Barrier Layers in the Tropical South Atlantic: Mean Dynamics and Submesoscale Effects*

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ABSTRACT

Barrier layers are generated when the surface mixed layer is shallower than the layer where temperature is well mixed, in geographical regions where salinity plays a key role in setting up upper-ocean density stratification. In the tropical oceans, thick barrier layers are also found in a latitude range where spiraling trajectories from surface in situ drifters suggest the presence of predominantly cyclonic submesoscale-like vortices. The authors explore these dynamical processes and their interplay in the present paper, focusing on the tropical South Atlantic Ocean and using a high-resolution modeling approach. The objective is threefold: to investigate the mean dynamics contributing to barrier-layer formation in this region, to study the distribution and seasonality of submesoscale features, and to verify whether and how the submesoscale impacts barrier-layer thickness. The model used is the Regional Ocean Modeling System (ROMS) in its Adaptive Grid Refinement in Fortran (AGRIF) online-nested configuration with a horizontal resolution ranging between 9 and 1 km. The simulated circulation is first described in terms of mean and submesoscale dynamics, and the associated seasonal cycle. Mechanisms for barrier-layer formation are then investigated. The results confirm previous hypotheses by Mignot et al. on the relevance of enhanced winter mixing deepening the isothermal layer, whereas the salinity stratification is sustained by advection of surface fresh waters and subsurface salinity maxima. Finally, submesoscale effects on barrier-layer thickness are studied, quantifying their contribution to vertical fluxes of temperature and salinity. Submesoscale vortices associated with salinity fronts are found to have a significant effect, producing thicker barrier layers (by ~20%–35%) and a shallower mixed layer because of their restratifying effect on salinity.

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1. Introduction

The role of salinity in upper-ocean stratification and air–sea interaction processes has been increasingly recognized in recent years (Vialard and Delecluse 1998a; Sprintall and Roemmich 1999; Delcroix and McPhaden 2002; Kara et al. 2003; Tailleux et al. 2005; de Boyer Montégut et al. 2007; Mignot et al. 2007; Helber et al. 2012). Observational studies based either on climatological or Argo float profiles

(Sprintall and Tomczak 1992; de Boyer Montégut et al. 2004; Sato et al. 2006; de Boyer Montégut et al. 2007; Liu et al. 2009) show that the mixed layer depth (MLD) does not coincide with the isothermal layer depth (ILD) in many regions of the world's oceans. The most common occurrence is when the isothermal layer extends deeper than the mixed layer because of the presence of a strong salinity-stratified region below the MLD, which is called the barrier layer (BL).¹ Since its discovery in the western equatorial Pacific Ocean (Godfrey and Lindstrom 1989; Lukas and Lindstrom 1991), the barrier layer has been studied extensively owing to its ability to oppose heat and momentum flux exchanges below the mixed layer (Vialard and Delecluse 1998a,b; Foltz and McPhaden 2009). Indeed, the uniform temperature (or even slight temperature inversion) that is found within the BL produces a drastic reduction in entrainment cooling (or even entrainment heating, especially near the equator), which in turn causes an increase in sea surface temperature (SST). Such a mechanism is particularly important in tropical and equatorial regions where (i) ocean–atmosphere coupled processes are strongest, and (ii) horizontal advection of heat is reduced because of the generally uniform temperature distribution. The impact of barrier layers on SST has been found to favor the intensification of tropical cyclones (Wang et al. 2011; Balaguru et al. 2012a), to increase precipitation and therefore produce an early monsoon season in the Arabian Sea (Masson et al. 2005), and to even influence the onset of El Niño–Southern Oscillation events in the western equatorial Pacific (Maes et al. 2005).

Thick barrier layers (>10 m) are found in the Arctic and Southern Oceans and in most equatorial and tropical regions. Their formation mechanism can be quite complex and varied, because it entails various processes affecting the evolution of salinity and temperature stratification at the surface as well as at depth. Main mechanisms have been described in Cronin and McPhaden (2002) and are principally divided between advection/subduction processes and the impact of surface forcing/river runoff. In the equatorial Pacific, for instance, both observational and model studies have shown that the dominant process of BL formation is the westward advection of surface central Pacific salty water and subsequent subduction below the freshwater pool of the western Pacific (Lukas and Lindstrom 1991; Vialard and Delecluse 1998b). South of the equator, an additional mechanism is the thinning of the mixed layer due to surface freshening by strong precipitation events under

the intertropical convergence zone (ITCZ; Vialard and Delecluse 1998b; Bosc et al. 2009). The effect of rain on barrier layers can be subtle, however, because particularly strong wind bursts, which can be concurrent with strong precipitation events, can also erode the salinity stratification and produce thin BLs (Zhang and McPhaden 2000). In tropical regions (between 10° and 20° latitude), various studies suggest the importance of the salinity maximum water, also known as subtropical underwater (STUW; Lambert and Sturges 1977; Stramma and Schott 1999), in setting up the subsurface salinity stratification and therefore providing favorable conditions for BL formation (Sprintall and Tomczak 1992; Mignot et al. 2007; Araujo et al. 2011). STUWs are found in all oceanic basins and are due to the subduction of high-salinity waters in the subtropical regions (Zhang et al. 2003; O'Connor et al. 2005). Mignot et al. (2007) suggest that tropical barrier layers found on the equatorward side of STUW regions are formed through two concurrent mechanisms: poleward advection of surface freshwater via Ekman transport, which produces a sharp halocline above the depth of the STUW, and vertical mixing in winter, which deepens the isothermal layer but is not strong enough to homogenize the salinity stratification as well. Sato et al. (2006), on the other hand, hypothesize the importance of synoptic, small-scale salinity fronts at the surface, which would cause subduction of high-salinity water below the MLD and therefore formation of a barrier layer. The presence of small-scale subduction processes is in agreement with the findings of Griffa et al. (2008), which use in situ satellite-tracked drifters to estimate the distribution of surface mesoscale and submesoscale eddies over the global ocean. As illustrated in Fig. 1, the authors find a predominantly cyclonic (consistent with subduction) polarity band around the 10°–20° latitude range in all oceans, which coincides quite precisely with the tropical regions of BL formation described by Sato et al. (2006), among other authors. This cyclonic band is mostly populated by eddies with scales comparable with the submesoscale [radius smaller than 15 km; see lower panel of Fig. 3 in Griffa et al. (2008)]. The idea is therefore that these submesoscale eddies may be related to the presence of synoptic salinity fronts mentioned by Sato et al. (2006) and possibly play a role in BL formation. Tanguy et al. (2010) hypothesizes that both this mechanism and the one put forward by Mignot et al. (2007) may be important for BL dynamics in the tropical Atlantic.

In this paper, we wish to shed light on the processes described above by adopting a high-resolution numerical modeling approach and focusing on the region of the tropical South Atlantic. We have three overarching objectives: (i) to investigate the mean dynamical processes that can contribute to BL formation in this region,

¹ The other case, when the ILD is shallower than the MLD, is referred to as compensated layer (CL) because temperature and salinity stratification compensate to make density well mixed below the isothermal layer (e.g., Liu et al. 2009).

FIG. 1. Results from Griffa et al. (2008) (modification of their Fig. 3, top): polarity distribution of looping (spiraling) drifter trajectories (each dot corresponds to the mean location of 20-day looping trajectory segments), where blue (red) dots identify cyclonic (anticyclonic) polarity, superimposed on the regions of barrier-layer formation from Sato et al. (2006) (black boxes).

(ii) to study the distribution and seasonality of the submesoscale features identified by Griffa et al. (2008), and (iii) to verify whether and how the submesoscale can impact barrier-layer thickness (BLT), therefore testing the hypotheses put forward by Mignot et al. (2007) and Sato et al. (2006) for BL dynamics. We use a nested-domain modeling configuration, so that a high horizontal resolution of up to 1 km can be achieved with reasonable computational resources. To our knowledge, the present study is the first attempt at providing a careful and detailed numerical investigation of BL formation in the tropical South Atlantic, with a high enough resolution to study possible submesoscale effects.

The paper is organized as follows. Section 2 describes the model configuration and the diagnostics used in the paper to quantify the eddy field and the contribution of both temperature and salinity to the evolution of the upper-ocean stratification. Section 3 presents results from the simulated mean circulation, stratification, and mesoscale/submesoscale processes. Section 4 is the focus of the paper and describes the mean and submesoscale dynamics influencing barrier-layer formation in our modeled tropical South Atlantic. Conclusions are discussed in section 5.

2. Methodology

a. The ocean model

The model used in this study is the Adaptive Grid Refinement in Fortran (AGRIF) version of the Regional Ocean Modeling System (ROMS; Shchepetkin and McWilliams 2005; Penven et al. 2006; Marchesiello et al. 2003). ROMS is a primitive equation, hydrostatic, free-surface ocean circulation model, with a terrain-following s -coordinate scheme that allows us to achieve a higher vertical resolution near the ocean surface and bottom, where small-scale turbulent dynamics mostly occur (Song and Haidvogel 1994; Shchepetkin and McWilliams 2005).

ROMS-AGRIF has an online, two-way nesting capability, which allows to seemly transfer physical and dynamical information from/to the nested domain and its parent (Penven et al. 2006; Debreu et al. 2012). This is a very desirable feature especially in studies such as this where a high horizontal resolution is achieved by considering a series of increasingly smaller nested domains with a progressively higher horizontal resolution.

The particular configuration used in this study has four nested domains, depicted in Fig. 2, with a resolution ranging between $1/4^\circ$ and $1/108^\circ$ (the four domains will be referred to as South Atlantic Ocean (SAO), T9km, T3km, T1km, respectively, throughout this work). The largest (and coarsest) domain covers most of the South Atlantic Ocean, from the South American to the African coast within the latitude range 5° – 48° S, while the nested domains are all situated in the tropical region between 7° and 20° S. There are 60 vertical s levels, which are distributed spatially according to a new stretching function for the vertical coordinate that allows us to achieve a higher vertical resolution in the mixed layer in deep regions (Shchepetkin and McWilliams 2009). The vertical resolution varies between ~ 2 m at the surface and ~ 400 m in the deep ocean in deep regions. Because the focus of this study is on open ocean dynamics, the model topography is smoothed [using 2-Minute Gridded Global Relief Data (ETOPO2); National Geophysical Data Center (2001)] in such a way that the depth steepness $|\delta h|/2h$ (the so called r factor, where h = depth) is everywhere less than 0.2. Other grid-related model parameters are as follows: $\theta_s = 5$, $\theta_b = 0$, $h_c = 10$ m, and $h_{\min} = 100$ m [for details on these parameters see, e.g., Shchepetkin and McWilliams (2009)].

The parameterization of turbulent phenomena is accomplished using the K -profile parameterization (KPP) mixing scheme in the vertical direction (Large et al. 1994), while the horizontal mixing of momentum, temperature, and salinity is performed along geopotential surfaces.

FIG. 2. South Atlantic parent domain (SAO, $1/4^\circ$ resolution) and location of the three nested grids in the tropical region, ranging between $1/12^\circ$, $1/36^\circ$, and $1/108^\circ$ resolution for the largest (T9km), medium (T3km), and smallest (T1km) grids, respectively. Shading is model bathymetry with a contour interval of 500 m.

The circulation of the parent domain (SAO) at the three open boundaries is constrained using Levitus climatological fields (Conkright et al. 1998) and a radiation open boundary algorithm (Marchesiello et al. 2001). A 300-km-wide sponge layer, with viscosities increasing up to $10^4 \text{ m}^2 \text{ s}^{-1}$, is also applied at the SAO boundaries to help diffuse the strong eddies that are formed in the equatorial region and at the Brazil–Malvinas confluence region near the southern boundary. Finally, momentum and tracers are slowly nudged toward their climatological boundary values within the sponge layer to reduce inconsistencies between the interior and boundary area circulation of the SAO domain.

The model simulation analyzed in this paper is performed as follows. First, the South Atlantic domain alone is spun up from rest for a period of 10 years (inspection of mean surface kinetic energy and barotropic kinetic energy suggest that a quasi equilibrium is reached after 6–7 years of simulation). Then, the full nested configuration is initialized from the last day of the spinup and run for the period 1 January 1999–31 August 2002, using 6-hourly, 1.125° surface forcing from the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) product (Uppala et al. 2005). The atmospheric surface fluxes are computed using an ocean–atmosphere boundary layer routine (Liu et al. 1979; Fairall et al. 1996a,b), and the following atmospheric fields: 2-m air temperature, 2-m dewpoint temperature (necessary to calculate relative humidity), and 10-m zonal and meridional wind velocities from the 6-hourly analysis data; and surface thermal radiation (net longwave), surface solar radiation (net shortwave), convective and stratiform (large scale) precipitation, interpolated at 6-h intervals from the 12- and 24-h forecast

steps data.² The latter procedure is performed to reduce well-known spin up problems of the ECMWF short-range (6 h) forecast fields (Buarque et al. 2004; Roske 2006).

Hereafter, we will discuss model results for the 3-yr period, 1 September 1999–31 August 2002, using the three tropical domains. Depending on the specific issues at hand, in the following sections we will consider either a particular domain or all three at once. In particular, comparing results at an increasing resolution will help quantify the role of submesoscale processes (Capet et al. 2008a,b,c), which are satisfactorily resolved only at the highest resolution.

b. Diagnostics

In this section, we have summarized the main diagnostics used throughout the paper to quantify the submesoscale/mesoscale eddy field and the separate contribution of temperature T and salinity S to the upper-ocean stratification.

1) TURBULENCE DIAGNOSTICS AND SUBMESOSCALE FIELD

A very common diagnostic used to characterize the turbulent eddy field is the computation of relative vorticity (Bracco and McWilliams 2010) and Rossby number ζ/f , where $\zeta = \partial v/\partial x - \partial u/\partial y$ is the relative vorticity, $f = 2\Omega \sin(\phi)$ is the vertical component of planetary vorticity, Ω is the mean Earth rotation, and ϕ is latitude. In particular, submesoscale features are characterized by small spatial scales on the order of the mixed layer Rossby deformation radius (Özgökmen et al. 2011) and by a high Rossby number,

²Total precipitation = convective + stratiform.

which are indicative of ageostrophic dynamics (Molemaker et al. 2005; Muller et al. 2005; Thomas et al. 2008). Therefore, in section 3c we compute the relative vorticity field at increasing resolution as a first step to investigate the presence and seasonality of submesoscale features. Emergence of small and intense features with a high Rossby number is the first qualitative indication of a significant submesoscale field.

Further insights into the turbulent field can be gained by considering the Okubo–Weiss parameter (Okubo 1970; Weiss 1991), $Q = s^2 - \zeta^2$, where s is the strain or deformation rate and its squared value is given by

$$s^2 = \left(\frac{\partial u}{\partial x} - \frac{\partial v}{\partial y} \right)^2 + \left(\frac{\partial v}{\partial x} + \frac{\partial u}{\partial y} \right)^2.$$

The Q parameter has been widely used in two-dimensional turbulence (Elhmaidi et al. 1993; Pasquero et al. 2001) and oceanic models (McWilliams 1984; Isern-Fontanet et al. 2003, 2006) to differentiate between elliptic regions, which are highly dominated by vorticity and where Q is negative and below a certain threshold $Q < -Q_0$, and hyperbolic regions, which are highly dominated by strain and where Q is positive and $Q > Q_0$. In 2D turbulence, regions where $|Q| \leq Q_0$ are considered as part of the background turbulent flow. A typically used value for Q_0 is $0.2\sigma_Q$, where σ_Q is the standard deviation of the

Okubo–Weiss parameter computed over the whole model domain for a given time step. Elliptic regions are considered representative of closed eddy or vortex structures, whereas hyperbolic regions are considered indicative of filaments and horizontally stretched features.

While Q is often used to identify specific elliptic and hyperbolic features within a given snapshot of the velocity field (e.g., Isern-Fontanet et al. 2003), it can also be used to provide general information on the average properties of the field by appropriately integrating it in space (Petersen et al. 2006; Zavala Sanson and Scheinbaum 2008). In section 3c, we compute time series of integral values of Q and Rossby number over elliptic and hyperbolic regions separately, at various resolutions. These diagnostics allow us to quantify the impact of the submesoscale, which is fully present at the highest resolution only, and to characterize its properties by comparing the results at different resolutions. In particular, the results can provide an indication of whether the submesoscale features are predominantly elliptic or hyperbolic.

2) TRACER EVOLUTION AND SUBMESOSCALE EDDY FLUXES

The evolution of the upper-ocean stratification can be investigated by quantifying the terms of the advection–diffusion equation, which, for any given tracer θ is given by

$$\underbrace{\frac{\partial \theta}{\partial t}}_{\text{rate}} = - \underbrace{\left(u \frac{\partial \theta}{\partial x} + v \frac{\partial \theta}{\partial y} \right)}_{\text{adv_h}} - \underbrace{w \frac{\partial \theta}{\partial z}}_{\text{adv_v}} + \underbrace{\left[\frac{\partial}{\partial x} \left(K_H \frac{\partial \theta}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_H \frac{\partial \theta}{\partial y} \right) \right]}_{\text{mix_h}} + \underbrace{\frac{\partial}{\partial z} \left(K_V \frac{\partial \theta}{\partial z} \right)}_{\text{mix_v}} + \underbrace{F}_{\text{forcing}}, \quad (1)$$

where the following balance terms have been identified: the tracer rate of change, the horizontal adv_h and vertical advection adv_v terms, the horizontal mix_h and vertical mix_v mixing due to the unresolved model scales and physics (K_H and K_V stand for horizontal and vertical diffusivity, respectively), and the tracer sources/sinks due mainly to surface atmospheric interactions F . For θ = salinity, F is surface freshwater flux at the surface, while for θ = temperature, F depends on surface heat flux at the surface and penetration of solar radiation below the surface, which in the model is parameterized according to a Jerlov type-I water (Jerlov 1968).³ Note that each term of the T

and S equation was computed online and stored every day during an additional simulation identical to the one discussed so far in this paper, but carried out for the period September 1999–August 2000 only.

In addition to considering the relative importance of the various terms in Eq. (1), we also quantify the contributions of the mesoscale and submesoscale flows to horizontal and vertical advection. The mesoscale and submesoscale components (indicated with a prime and a double prime, respectively) are computed by first defining a large-scale flow $\langle u \rangle$, as a monthly average [the variable u is used as an example, but all model variables (u , v , w , T , S , and ρ , where ρ is density) are similarly decomposed]. The monthly average is chosen because we wish to investigate the seasonal cycle of submesoscale effects. Then, a filtered flow $\tilde{u}$ is estimated to separate the mesoscale and submesoscale components, by applying a one-lobe sine window with a 60-km filter length. This length scale is chosen to be smaller than the internal Rossby radius of deformation, but large enough

³ The forcing F is computed as a vertical flux; therefore, in the upper layer, F for temperature is proportional to the difference between the surface heat flux (sum of short- and longwave radiations, latent and sensible heat) and the amount of shortwave radiation penetrating below the surface. In the second layer below the surface, F for temperature is proportional to the fraction of shortwave radiation penetrating between the second and first layer, and so on for deeper layers.

to include submesoscale processes at all times and latitudes. The mesoscale and submesoscale components are therefore defined as $u' = \tilde{u} - \langle u \rangle$ and $u'' = u - \tilde{u}$, respectively, so that $u = \langle u \rangle + u' + u''$.

$$\underbrace{\left(\frac{\partial \langle u' \theta' \rangle}{\partial x} + \frac{\partial \langle v' \theta' \rangle}{\partial y} \right)}_{\text{mesoscale horizontal}} - \underbrace{\left(\frac{\partial \langle u'' \theta'' \rangle}{\partial x} + \frac{\partial \langle v'' \theta'' \rangle}{\partial y} \right)}_{\text{submesoscale horizontal}} - \underbrace{\frac{\partial \langle w' \theta' \rangle}{\partial z}}_{\text{mesoscale vertical}} - \underbrace{\frac{\partial \langle w'' \theta'' \rangle}{\partial z}}_{\text{submesoscale vertical}} - \langle \text{mixed meso} - \text{submesoscale terms} \rangle. \quad (2)$$

These terms will be quantified and discussed in section 4b.

3. Model circulation

a. Mean

The purpose of this and the following subsection is to describe the simulated large-scale circulation with a focus on the South Atlantic tropical region within latitudes 8° – 18° S and longitudes 0° – 40° W. We will therefore discuss the results of the T9km domain, as mean quantities computed over the 3-yr period, 1 September 1999–31 August 2002. In this region, the tropical South Atlantic upper circulation consists of the southern branch of the South Equatorial Current (SEC) system, which flows westward until it bifurcates near the Brazilian coast to form the western boundary current system known as the Brazil Current/North Brazil Current (and Undercurrent) (e.g., Stramma and Schott 1999; Lumpkin and Garzoli 2005; Talley et al. 2011). The region is influenced by the tropical atmospheric seasonal cycle associated with the meridional shift of the ITCZ. As the ITCZ migrates to its equatorial, southernmost position in February–March, the southwesterly winds in the tropical South Atlantic reach their annual minimum strength, whereas as the ITCZ establishes itself around its northernmost location (at $\sim 10^{\circ}$ N) during the austral winter (July–September), the South Atlantic southwesterly winds strengthen and reach their maximum in August (Mitchell and Wallace 1992; Waliser and Gautier 1993; Biasutti et al. 2012). This seasonal cycle is present in the model wind stress field, shown in Figs. 3a and 3b, and, as we shall see in sections 3c and 4, also plays an important role in setting up the mean stratification and influencing the seasonal changes in submesoscale processes and barrier-layer thickness. Following the Sverdrup theory, on the other hand, the SEC circulation, and in particular the location of its bifurcation point near the Brazilian coast, seems to be mostly influenced by the meridional location of the zero wind stress curl curve (Rodrigues et al. 2007). This curve migrates meridionally

By substituting the above decomposition in Eq. (1) and taking the monthly mean $\langle \cdot \rangle$, we derive the following contributions of the mesoscale and submesoscale flows to horizontal and vertical advection:

in a similar fashion as the ITCZ, but following a slightly shifted seasonal cycle, reaching its southernmost (northernmost) location in July (November) and causing a corresponding meridional migration of the coastal SEC bifurcation point. Indeed, our simulated mean surface circulation presented in Figs. 3c and 3d shows an SEC bifurcation point near the Brazilian coast (indicated by a white arrow) at $\sim 12^{\circ}$ S in July and $\sim 7.5^{\circ}$ S in November, shifting to the south by about 1° – 4° at 100-m depth (not shown), thus exhibiting a seasonal cycle as well as depth variations that are similar to the results by Rodrigues et al. (2007).

Another important aspect that is well simulated by the model is the mean surface and subsurface salinity structure, which is presented in Fig. 4 and compared with *World Ocean Atlas* (WOA) climatological fields. Despite the model overestimating the upper salinity by ≈ 0.2 psu compared with WOA climatology, the spatial patterns are well reproduced both in the horizontal and vertical direction. Sea surface salinity (SSS; Fig. 4a) exhibits a maximum centered at $\sim 18^{\circ}$ S and a mean meridional gradient between 14° and 8° S separating the high-salinity waters of the subtropical gyre from the relatively fresher equatorial waters, in agreement with observations (Tsuchiya et al. 1994) and WOA climatology (Fig. 4d). The SSS seasonal cycle (not shown) is fairly weak, exhibiting values fresher than 36 psu only to the north of 8° S and to the east of 20° W during the austral winter (June–September). Indeed, as the climatological ITCZ remains centered around the equator or to the north (Waliser and Gautier 1993; Biasutti et al. 2012), our region of interest is very little affected by mean tropical precipitation. Thus, mean freshwater flux ($E-P$, where E = evaporation rate and P = precipitation rate) is always positive in the latitude range 8° – 18° S and to the east of 35° W (the $E-P$ seasonal cycle will be further discussed in section 4).

Model results also show the correct location of the salinity maximum at ~ 90 -m depth (Figs. 4b,c), which characterizes the subtropical underwater. South Atlantic STUW is found between 13° and 6° S and to the west

FIG. 3. Wind stress amplitude (note the log scale) and vectors for the months of (a) February and (b) August. The white line is the zero wind stress curl contour. Streamlines of surface model velocities for (c) July and (d) November. The white arrow indicates an estimate of the SEC bifurcation point at the coast. Note the different months depicted in (a) and (b), and in (c) and (d). Monthly averages are computed over the 3-yr period, 1 Sep 1999–31 Aug 2002. Results are for the T9km grid.

of $\sim 10^\circ\text{W}$, advected first westward by the SEC and then northward by the North Brazil Undercurrent [Blanke et al. 2002; Stramma et al. 2005; see also Fig. S9.28 of the online supplemental material of Talley et al. (2011)]. As mentioned in section 1, reproducing the STUW correctly is very important for a good representation of barrier-layer formation, as we shall see in section 4 and

as also predicted by previous investigators (Sprintall and Tomczak 1992; Mignot et al. 2007; Araujo et al. 2011). Finally, we have compared the annual model sea surface temperature (SST) with the Group for High Resolution Sea Surface Temperature (GHRSSST) 0.25° data product averaged over the same period, 1 September 1999–31 August 2002 [results not shown; data produced by the National

FIG. 4. Annual salinity at the surface, 90-, and 200-m depth for (a)–(c) the T9km grid and (d)–(f) WOA climatology.

Oceanic and Atmospheric Administration (NOAA) National Climatic Data Center and available online at http://podaac.jpl.nasa.gov/dataset/NCDC-L4LRblend-GLOB-AVHRR_OI. Similarly to SSS, model SST has a warm bias of $\sim 0.5^\circ\text{C}$ with respect to the observations (data error is generally lower than 0.2°C in the tropical South Atlantic), but the spatial structure is well reproduced.

b. Mean stratification and barrier-layer thickness computation

In this section, we take a closer look at the modeled mean stratification and the ability of the model to reproduce the observed barrier layers. Barrier-layer thickness is computed as the difference between the isothermal layer depth and the mixed layer depth, $\text{BLT} = \text{ILD} - \text{MLD}$, so that a nonzero BLT is found where the isothermal layer is deeper than the layer where density is well mixed. We adopt a commonly used method to calculate ILD [e.g., Carton et al. 2008; de Boyer Montégut et al. 2004; see also Kara et al. (2000) for a discussion on isothermal and mixed layer depth computation], that is the depth where temperature has changed by 0.2°C with respect to the 10-m reference level. All temperature and density changes in the upper 10 m of the water column are therefore neglected, to avoid contamination from the daily cycle. Similarly, the MLD is defined as the depth where a density change corresponding to a 0.2°C temperature change has occurred with respect to the

density at the 10-m reference level. If there is no salinity stratification and if density stratification is controlled entirely by temperature, then MLD is equivalent to ILD and $\text{BLT} = 0$. Note that, because ROMS sigma levels occur at varying depths and are characterized by very different thicknesses over the domain, we have first vertically interpolated the model temperature and salinity onto uniform depth levels before computing the ILD and MLD.⁴

The simulated model stratification is compared with WOA climatology (not shown) and with the stratification measured by the four Argo profiler floats that sampled the South Atlantic tropical region of interest between 1999 and 2002 (Fig. 5, top left). Figures 5a–c show the mean profiles of potential temperature, salinity, and density in the upper 300 m from Argo data (thick lines) and T9km model fields (thin lines; model quantities were interpolated onto the Argo data points, and both model and Argo data mean values were calculated within specified vertical bins). The difference between model and Argo data quantities is presented in Figs. 5d–f, in terms of bias (solid lines) and standard deviation (dashed lines). A model–data difference in terms of standard deviation signifies a difference due to the

⁴ The depth levels used for the interpolation are 0: 0.2: 20, 20: 1: 50, 50: 2: 100, 100: 5: 200, 200: 10: 300, 300: 20: 500, 500: 50: 1000, 1000: 100: 2000, and 2000: 200: 6000 m.

FIG. 5. Comparison between Argo and T9km model profiles in the upper 300 m. (top left) Argo trajectory data covering the area of interest between 1999 and 2002 (different colors for different floats). (a)–(c) Mean profiles of potential temperature, salinity, and density, respectively, from Argo (thick lines) and model (thin lines) data. Also shown is the mean isothermal layer depth (a) and the mean mixed layer depth (c) from Argo (thick lines) and model (thin lines). (d)–(f) Difference between model and Argo results for potential temperature, salinity, and density, respectively, in terms of bias (solid lines) and std dev (dashed lines).

model not predicting the T and S structures at the same time and location as the Argo data, for each depth range where statistics were calculated.

The model potential temperature and salinity profiles generally agree well with the Argo profiles. A bias in potential temperature of up to 1°C at 150 m is due to the fact that the model ILD is slightly shallower than in the data (by ≈ 10 m) and the model thermocline is sharper than in the Argo profiles. On average, the model surface and subsurface (100 m) salinity is overestimated by up to 0.2 psu, consistently with the WOA salinity comparison in Fig. 4. Because on average both the ILD and MLD are shallower than in the data, model and data BLTs are similar, within a ± 10 -m uncertainty.

c. Eddy field and submesoscale

In this section, the properties of the turbulent eddy field and the emergence of submesoscale processes in the tropical South Atlantic are studied by analyzing the results of the T9km, T3km, and T1km simulations. The internal Rossby radius of deformation—a typical spatial scale for mesoscale processes—ranges between 80 and 150 km at these latitudes (Chelton et al. 1998). A similar scale for submesoscale processes in the upper ocean is the mixed layer Rossby deformation radius (Özgökmen et al. 2011), equal to $\sqrt{\text{MLD}g\Delta\rho/\rho_0}/f$, where g is gravitational acceleration, $\Delta\rho$ is the density jump across the mixed layer, and ρ_0 is the averaged density within the mixed layer. Our model results show that, in the region of interest, this scale varies between 6 and 17 km in summer and between 6 and 22 km in winter, with variations mainly owing to latitude. Therefore, all the tropical model simulations are capable of solving the mesoscale, whereas only T1km can be considered as properly resolving the submesoscale at all times and latitudes.

This can be visualized in Fig. 6, which shows snapshots of relative vorticity $\zeta \text{sign}(\phi)$, at 0-, 100-, and 500-m depth, in austral winter and spring–summer. The contours of Rossby number equal to 1 and -1 are overlapped over the relative vorticity field, to highlight the regions characterized by ageostrophic dynamics. The following conclusions can be suggested from Fig. 6. (i) Small-scale structures, mainly small eddies and thin filaments, on the order of the submesoscale [$O(3\text{--}5\text{ km})$] emerge within the T1km simulation and are mostly seen during austral wintertime (represented by the 17 September 1999 snapshot in Figs. 6a–c). Relative vorticity on the spring snapshot (9 November 1999 in Figs. 6e–f) shows a much less turbulent field, characterized by larger eddies with scales closer to the mesoscale. (ii) The submesoscale-like structures in winter are found in the

upper ocean: while some small filaments are still visible at 100-m depth (Fig. 6b), coherent structures at 500 m are mostly mesoscale (Fig. 6c,f). (iii) Upper-ocean small-scale features are characterized by a high Rossby number (≥ 1) and by mostly cyclonic vorticity [$\zeta \text{sign}(\phi) > 0$ in the Southern Hemisphere], in agreement with the findings of Griffa et al. (2008).

A deeper understanding of the turbulent field can be obtained by considering the distribution of the Okubo–Weiss parameter [see section 2b(1)]. Figure 7 presents the daily variations of the Okubo–Weiss parameter averaged over elliptic (with changed sign; Q^-) and hyperbolic (Q^+) regions using $Q_0 = 0.2\sigma_Q$, for the T9km (Fig. 7a) and T1km (Fig. 7b) grids. In both simulations, Q^- is generally higher than Q^+ , indicating a stronger influence of eddies with respect to filaments/strain-dominated structures, and the difference, $Q^- - Q^+$, peaks in August–September while reaching a minimum value during the austral summer–fall (January–May). The relative difference, $(Q^- - Q^+)/[0.5(Q^- + Q^+)]$, exhibits a mean value that only changes between 47% and 34% from T9km to T1km, but the highest-resolution results have a much higher temporal variability (standard deviations = 4% and 20% for T9km and T1km, respectively). The predominance of eddy structures found here is in contrast to the findings of Mensa et al. (2013) in the Gulf Stream recirculation region, where submesoscale filaments and strained structures dominate over submesoscale vortices. The latter justifies the lack of submesoscale eddies observed from surface looping trajectories in the Gulf Stream recirculation by Griffa et al. (2008).

Further insights into the tropical South Atlantic submesoscale field are gained from the daily variations of $|\zeta/f|$ averaged over elliptic and hyperbolic regions, shown in Fig. 8. Values of $|\zeta/f|$ are higher for vorticity-dominated regions than for strain-dominated regions, on average by an approximately threefold factor. As already seen from the snapshots in Fig. 6, at a higher resolution the resolved smaller-scale structures are characterized by a high Rossby number, and Fig. 8 suggests that this is caused by eddy-like features. Both T9km and T1km results show a seasonal cycle in $|\zeta/f|$ for elliptic regions, although seasonal changes in the low-resolution grid are much weaker than in the high-resolution grid. This may be due to the fact that T9km is partially resolving submesoscale processes during winter, when submesoscale processes are characterized by a larger spatial scale because of the decreased upper-ocean stratification. On the other hand, the Rossby number in hyperbolic regions is always much smaller than 1 and exhibits a less evident seasonal cycle for both T9km and T1km.

FIG. 6. Relative vorticity at (a),(d) 0; (b),(e) 100; and (c),(f) 500 m; for 17 Sep 1999 (left) and 9 Nov 1999 (right) representative of austral winter and spring–summer conditions, respectively. The T1km snapshots are superimposed on the T9km snapshots (T3km not shown), to emphasize the different field resolution (the square indicates the boundaries of the T1km domain). White contours mark the regions where $\zeta/f \geq 1$ (regions where $\zeta/f \leq -1$ are not present).

FIG. 7. At 0 m, Q computed for the (a) T9km and (b) T1km simulations [note the different scales used in (a),(b)]. Thick lines indicate $-Q$ averaged over elliptic, vorticity-dominated regions ($Q < -Q_0$), while thin lines indicate Q averaged over hyperbolic, strain-dominated regions ($Q > Q_0$). The threshold Q_0 is equal to $0.2\sigma_Q$. Relative differences between the thick and thin lines range between 33% and 60% (mean = 47%, std dev = 4%) for the T9km case, and between -33% and 97% (mean = 34%, std dev = 20%) for the T1km case.

These results are confirmed by the histograms of ζ/f , presented in Fig. 9 and computed for all points in the domain for all three grids: T9km, T3km, and T1km. The histograms show the gradual spread of ζ/f toward higher values when resolution is increased (except that anticyclonic structures in T9km seem to have similar Rossby number as the higher-resolution grids), and also point to the asymmetric distribution of ζ/f toward positive values (cyclonic polarity). The asymmetry is present at all resolutions, with skewness values of 1.74, 1.78, and 2.43 for T9km, T3km, and T1km, respectively, and is therefore more pronounced and characterized by a longer tail in T1km. Partitioning the values of ζ/f into an austral winter (July–September) and summer subset (December–February), the corresponding histograms (not shown) reveal that the cyclonic asymmetry is always present, but decreases substantially at the lower resolutions during summer.

In summary, our simulations indicate the following eddy field characteristics in the tropical South Atlantic. (i) Eddy and filament structures at the scale of the submesoscale (3–15 km) emerge in T3km and T1km in the upper 100 m of the water column. Vorticity-dominated

regions are characterized by a high Rossby number [$O(1)$] at the higher resolutions, suggesting that they are associated with ageostrophic submesoscale vortices. Strain-dominated regions exhibit ~ 3 times smaller Rossby numbers. (ii) Submesoscale eddy structures at high ζ/f show a seasonal cycle with maximum values in the austral winter (August) and minimum values in summer (December–February). This seasonal cycle is strongest in T1km, where the submesoscale is best resolved, and coincides with the seasonal cycle of wind strength and vertical stratification as will be discussed in section 4. (iii) Both vorticity- and strain-dominated regions are mainly cyclonic (characterized by negative vorticity in the Southern Hemisphere), in agreement with what was found by Griffa et al. (2008). This is seen at all resolutions, but the eddies in T1km have higher ζ/f .

4. Barrier-layer formation and submesoscale effects

The mechanisms responsible for the formation of the simulated barrier layers in the tropical South Atlantic will be investigated in this section, focusing on

FIG. 8. At 0 m, $|\zeta/f|$ computed for the (a) T9km and (b) T1km simulations and averaged over elliptic (thick lines) and hyperbolic regions (thin lines).

the BL mean dynamics (section 4a) and the possible influence of submesoscale processes on BLT (section 4b). Because the thickest BLs are found in the region covered by the T3km domain, these grid results will be discussed in section 4a, whereas the impact of resolution will be illustrated in section 4b by considering results from all three tropical grids.

a. Dynamics of mean barrier-layer formation

As discussed in section 1, one suggested mechanism for tropical BL formation is subduction of surface high-salinity waters during strong SSS front events (Sato et al. 2006). To test this theory and verify whether salinity fronts are associated with precipitation events, we monitor the model solution by looking at animations over the 3-yr period, September 1999–August 2002, of the following quantities: 1) BLT, 2) SSS field superposed with contours of salinity gradients VSSS, and 3) freshwater flux $E-P$ for both the T3km and SAO domains (the animations for T3km are available as supplemental materials). Two snapshots from 1) and 2) for 17 September 1999 and 9 November 1999 are presented in Figs. 10a–d for the T3km grid (snapshots of $E-P$ are not shown for brevity because $E-P > 0$ everywhere in T3km at those times). The chosen snapshots are representative of winter and spring–summer conditions, respectively. The seasonal

cycle and general temporal variability of the quantities of interest are shown in Figs. 11a–c and 12a and 12b. In particular, Figs. 11a–c display daily values of, respectively: BLT averaged over regions where $BLT \geq 10$ m; area fraction where $10 \leq BLT < 40$ m and $BLT \geq 40$ m; and domain-averaged horizontal gradients of

FIG. 9. Histogram of the Rossby number at 0 m from all points in the domain, for the T9km (gray lines), T3km (thin black lines), and T1km simulation (thick black lines).

FIG. 10. Two model snapshots [(a),(c),(e) 17 Sep 1999 and (b),(d),(f) 9 Nov 1999] representative of winter and spring–summer conditions, respectively, of: (a),(b) BLT and contour of $VSSS = 1.5 \times 10^{-5} \text{ psu m}^{-1}$ (note that negative BLT corresponds to the compensated layer case mentioned briefly in section 1); (c),(d) SSS and contours of $VSSS = 1.5 \times 10^{-5}$ (thin lines) and $VSSS = 3 \times 10^{-5} \text{ psu m}^{-1}$ (thick lines); and (e),(f) ILD and contour of $BLT = 40 \text{ m}$. Results are for the T3km grid (axes are lat and lon).

SSS and surface density. Figures 12a and 12b show the climatological annual cycle of Figs. 11a and 11c, respectively.

Let us first summarize the scenario depicted by the above-mentioned figures (Figs. 10a–d; 11a–c; and 12a,b). A clear result is that austral winter months (July–September) mark the formation of thick barrier layers (Figs. 10a; 11a,b; and 12a), in agreement with previous

observational studies (e.g., Sato et al. 2006; Mignot et al. 2007). The thick winter BLs cover extensive areas of the domain, at times up to 60% of the total area (Fig. 11b). The high BLT values are found adjacent to regions of high $VSSS$ (salinity fronts), always on the fresh side of the fronts (animations available as supplemental materials and Figs. 10a,c), supporting the idea that the barrier layer is formed in conjunction with subduction

FIG. 11. Daily changes for the T3km grid of (a) BLT averaged over regions where $BLT \geq 10$ m, (b) area fraction where $10 \leq BLT < 40$ m (black line) and $BLT \geq 40$ m (gray line), (c) domain-averaged horizontal gradients of SSS (thin line) and surface density (thick line), and (d) domain-averaged ILD (thick line) and MLD (thin line).

of high-salinity waters. Nevertheless, similar VSSS, both in terms of 2D structure and front strength (animations available as supplemental materials and Figs. 10c,d; 11c; and 12b), are found later in the season, but barrier layers with only $BLT \leq 10$ m occur during the November–March months. Daily changes of VSSS and thick BLs are uncorrelated with each other between August 1999 and July 2000, and generally out of phase at the remaining

times. Furthermore, the freshwater flux is rarely negative within the T3km domain and does not influence the SSS structure in this region directly (daily $E-P$ not shown, climatological seasonal cycle in Fig. 12d). On the other hand, meridional (and subsequently zonal) advection of low-salinity equatorial waters is observed throughout the year (animations available as supplemental materials and Figs. 10c,d). Such findings suggest that the surface salinity structure and the presence of fronts may influence the BL horizontal distribution, but alone do not cause its formation.

Further investigation of the evolution of vertical stratification is accomplished by monitoring animations of the ILD horizontal structure with contours of $BLT = 40$ m superimposed, and vertical distributions of density and salinity at $12^\circ S$ (supplemental materials). As before, two snapshots for 17 September 1999 and 9 November 1999 are extracted from the three animations and presented in Figs. 10e and 10f and Fig. 13. Daily variations and climatological annual cycle of the averaged ILD (and MLD) are displayed in Figs. 11d and 12c, respectively. From these results, it is apparent that the isothermal-layer depth exhibits a very similar seasonal cycle as BLT, and daily ILD changes closely follow the BLT changes. The highest BLTs are found at times when thick (>60 m) ILDs are also present, whereas BLTs ≤ 10 m or no BLs at all are formed when $ILD < 50$ m (Fig. 10e,f).

The thickening of the isothermal and mixed layers in winter is likely due to the enhanced mixing induced by strong winds as well as increased buoyancy owing to the negative surface heat flux. This can be seen in Figs. 12e and 12f, which show the climatological annual cycle of wind strength and surface heat flux, averaged over the T3km domain. Maximum wind strength is found in June–August, and surface heat flux reaches maximum negative values in June, returning to positive values in September. A question arises at this point, however: why is mixing not as effective at deepening the mixed layer as much as the isothermal layer? Indeed, MLD values are on average always smaller than ILD (suggesting the presence of a climatological BL in the tropical South Atlantic, albeit with $BLT < 10$ m outside the July–September window). A possible answer is that salinity can only mix up to a certain extent because of the strong salinity stratification induced by the (i) advected fresh surface waters and (ii) subsurface salinity maximum associated with the South Atlantic subtropical underwater. This is seen in Figs. 13c and 13d, for example, where the presence of STUW is seen at both the 17 September 1999 and 9 November 1999 snapshots. Pockets of low or mixed salinity in the upper 40–60 m are visible at both snapshots, but only on 17 September do they correspond to regions

FIG. 12. Climatological annual cycle of (a)–(c) fields in Figs. 11a,c, and d, respectively; (d) surface freshwater flux (i.e., $E-P$); (e) wind stress strength; and (f) surface heat flux. Forcing fields are 10-day low-pass-filtered quantities.

of deep (up to 100 m) ILDs and consequent thick BLs. Furthermore, those regions where salinity does mix down to 100 m on 17 September (e.g., within longitude ranges $\sim 29.5^{\circ}$ – 31° W, 23.2° – 23.5° W, and to the east of $\sim 23^{\circ}$ W) are also characterized by the smallest BLTs.

The different evolution of the temperature and salinity stratification can be quantified by considering the terms of the tracer advection–diffusion equation [Eq. (1), see section 2b(2)]. To have an idea of the tracer equation seasonal cycle, we first consider the monthly variations of the temperature and salinity rate of change [first term in Eq. (1)] integrated over the mixed layer and over a region typically characterized by $BLT > 10$ m (10° – 15° S, 25° – 40° W; hereafter we will refer to this area as “region A”). Results are presented in Fig. 14. They show that, starting in April and continuing until August, the surface heat flux (Fig. 12f) induces a tendency to cool the mixed layer, and this tendency is enhanced in June–August by high winds (Fig. 12e). Positive heat flux between September and March produces a warm tendency in the upper ocean. The salinity rate of change, on the

other hand, is less dramatic and its seasonal cycle only resembles that of the freshwater flux (Fig. 12d). It shows a freshening tendency between August and October associated with horizontal advection of fresh equatorial water.

With these seasonal cycles in mind, we now proceed to analyze the seasonal changes of the dominant terms of Eq. (1) as a function of depth. For brevity, only results for the months of August and November have been included in Fig. 15, separately for temperature and salinity. In August (and starting in April with intense surface cooling), buoyancy- and wind-induced mixing dominates the balance within the IL, and its cooling tendency surpasses the warming tendency induced by horizontal advection of warm low-latitude waters and penetration of solar radiation. This induces a deepening of the isothermal layer starting in April and peaking in August (Fig. 15a). Following the arrival of austral spring conditions (positive heat flux and low winds), the isothermal layer warms and stratifies between September and March. Maximum stratification is reached in November (Fig. 15c),

FIG. 13. Vertical slice at 12°S for the two snapshots shown in Fig. 10 of (a),(b) density ($\rho - 1000$) and (c),(d) salinity (psu), superposed with the MLD (black line) and ILD (green line). Snapshot 17 Sep 1999 (9 Nov 1999) is presented in (a),(c) [(b),(d)]. Results are for the T3km grid (x axis is lon).

when warming due to advection and solar radiation dominates cooling effects due to vertical mixing throughout the isothermal layer.

The adjustment followed by upper-ocean salinity in August (Fig. 15b) is fairly different than that followed by temperature. Because of a positive surface input due to $E - P > 0$, vertical mixing of salinity in the upper 20–30 m of the water column induces a positive salinity change. Nevertheless, contrary to the effect on temperature, mixing never dominates the salinity balance. Above 30 m, freshening due to horizontal advection is slightly larger than mixing-induced salinification. Below 40 m, the vertical mixing changes sign, but so does the vertical advection (causes of vertical salinity advection at these depths will be discussed in section 4b), and an overall fresh tendency is established throughout the mixed layer. The reason why mixing does not play a major role in homogenizing the upper-ocean salinity in winter is likely due to the strong salinity stratification caused by the presence of advected upper-ocean freshwater and the subsurface salty STUW (Figs. 10c,d and 5b). Because the described salinity stratification prevails year-round, a climatological barrier layer is ever present in

the tropical South Atlantic, but it becomes much thicker during the austral winter because of a halocline sitting on a much deeper isothermal layer. These results confirm the theory put forward by Mignot et al. (2007) for BL formation in tropical regions located on the equatorward side of STUWs.

b. Effects of submesoscale on BLT

In this section, we focus on the interplay between submesoscale processes and barrier-layer formation. As a first step to investigate the effects of the submesoscale, we compare results from the two tropical domains, T9km and T1km, for which submesoscale processes are generally unresolved and well resolved, respectively. Note that, because our model configuration is two-way nested, all nested-grid solutions are affected by each other. A stricter assessment of model resolution on BLT formation would require independent simulations at different resolutions.

Figure 16 shows daily changes of BLT averaged over regions where $BLT \geq 10$ m, and domain-averaged MLD and ILD, from T9km and T1km. The BLT values in T9km have been calculated within region A (see section 4a for

FIG. 14. Climatological annual cycle of mixed layer rate of change of potential temperature (thick line) and salinity (thin line), averaged over the region 10°–15°S, 25°–40°W of the T3km domain (region A).

a definition of this region), where formation of the thickest barrier layers is seen. Figure 16a shows that thick BLs are generally higher in T1km than in T9km, with a difference factor of 20%–35% during the winter season. Evidence of this is also seen in the horizontally averaged MLD and ILD (Figs. 16b,c). The mean mixed layer is shallower in T1km than in T9km, generally by 10–20 m. On the other hand, horizontal resolution (i.e., submesoscale) does not seem to affect the isothermal layer depth as much as the MLD.

We have further investigated this issue by directly estimating the effect of the mesoscale and submesoscale flow on the tracer equation balance, as described in section 2b(2) [Eq. (2)]. The mixed meso–submesoscale eddy fluxes [last term in Eq. (2)] are found to be negligible at all times. The horizontal eddy fluxes [first two terms in Eq. (2)] are dominated by the mesoscale component (results not shown), as also found by Capet et al. (2008b). The vertical eddy fluxes, on the other hand, are dominated by the submesoscale contribution. This is apparent in Fig. 17, where the third and fourth terms of Eq. (2), computed from T3km and T1km, are plotted as a function of depth for the months of August and November (we have chosen to show T3km rather than T9km results in order to make a direct comparison with

Fig. 15). Results are horizontally averaged over region A in T3km and over the whole domain in T1km, and are presented separately for temperature and salinity.

Let us first discuss the results for August (Figs. 17a,b). The submesoscale contribution to the tracer vertical flux divergence is almost one order of magnitude higher than the mesoscale contribution in T1km; this difference is reduced in T3km, but still visible. The vertical patterns in Figs. 17a and 17b are very distinctive, showing a positive (negative) peak in temperature (salinity) flux divergence at ≈ 10 -m depth and a negative (positive) peak in temperature (salinity) flux divergence at ≈ 90 m for temperature and ≈ 70 m for salinity. The overall effect is therefore to produce a warmer, fresher, and more buoyant environment at the surface and a colder, saltier, and less buoyant environment near the base of the isothermal and mixed layers, thus inducing a tendency for a more stratified IL and ML. This is in agreement with what found by Fox-Kemper et al. (2008) in an idealized model configuration and by Mensa et al. (2013) in the Gulf Stream recirculation region. One interesting new result here is that the submesoscale flux divergence of salinity between 90 and 140 m tends to restratify the relatively homogeneous, salty layer associated with STUW and found at 90–110-m depth

FIG. 15. Dominant terms of the tracer equation balance [Eq. (1)] for (a),(c) potential temperature and (b),(d) salinity, averaged over the region 10° – 15° S, 25° – 40° W of the T3km domain (region A) and over the months of August (top) and November (bottom). The following terms are presented: meridional, zonal, and total horizontal advection (dot-dashed, dashed, and thick solid black line, respectively), vertical advection (thin black line), vertical mixing (thick blue line), forcing contribution (thin blue line), and rate of change (red line). The horizontal mixing terms are very small and not shown. Thin (thick) horizontal lines indicate the depths of the mixed (isothermal) layer.

(see for example Fig. 5b). Most importantly, if we compare Figs. 17a and 17b with Figs. 15a and 15b, we note that the submesoscale restratification effect has a much greater impact on salinity than temperature. In both cases, the sum of mesoscale and submesoscale contributions to the vertical flux divergence accounts for most of the total vertical advection below 20 m. But, as discussed in section 4a, temperature stratification in the isothermal layer is mostly controlled by vertical

mixing in winter, and the submesoscale restratification effect does not have the strength necessary to compete with it. In the upper 40 m (and below the very surface of the ocean), the submesoscale and mesoscale contribution to the temperature balance is at most $0.26^{\circ}\text{C month}^{-1}$ in the T3km region A (gray lines in Fig. 17a), whereas mixing ranges between 1° and $\approx 20^{\circ}\text{C month}^{-1}$ in the same region (thick blue line in Fig. 15a). Between 40 m and the base of the isothermal

FIG. 16. Daily changes of (a) BLT averaged over regions where BLT ≥ 10 m (and over region A in the T9km case), (b) domain-averaged MLD, and (c) domain-averaged ILD. Thin and thick lines indicate results from T9km and T1km, respectively.

layer, the maximum submesoscale/mesoscale contribution goes up to $0.5^{\circ}\text{C month}^{-1}$, but it produces the same effect as mixing and horizontal advection, that is, to keep the lower isothermal layer as cold as the surface and fairly homogenized. This may be the reason why ILD is not very affected by a change in resolution, as seen in Fig. 16c.

For salinity, on the other hand, mixing does not affect stratification as much (Fig. 15b). In the upper 40 m, the vertical flux divergence due to submesoscale processes acts in conjunction and with comparable strength as the salinity rate induced by horizontal advection (thick black line in Fig. 15b), that is, it acts to make the upper ocean even fresher. Below 40 m, the submesoscale vertical flux of salinity makes the base of the mixed layer even saltier, exacerbating the situation already induced by the STUW presence at ≈ 90 -m depth. Average BLT in T9km region

A is 13.56 m in August, while it becomes 17.05 m in T1km, with a nonnegligible relative difference of $\approx 23\%$.

In November (Figs. 17c,d), submesoscale vertical divergence is not only reduced, but also acts over the same depth range for both temperature and salinity. Furthermore, and as expected, T3km shows a noticeably less pronounced submesoscale contribution in spring.

5. Concluding remarks

In this paper, we presented the results from a new, very high-resolution model configuration of the tropical South Atlantic Ocean, with the objective of investigating submesoscale distribution, formation mechanism of barrier layers, and possible effects of submesoscale processes on BL thickness. We found that submesoscale in this region is mostly elliptically dominated, in contrast

FIG. 17. Vertical flux divergence of (a),(c) temperature and (b),(d) salinity due to the mesoscale ($-\partial\langle w'\theta'\rangle/\partial z$; dashed lines) and the submesoscale component ($-\partial\langle w''\theta''\rangle/\partial z$; solid lines), obtained from T3km region A (gray lines) and T1km (black lines). Results are for the months of August in (a),(b) and November in (c),(d).

to what was determined by Mensa et al. (2013) in the Gulf Stream recirculation region, where the submesoscale is strain and filament dominated. The differences are likely due to the fact that the tropical South Atlantic has a very weak large- and mesoscale flow compared to the northwestern Atlantic. Therefore, strong mesoscale fronts are absent and submesoscale features develop around much weaker large-scale fronts, which are most likely generated by meridional advection of warm and fresh equatorial waters into the relatively colder and saltier subtropics. Small-scale

eddies are predominantly cyclonic in this region, as also found by Griffa et al. (2008). Submesoscale processes are strongest during the austral winter (July–September), in agreement with results by Mensa et al. (2013), possibly because the thick mixed layer in winter provides the necessary available potential energy for mixed layer instabilities and submesoscale features to generate (e.g., Boccaletti et al. 2007). A more detailed investigation is necessary to assess submesoscale formation processes and associated seasonal cycle in the tropical South Atlantic.

Our study of the mechanisms causing the formation of barrier layers in tropical regions has confirmed the hypothesis put forward by Mignot et al. (2007), that is, that thick BLs are generated in winter because of enhanced buoyancy- and momentum-induced mixing, which deepens the isothermal layer more than the mixed layer. The presence of a strong salinity stratification, owing to upper-ocean advection of fresh waters and the subsurface location of salty STUW, ensures that mixing is not as effective at homogenizing salinity as much as temperature. Surface salinity fronts associated with the meridional advection of fresh equatorial waters were hypothesized to be important for mean BL formation by Sato et al. (2006). And indeed we find that BLs are formed near SSS fronts, but the presence of a deep isothermal layer is also instrumental for the formation of a thick barrier layer. Strong SSS fronts are generally found year-round, but only in winter does the IL deepen enough for thick BLs to emerge.

On the other hand, fronts and associated submesoscale dynamics still affect BLT. We have investigated this by comparing results from the finest grid, T1km, with those obtained from a region of T9km and T3km where the thickest BLs are formed (region A). We find the emergence of submesoscale features at higher resolution and an increasing effect of the submesoscale on the vertical fluxes of tracers from T9km to T1km. Submesoscale processes are known to restratify the mixed layer, opposing the homogenizing influence of vertical mixing. This restratification effect is seen in the Gulf Stream study by Mensa et al. (2013), but not in the California Current simulations by Capet et al. (2008b), possibly because of the use of a constant surface forcing inducing an overestimated vertical mixing. However, it is not straightforward to justify why submesoscale would restratify the isothermal layer differently from the mixed layer. Indeed, we find that the restratification effect is similar for temperature and salinity, but it is overcome by mixing in the isothermal layer. Instead, it contributes to strengthening the salinity stratification, freshening the upper ocean in concert with what already accomplished by horizontal advection of fresh equatorial waters, and making the base of the mixed layer saltier, contributing to the effect of the STUW presence. Thus, mixing is even less effective at homogenizing the mixed layer when submesoscale processes are present.

We can speculate that the findings of this paper, especially in terms of influence of the submesoscale on barrier-layer formation, could be valid also for other tropical areas and relevant for climate reasons. Indeed, the upper-ocean structure plays a fundamental role on air–sea interaction processes and on their climate

consequences, especially in the very active tropical regions (Blanke and Delecluse 1993). In the tropical Atlantic, BLs are thought to be strictly linked to SST patterns and the ITCZ location (Breugem et al. 2008), and it has been suggested that an increased realism in BL formation mechanisms in numerical models could help alleviate commonly found SST biases (Balaguru et al. 2012b). It should be noted that the thickest and most influential Atlantic BLs are actually found in the northwestern subtropical area, rather than in the southern region considered here. Nevertheless, the South Atlantic region is interesting because it allows us to focus on the specific submesoscale mechanisms, which are prevalent given the lack of an energetic mesoscale field. The present findings on submesoscale dynamics and impact are likely to be relevant also for other BL regions, as suggested by the results in Fig. 1 that show the presence of submesoscale eddies with dominant positive vorticity and elliptical properties in most of the BL tropical areas. The relative importance of the submesoscale impact with respect to other factors will probably depend on the specific region and will need further investigation.

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