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# ATMOSPHERIC FORCING AS A SOURCE FOR OCEAN VARIABILITY AT THE BRAZIL-MALVINAS CONFLUENCE ZONE

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## ABSTRACT

The goal of this study is to evaluate the impact of the atmospheric forcing of the scale of the storms on the oceanic circulation and temperature at the Brazil-Malvinas Confluence Zone. SST data and numerical simulations are used to get some insight on the mechanisms responsible for the establishment of the oceanic temperature and velocity fields response to the wind variability and to analyze the penetration depth, the time and the persistence of this response. Results show that atmospheric storms can be very effective on generating a SST persistent perturbation in the area and given their high frequency they can generate an important effect on the variability of the SST and its mean. Two possible mechanisms are considered for the oceanic response, the wind stress forcing and the heat exchange through the ocean-atmosphere interface. The latter probes to be the most important one. Even though the wind stress forcing has a rol on modifying the oceanic characteristics by means of advection, it is a slow process that will only be effective if the atmospheric perturbation remains for long periods. In contrast, the heat flux through the ocean surface is much more rapid and effective, particularly in cooling the ocean. The cooling of the ocean surface produces static instability and convection that rapidly transfers the cooling to lower layers. Vertical mixing relatively heats the ocean surface allowing for a new cooling by means of interaction with the atmosphere. Results show that even though most of the energy is lost as inertia-gravity waves, it takes to the ocean 24 hours to develop a geostrophically balanced response and it is proportional to the persistence of the atmospheric event. Once the response is setup into the ocean, it tends to remain even though the atmospheric disturbance disappears, because oceanic dissipative effects are of a much longer temporal scale than the atmospheric ones. The scale of penetration results to be of the order of 200 meters. Therefore, in a broad and shallow region such as the Argentinean continental shelf these events might be able to produce the mixing of the whole water column. Atmospheric storms in the area have a mean annual frequency higher than once a week and the signal is present in the ocean for a longer period. The results suggest that atmospheric variability on the scale of the storms can be very important on determining the local SST.
**Keywords:** Brazil-Malvinas Confluence – SST variability – Ocean-atmosphere interaction

## RESUMEN

El objetivo de este trabajo es evaluar el impacto del forzante atmosférico en la escala de las tormentas sobre la circulación y la temperatura del océano en la región de la Confluencia Brasil-Malvinas. Se utilizan datos de temperatura de la superficie del mar (TSM) y simulaciones numéricas con el objetivo de estudiar los mecanismos responsables del establecimiento de la respuesta de los campos de temperatura y velocidad en el océano a la perturbación atmosférica y analizar la profundidad de penetración, el tiempo y la persistencia de esta respuesta. Los resultados indican que las tormentas atmosféricas pueden ser muy efectivas en la generación de una perturbación persistente de la TSM en el área y, dada su alta frecuencia, pueden tener un efecto importante en la variabilidad de la TSM y su media. Se consideran dos mecanismos posibles en la generación de la respuesta, el forzante del viento y el intercambio de calor a través de la interfase mar-atmósfera. El último resulta ser el más importante. Aunque el esfuerzo del viento modifica el océano por advección, éste es un proceso muy lento, que sólo puede ser efectivo si la perturbación atmosférica permanece por períodos largos. En contraste, el flujo de calor a través de la superficie del mar es un mecanismo mucho más rápido y efectivo, especialmente enfriando el océano. Un enfriamiento de la superficie del mar produce inestabilidad estática y convección, que rápidamente transfiere el enfriamiento a capas más profundas. La mezcla vertical a su vez calienta relativamente la superficie del océano, facilitando un nuevo enfriamiento por interacción mar-atmósfera. Los resultados muestran que, aunque gran parte de la energía se pierde como ondas grávito-inerciales, en 24 horas el océano desarrolla una respuesta balanceada geostróficamente y que esta respuesta es proporcional a la persistencia del evento atmosférico. Una vez que la respuesta oceánica se establece, tiende a permanecer aunque la perturbación atmosférica desaparezca, porque los efectos disipativos tienen una escala temporal mucho más larga en el océano que en la atmósfera. La escala de penetración resulta ser del orden de 200 metros. Por lo tanto, en una región ancha y somera como la Plataforma Continental Argentina, estos eventos podrían producir la mezcla de toda la columna de agua. Las tormentas atmosféricas en el área tienen una frecuencia anual mayor a un evento semanal y la señal permanece en el océano durante un tiempo aún más largo. Los resultados sugieren que la variabilidad atmosférica en la escala de las tormentas puede tener un gran impacto en la TSM local.

**Palabras clave:** Confluencia Brasil-Malvinas – Variabilidad de la SST – Interacción mar-atmósfera

## 1. INTRODUCTION

A well-known feature of the ocean surface in the Southwestern Atlantic is the large small-scale variability observed on the velocities and temperature fields. For long this kind of variability was thought to be a consequence of internal processes by means of instabilities. Nevertheless, more recently it has been recognized the role of the atmospheric forcing in the production of long-lasting features in the ocean, not only on the planetary scale but also on smaller scales. Theoretical results (Orlanski and Polinski, 1983) indicate that the temporal scale required for an atmospheric storm to be effective in generating a permanent, geostrophically balanced response in the ocean is of the order of only 1 day. The response is proportional to the product between the intensity of the storm and its temporal scale. For storms with scales smaller than the (barotropic) Rossby radius of deformation, the factors that determine the penetration of the response are the stratification and the spatial scale of the atmospheric feature: the larger the stratification and the shorter the storm scale, the smaller the penetration. Once the oceanic response is established, it remains for a long time even though the event that originates the atmospheric forcing has dissipated, because the scales of atmospheric and oceanic dissipation are different.

Krauss (1981), Large and Crawford (1995) and Crawford and Large (1996) suggested that certain storms generate large inertial oscillations that may produce sudden upper layer cooling events. They found that even though the events occur as a response to the storms, they are not always related to intense winds and they may even take place when the heat flux is from the atmosphere to the ocean. They found that those cooling events in the mixed layer are related to a downward heat flux to the lower layer favored by the phase change with depth of inertial currents.

The southwestern Atlantic transition zone between the area of subtropical influence (Brazil Current with warm and salty waters) and the area of polar influence (Malvinas Current with cold and fresher waters) is the so-called Brazil-Malvinas Confluence. It is characterized by one of the largest temperature gradients in the world (Legeckis and Gordon, 1982) and also by a high variability over a wide range of spatial and temporal scales (Legeckis and Gordon, 1983; Olson et al., 1988; Garzoli and Garraffo, 1989; Garzoli and Simionato, 1990; Provost et al., 1992). The current system present in the area exhibits an intense seasonal cycle (Garzoli and Giulivi, 1993; Podestá et al., 1991; Provost et al., 1992, Matano et al., 1993), a semiannual signal (Olson et al., 1988; Provost and Le Traon, 1993) and high variability at higher frequencies (Garzoli and Simionato, 1990; Olson et al., 1988; Garzoli and Giulivi, 1993; Rivas and Piola, 2001). Part of this high frequency variability has been related to ocean-atmosphere interaction processes associated to variability of the wind field (Garzoli and Giulivi, 1993) and the heat flux through the ocean surface (Rivas and Piola, 2001).

This oceanic transition zone has its continental counterpart in the area located between 34º S and 38º S, over southern South America (Fernández, 2000a). Meridional humidity fluxes from the north and south converge over this region, and the zonal eastward fluxes are maximum at 38º S, transporting humidity to the ocean (Fernández, 1998; Fernández, 2000b). The area between 30°S and 45°S is characterized by one of the highest cyclogenetic activity within the Southern Hemisphere (Necco, 1982; Sinclair, 1994; Gan and Rao, 1991). Cyclones are generated in the Patagonic region and in the Argentinean Littoral (Eastern South America). Cyclogenesis on the area have a mean frequency of around 120 events by year (Gan and Rao, 1991), with higher frequency during winter and spring. Even though this kind of storms have a very explosive nature and their life time is relatively short, during these events winds easily reach 35 knots and the spatial scale of the storms is of several hundred of kilometers. As the storms develop, air is advected to the area from far regions and a strong and a time varying temperature gradient is established between the ocean and the atmosphere allowing intense vertical heat and momentum exchanges. On the scope of the formerly cited theoretical and observational results (Orlanski and Polinski, 1983; Krauss, 1981; Large and Crawford, 1995; Crawford and Large, 1996), one can expect that this kind of storms will introduce a strong signal into the ocean.

Understanding the role of these atmospheric features on the generation of an oceanic persistent response at the area of the Brazil-Malvinas Confluence is important. Due to their high frequency, they may able to introduce a persistent signal into the ocean and therefore they can constitute an important effect to explain part of the variability and its mean. Moreover, in spite of their frequency, these events are often filtered out in the computation of the wind stress when it is done by computing the wind average and then the wind stress (as done for example by Hellerman and

Rosenstein, 1983). Even though the littoral effects of this kind of storms are pretty well known, as they cause a sudden lowering on the air temperature, occasional floods over the Río de la Plata estuary area, etc., almost nothing is known about their effects on deeper waters.

The goal of this study is to evaluate the impact of the atmospheric forcing in the scale of the storms on the oceanic regional circulation and temperature. SST data and numerical simulations are used to get some insight on the mechanisms responsible for the establishment of the oceanic temperature and velocity fields response to the atmospheric perturbation and to analyze the penetration depth, the time and the persistence of this response.

This article is organized as follows. In section 2, SST data are used to determine the oceanic sea surface temperature perturbation in situations in which it is known that atmospheric storms were present and to quantify the magnitude of the SST variation that can be associated to the presence of the storms. An observation of the data suggests the mechanisms responsible for such a response. In section 3 a numerical case study that reproduces a particular situation occurred on November 1989 is described. The mechanisms responsible for the response are analyzed by means of sensitivity studies. In section 4 the persistence, time and penetration depth of the response are analyzed. Finally in section 5 results are summarized and main conclusions are drawn.

## 2. DATA

Data were used to observe the oceanic surface response on situations of intense Argentinean Littoral cyclogenesis over South American coast along several years identified by Seluchi (1995). Emphasis has been put in this kind of cyclogenesis because they have been well studied besides several cases along the years have been well documented.

These storms, that have a very explosive nature and that are strongly influenced by the presence of the Cordillera de Los Andes (Seluchi and Saulo, 1996), develop around 30° S. They are usually associated to a high pressure center that develops southernmost, over the Argentinean Continental Shelf, that increases the Southeastern winds over the Río de la Plata estuary region.

A typical intense situation occurred on November 11 to 12, 1989 is shown in figures 1b-e. Data derived from ECMWF analysis every 12 hours display the evolution of wind stress associated to the cyclone. Figure 1a shows the mean wind stress for November 1989. Vectors have been kept on the same scale to emphasize the intensity of the storm and to illustrate how strongly they are filtered out when computing the average. The main feature of the climatological winds at the Southwestern Atlantic is the presence of an intense zonal circulation southern 40º S in summer and 35º S in winter, related to the Westerlies (Hoffmann, et al., 1997). Northward these latitudes the mean atmospheric circulation is dominated by the South Atlantic Subtropical Anticyclone, with smaller wind velocities (Hoffmann, et al., 1997). Figure 1 clearly shows that the instantaneous wind field is dominated by the passage of storms. When the average is computed the effect of this variability is lost and the mean values result very small except at the region of the Westerlies (figure 1a).

It can be seen from figure 1 that when this kind of storms develop over the area they introduce cyclonic vorticity into the ocean northward the Río de la Plata estuary and anticyclonic vorticity southward from it. Simultaneously, cold air is rapidly advected northward, to the Río de la Plata region, and hot air is advected southward, to the same area. This produces a sudden change in the air temperature. The temperature variation observed from November 11 to November 12, 1989, as the situation shown in figures 1b-e developed, is shown in figure 2.

Gan and Rao (1991) analyzed the frequency of occurrence of these kind of atmospheric features over the area from ten years of data. They conclude that the events have a frequency of around 60 by year with higher frequency during winter and spring. These results are in good agreement with a study by Necco (1982) based on a smaller amount of data. This frequency implies an annual average higher than one event a week. If these storms introduce into the ocean a persistent signal, their effect can be very important on determining the variability and means of the oceanic variables in the region.

The southwestern Atlantic has a relatively poor observational coverage besides satellite observations. Because of the cloud cover associated to cyclones it is not possible to use satellite data to observe the situation in the ocean *during* these events. Therefore, the study of weekly analyzed data blended from bias corrected satellite, drifters and ships observations (IGOSS Data Set, Reynolds and Smith, 1994) with 1 degree resolution for the region was performed. Even though this data will not allow tracking the oceanic situation during the storm development, if it introduces a persistent signal into the ocean, it should be seen as a change in the weekly SST ensemble.

Data show that when a cyclone develops over the area, a variation in the oceanic surface temperature higher than 0.5° is observed in a week, suggesting that storms have an important and persistent effect on the oceanic variability. The nature of the SST perturbation that can be associated to atmospheric forcing seems to depend on the position of the cyclone, the thermal gradient in the area and the relative position between the oceanic front and the cyclone, not been possible to establish an exact pattern for the oceanic temperature response. Even though, in most of the cases a SST decrease is observed southward the atmospheric cyclone over the Malvinas Current and the Argentinean Continental Shelf waters.

One of those situations, corresponding to the storm of November 10-12, 1989 is shown in figure 3 displaying the SST difference between the week when the cyclone developed and the previous one. The atmospheric storm (figure 1) was located around 32° S, over the coast. A SST decrease of the order of 0.6°C is observed south of 35° S, in the area of anticyclonic atmospheric vorticity and a temperature increase of the order of 1.2°C occurs in the Brazil Current region, in the area of cyclonic atmospheric vorticity.

The temperature increase observed northernmost 35° S can be associated to the seasonal cycle, meanwhile the temperature decrease observed southward that latitude is in clear opposition with the seasonal variation for the area (figure 4). Figure 4 shows the climatological SST change for November from the IGOSS Data Set. During that month, the end of the austral spring, the whole system of currents is moving southward (Olson et al., 1988; Garzoli and Garraffo, 1989; Provost et al., 1992; Matano et al., 1993) and the area is being heated. The maximum heating occurs over the Río de la Plata estuary with values of around 2.7°C. This variation corresponds to a SST weekly change mean value of around 0.6°C. When comparing figures 3 and 4 it is found a perturbation in the ocean caused by the presence of the cyclone whose value is similar but whose sign is opposite to the seasonal change for the area.

The characteristics observed in figure 3, both heating and cooling, can be qualitatively associated to the presence of the cyclone (figure 1) as follows. Wind stress would produce advection of warm waters from the north (Brazil Current waters) by introducing cyclonic vorticity into the ocean, and advection of cold waters from the south (Malvinas Current and Argentinean Shelf waters) by introducing anticyclonic vorticity. Even though this sounds attractive, these cyclones have a short lifetime, of approximately 48 hours, that makes difficult to assess that the advection caused by momentum exchange can be responsible for the observed oceanic perturbation. There must be another mechanism able to produce such a fast oceanic response. It is interesting to note that the pattern of the air temperature perturbation associated to the cyclone (figure 2) resembles in a high extent the pattern of the observed oceanic SST variation (figure 3) related to the cyclone. This suggests that it can be the heat exchange between the ocean and the atmosphere the main mechanism that causes the oceanic perturbation. Heat flux can have a very rapid effect, particularly in cooling the ocean. A cooling of the ocean surface, even if small, will cause instantaneous vertical convection and mixing, that will transfer the signal deeper into the ocean and will relatively heat the surface allowing for a new cooling and convection. Heating, in turn, would be much slower and it would penetrate less into the ocean. Frankignoul and Reynolds (1983) used meteorological data to estimate the relative importance of the different mechanisms of SST anomaly generation in mid latitudes North Pacific Ocean. They considered forcing due to heat flux and Ekman advection anomalies and found that heat flux anomalies play a more important role than advection by anomalous Ekman currents. The importance of the heat flux on producing SST anomalies as a response to storms was also noted from models and data analysis by Krauss (1981), Large and Crawford (1995) and Crawford and Large (1996). They concluded that the temperature anomalies can result through downward heat flux even when winds are small. Similarly, analyzing two temperature time series collected off the northern Patagonia, at 43º S, Rivas and Piola (2001) found that frequent episodic mixed layer cooling events are produced mainly by sudden changes in the surface heat flux associated to northward penetrations of subpolar air.

The fact that the storm is responsible for the observed SST variation and the possible mechanisms for this response will be explored in the next section by means of numerical simulations.

## 3. NUMERICAL SIMULATION

A case study was performed to test the effect of a cyclone over the ocean waters. The cyclogenesis situation occurred on November 10 to 12, 1989, whose effect on the ocean was shown in Figure 3, was arbitrarily chosen for the analysis.

A version of the non-linear, primitive equation Bryan-Cox/GFDL (Cox, 1984) model in which simple open boundary conditions have been introduced to allow for regional modeling was used. The model solves the primitive equations in spherical coordinates with two basic assumptions: first, rigid lid approximation that filters external waves is used to reduce the timestep of the model; second, Boussinesq approximation is used. Boundary conditions have been introduced as follows. First, an open boundary condition on mass transport stream function was set; if model is initialized with a mass transport stream function that crosses the boundaries, it will give a solution that keeps the initial flux through the boundaries. The scheme was complemented by introducing sponge layers that force the values of the other variables to time varying specified values, either coming from observations or other numerical models.

A model for South Atlantic region was run to obtain an initial condition that reproduces the most relevant characteristics of the observed situation during the week previous to the cyclone development. The model domain, shown in figure 5, spans from 73.75° W to 23.75° E and from 66.25° S to 13.75° S, covering an area much larger than the one of interest to avoid boundary conditions influence the results. The model resolution is 2.5°, being the points coincident with ECMWF analysis to avoid the need for wind and air temperature data horizontal interpolation. The model has a realistic topography and 15 vertical levels (Table I) with higher resolution on the upper thermocline. Topography is shown in figure 5 and the model levels in Table I. Figure 5 shows the broad Argentinean Continental Shelf.

The model was initialized with the Levitus and Boyer (1994) temperatures and salinities; these values were preserved along the integration on the eastern and western boundaries through the Newtonian dumping terms of the sponge layers. Mass transport through those boundaries was fixed to 160 Sv. (1 Sv = $10^6$ $m^3s^{-1}$); this way, the model has a transport by Antarctic Circumpolar Current which is consistent with observations. A symmetry boundary condition was imposed in the northern boundary; even though model will not resolve well the deep circulation this is not a serious limitation for this study in which only the surface circulation is relevant.

The model was forced with wind stresses computed from ECMWF mean November 1989 winds; the wind speed values were converted to wind stresses by a simple approach suggested by Hellerman (1967) in which the drag coefficient is a function of the wind velocity and the air temperature. To include into the model the effect of high frequency wind variability, wind stress was first computed from ECMWF data and then averaged along the month.

The ocean-atmosphere heat exchange in the oceanic mixed layer is simulated by means of Newtonian terms that force oceanic temperatures in top of the model to observed satellite SST from IGOSS data with a relaxation time of 30 days. This heat is then distributed into the mixed layer, whose depth is 30 meters, giving a heat flux of 10 langleys day$^{-1}$. The model was run for a time equivalent to 30 years, after the solution reaches the steady state. As the model has been initialized with observational data, the main thermocline is already present in the initialization and it takes to the model a relatively short time to adjust to a steady state.

Results of this simulation are shown in figure 6, that exhibits the mass transport stream function (figure 6a) and the sea surface temperature (figure 6b) at the steady state for the region of interest. Model reproduces reasonably well the main features of the oceanic surface circulation in Southwestern Atlantic area; the Brazil-Malvinas Confluence is located around 40° S, in good agreement with observations for that time of the year. The sea surface temperature (figure 6b) reproduces the data observed on the week previous to the cyclone development due to the upper boundary condition imposed to the model.

Starting from this solution, the model is run for a time equivalent to 48 hours, forced with ECMWF winds and air temperatures for November 10-12, 1989 varying every 12 hours (figure 1b-e). Wind speed was converted to wind stress by the same technique explained before and ocean-atmosphere heat exchange is prescribed to 10 langleys day$^{-1}$.

Results of the numerical simulation are shown in figure 7, which presents the SST change obtained from the model in the 48 hours. The figure also shows that besides the low resolution model succeeds on representing qualitatively and quantitatively well the pattern (both, position and intensity) of the temperature reduction observed in the data (figure 3). A SST decrease southern of 35° S is obtained, with a maximum of 0.5°C at approximately 39° S. The temperature increase shown by the data northernmost 36º S is almost absent in the model (that only produces a change of 0.2°C), suggesting that other effects may be significant and they were not included on the simulation. As suggested by the analysis of the climatological seasonal SST variation of previous section (figure 4), it is probably related to the seasonal cycle.

The question that remains to answer is to define the mechanism that allows such a rapid response of the oceanic variables to the atmospheric perturbation. The air temperature change in the modeled 48 hours of section 2 (figure 2) exhibits a very good qualitative resemblance with the SST change observed in model results for that period (figure 7). This suggests that the heat flux through the ocean surface can be the principal responsible for the SST observed anomaly. Figure 8 shows the time evolution of the different terms of the equation for temperature time derivative (Cox, 1984) at the point of maximum variation of SST observed on model results. This equation is:

$$\frac{\partial T}{\partial t} = -\frac{\sec\varphi}{a}\left[\frac{\partial uT}{\partial\lambda} + \frac{\partial vT}{\sec\varphi\partial\varphi}\right] - \frac{\partial wT}{\partial z} + \frac{A_{TH}}{a^2}\nabla^2 T + \frac{\partial}{\partial z}\left(\frac{A_{TV}}{\delta}\frac{\partial T}{\partial z}\right) + HF \qquad (1)$$

where $T$ is the temperature, $(u,v,w)$ is the velocity vector, $\lambda$ and $\varphi$ are the longitude and latitude, $a$ is the radius of the Earth, $A_{TV}$ and $A_{TH}$ are the eddy vertical and horizontal diffusion coefficients, $HF$ represents the heat flux through the ocean surface and

$$\delta = \begin{cases} 1 \xrightarrow{if} \dfrac{\partial^2 \rho}{\partial z^2} < 0 \\ 0 \xrightarrow{if} \dfrac{\partial^2 \rho}{\partial z^2} > 0 \end{cases} \tag{2}$$

accounts for the vertical convection in the case of static instability conditions.

This way, the five terms represent respectively: a) the zonal flux divergence; b) the meridional flux divergence; c) the vertical flux divergence; d) the horizontal diffusion; e) the vertical diffusion and f) heat flux through the ocean surface. Results (figure 8) indicate that diffusion terms (figures 8 d and e) are of a lower order of magnitude than the other terms and divergence terms (figures 8 a to c) compensate each other, being the heat flux term (figure 8 f) the one that determines the value of the resulting temperature.

To further explore that possibility, two more cases were studied with the model. In the first one the model surface is forced with de ECMWF winds but heat flux between the ocean and the atmosphere is eliminated. In the second one heat exchange is allowed in the simulation but wind forcing is suppressed. Results are shown in figure 9a (in which wind forcing has been eliminated) and figure 9b (in which heat flux has been suppressed). Both figures display the SST change reproduced by the model after 48 hours of simulation. Figure 9a reproduces the main features observed on the complete simulation (figure 7), meanwhile in figure 9b the oceanic response is almost absent. It can be seen from the figures that even though wind stress forcing has a role in advecting warm waters from Brazil Current to the South (figure 9b), most of the SST change produced during the cyclone evolution is due to the ocean-atmosphere heat exchange. This emerges very clearly when comparing figure 9a with figure 7. This mechanism is much more effective for cooling the area south of 35° S than for heating it northernmost. As discussed in section 2, cooling is rapidly transferred to lower layers through vertical mixing due to static instability which in turn prepares the upper layers for a new and fast cooling.

## 4. THE PERSISTENCE OF THE RESPONSE

In previous sections it has been shown that storms can be effective on generating a rapid response from the ocean in the Southwestern Atlantic area, and the mechanisms for such a response were discussed. An obvious question now is how long the oceanic forced perturbation can remain in the ocean and how deep is its penetration. These questions will be explored in this section.

To improve the details of the solution, a higher resolution (0.375°) nested model is run in this case. Model is initialized from the steady state for average November 1989 (Section 3) and boundary conditions are fixed through Newtonian dumping terms to that solution. Model is then run until it reaches the steady state to get an initial condition to be forced with a cyclone.

For this analysis model is forced with a storm that is generated and decays in the same place in a matter of a few days; to do that, the wind stress forcing is chosen to be a separate function of space and time: $\tau(x,y,t) = \tau_0(x,y)\ T(t)$, and an analogous concept is applied to the heat flux through the ocean surface: $\theta(x,y,t) = \theta(x,y)\ T(t)$. $\tau_0(x,y)$ and $\theta_0(x,y)$ are chosen to be the anomalies respect to the mean November 1989 state introduced by the storm of November 10-12, 1989 at its maximum development. High resolution wind velocity and air temperature were provided by Seluchi and Saulo (1996), simulated with a high resolution atmospheric model, and $T(t)$ is written as:

$$T(t) = [\tanh \alpha t - \tanh \beta(t-\Delta t)H(t-\Delta t)] \qquad (3)$$

where $\alpha$ and $\beta$ are the generation and decay rates, respectively, $\Delta t$ is the duration time of the storm and $H$ is the Heaviside function. The shape of $T(t)$ will depend on the election of the parameters $\alpha$, $\beta$ and $\Delta t$. Figure 10 displays the shape of $T(t)$ vs. time chosen for our simulation: the storm develops in a period of two days and then it decays quickly. This way, even when reproducing the most important features of the storm, it is possible to isolate the effect of the wind stress and heat flux through the ocean surface from the displacement of the cyclone and other effects that would be present on data if many days were used.

The results of this experiment are shown in figures 11 and 12. Figure 11 exhibits the temperature change every 24 hours (referred to the initial state) in a zonal section across the maximum of the SST response to the atmospheric perturbation. Figures 12a and c show the meridional and zonal velocity fields change (referred to the initial state) at the latitude-longitude point of maximum SST reduction at level 1 of the model. Several different depths in the model are presented in figures 12b and d. Results are in good agreement with the theoretical ones by Orlanski and Polinski (1983). Even though most of the energy is lost as inertia-gravity waves (figures 12a and c), it takes to the ocean 24 hours to develop a geostrophically balanced response and this response is proportional to the duration of the atmospheric event. Once the perturbation is setup in the ocean, it tends to remain even though the atmospheric disturbance disappears, because oceanic dissipative effects are of a much longer temporal scale than the atmospheric ones. The scale of penetration is variable, probably affected by stratification (Orlanski and Polinski, 1983), but near the coast it reaches the 200 meters. Therefore, in a broad and shallow region such as the Argentinean continental shelf these events might be able to produce the mixing of the whole water column.

## 5. CONCLUSIONS

Our results indicate that atmospheric storms can be very effective on generating a SST persistent perturbation in the area, and given their frequency they can have an important effect on the mean SST values and on its variability. The oceanic response to the atmospheric perturbation is in very good agreement with Orlanski and Polinski (1983) theoretical results. Of the two possible mechanisms for the oceanic response, the wind stress forcing and the heat exchange through the ocean-atmosphere interface, the second one probes to be the most important one. This last result is in good agreement with Frankignoul and Reynolds (1983) research on mid latitudes North Pacific Ocean, with Rivas and Piola (2001) analysis at the Northern Patagonia, and with Krauss (1981), Large and Crawford (1995) and Crawford and Large (1996) models and data derived results. Even though the wind stress forcing has a rol on modifying the oceanic variables characteristics by means of advection, it is a slow process that will only be effective if the atmospheric perturbation remains for long periods. In contrast, the heat flux through the ocean surface is much more rapid and effective, particularly in cooling the ocean. It is the case in Littoral Cyclogenesis over the Southwestern Atlantic, because a cooling of the ocean surface will produce static instability and convection that rapidly transfers the response to lower layers. I will, in turn, generate vertical mixing that relatively heats the ocean surface allowing for a new cooling by means of interaction with the atmosphere.

Results show that even though most of the energy is lost as inertia-gravity waves, it takes to the ocean 24 hours to develop a geostrophically balanced response and it is proportional to the duration of the atmospheric event. Once the response is setup into the ocean, it tends to remain even though the atmospheric disturbance disappears, because oceanic dissipative effects are of a much longer temporal scale than the atmospheric ones. The scale of penetration results to be of the order of 200 meters near the coast in our simulation. So, in a broad and shallow region such as the Argentinean continental shelf these events might be able to produce the mixing of the whole water column.

Argentinean Littoral cyclogenesis have a mean annual frequency higher than once a week and the response signal is present in the ocean for a longer period. Patagonic region cyclogenesis has similar frequency and intensity and, as suggested by Rivas and Piola (2001) results, probably contributes to the local variability in a similar way. The results of the present study suggest that atmospheric variability in the scale of the storms can be very important on determining the local oceanic variability and then the means, and it should be included on the estimation of the wind stress and the heat fluxes through the ocean surface whenever they are used to determine the oceanic response.

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## REFERENCES

Cox, M., 1984. A primitive equation, 3-dimensional model of the ocean. GFDL Ocean Group Technical Report N°1. August 30, 1984. Princeton University.

Crawford, G. B. and W. G. Large, 1996. A numerical investigation of resonant inertial response of the ocean to wind forcing. J. Phys. Oceanog., 26(6), 873-891.
Fernández, A, 1998. Variación anual de los transportes de vapor de agua en áreas argentinas. CD del X Congreso Brasileño de Meteorología y VIII Congreso de FUSMET. SBMET.
Fernández, A., 2000a. Water vapor circulation: Spatial and annual variations over Argentina. Preprints 6th International Conference on Southern Hemisphere Meteorlogy and Oceanography. American Meteorological Society.
Fernández, A., 2000b. Vapor de agua: distribución espacial sobre Argentina. CD de la XX Reunión Científica de Geofísica y Geodesia. AAGG, 2000.
Frankignoul, C. and R. W. Reynolds, 1983. Testing a dynamical model for midlatitude sea surface temperature anomalies. J. Phys. Oceanogr., 13, 1131-1145.
Gan, M. A. and V. B. Rao, 1991. Surface cyclogenesis over South America. Mon. Wea. Rev. 119(5), 1293-1302.
Garzoli, S. L. and Z. D. Garraffo, 1989. Transports, frontal motions and eddies at the Brazil-Malvinas currents confluence. D. Sea Res. 36, 681-703.
Garzoli, S. L. and C. Giulivi, 1993. What forces the variability of the southwestern Atlantic Boundary Currents?. D. Sea Res., 41(10), 1527-1550.
Garzoli, S. L. and C. G. Simionato, 1990. Baroclinic Instabilities and forced oscillations at the Brazil-Malvinas Confluence Front. D. Sea Res., 37(6A), 1053-1074.
Hellerman, S., 1967. An updated estimate of the wind stress in the World Ocean. Mon. Wea. Rev. 95, 607-626.
Hellerman, S. and M. Rosenstein, 1983. Normal monthly wind stress over the World Ocean with error estimates. J. Phys. Oceanogr. 13, 1093-1104.
Hoffmann, J., M. Nuñez y M. C. Piccolo, 1997. Características climáticas del Atlántico Sudoccidental. En "El Mar Argentino y sus recursos pesqueros. Instituto Nacional de Investigación y Desarrollo Pesquero", Secretaría de Agricultura, Ganadería, Pesca y Alimentación, Mar del Plata, Argentina, 1997. 1, 163-193.
Krauss, W., 1981. The erosion of a thermocline. J. Phys. Oceanogr., 11(4), 415-433.
Large, W. G. and G. B. Crawford, 1995. Observations and simulations of upper-ocean response to wind events during the ocean Storms Experiment. J. Phys. Oceanogr., 25(11), 2831-2852.
Legeckis, R. and A. L. Gordon, 1982. Satellite observations of the Brazil and Falkland Currents. D. Sea Res., 29(3A), 375-401.
Levitus, S.. and T. P. Boyer, 1994. World Ocean Atlas 1994. NOAA Atlas U.S. De-partment of Commerce, Washington, D.C. 117.
Matano, R. P., M. G. Schlax and D. B. Chelton, 1993. Seasonal Variability in the southwestern Atlantic. J. Gephys. Res., 98, 18027-18035.
Necco, G. 1982. Comportamiento de vórtices ciclónicos en el área sudamericana durante FGGE: Ciclogénesis. Meteorológica, 13, 7-20.
Olson, D. B., G. P. Podesta, R. H. Evans and O. B. Brown, 1988. Temporal variations in the separation of Brazil and Falkland Currents. D. Sea Res., 35(12), 1971-1990.
Orlanski, I. and L. J. Polinski, 1983. Ocean response to mesoscale atmospheric forcing. Tellus, 35 A, 296-323.
Podesta, G. P., O. B. Brown and R. H. Evans, 1991. The annual cycle of satellite-derived sea surface temperature in the southwestern Atlantic Ocean. J. Climate, 4, 457-467.
Provost, C., O. Garcia and V. Garcon, 1992. Analysis of satellite sea surface tempera-ture time series in the Brazil-Malvinas Currents Confluence Region. J. Geophys. Res., 97, 17841-17858.

Provost, C. and P. Le Traon, 1993. Spatial and temporal scales in altimetric variability in the Brazil-Malvinas Currents Confluence region: dominance of the annual and semian-nual periods. J. Geophys. Res., 98, 18037-18052.
Reynolds, R. W. and T. M. Smith, 1994. Improved global sea surface temperature analyses. J. Climate, 7, 929-948.
Rivas, A. L. and A. R. Piola, 2001. Vertical stratification at the shelf off northern Patagonia. Cont. Shelf Res. In press.
Seluchi, M. E., 1995. Diagnóstico y pronóstico de situaciones sinópticas conducentes a ciclogénesis sobre el este de América. Geofisica Internacional, 34, 171-200.
Seluchi, M. E. and A. C. Saulo, 1996. Possible mechanisms .yielding an explosive coastal cyclogenesis over South America: experiments using a limited area model. Australian Meteorological Magazine, 47(4), 309-320.
Sinclair, M.R., 1994. An objective cyclone climatology for the southern hemisphere. Mon. Wea. Rev., 122, 2239-2256.

## FIGURE CAPTIONS

**Figure 1: Wind stress vectors computed from ECMWF wind analysis corresponding to a) November, 1989 average and b) through e) the period of the evolution of the cyclone of November 10-12, 1989 every twelve hours. Vectors have been kept in the same scale to emphasize the intensity of the cyclone and to illustrate how strongly they are filtered out when computing the average.**

**Figure 2: Air temperature variation observed from ECMWF analysis between November 10 and November 12, 1989 (the 48 hours of evolution of the cyclone shown in Figure 1)**

**Figure 3: SST variation observed between the week previous to the evolution of the cyclone of November 10-12, 1989 and the week of the cyclone. Data are blended from bias corrected satellite, ships and buoys observations (IGOSS Data Set, Reynolds and Smith, 1994).**

**Figure 4: Climatological SST variation for the month of November from IGOSS Data Set (Reynolds and Smith, 1994).**

**Figure 5: Model domain and topography.**

**Figure 6: Model integration results at the steady state: a) mass transport stream function in Sv. and b) temperature at model layer 1. Model is forced with November 1989 mean wind stress from ECMWF analysis and IGOSS Data SST.**

**Figure 7: Model integration results after 48 hours of simulation. The model is forced with ECMWF analysis winds and air temperatures.**

**Figure 8: Time evolution of the different terms of the equation for the temperature time variation at the point of maximum variation of SST observed on model results when it is forced with ECMWF analysis winds and air temperatures for the 48 hours of evolution of the cyclone: a) zonal flux divergence; b) meridional flux divergence; c) vertical flux divergence; d) horizontal diffusion; e) vertical diffusion and f) heat flux through the ocean surface.**

**Figure 9: 48 hours simulation similar to figure 8 but in which a) wind forcing has been eliminated and b) heat flux has been suppressed.**

**Figure 10: Time function used to modulate the wind stress anomaly introduced by the cyclone in the long term run.**

**Figure 11: Temperature change, referred to the initial state, every 24 hours on a zonal section across the maximum of the oceanic temperature response to the atmospheric storm for the long period run.**

**Figure 12: Velocity field change (referred to the initial state) at the latitude-longitude point of maximum SST reduction: a) meridional velocity at level 1 of the model; b)**

**meridional velocity at several other layers of the model; c) zonal velocity at level 1 of the model and d) zonal velocity at several different depths in the model.**

| Level | Layer thickness (m) | Grid Point Depht (m) |
|---|---|---|
| 1 | 30.00 | 15.00 |
| 2 | 46.30 | 53.15 |
| 3 | 67.45 | 110.03 |
| 4 | 94.46 | 190.98 |
| 5 | 128.46 | 302.44 |
| 6 | 170.57 | 451.96 |
| 7 | 221.91 | 648.20 |
| 8 | 283.51 | 900.91 |
| 9 | 356.21 | 1200.80 |
| 10 | 440.57 | 1619.20 |
| 11 | 536.78 | 2107.80 |
| 12 | 644.54 | 2698.50 |
| 13 | 763.00 | 3402.30 |
| 14 | 890.71 | 4229.10 |
| 15 | 1025.53 | 5187.20 |
| bottom | | 5700.00 |

**Table I: Model layer thickness and grid points depth.**