EDDY AND WIND FORCED SPICINESS VARIABILITY IN A NUMERICAL MODEL ENSEMBLE – FORMATION, PROPAGATION, AND PREDICTABILITY OF DENSITY COMPENSATED ANOMALIES IN THE PACIFIC THERMOCLINE

A THESIS SUBMITTED TO THE GRADUATE DIVISION OF THE UNIVERSITY OF HAWAI’I AT MĀNOA IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF

MASTER OF SCIENCE

IN

OCEANOGRAPHY

DECEMBER 2020

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Keywords: oceanography, spiciness, ocean-atmosphere interaction, Pacific
To my son,

Sebastian Ikaika Miranda,

The little man who has taught me so much.
ACKNOWLEDGMENTS

This work is part of a collaborative research project with the Japan Agency for Marine-Earth Science and Technology’s (JAMSTEC) and the International Pacific Research Center (IPRC). I would like to thank my principal investigator and advisor Niklas Schneider for his patience, persistence and at times perseverance with me as a student. It was a true honor to work and collaborate with such a world renowned scientist. I also thank my committee members. First, Dr. Ryo Furue for providing the genesis of this study with his OFES work. His attention to detail and breadth of knowledge in the field are unmatched. Dr. Jim Potemra, the first physical oceanographer I took a class from and who focused and grounded our committee meetings. Dr. Eric Firing, whose technical expertise, acumen, and mentorship fostered the data analysis in this project and for greatly assisting with my scientific writing. I specially thank Dr. Alma Carolina Castillo-Trujillo. One of my best and truest friends who without I would have stood no chance of making it out of physical oceanographer infancy.
ABSTRACT

Over the last few decades, research on the atmosphere’s response to internal ocean dynamics has matured significantly. This study uses the Japan Agency for Marine-Earth Science and Technology’s (JAMSTEC) Ocean General Circulation Model for the Earth Simulator (OFES) in a high-resolution, eddy resolving configuration, and expands upon published results of a single OFES run by analyzing a five-member ensemble hindcast from January 2005 to December 2014. We identify and describe spatial and temporal variability of density compensated potential temperature and salinity - called spiciness $\chi'$ - in the Pacific’s upper-thermocline below the mixed layer. On the $\sigma_\theta$25 potential density surface, we find i) large temporal spiciness variance in regions of prominent isopycnal temperature gradients of Baja and in the western and eastern Equatorial Pacific ii) propagation of spiciness signals along mean advective pathways, while iii) large eddy induced spiciness variance in the upper-thermocline is limited to areas off Baja and the western equatorial Pacific. Quantitatively, the standard deviation of spiciness over time $\sigma_t(\chi)$ off Baja California due to winds and eddies reach 1.4 K, and 0.8 K respectively. In the eastern equatorial Pacific, $\sigma_t(\chi)$ is up to 0.6 K, and predominantly results of atmospheric forcing associated with ENSO, while eddy induced variance is smaller than 0.1 K. In the western Pacific north of New Guinea $\sigma_t(\chi)$ is 0.5 K, and 0.3 K, due to winds and eddies, respectively.
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CHAPTER 1
INTRODUCTION

1.1 Motivation

The atmosphere’s response to internal ocean dynamics is a relatively new facet of oceanography that has only a few decades of maturity. In terms of climate variability, the primary focus of inquiries has been the rich mesoscale variability of western boundary currents and their impact on the atmosphere (Qiu and Chen, 2010; Ma et al., 2015, 2017). Here, we look at the thermocline of the subtropical and tropical Pacific, and explore low-frequency modulations of density-compensated temperature and salinity anomalies, so called spiciness anomalies, in the presence of mesoscale eddies and internal ocean dynamics.

This investigation is motivated by recent studies that show the tropical Pacific, which is dominated by ocean-atmosphere coupled dynamics associated with ENSO, acts as the primary driver of Pacific wide decadal variability (Zhang et al., 1997; Trenberth et al., 1998; Gershunov and Barnett, 1998). Oceanic processes may explain the underlying linkages between the ocean and the atmosphere as the ocean acts as a low band-pass filter to the atmosphere (Liu and Alexander, 2007; Kilpatrick et al., 2011). Specifically, on decadal (∼7–10 years) timescales, thermocline processes may be important in climate variability. In the North Pacific, for example, low-frequency variability has significant impacts on climate which influences marine ecosystems and associated economics (Miller and Schneider, 2000). In this deeper regime of geostrophic flow within the thermocline, where there is less mixing as compared to the mixed layer, Taguchi and Schneider (2014) hypothesize spiciness of subsurface isopycnals show better predictive potential than the sea surface in the North Pacific mid-latitudes. Temperature and salinity variations in the thermocline may also be a key to understanding low-frequency ocean variation such as the El Niño Southern Oscillation (ENSO) and Pacific Decadal Oscillation (PDO) and the associated effects on global climate.

1.2 Spiciness

Temperature and salinity anomalies can be separated into dynamical and spiciness components. The spiciness component is defined as temperature and salinity anomalies that are density-compensated, and encompass values of conservative temperature ($\theta$) and salinity ($S$) on a neutral surface (Vernis, 1972; Munk, 1981; Jackett and McDougall, 1985; Schneider, 2000; Kilpatrick et al., 2011). For flow confined to a neutral surface, spiciness is a passive tracer, and is useful for characterizing water masses and intrusions, and does not influence fluid flow (Jackett and McDougall, 1985; Lukas, 2001; Schneider, 2004; Lukas and Santiago-Mandujano, 2008; Williams and Follows, 2011; McDougall and Krzysik, 2015).
Constant potential density surfaces, known as isopycnals, are very good approximations of neutral surfaces, and therefore we use $\theta$ and $S$ metrics on an isopycnal for quantifying spiciness and describing how water masses with spiciness signatures spread as a tracer. The notation $(\sigma_\theta)$ represents the value of a water’s potential density minus 1000 kg m$^{-3}$ and is used to identify specific isopycnals. For example, $\sigma_\theta 25$ is density in 1000 kg m$^{-3} + 25$ kg m$^{-3}$. For a spice signature on an isopycnal, we identify negative and positive spiciness signals$^1$ which can be described as “cool-fresh” (negative spiciness) or “warm-salty” (positive spiciness).

As an example of a spiciness variation, consider Fig. 1.1 showing the Eulerian change between two temperature-salinity profiles at 19N 115W West of Baja California taken a year apart. From 2006 to 2007, there is an increase in spiciness as $\theta$ and $S$ increase and shift along the dashed lines of $\sigma_\theta 25$. Specifically in 2006, $\sigma_\theta 25$ is characterized by $\theta = 15^\circ$C and $S = 33.75$ psu, and a year later at the same location and season, $\sigma_\theta 25$ is both warmer with $\theta = 17^\circ$C, and saltier with $S = 34.45$ psu.

The dynamical component is observed as a density perturbation that impacts pressure (Schneider, 2000). Fig. 1.1 is useful for distinguishing the difference between dynamical and spiciness signals. A dynamical signal is observed in the figure by density differences at the model’s depth horizons (marked by dots).

$^1$ $\theta$ or $S$ anomalies on the isopycnal are used to identify spiciness where anomaly is defined as the difference from the mean annual cycle (climatology) in a respective timeseries. The symbol $\chi$ (pronounced kai) is used to denote either $S(t,x,y,\rho)$ or $\theta(t,x,y,\rho)$ depending on the context.
Figure 1.1: Spiciness Temperature-Salinity Diagram Example

Temperature-Salinity diagram at 19N 115W (West of Baja California) in July 2006 (blue) and July 2007 (red) from OFES. Values in T-S space are plotted at the same location, and colors identify different years. Each dot represents a depth level of the model output. A “spiciness” change is movement along the dashed lines. A “dynamical signal” is observed as a density perturbation that impacts pressure at a sample location (dots). Each of the dashed contours represent the boundary of a density layer written $\sigma_\theta$. Note the strong property shifts at the thermocline in vicinity of the $\sigma_\theta 25$ density surface and then beginning of the isothermal layer at larger isopycnal values greater than $\sim \sigma_\theta 26.5$.

1.3 Relevance

Increased attention in spiciness began with Gu and Philander (1997)’s seminal study suggesting that spiciness, or water mass anomalies, govern decadal climate variability and the decadal modula-
tion of ENSO. They hypothesized that advection\(^2\) of spiciness anomalies tied the subtropics to the low-latitudes where they affect the atmosphere and climate variations. Gu and Philander (1997) describe a thermal exchange between the ocean and atmosphere happening as a rapid poleward transport of heat by the atmosphere, and a slow, equatorward transport of thermal anomalies in the thermocline. The influx of water with spice signatures from higher latitudes into the thermocline changes the thermal structure of the tropical oceans downstream, implying that climate in the tropics is modulated to a degree by atmosphere-ocean exchange upstream. Schneider (2004) explored this adjustment and looked at the response of the coupled ocean-atmosphere system to equatorial emergence of upstream spiciness anomalies, artificially generated in the thermocline upstream of the Pacific equatorial upwelling region. The emergence at the surface of these spiciness anomalies was found to rearrange tropical sea surface temperature, winds and thermocline depth, and to modulate ENSO (Schneider, 2004).

### 1.4 Background of Spiciness and Subduction

Mixed layer \(\theta\) and \(S\) attributes are transferred to the interior thermocline at an isopycnal’s outcrop. Outcrops in Fig. 1.2 are where the mixed layer has the same potential density value as an underlying thermocline layer beneath and longitudinally at a given latitude across the ocean, with increasing \(\sigma_\theta\) values poleward.

Wind-driven subduction into the thermocline occurs through net downward pumping and equatorward advection underneath shallower mixed layer waters toward lower latitudes, as shown in Figs. 1.2–1.4 (Stommel, 1979; Luyten et al., 1983; Williams and Follows, 2011). Only waters subducted in late winter escape re-entrainment by the seasonal mixed layer deepening and meridional migration of the outcrop line. Stommel (1979) first described this process as a “Daemon” that subducts into the thermocline waters with late winter properties.

Once in the upper-thermocline, mixing is weak compared to the mixed layer, and the \(\theta\) and \(S\) are shielded from the surface mixed layer.

Once in the thermocline, the evolution of spiciness on isopycnal surfaces is governed by,

\[
\frac{\partial}{\partial t} \chi' + \vec{u}' \cdot \nabla \chi' \approx -\vec{u}' \cdot \nabla \bar{\chi} - \nabla \vec{u}' \chi' + \text{diabatic proc. (1.1)}
\]

where terms associated with the nonlinear pressure dependence of the equation of state for sea water have been omitted as they are small for motions between the sea surface and the upper thermocline. The left hand side (LHS) terms describe the propagation of spiciness anomalies following the mean flow. In Section 1.6 we show that the propagation of spiciness anomalies follow the time averaged geostrophic flow. The first term on the RHS describes how a time varying flow

\(^2\)Advection, by definition, is the movement of a fluid and its associated properties by the flow (Williams and Follows, 2011).
field can generate, \textit{in situ}, $\chi'$ anomalies from anomalous advection displacing property and mean spiciness fronts on isopycnal surfaces. These signals are a means of affecting climate on multi-year time scales, interacting with each-other in both positive and negative feedback loops upon returning to the surface in locations with strong vertical mixing and upwelling such as the equator (Zhang et al., 1998; Schneider, 2000, 2004; Johnson, 2006).

The second term on the RHS includes eddy-induced divergence of the spiciness flux. The effect of this term will be explored in Section 1.8. Diabatic processes play a part in the attenuation of spiciness anomalies but will not be explored and are outside the scope of this study.
Figure 1.3: Williams and Follows (2011) Ocean Ventilation Diagram

Williams and Follows (2011) Diagram depicting ocean ventilation for a subtropical gyre. The base of the ocean mixed layer-thermocline interface is represented by the dashed line. The outcropping of $\sigma_\theta$ surfaces from the thermocline to the mixed layer and atmosphere exposure are separated by thick solid lines, with lighter isopycnal surfaces ($\sigma_-$) lying on top of heavier isopycnals ($\sigma_+$) (Williams and Follows, 2011). Reproduced with permission of The Licensor through PLSclear.


1.5 Spiciness Anomaly Generation in the Mixed Layer

1.5.1 Deep mixed layer with diapycnal mixing

The largest spiciness signals may result from injection of anomalies directly into the upper-thermocline at the ocean mixed layer–thermocline interface. In a 40-year hindcast model, Yeager and Large (2004) studied isopycnal anomaly creation through late-winter diapycnal (between two isopycnals) mixing, and found that the largest magnitude spiciness anomalies were formed in the eastern Pacific between latitudes of 10° – 30°N(S) and equatorward of an isopycnal’s outcropping areas.

Yeager and Large (2004) identified two factors causing this phenomenon. The first is a salinity inversion caused by changes in surface salinity due to non-local evaporation and ocean surface transport which results in saltier waters lying on top of fresher waters in late-winter (Yeager and Large, 2004). This is observed in Fig. 1.5 where the y-axis represents the salinity anomaly on $\sigma_\theta 25.5$ in October, and the x-axis is a difference in salinity value on the surface and at a sample depth of 205 m. Scatter points to the left of $x(0)$ represent fresher surface waters on top of deeper, saltier waters. Points to the right of $x(0)$ represent saltier waters on top of fresher waters, with most of the samples lying on the right side of $x(0)$. 

Figure 1.4: Williams and Follows (2011) Ocean Subduction Diagram

Williams and Follows (2011) Diagram showing subduction of mixed layer waters into the thermocline. The arrows show time integrated flow rates. The difference in the thick, slanted dashed line is the ocean mixed-layer thermocline interface where the $S_{ann}$ represents the annual subduction rate. There is an Ekman down-welling rate ($-w_{ek}$) which accounts for the vertical flux per unit area, and the horizontal flux per unit area ($\vec{u}_H \cdot \nabla H$) passing through the ocean mixed layer - thermocline interface (Williams and Follows, 2011). Reproduced with permission of The Licensor through PLSclear.
The second factor contributing to direct spice injection into the thermocline occurs in winter when local surface cooling dissipates the summertime-formed stabilizing thermal stratification, and wind stress provides additional energy to mix the density field deeper into the pycnocline; eroding vertical temperature gradients (Yeager and Large, 2004).

![Figure 1.5: Yeager and Large (2004) Scatterplot of upper-ocean salinity gradient](image)

*Scatterplot showing the relation between the annual mean upper-ocean salinity gradient defined as $S(0m) - S(205m)$ and the post-mixing (Oct) salinity anomaly on $\sigma_\theta 25.5$. Each point represents a single horizontal grid location in the box $45^\circ - 15^\circ S, 120^\circ - 70^\circ W$ (Yeager and Large, 2004). Note the preponderance of points to the right of $x(0)$ on the $x$-axis representing saltier waters lying on top of fresher. ©American Meteorological Society. Used with permission.*

This diapycnal mixing process does not occur seasonally every year and is balanced by periods of inactivity whereby mean advection upstream results in climatological anomalies of the opposite sign (negative). Therefore in these formation zones positive and negative anomalies are produced by different physical processes on interannual timescales.

Yeager and Large (2004) found two areas in the southeast Pacific’s (SEP) subtropics where salinity on $\sigma_\theta 25.5$ had the largest interannual variability than anywhere else. These two locations, at approx $20^\circ S$ $100^\circ W$ and $20^\circ S$ $75^\circ W$ are well equatorward of the $\sigma_\theta 25.5$ seasonal outcrop region by about $10^\circ$ (~1000 km). These large amplitude spiciness signals on $\sigma_\theta 25.5$ converged to the equator along equatorial exchange pathways and remained as warm-salty or cool-fresh by .2 K and
1.5.2 Meridional shift of the outcrop line

Mixed layer processes affect spiciness of subducted waters. However, the sign of spiciness anomalies on isopycnal surfaces depends critically on the lateral movement of isopycnal outcrop lines induced by diabatic processes. For example, an observed warm-salty signal on $\sigma_\theta$ 24.5 in the western Pacific at 165°E was found by Kessler (1999) to be associated with upstream fresh surface signal in the southeast Pacific from the surface outcrop a few years earlier. This can be understood by considering displacements of the isopycnal outcrop. A high-salinity tongue extends from the $\sigma_\theta$ 24.5 surface outcrop at 15°S–20°S, 120°W–100°W in the eastern Pacific to 5°S–10°S at 165°E. Equatorward of this high salinity tongue in the eastern Pacific, waters are fresher. Therefore a surface flux induced fresh signal on $\sigma_\theta$ 24.5 indicates an equatorward displacement of the outcrop line.

Nonaka and Sasaki (2007) showed that cool sea surface temperature anomalies (SSTAs) formed in the subtropical South Pacific cause an equatorward shift of an isopycnal outcrop line of up to 5° (∼500 km) latitude to an area of climatologically warmer and saltier waters. This induces warm and salty anomalies on the corresponding density surface within the thermocline and explains Kessler (1999)’s observations. A diagram of this local surface cooling at an isopycnal outcrop line is shown in Fig. 1.6. Here, the dashed line is akin to the $\sigma_\theta$ contours in Fig. 1.1. Gray and black thick lines show the T-S relationship in warm and cold years respectively. The intersection of the water mass density with the outcropping isopycnal (dashed line) in a cold year results in a T-S anomaly on the isopycnal that is warmer and saltier than the warmer year (grey) line (Nonaka and Sasaki, 2007). This may be observed as spiciness signals on an isopycnal surface that are positively correlated with opposite-sign surface temperature and salinity anomalies (Nonaka and Sasaki, 2007).
Figure 1.6: Nonaka and Sasaki (2007) Effect of Local Cooling and Salinification at Isopycnal Outcrop

Diagram of the effect of local cooling and salinification on the surface at an isopycnal outcrop line. Notice the difference in the intersection of the dashed line with warmer (thick grey line) and cooler (thick black lines). In cooler years, the intersection of the water mass density with the outcropping isopycnal (dashed line) results in a T-S anomaly on the isopycnal that is warmer and saltier than higher (grey) line (Nonaka and Sasaki, 2007). Note the dashed line is akin to the $\sigma_\theta$ contours in Fig. 1.1. ©American Meteorological Society. Used with permission.

1.6 Signal Propagation and Paths

Large-scale flow below the surface mixed layer is approximately in geostrophic balance,

\begin{equation}
fv_g = \frac{1}{\rho} \frac{\partial p}{\partial x} \tag{1.2}
\end{equation}

\begin{equation}
fu_g = -\frac{1}{\rho} \frac{\partial p}{\partial y} \tag{1.3}
\end{equation}

with Coriolis parameter $f$ and meridional (zonal) velocity $v_g$ ($u_g$) balancing the pressure gradient force $\frac{1}{\rho} \frac{\partial p}{\partial x}$ ($-\frac{1}{\rho} \frac{\partial p}{\partial y}$) with longitude (latitude). This geostrophic approximation is valid except at very low latitudes where momentum advection becomes important, especially in the zonal momentum balance.

For geostrophic flow, pressure is the streamfunction for horizontal flow on level surfaces, and Montgomery potential ($M$) is the streamfunction for horizontal flow on potential density surfaces.
Water parcels move approximately along pressure-corrected potential density surfaces when the motion is adiabatic and changes in depth are small compared to length as found in basin-scale flows (Williams and Follows, 2011). Recent works from Laurian et al. (2009), Sasaki et al. (2010), and Katsura et al. (2013) use $M$ to show transportation of spiciness signals propagating along $M$ streamlines in Pacific Subtropical Gyres. Eqs. 1.4 and 1.5 show geostrophic balance in $M$ coordinates,

\begin{align}
  u_g &= -\frac{1}{f} \frac{\partial M}{\partial y} \\
  v_g &= \frac{1}{f} \frac{\partial M}{\partial x}
\end{align}

Another method of identifying mean flow paths in the insulated upper-thermocline is by calculating the potential vorticity as it is approximately conserved (Pedlosky, 1998; Williams and Follows, 2011). Therefore this variable can be used as a tracer for flow along isopycnals in the thermocline. Schneider et al. (1999a) used mean potential vorticity, calculated from coarse resolution XBT temperature collections from which isotherms were used to identify layer thicknesses,

\begin{equation}
  Q = \frac{\Delta \rho}{\rho_0} \frac{f}{\Delta H}
\end{equation}

where $\rho_0$ is the mean (reference) density, $\Delta \rho$ is the density difference between the two isotherm layers, $f$ is the Coriolis parameter, and $\Delta H$ is the mean vertical distance between the 12°C and 18°C isotherms, giving the isotherm layer thickness.

Schneider et al. (1999a) found temperature anomalies propagating from the isopycnal outcrop in the North Pacific through the thermocline along lines of constant potential vorticity with a speed and path consistent with mean advection and geostrophic flow (approximately 7 cm s$^{-1}$). Schneider et al. (1999a) found these propagation path dynamics along mean advective pathways were valid until approximately 18°N. Equatorward of this latitude, Schneider et al. (1999a) concluded thermal anomalies were a result of low-latitude wind stress curl. Schneider et al. (1999a) notes the sign of these signals was determined by ventilated thermocline dynamics consistent with Luyten et al. (1983)’s ventilated thermocline theory.

The study was not able to show that the advection of temperature anomalies from the mid-latitudes had an effect on the equatorial region. It suggested that anomalies did not propagate the full length of McCreary and Lu (1994)’s subtropical cell (STC), and countered Gu and Philander (1997)’s hypothesis that there was significant coupling between the North Pacific subtropics and the equatorial region through the ocean’s thermocline (Schneider et al., 1999a). Due to the absence of salinity data in the Schneider et al. (1999a) study, analysis of properties on isopycnals in the upper-thermocline was not conducted. This can be problematic, Lukas (2001) notes that conclusions regarding ocean dynamics and thermodynamics based only on temperature data “are potentially very misleading”.
The advent of the Argo array removed this limitation, when World Climate Research Programme’s (WCRP) Climate and Ocean – Variability, Predictability, and Change (CLIVAR) and the Global Ocean Data Assimilation Experiment (GODAE) began an international joint venture for the deployment of profiling floats in the 1990’s. The first float was launched in 1999 and the array has grown to more than 4000 floats providing near-global ocean coverage, with over half sourced by the United States (Gould et al., 2004). With the growth of the Argo array over the last two decades, there now exists far more abundant observations of the ocean’s interior. Each profiling float measures temperature and salinity to depths of 2000 m, and reports position every 10 days, providing unprecedented in-situ data of the ocean’s interior (Traon, 2013; Gould et al., 2004; Roemmich and Gilson, 2009). With Argo spatial coverage and salinity data, Sasaki et al. (2010) diagnostically calculated isopycnal surfaces on a basin scale in the North Pacific and looked at spiciness propagation. Using this technique, Sasaki et al. (2010), and Kolodziejczyk and Gaillard (2012), showed positive and negative spiciness signals with a path and speed along isopycnals via advection by mean geostrophic current. One example from Sasaki et al. (2010) is a cool/fresh $\chi'$ signal propagating from the eastern subtropical North Pacific to the western tropical Pacific shown in Figs. 1.7, 1.8, and 1.9 which was tracked for six years (2003–2008).
Figure 1.7: Sasaki et al. (2010) Spiciness observations comparison 1

An example of a cool/fresh spiciness signal outlined by blue contours at -.03 Practical Salinity Scale -1978 (PSS-78) from Sasaki et al. (2010) in a gridded Argo float product averaged on $25 < \sigma_\theta < 25.5$ isopycnals. The thin white lines represent mean Montgomery Potential isolines, with red dots are spaced at $10^3$ km from the outcrop line along the subduction path and colorbar shading showing average salinity values from 2001–2008. This cool/fresh signal is tracked for 6 years for the entirety of Sasaki et al. (2010)’s data set from 2003–2008. We observe the same subduction pathway east of 160°W and 30°N containing both positive and negative signals propagating at the speed of mean advection with the mean geostrophic current which are discussed in Section 3, and visible in Figs. 3.20 and 3.27. Also, our normalizing of the signals in Figs. 3.29 and 3.33 aid in visualizing signals further in time as they attenuate.
Sasaki et al. (2010)’s panels showing annual mean salinity anomalies ($\times 10$ PSS-78 in colorbar) averaged over $25 < \sigma_\theta < 25.5 \text{kg m}^{-3}$ isopycnals in the Pacific Ocean from an interpolated Argo product. Anomalies are deviations from the mean field computed for the period 2001–2008. Thin black contours denote $1.6 \text{ m}^2 \text{s}^{-2}$ and $3.4 \text{ m}^2 \text{s}^{-2}$ mean Montgomery potential isopleths, and white circles indicate tracer positions calculated from mean velocity fields on isopycnals (Sasaki et al., 2010). A warm/salty spiciness signal in OFES is discussed in Section 3 and observed in Fig. 3.29 on $\sigma_\theta 25$. It is also visible in Sasaki et al. (2010)’s 2007 panel in this figure.
Another comparison of the warm/salty spiciness signal appearing in Fig. 3.29 on $\sigma_\theta$ 25 in OFES with Sasaki et al. (2010)’s observational data from a gridded Argo float product vertically over $25 < \sigma_\theta < 25.5$. Our 2007 warm/salty signal is visible to the left (2007 on y-axis) of the described cool/fresh signal. This is a time-distance diagram of monthly salinity anomaly ($\times10$ PSS-78) averaged vertically over $25 < \sigma_\theta < 25.5$ kg m$^{-3}$ isopycnals and horizontally over the 1.6 m$^2$ s$^{-2}$ and 3.4 m$^2$ s$^{-2}$ mean Montgomery potential isopleths. Horizontal axis is the distance ($\times10^3$ km) from the outcrop line along the path. White lines indicate advection by the mean geostrophic current, and the black line describes movement of the center of mass of tracer particles from (Sasaki et al., 2010).

During the signals’ journey, it is attenuated and subjected to high frequency injection of spiciness anomalies by temporal changes in the velocity fields (Sasaki et al., 2010). However, despite the attenuation of $\chi'$, upon reaching the equatorial Pacific, model studies suggest these signals may affect equatorial upwelling source waters and tropical climate (Fukumori et al., 2004; Schneider, 2004). Sasaki et al. (2010) therefore suggests there is a continuous forcing on the equatorial region from subtropical North Pacific by spiciness variations.
1.7 Link of dynamical signals and spiciness: Anomalous advection forced by anomalous trades

Dynamical signals are governed by basin scale wave dynamics resulting in a change of a water mass' density and propagate at low frequencies via Rossby or Kelvin waves (Schneider et al., 1999a; Lazar et al., 2001; Furue et al., 2018). Temperature and salinity signals that are not density-compensated and can be observed as a pressure perturbation associated with heave or lateral displacements of density contours (Schneider et al., 1999a).

In areas of high spiciness gradients and fronts on the isopycnal these density perturbations and associated currents generate spiciness signals by anomalous advection, a process active along subsurface spiciness fronts in the north and southeast Pacific (Schneider, 2000; Yeager and Large, 2004; Kilpatrick et al., 2011). Lazar et al. (2001) found that these wave processes influenced the amplitude and spatial structure of subsurface anomalies to the extent that an understanding of both dynamical and spiciness signals is necessary to predict teleconnection patterns. In a fully coupled ocean-atmosphere model, Schneider (2000) shows subsurface spiciness generation from surface forcing via anomalous trade winds accelerating or decelerating ocean currents, which in turn displaces property fronts on an isopycnal. Schneider (2000) found $\chi'$ forming through this process in the Central Pacific tropical low latitudes 10–15°N(S) and traveling north(south)-westward with mean geostrophic flow. Upon reaching the Equatorial Undercurrent (EUC), the $\chi'$ signals reverse direction and are advected eastward.

If this anomalous advection occurs as a seasonal or interannual process, a closed cycle may be induced on a decadal time scale. If periodic, both positive and negative feedback loops can result when perturbations up-well to the surface and interaction with the atmosphere occurs (Gu and Philander, 1997; Schneider, 2000, 2004; Kilpatrick et al., 2011).

1.8 Mesoscale Eddy Variability

The dynamics of spiciness discussed so far involve the background large scale flow. However, mesoscale eddies may also impact spiciness evolution, a topic investigated in this thesis. Mesoscale eddies are extensive throughout the ocean and account for more than 50% of its variability (Chelton et al., 2007). Eddies have lifespans on the order of weeks to months and can span hundreds of kilometers. They transport momentum, heat, mass and oceanic tracers, and therefore contribute to general circulation and large-scale water mass distributions (Chelton et al., 2007; Williams and Follows, 2011).

Globally, mesoscale eddies are generated primarily from dynamical instabilities of jets and currents, and not directly as a result of external forcing (Williams and Follows, 2011). This instability in the background flow, which involves the flattening of sloping density surfaces, is termed baroclinic instability (Williams and Follows, 2011). Eddies also result from barotropic instabilities which
extract kinetic energy from horizontal shear flow. For the purposes this study, eddy generation is assumed to be a stochastic process uncorrelated with atmospheric forcing.

Eddy stirring\(^3\) affects tracer spread in the thermocline and occurs predominantly on isopycnals (Williams and Follows, 2011). A consequence of eddy stirring and eventual mixing is the diffusion of tracers down gradients from high to low concentrations, resulting in an increased spatial spread of the tracer.

Eddy motion along isopycnals may lead to eddy induced fluxes of spiciness if a large-scale spice gradient on an isopycnal surface is present. Total time-mean spiciness flux is made up of advection of mean spiciness by mean flow, and eddy flux of spiciness,

\[
\overline{u\chi} = \overline{\bar{u}\bar{\chi}} + \overline{\bar{u}'\chi'}
\]

where \(\overline{u\chi}\) is the total time-mean spiciness flux, \(\overline{\bar{u}\bar{\chi}}\) the advection of mean spiciness by mean flow, and \(\overline{\bar{u}'\chi'}\) the eddy flux.

Numerical investigation of the role mesoscale eddies in spiciness variability requires high resolution models that resolve the underlying dynamics. Furue et al. (2018) used Japan Agency for Marine-Earth Science and Technology’s (JAMSTEC) Ocean General Circulation Model for the Earth Simulator (OFES) to look at the impacts of sea surface salinity (SSS) in an eddy-resolving general circulation model. Using two single OFES model runs with different surface forcing parameters, Furue et al. (2018) found that the dominant response of the thermocline to observed SSS anomalies (\(S'\)) were spiciness variations in the tropical and subtropical pycnocline. These \(S'\) signals were found in OFES to subduct down from the surface into the ocean’s upper-thermocline and upon subduction these perturbations propagated equatorward and west-ward in the Pacific’s subtropical gyres, consistent with the processes described by Gu and Philander (1997) and Sasaki et al. (2010).

Furue et al. (2018) concluded that dynamical signals (changes in pycnocline depth, ultimately resulting from the change in SSS) must also be present but were obscured by mesoscale noise due to eddy-induced variance and therefore these signals were difficult to identify. Furue et al. (2018) explicitly noted that an ensemble of runs would be required to resolve the chaotic nature of the mesoscale eddies and reveal dynamical signals where density and pressure anomalies were present.

1.9 Climate variability and spiciness in the main pycnocline

The most dramatic form of Pacific variability occurs during ENSO events, which involve coupled changes in winds and upwelling (Williams and Follows, 2011). A characteristic during “normal” periods in the eastern Pacific are surface waters being relatively cool, which is sustained through

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\(^3\)Stirring as a physical process describes how patches of tracer are drawn into narrow filaments by larger scale time-varying current structure (Williams and Follows, 2011). When combined with mixing, this process is irreversible.
equatorial and coastal upwelling (Williams and Follows, 2011). During an El Niño event, SSTAs are positive, the thermocline is depressed, and there is a weakening of easterlies which in turn suppresses equatorial upwelling. There is also a decreased SST gradient between the western and eastern equatorial Pacific which further weakens the easterly trades (Wang et al., 2004b). During La Niña, the opposite occurs; in the eastern Pacific, the thermocline is elevated, and SSTAs are negative.

Wang et al. (2004a) hypothesizes that a change in advection is the most likely cause for T-S variability associated with El Niño events. For example, Wang et al. (2004a) found that El Niño is attributed to a larger convergence at the equator of warmer-saltier water from the southern hemisphere and a lesser convergence of cooler-fresher water from the Northern Hemisphere. Understanding the thermocline/pycnocline T-S relationship is important as it may alter near-surface circulation, which in turn affects SST and air-sea interaction such as wind changes and heat flux (Wang et al., 2004a).

Additionally, O’Kane et al. (2013) showed strong correlations in water mass properties affecting ENSO using a conglomeration of data from the Argo array, eXpendable BathyThermographs (XBTs), and conductivity, temperature, depth (CTD) profiles in an ocean general circulation model (OGCM) reanalysis product. Negative anomalies advecting along the $\sigma_\theta 25$ isopycnal were found to shoal the thermocline and correspond with El Niño events; positive anomalies were found to weaken the model’s ENSO and deepen the thermocline (O’Kane et al., 2013).

North Pacific, low-frequency climate variability includes the Pacific Decadal Oscillation (PDO), a well-established sea surface temperature anomaly (SSTA) index identified in the 1990’s, forced by a combination of variability of the Aleutian Low atmospheric system, ENSO, and oceanic zonal advection from the Kuroshio-Oyashio Extension (Schneider and Cornuelle, 2005; Newman et al., 2016). These processes span from high latitudes of the North Pacific (Aleutian Low) to the equatorial region (ENSO) and are most evident as SSTAs resulting from local North Pacific atmosphere-ocean interaction (Newman et al., 2016).

Miller and Schneider (2000) identified these surface-forced temperature anomalies in the central North Pacific resulting from large-scale perturbations of the Aleutian Low. These SSTAs are then subducted into the thermocline during late-winter and provide an influx of water to the thermocline which is then transported meridionally south (Newman et al., 2016).

McCreary and Lu (1994) described these ocean interior pathways with Subtropical Cell (STC) theory. The STC is driven by surface Ekman flow and is an important ocean tunnel connecting the subtropics to the tropics through meridional transport of water (McCreary and Lu, 1994; Liu and Alexander, 2007). Beginning at higher latitudes outside the tropics, exposed seasonally outcropping waters of isopycnals are subducted through downward Ekman pumping into the thermocline and over the course of several years these STC waters are advected to the equatorial region through the thermocline and then upwelled to the surface (McCreary and Lu, 1994). The water then flows
poleward in the surface layer creating a closed circulation cell.

There are two means by which the STC modulates equatorial water properties. Nonaka et al. (2002) differentiates these as mean advection of temperature anomalies (LHS Eq. 1.1), and anomalous advection of mean temperature by changes of the STC’s speed (first term RHS Eq. 1.1). Fukumori et al. (2004) notes this time-variability of subtropical to tropical circulation remains one of “leading uncertainty in recent studies”.

There are many questions to be answered regarding the precise role spiciness anomalies have in ENSO prediction and how SSTA’s of the PDO affect downstream dynamics at the equator. For example, can a statistical relation of PDO phase be made with the amplitude of ENSO events? Can predicted probability of an ENSO event be enhanced with improved understanding of internal ocean dynamics in the thermocline? How does the formation and propagation of density-compensated anomalies change the thermocline and what is the significance they hold in atmosphere-ocean dynamics and adjustments associated with ENSO? Answers to these questions can be extrapolated from an improved understanding of spiciness in the thermocline.

1.10 Present Study

Here, we use an ensemble of OFES model realizations that incorporate Argo data to clarify and quantify the role and importance of spiciness in the Pacific’s upper-thermocline by looking at $\theta$ and $S$ on isopycnals. Model parameters are the same as Furue et al. (2018). We analyse monthly averaged potential temperature, salinity, and sea surface height($\eta$) data from OFES.

Our primary focus is on the $\sigma_{\theta}$ 25 surface in the Pacific due to it being an upper-thermocline layer where some of the largest amplitude spiciness variations occur through different physical processes. $\sigma_{\theta}$ 25 receives seasonal exposure to the atmosphere through isopycnal surface outcrops between 60°N–30°N and 10°S–40°S. From these latitudes, $\sigma_{\theta}$ 25 waters subduct in both hemispheres and are within the domain of the subtropical gyres. This includes the equatorial region where prominent oceanic currents transport water and upwelling occurs. Additionally, $\sigma_{\theta}$ 25 is in a window of isopycnals that have the largest spice changes associated with ENSO (Wang et al., 2004b,a). Depending on the advective flow path, some $\sigma_{\theta}$ 25 waters reach the EUC, where advection occurs rapidly on the order of 1 m s$^{-1}$ eastward to the Pacific’s eastern boundary.

1.10.1 Thesis Objectives

The primary objective of this thesis is to identify and describe spatial and temporal spiciness variance in the Pacific thermocline. To achieve this we address the following questions by looking at $\sigma_{\theta}$ 25 in OFES:

1. What is the spatial dependence of spiciness variance? Addressing this question will provide a starting point to clarify the origin, pathway, and destination of spiciness signals on the
isopycnal. Locations of high spiciness variance may show where changes in spiciness affect near surface circulation, which in turn affects atmosphere-ocean interactions.

2. Where are mean spiciness gradients and fronts? Is spiciness variance co-located with subsurface mean spiciness gradients and fronts (\(\nabla \chi\))? This will test Schneider (2000)’s hypothesis of spiciness generation through anomalous advection along mean gradients and fronts. Areas where spiciness variance is co-located with mean gradients of \(\theta\) and \(S\), may support this hypothesis.

3. Is there a meridional flux of spiciness due to eddies? Fukumori et al. (2004) looked at the origins, pathway, and destinations of Niño-3 region (5°–5°N, 150°–90°W) surface waters, noting that approximately 20% of it flows from the extra-tropics directly toward the equator through the interior ocean. We investigate if there is the possibility of a meridional flux of spiciness from eddies “short-cutting” from mean flow paths of the subtropics to the equator.

4. What is the eddy-induced spiciness variance in individual OFES realizations? This will expand on Furue et al. (2018)’s single model run experiment by having multiple realizations from which approximations of eddy-induced spiciness variance can be made.

5. How much spiciness variance is attributed to external forcing? The OFES ensemble average dampens small-scale noise from stochastic mesoscale eddies and approximates spiciness variance in the thermocline resulting from the atmosphere’s external forcing. If winds are known, then this variability is deterministic from which predictions can be made.

6. Are there identifiable spiciness signals in OFES? How do these compare with observations? Recent studies from Sasaki et al. (2010) and Kolodziejczyk and Gaillard (2012) have identified spiciness signals propagating on neutral surfaces with observations from Argo data.

7. Is the transport of spiciness signals a result of mean advection? This question tests Gu and Philander (1997)’s and Sasaki et al. (2010) hypotheses of mean flow advection of spiciness on a neutral surface as a mechanism of spice signal propagation. Additionally, flow paths are compared between hemispheres to see if and where spiciness signals reach the equator.

8. To what degree does ENSO modulate spiciness in the thermocline? Wang et al. (2004a) correlated ENSO events with water properties from different locations reaching the equator. Results of ENSO modulation of spiciness in the upper-thermocline may be useful for studies of data assimilation that employ statistical relationships between spiciness variability and ENSO.

9. Are spiciness signals a result of anomalous advection due to Rossby Waves? Furue et al. (2018) noted that dynamical signals were present, but difficult to identify in individual realizations.
due to mesoscale eddy noise. We attempt to dampen this noise with the OFES ensemble average data set to search for evidence of dynamical signals.

In addition to our objectives above, we identify areas of discussion from our time looking at OFES data and provide recommendations for future work.

Chapter 1 provided our motivation, relevance and background of spiciness, a description of spiciness generation mechanisms, spice signal propagation, climate variability and spiciness, and described the present study and its objectives. Chapter 2 describes the OFES model and our approach and methodology of the experiment, data preparation, and analysis techniques. Chapter 3 provides results. Chapter 4 is discussion. Chapter 5 is conclusions and recommendations for future work.
CHAPTER 2
APPROACH AND METHODOLOGY

2.1 OFES Model and Experiment

OFES is a variant of the Modular Ocean Model 3 (MOM3) developed at the Geophysical Fluid Dynamics Laboratory/National Oceanic and Atmospheric Administration (GFDL/NOAA); it is a semi-global and eddy-resolving OGCM. The model implements the primitive equations of volume conservation, the momentum equation, and thermal energy, for approximate formulation of potential temperature and salinity onto grid boxes of simulated ocean (Masumoto et al., 2004). It is intended to realistically reproduce ocean circulations in the upper ocean, and is forced by National Centers for Environmental Prediction (NCEP)/National Centers for Atmospheric Research (NCAR) reanalysis data to include winds, heat flux, precipitation, and evaporation (Masumoto et al., 2004; Furue et al., 2018). Bottom topography was incorporated into the model using a 1/30° resolution data set created from Southampton Oceanography Centre’s Ocean Circulation and Climate Advanced Modeling (OCCAM) Project (Masumoto et al., 2004).

JAMSTEC conducted a unique OFES configuration of two separate ensemble experiments with different surface boundary conditions which branch from the initial model integration in 2005 and run to the end of 2014. Each of these ensemble experiments contain five model realizations, each having slightly altered initial conditions to provide independent mesoscale eddy fields.

JAMSTEC prepared monthly data files from January 2005 to December 2014 of θ, S, η. The data fields are 1/10° × 1/10° spatial resolution boxes comprised of 54 vertical levels, with 5-m interval resolution near the surface and increasing space between levels as depth increases. For this study the upper 31 depth levels are used which have improved resolution for upper-thermocline (0–600m) processes. The spatial focus is the subtropical and tropical Pacific.

The first experiment is a 5-member ensemble over 2005–2014 of the standard OFES run described above, surface forcing is based on a re-analysis product, and sea surface salinity (SSS) is strongly restored to the WOA98 monthly climatology (Boyer et al., 1998; Antonov et al., 1998; Masumoto et al., 2004; Furue et al., 2018)’s. This is hereto referred to as the NCEP experiment.

The second experiment is also a 5-member ensemble over 2005–2014 of the standard OFES run, however reference SSS is replaced with the SSS from an International Pacific Research Center (IPRC) gridded Argo product (http://apdrc.soest.hawaii.edu/projects/argo/), which hereto is referred to as the Argo experiment (Furue et al., 2018).

The salinity drift adjustment is important because the feedback mechanisms and dissipation of SST and SSS anomalies act on very different timescales (Williams et al., 2006). There is a rapid, negative feedback between SST anomalies and surface heat flux anomalies (ex. positive SSTA results in a negative heat flux from the surface). Salinity structure however is forced by freshwater flux with
the atmosphere through evaporation and precipitation, river run-off, and formation and melting of sea-ice. Therefore, there is no such feedback between SSS anomalies and surface freshwater flux anomalies (Williams et al., 2006).

2.2 Data Preparation

Raw data was received from JAMSTEC as 3-D monthly files of state variables of either $\theta$ or $S$, and 2-D data files of $\eta$. A control file (.ctl) for reading the files in Grid Analysis and Display System (GrADS) was also delivered by JAMSTEC, however it was decided to use the Python programming language for additional flexibility in converting raw files to prepared Network Common Data Form (netCDF), a machine independent, self-describing data format for ease of readability and analysis. With similar data preparation and little to no adjustment, our codes can be used for future work on analysis with comparable raw data.

2.2.1 Depth to density coordinate transformation

It is important to note that, in the OGCM run, the equation of state is approximated using a cubic approximation to the UNESCO equation of state (Pacanowski and Griffies, 2000). In our analysis we calculate potential density anomaly ($\sigma_\theta$, equal to potential density minus 1000 kg m$^{-3}$) of seawater at one standard atmosphere using Millero and Poisson (1981)'s UNESCO seawater equation of state (EOS),

$$\sigma_\theta(z) = \text{EOS}(S(z), \theta(z), p = 0)$$

(2.1)

from the prepared OFES output $\theta$ and $S$ netCDF files. The UNESCO EOS is appropriate from 0 to 40 °C and 0.5 to 43 Practical Salinity Units (psu) (Millero and Poisson, 1981).

After potential density is calculated for the entire domain of depth levels, a linear method is used to interpolate $\sigma_\theta 25(z)$. Then, $\theta$ and $S$ are interpolated onto $\sigma_\theta 25(z)$ to transition the state variables from depth to density coordinates. It is important to note that accuracy of this method is dependent on how well the vertical grid resolves the temperature and salinity structure.

This transformation from $z$-space to $\sigma_\theta$-space allows us to study properties as they appear on $\sigma_\theta 25$. The analysis ready data for this project contains isopycnal depth, salinity, and potential temperature.

2.2.2 Montgomery potential approximation

The Montgomery Potential ($M$) on a potential density surface is analogous to the pressure divided by density on a geopotential surface, with the geostrophic flow following contours of constant $M$ or pressure, respectively. We employ the hydrostatic equation for varying fluid density to calculate hydrostatic pressure ($p$) from OFES output of $\eta$, and potential density $\rho = \sigma_\theta + 1000$ kg m$^{-3}$
calculated previously,

\[ p(z) = g \int_{z}^{0} \rho(z) dz + \eta g \rho(0) \]  

(2.2)

where \( g \) is gravitational acceleration, \( z \) is isopycnal depth. After \( p \) is calculated using Eq. (2.2), \( M \) is diagnostically calculated to approximate geostrophic mean flow,

\[ M(\sigma_{\theta}) = \frac{p(\sigma_{\theta})}{\rho_{0}} + g z(\sigma_{\theta}) \]  

(2.3)

where \( M(\sigma_{\theta}) \) is Montgomery potential on the isopycnal surface, \( p(\sigma_{\theta}) \) is pressure on isopycnal surface, \( \rho_{0} = 1025 \text{ kg m}^{-3} \) is a reference density, \( g \) is gravitational acceleration, and \( z(\sigma_{\theta}) \) is depth of isopycnal.

### 2.2.3 Potential vorticity approximation

The potential vorticity (\( Q \)) of a water mass is largely determined at the sea surface when exposed to atmosphere’s winds and diabatic forcing. Once the water parcel is entrained in the thermocline, it is no longer exposed to wind stress and other fluxes and its \( Q \) value is approximately conserved and is proportional to the absolute angular momentum along the local vertical axis (Luyten et al., 1983; Pedlosky, 1986).

In ventilated thermocline models, Ertel’s potential vorticity is approximated as potential vorticity \( Q_{L} \) in an isopycnal layer,

\[ Q_{L} = (\zeta + f) \left( \frac{\partial \sigma_{\theta}}{\partial z} \right) \]  

(2.4)

where the \( (\zeta + f) \) term represents absolute vorticity of the fluid with \( \zeta \) the vertical component of vorticity, \( f \) the Coriolis parameter, and the \( \frac{\partial \sigma_{\theta}}{\partial z} \) term is the change of isopycnal density with respect to the vertical coordinate.

For our large scale, low Rossby number\(^1\) calculations, \( |\zeta| \ll |f| \), and the right hand side of Eq. (2.4) can be approximated to \( f \frac{\delta \sigma_{\theta}}{\delta z} \). Therefore \( Q \) is proportional to the difference between isopycnal depths of two \( \sigma_{\theta} \) layers,

\[ Q_{L} \approx f \left( \frac{\delta \sigma_{\theta}}{\delta z} \right) \]  

(2.6)

where \( \delta \sigma_{\theta} \) is the change of isopycnal density and \( \delta z \) is change of isopycnal depth.

In our idealized model, \( Q_{L} \) is established at the point of subduction and remains approximately unchanged until the parcel re-emerges into the mixed layer, potentially after a basin-scale transit of tens of thousands of kilometers. Therefore flow pathways can be elucidated as fluid trajectories on isopycnals follow lines of constant \( Q_{L} \) (Luyten et al., 1983). This term should hold valid as a tool

\[ R_{o} = \frac{U}{fL} \]  

(2.5)

where \( U \) represents the scale of velocity and \( L \) is the scale of length.

\(^{1}\)
for analysis of flow paths until reaching very low latitudes where \( f \) becomes small and the Rossby number of water parcels is no longer small compared to one.

## 2.3 Analysis Techniques

In this paper, spiciness is \( S \) or \( \theta \) mapped on an isopycnal and can be viewed mathematically as \( S(t, x, y, \rho) \) or \( \theta(t, x, y, \rho) \). The symbol \( \chi \) (pronounced \textit{kai}) is used to denote either \( S(t, x, y, \rho) \) or \( \theta(t, x, y, \rho) \) depending on the context.

While the forcing parameters within the experiment are the same as Furue et al. (2018) (aside from the aforementioned small alterations to initial conditions) our approach differs in how we identify and define anomaly. Furue et al. (2018) conducted two experiments that differed in the surface boundary condition for salinity. Anomalies were defined as the differences of salinity between the experiments:

\[
\delta S(t, x, y, z) \equiv S_{\text{Argo}}(t, x, y, z) - S_{\text{NCEP}}(t, x, y, z)
\] (2.7)

Where \( \delta S \) is the salinity difference, and \( S_{\text{Argo}} \) and \( S_{\text{NCEP}} \) is the salinity output of the OFES experiments with salinity boundary conditions based on Argo observations, or climatology used in the NCEP data. This anomaly is then separated into the spiciness and “heave” (or dynamical) components. Furue et al. (2018) found the addition of observed Argo sea surface salinity to OFES better represents the observational subsurface properties than the NCEP experiment. Specifically in the Argo experiment, subsurface salinity variability was improved below the mixed layer in the subtropics and the tropics (Furue et al., 2018). With this assessment, the majority of our analysis will focuses on the Argo experiment, with the NCEP ensemble and realizations used for secondary and comparative purposes only.

Here, we define spice anomalies \( \chi'_i \) for each realization \( i \) as the difference of \( \theta \) or \( S \) from the climatology of the realization \( i \) (mean annual cycle) over the duration of the OFES hindcast,

\[
\chi'_i(t) = \chi_i(t) - \bar{\chi}_i(t_m)
\] (2.8)

where \( \chi_i \) is the spiciness and \( \chi'_i(t) \) the anomaly for a given month \( t \) month of the ten year integration, and \( \bar{\chi}(t_m) \) is the average seasonal cycle of spiciness over all years, and depends on the calendar month \( t_m \). Throughout, an overbar will denote seasonal climatology unless noted otherwise.

### 2.3.1 Approximating variance of separate processes

We use the model ensemble to separate responses to atmospheric forcing from impacts of ocean eddies and intrinsic ocean variability. Ensemble members are forced by the same air-sea fluxes but are subject to different initial conditions. Ensemble averages therefore reflect the response to the common surface fluxes, while differences between realizations are a measure for the eddy-induced
variance. Please note that the limited ensemble size makes this separation approximate, as there are nonlinear processes coupling atmospheric forcing and ocean intrinsic variability. However, we assume these effects are secondary.

Ensemble averages of a spiciness \( \chi_\epsilon \), or any other variable, are indicated by a subscript \( \epsilon \), and are obtained from the realizations as,

\[
\chi_\epsilon(t) = \frac{1}{N} \sum_{i=1}^{N} \chi_i(t)
\]

where \( N = 5 \) the number of realizations in the ensemble. Ensemble averages depend on time \( t \).

Ensemble averaged spiciness anomalies \( \chi'_\epsilon(t) \),

\[
\chi'_\epsilon(t) = \chi_\epsilon(t) - \overline{\chi}(t_m)
\]

are obtained as the difference of ensemble averaged spiciness at a given month from the ensemble average climatology \( \overline{\chi}(t_m) \). Ensemble averaged spiciness anomalies reflect atmospheric forcing.

The difference \( \Delta \chi'_i \) between spiciness anomalies of individual realizations \( \chi'_i \) and the ensemble average \( \chi'_\epsilon \),

\[
\Delta \chi'_i = \chi'_i - \chi'_\epsilon
\]

approximates of stochastic mesoscale eddy activity.

This split of spiciness anomalies is visualized in Fig. 2.1. The top panel shows ensemble average spiciness anomaly \( \chi'_\epsilon \), the middle figure of the spiciness anomaly of a \( \chi'_i \) is a single realization, and the bottom panel shows \( \Delta \chi'_i \), the deviation of spiciness induced by eddies.

The structure of \( \chi'_\epsilon \) is much smoother than \( \chi'_i \). The warm/salty signal with an amplitude of up to \( \sim 1.5^\circ \text{K} \) appears in 2007 at 23°N 120°W off Baja California is present in both top and middle panels, with increased amplitude in the single realization showing \( \chi'_i \). In the bottom \( \Delta \chi'_i \) panel, no large scale signal is present, and the structure appears noisy, with smaller scale anomalies a result of stochastic mesoscale eddies.

The order of spiciness anomalies \( \chi_i(t) \) between realizations is represented by the variance \( \sigma^2_\epsilon \) and standard deviation \( \sigma_\epsilon \) over the realization number \( i \). Areas of high variance over time \( \sigma^2_t \) in both ensemble and individual realizations identify the spatial dependence of spiciness variance. Prominent fronts and gradients are identified from the temporal average of both ensemble \( \overline{\chi}(t_m) \) and individual realizations \( \overline{\chi_i}(t_m) \). Since seasonal cycles of spiciness in the thermocline are small on account of Stommel’s daemon discussed in the introduction, we present annual averages to characterize spiciness anomalies of a particular year and annual averages of the climatology.
Figure 2.1: Hovmöllers Comparing Ensemble Average and Single Realizations

Hovmöller panels showing, from top to bottom; $\chi_t'$, $\chi_i'$, and $\Delta \chi_i'$. The vertical axis is time from OFES Argo datasets and the horizontal is longitude. Longitude values have been averaged over 5° latitude, and the colorbar is $\theta'$ anomaly in K. Note the warm/salty signal with an amplitude of up to $\sim 1.5^\circ K$ in 2007 at 23° N 120° W in both ensemble (top) and single (middle) panels.
2.3.2 Cross-correlation analysis

Lagged correlation between timeseries and other timeseries for the entire spatial domain, or across a specified latitude longitudinally, is calculated to create correlation maps and diagrams. This method can be used to identify the path of a signal, with the local maximum of the lagged correlation coefficient (R) qualitatively acting as a passive tracer. Combining this path with a contour plot of isolines of $M$ gives insight into how closely the signal follows the mean flow paths.

Looking at the R value alone, signals can be followed for up to a few years. This approach is limited due to solely looking at the correlation qualitatively rather than a quantitative passive tracer.

Cross-correlation of $\chi'$ and $M'$ timeseries is calculated to investigate correlation between spiciness and pressure perturbations.

2.3.3 Normalized spiciness

For visualizing the propagation of spiciness signals and identifying low-frequency signals (interannual to decadal), a 12-month running average is taken of anomalies from the ensemble data set. These monthly running averages are then normalized by the $\sigma_t(\chi')$ spatially on $\sigma_0$ 25 to yield $\frac{\chi'}{\sigma}$ in Eq. (2.12),

$$\chi'_{normalized} = \frac{\chi^{year}(t, \vec{x})}{\sigma_t(\chi(t, \vec{x}))}$$ (2.12)

where $\chi'$ is the spiciness signal and $\sigma_t$ is the temporal standard deviation at a spatial location.
CHAPTER 3
RESULTS

3.1 Background State in OFES

To test the veracity of the OFES model and to provide the mean states for the following discussions, we show the depth, flow field, and T-S properties on the $\sigma_\theta = 25 \text{ kg m}^3$ isopycnal surface.

Figure 3.1: Ensemble average depth $z_t$ of $\sigma_\theta 25$

The depth $z_t$ of the $\sigma_\theta 25$ isopycnal from the OFES Argo ensemble simulation, averaged over all realizations and over the entire 10 years of the simulations. Colormap and contours show depth in meters, with a contour spacing of 10 m. Seasonal outcrop values have been masked.

Temporally averaged depth $z_t$ of $\sigma_\theta 25$ from the OFES Argo ensemble simulation in Fig. 3.1 reveals a structure consistent with $z_t$ of Johnson and McPhaden (1999)'s observations from CTD casts of neutral density anomaly $1 \gamma_n = 25 \text{ kg m}^3$ in Fig. 3.2. The isopycnal is deepest in the western Pacific, reaching depths of over 220 m in the Philippine Sea, then shoals eastward reaching its shallowest depths of $< 30$ m in vicinity of the Costa Rica Dome (90°W 8°N) and along the South American coastline.

The counter-current ridge in the north hemisphere is prominently reproduced in the model, beginning at $\sim 5^\circ \mathrm{N} 140^\circ \mathrm{E}$ and exceeding 150 m between 140°E–150°E. It shoals in OFES across the length of the Pacific ocean until it reaches depths of 40 m off the coasts of Ecuador and Colombia.

$^1$For our purposes at shallow depths in the upper-thermocline, isopycnal $\sigma_\theta 25$ closely resembles neutral density anomaly, $\gamma_n = 25 \text{ kg m}^3$, and therefore is useful for comparing observational data with OFES model output. As an example, off Baja, with $S=33.75, \theta=14, p=50\text{dbar}, \gamma_n = 25.25$ and $\sigma_\theta = 25.22$
Depth of the ridge is greater in observations at 175 m in the western Pacific than the model’s, however the overall structure and location in both model and observations in Fig. 3.2 are comparable.

Figure 3.2: Johnson and McPhaden (1999) average depth $z_t$ of $\gamma_n = 25 \text{ kg m}^3$.

Temporally averaged depth (m) of the neutral surface $\gamma_n = 25 \text{ kg m}^3$. Observational data consists of 15,693 individual CTD stations between 20.5°S and 20.5°N, 120°E and 70°W from 1967–1998. White contours are at 25 m depth intervals. The solid black dots are centered bins of the spatially averaged CTD casts of bins centered every 1° latitude and 5° longitude. ©American Meteorological Society. Used with permission.
Montgomery potential $M_t$ on $\sigma_\theta 25$ temporally averaged over all realizations and over the entire 10 years of the simulations. All seasonal values have been masked and black contours are at a spacing of 1 m$^2$s$^{-2}$. Note the bifurcation between the 3 and 4 m$^2$s$^{-2}$ isolines in the North Pacific’s western boundary near 15°N 130°W, where mean flow branches north to the Kuroshio Extension, or south into both the Indonesian Throughflow to the southwest and back eastward to the equator as observed in Fig. 3.11.

Fig. 3.3 shows the temporally averaged lines of $M$ at a spacing of 1 m$^2$s$^{-2}$ calculated from OFES Argo ensemble average data from 2005–2014. The flow field implied by the $M$ contours show the substantially different flow paths of $\sigma_\theta 25$ in the north and south hemispheres from the subtropics to the tropics. Between the 1 m$^2$s$^{-2}$ and 3 m$^2$s$^{-2}$ $M$ isolines in the North Pacific, the path of mean advection within the insulated thermocline beginning at 30°N 120°W is southwestward in a corridor approximately 500 km in width and do not converge to the equator until the international date line. This curvature around the eastern low latitude North Pacific is associated with the shadow zone, identified by high values of potential vorticity (Fig. 3.4). West of 180°, if signals reach the western boundary, two separate paths are possible with the bifurcation between the 3 m$^2$s$^{-2}$ and 4 m$^2$s$^{-2}$ $M$ isolines at 15°N 130°W where mean flow either turns south into the Indonesian Throughflow and equator, or north to the Kuroshio Extension.

In the Southern Hemisphere, the path between 1 m$^2$s$^{-2}$ and 3 m$^2$s$^{-2}$ $M$ isolines off of Peru at approximately 10°S 100°W, reaches the equator as far east as 130°W. A much shorter path for $\chi'$ signals to reach the equatorial region.

The mean flow paths implied by Fig. 3.3’s $M$ contours are consistent with Fukumori et al. (2004)’s description of a potential vorticity “barrier” which results in longer interior transport.
pathways in the North Pacific than in the South Pacific. Mean potential vorticity is calculated using Eq. (2.6) and averaged temporally on $\sigma_\theta 25$, with results shown in Fig. 3.4. Depths of $\sigma_\theta 24.9$ and $\sigma_\theta 25.1$ isopycnals in OFES are used to determine the $\left( \frac{\partial \sigma_\theta}{\partial z} \right)$ term. The strong gradients of potential vorticity and aforementioned shadow zone in the North Pacific are visible in the closeness of the black contours and colormap (units are $\times 10^{-5}$ kg $m^{-4}$) beginning near 20°N 120°W and extending west and south to 10°N and the date line. No such barrier exists in the southern hemisphere and therefore waters take a more direct path to the equator from origins off of S. America.

Figure 3.4: Average Potential Vorticity approximation $Q_L$ of $\sigma_\theta 25$
$Q_L$ temporally averaged over entire timeseries and all realizations on $\sigma_\theta 25$ from the OFES Argo ensemble in the Pacific Ocean. Seasonal outcrop values have been masked. $Q_L$ values are approximated using 2.4, with red color show positive $Q_L$, and blue values are negative $Q_L$. Contour units of $Q_L$ are $1 \times 10^{-5}$ kg $m^{-4}$. Note areas of high spatial gradients corresponding with Ekman suction in the eastern subtropics of the North Pacific.

In the western equatorial Pacific region, there is a confluence of westward basin-scale flows from both hemispheres which can be seen in Kashino et al. (1999)’s salinity section (Fig. 3.5). This cross-section compiled from CTD casts at 130°E shows two salinity maxima converging from North Pacific Tropical Water$^2$ (NPTW) and South Pacific Tropical Water$^3$ (SPTW) resulting in a strong salinity front at 5°N.

Fig. 3.5 also shows the vertical structure from the surface to 600 m, including isopycnal layers.

$^2$NPTW salinity maximum around $\sigma_\theta 24$.
$^3$SPTW salinity maximum around $\sigma_\theta 25$. 
Results shown in Fig. 3.1 of \( \sigma_{\theta} \) 25 depths of 100 m to 150 m in the WEP are consistent with Kashino et al. (1999)’s observations, seen as the thin dashed contour of \( \sigma_{\theta} \) 25 just below the 100 dbar tick mark in Fig. 3.5’s y-axis; averaging from just over 100 m to \( \sim \) 150 m from 8\(^{\circ}\)N–0\(^{\circ}\). Additionally, Fig. 3.5’s cross-section shows \( \sigma_{\theta} \) 25 crossing the salinity front at 5\(^{\circ}\)N and lying only tens of meters below the mixed layer, implying \( \sigma_{\theta} \) 25 variability provides a glimpse of the upper-thermocline changes directly beneath the WEP’s mixed layer.

Figure 3.5: Kashino et al. (1999) Salinity Front in the Western Equatorial Pacific

Salinity front in the WEP from Kashino et al. (1999). Cross-section compiled from CTD casts to show the vertical structure from the surface to 600 m depth along 130\(^{\circ}\)E during Kaiyo World Ocean Circulation Experiment (WOCE) II (February 1994). Solid contours have intervals of .1 psu and \( \sigma_{\theta} \) layers are shown with thin dashed lines. Note the strong salinity gradients present in the vertical with depth and also in vicinity of 5\(^{\circ}\)N.

Figs. 3.6 and 3.7 show the temporal averages from 2005–2014 of OFES \( \theta \) and \( S \) on \( \sigma_{\theta} \) 25 respectively. The seasonal values in both figures have been masked, leaving only areas that are insulated in the thermocline for the entire timeseries. The line of masked values can be interpreted in both figures as the outcrop line for the duration of the data set.

West of 150\(^{\circ}\)W the isopycnal outcrops at 25\(^{\circ}\)N–28\(^{\circ}\)N and 21\(^{\circ}\)S–22\(^{\circ}\)S. Further east, the outcrop reaches 20\(^{\circ}\)N in the Northern Hemisphere before swinging to 30\(^{\circ}\)N off the coast of Baja California. East of 150\(^{\circ}\)W on the Southern Hemisphere, the outcrop shifts to lower latitudes and reaches 8\(^{\circ}\)S off the coast of South America.
In Fig. 3.6, temperature on the isopycnal shows a marked gradient across the equator. In the Northern Hemisphere, temperatures are between 18–19°C. A tongue of cool waters as low as 14–15°C emanates from the California current region off Baja. Further offshore in the subtropical central Pacific, waters are warmer than 19°C. Waters are also warmer off the coast of Central America and connect to the high temperatures on this isopycnal in the Southern Hemisphere. There, temperatures there are dominated by a band of values higher than 22°C that stretches from the subduction regions between 150°W and 110°W and 12°S–21°S toward the western Pacific while shifting equatorward by 5 degrees of latitude. The sharp gradient on the equator separates warm Southern Hemisphere waters and cool Northern Hemisphere waters, with the difference in temperature reaching 3 K in the western Pacific. In the eastern equatorial Pacific, the difference is reduced by cooler waters subducted near S. America.

![Figure 3.6: Temperature on $\sigma_\theta$ 25](image)

*Temporally averaged $\theta$ on $\sigma_\theta$ 25 over the entire timeseries and realizations of the OFES Argo ensemble in the Pacific Ocean. All seasonal outcrop values have been masked, and the colormap shows $\theta_t$ values in °C from 2005–2014. Solid black lines are contours with interval length of 1 K in showing °C. Note areas of high spatial gradients off of Baja California, the Equator, and the Western Equatorial Pacific.*

The large-scale structure of averaged salinity values on $\sigma_\theta$ 25 in Fig. 3.7 nearly mirrors Fig. 3.6’s $\theta$. The fresh waters off Baja have salinity values as low as 33.6 psu, while the high salinity waters on the Southern Hemisphere subduction regions have values as high as 36.2 psu at 18°S 140°W and 35.4 psu at 5°S 100°W. These OFES values can be compared to Johnson and McPhaden (1999)’s Fig. 3.8 of temporally averaged salinity values from 1967–1998 of CTD casts. The low salinity tongue sweeping southwestward from Baja to the international date line is observed in both figures with very similar structure and magnitudes of salinity, for example at 20°N 120°W a
salinity minimum of 34 psu/PSS is present. Fig. 3.8’s tongue of 34.8 psu in observations reaches the western boundary between 5°N–10°N, while in OFES the 34.8 psu contour extends to the dateline only. This is despite the OFES simulation underestimating the low salinity off Baja at 20°N 120°W (34 psu contour in 3.7), and 34.1–34.2 psu in observations at the same location. In the Southern Hemisphere, values greater than 36 psu reach across the ocean to the western Pacific and the coast of Indonesia, while observations show that along 10°S waters are fresher than 36.0 psu west of the dateline.

Temperature and salinity gradients on the isopycnal surface decay faster than in Johnson and McPhaden (1999)’s Fig. 3.8, suggesting that mixing in OFES is more vigorous than in the ocean. For example, the tongue of low salinity off Baja decays rapidly so that in the west of the international date line at 6°N OFES shows salinity larger than 34.8 psu, while observed salinity values there are smaller at \( S < 34.8 \text{ psu} \).

Figure 3.7: Salinity on \( \sigma_\theta \) 25

*Temporally averaged salinity values on \( \sigma_\theta 25 \) over the entire timeseries and realizations from the OFES Argo ensemble in the Pacific Ocean. All seasonal outcrop values have been masked, and the colormap shows salinity values in practical salinity units (psu) averaged from 2005–2014. The solid black lines are contours with interval length of .2 psu. Note areas of high spatial gradients are similar to Fig. 3.6.*
Figure 3.8: (Johnson and McPhaden, 1999) Salinity Observations on $\gamma_n = 25 \text{ kg m}^{-3}$

Temporally and spatially averaged salinity values on $\gamma_n = 25 \text{ kg m}^{-3}$ from CTD casts in the Pacific Ocean. Observational data consists of 15,693 individual CTD stations between 20.5$^\circ$S and 20.5$^\circ$N, 120$^\circ$E and 708$^\circ$W from 1967–1998. The solid black dots are centered bins of the spatially averaged CTD casts of bins centered every 1$^\circ$ latitude and 5$^\circ$ longitude. White contours have a spacing of .1 PSS-78 with values in-line with contours. These salinity values can be compared to the OFES values in PSU in Fig. 3.7. Note areas of high spatial gradients along the equator and off Baja California, and similar structure and values to OFES Argo data. ©American Meteorological Society. Used with permission.

The distribution of temperature and salinity on $\sigma_\theta$25 is consistent with the mean advection indicated by the Montgomery streamfunction of Fig. 3.3. The cooler/fresher waters off Baja become warmer and saltier along lines of constant $M$ in the Northern Hemisphere; at 20$^\circ$N 130$^\circ$W, spiciness is relatively cool and fresh at 34.2 psu/17$^\circ$C, following the 2 m$^2$s$^{-2}$ $M$ contour west from this location to the international date line where the water warms and becomes saltier to > 34.8 psu/19$^\circ$C. This increase in $\Theta/S$ in the Northern Hemisphere along the streamfunction $M$ is indicative of mixing along the isopycnal that tends to diminish spatial gradients as water parcels move downstream.

In the Southern Hemisphere however, warm and salty waters remain fairly consistent along $M$ isolines from the high salinity tongue off South America. Waters from 12$^\circ$S 105$^\circ$W are 35.8 psu/21$^\circ$C and again, following the 2 m$^2$s$^{-2}$ $M$ isoline in the South Pacific, upon reaching 2$^\circ$S 180$^\circ$ are the same spiciness.
3.1.1 Annual individual and ensemble data in OFES

Fig. 3.9 reveals the background structure of $\chi$ on $\sigma_25$ in a single OFES realization from temporal averaging of the 10 year data set. Notice there is little difference on a large scale between the five-member ensemble average (Fig. 3.6) and the single realization of Fig. 3.9, however finer structure is revealed in spiciness contours of Fig. 3.9, particularly around the cold feature that extends off Baja southwestward into the mid-Pacific. The prominent fronts remain consistent with those in the ensemble average, such as the close contours along the equator between 19–21°C spanning the breadth of the Pacific along the equator. Plots of all individual realizations are in Appendix B, Figs. B.2–B.6. Appendix C, Figs. C.1–C.10 show raw $\theta'$ values on $\sigma_25$ annually for each realization, year, and ensemble-average from the OFES Argo data set.

The proximity of property contours in Figs. 3.6 and 3.7 are used to infer where subsurface spiciness gradients and fronts of $\sigma_25$ are. The temporally averaged ensemble mean $\overline{\chi_e(\theta)}$ and $\overline{\chi_e(S)}$ reveal the zonal spatial gradients west of Baja California and the meridional gradients spanning the equator from the WEP and EEP.

Figure 3.9: Average temperature on $\sigma_25$ in a single realization $\chi_1$ using OFES Argo $\theta$ data from a single realization from 2005–2014. Black lines are contoured at 1 K and show temperature in °C. Note small-scale structure differences in the spiciness contours between $\chi_1$ and $\chi_e$ (Fig. 3.6), particularly in the cool/fresh feature off Baja.

3.2 Spatial Dependence of Spiciness Variance

Surface flux induced spiciness change over time is revealed by calculating the temporal variance ($\sigma_t^2$) and temporal standard deviation ($\sigma_t$) of $\chi_e$ on $\sigma_25$. This is observed in Fig. 3.10, as an
approximation of externally driven thermocline spiciness. This is revealed spatially in the colormap showing areas of high \( \sigma_t(\chi_e) \) in K. Over-plotted are black contours of \( \overline{\theta} \) from Fig. 3.6.

Figure 3.10: Temporal Standard Deviation of Spiciness in Argo Experiment Ensemble \( \sigma_t(\chi_e) \)

\( \sigma_t(\chi_e) \) using OFES Argo ensemble \( \theta \) data from 2005–2014 on \( \sigma_\theta 25 \) in the Pacific. The black lines are average \( \theta \) contours from Fig. 3.6, with a spacing of 1 K, and colorbar is \( \sigma_t(\chi_e) \) in K. Observe how areas with close contours tend to correspond with areas of greater \( \sigma_t(\chi_e) \).

The Baja California Coast (BCC) at 23°N 110°W in Fig. 3.10 presents the largest externally forced spiciness variation where \( \sigma_t(\chi_e) \) is up to \( \sim 1.6 \) K. The contours of spiciness in vicinity of 23°N 110°W in Figs. 3.6 and 3.7 are very close, implying a large spiciness gradient at this location. This gradient lessens southwestward along the same cool/fresh tongue that reaches the international date line. With the reduced gradient and greater spiciness contour spacing, the amplitude of \( \sigma_t(\chi_e) \) decreases.

The western equatorial Pacific (WEP) north of New Guinea also has relatively large \( \sigma_t(\chi_e) \) of \( \sim 0.64 \) K. Beginning at approximately 5°N 130°W, a tail of \( \sigma_t(\chi_e) = \sim 0.4–0.6 \) K slides down to the equator and extends eastward to 180°, appearing to nearly follow the tightly packed contour lines of spiciness.

Fig. 3.11 from Lukas et al. (1991) is a schematic of the Pacific’s near-surface currents at the low-latitude western boundary region. This schematic depicts the two prominent eddies flanking the western edge of the strong meridional gradients in Figs. 3.6 and 3.7; the Mindanao Eddy to the north centered at 6°N 128°E and the Halmahera Eddy centered at 4°N 131°E at the entrance to the Indonesian Throughflow. As will be discussed below, spiciness variations here coincide with large changes of isopycnal depth, suggesting that forced fluctuations of these eddies cause the spiciness signal.
The area with the greatest spatial extent of large amplitude spiciness variance is the eastern equatorial Pacific (EEP) with $\sigma_t(\chi_x) = \sim \cdot6$ K. Beginning at the equator and 115°W, a coned contour of $\sigma_t(\chi_x) > .6$ K begins expanding meridionally and eastward to the coast of South America. In the EEP, the contours of spiciness are not as close together as off Baja California or WEP, with only a $\sim 1$ K change over in spiciness over $\sim 8^\circ$ latitude ($\sim 800$ km) centered at the equator from 110°W–85°W.

The high variance regions in Fig. 3.10 coincide spatially with the subsurface mean spiciness gradients and fronts implied in Figs. 3.6 and 3.7, which suggests that temporal spiciness variance is large in regions of high isopycnal $\theta$ and $S$ gradients. Or rather, as Schneider (2004) hypothesized, $\sigma_t(\chi)$ along temperature gradients is due to the movement of spiciness fronts.

The standard deviation of spiciness can be translated to lateral isopycnal displacement by dividing the standard deviation by the magnitude of the mean gradient. For example, off Baja at 110°W 20°N, $\sigma_t(\chi_x) = \sim 0.8$ K, and the gradient from Fig. 3.6 is approximately 3°C per 3° of latitude or $\sim 300$ km. Therefore, a meridional displacement of the mean isopycnal by about 100 km accounts for the spiciness variation at this location.

Locations of spiciness variance do not necessarily coincide with variance of isopycnal depth. This is quantified as the standard deviation in meters of isopycnal depth ($\sigma_t(z_e)$) integrated for the duration of the OFES ensemble. This is most apparent off of Baja California in Fig. 3.12, where the colormap shows little vertical movement of $\sigma_t 25$ with $\sigma_t(z_e) < 10$ m, and yet the highest variance in spiciness is present. Also, at the equator from 140°W–110°W $\sigma_t(z_e) > 25$ m, implying a large degree of vertical isopycnal movement and upwelling. In the eastern equatorial Pacific however,
spiciness variance increases in amplitude from $\sim 0.6 - 1$ K from 120°W to the coast of South America, yet isopycnal motion in the vertical diminishes from $\sigma_t(z_e) > 25$ m to $\sigma_t(z_e) < 10$ m.

The exception is in the WEP north of New Guinea from 130°E–150°E 3°N-8°N where $\sigma_t(z_e) > 30$ m and spice variance is also relatively large with an amplitude of 0.8–1.0 K.

![Figure 3.12: Temporal Standard Deviation of Ensemble Depth $\sigma_t(z_e)$](image)

$\sigma_t(z_e)$ of $\sigma_\theta 25$ from OFES ARGO ensemble in the Pacific Ocean. Black contour lines are $\sigma_t$ in meters. Colorbar units are $\sigma_t(z_e)$ (meters). Note the high variability of of depth in WEP and from 140°W–100°W. Seasonal outcrop values have been masked.

### 3.2.1 Baja California

Off the Baja California coast, $\sigma_\theta 25$ waters have the largest amplitude spiciness variations in this study.

The timeseries in Fig. 3.13 shows $\chi'$ variability in a 5° × 5° averaged box (19°N–24°N, 115°W–110°W) of OFES $\theta$ data west of Baja of $\sigma_\theta 25$. Individual realizations and the ensemble average from both experiments are shown, with thin colored lines in both top and middle panels representing individual realizations and the thicker black line the respective experiment’s ensemble average. Note that the ensemble oscillates up to ±1 K with the individual realizations deviating up to $\sim 0.5$ K from the ensemble in parts of the timeseries. The bottom panels of Figs. 3.13–3.15 show the standard deviation between the individual realizations ($\sigma_\epsilon$) for each experiment; notably the bottom panel of Fig. 3.13 $\sigma_\epsilon$ is up to nearly 0.5 K. Note that variations of standard deviations between Argo and NCEP runs appear uncorrelated, suggesting that fluctuations result from the limited sampling of the finite ensemble.
3.2.2 Western Pacific north of New Guinea

East of the Indonesian Throughflow, the western Pacific north of New Guinea (WEP), lies from the equator to $\sim 5^\circ N 130^\circ E$–$140^\circ E$. This area is also referred to as the West Caroline Basin.

The OFES $\theta$ ensemble timeseries data in Fig. 3.14 shows that spiciness oscillates up to $0.5 K$ in the WEP in both the NCEP and ARGO experiments. The thinner colored lines in the top two panels of Fig. 3.14 also show that there is deviation from the ensemble average in the individual realizations which is also reflected in the bottom panel with $\sigma_\epsilon$ values of up to $0.4 K$. Fig. 3.14 shows little difference between the Argo and NCEP experiment simulations suggesting that spiciness variations are not a result of surface freshwater flux, but rather from changes in SSTA’s as evinced by Nonaka and Sasaki (2007), or anomalous circulation.
Figure 3.14: Timeseries comparisons western equatorial Pacific north of New Guinea
Timeseries comparison of $\chi'$ in $5^\circ \times 5^\circ$ averaged box ($0^\circ$N-$5^\circ$N $135^\circ$E-$140^\circ$W) in the western equatorial Pacific north of New Guinea of $\sigma_\theta$25. Vertical axis is temperature anomaly and horizontal axis is time in months from beginning of timeseries in January 2005 ending December 2014 (120 months). The thicker black line represents the ensemble average of the five realizations for each respective experiment. Thin colored lines are individual realizations within each experiment. Bottom plot is the variance $\sigma_\epsilon$ of ARGO (red) and NCEP (blue) ensembles.

3.2.3 Eastern equatorial Pacific

The largest swath of surface induced spiciness flux in Fig. 3.10 is in the EEP, in the coned feature of $\sigma_t > .6$K that extends off the coast of South America west to $115^\circ$W. Taking a slice of the EEP’s waters in this swath at $2.5^\circ$S-$2.5^\circ$N $90^\circ$W-$95^\circ$W in Fig. 3.15 shows a markedly different result when compared to $\sigma_\theta$ 25 waters off of Baja and WEP in Figs. 3.13 and 3.14 respectively.

In both experiments, which total ten realizations, the timeseries of individual realizations in Fig. 3.15 deviates very little from the thick black ensemble line. There are large oscillations in spiciness of $\sim \pm 1$K in the top and middle timeseries panels of Fig. 3.15, and yet stochastic variation due to eddies is small, with $\sigma_\epsilon < .1$K in the bottom panel from both OFES experiments, suggesting that spiciness variability in the EEP is dominated by external forcing.
Figure 3.15: Timeseries comparisons in the Eastern equatorial Pacific

*Timeseries comparison of* $\chi'$ *in* $5^\circ \times 5^\circ$ *box in the EEP of* $\sigma_\theta$ *25* ($-2.5^\circ S–2.5^\circ N, 90^\circ W–95^\circ W$). *Vertical axis* is temperature anomaly and *horizontal axis* is time in months from beginning of timeseries in January 2005 ending December 2014 (120 months). *Thick black line* represents ensemble average of the five realizations for each respective experiment. *Thin colored lines* are individual realizations within each experiment. *Bottom plot* is $\sigma_\epsilon$ of Argo (red) and NCEP (blue) ensembles.

### 3.2.4 Tropical mid-Pacific

The timeseries in Fig. 3.16 of $5^\circ \times 5^\circ$ box (averaged between $17.5^\circ N–22.5^\circ N$ $167.5^\circ E–172.5^\circ E$) in the tropical mid-Pacific is an area with relatively low amplitude changes due to external forcing and eddy-induced spiciness observed in Figs. 3.10 and 3.18, where $\sigma_t < .2$K in both plots respectively. In Fig. 3.16 the timeseries shows $\theta'$ signals rarely exceeding $\pm .5$K, and a $\sigma_\epsilon(\chi') < .1$K in the bottom panel. Comparing the mid-Pacific and EEP timeseries, there exists relatively large amplitude external forcing of spiciness variability with low eddy-induced variability (Fig. 3.15), while in the mid-Pacific there is relatively low surface flux induced and eddy-induced spiciness variability (Fig. 3.16).
Figure 3.16: Timeseries comparisons Mid-Pacific

Timeseries comparison of $\chi'$ in a $5^\circ \times 5^\circ$ averaged box ($17.5^\circ N$–$22.5^\circ N$, $167.5^\circ E$–$172.5^\circ E$) of $\theta$ on $\sigma_\theta$ 25. Vertical axis is temperature anomaly and horizontal axis is time in months from beginning of timeseries in January 2005 ending December 2014 (120 months). Thick black line represents ensemble average of the five realizations for each respective experiment. Thin colored lines are individual realizations within each experiment. Bottom plot is the $\sigma_\epsilon$ of ARGO (red) and NCEP (blue) ensembles.

3.3 Experiment Comparison

An example of prepared $\chi'$ variation as measured by temperature $\theta'$ on $\sigma_\theta$ 25 from both experiments and the difference between them is shown in Fig. 3.17. Each individual realization (thin colored lines) and ensemble average (thick black line) are shown off Baja California in an averaged $5^\circ \times 5^\circ$ box, with units of $\theta$ in K. The ensemble average, and the five individual realizations as a comparison of amplitude and ensemble smoothing. The Niño 3.4 index (dashed blue lines) is shown as a reference for amplitude in K.

Argo and NCEP simulations both show consistent variations of spiciness with positive values of about 1 K warming after the 2006/2007 and 2009/2010 El Niño events, at months 24 and 26 of the model’s runs. Variability occurs consistently for all realizations and suggests atmospheric forcing as a cause. Differences between Argo and NCEP experiments reach values of less than 0.5 K, indicating that changes of anomalies of the surface fresh water are of secondary importance.
Figure 3.17: Timeseries and experiment comparisons west of Baja California

Timeseries comparison of \( \theta' \) in \( 5^\circ \times 5^\circ \) averaged box (19\(^\circ\)N–24\(^\circ\)N 115\(^\circ\)W–110\(^\circ\)W) west of Baja on \( \sigma_\theta \) 25 in the two experiment ensembles with monthly Niño 3.4 index superimposed (dashed blue line). Vertical axis is temperature anomaly (K) and horizontal axis is time in months from beginning of timeseries in January 2005 ending December 2014 (120 months). Thick black line represents ensemble average of the five realizations for each respective experiment. Thin colored lines are individual realizations within each experiment. Bottom plot is the difference of Argo ensemble and NCEP ensemble \( \chi'_{\text{Argo}} - \chi'_{\text{NCEP}} \). Note vertical scale difference in ARGO-NCEP panel.

3.4 Eddy Induced Spiciness Variance

The following section quantifies eddy-induced spiciness variability in OFES using the approaches described in Section 2.3.

Eddy-generated spiciness variability in an individual model run is estimated from the temporal standard deviation of the spiciness deviations, \( \sigma_t(\Delta \chi'_i) \). In Fig. 3.18 off the southern coast of Baja California at 25\(^\circ\)N 110\(^\circ\)W, and extending southwestward to 110\(^\circ\)W 25\(^\circ\)N, eddy-induced variance on \( \sigma_\theta \) 25 is \( \sigma_t(\Delta \chi'_i) = \sim .8 \text{ K} \). Directly to the northwest along the seasonally masked values \( \sigma_t(\Delta \chi'_i) \) are larger than 0.6 K. Similar values are observed adjacent to the masked outcrops \( \sigma_\theta \) 25 at 25\(^\circ\)N centered on 130\(^\circ\)E 5\(^\circ\)N. From these higher pockets of \( \sigma_t(\Delta \chi'_i) > .6 \text{ K} \), eddy-induced variance extends for thousands of kilometers from 8\(^\circ\)N–25\(^\circ\)N, 180\(^\circ\)–80\(^\circ\)W with values of \( \sigma_t(\Delta \chi'_i) = .2 – .4 \text{ K} \).

Fig. 3.18 also shows the expansive area of \( \sigma_t(\Delta \chi'_i) = .2 – .4 \text{ K} \) from 20\(^\circ\)S–5\(^\circ\)S, \( \sim 140 \circ \text{W–100} \circ \text{W} \) in the tropical eastern South Pacific.

In the WEP, \( \sigma_t(\Delta \chi'_i) \) values are higher. A tendril of \( \sigma_t(\Delta \chi'_i) = \sim .2 – .6 \text{ K} \) begins at 130\(^\circ\)E 5\(^\circ\)N and extends to 180\(^\circ\), having structure similar to the same feature in Fig. 3.10.
Areas where $\sigma_t(\Delta \chi') < .2$ K in Fig. 3.18 are in the North and South Pacific west of the international date line, and the EEP.

Figure 3.18: Temporal standard deviation of Spiciness Deviations $\sigma_t(\Delta \chi')$

Temporal standard deviation of spiciness deviations $\sigma_t(\Delta \chi')$ in Argo experiment where $\Delta \chi' = \chi'_i - \chi'_e$ is an approximation of eddy-induced spiciness variance. Black contour lines are spaced at 1°C. Note the variance in the EEP Fig. 3.10 is absent here and areas the eddy field impacts spiciness on the isopycnal surface are revealed. We hypothesize that the time variance visible in the ensemble average in (EEP) is due to what is common in all realizations, i.e. the wind forcing.

The different locations and amplitudes of eddy and wind generated spiciness variability on $\sigma_\theta$ 25 are revealed when Fig. 3.18 and Fig. 3.10 are compared. Notably off Baja, wind and eddies induce spiciness variability with $\sigma_t$ up to 1.4 K and .8 K, respectively. In the EEP, wind and eddies induce spiciness variability with $\sigma_t$ up to .6 K and <.1 K, respectively. In the WEP north of New Guinea, wind and eddies induce spiciness variability with $\sigma_t$ up to .5 K and .4 K, respectively. Below are a summary of findings:

<table>
<thead>
<tr>
<th>Location</th>
<th>Lat/Lon</th>
<th>Winds</th>
<th>Eddies</th>
</tr>
</thead>
<tbody>
<tr>
<td>Baja</td>
<td>19°N–24°N 115°W–110°W</td>
<td>up to 1.4 K</td>
<td>up to .8 K</td>
</tr>
<tr>
<td>EEP</td>
<td>2.5°S–2.5°N 90°W–95°W</td>
<td>up to .6 K</td>
<td>&lt;.1 K</td>
</tr>
<tr>
<td>WEP</td>
<td>0°N–5°N 135°E–140°E</td>
<td>.5 K</td>
<td>.4 K</td>
</tr>
</tbody>
</table>

Taking the standard deviation between realizations in the ensemble ($\sigma_e$) reveals similar spatial structure of where eddy variance is present, and also approximates the amplitude of eddy-induced spiciness variance. Fig. 3.19 shows $\sigma_e(\chi')$ integrated over the entire OFES timeseries, with black con-
tours representing mean \( \theta \) values in °C, and the colormap values are \( \sigma_e(\chi') \) in Kelvin on log\(_{10}\) scale. Note that the structure of the eddy-induced variance is similar to Fig. 3.18, with areas of greater variance represented by orange and red values with \( \sigma_e(\chi') = \sim 1-6 \) K. Consistent spatially and in amplitude with Fig. 3.10, eddy induced variance is observed off of Baja and the WEP-Indonesian Throughflow in Fig. 3.19. However, very little eddy-induced variance exists in the EEP or northwest Pacific with \( \sigma_e(\chi) = \sim <0.01-0.1 \) K.

\[ \text{Figure 3.19: Ensemble Standard Deviation of } \chi' \text{ between Argo Realizations} \]

Standard deviation between ARGO realizations of temperature anomalies (\( \sigma_e(\theta') \)) averaged over the entire period of the model runs. Black contour lines are spaced at 1° C. Note the EEP spiciness variance in Fig. 3.10) is not present and areas the eddies affect spiciness on the isopycnal surface are revealed. We hypothesize that the time variance visible in the ensemble average in EEP is due to what is common in all realizations, i.e. the wind forcing.

This pattern is also observed in the annual integrations of \( \sigma_e(\chi) \) in Figs. 3.20 and 3.21. These two series of panels show the standard deviation between realizations averaged for each year in the OFES Argo experiment using \( \theta \) data (\( \sigma_e(\chi^\text{year}) \)). Blue and purple areas representing high \( \sigma_e(\chi^\text{year}) \) are consistent each year off of Baja California and the WEP north of New Guinea. Areas that are low in variance are red and gray in the figures which is observed in the East China and Philippine seas and EPP, indicating low annual eddy-induced variance.
Figure 3.20: 2005–2009 Standard deviation of spiciness between ARGO realizations by year $\sigma_r(\chi^{\text{year}})$

The colormap is logscale of standard deviation in $K$. 
Figure 3.21: 2010–2014 Standard deviation of spiciness between ARGO realizations by year $\sigma_{\epsilon}(\chi^{\text{year}})$
The timeseries in Figs. 3.13–3.17 also help to visually identify and quantify eddy induced spiciness variance at locations of interest. The bottom panel for each location gives the standard deviation between realizations in respective Argo and NCEP experiments. Comparing Fig. 3.13 to Fig. 3.15 of Baja and EEP timeseries respectively, the difference in amplitude in eddy induced variance is evident as individual realizations deviate from the ensemble. The WEP timeseries however in Fig. 3.14 presents near equal contributions in wind and eddy-induced spiciness variance.

3.5 ENSO Modulation of Spiciness

This section identifies relationships in T-S variability between $\sigma_\theta$ 25 regions of high spice variance and ENSO using lagged correlation between averaged OFES Argo $\chi'_\epsilon$ timeseries and the Niño 3.4 index ($5^\circ$N–$5^\circ$S 170°W–120°W) (Rayner et al., 2003).

The top panels in Figs. 3.22, 3.23, and 3.24 are timeseries of Niño 3.4 index (dashed blue line) and timeseries of $\chi'_\epsilon$ on $\sigma_\theta$ 25 off Baja, the EEP, and WEP (thicker black lines) respectively. The vertical axis units of the top panels are $\chi'_\epsilon$ amplitude in K, and the horizontal are months beginning in Jan 2005(0)–Dec 2014(119). The bottom panels in Figs. 3.22–3.24 show the qualitative relationship between the Niño 3.4 index and respective OFES timeseries with correlation coefficient value (R) on the vertical (y-axis), and the lag value in months on the horizontal (x-axis).

The EEP has the highest correlation with the ENSO index of the locations shown in Figs. 3.22, 3.23, and 3.24. Results of the largest cross-correlation coefficient values of $\chi'_\epsilon$ timeseries on $\sigma_\theta$ 25 in the Pacific to the Niño 3.4 index summarized in Table 3.2 and here:

1. Off Baja California, maximum correlation of spiciness (R = $\sim$0.5) lags the Niño 3.4 index by 4 months Fig. 3.22
2. In the EEP, maximum correlation of spiciness (R = $\sim$0.8) lags the Niño 3.4 index by 1 month Fig. 3.23
3. WEP, maximum correlation of spiciness (R = $\sim$0.5) and maximum anti-correlation (R = $\sim$−0.6) lags the Niño 3.4 index by 38 months and 48 months respectively Fig. 3.24

<table>
<thead>
<tr>
<th>Location</th>
<th>Lat/Lon</th>
<th>Max Correlation</th>
<th>Lag (months)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Baja</td>
<td>N E</td>
<td>R = $\sim$0.5</td>
<td>4</td>
</tr>
<tr>
<td>East Eq. Pac</td>
<td>N E</td>
<td>R = $\sim$0.8</td>
<td>1</td>
</tr>
<tr>
<td>WEP</td>
<td>N W</td>
<td>R = 0.5 and $\sim$0.6</td>
<td>38 and 48</td>
</tr>
</tbody>
</table>
Figure 3.22: $\chi'$ Timeseries comparison west of Baja California and Niño 3.4 index

Timeseries comparison of anomalies of $5^\circ \times 5^\circ$ box West of Baja (removal of annual cycle) of $\sigma_0 25$ of OFES Argo ensemble and monthly Niño 3.4 index superimposed (dashed blue line). Vertical axis is temperature anomaly in K and horizontal axis is time in months from beginning of timeseries in January 2005 ending December 2014 (120 months) in upper panel. Lower panel is correlation coefficient “R” in legend with a maximum correlation at a lag of four months.
Figure 3.23: $\chi'$ Timeseries comparison in eastern equatorial Pacific and Niño 3.4 index

Timeseries comparison of anomalies of EEP box (removal of annual cycle) of $\sigma_\theta$ 25 of Argo ensemble and monthly Niño 3.4 anomalies superimposed (dashed blue line). Vertical axis is temperature anomaly and horizontal axis is time in months from beginning of timeseries in January 2005 ending December 2014 (120 months) in upper panel. Lower panel is correlation coefficient “R” in legend with a max lag correlation at one month.
Figure 3.24: Timeseries comparison of $\chi'_\epsilon$ and Niño 3.4 index north of New Guinea

*Top panel showing timeseries of $\chi'_\epsilon$ (using $\theta$ data from OFES Argo ensemble) on $\sigma_\theta$ 25 and Niño 3.4 index north of New Guinea (dashed blue line), with horizontal axis as time in months from beginning of timeseries in January 2005 ending December 2014 (120 months). Lower panel is correlation coefficient “$R$” in legend with a max lag correlation of 38 months. It is noted that a max lag of 38 months, for a low-frequency signal in a relatively short timeseries may not be a reliable indicator of correlation.*

3.6 Signal Propagation

Two large amplitude spiciness signals on $\sigma_\theta$ 25 are observed and described in this OFES experiment as they propagate along mean advective pathways from the subtropics to the tropics. The first is a positive warm-salty signal that appears at the outcrop in 2007 and is compared to observations of a signal described by Sasaki et al. (2010) and Kolodziejczyk and Gaillard (2012). The second, a cool-fresh negative signal appearing in 2012, is described for the first time in this paper. Both signals are described in greater detail in following sub-sections.

The Hovmöller diagrams of Figs. 3.25 and 3.26 provide an initial look at spiciness signal propagation across the Pacific using $\chi'_\epsilon$ normalized data. The top panel in both figures are $M$ contours and outlined boxes for the data used in the bottom panel Hovmöllers. The red boxes are Hovmöller’s with meridional averaging and the blue box is zonal averaging about the equator. The bottom Hovmöller panels are arranged south to north, left to right.

In Fig. 3.25, the meridionally averaged Hovmöller’s (red boxes) converge to the equator in the

As a general rule, and taken into account for this study, only lags/leads of no greater than $\frac{1}{3}$ the length of the timeseries (or 36 months) are considered reliable indicators of correlation.
western Pacific to the zonally averaged blue box. Fig. 3.26’s meridional Hovmöllers converge in the mid-Pacific (blue box) and capture data from the eastern Pacific.

In the bottom left of Fig. 3.25’s Hovmöller, a negative signal appears from the east in the Southern Hemisphere at 140°W in 2005, and propagates west and north towards the equator, reaching it in 2007 (middle Hovmöller). Also observed in the Southern Hemisphere is a positive signal in 2011 propagating west and north and reaching the equator in 2014 at the end of the experiment. As these signals of alternating sign spiciness reach the equator, they present an apparent “spine” feature at 0° in the middle Hovmöller.

In Fig. 3.25, spiciness signals are also observed reaching the equator from the Northern Hemisphere. The first is a positive signal reaching the equator in 2005, followed by a negative signal from mid-2007 to 2009. In 2009, a positive signal enters the right Hovmöller at 160°W and propagates west and equatorward until mid-2012, until it attenuates at ~7°N. The ends of the positive signals in 2007 and 2013 at 160°E in Fig. 3.25’s right Hovmöller implies a spiciness period of approximately 6 years.

The signals in the Hovmöllers of Fig. 3.26 tell a different story of signal propagation to the equator. The Southern Hemisphere signals show of low-frequency spiciness period and propagation west to the equator similar to Fig. 3.25. However, the right Hovmöller of the Northern Hemisphere is noisy and signals are difficult to discern, implying there is not a clear path to the equator in the limits of the right Hovmöller’s data field from 120°W–180°.
Figure 3.25: Signal Propagation to the Equator: West-Pacific

Combined Hovmöller diagrams of normalized spiciness in the Argo ensemble. Red boxes show Hovmöllers with meridional averaging and blue box is zonal averaging about the equator. The panels are arranged South to North, left to right. In the western Pacific signals from both hemispheres are observed reaching the equator.
Signals Reaching the Equatorial Region (MidPac)

Figure 3.26: Signal Propagation to the Equator: Central-Pacific

Combined Hovmöller diagrams of normalized spiciness in the Argo ensemble. Red boxes show Hovmöllers with meridional averaging and blue box is zonal averaging about the equator. The panels are arranged South to North, left to right. In the central Pacific clear signals from the southern hemisphere appear to reach the central Pacific while the Northern Hemisphere appears to be subject to more seasonal activity, especially to the East.

3.6.1 2007 Positive spice signal

The 2007 positive signal is observed in Figs. 3.27–3.29, appearing off Baja in 2007 and growing through 2008 as it propagates south and westward with mean advection until 2013. Figs. 3.27 and 3.28 display spiciness values alone in the ensemble and individual realization respectively with a contour threshold of ±0.15 psu, and M contours over-plotted as solid lines in Figs. 3.27 and 3.28 (dashed in Fig. 3.29). The positive signal emerges from the line of seasonally masked values in the 2007 panels of Figs. 3.27, 3.28, and 3.29, and moves southwestward along the dashed M lines in 2008, and reaches the Hawaiian Islands in 2009. As expected, the spatial extent in the ensemble panels of Fig. 3.27 is less than that in an individual realization Fig. 3.28. The signal is most prominently
observed in Fig. 3.29, where $\chi'$ is normalized by $\sigma_1(\chi)$ as the denominator, and a solid contour line has been set at $\chi'_\text{normalized} = 1.5$.

This positive signal is picked up on the right of Fig. 3.25 at 160°W in 2009 and is visible in the western equatorial region from 2011–2013. All years of normalized spiciness are available in Appendix F, Figs. F.1–F.10.
Figure 3.27: Positive spiciness signal in the North Pacific observed from $\chi'_e$

Annual mean spiciness from OFES ensemble average salinity anomalies on isopycnal $\sigma_\theta 25$. Each panel shows $\bar{\chi}'_{e,\text{year}}$ with thick, colored contour thresholds of $\pm 15$ psu. The thin black contours represent temporally averaged Montgomery Potential isolines, note the appearance of a positive spiciness signal in the North Pacific beginning in 2007, which propagates west and equatorward through 2009 before being lost using this threshold technique. It is still visible when normalized by the standard deviation at locations.
Figure 3.28: Positive spiciness signal in the North Pacific observed from $\chi'_i$

Annual mean spiciness from OFES single realization salinity anomalies on isopycnal $\sigma_\theta 25$. Each panel shows $\chi_i^{\text{year}}$ with thick, colored contour thresholds of $\pm 0.15$ psu. Thin black contours represent temporally averaged Montgomery Potential isolines, note same signal is visible as the $\chi'_e$ panels, however there is a wider spatial spread and larger amplitude at the barycenter of the signals.
The panels displaying annually averaged anomaly all show similar structure, with the ensemble
panel showing smoother structure, and less spatial spread and amplitude of mean spiciness.

Fig. 3.30 shows correlation of the Baja Timeseries spiciness (averaged within 19°N–24°N, 115°W–110°W) in the OFES Argo ensemble, with the ensemble average data set spatially across the domain on $\sigma_\theta$ 25. The panels of Fig. 3.30 are in sequence from lag 0 to 36 months lead. The highest correlation values are present in the top panel off Baja where the timeseries is taken at a lag of zero. The red values of high correlation move westward along M contours in the second and third from top panels, at 12 and 24 months lead respectively. The high correlation values are distinguishable for 36 months until a visually identifiable correlation is no longer observed. There are also high anti-correlation values (blue) to the west of the high correlation values which also move west in subsequent lead panels. Figures from 36 months lag through 36 months lead are available in Appendix D, Figs. D.2–D.3.
Figure 3.30: Baja cross correlation maps of $\chi'_{\text{BAJA}}(t)$ with $\chi'(t, x, y)$ from 0 to 36 months lead $\chi'_{\text{Baja}}(t)$ cross correlated with $\chi'(t, x, y)$ on $\sigma_\theta 25$. From top to bottom the panels show 0 to 36 months lead in six month intervals with positive correlation values in red and negative in blue (all panels use same colorbar as top map at lag = 0).
3.6.2 2012 Negative spice signal

In Figs. 3.31–3.33, a negative signal appears at the seasonal outcrop line west of Yeager and Large (2004)’s southeast Pacific spiciness generation zone in 2012. Beginning at \( \sim 15^\circ S \ 110^\circ W \), it moves north and westward along \( M \) isolines to the end of the timeseries in 2014 to \( \sim 10^\circ S \ 160^\circ W \). In Fig 3.31 of ensemble data, the \(-15\) psu contour spans \( \sim 2000 \) km at \( 120^\circ W \) when it appears in 2012, and decreases to \( < 500 \) km at \( 155^\circ W \) in 2014. In the single realization however, the signal strength at this contour threshold is much greater, with the \(-15\) psu contours covering \( \sim 4000 \) km in 2012, shrinking to \( \sim 2000 \) km in 2014. A glimpse of this cool/fresh signal also appears in Fig. 3.25 at the top left of the Hovmöller in 2013 at \( 130^\circ W \), and in Fig. 3.26 in 2012 at \( 120^\circ W \). It propagates until the end of the timeseries with a signature \( \chi'_{\text{normalized}} < -1.00 \).

At the time of this publication, the negative signal has not been described in other model or observation studies.
Figure 3.31: Ensemble Negative Spiciness Signal in the Southern Hemisphere observed from $\chi'$. Annual mean negative salinity anomalies on isopycnal $\sigma_\theta 25$ with thick contours at $\pm \pm15$ psu. Thin black lines represent temporally averaged Montgomery potential isolines. Note the appearance of a negative spiciness signal in the Southern Hemisphere beginning in 2012, which propagates westward through 2014 when our timeseries ends. There is also a growing large positive spiciness signature in the eastern Pacific.
Figure 3.32: Single Negative Spiciness Signal in the Southern Hemisphere observed from $\chi_i$

Annual mean negative salinity anomalies on isopycnal $\sigma_25$ with thick contours at $\pm 0.15$ psu. Thin black lines represent temporally averaged Montgomery Potential isolines. Note the appearance of a negative spiciness signal in the Southern Hemisphere beginning in 2012, which propagates westward through 2014 when our timeseries ends. There is also a growing positive spiciness signature in the eastern Pacific.
Figure 3.33: Annually Averaged Spiciness Anomaly normalized by Temporal Standard Deviation $\sigma_t$ (2012-2014)
3.6.3 Signal comparisons

For both positive and negative signals, the $\chi'_c$ closely follow the isolines of mean Montgomery Potential and are traceable for several years when looking at spiciness alone (no normalization). The individual realizations present greater spatial spread and higher amplitudes of the spiciness threshold than the ensemble average. Both Figs. 3.27 and 3.28 can be compared to Sasaki et al. (2010)’s plot reproduced as Fig. 1.8, lower right 2007 panel, and Fig. 1.9 visible next to y-axis in 2006 and left of cool/fresh signal. The individual realizations show some departure from these paths which can be seen even when comparing the 12-month running averages of $\bar{\chi}$ in both positive and negative signals of Figs. 3.27 and 3.28, and Figs. 3.31 and 3.32 respectively.

A positive signal in the EEP begins to grow in amplitude and spread spatially west and meridionally from 2012–2014. Figs. 3.31–3.33 show this feature well with contours, and it is the most pronounced signal in raw annual average spiciness plots in Appendix C, Figs. C.8–C.10.

3.6.4 Relation between spiciness anomalies and geostrophic current variability

Figs. 3.34 and 3.36 are used to see if basin scale wave processes are involved in the propagation of spice signals.

Fig.3.34 shows the temporal standard deviation of $M'_i$ on $\sigma_25$. Temporally averaged $M$ isolines are also plotted as an overlay to show mean flow paths. Areas of high variability in $M$ are visible in red and black on the log scale colorbar and areas of low variability are yellow. Zonal bands of $\sigma_t(M'_i)$ sweep across the Pacific basin; there is a low-latitude band at $\sim 5^\circ$N and a subtropical band at $\sim 15^\circ$N in the North Pacific.

Looking back at Fig.3.16 timeseries of $\chi'_i$ and $\chi'_c$ in the Mid-Pacific, the amplitude of $\chi'$ is smaller in these zonal bands of $M'$ than in locations with high spice variance with $|\chi'| < \sim .5$K and $\sigma_c(\chi') < .2$K. However, when normalized by standard deviation or cross-correlation the signal is more apparent. The areas of greater $\sigma_t(M'_i)$ can be compared to Appendix F, Figs. F.1–F.10, where oscillations from positive to negative spiciness across the Pacific subtropics, beginning in Hawaii at $\sim 160^\circ$W at 10$^\circ$N–25$^\circ$N and extending to the western boundary, are observed in the sequence of annual panels.

Harking back to Fig.3.30, this low-frequency pattern across the basin between 10$^\circ$N–25$^\circ$N is observed in the spatial cross-correlation of $\chi'_c$ across the Pacific domain with the Baja $\chi'_cBaja$ ensemble timeseries (Fig.3.17), with correlation coefficients of up to $\pm .8R$ in magnitude. In Appendix D, the pattern of correlation and anti-correlation moves across the Pacific subtropics is observed from three years lag to three years lead in Figs. D.2–D.3.

Fig.3.35’s top panel shows timeseries of the Argo ensemble’s $\sigma_25$ depth (shown as $\tilde{\rho}_25$ in dashed lines), $\chi'_c$ in K as the thick red line, and $M'_i$ in m$^2$s$^{-2}$, with months in the experiment on the x-axis (0 is Jan 2005). Note that this area of the mid-Pacific has relatively low amplitude spiciness fluctuations.
In the bottom panel of Fig. 3.35, cross correlation between $z'_\epsilon$ and spiciness $\chi'_\epsilon$, and depth and $M$ is shown with the x-axis representing months of lag (negative values represent lead). The blue line shows $R \simeq .5$ at lag = 0 when $z'_\epsilon$ is cross-correlated with $M$. The red line in the bottom panel of Fig. 3.35 is $z'_\epsilon$ cross-correlated with $\chi'_\epsilon$, with $R \simeq -.4$ at lag = 0.

Correlations of $\chi'_\epsilon$ and $M'_\epsilon$ timeseries in the mid-Pacific with $\chi'_\epsilon$ across a longitudinal domain are both displayed in Fig. 3.36’s Hovmoller diagram. The y-axis values represent lag and lead in months, the x-axis is East longitude. The colors in the Hovmoller represent $R$ values of $\chi'_\epsilon$ (mid-pac) correlated with $\chi'_\epsilon$ along 20°N across the basin. To the right of Fig. 3.36 in the eastern Pacific, any correlation is difficult to discern. However, west of 135°W (labeled 225 on x-axis), a clear pattern of correlation and anti-correlation values are present, and are suggestive of mean advection for spiciness. The speed of the wave can be estimated from the slope of a linear fit along the correlation values. The solid white line overlays maximum correlation values in the plot, spanning 50 degrees longitude at 20° which is \( \sim 5200 \) km. The time covered is approximately 30 months between 125°E and 175°E, therefore the slope of the line implies a propagation speed of approximately \( \sim 7 \) cm s\(^{-1}\).

The contour lines in Fig. 3.36 are $R$ values of $\chi'_\epsilon$ (mid-pac) correlated with $M$ along the latitudinal band at 10°N. The solid black line is drawn across these peak correlation values, which also moves east to west across the Pacific basin. The slope of the black line indicates a speed close to that of the white line, \( \sim 7 \) cm s\(^{-1}\), and the wave has an approximate 90° phase-shift relationship between maxima of correlation between $\chi'_\epsilon$ and $M'_\epsilon$ and the colormap $R$ values of the spiciness timeseries in the mid-Pacific correlated with spiciness of the ensemble longitudinally.

West of the dateline, this covariability of spiciness and Montgomery potential is consistent with Eq. (1.1) of anomalous advection acting on a spiciness anomaly propagating westward with the time mean flow (see also Kilpatrick et al., 2011). The temporal average here of spiciness is a positive, albeit weak, meridional gradient at 20°N transitioning from cool and fresh waters closer to the equator, to warm and salty waters at the outcrop lines at 25°–30°N (Fig. 3.6). This spatial gradient is consistent with Johnson and McPhaden (1999)’s observations (Fig.3.8), east of 170°W and north of \( \sim 5° \)N, the waters become saltier by up to .2 PSS between 10°–20°N at 180°.

The positive correlation of spiciness and the Montgomery streamfunction to its east in Fig. 3.36 implies a meridional geostrophic flow in the region between maxima of the spiciness and Montgomery potential anomalies. This implies that a warm (cool) spiciness anomaly is associated with poleward (equatorward) geostrophic currents, and a cooling (warming) tendency due to anomalous advection. Consistently, the correlation of spiciness with itself is negative at lags of about 24 months. The phase lag between $\chi'_\epsilon$ and $M'_\epsilon$ is observed, with $\chi'_\epsilon$ leading $M'_\epsilon$ by approx 24 months.
Figure 3.34: Temporal standard deviation of Montgomery Potential Anomalies $\sigma_t(M'_e)$ $\sigma_t(M'_e)$ in the Argo ensemble. The contours are temporally averaged $M$ isolines, with the colormap showing $\sigma_t(M'_e)$ on the isopycnal surface on a log$_{10}$ scale. Areas of red to black are higher temporal variance, while areas of yellow imply low variance in $M'$. Note the zonal bands of variance in the low and mid latitudes, particularly at $10^\circ$N–$20^\circ$N.

Figure 3.35: Mid-Pacific $\chi'_e$ cross-correlation of depth, spiciness and Montgomery potential

Top panel showing a timeseries of $z'_e$ (dashes), $\chi'_e$ (red), and $M'_e$ (blue) in the Mid-Pacific at $20^\circ$N $170^\circ$E. Correlation of the Mid-Pacific $\chi'_e$ timeseries with $\chi'_e$ and $M'_e$ anomalies with $\sigma_\theta$25 depth ($z'_e$). The bottom panel is lag and lead in months of the correlation value. Here we find the highest correlation ($R = \sim .5$) at lag 0 of the isopycnal depth perturbation and $M'$. Note that there is a lag and lead of approximately 24 months between $z'$ and $\chi'_e$ and the largest anti-correlation at lag 0.
Figure 3.36: Cross correlation of spiciness and Montgomery potential

Correlation of the Mid-Pacific $\chi'_e$ timeseries at $20^\circ N$ $170^\circ E$ with $\chi'_e$ and $M'_e$ anomalies at $20^\circ N$ across the Pacific. The vertical axis is lag and lead in months and horizontal axis is East longitude across the Pacific basin. Montgomery potential anomalies correlation coefficients across the Pacific are over-plotted contour lines. Solid contour lines represent positive correlation while dashed contour lines represent negative values. Note that there is a phase lag between $\chi'_e$ and $M'_e$ with $\chi'_e$ leading $M'_e$ by approx 24 months.
CHAPTER 4
DISCUSSION

This is the first study to separate and approximate spiciness variance from mesoscale eddies and external forcing in order to clarify their roles and importance in the Pacific upper-thermocline. Analysis of OFES individual realizations and the Argo ensemble average revealed that eddies account for up to $\sigma_t(\chi) \sim 0.6-0.8$ K in locations of high eddy-induced variance such as Baja and the WEP regions. In the EEP and middle-Pacific however eddy-induced spiciness variance is low with values of, $\sigma_t(\Delta\chi) < 0.1$ K and $\sigma_\epsilon < 0.1$ K.

The absence of eddy-induced spiciness variance in the EEP requires further investigation. Tropical instability waves (TIW) in the region may have an impact on the pycnocline/thermocline in the EEP (Willett et al., 2006). Another location where eddy-induced variance is not present in OFES is the eddy train from Tehuantepec at 15°N 95°W where eddies form from strong, intermittent wind jets resulting in eddies 180-500 km in diameter, have rotational speeds in excess of 1 m s$^{-1}$, and which depress the pycnocline from 60 to 145 m at the eddy center (Willett et al., 2006). The presence of a large pycnocline displacement signal implies they are highly baroclinic.

The absence of eddy-induced variance here may be due to $\sigma_\theta$ being too deep to be affected by TIWs in the EEP, however the shallowness of the EEP’s thermocline/pycnocline would lead us to hypothesize that this is not the case. A look at a shallower isopycnal may provide clarity. The TIW period of 20-40 days may play a factor by having the waves averaged out in our monthly averaged data-sets. Slightly south off Papagayo and Panama in Fig. 3.18 there does appear to some eddy-induced variance of 0.2–0.4 K in OFES, implying that stochastic baroclinic eddies are present.

Spiciness signals of both signs are observed in OFES propagating from the subtropics to the equator, and within the limits of the ten year timeseries, the western boundary. The 2007 warm/salty signal in the North Pacific’s follows $M$ contours in Fig. 3.3. The ensemble average signal appears to closely follow Johnson and McPhaden (1999)’s North Pacific mean flow paths taking nearly 7 years to cross the basin to reach the western boundary bifurcation near 15°N 125°E in Fig. 3.3, where it may go north to the Kuroshio, south to the Mindanao current, or perhaps a combination of both. The 2007 positive signal appears in Sasaki et al. (2010)’s Figs. 1.8 and 1.9 but is not described. Its presence in OFES adds further credence to the accuracy of OFES simulations with comparisons to observations. This signal is also consistent with the positive salinity anomaly Kolodziejczyk and Gaillard (2012) identified from Argo observations in the Northern Hemisphere during the same time period which follows a similar path on the deeper $\sigma_\theta$ 25.5 isopycnal. It moves from the northeast Pacific south and west-ward until the end of the timeseries in 2011, however there is significant attenuation after 2009 (Kolodziejczyk and Gaillard, 2012).

The 2012 cool/fresh signal described in Figs. 3.31 and 3.32, follows the more direct path described by Johnson and McPhaden (1999) northwestward to the equator. The signal appears to
arrive at the equatorial region in 3–4 years which is consistent with Furue et al. (2018)’s findings and Johnson and McPhaden (1999)’s described South Pacific advection path. Since this signal is negative and is coming from Yeager and Large (2004)’s SEP formation zone, it is possible that this signal is a result of the mean advection carrying a negative anomaly during a time when the SEP does not inject a positive anomaly at the base of the mixed layer as described in Section (1.5.1).

The sign and origin of spiciness signals reaching the equator may be important factors for statistical predictability of ENSO events. Our look at ENSO modulation of spiciness in the thermocline was consistent with Wang et al. (2004a,b)’s findings of warmer and saltier waters in the EEP corresponding with El Niño events. Also investigated were ENSO modulation of spiciness in the thermocline, and spice signal propagation through dynamic processes.
CHAPTER 5
CONCLUSIONS AND FUTURE WORK

From the results of this study we draw the following conclusions:

1. Temporal spiciness variance is large in regions of large isopycnal temperature gradients

2. Propagation of spiciness signals appears via mean advective pathways and possible wave processes

3. Eddies play a large role in spiciness variance in the upper-thermocline off Baja and WEP

4. Externally forced low-frequency spiciness variation occurs in Baja, WEP, and eastern equatorial Pacific

Additionally, our findings support the null hypothesis of spice signals advecting from the extratropics to the tropics where they may affect equatorial climate (Gu and Philander, 1997; Johnson and McPhaden, 1999; Kilpatrick et al., 2011).

A high degree of spiciness variance was found in the WEP with roughly equal amounts of variance from external forcing and eddies. Understanding spice variance in the WEP may help explain the origins of Indonesian Throughflow waters, lead to improved understanding of communication between the Pacific and Indian Oceans, and enhance ENSO predictability. While this work was primarily focused on where variance occurs, its amplitude, and how signals are propagated, more investigation of spiciness in specific high variance regions such as the WEP is required. Questions arise such as:

1. Do spiciness variations in the WEP propagate into the Indian Ocean via the Indonesian Throughflow?

2. Do large-scale seasonal eddies such as the Mindanao and Halmahera eddies affect spiciness in the Indonesian Throughflow and waters communicated between the Pacific and Indian Ocean?

Additional investigation is required to see if a subtropical-to-tropical pathway in the eastern North Pacific is significantly affected by eddy fluxes. This study does not present sufficient evidence to draw conclusions regarding a meridional shortcut from mean advective pathways for spice signals to reach the equator in the Northern Hemisphere. While we observe small differences in structure, and a greater spatial spread and amplitude in the signals from individual realizations, more analysis is required to draw conclusions on the presence of a shortened meridional transport path from established mean flow paths to the equator. The $\sigma_e(\bar{\chi})$ of the annual anomalies in Fig. 3.19 shows $\sigma_e(\bar{\chi}) < .1K$ south of 15°N in the eastern Pacific. This is in line with Schneider et al. (1999a)’s
findings of little eddy variance South(North) of 18°N(S). If there was a meridional corridor of significance providing a shortcut to the equator, greater variance in this region would be expected.

Off the Baja region, we find spiciness variance is subject to both externally forced and eddy-induced processes. Additionally we show that variability in areas of high property gradients is partially a result of eddy activity.

We identify clear, basin-wide bands of pressure perturbations at 5°N and 15°N and at 10°S–15°S in the Pacific. There is some correlation with the pressure perturbations and isopycnal heave. The ensemble average does reduce mesoscale eddy noise and increase the signal to noise ratio for identifying dynamical signals such as Rossby and Kelvin waves in OFES. The 90° phase shift between $\chi'_{\text{mid}}(\text{mid}-\text{pac}) - \chi'_{\text{mid}}$ and $\chi' - M'_{\text{mid}}(\text{mid}-\text{pac})$ in Fig. 3.36 requires further analysis to describe the precise dynamical processes and conduct a proper significance test of the correlation.

More investigation of spiciness variability in the EEP is required, specifically the variance eddies at low latitudes off of Tehuanatepec, Papagayo, and Panama where there is generation of eddies by wind stress curl in the lee of topography (Willett et al., 2006). This study concluded that variance in the EEP is caused almost entirely from atmospheric forcing. However questions pertaining to the eastern tropical Pacific arise such as:

1. To what degree is upper-thermocline spice variance in the eastern tropical Pacific due to barotropic instabilities of the shear in the equatorial current system?

2. Is OFES wind forcing with sufficiently high resolution in space and time to reproduce these eddies?

The thermocline T-S relationship variability and ENSO events also requires further investigation, such as identifying a statistical relationship between spiciness amplitude and origin, and ENSO events.
APPENDIX A
VARIABLES AND SYMBOLS
<table>
<thead>
<tr>
<th>Symbol</th>
<th>Nomenclature</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\rho$</td>
<td>rho</td>
<td>density in kg m$^{-3}$</td>
</tr>
<tr>
<td>$\sigma_\theta$</td>
<td>sigma-theta</td>
<td>density anomaly of seawater 1000 kg m$^{-3} + \sigma_\theta$</td>
</tr>
<tr>
<td>$\gamma_n$</td>
<td>$\gamma_n$</td>
<td>neutral density anomaly kg m$^3$</td>
</tr>
<tr>
<td>$\theta$</td>
<td>theta</td>
<td>Conservative Temperature in degrees Celcius</td>
</tr>
<tr>
<td>$S$</td>
<td>salinity</td>
<td>Salinity in Practical Salinity Units (PSU)</td>
</tr>
<tr>
<td>$\eta$</td>
<td>eta</td>
<td>Sea Surface Height</td>
</tr>
<tr>
<td>$\Delta$</td>
<td>Delta</td>
<td>Deviations</td>
</tr>
<tr>
<td>$\sigma^2$</td>
<td>sigma squared</td>
<td>Variance</td>
</tr>
<tr>
<td>$\sigma_\epsilon$</td>
<td>sigma epsilon</td>
<td>standard deviation between realizations in an ensemble</td>
</tr>
<tr>
<td>$\sigma_t$</td>
<td>sigma 't'</td>
<td>standard deviation over time</td>
</tr>
<tr>
<td>$M$</td>
<td>N/A</td>
<td>Montgomery Potential m$^2 s^{-2}$</td>
</tr>
<tr>
<td>$f$</td>
<td>N/A</td>
<td>Coriolis Parameter s$^{-1}$</td>
</tr>
<tr>
<td>$\chi$</td>
<td>chi</td>
<td>Spiciness in Practical Salinity Units or Kelvin</td>
</tr>
</tbody>
</table>
APPENDIX B
AVERAGE SPICINESS ON $\sigma_0^{25}$
Figure B.1: Average Temperature on Isopycnal $\bar{\theta}_t$ of individual realizations in ARGO experiment. Note small-scale differences in structure where temperature gradients of each $R_j$ model are, compared to the smoothing present in the ensemble run of Fig. 3.6.
Figure B.2: Average Temperature on $\sigma_{25}$ in ARGO Realization 1
$\bar{\theta}_i$ in individual realization. Black lines are contoured at 1°C. Note small structural differences in the spiciness contours between $\chi_i$ and the smoothing in the ensemble average $\chi_\epsilon$.

Figure B.3: Average Temperature on $\sigma_{25}$ in ARGO Realization 2
$\bar{\theta}_i$ in individual realizations. Black lines are contoured at 1°C. Note small structural differences in the spiciness contours between $\chi_i$ and the smoothing in the ensemble average $\chi_\epsilon$. 
Figure B.4: Average Temperature on $\sigma_{25}$ in ARGO Realization 3

$\bar{\theta}_i$ in individual realizations. Black lines are contoured at 1$^\circ$ C. Note small structural differences in the spiciness contours between $\chi_i$ and the smoothing in the ensemble average $\chi_\epsilon$.

Figure B.5: Average Temperature on $\sigma_{25}$ in a single ARGO Realization 4

$\bar{\theta}_i$ in individual realizations. Black lines are contoured at 1$^\circ$ C. Note small structural differences in the spiciness contours between $\chi_i$ and the smoothing in the ensemble average $\chi_\epsilon$. 
Figure B.6: Average Temperature on $\sigma_\theta 25$ in a single ARGO Realization 5 $\bar{\theta}_t$ in individual realizations. Black lines are contoured at 1°C. Note small structural differences in the spiciness contours between $\chi_i$ and the smoothing in the ensemble average $\chi_e$. 
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ANNUAL SPICINESS IN ARGO REALIZATIONS $\chi'(T_{YR})$
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CROSS CORRELATION OF TIMESERIES KEY WITH SPATIAL SPICINESS
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LAGGED CORRELATIONS OF $Z'$, $\chi'$, AND $M'$ AT LOCATIONS OF INTEREST
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ANNUALLY AVERAGED SPICINESS ANOMALY
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Figure F.4: Annually Averaged Spiciness Anomaly normalized by Temporal Standard Deviation $\sigma_t$ (2008)
Figure F.5: Annually Averaged Spiciness Anomaly normalized by Temporal Standard Deviation $\sigma_t$ (2009)

Figure F.6: Annually Averaged Spiciness Anomaly normalized by Temporal Standard Deviation $\sigma_t$ (2010)
Figure F.7: Annually Averaged Spiciness Anomaly normalized by Temporal Standard Deviation $\sigma_t$ (2011)

Figure F.8: Annually Averaged Spiciness Anomaly normalized by Temporal Standard Deviation $\sigma_t$ (2012)
Figure F.9: Annually Averaged Spiciness Anomaly normalized by Temporal Standard Deviation $\sigma_t$ (2013)

Figure F.10: Annually Averaged Spiciness Anomaly normalized by Temporal Standard Deviation $\sigma_t$ (2014)
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