Observational Evidence of Winter Spice Injection

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ABSTRACT

Temperature and salinity \((T-S)\) profiles from the global array of Argo floats support the existence of spice-formation regions in the subtropics of each ocean basin where large, destabilizing vertical salinity gradients coincide with weak stratification in winter. In these characteristic regions, convective boundary layer mixing generates a strongly density-compensated (SDC) layer at the base of the well-mixed layer. The degree of density compensation of the \(T-S\) gradients of an upper-ocean water column is quantified using a bulk vertical Turner angle \((\text{Tu}_b)\) between the surface and upper pycnocline. The winter generation of the SDC layer in spice-formation zones is clearly seen in Argo data as a large-amplitude seasonal cycle of \(\text{Tu}_b\) in regions of the subtropical oceans characterized by high mean \(\text{Tu}_b\). In formation regions, Argo floats provide ample evidence of large, abrupt spice injection \((T-S\) increase on subducted isopycnals due to vertical mixing) associated with the winter increase in \(\text{Tu}_b\). A simple conceptual model of the spice-injection mechanism is presented that is based on known behavior of convective boundary layers and supported by numerical model results. It suggests that penetrative convective mixing of a partially density-compensated water column will enhance the Turner angle within a transition layer between the mixed layer and the upper pycnocline, generating seasonal \(T-S\) increases on density surfaces below the mixed layer. Observations are consistent with this hypothesis. In OGCMs, regions showing high \(\text{Tu}_b\) mean and seasonal amplitude are also the sources of significant interannual spice variability in the permanent pycnocline. Decadal changes in the North Pacific of a model hindcast simulation show qualitative resemblance to the observed multiyear time series from the Hawaii Ocean Time series (HOT) station ALOHA. Modeled pycnocline variations near Hawaii can be linked to high \(\text{Tu}_b\), seasonality and winter spice injection within a formation region upstream of ALOHA, suggesting that spice injection may explain the origins of observed large, interannual variations on isopycnals in the ocean interior.

1. Introduction

One of the fundamental simplifications historically employed in studies of ocean ventilation is that the deepest winter mixed layer links the surface to the ocean interior, so that subducted properties reflect the late-winter sea surface temperature and salinity (Iselin 1939; Stommel 1979; Luyten et al. 1983; Woods 1985). Consequently, in an adiabatic ocean interior with geostrophic flow along isopycnals, variability on deep isopycnals should be traced to origins at the ocean surface. However, in some regions there have been noticeable failures to link observed trends in isopycnal salinity to upstream changes at winter density outcrops (Kessler 1999; Suga et al. 2000), which cannot entirely be explained by the poor knowledge of surface forcing fields and other mixed layer budget terms. The success of the theory of the ventilated thermocline suggests that coupling an immiscible geostrophic interior to a mixed layer of uniform temperature, salinity, and density is a powerful approximation, and yet “one need not expect such an elementary model to replicate all features in all geographic regions of the world” (Luyten et al. 1983).

In ocean general circulation models (OGCMs), the largest interannual variations in ocean spice arise as a result of vertical diapycnal mixing in the specific subtropical formation regions where large salinity inversions coincide with weak winter stratification (Yeager and Large 2004, hereafter YL04; Luo et al. 2005). The abrupt, positive changes in temperature and salinity \((T-S)\) on subducted density surfaces seen in such simulations are related to low density ratio \(R_p\) at the base of the winter mixed layer. Surface flux varia-

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tions are not good indicators of the magnitude and timing of these diapycnal injection events, which can significantly alter the $T-S$ characteristics on shallow isopycnals at locations well removed from their winter outcroppings. Therefore, an understanding of the origins of coincident $T-S$ variability on isopycnals (spice variability) may require explicit consideration of late-winter profiles in which a strongly density-compensated (SDC) layer exists between the mixed layer and the interior.

Isopycnal anomalies that subduct into the permanent pycnocline from such regions would trace their origins to diabatic processes that take place below the surface in a layer of large, compensating vertical $T-S$ gradients. This implies that that in some regions one might expect large discrepancies between isopycnal characteristics in the interior and surface conditions at outcrop locations. While recent studies of observed mixed layer depth have drawn attention to regions where density-compensation results in large discrepancies between density- and temperature-based mixed layer depth criteria (Kara et al. 2003; de Boyer Montégut et al. 2004), the significance of such regions in giving rise to ocean variability on density surfaces deserves more attention.

Direct observational evidence of large-scale change in ocean spice through the proposed injection mechanism (diapycnal mixing below the winter well-mixed layer that transmits high $T-S$ surface properties onto subducted density surfaces) has been lacking, but is essential to give credence to model-based findings. If the model results are accurate, then interpretation of deep-ocean variations may be complicated by the potentially significant role that diapycnal processes play in generating anomalies on isopycnal surfaces in some locations. There is mounting evidence that the southeast Pacific Ocean is one such region where winter spice injection is particularly active, as suggested by YL04. Hydrographic data have yielded evidence of a large body of thermocline water with $R_p < 2$ around 20°S, 110°W (Tsuchiya and Talley 1998; Wong and Johnson 2003), which is presumably a result of $T-S$ gradient alignment through convective mixing in winter. Recently, Johnson (2006) analyzed data from two Argo floats (see online at www.argo.net) from this region, World Meteorological Organization (WMO) 4900451 and WMO 4900454, which clearly show spice anomaly generation on subducted density surfaces associated with late-winter reduction in $R_p$. At roughly 200 m, $R_p$ is observed to decrease from a low background level near 1.8 to as low as 1.5 for a few weeks at the base of the winter mixed layer. Other observations of regional water mass variations, including the Atlantic studies of Jenkins (1982), Pollard and Pu (1985), and Gordon (1981), may offer indirect evidence of spice injection. Possible connections between these antecedent findings and the present study are discussed in the summary.

While the ocean has never been more extensively measured than at present, placing the spice-injection mechanism on a firm observational footing poses a considerable challenge. The most active spice-formation zones (YL04) are in the southeast Pacific and central South Atlantic Ocean regions, which are historically undersampled in salinity (Locarnini et al. 2002; de Boyer Montégut et al. 2004). Anomaly creation depends on the details of subsurface mixing on vertical scales of order 10 m and over monthly, weekly, or even finer time scales in winter. Thus, high-time-resolution repeat measurements of both temperature and salinity in remote regions in wintertime are needed. Not surprisingly, such measurements have been rare and usually limited to ocean weather stations in the Northern Hemisphere (e.g., Fofonoff and Tabata 1966; Lazier 1980). The geographical extent of newly formed anomalies cannot be determined without high spatial resolution sampling of the upper ocean, and tracking their subsequent evolution requires multiyear datasets. The present observational database is sufficient to corroborate some, though not all, aspects of the model results presented in YL04.

After describing the sources of data in section 2, we present a conceptual model of the spice-injection mechanism that demonstrates how convective boundary layer mixing can result in large seasonal spice increase on subsurface isopycnals (section 3). The model guides the search for observational evidence of spice injection, which begins in section 4 with an examination of winter mixing recorded by a single Argo float in the South Atlantic. The global Argo database is then analyzed to demonstrate the existence of large-scale convective spice-formation zones in the subtropics of each ocean basin (section 5). Before concluding (section 7), we draw a link between observed decadal variability on isopycnals at a repeat hydrostation in the North Pacific to spice injection occurring upstream in the northeast Pacific spice-formation zone (section 6). At present, the spatial and temporal resolution of ocean observations is not sufficient to demonstrate and clarify all aspects of the proposed mechanism of spice injection, and so results from two numerical models are included. The large-scale behavior of an OGCM hindcast simulation is compared to Argo observations in section 5, and the simulation is then used to support the arguments of section 6. Results from a one-dimensional (1D) model of vertical mixing are presented in the appendix as additional validation of the hypothesis of section 3.
2. Data and model

This study examines data from the global Argo array of profiling floats, which provides multiyear time series of upper-ocean variables, including both temperature and salinity, in all regions of the World Ocean (Roemmich et al. 2004). There are several reasons why this particular dataset is well suited for investigating the proposed mechanism of spice generation. The high-frequency temporal sampling every 10 days is adequate to detect abrupt changes in temperature, salinity, and density near the base of the mixed layer, which take place over the course of the winter months. The global coverage and spatial sampling density, while as yet deficient in some regions, are unprecedented and presently adequate to resolve the geographic foci of spice formation, especially key regions such as the southeast Pacific (SEP) and South Atlantic (SA). The dataset does not suffer from hemispheric or seasonal biases in sampling, and the vertical resolution is high at roughly 10 m. The quality of the data is considered better than that obtained from XBTs (Roemmich et al. 2004), and the fidelity of the raw Argo profile data remains uncompromised by postprocessing and remapping, which can result in the creation of spurious water masses on isopycnals (Lozier et al. 1994).

In section 5, a global analysis makes use of all Argo float measurements taken between January 1999 and September 2006. Both real-time and delayed-mode Argo data are included. Despite the potential for erroneous drift in salinity measurements (Wong et al. 2003), the large number of profilers included in the section 5 analysis (≥3500) meant that only cursory quality control could be performed, based entirely on the quality control (QC) flags that accompany the data. The drift issue is mitigated, however, by considering vertical differences only.

Another dataset examined is the Hawaii Ocean Time series (HOT) measurements from station ALOHA, located north of Oahu, Hawaii, at 22°45′N, 158°W. Near-monthly CTD samples of temperature and salinity to 1000 m and deep casts to 4750 m have been collected at this site between late 1988 and the present. The resulting time series shows large variations in salinity (±0.15) on upper-ocean density surfaces (Lucas 2001). The proximity of the ALOHA station to a potential spice-formation zone in the northeast Pacific makes it a particularly relevant dataset for the present study.

Observations are compared to results from an OGCM run in a fully prognostic mode while forced with historical (1958–2000) atmospheric state fields. The model is version 1.4 of the Parallel Ocean Program (POP; Smith et al. 1992), which was used as the ocean component of the Community Climate System Model, version 3 (CCSM3), global coupled climate model (Collins et al. 2006). An overview of the model physics including the parameterization of double diffusion can be found in Danabasoglu et al. (2006), and the bulk flux forcing methodology being used is described in Large et al. (1997). The primary forcing datasets are as follows: 6-hourly winds, near-surface air temperature, and humidity from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996), from 1958 to 2000; daily downwelling shortwave and longwave radiation from the International Satellite Cloud Climatology Project global radiative flux data product (ISCCP FD), spanning 1984–2000 (Zhang et al. 2004); monthly precipitation from a blend of the Xie and Arkin (1996) and Global Precipitation Climatology Project (GPCP; Huffman et al. 1997) products, from 1979 to 2000; and, daily observed sea ice fraction courtesy of the National Snow and Ice Data Center (NSIDC), from 1979 to 2000. These datasets have been modified to reduce known climatological biases, as detailed in Large and Yeager (2004). This POP model simulation is an updated version of the NCAR Community Ocean Model hindcast experiment analyzed in YL04 and described in detail in Doney et al. (2003, 2007). The OGCM incorporates the same K-profile parameterization (KPP) vertical mixing scheme (Large et al. 1994) used in a 1D mixing model whose behavior is discussed in the appendix. The OGCM implementation of KPP includes the effects of double diffusion, but the 1D model simulation does not.

3. Convective spice injection

We propose that positive late-winter spice injections in nature are the result of strong diapycnal mixing at the base of a penetrating convective boundary layer driven by surface cooling during fall and winter in regions where the vertical salinity gradient is destabilizing (∂S/∂z > 0). A simple seasonal mixing scenario can be sketched based on well-studied and understood boundary layer physics (e.g., Stull 1988), and constrained by appropriate assumptions, which demonstrates the key aspects of the proposed mechanism. This hypothesis will be supported by model simulation and tested against observations in later sections.

In a convective boundary layer, buoyancy loss from the surface penetrates further into stratification than demanded by conservation, so that there is a negative entrainment flux, which is a constant fraction of the (positive) surface buoyancy flux (Ball 1960). Observations (e.g., Caughey 1982) and large-eddy simulation results (Moeng and Wyngaard 1984) find this fraction...
to be roughly -0.2. Using this value for the entrainment of buoyancy (assumption 1), the evolution of the upper-ocean buoyancy profile $B(z)$ over a time interval $\Delta t$ of convection can be drawn to scale (Fig. 1a). The initial profile $B_i(z)$ is well mixed in buoyancy to depth $h_1$ and decreases linearly with depth below (depth $d = -z$ defined to be positive downward). If the surface buoyancy loss corresponds to an area $A_{e1}$, the penetration is such that the resultant buoyancy profile is $B_2(z)$, with $A_{e2} = A_1 = 0.2 A_{e1}$. We assume that the depth of the well-mixed layer doubles so that $h_2 = 2h_1$ (assumption 2). The entrainment depth, $h_e$, is shallower than $h_2$ and demarcates the regions of the boundary layer subject to buoyancy loss (above) and gain (below). The effects of mixing reach to depth $h_{e}$, below which the ocean is undisturbed and $\partial_z B$ remains constant at $N_1^2$. As a result of entrainment, there is a layer of enhanced stratification between $h_2$ and $h_e$, which we will refer to as the transition layer between the well-mixed layer and the interior. Isopycnals lighter than $\rho_o$ [corresponding to $B_2(0)$] outcrop, those between $\rho_o$ and $\rho_i$ (corresponding to the depth $h_i$) deepen, and those denser than $\rho_o$ remain at constant depth.

In general, buoyancy changes in space and time are due to both temperature and salinity contributions:

$$
\Delta B = \frac{-g \Delta \rho}{\rho_o} = g \alpha \Delta T - g \beta \Delta S, \tag{1}
$$

where $\rho_o = 1026 \text{ kg m}^{-3}$ is an ocean reference density, $g$ is gravitational acceleration, and $\alpha$ and $\beta$ are the coefficients of expansion for temperature and salinity, respectively. In regions characterized by large salinity inversion in the upper 200 m ($\Delta_{200} S = S_{200} - S_{0} = 1$), the stratifying effect of a positive $\Delta_{200} T$ will be significantly countered by the destabilizing salinity profile.

The degree of density compensation of vertical gradients of $T$ and $S$ is commonly quantified in terms of either the density ratio $R_p$ or the closely related Turner angle (Ruddick 1983):

$$
R_p = \frac{\alpha \partial_z T}{\beta \partial_z S} \quad \text{and} \quad \tag{2}
$$

$$
\text{Tu} = \tan^{-1} \left( \frac{\alpha \partial_z T + \beta \partial_z S}{\alpha \partial_z T - \beta \partial_z S} \right) \tag{3}
$$

$$
= \tan^{-1} \left( \frac{R_p + 1}{R_p - 1} \right) \tag{4}
$$

$$
= \tan^{-1} \left( 1 - 2 \rho_o \beta \frac{\partial S}{\partial \rho} \right), \tag{5}
$$

FIG. 1. Idealized profiles of upper-ocean (a) buoyancy and (b) temperature and salinity as a function of depth before (thick curves $B_1$, $T_1$, and $S_1$) and after (thin curves $B_2$, $T_2$, and $S_2$) a period $\Delta t$ of convective boundary layer mixing. Figure notations are as follows: $h_1$, initial depth of well-mixed layer; $h_2$, final depth of well-mixed layer; $h_e$, entrainment depth; $h_i$, boundary layer depth; $h_o$, maximum depth of penetrative buoyancy change; $\rho_o$, initial sea surface density; $\rho_e$, density at depth $h_e$; $A_{e1}$, buoyancy loss due to surface buoyancy flux; $A_{e2}$, buoyancy loss due to entrainment; $A_{\alpha}$, buoyancy gain due to entrainment; TL, transition layer; and SDC layer. The $x$ and $y$ axes are nondimensional buoyancy and depth, scaled by $N_1^2 h_1$ and $h_1$, respectively. The profile curves for buoyancy, temperature, and salinity are thus given by $[(B - B_i(0))]/(N_1^2 h_1)$, $[\alpha (T - T_i(0))]/(N_1^2 h_1)$, and $[\beta (S - S_i(0))]/(N_1^2 h_1)$.\[2898]
where (5) makes use of (1). The Turner angle is between ±45° when a water column is stabilized by both ∂zT > 0 and ∂zS < 0. A destabilizing salinity gradient in the presence of a stabilizing temperature gradient corresponds to Tu > 45°. As Tu → 90°, the buoyancy effects of ∂zT > 0 and ∂zS > 0 are opposite in sign and approach perfect density compensation (also indicated by Rn → 1).

The change in buoyancy from $B_1$ to $B_2$ (Fig. 1a) will be accompanied by changes in the near-surface temperature and salinity profiles (Fig. 1b). The vertical gradients of $T_1$ and $S_1$ are chosen to correspond to a density ratio of 2 ($Tu \approx 71.6°$) by setting $\partial z T_1 = 2N_1$ and $\beta \partial z S_1 = N_1^2$ (assumption 3). The vertical gradient of $B_1$ is recovered from $T_1$ and $S_1$ through (1): $N_1^2 = \partial z T_1 - \beta \partial z S_1$. In addition to those already mentioned, the following assumptions were needed to sketch the evolution of $B$, $T$, and $S$ depicted in Fig. 1: 4) the buoyancy loss $A_n$ is entirely due to surface heat loss; the winter flux of freshwater at the surface is negligible, 5) the transition layer has thickness $h_1 - h_2 = 0.3h_1$, and 6) $h_2$ and $h_1$ correspond to the depth of the well-mixed layer and the depth of deepest penetration of mixing, respectively, for both $T$ and $S$ as well as for $B$.

Assumptions 2 and 5 are based on a comparison of fall and late-winter profiles from an Argo float (WMO 5900117) in the South Atlantic, implying that the ΔT for this schematic is on the order of months. Assumption 4 is appropriate for low-precipitation regions such as the subtropics, where roughly 85% of the buoyancy loss due to evaporation is associated with cooling by the latent heat flux and only 15% to the increase in salinity. The final assumption is made for simplicity, but it is perhaps not necessary that the depths $h_2$ and $h_1$ defined from the buoyancy profile, also correspond to the new profiles of salinity and temperature. For example, it is possible that the depth of well-mixed $T$ and $S$ is shallower than the depth of well-mixed buoyancy ($h_2$), as has been observed in some regions (Kara et al. 2003; de Boyer Montégut et al. 2004), or that changes in $T$ and $S$ occur at depths slightly below $h_1$, if perfect density compensation is achieved.

Since the surface buoyancy loss and resulting convection is driven by winter latent cooling, $A_n$ is equivalent to the time-integrated surface heat loss (Fig. 1b). A consequence of assumption 4 is that the change from $S_1$ to $S_2$ is entirely due to entrainment mixing down to depth $h_2$, with no change in depth-integrated salinity. In order for the $T_2$ and $S_2$ well-mixed layers to reach $h_2$ (assumption 6), the entrainment of both $T$ and $S$ must be significantly greater than the entrainment of buoyancy that is set by assumption 1. The buoyancy-driven entrainment of $T$ and $S$ enhances the thermocline and halocline in the transition layer between $h_2$ and $h_1$. Within this layer, the relative buoyancy contribution of the destabilizing halocline is augmented since the flux of heat at the surface dominates the flux of freshwater (assumption 4). Thus, the Turner angle within the transition layer increases. The postmixing profiles are characterized by $Tu \approx 75.6°$ between $h_2$ and $h_1$, up from the initial value of 71.6°. (Note that in Fig. 1b, a Turner angle of 90° corresponds to $T$ and $S$ profiles with identical slope.) Convection generates an SDC layer just below the mixed layer. This zone of enhanced density compensation coincides with the transition layer in Fig. 1 but could extend deeper if assumption 6 were relaxed. The weakly density-compensated (WDC) temperature and salinity below the SDC layer represents a background state set by assumption 3.

Spice injection is synonymous with the creation of the SDC layer at the base of the mixed layer. The spice increase on density surfaces that have not outcropped is apparent when the evolution from $S_1(z)$ to $S_2(z)$ is transformed to density coordinates (Fig. 2). Salinity (and hence temperature) increases on a range of transition layer isopycnals between $p_b$ and $p_c$, and the salinity change is a decreasing function of density. At some point in late winter, the profile $B_2(z)$ represents a minimum annual surface buoyancy (maximum density), so that $h_2$ and $p_b$ correspond to the maximum winter mixed layer depth and outcrop density, respectively. Isopycnals denser than $p_b$ (deeper than $h_2$) at this time will remain below the surface at this location, but can have their tracer characteristics significantly altered as a result of the local convective mixing. The penetration into stratification of spice change can be quantified by the density range over which injection occurs:

$$p_{inj} = p_c - p_b,$$

where $p_b$ is now taken to represent the late-winter maximum sea surface density (SSD$_{max}$) and $p_c$ is the maximum density surface on which there is a nonzero salinity change.

The transformation from $S_1$ to $S_2$ increases the negative slope of the water column in $S$–$\rho$ coordinates (Fig. 2), both in bulk measure between the surface and the upper pycnocline and locally within the SDC layer. A steepening of the slope of the $S(\rho)$ profile is a reflection of enhanced density compensation; higher Turner

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1 It is a straightforward exercise to show that the ratio $\Delta T/\Delta S$ associated with a given latent heat flux is $-\Lambda(C_p S_e) \sim -18°\mathrm{C}$, where $\Lambda$ is the latent heat of vaporization, $C_p$ is the ocean heat capacity, and $S_e$ is a reference ocean salinity (taken to be 35). The approximate 85% thermal contribution to $\Delta B$ follows directly from (1).
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(0)]}/(\rho N h_1). From (5), the Turner angle as a function of profile slope reduces to Tu = \tan^{-1}[1 - 2(\delta^* / \rho^*)]. Spice injection occurs on isopycnals between \(h_b\) and \(h_s\). The small boxes denote the well-mixed layer.

angles are generated as \(\partial S / \partial \rho\) becomes more negative in (5). Although the SDC layer is a relatively thin layer at the base of the mixed layer, its signature should be evident in a bulk measure of the degree of density compensation of the upper-ocean water column:

\[
Tu_b = \tan^{-1} \left( \frac{\alpha \Delta_{200}* T + \beta \Delta_{200}* S}{\alpha \Delta_{200}* T - \beta \Delta_{200}* S} \right). 
\]  

The bulk vertical Turner angle \(Tu_b\) is a measure of the relative contribution of the vertical gradients of \(T\) and \(S\) to the density stratification between the surface and the upper pycnocline (here taken to be 200 m). Since the \(\Delta_{200}* T\) and \(\Delta_{200}* S\) in late winter are in essence measures of gradients that exist across tens of meters at the base of the mixed layer (Fig. 1), \(Tu_b\) is a clear and convenient indicator of the winter generation of the SDC layer and of spice injection on subducted isopycnals.

The evolution depicted in Figs. 1 and 2 follows directly from the penetrative convective mixing of buoyancy and the assumptions made about surface forcing and water column characteristics. Different scenarios will change the details of spice injection, but not the essential conclusion: penetrative convective mixing of a partially density compensated water column, driven by surface heat loss, will result in enhanced density compensation in a transition layer just below the mixed layer, generating spice increase on subsurface isopycnals.

Boundary layer mixing is the key generative mechanism and must include a significant diapycnal component in order to generate the transformation from \(S_1\) to \(S_2\) shown in Fig. 2. Large diffusivities and nonlocal transport are expected (Large et al. 1994) above the boundary layer depth, \(h\), which is maximum somewhere between \(h_s\) and \(h_b\), but that periodically retreats to depths within the well-mixed layer (Fig. 1a). Below \(h\), only interior mixing processes are active, so that the vertical flux of temperature \(\bar{w} T\) and salinity \(\bar{w} S\) can be described respectively by downgradient diffusion as

\[
\bar{w} T = -\kappa_T \partial_s T \quad \text{and} \quad \bar{w} S = -\kappa_S \partial_s S, \tag{8}
\]

where the lowercase quantities under the overbar represent fluctuations from the time mean. From (1), it follows that buoyancy fluxes \(\bar{w} B\) below \(h\) are given by

\[
\bar{w} B = -g \alpha \kappa_T \partial_s T + g \beta \kappa_S \partial_s S. \tag{9}
\]

In situations where \(\partial_s T > 0\) and \(\partial_s S < 0\), both constituents are stabilizing, and their fluxes and changes below \(h\) are strongly constrained by buoyancy effects. However, if \(\partial_s T > 0\) and \(\partial_s S > 0\) (as in Fig. 1), large diapycnal diffusive fluxes of \(T\) and \(S\) can occur if \(\partial_s T\) and \(\partial_s S\) are strongly compensating and the concomitant buoyancy flux remains small.

A high degree of density compensation of \(\partial_s T\) and \(\partial_s S\), corresponding to Turner angles between 71.6° and 90° (\(2 > R_p > 1\)), is associated with enhanced diffusivity of both temperature and salt due to double-diffusive salt fingers (Schmitt 1981), with a somewhat lower temperature effect (\(\kappa_T \approx 0.7 \kappa_S\)). Observed estimates (St. Laurent and Schmitt 1999) range from a maximum \(\kappa_S = 0.5 \text{ cm}^2 \text{s}^{-1}\) at \(Tu \approx 81°\) (\(R_p \approx 1.38\)) to less than 0.1 \(\text{cm}^2 \text{s}^{-1}\) for \(Tu < 74°\) (\(R_p > 1.8\)). In the permanent pycnocline, nondouble diffusive \(\kappa\) is of the same order (0.1 → 0.2 \(\text{cm}^2 \text{s}^{-1}\); Ledwell et al. 1993). However, diffusivities in the boundary layer can be up to three orders of magnitude greater (Large et al. 1994). During episodes of active convection in late winter, when \(h > h_s\), the enhanced diffusion associated with large boundary layer \(\kappa\) and sharp gradients of \(T\) and \(S\) in the transition layer would likely greatly overshadow any salt fingering effects. Thus, if we focus on generative processes taking place within the transition layer, it is sufficient to assume

\[
\kappa_T = \kappa_S = \kappa. \tag{10}
\]
The necessary condition for unconstrained diapycnal diffusion in (9) becomes \( \beta_0 S = \alpha_0 T \), which corresponds to \( R_s \approx 1 \) and \( Tu \approx 90^\circ \).

Although we propose that spice injection is directly related to the generation of high Turner angles within a SDC layer, salt fingering is not invoked here as a primary mechanism. Each episode of active convection deepens the mixed layer, increases \( \partial_z T \) and \( \partial_z S \) within the transition layer, and produces strong diapycnal diffusion in the vicinity of \( h \), which is described by (9) and (10) with \( w_b \approx 0 \). The spice injection is the cumulative effect of this cross-isopycnal flux, the net rotation of the \( S(\rho) \) profile. Double diffusion is implicated in the subsequent weakening of \( \partial_z T \), \( \partial_z S \), and \( \partial S/\partial \rho \) as injected spice anomalies are mixed deeper into the water column (in the WDC layer) on longer time scales.

Not all aspects of the spice-injection hypothesis can be verified with observations at the present time, but several key features of the mechanism described above will be apparent in ocean measurements. A 1D model of vertical mixing can provide further useful insight into and preliminary confirmation of the mechanism we have outlined. The appendix includes some results of a 1D simulation in order to 1) demonstrate that vertical boundary layer mixing is the key generative process, not double diffusion; 2) quantify the relevant fluxes; and 3) examine the details of water column evolution at higher vertical and temporal resolution.

4. Convective spice injection in the South Atlantic

The SA was identified in YL04 as a region where significant pycnocline variations in \( T-S \) on isopycnals are likely to originate because upper-ocean salinity inversions are large, while winter stratification is relatively weak (thus indicating large winter \( Tu_0 \)). Several Argo floats in the central South Atlantic can be found that record one or more full annual cycles of upper-ocean temperature and salinity in this location, so that the effects of convective mixing can be studied. We focus initially on data from a single float (WMO 5900117) that was deployed in September 2001 near the Greenwich meridian at 20°S. It then drifted westward (at a drift depth of \(~1000\) m) by about 22° and equatorward by nearly 4°, with data transmission ending in June of 2004 (Fig. 3). Contours of climatological salinity on \( \alpha_0 = 25.2 \) kg m\(^{-3}\) computed from the World Ocean Atlas 1998 (WOA98) climatology (Levitus et al. 1998) show that the trajectory of the float is upgradient through the well-known tongue of high salinity in the South Atlantic pycnocline (Zhang et al. 2003). The float drift is considerable and so the full data time series should not be regarded as fixed point measurements.

Nevertheless, if we assume that the seasonal variations associated with vertical mixing processes are much larger than changes due to lateral processes and drift, then the float data can still be useful for assessing the spice-altering effects of winter convection.

The float samples a region in which surface salinity exceeds that at 200 m by roughly 1, and this inversion grows larger as the float moves westward (Fig. 4). Winter cooling and convective mixing generate key features associated with seasonal spice injection that are consistent with those suggested by the conceptual model (section 3) and seen in the vertical mixing model (see the appendix). The stabilizing thermocline and destabilizing halocline become sharply focused at the base of the mixed layer during winter months. A salinity change of \(~0.8\) is measured across only tens of meters in late August 2002. The deep penetrative mixing that follows the erosion of the summer pycnocline creates a transition layer below the well-mixed layer in which the strong gradients of \( T \) and \( S \) become much larger than the corresponding density gradient at those depths. The density contour interval of 0.25 kg m\(^{-3}\) is approximately equal to the change in density expected from a 1°C temperature change at the mean temperature and salinity of the float data (0.27 kg m\(^{-3}\)). The deepening of isopycnals as winter mixing progresses is consistent with positive buoyancy change below the entrainment depth, and there is some indication of enhanced deepening just before isopycnal outcrop.

The winter generation of a SDC layer is evidenced by the existence of Turner angles higher than 77° below the mixed layer in both 2002 and 2003. There are brief episodes when Turner angle exceeds 85° at the base of the deepening mixed layer, which may be a signature of intense mixing events occurring in fall and winter. The upper pycnocline is characterized by a thick (50–100 m) and persistent background layer in which \( Tu > 71.6^\circ \) \( (R_s < 2) \), which appears to grow deeper and thicker as the float moves into the western South Atlantic. The sinuous time series of bulk vertical Turner angle reflects the seasonal regeneration of the SDC layer (Fig. 4d). It has been shown that at least one Argo float from the SEP region recorded winter \( Tu \) as high as 80° at the base of the mixed layer (Johnson 2006), and perhaps better examples of SDC layer renewal remain to be found in the Argo database.

Sharp increases in salinity are observed on a range of density surfaces in the SDC layer as the mixed layer deepens in 2002 and 2003 (Fig. 4e), and positive changes are seen even on isopycnals that remain subducted in 2003 at depths of over 100 m. Some effects of horizontal processes, including drift, are present. For example, between January 2002 and November of 2003,
the background climatological mean salinity on $\sigma_0 = 25.2$ increases by 0.6 (Fig. 3). Nevertheless, there are compelling reasons to believe that the sharp increases are primarily related to vertical mixing. The geographic displacement of the float is rather small during the 2–3-month winter periods when salinity increase is maximum, and the rapid salinity increases clearly coincide with the deepening of the mixed layer and the rise in $T_u$. The increase can be seen, to a lesser extent and with time delay, on deeper isopycnals as winter mixing progresses, which is a classic signature of vertical mixing (Large et al. 1994), although advection of ventilated signals from higher latitudes would also appear as a time delay. Finally, these abrupt increases in isopycnal salinity are qualitatively very similar to winter spice-injection events simulated in an OGCM (YL04) and in a 1D KPP model (see the appendix), whose origins can be definitively linked to vertical mixing.

Winter injection of positive spice onto subsurface isopycnals is seen in the evolution of the observed upper-ocean salinity–density relationship in the winter of 2003. Individual column profiles recorded by WMO 5900117 between March and November of this year show large changes in the $S$–$\sigma_0$ curve above 200 m (Fig. 5), with smaller changes on denser surfaces as would be expected from 1D vertical mixing. Winter cooling results in an increase of the sea surface density by nearly 1 kg m$^{-3}$, reaching a maximum in September just less than $\sigma_0 = 25.4$ kg m$^{-3}$. The sea surface salinity (SSS) decreases only slightly between March and September as lower-salinity water mixes up from below. Diapycnal mixing below the well-mixed layer (represented in Fig. 5 as clustered points of nearly equal $\sigma_0$, $S$) has a freshening effect on some isopycnals in early winter, but by late winter (July onward) salinity change is positive on near-surface isopycnals. The increase in negative slope of the $S$–$\sigma_0$ curve as winter cooling increases SSD is generally consistent with Fig. 2. From 28 July 2003 to 16 September 2003, a positive increase in isopycnal salinity of $\sim 0.2$ is recorded on a range of density surfaces between the SSD$_{\text{max}}$ and $\sim 25.55$ kg m$^{-3}$, and then smaller increases occur on denser surfaces up to $\sim 25.8$ kg m$^{-3}$.
Fig. 4. Time series from WMO 5900117 of upper-ocean (a) salinity contoured at 0.2, (b) temperature contoured at 1°C, (c) Turner angle Tu with density overlaid (contoured at 0.25 kg m$^{-3}$), (d) bulk vertical Turner angle, Tu$_b$, and (e) isopycnal salinity on five $\sigma_\theta$ surfaces: 25.2, 25.4, 25.5, 25.6, and 25.7 kg m$^{-3}$. In (c), Tu values of 60°, 71.6°, 77°, and 85° correspond approximately to $R_\sigma = 3.7, 2, 1.6, and 1.2$, respectively, and the cross hatch indicates that vertical gradients are too small to compute Tu. The density contours in (c) are chosen so that they would roughly coincide with temperature contours if salinity were constant ($dp \approx 1000\sigma_\theta dT$, $\alpha = 2.6 \times 10^{-4}$°C$^{-1}$).
The $\rho_{int}$ for this 2-month interval, over which the approximation of purely vertical processes is perhaps quite valid, is roughly 0.4 kg m$^{-3}$. Later profiles from October and November show evidence of spring re-stratification and slight reduction in near-surface Tu, which reaches a maximum in September. Salinity on the surface $\sigma_0 = 25.2$ kg m$^{-3}$ (which does outcrop) is almost 0.5 higher in November than in the previous March, and drift toward higher climatological salinity on the isopycnal would explain less than half of this increase (Fig. 3).

The winter spice injections recorded by this Argo float in 2002 and 2003 are not necessarily the result of anomalous thermohaline forcing at the surface. Normal winter cooling and convective mixing is sufficient to generate the steeper $S(\rho)$ profile of Fig. 2. Float WMO 5900117 recorded injections of comparable magnitude in two consecutive winters in the South Atlantic. While it seems likely that this fundamentally seasonal injection process could be damped or enhanced by anomalous wintertime forcing, there is no way of knowing whether the spice anomalies of Figs. 4 and 5 represent particularly strong or weak winter injections without precise knowledge of the surface fluxes and background climatological state. An indication, perhaps, that the 2003 injection was weaker than normal is the fact that SDC layer Tu values in (5) in late winter do not exceed the deeper Tu values in the upper pycnocline that are constant in each of the measurements (Fig. 5). The conceptual model of section 3 assumed constant background stratification, but in Fig. 5 there is a large seasonal pycnocline that impedes convective mixing so that the injection of spice does not appear to generate a significant anomaly relative to the interior background.

5. Observational evidence of spice-formation zones

The global Argo database includes many floats that show evidence of winter spice injections associated with vertical mixing in key formation regions, similar to the 2002–03 increases recorded by float WMO 5900117 in the South Atlantic. High interannual spice variability related to the injection mechanism is likely to be found.
wherever large upper-ocean salinity inversions exist together with weak winter stratification (YL04, their Fig. 15). These conditions imply the existence of an SDC layer and correspond to regions of high $T_u$, which becomes even more elevated in wintertime. Individual floats can indeed be found in each of the regions of interest [northeast Pacific (NEP), SEP, northeast Atlantic (NEA), SA, South Indian Ocean (SI), and Tasman Sea], which record large winter $T-S$ increases on subducted isopycnals. Example injection events will be presented below, but it is fitting to demonstrate the existence of large-scale spice-formation regions using $T_u$. The generation of positive spice through seasonal diapycnal injection should correspond to a significant increase in bulk upper-ocean density compensation between summer and winter.

The geographic distribution of modeled bulk Turner angle from the boreal winter and summer seasons highlights regions where an OGCM hindcast suggests spice injection is likely to occur (Fig. 6). Three-month averages are shown from a monthly climatology that is computed from years 1996 to 2000 of the POP ocean hindcast simulation. Regions where $T_u > 45^\circ$, indicating the presence of a destabilizing $\Delta_{200}S$, are prevalent throughout the subtropics and Tropics. Large values of $T_u (>60^\circ)$, however, exist only between approximately 10°–40°S and 20°–35°N, where stratification is relatively weak compared to the Tropics because of strong winter cooling. A large seasonal cycle is evident in all the regions identified as spice-formation zones in YL04, with winter mixing generating $T_u > 70^\circ$ in each, and there are particularly notable winter signals in the Southern Hemisphere regions: SEP, SA, and SI.

A seasonal composite of roughly 7 yr of Argo float data paints a very similar picture (Fig. 7). No smoothing or time averaging has been applied, and thus each dot represents a single, instantaneous water column measurement taken during one of the months indicated. A consequence of the lack of spatiotemporal averaging is that nearby data points are plotted on top of each other, obscuring some noteworthy measurements recorded by floats that do not drift much (e.g., given 10-day sampling, a single float can contribute up to 10 data points yr$^{-1}$ to each of the panels in Fig. 7).

The Argo data confirm that each of the formation regions identified above is in fact characterized by a significant increase in $T_u$ during winter months. Strong seasonality is seen poleward of 10°S and 20°N, and the seasonal range compares well with the modeled range throughout the globe, apart from western boundary current extension regions that are poorly represented in this class of OGCM (Large and Danabasoglu 2006). Winter bulk Turner angles $>65^\circ$ are observed over large subtropical areas in the SEP, SA, SI, NEA, and NEP. Low Argo float density in the SEP and SA regions occurs in precisely the locations believed to be the most conducive to spice injection and where the highest $T_u$ values are expected. Surrounding data points do strongly suggest that the high tendency for spice injection in the modeled SEP and SA is a realistic result. The data coverage is likely to improve with time, and it should be possible in future years to demonstrate with ever greater confidence that these regions are the two most active zones of spice generation via the injection mechanism.

We expect the high $T_u$ seasonality in the SEP and SA to be a consequence of penetrative convection and to correspond to large, positive winter spice intake on subducted isopycnals, as sketched in Figs. 1 and 2. For comparison with Figs. 6 and 7, the $T_u$ values before and after the convection described in section 3 are 71.6° and $\approx 74^\circ$, respectively. To confirm that spice injection is occurring, we will take a closer look at the late-winter evolution of water columns in these two regions by contrasting pairs of measurements from individual floats (Fig. 8). A postconvection measurement is defined to correspond to the highest winter sea surface density (between August and November), while a preconvection measurement is taken approximately 2 months prior by the same float. In the Southern Hemisphere subtropics, $SSD_{max}$ occurs predominately in either August or September, and so the changes plotted in Fig. 8 are mostly August–June or September–July differences. Unlike in Fig. 7, not all available Argo data have been included. Only measurements that satisfy the following criteria are shown: 1) the float must have recorded two good profiles from a given year corresponding to early and late winter (as defined above), 2) $\Delta_{200}S$ must be greater than 0.2 in both profiles, and 3) there must be an increase in $T_u$ from early to late winter.

A useful diagnostic is the mean salinity injection over the course of 2 months, defined as

$$\overline{S_{inj}} = \frac{1}{p_{inj}} \int_{p_b}^{p_c} [S_2(p) - S_1(p)] dp,$$

where $dp = 0.01$ kg m$^{-3}$, and the notation follows that of Figs. 1 and 2. The integration is performed in density coordinates from the late-winter sea surface density ($p_b \approx SSD_{max}$) down in depth to the isopycnal $p_c$ where the salinity change between the late-winter profiles becomes negligible [$S_2(p) - S_1(p) < 0.05$]. The integral salinity injection is normalized by the density range between the surface and the depth of maximum penetrative salinity change in (6). In Fig. 2, $\overline{S_{inj}}$ corresponds to
Fig. 6. Seasonal mean $T_u_b$ (bulk 200-m vertical Turner angle) computed from the POP ocean model hindcast. The climatological (years 1996–2000) 3-month means for (top) Northern and (bottom) Southern Hemisphere winter.
Fig. 7. Instantaneous $T_u$ (bulk 200-m vertical Turner angle) measured by Argo profilers, plotted as a composite of all available float data collected between 1999 and 2006. (top) Measurements from January, February, and March; (bottom) measurements from July, August, and September.
For a subset of Argo floats available in the given geographic domain (a) month between August and November in which the maximum sea surface density for a given year was recorded, (b) mean salinity injection on subducted isopycnals between a measurement taken during the month of SSD$_{\text{max}}$ and one taken roughly 2 months prior by the same float, (c) density range $\rho_{\text{inj}}$ of the spice injection, and (d) maximum depth $h_i$ of the spice injection. Quantities in (b), (c), and (d) are computed from the difference between two upper-ocean profile measurements recorded by the same float, as described in section 5; the latter taken during the month of SSD$_{\text{max}}$ indicated in (a) and the earlier one taken roughly 2 months prior in the same year. Values are plotted at the float location of the later measurement. Points that are circled/numbered are examined in detail in Fig. 9.
the area between the curves \( S_3 \) and \( S_1 \), normalized by the density range \( \rho_{\text{inj}} \) over which the two profiles differ. The mixed layer \((d\rho = 0)\) contributes nothing to the mean salinity injection, so a nonzero value indicates spice change below the mixed layer arising from diapycnal mixing. The choice of a 2-month interval between \( S_1 \) and \( S_2 \) is a compromise between maximizing signal strength and minimizing the possible effects of float drift. To prevent inclusion of differences that extend far below the mixed layer, which are likely attributable to nonvertical processes (including float drift) or measurement error, a further restriction on the points included in Fig. 8 is that \( h_i \) (corresponding to \( \rho_i \)) be less than 250 m.

The bulk Turner angle is highly informative and easily computed, but it does not take into account the details of the subsurface evolution of the water column from summer to winter. A positive value of \( \bar{S}_{\text{inj}} \), however, is an unambiguous indication of spice injection on a range \( \rho_{\text{inj}} \) of density surfaces that are not directly exposed to air-sea exchange. As defined in section 3, \( \rho_{\text{inj}} \) only measures the density penetration of spice at the time of maximum SSD; diapycnal mixing continues even after spring restratification has begun (Figs. 4 and 5 and Johnson 2006), and so \( \rho_{\text{inj}} \) is more a measure of the direct effects of penetrative convection than of subsequent mixing. Observed values of \( \rho_{\text{inj}} \) in the SEP and SA are routinely \(-0.25\) kg m\(^{-3}\) and quite a few examples greater than the \(-0.4\) kg m\(^{-3}\) recorded by WMO 5900117 (labeled 4) are found in both basins (Fig. 8c). Comparing these values to a plot of climatological winter SSD (e.g., YL04, their Fig. 15) gives an indication of the geographic mismatch between anomaly origination and isopycnal outcrop location that might occur as a result of winter spice injection.

The water column evolution recorded by WMO 5900117 in 2003 (Fig. 5) gives an \( \bar{S}_{\text{inj}} \) value of 0.2 from (11) between July and September measurements. The associated values of \( \rho_{\text{inj}} \) and \( h_i \) are 0.42 kg m\(^{-3}\) and 156 m, respectively. These numbers indicate that between July and September 2003, diapycnal fluxes increased the salinity on subducted isopycnals that were up to 0.42 kg m\(^{-3}\) denser than the maximum SSD of that year, with the isopycnal salinity increase occurring to depths of up to 156 m and an average increase of 0.2 on isopycnals in the SDC layer. The correspondence between positive values of \( \bar{S}_{\text{inj}} \) and higher late-winter \( T_u \) in upper-ocean \( S-\sigma_t \) curves (closer to vertical) is demonstrated for several example pairs of Argo float measurements (Fig. 9). The six \( \bar{S}_{\text{inj}} \) values labeled in Fig. 8 give the mean salinity increase between the two winter profiles plotted in Fig. 9, while \( \rho_{\text{inj}} \) indicates the abscissa range over which they both exist and differ.

The change in the \( S-\sigma_t \) relation of the late-winter water columns recorded in the SEP and SA (Fig. 9) show qualitative resemblance to the transformation expected from penetrative convection mixing (Fig. 2). As already noted (section 4), the assumption of constant background stratification in the conceptual model of spice injection is inappropriate if the profile \( B_1 \) is (still) characterized by a strong seasonal pycnocline. However, this approximation appears quite good for the profiles of examples 3, 5, and 6, which show the closest resemblance to the curves of Fig. 2. In all the examples, a clear steepening of the \( S-\sigma_t \) curve is evident. The conceptual model is validated as a useful tool, but it cannot explain the detailed structures seen in the observed transformations, which are not purely vertical.

Example 1 is associated with an increase in salinity on isopycnals down to nearly 200 m over a density range of over 0.6 kg m\(^{-3}\), making it one of the largest spice-injection events included in Fig. 8 (\( \bar{S}_{\text{inj}} > 0.3 \)). Examples 2 and 3 show mean spice injections of similar magnitude despite changes in SSS that are opposite in sign. While profile drift may be partly implicated, these observations appear to demonstrate that positive spice injection can occur regardless of the sign of winter surface freshwater flux. The float WMO 5900117 (4 in Figs. 8 and 9) is included to show that the 2003 injection event analyzed in section 4 was not an extraordinary event for the South Atlantic; many other winter injections of similar magnitude can be found in the region. Some example injections (1, 2, 3, 5) show evidence of late-winter near-surface \( T_u \) greater than deeper pycnocline values, perhaps indicating that anomalously strong salinity injection and associated regeneration of the SDC layer took place that winter. The final two examples (5 and 6) are from the SEP region from floats with particularly low drift displacements, which were analyzed in depth by Johnson (2006). They both exhibit exceptionally deep penetration of spice (large \( h_i \)) and also show opposite changes in SSS.

We conclude that large spice injections on subducted isopycnals, occurring over short time scales at the apex of winter mixing, have been recorded by Argo floats throughout the SEP and SA regions. Mean isopycnal salinity increases and density penetrations greater than 0.1 and 0.25 kg m\(^{-3}\), respectively, are commonplace. Water column evolution similar to that measured by WMO 5900117 (Figs. 4 and 5) appears to be pervasive in the South Atlantic and southeast Pacific, and by extension, in all the spice-formation regions in Fig. 7 that exhibit a large mean and seasonal amplitude of \( T_u \).
6. Decadal variability related to winter spice injection

Long-term ocean monitoring at repeat hydrostations has revealed large variations in salinity on pycnocline density surfaces in the subtropical Atlantic (Jenkins 1982) and in the North Pacific near Hawaii (Lukas 2001). The proximity of the HOT station location to the NEP spice-formation zone makes this a particularly relevant observational benchmark for this study. Pycnocline salinity evolution on density surfaces from the HOT station ALOHA (Fig. 10a) is marked by a freshening during the mid-1990s on surfaces lighter than $\sigma_0 = 25$ kg m$^{-3}$ followed by several years of higher than average salinity in the early part of this century. This plot replicates Fig. 2c from Lukas (2001), but uses a longer time series. Observed decadal variation in the density range ($\sigma_0 = 25–26$ kg m$^{-3}$) is lower in amplitude and opposite in sign compared to the variability seen on lighter surfaces. Isopycnal salinity anomalies from the POP hindcast simulation are plotted over equivalent years (Fig. 10b), after averaging over the four model grid cells nearest the ALOHA station location ($\sim 1^\circ \times 0.5^\circ$ average centered at 22.5$^\circ$N, 158$^\circ$W).

The POP model hindcast does not reproduce the HOT signals, but an exact comparison of tracer evolution on density surfaces at a particular location is not necessarily expected. The model is a coarse-resolution
OGCM with inherent biases, run without data assimilation, and forced by imperfect atmospheric fields, including a precipitation field that is highly uncertain (Large and Yeager 2004; Béranger et al. 2006). Consequently, there are errors in air–sea fluxes, parameterized mixing, and large-scale flow and density structure that lead to inaccuracy in the hindcast on small scales. The simulated sea surface is denser than observed at this location, and so outcropping is apparent in the model time series for the given density range. The model mean salinity as a function of density is higher at this location, by about 0.25 between $\sigma_0 = 24$ and 25 kg m$^{-3}$ and by about 0.1 between $\sigma_0 = 26$ and 27 kg m$^{-3}$ (not shown). Nevertheless, the model simulates vari-

**Fig. 10.** Time series of salinity anomaly on a range of isopycnal surfaces (top) from the HOT Aloha station and (bottom) from the POP model hindcast simulation.
not locally forced at POP–ALOHA, but instead advect to this location from an origin closer to the NEP spice-formation zone. The source region centered at ~26°N, 151°W (POP–source) was identified from monthly animations of salinity anomaly on $\sigma_0 = 25$ kg m$^{-3}$, which show that large spice anomalies are often generated in winter to the northeast of POP–ALOHA, close to the winter outcrop of the isopycnal, and subsequently propagate southwestward as they diminish. Of course, tracing the POP–ALOHA anomalies back to a single stationary source is a gross simplification; the outcrop location changes from year to year and the velocity field is not constant. The POP–source time series (Fig. 11a) must be interpreted as one example of the variations on $\sigma_0 = 25$ kg m$^{-3}$ upstream of the observation location, closer to its winter outcrop, which combine in a complex way to generate the signal seen at POP–ALOHA.

The propagation of a decade-long spice fluctuation is evident with a phase shift of a year or less from the source region to POP–ALOHA. Positive anomalies at the source in the early 1990s are followed by negative anomalies in the late nineties, and the decadal variation at POP–ALOHA appears to follow, with some diffusion of the signal (Fig. 11a). The monthly SSD time series at the two locations (Fig. 11b) confirms that while $\sigma_0 = 25$ kg m$^{-3}$ remains shielded from the surface at POP–ALOHA by more than 0.5 kg m$^{-3}$ throughout the time series, this isopycnal comes close to outcropping in winter at the source region location, especially in the early 1990s.

The POP–source isopycnal salinity time series shows more power at higher frequencies, and in particular, large, abrupt increases are evident in the winters of 1991 and 1994, when $\sigma_0 = 25$ kg m$^{-3}$ comes close to outcropping. The positive salinity pulses on the isopycnal do not correspond to anomalously high SSS at the source (Fig. 11c), which exhibits a steady mean and seasonal cycle from 1988 to 1994. Thus, the abrupt spice increases in the early 1990s seen on $\sigma_0 = 25$ kg m$^{-3}$ cannot be attributed to subduction of anomalous surface properties. Instead, these sharp increases are diapycnal injection events that occur when $\sigma_0 = 25$ kg m$^{-3}$ comes anomalously close to the base of the winter mixed layer in 1991 and 1994, so that salinity on the isopycnal becomes nearly equal to late-winter SSS (Fig. 11c). The source region upper ocean is characterized by an ability on isopycnals at this location of comparable magnitude and time scale to that seen in observations. For example, the HOT time series shows a salinity anomaly increase on $\sigma_0 = 24.2$ kg m$^{-3}$ of ~0.3 between 1997 and 2001, while the model simulates a decrease on $\sigma_0 = 25$ kg m$^{-3}$ of ~0.4 between 1995 and 1999. The similarities in the period and amplitude of signals suggests that the model may incorporate the correct generation mechanism, and may be a useful tool for interpreting the spice signals seen in the HOT time series.

The POP anomaly time series on $\sigma_0 = 25$ kg m$^{-3}$ at the point representing the model ALOHA location (POP–ALOHA) is dominated by a near-decadal oscillation of amplitude ~0.2 (Fig. 11a), and a similar signal is seen on the same isopycnal at a source location farther upstream. The spice anomalies on this surface are
ALOHA location in the mid-1990s (Fig. 10b) can be explained by advection of large positive anomalies originating within the NEP spice-formation zone in the early 1990s, when winter $T_u$$_b$ was anomalously high and pulselike positive injections abruptly increased the spice on isopycnals near $\sigma_u$ = 25 kg m$^{-3}$. The subsequent negative anomaly in the late 1990s (Fig. 10b) is consistent with reduced injection activity and reduced winter $T_u$$_b$.

7. Discussion and summary

The enhancement of upper-ocean vertical density compensation, which occurs when winter convective mixing focuses a stabilizing thermocline and a destabilizing halocline at the base of the mixed layer, is a seasonal phenomenon seen in subtropical locations throughout the globe, and to an extreme degree in select spice-formation regions (Fig. 7). The signature of such mixing is the creation of a SDC layer below the winter mixed layer. The formation regions where this occurs are identical to those identified by Kara et al. (2003) as locations where a particularly large $\Delta T$ criterion is needed in order to ensure agreement between isothermal layer depth and mixed layer depth based on $\sigma_t$ (their Fig. 7). This noted discrepancy between isothermal and isopycnal layers reflects the high degree of vertical density compensation, and thus high mean bulk vertical Turner angles, in these locations.

We expect spice injection resulting from wintertime penetrative convection (section 3) to produce significant variability on isopycnals in these locations. Whenever $T_u$$_b$ is large and even more elevated in winter, there will be large vertical $T$–$S$ gradients compressed into alignment in a SDC layer in wintertime. Such conditions force a vigorous diapycnal diffusion across the base of the boundary layer. The winter spice increase on subducted isopycnals recorded by WMO 5900117 in the subtropical South Atlantic (Figs. 4 and 5) is merely one example of many that can be found in the Argo collection. A multiyear composite of Argo float data from the SEP and SA regions shows ubiquitous positive spice injection occurring in late winter (Fig. 8) over time and space scales that must correspond to a wide range of anomalous winter forcing. Argo measurements provide compelling evidence that convective spice injection is a real, large-scale phenomenon in the subtropical ocean, and this confirmation of model findings is a testament to the increasing fidelity of present-generation OGCMs.

While the regions where subsurface spice injections can occur are limited geographically, model results suggest that this mechanism may account for a significant fraction of the interannual variance on deep isopycnals throughout the ocean (YL04). The background climatologies of salinity and density will in general be highly uncertain at a given float location, and so Argo is unable to confirm the finding from OGCMs that large interannual spice anomalies can persist and propagate across basin scales in the ocean interior, affecting decadal change in the deep ocean (Schneider 2000; Pierce et al. 2000; YL04). Such an approach may be possible at locations where repeat hydrography provides reliable information on the background state (King and McDonagh 2005). Another alternative may be to exploit the hypothesized link between anomalous spice injection and anomalous winter $T_u$$_b$; observations of global salinity (and hence density) at the sea surface and in the upper pycnocline could then provide insight into interannual variations of ocean spice.

Multiyear, fixed-location hydrostation observations, such as the HOT ALOHA data, show clear evidence of interannual variations on density surfaces, but the origins of such changes are difficult to pinpoint (Lukas 2001). Based on the similarity of model pycnocline variability to observed (Fig. 10), we speculate that the observed signals are related to winter spice-injection events that took place to the northeast of ALOHA, closer to where Argo floats indicate maximum winter $T_u$$_b$ in the NEP region (Fig. 7). Pycnocline spice fluctuations are likely to be related to overall trends in surface thermohaline forcing in source regions, as suggested by Lukas, but only indirectly and insofar as such trends effect late-winter mixing at the base of the boundary layer. The considerable magnitude of isopycnal anomaly signals recorded at HOT is presumably related to the proximity of this ocean station to the NEP spice-formation zone. An appropriately located ocean observing station in the South Pacific would probably record even larger variations on isopycnals whose origins could be traced to the exceptionally high winter $T_u$$_b$ in the SEP region (Fig. 7), with subsequent advection of anomalies toward the western equatorial Pacific (Kessler 1999).

The explanation of the origins of ocean spice changes presented herein is consistent with, and could offer fresh insight into, a number of observations that have demonstrated decadal-scale regional water mass variations showing a significant isopycnal component. Jenkins (1982) hypothesized that the positive decadal salinity changes on isopycnals recorded in the Sargasso Sea were linked to anomalously strong latent cooling and low $d\rho/dz$ in mode water formation regions. This conceptual model was adopted by Pollard and Pu (1985) and Pollard et al. (1996) to explain salinity deviations from the mean $T$–$S$ relationship observed...
northeast of the Azores, which were highly correlated with isopycnal thickness.

Latent cooling is also a key element in the spice-injection model; it drives the winter transformations depicted in Figs. 1 and 2. The correlation of isopycnal salinity with oxygen noted by Jenkins would also be seen if the high surface values were diapycnally injected in winter. However, the spice-injection hypothesis differs insofar as it places an emphasis on the role of boundary layer mixing in generating extremely large and abrupt winter spice increase on near-surface isopycnals, whether or not the latent cooling is anomalous. In spice-formation regions, large salinity inversions amplify the imprint on isopycnals of any anomalous surface forcing. The SSS increase associated with a given amount of latent cooling is much smaller than the potential isopycnal salinity increase that can arise when the $S$–$T$ curve steepens as a result of convective entrainment. The evaporative increase of salinity was not even included in Figs. 1 and 2. The existence of a SDC layer with a large salinity gradient (~1) and associated high $T_u_b$ implies that relatively small perturbations in water column density can produce large changes in isopycnal spice. The magnitude of injected spice anomalies depends on the details of diapycnal mixing below the boundary layer, which is certainly a function of anomalous surface forcing, but large interannual variations in mixed layer salinity are not required to generate significant interannual spice anomalies, as would be required in the Jenkins (1982) model.

Spice injection would be implicated more on the eastern side of the Atlantic basin (Fig. 7) than at the Panulirus station in the Sargasso Sea, but isopycnal anomaly signals generated near the Azores would likely propagate to some extent into the far western Atlantic (Laurian et al. 2006). An analog to Jenkins’s Fig. 2, for the injection model of spice genesis, would look more like the $T$–$S$ curve evolution shown in YL04 (their Fig. 6) or Johnson (2006, his Fig. 3), with a deep compensated layer indicated by a $T$–$S$ curve that is nearly parallel to isopycnals, a large and destabilizing salinity gradient throughout the water column in both summer and winter, and a seasonal pycnocline that collapses in late winter. The before and after $T$–$S$ profiles for a spice-injection event would show a much larger isopycnal salinity change than the one shown in Jenkins’s schematic. It may be that spice injection is the unknown mechanism invoked by Jenkins (1982) when he summarizes that the “high salinities are somehow driven by greater than normal heat loss.”

In the South Atlantic, Gordon (1981) argued that the many “warm, salty intrusions” seen in hydrographic station $T$–$S$ relationships along 38°S were signatures of relatively salty winter mixed layers injected into the fresher main thermocline. His discussion of the origins of these (isothermal) anomalies, which emphasized the role of winter cooling and deep convective mixing in generating routine positive salinity anomalies together with low-density ratios in the upper thermocline in this region, is entirely consistent with and anticipates the spice-injection mechanism outlined here as a global phenomenon. The same arguments were used to explain the density-compensated thermohaline structure of eddies in the vicinity of the Brazil–Malvinas Confluence (Gordon 1989). Other suggestive examples of large, low-frequency changes on isopycnals observed in the vicinity of convective spice-formation zones (Fig. 7) include NEA (Bryden et al. 1996); SI (Bryden et al. 2003); NEP, SEP, and SI (Wong et al. 1999); and SEP (Kessler 1999).

In Johnson’s (2006) examination of two Argo floats in the SEP spice-formation zone, he shows that the observed growth of spice anomalies on subsurface isopycnals is related to a strong diapycnal flux near the base of the winter mixed layer that transfers the relative warm/salty surface properties downward through the water column, continuing even after restratification has begun at the surface. The efficacy of the cross-isopycnal mixing and resulting depth/density penetration of spice is enhanced by the double diffusion associated with the observed persistence of winter density ratios as low as 1.6 for several months. His findings are consistent with the spice-injection hypothesis. However, his assumption that observed spice anomalies are surface-forced, by anomalous winter latent heat fluxes, seems unwarranted and deemphasizes the seasonality of the phenomenon. The isopycnal signals he presents are merely winter anomalies from the $\sigma$–$S$ relation of the previous summer. Without a comparison to background climatology, there is no way of knowing whether these winter anomalies are anomalously weak or strong.

The primary conclusion is that Argo observations confirm that spice is injected on subducted isopycnals each winter over widespread spice-formation regions of the subtropics. The observed winter increase in bulk upper-ocean Turner angle is a signature of the creation of a strongly density-compensated layer below the well-mixed layer. Penetrative convective mixing of a partially density-compensated water column will generate a layer of large, compensated $T$–$S$ gradients as a result of normal winter latent cooling. This SDC layer is conducive to large diapycnal fluxes that effectively ventilate isopycnals that do not outcrop locally. Isopycnal properties that are set in this way by injection need not match surface properties at the density outcrop location. Observations from HOT support the conclusion
that large pycnocline variations of ocean spice in disparate ocean regions can likely trace their origins to these spice-formation regions. Conceptual models that assume adiabatic interior flow from a late-winter mixed layer of uniform density, temperature, and salinity are ill suited for the study of spice variability, which originates in regions where significant diapycnal processes occur at the base of the mixed layer.

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APPENDIX

Convective Spice Injection in a 1D Model

A 1D numerical model of vertical mixing provides insight into the expected evolution of upper-ocean $T-S$ during winter convection in the presence of a large, destabilizing salinity gradient (outlined schematically in section 3). The model is based on the well-established KPP ocean boundary layer physics. The full details of the parameterization and model are described in Large et al. (1994). The model is run at 2-m vertical resolution down to 250 m, with a time step of 1 h, and the effects of double diffusion have not been included in (9). The identical model solution was also examined in appendix B of YL04, their Fig. B1b, with the initial condition taken from early April output of their global ocean hindcast simulation from the southeast Pacific location 23°S, 102°W. The water column at the start of integration is characterized by a salinity inversion of $>1$ in the upper 200 m, which is comparable to observed austral summer profiles taken from the southeast Pacific region (Johnson 2006). The application of daily turbulent and solar surface fluxes generates time evolution of the modeled water column due to purely vertical processes.

A seasonal pycnocline associated with high-temperature surface water exists at the start of the model integration (Fig. A1a), but this disappears with the onset of winter latent cooling. Mixed layer salinity decreases by at most 0.3 in late winter, so the upper-200-m salinity gradient maintains most of its strength but becomes sharply focused at the base of the mixed layer by convective entrainment, as shown in Fig. 1.

The concurrent development of an enhanced thermocline at the base of the mixed layer, $\partial_z T$, which is highly density compensated by $\partial_z S$ is indicated by the generation of a layer characterized by extremely high Turner angle ($\approx 88^\circ$) below the mixed layer between June and September. This winter-generated strongly density-compensated (SDC) layer persists throughout the year as a layer of Tu $\approx 80^\circ$ in the upper pycnocline (Fig. A1b). The high mean Tu, value (Fig. A1c) indicates a significant degree of upper-ocean compensation throughout the year, and its seasonal peak is clearly correlated with the late-winter renewal of the SDC layer.

Spice injection occurs within the SDC layer in late winter. Abrupt increases in salinity (and hence temperature) are seen on four of five example density surfaces plotted in Fig. A1d, while the fifth and deepest surface shows a relatively slow and delayed increase that continues into austral spring. The large injections are associated with episodes of vigorous turbulent vertical flux of salinity and buoyancy within the boundary layer (not shown), during which there is temporary enhancement of $\partial_z S$ at the base of the boundary layer. The surface $\sigma_0 = 25.6$ kg m$^{-3}$ remains subducted despite an injection event in early September during which $\partial_z S$ on the isopycnal exceeds 0.05 day$^{-1}$. By early spring, the salinity on this surface shows an overall increase of $\approx 0.7$ from its initial value even though there is negligible change in the depth of the isopycnal over the 200-day simulation.

A close inspection of the column profile near the boundary layer depth at the model time step $t = 244.0$ clarifies the origins of the early September spice injection on this isopycnal (Fig. A2). The profiles of density and salinity at this instant show a well-mixed layer to $\approx 189$ m and a boundary layer depth, $h$, which is slightly deeper at 193.5 m. Large, coincident gradients of temperature ($\approx 2^\circ$C, not shown) and salinity ($\approx 0.9$) exist over roughly 10 m, straddling the boundary layer depth. The depth of $\sigma_0 = 25.6$ kg m$^{-3}$ is just below the boundary layer, at 195.4 m based on linear interpolation between model depth levels. Slightly enhanced stratification below the well-mixed layer is evident, but small compared to the enhanced thermo/halocline. The expected enhancement of stratification near the boundary layer depth (Fig. 1) is most obvious in Fig. A2b as a local minimum in $\partial_z \rho$. The vertical gradients of both density and salinity are largest just below the boundary layer depth (Fig. A2b), although the vertical resolution of the model may be too coarse to adequately resolve such features, even at 2 m. The depth of the maximum
Fig. A1. Time series of output from a 1D model of vertical mixing after initialization at $t = 100$ days (10 Apr). (a) Upper-ocean salinity with $\alpha_0$ contours overlaid at 0.1 kg m$^{-3}$ intervals (additional contours at 25.55, 25.58, and 25.62 kg m$^{-3}$); (b) Turner angle, $\theta_u$, with density contours overlaid corresponding to the isopycnals in (d); (c) bulk 200-m Turner angle, $\theta_{u_b}$; and (d) salinity on the isopycnal surfaces $\sigma_0 = 25.50$, 25.55, 25.58, 25.60, and 25.62 kg m$^{-3}$. Dashed curves in (d) show $\delta S$ on the given density surfaces computed from daily mean values and refer to the scale on the right.
Fig. A2. Snapshot of 1D vertical mixing model variables near the base of the mixed layer at time step $t = 244.0$ days (2 Sep), at which time $\delta S$ on $\sigma_0 = 25.6$ is near maximum (cf. Fig. A1). (a) Salinity and density differences from surface values (unitless and kg m$^{-3}$); (b) Turner angle (Tu) and vertical gradients of salinity and density (°, m m$^{-1}$, and 10$^{-3}$ kg m$^{-3}$); (c) parameterized turbulent fluxes of salinity ($wS$) and buoyancy ($wB$) (10$^{-5}$ m s$^{-1}$ and 10$^{-8}$ m$^2$ s$^{-1}$); and (d) time tendencies of salinity ($\delta S$) and buoyancy ($\delta B$) computed as the vertical divergences of the fluxes in (c) (day$^{-1}$ and 10$^{-8}$ m s$^{-2}$ day$^{-1}$). In both (c) and (d), the diamond curve indicates the salinity contribution to the net buoyancy curve and has units matching the respective buoyancy curve. In (a)–(d), the closest axis corresponds to filled circles, the next closest to triangles, and the farthest (if needed) to diamonds; the dotted horizontal line marks the boundary layer depth, $h = 193.5$ m, and the long-dashed horizontal line marks the depth of the isopycnal $\sigma_0 = 25.6$ kg m$^{-3}$ (195.4 m).
Turner angle (−89°) is also found to nearly coincide with the boundary layer depth, and Tu > 88° on the isopycnal σ₀ = 25.6 kg m⁻³.

The negative vertical fluxes of buoyancy and salinity associated with entrainment remain elevated in a thin layer below the boundary layer depth (Fig. A2c) and decrease rapidly to much lower background values. The buoyancy flux profile is characterized by an entrainment zone of negative flux, as expected from numerous observational and numerical studies of convective boundary layers (e.g., Caughey 1982; Moeng and Wyngaard 1984). The net buoyancy flux at the depth of the isopycnal σ₀ = 25.6 kg m⁻³ is smaller by more than a factor of 10 than and opposite in sign to the salinity contribution to the buoyancy flux (−gβws), indicating that ws and wT are both large but compensating.

The time tendency of a property X is related to the vertical flux through

\[ \partial_t X = -\partial_z (\mathbf{w} T). \] (A1)

The vertical divergences of the fluxes in Fig. A2c generate positive time derivatives of both buoyancy and salinity in a layer roughly 10 m thick below h (Fig. A2d). At the depth of the isopycnal, \( \partial_z S = 0.05 \text{ day}^{-1} \), which is consistent with the peak injection rate on this isopycnal seen in Fig. A1. The \( \partial_z B > 0 \) on the isopycnal would tend to deepen it as the injection occurs, but the net buoyancy tendency is again much less than and opposite in sign to the haline component of the buoyancy tendency. We conclude that diapycnal diffusion below the boundary layer, described by (9) with \( \mathbf{w} \beta = 0 \), is causing the rapid rise in salinity on this isopycnal seen in Fig. A1.

The spice increase seen on the example surface σ₀ = 25.6 kg m⁻³ in fact occurs over a range of densities below the boundary layer, as well as on those surfaces that enter the boundary layer and outcrop over the course of winter. Our focus here, though, is spice generation on isopycnals that remain subducted. The \( S - \rho \) signature of spice injection (Fig. 2) has previously been demonstrated for this model integration in YL04’s Fig. B2 (curve b), which compares late-September and mid-April profiles. The SSDmax of the model simulation is σ₀ = 25.59 kg m⁻³, and there is penetration of spice over a range of \( \rho \approx 0.05 \text{ kg m}^{-3} \) below this density, even without the enhanced mixing expected with double diffusion.

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