Geostrophic Circulation in the Tropical North Pacific Ocean Based on Argo Profiles

Dongliang Yuan$^{1,2,*}$, Zhichun Zhang$^{1,3,4}$, Peter C. Chu$^5$, and William K., Dewar$^6$

$^1$Key Laboratory of Ocean Circulation and Waves, and Institute of Oceanology, Chinese Academy of Sciences, Qingdao, China

$^2$Qingdao Collaborative Innovation Center of Marine Science and Technology, Qingdao, China

$^3$Department of Ocean Science and Engineering, Zhejiang University, Hangzhou, China

$^4$State Key Laboratory of Tropical Oceanography, South China Sea Institute of Oceanology, Chinese Academy of Sciences, Guanzhou, China

$^5$Naval Ocean Analysis and Prediction (NOAP) Laboratory, Department of Oceanography, Naval Postgraduate School, Monterey, California, U.S.A

$^6$Department of Ocean, Atmosphere and Earth sciences, Florida State University, Tallahassee, Florida, USA

*Corresponding author

Email: dyuan@ms.qdio.ac.cn
Abstract

Absolute geostrophic currents in the North Pacific Ocean are calculated from the newly gridded Argo Profiling float data using the P-vector method for the period of 2004 through 2011. The zonal geostrophic currents based on the Argo profile data are found to be stronger than those based on the traditional WOA09 data. A westward mean geostrophic flow underneath the North Equatorial Countercurrent is identified using the Argo data, which is evidenced by sporadic direct current measurements and geostrophic calculations in history. This current originates east of the dateline and transports more than $4 \times 10^6$ m$^3$ s$^{-1}$ water westward in the subsurface northwestern tropical Pacific Ocean. We name this current the North Equatorial Subsurface Current.

The transport in the geostrophic currents is compared with the Sverdrup theory and found to differ significantly in several locations. Analyses have shown that errors of wind stress estimation cannot account for all of the differences. The largest differences are found in the area immediately north and south of the bifurcation latitude of the North Equatorial Current west of the dateline and in the recirculation area of the Kuroshio and its extension, where nonlinear activities are vigorous. It is, therefore, suggested that the linear dynamics of the Sverdrup theory is deficient in explaining the geostrophic transport of the tropical northwestern Pacific Ocean.

Key word: Sverdrup theory; absolute geostrophic current; P-vector
1. **Introduction**

The alternating zonal currents of the low-latitude North Pacific Ocean have long ago been recognized by the oceanographic community based on ship drift data and sporadic hydrographic surveys (Schott, 1939; Reid, 1961; Wyrtki, 1965). The upper ocean circulation was suggested by Sverdrup (1947) to be forced by the curl of the wind stress through the so-called Sverdrup balance. The theory assumes a linear dynamic framework and has obtained the meridional transport of the wind-driven ocean circulation by integrating the wind-stress curl without detailed information of the oceanic baroclinicity. The zonal transport of the wind-driven circulation is then obtained by integrating the streamfunction from the eastern boundary of the North Pacific Ocean. Due to the lack of sufficient measurements to estimate the mean circulation accurately, the validity of the theory has not previously been tested in the North Pacific Ocean.

A few studies have, however, attempted to verify the accuracy of the theory in the Atlantic Ocean. The first was that of Leetmaa et al. [1977], which inspired the study of Wunsch and Roemmich [1985], Böning et al. [1991], Schmitz et al. [1992], etc. Their results have shown that the Sverdrup meridional transport is generally consistent with the meridional transport calculated directly from the geostrophic currents based on hydrographic data in the northeastern subtropical North Atlantic Ocean, but is inconsistent with the geostrophic transports in the northwestern subtropical North Atlantic Ocean. The difference has been attributed to buoyancy-forced meridional overturning circulation in the North Atlantic Ocean.
In the Pacific Ocean, Meyers [1980] discussed the meridional transport of North Equatorial Countercurrent (NECC) and found significant inconsistency with the Sverdrup theory. Hautala et al. [1994] estimated the meridional transport of the North Pacific subtropical gyre along 24°N and noted that the Sverdrup balance is not valid in the western subtropical Pacific Ocean. The two studies are still very limited in disclosing the basin-scale differences from the Sverdrup theory, and neither of the studies has elaborated on the reasons of the inconsistency.

All of the above existing evaluations of the Sverdrup balance are based on one-time hydrographic measurements in a cross-basin section and have only been able to evaluate the accuracy of the theory in an integrated meridional transport from the eastern boundary. Lately, Wunsch [2011] has evaluated the accuracy of the Sverdrup theory using an assimilated global ocean dataset. A point-wise evaluation of the Sverdrup balance in the real ocean is highly desired but has not been fulfilled so far due to sparse and uneven distribution of the hydrographic casts in time and space of the world oceans, which will inevitably bring significant aliasing errors into the mean circulation and meridional transport.

By far the largest zonal surface current in the tropical North Pacific Ocean is the North Equatorial Current (NEC) flowing westward across the Pacific basin in the latitudinal band of roughly 7° ~ 20° N (Wyrtki, 1965; Nitani, 1975). On both sides of NEC are two eastward currents called the North Equatorial Countercurrent (NECC) in the south and the Subtropical Countercurrent (STCC) in the north. Existing studies of these zonal currents are based primarily on hydrography measurements conducted
with uneven temporal and spatial distributions. Good estimates of the
three-dimensional mean structure and the seasonal-to-interannual variations of these
currents at the basin scales have not been achieved in the past.

So far, there have been several methods to estimate geostrophic currents from the
temperature and salinity profile data, such as the traditional dynamic height
calculation in reference to a level of no motion, the $\beta$ spiral method of Stommel and
Schott (1977), the box inverse method of Wunsch (1978), and the P-vector method of
Chu (1995). The traditional dynamic height calculation is simple, but cannot obtain
the absolute geostrophic currents due to the use of a reference level. This major
disadvantage has promoted the invention of more objective methods, such as the $\beta$
spiral method, the box inverse method, and the P-vector method, to determine the
absolute geostrophic currents. The $\beta$ spiral method is based on the conservation of
mass, density, and planetary vorticity to estimate the reference velocities (Stommel
and Schott, 1977; Schott and Stommel, 1978). Although the algorithm forms an
over-determined problem when multiple levels of hydrographic data are used, the
calculated geostrophic currents are generally noisy due to the use of the second order
derivatives of the potential density surfaces in the calculation. The box inverse
method estimates the reference velocities by balancing the fluxes of water and tracers
into and out of a closed region (Wunsch, 1978). However, this method forms an
underdetermined system (the number of equations is less than the number of
unknowns) and is best applicable for ship-based surveys in a small region. The
P-vector method assumes conservation of potential vorticity and density and forms an
over-determined system if multiple levels of hydrographic data are used.

Dynamically, the P-vector method is equivalent to the $\beta$ spiral method under the
Boussinesq and geostrophic approximation (Appendix A), but is able to control the
errors of the calculation well through the use of the potential density gradients only

The advent of the Argo project in the world oceans has ushered in an
unprecedented era of sampling the world oceans with synchrony at basin and global
scales. These data can be used to study the general circulation at basin and global
scales and to evaluate the accuracy of the Sverdrup balance. In this study, we calculate
the absolute geostrophic currents (AGC) in the North Pacific Ocean based on the
newly gridded Argo Profiling float data using the P-vector method. The
three-dimensional structure of the mean circulation is then studied and the meridional
transport of the geostrophic currents is calculated to evaluate the accuracy of the
Sverdrup theory.

In section 2, the data and the P-vector method used in this study are introduced.
In Section 3, the accuracy of the AGC is assessed using the altimeter data and in situ
mooring measurements in the North Pacific Ocean. The three-dimensional structure
of the alternating zonal currents is studied and the transport of the mean geostrophic
currents is computed and compared with the Sverdrup theory. The error estimates
and the sensitivity of the geostrophic transports with respect to the depth of
integration and wind products are discussed in Section 4. Conclusions are
summarized in Section 5.
2. Data and method

2.1 Data

The gridded Argo data used in this study are obtained from the Argo page (http://www.argo.ucsd.edu/Gridded_fields.html). The product name is “Global gridded NetCDF Argo only dataset produced by optimal interpolation”, which includes salinity and temperature data on a $1^\circ \times 1^\circ$ horizontal grid and at 58 vertical levels. These temperature and salinity data are collected by Argo floats deployed into the world oceans, which drift for a number of years making measurements of temperature and salinity profiles of the ocean at 10 day cycles. These profiles are then quality controlled, interpolated onto the regular grid and archived in monthly mean files for scientific use. In this study, we use the profiles averaged from January 2004 to December 2011 (Roemmich and Gilson, 2009). Besides the Argo data, the climatological average of the World Ocean Atlas 2009 (called WOA09) temperature (Locarnini et al., 2009) and salinity (Antonov et al., 2009) data are also used for comparison.

The in situ measurements of ocean currents by Acoustic Doppler Current Profiler (ADCP) current meters deployed at the 10 m depth at the TAO/TRITON sites of ($8^\circ\ N\ , 137^\circ\ E$) and of ($8^\circ\ N\ , 156^\circ\ E$), respectively, are used to evaluate the accuracy of the AGC calculated by the P-vector method. Assuming a vertical viscosity of 0.012 m$^2$ s$^{-1}$, which corresponds to a surface Ekman layer thickness of 91 meter, the Ekman spiral is calculated, using the National Center for Environmental Prediction (NCEP) wind data and the drag coefficient of Large and Pond (1981), and is subtracted from
the current meter time series at 10 m. Because the Ekman layer thickness is much
larger than 10 m, the surface Ekman velocity is not sensitive to the choice of the
vertical viscosity coefficient. The hourly current meter time series are averaged into
monthly mean data to be compared with the altimeter and the P-vector geostrophic
currents.

In addition, the surface geostrophic currents based on the merged altimeter sea
level data of the French Archiving, Validation, and Interpolation of Satellite
Oceanographic Data (Aviso) project collected by the Topex/Poseidon, Jason-1, and
European Research Satellites are used to evaluate the P-vector AGC. The altimeter
data are archived on a global grid of $1/3^\circ$ resolution between $82^\circ$ S and $82^\circ$ N at
weekly intervals with tidal and sea level atmospheric pressure corrections
incorporated. The geostrophic currents are calculated from the absolute dynamic
topography, which consists of a mean dynamic topography (Rio and Hernandez, 2004)
and the anomalies of the altimeter sea level, and are averaged into monthly mean data
at $1^\circ$ longitude by $1^\circ$ latitude resolution to be compared with the AGC at the surface.
The Aviso sea level data can be accessed at http://www.aviso.oceanobs.com/.

The monthly climatological data of the Earth Simulator (OFES) (Masumoto et al.,
2004) averaged from the last 10-year simulation of a 50-year model spin-up forced by
the climatological NCEP/NCAR reanalysis data are used to examine the Sverdrup
theory as well. The OFES model domain covers the global ocean from $75^\circ$ S to $75^\circ$
N, with a horizontal resolution of $0.1^\circ$ longitude $\times 0.1^\circ$ latitude and a stretched
vertical coordinates of 54 levels from the sea surface (2.5 m) to a maximum depth of
6065 m.

The NCEP/NCAR reanalysis wind velocity products and the EAR-40 wind velocity products of the European Center for Medium-range Weather Forecast are used to calculate the Sverdrup transport in this study. The wind-stress is calculated based on the quadratic drag law using different drag coefficients of Garratt (1977), Large and Pond (1981), and Foreman and Emeis (2010), respectively, to test the sensitivity of the Sverdrup transport in this paper.

2.2 Absolute geostrophic currents

The AGC in this study are calculated from gridded temperature and salinity data using the P-Vector method (Appendix B), which is based on the conservation of potential density and potential vorticity under three approximations: the geostrophic balance, adiabatic flow, and the Boussinesq approximation (Chu, 1995, 2006). The intersection of isopycnal and equal-potential-vorticity surfaces determines the direction of the geostrophic currents, which is called the P-vector. The thermal wind vector between any two levels can be used to calculate the magnitudes of the geostrophic currents at the two levels. In practice, the geostrophic currents are determined by least-squares fitting to the data at multiple levels (Chu, 1995, 2000; Chu et al., 1998, 2001).

In the past (e.g. Chu, 1995), the P-vector geostrophic currents were computed by applying the least-squares fitting in the whole water column of the ocean. Since the conservation of density and potential vorticity is generally not accurate in the upper mixed layer of the ocean, we chose to construct the geostrophic currents only in the
intermediate layers (Zhang, 2011; Zhang et al., 2013), i.e. the P-vector method is used to calculate the geostrophic currents between 800 dbar and 2000 dbar. The geostrophic currents above the 800 dbar are determined by the dynamic height calculation, using the geostrophic currents at 800 dbar as the reference velocity. Experiments have shown that the AGC are not sensitive to the choices of the depth ranges mentioned above so long as the P-vector calculation is conducted significantly below the surface mixed layer.

The errors of the geostrophic currents are estimated based on the thermal wind relation and the standard deviation of the density, assuming that the largest errors of the currents occur when density errors are out of phase at the adjacent grid points. In addition, the errors of the mean circulation are augmented by the standard deviation of the P-vector geostrophic currents so that the time-dependency of the ocean circulation is included in the error estimate. Errors of the meridional transports at each grid are obtained by vertical integrations from 1900 m to the sea surface.

2.3 The Sverdrup balance

The Sverdrup theory assumes linear, geostrophic dynamics and the existence of a maximum depth H in the ocean, beyond which the horizontal and vertical velocity vanish (Sverdrup, 1947). The vertically integrated momentum and continuity are governed by the following equations.

\begin{align}
\frac{\partial P}{\partial x} &= -fV + \frac{\tau^x}{\rho} \\
\frac{\partial P}{\partial y} &= fU + \frac{\tau^y}{\rho}
\end{align}

(1a)

(1b)
where the notations are conventional and \((U, V, P)\) stand for vertically integrated horizontal velocities and pressure from \(-H\) to 0. Cross-differentiating the first two equations and using of the third, the Sverdrup relation is obtained as the following.

\[
P = \frac{\nabla \rho \times \nabla q}{|\nabla \rho \times \nabla q|} \tag{2}
\]

where \(\rho\) is the water density, \((\tau^x, \tau^y)\) the wind stress components, and \(\beta = df/dy\) the meridional gradient of the Coriolis parameter. Sverdrup (1947) assumes that the vertical velocity \(w\) vanishes at the depth \(z = -H\). In this paper, \(H\) is chosen to be the maximum depth of the Argo float measurements of 1900 m, or at some specified isopycnals (see Section 4.3). The vertically integrated meridional transport in Equation (2) includes the surface Ekman transport and the geostrophic transport.

The integration of the meridional geostrophic transport along a latitude \((y)\) from the eastern boundary \((x_E)\) to a longitude \((x)\) is calculated by

\[
\int_{x_E}^{x} \int_{0}^{-H} V dxdz = \frac{1}{\beta \rho} \int_{x_E}^{x} \left( \frac{\partial \tau^x}{\partial y} - \frac{\partial \tau^y}{\partial x} \right) dx + \frac{1}{f \rho} \int_{x_E}^{x} \tau^y dx \tag{3}
\]

Both sides of (3) vary with \((x, y)\). The left side is called the geostrophic meridional transport (calculated from ocean hydrographic data). The right side is the meridional Sverdrup-minus-Ekman transport (S-E transport in abbreviation) computed from surface wind stress. The traditional Sverdrup transport is the first term on the right side of the Equation (3).

3. Results

3.1 Validation of the P-vector AGC
Figure 1 shows the averaged surface dynamic topography with a spatial average removed and the averaged surface AGC for the period of 2004-2011 in the North Pacific Ocean. The geostrophic currents at the sea surface are characterized by NEC in the latitudinal range of 7° N-20° N, NECC south of 7° N, STCC along roughly 25° N, and the Kuroshio extension east of Japan, which are consistent with the historical understanding of the alternating zonal currents in the area. Due to the coarse resolution of the gridded data, the western boundary currents near the coasts of the Philippines and in the East China Sea are not resolved well by the calculation.

Comparisons of the P-vector AGC at the sea surface with the altimeter geostrophic currents and with the TRITON moored current meter data at 10 m at (8° N, 137° E) and at (8° N, 156° E) are shown in Figure 2. The current meter time series with and without the Ekman velocity are both shown, the difference of which is small due to the weak easterlies at this latitude. The agreement among the P-vector calculation, altimeter geostrophy, and the direct current measurement is reasonable. At (8° N, 137° E), the root-mean-square (RMS) difference of zonal currents between the AGC and current meter is 7.32 cm s⁻¹, larger than that of 5.75 cm s⁻¹ between the altimeter and the current meter. Also the correlation coefficient of zonal currents between the AGC and the current meter is 0.38, smaller than that of 0.755 between the altimeter and the current meter, but both correlations are close to or above the 95% significant level (Table 1). The significant level of the correlation in this paper is based on the decorrelation time scales estimated using the method of Davis (1976) and Kessler et al. (1996). In contrast, the meridional component of AGC is in better agreement with the
current meter data than are the altimetric geostrophic currents. The latter has
systematically underestimated the meridional transport of the ocean during the
observational period. The correlation coefficient of meridional currents between the
AGC and the current meter is 0.59, in comparison with the correlation coefficient of
0.559 between the altimeter and the current meter, both of which are above the 95%
significant level. The RMS difference of the AGC meridional currents from the
current meter data is 4.05 cm s\(^{-1}\), smaller than that of 4.55 cm s\(^{-1}\) between the altimeter
meridional currents and the current meter data.

At (8° N, 156° E), the RMS difference of zonal currents between the AGC and
current meter is 5.27 cm s\(^{-1}\), smaller than that of 6.49 cm s\(^{-1}\) between the altimeter and
the current meter. The correlation coefficient of zonal currents between the AGC
and the current meter is 0.66, above the 95% significant level and comparable to that
of 0.79, also above the 95% significant level, between the altimeter and the current
meter. The systematic difference between the altimeter and the AGC meridional
velocity is still glaringly large. During the winter of 2010-2011, the Argo meridional
currents show out-of-phase variations from the current meter and altimeter data,
which are induced by what appears to be a displaced stationary eddy in the vicinity of
8° N, 156° E in the Argo data (Figure 3). It is possible that the discrepancy is due
to the gridding and interpolation of the sparse Argo data in the area. Otherwise, the
limited current meter data seems to suggest that the AGC is a reasonable estimate of
the local geostrophic currents. The correlation coefficient of meridional currents
between the AGC and the current meter is 0.28, in comparison with the correlation
coefficient of 0.35 between the altimeter and the current meter, both of which are above the 95% significant level. The low correlations are partly attributed to the short and intermittent current meter time series at this location. The RMS difference of the AGC meridional currents from the current meter data is 3.17 cm s\(^{-1}\), much smaller than that of the altimeter meridional currents from the current meter data of 5.21 cm s\(^{-1}\).

The distribution of the RMS differences of surface geostrophic currents between the AGC and the altimeter in the North Pacific Ocean is shown in Figure 4. Over the majority of the North Pacific Ocean, the AGC are within a few centimeters per second from the altimeter geostrophic currents, except in the Kuroshio extension area, where meso-scale eddies are active. The RMS differences are comparable with the estimated errors of the altimeter geostrophic currents (Rio and Hernandez, 2004), suggesting the validity of the AGC in studying the three-dimensional transport of the ocean. In particular, the RMS difference of the meridional geostrophic currents is small over the entire tropical North Pacific Ocean, suggesting that the data can be used to assess the accuracy of the Sverdrup theory.

The correlation coefficients of geostrophic current variations between the AGC and the altimeter are mostly between 0.3 and 0.6 for the meridional velocity, above the 95% significance level (Figure 4). The correlations for zonal velocity are mostly between 0.3–0.9 in the western and tropical Pacific Ocean, also above the 95% significance level. Those in the eastern mid-latitude Pacific are a little small and not as significant, mostly due to the sparse data coverage and longer decorrelation time.
scales, but that area is not the focus of this study. The above comparisons give us confidence that the P-vector AGC can be used to evaluate the Sverdrup balance and to study the three-dimensional transports of the North Pacific Ocean circulation.

3.2 Comparison of the AGC meridional transport with the Sverdrup theory

The Sverdrup balance is examined using the AGC data. Since the Sverdrup theory is for steady-state circulation, the errors of the steady state estimation based on the 8-year data are first examined using the altimeter geostrophic currents. The upper and lower panels of Figure 5 show the time series of the zonally averaged meridional velocity between 120°E and 180° along 8°N and 18°N, respectively, calculated from the altimeter sea level. The time means of 1993-2011 and of 2004-2011 are plotted in thick solid and dash lines for comparison. The 8-year averages are close to the 20-year averages compared with the standard deviation of the time series. But, while the short-term mean at 8°N is within 92% confidence interval of the long-term mean, at 18°N, the short-term mean is merely within 74% confidence interval of the long-term mean. Hence the 8-year average at 8°N is a good approximation of the long term mean circulation and can be used to validate the Sverdrup theory, while at 18°N the 8-year mean still contains sizable temporal variations.

The mean geostrophic meridional transport (Figure 6a) and the S-E meridional transport (Figure 6b) in the North Pacific for the period of 2004-2011 are calculated based on the left and right sides of (3). Here, the geostrophic meridional transport is integrated from 1900 m to the sea surface and from the eastern boundary of the North
Pacific Ocean. The NCEP surface wind stress data based on the Garratt (1977) drag coefficient and averaged over 2004-2011 were used to compute the S-E transport. Subtracting the S-E meridional transport from the geostrophic meridional transport yields the meridional transport discrepancy shown in Figure 5c, which is the discrepancy from the Sverdrup theory. The S-E transport is generally in good agreement with the geostrophic meridional transport in the eastern subtropical North Pacific Ocean and in the area along roughly 15° N and 21-27° N west of the dateline.

The agreement of the S-E transport with the geostrophic meridional transport in the area between 21°N and 27°N is consistent with Hautala [1994]'s analysis along 24° N. The S-E transport failed to explain the geostrophic meridional transport in the recirculation regions of the Kuroshio and its extension, perhaps owing to the deficiency of the linear approximation of the Sverdrup theory. Significant differences between the geostrophic meridional transport and the S-E transport are also found in the region of [6°-14°N, 130°E-150°W] between NEC and NECC, and in the region of [15°-20°N, 120°E-150°W] between NEC and STCC as shown in Figure 5c. The maximum differences are larger than 20 Sv (1 Sv = 10^6 m^3 s^-1), suggesting that the deviation of the circulation dynamics from the Sverdrup theory in these two areas is significant. In particular, the S-E transport and the geostrophic meridional transport have opposite signs in the region between 6°N and 12°N in the western North Pacific Ocean, showing that the geostrophic currents there are not governed by the wind curl forcing at all.

3.3 Three-dimensional structure of NEC
The three-dimensional structure of the mean NEC in the North Pacific Ocean is examined using the P-vector AGC. Figure 7 shows the distribution of the mean zonal velocity in different meridional sections across the North Pacific Ocean. The signals smaller than the error bars are indicated by a dot at each point in the plots. The NEC transport is shown to increase toward the western Pacific Ocean (Table 2). The axis (maximum zonal velocity) of the NEC is shown to shift to the north at depth. The 2 cm s\(^{-1}\) zonal velocity contours have shifted beneath the STCC and far north beyond at the 600 m depth, most of which are larger than the error bars. The maximum mean zonal velocity of the NEC exceeds 28 cm s\(^{-1}\) in the western Pacific Ocean.

South of the NEC at the surface, the NECC flows eastward above the 200 m depth. Underneath the NECC is a weak westward flow originating from the eastern Pacific Ocean with increasing zonal velocity toward the western Pacific Ocean. This current has not been named or discussed in the published literature. We have examined historical data and found that this current was indeed indicated by direct ADCP measurements at 5\(^\circ\)N, 142\(^\circ\)E on December 22, 2001 and at 5\(^\circ\)N, 137\(^\circ\)E on October 27, 2002 archived by Japan Oceanographic Data Center (Figure 8). This current is evident in almost all of the monthly mean AGC maps with a velocity core separate from the NEC. Historically, mean current maps are difficult to obtain in the low-latitudes due to contamination by the strong intraseasonal signals. The basin-scale, high-frequency, and synchronous mapping of the Pacific Ocean by Argo profiling floats allow us to determine the mean currents with some confidence. We
shall call this current the North Equatorial Subsurface Current (NESC) in the following text. The maximum mean velocity of the NESC exceeds 2 cm/s in the western North Pacific Ocean according to the AGC calculation and the ADCP measurements. The westward transport of the NESC between 3° N and 7° N and at the depth range of 200 m to 600 m at 140° E is as large as 4.2 Sv (Table 2). However, both the NESC and the western Pacific Part of the NECC are within the error bars, due to the strong eddy activities there.

The northwestern Pacific Ocean has rich subsurface undercurrents, none of which have dynamical explanations so far. Recently, Cravette et al. (2013) have observed alternating zonal currents at the intermediate depths of 1000 m and 1500 m. Qiu et al. (2013) suggested that these currents are generated by nonlinear triad instability of the Rossby waves radiated from the eastern boundary. The theory didn’t explain stronger zonal currents in the western Pacific Ocean than in the east. Further studies are needed to understand the dynamics.

North of the NEC at the surface is a weak eastward current called the STCC. Further north of it is the Kuroshio recirculation flowing to the west. These currents are within the error bars due to the strong eddy activity. The strong Kuroshio and its extension, although below the measurement errors, are easily identified in the sectional distribution of the zonal currents north of about 30° N. The Kuroshio extension east of 160° E has double cores in the mean field, perhaps owing to an effect of the Shatsky Rise, located at about 158° E.

Directly underneath the NEC, the mean flow is extremely weak and within the
The P-vector calculation has not identified any mean eastward flowing undercurrent core beneath NEC east of 145° E across the Pacific basin.

A comparison of the Argo AGC and the WOA09 AGC suggests that the zonal patterns of the two sets of AGC are similar overall (Figure 9). The WOA fields have been smoothed in space to eliminate sub-gridscale structure. The AGC based on the WOA09 hydrography data have under-estimated the strength of the NESC beneath the NECC in the subsurface northwestern Pacific Ocean significantly. The difference explains why the NESC was not discussed by previous studies using synoptic surveys of the regional hydrography. In comparison, the Argo profiles show the existence of the NESC clearly. The NESC weakens in the Philippine Sea as it approaches the western boundary and eventually merges with the New Guinea Coastal Undercurrent at about 128° E according to the Argo AGC (figure omitted).

The strength of the NEC and the NECC increases westward in the Philippine Sea (Figure 9), the magnitudes of which in the northwestern Pacific Ocean are also significantly under-estimated by WOA09 data due to the smoothing. Two eastward undercurrent cores are identified underneath the westward flowing NEC in the Philippine Sea in the Argo AGC, which are called the North Equatorial Undercurrent by Hu et al. (1991) and by Qiu et al. (2013). The Argo data suggest that the North Equatorial Undercurrent is confined in the Philippine Sea west of about 140° E. None of the undercurrent cores are identified in the AGC inverted from the WOA09 data.

The combination of equations (2) and (3) yields the zonal wind-driven Sverdrup
transport. Following the convention used in this paper, the vertically integrated zonal transport is calculated from the AGC and is compared with the S-E zonal transport based on the Sverdrup theory (Figure 10). The upper and middle panels in Figure 10 show that the AGC and the Sverdrup zonal transports are significantly different from each other. The difference of the AGC from the S-E zonal transport are primarily in the same zonal bands as are the largest meridional transport differences in the tropical northwestern Pacific Ocean in Figure 5c. The comparisons suggest that the difference of the AGC transports from the Sverdrup theory is structural, which indicates a deficiency of the classical Sverdrup theory in estimating the zonal and meridional transports of the tropical North Pacific Ocean circulation.

4. Discussions

4.1 Error estimates

We have estimated the errors of the geostrophic currents based on the standard deviation of the density fields in combination with the thermal wind relation. The error of the least-square fitting between 800 m and 1900 m is believed to be smaller than this estimate. Thus our error estimate represents the upper bound of the geostrophic current errors induced by the density errors. In addition, the standard deviations of the AGC time series are used to estimate the errors induced by the time dependency of the currents. One standard deviation is taken into consideration of the error estimate. Figure 11 shows the errors of the surface zonal and meridional AGC and of the meridional transport of the AGC based on the above estimates. The
error bars of the AGC are comparable with those of the altimeter geostrophic currents
of a few centimeters per second according to Rio and Hernandez (2004). The error
bars for the meridional transport are generally smaller than 2 Sv, except near the
western boundary and in the Kuroshio extension. The small magnitudes of the error
bars in comparison with the structural differences from the Sverdrup relation suggest
that the identified differences of the AGC meridional transports from the S-E
meridional transport are statistically significant.

4.2 Geostrophic meridional transport discrepancy based on the WOA data

The above analyses suggest that the geostrophic transports in the upper ocean
differ significantly from the simple linear Sverdrup relation in some areas of the
tropical North Pacific. It is important to know if these differences are specific for
the Argo profile data. For this purpose, we compare the meridional transport of the
WOA09 AGC with the S-E transport in Figure 12. The calculation of the transports
in Figure 12 is exactly the same as in Figure 6, except using the WOA09 data. The
meridional transport of the geostrophic currents and its deviation from the S-E
transport is almost identical to those based on the Argo profiles, suggesting that the
deviation of the geostrophic transport from the linear Sverdrup theory is robust.

4.3 Independence of the meridional transport discrepancy on depth \( H \)

The Sverdrup relation (2) is established with the assumption of zero vertical
velocity at the depth \( z = -H \). We have conducted sensitivity tests and found that the
meridional transport discrepancy in Figure 5c is not sensitive to the selection of \( H \)
below, say, 1500 m (figures omitted). Modern discussions of the Sverdrup theory
have suggested that the assumption of zero vertical velocity at a constant depth $H$ is one of the primary restrictions of the theory (Marchuk and Sarkisyan, 1973; Wunsch and Roemmich, 1985). However, the vertical integration of the continuity equation over a varying depth $z=-H(x,y)$ will result in extra terms in equation (3), which will destroy the elegance of the Sverdrup relation in equation (2). An exception is if $z=-H(x,y)$ is a isopycnal surface. However, the cross-differentiation of the momentum equations will generate a Joint Baroclinicity and Relief (JBAR) term in the Sverdrup relation (Marchuk et al, 1973). Assuming the JBAR term is small due to the flat isopycnals in the abyssal tropical oceans (e.g. Figures 7 and 9), one can test the accuracy of the Sverdrup relation based on vertical integrations above different isopycnal surfaces.

The meridional transport differences between the left and right sides of (3) based on different values of $H$ corresponding to $\sigma_0$ levels of 26.5, 27.0, 27.2, and 27.5 for the left side of (3) are shown in Figure 13. The overall meridional transport discrepancies from the Sverdrup theory for different $\sigma_0$ levels are similar to those in Figure 6c and much larger than the error bar estimates in the areas of 6°-12°N and 15°-20°N, suggesting their significance and independence of the bottom limit of the vertical integration.

4.4. Independence of meridional transport discrepancy on surface wind data

To further understand the reason for the large differences of the geostrophic meridional transport from the Sverdrup linear theory in the areas of 6°-12°N and 15°-20°N of the western North Pacific, averaged ERA-40 surface wind stress data for
the period of 1961-2000 and averaged NCEP/NCAR reanalysis surface wind stress
data for the period of 1948-2011 are used to examine the sensitivity of the meridional
transport discrepancy on the surface wind products. These surface wind stress
products are calculated using the drag coefficient of *Large and Pond* (1981).

The differences between the left and the right sides of (3) for different wind
products are shown in Figures 14a and 14b. For these experiments, the lower limit
of the vertical integration of the geostrophic meridional transport is set at \( \sigma_0 = 27.2 \).
The spatial patterns of the deviation from the Sverdrup theory in Figures 13a and 13b
for the different wind products are similar to those in Figure 6c, with the maximum
differences larger than 20 Sv, suggesting that the deviation from the Sverdrup theory
in these two areas is robust. These results also suggest that the wind stress errors
cannot account for all of the meridional transport discrepancies from the Sverdrup
theory.

**4.5. Independence of meridional transport discrepancy on drag coefficients**

The Large and Pond formula suggests that the drag coefficient used to calculate
the wind stress increases with wind speed. Recent studies indicate that the drag
coefficient in the marine atmospheric boundary layer increases with the wind speed
for moderate winds, and levels out at high wind speed (Foreman and Emeis, 2010).
The use of the Foreman and Emeis drag coefficients has resulted in a similar pattern
of the geostrophic meridional transport discrepancies from the S-E transport as in
Figure 6c (Figures 14d). The magnitudes of the meridional transport discrepancies
using the Foreman and Emeis drag coefficients in Figure 14d are about 5 Sv larger
than those using the Large and Pond drag coefficients shown in Figure 14c, suggesting that the uncertainty of the drag coefficients is not the main reason of the geostrophic meridional transport discrepancies from the S-E transport.

4.6. The Sverdrup balance in a high-resolution ocean model

The high-resolution OFES model provides an opportunity to investigate the origin of the meridional transport discrepancy in the interior tropical and subtropical Northwestern Pacific Ocean. Here, the climatological annual mean NCEP wind stress used to drive the model is used to calculate the S-E transport. The geostrophic meridional transport is calculated from the OFES climatological annual mean temperature and salinity simulations using the P-vector method. The lower bound of the vertical integration $H$ is chosen to be 1900 m. Experiments using different isopycnic surfaces (such as $26.5\sigma_0$, $27\sigma_0$, $27.2\sigma_0$ and $27.5\sigma_0$) as the lower bound of the vertical integration of the geostrophic meridional transport calculation show essentially the same results (not shown). Figure 15a shows the significant meridional transport discrepancy in the latitudinal bands of 6°-12°N and 12°-20°N, the magnitude and area coverage being essentially the same as those based on the Argo data (as shown in Figure 6c). The maximum differences are ~5 Sv in 6°-12°N and more than 10 Sv in 12°-20°N. Since the OFES simulation is dynamically consistent with the wind forcing, it should satisfy the Sverdrup balance exactly should the dynamics be linear. The above comparison suggests strongly that the significant meridional transport discrepancy is due to nonlinear effects within the OFES model.

An alternative method of calculating the ocean meridional transport is to use the
OFES simulated velocity, which includes geostrophic and ageostrophic components. The Sverdrup transport, i.e. the first term on the right side of equation (3), is subtracted from the simulated meridional transport. The difference is shown in Figure 15b. The meridional transport discrepancy of the model is essentially the same as that of the AGC (Figure 15a) in the interior tropical northwestern Pacific Ocean, which suggests that the Sverdrup relation is not a valid approximation of the leading order general ocean circulation in the North Pacific Ocean.

Diagnosis of coarse-resolution ocean model simulation suggests that the simulated North Pacific Ocean circulation is generally in good agreement with the Sverdrup relation (e.g. Jiang et al., 2006). Considering that the OFES model simulation is in strict dynamic consistency with the NECP winds, the difference between the coarse- and fine-resolution models is suggested to be the reason of the discrepancy from the Sverdrup relation. Ocean nonlinearity is a natural candidate of the cause of the discrepancy.

Finally, as a demonstration of the insensitivity of the meridional transport discrepancy to the choice of $H$, we have calculated the depth of vanishing vertical velocity ($|w| < 10^{-9}$ cm s$^{-1}$) at each grid point in the OFES simulation. The function of $z=-H(x,y)$ over the tropical North Pacific is shown in Figure 14c. The meridional transport discrepancy based on this lower bound of vertical integration is shown in Figure 15d. Again, the same structure as in panels a and b re-emerges. This insensitivity to the choice of the vertical bound of integration suggests that the discovered discrepancy from the Sverdrup balance in the northwestern tropical Pacific
Ocean is robust.

5. Conclusions

In this study, absolute geostrophic currents in the North Pacific Ocean are calculated based on the gridded Argo profiling float data for the period of January 2004 to December 2011 using the P-vector method. The mean meridional transport of the geostrophic currents is compared with the Sverdrup transport in the North Pacific Ocean to assess the accuracy of the Sverdrup theory. The results have shown large differences from the Sverdrup balance in the regions of 6°N-12°N, 15°N-20°N, and in the recirculation and extension areas of the Kuroshio in the Northwestern Pacific Ocean. Analyses suggest that the deviation from the Sverdrup balance is statistically significant and is supported by both the Argo and the WOA09 data. The pattern of the discrepancy from the Sverdrup balance is independent of the depth of vertical integration and wind stress products used to estimate the Sverdrup transport. These facts suggest that the difference of the geostrophic meridional transport from the Sverdrup balance is robust and structural and that wind stress errors are not the main reason of the large differences. A comparison of the geostrophic meridional transport and the Sverdrup transport in the high-resolution OFES simulation shows a similar difference. Considering that the wind stress forcing is in strict dynamic consistency with the ocean circulation in the OFES model, it is speculated that nonlinear effects of the ocean circulation are the reason of the differences. The results suggest that the linear dynamics of the Sverdrup theory are deficient in explaining the meridional transport in the tropical northwestern Pacific Ocean.
The basin-scale synchronized profiling of the North Pacific Ocean temperature and salinity has been realized for the first time in history by the Argo project. The geostrophic currents based on these Argo profiles suggest stronger mean surface zonal currents of the NEC and the NECC than those based on the traditional ship-based WOA09 data. In particular, a subsurface westward current underneath the NECC is found in the Argo geostrophic currents, which is absent in the WOA09 data. The current originates east of the dateline and gets intensified toward the western North Pacific Ocean. The average core of this subsurface current is located at the depth range of 200 m-500 m and between 4° N and 6° N, with the maximum amplitude of the westward mean zonal current exceeding 2 cm s⁻¹ and a mean zonal transport of more than 4.2 Sv to the west between 3° N and 7° N. We name this current the North Equatorial Subsurface Current.

**Acknowledgments.**

This work is supported by National Basic Research Program of China (project 2012CB956000), by the NSFC Grants 40888001 and 41176019, by CMA Grant GYHY201306018, and by CAS Strategic Priority Project XDA11010301. Z. Zhang is supported by KLOCAW Grant 1208 and by Zhoushan Municipal Science and Technology Project (2011C32015, 2012C33033). P. C. Chu was supported by the Naval Oceanographic Office. WK Dewar is supported by NSF grants OCE-0960500 and OCE-1049131.
Appendix A: Consistency of P-Vector method with the $\beta$ spiral method

The P-vector method assumes steady state circulation along isopycnal surfaces, i.e.

\[
\frac{d\rho}{dt} = u \frac{\partial \rho}{\partial x} + v \frac{\partial \rho}{\partial y} + w \frac{\partial \rho}{\partial z} = 0. \tag{A1}
\]

Taking the Boussinesq and geostrophic approximations, the thermal wind relation suggests

\[
\frac{\partial u}{\partial z} \frac{\partial \rho}{\partial x} + \frac{\partial v}{\partial z} \frac{\partial \rho}{\partial y} = 0. \tag{A2}
\]

The conservation of potential vorticity gives

\[
\begin{aligned}
\frac{d}{dt} \left( \frac{f}{\rho} \frac{\partial \rho}{\partial z} \right) &= \frac{1}{\rho} \frac{\partial \rho}{\partial z} \beta v + \frac{f}{\rho} \left( \frac{\partial^2 \rho}{\partial x \partial z} + \frac{\partial^2 \rho}{\partial y \partial z} + \frac{\partial^2 \rho}{\partial z \partial z} \right) \\
&= \frac{1}{\rho} \frac{\partial \rho}{\partial z} \beta v + \frac{f}{\rho} \left[ \frac{\partial}{\partial z} \left( \frac{\partial \rho}{\partial x} + \frac{\partial \rho}{\partial y} + \frac{\partial \rho}{\partial z} \right) - \frac{\partial u}{\partial z} \frac{\partial \rho}{\partial x} - \frac{\partial v}{\partial z} \frac{\partial \rho}{\partial y} - \frac{\partial w}{\partial z} \frac{\partial \rho}{\partial z} \right] \\
&= \frac{1}{\rho} \frac{\partial \rho}{\partial z} \beta v + \frac{f}{\rho} \frac{\partial w}{\partial z} \frac{\partial \rho}{\partial z} = 0
\end{aligned}
\]

Thus in a stratified ocean, $\beta v = f \frac{\partial w}{\partial z}$, which is the vorticity balance of the $\beta$ spiral method.
Appendix B: The principles of the P-vector method

The P-Vector method is based on the conservation of potential density and potential vorticity (Equations B1 and B2) under two approximations: the geostrophic balance and the Boussinesq approximation. The thermal wind vector (Equations B4 and B5) between any two levels, \( z_k \) and \( z_m \), can be used to calculate the magnitudes of the geostrophic currents at the two levels.

\[
\begin{align*}
\mathbf{v} \cdot \nabla \rho &= 0, \\
\mathbf{v} \cdot \nabla q &= 0, \\
q &= \frac{f}{\rho} \frac{\partial \rho}{\partial z}, \\
\Delta u &= \frac{g}{f \rho_0} \int_{z_k}^{z_m} \frac{\partial \rho}{\partial y} dz, \\
\Delta v &= -\frac{g}{f \rho_0} \int_{z_k}^{z_m} \frac{\partial \rho}{\partial x} dz.
\end{align*}
\]

Two necessary conditions must be satisfied for validity of the P-vector method:

1. The \( \rho \) surface is not parallel to the \( q \) surface

\[
\nabla \rho \times \nabla q \neq 0, \quad (B6)
\]

2. The velocity( \( u, v \) ) should execute a \( \beta \) spiral at any two levels (\( z = z_k \), and \( z = z_m \))

\[
\begin{bmatrix}
    u^{(k)} \\
    v^{(k)} \\
    u^{(m)} \\
    v^{(m)}
\end{bmatrix} \neq 0. \quad (B7)
\]

Thus, a unit vector \( \mathbf{P} \) can be defined as

\[
P = \frac{\nabla \rho \times \nabla q}{|\nabla \rho \times \nabla q|} \quad (B8)
\]

which is the intersection of isopycnal and equal-potential-vorticity surfaces. The geostrophic currents, \( \mathbf{V} = (u, v, w) \), are assumed to follow the unit vector \( \mathbf{P} \), i.e.

\[
\mathbf{V} = \mathbf{r}(x, y, z) \mathbf{P} \quad (B9)
\]

where \( \mathbf{r} \) is the proportionality coefficients. The thermal wind relation at two different depths \( z_k \) and \( z_m \), suggests a set of algebraic equations to determine the coefficients \( \mathbf{r} \).
\[ r_{(k)}^{(k)} p_{x}^{(k)} - r_{(m)}^{(m)} p_{x}^{(m)} = \Delta u_{km} \quad \text{(B10)} \]
\[ r_{(k)}^{(k)} p_{y}^{(k)} - r_{(m)}^{(m)} p_{y}^{(m)} = \Delta v_{km} \]

The determinant of (B10)

\[
\begin{vmatrix}
  p_{x}^{k} & p_{x}^{m} \\
  p_{y}^{k} & p_{y}^{m}
\end{vmatrix} = \sin(\alpha_{km}), \quad \text{(B11)}
\]

where \( a_{km} \) is the \( \beta \) spiral turning angle between \( \mathbf{P}^{(k)} \) and \( \mathbf{P}^{(m)} \) at the two levels \( z_{k} \) and \( z_{m} \). In practice, the least squares fitting is used to determine the coefficients \( r \) using data of multiple levels.
References


Qiu, B., S. Chen, and H. Sasaki, (2013), Generation of the North Equatorial Undercurrent jets by triad baroclinic Rossby wave interactions, (accepted by JPO).


Wyrtki, K. (1961), Physical oceanography of the southeast Asian waters, NAGA Report 2, Scripps Inst. of Oceanogr., University of California, San Diego, La Jolla, CA,


Figure captions:

Figure 1. 2004-2011 mean Argo dynamic height (contours) relative to 1500 m depth with the spatial average removed and surface AGC (vectors). Unit of contours is cm. Unit of vectors is cm s⁻¹.

Figure 2. Comparison of Argo surface AGC (thick curve) with the direct current meter measurement of the TRITON array at 10 m depth (thin curve) and with geostrophic currents based on the absolute dynamic topography of satellite altimetry at the same location (asterisk curve). The upper two panels are at (8°N, 137°E) and the lower two panels are at (8°N, 156°E). The current meter velocity components with the surface Ekman velocity removed are shown in dash curves for comparison.

Figure 3. Comparison of the surface dynamic height based on Argo profiles in reference to the 1900 m (a) with the sea surface height of satellite altimeter (b) in December 2012, showing a dislocated eddy in the vicinity of the mooring site (marked with a dot) in the former. Spatial averages are removed in the plots. Unit is cm.

Figure 4. Root-mean-square differences and correlation coefficients between the Argo AGC and altimeter geostrophic currents at the surface of the North Pacific Ocean during 2004-2011. Upper panels are for the root-mean-square differences. Contour interval is 2 cm s⁻¹. Lower panels are for the correlation coefficients, with correlations below 95% significance level not shown. Contour interval is 0.2.
Figure 5. Time series of area averaged meridional geostrophic currents based on the satellite altimeter data between 120°E and the dateline along 8°N and 18°N. The short dash and long solid lines represent the mean of 2004-2011 and of 1993-2011, respectively. The standard deviations of 1993-2011 and the differences between the long-term and short-term means are marked with the bars.

Figure 6. Comparison of 2004-2011 mean meridional transport based on Argo AGC (a) and the S-E transport based on the annual mean (2004–2011) NCEP surface wind data (b) in the North Pacific Ocean. Their difference is shown in (c). Unit is 10^6 m^3 s^{-1}. The wind stress calculation is based on the Garratt drag coefficient. The gray shade is negative.

Figure 7. 2004-2011 mean zonal geostrophic velocity (thin contours) from Argo data in different meridional sections across the North Pacific Ocean. Unit is cm s^{-1}. Thick contours mark the potential density surfaces. The gray shading indicates westward currents. The dots at the grid points stand for signals below the error bars.

Figure 8. In-situ zonal velocity profiles measured by ADCP current meters at 5°N, 142°E on Dec 22, 2001 (a) and at 5°N, 137°E on Oct 27, 2002 (b).

Figure 9. Comparison of the AGC based on the Argo (upper three panels) and the WOA09 (lower three panels) hydrography data at different meridional sections across the tropical Northwestern Pacific Ocean. Contour interval is 2 cm s^{-1}. Thick contours mark the potential density surfaces. The gray shading indicates
westward currents.

Figure 10. Mean vertically integrated zonal transport of the AGC and of the Sverdrup theory (upper two panels), and their difference (lower panel). Unit is $m^2 \, s^{-1}$.

The gray shading indicates westward currents.

Figure 11. Error bars (see text) of the surface zonal (a) and meridional (b) geostrophic currents based on Argo data and of the grid meridional transport of the geostrophic currents smoothed by a $5^\circ$ longitude by $5^\circ$ latitude average filter (c). Contour intervals are $2 \, cm \, s^{-1}$ for the surface AGC and $0.5 \, Sv$ for the meridional transport. The gray shading indicates less than $4 \, cm \, s^{-1}$ for a and b, and $1.5 \, Sv$ for c.

Figure 12. Same as Figure 5, except that the AGC meridional transport is based on the WOA09 data. Unit is $10^6 \, m^3 \, s^{-1}$. The bottom limit of the vertical integration is $1900 \, m$. The gray shade is negative.

Figure 13. Dependence of the meridional transport discrepancy on the choice of the isopycnal surfaces of (a)$26.5 \sigma_0$, (b)$27 \sigma_0$, (c)$27.2 \sigma_0$, and (d)$27.5 \sigma_0$, respectively, as the bottom limits of the integration. The geostrophic transport is based on the Argo geostrophic currents. The Sverdrup balance is based on the NCEP wind. The gray area is negative. Contour interval is $5 \, Sv$.

Figure 14. Meridional transport discrepancy based on (a) ERA-40 surface mean (1961-2000) wind stress, (b) NCEP surface mean (1948-2011) wind stress, (c) NCEP surface mean (2004-2011) wind vector using the Large-Pond (1981) drag coefficients, and (d) NCEP surface mean (2004-2011) wind vector using the
Foreman-Emeis (2010) drag coefficients. The gray shade is negative.

Figure 15. Comparison of the meridional transport discrepancies of the OFES model from the wind-driven meridional transport based on Sverdrup Balance. (a) The OFES meridional transport is calculated based on the geostrophic currents calculated from the OFES simulated temperature and salinity fields using the P-vector method. (b) The OFES meridional transport is calculated based on the OFES simulated velocity. (c) Distribution of the depth of vanishing vertical velocity in the OFES simulation, Unit is km. (d) Meridional transport discrepancy based on the function of $z=-H(x,y)$ in Panel c. The gray shade is negative.
Table captions

Table 1. Correlation coefficients between Argo geostrophic currents and the currentmeter velocity and between the altimeter geostrophic currents and the currentmeter velocity at 137°E, 8°N and 156°E, 8°N, respectively. The value in the bracket denotes the RMS differences in cm s\(^{-1}\). All correlations are close to or above the 95% confidence level.

Table 2. Transports and standard deviations of major currents at different meridional sections in the North Pacific Ocean. (Unit is Sv)
Table 1. Correlation coefficients between Argo geostrophic currents and the current meter velocity and between the altimeter geostrophic currents and the current meter velocity at 137°E, 8°N and 156°E, 8°N, respectively. The value in the bracket denotes the RMS differences in cm s⁻¹. All correlations are close to or above the 95% confidence level.

<table>
<thead>
<tr>
<th></th>
<th>zonal AGC, currentmeter</th>
<th>Zonal altimeter, currentmeter</th>
<th>Meridional AGC, currentmeter</th>
<th>Meridional altimeter, currentmeter</th>
</tr>
</thead>
<tbody>
<tr>
<td>137°E, 8°N</td>
<td>0.38 (7.32)</td>
<td>0.755 (5.75)</td>
<td>0.59 (4.05)</td>
<td>0.5593 (4.55)</td>
</tr>
<tr>
<td>156°E, 8°N</td>
<td>0.66 (5.27)</td>
<td>0.79 (6.49)</td>
<td>0.28 (3.17)</td>
<td>0.35 (5.21)</td>
</tr>
</tbody>
</table>

The 95% confidential level of correlation coefficient is 0.40 for zonal velocity (with a decorrelation time scale of ~4 months) and 0.28 for meridional velocity (with a decorrelation time scale of ~2 months).

Table 2. Transports and standard deviations of major currents at different meridional sections in the North Pacific Ocean (Unit is Sv)

<table>
<thead>
<tr>
<th>Currents</th>
<th>Area of integration</th>
<th>140°E</th>
<th>145°E</th>
<th>160°E</th>
<th>180°E</th>
<th>170°W</th>
<th>155°W</th>
<th>140°W</th>
</tr>
</thead>
<tbody>
<tr>
<td>NECC 3-7N,0-500m u&gt;0m/s</td>
<td>23.4±8.3</td>
<td>23.4±7.7</td>
<td>17.7±6.3</td>
<td>21.4±8.6</td>
<td>20.4±8.3</td>
<td>23.9±9.0</td>
<td>23.4±7.2</td>
<td></td>
</tr>
<tr>
<td>NESC 3-7N,200-600m u&lt;0m/s</td>
<td>-4.2±1.2</td>
<td>-3.8±1.0</td>
<td>-3.7±0.9</td>
<td>-3.5±0.9</td>
<td>-3.2±0.9</td>
<td>-2.5±0.7</td>
<td>-3.3±09</td>
<td></td>
</tr>
<tr>
<td>NEC 7-21N,0-1200m u&lt;0m/s</td>
<td>-61.9±6.0</td>
<td>-56.8±4.7</td>
<td>-47.2±4.7</td>
<td>-37.1±4.3</td>
<td>-30.5±3.6</td>
<td>-29.5±4.5</td>
<td>-21.1±4.2</td>
<td></td>
</tr>
<tr>
<td>STCC 18-25N,0-150m u&lt;0</td>
<td>2.3±0.7</td>
<td>2.0±0.9</td>
<td>1.8±0.6</td>
<td>2.3±0.7</td>
<td>1.2±0.3</td>
<td>1.3±0.4</td>
<td>0.7±0.3</td>
<td></td>
</tr>
<tr>
<td>KR 26-35N,0-1200 u&lt;0</td>
<td>-21.4±9</td>
<td>-21.7±8.3</td>
<td>-16.8±7.5</td>
<td>-6.0±2.4</td>
<td>-3.8±1.2</td>
<td>-2.7±1.8</td>
<td>-2.48±1.1</td>
<td></td>
</tr>
<tr>
<td>KE 31-40N,0-1200 u&gt;0m/s</td>
<td>46.0±6.3</td>
<td>54.4±8.2</td>
<td>41±8.0</td>
<td>17.0±3.3</td>
<td>12.1±1.3</td>
<td>5.0±0.85</td>
<td>3.5±1.0</td>
<td></td>
</tr>
</tbody>
</table>

Figure 1. 2004-2011 mean Argo dynamic height (contours) relative to 1500 m depth with the spatial average removed and surface AGC (vectors). Unit of contours is cm. Unit of vectors is cm s$^{-1}$. 
Figure 2. Comparison of Argo surface AGC (thick curve) with the direct current meter measurement of the TRITON array at 10 m depth (thin curve) and with geostrophic currents based on the absolute dynamic topography of satellite altimetry at the same location (asterisk curve). The upper two panels are at (8°N, 137°E) and the lower two panels are at (8°N, 156°E). The current meter velocity components with the surface Ekman velocity removed are shown in dash curves for comparison.
Figure 3. Comparison of the surface dynamic height based on Argo profiles in reference to the 1900 m (a) with the sea surface height of satellite altimeter (b) in December 2012, showing a dislocated eddy in the vicinity of the mooring site (marked with a dot) in the former. Spatial averages are removed in the plots. Unit is cm.
Figure 4. Root-mean-square differences and correlation coefficients between the Argo AGC and altimeter geostrophic currents at the surface of the North Pacific Ocean during 2004-2011. Upper panels are for the root-mean-square differences. Contour interval is 2 cm s\(^{-1}\). Lower panels are for the correlation coefficients, with correlations below 95% significance level not shown. Contour interval is 0.2.
Figure 5. Time series of area averaged meridional geostrophic currents based on the satellite altimeter data between 120°E and the dateline along 8°N and 18°N. The short dash and long solid lines represent the mean of 2004-2011 and of 1993-2011, respectively. The standard deviations of 1993-2011 and the differences between the long-term and short-term means are marked with the bars.
Figure 6. Comparison of 2004-2011 mean meridional transport based on Argo AGC (a) and the S-E transport based on the annual mean (2004–2011) NCEP surface wind data (b) in the North Pacific Ocean. Their difference is shown in (c). Unit is $10^6$ m$^3$ s$^{-1}$. The wind stress calculation is based on the Garratt drag coefficient. The gray shade is negative.
Figure 7. 2004-2011 mean zonal geostrophic velocity (thin contours) from Argo data in different meridional sections across the North Pacific Ocean. Unit is cm s\(^{-1}\).

Thick contours mark the potential density surfaces. The gray shading indicates westward currents. The dots at the grid points stand for signals below the error bars.
Figure 8. In-situ zonal velocity profiles measured by ADCP current meters at 142° E, 5° N on Dec 22, 2001 (a) and at 137° E, 5° N on Oct 27, 2002 (b).
Figure 9. Comparison of the AGC based on the Argo (upper three panels) and the WOA09 (lower three panels) hydrography data at different meridional sections across the tropical Northwestern Pacific Ocean. Contour interval is 2 cm s$^{-1}$. Thick contours mark the potential density surfaces. The gray shading indicates westward currents.
Figure 10. Mean vertically integrated zonal transport of the AGC and of the Sverdrup theory (upper two panels), and their difference (lower panel). Unit is $m^2 s^{-1}$.

The gray shading indicates westward currents.
Figure 11. Error bars (see text) of the surface zonal (a) and meridional (b) geostrophic currents based on Argo data and of the grid meridional transport of the geostrophic currents smoothed by a 5° longitude by 5° latitude average filter (c). Contour intervals are 2 cm s$^{-1}$ for the surface AGC and 0.5 Sv for the meridional transport. The gray shading indicates less than 4 cm s$^{-1}$ for a and b, and 1.5 Sv for c.
Figure 12. Same as Figure 5, except that the AGC meridional transport is based on the WOA09 data. Unit is $10^6 \text{ m}^3 \text{ s}^{-1}$. The bottom limit of the vertical integration is 1900 m. The gray shade is negative.
Figure 13. Dependence of the meridional transport discrepancy on the choice of the isopycnal surfaces of (a) $26.5\sigma_0$, (b) $27\sigma_0$, (c) $27.2\sigma_0$, and (d) $27.5\sigma_0$, respectively, as the bottom limits of the integration. The geostrophic transport is based on the Argo geostrophic currents. The Sverdrup balance is based on the NCEP wind. The gray area is negative. Contour interval is 5 Sv.
Figure 15. Comparison of the meridional transport discrepancies of the OFES model from the wind-driven meridional transport based on Sverdrup Balance. (a) The OFES meridional transport is calculated based on the geostrophic currents calculated from the OFES simulated temperature and salinity fields using the P-vector method. (b) The OFES meridional transport is calculated based on the OFES simulated velocity. (c) Distribution of the depth of vanishing vertical velocity in the OFES simulation, Unit is km. (d) Meridional transport discrepancy based on the function of $z=-H(x,y)$ in Panel c. The gray shade is negative.
Supplemental Material
Click here to download Non-Rendered Figure: new--jpo904.doc