A model study of temperature anomaly propagation from the subtropics to tropics within the South Atlantic thermocline

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Abstract. A water mass flow formed by the Benguela Current and the South Equatorial Current connects the eastern subtropics to the western tropics through the Atlantic upper thermocline. We perform a process study with an Atlantic OGCM to examine whether and how synthetic subtropical mixed layer heat anomalies would employ this corridor to reach low-latitudes. Our results suggest that it is a realistic scenario and that the time scale and trajectory of the movement can be explained to first approximation by time-mean flow advection, since salinity compensation is tempering the density perturbation. In addition, wave processes seem to influence strongly the evolution of the intensity and the shape of the anomalies. It is apparent that an approach using both perspectives is necessary to fully understand and predict subsurface oceanic teleconnections.

1. Introduction

Recent studies suggest that subsurface equatorward displacement of heat anomalies originating in the subtropics could be one of the mechanisms involved in the decadal variability of the sea surface temperature (SST), particularly in the tropics. Most work along these lines has been in the Pacific Ocean because of possible decadal-scale links to the El Niño Southern Oscillation (e.g., Gu and Philander, 1997). In the Atlantic, subtropical-tropical communication has received less attention. Nevertheless, the decadal variability in this basin, in the tropics (e.g., Servain, 1991) and subtropics (e.g., Tourre et al., 1998), is essential to the climate of the surrounding continents (e.g., Fontaine et al., 1999; Hurrel, 1996) and it is important to look at this potential mechanism. In the South Atlantic below the mixed layer, the South Equatorial Current (SEC) forms a broad thermocline flow, fed by the Benguela Current, that provides the largest part of the shallow water of the tropical and equatorial regions (Stramma and England, 1999). It is therefore an interesting candidate to study as a subsurface pathway to the western low-latitudes for heat anomalies formed in the southeastern subtropics. More generally, the analyzis of propagation of signal along such a pathway is important since it provides insight to the thermocline ventilation mechanism and can be compared to Tritium studies on equator ventilation (e.g., Fine et al., 1987).

Regarding the possibility for heat anomalies to move along subsurface mean flow, Deser et al. (1996) studied one propagation of a subducted mid-latitude temperature anomaly (≈0.5ºC) on constant density surfaces (isopycnals) at the speed of the mean currents in the North Pacific. Schneider et al. (1999a) confirmed this finding, regardless of the sign of the anomalies. These authors suggest that perturbations follow the subtropical gyre circulation from around 30ºN to 15ºN-18ºN, whereas equatorward of this latitudes, the physics of heat anomalies is dominated by local wind stress curl. Hence, north of 18ºN at least and as invoked by Gu and Philander (1997), these results can be interpreted as advective processes within the ventilated thermocline. The mechanism responsible for the anomalies to behave like passive tracer could be a density compensation by salt anomalies (Deser et al., 1996; Schneider et al., 1999a,b). A competing theory is that such movements correspond to baroclinic planetary waves within a mean current, having trajectories and speeds comparable (but smaller) to the mean current (Liu, 1999; Huang and Pedlosky, 1999). Liu and Shin (1999) compared the subsurface evolution of a synthetic temperature anomaly and a companion passive tracer in an idealized North Atlantic OGCM. They found comparable pathways, but a propagation of the passive tracer two times faster, and interpreted this discrepancy as resulting from the propagation of a so-called « advective baroclinic mode ».

Several questions are addressed here: how does the subsurface pathways of heat anomalies compare to mean flow advection? How significant are the associated anomalous density signal and currents and what is the importance of salt compensation? What is the role of wave versus advective processes? And at last, are mid-latitudes anomalies capable of traveling several years and reaching the low latitudes in the South Atlantic Ocean as a coherent signal?

2. The numerical model and its pathways

We employ a reduced-gravity, primitive equation OGCM (Gent and Cane, 1989; Murtugudde et al., 1996) coupled to an advective atmospheric mixed layer model (AML, Seager et al., 1995). The model domain covers (50ºN,50ºS)x(100ºW,20ºE) with 20 sigma layers in the vertical and a horizontal resolution of 1/2º. Note that this is not an isopycnal ocean model. The ocean is forced with climatological winds (Hellerman and Rosenstein, 1983) and the AML is forced with climatological precipitation (Oberhuber, 1988), ERBE radiation and ISCCP cloudiness. The coupled model is spun up to steady state with annual mean forcing fields in order to study the physics of temperature anomaly propagation within a stationary circulation (reference run). To obtain a global view of the thermocline pathways, we assumed that turbulent mixing is weak enough that the waters conserve their density and Bernoulli function (hereafter *B*) to a good approximation (away from the surface, the equator, and the boundaries). Based on this assumption, the flow is accurately depicted by the isopycnal projection of horizontal velocity vectors and *B,* the latest defined as

$$B(\sigma) = \frac{1}{2}\rho_0(u^2+v^2) + \rho_0 g\eta + g\int_{z(\sigma)}^{0}\left[\rho - \rho(\sigma)\right]dz$$

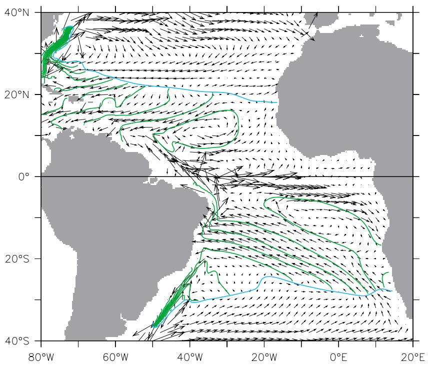

Figure 1. Bernoulli function and currents projected on σ=25.3, except poleward of the outcrop line (blue) where mixed layer currents are plotted.

where ρ, u and v are the density and the horizontal components of the velocity, σ=ρ-1000 is the isopycnal of interest and η the sea surface elevation. The relevant isopycnal range for our purpose is well represented by the σ=25.3 surface, which outcrops close to 30°S, in the heart of the subtropics (Fig. 1). The B isolines divide the southern basin into 3 main regions in agreement with ventilated thermocline theories: from the west to the east along the outcrop line one sees a small recirculation cell, the subtropical gyre ventilation region from 40°W to approximately 0°E, and our region of interest that feeds the Equatorial Undercurrent (EUC). Namely the Benguela Current, that becomes the SEC and, at the coast, the North Brazilian Undercurrent (NBUC). If we suppose that a temperature anomaly in the thermocline follows the mean circulation, it implies that a SST anomaly (SSTA) in the vicinity of 30°S will reach the low latitudes only if it occurs near 10°E. Note that since the simulated mixed layer in this region is characterized by downward Ekman pumping, the subduction of a SSTA can happen slowly without the seasonal cycle of the mixed layer.

3. The SST anomaly experiment.

A synthetic temperature perturbation ΔT is added to the SST field over a circular area of 20º diameter during a two-month

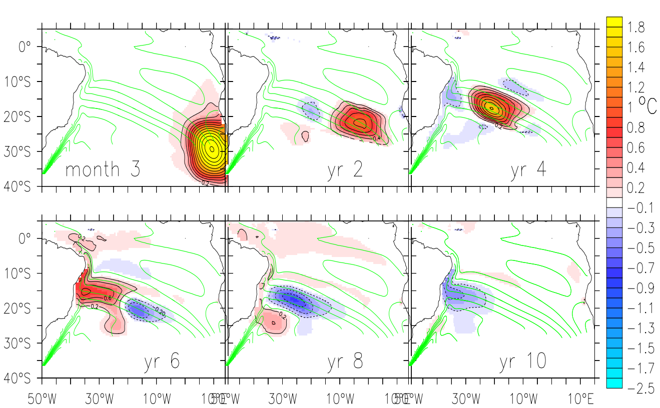

Figure 2. 10-year evolution of the temperature anomaly projected on σ = 25.3 of the reference run, for each other year in December, except for March of the first year; c.i.=0.2ºC. Bernoulli contour in green.

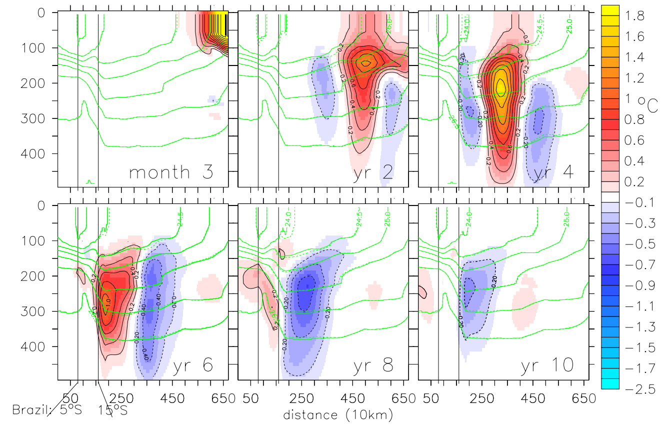

Figure 3. Same as Fig. 2 but in the vertical, along the trajectory of the core of the temperature anomaly. Density stratification of the reference (solid) and anomalous runs (dash) in green; c.i.=0.5 kg.m^{-3}.

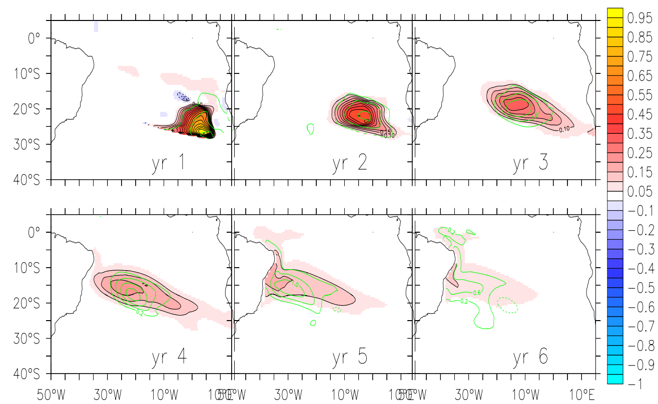

Figure 4. 5-year trajectory of the passive tracer (color) in December along σ = 25.3. Temperature anomaly in green, c.i.=0.04°C.

period t=τ, with the form:

$$\Delta T(x, y, t) \propto \cos[t\pi/2\tau]\cos[(x+10°)\pi/20°]\cos[(y+30°)\pi/20°] \quad (2)$$

Such a large spatial scale is of the order of magnitude of the observations in the Pacific. At the end of τ, the SSTA has reached a maximum amplitude of about +2.2ºC. We then switched off the SSTA and artificially held the SST field constant within the 20º diameter disc for four additional months, after which the SST was allowed to evolve freely. The

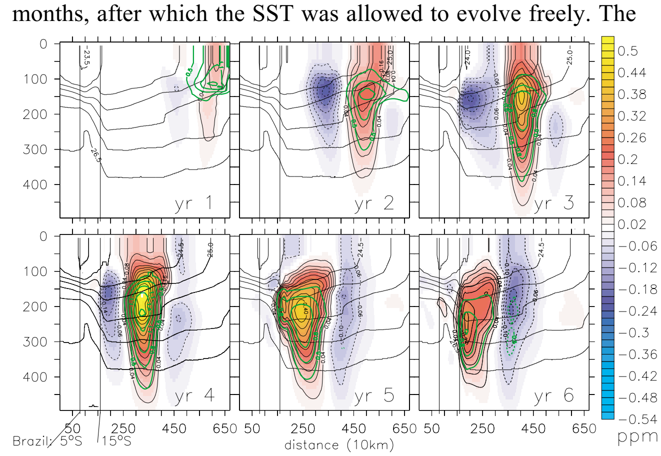

Figure 5. Same as Fig. 3 but for the salinity anomaly (color) over 6 yrs in December; c.i.=0.04 psu. Temperature anomaly in green; c.i.=0.04 °C.

time evolution of the maximum temperature anomaly over 10 years projected onto the isopycnal surface σ=25.3 of the reference run and along its path in the vertical are displayed in Fig. 2 and Fig. 3 respectively. In good agreement with the mean circulation, the pathway is first northwestward (along the Benguela Current and the SEC), then runs northward along the NBUC and finally spreads along the equator in the EUC. During the first year the main part of the warm perturbation subducts trough Ekman pumping and, at the end of year 1, the resulting subsurface anomaly reaches its maximum value of +1.2ºC. Later on, during the first 5 years corresponding to the interior ocean journey, the core of the perturbation moves within a rather constant density range (between σ=25 and 25.5, as seen on Fig. 3), supporting observational and modeling results in the Pacific (e.g., Lysne et al., 1997; Zhang et al., 1998). Its mean velocity is around 3 cm.s^{-1}, a value very close to the intensity of the mean circulation along its path. Meanwhile, the signal spreads in depth over time and, more importantly, its amplitude increases to 1.7ºC until the end of year 4 and decays afterwards. At the same time secondary side anomalies of opposite sign begin to appear as soon as the initial anomaly subducts and intensifies in time as well. From years 6 to 8, the anomaly core follows the Brazilian coast. In this region, the core seems to switch position to a deeper isopycnal range (between σ=26 and 26.5) and keeps a 1ºC intensity. From year 7 to 8, the core travels along isopycnals upward and equatorward within the NBUC and the following opposite sign anomaly reaches its maximum value of 0.7ºC. Meanwhile, a secondary patch has detached, intensified and moved southward in the Brazil Current. From year 9 to 10 the residue of the main core has left the coast towards the east at 5ºS and eventually reaches 0.2ºC (Fig. 3).

4. Physics of the trajectory

In order to compare the trajectory of the anomaly and the characteristics of the mean flow in the ocean interior, we plotted the *B* isolines of the reference run over Fig. 2. It shows that the path is close to the SEC but that the anomaly is slightly shifted southward of the *B* isolines (it reaches the Brazilian coast near 15°S instead of 12°S). To identify further the advective pathway, we released a passive tracer in the reference circulation, at the end of the first year below the mixed layer with a spatial distribution close to the heat anomaly. Fig. 4 shows the 5-year trajectory of the tracer on the σ=25.3 isopycnal at 1-year intervals, along with the distribution of the temperature anomaly. The two trajectories are characterized by pathways and propagation times comparable (we obtained similar results with Lagrangian floats). However the movement of the tracer is slightly faster and displays the small meridional shift seen before (tracer north of the temperature anomaly). These differences are small, and one possible cause, in term of mixing processes, could be densification by cabelling (i.e. moves water deeper and/or poleward).

The main discrepancies are found in the spatial structure of the passive tracer: no side anomalies of opposite sign are observed and the intensity decreases far more rapidly than the temperature anomaly. A last experiment was made to investigate the importance of the sign of the anomaly, where we added to the reference circulation a negative SSTA of similar structure as the positive one: the trajectory, the spatial structure and the time frame are almost identical (not shown). Our results contrast with the comparable study of Liu and Shin (1999) on several points. These authors find the trajectories of the passive and active tracers to depart strongly from each other (propagation speed more than two times larger for the passive tracer, trajectory parallel but significantly shifted equatorward). Furthermore, they observe the active tracer signal to weaken much faster than the passive tracer, as opposed to our simulations.

In the following, we further investigate the similitudes and discrepancies in the shape of the signal between the active and passive tracers. Since, except when approaching the coast, the movement of the maximum heat anomaly has a dominantly isopycnal component and a trajectory close to a passive tracer along an isopycnal, it could be interpreted as dominated by mean flow advection. If this is the case, there should be a mechanism maintaining a low intensity for the density currents perturbation and correspondingly for anomalous currents. First, using the thermal expansion coefficient, a temperature anomaly of +1ºC in the region of interest would produce roughly a +0.27 kg.m^{-3} density anomaly. Considering the size of our anomaly, it would correspond to anomalous density gradients of intensity comparable to the background state. However, we observe a compensating positive salt signal moving along with the temperature perturbation (Fig. 5). Accordingly, the anomalous currents are of the order of a few mm.s^{-1} (with some localized maxima reaching 1 cm.s^{-1}), several times smaller than the local mean circulation (2-3 cm.s-1). The salinity anomaly appears roughly at the same position as the temperature anomaly in the first months of the experiment and its compensating contribution increases in time. At the end of year 3, its core reaches around 0.4 psu and balances more than 75% of the perturbation due to the temperature anomaly (resulting in perturbation of 0.1 kg.m^{-3}). Hence the salinity field plays an important role in the physics of the displacement by compensating the temperature-related density anomaly. An additional experiment using a constant surface salt flux exhibits the same phenomenon (not shown), hence it is not related to air-sea interaction but to oceanic adjustments which physics we are currently studying.

Several aspects of the displacement tend to indicate that there are undulatory phenomena at play. First, the increase in intensity of the heat anomaly core occurring from year 2 to 5 resembles amplitude increase of a wave signal through interactions between different vertical modes (Huang and Pedlosky; 1998; Liu, 1999) or/and through instability in strong westward shear flow (Liu, 1999). Second, the rapid and notable vertical extension of the heat and salt anomalies (Fig 3 and 5, yr 2,3) as well as the switch to a deeper isopycnal is similarly suggestive of downward wave propagation. At last, we observe a clear anomaly over the whole water column propagating from the initial location essentially westward in the interior basin (around 25ºS) and southward along the Brazilian coast (not shown). It can be interpreted as a first mode baroclinic Rossby wave (interior basin speed of 8 cm.s^{-1} in agreement with observations of Qiu et al., 1997) creating a signal advected southward by the Brazilian current at the coast.

5. Conclusions

We performed a modeling study of the displacement of a subducted temperature anomaly in the South Atlantic moving from the eastern subtropics towards the western tropics and the Equator. The trajectory and speed of its core are in agreement with the mean flow to a first approximation, a result consistent with previous studies in the Pacific (Schneider et al., 1999a,b). Our analysis shows that the passive tracer-like characteristics of the trajectory are associated with a salinity-compensation phenomenon that balances strongly the temperature-related density perturbation and corresponds to weak anomalous velocities. Hence a greater effort should be made to take into account the effect of salinity in related model and observation studies.

At the same time, our understanding of some characteristics of the phenomenon is enhanced from a wave perspective. Most important are the remarkable increase of the intensity of the signal with time and its rapid and large vertical extension. Therefore, while we conclude that the subsurface displacement of the maximum temperature anomaly can be predicted to a good approximation with mean flow advection arguments, we suggest that a combination of advection and wave arguments is needed to fully assess temperature anomaly displacement in oceanic basins. The propagation of waves in a mean current should be further studied (like Liu, 1999) as well as the factors controlling the importance of one mechanism relative to the other.

Finally, whereas the Schneider et al. (1999a,b) study indicates no significant subtropics/tropics teleconnections via heat anomalies advection, our work suggest that such a process might be relevant in the Atlantic. By emphasizing the existence of equatorward advective pathways for tracers, it also corroborates early tritium studies (e.g.; Fine et al., 1987). At last, it supports studies (e.g.; Gu and Philander, 1997) that underlined the importance of inferring subsurface decadal displacement of subtropical heat anomalies from the knowledge of their initial location and the time-mean flow. This is an essential issue in the context of prediction efforts of decadal scale variability of the tropical oceans.

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