8.1 Orographic Precipitation and Airmass Transformation: An Alpine Example

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1. Introduction

Orographic Precipitation and Airmass Transformation (i.e. OPAT) plays an important role in the earth's climate because of its influence on inter-ocean and meridional water transport, water vapor feedback, coastal rain forests, dry continent interiors, and glacier mass balance. It is also important to water resources issues and to the natural hazard of flooding. The focus of this paper is the multi-scale aspect of OPAT. We consider the net effect of the Alps on airmass transformation on a scale of 100km, as well as scales as small as 5 km on which the significant dynamical and thermodynamical processes actually occur. In order to span these scales, we use several tools. Two mesoscale models are used to provide approximate air flow and cloud fields on scales of 5 km or less. Global forecast models and analysis (ECMWF) provides field on a 40 km scale. The Alpine raingauge network provide a resolution of 25 km. The Monte Lema Doppler Radar provides reflectivity, wind fields and precipitation estimates on scales of 3 to 6 km. An additional analysis tool is the high-resolution "upslope" model, based on the hypothesis that the precipitation is driven by local uplift (Smith 2002, see paper in this volume).

The analysis is divided into two parts: volumetric and flux integrals to characterize the water budget, and airmass transformation on individual air parcels crossing the Alps. Full details and a discussion of radar data are given in Smith et al. (2002).

2. The September 20th IOP2b case

Precipitation and foehn events and climatology in the Alpine region has been studied more intensely than in any other part of the world (e.g. Seibert(1990), Frei and Schar(1998), Dorninger(), Buzzi et al. (1998), Doswell et al.(1998), Ferretti et al.(2000), Schneidereit and Schar(2000), Mladek et al.(2000), Rotunno and Ferreti(2001). The Intensive Observing Period (IOP) 2b on the 20th of September, 1999 has been studied by several MAP researchers so a detailed synoptic review is not needed here. IOP2b included the strongest precipitation event that occurred in the southern Alps during the MAP field phase and it was associated with significant flooding. The salient aspect of the event was an eastward drifting front and moist southerly jet impinging on the Italian Alps. This airstream was moistened due to strong water

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vapour fluxes over the Mediterranean and Adriatic Seas as it moved northward. As it came ashore over the Italian coast, significant moisture was lost from the jet by orographic precipitation in the coastal Alps near Nice and the coastal Apennines near La Spezia, but the jet was still quite moist when it reached the main Alpine massif. In contrast, the air in the north of the Alps is dryer due to orographic air mass transformation (Figure 1). Over the Po Valley, the low level airflow was southeasterly, presumably due to a combination of frontal, frictional and mountain blocking effects.

3. Analysis of Water Flux and Volume Integrals

An understanding of OPAT requires quantitative estimates of the components of the water budget in the air column above the mountain range. The control volumes in this study are the large box between 45.5 to 48N and 8 to 13 E and the three sub-boxes A, B, and C shown in Figure 1. The dominant water vapor fluxes for Box A are given in Table 1. Because of the predominance of southerly flow in this case, the fluxes through the east and west boundaries are less significant that those through the north and south boundaries. The water flux through the box top is small and not discussed here.

The fluxes of water vapor in through the southern boundary (Row a) and out through the northern boundary (Row b) are rather consistent across the four models (Table 1). For Box A, estimates of the influx of water vapor vary only from 49 to 56 $*10^{11}$ kg. The model precipitation values vary more noticeably (Row c). The highest precipitation was predicted by COAMPS, followed by ECMWF and MC2. For Box A, COAMPS predicts a precipitation of $19*10^{11}$kg while MC2 predicts $12*10^{11}$ kg.

The precipitation estimate in Row e of Table 1 comes from the upslope equation using the low level specific humidity and the vertically integrated winds up to 5km from COAMPS and MC2. It is more accurate to consider Row e as an estimate of the generation rate of super-saturated vapor or cloud water. The MC2 model predicts slightly higher upslope cloud water generation rate due its stronger low level winds over the slopes.

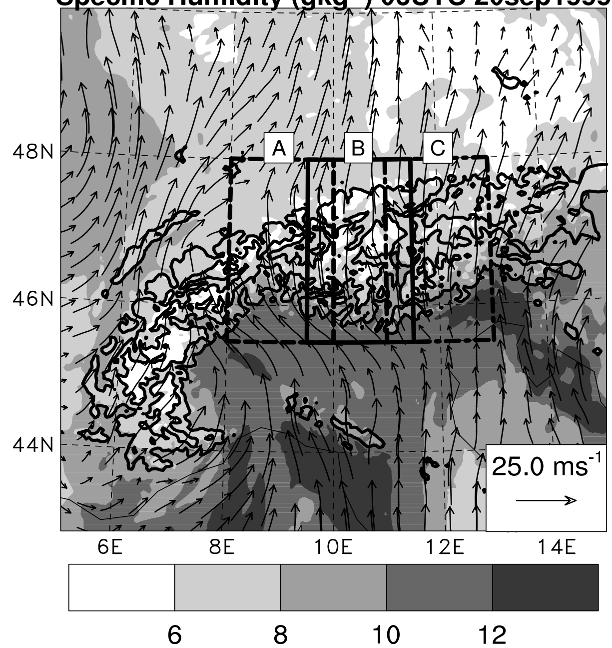

Figure 1: Low level wind and humidity at 06UTC on Sept. 20, 1999 from COAMPS

Table 1. Accumulated 24-hour fluxes and ratios for Box A derived from models and data for 20 September, 1999 (COAMPS/ MC2/ ECMWF forecast/ ECMWF analysis)

Row	Quantity	Box A
a	WV influx	56/53/49/53
b	WV outflux	33/35/39/35
c	Model Prec.	19/12/17
d	Actual Prec.	19 (19)
e	Upslope Prec.	95/108
f	DR%*(c/a)	34/23/36
g	DR%*(d/a)	33/36/39
h	PE%(d/.36a)	95/64/101
i	PE %(c/e)	20/11

Notes: Flux and precipitation values are in units 10^{11} kg .

Two values for actual precipitation are given (row d). The first comes from the analysis by Frei and Haeller (private communication from the MAP Data Center in Zurich). The second from an interpolation scheme, VERA, run by the University of Vienna. The two values agree in this case.

We define the drying ratio (DR) as the ratio of the precipitation to the incoming flux of water vapor. *DR = Precipitation / WV Influx* . Both of these input values are relatively well constrained. DR values are given in Row f of

Table 1 for three models. As the two mesoscale models have similar incoming fluxes but different precipitation rates, their DR values are different. The COAMPS values (i.e. 34 to 40%) are probably more believable as the COAMPS net precipitation is closer to the observed values. It also agrees well with the ECMWF forecast estimates, done on a much coarser grid. If DR is recalculated using observed precipitation (Row g) the values are quite consistent; ranging between 33% and 46%.

The precipitation efficiency (PE) is defined as the ratio of the rates of precipitation and cloud water generation. *PE = precipitation / cloud water generation .* This quantity is difficult to estimate and to some extent it is an ambiguous quantity. The simplest way to compute PE is to assume a smooth pseudo-adiabatic lifting of the incoming air mass by 2km. In the range of temperatures appropriate here, this process would condense about 36% of the water vapor. Using the incoming flux (Row a) and the actual precipitation (Row d), PE values are put in Row h. The values of PE computed in this way are equal to the drying ratios divided by 0.36. The values are nearly 100% for COAMPS.

An alternative way to compute PE is to use an "upslope" precipitation model to estimate the rate of cloud water generation. Values computed in this way are given in Row i of Table 1. These PE values are much smaller than Row h. To understand this result, notice that the upslope generation of cloud water (~$100*10^{11}$kg) is roughly double the incoming water vapor flux (~$55*10^{11}$kg). This apparent contradiction arises from the repeated ascent and descent over multiple ridges as the air climbs over the Alpine massif. This aspect is a unique feature of orographic precipitation over complex terrain. The same water molecules condense over and over. We have not attempted to compute the rate of cloud water generation from the models as this entails the same ambiguity as the upslope model in regions of descent. If descent is fully included in the generation of cloud water, the PE will be 100% by definition; a tautological result.

1. Trajectories and Airmass Transformation

The concept of orographic airmass transformation relates to the change in water vapor concentration associated with orographic precipitation and the change in potential temperature caused by latent heating. Bulk aspects of these changes were examined using flux calculations in Section 3, but these thermodynamic processes are better illustrated by following air parcels across the mountains. We computed several dozen air parcel trajectories from the COAMPS simulation, starting along the 45.5 latitude line at a variety of different longitudes (from 8 to 13E) and altitudes (from 500 to 6000m).

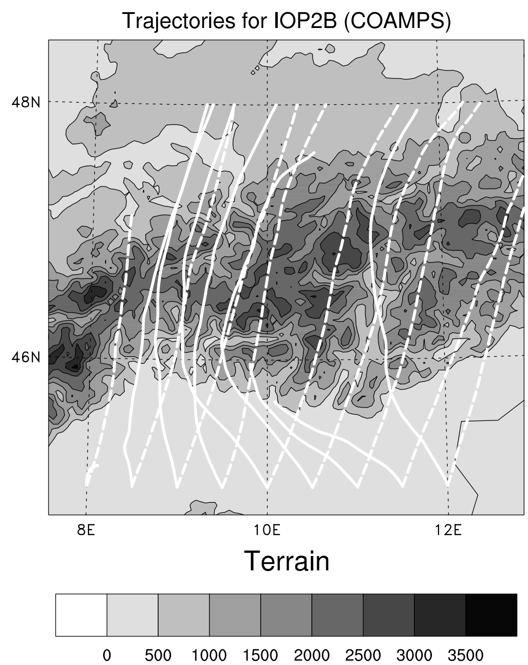

Figure 2: Trajectories launched at 06UTC from 1500 and 4000m.

The relationship between parcel warming and launch altitude for all the trajectories is rather interesting. The scatter is unexpectedly large due the differences between the western, central and eastern launches. The western launches show warming of nearly 10C from launch altitudes of 1000 to 4000 meters. Warming decreases to a small value at and above 6000m. The eastern launches show much less warming. Aloft, a significant number of parcels were cooled and moistened during their cross-mountain transit. Cooling as large as 4 degrees is noted for a few of the parcels launched between 4 and 6km. No abnormality is noted at or below the freezing level at 3000 meters. This may indicate that hydrometeor melting is not strongly influencing the temperature of air parcels.

The warming of air parcels is strongly correlated with the change in altitude during the transit. The slope of this correlation line can be expressed as a potential temperature lapse rate of 3.8 degrees per kilometer. A significant number of parcels have experienced a net cooling and descent. One parcel descended 1500 meters. The slope of 3.8 is approximately the slope of the moist adiabat for these temperatures. It is also close to the typical observed lapse rate in mid-latitudes. One can imagine that with a large variability in the warming and cooling over the Alps, the parcels must sort themselves out by potential temperature on the lee side. Parcels with a larger $\Delta\Theta$ will buoyantly rise relative to others, to find an equilibrium level. The variability in parcel thermodynamics, makes a "scrambling" of parcels part of the airmass transformation process

These results suggest some limits on the generation of foehn. Parcels which are strongly warmed and dried, find it impossible to descend (at least in a permanent way) over the lee slopes of the Alps. Only a temporary wave-induced lee descent is seen, immediately downstream of the peak. Parcels which have cooled or seen little potential temperature change, find it easier to descend (see Seibert, 1990, Doyle and Smith, 2002).

5. Conclusions

The objectives of the study are to quantify the orographic transformation of an airmass and clarify the processes involved. For the case considered, the ratio of precipitation to incoming water vapor flux is 35±5%. Precipitation efficiency values are ambiguous due to repeated small scale ascent and descent. Trajectories crossing the Alps integrate these small-scale processes to develop net warming and drying associated with airmass transformation. Strong airmass transformation extends above an altitude of 5km, including regions of net moistening and cooling aloft. Strongly warmed parcels are not able to descend along the Alpine lee slopes but rather continue to rise into the middle troposphere. Variable thermodynamics leads to parcel scrambling. Radar data (not shown) confirms the model prediction that the rainfall field is tightly controlled by local terrain on scales as small as 10km.

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