Impacts of the Andes cordillera on precipitation from a midlatitude cold front

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Abstract

While mesoscale details of frontal interactions with mountain ranges are complex, a commonly observed modification is the deformation and deceleration of surface fronts and precipitation upstream of the high terrain. However, the effects of the Andes Cordillera, the major mountain range in South America, on precipitation patterns of baroclinic systems approaching from the SE Pacific remain largely unstudied. This study focuses on a case in late May 2008 when an upper-level trough and surface cold front produced widespread precipitation in central Chile. The primary goal is to analyze the physical mechanisms responsible for the structure and evolution of the precipitation. The analysis is principally based on mesoscale numerical simulations with the Weather Research and Forecasting model (WRF), complemented by in-situ and satellite observations.

The WRF simulations indicate that as an upper-level trough approached central Chile, mid-tropospheric flow below 600 hPa was blocked by the high topography and deflected poleward in the form of a barrier jet. This northerly jet had wind maxima in excess of 15 m s\(^{-1}\), was centered around 925 hPa, and extended westward 200 km from the mountains. It intersected the cold front, which approached from the south near the coast, increasing convergence along the frontal surface, slowing its equatorward progress, and enhancing rainfall over central Chile. Another separate region of heavy precipitation formed over the upwind slopes of the cordillera. A trajectory analysis confirmed that the barrier jet moved low-level parcels from their origin in the moist SE Pacific boundary layer to the coast. When model topography was reduced to twenty percent of its original height, the cold front advanced more rapidly to the northeast, generated less precipitation in central Chile between 33°S and 36°S, and produced minimal orographic precipitation on the upwind Andean slopes. Based on these findings, we conclude
that the high topography was responsible for not only orographic precipitation but also for substantially increasing precipitation totals over the central coast and valley.
1. Introduction

The Andes cordillera is the major mountain range in the Southern Hemisphere. It extends for more than 5000 km along the west coast of South America, from north of the equator to the southern tip of the continent, with mean heights ranging from 1000-1500 m ASL in midlatitudes to up to 5000 m ASL at subtropical and equatorial latitudes, and its cross-mountain scale is about 200 km. The Andes impact tropospheric circulation on a broad range of scales, from the generation of mesoscale eddies (e.g., Garreaud et al. 2003) to the disruption of weather systems (e.g., Seluchi et al. 2006; Garreaud and Fuenzalida 2005) to the organization of the South American monsoon system (e.g., Vera et al. 2006). The systems impinging the southern portion of the continent are embedded in the south Pacific storm track whose axis is located between 40-50°S throughout the year (Trenberth 1991; Hoskins and Hodges 2005). Much attention has been devoted to the channeling effects of surface anticyclones along the east (downstream) side of the Andes that lead to rapid incursions of cold air that may reach as far north as the Amazon basin (Seluchi et al. 2006). The disruption of weather systems, particularly cold fronts, along the west (upstream) side of the extratropical Andes has been less studied, in part because of the lack data over the adjacent Pacific Ocean.

Orographic influences on frontal systems have been studied previously in other locations around the world. One of the most prominent effects of mountains on the lower atmosphere is “upstream blocking” (Yu and Smull 2000). For the case of a steep, nearly two-dimensional barrier, such as that presented by the central Andes, a major result of upstream blocking is the development of strong low-level flow parallel to the terrain axis (Parish 1982) resulting from downgradient ageostrophic acceleration in the along-barrier direction (Overland 1984; Lackmann and Overland 1989). These “barrier jet” wind maxima are often supergeostrophic
Low-level blocking is common when flow normal to the terrain axis is characterized by a large $Nh/U$ regime (a regime with a small Froude number), where $N$ is the buoyancy frequency, $h$ is the height of the mountain, and $U$ is the speed of the free airstream (Doyle 1997; Pierrehumbert and Wyman 1985; Overland and Bond 1995).

While the mesoscale details of topography-frontal interactions are complex, a commonly observed modification is the deformation and deceleration of surface fronts and precipitation upstream of the high terrain (Neiman et al. 2004). Bjerknes and Solberg (1921) were among the first to observe this phenomenon, noting that a warm front deformed as it approached coastal Norway, with the lower tropospheric portion precluded from crossing the complex topography. More recent observations in the Olympic Mountains of western Washington State documented the development of strong poleward surface winds as cold fronts approached the coastal range due to mesoscale pressure ridging upstream of the topography (Mass and Ferber 1990). A similar low-level, prefrontal wind maximum was observed adjacent to the steep terrain along the central California coast (Doyle 1997). Other observations of fronts decelerated by upstream flow blocked by topography have been made in the European Alps (e.g., Kurz 1990), the Appalachian Mountains (O’Handley and Bosart 1996), the coastal mountains of western Canada (Doyle and Bond 2001), and the Pacific Northwest (Braun et al. 1999). Finally, in addition to these observational studies, there are multiple theoretical experiments that describe the inverse relationship between the Froude number and the degree of frontal retardation (e.g., Egger and Hoinka 1992).

In the broadest sense, the Andes south of 30°S exhibit a typical rainfall pattern for a midlatitude mountain range: a relatively wet upwind side and relatively dry region downwind.
(e.g. Smith 1979; Roe 2005). Around 40°S, for instance, annual mean precipitation varies from 2000-3000 mm along the Pacific (Chilean) coast and western slope of the Andes to less than 200 mm just to the west of the continental divide over Argentinean Patagonia—a distance of less than 300 km in the zonal direction. Such a marked west-east gradient is often mentioned as a classical example of upslope rain enhancement and downslope rainshadow effects (Smith and Evans 2007). Furthermore, the local mid-tropospheric zonal flow exhibits a significant, positive (negative) correlation with precipitation west (east) of the Andes at both synoptic scale (Falvey and Garreaud 2007) and interannual time scale (Garreaud 2007).

A marked north-south precipitation gradient is also observed along the west side of the Andes cordillera. Records from coastal and inland stations in central Chile show a steep increase from about 100 mm year$^{-1}$ at 30°S up to 2000 mm year$^{-1}$ at 40°S, leading to a transition from arid conditions to rain forests. This meridional precipitation gradient is co-located with a gradient in terrain elevation: the Andes mean height decreases from about 4500 m ASL at 30°S down to 1500 m ASL at 40°S. Some authors in the geological community have speculated that the inverse relationship between precipitation and Andes’ elevation is rooted in a long-term relationship between climate, erosion and tectonics (see Farías et al. 2008 for a review). The narrow strip of land between the Andes and the Pacific coast also concentrates about 10 million inhabitants, about 70% of the Chilean population, so understanding the orographic control of precipitation along central Chile is an important practical issue.

To advance in our understanding of the upstream influence of the Andes upon midlatitude weather systems, and hence precipitation distribution, in this study we examine the structural evolution of a cold front that passes over central Chile in May (late fall) 2008. As shown later, the event had typical features and produced 30-50 mm of rainfall in central Chile. Our analysis is
based on in-situ and satellite data, as well as the results from two mesoscale numerical simulations (one control simulation and one with reduced topography) performed with the Weather and Research Forecast model (WRF; Skamarock et al. 2005). The remainder of the article is organized as follows. Synoptic and mesoscale observations of the event are presented in section 2. Section 3 describes the model control and sensitivity experiments, and model output is analyzed in section 4. Our conclusions are presented in section 5.

2. Episode overview

In this section, we present a surface and upper-air overview of the cold front passage in central Chile from 25 to 28 May 2008. We selected this case because it represents a common autumn-winter synoptic pattern whereby a positively-titled mid-level trough approached subtropical South America, promoted frontogenesis along and offshore of the Chilean coast, and led to widespread precipitation in central Chile. We base our analysis on GOES-12 visible images, upper-air data from the NCEP/NCAR Reanalysis (Kalnay et al. 1996), and surface observations from a total of 41 stations in central-south Chile operated by the National Weather Service (DMC; 28), National Water Bureau (DGA; 12) and Universidad de Chile (1). Figure 1 shows the locations of the observing stations, the total precipitation recorded at each station during the period, and the topography of the region.

At 1800 UTC 25 May there were two primary upper-air features over South America. First, a trough axis at 500 hPa extended from a vorticity maximum centered near 40°S, 85°W southeastward to the Antarctic peninsula (Fig. 2a), and second, zonal (cross-barrier) flow at 500 hPa between 15 and 30 m s⁻¹ was impinging on the central and south-central Chilean Andes. At the surface, a cold front was identifiable in visible satellite imagery as a band of low and middle
clouds extending from near 38°S, 72°W northwestward to 35°S, 85°W (Fig. 2a). The open-ocean portion of the front was characterized by a temperature gradient around 5°C/100 km. By May 26, an extensive area of shallow open cloud cells dominates off southern Chile, indicative of very cold low-level air moving behind the front (Fig. 2b). Surface reanalysis shows a large and intense (1035 hPa) cold-core (5-7°C) post-frontal anticyclone over the southeast Pacific. The intensity of the surface anticyclone is consistent with a marked mid-level ridge extending well into midlatitudes at 95°W (Fig. 2b). Near the ocean surface, the thermal contrast between the cold anticyclone and the warm subtropical air extended along an east-west baroclinic zone over 1,200 km into the southeast Pacific. The temperature gradient along the front tightened to reach its maximum value of 7.5°C/100 km on 1800 UTC 26 May. By 1800 UTC 27 May, the 500 hPa vorticity maximum over the SE Pacific had moved eastward to a position along the cordillera near 40°S and was in phase with a large closed low east of Argentina in the South Atlantic (Fig. 2b). A positively-tilted trough axis extended northwestward from this vorticity maximum into the SE Pacific. A well-defined band of cumulus clouds (Fig. 2b) identified the position of the cold front and extended from near 30°S, 80°W in the SE Pacific southeastward to the Andes cordillera, where it then continued eastward into western Argentina.

The cold front was also observed by the Quick Scatterometer (QuikSCAT; Ebuchi et al. 2002; Leslie et al. 2008) and Tropical Rainfall Measuring Mission (TRMM; Simpson et al. 1996) Microwave Imager (TMI; Kummerow et al. 1998) satellite instruments (Fig. 3a-d). The QuikSCAT data revealed a well-defined wind shift associated with the cold front, oriented from SE to NW offshore central Chile on May 25 (Fig. 3a), and confirmed that this wind shift progressed equatorward, passing north of 27°S after May 28 (Fig. 3d). The open-ocean section of this wind shift maintained a SE to NW orientation as it advanced northeastward. However,
the along-shore section, which was initially oriented SSE to NNW, deformed over time and became nearly parallel to the Chilean coast. As a result, this section of the front did not advance as rapidly equatorward as the open-ocean section. TMI rainfall data confirmed that a region of precipitation with intensities from 2 to 4 mm hr\(^{-1}\) was co-located with the wind shift over the southeast Pacific on May 26 (Fig. 3a). This precipitation region was observed to advance equatorward with the wind shift and diminish in intensity (Fig. 3b), eventually dissipating by May 27. In addition to the decline in open-ocean precipitation rates, the temperature gradient also weakened, decreasing from around 4°C 100 km\(^{-1}\) on May 27 to 1.5°C 100 km\(^{-1}\) by May 28.

The decline in precipitation intensity is consistent with the reversal of quasi-geostrophic forcing that occurred after the upper-level trough axis passed to the east of the surface front, and the decrease in the temperature gradient is likely aided by the passage of the cold, post-frontal air over progressively warmer sea surface temperatures as it moved northward.

We also examined the progress of the cold front in time series of temperature, dew point temperature, and air pressure at three surface stations. Hourly observations from Santiago (DGF), as well as synoptic observations from Concepción (CNP) and La Serena (LAS; see Fig. 1), show that although the cold front progressed northward along the coastal and central valley of Chile, its strength – measured by changes in temperature, dew point temperature, and pressure – and total precipitation gradually diminished. The front arrived first at CNP, evidenced by a 10°C drop in dew point temperatures as well as an 8 mb rise in surface pressure and a 3°C drop in surface temperature (Fig. 4a-c). A total of 63 mm fell at CNP, beginning by 1800 UTC 25 May and ending around 0600 UTC 27 May. Farther north in the valley at DGF, the cold front arrived around 1800 UTC 27 May, evidenced by a temperature drop from 11°C to 6°C, a 4 hPa rise in pressure, and a 5°C decrease in dew point temperature. Rainfall began shortly after 1200 UTC
26 May, and continued until about 0000 UTC 28 May, accumulating 43 mm in 36 hours. The rainfall totals, temperature and dew point change, and pressure rise were each less at DGF than at CNP, indicating that the cold front was weakening as it progressed northward. This trend continued at LAS, the northernmost of the three stations in this analysis, which measured a temperature decrease of only 4°C, dew point decrease of 2.5°C, pressure rise of 4 mb, and only 7 mm of precipitation. Other surface stations also reflect this precipitation pattern, with southern sites near Concepcion receiving 60-80 mm of event-total rainfall, locations near Santiago receiving between 40-60 mm, and stations near La Serena receiving less than 10 mm (Fig. 1). A maximum of 150 mm was measured at an elevation of 1500 m in the cordillera southeast of Santiago.

3. Model description

To better understand the temporal and spatial evolution of this precipitation event, we ran a 72 hr simulation using version 2.2 of the Advanced Research Weather Research and Forecasting Model (WRF-ARW, or simply WRF; Skamarock et al. 2005), initialized at 1200 UTC 25 May 2008, about 24 hours before the onset of precipitation in central Chile. The single-domain forecast was centered at 35°S, 75°W with 351 grid points in the x-direction and 350 grid points in the y-direction at a horizontal resolution of 10 km and a timestep of 50 s (see Fig. 1 for an outline of the model domain). The model top was prescribed as 50 hPa and vertical sigma-level resolution decreased from 15 hPa near the surface to 60 hPa at the model top, yielding 24 unevenly-spaced half-sigma levels. The initial and lateral boundary conditions were obtained every 6 h from the NCEP Global Forecasting System (GFS; Caplan and Pan 2000) with 1° horizontal resolution. Additional physical parameterization choices are summarized in Table 1.

To isolate the impact of the high Cordillera on the evolution of the front as well as the
magnitude and distribution of frontal precipitation, two simulations of the event were performed: a control (CTL) simulation, with full topography and all the choices discussed above and in Table 1, and an experiment (LOW) with topography over the entire domain reduced to 20 percent of the control. Other than the topography, all other model characteristics in LOW, including the physics options, were kept equal to CTL.

4. Results

a. Evolution of precipitation features (PFs)

A qualitative analysis of the temporal evolution of the 3-hr precipitation, 975 hPa wind, and 500 hPa geopotential height from the CTL simulation (Fig. 5) reveals that three primary precipitation features (PFs) were associated with this event: a “Pacific precipitation zone” (PPZ), located over the SE Pacific west of the Chilean coast; a “coastal precipitation zone” (CPZ), extending from the eastern edge of the PPZ along the Chilean coast eastward to the Andes Cordillera, and an “orographic precipitation zone” (OPZ), located over the high Andean terrain. As we will show, the structure and evolution of each PF was directly related to the structure and evolution of lower- and mid-tropospheric wind field, including a topographically-forced barrier jet.

At the start of the CTL simulation, the PPZ was located over the southeast Pacific west of Chile and was oriented from SE to NW with precipitation rates of 2-3 mm hr$^{-1}$ (Fig. 5a). The CPZ, in contrast, was located along the central Chilean coast and had a more meridional orientation and precipitation rates between 3 and 5 mm hr$^{-1}$. The PPZ joined the CPZ 50 km west of the coast, and the two features remained connected throughout the remainder of the CTL simulation. The PPZ maintained a SE to NW orientation and advanced steadily equatorward
(Figs. 5b-d), and, as discussed in Section 2, after passing north of 27°S on 28 May, its precipitation intensity diminished.

In contrast to the PPZ, the CPZ advanced equatorward more slowly, stalling over central Chile at about 36°S during 27 May with precipitation rates from 5-10 mm hr\(^{-1}\) (Fig. 5c). By the end of the simulation at 1200 UTC 28 May, the CPZ had become oriented nearly parallel to the coast (Fig. 5d). This change in orientation is physically consistent with similar cold front-terrain interactions observed off the California coast (e.g., Doyle 1997; James and House 2005) and suggests that features observed in that region, including a topographically-forced barrier jet, may have also been present in central Chile. The OPZ formed around 1200 UTC 26 May 2008 over the Chilean cordillera (Fig. 5b) as zonal wind speeds increased and static stability decreased ahead of the approaching upper-level trough. These two factors combined to increase Froude numbers above the critical threshold, particularly for parcels originating above 600 hPa, and resulted in cross-barrier orographic precipitation intensities ranging from 1-2 mm hr\(^{-1}\). By 1200 UTC 27 May, the OPZ covered a large north-south portion (over 600 km) of the western Cordillera above 3000 m with precipitation rates over 10 mm hr\(^{-1}\). Similar to the coastal CPZ, the OPZ did not advance northward as rapidly as the PPZ, remaining roughly between 30°S and 36°S throughout the remainder of the event as strong cross-barrier flow and low atmospheric stability prevailed ahead of the upper-level trough axis.

b. Time-space cross-sections

We now examine the structure and evolution of the three PFs through a series of time-longitude and time-latitude cross-sections. An east-west cross-section along 33°S (selected because that parallel intersects with each of the three PFs during the CTL simulation) clearly shows both the PPZ, which moved eastward in time, and the OPZ, which formed just after 0000
UTC 26 May and remained stationary over the mountains near 70°W throughout the remainder of the simulation (Fig. 6a). The slope of the isohyets shows that the core of the PPZ advanced eastward about 6° day$^{-1}$, or 600 km day$^{-1}$, and this speed compares well with mean post-frontal surface wind speeds that averaged between 7 and 8 m s$^{-1}$. Precipitation intensities along the PPZ remained steady for the first half of the simulation as the open-ocean portion of the front advanced equatorward. An east-west time-latitude cross section of 700 hPa $u$-wind component along 33°S (Fig. 6c) confirms the relationship between zonal wind speed and both the intensity and evolution of the PPZ and, especially, the OPZ (Fig. 6a). For example, the eastward progression of wind speeds above 15 m s$^{-1}$ matched almost exactly the location and eastward progression of the PPZ. Furthermore, 700 hPa $u$-wind maxima greater than 20 m s$^{-1}$ were aligned along the cordillera near 70°W, similar to the alignment of the OPZ in Fig. 6a (the region between 69°W and 70.5°W at 700 hPa is below ground, hence the small gap in Fig. 6c). This relationship between stronger mid-tropospheric westerly flow and surface precipitation agrees with Falvey and Garreaud (2007), who found that central Chile precipitation was most strongly correlated with the 700 hPa $u$-wind component.

A north-south cross-section of precipitation along the Chilean coast more clearly shows the full evolution of the CPZ. This PF advanced northward an average of 350 km day$^{-1}$ (half as fast as the PPZ), reaching 30°S by the end of the simulation (Fig. 6b). Note that the CPZ stalled around 36°S for 18 hours on May 26. During this period, precipitation intensities reached their maximum and ranged from 6 to 10 mm hr$^{-1}$. Surface stations in the vicinity of Concepción (located near 36°S) confirmed this peak in precipitation, recording nearly all their event total precipitation, between 40 and 80 mm, during this 18-h period. After resuming its northward progress along the coast, precipitation intensities peaked again near 33-34°S on May 27, and this
peak was also confirmed by surface observations in the region. After passing north of 33°S, precipitation along the CPZ diminished in intensity.

A north-south cross-section of 925 hPa $v$-wind component along the Chilean coast (Fig. 6d) shows strong agreement between precipitation intensity (Fig. 6b) and magnitude of the northerly low-level flow. The location, structure, and evolution of this wind maximum were very similar to those of the precipitation. For example, a maxima in northerly flow (speeds greater than 15 m s$^{-1}$) advanced steadily northward until 0600 UTC 26 May, stalled for 18 hours around 36°S, and then continued advancing northward (Fig. 6d), very similar to the evolution of the CPZ (Fig. 6b). Wind maxima in the northerly flow corresponded, in both timing and location, to the greatest precipitation intensity of the CPZ. Additionally, where northerly flow was weaker, including north of 33°S, precipitation intensity along the CPZ was also weaker. A longitude-height cross-section along 34°S reveals that the northerly flow was characterized by wind speeds above 15 m s$^{-1}$, that it extended westward from the cordillera by about 200 km, and that it was located above the frontal surface between 850 and 950 hPa (Fig. 7). A maximum in specific humidity (> 10 g kg$^{-1}$) was co-located with this wind maximum, indicating poleward transport of moisture from the subtropics. These details collectively describe a barrier jet that formed in response to lower-atmospheric flow blocking by the Andes and are evidence that the barrier jet clearly influenced precipitation in coastal Chile during the cold front event.

c. Trajectories

In this subsection we use a parcel trajectory analysis to further diagnose the forcing for precipitation, as well as airmass source region and vertical motion. Using a simple, tri-dimensional advective technique, we produced backward and forward trajectories at all model sigma levels for two locations, one along the coast at the northern edge of the CPZ (Fig. 8), and
the other along over the cordillera in the OPZ (Fig. 9). The trajectories were calculated backward for the 24 hours leading up 1800 UTC 26 May and forward for the 24 hours following. Parcels that approached the point just north of the cold front (Fig. 8) clustered into two main groups. Those parcels below 1500 m approached first from the northeast and then from the north, originally following the synoptic northwesterly flow over the SE Pacific before turning south as they encountered the northerly barrier jet close to the coast (Fig. 8a, open circles). These parcels either maintained their height or rose slightly as they traversed the SE Pacific northwest of the front, but once they reached the CPZ (rainfall along the front is indicated by the stippled region), they ascended several thousand meters (Fig. 8b) over a short distance. After passing through the frontal zone, the parcels turned easterly with the prevailing flow and continued across the southern cordillera into Argentina. Those parcels located above 1500 m approached generally from the west, following the prevailing mid- and upper-level flow (Fig. 8a, closed circles), and also converged and ascended upon reaching the front. We believe the CPZ trajectories confirm the role of the barrier jet as a “transport agent,” moving low-level parcels from their origin in the moist SE Pacific boundary layer to the coast. Once reaching the front, the moist parcels converged, quickly ascended, and produced heavy rainfall in the CPZ.

Farther north in the OPZ, the trajectories reveal additional information about the character and source of the region of orographic precipitation. Apart from a few parcels that appear to have been entrained in the flow from the lower troposphere, nearly all of the flow that crossed the Cordillera originated above 2500 m (Fig. 9b), including those few parcels whose trajectories brought them from the north and northwest (Fig. 9a). This result agrees with a Froude number analysis of the flow for this region, where Froude numbers greater than 0.9 (less than 0.9), which are favorable for cross-barrier (blocked) flow, were common above (below) 600
hPa. As the flow approached the cordillera, parcels began ascending several hundred kilometers west of the coast, and those parcels below 3500 m (open circles) that originated over the northern coast transported mid-level moisture that contributed to the orographic precipitation. In addition, it appears that the ascent of mid-troposphere parcels as they approached the cordillera from the west, including those above 3500 m (closed circles), augmented precipitation totals over central Chilean valley including at Santiago.

\textit{d. Evolution of LOW simulation}

For additional insight into the role of the Andean topography in the evolution of the frontal passage, we re-ran the CTL simulation with topography reduced to 20\% of the original. The change in topography had little impact on the large-scale mid- and upper-level synoptic development. However, the evolution of the surface fronts and their associated precipitation was markedly different in the two simulations. We identified three main differences in the structure and evolution of the PFs (Fig. 10). (1) The PPZ and CPZ both advanced equatorward faster in LOW (Fig. 10a-d) than in CTL (Fig. 5a-d). Specifically, the PPZ reached 28°S about 12 h earlier in the LOW (Fig. 10c) than in the CTL (Fig. 5d), and the CPZ did not stall over central Chile but instead moved steadily northeastward, crossing into Argentina during the simulation. The surface front itself also advanced northeastward much more rapidly in LOW than in CTL, reaching the edge of the model domain about 12 h earlier than the CTL (Fig. 10). As it advanced equatorward, it maintained its shape, even across the continent (Fig. 10b-c), and did not deform as it encountered the topography. This difference is seen clearly by comparing the slopes of the CPZ in the east-west (Figs. 6a and 11a) and north-south (Figs. 6b and 11b) cross-sections: the isohyets have a larger slope in the CTL than in the LOW, indicating that the CPZ moved.
equatorward more rapidly in the LOW experiment. This deformation appears to be forced by the barrier jet. (2) Precipitation intensities of the CPZ, particularly over central Chile between 33°S and 36°S, were much lower in the LOW (Fig. 11b) compared to the CTL (Fig. 6b), indicating that the total forcing for precipitation was much weaker in the LOW experiment. This difference is also reflected in storm-total precipitation (Fig. 12), where precipitation was much greater in the CTL than the LOW simulation, and concentrated between 33°S and 36°S. The LOW simulation captured neither the magnitude of either the coastal and valley rainfall between 36°S and 39°S nor the heavy precipitation along the Cordillera. This difference in precipitation totals between the two simulations is directly related to topography: as the northerly barrier jet slows the front’s equatorward progress, the jet increases convergence over the frontal slope, thus producing heavier precipitation (because of the greater convergence) that lasts longer (because of the slower frontal progression). (3) The OPZ, one of the three major PFs visible in the CTL simulation, was completely absent from the LOW experiment (Figs. 11a, 12a). The reduced topography was not high enough to generate significant orographic precipitation. These three differences clearly demonstrate that the high Andean terrain has a significant impact on both the structure and the magnitude of frontal precipitation features in central Chile.

5. Conclusions and discussion

The goal of this study was to improve understanding of precipitation patterns in central Chile during a cold front passage and examine the influence of the Andes mountains on the amount and distribution of rainfall. To achieve this goal, we simulated an event from 25-28 May 2008 with both full and reduced topography. The event produced widespread precipitation over central Chile, with totals between 40 and 80 mm at coastal and valley locations increasing to
over 150 mm in the pre-cordillera. Through a combination of surface measurements, satellite observations, and numerical simulations, we identified three precipitation features and categorized them by their location (Pacific, coastal, orographic). Trajectories over the cordillera revealed that nearly all the parcels crossing the high Andes barrier north of 35°S originated above 600 hPa, consistent with the Froude number profile of the approaching airmass. Sub-600 hPa flow was blocked by the topography and subsequently led to the formation of a northerly barrier jet. Farther south, trajectories calculated at points originating over the cold front indicate that the barrier jet transported moist air from the subtropical oceanic marine boundary layer southward along the Chilean coast an then up and over the frontal surface, enhancing convergence and rainfall over central and southern Chile. When topography was reduced, we found that the character of the precipitation changed dramatically. The cold front advanced more rapidly to the north and northeast, greatly reducing precipitation in the central valley between 33°S and 36°S and eliminating all of the precipitation along the Cordillera and most of it south of 36°.

We propose the following physical mechanism to explain the precipitation pattern and role of Andean topography. As the upper-level trough approached central Chile, mid-tropospheric flow was blocked by the high topography and deflected poleward in the form of a barrier jet. This northerly jet intersected the near-coast section of the cold front as it approached from the south, producing two related effects. First, the northerly flow increased convergence along the frontal surface, producing widespread rainfall totals of 40 to 80 mm over the region between 33° and 39°S. Second, the northerly flow impinged upon the frontal surface and slowed its equatorward progress, causing the front to deform as its open-ocean section continued northward. The slowed progress of the coastal front increased the duration of convergence of
northerly flow along the frontal surface and thus increased rainfall over the CPZ. Mid-tropospheric flow was able to cross the topography and subsequently produced up to 300 mm of orographic precipitation along the Cordillera.

Finally, precipitation totals from the LOW simulation were significantly lower than observed at nearly all of the 41 observing stations, and they were also much lower than totals from the CTL simulation. We therefore conclude that the high topography significantly augmented precipitation during this cold front passage through central Chile. The topography was directly responsible for orographic precipitation along the Cordillera, and by sustaining an enhanced northerly convergent flow over the frontal surface, it also substantially increased precipitation totals over the central coast and valley.

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Fig. 1. Terrain height, in meters above sea level (m ASL) for the control simulation. Black squares indicate the locations of the 41 observing stations, and square size indicates the amount of precipitation which fell during the cold front passage from 25-29 May 2008. Time series of temperature, dew point, and air pressure are presented in Fig. 4 for Concepcion, Santiago, and La Serena.

Fig. 2. Visible satellite image and 500 hPa reanalysis geopotential height overlay at 1800 UTC 25 May (a) and 1800 UTC 28 May (b). Heights are contoured every 60 m.

Fig. 3. QuikSCAT surface wind and TRMM 3-hr rainfall rate (in mm 3 hr$^{-1}$) observations every 12 hr from 1200 UTC 26 May to 1200 UTC 28 May. Wind speed is scaled to 10 m s$^{-1}$. The wind shift and precipitation associated with the cold front moved equatorward with time.

Fig. 4. Time series of (a) temperature in °C, dewpoint temperature (b) in °C, and station air pressure (c) in mb at Concepción (asterisks), Santiago (circles), and La Serena (diamonds), from 1200 UTC 25 May to 0000 UTC 29 May. Time is in UTC. Frontal passage, indicated by the dotted lines, occurred at Concepción first, then Santiago, and finally at La Serena. Gray sections in (b) indicate duration of observed rainfall. Station locations are given in Fig. 1.

Fig. 5. CTL simulation output of 3-hr precipitation totals, 975 hPa wind, and 500 hPa geopotential height. Wind speed is scaled to 10 m s$^{-1}$ and geopotential height is contoured every 60 m. Notice the equatorward progression of the cold front and precipitation features (PFs).
the three PFs are easily identifiable: Pacific precipitation zone (PPZ) located over the SE Pacific, the coastal precipitation zone (CPZ) located over the Chilean coast and central valley, and the orographic precipitation zone (OPZ) located over the high cordillera along the Chile-Argentina border.

Fig. 6. CTL simulation cross-sections of (a) east-west 3-hour precipitation (mm) along 33°S; (b) north-south 3-hour precipitation (mm) along the coast; (c) east-west 700 hPa $u$-wind along 33°S (m s$^{-1}$); and (d) north-south coastal 925 hPa $v$-wind (m s$^{-1}$). Wind speed in (c) and (d) are shaded every 5 m s$^{-1}$. In (c), part of the topography extends above 700 hPa and is indicated by the white bar. In (d), stippled contours indicate southerly flow.

Fig. 7. Longitude-height cross-section of horizontal wind (vectors, with $v$-wind shaded), potential temperature (thin black lines), and specific humidity (thick dashed lines), at 34°S on 27 May 2008. The near-surface northerly barrier jet is well-defined between the mountains and 75°W and the surface and 850 hPa. The western half of the barrier jet is co-located with a maximum in specific humidity, indicating moisture transport from the subtropical Pacific to the north.

Fig. 8. Simple advective trajectory analysis showing locations of all parcels that passed through a point located just north of the synoptic cold front at 1800 UTC 26 May. Open (filled) circles represent hourly parcel locations below (above) 1500 m, and in (a), model 3-h rainfall is stippled and terrain height is shaded. The cluster of sub-1500 m parcels arriving at the cold front from the north represents the barrier jet.
Fig. 9. Same as Fig. 8, but for a point over the cordillera. Open circles represent parcels below 3500 m. Nearly all parcels that cross the barrier originate at or above 2500 m, indicating the presence of strong blocking for parcels below 700 hPa.

Fig. 10. Same as Fig. 5, but for LOW simulation. Note the only PF in the LOW simulation is the PPZ, located east of the coast, and the orographic OPZ does not form without high terrain.

Fig. 11. LOW simulation cross-sections of (a) east-west 3-hour precipitation (mm) along 33°S; (b) north-south 3-hour precipitation (mm) along the coast; (c) east-west 700 hPa $u$-wind along 33°S (m s$^{-1}$); and (d) north-south coastal 925 hPa $v$-wind (m s$^{-1}$). Wind speed in (c) and (d) are shaded every 5 m s$^{-1}$. In (d), stippled contours indicate southerly flow.

Fig. 12. Total precipitation from the CTL (a) and LOW (b) simulations.
Table 1. Summary of physical parameterization selections used in the WRF control and low topography simulations.

<table>
<thead>
<tr>
<th>Physical parameterization</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Radiation</td>
<td></td>
</tr>
<tr>
<td>Shortwave: Goddard</td>
<td>Chou and Suarez (1994)</td>
</tr>
<tr>
<td>Longwave: rapid radiative</td>
<td>Mlawer et al. (1997)</td>
</tr>
<tr>
<td>transfer model (RRTM)</td>
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</tr>
<tr>
<td>Timestep: 30 minutes</td>
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</tr>
<tr>
<td>Microphysics</td>
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<tr>
<td>Purdue-Lin</td>
<td>Lin et al. (1983)</td>
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<tr>
<td>Subgrid-scale precipitation</td>
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<tr>
<td>Kain-Fritsch</td>
<td>Kain and Fritsch (1993)</td>
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<tr>
<td>Surface layer processes</td>
<td></td>
</tr>
<tr>
<td>Noah land-surface model</td>
<td>Chen and Dudhia (2001)</td>
</tr>
<tr>
<td>Turbulence kinetic energy</td>
<td></td>
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<tr>
<td>Mellor-Yamada-Janjić</td>
<td>Mellor and Yamada (1982);</td>
</tr>
<tr>
<td></td>
<td>Janjić (2001)</td>
</tr>
</tbody>
</table>
Fig. 1. Terrain height, in meters above sea level (m ASL) for the control simulation. Black squares indicate the locations of the 41 observing stations, and square size indicates the amount of precipitation which fell during the cold front passage from 25-29 May 2008. Time series of temperature, dew point, and air pressure are presented in Fig. 4 for Concepcion, Santiago, and La Serena.
Fig. 2. Visible satellite image and 500 hPa reanalysis geopotential height overlay at 1800 UTC 25 May (a) and 1800 UTC 28 May (b). Heights are contoured every 60 m.
Fig. 3. QuikSCAT surface wind and TRMM 3-hr rainfall rate (in mm 3 hr\(^{-1}\)) observations every 12 hr from 1200 UTC 26 May to 1200 UTC 28 May. Wind speed is scaled to 10 m s\(^{-1}\). The wind shift and precipitation associated with the cold front moved equatorward with time.
Fig. 4. Time series of (a) temperature in °C, dewpoint temperature (b) in °C, and station air pressure (c) in mb at Concepción (asterisks), Santiago (circles), and La Serena (diamonds), from 1200 UTC 25 May to 0000 UTC 29 May. Time is in UTC. Frontal passage, indicated by the dotted lines, occurred at Concepción first, then Santiago, and finally at La Serena. Gray sections in (b) indicate duration of observed rainfall. Station locations are given in Fig. 1.
Fig. 5. CTL simulation output of 3-hr precipitation totals, 975 hPa wind, and 500 hPa geopotential height. Wind speed is scaled to 10 m s\(^{-1}\) and geopotential height is contoured every 60 m. Notice the equatorward progression of the cold front and precipitation features (PFs). In panel (b) the three PFs are easily identifiable: Pacific precipitation zone (PPZ) located over the SE Pacific, the coastal precipitation zone (CPZ) located over the Chilean coast and central valley, and the orographic precipitation zone (OPZ) located over the high cordillera along the Chile-Argentina border.
Fig. 6. CTL simulation cross-sections of (a) east-west 3-hour precipitation (mm) along 33°S; (b) north-south 3-hour precipitation (mm) along the coast; (c) east-west 700 hPa $u$-wind along 33°S (m s$^{-1}$); and (d) north-south coastal 925 hPa $v$-wind (m s$^{-1}$). Wind speed in (c) and (d) are shaded every 5 m s$^{-1}$. In (c), part of the topography extends above 700 hPa and is indicated by the white bar. In (d), stippled contours indicate southerly flow.
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