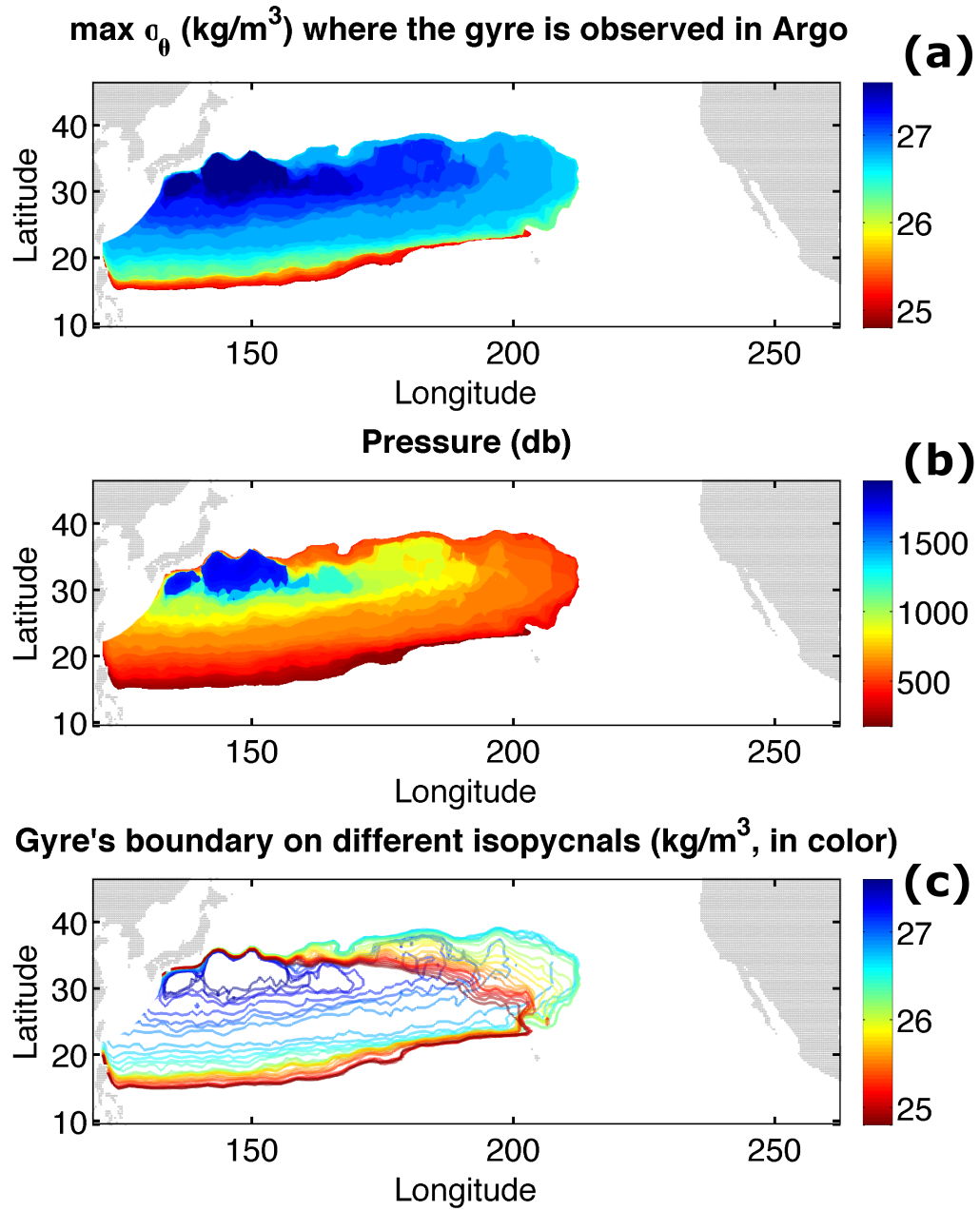


Figure 5.12: Map of (a) the maximum potential density surface (kg/m³) where the gyre circulation is observed in Argo and (b) the corresponding pressure. (c) Horizontal boundary of the gyre on different isopycnals.

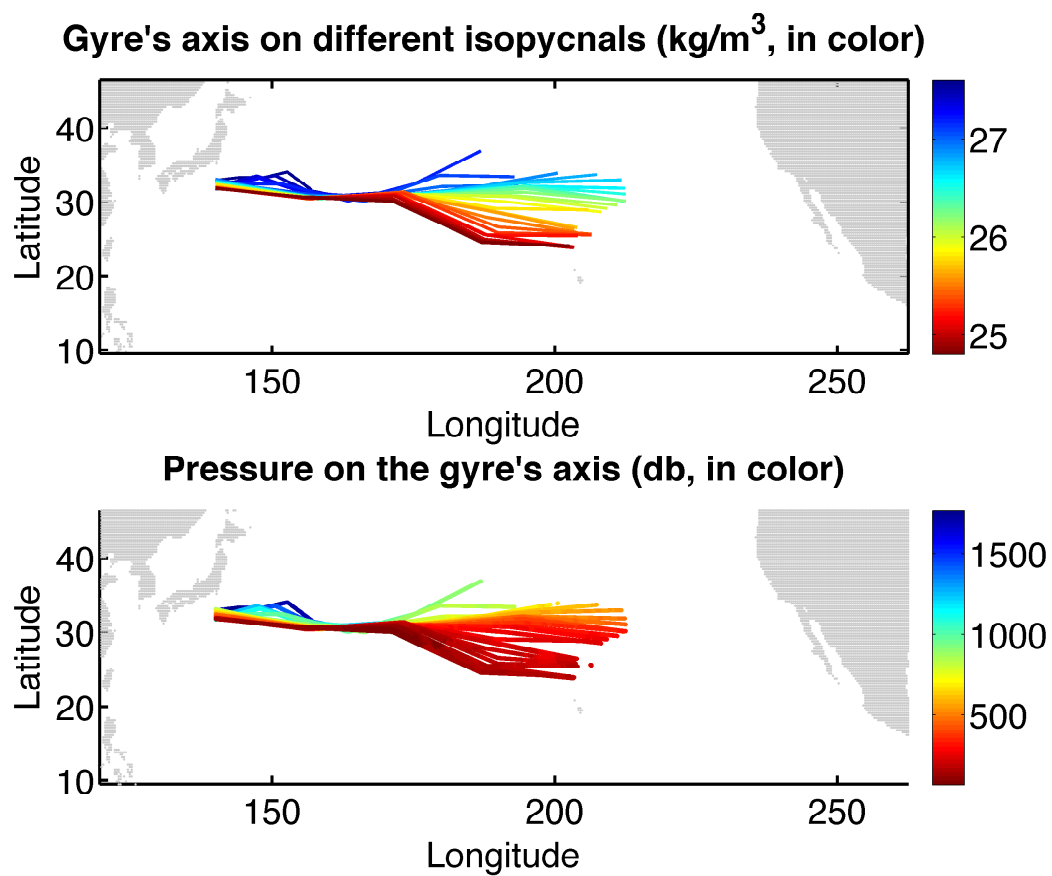


Figure 5.13: (a) Gyre's axis on different isopycnals (kg/m^3 , in color) and (b) corresponding pressure (db, in color).

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