Use of MM5 in 4DWX

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1. Introduction
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This document is intended to augment the material in the COMET NWP module. The module covers the basics of NWP models, physics in these models and limitations of such models. Herein we provide information, through the use of examples obtained from MM5 forecasts at each of the test ranges, that will assist the 4DWX user in interpreting MM5 forecasts while providing some important background information about mesoscale flows induced by terrain and land-surface property variation. In addition to working through the COMET module and reading the document below, the reader is strongly encouraged to read Chapters 10 and 11 of Whiteman (2000). These chapters contain an extremely well written and illustrated treatment of all types of terrain and diurnally induced circulations and will be useful for interpreting MM5 results in the context of local range topography.

2. Background of MM5 in 4DWX
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2.1 Model Computational Domain Configuration

The purpose of the MM5 in 4DWX is to provide forecast guidance for testing at ATEC ranges and represent the 3-dimensional, time varying atmospheric state adequately for post-test analysis. Most tests at these ranges are sensitive to the winds and stability within the planetary boundary layer. This means that, in addition to accurately predicting synoptic-scale and larger mesoscale motions, the MM5 model must account for local forcing, often manifested as motions induced by the coupling between complex terrain or land-sea contrasts and diurnal heating. The inclusion of high-resolution domains in the MM5 configuration compared to, say, the resolution used in the NCEP operational models (22 km and coarser) is predicated on the added predictive capability that is introduced because at least part of the forcing of small-scale motions is prescribed given terrain, surface condition, cloud cover and diurnal cycle. In places where surface forcing is weak (i.e. over the ocean or perhaps in certain regions where the topography is small and the surface condition nearly homogeneous), use of high resolution models without comparably high-resolution observations to initialize them is less beneficial.

Because of computational constraints (recall from the COMET module that a factor of 2 increase in horizontal resolution requires a factor of 8 increase in computing power), it is not possible to run high resolution (< 5 km grid spacing) everywhere. For instance, if we were to cover the entire U.S. with a 3 km domain, a 24-hour forecast would require about 2 months of continuous processing on the 8-processor Origin-2000 machines used in 4DWX. This does not even consider the enormous memory requirements to run such a simulation.

Because the emphasis is on local flows and the regions of interest tend to be geographically limited at any given time (i.e. the test ranges have well defined limits), there is no need to blanket the entire continent with high resolution. We employ grid nesting to achieve resolutions as fine as 1-3 km, yet remain able to integrate the models on high performance workstations or clusters of PCs in an acceptable time. The use of multiple nests allows one to transition gradually from the coarse resolution of the model that provides boundary conditions to the innermost region where high resolution is required (Figs. 1.1, 1.2 and 1.3).

Nests can either have prescribed boundaries or interact with their "mother" (coarser) domain (see COMET module for more details). In MM5, the resolution ratio between nests is 3:1. For instance, an interactive nest within a domain with 30 km grid spacing must have 10 km grid spacing. This number was chosen for MM5 because early tests showed optimal behavior when this 3:1 ration was imposed. Because the dimensions of each successive nest tend to be within a factor of 2 or so of the parent domain, there is a rapid reduction of the area covered within successive nests, giving the overall domain configuration a distinctive telescoping effect as shown in Figs. 1.1, 1.2 and 1.3 for DPG, WSMR and ATC, respectively.

The coarsest domain is often chosen to occupy a particularly large area because of the desire to move adverse effects of the lateral boundaries as far away from the areas of interest as possible. As detailed in Warner et al. (1997), lateral boundary conditions are often supplied by operational NWP models (Eta, AVN, MRF) and these typically have a time resolution of 6-12 h. This time resolution can be too coarse to allow correct propagation of features through the boundary, particularly upper-level vorticity extrema and attendant jet streaks.

In addition, differences in physics between the model providing boundary conditions and MM5 can yield spurious forcing on the boundary. For instance, MM5 in 4DWX uses a simple ice microphysics scheme where the mixing ratios of rain (snow) and cloud water (ice) are predicted when the temperature is above (below) freezing. The Eta model does not consider ice, hence is more likely to approach water saturation in the upper troposphere, supplying excessive water vapor which is quickly turned to ice by MM5 usually causing an overprediction of upper-tropospheric clouds. Another source of error is that cloud variables (cloud water or ice and rain water or snow) are not prescribed on lateral boundaries, hence are always initially zero within air flowing into the MM5 domain. Because it requires a few hours to fully spin up the cloud variables, there is an artificial lack of clouds and precipitation near the boundary which extends further into the domain near the inflow boundary than near the outflow boundary.

The size and location of the coarsest domain and for each of the ranges was determined roughly by the estimate that, in the cool season, when winds at the jet stream level average 20-30 m s\(^{-1}\) (more at ATC), air parcels move about 1000 km per 12 h. Synoptic-scale disturbances tend to move at about half this speed. This means that the upstream boundary should be a minimum of 1000 km away from the region of interest to avoid synoptic-scale features propagating from the upstream boundary to the region of interest during a 24-hour forecast. Thus, the distance to the upstream boundary is at least 1000 km in each of the domain configurations shown below. To avoid any air parcels moving from the boundary to the region of interest in 24 h, the upstream boundary would have to be at least 2000 km upstream and in some cases, 3000 km. Despite the fact that enlarging the coarsest domain is not computationally burdensome, the Eta model grid used as boundary conditions for MM5 is limited in its spatial extent and does not allow, for instance, a western boundary that is considerably further upstream than is shown in Figs. 1.1-1.3. Extensions of ATEC forecasts to 36 hours in 2000 have not revealed a sharp degradation beyond 24 h due to improperly specified lateral boundary conditions, but this issue requires daily monitoring.

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**DPG Domain Configuration**

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Figure 1.1 DPG Domain Configuration.

WSMR Domain Configuration

Figure 1.2 WSMR Domain Configuration.

ATC Domain Configuration

Figure 1.3 ATC Domain Configuration.

In each of the panels in Fig. 1, the terrain on domain 3 is contoured. Note that the other domains also have varying terrain (not shown), but since the resolution is coarser, fewer features are resolved. Where nest grid points coincide with mother domain grid points, the terrain must be identical. This means that the terrain field of a mother domain will be noisy in the overlap with the nest. This does not cause a problem because the solution in the nest overwrites the solution in the mother domain at each time step. However, if a nest is switched off during a simulation, the terrain on the mother domain is not altered and hence will retain the high-resolution terrain signature, possibly leading to undesirable noise.

The key difference between MM5 in 4DWX and the NCEP models is the high resolution terrain and land-surface fields, including the coastline definition. If the synoptic-scale flow is treated accurately, this means that the flow impinging on the terrain features will be well specified. The resulting effects of the terrain or land surface variation and diurnal heating can produce circulations which are not resolvable within NCEP models.

In most circumstances, there appears to be an optimal horizontal (and vertical) resolution such that the most important terrain or land-surface features are resolved but the forecast problem remains computationally tractable. For many areas, a horizontal grid spacing of less than 10 km is required to capture the important terrestrial features, sometimes as fine as 1 km in the case of small lakes, bays or isolated mountain peaks. An example motivating the use of a grid spacing for DPG is shown in Fig. 1.4. Here, an MM5 forecast of the salt breeze is shown for an innermost domain of 3.3 km (left or above) and 10 km (right or below). The 10 km solution has been interpolated to the 3.3 km domain for comparison. Note that the terrain is much better resolved at higher resolution to the point where valleys appear that are absent at coarser resolution. This is critical for simulating mountain valley circulations and flow blocking by terrain (to be discussed in section 3). Note that in some locations, the winds on the two domains are from opposing directions.

DPG Salt Breeze: Resolution Dependence

Figure 1.4. (a) Operational 10-h forecast initialized 1200 UTC 9 September, 1997; (b) same as (a) but with an innermost domain of 10 km grid spacing.

As with coarser resolution models, not all of the weather produced is correct and intelligent use of the model requires experience to decide what is probably realistic in addition to a physical understanding of the phenomena at scales much less than 100 km. The predictability of the smaller scale features is generally much less than that for synoptic-scale features. As a result, the total length of the model integration is less than that for the coarser Eta model (22 km grid spacing run for 60 h) and much less than for the AVN or MRF (roughly 60 km grid spacing run out to 10 days or more). Were it not for the influence of terrain and land-surface features, one could argue that 24-36 h integrations at 3.3
km resolution far exceeds the predictability of circulations at these scales. In fact, the small-scale wind circulations that are not part of the repeatable diurnally-forced and terrain-forced circulations are probably predictable for only a few hours. This limited predictability is part of the motivation for the real-time FDDA application being developed for 4DWX in which forecasts extend only 6 hours. The basic concepts of predictability imply that most features not tied to larger-scale features (e.g. fronts) or terrain features will not be predicted accurately in a 24-36 hour forecast. However, the model may produce highly realistic-looking features at these time ranges which may verify in a qualitative or statistical sense.

At small scales, the notion of geostrophic motion breaks down and quasi-geostrophic theory is rendered practically useless. The Coriolis force, generally responsible for adjusting motions toward a state of geostrophy, operates over the course of several hours. If we impose a mountain that is, say, 30 km wide and a wind that is 5 m s⁻¹, then it air flows past the mountain in only 1.5 hours. The flow never has a chance to become geostrophic as it is being altered by the mountain. As one moves from scales of 1000 km to less than 100 km, the pressure field becomes used increasingly as a diagnostic to assess flow acceleration (i.e. acceleration down the pressure gradient) rather than a streamfunction for the wind. Most forecasting techniques and experience rely, at least in part, on the notion of geostrophic balance. The lack thereof necessitates different approaches for diagnosing weather situations. The phenomena discussed in sections 3 and 4 serve as a set of mesoscale building blocks that can guide forecasters in their interpretation of high-resolution simulations.

2.2 Choice of Physics Options

As detailed in the COMET NWP module, there are numerous representations of physical processes in models that must be parameterized. This means that high complex processes, involving formation of precipitation, or molecular-level interaction of solar radiation with the atmosphere, or all unresolved scales of atmospheric turbulence must be represented in terms of tunable parameters and the variables predicted on the grid scale of the model.

Because computational speed and reliability are important factors in the design of MM5 for 4DWX, the choices for most schemes were typically of intermediate sophistication which required the least number of computations and were relatively applicable to all ranges. The most sophisticated scheme is the OSU Land Surface Model (LSM). This scheme represents the effects of varying surface characteristics and governs the amount of latent and sensible heat that can be transferred to the atmosphere as well as the amount of frictional drag on the near-surface flow. In general, the LSM, an addition to 4DWX MM5 in 1999, outperforms the simpler slab model that was used prior to the LSM because it:

1. Adjusts soil moisture during the forecast in accord with predicted precipitation.
2. Adjusts snow cover based on accumulating, sublimation and melting.
3. Incorporates an annual cycle of vegetation greenness (which improves treatment of evapotranspiration).

The LSM increases the run time of forecasts by 5-10%, but its added sophistication is deemed worth the additional processing time since knowledge near-surface conditions is so vital for ATEC testing. The LSM is initialized from output from the Eta model, which runs a nearly identical LSM. The user should note that while the Eta provides soil moisture information, there are no direct observations of this quantity that constrain the Eta model. Thus, while the input parameters to the LSM (wind, temperature, humidity, precipitation, etc.) are fairly well observed, there is room for error within the LSM itself.

Other physical parameterizations include radiation, cloud physics and boundary layer schemes. The radiation scheme is the Dudhia scheme which allows interaction of short and long-wave radiation with clouds. The scheme does not spectrally decompose the radiation further. Thus, it cannot handle situations where there is absorption or emission in selected wavelength bands.

Cloud physics is treated by a scheme which predicts a single variable for hydrometeors and a single variable for cloud particles. Rain water mixing ratio is predicted at temperatures above freezing and snow is predicted at subfreezing temperatures. The sudden freezing or melting at 0° can lead to abrupt heating or cooling which can produce some noise
around the freezing level. Perhaps a more important limitation of the scheme (and most other schemes) is that the vertical spacing of model levels is probably too coarse to properly resolve thin cloud layers. A spacing of 100 m or less is probably needed and this would require 5-6 times the run time of the present configuration at the ranges. Thus, thin cirrus, for example, is not well predicted. Cirrus forms in the model, but is usually optically thick whereas observed cirrus is often translucent, sometimes not even visible. Finally, the model also does not incorporate patchy cloudiness. The COMET module on cloud microphysics is an excellent source of more information on how cloud schemes work in NWP models.

The planetary boundary layer (PBL) is treated using a first-order closure scheme (the MRF PBL, adapted from the NCEP version). In first-order schemes, all turbulence characteristics are diagnosed from the grid-averaged state variables. As a result, most mixing of quantities is down-gradient, the effect being to remove sharp features in the vertical temperature and wind profile within the PBL. The nature of the mixing depends on the state or "regime" that the PBL is in. The regime is diagnosed from the bulk Richardson number (ratio of Brunt-vaisala frequency to vertical wind shear over the lowest few hundred meters). There are 4 regimes; highly stable, marginally stable, neutral (mechanically forced) and unstable (driven by surface heating). The mixing is prescribed based on regime and other factors such as surface roughness. Because the scheme only operates in a single column, it is possible for adjacent grid points to be in different regimes and have very different mixing. In some cases, this can lead to small-scale noise.

A unique feature of the MRF scheme is that in unstable conditions (daytime), explicit mixing occurs between the lowest model level and other levels within the PBL. That is, the potential temperature 5 layers above the ground can feel the surface temperature almost instantly. This is designed to mimic the effect of vigorous large eddies in the PBL. However, the mixing can become too vigorous and sometimes remove the daytime superadiabat near the ground. The MRF PBL is also known to have too much mixing in strong wind cases which can lead to an over-deepening of the PBL. For winds less than 10-15 m s\(^{-1}\), this is not a problem.

Again, the COMET module is very helpful to understand all the complex PBL processes that NWP models try to represent. Some additional points to keep in mind are that PBL errors in one region not only effect the diurnal cycle locally, but also affect local and regional wind circulations (see section 4). In addition, because PBL schemes are column-based, they assume all mixing is vertical. At high resolution (near 1 km), the lateral mixing through large eddies may become important.

2.3 Effect of Model Initialization

The NCEP Eta model serves as a first guess for the MM5 analyses that provide initial and boundary conditions for the model. A thorough explanation of how MM5 initial and boundary conditions are created can be found on the MM5 Web site (look for the tutorial class notes for MM5 Version 3 under Documentation).

Briefly, given a first guess from the Eta model, data are used to create an improved analysis for MM5. The scheme which analyzes the data is a univariate Cressman algorithm which analyzes analysis increments (first guess minus observations) interpolated to pressure surfaces. The fact that variables are analyzed independently means that a change in wind does not directly imply a change in temperature or pressure in the analysis unless there is a direct observation of such a change. Upon being analyzed, the analysis increments are added to the first guess and the result is vertically interpolated to the model terrain-following surfaces.

The NCEP Eta model is run 4 times per day, however, probably the best representation of the atmosphere is contained in the 00Z and 12Z runs because they are initialized with rawinsonde data. These two Eta runs are used in 4DWX. Time to completion of MM5 is critical for the 12Z forecast, so rather than wait for the 12Z Eta to finish, MM5 uses the 00Z Eta 12-h forecast as a first guess for conditions at 12Z, then uses soundings and surface observations to enhance the first guess. The enhancement from rawinsonde data is often crucial for obtaining a good forecast. Surface data, despite their frequency of reporting and relatively high spatial density, are generally not as important as rawinsonde data because they cannot give information about synoptic-scale features throughout the troposphere which determine much of the forcing for local-scale weather.

For the 00Z MM5 run (or 06Z run in the case of DPG), we wait for completion of the 00Z Eta before starting MM5. In
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In this section, we will provide examples and brief explanations for various terrain-induced phenomena that are observed at ATEC ranges. Nearly all the phenomena studied here have been given extensive treatment in various publications. Much of the following phenomenology is treated in greater detail in Whiteman (2000) or references therein.

3. Terrain Induced Flows

In nearly all attempts to forecast various mountain-induced phenomena, it is important to have a general idea of the vertical structure of the flow impinging on the mountains. Typically, this is done by examining soundings at some point upstream from the mountain. But, how far upstream should one look? How does one derive a relevant profile from the complicated flow that almost always exists even upstream from a mountain? The answers to these questions depend primarily on the horizontal scale of the orography considered and the character of the upstream orography. The answers also require some local experience, but here are a few guidelines.

1. The distance upstream should be roughly twice the width of the mountain or range of mountains under consideration. It should be in a flat area and not too near to any upstream mountains. The phrase "not too near" means a distance of several times the characteristic length scale of the upstream mountains.

2. The direction defining "upstream" is best given by the wind near the mountain top height, averaged spatially over
an area larger than the mountain itself.

3. Surface winds are seldom useful to indicate the speed of the incoming flow. Friction will make the winds weak and conditions may be calm late at night.

4. A "representative" sounding should be examined to determine the upstream conditions. The particular features in the sounding that are important will depend on the type of mountain-induced flow to be considered. The term "representative" here means not heavily influence by features with a scale similar to or less than the topography of interest.

3.2. Mountain Waves

Mountain waves are nearly always present near complex terrain, but tend to become more important in the lee of mountains as both the size of the mountain and speed of the incoming flow increase. It is common to have the strongest winds at the surface over and immediately leeward of the mountain and light surface winds on the upstream side. A mountain wave has a distinctive signal in the horizontal pressure gradient (high pressure upstream and low pressure downstream), vertical motion and sometimes cloudiness as well. Usually, there is strong subsidence in the lee with a tendency to remove all cloudiness in the middle and lower troposphere as a result. Upper-tropospheric clouds are often present, sometimes referred to as lenticular clouds because of their laminar, "lens-like" appearance. The appearance of the clouds will depend on the size and shape of the mountain inducing the wave, the amplitude of the wave and the relative humidity of the incoming flow.

The most common type of mountain wave that is resolved by mesoscale models is the hydrostatic mountain wave. These waves do not propagate downstream, but do propagate vertically, usually well into the stratosphere and even higher. The vertical wavelength of a mountain wave is dependent on the incoming flow and static stability and is roughly 6.3*U/N, where U is the wind speed approaching the mountain (the wind near mountain top height usually works well) and N is the Brunt Vaisala frequency averaged over the troposphere (usually about 0.01, less in summer and often more in winter). Because N varies less than U, the wavelength is mainly proportional to velocity. If U=20 m s\(^{-1}\), the wavelength is about 12.5 km. This means that, if there is downward motion at the surface in the lee of the mountain, there will be upward motion at a height of about 6.2 km (AGL) and downward motion at 12.5 km (typically in the stratosphere). The upward motion corresponds to the altitude of lenticular clouds. These clouds are important because they affect the surface temperature and often are not predicted by large scale models, but can appear in MM5.

An example of a hydrostatic mountain wave near DPG is shown in Fig. 3.1. Mountain waves are best seen through vertical cross sections of wind, vertical motion, potential temperature and relative humidity. This particular example is west-southwesterly flow over the Oquirr Mountains.

**Table 3.1**

| Figure 3.1. (a) Terrain and cross section location and (b) wind and potential temperature along cross section valid 2100 UTC 9 December obtained from RT-FDDA final analysis; (c) as in (b) but for wind and relative humidity and (d) as in (b) but for vertical velocity and wind normal to cross section (dashed for flow out of page). |

In Fig. 3.1b the winds are shown to be mainly southerly over the desert floor (center of domain), veering to westerly by 3 km MSL and increasing aloft. The westerly or west-southwesterly flow is oriented perpendicular to the Oquirr range, and therefore is the preferred direction to elicit the greatest wave response. The cross section of potential temperature (3.1.b) shows descending isentropes in the lee of the Oquirrs (their peaks reaching about 2.7 km MSL as represented in MM5) at low levels, transitioning to elevated isentropes at middle levels. Just upwind of the mountain, elevated isentropes at low levels transition to depressed isentropes in the middle troposphere. The depth of the cross section spans about one-half of a vertical wavelength, which we estimate to be about 10 km.
Corresponding to the areas of elevated isentropes, we expect to find clouds. The cross section of relative humidity shows nearly saturated air near the mountain top and also high humidity at about 5 km MSL in the lee. The large area of greater than 90% relative humidity at the right edge of the domain from 3-4 km MSL is not directly due to the Oquirr Mountains. The cloud pattern just associated with the Oquirr Mountains is a cap cloud over the summit and possibly a wave cloud emanating from the middle troposphere. The presence of either of these clouds depends mainly on the upstream relative humidity and amplitude of the mountain wave. In this case, a dry layer in the middle troposphere would make it more difficult for wave clouds to form.

The vertical motion within the wave (Fig. 3.1.d) indicates strong subsidence in the lee, responsible for the low-level drying there, and strong ascent in the middle troposphere above it, responsible for wave-induced lenticular clouds. While the model may have difficulty predicting these clouds explicitly (in the cloud water and cloud ice fields), the forecaster should be alerted to the possibility of their existence by the elevated humidity and overall mountain wave structure. Careful inspection of Fig. 3.1.d also suggests that the flow component in or out of the page can also be affected by the wave.

The example in Fig. 3.1 is of a wave of modest amplitude. Probably the most important effect of mountain waves is in the case that they reach large amplitude and create strong, even severe, winds in the lee. The conditions that determine whether strong winds occur at the surface are fairly subtle, but may be summarized as follows:

1. Strong winds near mountain top (>15 m s⁻¹).
2. A relative absence of forward vertical wind shear above mountain top. This means that the wind component at, say, 500 mb, in the direction of the flow at mountain top level should not be greater than about 1.5 times the mountain top flow. If the flow reverses direction with height above the mountain, this is particularly favorable for large amplitude waves.
3. A layer of enhanced stratification near or just above the mountain.
4. An absence of deep cold air near the surface in the lee.

Note that the example in Fig. 3.1 featured considerable shear through the lower and middle troposphere. Near the level of the mountain top, the shear is almost 10 m s⁻¹ per km. There is no unique definition of "strong" or "weak" shear, but practically, if the change in wind speed over half of the vertical wavelength of the mountain wave (2-7 km MSL in the example) is about as large as the average wind speed over the same depth, the shear will strongly diminish the amplitude of the wave.

Large-amplitude waves can also be discerned from x-y sections. An example is shown in Fig. 3.2, also taken from the DPG area, showing the signatures of a larger-amplitude mountain wave in the lee of the Deep Creek range. The geopotential heights at 700 mb (Fig. 3.2a) show a sharp deviation of the overall northwest-southeast orientation over and in the lee of the Deep Creek range. A weak ridge is evident on the upstream side of the range with a sharp trough immediately in the lee of the mountains. High humidity is evident over the summit and low humidity downstream. The high humidity indicates air that has risen and cooled while low humidity indicates air that has descended (Fig. 3.2b). The wind responds to the orographically imposed pressure gradient by accelerating into the trough, then decelerating back to its upstream value further downstream. The other "wiggles" in the 700 mb heights elsewhere are mostly smaller amplitude mountain wave signatures due to northwesterly flow impinging on smaller orographic features or passing over large features more nearly parallel to the major axis of the terrain feature, thus prompting a weak response. The mountain waves from other ranges are particularly well defined by low-level vertical velocity.

A word of caution is warranted when interpreting mountain-induced features at coarser (say 30 km) resolution. Similar patterns of low-level vertical velocity and height perturbation may be found in MM5 at 30 km resolution, but these are not indicative of a mountain wave and hence, will not have the vertical structure noted in Fig. 3.1. The ascent-descent couplet seen in Fig. 3.2b is present when air passes over terrain on nearly all scales. However, the response on scales of 100 km is inertia-gravity waves, which can propagate upward and downstream and hence affect a greater area. On still larger scales, the flow over the mountain is related to lee cyclogenesis, where the low pressure in the lee develops a
cyclonic circulation and an anticyclone forms over the mountain. As we move from mountains on scales of a few tens of km to mountains of 1000 km scale, the geopotential ridging switches from upstream to directly over the mountain and the lee troughing switches from being stationary in the lee to a feature that advects downstream. The winds change from blowing across the height contours to (at least above the PBL) blowing along the height contours. In general, a complex massif like the Rocky Mountains projects onto a range of scales, and all these motions are present simultaneously. Model grids of a given resolution tend to emphasize only one of these components at a time, so examination of all the domains in 4DWX MM5 is important to comprehend the full effect of terrain.

Mountain Wave: Plan View

Figure 3.2. (a) 700 mb wind, geopotential heights and relative humidity for 0600 UTC 11 December and (b) vertical velocity at 600 m for the same time as in (a).

It is often useful to conceptually relate the synoptic-scale flow pattern to the occurrence of mountain waves capable of producing severe winds. Often, these waves are tied to the presence of jet streaks aloft and near-surface low-pressure systems. For mountain ranges aligned north-to-south, the strongest waves typically occur when the surface low center is directly to the north of the mountain. This places the mountain range within the strong westerlies at low levels to the south of the low. The favored location is often near the right exit region of the jet streak, where subsidence occurs. The exact role of subsidence is unclear, but the most likely possibility is that subsidence creates an inversion in the middle troposphere, often colocated with a descending upper-level front. From the criteria above, this enhanced static stability within the inversion is favorable for waves to become large amplitude at low levels.

As a low pressure area moves to the northeast, there is a synoptic-scale pressure gradient force directed eastward across the mountains. Although conditions are often less favorable for mountain waves, due to increased vertical shear and lower static stability, the addition of synoptic and mountain-induced pressure gradients often produces widespread strong winds that are still greatest immediately in the lee of a mountain.

3.3. Flow blocking and lee vortices

A common mesoscale phenomenon near tall and steep orography characteristic of most of the western U.S. and even parts of the Appalachians is flow blocking. In general, air impinging on mountains has insufficient kinetic energy to overcome the negative buoyancy that arises through adiabatic cooling in a stably stratified atmosphere. Thus, instead of flowing over the mountain crest, air is forced around the mountain. A parameter which characterizes the flow impinging on a mountain is the Froude number $U/(NH)$, where $U$ is the strength of the wind (usually the average wind speed between the ground and mountain top, $H$ is the mountain height and $N$ is the Brunt Vaisala frequency. Thus, if the wind is weak, the mountain tall and the air strongly stratified, the Froude number will be small. If the Froude number (Fr) is less than about 0.7 (in essence, the definition of small), upwind stagnation occurs. One expects that the Froude number will generally be small with respect to most the mountain ranges near the ATEC ranges unless daytime boundary layer over the lower elevations has reached a depth comparable to the mountain height.

For a mountain that is fairly symmetric, meaning that, following the incoming flow, the mountain shape to the right of the crest is not too different that the shape to the left, there will be a region in line with the crest where the flow stagnates. Air to the left flows around the left side of the mountain; air to the right flows around the right side (Fig. 3.3). Air approaching a mountain at somewhat higher levels sees an effectively smaller obstacle and therefore is more likely to pass over it. Thus, there is a location on the upwind slope which separates air flowing over versus around the mountain.

Flow Stagnation and Lee Vortices
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Figure 3.3 Schematic of circulation around an isolated, nearly bell-shaped mountain in the case that the Froude number is small (after Whiteman, 2000).

Low-level air passing over the flanks of the mountain tends to accelerate. Horizontal vortices form on each flank of the mountain, cyclonic on the right and anticyclonic on the left. There are several distinct physical mechanisms which lead to the production of these circulations. In a frictional boundary layer, air near the surface is decelerated. If the surface is sloping, then there will be decelerated air near the surface but unaffected air at the same height further away from the mountain. The implication is a vertical component of relative vorticity, positive on the right and negative on the left. Another mechanism is vortex tilting wherein the subsidence in the lee of the mountain tilts vorticity that lay in the horizontal plane. Yet another mechanism invokes dissipation of the flow leeward from the peak which mechanically forces vorticity on the flanks of the obstacle. Regardless of the mechanism responsible, which is an ongoing research topic, the pattern in Fig. 3.3 is robust and nearly steady state.

Flow Blocking and Terrain Trapping

Figure 3.4 (a) Domain 1 surface wind barbs and temperature field; (b) Domain 2 surface wind barbs and temperature field.

Figure 3.4 (ctnd). (c) Domain 3 surface wind barbs and temperature field; (d) SAMS observations for 2040 UTC 1 December.

3.4 Gap Flows

In most areas of complex terrain there are well defined breaks in the relief which may be high passes or impressive valleys which cut through mountain range. The wind in these valleys is often substantially different than the wind over the adjacent peaks and even over the lower elevations away from the mountains. Winds in such gaps can become exceedingly strong. In fact, some of the strongest winds ever "recorded" are gap winds near the Taku Strait near Juneau, Alaska. (Actually, the wind speeds are not well known, due to destruction of the anemometers during events; speeds may exceed 100 m s⁻¹).

Strong gap winds result from the accumulation of a pressure gradient along the gap, often associated with a strong low-level temperature contrast. This is common near coastal mountain ranges in winter where cold air over the continent contrasts with mild marine air. High mountains block the bulk movement of the cold airmass, thus forcing it to flow through gaps. Air flowing down the pressure gradient accelerates, offset only by surface friction. Without friction, it is simple to estimate the upper bound on the strength of the wind at the outflow end of the gap. The kinetic energy change is just equal to the difference in pressure from one of the gap to the other, \( E = ru^2/2 = p_2 - p_1 \), where \( r \) is density, \( u \) is wind speed and \( p_2 \) is the pressure at the upwind end of the gap and \( p_1 \) is the pressure at the exit of the gap. If we assume a density of 1.2 kg/m³ and a pressure difference of 20 mb (2000 Pa), this yields about 60 m s⁻¹. Note that the length of the gap is not very important for determining the ultimate acceleration; it is only the pressure difference from one end to the other.

Most gaps are not flat, but are mountain passes where there is still a substantial elevation rise and fall along the gap. Flow over a pass can excite a mountain wave which can strengthen the surface winds near the exit of the gap. The extreme wind events in places like the Taku Strait can be aided by acceleration over higher elevations to the east.

Most gaps flows are not severe winds, but winds which nonetheless depart significantly from the surrounding flow. An example from the RT-FDDA system shows gap flow over relatively low elevation to the south of the Deep Creek range.
in Utah (Fig. 3.5). Note that there is an acceleration down a pressure gradient directed northwest to southeast. Note, too the tendency for the flow to "fan out" upon exiting the gap. This is a common aspect of gap flows.

**Gap Flow Over Snake Valley, Utah**

Figure 3.5 Example of surface winds over Snake Valley Utah. At left are surface streamlines and topography showing the northwesterly gap flow over lower elevations; at right is the 850 mb map of geopotential height, wind speed and wind barbs. Contour interval for height is 5 m. Light blue indicates winds less than 5 m s^{-1}. Medium blue indicates winds greater than 5 m s^{-1}.

### 4. Diurnally Forced Flows

In this section is summarized a variety of mesoscale circulations which arise from the horizontal contrast in surface heating or cooling that derives mainly from the solar cycle, but can also occur in response to differential heat fluxes arising from varying surface thermal characteristics. In the former category are sea breezes, the "salt breeze" and mountain-valley breezes. In the latter category is flow over localized warm water, such as results in autumn and winter when cold air passes over Great Salt lake or Chesapeake Bay.

#### 4.1. Mountain-Valley Circulation

The mountain-valley breeze is fundamentally tied to terrain slope and the diurnal cycle. Fig. 4.1 is a schematic which shows how it begins.

**Schematic of Mountain-Valley Breeze**

Figure 4.1. Schematic of mountain-valley breeze. Left panel denote daytime conditions, with heating making the near-surface air warmer than air at the same height away from the terrain. Right panel is for nighttime when the opposite occurs. The daytime upslope is typically 100-200 m deep whereas the nighttime drainage is often less than 50 m deep.

Daytime surface heating creates warmer air along the slope than air at the same altitude but further from the surface. This horizontal temperature contrast is associated with lower pressure near the terrain. Flow accelerates toward the terrain and is forced up the slope. Above the mountain top level there is a reversal of the flow. The circulation loop is closed by subsidence, which can be very weak, occurring to the right of the upslope in the diagram.

At night, the air adjacent to the slope becomes cooler than the air at the same altitude further from the slope. This drives downslope flow adjacent to the terrain. This current is extremely shallow, sometimes only 20-40 m deep. Because the lowest model layer in the terrain-following coordinates of 4DWX MM5 is at about 40 m AGL, the details of this drainage flow are not well resolved. Nevertheless, the model does produce the gross character of the daytime upslope and nocturnal drainage flows as shown in the example below.

**Mountain-Valley Circulation at Dugway**

Figure 4.2. Domain 3 surface streamlines for (left or above) 2000 UTC 5 December 2000 and (right or below) 1200 UTC 6 December. Also shown is terrain.

In Fig. 4.2a, representing daytime, the main signals are divergent flow over the salt flats (low lying area encompassing the northwest quadrant of the domain) and convergence lines atop each of the ridge lines. This is particularly pronounced for the three north-south oriented ridges in the northeastern part of the domain (Cedars, Oquirrs and Stansburys). The actual wind velocities are only 2-3 m s\(^{-1}\). Note that there are smaller terrain features which do not show convergence at the crest. For smaller ridges, the divergent flow off the salt flats passes over the feature. Consistent with arguments about flow blocking being proportional to the Froude number (Sec. 3.2), small terrain features which do not penetrate through the top of the daytime mixed layer are embedded entirely within a layer of zero stratification, hence there is no resistance to air ascending their slopes. The imposed flow off the salt flats easily passes over the smaller terrain features and disrupts the symmetric upslope pattern that would otherwise occur. The maximum convergence is typically found downstream from the ridge instead of atop the ridge as in the case of mountains tall enough to extend above the PBL.

At night (Fig. 4.2b) the pattern is nearly reversed. Note there is divergence from the ridge crests with flow convergence into the valleys. Near the Deep Creek mountains (left center of figure) blocking of the northwesterly flow is evident with flow on the east side veering to easterly, setting up a convergence boundary with the drainage northwesterlies off the mountain peak. The presence of a weak northerly synoptic-scale flow affects the orientation of the drainage flow, and its interaction with the mountains results in local flow blocking on the north side of mountain ranges and acceleration on the sides (provided the valley is relatively wide, such as to the west of the Cedar mountains).

In the instance where topography is isolated and simple in shape, the slope flows are the dominant response to diurnal heating. However, valleys are often present and a slope to the valley floor orthogonal to the slope of the sidewalls (which evince the slope flow response to diurnal heating) leads to up- and down-valley breezes. Often the slope of the valley floor is smaller than that of the slopes on the side, but this sloping surface typically maintains a greater length scale than do the sides. Thus, the overall magnitude of the up-valley daytime and down-valley nighttime wind is similar in magnitude to the slope flows on the sides of the valley. The confluence resulting from superposing the nighttime downslope and down-valley winds can give the impression of flow channeling, even though the acceleration due to constriction is not important.

Finally, there is a component to the mountain valley wind system forced by differential heating across the valley. This usually occurs when solar radiation is incident on one side of a valley, but the other is in shadow. Flow is directed across the valley toward the side in sunlight. This circulation is important only for fairly narrow valleys, typically not more than a few km wide. MM5 currently has no ability to capture this circulation because such valleys are marginally resolved and because the model does not account for shadowing effects due to terrain. This latter limitation is particularly important when one is concerned about the timing of slope-flow reversal near sunrise and sunset. For slopes that face west (east), MM5 will initiate daytime upslope too early (late). Nocturnal drainage flow tends to initiate too early in MM5 for all terrain orientations due to cessation of vertical mixing too soon after sunset.

4.2. Deep Convection

Typical summertime soundings exhibit convective potential provided that air can be lifted to its level of free convection. Often, a capping inversion is present above the boundary layer, thus requiring the air to be lifted adiabatically to become positively buoyant. The mountain-valley circulation is an important and pervasive mechanism which can provide this lifting and explain why convection tends to trigger first on the highest peaks. An example of the preference for convection forming over elevated terrain is shown in Fig. 4.3. This rendering from Vis-5d shows a strong preference for initiation to the east of White Sands over the Sacramento Mountains. The height combined with the size of mountain range (over 100 km in the north-south direction) make it the dominant terrain feature in the area.

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**Terrain-induced Convection over WSMR**

Figure 4.3. Vis-5d rendering of surfaces of constant rain water mixing ratio (red) and cloud water and ice (white) as
well as wind at $z=2.5$ km (MSL).

As seen in Fig. 4.3, the 3.3 km resolution domain is capable of producing deep convection (i.e. no cumulus parameterization is needed; see COMET module for more details of these schemes). However, the structure does not closely resemble that of individual cumulonimbus towers. At this horizontal resolution, MM5 is generally not capable of producing the variability that occurs in real convection. This overly-smoothed representation of convective cells leads to a distribution of precipitation that is smoother than the observed distribution (Fig. 4.4).

Convective Precipitation Forecast

Figure 4.4. (a) MM5 domain 3 and (b) observed (radar-derived) precipitation for a 24-h period 1200 UTC 12 August, 1997 to 1200 UTC 30 August, 1997. Values are in mm.

The challenge to forecasters attempting to use the MM5 forecast of convection results from the high temporal and spatial variability of the rainfall and the fact that the details are probably not predictable. However, there are some useful aspects of the forecast that could provide useful guidance

1. The timing of precipitation in the diurnal cycle is roughly correct. This tends to be generally true, especially of terrain-forced convection.

2. Heavy precipitation is predicted in the correct quadrant of the domain, i.e. near the southeastern part of WSMR. However, the details are wrong.

3. MM5 predicts widespread precipitation areas with more than 40 mm of rain, but this is not observed. There are possible problems with radar coverage, but overprediction of maximum precipitation amounts commonly occurs at this horizontal resolution.

4. There is little precipitation in the northern part of the range where some convection was observed. It is possible that the overprediction to the southeast affected precipitation further north. This kind of compensation is often seen.

4.3. Breezes Generated by Surface Heating Contrasts

A general class of mesoscale flows is driven by horizontal density contrasts arising from horizontal variations in surface heating. The most well known example is that of the sea or lake breeze. For an equal amount of solar radiation incident on land and water, there will be both sensible heating of the ground and evaporation of water from the top layer of soil. On a water surface the solar radiation produces almost no temperature change due to several factors: the large heat capacity and thermal inertia of water compared to land surface constituents, partial transmission of radiation through the water surface and absorption below and high albedo of water surfaces. This difference between the nature of solar radiation absorption and heating leads to differences in density between near surface air over land and over water. If the soil is particularly dry and dark, the temperature contrast with the surface layer above water is enhanced.

The atmospheric response to this density contrast is the accumulation of a surface pressure gradient force directed toward the land. Higher pressure over the water is associated with denser air. In most cases, density translates to temperature, however, the more accurate relationship is between density and virtual temperature. At the same temperature, moist air is less dense (higher virtual temperature) than dry air. In most situations the temperature gradients are large enough to dominate the contributions to density from moisture gradients. However, in extreme cases, variations in water vapor can represent an effective temperature contrast of 1-2°C.

The response to the horizontal pressure gradient is an acceleration of near-surface air toward the land. A return current exists aloft, often weaker and more spread out in the vertical. Typically, the entire circulation resides within the lowest 2-3 km above the ground. Especially when weak offshore flow precedes the development of the sea breeze, the leading
edge of the onshore flow can take on characteristics of a sharp cold front. The propagation and fine-scale structure of this front are well described as a density current, a shallow mass of cool (dense) air spreading under the influence of gravity. As the front propagates inland, it tends to weaken rapidly because of surface sensible heating over land behind the front. Clouds may be produced atop the updraft at the leading edge and in environments with conditional instability, deep convection can occur at the leading edge.

An example of a sea breeze (technically a bay breeze) emanating from the Chesapeake and Delaware Bays near ATC is shown in Fig. 4.5. There are three water bodies that force the generation of the sea breeze. Not apparent is the larger-scale affect of the Atlantic Ocean which forces a general southeasterly breeze to develop and propagate northward within each of the bays. Superposed on this are the individual breezes from each bay. Note the generally divergent character of the winds over the bay as cool air spreads laterally. Also note that the presence of a westerly ambient wind causes an asymmetry in the intensity of the convergence at the leading edge of the bay breezes, being stronger on the west (upwind) side of the bays and weaker on the east (downwind) sides. Thus, only on the upwind side of the bay is the onset of the sea breeze marked by a strong frontal passage. Note also that the northern end of Delaware Bay produces no bay breeze. This is probably because it is not well resolved by the model, but bodies of water that are less than a few kilometers wide seldom generate any appreciable sea or bay breeze.

**Bay Breezes Near ATC**

![Figure 4.5. Surface wind vectors and temperature from May 29, 1999 at 2200 UTC (1700 LST). Water is colored blue and elevated terrain appears red or pink. The countour interval is 1°C.](http://www2.mmm.ucar.edu/people/davis/files/4dwx_mm5.html)

The sea/lake breeze is influenced by the Coriolis force (at mid-latitudes) because it typically persists for several hours. The clockwise turning, which tends to make the flow parallel to the coast, is balanced by the continuing thermal forcing, which makes the flow onshore. The result can be a nearly steady state with the wind direction determined by the relative importance of Coriolis and thermal forcing. If the shoreline is oriented north-south, the steady-state breeze will usually be southeasterly. After the differential heating is removed, the sea breeze decays. When the land surface becomes cooler than the water, the flow becomes offshore. Typically, the offshore flow is weaker than the daytime onshore flow because the depth of cool air over land at night is shallower, owing to the shallowness of the radiatively cooled nocturnal boundary layer. Because the layer of cool air over the land is shallow compared to the depth of the cool air over water during the daytime, the pressure contrast between land and water will be relatively smaller, hence the land breeze is weaker than the seabreeze and seldom develops frontal characteristics.

Coastline shape is an important factor for determining preferential locations for convergence or divergence within the PBL when the land-sea breeze circulation exists. This, in turn can determine whether areas of low clouds and fog are preferentially thicker or thinner at different places along the coast and whether drizzle may form. Areas where coastlines are convex will have convergence over land during the day and divergence at night. Convergence over land may take the dramatic form of colliding sea breezes which can spawn deep convection, or more benign lifting which can initiate fog and drizzle. Concave coastlines (bays or lakes) will feature convergence over the water at night and divergence during the day. Highly concave coastlines near warm seas and warm inland lakes can occasionally focus nocturnal convection as land breezes from opposing shores converge over the water.

Over land, variations in surface characteristics can create differential heating and density-driven circulations that resemble the sea breeze. Typically, enhanced soil moisture content or higher surface albedo are responsible for locally reducing the sensible heating of the ground relative to the surroundings (see COMET module for more on land-surface physics). Higher pressure and cooler temperatures occur over moist or highly reflective land surfaces. Air flows from these toward drier or darker areas, creating horizontal convergence in the PBL and lifting on the warm side of the thermal contrast.

A particularly interesting example of the effect of varying land surfaces is found near the interface between the salt flats (playa) and desert in Utah (near DPG). Playa usually has higher moisture content, higher albedo and greater thermal
inertia (related to heat capacity) than the desert. As a result, playa warms more slowly during the day and cools more slowly at night than the desert. The thermal contrasts result in the "salt breeze", toward the desert during the afternoon and toward the salt flats at night.

5. Interpretation and Systematic Errors

Throughout this document, examples of various phenomena have been presented, many of which were taken directly from MM5 forecasts. The structure and behavior of the features agree quantitatively with what one expects from theory, however, the forecaster is presented with the challenge of deciding whether the feature present in a forecast exists in reality and whether it is quantitatively represented correctly. In general this challenge is made tractible through experience with using the model on a daily basis. However, there are a few general aspects about MM5 (and perhaps other models) that can assist the forecaster in interpreting model output.

1. The variability in the atmosphere is greater than in MM5.

Despite having high-resolution terrain and land use information, MM5 systematically underestimates the spatial and temporal variability observed in surface and upper air data at ranges. Even at horizontal resolutions of 1-3 km, there is a continuum of atmospheric motions at smaller scales that cannot be represented by the model. Furthermore, the model initial conditions are probably able to represent scales of motion no smaller than 100 km owing to sparse data coverage.

The lack of variability in MM5 means, for instance, that features appearing the model should be interpreted as time average behavior over time intervals that depend on the grid resolution. At 3 km resolution, the model is not able to resolve variability on time scales much less than 10 minutes or so; at 30 km, the effective time resolution is perhaps an hour. These time scales are considerably greater than the minimum resolvable time scales which are an order of magnitude shorter. These estimates refer to the variability that is actually present in the model simulations. In particular, these limitations imply that the high frequency variations in time series at selected points will not be captured by MM5.

As we have already noted (and as is discussed in detail in the COMET module), the effective horizontal scale resolved by the model is several grid lengths, or perhaps 15 km on domain 3 at ATEC ranges. Thus, differences occurring between stations separated by this distance or less will not appear in MM5. In practice, unless there are terrain or land-surface variations near the grid scale of the model, the effective minimum scale of PBL-forced features in the model will be considerably greater than 10 km.

2. Solutions near domain boundaries should be interpreted cautiously.

As discussed above and in the COMET module, there are two types of lateral boundary conditions; prescribed (time varying) and interactive. The latter is used on boundaries separating nests from the mother domain. Prescribed boundaries are used on the coarse domain (30 km in 4DWX). Within a band 5 grid points wide along the perimeter of the coarse domain, the solution is adjusted from the boundary condition to the interior solution. In this transition region, the model solution is often noisy and can be unrepresentative of the interior solution. This happens more often at the outflow boundary and especially when the model or analysis providing the boundary conditions implies a different solution than MM5. Near the boundaries of a nest, there can also be some inconsistencies between nest and mother domains which reveal themselves as large gradients or can result in a persistent anomalous forcing of any of the primary variables. In general, within 3-4 grid points of any boundary the solution should not be trusted. Near inflow boundaries, incorrect specification of boundary conditions results in errors which can be advected into the interior. Errors in upper level moisture are particularly problematic because the model dynamics are not able to compensate for such errors as might occur with temperature or wind, and because upper level winds are strong enough to transport the error 1000-2000 km into the domain in just 24 h.

3. MM5 predictive skill decreases with time, especially above the PBL.
This is a general feature of atmospheric prediction for all models on all scales. Curiously, during times of boundary layer decoupling at night, the predictive skill of MM5 can increase to the extent that local flows dominate and the model skillfully predicts these flows. In general, error growth rates are scale dependent, being larger for smaller-scale flows. Error doubling times for situations dominated by convection are as little as 30 minutes (the lifetime of a thunderstorm). For situations governed by synoptic-scales, the error doubling time is about 2 days. Elements of both rapid growth on small scales and slower error growth on large scales can be seen in MM5 ATEC forecasts. Boundary-layer features on scales of 10-20 km seem to have prediction errors that become large on time scales of 1-2 hours whereas upper-air data show error growth of perhaps 30-50% in 24-36 hours. Note that even if small-scale errors grow fast, it does not mean that the entire forecast deteriorates quickly. The overall error growth depends on the relative amplitude of smaller and larger-scale features. For instance, situations with important synoptic-scale features are well-known to be more predictable that situations where synoptic-scale forcing is weak, unless there are well defined terrain or land-surface features to focus diurnal circulations.

Predictability limits also affect the way in which MM5 output is interpreted and used. Features of only a few tens of kilometers in scale in a 24-36 hour forecast should not be interpreted literally unless they are tied to a dominant terrain or land-surface feature. However, smaller-scale features appearing in models are often suggestive that such features will exist and could be verified in a statistical sense given a sufficient number of similar cases. Limits of predictability mean that forecasters must engage in probabilistic thinking even if the ultimate goal is a decision-oriented product.

4. MM5 physics is a columnar, parameterization of processes in which lateral communication between grid points does not occur.

The radiation, cloud and boundary layer physics in MM5 operate in a single column of grid points. Communication from one grid point to another in the model is handled entirely by horizontal advection. For coarse resolution grids, this treatment is justified, but as one approaches 1 km grid spacing or less, these implicit assumptions of columnar physics start to break down. Below are listed two sources of error attributable to columnar physics.

- Explicit lateral mixing is not considered. In the PBL, for instance, all mixing occurs within a column. Over steeply sloping terrain, this assumption is particularly poor, but even over flat ground, horizontal mixing becomes important as the horizontal grid spacing approaches the PBL depth.

- Radiative transfer is assumed to occur entirely within a column. Shadowing effects are not considered. This includes shadowing by fixed objects (mountains) and by clouds when the clouds are not directly overhead. As discussed in Sec. 4, this eliminates cross-valley circulations driven by differential heating resulting from shadowing. It also produces biases of temperature changes at some stations near sunrise or sunset.

5. The vertical resolution in MM5 does not allow proper treatment of thin cloud layers.

Many cloud layers are only a few hundred meters thick. Typical vertical resolution of MM5 above the PBL is several hundred meters and is more than 500 meters in the upper troposphere. Because there is no fractional cloudiness parameterization in MM5, the implied thickness of clouds in MM5 is at least the thickness of the model layer. Thus, clouds in MM5 tend to be too thick optically. Also, MM5 does not allow patchy cloudiness in the horizontal; a grid cell is either saturated or it is not. Some new schemes do allow this, and therefore better handle solar radiation passing through these patchy layers, but such schemes are not used currently in 4DWX due to their greater computational cost and general lack of performance validation. The limitations of the scheme used in 4DWX MM5 mean that when thin cirrus is observed, for instance, MM5 is likely to treat the surface effects quantitatively incorrectly.

6. The nocturnal boundary layer transition in MM5 is too abrupt.

After sunset, there is often an elevated mixed layer that persists well into the evening and delays the decoupling of the surface layer from the interior. Although not fully understood, it appears that MM5 decays the mixing too
quickly after sunset and does not maintain a mixed layer into the evening. This results in a rapid decoupling of the surface layer, a cold bias and an early onset of drainage winds compared to reality. The problem is likely due to the MRF PBL scheme, which uses explicit mixing when the whole PBL is unstable, but this shuts off when the surface layer starts to cool regardless of the true vigor of mixing aloft.

There are other errors that can occur in the model, but these are more flow dependent and vary by location, hence, there are no general statements to be made. Forecaster experience is required to divulge these errors and facets of interpreting high-resolution forecasts.

From the examples presented in preceding sections, it may be apparent that the tools needed to examine high-resolution forecasts differ considerably from those used to examine coarser, mainly synoptic-scale, forecasts. One set of tools is phenomenological understanding, but in terms of the set of variables best suited for diagnosing such phenomena within MM5, we can make some recommendations. Some of these tools are available in UGUI or in the visualization package for RT-FDDA, but others are not.

1. Surface Potential Temperature

Although temperature is often plotted as a surface thermodynamic variable, potential temperature is often more useful in areas of complex terrain because it is unchanged by upslope and downslope motions. Even when there is daytime heating, because potential temperature is also vertically well mixed, it clearly shows horizontal density contrasts in the presence of complex topography.

2. Vertical cross sections (wind, potential temperature, relative humidity and vertical velocity).

Such cross sections are important for diagnosing the static stability of air approaching mountains, likely presence of clouds (humidity is important to view in conjunction with cloud water or cloud ice because if the atmosphere is subsaturated, it indicates whether the atmosphere is close to producing clouds), and vertical shear profiles. Skew-Ts can also be used to assess vertical profiles, but a cross section aligned along the flow is less sensitive to point-to-point variations.


4. Terrain and land-use fields.

It is critical to be familiar with the model-represented terrain and land-use fields and how they differ from reality. Terrain in MM5 will typically underestimate the true maximum elevations, and this can be important for determining whether air goes over or around the terrain feature.

5. Geopotential height

On coarse domains, the uses of this field are traditionally to diagnose synoptic-scale dynamics. On fine grids, this field is still useful provided that (a) one thinks more in terms of accelerations down height gradients; (b) one chooses a pressure surface near the average pressure at the surface in the region of interest (for instance, 850 mb works well at DPG); and (c) one looks at the highest resolution domain available for a particular area (coarse-domain heights over nest areas can look quite noisy).

6. Summary

This document has sought to provide additional information for users of MM5 in 4DWX beyond that found on the COMET NWP web site, on the MM5 web site and in the references listed at the bottom. The reader is strongly encouraged to become familiar with the salient parts of these guides. This will help in utilizing the material contained herein. We have used specific examples of flows that have been observed at some of the ATEC ranges. Together, the phenomena considered here represent the bulk of mesoscale flows that will likely be encountered on a daily basis. Because there is no unifying theory for the mesoscale as exists for synoptic-scales (quasigeostrophy), the user of
mesoscale models is forced to become familiar with a set of phenomenological "building blocks" through which complicated real-life scenarios can be interpreted.

7. References
