Katabatic Flows and Their Relation to the Formation of Convective Clouds—Idealized Case Studies

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ABSTRACT

The formation of a convective cloud system as a result of a katabatic-induced surface cold front at the eastern Andes Mountains of South America was investigated in a numerical model study. The occurrence of this cloud system is hypothesized to be a consequence of converging cold-air drainage from slopes and valleys resulting from the concave shape of the terrain. Simplified terrain configurations were applied to three different atmospheric experiments to determine the influence of the terrain and the ambient stratification on the underlying processes. The simulation demonstrated the occurrence of a convective cloud, but not in every simulation. The initial stable stratification experiment did not initiate convective activity. Further analysis of the development of the convective cells confirmed the terrain’s impact, but it also showed a dependence on atmospheric conditions. The katabatic flows and the surface fluxes from which they are induced are sensitive to the ambient stratification, which was seen when only a weak cold front developed in response to a decrease in the surface inversion and the downslope velocity. On the basis of specific characteristics of katabatic flows and the heat-exchange budget, the presence of these flows and their significance as the driving force behind the cloud-formation process were confirmed.

1. Introduction

Convective clouds are common features in tropical regions. They generally develop over land and maximize in the late afternoon in response to the diurnal cycle of insolation (Meisner and Arkin 1987; Yang and Slingo 2001; Sato et al. 2009). The land surface is heated by diabatic processes, which lead to sufficient positive buoyancy for strong upward vertical motion; the formation of convective clouds then proceeds from the release of latent heat (Mapes et al. 2003; Poveda et al. 2005).

The area of interest (Fig. 1) covers southern Ecuador and the adjacent northern Peruvian Amazon basin in the central Andes Mountains of South America. In this region, cloud and rainfall development are affected not only by diurnal variations of insolation but also by the complex terrain of the Andes—for example, local circulation systems such as windward and leeward effects as well as mountain–valley breezes (Lenters and Cook 1995). During the day, insolation produces differentially heated slopes with ascending air and convective cloud generation at higher elevations. After sunset, the situation is reversed. As a result of the negative net radiation budget at the surface, cold air descends along the slopes. These katabatic flows produce negative buoyancy by inducing an inversion layer near the surface, which usually inhibits convective activity (Defant 1949; Barry 2008).

In the Rio San Francisco valley, located in the highlands of southern Ecuador, observations of precipitation dynamics and other data revealed an unexpected early-morning rainfall peak at the Estacion Cientifica de San Francisco (ECSF) at 3°58′18″S, 79°4′45″W and an altitude of 1860 m MSL (see left panel of Fig. 1) (Bendix et al. 2006, 2009). Analysis of cloud patterns from corresponding Geostationary Operational Environmental Satellite imagery indicated the appearance of nocturnal mesoscale convective systems (MCS) at the foothills of the Andes in the Peruvian Amazon basin. The minimum infrared cloud-top temperature distributions gave evidence of high clouds, which led to the conclusion that the rainfall peak is produced by strong nocturnal convection. The occurrence of convection was hypothesized to be the result of a highland–lowland interaction: namely, nocturnal cold drainage of air from the Andean slopes
and valleys converges with warm, moist air in the Amazon basin. The katabatic flows act as a local surface cold front, which accounts for an atmospheric destabilization in the foothills with subsequent initiation of deep convection. The specific terrain configuration of the Andes Mountains (a quasi-concave geometry; see lower-right panel of Fig. 1) in the cloud-formation area is of particular importance because it contributes explicitly to this condition of atmospheric destabilization.

Previous work has found an association between the formation of nocturnal convective clouds (and subsequent rainfall) and lower-tropospheric flow systems in the eastern Andes (Lopez and Howell 1967; Mapes and Houze 1992). Garreaud and Wallace (1997) showed that nocturnal rainfall appears to arise from enhanced low-level convergence that is a result of the nocturnal circulation between the Andes and the Amazon region. De Angelis et al. (2004) described the convergence of cold-air drainage from the Andes and warm, moist air from the Amazon in terms of the katabatic flows that induce low-level instability by acting as cold fronts. Some authors have treated the katabatic flows as atmospheric density currents. They found evidence that the density currents trigger the lifting of air parcels, which results in condensation and cloud-formation processes as long as sufficient moisture is available (Moncrieff and Liu 1999; Cunningham 2007). Shapiro et al. (1985) showed that the head of a density current can trigger the development of mesoscale convective cloud systems.

In an idealized case study, Trachte et al. (2010) used approximated terrain models to examine how the shape of terrain modifies katabatic flows. They showed that a thermally induced current develops and that concave geometry forces the air to converge. The study presented here aims to examine the convergence of katabatic flows and the subsequent development of nocturnal convective clouds. A numerical gridbox model was used to analyze the hypothesized succession of processes described above. Idealized case studies have been applied 1) to demonstrate the initiation of deep convection and subsequent cloud formation due to the terrain configuration, 2) to show the occurrence of a local surface cold front, and 3) to show that this atmospheric front is induced by katabatic flows from the slopes and valleys of the simplified terrain.

The next section gives a brief description of the model settings that were used. The results of the convective cloud generation are then presented and are discussed in the context of the shape of the terrain and the evolution of a katabatic-induced surface cold front.

2. Model setup

For the simulation of the idealized formation of a convective cloud system, the Advanced Regional
AUGUST 2012  TRACHTE AND BENDIX  1533

Prediction System (ARPS) was used. It was developed at the Center for Analysis and Prediction of Storms at the University of Oklahoma. ARPS is a fully compressible, nonhydrostatic numerical model with a generalized terrain-following coordinate system and vertical stretched grid. For more details, see Xue et al. (1995, 2000, 2001).

The simulations were calculated on a domain with $211 \times 211$ grid points and a horizontal resolution of 500 m, which is sufficient to resolve explicit simulations of moist convection (Xue and Martin 2006). The 68 vertical layers have an average spacing of 250 m and are stretched to a minimum of 30 m near the ground by a hyperbolic function. The minimum vertical grid spacing is maintained for the lowest 300 m to resolve the formation of a thermally induced density current in the planetary boundary layer (PBL). As in a previous study (Trachte et al. 2010), the terrain is represented by simplified configurations that resemble the main structures of the Andes Mountains in the study area: a simple slope geometry (SLP), a uniformly concave ridgeline (BSN), and a concave ridgeline interrupted by several mountain peaks draining into a basin (BSNV), as shown in the upper-right panel of Fig. 1. These terrain configurations will aid in the demonstration that cold air from the slopes and valleys converges and initiates moist convection.

**a. Initialization and boundary conditions**

The simulations were horizontally homogeneously initialized with conditions derived from an idealized tropical atmosphere (McClatchey et al. 1972). Three different atmospheric conditions for the PBL were used. Each potential temperature profile begins at 300 K at the surface and is adjusted as follows: the first atmospheric conditions represent a situation at sunset with an approximately neutral temperature stratification beneath a stable layer (SC), the second conditions contain a stably stratified layer near the surface and a neutral profile above representing a fully developed nocturnal boundary layer over land (NC), and the third atmospheric conditions depict a continuously stratified profile (CC). The wind field at the beginning of the simulation is set to zero to avoid any stream flows overlaying the thermally induced flow. Nine numerical experiments were conducted by combining the three atmospheric conditions with the three terrain configurations (Table 1). The model was run for 23 400 s.

The boundary conditions at the lateral sides had radiation-open conditions of the Klemp and Wilhelmson type (Klemp and Wilhelmson 1978) to obtain a constant phase speed, which was computed and applied at each time step. At the top of the domain, a rigid wall is used with a damping layer at two-thirds of the atmosphere with five friction-absorbing layers to prevent unrealistic reflections.

**b. Physics**

This section describes the physics packages that were used with ARPS. For the turbulence parameterization, the 1.5-order turbulent kinetic energy (TKE) with the Deardorff closure scheme was used (Deardorff 1972). The parameterization consists of terms for advection, potential–kinetic energy conversion, shear production, dissipation, and diffusion of TKE (Xue et al. 1995, 2000). The surface fluxes are responsible for mass and heat exchanges with the atmosphere and are computed by a surface-flux model that is dependent on stability and the roughness length (Businger et al. 1971; Byun 1990). The fluxes are solved on the basis of the similarity theory of Monin and Obukhov. The surface momentum fluxes are defined by

$$\tau_{13}\text{surface} = -(\rho pw^r)\text{surface} = \rho C_{dm} V u \quad \text{(1)}$$

$$\tau_{23}\text{surface} = -(\rho pw^r)\text{surface} = \rho C_{dm} V v. \quad \text{(2)}$$

Here $\tau_{ij}$ is the Reynolds stress tensor, which is parameterized in terms of the resolvable scale quantities. The equations above use the drag coefficient $C_{dm}$ and the wind speed $V$, which contains components $u$ and $v$. Also, $\rho$ is air density and $w$ is the vertical component of the wind. The sensible and latent heat fluxes that account for the heat exchange are described by Eqs. (3) and (4), respectively:

$$H = \rho C_{dh} C_p V (T - T_s) \quad \text{and} \quad \text{(3)}$$

$$\text{LE} = L(E_g + E_{tr} + E_r), \quad \text{(4)}$$

with the sensible and latent heat flux $H$ and LE, the heat-exchange coefficient $C_{dh}$, the specific heat capacity $C_p$, the air temperature $T$ taken at the first level above ground, and the ground surface temperature $T_s$. The direction of the sensible heat flux is consequently dependent on the temperature gradient near the surface. The latent heat flux is the sum of the evaporation from the soil surface $E_g$, the transpiration $E_{tr}$, and the fluxes from canopy water evaporation $E_r$, where $L$ is the molar heat.

**Table 1. Matrix of numerical experiments showing the combinations of atmospheric conditions SC, NC, and CC and terrain configurations SLP, BSN, and BSNV, as defined in the text.**

<table>
<thead>
<tr>
<th></th>
<th>SC</th>
<th>NC</th>
<th>CC</th>
</tr>
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<tbody>
<tr>
<td>SLP</td>
<td>SLP-SC</td>
<td>SLP-NC</td>
<td>SLP-CC</td>
</tr>
<tr>
<td>BSN</td>
<td>BSN-SC</td>
<td>BSN-NC</td>
<td>BSN-CC</td>
</tr>
<tr>
<td>BSNV</td>
<td>BSNV-SC</td>
<td>BSNV-NC</td>
<td>BSNV-CC</td>
</tr>
</tbody>
</table>
In the simulations presented here, the surface fluxes were calculated using stability-dependent drag coefficients with the bulk Richardson number as the stability parameter. For more details, see Xue et al. (1995, 2001). The general surface characteristics of the soil model, such as soil and vegetation type, are provided by a force–restore, two-layer soil and vegetation model (Noilhan and Planton 1989). The input parameters for the soil and vegetation type are based on U.S. Department of Agriculture textural classes. Each model run in the current study is initialized with loam and rainforest soil types to match the conditions in Ecuador.

The radiative cooling necessary for the thermally driven flow is determined by the use of atmospheric radiation transfer parameterizations. It is the primary force of the heat energy budget [Eq. (5)] and includes the net radiation, sensible heat flux, and ground heat flux into the surface. The net radiative flux is given by

\[ R_n = R_{sw}(1 - \alpha_g) + \epsilon_g(R_A - \sigma T_s^4), \]

where \( R_{sw} \) is the shortwave radiation, the albedo is \( \alpha_g \), the ground surface emissivity is \( \epsilon_g \), incoming longwave radiation is \( R_A \), emitted longwave radiation from the ground surface is \( \sigma T_s^4 \), and \( \sigma \) is the Stefan–Boltzmann constant. If \( R_{sw} \) becomes zero and the second term becomes larger, the net radiation becomes negative.

An explicit microphysics parameterization scheme (Lin et al. 1983) was used to simulate the cloud processes. Six prognostic equations were solved for water vapor, cloud water, rainwater, cloud snow, cloud ice, and graupel. Six prognostic equations were solved for water vapor, cloud water, rainwater, cloud snow, cloud ice, and graupel. Because of the high horizontal resolution of the domain, the convection is resolved explicitly and no cumulus parameterization scheme was needed.

3. Results

The simplified terrain configurations presented above were used to examine the development of a convective cloud system that results from converging cold-air drainage flows using three different atmospheric stratification experiments. In addition to other parameters, the simulation runs were analyzed using the vertical equivalent potential temperature distribution as well as the main instability parameters that indicate the atmospheric environmental conditions. The influence of the terrain on cloud formation is illustrated by means of the horizontal divergence field. The characteristic features of the underlying processes are examined and discussed in the context of their driving forces. To anticipate a discussion about wind shear in the context of convection, a shear experiment on the basis of a katabatic flow and an opposing fast-moving current in the PBL is briefly shown in section 3a.

### a. A note regarding wind shear

With the SLP-NC case the influence of wind shear on the cloud formation is shown (e.g., Rotunno et al. 1988; Houston and Wilhelmson 2011; Moncrieff and Liu 1999; Richardson et al. 2007; Weisman and Klemp 1982) to consider an opposing flow without the impact of the shape of the terrain. The experiment proceeded in the following manner: at the time step 18 600 s, when the katabatic flow clearly is developed, the wind field in the \( u \) and \( v \) directions of the SLP case is substituted for the values of \( u \) and \( v \) from a wind field for all levels in a specific region (Fig. 2a).

Figure 2a displays the composite horizontal wind field (vectors) at the initial time step. The embedded wind field representing the opposing flow can be seen with a northeast direction and is clearly distinguishable from the outgoing katabatic flows by identification of its stronger velocities of approximately 5 m s\(^{-1}\). The wind field causes wind shear at the lateral sides of the jet and particularly at its head. After the 1500-s time step of the simulation, the wind shear demonstrates its influence (Fig. 2b): namely, the converging flows result in compensating upward vertical motions that are indicated by the increasing vertical velocities, which reach a maximum of 19.97 m s\(^{-1}\). The rising air reaches its level of free convection (LFC) and generates a cumulus tower (illustrated by the solid contour line of total condensed water) that develops further into a cloud cluster.

The results show the expected convective activity regardless of the configuration of the terrain, when the wind velocity of an opposing flow is strong enough. Therefore, this case is not further considered. The next sections examine whether wind shear generated by the interaction between katabatic flows and the geometry of the terrain is sufficient to initiate deep convection as well.

### b. Formation of a convective cloud cluster

First, we start with the examination of the occurrence of a convective cloud cluster due to the configuration of the terrain. Figure 3 illustrates a cross section of the horizontal moisture convergence (shaded), the wind field in the \( u-w \) direction (vectors), and the total condensed water (solid contour line) of BSNV at time step 22 800 s (convection started at 18 300 s) for the three different atmospheric stratification experiments mentioned above.

In Fig. 3a (BSNV-SC), a convective cloud cluster with a vertical depth of approximately 9 km is shown in its mature stage and can be identified by the convective and stratiform cloud regions. Two cumulus towers can be seen, which are characterized by strong updrafts with maximum vertical velocities of 15.0 m s\(^{-1}\) forming a
buoyant plume. The ascending air reaches its LFC, which can be identified by an acceleration of the parcels (the \( w \) vector increases with height). The lifting of the air is forced by strong low-level moisture convergence (approximately 96.0 \( 10^{-3} \) g kg\(^{-1}\) s\(^{-1}\)). This region of convergence is the location of the generation of a deep convective cloud (displayed by the solid contour line of total condensed water) because of the release of latent heat. In the lower troposphere, entrainment takes place as a result of the upward mass flux. Accordingly, divergent branches of air in the upper portion of the domain as well as detrainment at the lateral sides evolve with time. Strong downdrafts are located at the leading edge of the cloud cluster and indicate the precipitation area. The downbursts of cold-air parcels reach the surface, where they spread out and create a gust front. The gust front is an area characterized by turbulent eddies and strong horizontal and vertical wind shear as shown in Fig. 3a. The gust front achieves a depth of approximately 1.5 km and vertical velocities of 4.5 m s\(^{-1}\). The outflow of the cold air results in lifting of the environmental air and subsequent formation of a buoyant plume as suggested by the moisture convergence of 44.0 \( 10^{-3} \) g kg\(^{-1}\) s\(^{-1}\). In the center of the cloud system, the downbursts from both cumulus towers converge at the surface (120.0 \( 10^{-3} \) g kg\(^{-1}\) s\(^{-1}\)), which causes the initiation of a new convective cell. Because of the strong wind shear, the downbursts do not prevent the inflow of moist air, and cell regeneration can occur.

In contrast, the simulation results for the BSNV-NC case are displayed in Fig. 3b. The atmosphere is calm and shows neither considerable moisture convergence nor any vertical motion. If the BSNV-CC set of conditions is used, a convective system is again generated (Fig. 3c). As with the BSNV-SC case (Fig. 3a), the cloud cluster develops in the middle of the basin and reaches a height of approximately 9 km. Strong moisture convergence of up to 119.0 \( 10^{-3} \) g kg\(^{-1}\) s\(^{-1}\) occurs at the rear of the cloud cluster, where several cells develop. The air is lifted by strong updrafts with a maximum vertical velocity of 28.55 m s\(^{-1}\) that generate a deep cloud system. The air diverges at the top of the system, and detrainment evolves at the lateral sides. In the lower troposphere, entrainment and a low-level inflow occur to compensate for the strong upward mass flux. The center of the cloud system is dominated by descending air, which represents the precipitation area. The downbursts of cold-air parcels once again reach the surface and spread apart. As seen in Fig. 3a, a gust front with turbulent eddies and strong wind shear evolves at the leading edge of the downbursts. At the rear of the system, cell regeneration occurs because of the cold-air outflow. The outflow causes lifting of warmer air, which results in the initiation of new convective cells with strong updrafts (maximum 28.55 m s\(^{-1}\)). The BSNV-CC cloud cluster develops at the same location as the BSNV-SC cluster, and has a comparable depth and other similar features. Other terrain configurations (BSN and SLP) were also applied to
the different atmospheric conditions described above (Table 1). The results are not shown, because neither of the terrain geometries generated any convective activity or cloud formation. This discrepancy is examined in the next two sections.

c. Analysis of atmospheric conditions

In this section, the occurrence or absence of deep convection is analyzed using the thermodynamic structure of each atmospheric simulation. For this analysis, the equivalent potential temperature is used:

\[ \theta_e = \theta \exp \left( \frac{L_e}{c_p T_{LCL}} w \right), \]  

where \( \theta \) is the potential temperature, \( L_e \) is the latent heat of evaporation, \( c_p \) is the specific heat content, \( T_{LCL} \) is the temperature at the lifting condensation level (LCL), and
is the mixing ratio for water vapor. The $\theta_e$ represents a value that is composed of the air temperature and its moisture content and is an indicator for potential instability in the atmosphere. A criterion for potential instability is the decrease of $\theta_e$ with height, which signifies a decrease in the absolute humidity.

Figure 4 shows the vertical distribution of $\theta_e$ for each combination of stratification conditions and terrain experiments at location 80 km by 80 km with time steps of 17 100 s to examine conditions just before initiation of convection. It can been seen that the profiles have similar characteristics and that there is little variation among the terrain configurations: namely, a layer of warm, moist air is located in the lower troposphere and is overlaid by a layer of colder, drier air. Thus, $\theta_e$ decreases with height, and the result is an atmosphere with a potential instability. This instability condition may become conditionally unstable if the air is lifted and becomes saturated. In the upper troposphere, $\theta_e$ increases with height. A closer examination of the individual profiles reveals that several differences are present. For the BSNV-SC and BSNV-CC cases (Fig. 4c), the warm, moist layer near the surface is characterized by maximum $\theta_e$ values of 342 K within a depth of approximately 200 m, which describes the top of a surface inversion. Above the inversion, the cold layer in the middle of the troposphere reaches a minimum $\theta_e$ of 318 K, resulting in a gradient of 24 K. In contrast, the inversion near the surface for the BSNV-NC case is stronger than the previous two cases and achieves a depth of approximately 300 m with maximum $\theta_e$ values of 345 K. The decrease in $\theta_e$ with height is less (a minimum of only 325 K is reached), however, which results in a weaker gradient of only 20 K. For each atmospheric stratification experiment, the vertical distribution of $\theta_e$ has the same profile regardless of the shape of the terrain. Only the depth of the warm, moist inversion layer near the surface (100 m) is weaker for the SLP (Fig. 4a) and BSN (Fig. 4b) cases, but the vertical gradient of $\theta_e$ remains the same.

Additional parameters were used to evaluate the thermodynamics and whether convective activity is induced (Table 2). These parameters are the convective available potential energy (CAPE; Moncrieff and Liu 1999) and the convective inhibition (CIN; Colby 1984). CAPE is the vertically integrated positive buoyancy of a parcel between the height of the LFC $Z_{\text{LFC}}$ and the height of the equilibrium level $Z_{\text{EL}}$. CAPE represents the maximum energy available to an ascending air parcel and is an indicator of the potential for convective initiation:

$$\text{CAPE} = \int_{Z_{\text{LFC}}}^{Z_{\text{EL}}} g \left( \frac{\Theta_{\text{par}} - \Theta_{\text{env}}}{\Theta_{\text{env}}} \right) dz, \quad (7)$$

where $\Theta_{\text{par}}$ is the potential temperature of the parcel lifted from the surface $Z_{\text{sfc}}$ up to $Z_{\text{LFC}}$ and $\Theta_{\text{env}}$ is the ambient potential temperature. In contrast, CIN represents the energy required to lift a negatively buoyant parcel from the surface to the LFC.
For the same time, the average CIN is highest in the SC/CC and NC conditions because they are the strongest and weakest of the atmospheric experiments, respectively, and the SC case does not give additional information. In Fig. 5a the horizontal divergence field in the SLP-CC case is illustrated. It shows a very different pattern from that of the BSN-CC (Fig. 5b) and BSNV-CC (Fig. 5c) cases. The structure of the divergence field is organized in a pattern that follows the terrain without a centralized maximum. Nevertheless, an observable line of alternating positive and negative divergence is present, but its values \(10^{-3}\) and \(10^{-4}\) s\(^{-1}\) are considerably weaker than for the other two cases.

### Table 2: Environmental parameters CAPE, CIN, and LFC that indicate the stability of the atmosphere at time steps 0 and 18 000 s for the numerical experiments SLP, BSN, and BSNV combined with SC, NC, and CC from \(x = 80\) km, \(y = 80\) km.

<table>
<thead>
<tr>
<th></th>
<th>SC</th>
<th>NC</th>
<th>CC</th>
<th>SC</th>
<th>NC</th>
<th>CC</th>
</tr>
</thead>
<tbody>
<tr>
<td>CAPE ((J \text{ kg}^{-1}))</td>
<td>(3202)</td>
<td>(2041)</td>
<td>(3514)</td>
<td>(2020)</td>
<td>(1372)</td>
<td>(2296)</td>
</tr>
<tr>
<td>BSN</td>
<td>(3202)</td>
<td>(2041)</td>
<td>(3514)</td>
<td>(2007)</td>
<td>(1345)</td>
<td>(2418)</td>
</tr>
<tr>
<td>BSNV</td>
<td>(3202)</td>
<td>(2041)</td>
<td>(3514)</td>
<td>(1809)</td>
<td>(1332)</td>
<td>(2382)</td>
</tr>
<tr>
<td>CIN ((J \text{ kg}^{-1}))</td>
<td>(0)</td>
<td>(-61)</td>
<td>(0)</td>
<td>(-35)</td>
<td>(-109)</td>
<td>(-35)</td>
</tr>
<tr>
<td>BSN</td>
<td>(0)</td>
<td>(-61)</td>
<td>(0)</td>
<td>(-26)</td>
<td>(-107)</td>
<td>(-15)</td>
</tr>
<tr>
<td>BSNV</td>
<td>(0)</td>
<td>(-61)</td>
<td>(0)</td>
<td>(-35)</td>
<td>(-101)</td>
<td>(-17)</td>
</tr>
<tr>
<td>LFC ((\text{hPa}))</td>
<td>(911)</td>
<td>(836)</td>
<td>(910)</td>
<td>(884)</td>
<td>(820)</td>
<td>(891)</td>
</tr>
<tr>
<td>BSN</td>
<td>(911)</td>
<td>(836)</td>
<td>(910)</td>
<td>(884)</td>
<td>(820)</td>
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<tr>
<td>BSNV</td>
<td>(911)</td>
<td>(836)</td>
<td>(910)</td>
<td>(895)</td>
<td>(820)</td>
<td>(902)</td>
</tr>
</tbody>
</table>

Table 2 presents these parameters for each simulation run. The initial state is identical for each terrain configuration but varies between the different PBL structures. There is a noticeable distinction in CIN between the SC/CC and NC conditions. In the NC condition, the CIN is smaller than in the other two cases but still reflects a significant potential for deep convection. The formation of a convective cloud cluster was hypothesized to be a consequence of the concave terrain configuration that allowed for the convergence of katabatic flows. This convergence can be described in terms of the horizontal divergence. Figure 5 illustrates the horizontal divergence field amplified by a factor of 1000 for the SLP-CC, BSN-CC, and BSNV-CC cases with a time step of \(17\) \(100\) s (as in Fig. 4) at a height of \(50\) m above the surface. The following analyses are made with CC and NC conditions because they are the strongest and weakest of the atmospheric experiments, respectively, and the SC case does not give additional information. In Fig. 5a the horizontal divergence field in the SLP-CC case is illustrated. It shows a very different pattern from that of the BSN-CC (Fig. 5b) and BSNV-CC (Fig. 5c) cases. The structure of the divergence field is organized in a pattern that follows the terrain without a centralized maximum. Nevertheless, an observable line of alternating positive and negative divergence is present, but its values \(10^{-3}\) and \(10^{-4}\) s\(^{-1}\) are considerably weaker than for the other two cases.

### d. Impact of terrain configuration

The formation of a convective cloud cluster was hypothesized to be a consequence of the concave terrain configuration that allowed for the convergence of katabatic flows. This convergence can be described in terms of the horizontal divergence. Figure 5 illustrates the horizontal divergence field amplified by a factor of 1000 for the SLP-CC, BSN-CC, and BSNV-CC cases with a time step of \(17\) \(100\) s (as in Fig. 4) at a height of \(50\) m above the surface. The following analyses are made with CC and NC conditions because they are the strongest and weakest of the atmospheric experiments, respectively, and the SC case does not give additional information. In Fig. 5a the horizontal divergence field in the SLP-CC case is illustrated. It shows a very different pattern from that of the BSN-CC (Fig. 5b) and BSNV-CC (Fig. 5c) cases. The structure of the divergence field is organized in a pattern that follows the terrain without a centralized maximum. Nevertheless, an observable line of alternating positive and negative divergence is present, but its values \(10^{-3}\) and \(10^{-4}\) s\(^{-1}\) are considerably weaker than for the other two cases.

The horizontal divergence field for the BSN-CC case is displayed in Fig. 5b. An explicit line of positive and negative divergences in the middle of the basin can be seen in an otherwise calm environment. The patterns of convergence and divergence alternate, which indicates the development of frontal structures (Eliassen 1959). The highest divergent values \(3.2 \times 10^{-3}\) and \(3.3 \times 10^{-3}\) s\(^{-1}\) are focused in the center of the basin, a direct result of the concave shape of the terrain. A large increase in mass convergence is closely linked to vertical
velocities, which are a consequence of horizontal convergence. The higher LFC in the BSN-CC case relative to the BSNV case helps to explain why these horizontal convergence values lead to an insufficient increase of mass convergence, less ascent of air, and a failure to initiate deep convection.

The horizontal divergence field for the BSNV-CC case is demonstrated in Fig. 5c, in which two regions are of particular interest. For the BSN-CC case, there is a zone inside the basin that contains alternating positive and negative divergence values, indicating frontal structures. The frontal structure is in the region of convective cloud formation seen in Fig. 3. The minimum and maximum divergence values are \(-4.1 \times 10^{-3}\) and \(2.85 \times 10^{-3}\) s\(^{-1}\), respectively. These extreme values, which are stronger than in the BSN-CC case, are organized in the middle of the basin and reflect the mass contribution due to the terrain. Furthermore, in the exit regions of each valley, additional divergence and convergence patterns occur that strengthen the values to reach \(2.0 \times 10^{-3}\) and \(-2.0 \times 10^{-3}\) s\(^{-1}\), respectively. These values are caused by the assumed cold-air drainage advancing into the basin, which is discussed in the following section.

By using the horizontal divergence field, the impact of the concave terrain configuration can be demonstrated. Moreover, it can be shown that the draining valleys add an additional mass contribution, which is apparently essential for the initiation of deep convection. The BSNV conditions do not induce deep convection with an initially stably stratified lower troposphere (BSNV-NC), however. If we compare the impact of the terrain geometry with different PBL stratifications, the influence of the atmospheric conditions can also be seen.

Figure 6 illustrates the horizontal divergence fields for the cases of SLP-NC, BSN-NC, and BSNV-NC conditions. These divergence fields vary significantly from the previous results. Inside the basin (Fig. 6c), no comparable convergence or divergence structures are visible, and there is no focusing of mass flux in the center. Similar patterns occur at the exit regions of each valley, but those features are not important enough for the initiation of convective activity inside the basin. The SLP-NC and BSN-NC cases (Figs. 6a,b) show weaker patterns of horizontal divergence when compared with the SLP-CC and BSN-CC cases in Fig. 5. Because the downslope flows are the presumed driving force behind the cloud formation associated with the terrain configuration, both features must be available.

e. Katabatic-induced surface cold front

The next step is to demonstrate the occurrence of a katabatic-induced surface front. An atmospheric front is

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**FIG. 5.** Horizontal cross section (x–y plot at \(z = 50\) m AGL) of the divergence field “Div” amplified by a factor of 1000 (s\(^{-2}\); shaded) at time step 17 100 for (a) SLP-CC, (b) BSN-CC, and (c) BSNV-CC.
defined by a gradient of a scalar attribute such as the potential temperature $\theta$. The thermal gradient can result in dynamic effects such as deformation of the horizontal wind field, which activates a secondary circulation with ascending and descending branches. Because the SLP and BSN cases failed to produce a convective cloud in each of the experiments, we focus on modifications of the BSNV case. Figure 7 shows a cross section through the middle of the basin for the BSNV-NC and BSNV-CC conditions at a time step of 15 600 s to examine conditions unaffected by convection. The appearance of characteristic features of atmospheric fronts is examined through the use of the potential temperature distribution (contours) and the wind field in the $u-w$ direction (vectors).

In Fig. 7a, the simulation results from the BSNV-NC case show an along-slope $u$ distribution creating a weak temperature gradient (0.5 K) with the ambient air at the same height level (Fig. 7b). Inside the basin, the isotherms are approximately uniformly arranged and reveal no distinct inclination. The wind velocities in the $u$ direction have maximum values of approximately 2.5 m s$^{-1}$. Weak vertical motion developed because of the lack of a temperature gradient. Thus, on the basis of these features, no surface cold front could be identified.

In contrast, the BSNV-CC case is considerably different from the BSNV-NC case, as was already suggested by the horizontal divergence patterns (Figs. 5, 6). The simulation results reveal two major features. In the first feature, shown in Fig. 7a, $\theta$ is distributed parallel to the slope but with a stronger gradient to the ambient air (1 K). The second feature can be seen at the base of the slope. At the base, a potential temperature difference of 1 K is located inside the basin. The difference is generated in the lower troposphere where the strongest values of $\theta$ and a noticeable inclination of the isotherms are visible. The organization of the wind vectors correspond to the $\theta$ pattern. Along the slope, a downslope motion can be seen in the lowest levels. The downslope wind is driven by the evolved temperature gradient because the initial wind field is zero. As seen in the isotherm and wind vector fields, the flow drains cold air down the slope into the basin, where temperature and density gradients are created. In this area, which is the leading edge of the current, strongest downslope velocity values reach approximately 4 m s$^{-1}$, and a deformation of the horizontal wind field is seen. The result is a convergence zone ahead of the current and divergence behind the current (Fig. 5c). Compensating ascending and descending mass fluxes can be seen in the figure as vectors. These atmospheric parameters indicate the formation of a surface cold front.

To state with confidence that the described density current and surface cold front (Fig. 7b) are induced by
katabatic flows, we now take a closer look at typical characteristics of katabatic flows. Katabatic flows are thermally induced downslope winds that evolve in hilly regions on calm, clear nights through radiative surface cooling. A buoyancy deficit develops as a result of a net radiation loss [Eq. (5)] that is caused by radiative divergence. This divergence is associated with ground heat fluxes [Eqs. (1)–(4)] that result from mechanically induced TKE and latent and sensible heat fluxes (Prandtl 1942; Defant 1949).

Figure 8 shows the vertical profiles of potential temperature, the wind fields in the $u$ direction, and the TKE at the center of a slope $x = 25.0$ km and $y = 78.0$ km for both the BSNV-NC (top panels) and BSNV-CC (bottom panels) cases at 3600 and 15 600 s. The profiles (Figs. 8a–c) show the characteristics of katabatic flows for the BSNV-NC case. As expected, the $\theta$ profile at the beginning of the simulation shows a stably stratified PBL (indicated by an increase with height) (Fig. 8a). The profile changes to a quasi-neutral layer between 350 and 450 m. Above and below this layer, $\theta$ increases with height, and a low inversion near the surface can evolve. The maximum wind field in the $u$ direction (Fig. 8b) is located in this weak katabatic layer, with velocity values reaching 2.1 m s$^{-1}$ but with a typical jetlike profile, and TKE is scarcely present (Fig. 8c). Thus, a weak katabatic flow develops.

These features are considerably stronger for the BSNV-CC case. Figure 8d illustrates the development of a surface inversion, which was identified by a positive temperature gradient in the lower PBL. The $\theta$ values increase at a rate of 1 K (100 m)$^{-1}$, creating a layer with a depth of approximately 140 m. At its top, the buoyancy deficit is zero. Thus, the temperature adjacent to the ground and the temperature of the ambient air are equal at the same height. Above this katabatic layer, $\theta$ shifts into an isothermal state. Inside the katabatic layer, the wind field forms its typical jetlike profile (Fig. 8e). Because the influence of the positive temperature gradient is strongest near the surface, the maximum wind velocities of 4.3 m s$^{-1}$ are reached in this area. The positive temperature gradient decreases at values above the maximum wind velocity and below the ground the friction increases, producing this jetlike behavior. A return flow develops at the top of the inversion layer. The residual layer of the PBL is calm, and the maximum wind velocities reach values of approximately 0.3 m s$^{-1}$. Because of the vertical shear inside the katabatic layer, an exchange of momentum associated with the formation of turbulent eddies exists (Fig. 8f). Another major source of TKE is the mechanically induced turbulence from ground friction. Thus, an exchange with higher layers within the katabatic layer is induced. The TKE achieves maximum values of 0.3 m$^2$ s$^{-2}$ near the jetlike wind velocities, where the vertical wind shear is maximum (Horst and Doran 1986).

The reasons of the described differences are examined by means of the essential components of the surface heat budget as the driving mechanism for katabatic flows. Figure 9 illustrates the surface heat exchange as a function
of $R_n$, $H$, LE, and ground heat flux $G$ for the BSNV-NC and BSNV-CC cases. Cold-air drainage develops when the soil temperature falls to a value below the adjacent air temperature. This occurrence is followed by a sensible heat flux from the atmospheric boundary layer to the surface to balance the original heat loss from the surface.

In Fig. 9a (the BSNV-NC case), a negative net radiation can be seen during the simulation, with an average value of $-36 \text{ W m}^{-2}$. The negative value indicates that the emitted longwave radiation [second term on the right-hand side of Eq. (5)] is the dominant term. As a direct response, $H$ (average value of $-1.8 \text{ W m}^{-2}$) and $G$ (average value of $-16 \text{ W m}^{-2}$) also become negative. Thus, an energy flux from the atmosphere to the surface layer is induced by a constantly cooling PBL as seen in the negative radiation budget. This process can lead to the formation of a surface inversion (Fig. 8) that is indicated by negative buoyancy (Table 2) and produces a positive temperature gradient.

In contrast, Fig. 9b displays the surface energy budget for the BSNV-CC case, which exhibits variations in the strength of the fluxes. Similar to the previous results, the net radiation becomes negative (average value of $-50 \text{ W m}^{-2}$). The modification of $H$ is particularly different from the previous case. Whereas in the BSNV-NC case the values are marginally negative, the values in the BSNV-CC case achieve an average of $-5 \text{ W m}^{-2}$ because of a stronger energy flux from the PBL to the surface. This suggests the formation of a stronger inversion and thermally driven flow. The result is the
development of a distinct thermally induced katabatic flow associated with the described characteristics, given that inclined terrain is available.

4. Discussion

As expected the formation of a convective cloud cluster is affected by the shape of the terrain. Moreover, the results disclose the significance of the adequate development of the driving force for the convective initiation: namely, the katabatic-induced surface cold front. Only two combinations of terrain and stratification (BSNV-SC and BSNV-CC) produced a deep cloud cluster (Fig. 3). There was a potential in all cases for convective initiation, as indicated by the vertical distribution of $\theta_e$ (Fig. 4). For all of the terrain models, however, a weaker gradient (20 K) was observed in the NC case, whereas the BSNV-SC and BSNV-CC cases developed a deeper inversion near the ground (140 m). When the stable layer becomes unstable as it is lifted, a stronger vertical gradient provides a stronger environmental gradient and the equilibrium level moves to a higher altitude. As a result, the air can rise to higher levels and the developing cloud can grow in depth. This process was observed in the BSNV-SC and BSNV-CC cases and can be seen in Fig. 3. Crook (1996) also showed that variations in the PBL temperature and moisture strongly affected the initiation of convection, when forced by boundary layer processes. In all of the cases, the CAPE also indicated that there was a good potential for deep convection (Table 2). High values of CIN can suppress convection despite high values of CAPE, however, and this was seen in the BSN-NC ($-107$ J kg$^{-1}$) and SLP-NC ($-109$ J kg$^{-1}$) experiments by lack of clouds. It is because of high CIN values that, despite large amounts of CAPE in all of the atmospheric experiments, convection was not initiated for every case. Because CAPE only represents the potential for deep convection, a trigger must be available in the atmosphere for the thermal energy to be released and become kinetic energy. Seitter (1986) showed that the lifting of air by the head of a density current could result in cloud development but that the vertical displacement must be sufficient to saturate the air. Thus, an additional trigger is necessary to induce deep convection.

In this study, the trigger mechanism was assumed to be the concave terrain line (Figs. 5, 6). Although the SLP-CC case displayed weak convergence ($-1.93 \times 10^{-3}$ s$^{-1}$) and divergence ($1.14 \times 10^{-3}$ s$^{-1}$) fields, the maxima of the divergence fields of the BSN-CC ($-3.2 \times 10^{-3}$ and $3.3 \times 10^{-3}$ s$^{-1}$) and BSNV-CC ($-4.1 \times 10^{-3}$ and $2.85 \times 10^{-3}$ s$^{-1}$) cases had significantly stronger values with a centralized mass flux toward the middle of the basin. Nevertheless, the BSN-CC case lacks the production of a cumulus cloud. This is due to the lower LFC in the BSNV-CC case (902 hPa), which is 100 m lower than that for BSN-CC (Table 2). Because of weak convergence and negative buoyancy, the mass contribution in the BSN-CC case is insufficient for the air to become saturated. The several draining valleys in the BSNV case provide an essential increase in mass in the middle of the basin. Nevertheless, the BSN-CC case lacks the production of a cumulus cloud. This is due to the lower LFC in the BSNV-CC case (902 hPa), which is 100 m lower than that for BSN-CC (Table 2). Because of weak convergence and negative buoyancy, the mass contribution in the BSN-CC case is insufficient for the air to become saturated. The several draining valleys in the BSNV case provide an essential increase in mass in the middle of the basin that also affects the surface fluxes. Because the geometry of the basin is identical to the BSN case except for the valleys, this additional mass contribution (which actually occurs in the Andes Mountains) is a crucial factor in the cloud-formation process. In this process, the most accurate presentation of the complex terrain of the Andes in the idealized case studies is an important feature of this study. Laurent
et al. (2002) demonstrated that MCS initiation in South America is mostly driven by topography. Calbo and Millan (1998) also showed the role of the terrain and highlighted its influence on the choice of the model grid size.

No convective events were generated under the NC conditions—not even with the BSNV experiment. This was caused by absence of both any considerable mass flux into the basin and any frontal structures in the area (Figs. 6, 7). The BSNV-CC case contained the underlying process for convection, whereas the BSNV-NC case did not. The cold air is caused by the katabatic flows and is drained into the basin through the slopes and valleys. As the air drains down the slope because of the gravitational force, it is adiabatically heated. The warming is not sufficient to compensate for the diabatic cooling and mechanical thermal energy loss, thereby leading to a net cooling of the air adjacent to the surface. This cooling causes a horizontal temperature gradient to develop inside the basin, resulting in the formation of a surface cold front.

Examination of the cross section of the surface cold front (Fig. 7) and of the characteristics of katabatic flows (Fig. 8) shows that the BSNV-CC case produced a strong katabatic flow whereas only weak katabatic flow was produced in the BSNV-NC case. These differences occurred because of variations in the vertical surface fluxes (Fig. 9). The intensity of the flow in the BSNV-CC case was considerably stronger, particularly in $H$, which is essential for the development of a katabatic flow. The lower troposphere is very sensitive to modifications of the surface temperature and humidity affecting the PBL structure, which is organized by surface and turbulent fluxes. The vertical fluxes can result in modifications to these fields, which are sensitive to thermodynamic activities of this nature (Mapes and Houze 1992). The surface fluxes, however, are the main driver for the development of thermally induced katabatic flows, which were obviously modified in the BSNV-NC case. The initial nocturnal boundary layer created a stable ambient atmosphere causing the downslope velocity and the height of the katabatic layer to greatly decrease, affecting the subsequent upward vertical displacement (Axelsen and Dop 2009). The heat exchange coefficient $C_{th}$ [see Eq. (3)] is dependent on the thermal instability of the atmosphere. Thus, because of the initial stable profile in the BSNV-NC case, $C_{th}$ is less than that for the BSNV-CC case. This condition results in a weaker horizontal gradient of the surface air to the ambient air at any given height level that in turns affects the development and strength of the katabatic flow.

### 5. Summary and conclusions

In this study the ARPS numerical model was used to investigate the development of a convective cloud cluster under different terrain configurations. The main purpose of this study was to find the mechanism of cloud formation. It was hypothesized that katabatic flows from the slopes and valleys of the Andes Mountains converge at the foothills because of a quasi-concave terrain shape, where they generate a surface cold front and subsequent convective initiation.

In this study three simplified terrain models were used: a simple slope, a uniformly concave ridgeline, and a ridgeline with several valleys draining into a basin. The BSNV case most closely resembles the features of the Andes Mountains. The influence of the PBL structure on the evolution of the main driving feature for convection was investigated by applying three atmospheric stratification experiments (SC, NC, and CC) to the terrain models. This combination of terrain models and stratification experiments outlined the importance of the terrain and the PBL structure, respectively.

A comparison of the occurrence of clouds in each of the simulations gave a first insight into the process of convective initiation. Only the BSNV-SC and BSNV-CC cases produced a convective cloud, whereas the SLP and BSN cases failed to initiate any convective activity. An analysis of $\theta_e$ showed a potential instability in all of the cases, and the vertical gradient of $\theta_e$ in the NC case was the weakest. CAPE was largest in the SC and CC experiments, and CIN showed lower values than in the NC experiment. An inspection of the horizontal divergence field for the SLP, BSN, and BSNV cases provided additional information about the terrain’s impact, which is considered to be part of the main trigger mechanism for cloud formation. Neither the SLP-CC nor the BSN-CC case provided sufficient convergence to lift the air masses to their LFC. Only the BSNV-CC case generated a strong mass contribution inside the basin, which led to ascending air and the release of latent heat. A cross section through the areas of the BSNV-NC and BSNV-CC cases illustrated the reason for these differences. A cold front was present in the BSNV-CC case but was lacking in the BSNV-NC case. Evidence was found for a katabatic flow that produced frontal structures, and its features were compared for the BSNV-NC and BSNV-CC cases. Using typical characteristics such as a jetlike wind profile, the katabatic flow’s formation was shown. The driving force of the flow was demonstrated by the surface fluxes, which indicated an energy flux from the PBL to the surface layer. Furthermore, the katabatic features suggested that the marginal development of downslope flow in the BSNV-NC case was the cause of insufficient horizontal divergence and the absence of convective activity. It was found, on the basis of evidence of a smaller magnitude of the heat-exchange budget in the BSNV-NC case, that the driving forces for the mechanism of cloud formation are an
evolved katabatic flow and a complex terrain structure with a concave shape and an extensive drainage system. In the shear experiment, deep convection was initiated simply by the introduction of an opposing flow with strong wind shear. The convective initiation occurred regardless of the atmospheric stratification or terrain geometry.

The simulation results of this study support the hypothesized mechanism of convective initiation. Moreover, the results highlight the significance of the complex structure of the terrain in the Andes Mountains as well as the atmospheric conditions. Thus, the idealized case studies presented here verify our hypothesis that katabatic flows interacting with the concave terrain cause the development of surface cold fronts, which initiate lifting of air to its LFC and cloud cluster formation if the convergence is sufficient.

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