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Transport and thermohaline structure in the western tropical North Pacific

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

in

Oceanography

by

Martha Coakley Schönau

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2017
The dissertation of Martha Coakley Schönau is approved, and it is acceptable in quality and form for publication on microfilm and electronically:

Chair

University of California, San Diego

2017
DEDICATION

To my parents, for their support and encouragement.
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PUBLICATIONS


ABSTRACT OF THE DISSERTATION

Transport and thermohaline structure in the western tropical North Pacific

by

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Doctor of Philosophy in Oceanography

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Professor Daniel Rudnick, Chair

Transport and thermohaline structure of water masses and their respective variability are observed and modeled in the western tropical North Pacific using autonomous underwater gliders, Argo climatology and a numerical ocean state estimate. The North Equatorial Current (NEC) advects subtropical and subpolar water masses into the region that are transported equatorward by the Mindanao Current (MC). Continuous glider observations of these two currents from June 2009 to December 2013 provide
absolute geostrophic velocity, water mass structure, and transport. The observations are compared to Argo climatology (Roemmich and Gilson, 2009), wind and precipitation to assess forcing, and annual and interannual variability. Observations are assimilated into a regional ocean state estimate (1/6°) to examine regional transport variability and its relationship to the El Niño-Southern Oscillation phenomena (ENSO).

The NEC, described in Chapter 1, is observed along 134.3°E, from 8.5°N to 16.5°N. NEC thermocline transport is relatively constant, with a variable subthermocline transport that is distinguished by countercurrents centered at 9.6°N and 13.1°N. Correlation between thermocline and subthermocline transport is strong. Isopycnals with subducted water masses, the North Pacific Tropical Water and North Pacific Intermediate Water, have the greatest fine-scale thermohaline variance. The NEC advects water masses into the MC, described in Chapter 2, that flows equatorward along the coast of Mindanao. Gliders observed the MC at a mean latitude of 8.5°N. The Mindanao Undercurrent (MUC) persists in the subthermocline offshore of the MC, with a net poleward transport of intermediate water typical of South Pacific origin. The variable subthermocline transport in the MC/MUC has an inverse linear relationship with the Nino 3.4 index and strongly impacts total transport variability. For each the MC and NEC, surface salinity and thermocline depth have a strong relationship with ENSO, and there is relationship between the fine-scale and large-scale isopycnal thermohaline structure. In Chapter 3, a numerical ocean state estimates shows strong interannual variability of regional transport with ENSO. Prior to mature ENSO events, transport in each the NEC, MC and North Equatorial Counter Current (NECC) increase. The increase is from meridional gradients in isopycnal depth related to interannual wind anomalies.
Introduction

Large-scale ocean circulation moves heat, salt, and nutrients on basin wide-scales. The redistribution of heat between the tropics and poles forces the global climate through air-sea interactions on interannual and decadal time-scales. The western tropical Pacific, the location of the warmest, freshest pool of water, has a strong air-sea interaction, and provides an essential pathway for oceanic heat transport to the Indian Ocean, the only inter-basin transport in the tropical ocean (Gordon 1986; Ganachaud and Wunsch 2000). The warm water impacts monsoon variability (Torrence and Webster 1999) and is the location of westerly wind bursts that begin El Niño events (Bjerknes 1969; Clarke 2014).

The variability of the western tropical Pacific circulation and heat content can thus affect global rainfall, climate, and ocean temperatures, creating far ranging impacts on marine and terrestrial ecosystems.

The role of the western tropical Pacific in the global overturning circulation and the impact of strong El Niño-Southern Oscillation (ENSO) episodes led to several research initiatives to observe and model the regional circulation. In the 1980s, the Tropical Ocean – Global Atmosphere program (TOGA) was initiated to monitor equatorial temperatures using an array of moorings, that was later developed into the Tropical Atmosphere Ocean/Triangle Trans-Ocean Buoy Network (TAO/TRITON) array (McPhaden et al. 1998). Simultaneously, hydrographic surveys in the far western Pacific such as the Western Pacific Ocean Circulation Study (WEPOCS) (Lindstrom et al. 1987) and collaborative United States/Peoples Republic of China (US/PRC) cruises (Toole et al. 1988) sought to resolve the mean structure of the currents in the region. Observations revealed a confluence of Pacific tropical, subtropical and subpolar water masses from
each hemisphere (Fine et al. 1994), traceable by oxygen content and their thermohaline structure. These water masses mix along and across isopycnal as they are advected from the subtropical gyres into the tropical, low-latitude western boundary currents. Transport is affected by annual forcing by the East Asian Monsoon and arrival of Rossby waves from the central Pacific, and interannual forcing from ENSO (Kim et al. 2004). However, the frequency of eddies in the region and a lack of repeat observations have made annual and interannual fluctuations in transport difficult to resolve. The goal of recent research in the far western Pacific has been to understand the region’s role in the coupled ocean-atmosphere system and to improve models and climate predictability. Research in the western tropical South Pacific, under the name South Pacific ocean circulation and climate experiment (SPICE) (Ganachaud et al. 2014), focused on the pathway to the Equatorial Undercurrent, whose water mass transport was considered important to ENSO and may play a role decadal climate variability. In the North Pacific, the Origins of the Kuroshio and Mindanao Current initiative (OKMC) (Rudnick et al. 2015), and the Northwestern Pacific Ocean Circulation and Climate Experiment (NPOCE) (Hu et al. 2015) sought to find the mean structure and transport variability in the far western tropical North Pacific, the main source for the ITF.

During OKMC, from June 2009 to January 2014, Spray gliders made repeat transects across the North Equatorial Current (NEC), the southern arm of the North Pacific subtropical gyre that flows into the Philippine Sea, and the Mindanao Current (MC), the low-latitude western boundary current that feeds the ITF and the North Pacific tropical circulation. Glider observations of the NEC and MC establish autonomous underwater gliders as capable platforms to observe and monitor large-scale ocean current
transport. To our knowledge, the roughly 2000 km round-trip travel across each the NEC and MC are the longest, repeat transects of ocean currents by gliders. Referencing to the depth-average velocity of the glider provides an absolute measure of water mass transport. The horizontal resolution of the gliders resolves fine-scale water mass structure, structure near topography, and narrow geostrophic currents. Glider observations can be combined with those from Argo floats to observe the ocean over a large spatial range.

The focus of this dissertation is on water mass structure and transport in the far western tropical North Pacific from autonomous observations and analysis of an Ocean State Estimate that assimilates these observations. Chapter 1 describes glider observations across the NEC, and Chapter 2, glider observations across the MC. Analysis provides mean thermohaline structure and transport of these currents and their variability during the four-year period of observations. Argo climatology (Roemmich and Gilson 2009) and forcing from wind stress and precipitation are used to separate annual and interannual anomalies. Chapter 3 uses a data assimilating Ocean State Estimate, Argo climatology, and glider observations to understand the mean structure and annual and interannual transport variability in the far western North Pacific. The relationship between the interannual transport anomalies and ENSO is also examined.

The NEC and MC have relatively steady thermocline transport of subtropical water, identified by a subsurface salinity maximum. Beneath the dominant surface flow there are persistent subthermocline countercurrents that meander and vary in width and strength. North Pacific intermediate water is transported westward and equatorward in the subthermocline by the NEC and MC, whereas the undercurrents transport intermediate water typical of South Pacific poleward and eastward. The subthermocline transport
variability has a large impact on the total transport, and in the MC is correlated with the Nino 3.4 index. The fine-scale water mass structure of the NEC and MC is enhanced along isopycnals where large-scale gradients are created by subducted water masses. The ratio between the fine-scale and large-scale structure persists between the NEC and MC, and a comparison is made between their thermohaline structures. Argo Climatology (Roemmich and Gilson 2009) resolves the annual and interannual variability during glider observations, and observations are compared with surface forcing from wind and precipitation products. The interannual variability in surface salinity and thermocline depth associated with ENSO exceeds annual anomalies.

The strong interannual anomalies in thermocline depth, salinity, and transport motivate analysis of annual and interannual water mass transport variability in Chapter 3. Ocean state estimates, verified by glider and Argo observations of water mass transport, are used to resolve annual and interannual transport anomalies near topography and on the equator. The state estimate finds increases in the NEC, NECC, and MC, and their prior to the El Niño event in 2015. The lead in zonal transport differs from the lag in transports that has been observed in the South Pacific (Kessler and Cravatte 2013). The interannual transport anomalies affect the distribution of warm water in the tropical Pacific and Indian Ocean, and have some relationship to anomalous zonal wind stress in the tropics and Ekman pumping in the central Pacific.
References


Chapter 1

Glider observations of the North Equatorial Current in the western tropical Pacific

The North Equatorial Current (NEC) of the Pacific Ocean advects subtropical, subpolar and tropical water masses. Repeat underwater glider observations of the NEC from June 2009 to January 2014 along 134.3°E provide absolute zonal geostrophic velocity, transport, and water mass structure at length scales of 10 km to 1000 km. The NEC is strongest near the surface and persistent eastward undercurrents are identified deeper than potential density surface 26 kg m$^{-3}$ at 9.6°N and 13.1°N. Mean transport from the surface to 27.3 kg m$^{-3}$ and 8.5°N to 16.5°N is 37.6 Sv ($10^6$ m$^3$ s$^{-1}$), with a standard deviation of 15.6 Sv. The transport variability is greatest deeper than 26 kg m$^{-3}$ due to undercurrent variability. Wavelet analysis at scales of 10 km to 80 km reveals extrema of fine-scale salinity variance along isopycnals (spice variance). High spice variance is found in the North Pacific Tropical Water (NPTW) and the North Pacific Intermediate Water (NPIW), with a spice variance minimum between water masses at 25.5 kg m$^{-3}$. A horizontal Cox number, $C_H$, relates salinity variance at fine-scales (10 km to 80 km) to that at greater length scales (120 km to 200 km). As a function of density, $C_H$ is nearly vertically uniform, indicating that the stirring of mean salinity gradients enhances fine-scale salinity variance. NPTW, with an estimated horizontal eddy diffusivity of order $10^4$ m$^2$ s$^{-1}$, is a useful tracer for the region and may be used to relate the fine-scale salinity variance to an eddy diffusivity.
1.1 Introduction

The North Equatorial Current (NEC) in the Pacific Ocean transports subpolar, subtropical, and tropical water into the western tropical Pacific, an area known for water mass exchange and mixing (Tsuchiya 1968; Lukas et al. 1991; Fine et al. 1994; Johnson and McPhaden 1999). Past studies of the NEC have examined the structure and displacement of water masses in the region (Qiu and Joyce 1992; Fine et al. 1994; Qu et al. 1999), the relationship between wind stress curl, transport, and bifurcation latitude (Qiu and Lukas 1996; Qu and Lukas 2003), transport of the NEC/Mindanao Current (MC)/Kuroshio Current (KC) system (Toole et al. 1990; Qu et al. 1998), and the relationship between the NEC and El Niño-Southern Oscillation (ENSO) (Qiu and Lukas 1996; Kashino et al. 2009). Variability of the NEC is influenced by changes in basin wide wind stress, Rossby waves generated in the eastern North Pacific, and local wind stress in the Philippine Sea (Kessler 1990; Qiu and Joyce 1992; Qiu and Lukas 1996; Qu et al. 2008). In this work, underwater glider observations provide repeat measurements of the geostrophic velocity across the NEC and the horizontal thermohaline structure at scales of 10 to 1000 km. The motivation is to provide new insight into NEC mean structure and its variability and to examine the relationship between the large-scale and fine-scale water mass structure, the stirring of thermohaline gradients, and eddy diffusivity.

Past observational and model estimates of mean transport of the NEC range from -42 to -62 Sv (10⁶ m³ s⁻¹) (negative implies westward), depending on reference level and latitudinal range (Toole et al. 1990; Qiu and Joyce 1992; Qiu and Lukas 1996; Qu et al. 1998). Interannual transport variability is larger than annual variability and is commonly related to positive and negative phases of ENSO (Qiu and Joyce 1992; Qiu and Lukas
Persistent eastward undercurrents below the NEC have recently been examined using ARGO floats (Cravatte et al. 2012; Qiu et al. 2013b). Previous estimates of NEC mean geostrophic transport differ due to the range of integration or by referencing to a level of no-motion where eastward undercurrents persist. Repeat glider sections in this study reference to a directly measured depth-average velocity providing a new estimate of NEC mean transport and revealing a relationship between NEC transport variability and that of the persistent eastward undercurrents.

Subpolar, subtropical, and tropical water masses in the NEC have been well documented (Tsuchiya 1968; Wyrtki and Kilonsky 1984; Talley 1993; Fine et al. 1994). Subducted water masses, the North Pacific Intermediate Water (NPIW) and the North Pacific Tropical Water (NPTW), carry subpolar and subtropical signals to the tropics in the form of “spice”, compensating temperature and salinity fluctuations along isopycnal surfaces (Gu and Philander 1997; Zhang et al. 1998; Sasaki et al. 2010). Areas of high spice variance, where there are large temperature and salinity fluctuations along isopycnals, can indicate both water source and water that is actively stirred (Ferrari and Polzin 2005; Smith and Ferrari 2009; Rudnick et al. 2011; Cole and Rudnick 2012; Todd et al. 2012). The thermohaline structure of NPIW and NPTW are thus set during subduction and then modified by stirring and mixing along the path of the NEC. In this study, glider sections are used to identify water masses, quantify their variability during the period of observation, and provide thermohaline structure at length scales as small as 10 km in order to examine along-isopycnal stirring.
Here we present repeat glider sections of the NEC along 134.3°E in the Philippine Sea. Ten glider deployments occurred from June 2009 to January 2014, with each deployment providing two sections of the NEC. Glider observations have horizontal resolution of roughly 6 km and complete a section of the NEC in roughly two months. Absolute zonal geostrophic velocity and transport for each section give an estimate of the NEC mean velocity, variability, and a relationship between the westward NEC and eastward undercurrents. ARGO climatology (Roemmich and Gilson 2009) is combined with glider observations to calculate an effective horizontal diffusivity for NPTW. Wavelet transforms are used to quantify isopycnal thermohaline structure, or spice, at fine scales (10 km to 80 km) and at large scales (120 km to 200 km). A horizontal Cox number, the ratio between spice variance at different length scales, is used to examine stirring of water masses in the NEC and may be related to the effective horizontal diffusivity.

The paper proceeds as follows: section 2 provides an overview of data collection and analysis; section 3 discusses NEC mean geostrophic velocity structure, transport and their respective variability; section 4 describes water mass properties, large-scale variability and fine-scale thermohaline structure; section 5 discusses water mass advection, stirring, and mixing; and section 6 provides a summary and discussion.

1.2 Data collection and analysis

The Spray underwater glider is a buoyancy driven autonomous vehicle equipped with a Sea-Bird CTD and Seapoint fluorometer. The glider follows a horizontal saw-tooth pattern, descending to a depth of 1000 m over a horizontal distance of 3 km then
ascending over another 3 km. Data are collected during ascent, similar to a profiling float, leading to profiles that are roughly 6 km apart in the horizontal. During ascent the vertical sampling is at roughly 1 m and the resulting data are binned vertically into 10 m depth bins. Further details of Spray glider operations are described by Sherman, et al. (2001), Rudnick, et al. (2004), and Rudnick and Cole (2011).

To observe the NEC, Spray gliders was deployed from Palau in the Philippine Sea and sampled a trajectory along 134.3°E between 7.5°N and 17°N (Figure 1.1). During the 4.5 years of observations, from June 2009 to January 2014, there were 10 glider deployments. Gliders made over 5000 profiles along this trajectory, completing 19 sections of the NEC from 8°N to at least 15°N. A diagram of the glider trajectory by latitude shows the sections of the NEC made for each glider deployment (Figure 1.2a) and the annual coverage (Figure 1.2b). A section of the NEC was completed roughly every two to three months, with the goal of sustained coverage of the NEC. There were slightly more profiles from June to November. The region was well sampled at all times of the year from 10°N to 15°N where the NEC is strongest.

An example glider section of the NEC observed from June to July 2009 is shown in Figure 1.3 by contouring the vertically binned salinity and potential temperature by depth, by objectively mapping these variables, or by interpolating these variables to isopycnal surfaces. Owing to the relatively slow speed of the glider relative to water, roughly 0.25 m s⁻¹, high frequency internal wave motion appears as horizontal ocean structure (Rudnick and Cole 2011) (Figure 1.3a). To identify the large-scale water mass structure and filter out the effects of internal waves the sections were objectively mapped using a Gaussian autocovariance (Figure 1.3b) (Bretherton et al. 1976). An appropriate
decorrelation length scale was found by fitting the autocovariances of potential temperature and salinity at each depth to Gaussian functions. A mean length scale of 80 km was chosen for the objective maps. Mean sections of potential temperature, salinity, and geostrophic velocity for the region were made by objectively mapping all glider profiles taken during the 4.5 year period simultaneously. Profiles that were not on the longitudinal line of 134.3°E due to strong currents were projected onto this line for the objective map. The large-scale variability of the NEC was quantified by taking the standard deviation of potential temperature, salinity, and geostrophic velocity between individually mapped glider sections.

Mean, annual, and monthly ARGO climatology (Roemmich and Gilson 2009) provided additional measurements of temperature, salinity and geostrophic velocity in the region. Although at lower resolution than the glider, the climatology has a longer time series, from January 2004 to present. A mean ARGO climatology (1/6° resolution) was compared to the objectively mapped mean glider section and used to obtain ocean structure at other longitudes along the direction of flow of the NEC. Annual (1/2° resolution) and monthly (1° resolution) climatology provided the annual cycle and interannual anomalies of temperature and salinity.

Anomalies of precipitation and wind were used to diagnose surface forcing in the NEC. Precipitation was obtained from the Earth System Research Laboratory (ESRL) CPC Merged Analysis of Precipitation (CMAP) monthly precipitation data, gridded at 2.5° (Xie and Arkin 1997), and wind components were obtained from monthly mean NCEP/NCAR reanalysis wind product, gridded at 2.5° and available from ESRL (Kalnay et al. 1996).
Fine-scale thermohaline structure of the NEC was examined by linearly interpolating the glider profiles to isopycnal surfaces, as is shown for the example glider section (Figure 1.3c). Examining data on isopycnals intrinsically filters the heaving due to internal waves. Wavelet transforms were used to quantify spice variance on density and depth surfaces at a variety of length scales. A wavelet transform provides wavenumber information that is localized in space, allowing spice variance to be analyzed with reference to latitude. Examples of wavelet transforms in geosciences are found in Farge (1992), Ferrari and Rudnick (2000), Lau and Weng (1995), and Torrence and Campo (1998). The wavelet analysis in this study follows Todd, et al. (2012). Wavelet transforms were performed on salinity for each glider section. The mother wavelet is a Morlet wavelet given by
\[ \psi(x) = e^{i2\pi k} e^{-\frac{x^2}{2}} \]
with \( k = 1 \). Prior to taking the wavelet transform, the salinity was interpolated to a common 1 km horizontal grid and a linear trend was removed. Interpolation may reduce variance at the highest wave numbers but it facilitates an estimate of the mean variance over all glider sections. Convolving the respective Fourier transforms of the mother wavelet and salinity performed the transform. Summing over scales produces the energy content for the chosen wavelength range as salinity variance per kilometer. Calculations using the wavelet transform are further described in section 5.

1.3 Geostrophic velocity and transport

1.3.1 Geostrophic velocity
The NEC relative zonal geostrophic velocity was calculated from the objectively mapped mean potential density using the thermal wind equations. To obtain an absolute velocity, the relative velocity was referenced to the mean depth-average zonal velocity measured by the glider (Figure 1.1) (Davis et al. 2008; Todd et al. 2011; Davis et al. 2012; Rudnick et al. 2015). The mean absolute zonal geostrophic velocity is shown in Figure 1.4; negative (positive) values indicate westward (eastward) velocity. Westward geostrophic velocity extends from 8.5°N to 16.5°N and from the surface to potential density surface 26 kg m\(^{-3}\). Deeper than 26 kg m\(^{-3}\), transport alternates between westward flow and eastward flowing undercurrents.

The depth of the NEC increases with increasing latitude as the pycnocline deepens, maintaining a uniform potential vorticity across the NEC (Toole et al. 1990). Potential vorticity is examined in ocean basins due to its importance to large-scale oceanic circulation and its use as a dynamic tracer of water masses (Holland et al. 1984; Keffer 1985; Talley 1985, 1988). Here, glider observations of the NEC provide a high-resolution meridional section of potential vorticity in the tropics. The potential vorticity of the mean glider section is given by

\[ Q = \frac{N^2}{g} \left( -\frac{\partial u}{\partial y} + f \right), \]

where \( N^2 \) is the buoyancy frequency squared, proportional to the vertical buoyancy gradient, and \( f \) is the planetary vorticity. The relative vorticity is approximated by \(-\partial u / \partial y\) with the assumption that the meridional velocity and its gradient in the zonal direction are small. However, across the NEC the relative vorticity is much smaller than the planetary vorticity, as is typical in large-scale horizontal flow (Pedlosky 1996). The
buoyancy frequency squared and potential vorticity, normalized by their respective maxima, are shown in Figure 1.5. The roughly constant potential vorticity along isopycnals in the NEC is due to the compensation between vertical buoyancy gradient and planetary vorticity; the buoyancy frequency squared decreases by a factor of two and the planetary vorticity increases by a factor of two. The compensation shows the important contribution of a changing planetary vorticity across the NEC to achieve a relatively uniform potential vorticity along isopycnals.

The zonal geostrophic flow of the NEC is strongest in the upper 200 m from 9°N to 13°N, with a maximum mean surface velocity of 0.47 m s\(^{-1}\), and a maximum velocity during the time of observation of 0.69 m s\(^{-1}\). Variability of the current, as measured by the standard deviation between sections, is largest in the upper 150 m where the current is the strongest (Figure 1.4b). The variability may be due to the meandering of the current or mesoscale eddies, as large eddies have previously been reported in the NEC (Wyrtki 1982). A time series of the near surface velocity averaged from 10 m to 100 m confirms that there is considerable mesoscale variability (Figure 1.6a). The time series reveals several cores of intense westward flow, that vary in space and time, amid a broad slower current. Eastward surface flow is common between 13°N and 16.5°N.

Mean eastward flowing undercurrents deeper than potential density surface 26 kg m\(^{-3}\) are centered at 9.6°N and 13.1°N, respectively (Figure 1.4a). The standard deviation of geostrophic velocity is elevated at the edges of the undercurrents (Figure 1.4b). A time series of velocity averaged from 26.7 kg m\(^{-3}\) to 27.3 kg m\(^{-3}\) shows that the undercurrents are persistent features of the region with variable width and location (Figure 1.6b). The undercurrent cores, defined as the location of maximum eastward velocity, meander north
and south. The widths of the undercurrents, defined as the north/south extent of eastward velocity, undergo expansion and contraction. The undercurrent at 9.6°N had a core position between 8.8°N and 11.1°N, varied in width from 100 km to 350 km, and had a maximum velocity at potential density 26.9 kg m\(^{-3}\). The undercurrent often extended to the surface when the core meandered south of 9°N. During its maximum breadth (>300 km), the undercurrent had two cores of eastward velocity, as observed between June and October 2010, and may be an instance of undercurrents combining. The undercurrent at 13.1°N was narrower and more localized than that at 9.6°N, with a maximum core velocity at a 27.1 kg m\(^{-3}\), and a core position between 12.4°N and 13.8°N. The width ranged from 0 to 200 km. Twice during observations the undercurrent at 13.1°N connected with an eastward surface current, possibly a mesoscale eddy. Separate eastward undercurrents were intermittently observed south of 8.8°N and north of 14°N, suggesting the existence of undercurrents at other latitudes.

1.3.2 Transport

The mean zonal geostrophic transport was calculated by integrating each geostrophic velocity section from 8.5°N to 16.5°N and surface to potential density surface 27.3 kg m\(^{-3}\) and then averaging over all sections. Fifteen of the nineteen sections covered this range. The mean zonal geostrophic transport is -37.6 Sv with a standard deviation between sections of 15.6 Sv and a standard error of 4.1 Sv. The geostrophic transport is comparable to other estimates of NEC transport that are referenced to a level of no motion deeper than 1500 m (Toole et al. 1990; Qu et al. 1998). To compare, ARGO climatology (1/6°) referenced to 2000 m and integrated over this range has a zonal geostrophic transport of -37.7 Sv.
Integrating the mean geostrophic velocity section (Figure 1.4a) provides absolute zonal transport as a function of depth, density, and latitude (Figure 1.7), with a total integrated transport of -39.6 Sv. As a function of depth, the maximum transport is near the surface, with strong transport extending through the thermocline to 300 m depth and westward transport extending to 500 m (Figure 1.7a). As a function of potential density, the maximum in transport is between 21 kg m\(^{-3}\) and 22 kg m\(^{-3}\), with strong transport to 26.5 kg m\(^{-3}\) (Figure 1.7b). Deeper than 27 kg m\(^{-3}\), the net transport is eastward. As a function of latitude, total transport (black) is separated into transport with density shallower (blue) and deeper (red) than 26 kg m\(^{-3}\), where 26 kg m\(^{-3}\) denotes the mean location of the top of the undercurrents (Figure 1.7c). Dividing the transport separates the transport of the undercurrents from that of overlying NEC. Maxima in total transport are between 11°N and 12°N, corresponding to strong surface flow, and between 14°N and 16°N, where the current is weaker but extends to 1000 m. Minima in total transport are due to eastward flow deeper than 26 kg m\(^{-3}\) and are located near the latitudes of the undercurrent cores. The strongest flow of the NEC, represented by transport shallower than 26 kg m\(^{-3}\), is relatively uniform by latitude.

As a function of time, the transports for individual glider sections are again divided into transport shallower than and deeper than potential density surface 26 kg m\(^{-3}\) in order to separate the transport variability of the overlying NEC from that of the undercurrents (Figure 1.8). Annual variability in this region is weak relative to interannual variability, so the two cannot be distinguished in our 4.5-year record (Kessler 1990; Qiu and Lukas 1996). The transport and standard deviation for water shallower than 26 kg m\(^{-3}\) is -33.1 \(\pm\) 7 Sv, and the transport and standard deviation for water deeper
than 26 kg m$^{-3}$ is $-4.5 \pm 9.2$ Sv. Thus transport in the NEC, confined mostly to the upper 300 m ($<26$ kg m$^{-3}$), remained relatively constant over the 4.5 year period. Comparing the magnitudes and standard deviations of transport, the transport deeper than 26 kg m$^{-3}$ is much smaller than that shallower than 26 kg m$^{-3}$, but exhibits greater variability. Transport deeper than 26 kg m$^{-3}$ is correlated with transport shallower than 26 kg m$^{-3}$ ($R=0.87$), and with the width of the undercurrents ($R=0.93$). The relationship between the undercurrents and overlying NEC was highlighted in 2013, when both extrema in total transport occurred, -71.8 Sv (May) and -10.3 Sv (December). In May, the transport in the NEC shallower than 26 kg m$^{-3}$ was exceptionally strong and the northern undercurrent was not present, leading to strong westward (negative) transport over this range. In December, the transport in the NEC shallower than 26 kg m$^{-3}$ was weak and large undercurrents were observed. Thus when the NEC transport is strong in the upper 300 m, eastward undercurrents tended to be small or non-existent over the range of observation. During weak NEC transport, the undercurrents tended to have greater width over this range leading to an increase in eastward undercurrent transport.

1.4 Water masses

Water masses transported by the NEC are identified by their salinity extrema in the objectively mapped mean salinity section (Figure 1.9). Potential temperature/salinity (TS) diagrams colored by latitude and geostrophic velocity provide the respective location and advection of these salinity extrema (Figure 1.10). The TS geostrophic velocity diagram is constructed by averaging in 0.01 psu and 0.2 °C bins to show the zonal advection of different TS properties (Figure 1.10b). Water masses identified by
their salinity extrema are North Pacific Intermediate Water (NPIW), North Pacific Tropic Water (NPTW), and North Pacific Tropical Surface Water (NPTSW).

NPIW is a subsurface salinity minimum that is advected south in the subtropical gyre and has a typical density in the western Pacific between 26 and 27 kg m\(^{-3}\) (Talley 1993; Bingham and Lukas 1994; Fine et al. 1994). Although the core of the NPIW lies north of the glider section (Qu et al. 1999), a tongue of NPIW can be identified as the salinity minimum north of 14°N, with a potential density of 26.6 kg m\(^{-3}\), salinity of 34.3 psu, and potential temperature of 9 °C (Figures 1.9, 1.10a). There is westward advection of NPIW in the NEC (Figure 1.10b).

NPTW is a subsurface salinity maximum formed by subduction in the gyre center due to excess evaporation (Fine et al. 1994). At formation, the salinity maximum of NPTW is near 35.2 psu, but decreases as the water mass is advected westward due to freshening from the NPIW and surface waters (Tsuchiya 1968; Lukas et al. 1991; Johnson and McPhaden 1999). In the glider section, NPTW is identified as the salinity maximum between 11°N and 16°N near potential density 23.5 kg m\(^{-3}\) (Figures 1.9, 1.10a). The salinity is between 35.0 and 35.1 psu, with a potential temperature of 25 °C. The mean glider section encompasses the salinity maximum of the NPTW. The NEC advects the NPTW westward at 0.1 to 0.4 m s\(^{-1}\) making it an ideal tracer to examine diffusivity and mixing (Figure 1.10b).

NPTSW is a fresh, shallow layer covering the southern part of the NEC and the northern part of the NECC (Wyrtki and Kilonsky 1984); the origin is either from local rainfall or meridional advection (Delcroix and Hénin 1991; Li et al. 2013). In the glider sections, NPTSW is between 8°N and 13°N, with potential density less than 22 kg m\(^{-3}\),
salinity less than 34.1 psu, and a potential temperature of 29 °C (Figures 1.9, 1.10a). There is westward advection of NPTSW by the NEC north of 9°N and eastward advection south of 8.5°N (Figure 1.10b).

TS properties of the undercurrents were persistent between glider sections. Positive velocities at potential densities deeper than 26 kg m$^{-3}$ identify the undercurrents in the TS zonal geostrophic velocity (Figure 1.10b). The undercurrent at 9.6°N had a salinity of 34.5 psu and density of 27.2 kg m$^{-3}$ and also encompassed a local salinity maximum. The TS characteristics may indicate the presence of Antarctic Intermediate Water (AAIW) (Qu et al. 1999; Qu and Lindstrom 2004). The undercurrent at 13.1°N had a relatively lower salinity and greater density core than that at 9.6°N. The TS properties are consistent with the interleaving of water masses in the North Pacific. Recent research suggests that the undercurrents below the NEC are fed by the southward Luzon Undercurrent and northward Mindanao Undercurrent in the Philippine Sea (Wang et al. 2015). The visible separation of the undercurrents in the TS diagram supports the theory that the undercurrents have different water mass origins.

1.4.1 Large-scale variability

Mesoscale eddies and annual and interannual forcing cause variability between glider sections. Annual fluctuations in temperature and salinity can be of similar or lesser amplitude than interannual fluctuations caused by ENSO. During the time of observations, there was an ENSO positive phase (El Niño) from July 2009 to March 2010, and an ENSO negative phase (La Niña) from July 2010 to March 2012, according to the NINO 3.4 index. As explained in section 2, ARGO climatology (1/2° and 1° resolution) (Roemmich and Gilson 2009) were used to identify annual and interannual variability
observed in the glider sections, and monthly wind and precipitation products were used to identify annual and interannual surface forcing over the region.

The highest standard deviation of salinity is near the surface, ranging from 0.15 to 0.3 psu (Figure 1.11a). Annual and interannual surface salinity anomalies from ARGO climatology are 0.15 psu and 0.5 psu, respectively. Variability in surface salinity is related to both local and regional precipitation and evaporation. The annual minimum (maximum) in surface salinity in January (March) follows anomalously high (low) precipitation over the western Pacific. The annual salinity minimum lags local rainfall by four months, suggesting that advection contributes to the salinity minimum, as previous researchers have found (Delcroix and Hénin 1991; Li et al. 2013). The exceptionally high standard deviation in surface salinity, from the surface to 100 m and between 8°N and 13°N, is caused by the presence and absence of NPTSW due to interannual precipitation associated with ENSO. During La Niña and normal conditions salinity in this region was less than 33.6 psu; during El Niño the salinity exceeded 34.3 psu. The interannual salinity anomaly is consistent with the effect of ENSO on the western tropical Pacific; positive (negative) with decreased (increased) local precipitation following El Niño (La Niña) events. In contrast to the NPTSW, the salinity variability in the regions of the NPTW and the NPIW was relatively low, indicating persistence of these water masses during glider observations.

The largest standard deviation of potential temperature is in the thermocline from 8°N to 10°N (Figure 1.11b). At depth, potential temperature variability is due to the displacement of isopycnal / isothermal surfaces, as potential temperature tends to follow potential density surfaces in this region (Figure 1.9b). ARGO climatology shows annual
temperature fluctuations of 2.5 °C at 100 m due to shoaling (deepening) of the 
thermocline from December to March (July to November). Across the NEC, the 
thermocline depth has annual and semiannual variability due to both the propagation of 
Rossby waves generated in the central Pacific and local wind stress (Wang et al. 2000; 
Qu et al. 2008; Kashino et al. 2011). From ARGO climatology, the annual depth 
variability of the 20 °C isotherm is ± 12 m, with the minimum depth occurring in 
December north of 12°N and in February/March south of 12°N. Annual changes in 
thermocline depth at the southern end of the glider section are consistent with the East 
Asian Winter Monsoon and the build up of the Mindanao Dome due to southwesterlies 
over the region from July to September (Masumoto and Yamagata 1991). The large 
standard deviation in the thermocline temperature between glider sections exceeds known 
annual variability and was influenced by an extreme shoaling of the thermocline from 
8°N to 12°N during an El Niño (2009-2010). During this event there was a -3.5 °C 
temperature anomaly at 100 m (60 m shoaling of the 20 °C isotherm).

1.4.2 Fine-scale structure

The fine-scale variance of salinity or potential temperature along isopyncals, or 
spice variance, can be an indication of water mass source and the stirring of large-scale 
thermohaline gradients by mesoscale eddies (Ferrari and Polzin 2005; Smith and Ferrari 
2009; Rudnick et al. 2011; Cole and Rudnick 2012; Todd et al. 2012). Spice can be 
visualized as the fluctuations of potential density contours along potential temperature 
surfaces (Figure 1.12). Each contour, spaced at a 0.1 kg m⁻³, is an isopycnal surface. 
Spice appears as waviness, where fluctuations of potential temperature are compensated
by salinity along the isopycnal. Areas of high spice variance are between 22 kg m\(^{-3}\) and 25 kg m\(^{-3}\), and below 26 kg m\(^{-3}\).

To quantify the fine-scale spice variance the wavelet transform was performed on salinity along isopycnals for each glider section as described in the section 2. Wavelengths from 10 km to 80 km were included in the transform, where 80 km is the mean decorrelation length scale. The mean fine-scale salinity variance per kilometer averaged over all glider sections is shown in Figure 1.13a, where red (blue) indicates areas of high (low) variance. There is high spice variance along the same isopycnals as NPTW and NPIW. A variance minimum occurs between these water masses from 25 kg m\(^{-3}\) to 26 kg m\(^{-3}\). Here, both salinity and potential temperature are nearly uniform on isopycnal surfaces indicating that the water is thoroughly mixed. In contrast, performing the wavelet transform on salinity on depth surfaces reveals high salinity variance in areas of high stratification, such as the pycnocline, due to heaving by internal waves (Figure 1.13b).

The spectra of salinity at constant potential density quantify salinity variance as a function of wavelength. To calculate the spectra, the along-isopycnal salinity was interpolated to a 1 km grid for each glider section, a linear trend was removed, and the Fourier transform was performed. The Fourier coefficients were band averaged and the width of the band increased with increasing wave number. The spectra were cut off at a wavelength of 12.5 km, above the Nyquist frequency for 6 km resolution. The mean spectra for each isopycnal, averaged over all glider sections, are shown in increments of 1 kg m\(^{-3}\) in Figure 1.14. The spectra have similar magnitudes as a function of wavenumber, with the exception of a smaller salinity variance along 25.5 kg m\(^{-3}\) for all wavenumbers.
The spectral slopes fall between -2 and -1. A passive tracer spectral slope of -2 is typical at fronts in the upper ocean (Cole and Rudnick 2012; Callies and Ferrari 2013). Here, most of the isopycnals have spectral slopes near -2. A spectral slope approaching -1 is observed where there is a spice variance minimum along 25.5 kg m\(^{-3}\). A decreased slope means that there is relatively greater variance at shorter wavelengths along that isopycnal. Confidence intervals indicate that the difference in slope is a significant result. Several studies have used dimensional analysis to relate the spectral slope of passive tracers to different mixing regimes (Scott 2006; Vallis 2006; Callies and Ferrari 2013), and it is interesting that flattest spectral slope is observed along the same isopycnal where we observe a spice variance minimum across the NEC.

1.5 Advection and mixing

The relationship between spice variance at different length scales provides insight into mixing and stirring in the ocean interior. The fine-scale thermohaline variance can be related to that at larger scales by calculating a horizontal Cox number. A Cox number is the ratio between the average gradient of fine-scale salinity or temperature fluctuations squared and the square of the large-scale salinity or temperature gradient. The Cox number can be calculated in either the vertical or horizontal and is then typically related to turbulent eddy diffusivity (Osborn and Cox 1972; Ferrari and Polzin 2005). Here we define a horizontal Cox number, C\(_H\), as the ratio between the average fine-scale salinity variance and average large-scale salinity variance calculated using the wavelet transform,
\[ C_H = \frac{\left( \frac{\partial S'^2}{\partial y} \right)}{\left( \frac{\partial S_{\text{large}}'^2}{\partial y} \right)}. \]

The numerator is the mean salinity variance per kilometer at length scales of 10 km to 80 km (Figure 1.13a), obtained from the wavelet transform as previously described. The denominator is the large-scale salinity variance per kilometer at length scales of 120 km to 200 km (Figure 1.15a). We choose 120 km as a reasonable lower bound as it encompasses the e-folding envelope of the Morlet wavelet for a carrier wavelength of 80 km. The upper bound (200 km) was chosen to limit edge effects of the transform on the range of the glider observations. The over bar on the numerator and the denominator denotes a mean over all sections.

The horizontal Cox number is used here to relate the fine-scale salinity variance to the salinity gradients at larger scales and quantify the stirring along isopycnals. Averaging \( C_H \), the fine-scale salinity variance, and the large-scale salinity variance by latitude from 10°N to 13.5°N reveals a pattern of these fields as a function of potential density (Figure 1.15b). There are local maxima in fine-scale and large-scale variance between 22 kg m\(^{-3}\) and 24 kg m\(^{-3}\), encompassing the potential density of the NPTW, and between 26 kg m\(^{-3}\) and 27 kg m\(^{-3}\), the potential density of the NPIW. A minimum in both fine-scale and large-scale variance occurs at 25.5 kg m\(^{-3}\). The resulting \( C_H \) is nearly uniform throughout the water column, indicating that fine-scale salinity variance along isopycnals is proportional to the salinity gradient at larger scales. The absolute value of \( C_H \) depends on the wavelengths included in the wavelet transforms. Here, \( C_H \) is of order unity. The uniformity of \( C_H \) as a function of potential density indicates that stirring of
thermohaline gradients at large scales creates fine-scale thermohaline variance. By analogy to a vertical Cox number, this process may be related to a large-scale diffusivity.

It is interesting to find the isopycnal diffusivity needed to maintain the thermohaline structure in the presence of observed advection. To see how the water masses advect, TS zonal geostrophic velocity diagrams were made for ARGO climatology (1/6°) at longitudes east and west of the glider section (Figure 1.16). To choose a reference level, the zonal geostrophic velocity from the ARGO climatology at 134.25°E was compared to the absolute zonal geostrophic velocity of the glider at 134.3°E. Referenced to 2000 m, the zonal geostrophic velocity from ARGO climatology shows similar velocity features and magnitudes to those of the glider. As discussed in section 3.2, their integrated transports are also similar. The ARGO TS absolute zonal geostrophic velocity diagrams are thus all referenced to 2000 m and include a latitudinal range of 8°N to 16.5°N. The velocity is averaged over 0.01 psu and 0.2 °C bins. The ARGO TS geostrophic velocity diagram at 134.25°E (Figure 1.16c) is comparable to that of the glider (Figure 1.10b). Glider and ARGO diagrams have similar westward velocities for both NPTW and NPIW.

Tracking the salinity extrema between longitudes, the extrema of NPTW and NPIW decrease as the water masses move westward. At 146.25°E (Figure 1.16a), NPTW is near potential density surface 23.8 kg m⁻³, with salinity near 35.2 psu. At 128.25°E (Figure 1.16d), the maximum salinity along 23.8 kg m⁻³ is below 35 psu. The decreased salinity maximum in the direction of flow suggests mixing. An increase in salinity occurs in the NPIW, but the change in salinity is smaller than that for the NPTW.
The decrease of the NPTW salinity maximum along the pathway of the NEC suggests that this maximum is a good tracer to use in an estimate of large-scale diffusivity. The mean glider section encompassed the salinity maximum of the NPTW. NPTW has a strong zonal velocity (Figure 1.10b), and a standard deviation of salinity less than 0.1 psu (Figure 1.11), indicating the persistence of this water mass over time. An estimate of the large-scale diffusivity of salinity for NPTW can be made with the mean glider observations and ARGO climatology using the advection-diffusion equation,

$$\frac{\partial S}{\partial t} + u \cdot \nabla S = k_h \left( \frac{\partial^2 S}{\partial x^2} + \frac{\partial^2 S}{\partial y^2} \right) + k_\rho \left( \frac{\partial^2 S}{\partial z^2} \right),$$

where $S$ is the salinity, $k_h$ is horizontal diffusivity, and $k_\rho$ is diapycnal diffusivity. The horizontal gradients are taken to be along isopycnals. Rough estimates of $k_h$ and $k_\rho$ are made by assuming weak meridional advection, no temporal change in salinity, and either all isopycnal or all diapycnal diffusivity. These assumptions lead to either

$$u \frac{\partial S}{\partial x} = k_h \left( \frac{\partial^2 S}{\partial y^2} \right) \quad \text{or} \quad u \frac{\partial S}{\partial x} = k_\rho \left( \frac{\partial^2 S}{\partial z^2} \right).$$

Glider observations provide curvatures of salinity in the $y$- and $z$- directions. Central finite-difference of ARGO climatology (1/6°) provides the salinity gradient in the $x$-direction at 134.25°E. As this is a large-scale estimate, salinity fields were first smoothed in the horizontal with a 300 km Gaussian window to eliminate small changes in salinity gradients and curvature. Mean diffusivities are calculated for the core of the NPTW from 12.5°N to 13.5°N and potential density surfaces $23 \text{ kg m}^{-3}$ to $24 \text{ kg m}^{-3}$. The resulting diffusivities are $k_h = 9 \times 10^3 \text{ m}^2 \text{ s}^{-1}$ and $k_\rho = 8 \times 10^5 \text{ m}^2 \text{ s}^{-1}$. 
These estimates are slightly larger than previous estimates of diffusivity in the upper ocean (Ledwell et al. 1998; Whalen et al. 2012) because we assume that diffusion is either all in the horizontal or all in the vertical. For example, if we assume that diffusion is half isopycnal and half diapycnal these estimates would then be $k_h = 4.5 \times 10^3 \text{ m}^2 \text{ s}^{-1}$ and $k_\rho = 4 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$, and similar to other diffusivity estimates.

1.6 Conclusions and discussion

Glider observations of temperature and salinity provide NEC mean geostrophic velocity, water mass structure, variability, and a relationship between fine-scale and large-scale thermohaline structure. The NEC transport from 8.5°N to 16.5°N and the surface to 27.3 kg m\(^{-3}\) is 37.6 ± 15.6 Sv. Identified water masses are the subsurface NPTW and NPIW, and the surface NPTSW. The standard deviation between glider sections indicates the variability of NPTSW due to ENSO, but low variability of the NPTW and NPIW. There is high fine-scale salinity variance in both NPTW and NPIW. Averaged over latitude, the horizontal Cox number is uniform as a function of potential density, indicating that fine-scale salinity variance is enhanced along isopycnals with mean salinity gradients at larger scales.

Glider observations of the NEC provide an estimate of transport variability during the 4.5 year period. Time series of zonal geostrophic velocity shallower and deeper than 26 kg m\(^{-3}\) show that the NEC has transient shifts due to mesoscale eddies and that there are persistent undercurrents at 9.6°N and 13.1°N. The existence of eastward flowing undercurrents greater than 5° latitude from the equator have previously been observed with ARGO floats in both the North and South Pacific (Cravatte et al. 2012; Qiu et al.
2013b), and may be the result of Rossby wave interactions (Qiu et al. 2013a). However, observations of these undercurrents have been scarce, and details of their dynamics and the relationship to overlying currents have yet to be explored. The NEC transport shallower than 26 kg m$^{-3}$ remains relatively uniform by latitude and is less variable than the deeper transport. Deeper than 26 kg m$^{-3}$, the transport variability is related to the latitudinal extent of the undercurrents either from the changes in undercurrent breadth or the meandering of the undercurrent north and south of the glider range. The coherence between the transport of the overlying NEC and the eastward undercurrents suggests that their transports are dynamically linked.

The velocity of the NEC, the undercurrents, and mesoscale eddies modify the thermohaline structure of the water masses that are set at the surface (Schneider 2000; Sasaki et al. 2010). Here, spice variance is elevated along certain isopycnals. These isopycnals tend to have larger salinity variance at all wavenumbers, as indicated by the spectra (Figure 1.14). The horizontal Cox number quantifies a relationship between the salinity variance at different length scales. Both NPIW and NPTW have elevated fine-scale salinity variance, providing further evidence that spice variance can be used to identify water mass sources (Todd et al. 2012). These water masses exist along isopycnals with large-scale mean salinity gradients confirming that variance is caused by stirring of the large-scale gradients by mesoscale eddies (Ferrari and Polzin 2005; Smith and Ferrari 2009). Furthermore, averaging the horizontal Cox number by latitude shows that the stirring of large-scale gradients is relatively uniform throughout the water column in the NEC, leading to a proportional relationship between the fine-scale and large-scale salinity variance.
Between the water masses, along 25.5 kg m$^{-3}$, there is a spice variance minimum indicating that the water may be of older origin in the sense that it has been more thoroughly mixed. Spectral analysis shows that 25.5 kg m$^{-3}$ has decreased spice variance at all length scales and a flatter spectral slope in comparison to other isopycnals. Previous studies on the temporal evolution of spice in the North Pacific, at 3° resolution, have tracked salinity anomalies subducted and advected between 25 kg m$^{-3}$ and 26 kg m$^{-3}$ (Sasaki et al. 2010; Kolodziejczyk and Gaillard 2012), where we find this salinity variance minimum. It is noted that 25.5 kg m$^{-3}$ is also the potential density surface of Subtropical Mode Water (STMW) formation in the North Pacific (Masuzawa 1969; Talley 1988), however, there is no evidence of STMW as a potential vorticity minimum at this latitude (Figure 1.5b). Furthermore, it is expected that any properties of STMW would be eroded by thorough mixing in the gyre before arriving at this location (Bingham 1992; Huang and Qiu 1994). The minimum in salinity variance at 25.5 kg m$^{-3}$ may be due to diapycnal mixing with the more actively ventilated layers above and below and the lack of large-scale salinity gradients in the horizontal. The tightness of the TS diagrams along 25.5 kg m$^{-3}$ supports this interpretation (Figure 1.10).

Owing to its strong westward advection, persistence of its salinity maximum during observations, and its high fine-scale salinity variance, the salinity maximum of NPTW is identified as a useful tracer for the region. An estimate of isopycnal diffusivity for the NPTW is $4.5-9 \times 10^3$ m$^2$ s$^{-1}$, under the rough assumption that anywhere from half to all the erosion of NPTW is caused by isopycnal mixing. The diffusion of the salinity extrema of NPTW as it is advected and stirred by mesoscale eddies must be related to the observed fine-scale salinity variance. A Cox number may relate an eddy diffusivity to a
molecular diffusivity, \( k_h = D \times C \) with \( D \) the molecular diffusivity and \( C \) the Cox number (Osborn and Cox 1972; Moum 1990; Ferrari and Polzin 2005). Here, the horizontal Cox number is of order unity and the relationship between \( C_H \) and eddy diffusivity would be very roughly \( k_h \approx C_H \times 5 \times 10^3 \text{ m}^2 \text{ s}^{-1} \). This relationship is dependent on the length scales used to define \( C_H \), and on the accuracy of our estimate of isopycnal diffusivity.

Repeated observations of the fine-scale water mass structure along with the large-scale velocity fields that shear and strain these water masses provide insight into how thermohaline variability that is set at the surface is mixed into the ocean interior. It remains unclear what amount of spice variance is due to water mass origin and what amount is dynamically linked to mesoscale stirring. In future work, concurrent glider observations of the Mindanao current will be used to examine the downstream modification of NPTW thermohaline structure. Tracking water masses that have distinct salinity extrema and fine-scale salinity variance will advance knowledge of how mixing modifies thermohaline structure along the path of major currents.

1.7 Acknowledgements

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**Figure 1.1:** Trajectories of 10 glider deployments (left) displayed with topography. The glider traveled from 7.5°N to 17°N, completing a round trip in roughly four months. Two sections of the NEC are made for each glider deployment. Gliders made over 5000 profiles from June 2009 to January 2014. The mean depth-average velocity of the glider (right) shows primarily zonal flow; the depth-average velocity is obtained by dead reckoning of glider position.
Figure 1.2: Trajectory of the glider along longitudinal line 134.3°E. The diagrams display glider position as a function of latitude and (a) date and (b) day of year. There were 19 sections of the NEC from 7.5°N to at least 15°N, and from 10°N to 15°N the NEC was sampled at all times of the year.
Figure 1.3: A glider section of the NEC observed from June 2009 to July 2009, showing salinity and potential temperature contours (a) binned in 10 m vertical bins, (b) objectively mapped with an 80 km Gaussian length scale, and (c) interpolated to potential density surfaces. Data is collected on ascent. Tick marks at top indicate the mean locations of glider profiles for (a) and (c). The fluctuations of internal waves visible in the vertically binned data (a) are removed by either objective mapping (b) or interpolating to isopycnals (c).
Figure 1.4: (a) Objectively mapped mean absolute zonal geostrophic velocity with potential density contours (gray) and zero geostrophic velocity contour (black). Negative (blue) indicates westward velocity. (b) Standard deviation of the absolute zonal geostrophic velocity between glider sections with mean absolute zonal geostrophic velocity contours. The greatest strength and variability of the NEC are in the upper 200 m. Deeper than 26 kg m$^{-3}$ there are persistent eastward undercurrents.
Figure 1.5: (a) Buoyancy frequency squared, $N^2$, normalized by its maximum, $4.19 \times 10^{-4}$ \(s^{-2}\), and (b) potential vorticity normalized by its maximum, $9.9 \times 10^{-10}$ \(m^{-1} s^{-1}\), as functions of potential density. Black contours indicate the mean absolute zonal geostrophic velocity interpolated to isopycnals. Potential vorticity is roughly constant along isopycnals across the NEC within the glider range.
Figure 1.6: Time series of mean absolute zonal geostrophic velocity averaged from (a) surface to 100 m, to capture the variability of the near-surface NEC, and (b) 26.7 kg m$^{-3}$ to 27.3 kg m$^{-3}$, to show persistence of eastward flowing undercurrents. The near-surface NEC is affected by current meanders and eddies. Eastward undercurrents are observed in all glider sections.
Figure 1.7: The transport between 8.5°N and 16.5°N of the mean absolute zonal geostrophic velocity integrated by (a) depth, (b) density, and (c) latitude. By latitude, the total transport (black) is separated into transport shallower (blue) and deeper (red) than 26 kg m\(^{-3}\), which denotes the mean potential density surface at the top of the undercurrents. The main transport in the NEC is in the upper 500 m corresponding to potential densities shallower than 27 kg m\(^{-3}\). By latitude, eastward undercurrents cause minima in the total transport. The NEC transport shallower than 26 kg m\(^{-3}\) is fairly constant by latitude.
Figure 1.8: Integrated absolute zonal geostrophic transport for each glider section covering the range from 8.5°N to 16.5°N. Transport is divided into that shallower (blue) and deeper (red) than 26 kg m$^{-3}$. The correlation coefficient between the two transports is $R=0.87$. Total transport (not shown) is the sum of the two. Greater transport variability is observed deeper than 26 kg m$^{-3}$.
Figure 1.9: Objectively mapped mean salinity (a) with potential density contours and (b) interpolated to potential density surfaces with potential temperature contours. Surface and subsurface salinity extrema are used to identify water masses. The salinity maximum of the North Pacific Tropical Water (NPTW), located along potential density surface 23.5 kg m$^{-3}$, is encompassed by the glider range and is used to examine advection and diffusion in the region.
Figure 1.10: Potential temperature/salinity (TS) diagrams of the objectively mapped mean fields colored by (a) latitude and (b) absolute zonal geostrophic velocity, averaged in 0.01 psu and 0.2 °C bins. The diagrams indicate the respective location and zonal advection of the water masses. The TS properties of the undercurrents are visible as eastward advection in (b).
Figure 1.11: Standard deviation between objectively mapped glider sections with contours of the objectively mapped mean for (a) salinity and (b) potential temperature. The greatest standard deviation of salinity and potential temperature are related to interannual forcing associated with ENSO.
Figure 1.12: Fluctuations of potential density along potential temperature surfaces for glider observations from June to July 2009 (see Figure 1.3). Potential density contours are at 0.1 kg m$^{-3}$ increments, with 22 kg m$^{-3}$, 24 kg m$^{-3}$, and 26 kg m$^{-3}$ shown in bold. Tick marks at top indicate the mean location of each glider profile. Compensating temperature and salinity fluctuations, or spice, appear as waviness along a given potential density surface.
Figure 1.13: Mean fine-scale salinity variance per kilometer calculated using the wavelet transform on (a) potential density surfaces with mean salinity contours and (b) depth surfaces with mean potential density contours. The transform was performed from 8.5°N to 15°N on each salinity section, then averaged between sections. The transform includes wavelengths of 10 km to 80 km. The color bar is in logarithmic scale to indicate areas of high salinity variance (red) and low salinity variance (blue). On potential density surfaces, elevated salinity variance is found on the same isopycnals of the NPTW and the North Pacific Intermediate Water (NPIW).
Figure 1.14: Spectra of salinity along isopycnals averaged over all glider sections. Dashed lines indicate spectral slopes of -2 and -1, respectively. Black lines are the normalized upper, mean, and lower bounds of the 90% confidence interval. Potential density surface 25.5 kg m$^{-3}$ has a lower salinity variance at all wavelengths and a flatter spectral slope.
Figure 1.15: (a) Mean large-scale salinity variance per kilometer on potential density surfaces calculated using a wavelet transform. The transform includes wavelengths of 120 km to 200 km. Mean salinity contours are shown in black. (b) The average by latitude of (left) the mean fine-scale salinity variance per kilometer (Figure 1.13a), (center) the mean large-scale salinity variance per kilometer (Figure 1.15a) and (right) the horizontal Cox number, $C_H$, the ratio of the fine-scale to large-scale salinity variance. The average by latitude is taken from 10°N to 13.5°N to avoid edge effects of the wavelet transform. The uniformity of the Cox number through the water column indicates the relationship of fine-scale salinity variance to the stirring of mean salinity gradients.
Figure 1.16: Mean zonal geostrophic velocity from ARGO climatology (1/6° resolution) referenced to 2000 m and averaged in 0.01 psu and 0.2 °C bins at (a) 146.25°E, (b) 140.25°E, (c) 134.25°E, and (d) 128.25°E. Figures follow the flow of the NEC from east (top) to west (bottom). Figure (c) is near the longitude of the glider observations and is comparable to Figure 1.10b. The advection and diffusion of NPTW in the NEC is apparent in its westward velocity and decrease of the salinity maximum from east (a) to west (d).
1.8 References


Chapter 2:

**Mindanao Current and Undercurrent: Thermohaline structure and transport from repeat glider observations**

Autonomous underwater Spray gliders made repeat transects of the Mindanao Current (MC), a low-latitude western boundary current in the western tropical North Pacific Ocean, from September 2009 to October 2013. In the thermocline (<26 kg m\(^{-3}\)), the MC has a maximum velocity core of -0.95 m s\(^{-1}\), weakening with distance offshore until it intersects with the intermittent Mindanao Eddy (ME) at 129.25°E. In the subthermocline (>26 kg m\(^{-3}\)), a persistent Mindanao Undercurrent (MUC), with a velocity core of 0.2 m s\(^{-1}\) and mean net transport, flows poleward. Mean transport and standard deviation integrated from the coast to 130°E is -19 ± 3.1 Sv (10\(^{6}\) m s\(^{-3}\)) in the thermocline and -3 ± 12 Sv in the subthermocline. Subthermocline transport has an inverse linear relationship with the Nino 3.4 index and is the primary influence of total transport variability. Interannual anomalies during El Niño are greater than the annual cycle for sea surface salinity and thermocline depth. Water masses transported by the MC/MUC are identified by subsurface salinity extrema and are on isopycnals that have increased fine-scale salinity variance (spice variance) from eddy stirring. The MC/MUC spice variance is smaller in the thermocline and greater in the subthermocline when compared to the North Equatorial Current and its undercurrents.
2.1 Introduction and background

The Mindanao Current (MC) is a low-latitude western boundary current (LLWBC) in the western tropical North Pacific, formed by the bifurcation of the North Equatorial Current (NEC) in the Philippine Sea (Figure 2.1). The MC flows equatorward along the coast of Mindanao until splitting to contribute to the global overturning circulation through the Indonesian Throughflow (ITF) (Gordon 1986), and to close the interior North Pacific Sverdrup transport via the North Equatorial Counter Current (NECC) (Wyrtki 1961). The MC is the northern source of subtropical water in the western Pacific warm pool (Fine et al. 1994; Lukas et al. 1996) and thus impacts the tropical heat and freshwater budgets and the El Niño-Southern Oscillation phenomena (ENSO) (Gu and Philander 1997; Zhang et al. 1998). In the subthermocline, the MC transports North Pacific intermediate waters equatorward while a Mindanao Undercurrent (MUC) (Hu et al. 1991) flows poleward. The confluence of subtropical, tropical, and intermediate water masses of North and South Pacific origin and their role in salt and heat exchange of the tropical Pacific and Indian Oceans has led to several initiatives in the last two decades to observe and model the MC/MUC system. Observations of the MC/MUC include sea level gauges (Lukas 1988), hydrographic surveys (Hacker et al. 1989; Toole et al. 1990; Hu et al. 1991; Lukas et al. 1991; Wijffels et al. 1995; Kashino et al. 1996; Qu et al. 1998; Firing et al. 2005; Kashino et al. 2009; Kashino et al. 2013), moorings (Kashino 2005; Zhang et al. 2014; Kashino et al. 2015; Hu et al. 2016), and Argo floats (Qiu et al. 2015; Wang et al. 2015). However, MC/MUC water mass transport, coastal structure, and variability are not well understood owing to strong currents, eddy activity, and large annual and interannual changes in surface forcing.
From hydrographic surveys, the MC current is narrow at 10°N, then widens, shoals and increases in speed as it flows equatorward until it separates into the ITF and NECC between 7°N and 6°N (Lukas et al. 1991). Upper-ocean cyclonic recirculation in the Philippine Sea, such as that associated with the quasi-permanent Mindanao Eddy (ME), centered near 7°N, 130°E, causes the MC transport to increase as it flows equatorward (Lukas et al. 1991). At 8°N, the mean MC is in geostrophic balance, extending approximately 200 km offshore, with a subsurface velocity core of 0.8 to 1 m s\(^{-1}\) (Wijffels et al. 1995; Qu et al. 1998). Although the ME has been observed by surface drifters (Lukas et al. 1991), its sporadic presence is difficult to observe in hydrographic transects (Wijffels et al. 1995; Qu et al. 1998; Firing et al. 2005; Kashino et al. 2013). The velocity variability of the MC, its structure near the coast, and the variability of the ME remain inadequately observed.

In the subthermocline, the poleward MUC is roughly 50 to 80 km offshore of the coast of Mindanao near 500 m depth, and potential density of 27.2 kg m\(^{-3}\) (Hu et al. 1991; Lukas et al. 1991; Qu et al. 1998), with a second velocity core occasionally reported further offshore (Hu et al. 1991; Qu et al. 1998). The MUC could be a quasi-permanent recirculation or a poleward current transporting South Pacific water masses; intermittent subthermocline eddies that populate the region on intraseasonal time scales make previous observations inconclusive (Hu et al. 1991; Qu et al. 1998, 1999; Firing et al. 2005; Dutrieux 2009; Chiang and Qu 2013; Wang et al. 2014; Zhang et al. 2014; Chiang et al. 2015; Kashino et al. 2015; Schönau et al. 2015; Hu et al. 2016). The forcing, persistence, and connectivity of the MUC, and its relationship to the thermocline circulation are areas of ongoing research.
The transport of the MC/MUC forms a closed-mass system with the NEC and Kuroshio Current (KC), and their respective undercurrents (Toole et al. 1990; Qiu and Lukas 1996; Qu et al. 1998). The relative transports of the MC and KC depend on the bifurcation latitude of the wind-driven NEC. The bifurcation latitude, dictated by Sverdrup theory as the zero-line of zonally integrated wind-stress curl, is affected by local Ekman pumping and suction caused by shifts of the East Asian monsoon, and the arrival of annual Rossby waves from the Central Pacific (Qiu and Lukas 1996; Kim et al. 2004; Qiu and Chen 2010). Interannual variability of the MC is forced by basin-wide changes during ENSO. Numerical models suggest an increase in MC transport when the NEC bifurcates at higher latitude: annually in the fall and interannually during El Niño (Qiu and Lukas 1996; Kim et al. 2004; Qiu and Chen 2010). Observations have both verified (Lukas 1988; Kashino et al. 2009) and conflicted (Toole et al. 1990) as to the relationship of transport with ENSO. Less is known about the transport variability of the MUC.

Transport variability of the MC/MUC impacts the heat and salt balance in the tropical Pacific. Water masses transported equatorward by the MC are the North Pacific Tropical Water (NPTW), a subsurface salinity maximum in the thermocline, and the North Pacific Intermediate Water (NPIW), a salinity minimum in the subthermocline (Gordon 1986; Bingham and Lukas 1994; Fine et al. 1994; Kashino et al. 1996; Schönau et al. 2015; Wang et al. 2015). Antarctic Intermediate Water (AAIW), saltier than NPIW and of South Pacific origin, is also present in the Philippine Sea and may be carried northward by the MUC (Fine et al. 1994; Wijffels et al. 1995; Qu et al. 1999; Qu and Lindstrom 2004; Schönau et al. 2015; Wang et al. 2015). The volume transport and
modification of these water masses as they transit the Philippine Sea are less well understood. Isopycnal stirring of large-scale salinity gradients introduced by subsurface water masses creates fine-scale salinity variance that leads to isopycnal and diapycnal diffusion (Ferrari and Polzin 2005). Quantifying the relationship between stirring, which increases isopycnal salinity variance, and diffusion, which decreases it, will give insight into heat and salt transfer in the ocean interior and improve parameterizations of mixing in models.

Repeat glider observations of the MC/MUC provide mean water mass transport and variability during a 4-year period. Spray gliders made 16 sections of the MC/MUC from September 2009 to October 2013 under the Origins of Kuroshio and Mindanao Current initiative (OKMC) (Rudnick et al. 2015a). Gliders observed temperature, salinity and depth-average velocity from the surface to 1000 m depth, at horizontal length scales from 6 to 2000 km, during all seasons and one ENSO cycle. Objective maps of geostrophic velocity, salinity, and potential temperature resolve the mean structure and variability of the MC, the MUC, and the ME. Argo climatology is compared to glider observations to assess annual and interannual variability. The thermohaline structure is compared between the MC/MUC and the NEC (Schö nau and Rudnick 2015) to infer water mass connectivity and modification. Gliders provide new observations on the stability of the MC thermocline transport and the relationship between the variable MC/MUC subthermocline transport and ENSO. The paper proceeds with a description of data and analysis methods, the mean and variability of water masses, geostrophic velocity, and transport, and fine-scale thermohaline structure.
2.2 Data collection and analysis

To observe the MC/MUC, the autonomous underwater glider Spray (Sherman et al. 2001; Rudnick et al. 2004) made repeat transects from Palau to the coast of Mindanao. The Spray glider is equipped with a Sea-Bird CTD, Seapoint fluorometer, and is remotely controlled with Iridium satellite. The glider completes a dive from the surface to 1000 m and back over a horizontal distance of 6 km in a cycle time of 6 h. Data is collected during ascent at a vertical resolution of about 1 m and binned into 10 m vertical bins. The depth-average velocity of the glider is obtained from dead-reckoning (Davis et al. 2008; Todd et al. 2011; Davis et al. 2012).

For each mission the glider profiled westward from Palau (134.5°E, 7.5°N) until reaching the 1000 m isobath along the coast of Mindanao at a latitude between 7°N and 10°N, and then profiled on the return to Palau (Figure 2.2a). There were 10 missions, with 16 sections of the MC/MUC completed within the given latitude range (Figure 2.2b). The temporal coverage of the glider was sufficient to observe the MC at all times of the year, with ten sections from September to February, and six sections from March to August (Figure 2.2b). The ENSO phase, as measured by the Nino 3.4 index, was positive (El Niño) from September 2009 to April 2010, and negative (La Niña) from June 2010 to April 2011 and August 2011 to March 2012.

An example transect from the return of mission 11A03601 is shown in Figure 2.3. Salinity with potential temperature contours are plotted on depth and density surfaces as a function of along-track distance from the nearest position to the coast at 9.3°N, 126.31°E, to roughly 130°E. On depth surfaces, internal wave motion causes waviness of isotherms (Figure 2.3a) (Rudnick and Cole 2011). To filter out this high-frequency noise, salinity,
potential density and potential temperature are objectively mapped on depth surfaces (Figure 2.3b). Alternatively, to maintain the horizontal resolution of 6 km, interpolating data to density surfaces also filters out internal wave motion (Figure 2.3c).

To make a mean section, glider observations are linearly projected and objectively mapped onto a mean line extending from the coast of Mindanao, from 8.15°N, 126.61°E to 8.77°N, 130°E (Figure 2.1b, 2.2a), hereafter referred to as the ‘mean line’. Only observations within 80 km of the mean line are included in the map. The objective map is made using a Gaussian autocovariance (Bretherton et al. 1976) with a length scale of 80 km. This length scale is based on the autocovariance calculated from observations of temperature and salinity, and is identical to that used in Schönau and Rudnick (2015).

Geostrophic velocity, calculated from thermal wind equations, is referenced to the depth-average velocity of the glider (Rudnick et al. 2004; Todd et al. 2011; Pelland et al. 2013; Lien et al. 2014b; Pietri et al. 2014). The absolute geostrophic velocity from gliders compares favorably to ADCP measurements from a mooring (Lien et al 2014), satellite sea surface height (Rudnick et al. 2015b), and to Argo climatology referenced to a level of no motion at 2000 m in the tropical Pacific (Schönau and Rudnick 2015).

The mean line extends from the 1000 m isobath approximately 380 km offshore, and was chosen at an angle 80°E of North to be perpendicular to isobaths and contours of Mean Dynamic Topography (MDT) (Figure 1.1b) (Aviso 2014). The mean depth-average velocity is perpendicular to the mean line near the coast where the current is strongest (Figure 2.2a), supporting the choice of the line’s direction. Due to the curvature of the coast, 80 km is the cut-off for inclusion of profiles in the objective map (Figure 2.2a). The inshore coordinates of the mean line were chosen to maximize the number profiles
indicates that there is alongshore shoaling of the thermocline of 20 m within this 80 km range but that it is less than the annual thermocline displacement (Argo 2009; Roemmich and Gilson 2009). There were roughly an equal number of profiles north and south of the mean line to accommodate for the spatial variability. Removing a linear along-shore gradient to account for shoaling does not affect the results of the objective map. The objectively mapped mean is thus a spatial and temporal average.

The variability of the MC/MUC is assessed from the standard deviation between the 10 glider missions (Figure 2.2b), each objectively mapped onto the mean line. Linearly projecting and mapping each glider mission together averages out the linear along-shore gradient, as gliders tended to traverse north then south of the mean line. On four of the missions, the glider trajectory exceeded 220 km from the mean line, and only the nearer transect was included in the map. For these single transects, along-shore thermocline depth variability was less than 10 m. Since strong currents and meanders influence the glider path, excluding observations because of distance from the mean line would exclude temporal variability. The standard deviation between glider missions thus reflects spatial and temporal variability. However, comparisons to Argo climatology in Section 3 and to mooring results in the conclusion indicate that temporal variance exceeds spatial variability within the region of observations.

An example of geostrophic velocity and salinity objectively mapped on the mean line is shown for mission 11A03601 in Figure 2.3d. The MC flows equatorward near the coast (blue) and there are two subthermocline poleward (red) cores offshore that may be expressions of the MUC. The range is extended to 134°E to show the small velocities
east of the mean line, although these may vary by section. The objectively mapped glider missions are used to make composite potential temperature-salinity (TS) diagrams to assess water mass transport, and they are compared to monthly (1°) and annual (1/2°) Argo climatology (Argo 2009; Roemmich and Gilson 2009), and forcing from wind and precipitation to assess annual and interannual variability. Wind products are available from NCEP/NCAR reanalysis (Kalnay et al. 1996; NCEP 1996) and precipitation from CPC Merged Analysis of Precipitation (CMAP) (CPC 1997; Xie and Arkin 1997), each gridded at 2.5° and available from the Earth System Research Laboratory (ESRL).

Isopycnal salinity variance, obtained using a wavelet transform, is used to examine thermohaline structure of the MC/MUC at different length scales and can be used to trace water masses (Todd et al. 2012; Schönau and Rudnick 2015). Following Todd et. al (2012), we use a Morlet wavelet, \( \psi(x) = e^{i2\pi k x} e^{-x^2/2} \), with \( k = 1 \). For each deployment, the transform is performed on the along-track isopycnal salinity that is detrended and interpolated to a 1 km grid. Performing the wavelet transform on the entire mission avoids cutting off low-wavenumber variance near the coast of Mindanao where the glider turns around. The resulting salinity variance is projected for each transect (16) and averaged in 10 km bins along the mean line. The mean salinity variance is an average over all transects. To separate length-scales, different ranges of wavelengths are included in the wavelet transform for fine-scales (10 to 80 km) and large-scales (120 to 200 km). The fine-scale scale ranges from the nearest spacing of glider observations to the decorrelation length scale, and the large-scale begins at the e-folding envelope of the chosen Morlet wavelet.
2.3 Mean structure

2.3.1 Water masses

Water masses are identified by their salinity extrema. The mean salinity has two haloclines (Figure 2.4a); first from the surface fresh layer that overlies the mean section to a subducted subtropical salinity maximum, and second in the transition to fresh intermediate water below the thermocline. The subsurface salinity extrema are greatest near the coast as displayed in a TS diagram (Figure 2.5a). The surface fresh layer that overlies the section is North Pacific Tropical Surface Water (NPTSW) (<34.1 psu and >28 °C) (Wyrtki and Kilonsky 1984), formed by local precipitation and advection (Delcroix and Hénin 1991). NPTSW is influenced by annual variability of the East Asian monsoon and interannual variability associated with ENSO (Delcroix and Hénin 1991; Li et al. 2013). NPTSW creates a large vertical salinity gradient within the upper 50 m that strongly impacts the western Pacific warm/fresh pool (Cronin and McPhaden 1998; Delcroix and McPhaden 2002). Beneath NPTSW is the subsurface salinity maximum of North Pacific Tropical Water (NPTW) (> 34.95 psu), near the coast and centered at potential density 23.5 kg m\(^{-3}\). NPTW is formed in the subtropics due to excess evaporation (Tsuchiya 1968; Katsura et al. 2013) and advected westward in the NEC and into the MC, freshening as it mixes with surface and intermediate waters along the path of flow (Lukas et al. 1991; Fine et al. 1994; Qu et al. 1999; Li and Wang 2012; Schönau and Rudnick 2015). In the subthermocline, the subsurface salinity minimum of North Pacific Intermediate Water (NPIW) (<34.4 psu), formed in the Okhotsk Sea (Talley 1993), is at potential density 26.55 kg m\(^{-3}\) in the MC (Bingham and Lukas 1994). Like NPTW, the extrema of NPIW is near the coast, however, its source may be either from
the NEC or more directly from the equatorward Luzon Undercurrent beneath the KC (Bingham and Lukas 1994; Fine et al. 1994; Qu et al. 1997; Schönau et al. 2015). Antarctic Intermediate Water (AAIW), potential density of 27.2 kg m$^{-3}$ and salinity greater than 34.5 psu, is denser and saltier than NPIW (Figure 2.5a) and is commonly identified by its higher oxygen (Qu and Lindstrom 2004). The presence of AAIW and its pathway into the Philippine Sea has been recorded by several observational studies (Fine et al. 1994; Wijffels et al. 1995; Qu et al. 1999; Qu and Lindstrom 2004; Zenk et al. 2005), and a similar TS curve for AAIW is observed here. Lower oxygen intermediate waters that have mixed in the tropics, such as the North Pacific Tropical Intermediate Water (NPTIW), are found east of 130°E (Bingham and Lukas 1995). The poleward transport of AAIW has been used to verify some connectivity between the MUC and southern North Equatorial Undercurrent (NEUC) (Schönau et al. 2015; Wang et al. 2015).

2.3.2 Geostrophic velocity

The mean absolute geostrophic velocity of the MC, referenced to the depth-average velocity, is equatorward, with a subsurface velocity maximum of -1.0 m s$^{-1}$ at 100 m depth (22 kg m$^{-3}$) and 50 km offshore (127°E) (Figure 2.4b,c). Near the coast, where the MC is the strongest, there is advection of NPTW and NPIW (Figure 2.4c). The MC decreases in strength with increasing distance offshore and intersects at the surface with what could be a cyclonic ME roughly 250 km offshore (128.75°E). The core of the possible ME extends through the thermocline to 250 m depth, with similar poleward and equatorward velocities and symmetric isohalines (Figure 2.4c). The mean depth-average velocity is cyclonic over this longitude range (Figure 2.2a). Mean hydrographic sections from Wijffels, et al. (1995) and Qu, et al. (1998) have similar velocity structures as the
glider observed mean MC, although neither of these studies found evidence of a mean ME. The width of the possible ME is roughly 250 km, consistent with that found by surface drifters at this location (Lukas et al. 1991).

In the subthermocline, the equatorward MC extends to 1000 m near the coast, decreasing in velocity with increasing depth and distance offshore (Figure 2.4b). Offshore of the equatorward subthermocline MC is a poleward MUC. The MUC has a maximum velocity of 0.17 m s\(^{-1}\) located 75 km offshore (127.75°E) and 650 m depth (27.2 kg m\(^{-3}\)). East of the MUC there is weak equatorward flow until 128.5°E, then mean poleward flow exists through the offshore edge of the mean line. The coastal structure of the equatorward MC, and mean location and width of the MUC referenced to an absolute velocity are a fundamental improvement to previous observations, which have either been time series at single locations or referenced to arbitrary levels of no-motion. The mean MC/MUC compares favorably with numerical results from Estimating the Circulation and Climate of the Ocean (ECCO) and the Parallel Ocean Program (POP) simulation (Schönau et al. 2015).

2.3.3 Transport

The transport of the MC/MUC across the mean line and from the surface to 27.3 kg m\(^{-3}\) is -22.4 Sv, with a thermocline transport (surface to 26 kg m\(^{-3}\)) of -19 Sv and subthermocline transport (26 kg m\(^{-3}\) to 27.3 kg m\(^{-3}\)) of -3 Sv. Potential density 27.3 kg m\(^{-3}\) is the densest isopycnal that is present in all sections. Potential density 26 kg m\(^{-3}\), at the base of the thermocline, divides the thermocline and subthermocline transport, which encompasses the intermediate water and the MUC (Figure 2.4b,c). This isopycnal has
been used as a reasonable separation of thermocline and subthermocline transports for the NEC (Schönau and Rudnick 2015) and NECC (Johnson et al. 2002; Hsin and Qiu 2012). The Sverdrup transport from the zonally averaged wind-stress curl was -28.3 Sv during this time period, and hydrographic estimates of the mean transport from the surface to 1000 m and coast to 130°E are around -27 Sv (Wijffels et al. 1995; Qu et al. 1998). These estimates exceed the glider transport as they do not consider or resolve the poleward undercurrent. To compare, glider transport summed from the coast approaches -30 Sv before decreasing offshore (Figure 2.6a). The sum of equatorward transport from the coast to 128.5°E and surface to 27.3 kg m⁻³ is -30.4 Sv. In the subthermocline, the poleward transport of the inshore MUC (127.25°E) is 3.6 Sv and the poleward subthermocline transport across the mean line is 8.9 Sv.

As a function of longitude, mean transport is greatest in the thermocline at the location of the MC core, then decreases to zero at 128.5°E (Figure 2.6b). The possible ME, centered at 129.25°E, has a cyclonic circulation of roughly 2 Sv, with 0.5 Sv greater transport poleward than equatorward. The subthermocline transport has an equatorward maximum near the coast, consistent with the subthermocline equatorward MC, then becomes poleward between 127°E and 127.75°E at the mean location of the MUC. As a function of density, mean transport is greatest at the surface between potential density 21 and 22 kg m⁻³ and uniform through the thermocline. Deeper than 27 kg m⁻³, the large standard error indicates that direction of mean net transport is inconclusive (Figure 2.6c).

Binning the volume transport of each objectively mapped mission by potential temperature and salinity, summing over all sections, and normalizing by bin size and the number of sections creates a TS diagram of water mass transport (Figure 2.5b). The TS
transport integrates to the average transport over the 10 missions from the surface to 1000 m depth. Both NPTW and NPIW have equatorward transport. In the subthermocline, there is a net poleward transport of water with potential density 27 to 27.5 kg m\(^{-3}\) and salinity greater than 34.5 psu, typical of AAIW. The poleward subthermocline transport shallower than 27.3 kg m\(^{-3}\) is 5 Sv, earning the MUC the designation as a current, as there is a mean net transport of a distinct water mass.

### 2.4 Variability

Variability in the MC/MUC is forced by basin-wide wind stress, transient eddies, Rossby waves, annual changes in the East Asian monsoon, and interannual changes associated with ENSO. The forcing creates water mass and geostrophic velocity variability between glider missions. Variability is greatest in thermocline depth, surface salinity, and subthermocline transport. The following sections assess the magnitude of variability and the responsible forcing.

#### 2.4.1 Thermocline depth

The standard deviation in potential temperature is greatest in the thermocline, between depths of 100 to 200 m, where the vertical temperature gradient is also a maximum (Figure 2.7a), making heaving of isopycnals the likely cause. Below the temperature mixed layer, which extends to 80 m, isotherms tend to follow isopycnals, causing isopycnal displacement to appear as temperature variability. The thermocline depth (measured as the depth of the layer 23-24 kg m\(^{-3}\)) is at a minimum in March 2010, coinciding with El Niño, and at a maximum in 2013 (Figure 2.8a). Here, a positive
anomaly is shoaling of the thermocline whereas a negative anomaly is deepening of the thermocline.

Annual changes in thermocline depth are forced by the local wind and by Rossby waves generated by wind over the central Pacific (Kim et al. 2004). During the winter, strong northeasterlies over the region create positive wind stress curl and Ekman suction that shoals the thermocline, leading to a cold layer at 100 m depth known as the Mindanao Dome (MD) that extends east to around 150°E between the MC, NEC and NECC (Masumoto and Yamagata 1991). The MD decays in May as the East Asian monsoon shifts to westerlies and the annual downwelling Rossby wave arrives from the central Pacific (Tozuka et al. 2002). Annual thermocline depth variability, taken as the depth of the 21°C isotherm from Argo climatology at 8.5°N is consistent with this forcing (Figure 2.8b). At 128.5°E, the minimum depth is from January to April and maximum from May to November. The annual Rossby wave arrives at this longitude in late summer.

To assess annual and interannual forcing, glider observations can be compared to Argo climatology and a linear, 1.5 layer linear model that allows the propagation of Rossby waves (Kessler 2006; Qiu and Chen 2010). Following Kessler (2006), the depth anomaly of the thermocline is governed by:

$$\frac{\partial h}{\partial t} + c_r \frac{\partial h}{\partial x} + Rh = -\nabla \times \left( \frac{\tau}{f \rho} \right).$$

Here, $R$ is a damping coefficient timescale of roughly 9 months, the long Rossby wave speed is $c_r = -\beta c^2 / f$, and $c$ is the internal gravity wave speed which is taken to be $\sim 2.6-3 \text{ m s}^{-1}$ across the Pacific Basin (Chelton et al. 1998). The solution is given by
where $h_E$ is the depth anomaly on the eastern boundary, and $h$ the depth anomaly on the western boundary. The thermocline depth thus depends on local wind, the propagation of Rossby waves forced by wind-stress curl in the central and eastern Pacific, and Rossby wave radiation from the eastern boundary from coastally trapped Kelvin waves; however, the dominant term is the propagation of Rossby waves generated in the central Pacific (Qiu and Chen 2010). At 128.5°E, the model shows depth variance consistent with annual forcing: deep in summer and fall, and shallow during winter (Figure 2.8c).

The model, Argo climatology, and glider results are compared at 8.5°N and 128.5°E, where the glider and Argo climatology have the same mean thermocline depth (Figure 2.8c). Nearer to the coast, the glider better resolves the coastal structure and has a deeper thermocline than the climatology. The model and Argo climatology have roughly the same magnitude but differ in phase. The model has a stronger annual cycle than Argo climatology (Figure 2.8c) but weaker interannual variability. The glider has the greatest extremes in shoaling and deepening as is expected of synoptic measurements when compared to gridded climatology. Both Argo climatology and gliders observed shoaling of the thermocline during the El Niño in 2009/2010, and deepening in 2013. The shoaling in 2009/2010 is consistent with forcing during El Niño, but the deepening of the thermocline in 2013 is of unknown origin.
2.4.2 Salinity

The standard deviation in salinity is greatest in the upper 100 m (> 0.2 psu), with a secondary variance maximum in the subsurface halocline between the NPTW and NPIW (Figure 2.7b). Where the vertical salinity gradient is large, isopycnal heaving is sufficient to account for the observed salinity variability, as calculated from the depth variance of isopycnals and their mean salinities. The vertical minimum in salinity variance coincides with the low vertical salinity gradient between the two haloclines.

The surface salinity is saltiest during the 2010/2011 El Niño, and freshest at the end of the 2010/2011 La Niña, with low variability in intermittent years (Figure 2.9a).

The annual cycle in surface salinity is less than 0.1 psu (Argo climatology), fresh from August to February and salty from July to August (Figure 2.9b). As evaporation tends to be weaker than precipitation in this region (Cronin and McPhaden 1998), the surface salinity is compared to the regional precipitation (Figure 2.9b). A positive (negative) salinity anomaly lags anomalously low (high) regional precipitation, consistent with the annual migration of the ITCZ with the East Asian Monsoon.

The relatively small magnitude of the annual cycle suggests that glider observed extremes are related to interannual precipitation forcing. Removing the annual cycle, interannual anomalies of precipitation and surface salinity are each large following El Niño and La Niña. The precipitation anomaly lags the Nino 3.4 index, negative following the 2009/2010 El Niño and positive following the 2010/2011 La Niña (Figure 2.9c). The interannual anomaly from the glider (where the annual cycle from Argo climatology has been removed) is correspondingly positive during lack of rainfall and negative with excessive rainfall. The magnitude of the interannual surface salinity anomaly is roughly
five times greater than that of the annual anomaly. During ENSO neutral conditions, when the Nino 3.4 index indicates neither an El Niño nor La Niña state, interannual anomalies of precipitation and surface salinity are small by comparison.

Although these results assess the phase of surface salinity anomalies to precipitation, they are not entirely independent of mixing and advection caused by wind and wind-stress curl. Annually, precipitation follows the East Asian monsoon (Figure 2.9b) and is in phase with wind forcing (Masumoto and Yamagata 1991). In winter and early spring the ITCZ is south of the region, precipitation is small, and positive wind-stress curl forces upwelling. In the summer, the ITCZ creates large precipitation over the region, wind-stress curl changes, and the annual Rossby wave arrives to force downwelling (Figure 2.8b). Interannually, low precipitation and upwelling each have extremes during the El Niño in 2009/2010 (Figures 2.8c, 2.9c). Thus salty anomalies from “outcropping” of denser, saltier water occur when a fresh layer is advected away or from upwelling and mixing with underlying isopycnals. The relationship between surface salinity, upwelling, and advection thus requires a more complete salinity budget than presented here.

2.4.3 Geostrophic velocity and transport

The MC/MUC geostrophic velocity and transport variability are assessed by the standard deviation of geostrophic velocity (Figure 2.7c), and transport as a function of longitude (Figure 2.10) and time (Figure 2.11). The MC core velocity and transport are stable in the thermocline, whereas the subthermocline has a large transport range. Interannual transport variability is more pronounced than annual variability and a relationship between the subthermocline transport and ENSO emerges.
The standard deviation of geostrophic velocity is large where the mean current velocity is strongest, approaching 50% of the mean velocity near the core of the MC and also large in the subthermocline by the coast (Figure 2.7c). However, the standard deviation is less than the mean equatorward MC in both the thermocline and subthermocline. Near the core of the ME (129.25°E) the variability is roughly equal to the mean, suggesting the ME was intermittently observed. There is a vertical minimum in standard deviation near potential density 26 kg m$^{-3}$ that separates the thermocline and subthermocline transport, occurring even near the coast where the current is strong. In the subthermocline, the standard deviation is large on either side of the MUC, typical of changes in location and breadth. The velocity variability in the subthermocline east of the MUC exceeds the mean.

Velocity is integrated over density surfaces to yield thermocline and subthermocline transport for each glider mission (Figure 2.10). In the thermocline, transport is greatest near the coast, tapering to zero between 127.5°E and 129.5°E, and becomes poleward in roughly half the sections (Figure 2.10a). The poleward transport may be an expression of the ME as it is at the same location of large variability near the ME core (Figure 2.7c). The transport in the subthermocline is variable; persistently equatorward near the coast with poleward transport in all sections (Figure 2.10b). Roughly half the sections have multiple poleward cores that previously have been reported as double MUC cores (Hu et al. 1991). However, it is difficult to define multiple cores of the MUC, as it’s unclear if each has net water mass transport or recirculates. For example, mission 11A03601 has symmetric isohalines in the subthermocline between poleward and equatorward velocities, centered at 129°E, suggesting water mass
recirculation (Figure 2.3d). TS transport diagrams are used to assess the net transport of water masses (Figure 2.10c, d, e). Subthermocline transport summed from the coast to 128°E has equatorward transport of NPIW, and a poleward transport of water typical of AAIW (>34.5 psu, 27.2 kg m\(^{-3}\)) (Figure 2.10c). Summing transport from the coast to 129°E (Figure 2.10d) now includes saltier equatorward subthermocline transport (~27 kg m\(^{-3}\), >34.5 psu). However, this TS range has net poleward transport when integrating to 130°E (Figure 2.10e). Thus the subthermocline equatorward flow between 128°E and 129°E (Figure 2.10b) is likely a partial cyclonic circulation from further offshore, as confirmed by cyclonic depth-average velocity centered at 129°E (Figure 2.3e). Such recirculation, typical of eddies, occurs during other glider missions. However, all glider missions had a net poleward subthermocline transport of intermediate water that is saltier than NPIW and typical of that found in the mean (Figure 2.5b). The MUC is thus a persistent current by its net poleward transport of a distinct water mass even if at times it meanders, partially recirculates or interacts with eddies.

Integrating the velocity over the mean line for each glider mission and for each the thermocline and subthermocline layers (Figure 2.11) provides a transport time series. The relative stability of the thermocline transport and variability of the subthermocline transport is apparent in their ranges: about 10 Sv (-24.0 to -13.6 Sv) for the thermocline and 40 Sv (-26.2 Sv to 22.1 Sv) for the subthermocline, four times as larger. The total transport ranges from 4.5 Sv to -40 Sv, fluctuating with the subthermocline (Figure 2.11a). This is in accordance with previous ranges of transport from hydrographic sections of 8 to -40 Sv (Toole et al. 1990; Lukas et al. 1991). Unlike in the NEC where thermocline and subthermocline transport were highly correlated (Schönau and Rudnick
2015), there does not appear to be a coherent relationship between these transports in the MC/MUC.

The MC/MUC transport may have annual and interannual variability (Lukas 1988; Toole et al. 1990; Qiu and Lukas 1996; Qu et al. 2008; Kashino et al. 2009). Plotting transport by day-of-year (Figure 2.11b) the annual cycle is not discernable. The observed range of transport is larger than a model estimated annual transport range of 10 Sv (Qiu and Lukas 1996). Thus the annual cycle may be masked by variability at other time scales. Plotting by Nino 3.4 index, an inverse linear relationship exists between the Nino 3.4 index and subthermocline transport (Figure 2.11c). Transport in the subthermocline is strongly poleward during La Niña and equatorward during El Niño. During ENSO neutral conditions the direction of subthermocline transport tends to be equatorward and less than 8 Sv. The correlation coefficient between the Nino 3.4 index and the subthermocline transport is -0.88, and that between the Nino 3.4 index and the total transport is -0.93. Assuming each glider mission is an independent degree of freedom the correlation coefficient is significant within a 1% confidence interval. It should be noted that only one ENSO cycle was observed and these observations only describe circulation near the coast. However, during this time period subthermocline fluctuations were correlated with the Nino 3.4 index and were the leading cause of total transport variability.

The depth penetration of the equatorward MC appears to be the source of the subthermocline transport difference between El Niño and La Niña. Both directions of subthermocline flow were observed during each event (Figure 2.10b), but net equatorward transport was 24 Sv greater on average during El Niño than during La Niña,
compared to a 9 Sv difference in poleward transport. Composite TS transport diagrams during each ENSO phase show the distribution of water mass transport (Figure 2.12). During neutral conditions (Figure 2.12a) transport is similar to the mean (Figure 2.5b). El Niño (Figure 2.12b) and La Niña (Figure 2.12c) have similar thermocline transports and notably different surface and subthermocline transports. During El Niño there is a lack of fresh NPTSW, and greater equatorward than poleward transport in the subthermocline. Although there was poleward subthermocline velocity during El Niño (June and December 2009, Figure 2.10b), the lack of net poleward transport (Figure 2.12b) suggests that there is recirculation or that poleward transport moved offshore of the mean line. During La Niña there was weak equatorward flow of NPIW and large poleward transport of saltier intermediate water across the mean line. The subthermocline transport was thus negative during El Niño because of net equatorward transport of NPIW, and positive during La Niña because of net poleward transport of water typical of AAIW, a significant insight into the interannual transport variability of the MC/MUC system.

2.5 Isopycnal salinity variance

The thermohaline structure of the MC/MUC at fine-scales can be related to large-scale gradients by comparing salinity variance on isopycnals, separated by length scale with a wavelet transform. At fine-scales (10 < 80 km) (Figure 2.13a) salinity variance per kilometer is large from the surface to 25 kg m$^{-3}$, encompassing the NPTW, a minimum at the base of the thermocline from 25.2-25.5 kg m$^{-3}$, and large in intermediate water deeper than 26 kg m$^{-3}$. The large-scale salinity variance (120 < 200 km) has a similar vertical structure (Figure 2.13b). At all scales, salinity variance is greatest near the coast, where
the subsurface salinity extremes of the NPTW and the NPIW create large-scale horizontal salinity gradients.

The ratio of fine-scale to large-scale salinity variance, a type of horizontal Cox number (Osborn and Cox 1972), is relatively constant through the water column, with the absolute value dependent on the range of wavelengths included in the transform (Figure 2.13c). A vertically constant horizontal Cox number was also observed across the NEC (Schönau and Rudnick 2015). Both results confirm a direct relationship between the fine-scale and large-scale thermohaline gradients, where the stirring of large-scale gradients causes fine-scale salinity variance.

### 2.6 Water mass modification

Water masses are modified as the NEC advects them into the MC and the MUC advects them into the southern NEUC (Schönau et al. 2015; Wang et al. 2015). Changes in salinity are caused by horizontal and diapycnal diffusion. Salinity decreases and density increases for the subsurface salinity maximum of NPTW (Figure 2.14a). Similarly, the salinity minimum of the NPIW becomes saltier and shifts to a denser isopycnal. The downward flux of salt and density between the warm, salty NPTW and cool, fresh NPIW is typical of double diffusion (Schmitt 1981). However, the well-mixed layer between the two (shaded) has a density ratio such that turbulence caused by internal wave shear is likely a dominant mechanism of diapycnal mixing.

Changes in fine-scale salinity variance can be used to track diffusion of water masses. Fine-scale salinity variance decreases along the isopycnals of NPTW between the NEC and MC (Figure 2.14b). In the subthermocline, salinity variance is greater in the
MC/MUC than in the NEC/NEUC. This corresponds to fine-scale salinity gradients that are introduced by the stirring of the fresh NPIW with saltier intermediate water. A salinity variance minimum near 25.5 kg m\(^{-3}\) between subtropical and intermediate water masses is a notable feature in both the NEC and MC (Figure 2.14b). The salinity variance minimum decreases along the path of flow, similar to NPTW, but shifts from 25.6 kg m\(^{-3}\) to 25.4 kg m\(^{-3}\), opposite the direction of flux by the salinity extrema, which shifts to denser isopycnals (Figure 2.14a). It’s unclear if this is remnant of sampling or would apply in other regions. The cause of the variance minimum is unknown. It is within an isopycnal range that has a low horizontal salinity gradient across much of the North Pacific, as observed by Argo climatology, and where interannual climatic variability at the surface has been observed to cause temporal spice variance (Yeager and Large 2007; Sasaki et al. 2010). A step in salinity of 0.08 psu was observed in this isopycnal range in 2012 but was likely observable only because of the small spatial variance along this isopycnal. The persistence of the variance minimum, and its shift to a less dense isopycnal, would make it an interesting layer to study mixing.

2.7 Conclusion and discussion

Repeat glider observations resolve the mean thermohaline structure and transport of the MC/MUC and their variability during a four-year period. The results provide new insight into the structure of the currents near the coast, water mass transport, variability and fine-scale structure. The MC is a strong current with a persistent transport of subtropical water masses, extending to a depth of 1000 m near the coast and decreasing in
velocity with distance offshore. In the subthermocline, the MUC is a persistent poleward flow that meanders offshore of the MC, with mean transport of a distinct water mass.

The range of subthermocline transport is four times greater than that of the thermocline and strongly influences the total transport variability. During the ENSO cycle observed, subthermocline transport along the mean line had an inverse linear relationship with the Nino 3.4 index. There was no evident relationship between the thermocline and subthermocline transport and no discernable annual cycle for total transport, although it may have been dwarfed by strong interannual variability. The subsurface salinity extrema of NPTW and NPIW decrease in magnitude between the NEC and MC along the path of flow, as does the salinity variance in the thermocline. The salinity variance of subthermocline water in the MC/MUC is greater than that in the NEC/NEUCs as fresh NPIW is stirred with saltier intermediate water in the tropical Pacific. The relationship between isopycnal salinity at large and small scales is consistent with stirring and diffusion of salinity extrema.

Glider observed mean velocity and standard deviation compare favorably to observations by a single mooring with an ADCP, located at 127.3°E, 8°N from 2010-2012 (Zhang et al. 2014). The mooring was offshore of the MC core, and had a mean velocity and standard deviation of 0.7 ± 0.2 m s⁻¹ at 100 m depth, nearly identical to that in the glider (Figure 2.4b, 2.7c). The inclusion of spatial variability between glider missions thus had minimal impact on the mean velocity and velocity variance. The gliders passed the mooring only a few times making a direct velocity comparison unreasonable, however, velocity comparisons between gliders and a mooring array near
the Luzon Strait show good agreement between glider derived absolute geostrophic velocity and mooring ADCP (Lien et al. 2014a).

Single moorings in the MC/MUC have observed intraseasonal velocity variability on timescales of 50 – 80 days (Kashino 2005; Zhang et al. 2014; Kashino et al. 2015). In the subthermocline, mooring observations and model results suggest the MUC velocity varies on even shorter timescales and is also affected by pressure gradients from Rossby waves and ENSO (Hu et al. 2016). Subthermocline eddies that have been observed and modeled in the region at these time-scales (Firing et al. 2005; Dutrieux 2009; Chiang and Qu 2013; Wang et al. 2014; Chiang et al. 2015) were observed by the glider (Figure 2.10). However, the relationship between intraseasonal velocity variability and transport variability cannot be determined from the temporal resolution of glider observations.

The relationship between the interannual subthermocline transport variability and ENSO is difficult to assess as only one ENSO cycle was observed. The increase in equatorward subthermocline transport during the El Niño may be related to the vertical structure of NEC bifurcation, the strength of the NEC, recirculation of the NECC around the MD, or forcing from the equator. The bifurcation latitude of the NEC increases with increasing depth, leading to a more northerly origin of the MC subthermocline water (Qu and Lukas 2003). As the bifurcation latitude shifts the vertical structure of the bifurcation also changes, with greater changes in the subthermocline. When the bifurcation occurs at the highest latitude, the subthermocline bifurcates at higher latitude than at the surface (Qu and Lukas 2003; Kim et al. 2004), possibly causing poleward transport to increase more in the subthermocline than in the thermocline. Preliminary investigations with Argo climatology verify that although this occurs during strong ENSO events, there was only a
small northward shift of bifurcation latitude in the subthermocline and no change in the thermocline during the 2009/2010 El Niño. The vertical structure of the MC/MUC transport and its relationship to ENSO has implications for the ITF transport, which is at a minimum and during El Niño and maximum during La Niña (Gordon et al. 1999; Hu et al. 2015). Regional transport changes over a longer temporal range could be investigated with Argo, but would be limited near the coasts.

The high horizontal resolution and absolute reference velocity from repeat gliders observations have provided accurate water mass transport and exchange in the tropics. These observations are useful in a western boundary current where deep, strong currents near the coast pose a challenge to moorings, floats and satellite altimetry. Combining such glider observations with Argo and satellite observations, and incorporating these observations into numerical models would be a powerful tool to examine large-scale water mass transport and heat budgets in the tropical Pacific.

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Chapter 2, in full, has been accepted for publication of the material in the Journal of Physical Oceanography, 2017. The dissertation author was the primary investigator and author of this paper:


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Figure 2.1: (a) Schematic of circulation in the tropical northwestern Pacific. Red is circulation from the surface to the bottom of the thermocline, typically 300 m depth. Main features are the North Equatorial Current (NEC), the Mindanao Current (MC), the Kuroshio, the Indonesian Throughflow (ITF), the North Equatorial Counter Current (NECC), and the Mindanao Eddy (ME). Subthermocline currents (blue) that oppose the direction of thermocline flow are the Luzon Undercurrent (LUC), the Mindanao Undercurrent (MUC) and the North Equatorial Currents (NEUCs). (b) Mean Dynamic Topography (MDT) from Aviso with contours at 0.015 m intervals. Glider observations are objectively mapped onto a mean line from 8.15°N, 126.61°E to 8.77°N, 130°E (thick black line). The mean line is perpendicular to coastal bathymetry and MDT near the coast.
Figure 2.2: (a) Trajectories of the 10 gliders deployed from Palau to the 1000 m isobath along the coast of Mindanao. Arrows show the mean depth-average velocity from the objective map of glider profiles within 80 km of the mean line (dashed box). There are 968 profiles included. (b) Glider missions coded by date and glider serial number: abbreviated year, hexadecimal month, and serial number. The dashed box denotes a 220 km range where observations from each mission are linearly projected and objectively mapped onto the mean line (solid black) to assess variability between missions. Most missions included observations north and south of mean line. Mission 11A03601 (thick) is featured as an example in Figures 3 and 10. (c) Annual coverage of glider missions colored by the Nino 3.4 index. Observations are at all times of the year and ENSO phases.
Figure 2.3: Example section from glider mission 11A03601 during its return from near the coast of Mindanao to Palau (see Figure 2a). Along-track salinity (color) and temperature (contours) from 9.3°N, 126.31°E to 8.22°N, 130°E: (a) on depth surfaces (b) objectively mapped onto depth surfaces and (c) interpolated to potential density surfaces. Tick marks in (a) are at the location of each profile. (d) Geostrophic velocity referenced to the depth-average velocity (color) and salinity (contours) from an objective map of mission 11A03601 onto the mean line. Dark grey lines denote 26 kg m\(^{-3}\) and 27 kg m\(^{-3}\), respectively. The range is extended to 9.15°N, 134°E to show the relatively weak velocities beyond 130°E, denoted by the dashed line. Negative (positive) velocity is equatorward (poleward). (e) Objectively mapped depth-average velocity for mission 11A03601.
Figure 2.4: Mean structure of the MC/MUC from an objective map of all glider profiles within 80 km of the mean line: (a) salinity with potential temperature contours, (b) geostrophic velocity referenced to the depth-average velocity on depth surfaces with potential density contours (gray), and (c) geostrophic velocity referenced to the depth-average velocity interpolated to potential density surfaces with salinity contours. In (b) and (c) equatorward velocity is negative (blue) and poleward velocity is positive (red). Subsurface salinity extrema are nearest to the coast and advected equatorward
Figure 2.5: Potential temperature-salinity (TS) diagrams colored by (a) mean longitude and (b) net transport over the 10 glider missions, objectively mapped onto the mean line and binned by 0.2 °C and 0.01 psu. Water masses are the North Pacific Tropical Surface Water (NPTSW), North Pacific Tropical Water (NPTW), North Pacific Intermediate Water (NPIW), and Antarctic Intermediate Water (AAIW). In (b) transport is summed in each bin then normalized by the bin size and number of sections. Equatorward transport is negative (blue) and poleward transport is positive (red). Water masses that had small net transport from recirculation or were measured infrequently appear as white. The MC has a net equatorward transport of NPTW and NPIW, and the MUC has a net poleward transport of water typical of AAIW.
Figure 2.6: (a) Total transport (black) summed along the mean line and separated into thermocline (blue) (surface to 26 kg m\(^{-3}\)) and subthermocline (red) (26 kg m\(^{-3}\) to 27.3 kg m\(^{-3}\)) transports. Shading is the standard deviation between the 10 missions, (b) Transport per distance for the thermocline (blue), subthermocline (red), and total (black) as a function of longitude. (c) Transport per density along the mean line. For (b) and (c) shading is the standard error, and the legend gives the mean and standard deviation of each layer.
Figure 2.7: Standard deviation of the objectively mapped glider missions (color) with mean contours (black) for each (a) salinity, (b) potential temperature, and (c) geostrophic velocity. Potential density contours (white) are from (a) 22-27 kg m\(^{-3}\) at 1 kg m\(^{-3}\) increments and at (c) 26 kg m\(^{-3}\). The standard deviation includes both temporal and spatial variability over the range of the glider missions in Figure 2b.
Figure 2.8: Observations and model of thermocline depth variability. For each, the depth anomaly is positive for shoaling and negative for deepening. (a) Depth anomaly of potential density layer 23-24 kg m\(^{-3}\) for each glider mission. Shading on date axis indicates observations during El Niño (red) and La Niña (blue) according to the Niño 3.4 index. (b) Annual cycle of 21°C isotherm depth from Argo climatology (Roemmich and Gilson, 2009) along 8.5°N. (c) Depth anomaly of the 21 °C isotherm for each a linear 1.5 layer wind forced model that allows Rossby wave propagation, monthly Argo climatology, and glider observations, all at 8.5°N, 128.5°E. The mean from June 2009 to October 2013 has been removed from each but not an annual cycle. The linear model and Argo climatology were smoothed with a 3-month running mean. The correlation coefficient is 0.6 between the glider and Argo climatology, 0.59 between the glider and model, and 0.52 between the model and Argo climatology.
Figure 2.9: Sea surface salinity and precipitation variability. (a) Salinity anomaly (0-50 m) from each glider mission. (b) Annual anomaly of precipitation and surface salinity (0-50 m) from Argo climatology (Roemmich and Gilson, 2009), each with a 3-month running mean and averaged from 2004-2014. The Argo anomaly is at 8.5°N, 127.5°E, and precipitation is averaged over the region 3.75-13.75°N and 126.25-136.25°E, encompassing rainfall over the NEC, MC, and local recirculation from the NECC. (c) Nino 3.4 index and interannual anomalies of precipitation and glider surface salinity (0-50 m, 127.5°E), where the glider mean and annual Argo anomaly are removed from glider observations. Precipitation is averaged over the same region as in (b) and filtered with a 5-month running mean. The maximum, lagged correlation coefficient between Nino 3.4 index and the salinity anomaly is 0.8, and that between the Nino 3.4 index and precipitation anomaly is -0.75, with the Nino 3.4 index leading the salinity anomaly and precipitation by 1 month. The precipitation anomaly and salinity anomaly have a maximum correlation of -0.65 with a zero time lag. Results depend on the area over which precipitation is averaged.
Figure 2.10: Transport per distance for each glider mission summed from (a) the surface to 26 kg m$^{-3}$ and (b) 26 kg m$^{-3}$ to 27.3 kg m$^{-3}$, where 26 kg m$^{-3}$ separates the thermocline and subthermocline transports. Note the magnitude of the color bars. Black bars in (b) denote glider mission 11A03601, used in (c)-(e). (c) TS transport diagram summed along the mean line from the coast to 128°E to encompass the equatorward MC and first poleward MUC core, (d) from the coast to 129°E to encompass the equatorward MC, first poleward MUC core, and offshore equatorward flow, and (e) from the coast to 130°E. See Figure 3 for a geostrophic cross-section and depth-average velocity of mission 11A03601. See Figure 5b for bin size and normalization of (c)-(e).
Figure 2.11: Transport through the mean line for each glider mission from the surface to 27.3 kg m$^{-3}$ (black) and divided into thermocline (blue) and subthermocline (red) transport by potential density 26 kg m$^{-3}$. Transports for each are shown as a function of (a) time, (b) day of year, and (c) Nino 3.4 index. The thermocline transport is relatively constant compared to the subthermocline transport, which has an inverse linear relationship with the Nino 3.4 index.
Figure 2.12: Composite TS transport diagrams for glider observations during ENSO (a) neutral (b) positive and (c) negative phases. The greatest transport difference between El Niño and La Niña states was in the subthermocline, where equatorward (poleward) transport dominated during El Niño (La Niña). See Figure 5b for bin size and normalization.
Figure 2.13: (a) Fine-scale (10 < 80 km) and (b) large-scale (120 < 200 km) isopycnal salinity variance from the wavelet transform projected, binned, and averaged over glider missions. Black contours are mean salinity. Salinity variance is large in the subtropical and intermediate water masses, separated by a salinity variance minimum. (c) Average over the mean line for each the fine-scale and large-scale salinity variance and their ratio.
Figure 2.14: (a) Mean salinity averaged on isopycnal surfaces for the MC (Figure 4) and
the mean NEC (Schönau and Rudnick 2015). Salinity extremes of NPTW and NPIW are
smaller in the MC than in the NEC. Gray shaded area indicates a density ratio $5 < R_\rho < 10$,
where $R_\rho$ is defined as $\alpha \theta_z / \beta S_z$ and $\alpha$ and $\beta$ are the thermohaline expansion and saline
contraction coefficients, respectively. (b) Fine-scale ($10 < 80$ km) isopycnal salinity
variance for the MC and NEC. Variance is smaller (larger) for the thermocline
(subthermocline) transport in the MC than the NEC. For each (a) and (b) the average for
the MC is across the mean line and that for the NEC is from 9°N to 15°N along 134.3°E.
2.9 References


Chapter 3: Annual and interannual transport variability in the western tropical North Pacific from observations and ocean state estimates

Argo climatology (Roemmich and Gilson 2009) (ARGO) and a Western Pacific Ocean numerical State Estimate (WPOSE) resolve annual and interannual volume transport anomalies in the far western tropical North Pacific in a region bounded by 125°E to 145°E and equator to 16.5°N, covering the regional circulation through the region from January 2009 to August 2016. The mean velocity and thermohaline structure for the WPOSE is verified using ARGO and autonomous glider observations (GO) along 145°E, 134.3°E, and 8.5°N. Annual volume transport anomalies are small north of 8.5°N but can vary by 50% of the mean volume transport between 8.5°N and the equator. Interannual anomalies exceed annual anomalies through the region. An increase in volume transport was observed in the North Equatorial Current (NEC), the Mindanao Current (MC), and the South Equatorial Current/New Guinea Coastal Undercurrent (SEC/NGCUC) beginning in January 2014, strengthening the North Equatorial Counter Current (NECC) until the mature El Niño in 2015/2016. The increase in transport is related to the meridional gradient in thermocline depth centered at 5°N at the southern part of the Mindanao Dome (MD), where extreme shoaling of the thermocline took place beginning with westerly wind bursts in January 2014. A 1.5-layer Rossby wave model captures the depth variability near the equator and north of 8°N but does poorly in the region of the MD.
3.1 Introduction

The western tropical Pacific is the location of the warmest, freshest pool of water in the global ocean. At the surface, high ocean temperatures lead to deep atmospheric convection responsible for westerly wind-bursts that begin interannual El Niño events (Clarke 2014). Beneath the warm, fresh surface layer there is an intersection of subtropical, and intermediate waters of North and South Pacific origin (Fine et al. 1994). Low-latitude western boundary currents carry subducted spice anomalies in the thermocline from the subtropical gyres of the North and South Pacific to the equator, which may additionally force decadal climate variability as they upwell (Gu and Philander 1997; Zhang et al. 1998). The response of the western tropical Pacific to interannual variability has large implications for the Indonesian Throughflow (ITF). The ITF is the only inter-basin tropical exchange of water and contributes to the global overturning circulation (Gordon 1986). The location of warm water between basins has a strong impact on the location of atmospheric convection and thus tropical and global climate systems (Schneider 1998). Therefore seasonal and interannual variability of water transport through the western tropical Pacific strongly affects the heat balance in the tropical Pacific and Indian Ocean.

Large, wind-driven currents transport water through the western tropical north Pacific (Figure 3.1). The North Equatorial Current (NEC) feeds into the region bifurcating near the Philippines into the equatorward Mindanao Current (MC) and poleward Kuroshio Current. The MC is a low-latitude western boundary current carrying subtropical thermocline and subpolar intermediate water along the island of the Mindanao until it splits into the ITF and the North Equatorial Counter Current (NECC)
The NECC, centered near 5°N, flows eastward into the tropical North Pacific and has large annual and interannual variability (Hsin and Qiu 2012b, 2012a; Zhao et al. 2013). The surface-circulation is forced by basin-wide wind stress that creates Ekman pumping leading to Sverdrup transport and by the propagation of Rossby waves from the central Pacific. The annual shift in the East Asian monsoon over the region causes anomalies in thermocline depth from local Ekman pumping and in the surface freshwater layer from the movement of the ITCZ. These annual anomalies are dwarfed by interannual anomalies associated with ENSO that create large shifts in the thermocline depth and surface fresh water.

Beneath the surface and thermocline circulation are notable subthermocline currents (Figure 3.1). The NGCUC feeds northward into the North Pacific and the Equatorial Undercurrent (EUC) flows eastward along the equator. In the Philippine Sea there is a poleward Mindanao Undercurrent (MUC) (Hu et al. 1991), equatorward Luzon Undercurrent (LUC) (Bingham and Lukas 1994; Qu et al. 1997), and eastward North Equatorial Undercurrents (NEUCs) (Qiu et al. 2013). Little is known what part these undercurrents play in volume and heat transport of the region particularly on annual and interannual time scales.

This work is motivated by the strong interannual variability of the region (Roemmich and Gilson 2011; Schönau and Rudnick 2015) and the possible role that transport may play in the warm water build-up along the equator that is considered a precursor to El Niño events (Jin 1997; Meinen and McPhaden 2000; Lu et al. 2017). Interannual thermocline depth anomalies from monthly Argo climatology (Roemmich and Gilson 2009) and interannual anomalies in absolute dynamic topography from Aviso
(Figure 3.2), are large in the western tropical North Pacific; twice the magnitude of annual anomalies. These anomalies show some symmetry around the equator but tend to be greatest in the North Pacific.

During an El Niño event the equatorial thermocline shoals in the western Pacific and deepens in the eastern Pacific. The east-west tilt is the leading mode of thermocline depth variability, and is typical of volume transport into and out of the region; the second mode is asymmetric around the equator and typical of warm water volume (WWV) build up and discharge at the equator (Meinen and McPhaden 2000, Figure 3). However, it is unclear how these modes affect volume transport in the western tropical Pacific. In the South Pacific, thermocline shoaling and increases in zonal volume transport lag the Southern Oscillation Index (SOI) (Kessler and Cravatte 2013). In the North Pacific, increased eastward transport across 156°E cause WWV to build in the equatorial region (Meinen and McPhaden 2001; Lu et al. 2017). The transport increase may be from the NECC, which has sources in both the North and South Pacific, thus playing important role in the distribution of heat in the equatorial region and El Niño events.

This work utilizes observations collected with autonomous underwater gliders (Schönau and Rudnick 2015), Argo climatology (Roemmich and Gilson 2009), and a data assimilative ocean state estimate based on the MITgcm to quantify the mean, annual and interannual volume transport in the far western tropical North Pacific, from January 2009 to June 2016, encompassing two El Niño episodes. The region of interest spans from 125°E to 145°E and equator to 16.5°N, the area with the greatest interannual variability in sea surface height and thermocline depth (Figure 3.2a,b). Glider observations and Argo climatology verify the volume transport and water mass structure from the ocean state
estimates for the NEC, NECC and MC. Transport across currents not resolved by observations, the ITF, South Equatorial Current (SEC), New Guinea Coastal Current (NGCC) and Undercurrent (NGCUC), and Equatorial Undercurrent (EUC), are examined using the ocean state estimates. Mean transport and annual anomalies are compared for transects across large currents. Interannual transport anomalies are compared to the Nino 3.4 index, wind forcing and thermocline depth variability, to understand their relationship with ENSO.

3.2 Data and analysis

Glider observations, Argo climatology and the ocean state estimates are compared along meridional lines in the western tropical North Pacific at 134.3°E and 145°E (Figure 3.2a), capturing transport for the NEC and NECC, and along 8.5°N for the MC transport. The lines, based on locations of glider observations (Figure 3.2b), capture the zonal volume transport into and through the region. The state estimates provide transport in regions not covered by glider observations or Argo climatology, particularly along the equator where geostrophy fails. Monthly wind forcing, gridded at 3/4°, is available from the ERA-Interim reanalysis from European Centre for Medium-Range Weather Forecasts (ECMWF) to examine relevant forcing. Aviso absolute dynamic topography, available from 1993-2015 from CNES (Aviso 2014) is used for interannual sea surface height anomalies.

3.2.1 Glider observations (GO)

Autonomous underwater Spray gliders were deployed from Palau at 7.5°, 134.3°E from June 2009 to January 2014 and made repeat transects across the NEC and MC
The Spray glider is equipped with a Sea-Bird CTD and Seapoint fluorometer. To profile, the glider dives from the surface to a depth of 1000 m over the horizontal distance of 3 km then surfaces over the distance of another 3 km, collecting data every meter during ascent. Observations are binned vertically by 10 m and have a horizontal resolution of roughly 6 km. Depth-average velocity is obtained from dead-reckoning (Davis et al. 2008; Todd et al. 2011) and used as a reference velocity for geostrophic velocity calculations. Gliders made 19 transects of the NEC and 16 transects of the MC at all times of the year and during one full ENSO cycle, making roughly 11,000 profiles and traveling 46,200 km during 4.5 years. Glider observations of the mean water mass structure and transport and their respective variability are described for the NEC in Schönau and Rudnick (2015), and for the MC in Schönau and Rudnick (2017). The mean fields of potential temperature, salinity, potential density and geostrophic velocity are objectively mapped with a Gaussian autocovariance with a length scale of 80 km, and compared to Argo climatology and ocean state estimates. For short, we will refer to glider observations as GO.

3.2.2 Argo climatology (ARGO)

Observations by Argo floats are mapped in space and time as part of the Roemmich and Gilson Argo climatology, described by Roemmich and Gilson (2009). Argo floats are equipped with CTDs, and maintain a drift depth of 1000 m for nine days before making a profile from 2000 m depth to the surface. There have been roughly 25,600 profiles in this region from January 2004 to August 2016, the period over which we examine the climatology. This is over a greater temporal and spatial range than GO, but with lower resolution. The mean climatology for temperature, salinity and depth is
available at 1/6° resolution, annual cycle at 1/2° and monthly anomalies at 1°. For short, we will refer to the Roemmich and Gilson Argo climatology as ARGO. All ARGO geostrophic velocities are referenced to a level of motion at 2000 m, which provides accurate transport in the subtropical gyre (Schönau and Rudnick 2015), but may underestimate transport nearer to topography (Schonau et al. 2015). ARGO geostrophic velocity and volume transport are compared to the ocean state estimates.

3.2.3 Western Pacific Ocean State Estimates (WPOSE)

The Western Pacific Ocean StateEstimates (WPOSE) are made from January 1, 2009 to August 31, 2016 using the Estimating the Circulation and Climate of the Ocean (ECCO) package (Stammer et al. 2002) which uses the MIT General Circulation Model (MITgcm) (Marshall et al. 1997) and its adjoint. The domain extends from 115°E to 170°E, 15°S to 27°N, but the assimilation of temperature and salinity observations from Argo, CTD, XBT and gliders, satellite derived along-track sea surface height (SSH), and gridded sea surface temperature (SST) is limited to the region 122°E to 170°E, 5°N to 20°N. The model resolution is 1/6° with 50 vertical levels. The vertical level spacing varies by depth, increasing from 2.5 m at the surface to 300 m near the bottom, with a maximum depth of 6,500 m. The model open boundary conditions for horizontal velocities, temperature, and salinity are interpolated from the data assimilative Hybrid Coordinate Ocean Model solutions (HYCOM + NCODA 1/12° global analysis) and model forcing fields are obtained from the NCEP/NCAR reanalysis project (Kalnay et al. 1996). The wind, specific humidity, air temperature, precipitation and net downward shortwave and longwave radiation are prescribed at the surface.
ECCO uses a four-dimensional variational (4D-Var) method (Heimbach et al. 2002) that adjusts model control variables to create a model simulation that minimizes the misfit between the simulation and real observations at the places and times of the observations. The model controls are temperature and salinity initial conditions, open boundary conditions, and atmospheric forcing. Each state estimate is a free-running, dynamically consistent simulation conserving heat, salt, mass and momentum. Each state estimate covered 4 months, with all iterations starting from HYCOM/NCODA solutions instead of the optimized state from the previous estimate. The model results are archived as daily averages for temperature, salinity, and absolute horizontal and vertical velocities. The daily output is averaged by month to compare to ARGO and to extract annual and interannual transport anomalies.

3.3 Mean geostrophic velocity and transport

Mean geostrophic velocity, thermohaline structure and volume transport from ARGO, GO and WPOSE are compared for major currents in the far western tropical North Pacific (Figure 3.3a). The comparison is used to validate WPOSE and establish the strength and vertical structure of the mean flow. Cross-sections at 145°E (Figure 3.4) and 134.3°E (Figure 3.5) capture the zonal velocity of the NEC, NECC, SEC/NGCUC and EUC, and at 8.5°N, the meridional velocity of the MC (Figure 3.6). Mean transports and their variability across these lines are summarized in Table 3.1, with thermocline and subthermocline transports separated at 26 kg m\(^{-3}\) (Johnson et al. 2002; Hsin and Qiu 2012b; Schönau and Rudnick 2015). The subthermocline transport includes transport from the 26 kg m\(^{-3}\) to 27.3 kg m\(^{-3}\), where 27.3 kg m\(^{-3}\) is the deepest isopycnal surface that
is consistently shallower than 1000 m across the NEC and MC (Schönau and Rudnick 2015; Schönau and Rudnick 2017). The ITF volume transports across 125°E and 1°N are calculated for WPOSE but not shown.

For each line, transport is binned by potential temperature – salinity (TS) to show the relative transport of water masses (Figures 3.4, 3.5 and 3.6). For ARGO and WPOSE, transport from the surface to 1000 m depth is binned by 0.02 psu and 0.4 °C for each month of the annual cycle. The binned transport is summed over all bins and normalized by the bin size and number of months (12). The TS transport diagram is the expected distribution of water mass transport over the course of a typical year and integrates to the mean transport. For GO, transport from the surface to 1000 m depth for each synoptic transect is binned by 0.02 psu and 0.4 °C, summed, and normalized by bin size and the number of transects, 19 across the NEC at 134.3°E and 16 across the MC near 8.5°N. The TS transport diagrams for GO are thus expected to show greater salinity variance than those from the annual cycle of ARGO or WPOSE as they include a greater number of transects and use synoptic data instead of averages.

3.3.1 145°E

At 145°E, zonal geostrophic velocity shows the major currents that carry water into and out of the far western tropical Pacific (Figure 3.4). The thermocline makes a ridge around 8°N (Figure 3.4a,b). North of the ridge is the westward, surface intensified NEC. South of the ridge, extending to 2°N, is the eastward NECC, cut-off in ARGO (Figure 3.4a) by requirements of geostrophy. In WPOSE (Figure 3.4b) there is westward surface velocity near the equator from the SEC, and further to the south an intensified westward, subsurface core at 200 m from the New Guinea Coastal Current Undercurrent
Eastward flow along the equator in the lower part of the pycnocline is the Equatorial Undercurrent (EUC) (Tsuchiya et al. 1989). There is a subsurface westward jet centered at 3°N below the NECC. Currents in ARGO are slightly stronger than those in WPOSE, but they have similar features.

From 8.5-16.5°N there is a transport of -35.1 Sv for WPOSE and -35.0 for ARGO (Table 3.1). The TS transport diagrams across the NEC have westward advection of subducted salinity extrema (Figure 3.4c,d). In the thermocline, the salinity maximum is from the North Pacific Tropical Water (NPTW) (>35 psu) (Tsuchiya 1968), and in the subthermocline, the salinity minimum is from North Pacific Intermediate Water (NPIW) (<34.4) (Fine et al. 1994). From 2.5-8.5°N, WPOSE has a mean transport of 21.3 Sv and ARGO has a mean transport of 18.9 Sv, with the greatest difference in the subthermocline (Table 3.1). Each has eastward advection of salinity extrema (Figure 3.4e,f). At this latitude, the subducted salinity maximum (>35 psu) may be from saltier South Pacific Tropical Water (SPTW) from the New Guinea Coastal Undercurrent (Fine et al. 1994), as the salinity maximum from NPTW has decayed. Small westward transport deeper than 27 kg m^-3 is observed in each. South of 2.5°N, WPOSE has a mean transport of -5.2 Sv, with net transport westward in the thermocline and eastward in the subthermocline (Table 3.1). The TS transport diagram shows strong westward transport of the saltier thermocline and subthermocline water from the South Pacific (Figure 3.4g), with eastward flow likely a mixture of North and South Pacific water masses.

3.3.2 134.3°E

At 134.3°E, ARGO (Figure 3.5a), WPOSE (Figure 3.5b) and GO (Figure 3.5c) each have a westward NEC north of 8°N. ARGO and WPOSE have an eastward NECC
south of 8°N, and WPOSE shows the westward inflow from the combined SEC/NGCUC along the equator. The yellow marker in Figure 3.5a,b is at the location of Palau, coincidentally at the shallowest thermocline depth between the NEC and NECC. The mean transports between ARGO, GO and WPOSE are in agreement for the NEC (-37.6 to -40.9 Sv) and the NECC (~31 Sv).

TS transport diagrams are compared north of 8.5°N (Figure 3.5d-f) and are also in agreement. Similar to 145°E, there is westward transport of NPTW in the thermocline and NPIW in the subthermocline. In the subthermocline ARGO (Figure 3.5a,d) and GO (Figure 3.5c,d) have net transport eastward in two separate undercurrents. Eastward subthermocline transport in the mean of WPOSE is not clearly distinguishable (Figure 5b,c). As expected, greater salinity variance and extremes are observed in GO as the TS transport diagram includes more transects and synoptic data instead of an average annual cycle used for ARGO and WPOSE.

3.3.3 MC at 8.5°N

At 8.5°N, the MC has strong meridional velocities near the coast of Mindanao (Figure 3.6). The cut-off longitude for velocity and transport is at 130°E. East of this longitude net transport is small (<4 Sv) with large eddy variability (Schönau and Rudnick 2017). ARGO (Figure 3.6a) has a weaker MC than either WPOSE (Figure 3.6b) or GO (Figure 3.6c), and does not resolve coastal structure. Referencing to a level of no motion at 2000 m underestimates the current strength (Schonau et al. 2015). Mean transport is correspondingly low, -14.9 Sv for ARGO (Table 3.1), half of previous observational estimates (Wijffels et al. 1995; Qu et al. 1998). MC velocities are stronger for WPOSE and GO; in each, equatorward flow extends into the subthermocline near the coast.
Offshore, a subthermocline poleward MUC is evident in GO and weaker in WPOSE, as was typical in undercurrents beneath the NEC.

MC mean transport is -25.6 Sv for GO and -31.5 Sv for WPOSE (Table 3.1), in agreement with previous estimates. The difference is in the subthermocline; WPOSE does not resolve the roughly 5 Sv of poleward transport by the MUC (Schönau and Rudnick 2017). GO (Figure 3.6f) and WPOSE (Figure 3.6e) have similar equatorward TS transports for NPIW and NPTW.

3.3.4 ITF at 125°E and 1°N

ITF transport is available from WPOSE as neither GO nor ARGO extend into the region. Across 125°E, where the MC feeds directly into the ITF, there is a mean transport of -9.8 Sv distributed equally between the thermocline and subthermocline (Table 3.1). The mean maximum velocity near -0.7 m s⁻¹ is surface intensified, with westward flow between 4°N and 6°N as the MC turns around the corner of Mindanao (not shown). Weaker eastward flow of roughly 0.2 m s⁻¹ returns some of the flow out of the Celebes Sea into the tropical Pacific. The transport is in rough agreement with that observed through the Makassar Strait, -13.3 Sv from 2004-2009 (Susanto et al. 2012). During El Niño, which occurred twice during WPOSE, the observed transport was -9.2 Sv. The net transport poleward across 1°N is smaller, roughly 2 Sv. Annual and interannual transport anomalies through the ITF will be discussed in the following sections, with a focus on the dominant flow through 125°E.

3.3.5 Summary

There is agreement between mean geostrophic velocities, transport and thermohaline structure between ARGO, GO and WPOSE across the NEC and NECC.
Across the MC, GO and WPOSE are in good agreement. WPOSE does not resolve the subthermocline undercurrent structure observed by ARGO and GO. In the MC, ARGO does poorly near the coast, and underestimates the mean current strength by referencing to a level of no motion where flow may persist; it will be disregarded when considering annual and interannual transport anomalies.

A horizontal view of mean circulation from WPOSE summarizes velocity and salinity in the thermocline (surface-26 kg m\(^{-3}\)) (Figure 3.7a) and subthermocline (26-27.3 kg m\(^{-3}\)) layers (Figure 3.7b). In the thermocline, the NEC flows westward north of 8.5°N. The salinity maximum of NPTW is the most visible in the far northeast and decays as it is advected eastward. The NEC bifurcates between 13°N and 14°N, and velocity increases in the equatorward MC. At the tip of Mindanao, the circulation splits into the ITF and the NECC. The NECC begins in a strong anti-cyclonic structure of the HE. In the HE, the North Pacific flow is met by saltier water coming from the South Pacific, creating a large salinity gradient. As the NECC flows eastward a part recirculates northward back into the NEC, while the main part flows eastward just north of the equator. The general, circulation in the thermocline is in good agreement with previous observational studies of the region.

In the subthermocline, current location and salinity gradients differ. Again, water is advected from the NEC to the MC, ITF, and NECC. The HE is present, but the circulation from the South Pacific doesn't penetrate as far west. At 134.3° the flow splits into a southern branch that feeds to the EUC along the equator, and a northern branch that continues on an eastward trajectory at a latitude of 7°N. The north-south salinity gradient is the meeting of fresh NPIW with intermediate water from the tropical North Pacific.
(NPTIW) and Antarctic intermediate water (AAIW) from the South Pacific (Bingham and Lukas 1995). However, from salinity alone it is difficult to discern the differences between the latter two. WPOSE captures the major currents, but lacks strong undercurrents that have previously been observed. There have been fewer observations of the subthermocline circulation in the region, and it is less well understood than the dominant surface circulation.

3.4 Annual transport and velocity variability

The annual variability in volume transport is assessed by the standard deviation over the annual cycle (Table 3.1). The annual variability of the NEC is small, with the standard deviation less than 15% of the mean transport. Transport variability is lower at 145°E than at 134.3°E, likely from annual recirculation around the Mindanao Dome west of 145°E. Annual variability of the NECC is much greater, with a standard deviation roughly 40% of the mean transport at 134.3°E and roughly 65% at 145°E. The large annual variability in the NECC is likely influenced by SEC/NGCUC transport from the equatorial region, which has a standard deviation twice the magnitude of mean (Table 3.1), suggesting that the net flow from the surface to 27.3 kg m$^{-3}$ alters direction. In WPOSE, the annual variability in MC is small, similar to that of the NEC. The ITF is more variable with a standard deviation of roughly 40% of the mean transport across 125°E. In general, the magnitudes of transport variability from each ARGO and WPOSE are in reasonable agreement.

The annual transport from WPOSE is explored in more detail to look at the relative phase of transport into and out of the region (Figure 3.7c-f). The region is
subdivided into a grid with boundaries along longitudinal lines 125°E, 134.3°E, and 145°E, and latitudinal lines 0°N, 8.5°N and 16.5°N (Figure 3.7a,b). Horizontal transports, summed from the surface to 943 m (depth of the closest grid box near 1000 m), and the vertical transport through the bottom grid box are plotted for each month. The transport convergence in each box is less than 1 Sv. Boxes 1 and 2 are the annual transports into the region north of 8.5°N, and Boxes 3 and 4 are the annual transport between 8.5°N and the equator. The annual cycles in transports (Figure 3.7) are compared to the mean velocity and depths of the isopycnals mapped every third month of the year (Figure 3.8).

Beginning with the NEC, transport westward into 145°E in Box 2 is relatively constant, a minimum in fall and maximum in the first half of the year (Figure 3.7d). NEC transport through 134.3°E has a greater annual cycle from recirculation around the Mindanao Dome (Tozuka et al. 2002), observed in the cyclonic circulation around Palau from January to April caused by shoaling of the thermocline (Figure 3.8a,b). By July (Figure 3.8c), the thermocline in the far western Pacific has deepened and recirculation from the NECC has abated. The increase in transport of the NEC in Box 1 is balanced between transports into the Kuroshio across 125°E, and the MC across 8.5°N (Figure 3.7c).

The annual cycles south of 8.5°N have greater variance (Figure 3.7e,f), with the exception that the inflow from the MC across 8.5°N into Box 2 is relatively constant (Figure 3.7e). The outflow from Box 2 oscillates between the ITF and NECC. In the early spring, there is lower transport eastward and an increase in the ITF across 125°E (Figure 3.7e). Results are consistent with fluctuations in transport through the Makassar Strait from direct measurements, typically smaller from October to December and peak in the
summer months (Susanto et al. 2012). Transport across the equator changes from southward in early spring to northward in the summer months and fall (Figure 3.7e). The change in transport is from subthermocline variability. In the thermocline, SEC/NGCUC velocities are persistently northward across the equator (Figure 3.8a-d). In the subthermocline, weak southward flow from January to April occurs during the weakest thermocline transport. The increase in SEC/NGCUC transport and decrease in ITF transport in late fall causes the NECC transport to increase substantially in the late fall (Figure 3.6e).

The greatest changes in annual transport are in Box 4 (Figure 3.7f). Both zonal and meridional inflow have significant annual cycles which cause the NECC outflow across 145°E to increase from 20 Sv in January to close to 50 Sv by July, with maximum transport sustained from mid-summer to late fall. This increase is in phase with transport increases across 134.3°E and 0°N, and the breakdown of the Mindanao Dome causing water to flow eastward instead of recirculate northward across 8.5°N. The strengthening of the NECC begins in the summer and fall in the thermocline (Figure 3.8e,d) and peaks earlier, between April and July, in the subthermocline (Figure 3.8f,g). The outflow for the thermocline (Figure 3.8b) and subthermocline (Figure 3.8f) are at different latitudes in April, with the surface NECC near 5°N, and the subthermocline contributing to the EUC near the equator. By July and October, the NECC has moved further south and strengthened so that the horizontal structure in each the thermocline and subthermocline layers are more similar (Figure 3.8d,h). The movement and strengthening of the surface NECC is consistent with previous observations of the annual cycle from satellite derived surface velocity (Hsin and Qiu 2012a). There are fewer observations of the
subthermocline circulation in this region but the change in source of the EUC between the north and south Pacific is interesting.

Annual changes in transport are caused by isopycnal gradients that are forced annually by the East Asian Monsoon and the arrival of Rossby waves from the central Pacific (Tozuka et al. 2002; Kim et al. 2004; Susanto et al. 2012). Local wind-stress is northeasterly during the late fall and winter, causing positive wind-stress curl and Ekman suction over the region, and southwesterly during the late spring and summer (Masumoto and Yamagata 1991). The southwesterlies are in phase with the arrival of a downwelling Rossby wave, which deepens the thermocline and thus influences transport (Hsin and Qiu 2012a). Annually, the meridional gradient in isopycnal depth has the greatest variance around 5°N, near the ridge that forms the NECC.

3.5 Interannual variability

3.5.1 Transport

The standard deviation of interannual transport anomalies are given at 145°E and 134.3°E for ARGO and WPOSE, and for WPOSE at 8.5°N, 125°E and 1°N (Table 3.1). In most cases, the interannual variability is equal to or exceeds annual variability, and is greater for WPOSE than for ARGO. For the NEC, the standard deviation of interannual transport is 15-25% of the mean transport, and in the MC it is 25% of mean transport, roughly twice that of the annual cycle. For the NECC, the ARGO shows similar magnitudes of interannual and annual transport anomalies; however, WPOSE anomalies are greater, with a standard deviation of 90% of the total mean transport. In the ITF and
inflow from the equator, annual and interannual anomalies are of roughly the same magnitude.

Interannual transport anomalies are likely related to ENSO, the leading cause of interannual variability in the region. Figure 3.9 shows the interannual transport anomalies across each section together with the Nino 3.4 index. The Nino 3.4 index is defined as the 3-month running mean of the temperature anomaly in the Nino 3.4 region, 5°S-5°N and 190°E-240°E. A correlation between the Nino 3.4 index and each of the interannual transport anomalies is performed at various time lags for each transport anomaly. At 145°E (Figure 3.9a) and 134.3°E (Figure 3.9b) interannual transport anomalies for the NECC (2.5-8.5°N) and the NEC (8.5-16.5°N) lead the Nino 3.4 index, as does the transport from the SEC/NGCUC (5°S-2.5°N). The correlation in the NECC is greatest, approaching 0.6 for ARGO and 0.8 for WPOSE, with maximum correlation leading the Nino 3.4 index at 2-3 month. Transport from the NEC and SEC/NGCUC have lower correlations, with the NEC leading the Nino 3.4 index by 3-4 months, and the SEC/NGCUC leading it by 2 months.

The lead in interannual transport anomalies is clearest for the 2015/2016 El Niño. The transports for the NEC, NECC, and inflow from the equator (SEC/NGCUC) are anomalously large (negative/westward for the NEC and SEC/NGCUC, and positive/eastward for the NECC) between January and June 2014. The increase in net inflow from the SEC/NGCUC was also observed during the 1997/98 El Niño (Ueki 2003). The NECC and equatorial inflow mirror one another and peak in July 2015, and the NEC transport peaks in June 2015 (Figure 3.9a,b), prior to the maximum Nino 3.4
index in late 2015. The NEC at 134.3°E becomes anomalously low (positive anomaly) in late 2015 just as the Nino 3.4 index peaks.

In 2009/2010 El Niño, transport anomalies of the NEC, NECC and SEC/NGCUC were of similar sign, but were in phase or only slightly leading the Nino 3.4 index, suggesting that there were substantial differences between the 2009/2010 and 2015/2016 El Niño events. During La Niña, there was on average weaker transport in the NEC, NECC and equatorial inflow. The MC also leads the Nino 3.4 index with a low correlation. Similar to the NEC, transport anomaly peaked prior to the 2015/2016 El Niño and is in phase during the 2009/2010 El Niño. The ITF had lower transport during El Niño, and was greatest during the ENSO neutral phase in 2012/2013. It was the only section where the transport anomaly did not lead the Nino 3.4 index.

3.5.2 Velocity

The zonal circulation north of 8.5°N has a greater role in the interannual transport variability than in annual transport variability. Annually, the transport variability depends on recirculation in the Mindanao Dome and strengthening of the equatorial inflow, and changes in the ITF. Prior to and during the 2015/2016 El Niño both the northern and southern zonal currents increased in transport to increase the outflow in the NECC. Composites of velocity and thermocline depth averaged from the surface to 27.3 kg m$^{-3}$ for periods of anomalous interannual transport provide insight into the sources of the interannual transport anomalies (Figure 3.10). There is some depth variability in circulation but averaging over the upper ocean layer captures the main features of the circulation.
In the mean (Figure 3.10a), the NEC flows into the MC, which splits into the ITF and NECC. The SEC/NGCUC flows into the NECC from the south around the HE. The NECC is centered at around 4°N, widening to the south as it flows eastward. There is a meridional ridge in isopycnals between 5°N and 10°N, with a slight zonal maximum from the coast of Mindanao to Palau that causes some recirculation by the NECC around Palau. From Jan. 2009- Feb. 2016 (Figure 3.10b), the period leading up to and during the 2009/2010 El Niño, the isopycnal ridge is slightly deeper and wider than in the mean, causing a more southerly route for the NECC. From March 2010-March 2012 (3.10c), during an extended La Niña like conditions, the composite is almost identical to the mean. From March 2012-April 2014 (Figure 3.10d), ENSO neutral conditions, the region is deeper than normal and the isopycnal ridge is not as zonally coherent, leading to two branches of the NECC, and greater recirculation around Palau.

Leading up to and beginning the 2015/2016 El Niño, there are big changes in circulation that correspond to the large transport anomalies described in the previous section (Figure 3.9e). From April 2014-Dec. 2015 (Figure 3.10e), isopycnals shoal between 5°N and 10°N and deepen between the equator and 5°N, creating strong meridional gradients that increase the zonal transport. The NEC, SEC/NGCUC and NECC are notably stronger. From January 2016 to August 2016 (Figure 3.10f), as interannual transport anomalies weaken and change sign, inflow into the NECC from the SEC/NGCUC and NEC decrease. The isopycnal layers deepen in the west, likely from the arrival of a downwelling Rossby wave. Recirculation around Palau increases as the isopycnals between the equator and 5°N shoal, reducing the meridional ridge. The NECC shifts northward and weakens in response.
3.5.3 Thermocline depth

The relative changes in isopycnal depths in the MD region are the likely source of transport increases in the NEC and NECC prior to the mature El Niño in 2015/2016. The meridional gradient in thermocline depth sharpens, deepening between 0-5°N and shoaling between 5-10°N (3.10e). Much of the variance in the western tropical Pacific thermocline depth has been explained by the propagation of Rossby waves into the region, forced by wind-stress curl over the central Pacific (Kessler 1990; Tozuka et al. 2002; Kim et al. 2004). A simple, linear 1.5 layer model that captures these Rossby waves is in good agreement with transport anomalies in the South Pacific (Kessler and Cravatte 2013), and sea surface height anomalies in the bifurcation region of the NEC near 13°N (Qiu and Chen 2010). Following Kessler (2006), the thermocline depth anomaly \( h \) is from a 1.5 layer Rossby wave model is given by

\[
\frac{\partial h}{\partial t} + c_r \frac{\partial h}{\partial x} + Rh = -\nabla \times \left( \frac{\tau}{f \rho} \right),
\]

where the depth is defined as positive downward (for our convention, positive is upward). Here, \( R \) is a damping time-scale typically chosen between 6 months to 2 years and the Rossby wave speed is \( c_r = -\beta c^2 / f \), where \( c \) is \(~2.8\) m s\(^{-1}\) (Chelton et al. 1998). The solution accounts for wave reflection and radiation from the boundary, however, these decay and have little effect on the western Pacific (Fu and Qiu 2002). Western Pacific depth anomalies have the greatest dependence on the propagation of waves forced by wind-stress curl anomalies in the central Pacific. Removing the dominant annual cycle, interannual thermocline depth anomalies from the Rossby wave model are shown in
Figure 3.11a, averaged over given latitude ranges. The waves travel faster near the equator and slower near the poles, seen by the slope of the anomalies. The greatest signal is the deepening of the thermocline in January 2011, concurrent with the beginning of the La Niña event. Shoaling was observed in the MD region (5.25-8.25°N) beginning in January 2013, and persisting through July 2016. Nearer to the equator (5.25-8.25°N) shoaling did not begin until the mature phase of the El Niño, with a persistently deep thermocline from January 2013 to July 2015. The mismatch in thermocline depth anomaly is consistent with a stronger NECC.

Comparisons are made between interannual depth anomalies at 145°E as a function of latitude for ARGO, WPOSE, and the Rossby wave model (Figure 3.11b). For ARGO and WPOSE the depth anomaly is that for the entire upper ocean layer, from the surface to 27.3 kg m$^{-3}$. Similar comparisons could be made for the thermocline (23-24.5 kg m$^{-3}$) but results are generally insensitive to the choice of upper ocean isopycnal layers. Beginning in July 2010, there is deepening that continues during the 2011/2012 La Niña. The depth anomaly in WPOSE and ARGO are stronger than that of the Rossby wave model and continue until January 2014. In January 2014, the depth anomaly shifts from negative (deep) to positive (shallow). WPOSE and the Rossby wave model show the difference in depth anomaly north and south of 4°N, with that to the south remaining deeper than average, and that to the north shoaling. ARGO shoals anomalously across all latitudes, although less south of 4°N.

The agreement between the interannual depth anomalies for ARGO, WPOSE, and the Rossby wave model is latitude dependent. At 145°E, ARGO and WPOSE are in good agreement with a correlation coefficient of 0.83 south of 5.25°N, and equal to or greater
than 0.9 to the north. As WPOSE only assimilates data north of 5°N, it may be expected that there is better agreement north of this limit. The agreement between ARGO and the Rossby wave model is relatively good prior to January 2010, but differences begin to merge beginning with the La Niña in June 2010.

From 5.25-8.25°N, the Rossby model lags ARGO and WPOSE, and lacks the extreme shoaling that begins in January 2014. The correlation coefficient between ARGO and the Rossby wave model is lowest in this latitude range, near 0.4. Similarly, the Rossby wave model does not capture the deepening of isopycnals from 8.25-11.25°N between January 2012 to January 2014 that suggests an increase in upper ocean heat content. ARGO and Rossby wave model have the best agreement at higher latitudes, with a correlation coefficient approaching 0.75. The mismatch is greatest at lower latitudes, and especially pronounced between 5.25-8.25°N, the center of the MD region.

3.6 Discussion

The build-up of warm water in the western Pacific has been thought to be a precursor for El Niño events (Wyrtki 1975), and an increase in transport from the NECC prior to El Niño has been reported in observations of equatorial WWV (Meinen and McPhaden 2001; Lu et al. 2017). However, all El Niño events are somewhat unique and there are several mechanisms that may play a role in WWV accumulation, movement, discharge and recharge along the equator (Wang and Picaut 2004). Some of these mechanisms were observed in the far western tropical Pacific in the lead up to the 2015/2016 El Niño.
Beginning with the 2011/2012 La Niña there were anomalous trade winds across the Pacific that continued until 2013 (Figure 3.12a). This spurred an anomalously deep thermocline (Figure 3.11b) that continued to deepen following the La Niña (Schöna and Rudnick 2017), leading to a large volume of warm water. This deepening was not well captured by the Rossby wave model (Figure 3.11c). In January 2014, westerly wind bursts along the equator began at 150°E, extending briefly to 170°E (Figure 3.12a). The large WWV and the wind bursts suggest that an El Niño would occur. However, warm water over the Indonesian archipelago kept the Walker circulation rooted to the western Pacific, and the reappearance of easterlies later in 2014 (Figure 3.12a) ended the development of the El Niño (McPhaden 2015). Despite the switch in wind anomalies across the rest of the Pacific, the thermocline shoaled in the far western tropical Pacific in January 2014, in phase with the initial westerly wind bursts, and transport of warm water out of the region in the NECC began. By June 2014, the transport anomalies in the NECC were of similar magnitude to the anomaly during the 2009/2010 El Niño, and two times in the SEC/NGCUC (Figure 3.9). The increase in the NECC transport in 2014 came from the SEC/NGCUC, as the NEC did not increase in transport until 2015 (Figure 3.9a). Anomalous NECC transport continued through 2014 and 2015, peaking prior to the Nino 3.4 anomaly in late 2015. The anomalous transport from the NECC continued to force warm water anomalies along the equator into 2015, which may have induced the reemergence of the El Niño in 2015 (McPhaden 2015).

The discrepancy between the equatorial forcing and thermocline depth in the far western tropical Pacific may be from the relationship with ENSO with the annual cycles in wind stress caused by the shift of the East Asian monsoon and the ITCZ (Tozuka et al.
Strong El Niño events are often tied to the second half of the year, typically growing from April when the equatorial region is warmest (Clarke 2014), and tend to follow strong East Asian Monsoon winter events (Yasunari 1987). The NECC transport anomalies observed here began in the spring concurrent with typical increases in the annual cycle. There is also considerable variability of the MD with ENSO events (Tozuka et al. 2002). Warming over the rest of the sub-polar and subtropical Pacific beginning in 2013/2014 could have had some effect on the relationship between the East Asian monsoon and ENSO (Alexander et al. 2002; Bond et al. 2015). However, it’s unclear how these relationships may have influenced the thermocline. Further research into the depth dependence of the transport may yield more insight into the forcing.

The lead in thermocline depth and transport in the far western Pacific is consistent with the correlation of interannual AVISO sea surface height (ADT) anomalies (Figure 3.12b), and interannual ARGO thermocline depth anomalies (Figure 3.12c) with the Nino 3.4 index. Aviso ADT or ARGO thermocline depth (23-24.5 kg m$^{-3}$), each from a 1° by 1° grid, was correlated with the Nino 3.4 index at various offsets ranging from -6 to 6 months. The result is asymmetric around the equator. In the tropical western North (South) Pacific, the thermocline depth leads (lags) the Nino 3.4 index. Both have positive correlations; a shoaling western Pacific is consistent with an increase in temperature in the Nino 3.4 region. Prior to El Niño, eastward travelling Kelvin waves set off by the westerly wind bursts raise the thermocline, reducing the east-west tilt of the thermocline. The lag in equatorial thermocline depth with the Nino 3.4 region likely comes from the second phase of the El Niño event, in which warm water discharges to higher latitudes and cooling at depth creates a zonal lift in the entire thermocline (Jin 1997; Clarke et al. 2002; Clarke 2014).
Thus the phase between the Kelvin waves and the discharge of warm water is not discernable from this type of map.

The same lead-lag correlation can be performed between the Nino 3.4 index and the interannual anomaly in zonal geostrophic velocity, derived from ARGO and referenced to 2000 m (Figure 3.12d). Again there is a north-south asymmetry. The correlation is small in the South Pacific and zonal velocity tends to lag the Nino 3.4 index. In the North Pacific there is greater correlation over the region of the NEC and NECC, approaching 0.6. The correlation coefficient is negative for the NEC, which is a westward current, and positive for the NECC, an eastward current. These results are consistent with an increase in transport observed in the NEC and NECC prior to the large El Niño in 2015/2016, and the lag in transport to the SOI in the South Pacific found by Kessler and Cravatte (2013). In the South Pacific, the variability in transport was well explained by the forcing of winds in the central Pacific using the 1.5 layer Rossby wave model. In the North Pacific, the Rossby wave model was not sufficient to explain the lead in transport that began in January 2014 particularly in the region of the Mindanao Dome.

3.7 Conclusions

ARGO, GO and WPOSE agree on the mean structure, transport, and thermohaline properties across the major currents in the far western tropical North Pacific. WPOSE lacks the subthermocline undercurrents in the region observed by ARGO and GO that may play an important role in variability in the region. However, annual and interannual transport anomalies between ARGO and WPOSE are similar, with WPOSE showing slightly greater interannual variability. The annual transport anomalies are greatest
between the equator and 8.5°N, where changes in the ITF and SEC/NGCUC transports have the greatest impact on the NECC. Annual changes in circulation include the springtime recirculation of the NECC into the NEC around the MD and the increase in NECC transport as the ITF decreases and SEC/NGCUC increases in the fall. Interannually, transport anomalies in the NEC and MC are 2-3 times the magnitude of annual transport anomalies, and in the ITF, SEC/NGCUC and NECC are equal to or exceed them. Increases in transport in the NEC, MC, and SEC/NGCUC drive an exceptionally strong NECC eastward, leading the NINO 3.4 index. The changes in interannual transport depend on the relative meridional and displacement of the thermocline.

The NEC, MC, SEC/NGCUC and the NECC all increased in transport beginning in January 2014, a whole year and a half earlier than the mature 2015/2016 El Niño event. The transport increase followed two years of an exceptionally deep thermocline in the region. In 2014, westerly wind bursts occurred along with strong shoaling of the thermocline centered at 5°N, the southern edge of the MD region. The persistent eastward transport of warm water by the NECC from 2014-2015 may have sustained WWV on the equator and continued to push the equatorial ocean into an El Niño in 2015. The shoaled thermocline and increase in transports began to decay during the peak of El Niño in October 2015. There was poor agreement between the depth anomalies in ARGO and the propagation of 1.5-layer Rossby wave model, leaving an intriguing question as to their source.
3.8 Acknowledgements

Glider observations and analysis were funded by the Office of Naval Research as part of the Origins of Kuroshio and Mindanao Current (OKMC) project through grants N00014-10-1-0273 and N00014-11-1-0429, and Flow Encountering Abrupt Topography (FLEAT) project grant N0014-15-1-2488. We thank the Instrument Development Group at Scripps Institution of Oceanography for all facets of the operations of Spray underwater gliders. We thank Pat and Lori Colin at the Coral Reef Research Foundation for their support in Palau.

The WPOSE was made available by B. Cornuelle and G. Gopalakrishnan and is available upon request. The 1/12 deg global HYCOM+NCODA Ocean Reanalysis was funded by the U.S. Navy and the Modeling and Simulation Coordination Office. Computer time was made available by the DoD High Performance Computing Modernization Program. The output is publicly available at http://hycom.org. Wind-stress from ECMWF ERA-Interim reanalysis is available at http://apps.ecmwf.int/datasets/. Roemmich and Gilson Argo Climatology is available at http://sio-argo.ucsd.edu/RG_Climatology.html. Aviso absolute dynamic topography MADT-H, is available from CLS Space Oceanography Division at http://www.aviso.altimetry.fr/.

Chapter 3, in part is currently being prepared for submission for publication of the material. Schönau, Martha; Rudnick, Daniel L.; Gopalakrishnan, Ganesh; Cornuelle, Bruce D. The dissertation author was the primary investigator and author of this material.
Figure 3.1: Schematic of circulation in the tropical northwestern Pacific. Blue is the circulation from the surface to the bottom of the thermocline, typically 300 m depth at 26 kg m$^{-3}$. Main currents are the North Equatorial Current (NEC), the Mindanao Current (MC), the Kuroshio (KC), the Indonesian Throughflow (ITF), and the North Equatorial Counter Current (NECC). At the equator, the South Equatorial Current (SEC) and seasonally reversing New Guinea Coastal Current (NGCC) enter the region. Subthermocline currents (red, >26 kg m$^{-3}$) that counter the thermocline circulation are the Luzon Undercurrent (LUC), the Mindanao Undercurrent (MUC), and the North Equatorial Currents (NEUCs). In the South Pacific the New Guinea Coastal Undercurrent (NGCUC) and the Equatorial Undercurrent (EUC) (purple), span from the lower part of the thermocline into the subthermocline. Cyclonic circulation occurs around the variable Mindanao Eddy (ME), and persistent anti-cyclonic Halmahera Eddy (HE). In the spring the NECC recirculates into the NEC around the Mindanao Dome (MD). The NGCUC surfaces in the North Pacific with the SEC to feed into the NECC around the HE.
Figure 3.2: Interannual standard deviations of (a) thermocline depth (average depth of the 23-24 kg m$^{-3}$ from monthly Argo climatology, and (b) Aviso Absolute Dynamic Topography (ADT). The black lines indicate where volume transport and thermohaline structure is compared between Argo climatology, glider observations and ocean state estimates.
Figure 3.3: (a) Lines in the western tropical North Pacific where Roemmich and Gilson (2009) Argo climatology (ARGO) (2004-2016), Glider Observations (GO) (2009-2014) and the Western Pacific Ocean State Estimate (WPOSE) (2009-2016) are compared. (b) Trajectories of Spray gliders across the NEC (Schönau and Rudnick 2015) and MC (Schönau and Rudnick 2017) from June 2009 to January 2014.
Figure 3.4: Zonal geostrophic velocity at 145°E across the NEC and NECC for (a) ARGO (ref. to 2000 m), and (b) WPOSE. Composite TS transport diagrams for ARGO for (c) 8.5-16.5°N and (e) 2.5-8.5°N, and WPOSE for (d) 8.5-16.5°N, (f) 2.5-8.5°N and (g) 5°S-2.5°N. Potential temperature and salinity are binned by 0.02 psu and 0.4°C for each month in the annual cycle, summed, and divided by bin size and number of months (12). Positive (red) indicates eastward velocity or transport. Negative (blue) indicates westward velocity or transport.
Figure 3.5: Zonal geostrophic velocity at 134.3°E across the NEC and NECC for (a) ARGO, (b) WPOSE and (c) GO (ref. to depth-average velocity). (d)-(f) Composite Potential Temperature - Salinity (TS) transport diagrams for 8.5-16.5°N encompassing the transport of the NEC. Transport for ARGO and WPOSE are binned according to Figure 3.4. GO TS transport diagram from the 19 objectively mapped sections across the NEC, divided by the bin size (0.02 psu and 0.4 °C) and number of sections. Positive (red) is eastward velocity or transport. Negative (blue) indicates westward velocity or transport.
Figure 3.6: Meridional geostrophic velocity at 8.5°N across the MC for (a) ARGO, (b) WPOSE, and (c) GO. Composite TS transport diagrams for (a) ARGO, (b) WPOSE, and (c) GO from 126.61-130°E. See Figures 3.4 and 3.5 for binning and normalization. Positive (red) indicates eastward velocity or transport. Negative (blue) indicates southward velocity or transport.
Figure 3.7: WPOSE mean velocity and salinity for the (a) thermocline (surface-26 kg m\(^{-3}\)) and (b) subthermocline (26-27.3 kg m\(^{-3}\)). (c)-(f) WPOSE annual transport from surface to 943 m into Box (1)-(4) shown in (a) and (b).
Figure 3.8: WPOSE annual velocity vectors (m s\(^{-1}\)) and isopycnal depth (color (m)) for (a)-(d) the thermocline (surface- 26 kg m\(^{-3}\)) and (b)-(h) the subthermocline (26-27.3 kg m\(^{-3}\)). Every third month is shown. Zonal circulation is dominated by the meridional gradient in isopycnals, with a ridge between the NEC and NECC.
Figure 3.9: ARGO and WPOSE interannual transport anomalies (surface-27.3 kg m$^{-3}$) the Nino 3.4 index and their lagged correlation. Transports are summed along (a) 145°E, (b) 134.3°E, and (c) 8.5°N and 125°E, and are smoothed with a 5-month running mean. Panels give range in longitude or latitude of summation. Colored bars indicate El Niño (red) and La Niña (blue) conditions. The correlation coefficients (ii) and (iv) are the correlation between the Nino 3.4 index and the corresponding interannual transport anomaly in the left panel. Negative (positive) correlation indicates the interannual transport anomaly leads (lags) the Nino 3.4 index. The NEC, NECC, and MC transports all lead the Nino 3.4 index.
Figure 3.10: WPOSE velocity and depth (surface-27.3 kg m$^{-3}$) averaged over periods of anomalous interannual transport (see Figure 3.9). (a) January 2009-August 2016: Mean. (b) January 2009-February 2010: Elevated transport in the NEC and NECC prior to and during the 2009/2010 El Niño. (c) March 2010-March 2012: Extended La Niña. (d) April 2012-March 2014: ENSO neutral conditions and small interannual transport anomalies. (e) April 2014-Dec. 2015: Increase in transport in NEC, NECC and equatorial inflow prior to and during the 2015/2016 El Niño (f) January-August 2016: Return to normal transport conditions following the peak and decay of 2015/2016 El Niño.
Rossby wave model it is 0.55. Between ARGO and the Rossby wave applied to each. The average correlation coefficient between ARGO and WPOSE is 0.9, averaged over the latitude range in each sub-panel.

Figure 3.11: (a) Interannual thermocline depth anomalies (positive-shallower, negative-deeper) for a 1.5 layer Rossby wave model, averaged over the latitude range in each sub-panel. (b) Interannual depth anomalies (m) from (i) ARGO, (ii) WPOSE and (iii) Rossby wave model at 145°E. ARGO and WPOSE are averaged from 21-27.3 kg m⁻³. (c) Comparison of depth anomalies for ARGO, WPOSE and Rossby wave model at 145°E, averaged over the latitude range in each sub-panel. A running mean of 3-months has been applied to each. The average correlation coefficient between ARGO and WPOSE is 0.9, between ARGO and the Rossby wave model it is 0.59, and between WPOSE and the Rossby wave model it is 0.55.
Figure 3.12: (a) ECMWF interannual zonal wind anomalies as a function of latitude. (b) Lead/lag correlation between Aviso Absolute Dynamic Topography (ADT) and Nino 3.4 index. (c) Lead/lag correlation between ARGO thermocline depth (23-24 kg m\(^{-3}\)) and Nino 3.4 index. (d) Lead/lag correlation between ARGO zonal geostrophic velocity (23-24 kg m\(^{-3}\)) and Nino 3.4 index. For (b)-(d) blue (red) indicates the thermocline depth leading (lagging) the Nino 3.4 index. Contours give the correlation, with black positive and yellow negative. For (b) and (c) correlations are only shown if greater than 0.6, and in (d) if greater than 0.3. For (c)-(d) there is an asymmetry between the North and South Pacific.
Table 3.1. Mean transport and transport variability across transects in the western tropical North Pacific

<table>
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<th>Interannual Standard Deviation</th>
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3.9 References


