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INTENSE SNOWSTORM IN THE SOUTHERN MOUNTAINS OF PERU ASSOCIATED TO THE INCURSION OF CUT-OFF LOW-PRESSURE SYSTEMS AT UPPER LEVEL

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ABSTRACT

This paper describes the formation and evolution of a Cut-off low over the Tropical East Pacific (southwestern Peru), occurred between June 29 and July 08 in 2004. The snowstorm associated to this system, affected the poorest regions of Peru (South Mountain) leaving a tragic balance of 4766 destroyed houses, 1703 Has of lost crops, 260505 dead animals (cattle, ovine, goat and mainly auquénidos) and deterioration of the settlers health (INDECI, 2004).

There were two components in this study: 1) the precursory formation flows of the Cut-off low, were analyzed and 2) it was determined the contribution of the dynamic and thermal forcing of high and middle troposphere. The analysis of flows shows that the genesis of the Cut-off low, in its initial stage was preceded by the amplification of one ridge on the Tropical and Subtropical Pacific, later by the northeast side of this ridge a short-wave trough- Jet Streak system propagates through the longer-wave-amplyfing ridge, favoring the incursion of stratospheric air. Meanwhile, on continent the formation of a anticyclonic circulation at middle and low levels with northeast winds on the Amazonia, supported the warm and humid air advection towards the South mountain range and Altiplano Peruvian-Bolivian that when interacting with the above mentioned mechanisms, favored the occurrence of intense snowstorms over a great extension of the Peruvian Andes.

Keywords: cut-off low; South America, Jet Streak, Peru

1. Introduction

The Cut off low pressure (COL) are system generated at upper levels troposphere associated to strong wave processes and isolation of jet stream (JS) circulation to be propagated in middle levels and in most cases to low levels. In one case study of a cut off low pressure diagnosis, Bell and Bosart, (1992) observed that at the initial stage of a ridge-trough system, the ridge is amplified a short wave trough is displaced associated to a jet streak that moves embedded in the long wave trough from middle latitudes towards warmer regions.

Hoskins et al., (1985); Hoskins (1991), defines the COL, as an isolated system of high cyclonic potential vorticity, that extends from the low stratosphere to the high level troposphere; whilst Van Delden A. y Negger R., (2003), define COLs or tropopause cyclon as a closed circulation of high troposphere and low stratosphere; besides they mention that at middle latitudes are the result of dynamical processes, in which air mass, with high potential vorticity, is cutted off from the polar stratosphere, getting isolated in the troposphere of middle latitudes.

Price y Vaughan, (1992), classify three classes of COLs, depending of nature of Jet Streak and the place where they are formed, these types are determined as: Polar, Subtropical y Polar Vortex, each one have their own characteristics. Pizarro, J. and Montesinos A., (2000), classify three types of COLs, the Chilean coast, being this the A type: the characteristic of this type is cyclonic circulation of the deep trough following by a oriented fluent NW-SW of B type, is similar to the A type, but in the center of the lower pressure is observed in the surface under of the trough and the characteristic of C type is a trough in the middle of the troposphere with an NE-SW orientation.

Van Haver et al., (1996), concluded that trpopause slopes are the highest agents of interchange between stratosphere and troposphere and the outcomes obtained for the Northern Hemisphere (NH), in individual events, they show more efficiency in spring than in autumn.

The studies made for the southern hemisphere are very scarce (SH), Kousky y Gan (1981), studied the cyclonics vortices of high troposphere during a five year period for South America and adjacent oceans; Qi et al. (1999), analyzed the climatology of the COLs over the Southern Australia and one case study; J. L. Baray et al. (2003), analyzed the dynamic of one COL over South Africa associated to the impact in the ozone and the last study carried out by Fuenzalina et al. (2005), show the climatology of the COLs in the SH.

In Peru, the COL causes different meteorological phenomena such as intense rains in summer; strong wind and snowstorm in winter causing loss of human lifes, serious crops damages in agriculture, livestock and infrastructure. The Highlands of the southern of Peru is the most affected.

Due to the severity of the socio-economic impact in Peru, its necessary to investigate this event, that is why it is important to know the dynamic of the COL in the SH.

2. Data source and analysis procedures

The large-escale environment is diagnosed using the National center for environmental Prediction (NCEP) reanalyzed meteorological field described in detail by Kalnay et al. (1996) to characterize the large-scale tropospheric circulation and termodynamic structure. The reanalysis fields have a 6-hour resolution on a 2.5° latitude x 2.5° longitude grid and include all mandatory levels from 1000 to 100 hPa.

The analysis was carried out through two phases which were divided in: the precursory and formation phase. The first one or the precursory phase (29 –30 June 2004) was analyzed through the observation of circulation patterns at high and low troposphere level such as the JS, sea level pressure, field of thickness of 500/1000 hPa and the identification of Jet Streak through the methodology established by Nielsen, (*Comet - Program*); whilst the formation phase of COL was analyzed through quasi-geostrophic equations of potential vorticity, hydrostatic, thermodynamic and Q-vector, as well as the analysis of satellite GOES 8 and the estimated precipitation by the TRMM/NASA.

3. Synoptic overview of the Cut Off Low

The patterns of circulation in the austral winter in South America are defined by passage of long wave trough which are conductors of the formation and intensification of migratory anti-cyclones, cyclogenesis amongst others, over the continent. Another important meteorological phenomena that normally occurs, specially in the south east Pacific is the cut off low pressure system (COL) there are fewer situations that occur over tropical latitudes, taking place one of them between June and July 2004, situated between latitudes 20º S .

3.1 Precursors phase

The precursory phase was considered between June 29 – 30, 2004 , this synoptic situation was characterized by the amplification of a ridge over the south-eastern Pacific. By June 29th, a ridge axis between 20°S 120°W to 40°S 90°W, started to propagate, posing its axis over 20°S 110°W to 40°S 80°W, this amplification of the ridge is associated to warm air incursion from the north so this both increase the deep trough (fig. 1.a and 1.b); likewise this trough associated with the displace of JS modulated to zonal flows in latitudes over 30°S, meanwhile to the north of the 30°S a southern flow separation with cold characteristics that entered into tropical latitudes. Likewise, this modulation in high level conducted at middle and low level where the clear incursion of the del JS from Subtropical latitude to tropical latitude help the incursion of cold mass from high levels low levels, this situation is reflected in the thickness analysis showed in fig. 2.a and 2.b where the dotted line is the thickness of 500/1000 hPa and the thick line is the isobar, in these it is observed the incursion of cold air to the north that occur on 29- 30 June 1200 UTC, this cold air advection is supported by the hydrostatic equation Holton, (1992), in the same way the anticyclone over the Pacific extends a light ridge to the northeast (near southwest of Peru) induced by high troposphere circulation patterns.

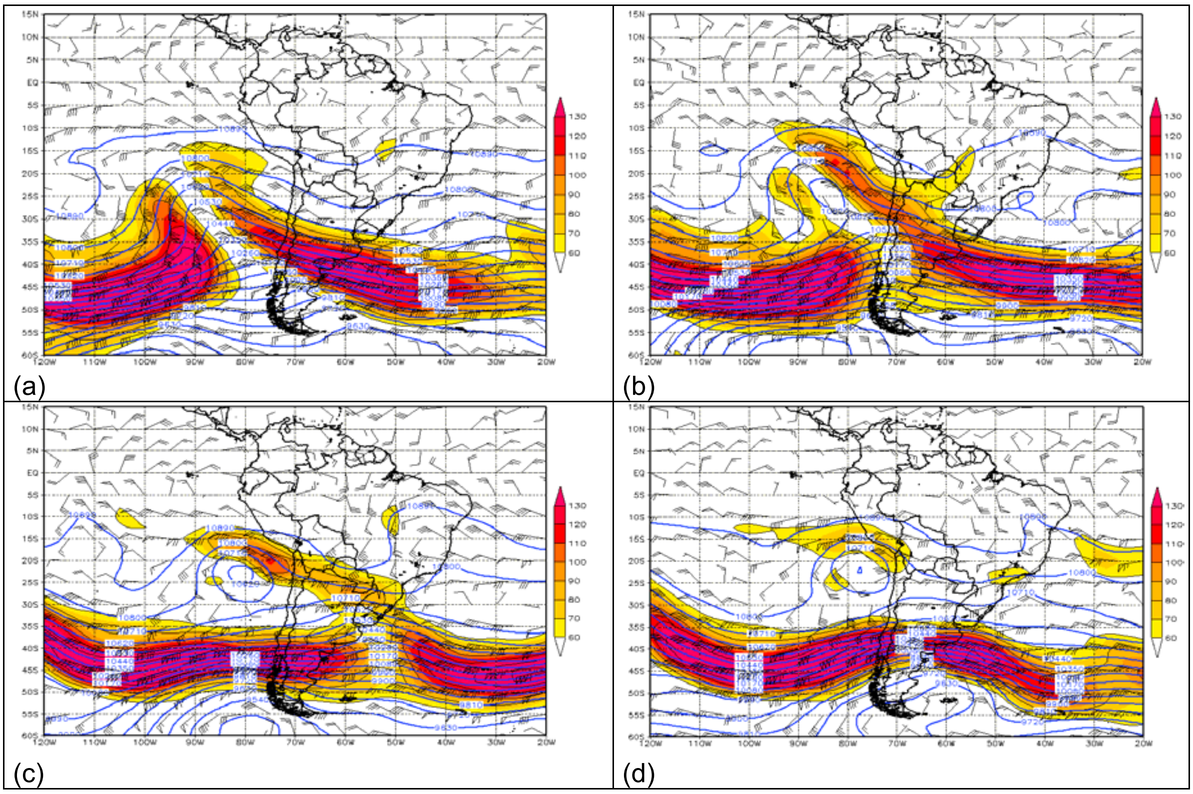


Fig. 1: 250 hPa height (solid, interval 90 m) and wind barbs and Jet stream (shaded, kt) for (a) 1200 UTC 29 June 2004 (b) 1200 UTC 30 June 2004 (c) 1200 UTC 01 July 2004 (d) 1200 UTC 02 July 2004

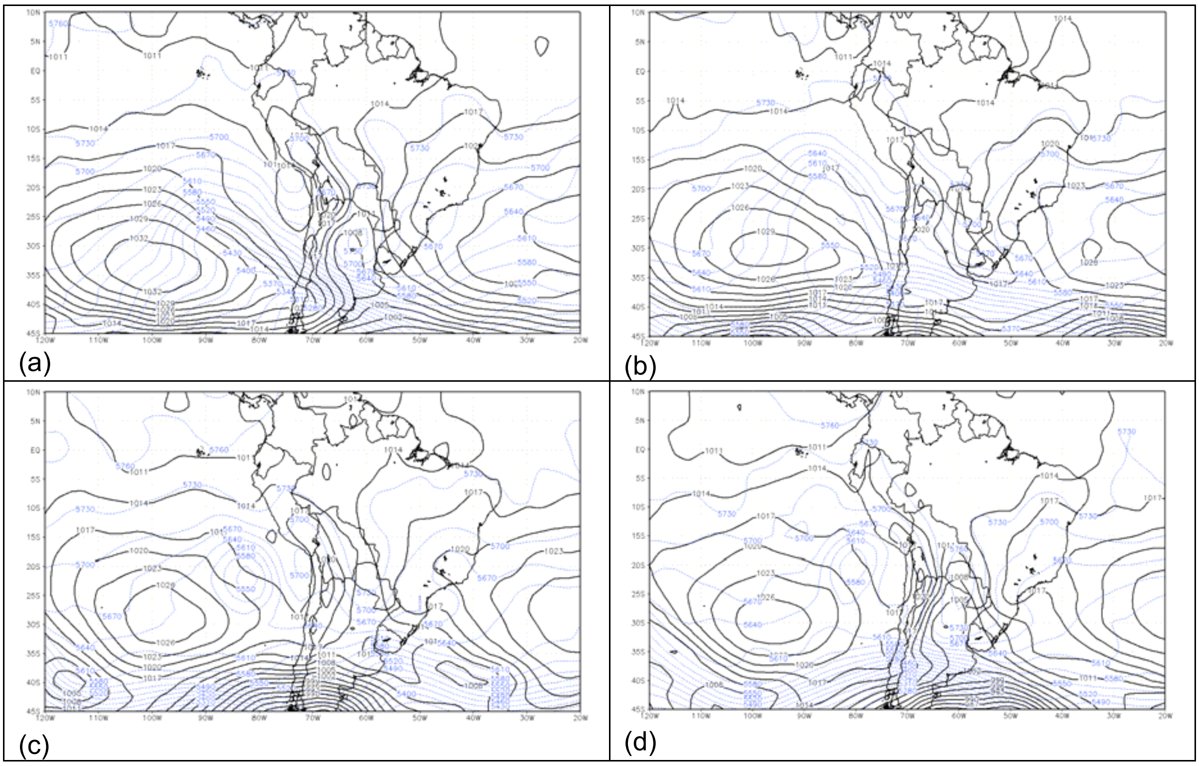


Fig. 2: Sea level pressure (solid, interval is 3 hPa) and 500/1000 hPa thickness (dasheds, interval is 40 m) at (a) 1200 UTC 29 june 2004 (b) 1200 UTC 30 june 2004 (c) 1200 UTC 01 july 2004 (d) 1200 UTC 02 july 2004

The northward displacement of the JS in the precursor phase induced the intensification of the wind vector rotational at high level, likewise, this vorticity increase in high level is supported by the vorticity equation (equation 1), where the geopotential change is proportional to vorticity change with an opposite sign, that is to say that the increase of vorticity corresponds to a decrease of the geopotential or it is understood as the cooling of the atmospheric system. Figure 3 corresponds to a cross section for June 29, 1200 UTC (precursor phase), it can be observe that the increase of relative vorticity (Blue lines dots) up to 300 hPa and a diminishing with a higher height, this change in cyclonic vorticity indicates the presence of a cold system (cold trough) supported in the vorticity equation. Another practical way to find cold conditions of the system is by observing isentropic that forms a bell-shaped curve between the trough position associated to the JS (in shaded).

$$\frac{\partial \zeta_g}{\partial t} = -\boldsymbol{Vg} \cdot \nabla(\zeta_g + f_0) - f_0 \nabla \cdot \boldsymbol{V} \qquad \text{.................ecu. 1}$$

Where $\zeta_g = \nabla^2 \phi / f_0$, ϕ is high geopotential, ***Vg*** is geostrophic wind

In the precursory phase we observed (fig. 1.a y 1.b), a trough deepness associated to southern intensification of JS, that is composed by the subtropical Jet Streak (JST) and North Polar Branch (JPN), following the methodology proposed by Nielsen (*Comet - Program*), such JST and JPN was identified in the isentropic separation 340 and 330ºK respectively (fig 3). These JST and JPN are associated to the displacement of short waves embedded in a long wave trough composed by the JS, similar characteristics were observed in the Northern Hemisphere by Bell y Bosart (1992).

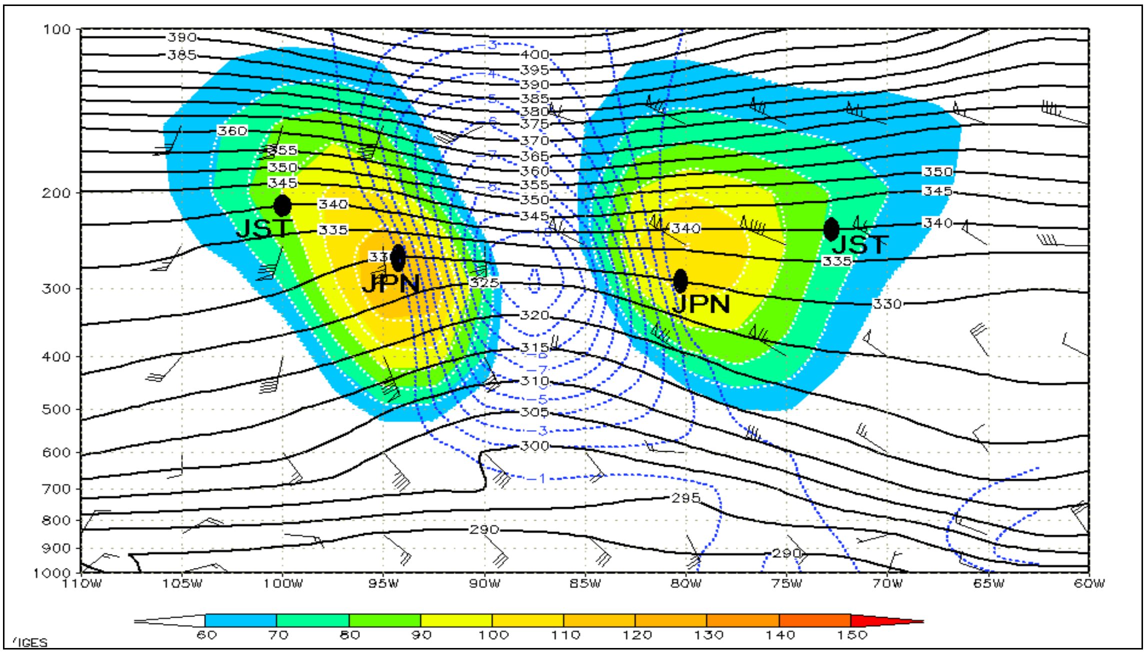


Fig. 3: West-East vertical cross sections along 30°S for 29 june 2004 (see Fig 1.b). Jet Stream (shaded, kt) and wind barbs and Potential temperature (solid line, kelvin) and relative vorticity (dots line, 1/s)

3.2 Formation phase

Once the trough is deepened west of Chile, a portion of JS accompanied by a JST and JPN it is separated from the total system forming a cut off cyclonic vortex with a nucleus between 23°S 82°W, this high level cyclonic formation reflects in middle levels with cool anomalies. In order to identify this cool situation it is observed the geopotential thickness of 500/1000 hPa (fig. 2.c and 2.d) with formation of closed nucleus of 5550 m of thickness, beneath the high level cyclonic vortex. At sea level pressure the effect is less perceived , showing a slight expansion of the ridge that the Pacific southeast anticyclon (APS) towards northeast of its former location (southwest of Peru) the following days the displacement of the COL is slow to the east and southeast and then by July 7-9 2004 it crosses the Andes (fig not shown).

3.3 Mean anomalies field

Fig 4 shows the geopotential anomalies for June 29 (precursor phase), where it can be seen a strong anomaly that exceeds 150 m , which extends from 10ºS towards high latitudes and from 130º W up to 90º W ; this configuration the entry a warm tongue of middle levels in the same way the negative anomaly fields expands to same latitudes and in the longitude of 90º W to 60º W. By June 30th (fig 4.b) the warm tongue expands southward from its initial location comprising 30º to 4º S favoring the formation of a separate cold nucleus (negative anomaly) towards north with values of –105m in the 20º S and 85ºW; whilst the positive anomaly keeps the same magnitude.

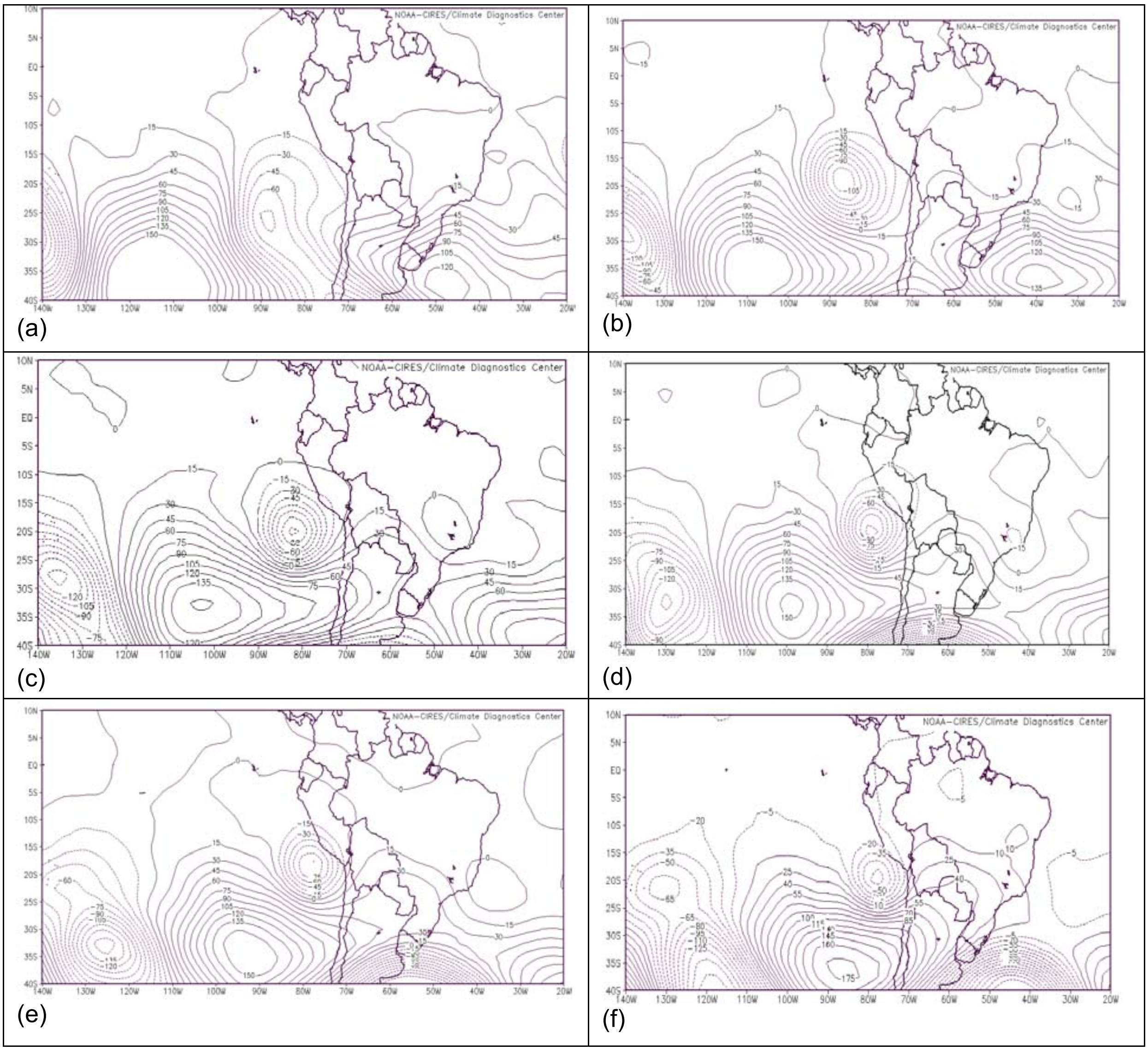


Fig. 4: Geopotential height anomaly (m) for (a) 29 june (b) 30 june (c) 01 july (d) 02 july (e) 03 july and (d) 04 july of 2004 at 500 hPa

By July 1st, (formation phase) the positive anomaly continues to expand eastward passing the Central Andes of Chile with an anomaly of 45 to 75 m; whilst the cold nucleus is totally isolated with a value of –90m (fig 4c) located between 20°S to 83°W, for such isolation configuration in its totality we considered the initial formation stage. The negative anomaly nucleus corresponds to the passage of a wave that presents a longer wave than usual, for which it can be concluded that the frontogenesis formation in high level and in tropical latitudes is directly associated to the passage of a long wave trough. Similar configuration was observed by Seluchi (1995) in the study carried out on synoptic situations of ciclogenesis formation east of South America.

By July 2nd the cold nucleus configuration keeps a value of –95m between 20°S a 80°W expanding towards the southern central part of Peru (fig 4d); whilst the warm anomaly over Argentina slightly diminishes; but not the positive nucleus over the southeast Pacific where it keeps an anomaly of +150m. Between the 3rd and 4th of July (fig 4e and 4f) the cold nucleus diminishes slowly presenting values up to –55m keeping its position at barlovent , this situation favors the sustained entrance of cold air over the Highlands of the southern region and for the next days the nucleus slowly moves towards southeast crossing at leeward between July 7-8 configuring a short wave trough over the continent (figure not shown).

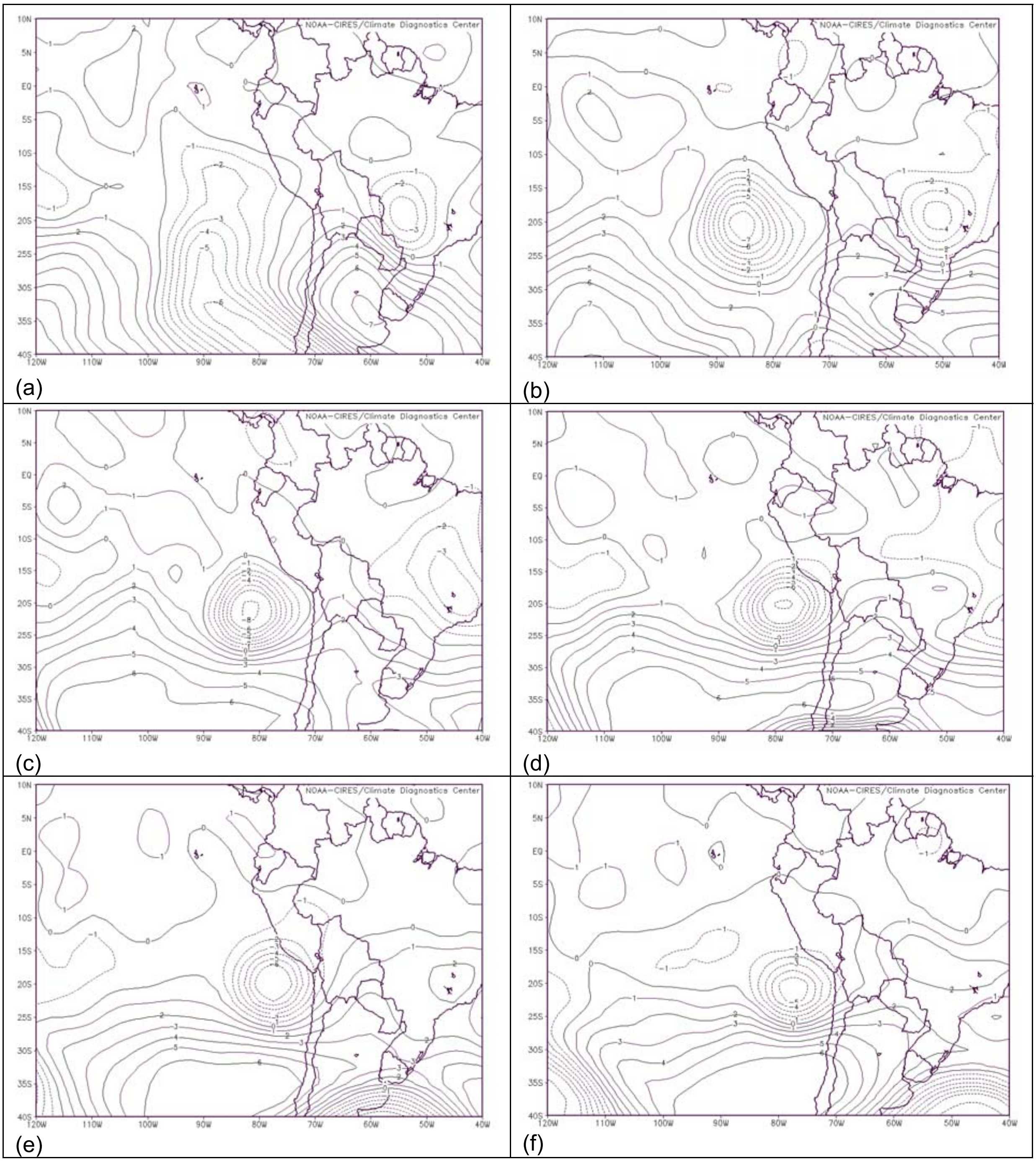


Fig. 5: Temperature anomaly (m) for (a) 29 june (b) 30 june (c) 01 july (d) 02 july (e) 03 july and (d) 04 july of 2004 at 500 hPa

The configuration pattern the trough-ridge was vital, in the formation of a COL, this situation of thermal variation of both systems was predominant in the intensification of the baroclinic region associated to the incursion of the JS at high levels. Likewise, it is evident the relation existing of warm and cold anomaly associated to the amplification and intensification of the trough ridge system.

In fig 5a, it can be seen the extension of a cold system (cold trough) along the southeast Pacific and its further isolation from a cold nucleus associated to the systems described above. The presence of a cold nucleus at barlovent kept the COL system approximately 6 days over the Pacific, getting to inject enough cold air over the Higlands of the southern region of Peru and the High Plateau.

3.3 Termodynamic structure

The evolution of COL has an stressed vertical structure in the troposphere, described by Bell and Bosart (1993). This behavior can be understood using the geopotential thickness at high, middle and low troposphere levels.

In high troposphere, the thermal evolution can be explained by the 100/500 hPa thickness (fig. 6a) in the temporary analysis in the central point of the nucleus of the COL in the 20°S 80°W for the precursory stage it maintains a slight variation until finishes this phase, then presents a mass expansion between July 1-3; this behavior is associated to the dominion of hot air in the trough axis, of high troposphere , compressing to the basement of the trough, and moving eastward, the point in analysis returns slowly to the initial condition on July 8th. This behavior can be explained by the mass divergence and convergence in the evolution of a closed circulation detailed by Bell and Bosart (1993); likewise it is explained by equation 1, represented in fig 2, that shows the decrease of negative vorticity from 300 hPa level up to 100 hPa that indicate an increase of geopotential height.

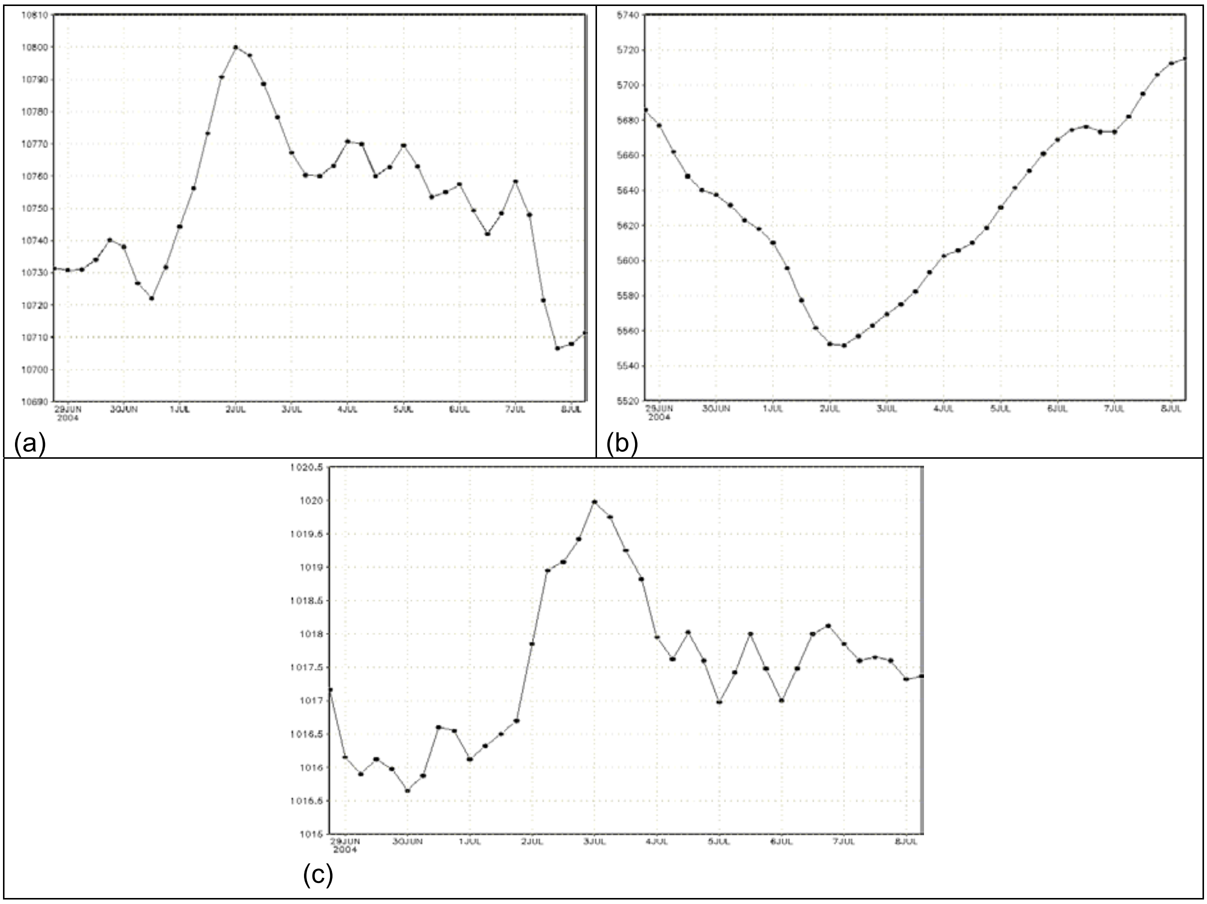


Fig. 6: Time series of (a) 500/100 hPa thickness; (b) 1000/500 hPa thickness; and (c) sea level pressure. Absissa indicates time (june and july, 2004) at 6 h intervals for 20°S 80°W.

In the middle troposphere, it is observed a cold nucleus identified by the geopotential thickness of 500/1000 hPa located beneath the cyclonic circulation axis of high levels (fig. 6.b). This nucleus of cold anomaly is compressed by the 100/500 hPa thickness, showing itself as a cold isolated system, the behavior of the structure under analysis is the evidence of the geopotential vorticity in high troposphere associated to the intensification of the trough.

In low troposphere, the analysis of sea level pressure presents different behavior from the one observed by Bell and Bosart (1993), showing slight variations in the precursory phase and increasing the atmospheric pressure an 4 hPa between July 1-3 2004 then decrease between July 3-5, and maintain in its daylight variability during the following days (fig 6.c and the sequences of fig. 2). This behavior can be explained by the presence of cold air in middle levels showing compression in low levels and extending in the ridge of APS towards northeast.

On the other hand, on previous days to the formation of the COL, it presented anticyclonic circulation patterns at high, medium and lower levels (300,500,700 hPa) over the continent with a nucleus located to the northeast of Bolivia (figure not shown), this circulation favored the advection of warm and humid air coming from the Atlantic and Amazonic Basin towards the Andes of Peru and Bolivia, whilst the contribution of humidity from the Equatorial Pacific was minimum. Garreaud (1999), described similar conditions when analyzing mechanism of precipitation production in the Central Andes.

In the temporary evolution of the averaged atmospheric humidity (mixture relation) between the 20-10ºS ant 600hPa level (fig 7a), it can be seen an increase of humidity between June 27 and July 1st, 2004, getting to occupy 75º W of the Central Andes and southern part of Peru, whilst for the second increase period from 1 to 8 July, it reached 80º W, thus the maximum nucleus, shown in fig 7a. present values of 2-3g/kg, on previous days to the occurrence of snowstorm over the mountains of Peru – Bolivia and on the day of occurrence of snowstorm it presents a maximum of 5 g/kg. In agreement to what Garreaud observed (2001); who says that it is necessary a minimum of 5-6 g/Kg near the High Plateau surface in order to have precipitation events. Taking into account the outcomes of Reanalysis (NCEP/NCAR) a 2.5 x 2.5° resolution represents a good indicator of high humidity over the region under study..

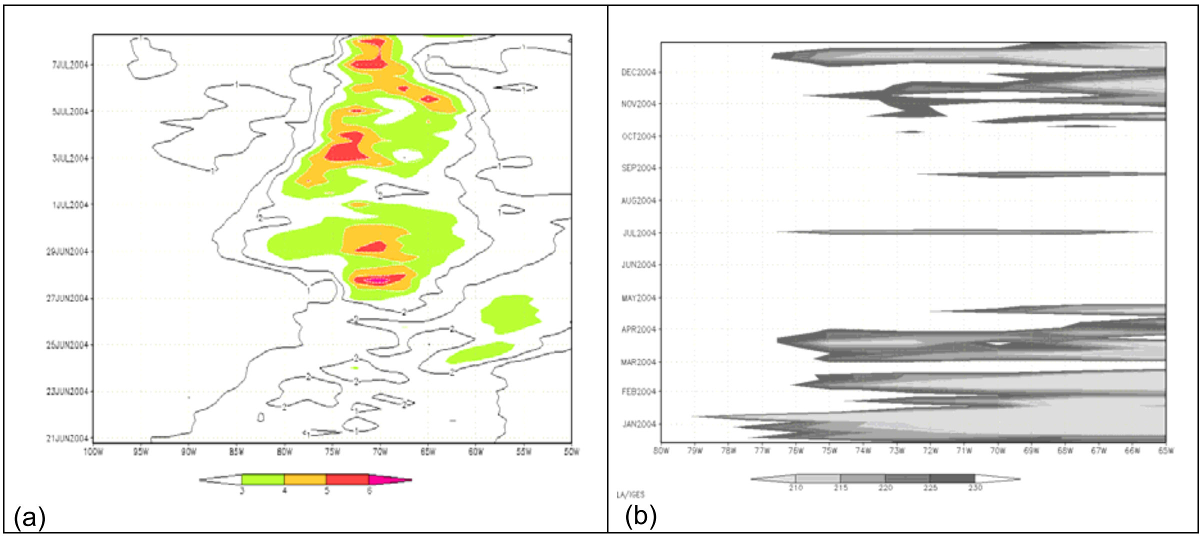


Fig. 7: Evolution (Hovmoeller diagram) of (a) mixture ratio (g/kg) 600 hPa (b) outgoing longwave radiation shading indicates OLR less than 230 W/m2. Average lat.: 20°S and 10°S.

The outgoing wave Resolution ORL identifies cloudiness coverage, different studies recognize ORL values equal or lower than 230 Wm-2 considered as the threshold of deep wet convection that derive from extreme clouds over the 300hPa. Garreaud (2001) considered to use a threshold of 220 Wm-2 due to the characteristics that the formation of convection presents at higher altitude. In the annual evolution showed in fig.7b it can be seen periods that exceed the threshold point, this behavior is a typical pattern of the rainy period in Peru; whilst in the 1st week of July 2004, it presented values that exceeded the threshold independently from the annual evolution. This behavior is associated to cloudiness coverage caused by the interaction of humidity advection coming from the eastern side and the entry of cold air mass from the Pacific caused by COL:

With regard to the stability and intensity mechanisms of winds within the system structure in analysis (COL), it is presented the sounding information (fig 8) for the Antofagasta station (northern of Chile). By July 1st the JS moves over the northern part of Chile with strong winds that exceed 120 knots over 200hPa., evidently diminishing at middle levels. Likewise, it is observed a storage of humidity over 500-300 hPa y due to the movement of the system towards the east, saturation decreases by July 2nd. It is evident the stability at lower levels (925-850 hPa), is associated to the intensification of the anti-cyclonic system on surface.

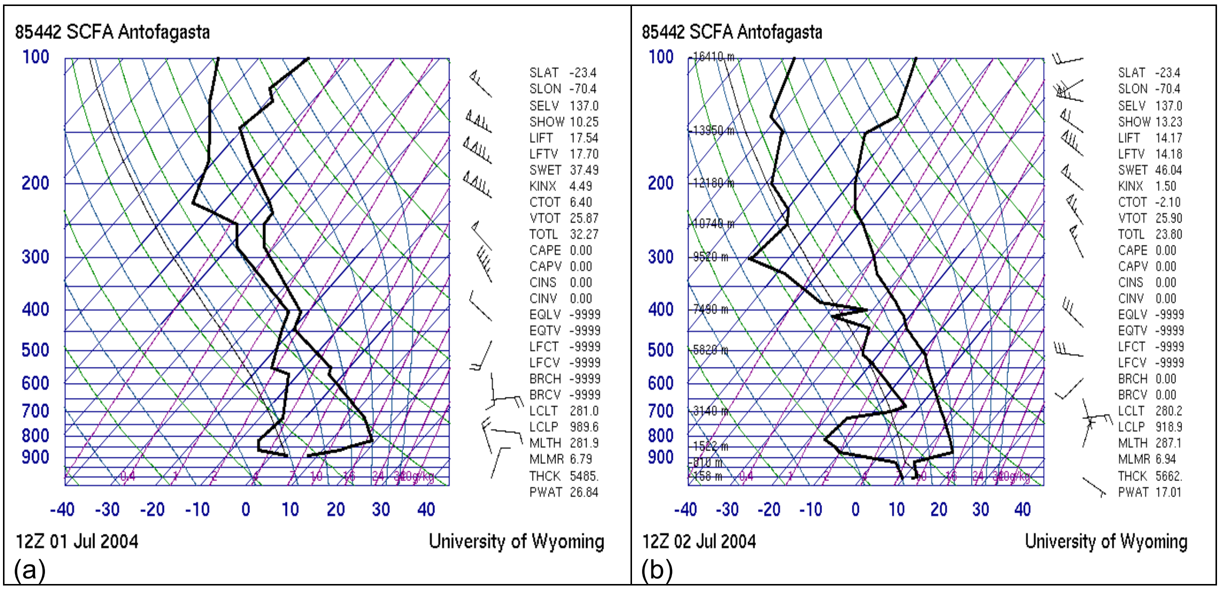


Fig. 8: Skew-T diagram (temperature solid, degrees Celsius; dewpoint temperature solid) and sounding winds (kt) for Antofagasta at 1200 UTC 01 and 02 july 2004.

3.4 Potential vorticity structure

Hoskins (1985), define the isentropic potential vorticity as

$$PV = -g(\zeta + f)\frac{\partial \theta}{\partial p}$$

……………………………………….ecu. 2

Following Bell y Bosart (1993), we will use the isobaric formulation of Ertel potential vorticity which can be written as:

$$PV = -g\left[(\zeta + f)\frac{\partial \theta}{\partial p} + \left(k \times \frac{\partial V}{\partial p}\right) \cdot \nabla_p \theta\right]$$

....Ecu.3

Where ζ is the relative vorticity, ***f*** is coriolis parameter, **g** is the acceleration due the gravity, **p** is the pressure, ∇_p is the horizontal gradient operator computed on the isobaric surface, V is the horizontal velocity vector and θ is potential temperature.

Hoskins et al., (1985), observed that on the surface of 1.5 - 2.0 units of potential vorticity (1PVU=-10^{-6}.m^{2}.K.s^{-1}.kg) that dynamic tropopause appears increasing quickly in the stratospheric air, thus also the WMO (1986), defines dynamic tropopause with a minimum of 1,6 PVU; whereas Nielsen-Gammon (1995), developed maps of bands of tropopause that are represented by values between 1,5 - 3,0 PVU. Following the works made by Hoskin (1985), Morgand and Nielsen-Gammon (1998) and Bell-Bosart (1993), the PV in the isobaric level of 300 hPa was determined using finite differences between the superior and inferior level (200 and 400 hPa); similar calculation was made for the inferior levels.

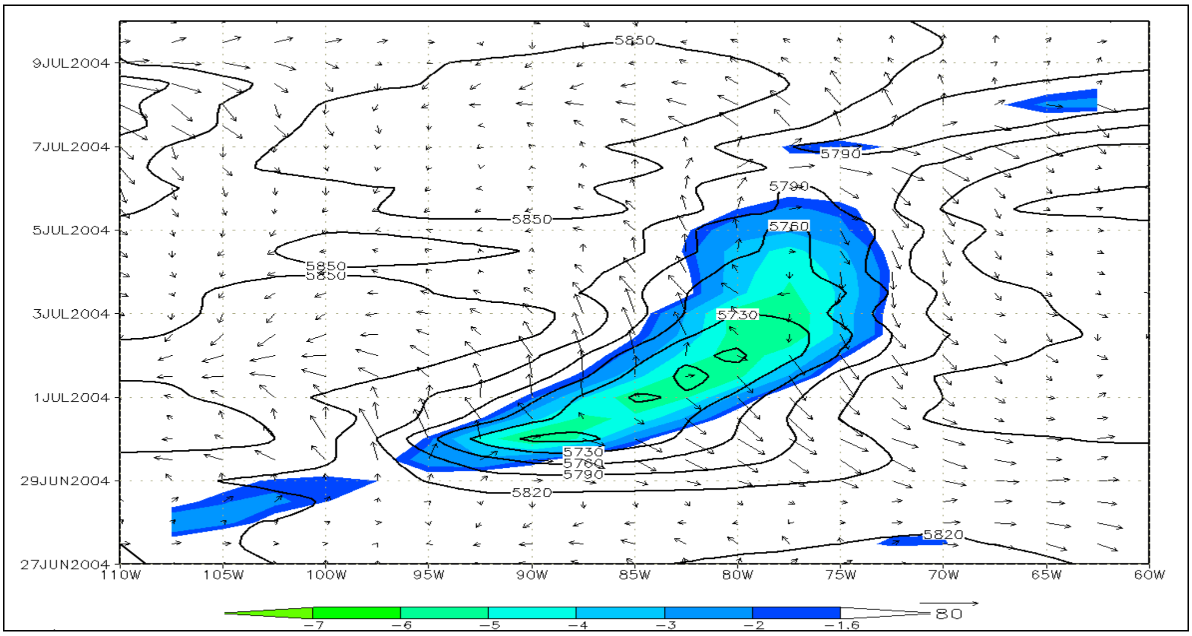
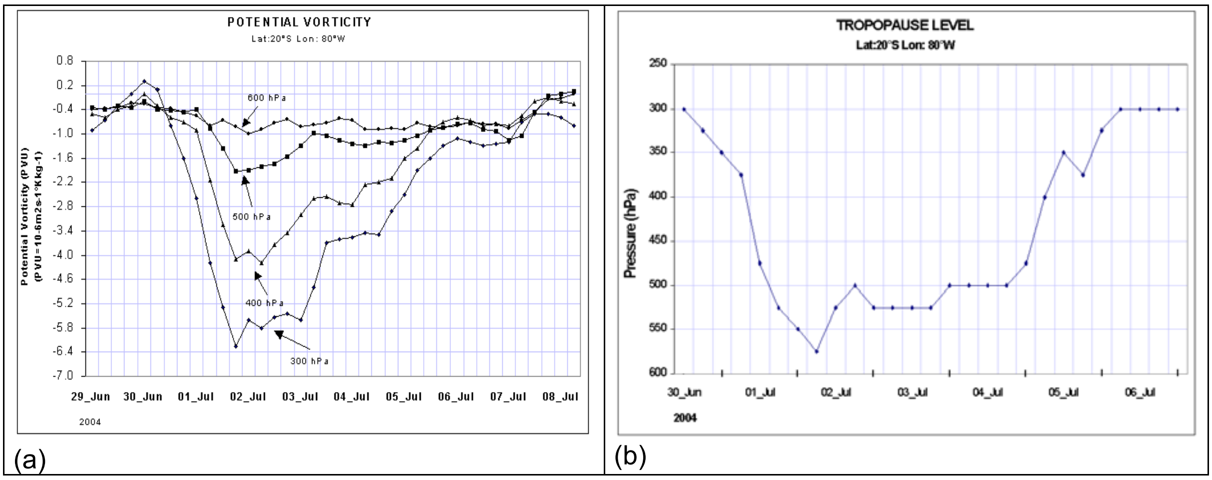


Fig. 9: Evolution (Hovmoeller diagram) of vorticity potential 300 hPa (shaded areas represent, PVU), geopotential height 500 hPa (solid line, m) and horizontal wind field at 500 hPa (vector, kt) for 20°S

In the temporary evolution it was observed and increased of the PV at high levels troposphere (300 hPa). This stressed increase is observed between 95-73ºW longitude at 20ºS since June 29 to July 5th (fig.9), in the same way beneath the maximum value of PV, it was formed a low pressure nucleus expressed in geopotential height (5730 m) at 500 hPa associated to strong winds that exceeded 50 knots, specially to the west side of the cyclonic system, that keeps alive circulation, likewise this cyclonic nucleus is explained by the deepness of a extratropical trough at high levels associated to Subtropical and North Polar Branch over tropical latitudes causing cooling from high levels to middle levels, causing the collapse of the tropopause due to the presence of a cold thermal anomaly in the middle tropopause.

Considering the dynamic tropopause as defined by WMO (1986) a temporary analysis was made in the central point of a COL (fig 8a) determining the PV in each isobaric level, same one that presents the maximum values between 2 –3 July 2004, with a maximum of 6.4 PVU at 300 hPa., and then it slowly diminishes towards July 6, when it presents values that do not represent the tropopause line.



(a) (b)

Fig. 10: Time series (a) maximun PV value within the developing trough for selected pressure level (b) level tropopause of the 1.6 PVU surface. Abscissa indicates times (june and july, 2004) at 6 h intervals.

The presence of a cold anomaly at middle levels caused the sloping of the tropopause, which is reflected in the increase of the PV from high levels to middle levels, this intrusion of stratospheric air at high and middle levels is review in figure 10b, where the limits of the tropopause line is determined at 550 hPa approximately by July 2nd, 0600 UTC 2004, this tropopause sloping sustained the geopotential thickness diminish from 500/1000 hPa allowing a greater entry of cold air at middle levels, which slowly came in through the southern Andes mountains of Peru favoring the occurrence of snowstorm.

3.7 The Q-vector analysis

Holton (1992), mentions that the Q-vector, it is physically interpreted taking into account the baroclinic movement, which is purely geostrophic so that the vertical speed vanish. Considering the equation Omega it is denoted by the following expression

$$\sigma \nabla^2 \omega + f_0^2 \frac{\partial^2 \omega}{\partial p^2} = -2\nabla \bullet \vec{Q}$$ Ecu.4

$$\vec{Q} \equiv \left(-\frac{R_d}{p}\frac{\partial \vec{V}_g}{\partial x} \bullet \nabla T, -\frac{R_d}{p}\frac{\partial \vec{V}_g}{\partial y} \bullet \nabla T \right)$$ Ecu.5

Where: σ: static stability, ω: omega, R_d: dry air constant, P: atmospheric pressure and Vg: geostrophic wind.

The ascent of air parcels is associated to the convergence of low and middle levels that exert heaves. In figure 11, it shows the geostrophic forcing term in the level of 400 hPa (Q-vector), this pattern of ascent is observed for days July 2 - 3 2004, day of occurrence of the snowstrom on the high zones of the south of Peru. The development of geostrofic forcing is associate directly to the evolution of the PV over the Eastern Pacific that favored the sloping of tropopausa and by compensation of thermal anomaly it stimulated the expansion of the tropospheric thickness (Hirschberg and Frisch, 1991) over the South region of Peru, thus also the divergent region of the entrance of the Jet Stream entered into phase. This conspiracy of the strong thermal gradient and the presented atmospheric flows favored the direct ageostrophic circulation (Uccellini and Kocin, 1987), organizing the frontogenesis of mean levels and high and then present heavy snow and snowstorm.

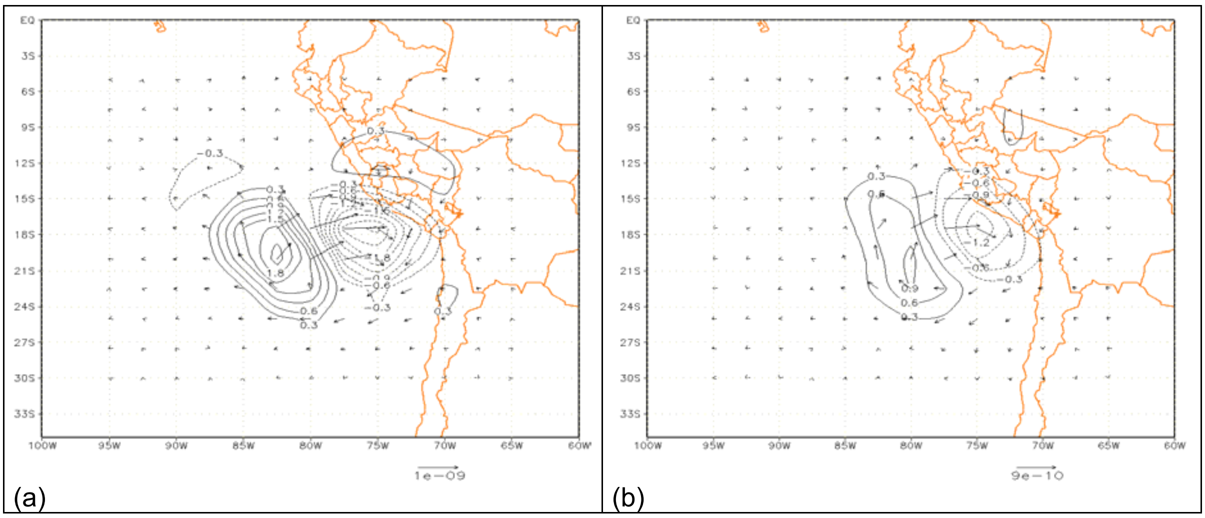


Fig. 11: Quasigeostrophic Q-vector (reference vector (located to right of figure title) length $1x10^{-10}$ Ks^{-1}) and Q-vector convergence (contour interval $0.3x10^{-15}$ K m^{-1} s^{-1}) for 1200 UTC 02 y 03 july 2004 400 hPa.

3.6 Satellite imagery

The satellite image of water vapour (fig 12.a) shows a band with gray tone forming a vortex-shaped southwest of Peru, which is associate to the formation of the COL, thus also a dark band is observed that waves, which is a clear indication of cold air entrance (see red arrow in fig 12.a). This contrast in colors its due to the slope of tropopause with presence of warm anomalies at high troposphere that contrast with the anomalies cold in mean levels and special in the side of the vertical shear, west side of the COL where the most intense wind appears, presenting a darker colors in the WV image.

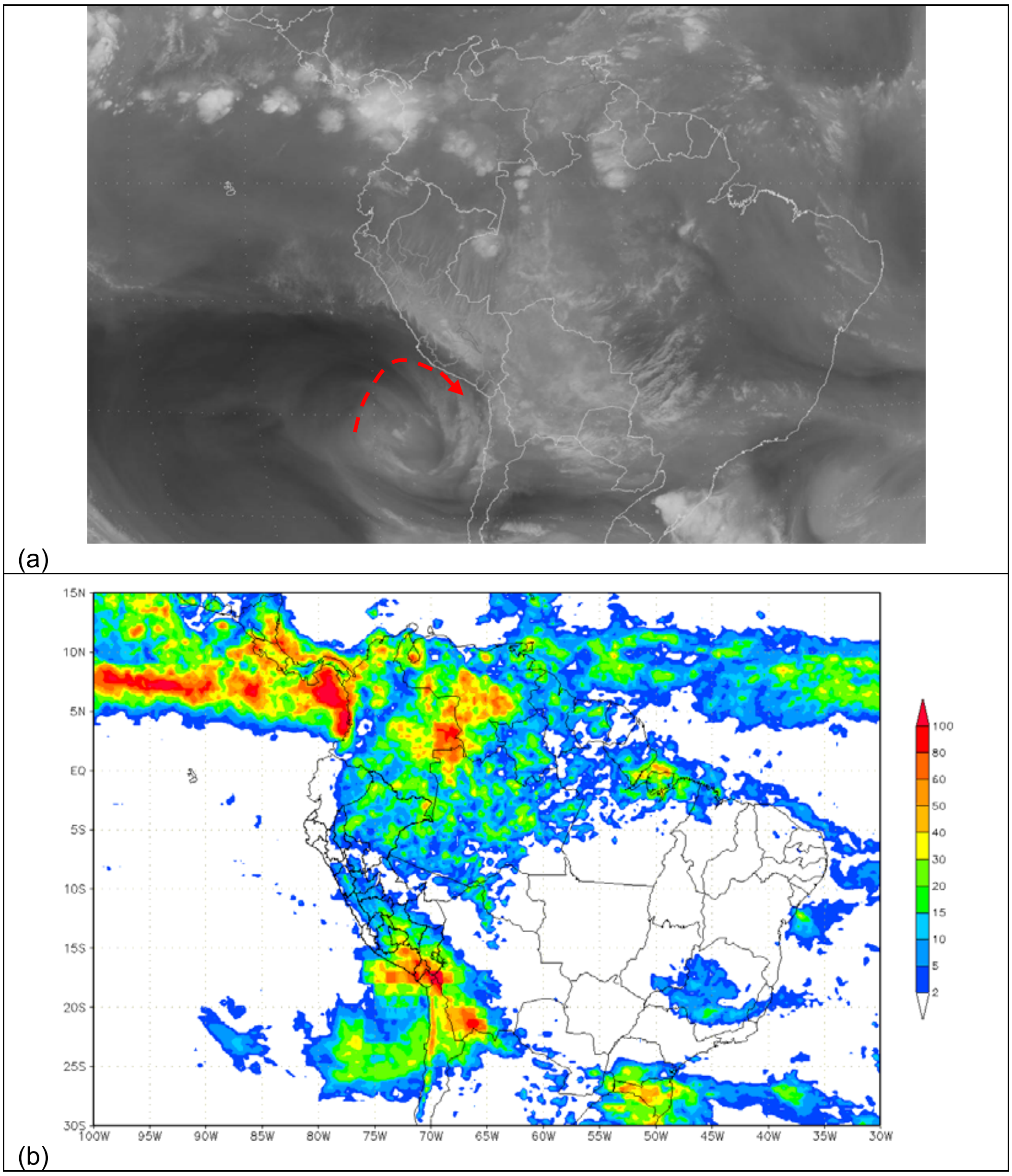


Fig. 12: The satellite analisis (a) Images from GOES 08, water vapor of 03 july 2004 valid at 1145 UTC (b) estimation precipitation (mm) on TRMM for total accumulation 29 june at 08 july 2004.

On the right side of the cyclonic vortex a speckled white strip is observed that extends from the central region of the Andes of Peru towards the west of Bolivia and north of Chile, this behavior is associate to the greater activity of cloudiness formation that expands towards high levels by compensation for sloping of tropopause generated by the presence of the COL (Hirschberg and Frisch, 1991).

The formation of cloudiness on the South region of Peru was associate to humidity advection coming from of the Amazonian basin which stimulated greater convective activity in the Intertropical Convergence Zone (ITCZ) on Colombia, Venezuela and north of the Peruvian forest; whereas the humidity contribution of the Eastern Pacific towards continent was minimum. In the course of the precursory phase and formation (29 June - 08 of July 2004), important liquid and solid precipitations took place and in special appeared in the South region of Peru, the west of Bolivia, north of Chile and in smaller intensity the Eastern Pacific in front of to the coasts of Perú and Chile, this behavior is reflected in figure 12.b that shows the precipitation estimated by Tropical Rainfall Measuring Mission (TRMM), according to the experience of the analyses made in the meteorological service of Peru, the results of the TRMM, provide good space configuration and approximate of 30 to 50 % of the real accumulated of precipitation over the Peruvian region.

4 Conclusions

The results of this investigation, suggest a fundamental pattern that one begins with the amplification of one ridge, in middle and high levels, positioned in the equatorial Eastern Pacific, which stimulated the deepening of a trough located southeast of such ridge. This trough was composed by the Subtropical jet and north polar branch jet embedded in the long wave trough that later entered to subtropical latitudes isolating from the original system as a cold bubble in warmer region, this formation is well-known as Cut Off Low Cyclon; also it was observed that the cold air entrance in middle levels of the atmosphere was stimulated by the incursion of the Jet Stream which increased the PV in high troposphere in the axis trough that was reflected in the sloping of tropopausa down to 550 hPa, in latitude of 20° S on the other hand the entrance of humidity coming from the Amazonian Basin was advected towards the South region of the Peru stimulated by the anticyclonic circulation of middle and low levels over continent (Brazil and Bolivia). This meterological conspiracy in tropical region derive into the formation of a frontogenesis and later the occurrence of snowstorm in high zones (above 3500 m) in the southern region of Peru.

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