

Air motions and precipitation growth in Alpine storms

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(Received 21 January 2002; revised 15 July 2002)

SUMMARY

This study documents two strong Mesoscale Alpine Programme (MAP) storms. Each occurred ahead of a strong baroclinic trough. Even though their 500 and 850 hPa flow patterns were strikingly similar upstream of the Alps, their flow at lower levels was different and resulted in different forms of orographic enhancement of the precipitation associated with baroclinic systems by the Alpine terrain. During Intensive Observing Period (IOP) 2b, the low-level flow and thermodynamic conditions over the Alps north-west of the Lago Maggiore region constituted an unstable atmosphere with Froude number $Fr > 1$ (i.e. unblocked or flow-over regime), while in IOP 8 the flow was stable with $Fr < 1$ (i.e. blocked or flow-around regime in the region immediately upstream of the slopes on the western side of Lago Maggiore). For IOP 2b (IOP 8) the wind field and precipitation patterns north-west of the Lago Maggiore were very similar to the high- (low-) Fr autumn climatology. Thus the two cases represent fundamentally distinct regimes of orographic modification of the baroclinic precipitation.

During IOP 8 (blocked case), the strong stability and weak wind speed at low levels prevented the airflow from rising over the slopes north-west of the Lago Maggiore region and forced it to turn away from the mountains. The precipitation over the mountains was produced as the strong flow above 900 hPa was forced over the blocked low-level air as well as the terrain. In contrast, during IOP 2b (unblocked case) the flow was strong at all levels with low static stability; therefore the low-level air rose easily over the abruptly rising terrain. Since the lower-level air rose together with the upper-level air, it could transport upward moisture unavailable in the blocked case; this moisture condensed to add significantly to the precipitation production in the unblocked case. In addition to the high Fr , the air was potentially unstable; the slight instability of the airstream impinging upon the upslopes favoured the development of convective cells over the lower slopes which were embedded in the stratiform background lifting, thus further enhancing the formation of cloud and precipitation on the lower windward slopes.

Examination of all the cases observed by polarimetric radar in MAP confirmed the microphysical processes seen in IOPs 2b and 8, and suggests fundamental microphysical differences between the unstable unblocked, and stable blocked cases. In both cases precipitation forms by a simple stratiform process in the form of dry snow aloft, becoming wet snow at the melting layer and falling out as rain below. However, in the unstable unblocked cases rainfall is enhanced by the participation of the low-level flow in the orographic lifting; when the low-level air rises easily over the first high peaks of terrain, raindrops grow rapidly by coalescence at low levels, and graupel forms just above the 0°C level. The coalescence-produced drops and melted graupel particles contribute to heavy rain on the lower Alpine slopes. The orographic enhancement in the stable blocked cases is fundamentally limited since these low-altitude precipitation processes cannot add significantly to the background stratiform precipitation.

KEYWORDS: Mesoscale Alpine Programme Microphysics Orographic precipitation Polarimetric Radar

1. INTRODUCTION

The European Alps constitute a long, tall and narrow mountain range over central Europe, located to the north of the Ligurian Sea (Fig. 1). When a baroclinic trough approaches the region from the west, strong southerly low-level flow advects moisture

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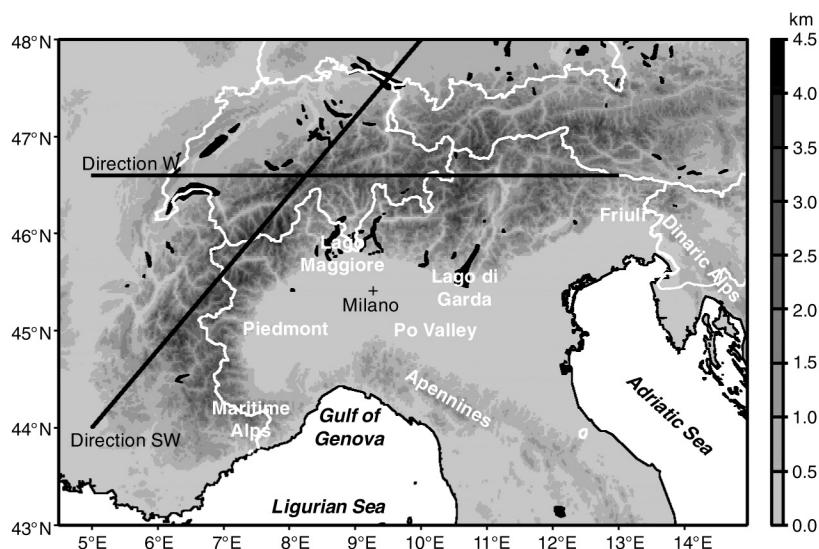


Figure 1. Alpine topography with geographic features. Direction lines are discussed in the text.

toward the Alps. During trough passages in the autumn, heavy precipitation and flooding can occur on the Mediterranean side of the Alps (Massacand *et al.* 1998). The Piedmont flood that occurred in the region of that name (Fig. 1) in November 1994 (Buzzi *et al.* 1998; Ferretti *et al.* 2000; Rotunno and Ferretti 2001) produced rain accumulations of more than 300 mm over a 36-hour period at some stations (Lionetti 1996). The devastating flood that occurred over the same region in October 2000 was a similar event (Gabella and Mantovani 2001).

A major objective of the Mesoscale Alpine Programme (MAP) held over the European Alps from 7 September to 15 November 1999 was to gain a better understanding of orographic precipitation mechanisms by documenting in detail several major rain events in the Alps. One of the main target areas of the experiment was the Lago Maggiore region (Fig. 1) whose precipitation climatology is characterized by an autumn (September, October and November) maximum (Frei and Schaer 1998). Houze *et al.* (2001) used radar data to construct a high-resolution climatology of the spatial characteristics of the precipitation and airflow over this region. Their results suggest that most of the precipitation growth occurs at altitudes below the Alpine crest. They also found that the precipitation over the Lago Maggiore region was greatest when the wind direction at the 2 km* level was southerly or south-easterly, and that the nature of the precipitation was a strong function of the Froude number, Fr , of the flow, which determines whether the flow is blocked (low Fr) or rises easily over the terrain (high Fr). The present paper extends this work by exploring the microphysical processes in low- and high- Fr regimes.

Storms that produced large precipitation amounts over the Alpine region were documented in fourteen MAP Intensive Observing Periods (IOPs). Two particularly instructive IOPs were IOP 2b (a high- Fr case occurring between 1300 UTC 19 September and 0100 UTC 21 September 1999) and IOP 8 (a low- Fr case occurring between 1200 UTC 20 October and 2200 UTC 21 October 1999). Each event occurred ahead of a strong

* All heights are given above mean sea level.

baroclinic trough. The 850 hPa flow in each case had a strong southerly component, in a pattern that has been identified as being a precursor of heavy precipitation (Massacand *et al.* 1998). Despite this basic large-scale similarity upstream of the Alps at 850 hPa and above, the precipitation distribution differed markedly between the two cases, indicating that differences in the flow regimes at the lowest levels (below ~ 900 hPa) were key processes explaining the differences in the character of the precipitation. This paper describes the differences in low-level flow and thermodynamics between these two cases that account for the different *Fr* characterizations of each case.

Several radars were deployed in MAP to help determine the airflow and precipitation structure in detail. These radars document the spatial structure of the airflow and reflectivity fields on scales small enough to relate them to the detailed topography of individual peaks and valleys on the windward slopes of the Alpine terrain. Since one of the radars used in MAP (the National Center for Atmospheric Research (NCAR) S-Pol radar) obtained measurements at both horizontal and vertical polarization, this study is able to determine the predominant types of hydrometeors associated with the fine-scale structure of the airflow and radar reflectivity over specific mountain peaks and valleys. Supplemental upper-air soundings and high-resolution model runs carried out in the project further document the thermodynamics and airflow. Incorporation of the polarimetric radar data into the detailed analysis of air motion and reflectivity allows us to extend the results of Houze *et al.* (2001) by including the microphysical processes in conceptual models of the interactive air flow and microphysics producing heavy orographic precipitation under both low- and high-*Fr* conditions. Most of the analysis presented below is of the two case-studies (IOP 2b and IOP 8) representing high and low *Fr*s. However, we also present the overall statistics of the polarimetric radar data for the whole MAP period to determine the representativeness of the two cases.

2. DATA

This study emphasizes the use of radar to determine the relationships of airflow and precipitation microphysics to the Alpine topography in the Lago Maggiore region (Figs. 1 and 2). MAP employed seven ground-based radars and two airborne radars in a co-ordinated network. Rain-gauge data, sounding data, wind profiler data, and output from numerical forecast models provide the context of the detailed radar observations.

(a) Radar data

The primary data for this study are from three of the ground-based scanning radars deployed in MAP: the NCAR S-Pol radar, the Swiss Monte Lema radar, and the French RONSARD. The location of these radars in relation to the terrain is shown in Fig. 2. Table 1 lists the principal characteristics of the radars. The S-Pol and RONSARD were installed especially for the MAP field phase, while the Monte Lema radar is a permanent facility of MeteoSwiss. Since the Monte Lema radar antenna is located at an altitude of 1.63 km, the lowest scan is generally above 1.5 km. Hence the 2 km level is about the lowest available Cartesian grid level that can be obtained using bilinear interpolation. The S-Pol radar was located at an altitude of 0.28 km and the RONSARD at 0.155 km. Data from these two radars supplemented the Monte Lema radar by providing information at lower altitudes and by scanning farther south over the Po Valley (Fig. 2). All these radars measured reflectivity and radial velocity. The S-Pol is a dual-polarization radar, which provides information on particle type (Vivekanandan *et al.* 1999; Zeng *et al.* 2001).

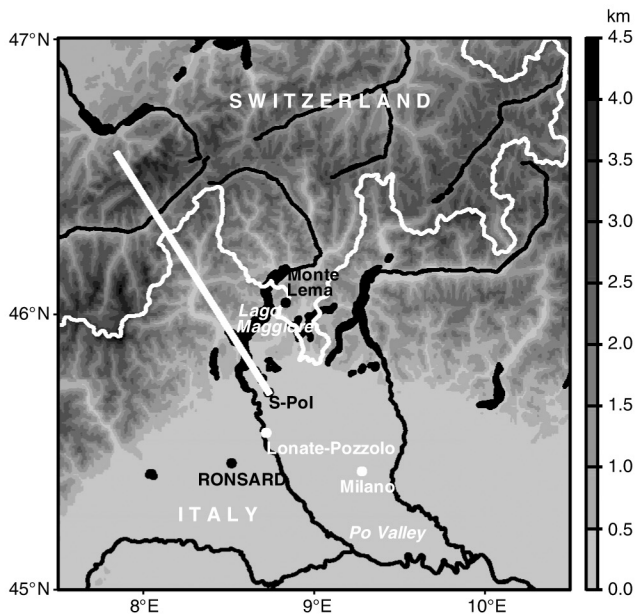


Figure 2. The Lago Maggiore region, with bodies of water (black), geographic features, radar locations and other observational sites. The straight white line shows the location of the cross-sections in Figs. 12 and 14.

TABLE 1. MAP RADAR CHARACTERISTICS

Radar	Monte Lema	RONSARD	S-Pol
Wavelength (cm)	5.6	5.4	10
Peak power (kW)	251	250	1000
Beam width (deg)	1.0	0.89	0.91
Unambiguous velocity interval (m s ⁻¹)	8.27–16.54 ¹	9.8 or 19.6 ²	22.4

¹Changes with increasing elevation angle.
²Depending on the chosen pulse repetition frequency.

Three-dimensional (3D) scans of radar data were recorded every 5 minutes for Monte Lema, every 15 minutes for RONSARD and independently for S-Pol. The Monte Lema and RONSARD radars scanned a full 360 degree cylindrical volume centred on the radar. S-Pol performed azimuthal sector scans in the region of heaviest orographic precipitation. The S-Pol scans alternated Plan Position Indicator (PPI) sectors with Range Height Indicator (RHI) sectors. The PPI volumes have been bi-linearly interpolated to a Cartesian grid with a resolution of 2 km × 2 km in the horizontal and 0.5 km in the vertical using NCAR’s SPRINT software (Mohr and Vaughan 1979). The interpolation allows the data to be displayed in ‘Mountain Zebra’ (James *et al.* 2000), which is a version of the NCAR Zebra display software (Corbet *et al.* 1994). In Mountain Zebra the data fields can be visualized, overlaid and interpreted relative to details of the topography. Following Houze *et al.* (2001), mean fields of reflectivity and radial velocity were constructed for individual precipitation events or IOPs. For each IOP, the averaged reflectivity and flow velocity fields were computed from all the available volumes within the time window in which the events occurred.

The S-Pol RHI sectors provided sufficient resolution to determine microphysical fields from measured polarimetric radar parameters. To calculate statistics on the microphysical particle data, mean patterns of frequency of occurrence of hydrometeor types were constructed by keeping track of the number of times that a particular hydrometeor type appeared in a $2 \text{ km} \times 2 \text{ km} \times 0.5 \text{ km}$ Cartesian grid volume during a certain time interval, and dividing by the total number of volumes considered. The data were further normalized to account for the fact that the number of beams in each Cartesian grid volume close to the radar is larger than further out.

(b) Rain-gauge, sounding and wind profiler data

To determine the larger-scale context of the precipitation detected by the three radars in the Lago Maggiore region, this study uses rain-gauge data from northern Italy and southern Switzerland to produce larger-scale maps of the rain accumulations in each IOP examined. The gauges were sparsely distributed and subject to local and other sampling problems. Nevertheless, they provide a general idea of the overall rain pattern in each storm. The precipitation maps presented in this study show accumulations during each precipitation event.

To determine the characteristics of the flow upstream of the radar-observed precipitation, this study uses data from the upper-air sounding at Milano (Fig. 2). During MAP, the observational frequency at this operational site was four per day (0000, 0600, 1200 and 1800 UTC). Storm-mean soundings were constructed by averaging all the available Milano soundings during each storm. A mean IOP 2b sounding was constructed by averaging all the available Milano soundings between 1300 UTC 19 September and 0100 UTC 21 September 1999, while the mean IOP 8 Milano sounding used data between 1200 UTC 20 October and 2200 UTC 21 October 1999. The flow inside the Po Valley was sampled at high-resolution in both time and height by observations from a French UHF/VHF (ultra high frequency/very high frequency) wind profiler located at Lonate-Pozzolo (see www.map.ethz.ch). These data were averaged over the event intervals given above to obtain storm-mean wind profiles for IOP 2b and IOP 8.

(c) Model output

During MAP, the atmospheric Mesoscale Compressible Community numerical forecast model (MC2, Benoit *et al.* 1997; Benoit *et al.* 2002) was run in real time to aid scientists in scheduling aircraft missions, radar operations and other special measurement activities. We obtained hourly output and constructed storm-mean fields by averaging these hourly fields. In addition, output from the European Centre for Medium-Range Weather Forecasts (ECMWF) model is presented to illustrate the larger-scale synoptic characteristics prevailing during heavy rain events.

3. AIRFLOW AND THERMODYNAMICS IN TWO MAJOR ALPINE RAIN EVENTS

This section compares two major MAP rain events (IOP 2b and IOP 8). We show that while at levels above $\sim 900 \text{ hPa}$ the events had common characteristics, the low-level flow and thermodynamics differed markedly, evidently resulting in strikingly different precipitation amounts and distributions.

(a) Synoptic conditions

During IOP 2b a trough moved over northern Italy producing heavy precipitation in several regions. The ECMWF model forecast for 1200 UTC 20 September 1999

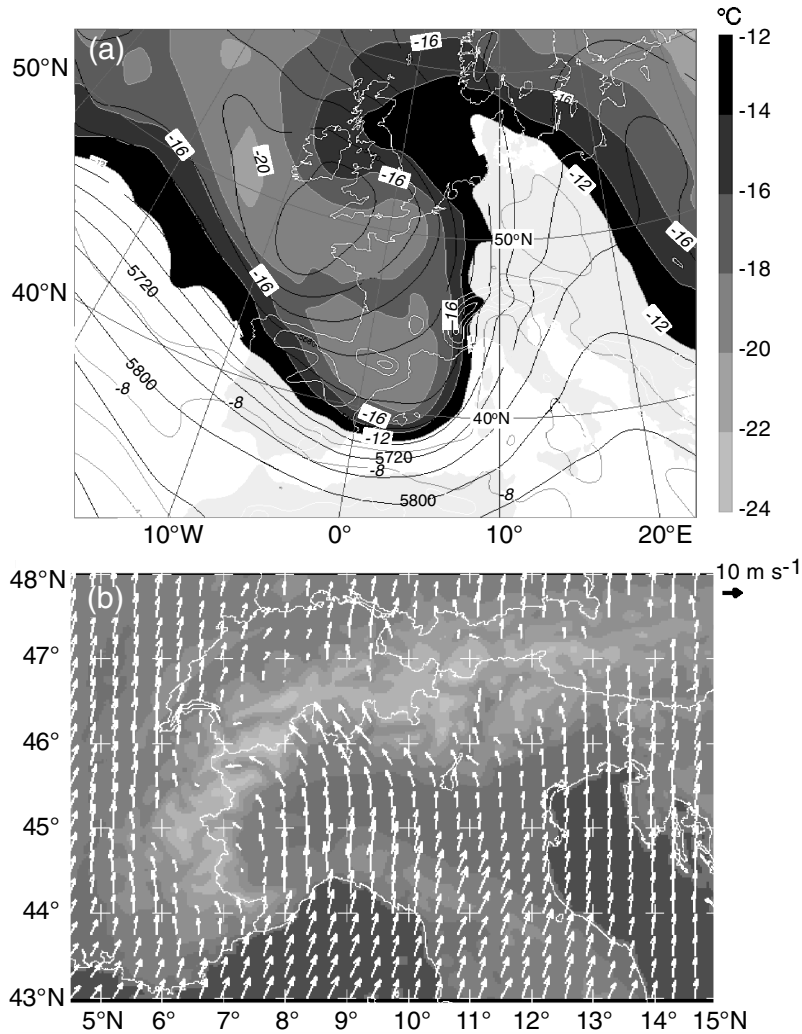


Figure 3. (a) European Centre for Medium-Range Weather Forecasts 500 hPa geopotential height (black lines) and air temperature (shading) forecasts over Europe for 1200 UTC 20 September 1999. (b) The atmospheric Mesoscale Compressible Community numerical forecast model (MC2) topography and storm-mean 840 hPa wind field around the Alps during IOP 2b (1300 UTC 19 September to 0100 UTC 21 September 1999).

indicated a deep trough extending south to northern Africa, with southerly flow reaching the Lago Maggiore region and crossing the Alps (Fig. 3(a)). To investigate the modelled flow on the scale of the whole Alpine barrier and to examine its 3D structure we present MC2 mean wind fields for the whole storm. The mean field of the 840 hPa wind field according to the MC2 (Fig. 3(b)) showed south-south-westerly flow over the Ligurian Sea, turning southerly over the Po Valley and south-south-easterly over the Lago Maggiore region. During IOP 8, another major trough located over western Europe produced intense precipitation in northern Italy. The ECMWF model forecast a 500 hPa geopotential-height field with a low pressure system located over the Mediterranean Sea off the south-east coast of France. At 1200 UTC 21 October 1999 the flow was again southerly towards the Alps over the Lago Maggiore region (Fig. 4(a)). The MC2 mean

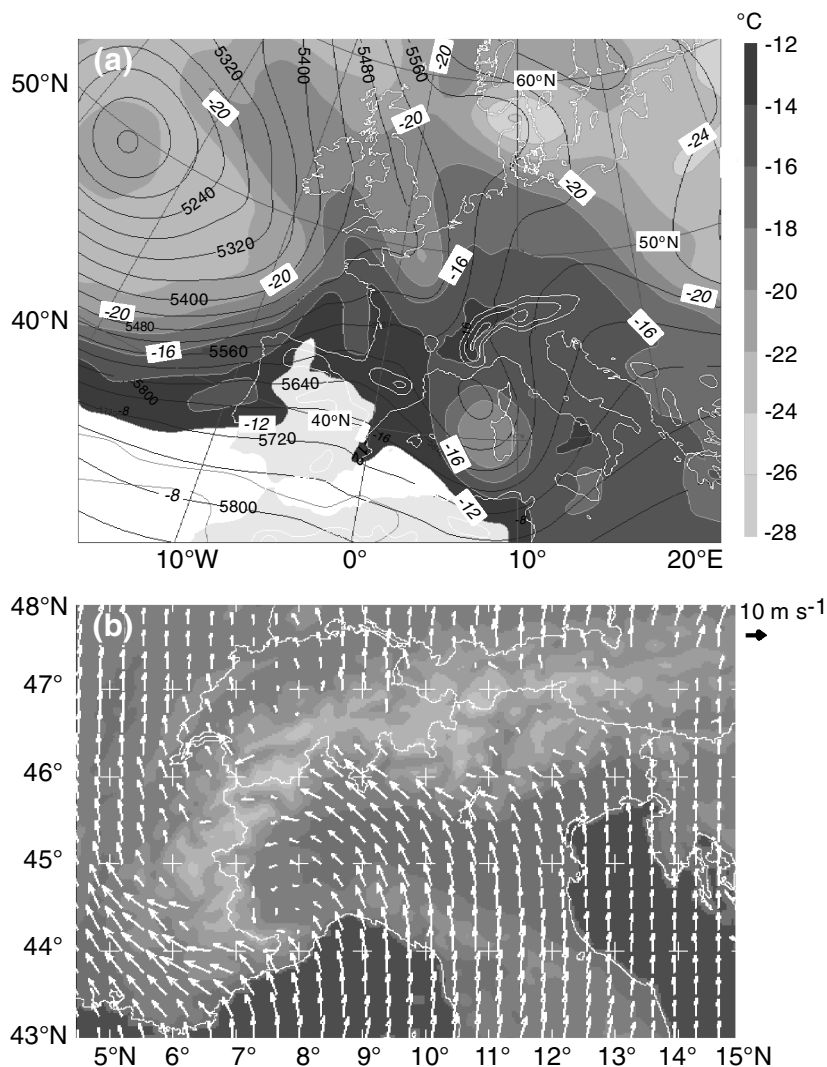


Figure 4. As Fig. 3 but (a) for 1200 UTC 21 October 1999, and (b) during IOP 8 (1200 UTC 20 October to 2200 UTC 21 October 1999).

840 hPa wind field depicted southerlies well upstream of the Alps and south-easterlies over Lago Maggiore (Fig. 4(b)).

A deep upper-level trough with southerly moist flow over the Alps has been identified as one of the important aspects of heavy floods over northern Italy (e.g. Massacand *et al.* 1998). Therefore, these two events were expected to produce intense rainfall over the southern flank of the Alps.

(b) Low-level winds

During IOP 2b, the MC2 forecast southerly flow at 940 hPa upstream of the Alps (Fig. 5(a)). The southerly flow advected moisture from the Ligurian Sea and turned to the west as it got closer to, and then over the mountains in the vicinity of Lago Maggiore. The surface winds (not shown) were similar to the 940 hPa mean wind pattern, and

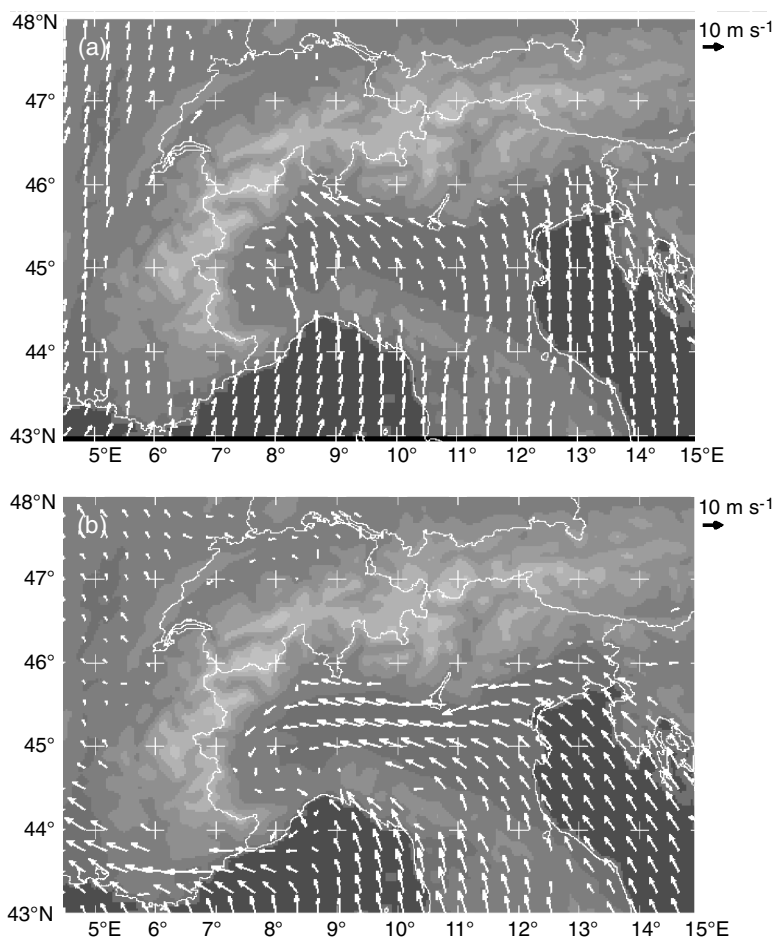


Figure 5. The atmospheric Mesoscale Compressible Community numerical forecast model (MC2) topography and storm-mean wind field around the Alps at 940 hPa during: (a) IOP 2b (1300 UTC 19 September to 0100 UTC 21 September 1999), and (b) IOP 8 (1200 UTC 20 October to 2200 UTC 21 October 1999).

clearly indicated that the south-easterly flow over the Adriatic Sea turned east when it entered the Po Valley and converged with a southerly airstream from the Ligurian Sea. This pattern is reminiscent of the surface winds during the Piedmont storm (Buzzi *et al.* 1998). During IOP 8 the MC2 940 hPa mean winds had a strong easterly component over the Po Valley (Fig. 5(b)). This flow decelerated south of the Lago Maggiore region as it turned toward the gap between the Maritime Alps and the Apennines (Fig. 5(b)). The moist southerly airflow from the Ligurian Sea never reached Lago Maggiore in this case. Figure 5 suggests that the Apennines could have played a role during this storm by preventing the southerly flow from reaching the Po Valley and favouring easterly winds over the Po Valley. It also suggests that the vertical extent of the easterly flow may be related to the height of the Apennines (B. Smull, 2000, personal communication). The surface winds (not shown) show essentially the same features as the 940 hPa wind field, except that the blocking effect and the escaping of the flow out of the Po Valley through the gap between the Maritime Alps and the Apennines is better defined. These characteristics of IOP 8 are very similar to the experiment conducted with no latent heating by Buzzi *et al.* (1998), in the sense that the easterly flow over the Po

Valley turned northerly toward the gap between the Maritime Alps and the Apennines. Rotunno and Ferretti (2002) have found, using numerical simulations, that the easterly flow during IOP 8 was deeper than during IOP 2b.

The MC2 mean wind patterns near the surface suggest that while the flow north-west of the Lago Maggiore was able to rise over the mountains during IOP 2b, the flow during IOP 8 was turned away from this part of the mountain barrier. IOP 2b resembled the Piedmont storm in the sense that during both storms the low-level flow rose over the terrain in the region where the precipitation was maximum. IOP 8 also had some common aspects with the Piedmont flood, namely a clearly defined easterly flow over the Po Valley and southerly flow above. However, during the Piedmont storm the easterly flow did not turn toward the south-west but rose over the terrain of the region, where the precipitation fallout was maximum (Buzzi *et al.* 1998; Ferretti *et al.* 2000).

Smull *et al.* (2001) evaluated the performance of the MC2 during IOP 8. They found that the extent and depth of the low-level easterly and northerly return flows over the Po Valley predicted by the model were underestimated. Additionally, from a comparison between radial velocity patterns predicted by the MC2 at the location of the Monte Lema and RONSARD radar sites (not shown) and the radial velocity observations, it is clear that the simulated flow below 2 km did not compare well with the observations during IOP 8. However, above this level the MC2 correctly depicted the flow direction. Therefore, in this study we use the MC2 output as a proxy of the flow on the whole Alpine scale; to carry out detailed analysis in the Lago Maggiore region we use observational data.

(c) *Accumulated precipitation*

The impact that the different flow regimes had on the accumulated precipitation amounts during the storms is evident from the patterns of rain-gauge data. Figure 6(a) shows the rain accumulated at gauges during IOP 2b. Accumulations of over 100 mm of rain were measured at sites located on the western slopes of the Lago Maggiore region, with one site recording 265 mm. To the south-west of Lago Maggiore, there was also enhancement of the rainfall at the foot of the mountains (Fig. 6(a)). High accumulations were also seen on the Alpine slopes to the west of the Friuli region.

In IOP 8, accumulations in excess of 100 mm occurred at a few locations around Lago di Garda (Fig. 6(b)). However, the amounts of precipitation accumulated over the Lago Maggiore region were below 100 mm, no greater over the lower slopes than over the Po valley.

The precipitation patterns of the two events were thus strikingly different, and consistent with the hypothesis that the flow immediately upstream of the western slopes of the Lago Maggiore region corresponded to different regimes: unblocked flow up and over the mountains during IOP 2b and flow around or blocked during IOP 8. These hypotheses are explored in the rest of this section.

(d) *Temperature and stability upstream of the Alps*

An ideal 2D obstacle will block an impinging flow whenever the Froude number of the upstream flow is <1 (Durrán 1990; Houze 1993). The Froude number is given by $Fr = U/(NH)$, where U is the upstream flow speed perpendicular to the terrain, N is the Brunt–Väisälä frequency and H is the height of the mountain barrier. The Alps are not an ideal 2D barrier. Nevertheless, we may gain insight into the dynamics of the flow over the Alpine barrier by examining the individual components that contribute to the Fr (i.e. the stability and the speed of the flow perpendicular to the terrain.)

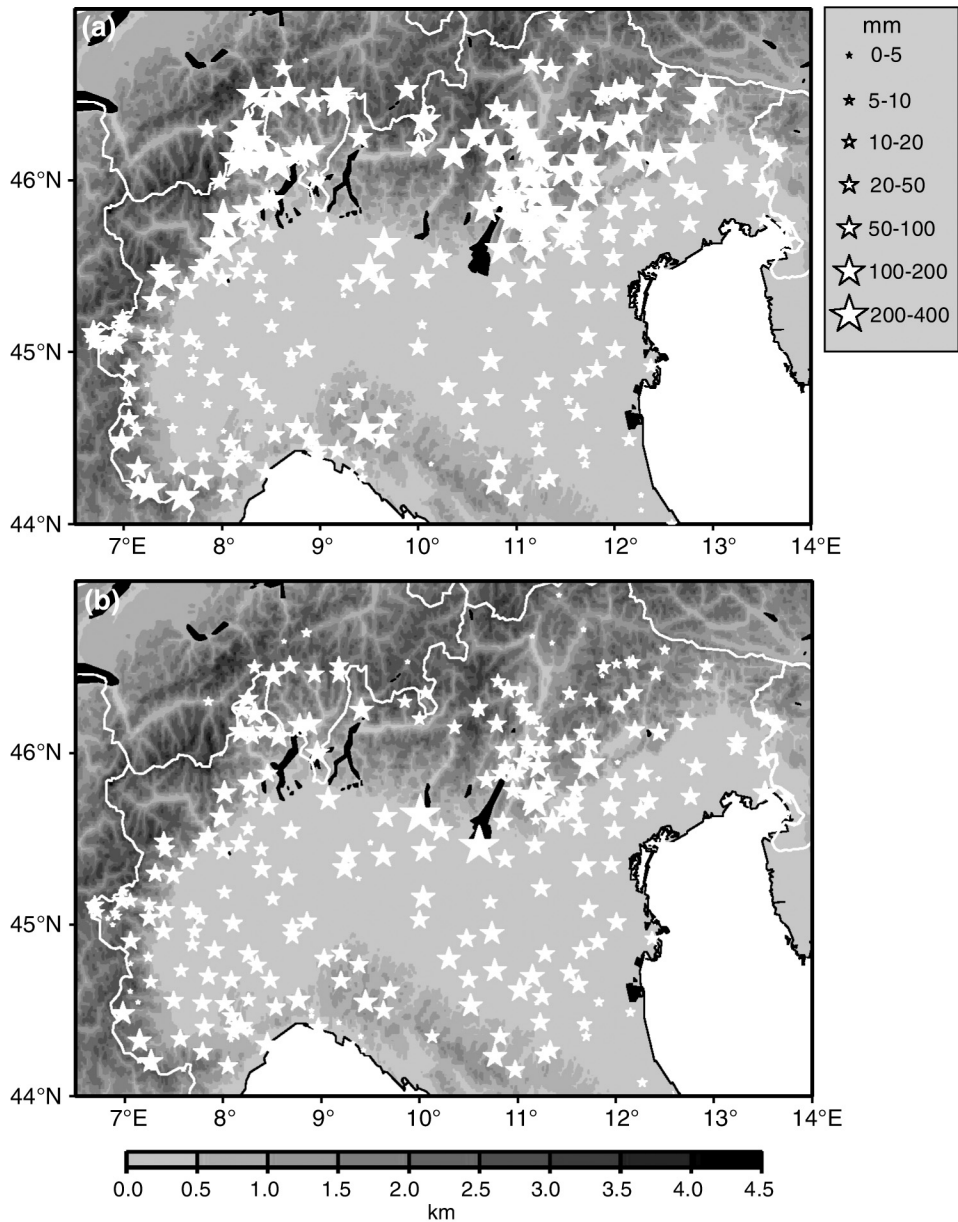


Figure 6. Accumulated precipitation over northern Italy and southern Switzerland during: (a) IOP 2b (1300 UTC 19 September to 0100 UTC 21 September 1999), and (b) IOP 8 (1200 UTC 20 October to 2200 UTC 21 October 1999).

Low-level mean upstream soundings from Milano were constructed for each event (Fig. 7). During IOP 2b, the atmosphere was unsaturated and conditionally unstable above 900 hPa and below 950 hPa, with the 900–950 hPa layer behaving as a weak capping inversion. There was also a negative vertical gradient of equivalent potential temperature from 710 to 950 hPa, indicating that this layer was potentially unstable (Fig. 8(a)). The surface temperature and mixing ratio were $\sim 19^{\circ}\text{C}$ and $\sim 12\text{ g kg}^{-1}$,

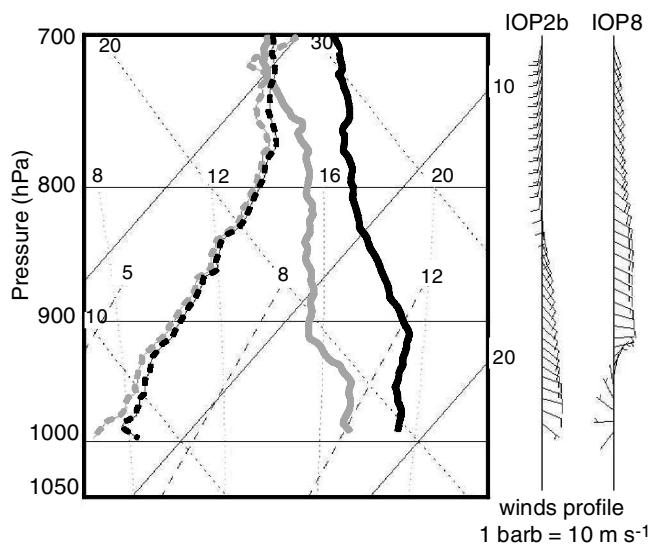


Figure 7. Temperature (black), dew-point (grey) and wind vectors for mean-storm soundings from Milano for IOP 2b (heavy solid, 1300 UTC 19 September to 0100 UTC 21 September 1999) and IOP 8 (heavy dashed, 1200 UTC 20 October to 2200 UTC 21 October 1999) plotted on a skew T -log p diagram. Temperature, wet-bulb potential temperature and dry adiabats are in $^{\circ}\text{C}$, saturation mixing ratio is in g kg^{-1} .

respectively (Fig. 7). The environment during IOP 8 was very close to saturation (Fig. 7) and stable everywhere except in a narrow layer below 975 hPa (Fig. 8(a)) with surface temperature $\sim 8^{\circ}\text{C}$ and surface mixing ratio $\sim 6 \text{ g kg}^{-1}$ (Fig. 7). To indicate stability quantitatively, we calculated the dry (N_d) and moist (N_m) Brunt-Väisälä frequencies by the method of Durran and Klemp (1982). Figure 8(b) compares the stabilities calculated from the Milano mean soundings. IOP 8 was absolutely stable, with both N_d^2 and $N_m^2 > 0$. IOP 2b was slightly unstable in terms of N_m^2 , which was slightly negative. This slight instability was apparently extremely important and it had two effects. Firstly, the higher Fr implied by the low static stability indicates that the air easily ascended the mountain slopes, and that orographic lifting and condensation were easily achieved. Secondly, the instability indicated by the negative value of N_m^2 accounted for the occurrence of embedded convective cells over local peaks and ridges in the terrain, which further enhanced the rainfall in IOP 2b (section 4(c)(i)). This marginally unstable condition is one factor that made IOP 2b similar to the Piedmont flood case (Buzzi *et al.* 1998; Doswell *et al.* 1998); in addition, the sounding information collected at Milano during the Piedmont flood (Frontero *et al.* 1996) indicates surface values of temperature and mixing ratio very similar to those observed during IOP 2b.

In contrast, the stable conditions in IOP 8 made it: (i) difficult for the low-level flow to ascend over the barrier; and (ii) impossible for embedded cellular convection to occur (section 4(c)(ii)). These factors combined with cooler conditions (implying lower saturation mixing ratios) in IOP 8 evidently led to lower precipitation amounts in the Lago Maggiore region.

(e) Wind structure upstream of the Alps

To determine the potential for blocking associated with the unimpeded rise of the upstream flow over the terrain, the magnitude of the wind in the direction perpendicular to the barrier (U) must be taken into account in addition to the stability. Over the Lago

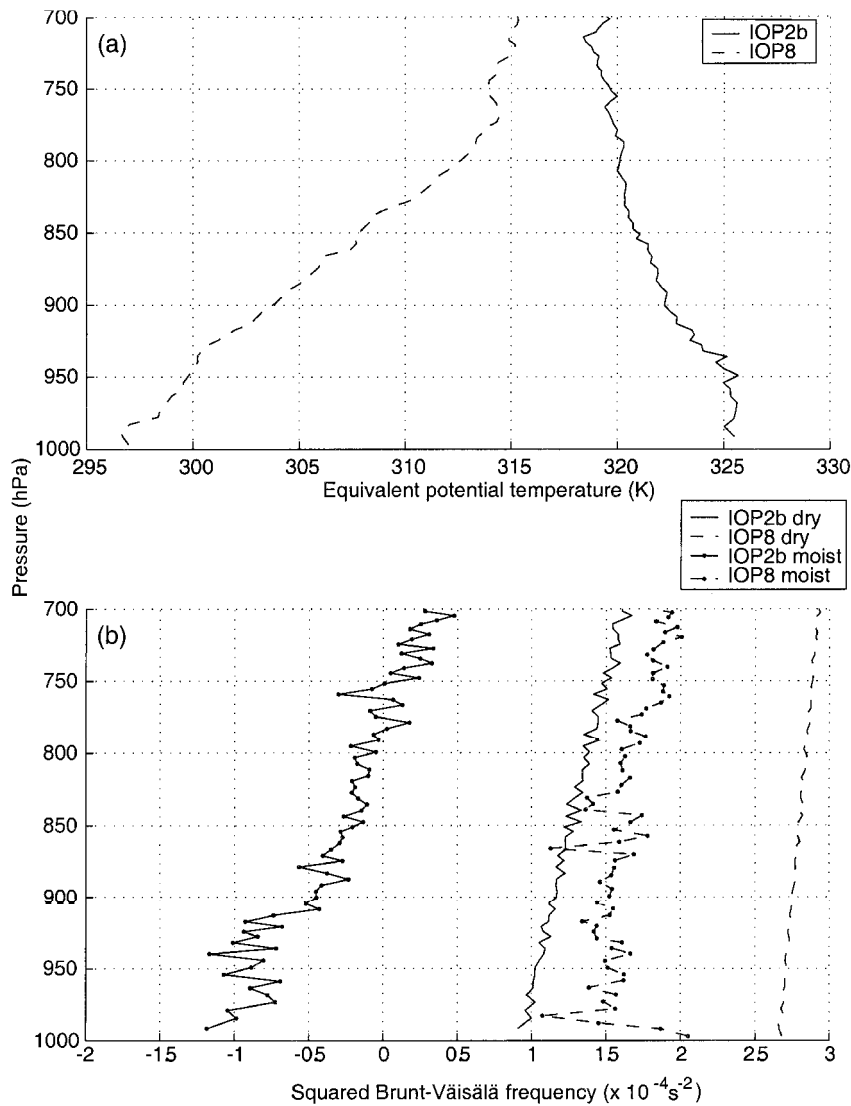


Figure 8. Mean-storm: (a) equivalent potential temperature, and (b) dry and moist squared Brunt–Väisälä frequency for IOP 2b (solid lines, 1300 UTC 19 September to 0100 UTC 21 September 1999) and IOP 8 (dashed lines, 1200 UTC 20 October to 2200 UTC 21 October 1999) calculated from Milano soundings.

Maggiore region it would be difficult to choose one orientation as representative of the whole barrier. We approximate it by two distinct directions. To identify them easily, we refer to the zonal direction as ‘Direction W’ and to the slanted one as ‘Direction SW’ (Fig. 1). According to wind profiles from the Milano sounding, the wind components perpendicular to both orientations of the barrier were much stronger at low levels (below 900 hPa) in IOP 2b (Figs. 9(a) and (b)). Only above 900 hPa and only perpendicular to Direction SW did IOP 8 have a significantly stronger cross-barrier flow than IOP 2b (Fig. 9(b)). Note also that IOP 8 had components away from the barrier below 925 hPa, that is to say, whilst the moist boundary-layer flow during IOP 2b rose over the terrain, during IOP 8 it evidently did not. Figures 9(c) and (d) also show storm-mean wind

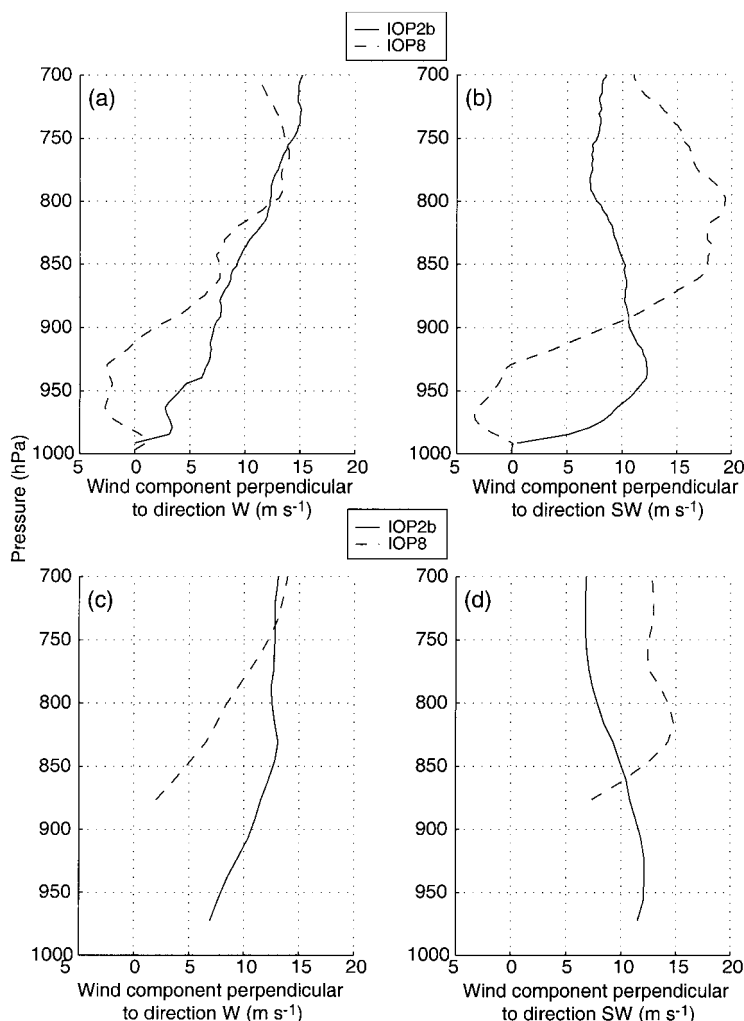


Figure 9. Storm-mean components of the wind perpendicular to: (a) Direction W from the Milano sounding; (b) Direction SW from the Milano sounding; (c) Direction W from the Lonate-Pozzolo wind profiler; and (d) Direction SW from the Lonate-Pozzolo wind profiler. Directions are as indicated in Fig. 1. Solid (dashed) lines denote IOP 2b (IOP 8) storm-mean from 1300 UTC 19 September to 0100 UTC 21 September 1999 (1200 UTC 20 October to 2200 UTC 21 October 1999). See text for further details.

data from the wind profiler located at Lonate-Pozzolo (46 km to the west-north-west of Milano). The wind components perpendicular to Directions W and SW according to the Lonate wind profiler showed a slightly different structure than the components measured at Milano (compare Figs. 9(a) and (b) with Figs. 9(c) and (d)). This was probably due in part to the different location of the observations, and partly due to the different sampling intervals (the Milano soundings were taken every 6 hours, while the wind profiler took data every 15 or 30 minutes). However, the relative structure of IOP 2b in relation to IOP 8 was consistent with the Milano observations.

According to the definition of the *Fr*, the weak flow (sometimes negative) normal to the barrier at low levels in the Lago Maggiore region during IOP 8, together

with its stronger stability, promoted a low-level blocked flow over the terrain north-west of the Lago Maggiore. Conversely, IOP 2b was more likely than IOP 8 to have produced upslope enhancement of precipitation by easily lifting the lowest layer of the atmosphere, where the moisture content was greatest. Thus, both the low-level wind and the low-level stability contributed to the easier lifting of the lower layer of air in IOP2b and the blocking of the lower layer immediately upstream of the Alps in IOP 8 (sections 4(a) and (c)).

(f) Trajectories

Rotunno and Ferretti's (2003) Fig. 8 shows backward and forward trajectories of air parcels that reached the region of maximum simulated rain (approx. 45.9°N, 8°E) at 1200 UTC 20 September 1999 (for IOP 2b) and 1200 UTC 21 October 1999 (for IOP 8). Here we refer to their figure. Parcels arriving in the Lago Maggiore region at altitudes of 1.5, 2.5 and 3.0 km at 1200 UTC in IOP 2b originated to the south of the Alps, flowed towards the terrain and rose over it, with the lowest trajectories showing a stronger westward deflection. This result is consistent with the flow being unblocked and the entire moist layer rising over the terrain (section 4(a)). During IOP 8 parcels arriving in the region at altitudes of 1.5, 2.5 and 3.0 km at 1200 UTC 21 October 1999 started on the Mediterranean side of the Alps, with the lowest trajectory being blocked by the terrain. The trajectories that ascended over the Alps followed a track over the Apennines upstream of the Lago Maggiore. It is likely that the air released its instability during this passage over the Apennines and was stable by the time it reached the Lago Maggiore region. During this event, convective precipitation was observed by aircraft radar over the Gulf of Genoa (Bousquet and Smull 2001) whereas the ground-based and airborne radar data taken over the Lago Maggiore region showed a completely stable and stratiform structure (see next section).

4. PRECIPITATION PROCESSES RELATED TO FINE-SCALE TOPOGRAPHY

In this section we present a description of the rain events of IOPs 2b and 8 as documented by the MAP radar network located in the Lago Maggiore region. The reflectivity and radial velocity patterns document the substantial differences in airflow and gross precipitation structure that existed between the two events. The hydrometeor types indicated by polarimetric radar observations obtained during each event allow us to draw further conclusions about the precipitation growth mechanisms operative in each case.

(a) Wind patterns over Lago Maggiore

The mean patterns of radial velocity detected by radar clearly confirm that different flow conditions prevailed during each event. Figure 10 shows the radial velocity at different levels along with the 800 m terrain contour. In Figs. 10(a) and (c) data from the Monte Lema radar show the radial velocity pattern at 2 km; in Figs. 10(b) and (d) data from the RONSARD radar show the radial velocity at 0.5 km.

During IOP 2b, the Monte Lema 2 km flow was south-south-easterly towards the convex indentation of the terrain (Fig. 10(a)). At 0.5 km, the RONSARD radar showed south-easterly flow toward the barrier, perpendicular to the local orography (Fig. 10(b)). Figures 10(a) and (b) indicate that during this event the entire lower layer moved as a whole unit and rose over the terrain, thus producing large local precipitation amounts in the Lago Maggiore region (Fig. 6(a)).

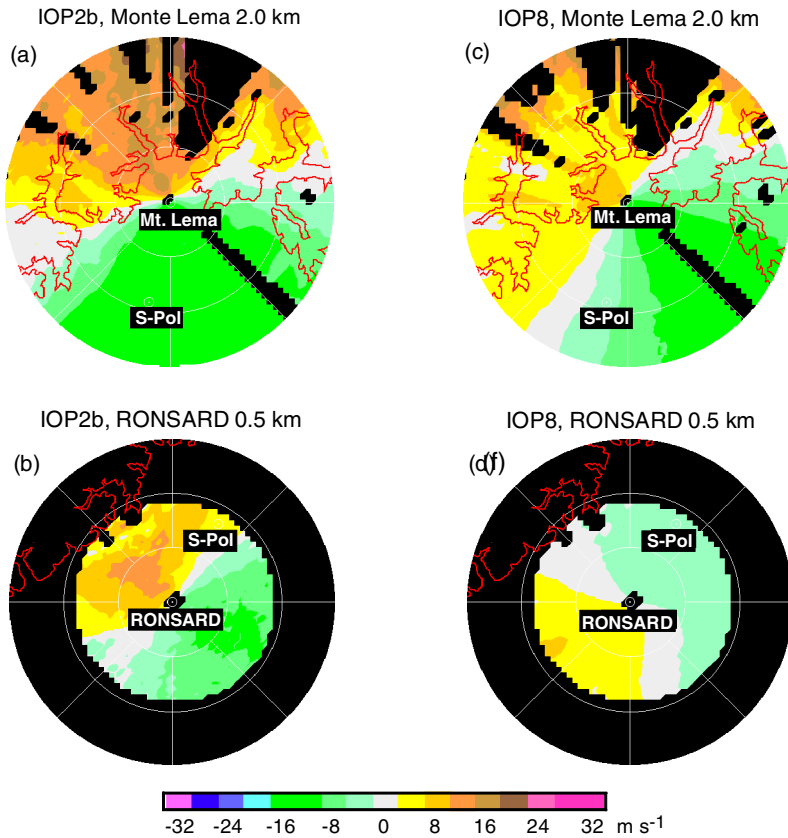


Figure 10. Constant-altitude plots showing the storm-mean radial velocity observed by: (a) Monte Lema radar at 2 km in IOP 2b (1300 UTC 19 September to 0100 UTC 21 September 1999); (b) RONSARD radar at 0.5 km in IOP 2b; (c) as (a) but in IOP 8 (1200 UTC 20 October to 2200 UTC 21 October 1999); (d) as (b) but in IOP 8. The 800 m terrain contour is shown in red. Range ring spacing is 20 km. Positive (negative) radial velocities denote outbound (inbound) flow. Heights are a.m.s.l.

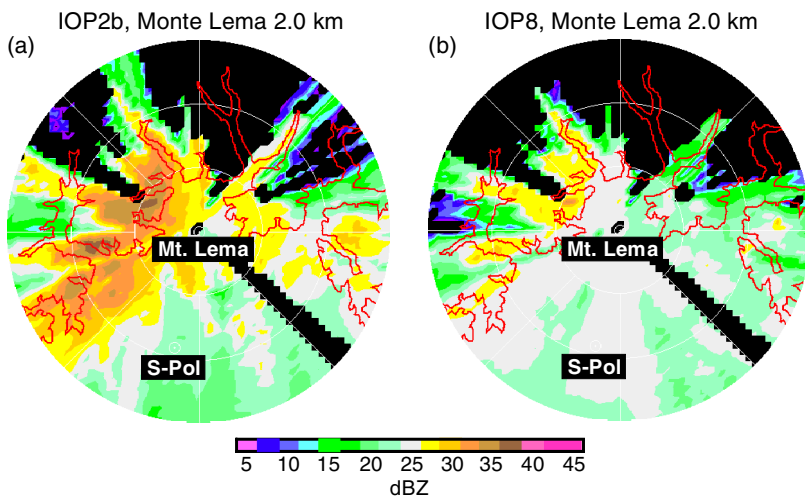


Figure 11. Constant-altitude plots at 2 km containing the storm-mean reflectivity observed by Monte Lema radar for (a) IOP 2b (1300 UTC 19 September to 0100 UTC 21 September 1999) and (b) IOP 8 (1200 UTC 20 October to 2200 UTC 21 October 1999). The 800 m terrain contour is shown in red. Range ring spacing is 20 km. Heights are a.m.s.l.

During IOP 8, the flow at 2 km had a south-easterly direction (Fig. 10(c)), in contrast to the south-south-easterly flow in IOP 2b (Fig. 10(a)). At 0.5 km, the wind direction during IOP 8 varied across the region covered by the RONSARD radar. To the south of the radar, the flow was from the east, while north of the radar it was from the north-east, i.e. parallel to the local terrain (Fig. 10(d)). This case shows a clear difference between the apparently blocked flow at the 0.5 km level, where air was flowing parallel to the mountains in the region immediately upstream of the western slopes of the Lago Maggiore region, and the flow at 1.0 km (not shown) and above which was flowing toward the barrier.

These results suggest that the type of flow regime (blocked or unblocked) was most apparent at the very lowest levels (0.5 km and below). The 4 km radial velocity fields for IOPs 2b and 8 (not shown) were quite similar, further suggesting that the blocked flow regime was constrained to the very lowest levels. In the next section, we explore how the flow regime and the local topography impacted the precipitation distribution and accumulation in each case.

(b) *Reflectivity patterns over the Lago Maggiore region*

The mean reflectivity field for the Monte Lema radar at the 2 km level for IOP 2b (Fig. 11(a)) had maximum values over the lower slopes, the strongest directly to the north-west (i.e. downstream) of the radar. This pattern is mirrored in the frequency of occurrence of precipitation at the same altitude (not shown), which demonstrates that the pattern seen in Fig. 11(a) was representative of the whole storm. The 0.5 km flow (shown in Fig. 10(b) along with the 800 m terrain contour) was perpendicular to the terrain located to the west of the radars. Hence the flow at the very lowest levels impinged on the slopes to the north-west of Lago Maggiore and produced large amounts of precipitation there (Fig. 11(a)).

In IOP 8 (Fig. 11(b)), the echoes were in general weaker* than in IOP 2b and much more uniform (except for the ground-clutter echo north-west of the radar). The enhancement of precipitation over the windward slopes in this case was apparently the result of forced ascent of the layer above about 1.0 km over the highly stable easterly and north-easterly flowing air near the surface (section 4(a)). The less extreme precipitation amounts in IOP 8, compared to IOP 2b, were partly the result of the lower layer of air not participating in the orographic uplift and partly the result of the lifted air being cooler and more stable in IOP 8. In the next section we describe some of the microphysical mechanisms that further account for the precipitation differences between the two storms.

(c) *Precipitation mechanisms inferred from polarimetric radar observations*

Data in MAP were collected by the NCAR S-Pol linearly polarimetric radar, which transmits and receives both horizontally and vertically polarized radiation. The polarimetric observables depend on (and hence also give information about) the physical characteristics of the hydrometeors, such as particle size, shape, thermodynamic phase and spatial orientation. Algorithms developed by Vivekanandan *et al.* (1999) and Zeng *et al.* (2001) were applied to the polarimetric measurements to identify the types of hydrometeors detected by the radar echoes. These algorithms are based primarily on four polarimetric radar variables: reflectivity (*dBZ*), differential reflectivity (*ZDR*), linear depolarization ratio (*LDR*), and specific differential propagation phase (*KDP*). By

* The spot of abnormally high reflectivity ~ 19 km north-west of the radar was produced by anomalous propagation due to the strong stability of the atmosphere during IOP8.

considering all the variables at once, one may infer certain characteristics of the precipitation particles producing the radar echoes. The algorithms used to classify particle type either set thresholds or hard boundaries (Zeng *et al.* 2001), or use a fuzzy-logic approach (Vivekanandan *et al.* 1999). This paper presents the results from both: we refer to the first as the 'UW' algorithm and to the second as the 'NCAR' algorithm.

(i) *IOP 2b microphysics.* Figure 12 shows a storm-mean cross-section along the white line in Fig. 2 during IOP 2b. This cross-section approximately parallels the flow at the 2 km level. The reflectivity field (Fig. 12(a)) shows a convective-like echo structure over the first peak of the mountain range, where the horizontal gradient of elevation first becomes very large. The low altitude of the maximum of reflectivity in the primary echo core tied to the first major mountain peak suggests that coalescence may have been important in the orographic enhancement of the convective precipitation forming over the peak of the terrain. Caracena *et al.* (1979) put forward this idea to explain the low-altitude echo maximum in the orographic convective storm that produced the Big Thompson flood in the Rocky Mountains in 1976. This process is evidently important in the seasonal rainfall over the Alps. Houze *et al.* (2001) found that a structure like that seen in Fig. 12(a) dominates the radar climatology of the entire autumn season over the Alps. For IOP 2b, Georgis *et al.* (2003) found that the precipitation maximum on the windward slopes was produced in intermittent pulses in the form of cells moving from the Po Valley to favoured locations in the lower Alps. Coalescence growth of raindrops would be expected to be large at low levels in these cells. Their research indicates that the characteristic time-scale for the advection of these cells from the Po Valley to the Alps was about an hour. They also showed that the maximum condensation occurred over the lower windward slopes of the Alps, ~ 8 km upwind from the precipitation maximum. This relatively small distance implies very efficient precipitation growth mechanisms, which is consistent with growth by coalescence (and possibly riming, see section 4(c)(i)).

However, ice microphysics also appears to be important to the growth and rapid fallout of precipitation on the windward slope in IOP 2b. Yuter and Houze (2001, 2003) obtained results consistent with these conclusions. They used data from a vertically pointing radar and a 1D parcel model with parametrized microphysics to show that under the conditions of IOP 2b riming and hence graupel formation could be expected above the 0°C level, while warm coalescence enhanced rain formation below the 0°C level. The S-Pol radar gives further insight into the role of ice in the orographic precipitation enhancement process. The radial velocity cross-section in Fig. 12(b) shows that during IOP 2b a low-level jet rose abruptly as the airflow encountered the first large peak of the topography. This dynamical mechanism, made possible by the high-*Fr* conditions, efficiently transported low-level moisture to higher levels. Figure 12(c) shows frequency of occurrence of hydrometeor types in this cross-section of the storm according to the NCAR algorithm (similar results were obtained with the UW algorithm). The contours surround the regions of maximum occurrence of three types of ice particles identified by the polarimetric-radar particle-identification algorithm: graupel, wet snow and dry snow. According to Straka *et al.* (2000) dry snow is characterized by randomly oriented ice particles with a low dielectric constant, while wet snow has an outer layer of water with a large complex refraction index. The particle-identification algorithm detects dry snow when all four polarimetric variables (*dBZ*, *ZDR*, *LDR*, and *KDP*) are low; for wet snow they are all high, in particular *ZDR* has a very strong signal (0.5 to 3 dB according to Straka *et al.* 2000). In Fig. 12(c) the red contour indicates that graupel occurred preferentially above the first major mountain peak, directly above

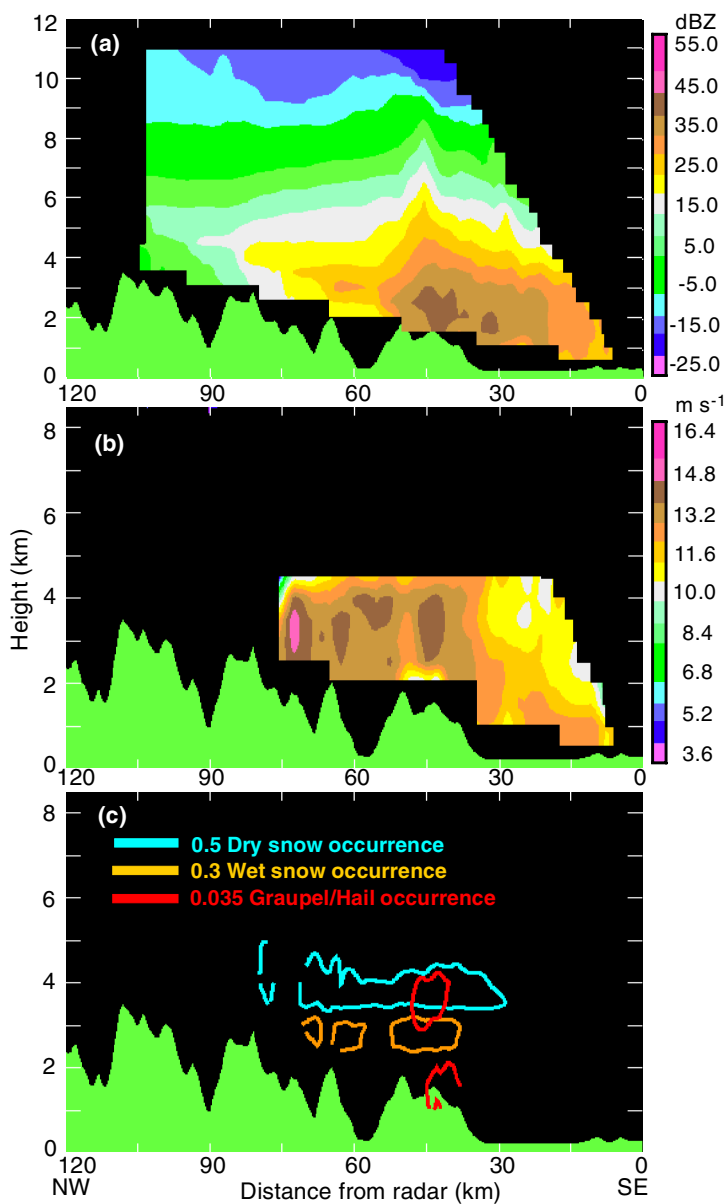


Figure 12. Vertical cross-section along the straight white line in Fig. 2 of the S-Pol radar storm-mean fields during IOP 2b (1300 UTC 19 September to 0100 UTC 21 September 1999). (a) Mean reflectivity, (b) mean radial velocity (positive values denote outbound flow), and (c) frequency of occurrence of particle types identified by polarimetric radar algorithms.

the reflectivity maximum (Fig. 12(a)), and also at the downwind location where the radial velocity jet reached its maximum altitude directly over top of the the first large peak of terrain (Fig. 12(b)). The smaller graupel contour over the first peak at the ground level is an artifact produced by ground clutter and should be ignored. The upper graupel contour indicates a real maximum; it was embedded in a broad layer of persistent dry snow (cyan contour), which was melting and falling into a layer of

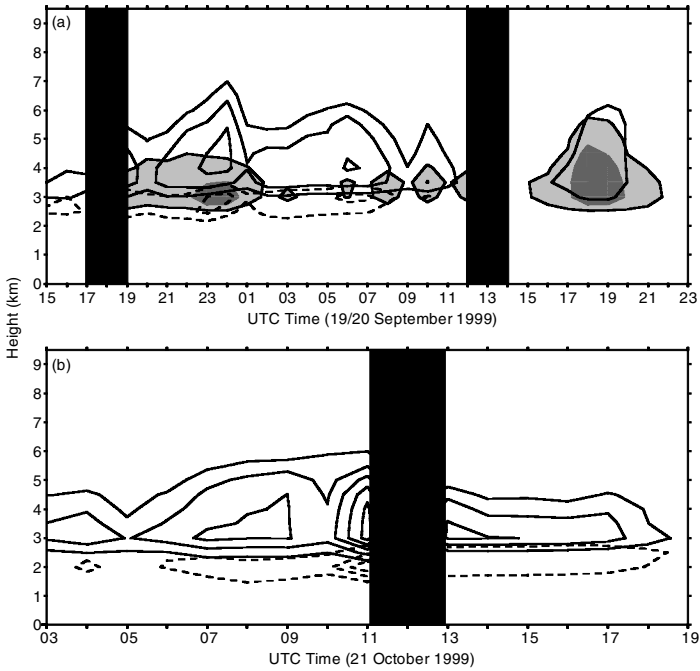


Figure 13. Evolution of mean frequency of occurrence of particle types identified by the NCAR algorithm within 20 degrees elevation and 60 km range from the S-Pol radar for: (a) IOP 2b and (b) IOP 8. Solid (dashed) lines denote frequency of occurrence of dry (wet) snow at 0.1 contour intervals. Dark (light) shading denotes regions where the frequency of occurrence of graupel/hail is greater than 0.005 (0.0006). The black bars denote regions of missing data. See text for further details.

wet snow (orange contour). The region of wet snow identified by the polarimetric radar was better defined below the maximum of graupel, suggesting that the melting of the graupel is increasing the amount of wet snow. The maximum in the occurrence of graupel directly over the precipitation maximum seen in the reflectivity suggests that riming of ice particles just above the 0°C level and their subsequent fallout and melting may have been another major factor, besides coalescence, contributing to the reflectivity maximum at lower levels. The low-level jet transported moisture above the 0°C level, efficiently saturating and condensing cloud liquid water, which was both collected by raindrops below the 0°C level and accreted by ice particles above the 0°C level. The accretion by ice particles of supercooled cloud liquid water, condensed and transported by the jet up the terrain, led to the maximum of graupel occurrence.

The particle-identification results from the S-Pol radar can also be viewed in time series, and for the whole domain scanned by S-Pol (as opposed to the single cross-section discussed above). The frequency of occurrence of hydrometeor types over the S-Pol domain (which consisted roughly of the quadrant north-west of S-Pol) was calculated by averaging all the pixels at elevation angles less than 20 degrees out to a range of 60 km. The performance of the particle-identification algorithms degrades outside these ranges. The hourly evolution of hydrometeors in this region during IOP 2b confirms the stratiform background seen in Fig. 12(c) consisting of dry snow overriding a layer of wet snow (Fig. 13(a)). Graupel appeared embedded between these two layers, steadily at times (e.g. 1900 UTC 19 September to 0200 UTC 20 September 1999, 1600

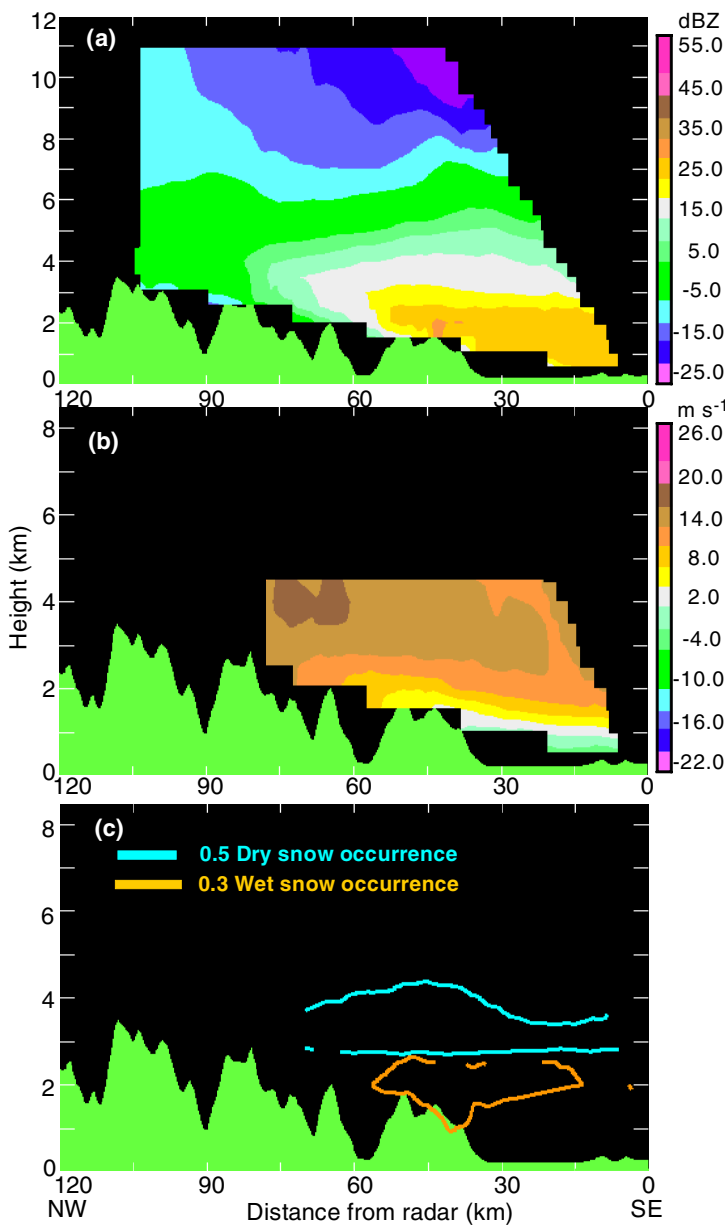


Figure 14. Vertical cross-section along the white line in Fig. 2 of the S-Pol radar storm-mean fields during IOP 8 (1200 UTC 20 October to 2200 UTC 21 October 1999). (a) Mean reflectivity, (b) mean radial velocity (positive (negative) values denote outbound (inbound) flow), and (c) frequency of occurrence of particle types identified by polarimetric radar algorithms.

to 2100 UTC 20 September 1999) and intermittently at other periods (e.g. 0200 to 1200 UTC 20 September 1999), indicating that riming was occurring during IOP 2b.

(ii) *IOP 8 microphysics.* A cross-section along the white line shown in Fig. 2 shows that during IOP 8 the precipitation had a stratiform structure with a distinct bright band at the 2 km level (Fig. 14(a)). This structure contrasts sharply with that seen in IOP 2b.

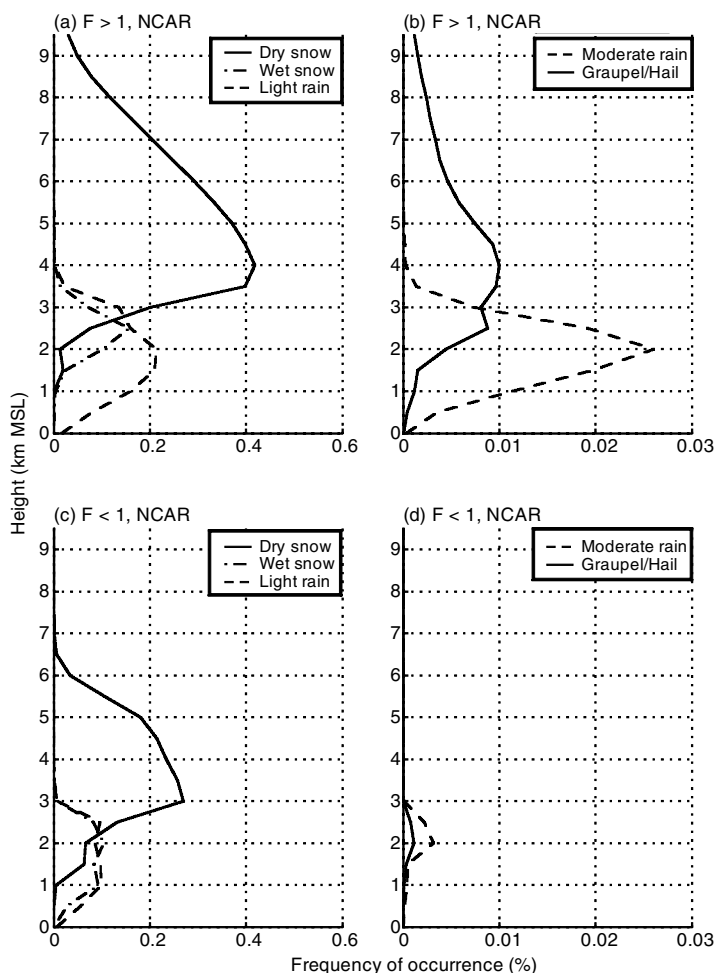


Figure 15. Mean frequency of occurrence of particle types identified by the NCAR algorithm (see text) within 20 degrees elevation and 60 km range from the S-Pol radar, when the layer-averaged 925–700 hPa flow direction from the Milano sounding was between 112.5–202.5 degrees azimuth and the Froude number was (a) and (b) > 1 and (c) and (d) < 1 during MAP.

The radial velocity structure in IOP 8 (Fig. 14(b)) contained an elevated jet around 3 km, which sloped upward more gradually than the jet in IOP 2b. At low levels (below 1 km) the radial component of the flow was toward the radar, indicating blocked flow over the Po Valley along this particular cross-section which crossed over the opening to the Toce Valley. Steiner *et al.* (2003) showed that the Toce Valley was characterized by down-valley flow during the time period of this cross-section.

The particle identification during IOP 8 along the same cross-section discussed above exhibited a horizontally layered structure, characterized by wet snow in the melting layer with dry snow above (Fig. 14(c)). The dominant particle-growth mechanism at upper levels was evidently vapour diffusion onto ice particles. The S-Pol radar detected no evidence of graupel whatsoever in this case. Fig. 13(b) shows the hourly evolution of the frequency of occurrence of hydrometeors during IOP 8 averaged over the S-Pol domain. The evolution showed a stratiform background with dry snow over a layer of

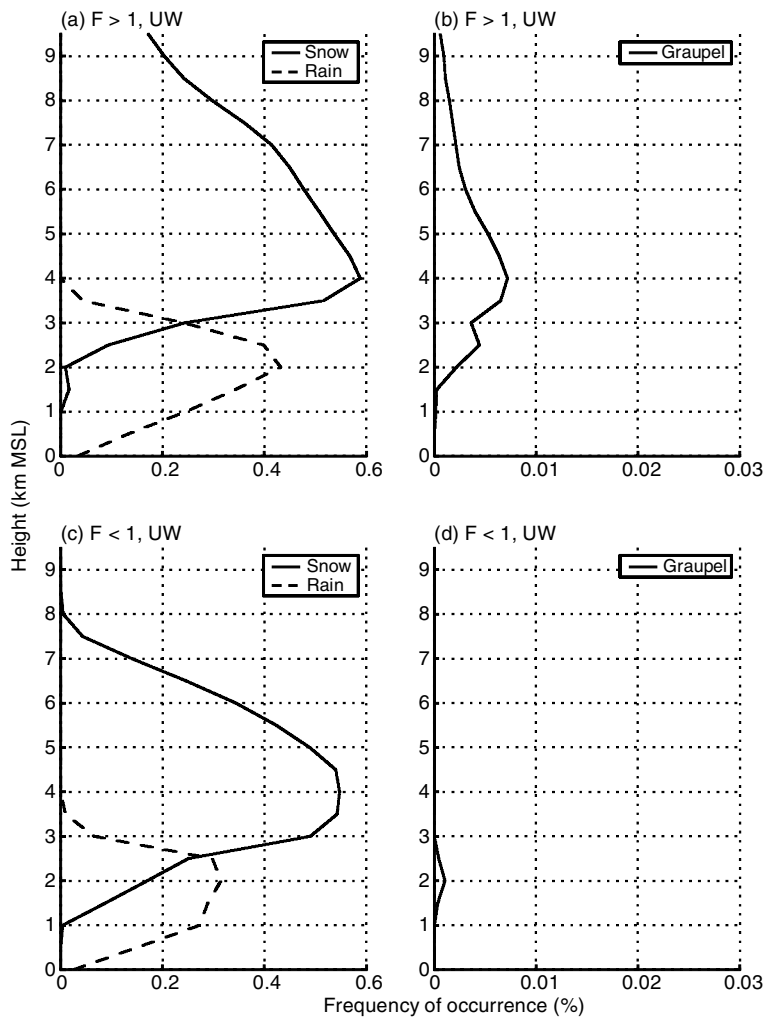


Figure 16. As Fig. 15 but particle types identified by the UW algorithm (see text).

wet snow. This stratiform structure was steady throughout the storm, with graupel and growth by riming practically absent.

(iii) *Microphysics observed by S-Pol in all MAP IOPs.* To determine whether the characteristic growth mechanisms seen in the two cases analysed here apply throughout the autumn season, we averaged the microphysical data types from the S-Pol radar over the whole area covered by S-Pol and over the whole time period of the MAP Special Observing Period (SOP), the results for both the NCAR (Fig. 15) and UW (Fig. 16) algorithms are presented. To construct these vertical profiles, only quasi-horizontal elevation angles (less than 20 degrees) and pixels only out to a range of 60 km were considered. The results are subdivided into low- Fr (<1) and high- Fr (>1) cases. Following Houze *et al.* (2001), the radar data were composited according to the Milano sounding. The Fr was calculated using the 925–700 hPa layer. If no sounding was available within 3 h of a radar scan time, the scan was not used for the Fr composite.

To restrict the analysis to cases with flow perpendicular to the terrain, only south-easterly and southerly winds were considered, i.e. soundings with the layer-averaged 925–700 hPa flow direction between 112.5 and 202.5 degrees azimuth. The low-*Fr* cases are likely to exhibit blocked-flow characteristics while the high-*Fr* cases exhibit unblocked flow. Low-*Fr* cases had a simple stratiform structure with a layer of dry snow, occurring most often at 3–5 km, overriding a layer of wet snow and light rain (Fig. 15(c)). The snow layer topped at around 7 km. The polarimetric data indicated practically no moderate rain or graupel in the blocked cases (Fig. 15(d)). The high-*Fr* cases had a deeper layer of precipitation, with dry snow extending up to about 9 km (Fig. 15(a)). The layer of dry snow dominated the upper levels 2.5–8 km, peaking in frequency at the 4 km level. There was a layer of wet snow peaking at about 2.5 km and a layer of light rain peaking at 1.5–2 km. The high-*Fr* cases differed most notably from the low-*Fr* cases by the intermittent occurrence of moderate rain at lower levels (peaking in frequency at the 2 km level). Graupel occurred intermittently at levels between 2.5 and 6.5 km, with a maximum frequency at about 4 km (Fig. 15(b)). The results from the UW algorithm (Fig. 16) coincide with Fig. 15, making our conclusions more robust.

Dry snow (solid lines in Figs. 15(a) and (c), and 16(a) and (c)) occurred more frequently at 4 km, where the particles grew by vapour diffusion. Graupel (solid lines in Figs. 15(b) and 16(b)) was only present in the high-*Fr* cases, where it reached its maximum around 4 km where the dry snow was also maximum. The graupel particles occurred much less frequently overall than the dry snow, forming intermittently in small cells embedded in the snow.

5. CONCEPTUAL MODEL

Figure 17 proposes conceptual models for the orographic precipitation mechanisms in the stable blocked, and unstable unblocked flows. The bases for these models are threefold: the results from IOPs 2b and 8 (described in sections 3 and 4); all the particle identification inferred from the S-Pol radar data in other MAP IOPs (section 4(c)(iii)); and all the radar data obtained in the Lago Maggiore region in 1998 and 1999 (Houze *et al.* 2001). The first conceptual case represents a stable blocked flow (similar to IOP 8) over the windward slopes north-west of the Lago Maggiore (Fig. 17(a)). In this case, the lowest-level flow does not rise over the terrain; only the flow above about the 1 km level (~ 900 hPa) rises over the windward slope. However, without the lowest 1 km layer rising much of the potential condensation and precipitation could not occur over the lower windward slopes. The lifting of the layer above 1 km is then too gentle to produce liquid water for riming. If this rising layer of air is also stable, no convective cells can form and make liquid water available for riming. The resulting precipitation is stratiform. Ice particles simply form in the ascending flow, drift downward, grow by vapour diffusion, and fall to the ground as stratiform rain. The dominant mechanism for increasing the mass of precipitate in this scenario is vapour diffusion onto ice particles (aggregation does not increase the mass of precipitation).

The unblocked case (Fig. 17(b)) has a general background of stratiform cloud and precipitation, similar to that in the stable blocked case. However, under conditions of high *Fr* (unblocked) the entire layer of air, including the surface layer, rises easily up the terrain. The inclusion of the moist low-level air in the air ascending over the windward slope of the range, makes the liquid-water content over the first peak of the terrain higher than in the blocked case. The resultant riming of ice particles has two important effects in increasing the efficiency of the orographic enhancement on the windward slope: (i) to increase the bulk mass content (i.e. mixing ratio) of the precipitation over the first

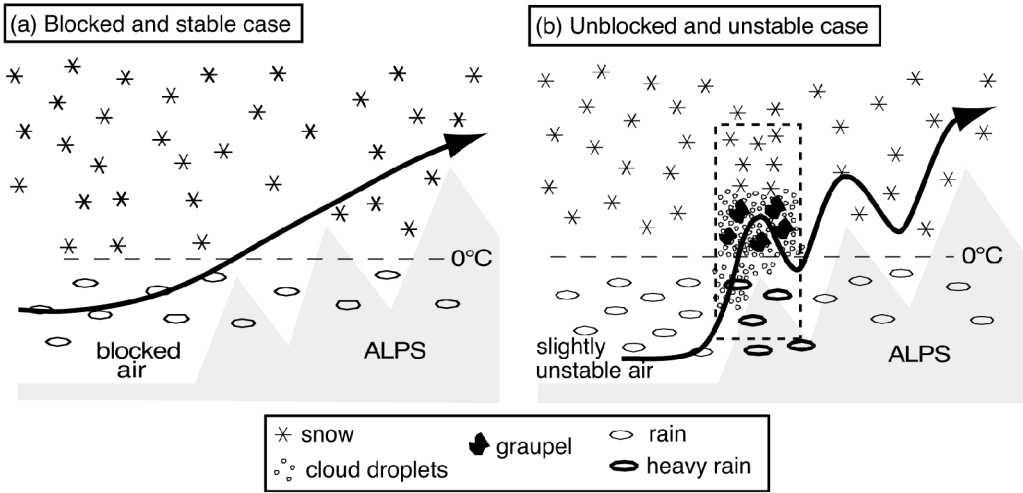


Figure 17. Conceptual model of the orographic precipitation mechanisms active in (a) stable blocked, and (b) unstable unblocked flows. The diagrams show the types of hydrometeors present in each case, along with the behaviour of the flow. The dashed box in (b) indicates the position of the embedded convective rain shower.

peak; and (ii) to produce individual precipitation particles (i.e. graupel) that have large enough fall velocities to fall out quickly (i.e. release the precipitation) over the lower windward slopes. If, in addition, the upstream moist airstream is potentially unstable (as in IOP 2b), convective cells may be triggered in the sudden upslope ascent facilitated by the high *Fr*. These cells embedded in the background stratiform cloud and precipitation produce pockets of especially high concentrations of cloud liquid water. The embedded buoyant convective cells thus accentuate the coalescence and riming processes occurring in the high-*Fr* flow. The cumulative effect of the embedded buoyant convective cells is to deepen the core of high storm-mean reflectivity seen over the first peak of terrain (e.g. Fig. 12(a)). With the help of the embedded convection, raindrops grow even more rapidly by coalescence at low-levels, and ice particles grow more rapidly by riming above the 0 °C level. The windward-slope precipitation enhancement in the unblocked case is most extreme when there is an apparent combination of high-*Fr* flow bringing low-level moist air rapidly up over the first major peak of terrain, and embedded buoyant convective cells occur, producing pockets of increased cloud liquid-water content and extending the depth over which the lifting and liquid-water production occur.

6. CONCLUSIONS

Two major rainstorms in MAP (IOPs 2b and 8) occurred with the passage of strong baroclinic waves, and in each case the precipitation patterns on the Mediterranean side of the Alps were highly modified by the Alpine terrain. However, the modification of the large-scale flow differed markedly between the two cases. Our analysis suggests a strong link between the orographically modified dynamics of the storm and the microphysical growth mechanisms.

Unstable unblocked (or high-*Fr*) flow conditions (represented by IOP 2b) produced more cloud water both below and just above the 0 °C level, which favoured coalescence below the 0 °C level and graupel production just above. Both the coalescence and the graupel production contributed to locally heavy rain on the lower windward slopes of

the Alps in the region observed by the S-Pol radar. During IOP 2b, the flow toward the barrier was strong at all levels. From a fluid dynamical standpoint, the air rose easily over the terrain because the upstream flow had a high Fr (low stability, with a strong wind toward the barrier). As a result of this lifting of moisture-laden low-level air, precipitation amounts in the Lago Maggiore region exceeded 100 mm at numerous locations and were more than double that in some spots along the lower slopes of the Alps. In addition, the slight instability of the upstream flow reaching the windward slopes favoured the occurrence of embedded convective cells over the lower slopes. These cells provided locally strong updraughts (Yuter and Houze 2001, 2003), which produced locally large concentrations of cloud liquid water. The S-Pol radar observations of a low-level reflectivity maximum (reminiscent of that described by Caracena *et al.* 1979) suggested that these conditions favoured rapid drop growth by coalescence at lower levels. The S-Pol polarimetric data further indicated a maximum frequency of occurrence of graupel just above the reflectivity maximum, suggesting that the concentrated cloud liquid water in the cells promoted growth of precipitation particles by riming above the 0 °C level. Fallout of both the graupel and the coalescence-produced raindrops, favoured locally heavy precipitation at the top point of the rise of the radial velocity jet over the first major peak of terrain encountered by the upstream flow.

In the region north-west of the S-Pol radar, stable, blocked conditions (represented by IOP 8), inhibited the lowest-level air from rising over the terrain. Therefore, air above about 1 km (900 hPa) rose over the terrain, and the air below that level backed away from the mountains in the region north-west of the Lago Maggiore. Clouds and precipitation over the mountains thus depended entirely on condensation of moisture above the 900 hPa level. The stability of the rising layer of air prevented cellular convection from forming and enhancing the orographic rainfall. Despite these handicaps, precipitation amounts were large in IOP 8, though not as large as in the unblocked case of IOP 2b, and were generally under 100 mm. The radars showed that the rain in IOP 8 was produced entirely by a simple stratiform process over the windward slopes. The polarimetric radar observations indicated only dry snow aloft, with wet snow in a well-defined melting layer and rain below. In the absence of cellular convection, graupel was not detected at any time during the storm.

The radar observations at the Lago Maggiore region for two autumn seasons suggest that under southerly and south-easterly low-level winds, the precipitation events in the region subdivide into high- and low- Fr events (Houze *et al.* 2001). IOPs 2b and 8 have proved to be instructive proxy cases for these two storm categories, respectively. The reflectivity and radial velocity patterns stratified by Fr in these two case-studies mirror the conditions seen in the corresponding Fr categories in the Lago Maggiore radar climatology. The S-Pol data collected in IOPs 2b and 8 further suggest fundamental microphysical differences between the unstable unblocked, and the stable blocked cases. A basic widespread stratiform structure, consisting of dry snow aloft growing by deposition, melting and then falling out as rain, prevailed in both unblocked and blocked cases (Figs. 17(a) and (b)). In the unstable, unblocked cases the basic stratiform structure was enhanced, since the lower layer of upstream flow rose up the barrier as a result of its high Fr , and further because cellular convection was embedded in the stratiform background precipitation (Fig. 17(b)). These enhancements produced graupel above the 0 °C level and heavier rain below, directly over the major peaks of the windward-side terrain.

Further examination of the S-Pol particle-identification fields shows that the microphysical characteristics of IOPs 2b and 8 were representative of the accumulated statistics for all the IOPs in the MAP SOP, i.e. for the whole autumn season. Statistics

of the frequency of occurrence of hydrometeors during the season indicate that in high-*Fr* cases, the general background precipitation was stratiform. A deep layer of dry snow overlaid a layer of wet snow, which overlaid a layer of light rain. Moderate rain occurred intermittently within this background stratiform structure below the 0 °C level. Graupel appeared intermittently above the 0 °C level. The graupel particles and moderate rain apparently occurred in convective cells over peaks in the terrain. The statistics show further that the graupel was absent in the blocked cases. Those cases had only the background stratiform layering of precipitation, and the layer tended to be shallower. This stratiform precipitation was copious as a result of its persistence over the windward slopes, as in IOP 8, but limited in intensity since the lower layer of air was blocked and did not participate in the lifting, and because embedded cells tended not to occur under these more stable conditions.

The case-studies and statistics of the particle types identified by polarimetric radar in MAP add to the picture of Mediterranean Alpine rainfall provided by the autumn radar climatology of Houze *et al.* (2001). The two basic *Fr* dynamic regimes not only have distinct reflectivity and radial velocity patterns on radar but also display distinctly different microphysics consistent with those basic dynamic regimes.

ACKNOWLEDGEMENTS

Fruitful discussions about the case-studies presented were held with Bradley Smull, Olivier Bousquet and Richard Rotunno. Daniel Luethi kindly provided the MC2 data. Rossella Ferretti provided the rain-gauge data for northern Italy. Wind profiler data was obtained from the MAP Data Centre. Curtis James's software was used in constructing the mean radar fields. Stacy Brodzik provided engineering support. Candace Gudmundson, Kay Dewar and Jill Campbell edited the manuscript and figures. Two anonymous reviewers provided several helpful suggestions. This research was supported by National Science Foundation grant ATM-9817700.

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