Radar observations of precipitation and airflow on the Mediterranean side of the Alps: Autumn 1998 and 1999

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SUMMARY

This study constructs and analyses composite three-dimensional fields of Doppler-radar observed radial velocity and reflectivity for all precipitation events occurring in the Lago Maggiore region on the Mediterranean side of the Alps during autumn 1998 and 1999. Mean patterns for the two years are in close agreement with each other. The radar data are consistent with previous rain-gauge studies in showing that the rain was heaviest over the lower windward slopes, and decreased toward higher terrain. The three-dimensional reflectivity fields show that precipitation growth occurred mainly at low altitudes. The composite radar data show that the precipitation was most intense when the mean flow around the 2 km level was southerly or south-easterly, i.e. when the mean flow was most perpendicular to the Alpine barrier.

Soundings from Milan indicated the Froude number of the flow upstream of the Lago Maggiore region. When the Froude number was high, the flow proceeded directly up and over the terrain of the lower Alpine slopes. Under these unblocked conditions, the low-level flow (including the layer from the surface to 2 km above mean sea level) rose directly up and over the terrain, and the precipitation was greatly enhanced over the lower windward slopes and over the portions of the Po Valley just upstream of the mountains. Under unblocked conditions, the precipitation enhancement only extended a short distance (a few tens of kilometres) upstream of the Alps. When the upstream Froude number computed from the Milan sounding was low (blocked conditions), the Doppler radial velocities indicated that the low-level flow (in the layer below 2 km above mean sea level) turned cyclonically as it approached the Alpine barrier, instead of rising over the terrain. The composite radar reflectivity data showed less precipitation enhancement directly over the windward slope, but in contrast to the unblocked case showed that the precipitation was enhanced 140 km or more upstream of the terrain. Evidently, the low-level flow began rising far in advance of blocked conditions.

The 1998 and 1999 autumn data sample further indicates the relative roles of wind speed and stability, which are combined in the Froude number. When the wind speed upstream was strong (>8 m s⁻¹), significant precipitation enhancement occurred on the windward slope of the Alps in the Lago Maggiore region, regardless of the static stability. However, the enhancement was far greater under unstable conditions. When the wind speed was weak (<8 m s⁻¹), the precipitation was generally near or below average, except when the stability was low and some patchy enhancement occurred over the Po Valley just upwind of the Alps. A diurnal precipitation maximum occurred in the early morning hours (0700–1000 LST), possibly where down-valley flow converged with synoptic-scale up-valley flow.

KEYWORDS: MAP Orographic precipitation

1. INTRODUCTION

The European Alps are notorious for heavy rains and floods (Lionetti 1996; Buzzi et al. 1998; Doswell et al. 1998; Ferretti et al. 2000; Rotunno and Ferretti 2001). These events occur primarily in the autumn on the Mediterranean side of the Alps, when moist Mediterranean air ahead of baroclinic waves impinges on the barrier. Orographic air motions modify the baroclinic circulation to produce locally heavy and persistent rain, which runs off rapidly in deep rocky narrow river valleys emptying into the Po Valley of northern Italy. The high Alpine mountain barrier juxtaposed with the Mediterranean moisture source make the region a natural laboratory for studying orographic precipitation. The Mesoscale Alpine Programme (MAP) (Bougeault et al. 2001) took advantage of these characteristics of the Alps by organizing a field programme to investigate orographic precipitation on the Mediterranean side of the Alps in autumn 1999.

Frei and Schär (1998) analysed data from rain-gauges to map the mean pattern of precipitation over the Alps. Figure 1 shows the pattern for the autumn season. Prominent
mesoscale maxima of rainfall accumulation occur on the southern slopes of the Alps collocated with indentations or concavities in the terrain. These maxima are mesoscale in the sense that they are much smaller than the whole Alpine massif yet much larger than individual river valleys. Two of these prominent mesoscale rainfall maxima are at 8.5°E, 46.2°N and 13.8°E, 46.2°N. At each location, the 800 m terrain contour (thick line) curves inward. (Note: all heights quoted in this study are above mean sea level.) The mesoscale rainfall maximum outlined by the rectangle in Fig. 1 is the region we examine in this study.

Figure 2 shows details of the topography within the region of the rectangle. At smaller scales, within each mesoscale maximum, the multiple ridges, peaks and river valleys further modify precipitation and runoff. The mesoscale indentations in the terrain evidently focus the large-scale moist flow over the barrier so as to maximize rain in these regions. The effects on the flow at these scales are probably in the category of quasi-balanced adjustments of the large-scale flow to the curvature of the terrain barrier. The smaller-scale peaks, ridges and river valleys produce more localized effects, where the Coriolis force is negligible.

One of the scientific objectives of the MAP Special Observing Period (SOP), conducted from 7 September to 15 November 1999 in the region shown in Fig. 2, was to investigate how the Alpine barrier affects precipitation on a descending cascade of scales from the scale of the entire barrier, to the mesoscale indentations of the barrier, to individual ridges, peaks and river valleys (Houze et al. 1998). For the SOP, seven ground-based research radars and two aircraft-borne radars collected data on the Mediterranean side of the Alps. The ground-based radar network was configured to sample the mesoscale precipitation within the area enclosed by the rectangle in Figs. 1 and 2. For the remainder of the paper we will refer to this area as the Lago Maggiore region.

We examine rain events that occurred in the Lago Maggiore region during autumn in 1998 and 1999, the year before and the year of the MAP SOP. The primary data of the study are the Swiss Meteorological Institute’s Doppler weather radar located at the
top of Monte Lema (altitude = 1.63 km, Fig. 2). This radar was the only MAP radar that operated in the Lago Maggiore region during both years. In addition we use data taken during the 1999 MAP SOP by the French RONSARD and the US National Center for Atmospheric Research (NCAR) S-Pol radars (also located in Fig. 2). We take advantage of the radar's ability to provide four-dimensional fields of radar reflectivity (a metric of the precipitation intensity) and radial velocity (an indicator of the airflow producing the precipitation) in high resolution over the whole autumn season. We analyse the data climatologically (statistically) rather than by cases to obtain the most general picture possible of the airflow and precipitation processes at work in this critical region. We make use of sounding data to characterize the upstream conditions and to stratify the data by Froude number, wind speed and stability. Specifically, the objective of this study is to use the combined analysis of the radar data from the autumn of 1998 and 1999 (hereafter called the superposed epoch analysis) to help establish the relative importance of these parameters in determining the orographic modification of the precipitation produced by the storms affecting the Lago Maggiore region.
2. Data and Methods of Analysis

(a) The Monte Lema radar

Joss et al. (1998) describe the Monte Lema C-band radar in detail. It is an operational radar, and the elevation angle (tilt) sequence is fixed and includes 20 tilts per 5-minute period. The reflectivity data undergo rigorous quality control within the radar processor. A decision tree algorithm consisting of a reflectivity threshold, a wide-band noise threshold, a Doppler velocity threshold, two statistical filters, a vertical reflectivity gradient test and a clutter map was incorporated to classify echoes as either clutter, system noise or precipitation. The technique removed virtually all of the reflectivity gates containing noise or terrain backscatter that could adversely affect statistical computations in this study. One notable exception occurred on 21 October 1999 (IOP*8 of MAP), when there was a persistent stable layer of air near the surface, and the lowest beam intercepted one mountain peak in northern Italy. This one case affected the reflectivity statistics at this point in 1999, as will be pointed out in section 3(b). Nevertheless, as with other radars, terrain shadowing, bright-band contamination, increasing sample volume size with range, attenuation and wet radome effects are potential sources of error when interpreting the data (Houze 1993; Doviak and Zrnić 1993; Joss et al. 1998).

The Monte Lema radar operates with a Nyquist velocity of only 8.27 m s⁻¹. Consequently, the radial velocity field contained extensive aliasing, which occurs whenever the magnitude of the radial component of the wind exceeds the Nyquist velocity (Doviak and Zrnić 1993). Since the dataset we use in this study contains over 10 000 individual three-dimensional volumes of radar data, it was impractical to de-alias the data by hand. We therefore used a new software package called Four-Dimensional De-aliasing (4DD), an efficient de-aliasing algorithm, which was used in real time during the MAP SOP (James and Houze 2001). 4DD successfully de-aliased the radar volumes with only occasional errors. Even in extreme events, these errors were generally very localized. They were so few as to have a minimal effect on the statistical results of this study.

We consider data taken by the Monte Lema radar during all precipitation events in the Lago Maggiore region from 1 September to 30 November 1998 and from 7 September to 15 November 1999. Three-dimensional volumes of radar data were recorded every 5 min. Following 4DD, the time interval between volumes was increased from 5 min to 1 h to reduce the autocorrelation between volumes and ease data storage requirements. The result was a reduction of the data archive from more than 10 000 radar volumes for the two seasons to 554 volumes for 1998 and 598 for 1999, for a total of 1152 volumes. The remaining volumes were then bilinearly interpolated to a Cartesian grid with a resolution of 2 km × 2 km × 0.5 km using NCAR’s SPRINT software (Mohr and Vaughan 1979), and finally converted to Unidata’s Network Common Data Format for visualization (James et al. 2000). Radar-data interpolation smoothes noise and facilitates computations with the data.

The Monte Lema radar antenna is located at an altitude of 1.63 km. The 2 km level is about the lowest available grid level that can be obtained using bilinear interpolation. In addition, the bilinear interpolation cannot produce an estimate of the conditions beyond a range of about 60 km. Therefore, we consider results for the Monte Lema radar only at or above the 2 km level and only out to a range of 60 km.

(b) The RONSARD and S-Pol radars

During the 1999 MAP season, two other scanning Doppler radars were installed at fixed locations at lower altitudes over the Po Valley (Bougeault et al. 2001).

* Intensive Observing Period.
The NCAR dual-polarization S-Pol radar (linus.atd.ucar.edu/rsf/spol/spol.html) and the French RONSARD radar (Chong et al. 2000) were at the locations shown in Fig. 2. The S-Pol was at an altitude of 0.28 km, and the RONSARD was at 0.155 km. After reducing the data to 1 h resolution, 320 radar volumes remained for S-Pol and 213 for RONSARD. Results from these radars supplement the Monte Lema results by providing information at altitudes below 2 km. The RONSARD also provides information on the radial velocity and reflectivity patterns over the Po Valley, south of the Lago Maggiore region.

(c) Calculations based on the radar data

It has been shown that sampling over long time periods, such as the autumn seasons considered in this study, greatly reduces the large uncertainties in radar rainfall estimates by removing the scatter of individual measurements (Cain and Smith 1976; Joss and Waldvogel 1990). We therefore convert horizontal patterns of radar reflectivity to patterns of rain rate to attach a meteorological unit to the reflectivity values. However, this paper does not require highly precise estimates of rainfall rates or amounts. Rather, it is the variability of the rates (or reflectivity values) in time, in space and by meteorological events that are the focus of this study. In vertical cross-sections we display the data in reflectivity units since rain rate has no meaning above the 0 °C level.

We obtain the seasonal mean rainfall rate and radial velocity during all storm events in the 1998 and 1999 MAP seasons by averaging the radar-derived rainfall rate and radial velocity fields. The rainfall rate was estimated using the Marshall and Palmer (1948) Z-R relation:

\[ Z = 200R^{1.6}, \]  

(1)

where \( Z \) is the equivalent reflectivity in \( \text{mm}^6\text{m}^{-3} \) and \( R \) is the rainfall rate in \( \text{mm h}^{-1} \).

The rain rate and radial velocity calculations treated missing echoes differently. We assumed that missing reflectivity meant that no precipitation was occurring, whereas a similar assumption could not be made with missing radial velocity gates. Therefore, the mean rain rate \( \overline{R} \) and radial velocity \( \overline{v} \) were computed as

\[ \overline{R} = \frac{1}{N} \sum_{i=1}^{N} R_i, \]  

(2)

and

\[ \overline{v} = \frac{1}{n} \sum_{j=1}^{n} v_j, \]  

(3)

where \( N \) is the total number of volumes, and \( n (\leq N) \) is the number of volumes with available radial velocity values at the grid point in question. From (2), it is evident that maxima in the mean rain-rate field indicated regions where the precipitation was either more frequent, more intense or both. To remove some of this ambiguity, a third field called the precipitation frequency was calculated and defined as the percentage of the total volumes in which reflectivity \( \geq 13 \text{ dBZ} \) (\(~0.16 \text{ mm h}^{-1}\)) was observed at each grid point.

In order to investigate the sensitivity of precipitation to a given variable, superposed epoch analyses (e.g. Reed and Recker 1971) were performed. This analysis extracted
all volumes from the archive whose time stamps (beginning time of data recorded in a volume) corresponded to epochs, defined by some specified condition. Then, the mean rainfall rate, radial velocity and standard deviations were computed at each grid point in the sample of extracted volumes.

\( (d) \) Statistical testing

To indicate the statistical significance of our maps of rainfall rate over seasons and individual superposed epochs within seasons we have computed a standard \( t \)-statistic and exercised the Student’s \( t \)-test at each pixel in the domain (see the appendix). The \( t \)-statistic was plotted as a horizontal field for each map that we constructed. The most informative \( t \)-statistic maps are included and will be discussed where relevant in section 4.

\( (e) \) Variables characterizing the upstream flow

We define meteorological events within the autumn season primarily by characteristics of the upstream flow. The flow into the region of precipitation observed by radar was sampled every six hours by radiosonde measurements from the Milano-Linate Airport (labelled ‘Milano’ in Fig. 2). From these data we calculated wind direction, wind speed, static stability and Froude number to characterize the environment immediately upstream of the radar-observed area. The values of these variables defined data subsets used in the superposed epoch analysis. Each variable was represented by its average value within the 925–700 hPa layer (corresponding roughly in altitude to 0.75–3 km) in the sounding. If no sounding was available within 3 h of a radar volume time stamp, the volume was not used for those superposed epoch analyses that required sounding information. Sounding information was available for only 480 of the 554 radar volumes for the 1998 MAP season, for 549 of the 598 Monte Lema radar volumes, and for 190 of the 213 RONSARD radar volumes for the 1999 season.

The Froude number \( Fr = U/(NH) \), where \( U \) is the mean (surface to \( H \)) upstream flow speed, \( N \) is the Brunt–Väisälä frequency and \( H \) is the height of the mountain barrier. We let the terrain height \( H = 2.5 \) km, the characteristic elevation of mountain passes along the north side of the Lago Maggiore region. We performed superposed epoch analyses based on the flow strength, static stability and Froude number for southerly and south-easterly flow events (specifically, those in which the direction of the mean flow in the 925–700 hPa layer was between 112.5 and 202.5° azimuth). The Brunt–Väisälä frequency was computed using finite differences over the 925–700 hPa layer. We will use the moist Brunt–Väisälä frequency (Durran and Klemp 1982). This choice is consistent with the fact that we are considering only cases of precipitation seen by radar. We assume that even if the upstream Milano sounding was less than saturated, the air was saturated by the time it was producing rainfall over the Alps. Experiments assuming dry or moist Brunt–Väisälä frequency based on a humidity threshold (e.g. Mass and Ferber 1990) led to erratic results.

3. OVERALL AVERAGE PATTERNS OF RADIAL VELOCITY AND REFLECTIVITY

\( (a) \) Mean radial velocity pattern

Figures 3(a) and (b) show the Monte Lema radar’s mean radial velocity field for both 1998 and 1999. The mean radial velocity field for the 1998 season (Fig. 3(a)) indicates that the prevailing wind direction at 2 km was south-south-easterly and its magnitude was 4–6 m s\(^{-1}\). The 1999 season mean radial velocity (Fig. 3(b)) compares well with
Figure 3. Constant-altitude plots containing the mean radial velocity during all precipitation events observed during two Mesoscale Alpine Programme seasons by the Monte Lema radar: (a) 1998 (554 radar volumes included in the calculation) and (b) 1999 (598 radar volumes included in the calculation) at 2 km above mean sea level. (c) to (e) 1999 RONSARD radar (213 radar volumes included in the calculation) at (c) 2 km, (d) 1 km and (e) 0.5 km above mean sea level. The range ring spacing is 20 km. Note that by Swiss convention, negative (positive) radial velocities denote outbound (inbound) flow.
that of 1998, except that the flow was slightly more intense (6–8 m s\(^{-1}\)) and had a slightly more easterly direction. For 1999, the RONSARD and S-Pol radars deployed for MAP provided low-level radial velocity data not available in the Monte Lema data. The mean radial velocity observed by the RONSARD radar during autumn 1999 at an altitude of 2 km (Fig. 3(c)) was similar to that of Monte Lema at 2 km. The wind direction shifted to east-south-easterly at 1 km (Fig. 3(d)) and to easterly at 0.5 km (Fig. 3(e)).

A change from easterly to southerly with increasing height at low levels was also a characteristic of the Piedmont flood on the Mediterranean side of the Alps in November 1994 (Buzzi et al. 1998; Doswell et al. 1998; Ferretti et al. 2000; Rotunno and Ferretti 2001).

The predominance of south-easterly flow at 2 km, as indicated by the radial velocity fields in Figs. 3(a) to (c), agrees with a longer-term rawinsonde climatology compiled by Kappenberger and Kerkmann (1997) for autumn precipitation events over the Lago Maggiore region. They found that the synoptic setting most favourable for precipitation over the Lago Maggiore region was amid south-westerly flow at 500 hPa, ahead of an approaching upper-level trough, while at 850 hPa the wind directions ranged from east through south to south-west. Inspection of the average radial velocity field at higher levels (not shown) indicates that the wind veered with height during precipitation events in the Lago Maggiore region, becoming south-westerly at 5 km, consistent with Kappenberger and Kerkmann’s climatology.

(b) Mean radar reflectivity pattern

Figures 4(a) and (b) show the mean rainfall-rate field derived from the Monte Lema radar data for the 2 km level for September–November 1998 and 1999. The rain-rate fields for the two years are consistent with each other, which indicates that the MAP SOP comprised a dataset representative of the autumn Alpine climatology. The spot of abnormally high reflectivity ~19 km north-west of the radar in the 1999 reflectivity map (Fig. 4(b)) was produced by anomalous propagation in IOP8, as mentioned in section 2(a). This point was the only one in either year that was obviously affected by clutter. The influence of this clutter is evident in several of the figures to be discussed later; this point should be ignored.

The rain-rate fields in 1998 and 1999 are also consistent with the rain-gauge-based climatology of Frei and Schär (1998; Fig. 1) but provide additional local detail plus three dimensionality. A vertical cross-section through the mean reflectivity field for the 1998 season (Figs. 4(c) and (e)) is consistent with the results of Frei and Schär (1998) in that the maximum precipitation occurred over the windward slopes of the Alpine range rather than over the highest terrain. The melting level was nearly always between 2 and 3 km. The radar bright band may have caused the altitude of maximum reflectivity implied by radar to be located somewhat higher than its actual location on the windward slope, but still well below the maximum height of the Alpine barrier (~4 km). The maximum in the vertical cross-section occurred between x = 25 and x = 40 km on the horizontal axis. Terrain clutter may have caused some slight overestimation of the radar-echo intensity as a result of residual clutter or side-lobe echoes in this region; however, the Swiss Meteorological Institute’s intensive quality control of the raw data (section 2(a)) all but eliminates this possibility. Echoes occurring downstream, over the highest peaks (~x = 10 km) and upstream over the Po Valley, were of consistently lower intensity. The vertical cross-section of precipitation frequency (Figs. 4(d) and (f)), defined as the percentage of volumes in which the reflectivity equaled or exceeded 15 dBZ, was also maximum on the lower windward slopes, where the average reflectivity (Figs. 4(c) and
Figure 4. Constant-altitude plots at 2 km above mean sea level containing the mean radar-derived rainfall rate observed by the Monte Lema radar during all precipitation events during (a) the 1998 Mesoscale Alpine Programme (MAP) season (554 radar volumes included in the calculation) and (b) the 1999 MAP season (598 radar volumes included in the calculation). Range ring spacing is 20 km. Vertical cross-section along the red line segment in (a) depicts: (c) radar-derived reflectivity and (d) precipitation frequency (i.e. the percentage of volumes for which the reflectivity ≥13 dBZ at each grid point) during the 1998 MAP season as observed by the Monte Lema radar (554 radar volumes included in the calculation), and (e) radar-derived reflectivity and (f) precipitation frequency during the 1999 MAP season as observed by the S-Pol radar (320 radar volumes included in the calculation). The green contour represents the terrain profile.
was maximum; hence this maximum was not a transient feature but a robust characteristic throughout the season.

Since the S-Pol radar's beam was not blocked at low levels as much as that of the Monte Lema radar, the vertical cross-sections of reflectivity in Figs. 4(e) and (f) further show how the mean echo pattern extended below 2 km down to near the surface of the terrain. The height of the maximum reflectivity (slightly above ~2 km) was the same in both years. By examining the individual storms making up the seasonal mean pattern, we determined that sometimes the precipitation over the windward slopes was purely stratiform, while in other cases the convective cells were embedded in a stratiform background. In the stratiform cases, the maximum reflectivity at 2–3 km was evidently determined by melting; the 0 °C level was usually at 3 km, sometimes as low as 2 km. When the stratiform precipitation was enhanced by embedded convective cells, the reflectivity pattern showed cells with maximum intensities in the 2–4 km altitude range against the background of stratiform radar echo. The cellular convective echo maxima were likely produced by coalescence of drops and/or riming growth of the precipitation particles (Houze and Medina 2001). The echo pattern in Fig. 4(e) was the combination over the whole season of the stratiform and convective precipitation mechanisms, with a bright band extending across the section at the 2–3 km level, out over the Po Valley, with embedded mean cells over the lower Alpine terrain, where the embedded convective echoes preferred to occur.

In the 1999 S-Pol data, the maximum echo was shifted ~10 km to the south-south-east (Fig. 4(e)), a feature that was also seen in the 1999 Monte Lema data (not shown). The precipitation frequency seen in S-Pol in 1999 (Fig. 4(f)) was also consistent with the pattern seen in the 1998 Monte Lema observations, and also shifted to the south-south-east.

The vertical cross-sections of reflectivity (Figs. 4(c) and (e)) both indicate that the maximum reflectivity consistently occurs at low altitude, below ~4 km. This altitude is approximately the height of the highest terrain downstream. This persistent feature of the echo pattern indicates that the precipitation particles form, grow and fall out quickly and efficiently over the lower slopes of the Alpine barrier, with most of the precipitation particle growth occurring at low levels. The 0 °C isotherm in this season was typically at an altitude of 3 km or lower. The echo maxima extend above this height. Therefore, the precipitation growth mechanisms are not entirely warm coalescence of liquid drops. Ice evidently also plays a role in the growth. Polarimetric measurements with the S-Pol radar in the MAP are consistent with these inferences (Houze and Medina 2001). Further study of the MAP data will focus on identification of these low-level growth mechanisms over the lower windward slopes of the terrain.

Since the 1998 and 1999 autumn seasons are in general agreement, we have merged the data for the two years in the analyses described in the remainder of this paper.

4. REFLECTIVITY AND VELOCITY PATTERNS BY SUPERPOSED EPOCH

(a) Rain distribution as a function of direction of impinging flow

As mentioned in section 1, the terrain curves inward over the Lago Maggiore region. Figure 5 indicates the influence of the concave shape of the topography on the precipitation in relation to the upstream wind direction. The individual panels of this figure present the radar data for autumn 1998 and 1999 according to the average wind direction in the 925–700 hPa layer, as computed from the Milano soundings (section 2(e)). For each wind-direction category (east, south-east, south, and southwest), the panels on the left show the distribution of radar-derived rain rate while the
Figure 5. Constant-altitude plots of mean rainfall rates (left column) and $t$-statistics of rain rate (right column) at 2 km above mean sea level (MSL) for all precipitation events in which the layer-averaged 925–700 hPa wind in the Milano-Linate sounding during the 1998 and 1999 Mesoscale Alpine Programme seasons indicated flow from (a) and (b) 67.5°–112.5° (91 radar volumes included in the calculation), (c) and (d) 112.5°–157.5° (183 radar volumes included in the calculation), (e) and (f) 157.5°–202.5° (248 radar volumes included in the calculation); and (g) and (h) 202.5°–247.5° azimuth (284 radar volumes included in the calculation). The 800 m MSL terrain contour is shown. Range ring spacing is 20 km. Red (blue) contours indicate regions where the $t$-statistic is greater than or equal to 1.96 (less than or equal to −1.96) and the null hypothesis can be rejected with a 95% confidence level.
panels on the right show the statistic of the rain pattern (computed as described in section 2(d) and the appendix). Each panel also shows the 800 m topographic contour. This contour indicates both the fine-scale pattern of ridges and valleys in the lower portion of the Alps and the broader outline of the mesoscale concave indentation of the Alps surrounding the Lago Maggiore region (cf. Figs. 1 and 2). The contour lies generally east, north, and west of the Monte Lema radar, with the lowland region of the Po Valley lying to the south. Thus, winds over this region from the east, south-east, south, and south-west each encounter rising terrain.

Figure 5(a) shows the rain-rate field at 2 km for the easterly flow cases (wind direction between 67.5 and 112.5°), which favour upslope enhancement of precipitation over the western slopes of the mountains surrounding the area covered by the radar. The statistical significance of this result is shown by the field of the t-statistic of the rain rate at the 2 km level (Fig. 5(b)). The red areas, where \( t > 1.96 \), indicate where the sample mean is significantly above the seasonal mean. Blue areas, where \( t < -1.96 \), indicate where the sample mean is significantly below the seasonal mean. Over most of the domain, the precipitation rate was not significantly different from the seasonal climatology. Rates were significantly above the seasonal mean over some regions to the west, where the flow was upslope, and significantly below over a few slopes of the eastern side, where the flow was downslope. The composite for south-easterly flow (112.5–157.5°, Fig. 5(c)) shows intense precipitation over most of the domain with a statistically significant precipitation increase over the western slopes and over the Po Valley (Fig. 5(d)). The cases with southerly flow (157.5–202.5°, Figs. 5(e) and (f)) had intense, significantly increased precipitation over all the lower slopes of the Lago Maggiore area. Comparison of Figs. 5(c) and (e) indicates that both south-easterly and southerly flow produced large orographic enhancement of rainfall on the lower slopes of the Alps, with the latter providing the strongest upslope component. In south-westerly flow, precipitation was below the seasonal mean, especially in the western portion of the radar domain, evidently because of downslope flow (Figs. 5(g) and (h)).

(b) Froude number

From section 4(a) it is evident that the south-easterly and southerly cases are associated with the most intensely orographically enhanced precipitation. Therefore, we will restrict the Froude-number analysis to cases within these two categories.

The Froude number, \( Fr \), combines the influences of stability, wind speed and terrain height. It indicates whether or not the flow has enough kinetic energy to rise over the barrier (Durran 1990; Houze 1993, chapter 12). It therefore suggests whether or not the orographic lifting will occur directly over the terrain or upstream. High \( Fr \) flow rises easily over a mountain barrier and robust upslope flow occurs, which can enhance precipitation over the windward side of a mountain barrier. When low \( Fr \) flow is blocked by the terrain, lifting and enhancement of rainfall can occur upstream of the barrier (e.g. Grossman and Durran 1984).

Figure 6 shows the fields of rain rate and t-statistic at the 2 km level for the \( Fr \) composite analyses. These patterns contain all radar data obtained when the flow into the Lago Maggiore region, as measured by the Milano sounding, was south-easterly or southerly and not blocked (\( Fr > 1 \), Figs. 6(a) and (b)) and blocked (\( Fr < 1 \), Figs. 6(c) and (d)). The t-statistic calculations show that when \( Fr \) was high, the precipitation rates over the foothills, lower slopes and western Po valley were significantly stronger.

When \( Fr \) was low, the precipitation rates were very close to the climatology, with a suggestion of blocking far upstream, near the south-east corner of the radar domain at a
Figure 6. Mean 2 km fields observed by the Monte Lema radar during the 1998 and 1999 Mesoscale Alpine Programme seasons when the layer-averaged 925–700 hPa flow direction was between 112.5°–202.5° azimuth, and the Froude number was (a) and (b) \( > 1 \) and (c) and (d) \( < 1 \). (a) Rainfall rate and (b) \( t \)-statistic (291 radar volumes included in the calculation), and (c) rainfall rate and (d) \( t \)-statistic (140 radar volumes included in the calculation). The 800 m mean sea level terrain contour is shown. Range ring spacing is 20 km. Red (blue) contours indicate regions where the \( t \)-statistic is greater than or equal to 1.96 (less than or equal to –1.96) and the null hypothesis can be rejected with a 95% confidence level.

range of about 80 km from the radar. The Alpine crest is located 60 km to the north and north-west of the radar site. The characteristic horizontal scale over which the effects of blocking occur in the case of low \( Fr \) flow is the Rossby radius of deformation, or

\[
L_R = NHf^{-1},
\]

where \( N \) is the moist Brunt–Väisälä frequency, \( H \) is the characteristic barrier height and \( f \) is the Coriolis parameter. The average value of \( N \) was about 0.008 s\(^{-1}\) when \( Fr \) was <1. Using the same value of \( H \) as for the \( Fr \) calculations (2.5 km) and \( f = 1.05 \times 10^{-4} \) s\(^{-1}\) for 46°N latitude, \( L_R \approx 190 \) km. The occurrence of upstream enhancement at a distance of 140 km of the barrier crest is thus consistent with theory.

If the upstream \( Fr \) computed from sounding data truly indicates whether or not the low-level flow is blocked, we would expect the radial velocity field observed by radar to be consistent with relatively unimpeded upslope flow in the high \( Fr \) cases, as well as with blocking in the low \( Fr \) cases. The lowest level we can examine in the Monte Lema radar data is 2 km (section 2(a)). The flow direction at the 2 km level tended toward
Figure 7. Mean 2 km radial velocity observed by the Monte Lema radar during the 1998 and 1999 Mesoscale Alpine Programme seasons when the layer-averaged 927–700 hPa flow direction was between 112.5°–202.5° azimuth, and the Froude number was (a) >1 (291 radar volumes included in the calculation) and (b) <1 (140 radar volumes included in the calculation). The 800 m mean sea level terrain contour is shown. Range ring spacing is 20 km. Negative (positive) radial velocities denote outbound (inbound) flow.

Figure 8. Mean 0.5 km radial velocity observed by the RONSARD radar during the 1999 Mesoscale Alpine Programme season when the layer-averaged 925–700 hPa flow direction was between 112.5°–202.5° azimuth, and the Froude number was (a) >1 (72 radar volumes included in the calculation) and (b) <1 (52 radar volumes included in the calculation). The 800 m mean sea level terrain contour is shown. Range ring spacing is 20 km. Negative (positive) radial velocities denote outbound (inbound) flow.

A more southerly direction in the high Fr cases (Fig. 7(a)) and toward a more south-easterly direction in the low Fr cases (Fig. 7(b)). This directional difference would be consistent with the impinging flow at lower Fr turning cyclonically as it approached the barrier, as would be expected in a blocking scenario. However, it is only a slight directional difference. Also, the strength of the flow is comparable for the two cases. The effect of blocking was apparently felt more strongly at levels below 2 km.
The RONSARD radar (section 2(b)) was able to observe the radial velocity at altitudes below 2 km because of its location in the lowlands of the Po Valley. The 1999 data from RONSARD show that the flow turned sharply cyclonically from east-southeast at high \( Fr \) (Fig. 8(a)) to north-easterly at low \( Fr \) (Fig. 8(b)). Again, this behaviour is consistent with the blocking implied by the precipitation patterns in Fig. 6.

Because of its more southern location (Fig. 2), RONSARD radar could see further upstream over the Po Valley than could the Monte Lema radar. Therefore, it can be used to investigate the upstream blocking for the \( Fr < 1 \) cases suggested by the Monte Lema radar in Fig. 6(d). \( Fr \) composites of the RONSARD rain-rate data (Fig. 9) show enhanced precipitation over the Po Valley for the \( Fr < 1 \) cases (cf. Figs. 9(a) and (d)). The area of red shading 60 km to the south of RONSARD on the \( t \)-statistic analysis of the rain rate for the \( Fr < 1 \) case (Fig. 9(e)) indicates the statistical significance of the blocking. A vertical cross-section of the RONSARD data along the red line in Fig. 9(a) shows that the precipitation frequency over the Po plain was considerably higher for the \( Fr < 1 \) cases.

\( Fr \), and hence the tendency toward blocking or not, depends on both the strength and stability of the flow. To separate these two influences, we subdivided the \( Fr \) composites according to the wind speed (Fig. 10). The cases with upstream wind speed \( > 8 \) m s\(^{-1}\) had the heaviest precipitation on the lower windward slopes of the Alps, especially north-west of the radar (Figs. 10(c) and (g)). We calculated \( U \) and \( N \) for all cases. The upstream wind speeds for the two strong-flow cases were comparable (11 m s\(^{-1}\)), hence any difference in the patterns was due to the stability. Very stable cases constitute the \( Fr < 1 \) strong-flow event (\( N = 8.7 \times 10^{-3} \) s\(^{-1}\)) and thus produce the blocking signal seen as the red (statistically high) area upstream in the \( t \)-statistic field (Fig. 10(h)). Much less stable cases constitute the \( Fr > 1 \) strong-flow composite (\( N = 1.0 \times 10^{-3} \) s\(^{-1}\)), and the airstream easily rose over the terrain. The \( t \)-statistic showed significantly enhanced rain rate both over and out to about 40 km upstream of the lower slopes of the Alps (red area in Fig. 10(d)).

In contrast to the strong-flow cases, the weak-flow composites (upstream wind speed 0–8 m s\(^{-1}\)) have very little precipitation (Figs. 10(a) and (e)). The corresponding \( t \)-statistic fields indicate that the precipitation over the lower slopes of the Alps was significantly suppressed in these cases (see the blue patches over the lower slopes in both Fig. 10(b) and (f)). The mean wind speed for the two weak-flow regimes was comparable (\( \sim 5.5 \) m s\(^{-1}\)). The extremely low mean stability for the higher \( Fr \) (\( Fr > 1 \)) weak-flow case (\( 0.2 \times 10^{-3} \) s\(^{-1}\)) evidently accounted for isolated patchy (probably convective) precipitation over the Po Valley just upstream of the slopes (Figs. 10(a) and (b)). The lower \( Fr \) (\( Fr < 1 \)) weak-flow case (Fig. 10(e)) was very stable (\( N = 7.6 \times 10^{-3} \) s\(^{-1}\)) and it produced barely any precipitation.

(c) Diurnal cycle

Figure 11 shows the total radar-estimated rainfall over the Lago Maggiore region (as defined in Fig. 2) for each hour of the day for autumn 1998 and 1999. A prominent maximum between 0700 and 1000 LST (0600 and 0900 UTC) occurred in 1998. A maximum occurred during the same time of day in autumn 1999, but it was not as prominent. It is possible that this morning maximum was a sampling fluctuation of these two particular years. However, a physical basis might exist for such a maximum. In autumn 1999, during several MAP IOPs, the mobile Doppler on Wheels (DOW) radar was deployed in the Ticino and Ticino River valleys in the Lago Maggiore region. During stable upslope flow events, the DOW observed persistent return flow within the valley.
Figure 9. Mean fields observed by the RONSARD radar during the 1999 Mesoscale Alpine Programme season when the layer-averaged 925–700 hPa flow direction was between 112.5–202.5° azimuth and the Froude number was (a) to (c) > 1 and (d) to (f) < 1. (a) 2 km rainfall rate, (b) 2 km t-statistics, and (c) vertical cross-section along the red line in (a) of the precipitation frequency (72 radar volumes included in the calculation). (d) to (f) As (a) to (c) but with 52 radar volumes included in the calculation. Range ring spacing is 20 km. Red (blue) contours indicate regions where the t-statistic is greater than or equal to 1.96 (less than or equal to −1.96) and the null hypothesis can be rejected with a 95% confidence level.
Figure 10. Mean 2 km rainfall rate (left column) and \( t \)-statistic (right column) observed by the Monte Lema radar during the 1998 and 1999 Mesoscale Alpine Programme seasons when the layer-averaged 925–700 hPa flow direction was between 112.5°–202.5° azimuth, and the Froude number was (a) to (d) >1 and (e) to (h) <1. (a) and (b) Flow below 8 m s\(^{-1}\) (121 radar volumes included in the calculation), (c) and (d) flow above 8 m s\(^{-1}\) (68 radar volumes included in the calculation), (e) and (f) flow below 8 m s\(^{-1}\) (72 radar volumes included in the calculation) and (g) and (h) flow above 8 m s\(^{-1}\) (72 radar volumes included in the calculation). The 800 m mean sea level terrain contour is shown. Range ring spacing is 20 km. Red (blue) contours indicate regions where the \( t \)-statistic is greater than or equal to 1.96 (less than or equal to −1.96) and the null hypothesis can be rejected with a 95% confidence level.
Figure 11. Total Monte Lema radar-derived rainfall estimate by the hour over the Northwest Target Area (rectangle in Figs. 1 and 2) at an altitude of 2 km above mean sea level during the autumn Mesoscale Alpine Programme seasons for (a) 1998 and (b) 1999.

in a shallow surface layer less than 2 km deep (Steiner et al. 2000; Houze et al. 2000). Airborne Doppler radar in the MAP also show this down-valley flow (Bousquet and Smull 2001). This down-valley flow could have been strongest when the atmosphere was most stable (e.g. in the early morning). One could speculate that as outflow from the valleys emptied into the Po Valley it converged with the synoptic-scale flow and thus enhanced the upward air motion and precipitation within the Lago Maggiore region.

5. CONCLUSIONS

During the autumn season in the Lago Maggiore region heavy rains and floods can occur when baroclinic waves bring Mediterranean air into northern Italy at low levels. Whenever this air impinges on the Alpine massif there is the potential for a
major rain or flood event. Seasonal climatology of Doppler-radar data for the autumn seasons of 1998 and 1999 document detailed three-dimensional characteristics of the precipitation structure and accompanying airflow. Radar reflectivity and Doppler radial velocity patterns observed in the two seasons are in close agreement with each other. One important implication of this result is that the MAP season (Bougeault et al. 2001) was generally representative of autumn rainfall on the Mediterranean side of the Alps.

Consistent with previous studies of rain-gauges in the Alps (Frei and Schär 1998), the radar reflectivity was generally strongest over the lower windward slopes and decreased toward higher terrain. Vertical cross-sections of the mean three-dimensional reflectivity patterns for both years show that precipitation generally developed at low altitudes, with most of the precipitation growth occurring at altitudes below the Alpine crest. We suggest that coalescence and/or riming led to rapid particle growth and fallout at low levels. This behaviour would be consistent with the findings of previous studies of orographic precipitation. Hobbs et al. (1973) and Hobbs (1975) concluded, both theoretically and empirically, that precipitation particle growth by riming at low levels led to quick and efficient fallout of precipitation on the windward side of the Cascade Mountains. Caracena et al. (1979) concluded that growth by coalescence of drops at low levels made the fallout of rainfall efficient in the Big Thompson Canyon flood of 1976 in the Rocky Mountains.

Superposed epoch analyses of the Doppler-radar data collected in the Lago Maggiore region during autumn 1998 and 1999 indicate that there is a clear relationship between the upstream flow direction and the intensity of precipitation over the Lago Maggiore region. These analyses further show the role of the perpendicularity between the local topography and the air stream. Over the Lago Maggiore region the precipitation was significantly greater when the wind direction around the 2 km level was southerly or south-easterly. The rainfall over the lower slopes rapidly dropped off when the flow became either easterly or westerly. When the southerly and south-easterly flows had a high Froude number, the flow proceeded directly up and over the terrain, and the precipitation was greatly enhanced over the lower windward slopes and over the portions of the Po Valley just upstream of the mountains. In extreme cases, this type of flow can lead to flooding in the Alps (Buzzi et al. 1998; Doswell et al. 1998; Ferretti et al. 2000; Rotunno and Ferretti 2001). With low Fr, southerly and south-easterly flows were strongly blocked below the 2 km level. However, at higher elevations the air stream rose over the terrain fairly easily. Thus, the enhancement of precipitation directly over the lower mountain slopes was denied the participation of the air in the lowest 2 km, which turned eastward in response to blocking. Apparently the lifting of the low-level air stream was shifted upstream, as precipitation enhancement occurred ~140 km (approximately one Rossby radius) upstream of the barrier crest in the blocked cases.

Composite analysis indicates that the speed of the flow strongly affected whether or not orographic enhancement of the precipitation occurred on the lower windward slopes of the Lago Maggiore region during autumn 1998 and 1999. The role of stability was to determine the location of the precipitation with respect to the topography. When the flow was strong and the stability low, the precipitation was greater over the lower slopes of the Alps and plains close to the mountains. During strong and very stable flow conditions (blocked case), the precipitation was also enhanced but some of the enhancement occurred upstream of the Alps. When the flow was weak the precipitation was in general equal or slightly below the mean, except for the low-stability case, when there was some enhancement in the form of patchy cells over the Po Valley.

Most of our physical interpretation of the climatological behaviour of the Doppler-radar data as a function of upstream wind speed and stability and perpendicularity
to the terrain derives from simple basic principles. Stronger wind and lower stability favour flow rising easily over terrain, with most of the precipitation enhancement occurring directly over the lower slopes of the terrain (Figs. 10(c) and (d)). If the stability is low enough, further enhancement may occur by the release of buoyant instability (Figs. 10(a) and (b)). Higher stability favours blocking, which shifts some of the orographic lifting upstream of the mountain barrier (Figs. 10(g) and (h)). These basic principles are discussed ideally in pedagogical references on orographic precipitation (e.g. Smith 1979; Houze 1993). The same principles have been used in various combinations to explain the specific behaviour of the Alpine orographic precipitation leading to the famous Piedmont flood of November 1994 (Buzzi et al. 1998; Doswell et al. 1998; Ferretti et al. 2000; Rotunno and Ferretti 2001). Our study shows how these straightforward basic principles clearly relate to the overall average behaviour of the orographic precipitation on the Mediterranean side of the Alps. Moreover, the radar climatology shows that the orographic enhancement is extremely sensitive to the basic upstream flow properties on the detailed scale of the Lago Maggiore region. The upstream flow speed and stability evidently determines whether heavy rains will be directly over the slopes or will be partially shifted upstream, over the Po Valley. Rather slight differences in prevailing upstream wind direction determine on which side of the convex indentation of the Alpine terrain bounding the Lago Maggiore region the heaviest rains and runoff will occur.

Finally, we have found that a diurnal precipitation maximum occurred between 0600 and 0900 UTC over the Lago Maggiore region. To produce this maximum, one may speculate that down-valley flow inside deep river valleys emptying into the larger Po Valley and converging with the synoptic-scale flow in the Po Valley was maximum in the early morning when the air was most stable.

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APPENDIX

In all our maps of rainfall rate, over any given time period or superposed epoch, we apply the Student’s t difference-of-means test at each grid point to indicate the statistical significance of the sample mean. An a priori confidence level of 95% rejects the null hypothesis that the mean rainfall rate at a given grid point did not differ significantly from the seasonal mean. Two-sided difference-of-means tests uses the expression,

\[
t = -\frac{\bar{x}_1 - \bar{x}_2}{\sqrt{\left(\frac{1}{N_1} + \frac{1}{N_2}\right)\left(\frac{N_1 s_1^2 + N_2 s_2^2}{N_1 + N_2 - 2}\right)}}
\]

(Spiegel 1972), where \(\bar{x}_1, s_1\) and \(N_1\) are the mean, standard deviation and number of volumes in the sample, and \(\bar{x}_2, s_2\) and \(N_2\) are the seasonal mean, standard deviation and
The total number of volumes in the archive. The null hypothesis is rejected in regions where $|r| > 1.96$, corresponding to a 95% confidence level.

The difference-of-means test requires that the volumes in the archive be mutually independent (Wilks 1995). To reduce the statistical dependence between radar samples, we used a 1 h time interval between successive volume scans. The 1 h time lag reduced the volume-to-volume autocorrelation at each grid point to an average of 0.3, with local autocorrelation minima less than 0.1 over the lower terrain, and local maxima up to 0.6 in areas where persistent orographic uplift was observed. Because the autocorrelation was higher over the higher terrain, the difference-of-means statistics in these areas may have been quantitatively exaggerated in those regions. In addition, precipitation is a highly skewed quantity; however, the difference-of-means test does not require a Gaussian distribution as long as the sample sizes in all of the analyses are sufficiently large (Wilks 1995). The sample sizes in this study are large enough to meet this condition.

The most informative maps of $r$-statistic that we have computed by these methods are discussed where relevant in the body of the paper.

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