Deep convection east of the Andes Cordillera: four hailstorm cases

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ABSTRACT
An analysis of four cases of severe hailstorms that occurred east of the Andes Cordillera in the northern part of Argentina is presented in this article. Analysed and observed data, as well as mesoscale model integrations, are used to evidentiate the underlying physical mechanism. The formation of heavy hailstones is the consequence of intense deep moist convection, with sufficiently high updraft speed to produce super-cooled water. The presence of warm and moist air is found. It appears to be a necessary but not sufficient condition to generate such intense convection. Convergence of moist enthalpy near the ground is also found. The passage of a cold front creates instabilities and causes upward motion on the warm side of the front, facilitating the development of deep convection. In some cases, low-level flow around a mountain creates wind convergence on the lee side, associated with lifting of air parcels. Although mountain waves are very frequent in the region, they do not appear determinant for the triggering of convection in the cases studied. It must be stressed that the results of this study are the consequences of the climatology of the region and cannot be trivially extrapolated to other regions.

Keywords: deep convection, storm, hailstone

1. Introduction
This study presents four cases of deep convection that occurred east of the Andes Cordillera in the northwestern part of Argentina. These events are analysed to put in evidence the physical mechanism driving them. In this region, in the period going from October to March, deep convection occurs frequently, sometimes producing heavy hailstones that reach the ground and produce considerable damage on cultivated areas. Nevertheless, not much study has been conducted from the physical point of view.

A full extent discussion on the physical process of moist convection in all its complexity would be overly ambitious and outside the scope of this article; a good review on the subject can be found, for example, in Stevens (2005). We will just make the simple consideration that one of the principal conditions that are required to generate deep moist convection is obviously the presence of a sufficient amount of moisture.

The height of the Andes is such that moisture cannot originate from the Pacific Ocean: crossing the Andes results in very strong drying of air masses penetrating the continent (Wang and Fu, 2004). In fact, the main source of warm and moist air for central South America is the low-level jet (LLJ; Stensrud, 1996; Nogues-Peagle and Mo, 1997; Li and Le Treut, 1999; Berbery and Collini, 2000; Salio et al., 2002). The LLJ consists of a low-level northerly circulation that is concentrated in a relatively narrow region located towards the east of the Andes Cordillera. The air comes from across the Amazon River Basin, which explains its high humidity. Nevertheless, as shown by Teitelbaum et al. (2008), the LLJ can weaken or even change its direction and as a consequence of the temperature decrease and moisture depletion – deep convection is suppressed. This happens when the South Pacific Anticyclone (SPA) is modified at its southern domain by transient synoptic systems, which can penetrate the continent and propagate towards lower latitudes arriving eventually to Mendoza (Garreaud, 2000). In some cases, when the LLJ has been interrupted, the moisture can come from a more easterly route, over Uruguay, Brazil, and even from the Atlantic Ocean due to the presence of the South Atlantic anticyclone (SAA). This moisture source can replace the LLJ and convection reappears. The approximate trajectory of these two possible moisture sources are illustrated in Fig. 1a, which also shows the geography of the region.

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In the climatology of this region, topography must also be taken into account: very high mountains can influence the appearance of convection. Convection over mountains has been the subject of numerous investigations. The initiation of cumulus clouds over mountains was studied numerically by Orville (1968), who showed that when evaporation is taken into account the upslope motion develops faster. Various sources of waves, including mountain waves, have been identified by Spiga et al. (2008) in South America. A mountain wave of large amplitude has been observed in the Antarctic Peninsula by Alexander and Teitelbaum (2007). In the case of a conditionally unstable flow but with a strong convective inhibition (CIN), horizontal convergence over the mountain weakens the inhibition and convection begins (Chen and Orville, 1980; Kirshbaum, 2011); additionally, mountain waves can start deep convection by lifting air parcels up to the level of free convection (LFC). The effects of the low-level flow around the mountains, which can create a convergence zone in its lee side, have been investigate by Lin et al. (1992), Hunt and Snyder (1980), Hozumi and Ueda (2005), and Yang et al. (2008). Horizontal mass convergence is hence an appropriate ingredient for forecasting convective initiation (Klupfel et al., 2011; Crook and Moncrief, 1988). A recent review on orographic effects on precipitation and clouds can also be found in Houze (2012).

On flat ground, moist enthalpy convergence (MEC) indicates the site where deep convection can start. One such case of convection initiated by MEC was shown by Ovarlez et al. (1996). The areal coverage of convection over flat ground also shows significant sensitivity to the details of the land surface, including land-use and soil moisture (Lanicci et al., 1987; Trier et al., 2004; Taylor et al., 2012, and references therein). Moisture convergence is also an essential ingredient of most convection parameterization in weather prediction or climate models. It is used as a predictor for triggering deep convection, or also in the closure schemes. A review can be found, for example, in Arakawa (2004).

A cold front in the region under consideration is sometimes regarded as a forecast of convection and storm. In fact, the passage of a cold front creates instabilities facilitating the development of deep convection (Crook, 1987; Joly and Thorpe, 1990; Schar and Davies, 1990; Trier et al., 2004; Siqueira et al., 2005; Seluchi et al., 2006; Cao et al., 2013). Additionally, temperature inversion weakening the CIN can occur during the passage of a cold front, allowing air parcels to rise up to the LFC.

At the microphysical levels, the formation of hailstones requires the presence of super-cooled liquid water (SCLW) in the cloud, which is possible when the updraft velocity is very strong. Rosenfeld and Woodley (2000) analysed high altitude measurements at the top of vigorous convective elements of cumulonimbus clouds obtained with aircraft measurements over Texas. An amount of SCLW of 1.8 g m$^{-3}$ was observed at temperature as low as $-37.5^\circ$C. With the same aircraft, Rosenfeld et al. (2006) observed SCLW droplets up to 4 g m$^{-3}$ for temperatures down to nearly $-38^\circ$C in the Mendoza region (west Argentina, on the east side of the Andes). The results obtained by Rosenfeld and Woodley (2000) were reproduced by a numerical cloud model by Khain et al. (2001). A case of SCLW, generated by mountains waves, was found by Heymsfield and Miloshevich (1993) along the Front Range of the Rocky Mountains. Another case of SCLW of orographic
The best-known process of growth of ice particles is the Bergeron–Findeisen process: when ice crystals fall into a super-cooled water cloud, they find themselves in an environment saturated with respect to liquid, which is supersaturated with respect to ice, the result is rapid growth of the ice crystals at the expense of water droplets. In large thunderstorms, ice crystals that collide with each other may stick together producing hailstones. According to Korolev (2007), the Bergeron–Findeisen process is just one of the three possible scenarios. The other two involve simultaneous growth or evaporation of liquid droplets and ice particles. The microphysics of hail formation has been intensively studied both theoretically and by laboratory experiments. The capture and freezing of super-cooled water droplets by falling ice crystals (riming) were quantitatively investigated in a vertical super-cooled cloud chamber of 10 m length at temperatures from $-5^\circ$C to $-25^\circ$C by Avila et al. (2009). They present evidence that ice crystals can start the riming process at smaller sizes than previously reported. Kay et al. (2003) tried to determine whether the homogeneous ice nucleation starts in the volume or at the surface of super-cooled water drops, but failed to reach a conclusive result.

In Mendoza, which is the region under consideration, the storm season extends from mid-October until the end of March. Rosenfeld et al. (2006) statistically studied this season during 4 yr (2000–2003). They found that storms in Mendoza are often severe, with 18% of 623 days with hailstorm and 8% with a hailstone diameter $>2$ cm.

The motivation of this study is to determine the factors involved in the process leading to these severe hailstorms, taking into account the specific conditions of the region. We analyse four cases between years 2009 and 2011 in which the storms were particularly strong with more than 70% destruction of crops. Following the methodology suggested by Doswell et al. (1996), we analyse the relative importance of moisture availability, MEC and passage of a cold front. As mountain waves propagate often in the atmosphere due to the orography of the region, we also study the role of mountain waves in the generation of the storms.

The remainder of this paper is organized as follows: in Section 2, we outline the methodology and the data used. In Section 3, a storm event is analysed in detail to identify the causes of the phenomenon. When mountain waves are present in the proximity (in space and time) of the region of the storm, their role in the generation of the deep convection is assessed. In light of what has been found in this first case, three other storms are studied in Sections 4–6. Section 7 summarizes and concludes.

2. Data and method

In this study, we use data from various sources. The analyses of the European Centre for Medium-range Weather Forecasts (ECMWF) provide an important source of information. The analysis has an original horizontal resolution at T213, which we reprojected on a $1^\circ \times 1^\circ$ grid. Vertically, there are 91 levels from the ground to 1 hPa. The analysis is provided every 6 hours, while a short range forecast can be used every 3 hours.

Since for this study we need higher spatial and temporal resolution, we also performed integrations of the Weather Research and Forecasting Model (WRF; Skamarok et al., 2005) in a three-domain nested version. WRF is a current generation mesoscale model that can simulate regional weather systems. Initial and boundary conditions are provided by the ECMWF analysis. The outer domain has a horizontal resolution of 18 km over a surface of 1800 $\times$ 1800 km. The first inner domain has 6 km horizontal resolution over a surface of 600 $\times$ 600 km, and the second inner domain a resolution of 2 km over a surface of 200 $\times$ 200 km. In all the cases analysed, the three domains are centred very near the site of the storm. The model is initialized 48 hours before the storm. Vertically, the model uses 125 levels from the ground up to 10 hPa. The time step is 30 seconds for the outer domain and 10 seconds for the innermost nested domain. The integration domain is centred on the site of the storm. The model uses convection parameterization for the outer and first inner domains, while convection is not parameterized in the second inner domain. The new Thompson microphysics scheme is used (Thompson et al., 2004; Hall et al., 2005), which simulates ice, graupel, snow and especially SCLW, which is essential in this study. The Thompson scheme misses an explicit parameterization of hail. This is a limitation because hail-containing downdrafts can be stronger than downdrafts containing only graupel and can lead to different cloud dynamics. Nevertheless, these effects should be negligible at the employed model grid size of 2 km. Figure 1b shows the outer WRF domain, including the orography and the sites where the four storms studied occurred.

To justify the use of maximum horizontal resolution of 2 km, we made a sensitivity test for one of the cases of study (the one of 12 February, see below). We tested cloud top height and maximum vertical velocity at the three WRF resolutions, plus additional integration at 1 km. The cloud top barely changes between the different integrations and – as will be shown below – it is practically the same as that obtained by satellite measurement. The vertical velocity, as expected (see, e.g., Miyamoto et al., 2013), increases with increased resolution. In the outer domain, the maximum upward velocity is 5 m s$^{-1}$; in the first inner domain it is 25 m s$^{-1}$ and in the second inner domain
The vertical velocity shown in Fig. 2d has a maximum of more than 40 m s⁻¹ between 9 and 11 km height. Such strong updrafts are not unusual in cases of deep convection. From the data obtained by an airliner, Lane et al. (2003) deduced a peak of vertical velocity of 40 m s⁻¹, with an uncertainty of about 30%. Musil et al. (1991) penetrated a hailstorm in southeastern Montana and found vertical velocities higher than 50 m s⁻¹. Rosenfeld et al. (2006) from measurements made by aircraft in Mendoza found updrafts of 40 m s⁻¹.

The high horizontal resolution obtained with WRF is a considerable advantage here: areas of high speeds can be resolved even when they have a very small scale. As an illustration, Fig. 2e shows a latitude/longitude cross-section of vertical velocity at 10 km altitude. The structure shown has a dimension of the order of 5–10 km across; a lower resolution averaging over a larger area would indicate a much lower vertical velocity. The figure shows an irregular zone where the vertical velocity is negative. The strong convective updraft is accompanied by downdrafts. The maximum downdraft speed is found 10 km to the south of the updraft. In Fig. 2f, the vertical speed profiles at the location of the maximum updraft and downdraft are shown. The maximum updraft speed (43 m s⁻¹) is found at 10 km altitude, while the maximum downdraft speed (−11 m s⁻¹) is at 14 km altitude. The fact that the maximum downdraft is at higher altitude than the maximum updraft is a robust feature of land convection (Heymsfield et al., 2010).

According to Waldstreicher (1989), moisture flux convergence in the boundary layer is a predictor of deep convection and even more so MEC which involves temperature rise. We diagnose here the MEC as

$$\text{MEC} = -\text{div}\left\{ \rho \left[ (c_{pd} + r_c) T + L_v r \right] \right\}$$

where $c_{pd}$ is the heat capacity at constant pressure for dry air, $r_c$ the total water mixing ratio, $c_l$ the heat capacity of liquid water, $T$ the temperature, $L_v$ the latent heat of vaporization, and $r$ the mixing ratio. Figure 2g shows that at the site of the convection, at low levels (2.5 km), there is a very localized region where the MEC reaches values as

(2 km) it is 45 m s⁻¹. At 1 km resolution, the vertical velocity attains saturation, the maximum value being 47 m s⁻¹.

Another dynamical tool used is FLEXTRA, a Lagrangian kinematic trajectory model that computes different types of air parcel trajectories (three-dimensional, isobaric, isentropic, etc.) to determine the origin of air masses (Stohl et al., 1995; Stohl and Wotawa, 2005). The time step of the FLEXTRA is continuously adjusted, convergence being assumed when the trajectory position between two subsequent iterations differ by $< 10^{-5}$ times the dimension of a grid point.

To analyse the occurrence and characteristics of convective systems, we also need to resort to satellite measurements. We determine the temperature at the top of clouds, a key diagnostics of convection intensity, from observations provided by the Service d’Archivage et de Traitement Météorologique des Observations Satellitaires (SATMOS, Sèze and Desbois, 1987). The system was developed by Météo-France and the Institut National des Sciences de l’Univers and provides an analysis of Geostationary Operational Environmental Satellite satellite data. The temperature at the top of the clouds together with temperature profiles from the WRF simulations is used to estimate the altitudes of the clouds.

S-band radar imagery is provided by the Servicio Meteorologico Nacional de Argentina. It indicates the presence of hailstones when the reflectivity is $> 57$ dBZ (Gosset et al., 1992; Férał et al., 2003). The day, the place and the percentage of destruction of crops can be obtained from maps of damage, provided by the Dirección de Agricultura y Contingencias Climáticas of Mendoza.

3. The storm of 2 February 2010

3.1. Deep convection

On 2 February 2010, between 22 and 23 UTC, an intense hailstorm lashed the district of Rivadavia (Mendoza). The radar image (Fig. 2a) shows that at 22:40 UTC, there are areas where the reflectivity is $> 57$ dBZ, indicating the presence of hailstones. The most affected site is located at 68°W, 33.3°S (La Central), where the destruction of crops reached 70%. To describe the origin of this phenomenon, we analyse the local conditions near the site and around the time of the event. This is done with the help of the WRF model, using the innermost domain of integration. The horizontal grid spacing of the second inner domain, 2 km, is a resolution similar to the 1.5 km resolution used by Wang and Liu (2009), which they found accurate enough to resolve deep convective clouds.

The results shown in the figures, unless differently indicated, come from the WRF second inner domain.
Fig. 2. Storm of 2 February 2010: (a) radar reflectivity (dBZ). At 22:40 UTC, the maximum reflectivity is found over La Central (68°W, 33.3°S). (b) Longitude/altitude cross-section (33°S) of relative humidity (colours) and temperature (°C, black lines). (c) Cloud top temperature (K). (d) Longitude/altitude cross-section (33°S) of vertical air velocity (m s$^{-1}$, colours) and temperature (°C, black line). Panel a is provided from the Argentinian Meteorological service, panels b and d are issued from the inner domain of the WRF integration and panel c is issued by Service d’Archivage et de Traitement Météorologique des Observations Satellitaires. See text for details. All figures are at 23 UTC. These and all following figures, unless specified differently, are obtained from the highest resolution WRF integration and show the whole inner domain. (e) Longitude/latitude cross-section of vertical air velocity (m s$^{-1}$) at 10 km altitude. (f) Vertical profiles of vertical velocity (m s$^{-1}$) at the locations of the maximum updraft and downdraft. (g) Longitude/latitude cross-section of MEC (W kg$^{-1}$, colours) at 2.5 km altitude and horizontal wind vectors (m s$^{-1}$) at the same altitude.
high as 1400 W kg\(^{-1}\). Note that the MEC, although it is a predictor of deep convection, as stated, can also be amplified at the site of convection by the local increase of convergence. However, the horizontal wind convergence seems to be of larger scale than the location of the convection. This shows that the wind convergence is not a consequence of the convection, but rather a factor that causes it.

This high concentration of MEC requires horizontal mass convergence and moisture supply. In the present case, the convergence can be seen in Fig. 3a, which shows surface pressure and wind at 850 hPa obtained from ECMWF forecast on 2 February at 21 UTC. We use the wind at 850 hPa because this level is assumed as representative of the maximum wind level related to the LLJ (Marengo et al., 2004). In fact, the LLJ can be seen in this figure, contributing to the formation of wind convergence at the site marked with a cross, where the deep convection occurs. To identify the moisture supply, we use the FLEXTRA kinematic trajectory model. The result is shown in Fig. 3b, which displays a set of 3-D 6-d back-trajectories. The back-trajectories have been calculated as reaching the site of the convection at 3 km altitude on day 2 at 23 UTC, the arriving point is slightly perturbed from one trajectory to the other. Together with the trajectories, the figure also shows the distribution of specific humidity near the ground on 28 January at 00 UTC. It can be seen that the air parcels arriving at the site of the convection come from a high humidity zone. Most trajectories by and large coincide with the LLJ, which transports warm and moist air from the Amazon River basin towards central South America (Marengo et al., 2004).

Fig. 3. Storm of 2 February 2010: (a) ground pressure (hPa, colours) and horizontal wind vectors (m s\(^{-1}\)) at 850 hPa obtained from ECMWF forecast at 21 UTC. (b) 3-D 6-d back-trajectories starting from the convective area at 3 km altitude, and specific humidity (g kg\(^{-1}\), colours) near the ground on 28 January at 00 UTC. (c) Longitude/altitude cross-section (33°S) of SCLW (g m\(^{-3}\), colours) and temperature (°C, black lines).
As already mentioned, as a result of deep convection, and especially when vertical speed is very fast, SCLW sometimes appears. This is what happens in this case as seen in Fig. 3c, which shows a longitude/altitude cross-section of SCLW at 33°S. The presence or absence of SCLW is a critical point in this study. Down to −10°C the SCLW is higher than 2 g m⁻³, at −30°C it is of 1.2 g m⁻³, and then below −39°C there is a sharp drop indicating that we are in the presence of homogeneous freezing (Wood et al., 2002). Such high SCLW represents favourable conditions for the growth of large hailstones (Rosenfeld and Woodley, 2000).

During its lifecycle, the storm travelled eastward; this is illustrated in Fig. 4a-c, where the MEC and the horizontal wind at 2.5 km altitude are shown. It can be seen that the storm (localized by the MEC maxima) covered around 1° of longitude between 22 hours of day 2 and 00 hours of day 3. To compare this displacement with the simple advection by the horizontal wind, Fig. 4d displays three-dimensional forward trajectories of air parcels starting at 22 hours from the location of the cloud at that time. The trajectories were calculated with the FLEXTRA code starting at five different heights. All trajectories are shifted southward with respect to the cloud locations; the storm is hence not simply advected by the wind. What happens is that the conditions of generation are shifted by the passage of a cold front. This is shown in Fig. 4e-g, where the front is identified by the magnitude of the temperature gradient. In each figure, a cross indicates the position of the convective column.

A skew T–log p diagram obtained from WRF outputs at 22 hours is shown in Fig. 4h. The CAPE computed for this case is as high as 1220 J kg⁻¹. Also, the figure shows that the level of neutral buoyancy (LNB) is at high altitude, near 200 hPa. The front lifting allows air parcels to reach LFC.

3.2. Mountain waves

We use a 3-D field of vertical velocities provided by the second inner domain of WRF model to identify the presence of mountain waves, to determine their influence in the generation of deep convection and storms. By visual inspection of vertical velocity maps, we search for a maximum over a mountain slope at a distance no more than 200 km from the storm, and at the same moment. On 2 February 2010, at 23 UTC, the higher vertical velocity over a mountain slope is found at 69.5°W 34.5°S.

Figure 5a shows the latitude/altitude cross-section of the vertical velocity along with the temperature isolines. The vertical velocity has a maximum of 12 m s⁻¹ at 5 km altitude, and the isoline of zero temperature shows oscillations characteristic of mountain waves. There are fundamental differences between the structure of the vertical velocity when it is the result of deep convection as shown above and when it is produced by a mountain wave. In the first case, we can see a positive velocity corresponding to the updraft and a negative velocity in the vicinity corresponding to the downdraft (compare Fig. 2e). In the case of a mountain wave, we can see a succession of alternating sign velocities, which follow the fluctuations of the zero-temperature isoline. These features are sufficient to distinguish between the two phenomena. The longitude/altitude cross-section of the relative humidity (Fig. 5b) shows a small cloud at 10 km altitude. This type of small cloud is a feature often encountered due to the presence of mountain waves. In Fig. 5c, which is a zoom around 8 km height of temperature and water vapour, we can see that they are modulated by the vertical transport induced by the mountain wave.

To understand the nature of the small cloud seen in Fig. 5b, we use the vertical profiles of temperature (T), pressure (p) and water vapour pressure (e) issued from the WRF integration to compute the effect of an adiabatic uplift. We select a longitude where there are no mountain waves (i.e. 70.6°W). Taking into account the values at 11 and 12 km, we can calculate the temperature that a parcel located at 11 km would have if it had to be moved adiabatically up to 12 km. This is given by

\[
\frac{T_{12}}{T_{11}} = \left(\frac{P_{12}}{P_{11}}\right)^{\gamma}
\]

where \(\gamma\) is the adiabatic coefficient, and the suffixes 12 and 11 indicate temperature and pressure at 12 and 11 km, respectively. At this point, using the following approximate expression:

\[
\ln(e^\theta) = 23.33086 - \frac{6111.72784}{T} + 0.15215\ln(T),
\]

we can compute the saturation vapour pressure value (in hPa) that the parcel would have after the adiabatic expansion. The saturation vapour pressure before the uplift from 11 to 12 km is 0.055, while after the uplift the formula gives a value of 0.017. This is very close to the actual vapour pressure 0.016, given independently by WRF at the site of the mountain wave. This result shows that the cloud is created by adiabatic lifting due to the mountain wave without contribution of humidity from lower layers.

To explain why mountain waves like this are not likely to produce deep convection, we shall analyse some details of the mountain climatology in the region. Figure 6a displays the wind vector (zonal–vertical) on a longitude/height cross-section. Note that the ratio between the components, and the form of the mountains, are adjusted following the ratio between the longitude and the altitude of the plot.
Fig. 4. Displacement of the storm of 2 February 2010. MEC (W kg\(^{-1}\), colours) and wind vectors (m s\(^{-1}\)) at 2300 m on (a) day 2 at 22 UTC, (b) day 2 at 23 UTC, (c) day 3 at 00 UTC and (d) forward trajectories starting from the place of the storm at five different altitudes on day 2 at 22 UTC. Storm of 2 February 2010: passage of a cold front. Magnitude of the temperature gradient (K km\(^{-1}\), colours) at 3 km height on day 2 at 22 UTC (e); on day 2 at 23 UTC (f) and on day 3 at 00 UTC (g). (h) CAPE (j kg\(^{-1}\)) obtained from WRF model and the \(T_p = \log p\) diagram.
The zonal–vertical component of the wind vector is parallel to the slope of the mountain except at the longitude where the vertical velocity increases abruptly and where the mountain wave is generated. Over the lee side of the mountain, the air descends and is adiabatically heated as shown in Fig. 6b, (Durran, 2003; Gaffin, 2002). Sometime the wind becomes extremely dry and warm by the Foehn effect (zonda wind in Argentina; Seluchi et al., 2003).

Much of the airborne moisture falls as rain on the windward side, near the top of the mountains. This often means that the land on the other side (the leeward side) is drier: an effect called ‘rain shadow’ typical of northeasterly situations in over Argentina. The higher the mountain, the more pronounced the rain shadow effect is. The same thing occurs in the case studied. In fact, as shown in Fig. 6c, on the windward slope of the mountain the relative humidity decreases with height although saturation vapour pressure also decreases due to the decrease in temperature. This is possible because the total amount of water is depleted along the windward slope (Fig. 6d) by rainfall, as shown in Fig. 6e. Therefore, at the site where the mountain wave is generated the moisture is insufficient and vertical velocity is weak to provide the conditions of generation of deep convection and super-cooled water.

Another more intense mountain wave appears near the same site at 22 UTC. Figure 7a and b shows the vertical wind and the relative humidity. The vertical wind has the characteristics of a mountain wave: a succession of alternating sign velocities with a maximum of 14 m s$^{-1}$ There is an increase in relative humidity at 11 km altitude but not enough to produce condensation. The mountain wave is far from where deep convection occurs. Figure 7c shows six forward trajectories starting at 22 UTC from the site where the mountain waves occurs; all the trajectories go southward, while the convective phenomena at 22 and 23 UTC are located to the north. As expected, there are no traces of SCLW. Unlike deep convection, the mountain waves are not effective in producing a major storm and fall of hailstones.

In conclusion, the analysis of local and regional conditions shows that the storm is mainly the consequence of (1) the presence of the LLJ, which provides warm and moist air, and (2) the passage of cold front, which uplifts these warm and moist air creating instabilities. These conditions are independent of the mountain waves, which play no role in the generation of the storm.

4. The storm of 12 February 2010

4.1. Deep convection

On 12 February 2010, at around 23 UTC, there were at least two severe storms that hit the region of General...
Alvear (67.7°W, 35°S), which destroyed 2300 ha of cultivated areas, including 300 ha where over 80% of crops were lost. We show some details of this storm here.

The radar image (Fig. 8a) shows a reflectivity greater than 57 dBZ over the region, indicating fall of hailstones. Fig. 8b displays relative humidity and temperature in a longitude/altitude cross-section. The top of the cloud reaches an altitude of 14 km. The temperature of the cloud top obtained from SATMOS (Fig. 8c) is between 205 and 210 K. This temperature compared with the WRF temperature profile gives practically the same altitude. Figure 8d shows the vertical wind in a longitude/altitude cross-section: there is a very strong convective updraft reaching 45 m s$^{-1}$ at 11 km. Figure 8e shows a convergence zone on the site where deep convection occurs (marked by the cross). The mechanism that can create the wind convergence locally is the low-level flow around a mountain, which generates convergence in the lee side (Hozumi and Ueda, 2005; Yang et al., 2008). This phenomenon is observed in the present case. Figure 9a shows the orography and the wind direction. A mountain over 2000 m high (marked by the star) is situated immediately to the west of the site where the convection takes place (the cross), a convergence area is visible in its proximity.

Figure 9b shows three locations where the MEC is greater than 1000 W kg$^{-1}$, including the site where the storm takes place (the cross). Over the other two, there are also convective columns (not shown). These three sites are...
nearly aligned with a cold front, shown by the magnitude of
the temperature gradient (Fig. 9c) and by the increase of
water vapour on its warm side (Fig. 9d). The longitude/
altoititude cross-section of SCLW is shown in Fig. 9e. A maxi-
mum of 1.6 g m$^{-3}$ found at 9 km where the temperature is
$-20\degree C$ is the result of the strong convection and the cause
of the hailstone formation. The $T_p-\log p$ diagram plotted
for the site of the convection 1 hour before (Fig. 9f) shows
the LNB near 200 hPa and a CAPE of 1730 J kg$^{-1}$, which
explains the high intensity of the updraft. Due to the
convergence of the low wind entering the mountain,
the passage of the cold front and high value of CAPE,
the conditions for the generation of deep convection are
fulfilled.

4.2. Mountain waves

As for the preceding case, we search for the presence
of mountain waves. At the same time as the convection
event there is a wave on the lee side of the mountain at
69.45$W$, 34.45$S$. Figure 10a and b shows longitude/
altoititude cross-section of the vertical wind and relative
humidity, respectively. The vertical speed shows a sequence
of upward and downward velocities with maxima of
6 and 4 m s$^{-1}$. The relative humidity indicates the presence
of a cloud at an altitude of around 12 km, and very low
relative humidity at the ground where the mountain wave
is generated. As expected, there are no traces of SCLW:
the mountain wave plays no role in the generation of this
storm.

5. The storm of 17 December 2009

On 17 December 2009, 2009a storm swept the region of San
Rafael with extremely serious consequences. The destruction
of crops exceeded 90% in some areas. The fall of hail
was accompanied by strong winds and rain that resulted in
deaths and injuries. The storm was particularly intense
around 20:30 UTC: in Fig. 11a, a radar reflectivity greater
than 57 dBZ is shown, indicating fall of hailstones in the
locality of Rama Caida (34.7$S$, 68.4$W$). Figure 11b and c
shows longitude/altitude cross-sections of relative humidity
and vertical velocity, respectively. The top of the cloud
reaches a height of 14 km, and the vertical wind has a
maximum of 40 m s$^{-1}$ at 10 km altitude.

The shape and place of the SAA (Fig. 11d) differ from
the previous cases and suggest that the warm and moist air
comes from the east, instead of being provided by the LLJ.
In fact, the 6 d back-trajectories (Fig. 11e) show that humidity comes from Uruguay and Southeast Brazil. Figure 12a shows several maxima of MEC, with values higher than 400 W kg\(^{-1}\), particularly at the site of deep convection (marked by the cross), where the value is above 800 W kg\(^{-1}\). The whole area of high MEC is a wind convergence zone: this is due to the presence of a front, visualized in Fig. 12b as the magnitude of temperature gradient near the ground. Figure 12c displays the SCLW over the site of the deep convection. There is a maximum of 1.6 g m\(^{-3}\) above the 0°C and a sharp drop near −40°C.

On that day, there are mountain waves, but their induced velocity is less than 3 m s\(^{-1}\), so that their role can be excluded. In this case, the warm and moist air comes from Uruguay and southern Brazil as shown by the back-trajectories. Its convergence and the passage of a front explain the generation of deep convection.

6. The storm of 22 February 2011

On 22 February 2011, at 20:30 UTC, hail fell again in a cultivated area near San Rafael at 34.5°S and 68.3°W. The radar reflectivity on the site of the storm is greater than 57 dB (Fig. 13a) indicating the formation and fall of hailstones.
Fig. 9. Storm of 12 February 2010: (a) orography (m, colours) and wind vector (m s\(^{-1}\)) at 3 km altitude using the WRF first inner domain; (b) MEC (W kg\(^{-1}\), colours) and wind vector (m s\(^{-1}\)) at 2300 m; (c) cold front identified by the magnitude of temperature gradient (K km\(^{-1}\)) at 2500 m; (d) longitude/latitude cross-section of water vapour concentration (g m\(^{-3}\)); (e) longitude/altitude cross-section of SCLW (g m\(^{-3}\), colours) and temperature (°C, black lines) at latitude 35°S and (f) the CAPE (J kg\(^{-1}\)) obtained from WRF model and the \(T_q = \log p\) diagram. All figures are for 23 UTC.

Fig. 10. Storm of 12 February 2010; mountain wave. Longitude/altitude cross-section at latitude 34.5°S and at 23 UTC of (a) vertical wind (m, colours) and temperature (°C, black lines) and (b) relative humidity (colours) and temperature (°C, black lines).
The synoptic conditions determined by the shape and extension of the SPA (Fig. 13b) are different from the previous cases. The SPA extends all the way into the Atlantic after being deformed by the crossing of the Andes between 50°S and 60°S, while the SAA appears of weaker amplitude than usual. While the explanation of this structure is beyond the scope of this paper, its consequences are of interest. This synoptic situation generates two groups of trajectories as seen in Fig. 13c: one coming from a southeastern route over the Atlantic Ocean and the other coming from inland Brazil and Uruguay. The latter crosses over more humid areas and are those that bring most of the moisture. Fig. 13d displays the relationship between the orographic elevation and the low-level horizontal wind. The wind deflection around the mountains (marked by the star) creates a convergence zone at the site where deep convection is generated (the cross). Figure 14a shows the wind convergence and the MEC maximum localized at the site of deep convection.

Figure 14b and c shows longitude/altitude cross-sections of vertical wind and humidity. The relative humidity shows saturation up to 10 km altitude. The vertical wind, although not very strong, has the features of a convective updraft.
This case is different from the others because a front is not present. As a consequence, the convection is weaker, as illustrated by the amplitude of the vertical wind and by the altitude of the top of the cloud. In the absence of a front, it is the horizontal wind convergence that supplies the force needed to lift the fluid parcels up to the LFC in this case. These conditions are sufficient to produce deep convection. Moreover, as shown in Fig. 14d, the SCLW reached values of 2.2 g m$^{-3}$. Finally, there are no mountain waves whose vertical wind was greater than 3 m s$^{-1}$ near the place and time of the storm.

7. Summary and conclusion

Deep convection occurs frequently in the region of Mendoza, in Central Argentina. From October to March, it creates the conditions of formation of heavy hailstones that hit the ground and produces severe damage to cultivated land. In this paper, we describe four cases of very severe hailstorms that hit the region between 2009 and 2011. We analyse four main factors that are likely to be at the origin of the storms: (1) the presence of a moisture source, (2) the presence of a cold front, (3) the presence of horizontal convergence due to the synoptic condition or to the flow around a mountain, and (4) the presence of a mountain wave.

The warm and moist air needed to generate the convection comes mainly from two sources: from the Amazon transported by the LLJ and from southern Brazil or even from the Atlantic Ocean as has been shown by Teitelbaum et al. (2008). Of the four cases presented here, two presented a moisture flux carried by the LLJ and the other two had a source of moisture coming from a more easterly route.

The arrival of warm and moist air is a necessary but not sufficient condition to generate deep convection. In the particularly intense cases analysed here, with the fall of hailstones and destruction of crops, convergence of moist enthalpy near the ground is also found. In two cases, there is a low-level flow around a mountain, and convergence appears near the lee side of the mountain. This near-ground convergence is associated with lifting of air parcels. In one of the cases, such lifting was sufficient to trigger convection, in another, it was accompanied by large-scale convergence due to the synoptic situation. Another important factor that often plays the main role in producing a storm is the passage of a cold front. In three of the four cases, maxima of MEC were co-located with a cold front. Moreover, the front facilitates the development of deep convection: air rises on the warm side of the front, making it easier to attain the LFC.

The complex orography of the region results in the generation of mountain waves. However, contrary to the studies on the Rocky Mountains cited above, our study
Fig. 13. Storm of 22 February 2011: (a) radar reflectivity (dBZ). At 20:30 UTC, the maximum reflectivity is found over San Rafael (68.3°W, 34.5°S); (b) ground pressure (hPa, colours) and horizontal wind vector (m s$^{-1}$) at 850 hPa obtained from ECMWF forecast on day 22 at 21 UTC; (c) 3-D 6-d back-trajectories starting from the convective area at 3 km altitude, and specific humidity (g kg$^{-1}$, colours) near the ground on 16 February at 21 UTC; (d) orography (m, colours) and wind vector (m s$^{-1}$) at 3 km altitude using the WRF first inner domain.

Fig. 14. Storm of 22 February 2011. (a) MEC (W kg$^{-1}$, colours) and horizontal wind vector (m s$^{-1}$) at 2300 m; longitude/altitude cross-section at latitude 34.5°S and at 20:30 UTC of (b) vertical wind (m s$^{-1}$, colours) and temperature (°C, black lines) and (c) relative humidity (colours) and temperature (°C, black lines); (d) longitude/altitude cross-section of SCLW (g m$^{-3}$, colours) and temperature (°C, black lines).
shows that these waves do not play any role in the production of deep convection and hailstorm, at least in the cases presented. Mountain waves are found in the vicinity, in space and time, of the storm. In all cases, however, the westerly wind component going up the windward side of the mountain provokes precipitation and hence makes the air less humid at the site where the mountain wave is generated. Additionally, the vertical wind induced by the mountain wave does not appear to be strong enough to produce deep convection. For both reasons, SCLW, which is essential for the formation and fall of hailstones, does not appear. It must be stressed that the results of this study are specific to the climatology of the region and cannot be simply extrapolated to other regions.

While ruling out mountain waves as a cause for hailstorms is a useful result, further work is needed to appreciate the predictive power of the other three conditions listed above. In the future, sensitivity experiments with a regional model will be conducted excluding one of them and comparing their respective importance.

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