Abstract

SHU-YUN CHEN. SYNOPTIC AND MESOSCALE ENVIRONMENTS FOR OROGRAPHIC RAINFALL ASSOCIATED WITH MAP IOP-8 (under the direction of Dr. Yuh-Lang Lin)

In this study, we have adopted Penn State/NCAR Mesoscale Model version 5 (MM5) to simulate the synoptic and mesoscale environments conducive to orographic rainfall associated with Mesoscale Alpine Programme (MAP) Intensive Observation Period 8 (IOP-8). The model sensitivity tests on cumulus parameterization schemes, microphysical parameterization schemes, and terrain resolution were also included in this study.

A deep trough system associated with low-level jet approached the Lago Maggiore target area at 0000UTC 20 October 1999. During the same time period, a high pressure system was located to the east of the trough system at the same time. The high-low pressure system then remained quasi-stationary through 1200 UTC 20 October and 1200UTC 21 October. The southerly flow advected conditionally unstable air, i.e. high $\theta_e$, up to the Po Valley and the southern Alpine slopes. The sounding upstream of the Ligurian Apennines appears to contain high convective available potential energy (CAPE). Meanwhile, an easterly flow penetrated the Po Valley along the foothill of the southern Alps. The easterly flow met the southerly flow near the northern coast of the Adriatic Sea and Ligurian Sea to help enhance the orographically induced low-level convergence. As a result, the low-level convergence near the Ligurian Apennines was stronger. The easterly flow was confined in the Po Valley between Alps and Apennines and kept moving toward the west. Eventually, it flowed out through the gap between Maritime Alps and Ligurian Apennines and formed a mesoscale vortex with the southerly flow around the western Po Valley. The relative cold and stable easterly flow then piled beneath to provide a stable environment. It was proved that the cold air serves as a cold dome to make the southerly flow easily ride on it. Therefore, the upward motion near the southern Alpine slopes was very weak and not able to produce convective
rainfall. Only shallow clouds developed and stratiform precipitation was shown from both model results and observations.

On the other hand, the southerly flow produced heavy orographic rainfall over the Ligurian Apennines. Along with the low-level convergence, which was enhanced by the confluence of easterly and southerly flow near the Ligurian Sea and Apennines, the upper-level divergence also played an important role in triggering and maintaining the convective systems near this region. The right entrance of jet streak was co-located with the Ligurian Apennines surrounding area through the model integration. As seen in the model simulation, the coupling of upper-level and lower-level forcing was essential for producing rainfall over Ligurian Apennines and Ligurian Sea during IOP-8.

Based on model sensitivity tests on microphysical parameterization schemes, we found that the Reisner scheme tended to underpredict the precipitation over the southern Alpine slopes. The Goddard LFO scheme produced a reasonable amount of snow particles but overpredicted the amount of graupel, which may help explain the overprediction of rainfall over the southern Alpine slopes. As shown by the cumulus parameterization schemes sensitivity tests, the Grell scheme was not active enough to produce enough rainfall. Most of the rainfall was produced via microphysical parameterization scheme. In comparison, the Kain-Fritsch scheme did produce a fair amount of precipitation near the southern Alps and the ocean. However, it appears the Grell scheme is more suitable for the stable environment near the Lago Maggiore target area because it produced a reasonable amount of rainfall.

For the sensitivity tests on Ligurian Apennines, we found that it has a significant impact on the rainfall distribution associated with MAP IOP-8. Without the Ligurian Apennines, the cold air was spread toward the ocean and the heavy rainfall was shifted toward the southern Alpine slopes.
SYNOPTIC AND MESOSCALE ENVIRONMENTS FOR OROGRAPHIC RAINFALL ASSOCIATED WITH MAP IOP-8

by

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BIOGRAPHY

Shu-Yun Chen was born in Taipei, Taiwan in February, 1975. She received Bachelor of Science degree at Chinese Culture University in 1997. She then attended graduate program at Chinese Culture University in September 1997. She graduated with a Master of Science degree in 1999. She worked as a full time research assistant for 2 years thereafter.

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I would like to express my appreciation to Dr. Y.-L. Lin for the guidance during the past two years. I would also like to thank Dr. Carey and Dr. Semazzi for their willingness to be on my advisory committee.

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CHAPTER 1

INTRODUCTION

Topography acts as a strong permanent modifier of atmospheric circulation on wide variety of scales. A complex mountain chain like Alps has great influences on weather system. The Alps have a length of 800 km, a width of 200 km, and numerous peaks and valleys with the highest peak reaching 4.8 km. The Alpine mountains are arc-shaped and extend from an area including France, Switzerland, Austria, Italy and Slovenia to the northern coast of Mediterranean Sea and connects to the Ligurian Apennines (Fig. 1a). This complicated topography plays a significant role in influencing the surrounding flow not only in large scales but also in local scale. The intense rainfall over the Alpine mountain in fall season often causes high cost to society, and thus brings the international focus to improve the understanding of heavy orographic rainfall.

The Mesoscale Alpine Programme (MAP) field experiment took place from 7 September to 15 November 1999 in order to expand the knowledge of orographic effects. One of the major objectives of the MAP (see Binder and Schär 1996; Bougeault et al. 2001) is to understand and improve the prediction of the orographic precipitation. There were fourteen Intensive Observation Periods (IOPs) producing heavy precipitation over the Alpine region. Among them, IOP-2B and IOP-8 gain most of the attention in the MAP related studies. The synoptic and mesoscale environments of both IOPs are similar. The precipitation amount and distribution, however, had remarkable differences (e.g. Medina and Houze 2003; Rotunno and Ferretti 2003). Lin et al. (2001) summarized the following common synoptic and
mesoscale environments that are conducive to heavy orographic rainfall: (1) a conditionally
or potentially unstable air stream impinging on the mountains, (2) a moist low-level jet, (3)
quasi-stationary synoptic system to slow the convective systems, and (4) a steep mountain.

The study (Rotunno, 2000) of Piedmont flood showed that the orographically modified
flow was a crucial element for producing extraordinary rainfall. Orographically modified flow
made two important effects on the total rainfall. First, the vertical motion has the major
influence on the local rain rate. Second, the horizontal motion helps determine where and
for how long the rain falls.

Both IOP-2B and IOP-8 were associated with an eastward moving deep trough and low-
level jet ahead of this deep trough, which advected warm and moist air, i.e. high $\theta_e$ from
the Mediterranean Sea, up to the Alpine region. The real-time Mesoscale Compressible
Community (MC2) model and Swiss Model (SM) predicted heavy rainfall over the LMTA.
Indeed, the heavy rainfall did occur during the IOP-2B with the maximum amount located
over the southern Alpine slopes. However, there was only relatively light precipitation near
the southern Alpine slopes and Po Valley during IOP-8.

According to observational and numerical modeling studies (e.g. Medina and Houze
2003; Lin et al. 2002; Rotunno and Ferretti 2003), one of the major differences of the IOP-
2B and IOP-8 was the upstream condition. Based on Medina and Houze (2003), a high
Froude number was associated with IOP-2B, while a low Froude number was associated
with IOP-8. In other words, the flow belonged to the unblocked (flow over) regime (Froude
number, $Fr > 1$) for IOP-2B and the flow for IOP-8 belonged to blocked (flow around)
regime ($Fr < 1$). However, the real atmosphere is more complicated. For example, Chu and
Lin (2000) identified three flow regimes for two-dimensional conditionally unstable flow over
a two-dimension mountain ridge: (1) upstream propagation convective systems (small $F_w$),
(2) quasi-stationary convective systems (moderate $F_w$), and (3) both quasi-stationary (large
$F_w$) and downstream propagation convective systems. Those flow regimes are controlled by
the moist Froude number, $F_w = \frac{U}{N_w}$, where $N_w = \frac{g}{\theta_v} \frac{\partial \theta_v}{\partial Z}$ is the Brunt-Väisälä frequency, $\theta_v$ is
the virtual potential temperature, and $U$ is the uniform incoming flow speed.
Based on Doppler radar observation, Bousquet and Smull (2003) found that during IOP-8, there were convective activities near the Gulf of Genoa. Some convective cells extended above 10 km Mean Sea Level (MSL) over the Mediterranean Sea, which indicated that upstream conditions were unstable. In this study, we are interested in understanding the formation and propagating mechanisms of the heavy rainfall over the Gulf of Genoa and Ligurian Apennines. We hypothesize that the coupling of low-level and upper-level forcing is the crucial factor to produce heavy rainfall over the Gulf of Genoa and Ligurian Apennines surrounding areas. Due to the limitation of available observational data over the ocean in both time and space, the mesoscale model will be adopted to simulate the flow field and the precipitating systems to expand our knowledge of orographic rainfall associated with MAP IOP-8.

In numerical models, the moist processes are represented by both grid-explicit microphysical and subgrid-scale cumulus parameterization schemes. These treatments are related and have strong interaction between themselves during the model integration. Kuo et al. (1996) assessed the performance of various grid explicit microphysics and subgrid-scale cumulus parameterization schemes in the simulation of a cyclone using Penn State-NCAR mesoscale model version 5 (MM5). Wang and Seaman (1997) presented an intercomparison of a few cumulus parameterization schemes associated with precipitation events over the continental United States for both warm and cold seasons. Ferretti et al. (2000) further confirmed the above findings (Kuo et al. 1996; Wang and Seaman 1997) that the hybrid schemes should be used in 30 km resolution. With the main forcing from complex orography and ocean, the convective precipitation over the Alpine region poses a challenging problem for rainfall prediction. Based on previous studies (e.g. Rotunno and Ferretti 2003; Richard et al. 2003; Kuo et al 1996; Ferretti et al. 2000), we conducted a series of numerical experiments for MAP IOP-8 to examine the performance of cumulus and microphysical parameterization schemes in rainfall prediction. In this study, our focus will be on the rainfall prediction. Therefore, the criterion of successful prediction is based on the quantity and distribution of precipitation produced by various parameterization schemes. Thus, the comparison of
predicted rainfall with observations is essential.

Furthermore, we take use of Penn State/NCAR MM5 to assess the influence of Ligurian Apennines on the rainfall. Based on a numerical simulation of MAP IOP-2B conducted by Stein (2002), when the Apennines were removed (Alps alone experiment), the accumulated rainfall was weaker everywhere by 4 mm. However, the terrain resolution in his simulation was relatively coarser. As mentioned before, there were convective activities over the Ligurian Sea and Apennines, which remains to be explained. We hypothesize that the Ligurian Apennines play a significant role in triggering and enhancing the convective systems over the ocean. In order to isolate the impact of Apennines on the convective systems over its surrounding area and Ligurian Sea, a simulation without Ligurian Apennines will be conducted to prove our hypothesis.

This study is organized as the following. The model and experiment design will be described in Chapter 2. Chapter 3 will describe the synoptic and mesoscale environments associated with MAP IOP-8 and rainfall distribution from both available observational data and model results. Chapter 4 will further discuss the low-level and upper-level forcing and model sensitivity tests of cumulus and microphysical parameterization schemes and terrain resolution. The conclusion is given in the last Chapter.
CHAPTER 2
MODEL DESCRIPTION AND
EXPERIMENT DESIGN

2.1 Model Description

The Penn State-NCAR MM5 version 3, which is nonhydrostatic and based on terrain-following (σ) vertical coordinates, is utilized to simulate MAP IOP-8 orographic rainfall event in this study. A two-way interaction with three nested domains is used for all the simulations. For comparison purpose, all simulations are initialized at 1999/10/19/12Z and the forecast period is 48 hours end at 1999/10/21/12Z. The initial condition is provided by the NCEP 2.5° × 2.5° reanalysis data. In addition, surface observation and sounding data are also included to improve the initial fields.

A high resolution boundary layer parameterization scheme, the Blackadar (1979) scheme (also see Zhang and Anthes, 1982), is used for all simulations. There are two modules which represent two different regimes of turbulent mixing. One is a nocturnal module in which K theory is used. The second is a free convection module which proposed by Estoque (1968) and developed by Blackadar (1976, 1978). In addition, the model topography was obtained by interpolating a 30s topography dataset. The outermost domain has 85x91 grid points with horizontal resolution of 45 km, which is followed by 15 km grid resolution with 121 × 121 grid points and 5 km grid resolution with 121 × 121 grid points (Figs. 1a-c). Forty-five unevenly
spaced full-sigma levels, i.e. $\sigma = 1.00, 0.99, 0.98, 0.97, 0.96, 0.95, 0.9375, 0.925, 0.915, 0.9,
0.885, 0.87, 0.865, 0.85, 0.835, 0.82, 0.805, 0.79, 0.775, 0.76, 0.745, 0.73, 0.71, 0.69, 0.67,
0.65, 0.63, 0.61, 0.59, 0.57, 0.55, 0.53, 0.51, 0.49, 0.47, 0.44, 0.41, 0.37, 0.33, 0.29, 0.25, 0.2,
0.15, 0.1, 0.05$, are used in vertical with the maximum resolution in the boundary layer. The
total domain of 45 km resolution covers the entire Europe, eastern Atlantic Ocean, Medi-
erranean Sea and north Africa, which includes origins of incoming flow, while the inner domain
is placed on northern Italy. The time steps for these three domains are 90s, 30s, and 10s,
respectively. Various combinations of the cumulus parameterization schemes (CPSs) and
microphysical parameterization schemes (MPSs) have been adopted. These schemes will be
briefly reviewed in section 2.3. Details of the MM5 model can be found in Dudhia (1993)
and Grell et al. (1994).

2.2 Experiment Design

In order to investigate the synoptic and mesoscale environments associated with MAP
IOP-8, a series of experiments is designed. They are summarized in Table 1. Two sets of
experiments are conducted for different purposes.

The first set of simulations aims at the sensitivity of the model to various cumulus
and microphysical parameterization schemes. The sensitivity tests of various schemes help
evaluate the ability of these schemes and enable us to successfully reproduce the MAP IOP-
8 event. All simulations make use of Blackadar scheme to parameterize boundary layer
processes. CNTL is the control case with Betts-Miller CPS for 45 km resolution and Grell
CPS for 15 and 5 km resolution simulations while the LFO scheme is used to parameterize the
moist processes. RESN is the same as CNTL but substituted the LFO scheme with Reisner
scheme for microphysical parameterization. In addition to MPSs, we have also tested the
sensitivity of the model to other CPS schemes, BMKF, in which the CPS for 15 and 5 km
resolution domains are replaced by the Kain-Fritsch scheme. The result of BMKF will be
examined along with CNTL for individual CPSs rainfall prediction. Case EXPT is a 5 km grid explicit simulation directly initialized by 45 km resolution domain, in which only LFO MPS is retained. The performance of MPSs associated with MAP IOP-8 will be manifested by comparing the results of CNTL, RESN and EXPT runs. A brief review of CPSs and MPSs will be given in the next section.

The second set is for sensitivity test of terrain resolution in rainfall distributions. FTRN uses finer resolution with 5 km terrain data which has the feedback from 1 km resolution terrain. More detailed terrain resolution is expected to have stronger influence on mesoscale convective systems. As mentioned in the introduction, we are interested in the mechanism of rainfall over the Ligurian Apennines. In order to isolate the impacts of Ligurian Apennines, NOAP is conducted in which the Ligurian Apennines has been removed from the terrain input file. The results of sensitivity tests will be discussed in Chapter 4.

2.3 Brief Reviews of Parameterization Schemes

Quantitative precipitation forecast is known as one of the most difficult tasks in numerical weather prediction. Due to the improvement of observational equipment and techniques, the understanding of microphysical and convective processes has advanced significantly in the last several decades. However, in most numerical weather prediction models, the majority of clouds, especially convective clouds, cannot be resolved by grid mesh. Thus, the moist processes need to be parameterized by the grid-volume mean variables in order to represent the moist processes in the atmosphere. In mesoscale models, the treatments of moist processes can be categorized in two major ways, cloud microphysical and cumulus parameterizations. In fact, these two categories of parameterization schemes are related and have strong interaction between themselves during the model integration period. The characteristics of MPSs and CPSs used in this study will be briefly reviewed in the following.
2.3.1 Microphysical Parameterization Schemes

Two different grid-explicit microphysical parameterization schemes are used in this study to parameterize the microphysical processes within a cloud. Instead of explicit representation of microphysical processes, bulk parameterization approach is adopted in most cloud models. In bulk parameterization schemes, the water substance is divided into different categories according to its size and phase. Each category of water substance is governed by its own continuity equation, based on the conservation of mass. Every equation has its own sink and source terms to satisfy the conservation law. In order to avoid calculation of complicated interactions among different size of hydrometeor particles, the shape and size distributions are assumed a priori and the basic microphysical processes are parameterized.

The Goddard LFO scheme is based on Lin-Farley-Orville (Lin et al., 1983, hereafter denoted as LFO) scheme and developed by Tao and Simpson (1993). It has six classes of water substances including water vapor, cloud water, rain water, cloud ice, snow and graupel or hail. In LFO scheme, the shape of liquid water and ice are assumed to be spherical. The size distributions of rain, snow, and graupel or hail are hypothesized as $N_k(D) = N_{ok} \exp(-\lambda_k D_k)$, where $k$ denotes different hydrometeor particles and $N_{ok}$ is based on observations, $\lambda_k$ is the slope parameter of size distribution, and $D_k$ is the diameter of the water substance. Tao and Simpson (1993) modified the LFO scheme to include (1) the option to choose either graupel or hail as the third class of ice to be more suitable for tropical convection or tropical cyclone simulations, (2) the saturation technique to ensure that supersaturation (subsaturation) cannot exist at a grid point that is clear (cloudy), and (3) the simultaneous calculation of all microphysical processes, except melting, evaporation and sublimation, based on one thermodynamic state.

The Reisner (1998) scheme is also based on LFO scheme, which includes mixing phases and ice and graupel number concentration prediction equation. It also allows coexistence of supercool water, ice above freezing level and the melting of snow. In addition, the number of concentrations are predicted.
2.3.2 Cumulus Parameterization Schemes

Three different cumulus parameterization schemes are selected in this study, which include Betts-Miller scheme, Kain-Fritsch scheme and Grell scheme. The basic characteristics of these CPSs are summarized as follows.

Betts and Miller (1986) proposed a cumulus parameterization for deep and shallow convection, which was based on model sounding adjustment toward the reference profiler over a given time period. The instability and vertical motion are relaxed toward quasi-equilibrium states, which has been well established on larger space and longer time scales (Lord and Arakawa 1980; Lord 1982; Arakawa and Chen 1987). It does not account for subgrid processes such as updraft, downdraft and microphysical processes. This scheme is considered to be suitable for grid size larger than 30 km (MM5 Online User’s Guide, available at http://www.mmm.ucar.edu/mm5/mm5-home.html and Wang and Seaman, 1997).

The Grell scheme (1993) is a simplified version of Arakawa and Schubert (1974) scheme, which assumes deep convective clouds have the same size. In fact, the Grell scheme is a simple-cloud scheme based on the quasi-equilibrium assumption. Updraft and downdraft fluxes are considered but entrainment (detrainment) is ignored within the cloud layer. There is no lateral mixing allowed except at the top and bottom of the cloud. This permits the scheme to operate easily. When a lifting depth criterion is met to access the available buoyant energy, the Grell scheme is activated. This scheme is considered to be suitable for 10-30 km grid resolution (MM5 User’s Guide; Grell et al. 1994).

The Kain-Fritsch (1993) scheme is CAPE-based convective adjustment scheme modified from Fritsch-Chappell (1980) scheme. It includes entrainment or detrainment and the associated evaporation (sublimation) effects. Shear effects are considered and it is especially developed for grid size on the order of tens of kilometers. When the grid-resolved vertical motion is able to overcome the negative buoyancy term, the convection is initiated. Furthermore, the convection will consume the CAPE in the relaxation time of approximately 1 hour.

The Kain-Fritsch scheme tends to handle precipitation at subgrid scale for cases associ-
ated with rising motion due to strong frontal instability and deep local convection. Similarly, Grell scheme shows the same results for local convection and weak frontal instability (Ferretti et al. 2000).
CHAPTER 3
SYNOPTIC AND MESOSCALE ENvironments AND RAINFALL DISTRIBUTION OF MAP IOP-8

3.1 Synoptic Environment of MAP IOP-8

MAP IOP-8 took place between October 19 to 22, 1999. From 45 km resolution simulation of IOP-8, a baroclinic cyclone (Bousquet and Smull 2003) with minimum pressure of 990 hPa, which originated from the Atlantic Ocean, approached Spain at 0000 UTC 20 October (denoted as 10/20/00Z, see Fig. 2a). At the same time, a high pressure system located in the northern Europe with a maximum pressure of 1032 hPa. The low-pressure deepened slightly to 988 hPa at 10/20/12Z (Fig. 2b) and kept moving toward east. During this time period, the easterly flow dominated the entire Lago Maggiore target area (LMTA) over the northern Italy near the southern Alpine slopes. Notice that the easterly flow from the east of the Alps was split into two branches at the eastern tip. Part of the southern branch of the easterly flow turned to southeast and passed over the Adriatic Sea. Another part of the southern branch of the easterly flow pushed its way toward the concave region and then deflected to the Mediterranean Sea.

At 10/21/00Z a tight cut-off low, which separated from the original low pressure system,
touched the fringe of western Alps (Fig. 2c). The low-level jet (LLJ) associated with this low brought moist and warm air mass up to the Alps from the Mediterranean Sea. Since then, the surface environment at Lago Maggiore area changed. Except the easterly flow, the southerly flow from Mediterranean Sea and southeasterly flow from Adriatic Sea, which eroded the opposite direction deflected flow shown in Fig. 2b, all converged at LMTA. A convergence zone near the Lago Maggiore region was formed which is favorable for inducing lower level upward motion besides the orographic lifting. As the deep trough advanced eastward, the LLJ ahead of it continuously transported conditionally unstable air up to the Lago Maggiore area. Based on previous studies (e.g. Buzzi and Foschini 2000, Schneidereit and Schär 2000), LLJ plays an important role in transporting moist and warm air from Mediterranean Sea toward Alps and triggering the mesoscale convective systems over the southern Alpine slopes. LLJ is also one of the common ingredients of heavy orographic rainfall (Lin et al. 2001). In the mean time, the high pressure system remained about the same strength and traveled eastward very slowly, at about 2.78 $ms^{-1}$. As will be demonstrated below that the high-low pressure system became quasi-stationary and deepened during the period of 10/20/12Z to 10/21/12Z. The longer the baroclinic disturbance stayed at the target area, the more precipitation could be expected.

Moreover, this low-pressure system extended to the upper troposphere and became a trough, as shown in Fig. 3. Before the deep trough system arrived at the Alps, a jet streak with the maximum wind speed exceeding 40 $ms^{-1}$ laid across the center of the domain (Fig. 3a). At 10/20/12Z, the system deepened to become an upper-level trough. Meanwhile, the high pressure ridge to the east of this trough started to build up. During the next 24 hours, the high pressure ridge stayed almost at the same location and gradually built up. The upper-level divergence associated with the difluence near the curving exit of the jet streak can be seen near the eastern edge of the trough (Figs. 3b-d). Besides the orographic lifting, as will be shown later, this upper-level divergence might help induce mid to low-level upward motion. In order to examine the development of this trough-ridge system at 300 hPa, the evolution of constant geopotential height of 9240 m in 6h interval was illustrated.
in Fig. 4. Through the simulation period of IOP-8, the upper-level trough deepened and propagated eastward, while the high pressure ridge remained quasi-stationary. The upper-level divergence, quasi-stationary high pressure to the east of the deepening trough and LLJ, i.e. three common ingredients as proposed by Lin et al. (2001), are similar to those in IOP-2B and appear to favor the heavy orographic rainfall event of MAP IOP-8.

These simulated flow fields in both surface and 300 hPa are able to capture the major features of those observed (e.g. Bousquet and Smull, 2003) and the NCEP 2.5° × 2.5° reanalysis fields shown in Fig. 5. The location and strength of the high and low pressure system are all reasonably simulated by MM5. However, there were some minor differences between model simulation and NCEP reanalysis data. The high pressure at surface to the northeast of Alps from model simulation was lower by 2 hPa at 10/21/00Z. Besides, the simulated cut-off low slightly fell behind, which was located to the northwest of the Alps, while it already extended to the Ligurian Sea at 10/21/12Z. At 300 hPa, the trough extended farther south and the geopotential height contour of 9240 m was already past Sardinia.

In addition to the surface and upper-level analyses, the synoptic structures can also be revealed in satellite imagery. In Fig. 6a, the IR satellite imagery showed a broad cloud band associated with the deep trough to the west of Italy at 10/20/18Z. Most of the cold cloud top regions were located over the northern Mediterranean coast and extended to North Africa. As this cyclone propagated toward east, the comma-shaped cloud band moved to Italy and Gulf of Genoa between 10/21/06Z and 10/21/12Z (Figs. 6c and 6d). The orographic precipitation system was mainly associated with this cloud band. Another cloud band followed behind, which was responsible for MAP IOP-9 event and would soon have influence on Alpine region, The ocean provides abundant moisture and sensible heat fluxes to support and strengthen the convective activities. The enhanced cloud regions evidently indicated the ongoing convective activities near the Corsica, Sardinia and southern Alpine slopes. At 10/21/12Z (Figs. 6c and 6d), the cloud band moved to central Italy and gradually passed the Lago Maggiore target area. From the infrared satellite imagery we could trace the precipitation systems. However, lower clouds might play important roles in producing
the orographic precipitation, but they could not be detected from satellite imagery if they are covered by the upper-level clouds. Other observational data should be also considered, such as radar observations, in order to make a full judgement.

3.2 Rainfall Distribution and Radar Observation

3.2.1 Rainfall Distributions

As mentioned earlier, during the passage of the deep trough system, the LLJ ahead of the trough might help advect the conditionally unstable air up to the southern Alps. The 6-h accumulated precipitation from both 5 km resolution simulation and rain gauge data are shown in Fig. 7. The regions of 6-h simulated accumulated precipitation exceeding 50 mm were located near the Ligurian Apennines, Ligurian Sea and the Po Valley (south of the Alps). Part of the precipitation was near the east side of Po Valley near the foothill of Alps. We hypothesize that the southeasterly flow from Adriatic Sea and easterly flow formed a convergence zone which was responsible for this portion of early rainfall even when the trough system and LLJ were still far away from the rainfall region. Figure 7a shows that the surface wind in the vicinity of the LMTA came from three different origins: (1) eastern Po Valley, (2) Mediterranean Sea and (3) Adriatic Sea. The influences of the flow from different origins will be discussed in Chapter 4. As the LLJ and the trough propagated eastward, the maximum 6-h accumulated precipitation also increased to 80 mm. The maximum rainfall were concentrated on Ligurian Apennines surrounding the Gulf of Genoa and upstream of southern slope of the Alps near the concave region (Fig. 7b). The LLJ from the Mediterranean Sea helped transport warm and moist, i.e. high \( \theta_e \) (will be shown later), airstream to the Gulf of Genoa, Po Valley and Alps which was lifted by the Ligurian Apennines and produced heavy rainfall in the coastal area.

The rain gauge dataset used for this paper was obtained from MAP database which is available from the MAP website (http://www.map.ethz.ch). Three data providing countries
within the Lago Maggiore target area were Austria, Italy and Switzerland. Figures 7c and 7d show the 6-h accumulated rainfall at 10/21/06Z and 10/21/12Z, respectively. Before the LLJ established over the Lago Maggiore area, only light rain was observed by the rain gauge (not shown). However, moderate precipitation took place after 10/21/00Z and gradually intensified as the LLJ strengthened. The stratiform rainfall (Houze and Medina 2001) spread all over the Piedmont and the east side of Po Valley within the next 12 hours. The maximum 6-h accumulated rainfall increased to 60 mm ending at 10/21/06Z (Fig. 7c), which was located near the southern Po Valley. At 10/21/12Z, the rainfall accumulation went up to 70 mm and heavy rain lined up along the LLJ (Fig. 7d). Notice that the rainfall was confined within the Po Valley and the southern Alpine slopes, but did not reach the Alps crest. In general, the model did catch the major observed features reasonably well, such as the precipitation amount and the location. However, the model did not produce the rainfall near the plain area (in the southeast and center of Po Valley, see Figs. 7c and 7d). Both model and rain gauge observations showed that fair amount of rainfall distributed over the western flank of Alps, Ligurian Apennines, Lago Maggiore area and southern Alpine slopes. However, there is no station over the Ligurian Sea to verify the model simulated rainfall. Therefore, the radar and lightning data will be used to verify the model results.

3.2.2 Radar Observations

During the period of the MAP field experiment, many ground based and airborne Doppler radars were deployed around the Lago Maggiore area to collect high resolution data (Binder and Schär 1996). Figure 8 showed the Alpine radar composite images of IOP-8 from 10/21/03Z to 12Z which was adopted from the MAP Database center. At 10/21/03Z, a broad band-like feature of rainfall rate exceeding 30 $mm h^{-1}$, which was associated with the cyclone, entered the western boundary of the domain. Only few spots of rainfall rate higher than 1 $mm h^{-1}$ were detected by the radars at Lago Maggiore. As this elongated system marched on within the next few hours, the entire Po Valley was filled with precipi-
The maximum rainfall rate of 30 $mmh^{-1}$ was observed at the foothill of Alps during 10/21/06-09Z, after 10/21/12Z, it gradually weaken as LLJ passed through the Po Valley area.

According Houze and Medina (2001), the southeast to northwest cross section took from S-Pol radar (the location of S-Pol radar was denoted in Fig. 1c) showed a stratiform structure (Fig. 9a). The maximum reflectivity was about 34 dBZ, and a distinct bright band at 2 km level extended upstream of the Alpine mountain range. In order to examine the vertical structure of the model simulation, a model derived southeast to northwest cross section (denoted in Fig. 1c) of radar reflectivity from the CNTL simulation with 5 km resolution is presented in Fig. 9b. The similar stratiform feature with the maximum reflectivity below 40 dBZ evidently showed a successful simulation. However, the model did not predict a clear appearance of the bright band. This may be due to the cumulus and microphysical parameterizations in the model, which were not able to accurately represent the complicated cloud microphysical processes.

Bousquet and Smull (2003) analysed various radar data sets of the MAP IOP-8 and P3 airborne Doppler radar data. They found that the tallest echoes over the Gulf of Genoa and the Apennine mountain chain during the MAP IOP-8 (Fig. 8 of their paper). Their figure indicates that there were convective activities over the ocean which is consistent with the cloud band in satellite imagery (Fig. 6). On the contrary, precipitation over the Piedmont and Po Valley was more stratiform even through the mid-level convective enhancements. Their result strongly supports our model simulated rainfall over the ocean in Figs. 7a and 7b.
3.3 Mesoscale Environments

3.3.1 Stability of Upstream Airstream

Three sounding stations at different locations were selected in the south-north direction from far upstream to the foothill of the Alps to represent the transition of the mesoscale environment from the ocean to the foothill of Alps (radiosonde launching locations were denoted in Fig. 1b).

Figures 10a and 10b are the sounding profiles of Cagliari at 10/21/00Z and 06Z, respectively. Cagliari is located at the southern end of Sardinia, which may represent the upstream environment over the Mediterranean Sea. Before the wind strengthened at Cagliari, there was a thin layer of temperature inversion at about 870 hPa. Mid-level dry air was located between 800 hPa and 500 hPa. However, the inversion layer disappeared at 10/21/06Z. The Convective Available Potential Energy (CAPE) was $2304 J/kg^{-1}$ and the lifting condensation level (LCL) and level of free convection (LFC) were 971 hPa and 844 hPa, respectively. Because both LCL and LFC were low, only weak lifting is needed to trigger convection by the release of conditional instability. Below 600 hPa, the environment is relatively dry because it was originated from North Africa (Lin et al. 2002).

Because Genoa is located by the coast of northern Mediterranean Sea, the low to mid-level troposphere was near saturated as can be seen in the sounding profile of Figs. 10c and 10d. However, the air was relatively dry near the surface with temperature of 9°C and dew point of 3.8°C (RH=69.6%) at 10/21/00Z. Between 900 hPa and 850 hPa, temperature inversion was observed. At 10/21/06Z, the depth of inversion layer decreased and the air between 900 hPa and 800 hPa became almost neutral (Fig. 10d).

Milan sits right at the foothill of the Alps. The environment was already disturbed by the mountain and the approaching cyclone. In Fig. 10e, a deep layer of moist air accompanied by a temperature inversion layer between 850 hPa to 750 hPa can be seen. The easterly flow occupied the lower layer (below 850 hPa) at Milan, which brought in the cold air situated near the surface (Bousquet and Smull 2003; Rotunno and Ferretti 2003). At the surface, the
temperature and dew point were 7.2°C and 4.2°C (RH=81.1%), respectively. At 10/21/06Z, the wind turned to southeasterly at lower level. From the sounding profile (Fig. 10f), the layer below 650 hPa was saturated. This was exactly the time that the cloud band reached southern Alpine slopes (Fig. 6b).

Both sounding profiles of Genoa and Milan indicated that there was a layer of cold air at the surface and a temperature inversion existed between 850 hPa and 750 hPa. This revealed that the air upstream (south) of the Alps at foothill and in Po Valley was stable before the LLJ approaching this region. The cold air mass at the surface played a key factor for the IOP-8 event. This had been recognized in previous studies of Rotunno and Ferretti (2002) and Bousquet and Smull (2003), however no discussions have been made for the stability of the air over the Mediterranean Sea.

3.3.2 Upstream Convective Activities

Due to the coincidence of the IR cloud band and lightning activity, lightning serves as a good indicator of convective activities. The transfer of charge that occurs when graupel particles produced in the region of strong updraft collide with smaller ice particles. The polarity of the charge transfer in the collisions is dependent on temperature and liquid water content. In the cold upper levels, the graupel particles take on negative charge and leave behind a cloud of small nonfalling particles, which become positively charge when negative charge is transferred to the graupel during the collisions. The lower part of the cloud becomes dominated by the negative charge of the graupel particles descending from upper levels. Figure 11 shows the cloud to ground lightning observational data from 10/21/03Z to 12Z in a 3 hour interval. The distribution of the lightning was very consistent with the cloud bands shown in Fig. 6. Most of them were located over the ocean and the Gulf of Genoa. This also supports the model simulated rainfall over the ocean (Figs. 7a and 7b). Notice that there were few lightning occurrences near the Lago Maggiore area, even though the cloud bands passed over the Po Valley. The convective activities were clearly shown over
the Mediterranean Sea, Corsica, Sardinia and later on over central Italy, but they did not reach the southern Alpine slopes during the IOP-8 period.

### 3.3.3 Wind Patterns

The evolution of the deep trough system also can be captured by the windprofilers during the IOP-8 period. Figures 12 and 13 are the wind fields for 900, 850, 500 and 300 hPa at 10/21/00Z and 06Z from MAP SOP data center. At 10/21/00Z, the low-level flow was mainly easterly at LMTA and veered with height. This feature was also captured by our simulation (Figs. 2c and 3c). The LLJ, associated with the upper level trough, was located near the border of France and Italy. At 10/21/06Z, the LLJ arrived at the Piedmont and the easterly flow shifted to a southerly/southeasterly flow. At 500 and 300 hPa, the curving exit of the jet streak was over the surrounding areas of the Ligurian Apennines. The directional changes of the surface and 300 hPa wind fields were also captured by the model simulation.
CHAPTER 4

MODEL SENSITIVITY TESTS

4.1 Low-Level Forcing

As revealed in the previous chapter, a deep trough system approached the LMTA (Figs. 2 and 3). The surface wind in the vicinity of the LMTA came from three different origins: (1) eastern Po Valley, (2) Mediterranean Sea and (3) Adriatic Sea. According to Bousquet and Smull (2003), the progressive erosion of the cold air mass from the eastern Po Valley produced stable lifting of the strong southeasterly flow near the Po Valley and maintained or generated the localized cooling near the surface. Moreover, the easterly flow tended to escape through the gap between Maritime Alps and Ligurian Apennines. It resulted in a narrow jet-like zone of northerly/northeasterly flow extending out over the Mediterranean Sea, which was shown in our simulated results (Figs. 7a and 7b).

At 10/20/18Z, a pool of high $\theta_e$ air at 850 hPa extended from northern Africa to the Mediterranean Sea (Fig. 14a). This high $\theta_e$ tongue was associated with the warm and moist air over the ocean. At the same time, the $\theta_e$ was below 308K over the LMTA which indicated a cold and dry condition. This elevated $\theta_e$ air was associated with the eastward moving low-level jet (Fig. 2). Within the next 18 hours (Figs. 14b-d), the southerly flow (LLJ) advected the warm and moist air up to the central and east Po Valley with maximum $\theta_e$ regions located near the coast and in the northern Italian Peninsula. Note that the high $\theta_e$ airstream did not penetrate to the LMTA during IOP-8. Another prominent difference to
the south of the east-west main ridge of the Alps between 850 hPa and surface wind field, the 850 hPa winds were from south, which overrides on the surface easterly and southeasterly cool stable air.

Figure 15 shows the model simulated convergence field superimposed on the upward motion \( w \geq 2 \text{cms}^{-1} \) and wind fields at 1000 hPa. A 15 km resolution was used for this simulation, which illustrates the low-level structures surrounding the LMTA. There were three sources that contribute to the low-level convergence. When the southerly flow approached the Alps, the flow was decelerated near the barrier. Thus, the low-level convergence formed along the upwind (southern) slope of the mountain. In Fig. 15, the convergence adjacent the eastern portion of the southern Alpine slopes, Apennines and French Alps were all due to the barrier effect. As this easterly LLJ flowed toward the west, the convergence along the upwind (southern) slopes of the mountain increased. Moreover, when the easterly flow was blocked by the arc-shaped portion of the Alps, it made an anticyclonic turn along the concave region and flowed out to the sea between French Alps and Ligurian Apennines. This deflected flow collided with the southerly LLJ near the Gulf of Genoa to form another convergence zone. In fact, a mesoscale vortex was formed by the deflected flow and southerly flow (LLJ) centered on the leeside of the Ligurian Apennines. This vortex was clearly shown in 5 km resolution simulation (Figs. 7a and 7b). This near surface mesoscale vortex has also been observed by Doppler radars (Fig. 7a of Bousquet and Smull 2003).

The third contributor was the easterly flow merging with the southerly LLJ near the northern coast of the Adriatic Sea and southern Po Valley. The convergence field formed along the boundary of easterly and southeasterly flows. It also enhanced the convergence caused by the barrier effect. Consequently, the convergence at the surface was able to produce upward motion. However, because of the stable environment near surface the upward motion at low-level was small with the maximum vertical velocity around 0.4 \( ms^{-1} \) near the Apennines. Overall, the local maximum low-level convergence was located over the eastern portion of the southern Alpine slopes, Ligurian Apennines and the elongated area associated with the southerly LLJ. As revealed in Figs. 7a and 7b, the distributions of the 6-h simulated
accumulated rainfall were co-located with the low-level convergence and vertical motion.

It appears that two mechanisms can be identified for producing rainfall. The first type is the convergence and upward motion induced by orographic blocking and lifting. This type of rain is more appropriately called orographic rain, which was distributed over the southern Alpine slopes and the Ligurian Apennines and their surrounding areas. The second type was caused by the convergence associated with the eastward propagation of the deep trough system. Hence, the associated rainfall was mainly over the ocean and coastal area.

Lin et al. (2001) proposed that the orographic rainfall distribution may be diagnosed by the general moisture flux and orographic moisture flux. The general moisture flux is defined as $wq$, where $w$ is the vertical velocity and $q$ is the mixing ratio. The orographic moisture flux, which is associated with orographic lifting, was defined as $(V_H \cdot \nabla h)q$, where $V_H$ is the horizontal wind velocity, $h$ is the mountain height, and $q$ is the water vapor mixing ratio. Calculated from the 15 km resolution simulation, figures 16a and 16b show the general moisture flux fields at 10/21/06Z and 10/21/12Z, respectively, and Figs. 16c and 16d are the orographic moisture flux fields at 10/21/06Z and 10/21/12Z, respectively. At 10/21/06Z, the maximum $wq$ regions were located over the Gulf of Genoa, southeast of the massif central and over the Ligurian Sea in between southeast coast of France and Sardinia. In addition, a small area of higher $wq$ was over the northern coast of the Adriatic Sea. At 10/21/12Z, the maximum $wq$ region moved across Corsica to the east coast of Gulf of Genoa, where the maximum rainfall occurred (Fig. 7b). As the 850 hPa LLJ pushed over to the southern Alpine area (Fig. 14b), the $wq$ also increased at LMTA. These locations were also co-located with the convergence region (Fig. 15b). Note that the orographic moisture flux reflected the rainfall directly associated with orographic lifting, unlike the general moisture flux which also reflected rainfall over flat surface and ocean. Compared with the rainfall distributions shown in Fig. 7, the terrain induced vertical moisture flux was able to capture upslope precipitation distribution reasonably well. Thus, these moisture fluxes may help predict rainfall distribution, as suggested by Lin et al. (2001).

Figure 17 shows the evolution of $\theta_e$ and vector wind fields along the south-north vertical
cross section perpendicular to the Alps (see Fig. 1b) from 10/20/18Z to 10/21/12Z for every 6h. The lower mountain located in the center of the domain is the Ligurian Apennines and the higher mountain to the right is the Alps. Between these two mountains is the Po Valley. With the advancement of the LLJ, a tongue of moist and warm air (shaded for $\theta_e \geq 322K$) started to established over the Gulf of Genoa from 10/20/18Z to 10/21/00Z (Figs. 17a and 17b). During this period, the low-level air to the south and over the upwind (southern) slopes of the Ligurian Apennines was cold and very stable. The Po Valley in between Apennines and Alps contained a relatively cold air pool, too. The $\theta_e$ was below 296K near the surface (below 900 hPa) in Po Valley, which was much colder than over the Gulf of Genoa. There was a convectively unstable layer over Ligurian Sea. The southerly flow from the ocean was able to pass over the Ligurian Apennines and produced orographic precipitation (Fig. 17a). Some shallow convective clouds developed upstream of the Ligurian Apennines. In fact, there was a stable dome over the upslope of the Ligurian Apennines, as evidenced by $\frac{\partial \theta_e}{\partial z} > 0$ (Fig. 17a). The cloud regions ($q_v = 0.1 \ gkg^{-1}$) extended downstream of the Apennines which were responsible for producing lighter rainfall in Po Valley. There was no downward motion on the leeside of Apennines after the southerly flow passed over the mountain. Instead, the cold air associated with the easterly flow piled beneath as a cold dome forced the warm air (LLJ) to override it. The cloud system over the Ligurian Apennines was composed of the convective clouds over the upstream of the Apennines in upper layer and the stable clouds due to the stable ascent (Smith 1979; Chu and Lin 2003; Lin 2003) in the lower layer over the Apennines and Po Valley (Fig. 17a).

The existence of such cool airmasses and the role of the peculiar juxtaposition of the multiple mountain barriers in their maintenance is similar to the early studies (e.g. Medina and Houze 2003; Bousquet and Smull 2003; Rotunno and Ferretti 2003) and was well recognized by Seibert (1990) in his study of south Foehn events. The cool airmasses acted like part of a barrier linking the Ligurian Apennines and the Alps, similar to the effective mountain idea proposed by Rottman and Smith (1989). The cold easterly flow near the surface acted as a barrier preventing the precipitation system from propagating toward Alps. At 10/21/00Z,
the high $\theta_e$ air (shaded in Fig. 17b) advanced toward the Gulf of Genoa due to diurnal cooling (Fig. 7b). The low-level easterly flow was still pronounced before 10/21/06Z. As the more southerly flow climbed over Apennines, the $\theta_e$ over the Po Valley gradually increased and the convectively unstable layer extended to a deep layer (e.g. 700 hPa) over the upslope of Apennines (Figs. 17c and 17d). The convective instability associated with this marine boundary layer provided a strong low-level forcing in triggering the convection and precipitation over the coastal Apennines. At 10/21/12Z, the high $\theta_e$ tongue penetrated Po Valley and the surface cold dome was eroded by the warm and moist air associated with the strong southerly wind.

4.2 Upper-Level Forcing

Along with the orographic forcing and convergence at lower levels, which help trigger the convective systems, the upper-level forcing may also play an important role in inducing the middle and lower level upward motion to trigger and enhance the orographic rainfall. Figure 18 is the 300 hPa divergence calculated from the 15 km resolution simulation superimposed with the 850 hPa upward motion starting every 6 hour from 10/20/18Z. As shown in Figs. 3, 5b and 5d, strong diffuence was near the jet streak exit region. During the passage of the deep trough system, the upper-level maximum divergence regions were scattered over the Ligurian Sea, eastern-northeastern Alps, Maritime Alps, Apennines and southern Alpine slopes. The 850 hPa maximum upward motion regions were not quite in phase with the 300 hPa divergence at earlier times such as 10/20/18Z and 10/21/00Z (Figs. 18a and 18b). However, they were approximately in phase during the period of initial rain formation and heavy rainfall (Fig.s 18c and 18d). Thus, the divergence field associated with the deep trough system indeed was able to enhance the lower-level upward motion. Furthermore, the divergence adjacent to the Ligurian Sea and Gulf of Genoa was stronger than in the southern Alps. This implies that the upper-level forcing (divergence) was able to couple
with the low-level forcing (convergence and orographic lifting) (Fig. 15).

According to Massacand et al. (1998), upper tropospheric positive PV streamer can enhance convection produced by orographically induced upward motion and is closely linked to the Alpine precipitation. It can serve as a precursor for predicting heavy orographic rainfall. In order to study the relationship, we examine the 300 hPa potential vorticity superimposed with horizontal wind vector field. The maximum horizontal velocity was located on the forward flank of the eastward propagating deep trough system (Fig. 19). Based on the jet streak dynamics, the right entrance region of the jet streak favors the divergence, which produces the thermally direct circulation and upward motion. A 300 hPa maximum divergence region (Fig. 18) was approximately located to the right entrance region of the jet streak. It was located to the southwest of the Maritime Alps at 10/20/18Z (Figs. 18a and 19a), then moved eastward to be approximately over the Gulf of Genoa at 10/21/06Z (Figs. 18b-c and 19b-c). Note that this was also the time and location of heavy rainfall occurred (Fig. 7). At 10/21/12 (Figs 18d and 19d), then moved over the northern Italian Peninsula. Again, it co-located with the heaviest rainfall region. In short, Figs. 18 and 19, the 300 hPa divergence and the right entrance region of the jet streak were co-located during IOP-8. Consequently, the upper-level environment was favorable in inducing the upward motion at lower levels. The upper-level PV was approximately in phase with the upper-level trough system and located on the cyclonic curvature region of the jet stream.

With the deep trough propagating eastward, the positive PV streamer extended to the Ligurian Sea and Mediterranean Sea. In order to further investigate the influences of the PV streamer, three cross sections AB, CD and NS (AB was along the southern Alps; CD was cut through French Alps and Ligurian Apennines; NS was in north-south direction. They are denoted in Fig. 1b) at 10/21/06Z and 12Z were presented in Fig. 20. Along the cross-section AB (Figs. 19a and 19b), the profound easterly flow at low-level continually brought in a cold air mass (Fig. 14) to form a cold dome over the Po Valley (Fig. 17). Along the cross section AB (Fig. 20a and 20b), there was not much interaction between the upper-level and low-level forcing, as shown by the lack of link between PV’s. Moreover, the upward motion
near the surface was weak because of the cold and stable air mass. The lower level stable condition was shown by the sounding at Milan in Figs. 10e and 10f. The PV maximum region was located within 800 hPa and 650 hPa near the western Alps, which was mainly associated with elevated latent heating. Along the CD cross section (Figs. 20c and 20d), the low-level easterly flow was weaker and blocked to the east by the Apennines. There was a maximum PV at the lower level adjacent to the Apennines. After the PV streamer associated with the deep trough passed the western Alps, the mid-upper level PV was linked to the low-level PV near the Apennines. Thus, the terrain uplifting, low-level convergence upper-level divergence, and LLJ were able to work together to trigger the convective system near the Apennines region. Along the cross section NS, there was a PV maximum at 10/21/06Z below 900 hPa over the Ligurian Sea which was mainly due to the latent heat released over the ocean (Fig. 20e). When the upward motion increased in order to climb over the Ligurian Apennines, the PV also increased. After the airstream passed over the Ligurian Apennines, the vorticity increased on the leeside which also help the increase of PV. At 10/21/12Z, the depth of maximum PV over the ocean was increased associated with the unstable air, i.e. $\frac{\partial \psi}{\partial z} < 0$ (Fig. 17c).

### 4.3 Sensitivity Tests on Physical Parameterization Schemes

In this section, the first set of simulations aims at the performance of various cumulus and microphysical parameterization schemes associated with MAP IOP-8. A detailed description of the experiment design of each simulation was given in Chapter 2 and a summary can be found in table 1.

#### 4.3.1 Characteristics of MPS Precipitation

We compared three model results, CNTL, RESN and EXPT in order to evaluate the performance of LFO and Reisner schemes in producing rainfall associated with IOP-8. The
simulated results of surface pressure, surface wind, 300 hPa geopotential height, and 300 hPa wind fields from RESN (figures not shown) were almost identical to the control simulation (Figs. 3 and 4). Therefore, the synoptic structure was not sensitive to the change of microphysical scheme. However, the precipitation was significantly affected by the MPSs. The 6h simulated rainfall from RESN and EXPT were shown in Fig. 21.

The simulated rainfall of CNTL simulation with 5 km resolution was shown in Figs. 7a and 7b. Heavier rainfall was located in the vicinity of the southern Alps and Ligurian Apennines for both 10/21/06Z and 12Z. The maximum 6h accumulated rainfall amount at 10/21/06Z was 53 mm to the southeast of the Maritime Alps. Secondary maximum rainfall region was near the windward slope of Apennines. At 10/21/12Z, the maximum accumulated rainfall increased to 82 mm, which ranged over the southern slope of Alps and Apennines. The 6h simulated rainfall pattern of RESN simulation was similar to the CNTL but less in quantity. The maximum 6h accumulated rainfall at 10/21/06Z was 62 mm over the Ligurian Sea (Fig. 21a). At 10/21/12Z, the 6h accumulated rainfall increased to 73 mm near the windward slopes of Alps (Fig. 21b) and there was only 59 mm over the upslopes of Alps at 10/21/12Z. From rain gauge data shown in Fig. 7d, the heavier rainfall with 70 mm was near the foothills of the Alps. The maximum rainfall for RESN shifted to the Apennines surrounding area which was opposite to the observation. The observational rainfall amount over the southern slopes of Alps was larger than at the Apennines. Thus, the model appeared to underestimate the 6h accumulated rainfall over the southern Alps for RESN case to be 59 mm at 10/21/12Z.

The explicit simulation (EXPT) produced heaviest rainfall of all cases in the first set of simulations. At 10/21/06Z, the model predicted 6h accumulated rainfall already reached 100 mm over the western end of the Alps and 89 mm over the ocean (Fig. 21b). After the southerly flow (LLJ) gradually eroded the easterly flow and pushed near the Po Valley, the rainfall increased to 84 mm, which was compatible to CNTL run. However, the EXPT produced an excessive amount of precipitation over the ocean (about 120 mm) and Apennines (about 118 mm). Based on Wang and Seaman (1997), using explicit scheme alone tends to
overpredict in both cold and warm seasons rainfall events in the continental United States. Even though the EXPT simulation does pick up the rainfall at the central Po Valley, which is absent in the other simulations.

Since the mesoscale models use both MPS (grid-explicit scale) and CPS (subgrid scale) to produce rainfall. An appropriate partition between CPS and MPS is important for producing realistic precipitation in numerical models. To further illustrate the precipitation characteristics associated with different microphysical parameterization schemes, 6-h accumulated rainfall from CNTL and RESN were separated into CPS and MPS produced rainfall (Figs. 23 and 24). There was no CPS in for EXPT case, therefore the rainfall was purely from MPS. For CNTL simulation, most of the precipitation at the LMTA was generated by a grid explicit scheme (MPS) in both 10/21/06Z and 12Z while the subgrid scheme (CPS) only produced rainfall over the Ligurian Sea and Apennines.

Similarly, the most rainfall over the LMTA was produced via MPS for RESN simulation (Fig. 24b and 24d). The Reisner scheme generated relatively smaller rainfall amount compared with the LFO scheme in the CNTL simulation (Fig. 22). In order to further explore the difference of LFO and Reisner schemes, the model derived reflectivity for RESN and EXPT was shown in Fig. 24. The reflectivity of RESN and EXPT was weaker (about 30 dBZ) in comparison with observation and the CNTL simulation (Fig. 9). This may explain the weaker precipitation over the southern slopes of the Alps for RESN case. Still, the reflectivity showed a stratiform precipitating structure but there was no clear appearance of a bright band for either RESN or EXPT simulation. Figures 25a-d depict the particles types of snow, ice, and graupel from S-Pol radar, CNTL, RESN and EXPT simulations, respectively. Compared with frequency of occurrence of snow from observation (Fig. 25a, adopted from Medina and Houze, 2003), the CNTL simulation did a better job than RESN and EXPT in predicting the production of snow. The particles at upper-levels were evidently ice and the lower level was snow and graupel. The snow was above the freezing level in both simulations. The RESN simulation only had a small portion of snow particles above 4 km height and the predicted rainfall was relatively smaller. The EXPT also predicted snow and graupel near
the second peak of the Alps in this cross section. However, there was no graupel observed while both simulations predicted fair amount of graupel particles. This may explain the rainfall overprediction problem in LFO and Reisner schemes. Another reason for the lower amount of precipitation maybe due to the larger slope intercept for rain, snow and graupel in Reisner scheme \((RON = 1.E10, \text{ before was } RON = 8.E6, \text{ MM5-news, 2002})\) in MM5 Version 3.

### 4.3.2 Characteristics of CPS Precipitation

The different CPSs also have great impact in rainfall prediction. Most CPSs are designed to parameterize the convection system above 20 km. Especially the Kain-Fritsch and Grell schemes were tested for 20-25 km resolution (Wang and Seaman, 1997). There were still some localized forcing (\(\leq 5 \text{ km}\)) due to the complicated terrain. Therefore, in CNTL, both Grell and LFO schemes were turned on at 5 km grid resolution simulation and for BMKF simulation, the Kain-Fritsch and LFO were also turned on for 5 km resolution domain. Figures 22c and 22d were the BMKF simulated 6-h rainfall. The heavier precipitation at 10/21/06Z was near the southern Maritime Alps with the 6h accumulated rainfall 100 \(mm\) and only 35 \(mm\) in Apennines surrounding area. At 10/21/12Z, 6h accumulated rainfall increased to 70 \(mm\) at Apennines region and 90 \(mm\) near the foothill of the southern Alps.

Similar to previous comparison, we also broke up the CPS and MPS rainfall for BMKF case and compared with CNTL case (Fig. 22). As mentioned before, the majority of rainfall especially over the LMTA was produced by MPS for CNTL simulation. The CPS precipitation, however, was mainly located over the ocean and Ligurian Apennines. For the Grell scheme in CNTL simulation, the available buoyant energy becomes accessible, the CPS is activated. As revealed in the previous chapter, the cold and stable condition at the LMTA may not be able to fully activate the Grell scheme. Hence, there was lack of CPS rainfall in CNTL simulation. In addition, according to Wang and Seaman (1997), even Grell scheme perform well in predicting the life cyclone and the total volume of precipitation events, it tends to have low precipitation efficiency. Adjustment of this parameter can have a signifi-
cant increase in convective rainfall.

The BMKF, on the other hand, produced fair amount of rainfall via CPS (Fig. 27). The most precipitation generated by CPS was located over the Ligurian Sea, upwind slope of Apennines western Po Vally near the coast of the Adriatic Sea and the southern Alps. The MPS still contributed the most rainfall at LMTA. This is because the CPSs represent a more rapid upward transport of heat and moisture to adjust the environment. In order for CPSs to function properly, the different trigger functions were embedded in various CPSs based on different dynamic control assumptions. The convection is determined by CAPE in Kain-Fritsch scheme. As shown in Fig. 10e and 10f, there was no CAPE at Milan to fully activate the KF scheme. Hence, the CPS rainfall was relatively low compared with MPS at the southern slopes of the Alps.

4.4 Terrain Sensitivity Tests

As revealed in the previous section, the MAP IOP-8 has a stable environment. The baroclinic disturbance alone would not contribute to the rainfall event. The terrain not only acts as a permanent modifier to the incoming flow, but also provides lower level forcing to trigger the convection when the synoptic and mesoscale environments are suitable. In order to further illustrate how the topography affects the rainfall distribution, FTRN and NOAP were performed.

Based on numerical simulation of IOP-2B conducted by Stein (2002) with the surrounding mountains removed (Alps alone experiment), the accumulated rain is everywhere weaker than 4 mm. Notice that, in his simulation, the terrain resolution was relatively coarser. For IOP-8, the southerly LLJ was able to pass over the Ligurian Apennines and produced a significant amount of rainfall near the upwind slopes. The presence of the cold and stable air associated with easterly flow near the surface not only prevented the LLJ from directly impinging Alps at the low-level, but also enhanced the convergence near the northern coast of Ligurian Sea.
Thus, the small mountain like Apennines could easily trigger convection. In order to isolate the impact of Apennines for MAP IOP-8, the mountain within 44°N to 45°N and 9°E to 11°E (near the northern coast of Ligurian Sea) was removed.

Without the Ligurian Apennines, the wind speed of easterly flow was even stronger (Figs. 28a and 28b). There was a very clear boundary between easterly and southerly flows. Moreover, the easterly flow, which was blocked by the French Alps and turned to south afterward, was no longer confined in the narrow gap between Apennines and Maritime Alps. Instead, more broad blocking flow rushed out to Ligurian Sea (Fig. 28b). At 10/21/12Z, as the LLJ strengthened, it was able to push toward the north. Most of the rain poured upon the southern slopes of the Alps (Fig. 28b). The width of the easterly flow shrunk. Figures 28c and 28d showed the north-south cross section (Fig. 1b) of $\theta_e$, cloud boundaries and vector wind fields at 10/21/12Z. Without Apennines, the cold dome over the Po Valley spread to the coast of Ligurian Sea (Fig. 28c).

On the other hand, the LLJ directly intruded to the Po Valley without blocking by the Ligurian Apennines. Because the lower level forcing (Apennines) had been removed, there was very weak vertical motion over the ocean which resulted in the limitation of cloud development and convective activity near the Gulf of Genoa. Hence, the rainfall near the Apennines significantly reduced. At 10/21/12Z, the cold air pushed further south and the depth was also decreased. The shallow clouds developed above the boundary between the cold dome and the southerly flow. Even though there was still strong lower level convergence due to the confluence of southerly LLJ and deflected easterly flow (Figs. 29a and 29c), the low-level upward motion over the ocean was very weak. However, the 300 hPa divergence and 850 upward motion (Figs. 29b and 29d) were not sensitive to the Apennines. The presence of Ligurian Apennines proved to have a significant influence on triggering the convective systems near the Gulf of Genoa. Therefore, the rainfall distribution associated with MAP IOP-8 proved to be sensitive to the terrain.

With the feedback of the 1 km terrain resolution in FTRN, the terrain feature in the simulation was more detailed and the slope was steep. The general rainfall pattern of FTRN
was similar to CNTL but the 6h accumulated rainfall was higher at the upwind slopes of Apennines. The maximum 6-h rainfall went up to 55 mm near the Apennines at 10/21/06Z (Fig. 30a). At 10/21/12Z, the maximum 6h accumulated rainfall shifted to the windward slopes of Apennines with the amount of 96 mm. However, the precipitation near the southern Alps was about the same magnitude as the CNTL case. Because of the detailed terrain from a 1 km resolution, the terrain would provide more localized interaction and forcing for incoming flow.

Hence, the precipitation was responding to the upstream finer terrain resolution. Figures 30c and 30d are the north-south cross section denoted in Fig. 1b. Comparing Fig. 30c with CNTL run in Fig. 17c, the clouds developed at the upwind slopes of Apennines and extended above 600 hPa then linked with high clouds. From Fig. 30c, the cloud boundary extended to the upper level which indicated the convective systems developed near the upslopes of Apennines. The upward motion in front of Apennines was also stronger. However, the wind flow and rainfall distribution over the Po Valley and southern alps were not significantly affected by finer resolution of terrain.
CHAPTER 5

CONCLUSION

In this study, we have adopted a mesoscale model (Penn State/NCAR MM5) to simulate the synoptic and mesoscale environments conducive to orographic rainfall associated with Mesoscale Alpine Programme (MAP) Intensive Observation Period 8 (IOP-8) and performed model sensitivity tests on cumulus and microphysical parameterization schemes and terrain resolution.

A deep trough system associated with low-level jet approached the Lago Maggiore Target Area (LMTA) and a high pressure system located in the northern Europe was gradually established. This high pressure system remained about the same strength and traveled eastward very slow at about 2.78 $ms^{-1}$. It appears that the high-low pressure system became quasi-stationary during 10/20/12Z to 10/21/12Z. The southerly low-level jet (LLJ) advected warm and moist air up to the Po Valley and the southern Alpine slopes from the Mediterranean Sea. Meanwhile, a relatively cold and stable air near the surface which was associated with the easterly flow to the east of the Alps piled beneath. The easterly flow was split into two branches at the eastern tip of the Alps. Part of the easterly flow turned to southeasterly and passed over the Adriatic Sea. Another part of the southern branch kept moving along the foothills of the southern Alps and deflected to the Ligurian Sea between Maritime Alps and Ligurian Apennines. This deflected flow met the southerly LLJ near the Gulf of Genoa to help enhance the low-level convergence due to the barrier effect near the eastern Po Valley and Ligurian Apennines area. In fact, there was a mesoscale vortex formed by the
deflected flow and the southerly LLJ which also helped trigger the convective systems in this region. In this study, we further confirmed the idea proposed by Lin et al. (2001) that the orographic rainfall can be diagnosed by the general moisture flux and orographic moisture flux. Especially for orographic rainfall distribution, the orographic moisture flux reflected the rainfall distribution associated with the orographic uplifting, while the general moisture flux also reflected the rainfall distribution over the flat surface and ocean.

On the other hand, the cold and stable air associated with the easterly flow penetrated Po Valley near the surface and stabilized the low-level environment. Therefore, the upward motion was very weak and not able to produce convective systems. The southerly flow was able to pass over Apennines and produce orographic rainfall. However, there were only shallow clouds developed over the Po Valley. It was proved that the cold air serves as a cold dome to make the southerly LLJ easily override it. Both model simulations and radar echoes showed a stratiform cloud structure over the southern Alpine slopes.

Besides the low-level convergence, upper-level divergence also played some roles in orographic rainfall associated with IOP-8. The 300 hPa divergence field was approximately in phase with the low-level upward motion. It evidently manifested the mechanism of the low-level and upper-level coupling to trigger the convective systems over the Ligurian Apennines and its surrounding area. Moreover, the right entrance region of jet streak favored the divergence, which might help produce the thermally direct circulation and upward motion. Through the model integration period, the right entrance of jet streak was co-located with 300 hPa divergence field. The upper-level divergence and low-level upward motion were also well coupled. Consequently, the upper-level forcing helped trigger and enhance the convective systems over the Ligurian Sea and Ligurian Apennines. In addition, the west-east cross section, the mid and upper-level PV streamer was linked to the low-level PV streamer near the Apennines. Hence, the low-level and upper-level coupling did help produce the heavy rainfall over the ocean and Ligurian Apennines. However, this coupling forcing is weaker than that associated with IOP-2B (Lin 2003).

Based on model sensitivity tests on microphysical parameterization schemes, the Reis-
ner scheme tended to underpredict the rainfall over the southern Alpine slopes. We further confirmed the previous study (Wang and Seaman 1997) that the grid explicit scheme alone tended to overpredict the precipitation. Therefore, an appropriate partition between cumulus and microphysical parameterization schemes is important for producing the realistic precipitation. The control simulation predicted a reasonable amount of snow compared with observations. However, both the CNTL and RESN cases predicted too much graupel over the southern slopes of the Alps, which may help explain the overprediction problem for both LFO and Reisner schemes.

It appears that the Grell scheme is more suitable for the stable environment near the Lago Maggiore target area because it produced a reasonable amount of rainfall. As a result, most of the rainfall near the southern Alps was produced via microphysical parametrization schemes. On the other hand, the cumulus parameterization schemes produced a fair amount of precipitation over the ocean and northern coast of the Adriatic Sea. Similarly, the lack of CAPE over the Po Valley was not able to fully activate the Kain-Fritsch scheme.

Without Ligurian Apennines, the precipitation distribution was dramatically and significantly altered, the vertical motion over the ocean and Ligurian Sea was very weak without the low-level forcing (Ligurian Apennines). It resulted in limited cloud development. Still there were strong low-level convergence and upper level divergence fields over the Ligurian Sea but no convective activities shown and the rainfall amounts were significantly reduced. It was proved that the presence of upstream mountain, i.e. Ligurian Apennines, have great influence on the rainfall distribution associated with IOP-8.

Basically, the model simulation did a reasonable job in reproducing the synoptic and mesoscale environments and orographic rainfall associated with the MAP IOP-8. This study clearly illustrates how the synoptic and mesoscale systems worked together for IOP-8 case. On the other hand, there are still some important issues, that need to be further addressed, such as the formation of the cold and stable air mass from east. Moreover, we would like to improve our model initial condition by using higher resolution data, such as ECMWF 0.5° × 0.5° data for more accurate simulations.
CHAPTER 6 REFERENCE


Table 1: Summary of MM5 simulations

<table>
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<th>Grid size</th>
<th>CPSs</th>
<th>MPSs</th>
<th>Special Features</th>
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<td><strong>Set 1: Sensitivity tests of CPSs and MPSs</strong></td>
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<td>LFO</td>
<td>no Apennines</td>
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Fig. 1: (a) Alps topography and geographic map from model with 45 km grid resolution domain. (b) 15 km grid resolution domain used for two-way nested simulations. The AB, CD and NS denote the vertical cross section, while three dots indicate three sounding stations used in this paper. (c) Same as (a) but for 5 km resolution domain. The dot indicates the position of S-Pol radar station and the solid line denotes the vertical cross section used in this paper. The terrain shaded in 400 m interval.
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Fig. 7: Six-hour accumulated rainfall simulated by the model with 5 km resolution are shown in (a) and (b) panels. (c) and (d) were plotted from rain gauge data. The shading interval is 10 mm. The 1000 m terrain contour is denoted by the solid line. The surface wind vector wind field (1 full barb = 10 ms⁻¹) are also plotted in (a) and (b).
Fig. 8: MAP radar composite image from MAP Data Center in 3-hour interval.
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Fig. 15: The 15 km resolution simulated 1000 hPa convergence field superimposed on the upward motion ($\geq 2$ $cms^{-1}$ shaded) and horizontal vector wind fields (1 full barb = 10 $ms^{-1}$). The convergence contour interval is $8 \times 10^{-5}$ $s^{-1}$. The dash line is the 1000 m terrain height.
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Fig. 21: The six-hour accumulated rainfall from 5 km resolution simulation in 10 mm interval ended at 10/21/06Z and 12Z. The solid line is the 1000 m terrain contour. (a), (b) are for case RESN, and (c), (d) are for case EXPT.
Fig. 22: The 6-h accumulated precipitation produced by cumulus and microphysical parameterization schemes from CNTL simulation with 5 km resolution ended at 10/21/06Z and 12Z. The solid line is the 1000 m terrain contour.
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Fig. 25: (a) The particle types identified by polarimetric radar algorithms (adopted from Houze 2001). Particle types of snow, ice and graupel are calculated from the 5 km resolution simulation. The shading area is the temperature above 0°C in 2°C interval. The heavy solid line denotes the snow (0.4 $g�^{-1}$), the dot line is the cloud ice (0.1 $g�^{-1}$) and the dashed line denotes the graupel (0.04 $g�^{-1}$). Panels (b), (c) and (d) are for CNTL, RESN and EXPT simulations, respectively.
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Fig. 29: (a) The 1000 hPa convergence field superimposed the 1000 hPa upward motion (shaded $\geq 2\text{cms}^{-1}$) at 10/21/06Z, (b) same as (a) but at 10/21/12Z, (c) 300 hPa divergence (solid line) and convergence (dot line) fields superimposed on the 850 hPa upward motion shaded with upward motion $\geq 5\text{cms}^{-1}$, and (d) same as (c) but at 10/21/12Z. The interval for 1000 hPa convergence is $8 \times 10^{-5}\text{s}^{-1}$ and for 300 hPa convergence and divergence both are $6 \times 10^{-5}\text{s}^{-1}$. The heavy dash line denotes the 1000 m terrain contour.
Fig. 30: The simulation results of FTRN from 5 km resolution. (a) 6-h accumulated rainfall ended at 10/21/06Z, (b) 6-h accumulated rainfall ended at 10/21/12Z, (c) The north-south cross section (Fig. 1b) of $\theta_e$, cloud boundaries, and vector wind fields at 10/21/06Z. The shaded region depicted the $\theta_e \geq 322K$, and (d) Same as (c) except at 10/21/12Z.