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Possible mechanisms yielding an explosive coastal cyclogenesis over South America: experiments using a limited area model

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The LAHM/GFDL-CIMA model, has been used to simulate an explosive cyclogenesis which produced a strong southeast gale over the Río de la Plata (a phenomenon locally known as 'sudestada'). The main purpose of this work is to analyse the physical processes related to this development, after having verified the model's ability to reproduce the situation.

Through different experiments, the model's potential to represent the event is demonstrated, while two other experiments have been designed in order to analyse the impact of topography and of condensation/evaporation processes during the cyclogenesis.

Latent heat release resulting from precipitation processes is proposed as the main mechanism acting to intensify the system. The major air mass instability over the southeastern South American region was triggered by the entrance of a rather weak upper-level trough, and later reinforced by diabatic processes. By running the experiment without topography it was shown that, though the Andes did not cause the cyclonic development, they may have had a significant role during the early stages of the system's evolution.

The processes undergone by this system might be summarised as follows: as the cyclonic disturbance approaches the Andes, it is blocked at lower levels, but continues propagating in the middle and upper troposphere, becoming more vertical and barotropic. As a consequence, increased westerly winds lead to a reduction in static stability leeward of the Andes which contributes to the strengthening of a thermal-orographic low pressure system and favours a sustained warm and moist advection over the region. Finally, once the depression traverses the mountains and recovers its tilt, classical baroclinic development starts, dramatically strengthened by latent heat release.

Introduction

One of the regions which experiences enhanced cyclogenesis over South America is located near its eastern coast, between 25ºS and 40ºS (Gan and Rao 1991; Sinclair 1995). In particular, whenever a deep low pressure centre is located close to the coast near 30°S, strong southeasterly winds push Río de la Plata's waters upriver, causing floods at Buenos Aires City and its surroundings. These extreme phenomena,

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which affect the most densely populated area in Argentina (more than 15 million people), are locally known as 'sudestadas' (Spanish for southeasterly gale). Because of the smoothness and funnel shape of the Río de la Plata's estuary, this great basin is easily flooded when strong and/or persistent southeasterly winds blow. In addition, frontal depressions affecting this kind of phenomenon are usually accompanied by an anticyclone lying to the South, which enhances the meridional pressure gradient and consequently intensifies southeasterly gales. As well, these cyclones are often associated with rainfall, sometimes intense, which acts to reinforce the floods.

In spite of the severity of these cyclogenesis events and the resulting 'sudestadas', few research studies have been carried out. The first ones, by Hessling (1923) and Schwerdtfeger (1954), showed that these cyclones develop when an upper-level trough, associated with a warm or a quasi-stationary front over this region, encounters warm and moist low-level air. Necco (1989) found that these systems are strongly influenced by upper diffluent mass fluxes, suggesting that cyclonic vorticity advection is the mechanism responsible for this kind of development. More recently, Jusem and Atlas (1991) analysed one explosive cyclogenesis over the Río de la Plata, evaluating the terms in the pressure tendency equation produced by a primitive equation forecast model. They found that thermal advection is the primary cyclogenetic effect, while vertically integrated horizontal convergence and vertical motion exert a damping effect on the pressure changes.

Using a five-year statistic (1980-1984), including 55 cases, Seluchi (1993) concluded that cyclogenesis events over eastern South America are caused by the arrival of an upper-level cyclonic disturbance, which propagates across lower latitudes than usual (frequently as a cut-off vortex) triggering a quasi-stationary front located over eastern South America. Coincidentally, Gan and Rao (1991) and Sinclair (1995) proposed baroclinic instability in the westerlies as the primary mechanism for South American coastal cyclogenesis. In both studies, however, the role of enhanced moisture and associated latent heat release are also mentioned as plausible contributors.

Furthermore, as these developments are sometimes explosive and strongly influenced by the Andes, low resolution global models are, sometimes, not suitable for forecasting them. This is particularly true over the southern hemisphere, where the models show less skill than over the northern hemisphere. For this reason, it is considered that higher resolution limited area models with full physics are more appropriate for the simulation of these systems, providing datasets with greater internal consistency, more adequate to perform physical analyses than the available observational datasets.

The purpose of this work is to analyse the synoptic situation associated with an explosive cyclogenesis event which developed over eastern Argentina and Uruguay, on 12 November 1989, using the LAHM/GFDL-CIMA model (version improved at CIMA (Orlanski et al. 1989; Nicolini and Saulo 1995) of the Limited Area Hibu Model/Geophysical Fluid Dynamics Laboratory). First, the model's ability to reproduce the situation will be assessed and then possible physical processes responsible for its development will be discussed.

The LAHM/GFDL hydrostatic model (Orlanski and Katzfey 1987) used for these experiments, is currently implemented for research purposes and contains bulk physical parametrisations of various subgrid-scale processes (radiation, planetary boundary-layer fluxes, deep convection). It has 18 sigma levels in the vertical while both domain and horizontal grid spacing can be designed for each run.

Moist convective parametrisation is currently introduced either using the cumulus cloud ensemble model developed by Arakawa and Schubert (1974), or representing condensation/evaporation processes directly by solving the corresponding model equations, this last option being the one chosen for this work (see Saulo and Nicolini (1997) for further details). This last approach (known as 'explicit moist physics') means that both stable rain and unstable moist convection are explicitly represented (i.e. at horizontal scales explicitly resolved by the model's grid size), handling phase changes between vapour and liquid states with an adjustment to saturation assumption when a relative humidity value of 95 per cent is exceeded. This treatment has been adapted by Saulo and Nicolini (1996), following the original proposals of Ross and Orlanski (1978, 1982). Similar formulations have been used extensively by a number of authors (see Molinari and Dudek (1992), for a review on this topic).

Initial and boundary conditions for experiments described here are provided by the ECMWF (European Centre for Medium-range Weather Forecasts) analyses, at six-hour intervals for seven standard pressure levels (1000, 850, 700, 500, 300, 200, and 100 hPa) with 1.125 degrees resolution, in both latitude and longitude.

The synoptic situation

General remarks

The 12 November 1989 event has been selected because it produced one of the most severe coastal cyclogenesis events in recent decades over eastern South America, yielding strong southeasterly gales over Buenos Aires. Surface winds up to 50 knots (at Reconquista station, 27.9°S, 59.4°W), and very intense precipitation over eastern Argentina and Uruguay were observed. The Río de la Plata overflowed and the water level exceeded the evacuation limit (2.8 m), reaching a peak value of 4.1 m, producing extensive floods which persisted more than 18 hours.

Following the criterion of Sanders and Gyakum (1980), this event can be considered a 'bomb', with a pressure decrease greater than a Bergeron. The observed pressure drop was around 26 hPa, between 1200 UTC 11 November and 1200 UTC 12 November. It is interesting to note that this cyclogenesis was not forecasted by the Argentine National Meteorological Service (NMS) nor by the global systems of NCEP (National Centre for Environmental Prediction) and ECMWF.

Synoptic overview

On 10 November at 1200 UTC (Fig. 1(a)) a cold front propagating from southern Argentina was located around 62°W, associated with an incipient cyclone north of the Malvinas (Falkland) Islands. Another low pressure centre can also be observed, frequently detectable at low levels in the lee of the Andes, now apparent around 25°S. This system, known as the Northwestern Argentine's Low (NAL), has a thermal-orographic origin, and has been extensively studied by Lichtenstein (1980, 1989). On this occasion, the NAL was merged with a quasi-stationary front weakly depicted by the contour's cyclonic curvature in Fig. 1(a), but more recognisable in NMS manual analyses (not shown). Between the cold and the quasi-stationary fronts, isolated convection was reported. At 500 hPa, a latitudinally extended trough (solid contours) was close to the occidental coast, with its corresponding ridge located over central Argentina.

In the subsequent 12 hours, the cold front moved northward approaching the quasi-stationary front (not shown), while the NAL deepened slightly and convection (not shown) began to affect the area between approximately 60°W, 25°S and 55°W, 37°S.

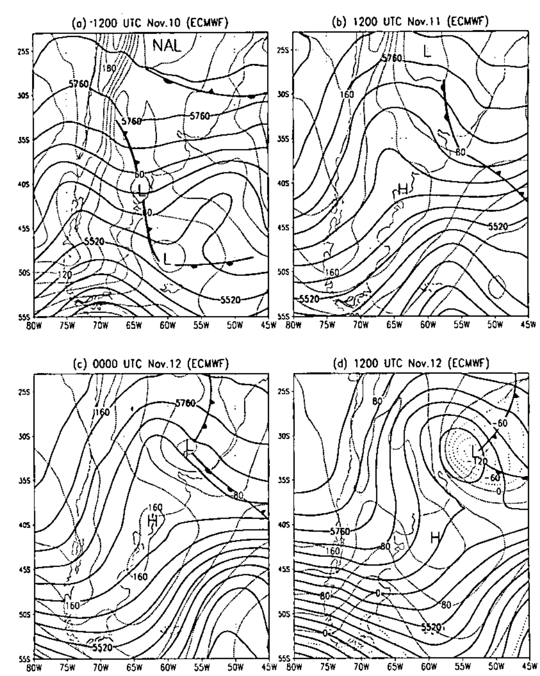
On 11 November at 1200 UTC (Fig. 1(b)) a weak frontal wave could be detected near Uruguay (denoted by 'L'), simultaneously with the arrival of a migratory anticyclone over Argentina, both features contributing to the development of light southeasterly winds over Buenos Aires province and the Río de la Plata. Deep convection was still confined to eastern Argentina and Uruguay, with considerable intensity at certain locations where thunderstorms and strong surface winds were reported. The 500 hPa trough was located at this time to the lee side of the Andes, with stronger cyclonic curvature and shear than in Fig. 1(a), both factors leading to an increase in cyclonic vorticity upwind of the low level frontal wave.

Figure 1(c) (0000 UTC, 12 November) clearly depicts a frontal cyclone, north of Río de la Plata, with a central pressure of 999.5 hPa (near Concordia station, 32.1°S, 56.0°W, according to NMS data). The intensification of southeasterly winds was significant, reaching from 20 knots to 35 knots all over the riparian area. This increase was favoured by the approach of the migratory anticyclone from the west, associated with a major ridge propagating from the Pacific Ocean. At this time, the frontal cyclone was approximately located below the inflection point of the 500 hPa trough (which is slightly poleward diffluent), denoting the beginning of the maximum intensification period (Petterssen 1956; Bluestein 1993).

The system reached its maximum intensity at approximately 1200 UTC (Fig. 1(d)) 12 November near the Atlantic coast, with a distinct closed cyclonic circulation from the surface to 300 hPa. Due to the lack of surface data from Brazil and Uruguay, we do not know the lowest pressure attained by this system, which, according to ECMWF data could have been lower than 985 hPa.

After November 13, the cyclone entered its decay phase, moving slowly southeastward to the Atlantic Ocean (not shown).

Fig. 1 Observed geopotential height in meters (ECMWF analyses) at 1000 hPa (short dashed lines) and 500 hPa (solid thick lines) for: (a) 1200 UTC 10 November; (b) 1200 UTC 11 November; (c) 0000 UTC 12 November and (d) 1200 UTC 12 November. Frontal zones are shown, low (high) pressure centres are denoted by 'L'('H') and the Northwestern Argentine's Low is denoted by the symbol 'NAL'.



Experimental design

In order to evaluate the model's capability to reproduce this event and to assess whether changes in resolution and/or initialisation time affect the results, three different experiments have been carried out. It should be pointed out that this study is not focused on the model's capability to forecast the situation, since real data has been used to update boundary conditions. All the experiments, which are listed below, have been integrated using the explicit representation of convection over an area encompassed by 10°S, 60°S and 100°W, 40°W for a period of 48 hours:

1. LR1012: This experiment has been initialised at 1200 UTC 10 November (see Fig. 1(a) for initial conditions), when there was very little evidence for the impending cyclogenesis. Its resolution has been fixed at 1.125 degrees (in latitude and longitude), taking less than two hours to be integrated on a Sun Sparcstation 10.
2. HR1012: The same as LR1012, except for the horizontal resolution, which has been increased to 0.375 degrees. This experiment took 30 hours to be completed using the same machine.
3. HR1100: This experiment has been initialised 12 hours later than experiment HR1012, at 0000 UTC 11 November.

Figure 2 shows 1000 and 500 hPa contours at three different times for all the experiments. It can be seen that the model simulates the synoptic situation fairly well, with very small differences between the experiments. As a whole, the temporal evolutions of middle and lower tropospheric patterns are broadly similar. In particular, the surface cyclone at the time of maximum intensity (see Figs 2(c), (f) and (i)) is generally well captured, both in magnitude and location (see Fig. 1(d) for comparison). However, comparing the simulations at 0000 UTC 12 November (Figs 2(b), (e) and (h)) the surface cyclone and its associated circulation pattern are better positioned in HR1012, being rather north in LR1012 and rather south in HR1100; hence no southeasterly winds can be observed at the Río de la Plata at this time in these two experiments.

Taking into consideration that HR1012 portrays the selected event quite well, we have chosen HR1012 instead of HR1100 as the control experiment. This selection is made for two main reasons. First, in order to focus on the physical processes involving the Andes, the experiment that more accurately represents the system evolution should be kept. In particular, the time at which the perturbation is windward and its subsequent displacement across the mountain barrier should be retained. Second, the lack of surface data and the absence of upper air observations at 0000 UTC over most of South America may adversely affect the quality of 0000 UTC analysed fields, and possibly impact on the accuracy of the HR1100 forecast.

It must be kept in mind that the key assumption is that the basic physical mechanisms responsible for the generation of this system may be properly described by the model.

Possible mechanisms yielding this explosive cyclogenesis

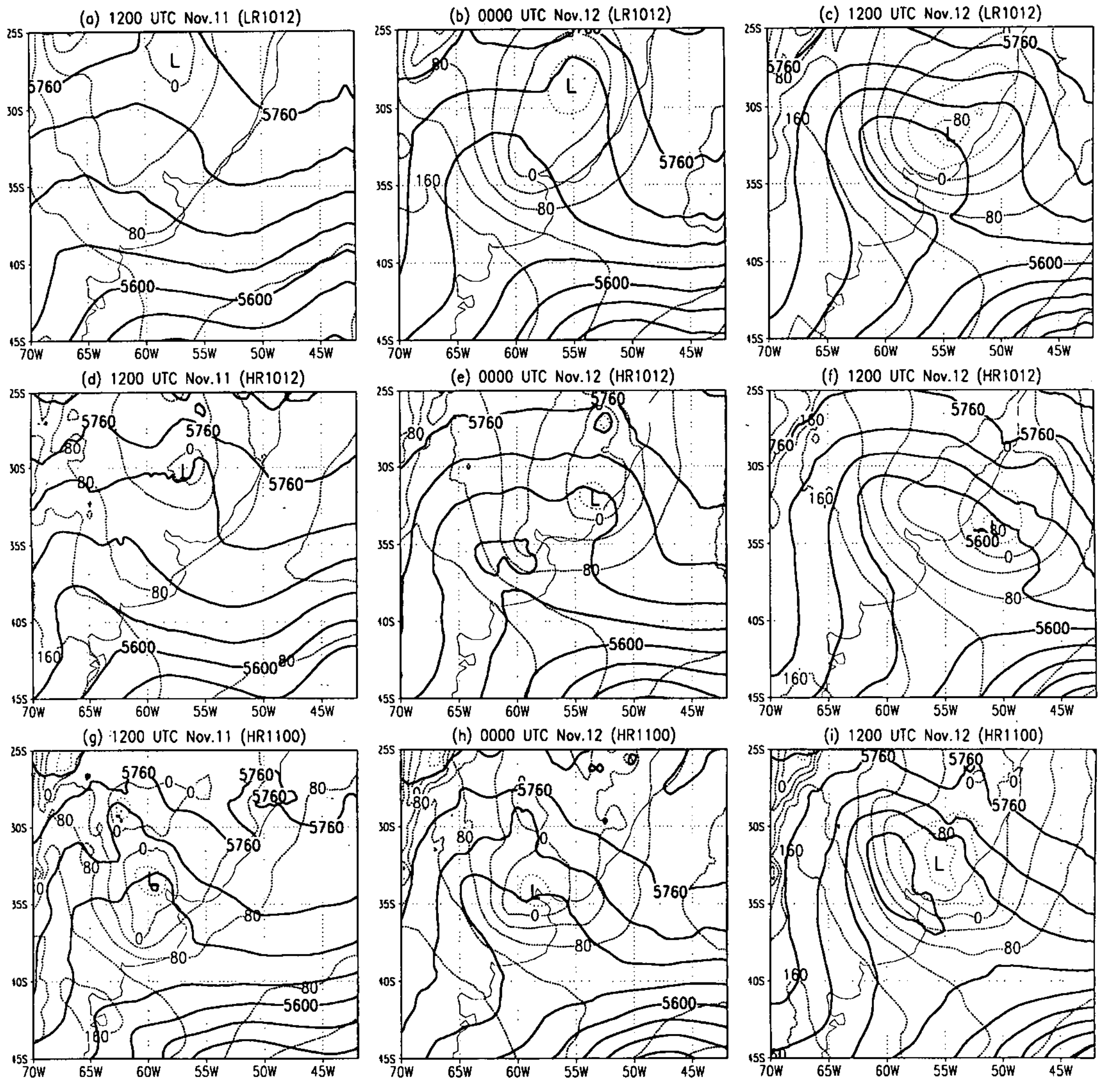
Dynamical considerations

As mentioned previously, one of the most frequent patterns yielding deep surface cyclogenesis is the presence of a strong upper-level cyclonic disturbance, usually associated with a tropopause undulation (Hirschberg and Fritsch 1991b) and an extrusion of stratospheric air. These mechanisms lead to an increase of upper-level thickness and, subsequently, to a drop in surface pressure. Jusem and Atlas (1991) regarded this mechanism as one of the most important in the generation of the 29 May 1984 storm. A similar conclusion was reached by Gan and Rao (1996) in their study of the 8 June 1981 coastal cyclogenesis event. Nevertheless, Bell and Bosart (1993) in a case study over the United States, have found that a strong vortex in the upper troposphere was associated with only a moderate surface cyclone. They concluded that baroclinicity at low and middle levels is also necessary to ensure a vertically extended low pressure system.

In order to evaluate whether this system behaves as those previously referred to, the upper-level disturbance will be followed during the first stages of its evolution. Figure 3(a) (valid through 0000 UTC 12 November), shows a west-east vertical cross-section at 33°S obtained with HR1012, at the beginning of the maximum growth period. Looking at the v-wind component, a cyclonic shear associated with an upper-level vortex can be detected, with its maximum located near 300 hPa. According to the vertical gradient of isentropes, the tropopause level can be detected approximately at 200 hPa with a weak undulation around 70°W, associated with a stratospheric air intrusion depicted by the increase of static stability between 300 and 200 hPa. Even though these patterns show the existence of an upper-level disturbance, this dynamical forcing seems to be weaker than that documented by other authors (Jusem and Atlas 1991; Bluestein 1993; Seluchi 1993) who show tropopause undulations reaching the 450 hPa level with an accordingly strong warm air intrusion.

Vertical cross-sections of geopotential height change during the maximum growth period (between 0000 and 1200 UTC 12 November) are shown in Fig. 3(b) through 3(d). In Fig. 3(b) geopotential decreases of similar magnitude over the entire troposphere can be found,

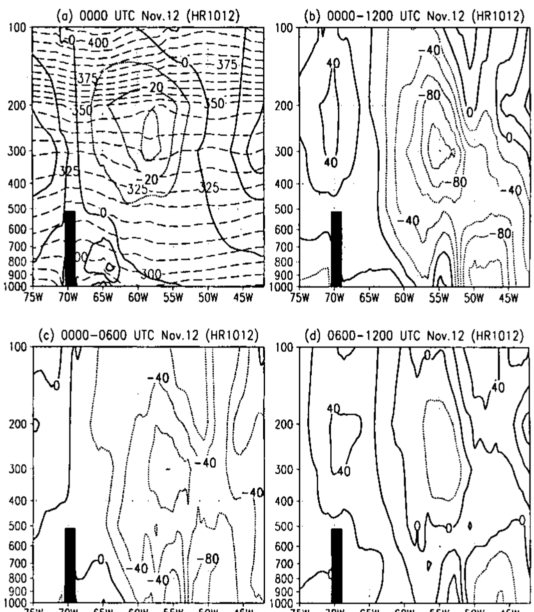
Fig. 2 **Geopotential height in meters at 1000 hPa (short dashed lines) and 500 hPa (solid thick lines) for: (a) 1200 UTC 11 November; (b) 0000 UTC 12 November and (c) 1200 UTC 12 November, simulated with LR1012. (d) (e) and (f) same as (a), (b) and (c) but simulated with HR1012. (g) (h) and (i) same as (a), (b) and (c) but simulated with HR1100. Low pressure centres are denoted by 'L'.**



with two relative maxima at lower and upper levels. In particular, the upper-level height fall is almost half that found by other authors. Moreover, this vertical structure does not appear as the classical pattern obtained by many previous studies (Hirschberg and Fritsch 1991a, Bluestein 1993; Seluchi 1993; Gan and Rao 1996) which show a strong upper-level vortex which subsequently propagates toward lower levels (classified as 'type-B' cyclogenesis by Petterssen and Smebye (1971)). This can be more clearly verified with the aid of Figs 3(c) and (d), where partial time contributions to geopotential decrease are shown. Recall that no distinct height loss can be tracked from upper to lower levels.

From this analysis, there does appear to be some dynamic forcing at the beginning of the explosive cyclogenetic phase. Other related processes, which could have helped to reinforce this development, will be investigated in the following sections by separately testing model sensitivity to topographic and latent heating processes.

Fig. 3 **West-east cross-sections at 33°S for the control experiment (HR1012). (a) Potential temperature (long dashed lines, contours drawn every 5°) and v-wind component at 0000 UTC 12 November (positive contours shown as thick lines and negative ones dotted. Contours drawn every 10 m/s.) (b) Geopotential height change for 0000-1200 UTC 12 November. (HR1012) (contoured every 20 m); (c) same as (b) but for 0000-0600 UTC 12 November; (d) same as (b) but for 0600-1200 UTC 12 November. Dark rectangles represent the approximate location and height of the Andes.**



Topographical effects

It has long been observed that cyclogenetic regions are preferentially located at the lee side of mountain barriers (Petterssen 1941; Miller and Mantis 1947); moreover, Trevisan (1976) and Chung (1977) showed that topographic shape, height and orientation with respect to wind direction, greatly influence some aspects of these cyclogenesis events. One of the first studies of Rocky Mountains lee cyclogenesis (Newton 1956) concludes that they are due to strong heating caused by forced descent over the eastern slope. Alpine cyclogenesis was studied at the beginning of this century by Von Ficker (1920), and later, among many others, by Buzzi and Tibaldi (1978) who showed that the Alps delay cold fronts, distorting thickness fields and generating a foëhn effect.

The Andes role on cyclogenesis has been less studied, though it is known that many cyclones develop about 1000 km to the east (Lichtenstein 1989; Gan and Rao 1991; Sinclair 1995), not matching, strictly speaking, what is known as lee cyclogenesis. Nevertheless, the Andes exert a distinct influence, being responsible for particular patterns in the lower-level circulation. For example, low pressure systems propagating from the Pacific Ocean are delayed at low levels, yielding a sustained warm and moist advection leeward.

In order to analyse the topographic effect upon the system, an additional experiment has been carried out, identical to the control (HR1012) but without mountains (NT1012, where surface height has been set to zero). Though it can be considered that the initial condition contains information about topography, it has been verified that after six hours, the patterns typically associated with the Andes proximity have been diminished. For example, contrasting with patterns seen in Fig. 1, lower level pressure contours become no longer parallel to the Andes. Nevertheless, it must be kept in mind that orographic effects cannot be completely removed in these kinds of experiments (where shorter time-scale signals are to be retained), though they can indeed be minimised. In this sense, results obtained from NT1012 should be interpreted carefully.

Figure 4 shows the 1000 hPa pressure field (dotted line) and the difference between the experiment without topography and the control (i.e. NT1012 minus HR1012). It can be seen that, even without topography, the low pressure centre is still present. However, the surface cyclone is slightly weaker and its development is not as explosive as in the control run. A closer inspection reveals that a difference dipole strengthens with time and its axis rotates from a SW-NE direction towards a W-E one, suggesting that the NT1012 cyclone is initially located southward and becomes weaker and somewhat delayed compared with the cyclone from the control run.

These results suggest that, even though the Andes do not strongly affect the behaviour of the surface cyclone, they have modified some of its characteristics. This will later be discussed in more detail.

Latent heat release impact

Cyclogenesis at the southeastern South American coast is sometimes accompanied by heavy rains and storms, which contribute (through latent heat release) to the system's growth. According to authors who have studied cyclogenesis over the east coast of North America (Smith et al. 1984; Manobianco 1989, among many others), latent heat release is the key factor to account for explosive deepening.

Figure 5 shows 1000 and 500 hPa geopotential heights at 1200 UTC 12 November obtained with NH1012, an experiment identical to the control, except that heat released (absorbed) by condensation (evapora-

Fig. 4 **Geopotential height in meters at 1000 hPa (dotted lines) simulated with NT1012 and differences in geopotential height between NT1012 and HR1012 (in meters, positive (negative) contours shown as thick (short dashed) lines) for: (a) 1200 UTC 11 November; (b) 0000 UTC 12 November and (c) 1200 UTC 12 November.**

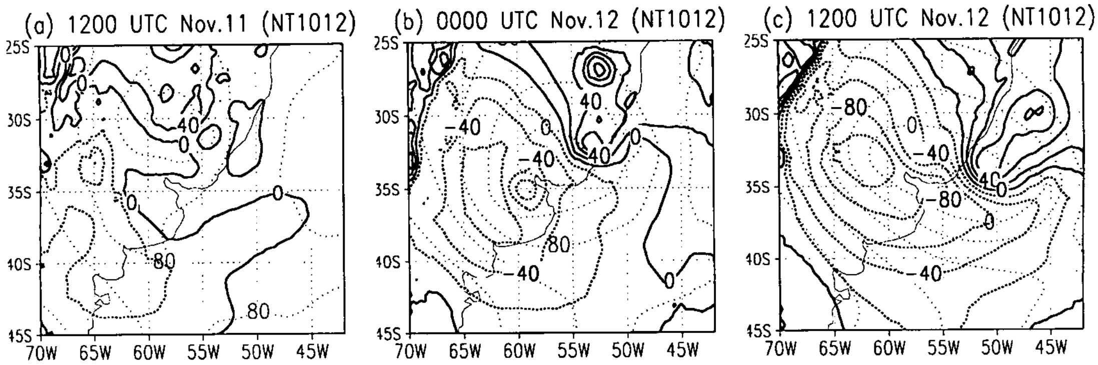
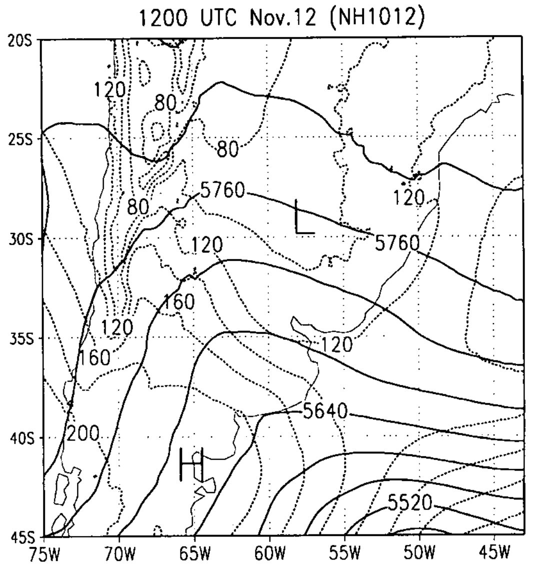


Fig. 5 **Geopotential height in meters at 1000 hPa (short dashed lines) and 500 hPa (solid thick lines) for 1200 UTC 12 November, simulated with NH1012.**



tion) processes has been removed. Almost no cyclonic development can be found, as indicated by the weak frontal wave and the smooth pressure field north of 35°S. The associated trough at middle levels is quite weak, and so is its relative vorticity. As the migratory anticyclone was little affected, southeasterly winds were still forecast over the Río de la Plata.

Figure 6 depicts a vertical cross-section of geopotential height changes (between 0 and 1200 UTC 12 November) calculated as follows:

$$\{z_{HR1012}(t_2) - z_{HR1012}(t_1)\} - \{z_{NR1012}(t_2) - z_{NR1012}(t_1)\}$$

where z is geopotential height, t_2 corresponds to 1200 UTC 12 November and t_1 to 0000 UTC 12 November. This field may be interpreted as a measure of geopotential height changes induced by latent heat release during the whole integration period.

It can be seen that the cross-section presented in Fig. 6, closely resembles the one already shown in Fig. 3(b), thus suggesting that height variations during this 12-hour period are mainly explained by latent heat release. Furthermore, diabatic effects act so as to enhance baroclinicity, as can be implied by the tilt with height of the minimum tendency axis. Therefore, it can be inferred that the system under study has been strongly reinforced by diabatic processes.

To further assess the significance of latent heat release processes, the precipitation associated with this system will be analysed. The observed accumulated precipitation field (between 1200 UTC 11 November and 1200 UTC 12 November) is presented in Fig. 7(a). Heavy rains over eastern Argentina and Uruguay, exceeding 100 mm up to 216 mm at Piedras Coloradas (Uruguay, at 32.2°S, 57.5°W) can be found, confirming that latent heat release could have indeed been a key factor in this case. In addition, Fig. 7(b) shows simulated precipitation over the same period where it can be seen that there is a satisfactory similarity between modelled and observed fields, thus supporting the idea that condensation/evaporation processes have been properly simulated.

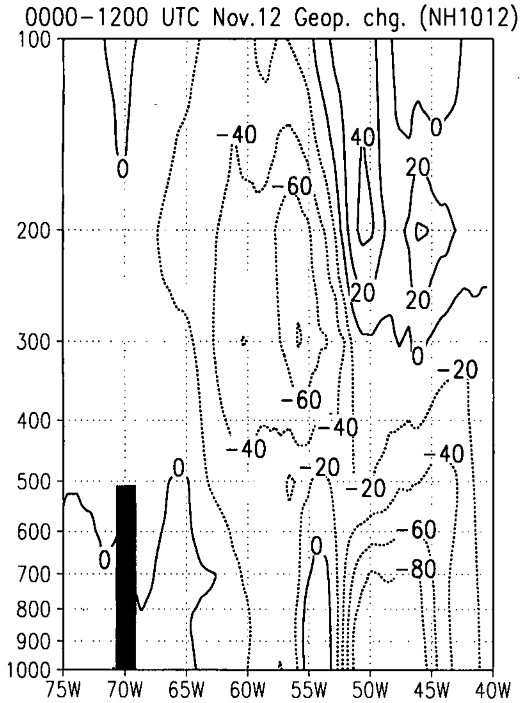


Fig. 6 **Temporal geopotential height changes in metres between 0000 and 1200 UTC, 12 November (see text for explanation). Dark rectangles represent the approximate location and height of the Andes.**

Discussion of results

From the results previously shown it can be proposed that, among those processes analysed, diabatic heating due to condensation is the main mechanism responsible for the explosive development of this system. This section will mainly deal with the Andes's influence upon the system, examining certain circulation patterns that could have played an important role during the life cycle of this event. At this stage it is worthwhile to note the particular shape of this mountain barrier, characterised by high tops (above 6000 m), modest broadness and steep slopes, which may produce differences in local circulations from those observed over other orographic obstacles.

In general, the classical mechanism resulting in lee-cyclogenesis begins with a decrease of low/middle-level static stability produced by adiabatic descent on the lee. This process establishes the surface cyclone, and baroclinic development starts when the upper-level disturbance approaches. This behaviour has been well recognised in the lee of the Rocky Mountains, and documented since Newton (1956).

As inferred by the analysis of NT1012, the Andes do not determine the occurrence of this cyclogenesis, but modify some aspects concerning its intensity and explosiveness. In order to better display the topography's role, Figs 8(a), (b) and (c), show the relative vorticity vertical cross-section at 33°S (obtained with HR1012), when the disturbance is (a) windward; (b) directly above the mountains and (c) leeward. On the other hand, Fig. 8(d) (lower left panel) is at the same time as 8(a) but when no topography has been considered (obtained with NT1012).

The process undergone by this system might be summarised as follows: as the depression coming from the Pacific Ocean approaches the Andes, it is blocked by the mountain barrier at lower levels, while the disturbance continues propagating at middle and upper levels. This causes the perturbation axis to become more vertical (Fig. 8(a)), delaying the system until it becomes nearly stationary and quasi-barotropic (Orlanski et al. 1987). When the upper-level trough approaches the Andes crest, westerly winds increase and leeward adiabatic descent leads to a decrease in static stability. This destabilisation of the lower levels can be inferred from Fig. 9 (particularly between 1200 UTC 10 November and 0600 UTC 11 November) where subsidence and the resulting decrease in the vertical gradient of potential temperature are shown (see figure caption for more detail). Because of this process, the thermal-orographic low pressure system (NAL -see Fig. 1(a)), which is rather intermittent at this time of the year (Lichtenstein 1989), becomes better organised somewhat eastward of the lee slope. One of the most important consequences of this low pressure system development is the beginning of sustained warm and moist advection over eastern South America. Figure 8(b) depicts two cyclonic vorticity maxima in the lower and middle troposphere (windward and leeward respectively), suggesting that the trough's continuous advance at higher levels finally causes a development in the lee of the Andes. The superposition of the system at the lee, with the NAL, results in a deeper single system which further enhances northerly winds and, subsequently, moist and warm advection. Once the depression passes over the mountains and recovers its tilt (Fig. 8(c)), classical baroclinic development starts, dramatically strengthened by latent heat release (note the number of contours indicating the greater intensity of cyclonic vorticity). On the other hand, Fig. 8(d) shows that, without topography, the disturbance holds its tilt throughout the troposphere. This fact may help explain that, compared with Fig. 8(a) (control run), the system attains a lower intensity at low levels, and locates around 65°W earlier than in the control run.

Fig. 7 **(a) Observed 24-hour accumulated precipitation (from 1200 UTC November 11 to 1200 UTC November 12) in mm. Shaded area corresponds to precipitation amounts greater than 5 mm. Contours are drawn every 25 mm; (b) Same as (a) but simulated with control experiment (HR1012).**

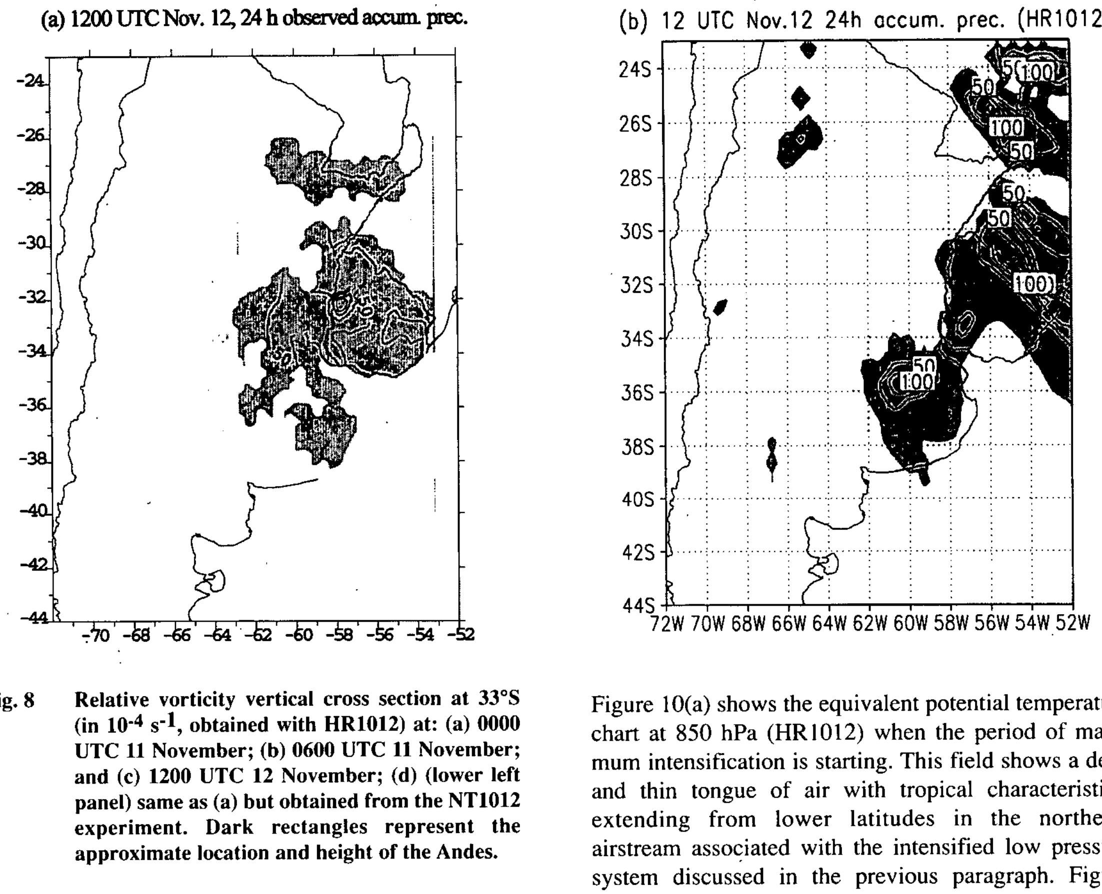


Fig. 8 **Relative vorticity vertical cross section at 33°S (in 10^{-4} s^{-1}, obtained with HR1012) at: (a) 0000 UTC 11 November; (b) 0600 UTC 11 November; and (c) 1200 UTC 12 November; (d) (lower left panel) same as (a) but obtained from the NT1012 experiment. Dark rectangles represent the approximate location and height of the Andes.**

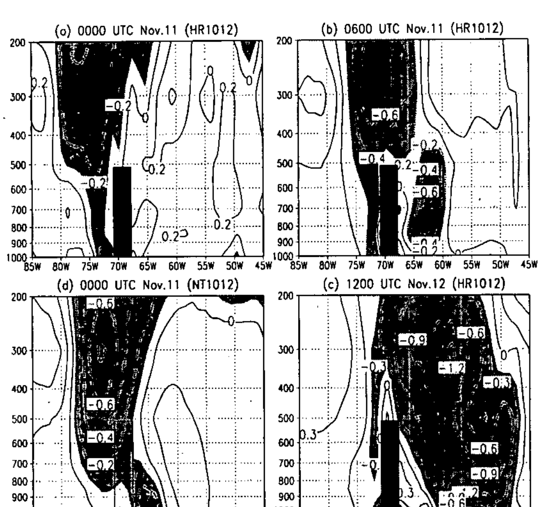


Figure 10(a) shows the equivalent potential temperature chart at 850 hPa (HR1012) when the period of maximum intensification is starting. This field shows a deep and thin tongue of air with tropical characteristics, extending from lower latitudes in the northerly airstream associated with the intensified low pressure system discussed in the previous paragraph. Figure 10(b) shows an equivalent potential temperature vertical cross-section at 33°S, at the same time. It can be seen that the air mass is clearly convectively unstable, with an equivalent potential temperature decrease with height of more than 20°, sufficient to provide the necessary energy to support convection. In contrast, Figs 10(c) and (d) show the same fields but obtained with NT1012, where lesser static instability and a smoother θ_e-tongue are detected, providing further evidence of the Andes influence upon this development.

This case study differs somewhat from the typical cyclogenesis events characterised by Seluchi (1993). In particular, on 10 November, the atmospheric circulation showed little evidence of the forthcoming cyclogenesis because the trough at middle and upper levels was less intense than in typical cases, and the zonal circulation index, usually very small for cyclogenesis cases, was almost normal. Moreover, using the forecast method proposed by Seluchi (1993), which does not consider moisture predictors, only a 15 per cent probability of

Fig. 9 **Cross section of area averaged (between 20-30°S and 60-70°W) potential temperature (K) (simulated with HR1012) for the period 1200 UTC, November 10 through 1200 UTC, November 12. Downward vertical motion (shaded) is also indicated.**

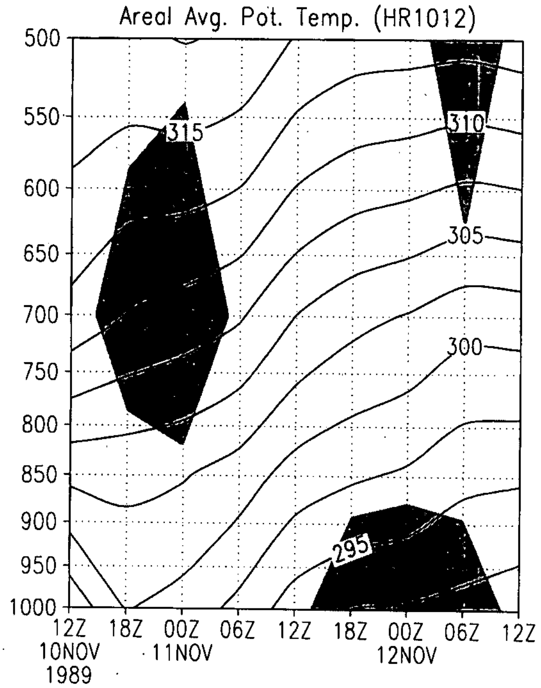
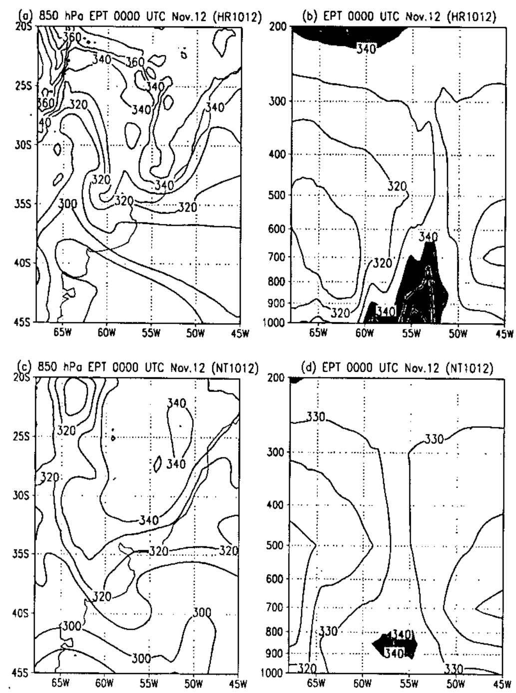


Fig. 10 **(a) Equivalent potential temperature (K) at 850 hPa, simulated with control experiment (HR1012) at 0000 UTC, November 12; (b) Vertical cross section of equivalent potential temperature at 33°S, at the same time as (a); (c)and (d) same as (a) and (b) respectively but for NT1012.**



cyclogenesis occurrence was obtained. Nevertheless, the existence of highly unstable conditions in conjunction with an initially weak trough, were enough to trigger convection and, consequently, generate a suitable environment for an explosive event.

Conclusions

The 12 November 1989 cyclogenesis event has been studied in order to (a) assess the LAHM/GFDL-CIMA regional model's performance on this event and (b) analyse the processes responsible for the cyclonic development.

The model's ability to simulate this situation has been evaluated through different experiments that showed its potential to predict the event 48 hours in advance, with a reasonable computational requirement. Using these results, a control experiment has been designed, for which horizontal resolution has been fixed at 0.375° (both in latitude and longitude) and its initial condition at 1200 UTC 10 November, to give 48 hours of simulated time. This experiment showed a satisfactory agreement with observed fields.

Latent heat release resulting from precipitation processes is proposed as the main mechanism acting to intensify this system. Major air mass instability over the southeastern South American region has been triggered by the entrance of a weak upper-level trough which could be later reinforced by diabatic processes. The first stages in this development could have been influenced by topography, through a delay in the system's propagation in the lower tropospheric layers which, in turn, favoured those mechanisms reinforcing air mass instability. In terms of development theory, adiabatic cooling could not counteract heating by condensation during this period. Consequently, the system entered its decay phase when the surface depression moved to the east far from the source of unstable air.

It is interesting to note that, according to Sinclair (1995) there are very few 'bombs' north of 40°S. This is possibly due to the fact that during winter, when baroclinic disturbances extend further north, they do not usually encounter air masses with enough convective instability. On the other hand, in summertime, even though conditionally unstable air masses are sometimes present at these latitudes, transient perturbations are confined to latitudes south of 40°S. Therefore, although infrequent, both effects can occur in the transitional season, as in the case chosen for this work.

Daily inspection of global forecasts suggests that, at least over South America, global models show some difficulties in forecasting explosive events like the one studied in this work. Taking into account the results of this research it can be suggested that such deficiencies arise for two main reasons; first, uncertainties in the representation of convection and the sensitivity to the moisture fields, and second, the ability of global models with limited resolution to represent topographic effects, especially over the Andes which are characterised by a narrow and steep mountain barrier, with very high peaks. It is believed that both of these aspects are more adequately handled by the regional model used here.

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