Investigation of the Dynamical and Thermodynamical Ingredients for Mid-Latitude Winter Season Precipitation

by

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Abstract

Five fundamental physical ingredients — forcing for ascent, moisture, instability, precipitation efficiency, and temperature — are incorporated into an ingredients-based methodology (IBM) for use as an operational tool in the analysis and prediction of mid-latitude winter season precipitation. The forcing ingredient is combined with the instability ingredient to form a new parameter, QPV, which serves as an indicator of heavy precipitation potential by identifying regions where these two ingredients co-exist. The ingredient diagnostics and QPV are incorporated into ingredients maps which facilitate a systematic approach to forecasting the duration, intensity, and type of precipitation. Examples of the application of the IBM to forecasting winter precipitation events are presented, including a case study of a heavy snow event which occurred on March 13-14, 1997 in the upper Midwest.

A detailed investigation of the instability ingredient is also performed. The evolution of conditional (potential) instability in winter cyclones is examined by considering the local rate of change of $PV_{cs}$ ($PV_e$). Although horizontal advection is the dominant mechanism by which $PV_e$ is changed, adiabatic generation, controlled by thermal wind advection of $\theta_e$, also contributes to the reduction of $PV_e$, primarily upshear of the thermal ridge at the northern end of the cold front in a mature or decaying cyclone. No analogous adiabatic generation term exists for $PV_{cs}$, however, the evolution of $PV_{cs}$ is shown to be influenced by what is termed the saturation deficit term. The contribution to the Eulerian decrease in $PV_{cs}$ made by the saturation deficit term is most significant where the relative humidity is low and horizontal temperature gradients are large. The saturation deficit term is generally of secondary importance to the advection of $PV_{cs}$, though it may contribute significantly in the dry regions behind a strong cold front and upshear of a thermal ridge in mature and decaying cyclones.
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Chapter 1

Introduction

Significant technological advances have occurred over the past decade in the National Weather Service Forecast Offices (NWSFOs) which have improved the quality of numerical weather prediction model forecasts and facilitated the analysis of gridded model data and observations. However, developments in forecasting winter season precipitation have not paralleled these technological advances, and there remains considerable reliance on empirically-derived rules of thumb. Many of the techniques for predicting snowfall accumulations currently used at NWSFOs were developed prior to the availability of sophisticated gridded data analysis programs, and generate 12- or 24-hour forecasts by simply extrapolating current observations or numerical model forecasts of synoptic variables (e. g., 200 hPa temperature, 500 hPa vorticity, or 700 hPa mixing ratio). Basing forecasts on extrapolation, never very satisfactory beyond the 3- to 6-hour range, is now unnecessary given the analysis tools currently available to forecasters. Instead, the fundamental elements, or ingredients, involved in a winter precipitation
event can be analyzed. An ingredients-based analysis allows forecasters to base their predictions more directly on the physical processes involved, thus enabling them to tailor a forecast to the unique conditions characteristic of each event. In contrast, empirically-derived rules of thumb are generally based on canonical scenarios that assume set conditions, and the reliability of the techniques diminishes as conditions vary from these scenarios.

The first objective of this thesis is to develop an ingredients-based methodology (IBM) as an operational tool to help forecasters analyze and predict winter precipitation events. The IBM provides a systematic approach to forecasting winter weather by establishing a framework for interpreting numerical forecast model output and observations. The IBM presented here diagnoses five key ingredients for a winter precipitation event: quasi-geostrophic forcing for ascent, moisture, instability (i.e., gravitational, inertial, or slantwise instability), precipitation efficiency (specifically cloud microphysical properties), and temperature.

The second objective of this thesis is to investigate the instantaneous distribution and temporal evolution of instability in cold-season mid-latitude cyclones. Instability is arguably the least understood and most often overlooked ingredient in winter storm events. Though neither a necessary nor sufficient condition for precipitation to occur, instability modulates the response of an air column to forcing for vertical motion. A given forcing will produce considerably more vertical motion if it coincides with a region of instability. As a result, forecasts tend to underpredict snow amounts when convection, the release of instability, enhances the precipitation rate. Thus, this investigation focuses
on evaluating the relative importance of the processes that govern the evolution of equivalent potential vorticity ($PV_e$) and saturated equivalent potential vorticity ($PV_{es}$), two diagnostic tools employed in the identification of regions of instability.

Chapter 1 provides a general introduction to the IBM for winter weather forecasting, discusses some of the advantages of this approach, and presents the ingredients that are implicitly included in a few of the traditional winter weather forecasting techniques. Chapter 2 presents a detailed evaluation of the instability ingredient, including an analysis of the mechanisms that influence the evolution of regions of instability in a mid-latitude cyclone. Chapter 3 presents the parameters developed for diagnosing each of the five winter weather forecasting ingredients, and discusses the application of the IBM. This application is illustrated with a case study of a winter season precipitation event that occurred on March 13-14, 1997.

1.1 Background

The ingredients-based forecast methodology has been employed operationally for more than two decades in the context of warm-season severe weather. It was originally developed to forecast the initiation of deep moist convection associated with warm-season thunderstorms (McNulty 1978, 1995; Doswell, 1987; Johns and Doswell, 1992). This methodology included three ingredients—instability, moisture, and lift—and looked for all three to be present in order for deep moist convection to occur. More recently, Doswell et al. (1996) proposed an ingredients basis for the prediction of rainfall associated with
flash floods. Starting with the premise that heavy precipitation is the result of sustained high rainfall rates which are a direct consequence of the rapid ascent of moist air, the authors qualitatively predicted the instantaneous rainfall rate \( R \) by assuming that it is proportional to the vertical flux of moisture. This notion is formalized in the relationship

\[ R = E w q \]

where precipitation efficiency (E), ascent rate \( w \), and mixing ratio \( q \), constitute three ingredients in this approach. Precipitation efficiency serves as the constant of proportionality and is defined as the ratio of the mass of water falling as precipitation to the influx of water vapor mass into the cloud. Using a fourth ingredient, precipitation duration \( t_{duration} \) and estimates of rainfall rate \( R \), they predicted total precipitation (P) as \( P = R t_{duration} \).

Nietfeld and Kennedy (1998) adjusted the approach of Doswell et al. (1996) for application in forecasting snowfall amounts. They proposed air temperature, snowfall rate, and snowfall duration as the three ingredients in a snow event. The snowfall rate \( R \) ingredient is further described as \( R = E w q \), where \( q \) includes the mixing ratio anticipated by moisture advection. The ascent rate \( w \) is diagnosed by considering the synoptic and sub-synoptic scale mechanisms for lift, and \( E \) describes the degree of saturation of the air mass, cloud physics pertaining to snowflake formation, and the ratio of snow to liquid water. Although Nietfeld and Kennedy (1998) use the ingredients terminology, their approach was essentially designed to serve as a conceptual model and was not developed to have operational utility.
The IBM presented in this thesis is based on a stricter definition of ingredient than that employed in previous studies. Here, an ingredient is defined as a fundamental physical component or process that contributes to the development of a meteorological event. This definition excludes intermediate parameters such as precipitation rate and duration which are important in the diagnosis of a precipitation event, but are dependent on the more elementary physical ingredients. Because precipitation rate and duration are derived from fundamental ingredients, they do not lend themselves for use in a physically-based forecast that can be easily tailored to event-specific conditions. Use of duration as an ingredient in the manner of Doswell et al. (1996) and Nietfeld and Kennedy (1998) implicitly assumes that the rainfall rate will remain constant throughout the duration of the event. Instead, using the definition of an ingredient employed here, the storm duration can be assessed through an evaluation of the selected ingredients through all forecast hours of a numerical model. If the necessary ingredients are expected to be present at a given forecast hour, then forecasters can predict a high precipitation potential. If an important ingredient is not expected to be present, the precipitation potential will be low.

This thesis also makes a clear distinction between ingredient and diagnostic. Ingredients represent the physical components or processes directly involved in a meteorological event, while diagnostics are the observable or computed quantities which can be used to assess the presence and strength of an ingredient. Previous work has often blurred this distinction, as illustrated by the use of the mixing ratio "ingredient" by Doswell et al. (1996) and Nietfeld and Kennedy (1998). Mixing ratio is actually only one of a
number of parameters that can be used to quantify the moisture availability, and thus is more appropriately considered a diagnostic of the moisture ingredient. In the IBM for forecasting winter season precipitation presented here, parameters will be introduced to diagnose each ingredient; however, the IBM is not dependent on these specific diagnostics. Because this IBM is grounded on the physical components and processes involved, it has the flexibility to incorporate new diagnostics for these ingredients as theoretical and technological advances make them available.

The following five ingredients are employed in the IBM for forecasting winter season precipitation in this thesis:

1. Forcing for ascent: Where and how strong is the forcing?

2. Moisture: Where and how much moisture is available?

3. Atmospheric instability: How strong will the response be to the forcing?

4. Precipitation efficiency: How will cloud microphysical characteristics affect the precipitation rate?

5. Temperature: What form will the precipitation take?

The first, second, and fourth ingredients are similar to ingredients used by Nietfeld and Kennedy (1998). The fifth ingredient was implicit in their study which considered only snow events. The third ingredient, however, has not been previously considered as an ingredient for winter season precipitation. This omission has likely occurred because instability is not a necessary ingredient for precipitation, and because until
recently convenient diagnostics for identifying winter season instability were not readily available nor well understood by forecasters. As previously mentioned, reduced stability is not a necessary condition for winter season precipitation, but can significantly amplify the response of an air column to forcing for vertical motion. Thus, in this thesis, the instability ingredient is included in the IBM using convenient diagnostics which are readily available to operational forecasters.

In summary, the IBM presented in this thesis follows directly from the following physical processes involved in a winter season precipitation event. To generate any amount and type of precipitation, some mechanism to force ascent (ingredient 1) in a region with sufficient moisture availability (ingredient 2) is required. The intensity of the ensuing precipitation can be modulated by the presence of instability (ingredient 3) and the cloud microphysical properties (ingredient 4). Finally, the precipitation type is related to the temperature profile (ingredient 5). A forecast using this IBM involves evaluating each ingredient at various levels in the atmosphere for every forecast hour for which gridded data are available, to determine which ingredients are present over the forecast area. The application of the IBM to forecasting winter season precipitation is discussed in more depth in Chapter 3.

1.2 Traditional Forecast Techniques from an Ingredients Perspective

The traditional techniques used for forecasting winter precipitation events are largely empirical relationships established from observations of consistent patterns in the de-
velopment of weather systems. Because of the abundance of observational evidence on which these techniques are based, there is good reason to trust their prognostic accuracy in similarly configured synoptic situations. However, there are many circumstances in which these techniques have failed to provide accurate forecasts. An investigation of each technique from an ingredients perspective can assist in determining the reason for the failure and help to identify the conditions under which it should or should not be applied. For example, a technique that does not consider variations in the strength of the forcing for ascent should only be applied under conditions characterized by "normal" forcing. Here, normal is defined as the strength of forcing present in the cases from which the empirical technique was derived, or for those cases where the technique's reliability has been demonstrated in operational situations. This section presents a discussion of the ingredients considered by some of the popular traditional techniques in operation at NWSFOs. Section 3.2.3 presents an investigation of the normal conditions implied by one technique, the Garcia Method (Garcia, 1994).

Forecasts for snowfall accumulation are frequently prepared in NWSFOs with the use of the synoptic climatology method (Goree and Younkin, 1966; Browne and Younkin, 1970), the Cook method (Cook, 1984), the Garcia Method (Garcia, 1994), the Magic Chart (Sangster and Jagler, 1985; Chaston, 1989), and the LEMO technique (Gordon, 1998). In the synoptic climatology method (Goree and Younkin, 1966; Browne and Younkin, 1970), the most favorable location for the occurrence of heavy snow is predicted to be 2.5° latitude to the left of the track of the 500 hPa vorticity maximum, or 90 nautical miles to the left of the track of the 850 hPa pressure minimum. The Cook
method (Cook, 1984) predicts that the average 24-hour snowfall in inches is half of
the maximum "indicated warm advection" in °C at 200 hPa. Here, warm advection is
defined as the difference between the warmest temperature at 200 hPa and the coldest
temperature within a distance of 15° latitude upstream of the forecast area along the
height contours. The Garcia Method (Garcia, 1994) predicts the maximum 12-hour
snowfall in inches will be twice the average mixing ratio (in g kg⁻¹) on an isentropic
surface which intersects the forecast area between 700-750 hPa. Average mixing ratio
is defined as the average between the mixing ratio at the initial time and the maximum
mixing ratio which could be advected in during the subsequent 12 hours. Construction
of the Magic Chart (Sangster and Jagler, 1985; Chaston, 1989) involves overlaying the
24-hour net vertical displacement (NVD) of air that will arrive at the 700 hPa layer with
the 12-hour forecasted 850 hPa temperature. The location of the snowfall is determined
by identifying the regions where the greatest 700 hPa NVD coincides with an 850 hPa
between -3 °C and -5 °C. Furthermore, the Magic Chart predicts that the 12-hour
snowfall accumulation in inches (for the period between the 12- and 24-hour forecasts)
is equal to the NVD (in mb) divided by 10. Finally, the LEMO technique (Gordon,
1998) predicts that the maximum snowfall in inches is a function of the magnitude and
speed of the 500 hPa vorticity maximum.

Table 1.1 identifies the ingredients addressed in these empirical methods. Some
techniques do not directly include an ingredient, but acknowledge its importance by
instructing the forecaster to consider it independently (indicated in Table 1.1 as "LtF,
for Left to Forecaster). As shown in Table 1.1, no technique considers more than two
Table 1.1: Ingredients included in traditional snow amount forecast techniques. See text for explanation.

ingredients, and the most commonly overlooked element in forecasting winter precipitation events is the instability ingredient. Although it is not a necessary ingredient for snowfall, reduced stability can strongly influence precipitation rates. Instability is investigated further in Chapter 2 before discussing the other ingredients used in the IBM and presenting case examples of the ingredients approach in Chapter 3.

There is some room for interpretation in identifying the ingredients used in the traditional forecast techniques presented in Table 1.1. Because these techniques were derived empirically, certain ingredients may be included implicitly. For example, the Magic Chart, which predicts snowfall based on the NVD of air parcels reaching the 700 hPa surface, may implicitly consider stability because air parcels in an unstable environ-
ment will experience greater vertical displacement in a given time interval than those in a stable environment. Additionally, any empirical technique may include efficiency mechanisms implicitly because the synoptic patterns identified by an empirical technique may have specific characteristic microphysical properties which are not directly accounted for but nonetheless exist in a majority of similar cases.

If modern dynamical theory is applied to interpret the physical basis underlying the empirical relationships, the implicit presence of some ingredients may be revealed, although they remain concealed from the forecaster. For example, the LEMO technique relates the strength of the 500 hPa vorticity maximum to snowfall accumulations. If this technique is put in the context of the dynamical relationship between vorticity advection and vertical motion (Sutcliffe, 1947; Trenberth, 1978), one could maintain that the forcing ingredient is included in the LEMO technique. The Cook method, which bases its forecast of snowfall accumulation on the magnitude and location of an upper level temperature anomaly, provides another example. In the context of potential vorticity (PV) thinking, an upper level temperature maximum is associated with a tropopause-level potential vorticity maximum. The synoptic-scale features of vertical motion and thermal structure could be inferred from the typical characteristics of a PV anomaly. In both cases, however, the snowfall predictions rely on the assumption that the storm behaves according to some pre-defined and technique-specific empirical model of precipitation systems. Without directly including the ingredients, these techniques do not give forecasters the capability to tailor a forecast to case-specific conditions.
1.3 Advantages of an Ingredients-Based Methodology

Empirical snowfall prediction techniques which base their forecasts on observations of consistent patterns can be thought of as conceptual models developed from the average behavior of many events. Because of the natural variability in the development of weather systems, there will always be synoptic or thermodynamic conditions which do not fit the empirical model and thus will not be properly forecasted. In contrast, one of the primary advantages of the IBM for winter season precipitation is that its validity is not restricted to specific synoptic or thermodynamic conditions. Because the IBM is based on a physical understanding of the processes involved in a precipitation event instead of empirically-derived formulae, it provides forecasters with the flexibility to accommodate a variety of conditions.

Other advantages of the IBM are associated with its utility in the interpretation of quantitative precipitation forecasts (QPFs) generated by numerical models. Instead of relying on a “black box” utilization of QPF estimates, forecasters can use the IBM to improve upon these predictions by diagnosing the mechanisms responsible for an event. This is especially important in situations modulated by instability. Under these conditions, if forecasters are aware of the potential for convective precipitation, higher accumulations can be anticipated and included in a forecast. Additionally, the IBM provides a means of comparing the forecasts of different model runs. By identifying the roles played by each ingredient, one can evaluate the differences in the QPFs generated from different models. For example, forecasters can evaluate whether there are
differences in forcing patterns or moisture, or whether any forecasts are modulated by instability. As conditions begin to evolve, the modeled ingredients can be compared with observed values of the ingredients and actual precipitation patterns upstream of the forecast area to assist in deciding which model to choose.

The IBM can also be used to identify regional differences in precipitation. In contrast, many traditional forecast techniques focus on one station or a single cross-section and predict the snowfall for that small area; thus, they do not provide information about neighboring areas or the overall distribution of precipitation without tedious repetition of the technique. Using the IBM, forecasters can examine ingredients on isobaric surfaces and identify localized regions of stronger forcing for ascent, isolated areas of instability, or boundaries of moisture, and thus anticipate spatial variations in precipitation accumulation. This is particularly important when a forecast area is near the boundary of a region of high precipitation potential. In this situation, if the synoptic features evolve slightly differently from the predictions of the numerical forecast models, there may be considerably more or less snowfall over the forecast area.

Finally, the IBM can be used to improve upon empirical techniques for estimating snowfall accumulation. Although the IBM does not independently provide a quantitative estimate of snowfall accumulation, it can be coupled with an existing empirical snowfall prediction technique to provide an estimate of the accumulation. The Garcia Method (Garcia, 1994) is employed for this purpose in section 3.2.3. Further analysis of case studies and real-time applications may lead to the development of a quantitative relationship between the magnitudes of the ingredients and the expected snowfall totals.
However, development of such a quantitative relationship may compromise the physical basis and flexibility inherent in this approach, and would be useful only when employed in conjunction with an analysis of the individual ingredients.

The ingredients approach is slightly more cumbersome than many of the traditional techniques. However, as will be shown in Chapter 3, diagnostic tools have been developed to facilitate the application of the IBM and minimize the extra effort required. The benefits of improved forecast quality undoubtedly justify this effort.
Chapter 2

Investigation of the Instability Ingredient in Mid-latitude Cyclones

Instability in winter season weather systems has recently been the subject of increased attention among research and operational communities (e. g., Wiesmueller and Zubrick, 1998; Holsten and Hendricks, 1997; Nicosia and Grumm, 1999; SS). However, precipitation events modulated by instability remain one of the most poorly forecasted wintertime events. This chapter investigates the nature of moist instability in mid-latitude cyclones and discusses appropriate techniques for forecasting the release of gravitational or symmetric instability in a winter precipitation event. An analysis of the mechanisms responsible for the evolution of instability throughout the life cycle of a mid-latitude cyclone is presented, including a diagnosis of geostrophic equivalent and saturated geostrophic equivalent potential vorticity, $PV_{eg}$ and $PV_{egs}$, respectively.

The adiabatic and frictionless quasi-geostrophic $\omega$ equation (Bluestein, 1992) allows for the instantaneous diagnosis of vertical velocity ($\omega$) based on geopotential ($\phi$) and
temperature \((T)\). This equation can be written:

\[
\left(\sigma \nabla^2_p + f^2 \frac{\partial^2}{\partial p^2}\right) \omega = -\frac{R}{p} \nabla^2_p \left(-V_g \cdot \nabla_p T\right) + f \frac{\partial}{\partial p} \left[V_g \cdot \nabla_p (\xi_g + f)\right]
\]

(2.1)

where \(\omega = \frac{\partial p}{\partial t}\), \(\sigma = -\frac{RT}{\rho g} \frac{\partial \rho}{\partial p}\), \(\nabla_p = \left(\frac{\partial}{\partial x}, \frac{\partial}{\partial y}\right)\) on a pressure surface, \(f\) is the Coriolis parameter, \(p\) is pressure, \(R\) is the gas constant for dry air, \(V_g\) is the geostrophic wind \((V_g = \frac{1}{f} \hat{k} \times \nabla_p \phi)\), and \(\xi_g\) is the vertical component of the geostrophic relative vorticity vector, \(\xi_g = \hat{k} \cdot (\nabla \times V_g)\). Terms 2 and 3 represent the quasi-geostrophic forcing for vertical motion, and have traditionally been considered as the separate physical processes of the Laplacian of horizontal temperature advection and the rate of change of vorticity advection with height. Term 1 is similar to a three-dimensional Laplacian of \(\omega\), \(\nabla^2 \omega = \left(\frac{\partial^2 \omega}{\partial x^2}, \frac{\partial^2 \omega}{\partial y^2}, \frac{\partial^2 \omega}{\partial z^2}\right)\), except that this pseudo-Laplacian operator is modulated by a stability parameter \(\sigma\) and the Coriolis parameter \(f\).

Equation 2.1 illustrates that the response \(\omega\) to a given forcing is inextricably tied to atmospheric stability \(\sigma\). For a given forcing (terms 2 and 3) and a small value of \(\sigma\), \(\nabla^2 \omega\) must be large (provided \(\omega\) has an \(x\) or \(y\) dependence). A large \(\nabla^2 \omega\) implies enhanced differences between the maxima and minima in the \(\omega\) field and either increased vertical motions or a decreased horizontal scale of the vertical motion field (Bluestein, 1992). Thus, the stability of the atmosphere in an area of forcing for vertical motion affects both the location and the intensity of the subsequent vertical motion.
2.1 Types of Moist Instability

An important distinction in evaluating moist instability is whether it is conditional or potential (convective) instability. Rogers and Yau (1989) note, “Convective [potential] instability has to do with the lifting of layers and should not be confused with conditional instability which applies to an undisplaced layer.” An air column that is conditionally unstable requires only an infinitesimal displacement to realize the instability, provided that the air is saturated locally. An air column that is potentially (convectively) unstable requires a finite vertical displacement to reach saturation and realize the instability (Schultz and Schumacher, 1999; hereafter SS).

Instability can be further distinguished as gravitational, inertial, and symmetrical based on the direction in which a displaced air parcel is compelled to move when the instability is realized. With gravitational instability, an air parcel accelerates vertically once displaced up or down. With inertial instability, an atmosphere is similarly unstable to horizontal displacements. Inertial instability occurs when the absolute vorticity is negative, a condition that is only occasionally observed in mid-latitudes and thus will not be considered in this thesis. Symmetrical instability exists in a two-dimensional atmosphere when it is stable to both vertical and horizontal displacements is unstable to displacements along slanted paths. These classifications of instability can be combined to distinguish four types of moist instability: conditional gravitational instability (CI), conditional symmetric instability (CSI), potential gravitational instability (PI), and potential symmetric instability (PSI). These types are summarized in Table 2.1.
<table>
<thead>
<tr>
<th></th>
<th>Conditional ($\theta_{es}$)</th>
<th>Potential ($\theta_e$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gravitational</td>
<td>CI</td>
<td>PI</td>
</tr>
<tr>
<td>Symmetric</td>
<td>CSI</td>
<td>PSI</td>
</tr>
</tbody>
</table>

Table 2.1: Summary of four types of moist instability: conditional gravitational (CI), conditional symmetric (CSI), potential gravitational (PI), and potential symmetric (PSI) instabilities.

An atmosphere is characterized by CI if the environmental lapse rate ($\gamma$) is greater than the pseudoadiabatic (or saturated adiabatic) lapse rate ($\Gamma_m$) and less than the dry adiabatic lapse rate ($\Gamma_d$): $\Gamma_m < \gamma < \Gamma_d$ (Rogers and Yau, 1989). This requirement is equivalent to the condition that the saturated equivalent potential temperature ($\theta_{es}$) decreases with height, $-\frac{\partial \theta_{es}}{\partial p} < 0$. CSI exists when the environmental lapse rate along a surface of constant absolute geostrophic momentum ($M_g = V_g + f x$, where the y-direction is taken to be in the direction of the geostrophic vertical shear) lies between the moist and dry adiabatic lapse rates (i.e., the atmosphere is conditionally unstable along an $M_g$ surface). For a two dimensional atmosphere, CSI can be diagnosed where contours of $\theta_{es}$ are more steeply sloped than contours of $M_g$ in a cross-section drawn perpendicular to the thermal wind (i.e., the $M_g - \theta_{es}$ relationship, e.g., Snook, 1992).

PI and PSI can be diagnosed similarly to CI and CSI, except that equivalent potential temperature ($\theta_e$) is substituted for $\theta_{es}$. Thus, an atmospheric layer is potentially unstable where $-\frac{\partial \theta_e}{\partial p} < 0$, and PSI is diagnosed by the $M_g - \theta_e$ relationship. Note that if the atmosphere is locally saturated, $\theta_e = \theta_{es}$ and potential and conditional instabilities are equivalent (PI $\equiv$ CI and PSI $\equiv$ CSI).
Because an atmospheric layer in an unsaturated potentially unstable atmosphere (PI or PSI) must undergo a finite vertical displacement to reach saturation and release the instability, potential instability is not a sufficient condition for moist convection. The magnitude of the required displacement will vary considerably under different thermodynamic conditions, and many regions of potential instability are never realized. In contrast, conditional instabilities (CI or CSI) only require an infinitesimal displacement and local saturation to be realized. Therefore, conditional instabilities are a more "appropriate measure of the susceptibility of the atmosphere to [gravitational and symmetric] convection" (SS).

One drawback to evaluating instability using the $M_g - \theta_c$ and $M_g - \theta_{es}$ relationships is that developing cross-sections can be cumbersome and time-consuming. In addition, this approach requires an atmosphere which can be accurately described in two dimensions. A more convenient and also more flexible technique for diagnosing moist instability on isobaric surfaces involves the use of geostrophic equivalent potential vorticity ($PV_{eg}$) and saturated geostrophic equivalent potential vorticity ($PV_{es}$). The application of $PV_c$ and $PV_{es}$ to diagnose moist instabilities in cyclones is discussed in the next section.

### 2.2 Diagnosing Moist Instability

Geostrophic equivalent potential vorticity ($PV_{eg}$) is defined as

$$PV_{eg} = -g \mathbf{\zeta}_g \cdot \nabla \theta_c$$  \hfill (2.2)

where $g$ is the gravitational constant, $\mathbf{\zeta}_g$ is the three-dimensional absolute geostrophic vorticity vector, $\nabla$ is the gradient operator in $x$, $y$, and $p$ coordinates, and $\theta_c$ is the
equivalent potential temperature. $PV_{e_g}$ has traditional potential vorticity units, PVU, where 1 PVU $= 1 \times 10^{-6} m^2 s^{-1} kg^{-1}$. For a zonal flow, Martin et al. (1992) showed that the equivalent potential vorticity is

$$PV_{e_g} = \frac{\partial M_g \partial \theta_e}{\partial p} - \frac{\partial M_g \partial \theta_e}{\partial x} \frac{\partial x}{\partial p}$$  \hspace{1cm} (2.3)

Regions of symmetric or gravitational instability can be identified where the quantity $PV_{e_g}$ is negative. Use of the two-dimensional $PV_{e_g}$ (equation 2.3) for this diagnosis, however, is limited to vertical cross-sections that meet the following conditions (SS): (1) the geostrophic wind direction doesn’t vary with height (i.e., it is two-dimensional); (2) the cross-section is perpendicular to the thermal wind (the shear of the geostrophic wind), and (3) the along-flow ageostrophic wind is small so that the assumption of a geostrophic wind is reasonable. This excludes, for example, regions of curvature (violation of the 2-D assumption) or wind acceleration (violation of the assumption of small along-front ageostrophy). In cyclones, flow is rarely constrained to two dimensions, particularly in the occluded quadrant (the area northwest of the sea level pressure minimum where many traditional forecasting methods suggest heavy snow often falls) because the flow is highly curved. In areas such as fronts where the along-front variations are negligible and the flow is nearly two-dimensional, the three conditions listed above may occasionally be met. However, even in cases characterized by two-dimensional dynamics, the use of two-dimensional $PV_{e_g}$ to diagnose instability still requires the tedious inspection of multiple cross-sections to assess the horizontal extent of instability, and the results are dependent on the orientation of the cross-section.
McCann (1995) observed that the complete three-dimensional $PV_e$ (equation 2.2) can be used to identify regions of general instability without assuming a two-dimensional flow. By recognizing that the geostrophic wind shear $-\frac{\partial V}{\partial p}$ is related to the temperature gradient through the thermal wind balance, he showed that

$$PV_e = g \left( -\frac{1}{f \rho \theta_e} \Gamma_m \nabla \rho \theta_e \right)^2 - \hat{k} \cdot \zeta \frac{\partial \theta_e}{\partial p} \right)$$

(2.4)

where $\Gamma_d$ and $\Gamma_m$ are the dry and adiabatic lapse rates, respectively. The first term on the right hand side is always negative, and its magnitude is determined by the horizontal temperature gradient. The second term is a measure of gravitational instability. McCann considered three possible scenarios. If the second term is negative (unstable to moist convection, $-\frac{\partial \theta}{\partial p} < 0$), $PV_e$ is negative and vertical (gravitational) instability dominates regardless of the magnitude of the first term. If the second term is positive (stable to moist convection, $-\frac{\partial \theta}{\partial p} > 0$) and the temperature gradient is weak enough that the first term is smaller than the second, then $PV_e$ is positive and the environment is stable. However, if the lapse rate is slightly stable ($-\frac{\partial \theta}{\partial p}$ only slightly greater than zero) and the horizontal temperature gradient is strong, the second term is positive but of lesser magnitude than the first term so that $PV_e$ is negative. Under these conditions the result is slantwise (symmetric) instability. Based on these observations, McCann concluded that the three-dimensional $PV_e$ combines gravitational and symmetric instabilities and becomes an all-purpose convection potential tool.

Use of a three-dimensional $PV_e$ permits a quasi-horizontal analysis of instability on pressure surfaces and a cross-sectional analysis independent of the orientation of the
cross-section. However, since symmetric instabilities remain strictly two-dimensional phenomena, the three-dimensional \( PV_{eg} \) does not offer any information about symmetric instabilities unless the flow is two-dimensional. In contrast, gravitational instability can be diagnosed by negative \( PV_{eg} \) (provided \( \frac{\partial P}{\partial p} < 0 \)) regardless of the curvature of the flow.

With virtually universal access to gridded data analysis programs (e. g., GEMPAK, AWIPS, Vis5D, GARP, NTRANS), the \( M_g - \theta_c \) relationships and \( PV_{eg} \) have increasingly been employed in operational forecasting and research for the diagnosis and evaluation of CSI (e. g., Snook, 1992; Moore and Lambert, 1993; Wiesmueller and Zubrick, 1998). However, SS identified a misconception behind the conventional application of \( PV_{eg} \) to diagnose CSI. The authors noted that unless the air parcel is saturated, \( PV_{eg} \) is strictly an indicator of potential instabilities (PI or PSI), and not conditional instabilities such as CSI. Instead, saturated equivalent geostrophic potential vorticity \( (PV_{esg}) \) should be used to diagnose CI and CSI. \( PV_{esg} \) is defined in the same manner as \( PV_{eg} \) but with saturated equivalent potential temperature \( \theta_{es} \) substituted for \( \theta_c \):

\[
PV_{esg} = -g \vec{z} \cdot \nabla \theta_{es} \tag{2.5}
\]

Saturated equivalent potential temperature is simply the equivalent potential temperature that an air parcel (at a given air temperature \( T \) and pressure \( P \) ) would have if it were saturated at that same \( T \) and \( P \). Mathematically,

\[
\theta_{es} = \theta_c \frac{L}{L_s c_p} + \theta_s \tag{2.6}
\]

where \( L \) is the latent heat of vaporization, \( l_s \) is the saturation mixing ratio at \( T \) and \( P \),
and $C_p$ is the specific heat capacity of dry air at constant pressure.

Bennetts and Hoskins (1979) noted when they introduced their theory applying wet-bulb potential vorticity (analogous to $PV_{eg}$ in this context) to the diagnosis of CSI that such an approach is only valid in the presence of saturation. This assumption has often been neglected in the subsequent use of $PV_{eg}$ to diagnose CSI. At saturation, $\theta_e = \theta_{es}$ and $PV_{eg} = PV_{esg}$, and either quantity can be used to diagnose CSI. However, when conditions are unsaturated, the quantities are not equal and cannot be used interchangeably. SS pointed out the misuse of $PV_{eg}$ resulting from inappropriate applications of the original theory by Bennetts and Hoskins (1979) and suggested that “in the future, we, as a meteorological community, use the term CSI only when employing $\theta_e^*$ [their notation for $\theta_{es}$] and use the term PSI only when employing $\theta_e$.” The effort of SS to reach the operational and academic communities is reflected in the increasing use of $PV_{esg}$ instead of $PV_{eg}$ (or $M_g - \theta_{es}$ instead of $M_g - \theta_e$) in the literature and among operational forecasters to diagnose CI and CSI (e. g., Nicosia and Grumm, 1999; Clark and Nicosia, 1999; Banitt, 1999; McCann 1999).

Hereafter, the moist potential vorticity quantities will be abbreviated as $PV_e$ and $PV_{es}$ for geostrophic equivalent potential vorticity and geostrophic saturated equivalent potential vorticity, respectively. Notice that although the “g” has been omitted from the abbreviation, geostrophic absolute vorticity is used in the calculation of these quantities.
Chapter 3

The Ingredients-Based Methodology for Forecasting Winter Season Precipitation Events

The IBM for forecasting winter season precipitation events was introduced in Chapter 1, and the five key ingredients were identified to be QG forcing for vertical motion, moisture, instability, precipitation efficiency, and temperature. Instability was investigated in Chapter 2, including an analysis of the mechanisms that govern the evolution of instabilities in mid-latitude cyclones. In Chapter 3, the five ingredients are formally described and the diagnostics used to quantify their contributions are introduced. The application of the IBM to forecasting winter precipitation is also discussed, and a case study of a strong midwestern snow storm is presented.
3.1 Ingredient Descriptions

3.1.1 Forcing for Ascent

As an initially unsaturated column of air is lifted by some forcing mechanism, its temperature decreases until saturation occurs. Provided sufficient condensation or ice nuclei are present, any additional lifting after saturation is achieved results in the condensation of water vapor to liquid or ice, which can eventually fall out as precipitation. This section addresses the forcing ingredient and presents a diagnostic for quantifying the degree of quasi-geostrophic (QG) forcing.

Most traditional forecast techniques for the prediction of winter season precipitation do not specifically consider forcing for ascent. The Magic Chart (Sangster and Jagler, 1985; Chastan, 1989) is the only technique listed in Table 1.1 that directly includes vertical motion in its forecast algorithm. Instead of considering forcing for ascent, however, the Magic Chart uses the net vertical displacement (NVD) of air parcels reaching the 700 hPa level at the end of the 12-hour forecast period (Reap, 1990). The Garcia Method (Garcia, 1994) does not include forcing for ascent directly, however it explicitly states that the “area of concern” chosen by the forecaster must coincide with some forcing mechanism before the technique can be applied. Since it already prescribes an independent analysis of the forcing ingredient, the Garcia Method (Garcia, 1994) can be integrated nicely into the IBM, as will be shown later.

With modern tools available for the analysis of gridded data, it is not difficult to compute diagnostics of QG forcing for vertical motion. One such diagnostic uses the
Q-vector form of the adiabatic and inviscid QG-\( \omega \) equation (Hoskins et al., 1978):

\[
\left( \sigma \nabla^2 + f^2 \frac{\partial^2}{\partial p^2} \right) \omega = -2 \nabla \cdot \vec{Q}
\]  

(3.1)

where

\[
\vec{Q} = \left[ -\frac{\partial V_x^g}{\partial x} \left( -\frac{\partial \phi}{\partial p} \right) i - \frac{\partial V_y^g}{\partial y} \left( -\frac{\partial \phi}{\partial p} \right) j \right]
\]

and

\[
\omega = \frac{dp}{dt}
\]

and the variables are used as defined in Chapter 2. As discussed in Chapter 2, the left-hand side is similar to a 3-D Laplacian, except that this pseudo-Laplacian operator is modulated by a stability parameter \( \sigma \) and the Coriolis parameter \( f \). A formal solution of equation 3.1 would require a distribution of \( \vec{Q} \) in an isobaric layer and boundary conditions for \( \omega \).

Often a qualitative estimate of forcing for vertical motion is all that is necessary in an operational environment. Recognizing that the forcing \( \nabla \cdot \vec{Q} \) and the response \( \omega \) are related by the pseudo-Laplacian operator, equation 3.1 provides a means of making a qualitative estimate. Convergence of the Q-vector (\( \nabla \cdot \vec{Q} < 0 \)) corresponds to a negative \( \omega \) and thus forcing for upward vertical motion. Divergence of the Q-vector (\( \nabla \cdot \vec{Q} > 0 \)) corresponds to forcing for downward vertical motion. Thus, isobaric contours of \( \nabla \cdot \vec{Q} \) computed at a number of levels, in conjunction with a cross-sectional analysis, provide a qualitative estimate of the direction and magnitude of the vertical motion forcing throughout the atmosphere.

Based on an evaluation of a number of case studies, a classification for describing
the QG magnitude of forcing was developed and is shown in Table 3.1. These ranges are

<table>
<thead>
<tr>
<th>( \nabla \cdot \vec{Q} \ (Km^{-2}s^{-1} \times 10^{-15}) )</th>
<th>Classification</th>
</tr>
</thead>
<tbody>
<tr>
<td>-1 to -5</td>
<td>Weak Forcing</td>
</tr>
<tr>
<td>-5 to -15</td>
<td>Moderate Forcing</td>
</tr>
<tr>
<td>&lt; -15</td>
<td>Strong Forcing</td>
</tr>
</tbody>
</table>

Table 3.1: Classification of Q-vector forcing.

intended to provide a consistent terminology for discussing the strength of QG forcing for ascent. Provided that the atmosphere is well-described by the QG equations (i.e., low Rossby number flow), there is ample moisture throughout the lower- to mid-troposphere, and no instability, the categories in Table 3.1 roughly describe the intensity1 of the ensuing precipitation. Weak forcing corresponds to light precipitation, moderate forcing to moderate precipitation, and strong forcing to heavy precipitation.

However, this approach to estimating the QG forcing for ascent neglects any modulation by the stability parameter, \( \sigma \), because it involves approximating the pseudo-Laplacian operator \( \left( \sigma \nabla^2 + f^2 \frac{\partial^2}{\partial \eta^2} \right) \) with the true Laplacian. Since \( \sigma \) appears in the full form of the QG-\( \omega \) equation (equation 3.1), accurate estimates of the magnitude of the upward vertical motions resulting from a given Q-vector convergence should include consideration of the atmospheric stability. Section 3.1.3 and chapter 2 will address the instability ingredient in more detail.

---

1 Precipitation intensity is defined according to the system used for surface station observations: light precipitation corresponds to -SN or -RA (for snow or rain, respectively), moderate precipitation to SN or RA, and heavy precipitation to +SN or +RA.
It is important to remember that the use of the Q-vector diagnostic as a sole means for vertical motion forcing does impose limits on the analysis, namely, that it is only applicable in those regions where the Rossby number is low. No accommodation for the role of non-QG motions in modulating the observed vertical motions is made in the approach presented in this thesis. Such effects may be particularly significant in the vicinity of frontal zones or jet streaks. A useful addition to the present analysis would include a quantification of the degree to which each case agrees with its QG description.

### 3.1.2 Moisture

Given ample moisture, even very weak forcing can be sufficient to generate precipitation. However, precipitation will not be produced in a dry atmosphere even in the presence of strong forcing and instability. Thus, an evaluation of available moisture should be included in any forecast for precipitation.

There are a variety of ways to assess the moisture availability in a system. The Magic Chart (Sangster and Jagler, 1985; Chastn, 1989) leaves the evaluation of the moisture availability up to the forecaster, recommending an examination of the 500-1000 hPa layer-averaged relative humidity. The Garcia Method (Garcia, 1994) is based on the assumption that moisture is often the limiting factor in a snow event. It predicts a direct relationship between the mixing ratio on a mid-tropospheric isentropic surface and snowfall amounts. For a 12-hour period, in a region where snowfall is expected based on an evaluation of forcing mechanisms, the Garcia Method predicts that the maximum amount of snow (in inches) will equal twice the maximum mixing ratio (in g
kg\(^{-1}\)) on an isentropic surface which intersects the 700-750 hPa layer over the forecast area. The Cook Method (Cook, 1980) draws from the theory of Jacobson et al. (1956) which holds that it isn’t necessary to explicitly worry about moisture in a developing storm because if the dynamics are strong moisture will find its way into the storm.

The technique for evaluating moisture availability in this thesis first involves the inspection of relative humidity at a number of levels throughout the lower and mid-troposphere to determine the degree of saturation. Relative humidity is defined as the ratio of the actual mixing ratio (\(l\)) to the saturation mixing ratio with respect to water at the same temperature and pressure (\(l_s\)) and is typically expressed as a percent (Wallace and Hobbs, 1977):

\[
RH = 100 \frac{l}{l_s}
\]  
(3.2)

After determining the degree of saturation, the available moisture must be quantified because relative humidity does not provide information about the absolute moisture content and, thus, cannot describe the amount or intensity of the resulting precipitation. For this purpose, we examine the mixing ratio \(l\), defined as the ratio of the mass of water vapor (\(M_w\)) to the mass of dry air (\(M_d\)) in a parcel:

\[
l = \frac{M_w}{M_d}
\]  
(3.3)

Mixing ratio was chosen for this technique because of our intent to integrate the ingredients technique with the Garcia Method for estimating snowfall accumulation. The Garcia Method is well suited for this purpose due to its comprehensive treatment of the moisture ingredient, including an analysis of mixing ratios on isentropic surfaces.
However, isentropic analysis is not appropriate for neutral or unstable conditions, nor regions where diabatic heating or cooling is occurring. This excludes any areas of instability and regions characterized by a phase change. By integrating the Garcia Method into the ingredients technique, situations when isentropic analysis cannot be accurately or prudently applied can be accommodated. This is discussed further in section 3.2.3.

3.1.3 Instability

In section 3.1.1, $\nabla \cdot Q$ was discussed as a technique for identifying regions of forcing for ascent. However, equation 3.1 reveals that an analysis of $\nabla \cdot Q$ alone will not account for the influence of atmospheric stability ($\sigma$) on the magnitude of the response ($\omega$) to a forcing. Thus, in this thesis, the instability instability ingredient is considered whenever the $Q$-vector diagnostic is employed to account for the modulation of the response to a given forcing for ascent.

The stability parameter $\sigma$ considers only gravitational instabilities and doesn’t identify regions of symmetric instabilities. In order to allow for consideration of both types, we use saturated geostrophic equivalent potential vorticity, $PV_{es}$, in the IBM to diagnose the instability ingredient, where $PV_{es}$ is defined by equation 2.5. By choosing to use $PV_{es}$ rather than $PV_e$, only conditional gravitational (CI) and conditional symmetric instabilities (CSI) are considered, rather than potential gravitational (PI) and potential symmetric instabilities (PSI). Where $PV_{es}$ is negative, CI exists if $-\frac{\partial \theta}{\partial p} < 0$ and CSI exists if $-\frac{\partial \theta}{\partial p} > 0$ and the atmosphere is characterized by two-dimensional flow. CI
and CSI are measures of the susceptibility of the atmosphere to moist gravitational and moist symmetric convection, not indicators of the existence of such convection (SS). Thus, for either instability to be realized and convection to occur, saturation must be present locally and a mechanism to force an infinitesimal upward vertical motion must exist. It is important to distinguish between gravitational and symmetric instabilities only insofar as the atmosphere may respond differently, particularly with respect to the organization of precipitation bands in each type of instability. However, this distinction is not emphasized in the forecast technique presented in this thesis because both instability mechanisms have similar implications, namely increased snowfall amounts and the potential for lightning and thunder. Instead, the IBM will focus on identifying a region of instability, then leave it to forecasters to analyze cross-sections and decide if a vertical or slantwise response can be expected.

It would be useful to derive a relationship between $\nabla \cdot \vec{Q}$, $PV_{es}$, and $\omega$ to provide quantitative estimates of the vertical motion; however, this derivation is beyond the scope of this thesis. Instead, regions where enhanced vertical motions can be expected are identified based on the co-location of both forcing and instability. Since both negative $\nabla \cdot \vec{Q}$ (associated with upward vertical motion) and negative $PV_{es}$ (associated with conditional instabilities) are indicative of a strong precipitation potential, regions with large positive values of the product $(\nabla \cdot \vec{Q})(PV_{es})$, computed where both terms are negative, have a strong likelihood for experiencing extreme precipitation events provided that sufficient moisture is available. Here, a new diagnostic parameter, QPV, is
introduced to capture this effect. QPV is defined as

\[
QPV = \begin{cases} 
(\nabla \cdot \vec{Q}) (PV_{es}) & \text{for negative } \nabla \cdot \vec{Q} \text{ and negative } PV_{es} \\
0 & \text{for positive } \nabla \cdot \vec{Q} \text{ and/or positive } PV_{es}
\end{cases}
\] (3.4)

and computed as

\[
QPV = \left( \frac{\nabla \cdot \vec{Q} - |\nabla \cdot \vec{Q}|}{2} \right) \left( \frac{PV_{es} - |PV_{es}|}{2} \right)
\] (3.5)

Where this positive product is large, there is forcing in the presence of instability, and strong upward vertical motion should be anticipated. Because the two quantities \( \nabla \cdot \vec{Q} \) and \( PV_{es} \) span a different range of values, the absolute magnitude of the quantity QPV may be more sensitive to one than the other. Thus, we employ QPV as an indicator of the potential for convective precipitation without regard to its absolute magnitude. It is important to remember that \( PV_{es} \) need not be negative for precipitation to fall. In fact, heavy snow often results from strong forcing alone. Therefore, contours of QPV should only be used to identify areas of potentially convective snowfall. Many winter season precipitation events with sufficient forcing and moisture will be associated with positive \( PV_{es} \) and, thus, zero QPV.

### 3.1.4 Efficiency

Doswell et al. (1996) defined efficiency as the ratio of the mass of water falling as precipitation to the influx of water vapor mass into the cloud in their ingredients-based technique for forecasting precipitation associated with flash flooding, (see section 1.1). This ingredient served as their constant of proportionality in a relationship between
rainfall rate, ascent rate, and mixing ratio. With knowledge of the processes governing the cloud microphysics, efficiency ingredient can be assessed in a more sophisticated manner. In this section the cloud microphysical processes that influence the efficiency of a winter precipitation event are discussed. First, the processes involved in the formation of ice from supercooled liquid droplets are examined. Then, the growth of ice particles following initiation, and the conditions that lead to maximum growth are discussed. Finally, ways to incorporate precipitation efficiency into the ingredients-based approach are identified.

**Ice Nucleation**

To initiate the precipitation process in cold clouds, supercooled liquid must freeze into an ice crystal. This can occur in three ways: heterogeneous nucleation, homogeneous nucleation, or deposition onto a pre-existing ice crystal. Heterogeneous nucleation is the process by which supercooled liquid condenses onto an ice nuclei to form an ice crystal. In homogeneous nucleation, supercooled liquid can form an ice crystal without any nuclei present, however this requires a temperature below $-40^\circ$C. Homogeneous nucleation is not considered here because it is not typically characteristic of winter storms at mid-latitudes. Furthermore, it is assumed that there are no ice crystals introduced into the layer of supercooled liquid, thus limiting the discussion to the initiation of ice crystals by heterogeneous nucleation.

The initiation of an ice crystal by heterogeneous nucleation requires a small particle, or nuclei. The ability of supercooled liquid to condense on the nuclei and form an
ice crystal depends on the chemical makeup of the particle, and the temperature and relative humidity of the ambient air (Rogers and Yau, 1989). Over the upper Midwest, 80-90% of the ice nuclei (IN) are some form of clay, primarily vermiculite. Vermiculite requires an air temperature of $-15^\circ C$ for ice nucleation to occur. Table 3.2 shows some typical IN and their corresponding activation temperatures. Because of the presence of multiple types of IN in a cloud, and the variation of initiation temperature with ambient relative humidity and surface characteristics of the IN, a specific value cannot be assigned to the temperature at which ice nucleation is always initiated. Instead, guidelines determined by statistical studies must be used to identify a range of temperatures for which ice nucleation can be expected. Based on the results of a number of studies (e.g., Borovikov et al., 1963; Mossop et al., 1970), Baumgardt (1999) prepared guidelines for determining whether ice nuclei might be initiated within a cloud. These are shown in Table 3.3. Furthermore, Baumgardt (1999) concluded that $-12^\circ C$ to $-14^\circ C$ should be considered the minimum range for a good chance of ice being in a cloud, and offered

<table>
<thead>
<tr>
<th>IN Composition</th>
<th>Activation Temperature</th>
</tr>
</thead>
<tbody>
<tr>
<td>Silver Iodide</td>
<td>$-4^\circ C$</td>
</tr>
<tr>
<td>Kaolinite</td>
<td>$-9^\circ C$</td>
</tr>
<tr>
<td>Volcanic Ash</td>
<td>$-13^\circ C$</td>
</tr>
<tr>
<td>Vermiculite</td>
<td>$-15^\circ C$</td>
</tr>
</tbody>
</table>

Table 3.2: Examples of some typical ice nuclei (IN) and the temperature required for ice crystal activation on their surfaces, from Rogers and Yau (1989).
Cloud Temperature | Chance of Ice in the Cloud
---|---
≤ -4°C | 0% (No Ice)
-10°C | 60%
-12°C | 70%
-15°C | 90%
-20°C | 100%

Table 3.3: Guidelines for determining if ice nucleation has occurred in a cloud, adapted from Baumgardt (1999).

-10°C as the operational cutoff point for no ice.

**Ice Particle Growth**

After initiation, ice particles can grow in the cloud through the following mechanisms, deposition, accretion, or aggregation. Growth by deposition involves the change from water vapor to ice when water vapor is deposited on ice or ice nuclei. This is governed by the Bergeron-Findeisen process in which ice crystals grow at the expense of liquid droplets in environments for which the relative humidity is 100%. Since the saturation vapor pressure over ice is less than that over water, the vapor is compelled toward the ice, depositing on the ice when it makes contact. The growth rate by deposition is greater at colder temperatures, with a maximum growth rate around -15°C (Rogers and Yau, 1989, see Fig. 3.1).

Growth by accretion occurs when an ice particle overtakes or captures supercooled
Figure 3.1: Normalized ice crystal growth rate by deposition as a function of temperature, for 50 kPa and 100 kPa. Adapted from Rogers and Yau (1989).

liquid droplets. This occurs mainly after the ice particle has grown to a sufficient size to begin to fall and collect the supercooled droplets. Hence, ice must initially grow by deposition at the top or mid-level of the cloud and only later by accretion as it falls through the cloud. Aggregation is the joining of multiple ice particles to form one large snowflake. This process generally occurs when temperatures are warmer than $-10^\circ$C, with the largest snowflakes occurring when the temperature is near $0^\circ$C.

In this thesis, temperature is included at each level in an analysis of the ingredients in order to assess the likelihood of ice nucleation and to anticipate enhanced precipitation rates associated with rapid snowflake growth. The minimum cloud-top temperature component of the efficiency analysis is best performed using satellite-derived cloud-top temperature data and inspection of soundings during the event; however, forecast
soundings and temperature on isobaric surfaces can provide an estimate of the likelihood that ice will be initiated in an event. Snowflake growth can be assessed in the IBM using model-predicted temperature by evaluating whether an area of strong forcing for ascent coincides with a region of maximum depositional growth \( T \approx -15^\circ \text{C} \). In this situation, heavy precipitation is possible provided sufficient moisture is present (Auer and White, 1982).

### 3.1.5 Temperature

An analysis of the temperature ingredient in a winter season mid-latitude cyclone addresses the precipitation type, or whether the precipitation will fall as rain, snow, ice pellets, or freezing rain\(^2\). The temperature ingredient is best diagnosed through the careful monitoring of observed temperature profiles as the storm develops; however, isobaric analyses of model-predicted temperature and moisture fields can help forecasters to anticipate precipitation type well in advance of the event. As with other forecasting rules of thumb, the conventional guidelines for predicting precipitation type are best applied in the context of a physical understanding of the thermodynamics governing precipitation type.

In a winter season weather system, precipitation is usually formed initially as snow and the thermal structure of the layer through which it falls determines the type of precipitation that will reach the ground. The initiation and growth of the original

\(^2\) Freezing rain is defined as rain that falls in liquid form but freezes upon impact, and ice pellets (or sleet) are transparent or translucent pellets of ice, 5mm or less in diameter, which form from the freezing of rain drops or the refreezing of largely melted snowflakes when falling through a below-freezing layer of air near the earth's surface (Huschke, 1959).
ice crystals requires special consideration and was the subject of section 3.1.4, which focused on the efficiency ingredient. The discussion begins here with ice crystals that have grown large enough to fall from the cloud.

The amount of heat transferred to a melting ice crystal is proportional to (1) the difference between the temperature of the crystal (generally 0°C) and the surrounding air, and (2) the amount of time the crystal spends at that temperature. Therefore, the ability of an ice crystal to melt depends on its fall velocity (related to its size and shape), crystal structure, and the temperature of the air through which it falls. If the air temperature remains below freezing between the level of ice crystal initiation and the ground, the precipitation will fall to the ground as snow. However, if a layer of warm (above freezing) air exists, either elevated or extending to the surface, there is the potential for one or more phase changes to occur before the precipitation reaches the ground.

**Monotonically Decreasing Temperature Profile**

For an atmospheric temperature profile which decreases monotonically with height, with an above-freezing layer extending to the surface, the precipitation will remain frozen and fall as snow if the wet bulb temperature is below 0°C throughout the column. This is because ice (liquid) begins to sublimate (evaporate) if it falls through an unsaturated layer. The environmental temperature will cool due to this sublimation (evaporation) and the frost point (dew point) will rise, so that both approach the wet-bulb temperature. For the case of a surface warm layer wet-bulb temperature greater than 0°C,
there is the potential for partial or complete melting of the ice crystal. The probability that the precipitation will be snow has previously been expressed simply in terms of the depth of the warm layer (or, the height of the freezing level above ground). Table 3.4 presents the work of McNulty (1988) describing the likelihood as a function of warm layer depth that the precipitation type will be snow. Notice that the temperature of the warm layer is not considered here. McNulty (1988) also notes, “On the average,

<table>
<thead>
<tr>
<th>Freezing Level</th>
<th>Chance of Snow</th>
</tr>
</thead>
<tbody>
<tr>
<td>35 hPa or 290 m</td>
<td>50 %</td>
</tr>
<tr>
<td>25 hPa or 201 m</td>
<td>70 %</td>
</tr>
<tr>
<td>12 hPa or 96 m</td>
<td>90 %</td>
</tr>
</tbody>
</table>

Table 3.4: Relationship between the freezing level for a temperature profile decreasing monotonically with height and the likelihood that the precipitation type will be snow, adapted from McNulty (1988).

the freezing level must be at least 1200 feet above the surface to insure that the snow will melt to rain.” This statement can be extended by recognizing that in the Western Plains regions of the United States, the 850 hPa surface is approximately 1200 feet above ground. Therefore, the 0°C 850 hPa isotherm marks approximately the “rain edge” of the rain-snow boundary in these locations. Areas with an 850 hPa temperature greater than 0°C are nearly certain to experience rain, however a close inspection of McNulty’s statement reveals that it does not comment on the likelihood of precipitation falling as snow for temperatures less than 0 °C. Moreover, Table 3.4 shows that even for a freezing level at 96 m (or approximately 12 hPa) above ground, there is still a 10% chance
that the precipitation will not be snow. Therefore, according to this interpretation of McNulty's work, the conventional use of the 850 hPa 0°C contour as the "rain-snow line" must be applied with caution, particularly in the "snow regime."

**Temperature Profile with an Elevated Warm Layer**

In the event of an elevated layer of above-freezing temperature with below-freezing temperatures underneath it (Fig. 3.2), the potential exists for freezing rain or sleet/ice pellets. The degree of melting that occurs in the warm layer is the key difference

---

**Figure 3.2:** Schematic of an Elevated Warm Layer. Thick solid line is ambient air temperature, thin solid lines are -10 °C, 0 °C, and 10 °C isotherms. Adapted from Czys et al. (1996).

between conditions characteristic of freezing rain and those of sleet (Penn, 1957). If the
snow completely melts to a liquid in the warm layer, then freezing rain or rain is likely. However, if the snow is only partially melted when it reaches the cold air beneath the warm layer, it begins to re-freeze immediately due to the presence of the remaining ice nuclei and falls to the ground as sleet or snow.

Czys et al. (1996) present a means of determining the degree of melting in an elevated warm layer based on the temperature of the elevated warm layer and the depth of that layer. The authors introduce a nondimensional parameter \( \tau \) to assist in the determination of whether complete melting can be expected in an elevated warm layer and thus, in the diagnosis of freezing rain and ice pellets. In their study, \( \tau \) is defined for a particle of a given radius as the ratio between the residence time (\( t_{res} \)) of the particle in the warm layer and the time required for complete melting (\( t_{melt} \)). \( t_{res} \) is based on the vertical distance the particle must fall (i.e., depth of the warm layer) and the velocity at which it falls. The relationship describing \( t_{melt} \) is derived through a balance between the rate of energy required to transform the solid to a liquid and the rate that the environment can supply the needed energy.

The resulting nondimensional parameter \( \tau \), can be used to determine the degree of melting in an elevated warm layer for a given maximum particle radius. For \( \tau \geq 1 \), conditions are characteristic of complete melting and freezing rain is favored. For \( \tau < 1 \), only partial melting of the largest particles is expected and precipitation will likely fall as ice pellets or snow. Figure 3.3 shows isopleths of the threshold value, \( \tau = 1 \), for a range of ice particle radii, warm layer depths, and mean warm layer temperatures. Based on 17 case studies in the Midwestern winter season of 1995-1996, Czys et al. (1996) found
that a critical ice particle radius of 400 $\mu m$ provides a reasonable level of accuracy. Thus, if the mean warm layer temperature and warm layer depth intersect in Fig. 3.3 above the 400 $\mu m$ $\tau = 1$ isopleth, complete melting and freezing rain can be expected provided the surface temperature is below 0°C. If these parameters intersect below the 400 $\mu m$ $\tau = 1$ isopleth, only partial melting is predicted.

An alternative to the Czys et al. (1996) technique considers only the warm layer maximum temperature. Because the depth of the warm layer is generally proportional to the maximum warm layer temperature, this simplification has proven to be a good diagnostic to determine the degree of melting (Baumgardt, 1999). Stewart and King
(1987) modeled the melting of snow as it falls through an elevated warm layer based on the empirical rates of snowflake melting described by Stewart (1985). The authors ran this model for various snowflake sizes, densities, maximum warm layer temperatures, and lapse rates. Evaporation and condensation were not considered, as the layer was assumed just saturated. Figure 3.4 presents these results and Table 3.5 summarizes the degree of melting as a function of maximum warm layer temperature. For the

![Figure 3.4](image_url)

Figure 3.4: The degree of snowflake melting as a function of maximum warm layer temperature. The curves correspond to particle size expressed in terms of the raindrop-equivalent diameter, and for two different initial snowflake densities. Adapted from Stewart and King (1987).

... case of complete melting (corresponding to a warm layer temperature > 3 – 4°C), it is
<table>
<thead>
<tr>
<th>Max. T of Warm Layer</th>
<th>Degree of Melting</th>
</tr>
</thead>
<tbody>
<tr>
<td>&gt; 3 – 4°C</td>
<td>Melts Completely</td>
</tr>
<tr>
<td>1 – 3°C</td>
<td>Partial Melting</td>
</tr>
<tr>
<td>&lt; 1°C</td>
<td>Very Little Melting with Re-freezing to Snow</td>
</tr>
</tbody>
</table>

Table 3.5: Summary of results from Stewart and King (1987), relating the degree of snowflake melting in an elevated warm layer to the maximum temperature of the warm layer.

also important to consider the temperature of the cold layer beneath the warm layer (McNulty, 1988). If the minimum temperature of the cold layer is greater than $-10^\circ$ C, it is unlikely that the freezing process can be initiated so the precipitation will remain as a liquid until it reaches the ground. Provided the ground temperature is below freezing, this will result in freezing rain. When the minimum temperature of the cold layer is less than $-10^\circ$ C, it is more likely that freezing will be initiated and snow or sleet will result provided there are enough ice nuclei. The depth of the cold air will determine the subsequent degree of re-freezing and, thus, whether the precipitation will be snow or sleet.

Summary

The precipitation type analyses described above are best performed with the use of modeled or observed soundings. Because observed sounding data is usually only available at 12-hour intervals, it may be useful to infer changes in the lower tropospheric thermal structure between balloon launches. Penn (1957) discussed the influences on
the evolution of a temperature profile: thermal advection, adiabatic temperature change associated with vertical displacement, and non-adiabatic heating or cooling. Thermal advection by the horizontal wind is generally the dominant process, however this can act in opposition to an adiabatic temperature change. Because the horizontal advection of warm air is generally associated with upward vertical motion, warming associated with the advection can be reduced by adiabatic cooling associated with the rising air parcels. The net effect can be a reduction in the rate of warming by advection alone by roughly 50%. Additionally, diabatic processes can influence the temperature profile. Evaporational cooling when precipitation falls through a dry layer can reduce the temperature by 5-10°C per hour until the air is saturated and the temperature reaches the wet bulb temperature. In very heavy snow events, the freezing level may also be altered by the absorption of latent heat when snow is melted to rain in a warm layer.

Figure 3.5 presents a summary of the steps involved in the diagnosis of precipitation type. This figure illustrates the complexity of the physical processes involved in determining precipitation type. Given forecasted and observed soundings, the ideas presented above will allow forecasters to perform a reasonable assessment of the likelihood that precipitation will be snow, ice pellets, freezing rain, or rain. However, in order to include a determination of precipitation type on isobaric surfaces with the other ingredients, the analysis is limited to diagnostics that can be computed at one level or layer in the atmosphere. Thus, in the ingredients maps, the 850 hPa 0°C “rain-snow line” will be used as a rough guideline for a determining precipitation type. This guideline is only reliable when the forecast region is more than a few hundred kilometers to either
1. Temperature Decreases Monotonically with Height

\[
\begin{align*}
\text{Wet Bulb Temperature} & < 0 \text{ C} & \text{Wet Bulb Temperature} & > 0 \text{ C} \\
\text{everywhere} & & \text{anywhere} \\
\downarrow & & \downarrow \\
\text{SNOW} & & \text{RAIN or SNOW}
\end{align*}
\]

McNulty (1988):

<table>
<thead>
<tr>
<th>Freezing Level</th>
<th>Chance that Precipitation Will be Snow</th>
</tr>
</thead>
<tbody>
<tr>
<td>35 mb or 950 ft</td>
<td>50 %</td>
</tr>
<tr>
<td>25 mb or 660 ft</td>
<td>70 %</td>
</tr>
<tr>
<td>12 mb or 315 ft</td>
<td>90 %</td>
</tr>
</tbody>
</table>

2. Elevated Warm Layer

\[
\begin{align*}
\text{Complete Melting*} & & \text{Partial Melting*} \\
\text{Cold Layer T} & > -10 \text{ C} & \text{Cold Layer T} & < -10 \text{ C} \\
& \text{(no ice nucleation)} & \text{(possible ice nucleation)} \\
\downarrow & & \downarrow \\
\text{Sfc T} & > 0 \text{ C} & \text{Sfc T} & < 0 \text{ C} \\
\downarrow & & \downarrow \downarrow \\
\text{RAIN} & & \text{FRZ RAIN} & & \text{SNOW or ICE PELLETS}
\end{align*}
\]

* The degree of melting is determined by the Cyzs et al. (1996) or Stewart and King (1988) relationships based on warm layer temperature and warm layer depth.

Figure 3.5: Summary of the steps involved in a diagnosis of precipitation type. See text for explanation.

side of the 0°C isotherm and more than a few degrees cooler than 0°C, and will be used in conjunction with an analysis of forecasted and observed soundings when the forecast area is any closer or any warmer.
3.2 Application of the Ingredients-Based Methodology

The five ingredients for winter season precipitation examined in this thesis (QG forcing for ascent, moisture, instability, efficiency, and temperature) can help forecasters predict the precipitation potential, intensity, duration, and type. Table 3.6 summarizes the diagnostics developed in the previous section for evaluating each ingredient. The quantities required to compute these diagnostics are readily available to the operational community\(^3\), and the IBM is in operation at several NWSFOs in the central U.S. Appendix A contains a few examples of forecast discussions from the NWSFO in Dousman, Wisconsin, illustrating the incorporation of the IBM into daily operations.

Section 3.2 discusses the application of the IBM through the use of isobaric ingredients maps (section 3.2.1) and ingredients cross-sections (section 3.2.2). Once the ingredients have been analyzed on pressure surfaces and for the necessary cross-sections, the IBM enables forecasters to address the following questions:

- When will precipitation begin? When will it end?
- What will be the intensity for each 6-hour period? Are potentially higher intensities possible or expected in certain regions due to areas of instability or localized stronger forcing?
- What form will the precipitation take?

\(^3\) A web page, http://speedy.meteor.wisc.edu/~swetzel/winter/winter.html, has been constructed to facilitate operational use of the IBM. This site includes descriptions of the five ingredients, a detailed discussion of the steps required to prepare an ingredients-based forecast for winter precipitation, and case studies which illustrate this application of the IBM. It also provides access to GEMPAK, NTRANS, and UNIX scripts to create the ingredients maps and cross-sections.
<table>
<thead>
<tr>
<th>Ingredient</th>
<th>Diagnostic</th>
<th>Threshold Values</th>
</tr>
</thead>
</table>
| Forcing    | $\nabla \cdot \vec{Q}$ \( (Km^{-2}s^{-1} \times 10^{-15}) \) | -1 to -5 $\rightarrow$ weak forcing  
-5 to -15 $\rightarrow$ moderate forcing  
$< -15$ $\rightarrow$ strong forcing |
| Moisture   | Relative Humidity 700-750 hPa Mixing Ratio | $> 80 \%$ 
x 2 = Garcia “normal conditions” 12-hour snowfall |
| Stability  | $PV_{cs}$ | $< 0$ for instability (CI or CSI) |
| Efficiency | Cloud-level Temperature | Maximum depositional growth at -15 °C  
Ice crystal nucleation at -10 to -15 °C |
| Temperature | 850 hPa T | 0° C very rough rain/snow line |

Table 3.6: Summary of ingredients and diagnostics for forecasting winter season precipitation.

- What is the areal extent of the precipitation?

- Are there boundaries of moisture, forcing, or instability that could dramatically affect the precipitation distribution if the actual features shift slightly from the model-predicted values?

In addition, the IBM can be integrated with a traditional technique, such as the Garcia Method, to estimate snowfall accumulation (section 3.2.3).

In order to take advantage of the flexibility provided by the IBM, it is important
that forecasters approach this technique from a physical perspective by developing an understanding of the processes involved in every forecast. Additionally, because the vertical motion ingredient diagnostic relies on the quasi-geostrophic assumption and on all ingredient diagnostics rely on the accuracy of the model forecast, one must continually compare the model-generated ingredient parameters with observations. Since the precipitation generation ingredients are so closely tied to the synoptic scale, quantities such as sea level pressure, thickness, and temperature should also be monitored to assess model verification. Variations in these quantities from the model-predicted values could lead to changes in the strength, timing, or location of the precipitation patterns.

3.2.1 Isobaric Ingredients Maps

Use of the IBM can be facilitated by the construction of ingredients maps that display all ingredient diagnostics together in a convenient manner. This section introduces the ingredients maps with an example from a convective snow event that occurred in southeastern Wisconsin on January 26-27, 1996. Experience has shown that the application of the IBM for three thin layers in the lower- to mid-troposphere (800-850 hPa, 700-750 hPa, and 600-650 hPa) best captures the distribution of the ingredient parameters. For storms with a deep sea level pressure minimum or intense upper level dynamics, the 500-550 hPa analysis has proven informative. However, there may be features in between these levels that are not captured by such an analysis. These features can often be identified by evaluation of ingredient cross-sections as discussed in section 3.2.2.
Figures 3.6-3.8 show the ingredients maps for the 24-hour NCEP-ETA model forecast valid at 0Z on January 27, 1996, for 800-850 hPa, 700-750 hPa, and 600-650 hPa, respectively. These maps include the model-predicted QPF as well as information about the ingredients for winter season precipitation that can help diagnose the mechanisms responsible for this QPF. The maps enable forecasters to determine where precipitation will occur and the intensity of the precipitation, and to obtain an estimate of precipitation type. Also, the forecasted mean sea level pressure on the ingredients maps provides a useful reference for relating synoptic-scale features to the ingredient diagnostics. The 24-hour forecast of mean sea level pressure valid at 0Z on January 27, 1996 (shown in Figs. 3.6c, 3.7c, and 3.8c), contained a well-developed mid-latitude cyclone centered just south of the Wisconsin-Illinois border. This forecast will be used in the remainder of this section to introduce the use of ingredients maps in the IBM for winter season precipitation.

If an area of vertical motion forcing coincides with relative humidity values of 80% or greater, some precipitation is likely. The ingredients maps valid at 0Z on January 27 (Figs. 3.6-3.8) show that QG forcing overlapped the contours of relative humidity greater than 80% over all of Wisconsin except the northwestern corner. For this storm, this agreement was observed throughout the 850-600 hPa atmospheric layer. In some other cases, significant variation in the vertical distribution of moisture requires additional consideration. Figures 3.9a-b show that at 0Z and 1Z on January 27, 1996, precipitation was indeed reported throughout most of Wisconsin with the exception of the far northwestern portion of the state.
Figure 3.6: 800-850 hPa ingredients map from the 24-hour forecast of the NCEP-ETA model run valid at 0Z on January 27, 1996. (a) Model-predicted precipitation labeled in inches of liquid equivalent (shaded contours). 850 hPa +4 °C, 0°C, and -4 °C isotherms in blue. (b) Q-vector convergence (solid lines) and negative $PV_{cv}$ (shaded) in the 800-850 hPa layer. Q-vector convergence is contoured every $-5 \times 10^{-15} \text{Km}^{-2} \text{s}^{-1}$ beginning with -5 (the -1 contour is also included). $PV_{cv}$ is shaded every 0.15 PVU beginning at 0 PVU. (c) Mean sea level isobars (solid lines) and 800-850 hPa layer average temperature (shaded). Isobars contoured every 4 hPa. Isotherms shaded every 2 °C from 0 °C to -20 °C. (d) Green contours are QPV (see text for explanation). Layer average relative humidity is shaded for values $\geq 70\%$ (contour interval 10%), and layer average mixing ratio (red lines) is 1 g kg$^{-1}$. 
Figure 3.7: As in Fig. 3.6 except for the 700-750 hPa layer.
Figure 3.8: As in Fig. 3.6 except for the 600-650 hPa layer.
Figure 3.9: Surface observations for Wisconsin at a) 0Z and b) 1Z on January 27, 1996. Surface temperature (dew point) is shown in °C at the upper (lower) left of each station marker, wind speed is indicated by the barbs where each barb is 10 ms$^{-1}$, and precipitation type and intensity is shown directly to the left of the station, using conventional symbols.
The intensity of precipitation is related to the strength of the forcing and may be limited by the availability of moisture. Additionally, if the forcing coincides with an area of weak stability, an enhanced response to the forcing with higher precipitation rates can be expected. An area of instability at 600-650 hPa in southeastern Wisconsin (Fig. 3.8b) coincided with moderate to strong forcing at 0Z on January 27 and can be seen as a QPV maximum in Fig. 3.8d.

Where nonzero QPV overlaps sufficient moisture, heavy precipitation and possibly thunderstorms can occur. Although the region of positive QPV in Fig. 3.8d was close to the boundary of sufficient moisture, with only 70-80% relative humidity predicted for Milwaukee, moisture was abundant at lower levels and the strong vertical motions at 850 hPa and 700 hPa would have supplied the 600 hPa layer with ample moisture. Thus, based on these ingredients maps, heavy precipitation and possible convection would be expected in southeast Wisconsin. Observations from this time indicate that thundersnow and heavy snow were indeed reported in the Milwaukee area at 0Z and 1Z (Fig. 3.9).

Precipitation intensity can also be modulated by the efficiency ingredient. The 600 hPa and 700 hPa temperature in the ingredients maps can be used to assess the microphysical characteristics. If a region with sufficient moisture and upward vertical motion coincides with the temperature of maximum depositional ice crystal growth (−15°C) enhanced precipitation rates may result. In this case, the 600-650 hPa layer average air temperature (Fig. 3.8c) in the vicinity of the QPV maximum in southeast Wisconsin was -15°C to -16°C, providing additional evidence of the potential for high
precipitation intensity.

The temperature profile will influence the type of precipitation, however a rough characterization of the precipitation type can be inferred from the 850 hPa 0°C isotherm. A more rigorous approach to determining precipitation type would involve an analysis of forecast and observed soundings. The location of the 850 hPa 0°C rain-snow line at 0Z on January 27, 1999, predicted that the precipitation at this time should be snow throughout Wisconsin. Observations show that the precipitation was indeed snow throughout the state. However, the surface temperature at 1Z was +1 °C at a few locations in southeastern Wisconsin (Fig. 3.9b), and Milwaukee experienced a brief changeover to light rain from 2Z to 3Z with a temperature of +3°C at 2Z. This occurrence highlights the importance of evaluating soundings and other upper air observations throughout the forecast period to anticipate changes in precipitation type.

By inspecting ingredients maps frequently at a range of atmospheric levels, forecasters can use these diagnostics to help anticipate the intensity, onset time, and end time of a winter precipitation event. Because of the amount of information contained in the ingredients maps, it may be helpful to use a table to organize the ingredient parameters and to insure a systematic evaluation of each. Table 3.7 shows an example of such a chart (for a 24 hour forecast) which could be used for this purpose. Such a table is also useful as a means of documenting the storm for archival investigations of the IBM.

The analysis of the ingredients maps requires considerable subjective judgment, however, certain guidelines have been found to apply in most situations. Relative humidity greater than 80% is generally sufficient moisture to generate at least some precipitation,
### Table 3.7: Example of a table for organizing the values of some winter precipitation ingredient parameters for the 0- to 24-hour forecasts of a numerical model.

<table>
<thead>
<tr>
<th>Model Run</th>
<th>Forecast Area</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Forcing</td>
</tr>
<tr>
<td>UTC</td>
<td>Level</td>
</tr>
<tr>
<td>(hPa)</td>
<td>(g kg$^{-1}$)</td>
</tr>
<tr>
<td>f00</td>
<td>850</td>
</tr>
<tr>
<td></td>
<td>700</td>
</tr>
<tr>
<td></td>
<td>600</td>
</tr>
<tr>
<td>f06</td>
<td>850</td>
</tr>
<tr>
<td></td>
<td>700</td>
</tr>
<tr>
<td></td>
<td>600</td>
</tr>
<tr>
<td>f12</td>
<td>850</td>
</tr>
<tr>
<td></td>
<td>700</td>
</tr>
<tr>
<td></td>
<td>600</td>
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</tr>
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<tr>
<td></td>
<td>700</td>
</tr>
<tr>
<td></td>
<td>600</td>
</tr>
</tbody>
</table>
provided a forcing mechanism is in place. QG forcing as diagnosed by the Q-vector divergence can be classified as weak, moderate, or strong as defined in Table 3.1. With sufficient moisture and no instability, weak, moderate, and strong forcing will generally correspond to light, moderate, and heavy precipitation, respectively. However, the intensity of precipitation will vary considerably in the presence of instability or with moisture limitations. Instability at any level with ample moisture and at least weak forcing can result in heavy precipitation, possibly accompanied by thunder and lightning. Additionally, the depth of the atmospheric layer involved in the precipitation event may have a significant impact on the intensity of the precipitation. A very shallow moist layer will typically produce less precipitation than a situation with ample moisture through 500 hPa.

3.2.2 Cross-Sectional Analysis of the Ingredients

In addition to using the IBM on pressure surfaces, it is important to consider the vertical distribution of ingredients because a layer of instability or forcing could exist at a level that is not captured by any of the standard ingredients maps generated regularly (600-650 hPa, 700-750 hPa, 800-850 hPa). A diagnosis of such a situation is presented below, followed by an example employing a cross-sectional analysis to determine whether to expect a conditional gravitational or conditional symmetric response in a region of instability.
**Ingredients Cross-Section for a Low-Level Instability**

On February 12, 1999, following the passage of a strong cold front through the Midwest, very cold air (-18 °C at 850 hPa) flowed over a residually warm and moist ground from the previous day’s storm (Milwaukee, Wisconsin had a high temperature of 68 °F and 0.55 inches of rain on February 11, 1999). The warm surface temperature beneath cold air aloft established a shallow low-level instability as will be shown in this section. The dry air and high winds facilitated the transfer of moisture from the wet ground to the atmosphere, providing a layer of ample moisture in the vicinity of the instability. With weak forcing the instability was realized, resulting in a convective snow event. It is interesting to note that this event was forced by mechanisms similar to those associated with lake-effect snow, namely a large air-surface temperature difference and a source of moisture at the ground⁴. The radar from 14Z on February 12 (Fig. 3.10) shows isolated bands of heavy snow (near white-out conditions were observed in Madison at this time) and the satellite image from 19Z (Fig. 3.11) indicates an unstable convective boundary layer with cellular convection and roll clouds.

The instability associated with this event in southern Wisconsin was confined to a very low layer and did not appear on the standard level ingredients maps. Figure 3.12b shows that the 800-850 hPa 18-hour ETA model ingredients map valid at 18Z on February 12 did not indicate any areas of negative \( PV_{es} \) over Wisconsin and only weak scattered QG forcing. However, a cross-section drawn across southern Wisconsin from

---

⁴ Events of this nature were the subject of a study in New England (Lundstedt, 1992). This led to the development of a Wintertime Instability inDEX (WINDEX) that predicts the potential for snow squalls over land based on the low-level temperature profile, boundary layer relative humidity, and lifted index.
Figure 3.10: WSR-88D radar reflectivity from Madison, Wisconsin, at 14Z on February 12, 1999. Echo intensity contoured by the legend shown.
Figure 3.11: Visible satellite image of the Upper Midwest from 19Z on February 12, 1999.
Figure 3.12: As in Fig. 3.6, except from the 18-hour NCEP-ETA model forecast valid at 18Z on February 12, 1999. The thick solid line in (a) represents the location of the cross-section shown in Fig. 3.13.
LaCrosse to Milwaukee (location shown in Fig. 3.12a) for 18Z on February 12 shown in Fig. 3.13 reveals a layer of negative $PV_{es}$ between 870-950 hPa, with $-\frac{\partial PV_{es}}{\partial p} < -5 \text{Km}^{-1}$ just above the surface and low level relative humidity greater than 70%. Given even a slight forcing, such instability could have been realized. Although little QG forcing is predicted in this cross-section, scattered weak QG forcing can be seen on the 950 hPa surface (not shown) throughout southern Wisconsin around this time. Furthermore, the air temperature just above 900 hPa (which corresponded to the level of instability) was -15 °C, the ideal temperature for maximum depositional growth of ice crystals. If the instability was realized, strong vertical motions occurred in a region susceptible to rapid ice crystal growth. Thus, when analyzed from an ingredients perspective using cross-sections, this 18-hour forecast provides significant clues to the potential for the scattered heavy snow showers that passed through Wisconsin in the wake of the strong cold front.

**Ingredients Cross-Section for a CSI Case**

Cross-sections can also be used to distinguish between regions of CI and CSI as will be illustrated for a snowstorm which covered the southern half of Wisconsin with 3-6 inches of snow in the 12 hours between 18Z on January 23, 1996 and 6Z on January 24, 1996. Up to eight inches were reported in a few east-central Wisconsin counties. Convective bands of moderate to heavy snow were reported during the afternoon of January 23 in southern Wisconsin, spreading to central Wisconsin during the evening. This was documented in the Dousman, Wisconsin NWSFO storm report as a “CSI event” (Haase,
Figure 3.13: Ingredients cross-section from LaCrosse to Milwaukee, Wisconsin (location shown in Fig. 3.12a) from the 18-hour forecast of the NCEP-ETA model valid at 18Z on February 12, 1999. a) Negative $P' \rho c$ shaded (in PVU according to the legend at right) and Q-convergence contoured every $-1 \times 10^{-15}$ K m s$^{-2}$ s$^{-1}$. b) $-\frac{\partial \theta_{es}}{\partial z}$ shaded (in K m$^{-1}$), $M_g$ solid (in m s$^{-1}$), and $\theta_{es}$ dashed (in °K). c) RH dotted, and shaded where $> 70\%$, mixing ratio solid, contoured every 0.5 g kg$^{-1}$. d) Temperature solid, contoured every 2 °C, and shaded where $>-12$ °C.
An inspection of ingredients maps from the NCEP-ETA model run initialized at 0Z on January 23, 1996 provides clear signs of the potential for CSI during this period. Figure 3.14 shows the 800-850 hPa ingredients map from the 18-hour forecast valid at 18Z on January 23. A region of negative $PV_{es}$ was present in the southern third of Wisconsin at this time (Fig. 3.14b). Toward western Wisconsin, there was weak to moderate forcing for ascent (Fig. 3.14b) coincident with the region of instability and thus a nonzero value of QPV (Fig. 3.14d). In a northwest to southeast-oriented cross-section (drawn perpendicular to the 500-1000 hPa thickness contours from Richland to Walworth counties in southern Wisconsin (location shown in Fig. 3.14a), two layers of instability can be seen (Fig. 3.15a). The lower layer of instability from 800-850 hPa corresponded to a region of gravitational stability since $-\frac{\partial M_g}{\partial p} > 0$ (Fig. 3.15b). Furthermore, the contours of geostrophic momentum ($M_g$) were less steeply sloped than the contours of $\theta_{es}$, indicating that CSI was possible because the flow was nearly two-dimensional. With weak to moderate forcing (Fig. 3.15a) and relative humidity of 90-100% (Fig. 3.15c), this instability was realized. Due to the limited moisture above 750 hPa, however, very heavy precipitation or thundersnow is not likely. Instead, banded shallow moist convection was expected. This is precisely what was observed, as shown by the radar from 18Z on January 23 in Fig. 3.16. The upper layer of negative $PV_{es}$ between 550 and 650 hPa corresponds to CI because $\theta_{es}$ decreases with height in this layer as shown in Fig. 3.15. However, the relative humidity at 600 hPa is only 50-60% so it is unlikely that this instability was realized.
Figure 3.14: As in Fig. 3.6 except from the 18-hour forecast of the NCEP-ETA model valid at 18Z on January 23, 1996. The thick solid line in (a) represents the location of the cross-section shown in Fig. 3.15.
Figure 3.15: As in Fig. 3.13 except from Richland to Walworth counties, Wisconsin (location shown in Fig. 3.14a) from the 18 hour NCEP-ETA model forecast valid at 18Z on January 23, 1996.
Figure 3.16: WSR-88D radar reflectivity from Moline, Illinois, at 18Z on January 23, 1996. Echo intensity contoured by the legend shown.
3.2.3 Snowfall Accumulation Estimates

The IBM for forecasting winter season precipitation does not independently provide a prediction of snowfall accumulation. In order to make a quantitative snowfall estimate, the ingredient information must be used in conjunction with a traditional technique. The Garcia Method (Garcia, 1994; hereafter GM) was chosen for this purpose because of its comprehensive treatment of the moisture ingredient and because it has been widely accepted as a reliable technique for heavy snow prediction (e.g., Nietfeld and Kennedy, 1998; Gordon, 1998; Gerard et al., 1998; Cobb and Albright, 1996).

Despite its popularity, the applicability of GM is somewhat limited because it is an empirical technique and fails to accurately predict the snowfall accumulation in many situations. Furthermore, GM was designed to answer the question of “how much,” but not the question of “where” (Garcia, 1994). It predicts the maximum accumulation for a region that has been pre-defined as an “area of concern” for snow. The rules of thumb recommended by GM for placement of the “area of concern,” and, thus, the location of the heaviest snow band, include:

- 120 to 150 miles left of the surface low track.
- Along the track of the 700 hPa and/or 500 hPa low.
- Between the -2°C to -5°C isotherms at 850 hPa.
- Near the mean 500 hPa temperature of -30°C.
- Just north of the 164 height contour at 200 hPa.
From an ingredients perspective, this area of concern constitutes the area where forcing for ascent is expected. By leaving the diagnosis of this element to forecasters, GM lends itself nicely to an incorporation with the IBM. In fact, Garcia (1994) states that “the isentropic forecast procedure outlined in this paper is not a stand-alone technique but should be part of a comprehensive approach.” In this thesis, the physical basis and flexibility of the IBM is incorporated with the quantitative prediction ability of GM to obtain reasonable estimates of snowfall over a broad region. Before outlining our procedure for integrating GM with the IBM, a brief discussion of GM, its limitations, and some proposed modifications are presented.

The Modified Garcia Method

The GM predicts that the maximum snowfall (measured in inches) over a 12-hour period in the area of concern is simply equal to twice the average mixing ratio (in g kg$^{-1}$) in the 700-750 hPa layer that area. An operational limitation of this approach arises from the ambiguity present in the definition of the 12-hour “average mixing ratio.” Strictly speaking, this technique prescribes that the current mixing ratio ($l_{t\rightarrow t_0}$) should be averaged with the maximum mixing ratio ($l_{\text{max}}$) that could be advected into the area of concern over a 12-hour period. $l_{\text{max}}$ is found by locating the maximum mixing ratio upstream of the area of concern on the isentropic surface that intersects the area of concern between 700-750 hPa within a distance $D$, where $D$ is defined as (wind speed at $t = t_0$)$x$(12 hours).

Because gridded model data and analysis programs are now readily available, the
extrapolation from current observations for a 12-hour period is no longer necessary. Instead, model-predicted values of mixing ratio can be evaluated at the shortest time interval provided by the model throughout the 12-hour forecast period to estimate the maximum forecasted mixing ratio within that 12-hour period. Results from this modified approach can differ significantly from those computed according to the technique originally proposed by Garcia (1994). Figure 3.17 illustrates this difference using 30- and 42-hour ETA model forecasts valid at 6Z and 18Z on December 15, 1996. Figure 3.17a shows the change in mixing ratio on the 290 °K isentropic surface between the initial value (at 6Z on December 15, 1996) and the model-predicted mixing ratios on the same isentropic surface 12 hours later (18Z on December 15, 1996), where

\[ \Delta l = l_{t=t_0+12} - l_{t=t_0} \]

The 290°K isentropic surface was chosen because it intersects northwestern Wisconsin between 700 and 750 hPa. Figure 3.17a indicates a model-predicted change (\( \Delta l \)) in mixing ratio over northwest Wisconsin of 0 to -1 g kg\(^{-1}\). In contrast, the original GM predicts an increase of 3-4 g kg\(^{-1}\) over the 12-hour period, as illustrated in Fig. 3.17b. This GM mixing ratio change was computed by multiplying the initial mixing ratio advection on the 290°K isentropic surface by 12 hours, an equivalent procedure to the extrapolation suggested by GM:

\[ \Delta l_{GM} = -\vec{V} \cdot \nabla l_{t=t_0} \times (12)(3600) \]

While the original GM assumes that the initial mixing ratio advection remains constant over a 12-hour period, the modified approach includes storm motion indirectly
Figure 3.17: a) 12-hour change in mixing ratio on the 290 °K isentropic surface between 6Z and 18Z on December 15, 1996. Change is labeled in g kg\(^{-1}\) and shaded at intervals of 1 g kg\(^{-1}\) according to legend at left. b) (12-hours)x(mixing ratio advection at 6Z on December 15, 1996), labeled as shaded in (a). Values computed from the NCEP-ETA model forecast initialized at 0Z on December 14, 1996. Isobars (solid black) are labeled in hPa and contoured every 50 hPa. Wind vectors on the 290 °K surface are shown as white arrows where speed is proportional to vector length.
by computing the model-predicted change over 12 hours. It follows that the differences between the model-predicted mixing ratio change ($\Delta l$, Fig. 3.17a) and the mixing ratio change by advection ($\Delta l_{GM}$, Fig. 3.17b) can be largely attributed to the storm motion during the 12-hour period. The differences between the two approaches are often the most notable in the northwest quadrant of the storm. In this region, the initially positive mixing ratio advection becomes negative during the intervening 12 hours as the storm moves past the forecast area and northwest winds usher in drier air.

The ambiguities associated with the original GM mixing ratio computation result in its being applied in a non-uniform manner among forecasters. In fact, many forecasters use only the initial mixing ratio to perform the analysis, dismissing the calculation of average mixing ratio over a 12-hour period altogether. In the interest of defining a consistent approach to computing average mixing ratio in this thesis, an average between the initial mixing ratio and the maximum mixing ratio forecasted during the 12-hour period over the area of concern will be used$^5$. Experience has shown that GM provides more reliable estimates of snowfall accumulation when the mixing ratio is computed in this manner. Hereafter, references to GM will implicitly include this modified mixing ratio calculation.

Isentropic surfaces are a valuable analysis tool because mixing ratio is conserved on such surfaces in the absence of precipitation. However, with the modification to the mixing ratio calculation just described the GM no longer involves manual extrapolation.

$^5$ The maximum forecasted mixing ratio is determined based on an inspection of gridded model data at the temporal resolution available. Typically, this will be every six hours. Although mixing ratio may be higher between these time intervals, this has proven to be effective.
of the mixing ratio advection, so isentropic analysis is not strictly necessary. Instead, the 700 hPa isobaric surface can be used to evaluate the mixing ratio\(^6\). By analyzing mixing ratio on an isobaric surface, the analysis procedure is simplified considerably because mixing ratio can be displayed along with the other ingredients. Additionally, the use of isobaric surfaces for this analysis is advantageous because it provides forecasters with an overview of the spatial distribution of moisture. The isentropic method provides a forecast for only a small region where the isentropic surfaces intersect the 700-750 hPa layer.

**Using the IBM with the Garcia Method**

Based on the preceding discussion, the following procedure incorporates GM into the IBM for estimating snowfall accumulation. First, the maximum forecasted mixing ratio at 700 hPa over the forecast area during a 12-hour period, from \(t = t_o\) to \(t = t_o + 12\) must be determined and averaged with the mixing ratio at \(t = t_o\). Then one can compute the GM “normal conditions” accumulated precipitation as:

\[
\text{Maximum Snowfall in inches} = 2 \times \text{(Average Forecasted Mixing Ratio)}
\]

Because it was determined empirically, this relationship holds only under “normal conditions” similar to those for the events used to formulate the original GM. Based on an evaluation of approximately ten case studies and close inspection of the original technique, it appears that these “normal conditions,” under which snow accumulation

\(^6\) The issue of whether the 700 hPa surface is well suited for this purpose has not been investigated in this study. Future work could investigate varying this level depending on individual storm characteristics.
is properly forecast by GM, are generally characterized by moderate forcing, no instability, a temperature profile corresponding to all snow (no rain or sleet), and a 10:1 snow to water ratio. The IBM can be used to assess whether a precipitation event meets these conditions; and, thus, whether it is necessary to adjust the original GM snowfall estimate. For example, a GM “normal conditions” snowfall estimate will be too low if the storm is characterized by strong forcing for ascent, instability, or unusually cold temperatures, and additional accumulations should be anticipated under these conditions. With only weak forcing for ascent, GM may overestimate snowfall and its prediction should be decreased.

To be of operational use, these ingredients-based adjustments to GM forecasts must be quantified. In a recent conference presentation, Garcia (1999) recommended an update of the original Garcia Method (Garcia, 1984) for “unusual cases,” i.e., very cold conditions, convective snowstorms, or jet couplet-induced heavy snow events. He suggested the use of the NWS “New Snowfall to Estimated Meltwater Conversion Table” (Friday, 1997), summarized in Table 3.8, to adjust the snowfall accumulation for very cold events. For example, if the surface temperature is between 19 °F and 15 °F (corresponding to a 20:1 snow to water ratio) for the duration of a snow event, the Garcia “normal conditions” prediction should be doubled. Furthermore, Garcia (1999) recommended a 4:1 ratio of snowfall to average mixing ratio for the coupled jet or convective events:

\[ \text{Maximum Snowfall in inches} = 4 \times \text{(Average Mixing Ratio)} \]

He specified that even though these events may not last a full 12 hours, the 4:1 ratio
Table 3.8: Snow to water ratio of frozen precipitation as a function of surface temperature, derived from the NWS “New snowfall to estimated melt water conversion table,” adapted from Friday (1997).

<table>
<thead>
<tr>
<th>Surface Temperature</th>
</tr>
</thead>
<tbody>
<tr>
<td>(° F)</td>
</tr>
<tr>
<td>(° C)</td>
</tr>
<tr>
<td>34 to 28</td>
</tr>
<tr>
<td>27 to 20</td>
</tr>
<tr>
<td>19 to 15</td>
</tr>
<tr>
<td>14 to 10</td>
</tr>
<tr>
<td>9 to 0</td>
</tr>
<tr>
<td>-1 to -20</td>
</tr>
</tbody>
</table>

However, this updated GM remains an empirical technique and is subject to many of the same limitations as the original GM because it still does not explicitly include a physical basis. We can combine the updated GM with the IBM by inferring that the jet couplet-induced heavy snow corresponds to strong forcing, and the convective snow is indicative of instability. This explicitly considers the physical reasoning underlying the 4:1 modified snowfall to mixing ratio relationship and assists in evaluating whether this adjustment is necessary. For example, when the forcing or instability ingredients indicate the potential for heavy snow \((\nabla \cdot \vec{Q} < -15 \times 10^{-15} \text{Km}^{-2}\text{s}^{-1})\) or \(PV_{es} < 0\) the original Garcia “normal conditions” estimate may underpredict the snowfall accumulations so a 4:1 ratio of snowfall to mixing ratio should be used. In this manner, the Garcia
(1999) update provides a convenient framework for quantifying the ingredients-based adjustments to the original GM predictions.

For the example of convective precipitation presented in section 3.2.2 (Fig. 3.14), the 700 hPa mixing ratio at 18Z on January 23, 1999 was 1-2 g kg^{-1} in the region of the snow band. The mixing ratio remained fairly constant at 700 hPa through the next 12 hours, so the average mixing ratio was also 1-2 g kg^{-1}. This corresponds to a traditional GM maximum snowfall forecast of 2-4 inches. However, because the IBM indicates the potential for convective snowfall, the GM predicted accumulations should be increased. If the 4:1 ratio of snowfall to mixing ratio is applied instead of the original 2:1 ratio, 4-8 inches is predicted. This prediction is in good agreement with the observed snowfall.

3.3 Case Study – March 13-14, 1997

On March 13, 1999, a late winter snowstorm surprised forecasters in Wisconsin by producing a narrow band of up to 13 inches of snow between 4Z and 12Z, several hours before a more significant event was predicted to begin. In the next 24 hours, moderate and occasionally heavy snow with isolated convection was observed, contributing to storm total accumulations of 15 to 30 inches over a large portion of the state. Figures 3.18 a-c show the observed snowfall accumulations in Wisconsin over 3 periods: before 12Z on March 13, between 12Z on March 13 and 12Z on March 14, and for the whole 32-hour event, respectively. The snowfall amounts shown in Fig. 3.18 are considerably
Figure 3.18: Wisconsin snowfall total (in inches) for a) the 8-hour period ending at 12Z on March 13, 1997, b) the 24-hour period between 12Z on March 13 to 12Z on March 14, and c) the full 32-hour event from 4Z on March 13 to 12Z on March 14. Snowfall contoured at 4-inch intervals in (a) and 6-inch intervals in (b) and (c).
higher than predicted\(^7\), and the NWSFO storm report for this event noted that “forecasters seem to have a difficult time with these [convective] situations, as history has shown, and snowfall amounts were well under-forecasted” (Haase, 1997).

The remainder of this section uses the March 13-14, 1997 snowstorm (hereafter, the March 1997 storm) to illustrate the application of the IBM to forecasting winter season precipitation. Following a short synoptic overview, the early onset of precipitation that led to the initial unpredicted band of snow is investigated from an ingredients perspective. The IBM will then be used to deduce a time series of precipitation type and intensity for this case. Finally, the transition from snow to freezing rain for a station in northeastern Wisconsin will be investigated.

### 3.3.1 Synoptic Overview

Surface cyclones that originate over the Oklahoma panhandle and track through Missouri, Indiana, and Michigan are called “panhandle-hook” type storms. The surface cyclone associated with the March 1997 storm followed such a path, as shown in Fig. 3.19. A standard rule of thumb used at the NWSFO located in Dousman, Wisconsin, for panhandle-hook type storms predicts that snow will begin falling in southern Wisconsin when the sea level pressure minimum reaches central Missouri. However, in this case the main surface low was still developing over the Oklahoma panhandle by the time Wisconsin had received up to 13 inches of snow (12Z on March 13). An analysis

\(^7\) The winter storm warning issued at 22Z on March 12 by the NWSFO located in Dousman, Wisconsin, called for a storm total of 6-12 inches of snow for the northern half of the state, to begin in the morning of March 13.
Figure 3.19: Observed track and minimum central pressure of the cyclone from 0Z on March 13 to 0Z on March 15, 1997.

of the mid- and upper-level dynamics along with an application of the IBM provides insight into the processes involved in this atypical storm evolution. Output from the operational NCEP-ETA model initialized at 0Z on March 13 was used throughout this investigation. The synoptic features in this model run (Figs. 3.20-3.23) verified well with observations (not shown) of a number of variables including sea level pressure, 850 hPa geopotential height and temperature, and 300 hPa geopotential height and wind speed. By using only NCEP-ETA model output, our analysis is constructed using information that was available to forecasters at the time of the snow event.

At 6Z on March 13, there were two distinct synoptic features: (1) a southern disturbance reflected as a weak surface low pressure minimum over Texas and Oklahoma, and (2) a northern system with a less obvious signature at the surface—only a weak
depression in the mean sea level pressure over Wyoming (Fig. 3.20c). However, there was a 300 hPa wind speed maximum (Fig. 3.20a), and a 500 hPa vorticity maximum (Fig. 3.20b) associated with this northern feature. The southern system was associated with a separate, but weak and poorly organized 500 hPa absolute vorticity maximum. The 300 hPa geopotential height field at 6Z 13 March reflected the presence of this southern disturbance through the trough in the height field, however it is clearly a separate feature from the more northern trough and its wind speed maximum.

By 0Z on March 14, the 300 hPa northern trough and wind speed maximum had intensified (Fig. 3.21a), and the associated 500 hPa vorticity maximum over North and South Dakota had similarly intensified (Fig. 3.21b). The sea-level pressure trough associated with these features had become slightly more pronounced as well, with an axis extending from Minnesota to northern Missouri. The Oklahoma cyclone had deepened slightly but remained more or less in the same location, a behavior that is consistent with the slow propagation of its associated weak upper level short wave trough.

Six hours later, at 6Z on March 14, the two systems had begun to interact. The 300 hPa trough over Nebraska continued to sharpen (Fig. 3.22a), and the northern and southern 500 hPa vorticity maxima remained relatively independent (Fig. 3.22b). However, by this time, the primary low pressure system was in a favorable region for development as it was located downstream of the intensifying upper trough (Fig. 3.22c). The minimum sea level pressure had fallen to 998 hPa, and the storm was centered over southern Illinois.

Finally, by 12Z on March 14, there was a well-organized mid latitude cyclone centered
Figure 3.20: 6-hour NCEP-ETA model forecast, valid at 06Z March 13, 1997 of a) 300 hPa wind speed (shaded in $\text{m s}^{-1}$ according to legend at left) and geopotential height (solid lines, contoured every 120 m). b) 500 hPa absolute vorticity (labeled in units of $10^{-15} \text{ s}^{-1}$ and shaded according to legend at left), and geopotential height (solid lines, contoured every 60 m), and c) 1000 hPa temperature (shaded every 4 $^\circ\text{C}$) and mean sea level pressure (solid lines contoured every 4 hPa).
Figure 3.21: As in Fig. 3.20 except for the 24-hour NCEP-ETA model forecast valid at 0Z on March 14, 1997.
Figure 3.22: As in Fig. 3.20 except for the 30-hour NCEP-ETA model forecast valid at 6Z on March 14, 1997.
over northeastern Indiana with a sea-level pressure minimum of 996 hPa (Fig. 3.23c). This feature was located downstream of the sharp 300 hPa and 500 hPa upper trough axes (Figs. 3.23 a and b). The 1000 hPa thermal structure also reflected the structure of this system, with the cold and warm frontal regions clearly visible (Fig. 3.23c).

### 3.3.2 Investigation of the Early Onset of Precipitation

The narrow band of snow which fell unexpectedly in southern Wisconsin before 12Z on March 12 covered a region 30 miles wide, from just south of LaCrosse to just south of Sheboygan (see Fig. 3.18). As mentioned earlier, the timing of this period of snow was not consistent with the traditional rules of thumb used to predict the onset of precipitation in panhandle-hook type storms. Application of these traditional rules assumes that the snowfall will be associated exclusively with the dominant low pressure system. In this case, the initial 12-18 hours of snowfall were largely associated with an upper level feature that was independent of the sea level pressure minimum.

The NCEP-ETA model relative humidity fields valid at 6Z on March 13, 1997 provide an excellent illustration of the existence of two distinct systems in the early stages of the storm (Fig. 3.24 a and b). In both the 700-750 hPa and 600-650 hPa layers, the relative humidity fields clearly indicate two separate features. One area of nearly saturated conditions in southern Missouri and Arkansas was associated with the developing surface low over the Oklahoma panhandle and its 500 hPa vorticity maximum over eastern Texas (see Fig. 3.20). This moisture likely originated in the Gulf of Mexico. In contrast, the high relative humidity air over Wisconsin, Minnesota, and North Dakota was associated
Figure 3.23: As in Fig. 3.20, except for the 36-hour NCEP-ETA model forecast valid at 12Z on March 14, 1997.
Figure 3.24: (a) As in Fig. 3.6d except without mixing ratio lines and for the 700-750 hPa layer from the 6-hour NCEP-ETA model forecast valid at 06Z March 13, 1997. The thick solid line in (a) represents the location of the cross-section shown in Fig. 3.25. (b) As for (a) but for the 600-650 hPa layer. (c) As in Fig. 3.6b except for the 700-750 hPa layer from the 6-hour NCEP-ETA model forecast valid at 06Z March 13, 1997. (d) As for (c) but for the 600-650 hPa layer.
with the northern system and was presumably of Pacific origin.

The relative humidity fields indicate that there was ample moisture at 700-750 hPa and 600-650 hPa in southern Wisconsin for precipitation (Figs. 3.24 a and b). However, the QG forcing and instability ingredients on the standard isobaric levels at this same time (Figs. 3.24 c and d) do not indicate a significant potential for heavy snow before 12Z on March 13. The mainly weak to moderate QG forcing ($\nabla \cdot Q = -1$ to $-5 \times 10^{-15} Km^{-2}s^{-1}$) over Wisconsin and the absence of instability ($PV_{cs} > 0$) suggested that any precipitation would probably be of light intensity.

A more complete picture of the ingredient distributions at 6Z on March 13 was obtained through a cross-sectional analysis. The cross-section, drawn from south of LaCrosse to south of Sheboygan in Wisconsin (location shown in Fig. 3.24a), reveals features that were not apparent on the isobaric ingredients maps and provide significant clues to the potential for high snowfall rates. Figure 3.25a shows that a layer of conditional instability ($PV_{cs} < -0.15$ PVU) existed in the cross-section between 600-500 hPa at 6Z on March 13. With sufficient moisture at these levels (RH = 90-100% at 600 hPa) and weak to moderate forcing between 500-700 hPa, this instability could be realized and heavy precipitation is possible. Furthermore, a consideration of the efficiency ingredient shows that the temperature in the 600-550 hPa layer (-12 °C to -18 °C) enabled the maximum depositional growth and condensation rates of ice crystals to occur, providing more support for a possible heavy snowfall event.

For this case, much of the information indicating the potential for heavy snow could have also been obtained by analyzing an isobaric ingredients map in the non-standard
Figure 3.25: As in Fig. 3.13 except from Lacrosse to south of Sheboygan, Wisconsin, (location shown in Fig. 3.24a) from the 6-hour forecast of the NCEP-ETA model data valid at 6Z March 13 1997.
layer, 550-600 hPa. Such an analysis reveals a QPV maximum over western Wisconsin at 6Z on March 13 (Fig. 3.26a), corresponding to a region of weak forcing and negative $PV_{es}$ (Fig. 3.26b). The 550-600 hPa QPV maximum coincided with ample moisture in west-central Wisconsin, indicating the potential for a high precipitation rate and convection at this time. Although much of the state was characterized by conditional instability in the 550-600 hPa layer, the heaviest snow occurred in a narrow band in southern Wisconsin where the relative humidity was greater than 80% throughout the 850-550 hPa layer.

The preceding application of the IBM to the diagnosis of the initial snow band in the March 1997 storm revealed clues for the potential for moderate to heavy snow in Wisconsin prior to 12Z on March 13. This analysis indicates that the NCEP-ETA model data available to forecasters prior to the storm contained information that could have alerted them to the possibility of the early snow event. Furthermore, the analysis provided another example of the importance of inspecting ingredients cross-sections in conjunction with the standard pressure level ingredients maps.

### 3.3.3 Ingredients-Based Forecast

The exercise of preparing a forecast for an event that has already occurred is invariably influenced by the hindsight knowledge of the actual verified conditions. However, there is value in discussing such a forecast because it illustrates techniques that could be used to anticipate the development of future storms. In this section, we discuss how an accurate forecast for precipitation duration, intensity, accumulation, and type could
Figure 3.26: (a) As in Fig. 3.24a except for the 550-600 hPa layer. (b) As for Fig. 3.24c except for the 550-600 hPa layer.
have been prepared prior to the March 1997 storm by applying the IBM. The forecast will focus on a region of central Wisconsin located 40 miles southwest of Green Bay, including the city of Appleton which received 21.7 inches of snow. Appleton is located just outside of the area affected by the early snow event, and 21 inches of its storm total accumulation fell in the 24 hours between 12Z on March 13 and 12Z on March 14.

**Forecast for Precipitation Duration and Intensity**

In order to prepare a forecast for the March 1997 storm, an evaluation of ingredient parameters from the NCEP-ETA model run initialized at 0Z on March 13, 1997 is performed. A complete application of the IBM would involve inspecting the ingredients maps at all forecast hours in the 800-850 hPa, 700-750 hPa, and 600-650 hPa layers. However, in this thesis, only a few ingredients maps are presented, and the remaining information is summarized in an ingredients table for the Appleton forecast area (Fig. 3.27).

The ingredients table in Fig. 3.27 shows that ample moisture (RH ≥ 80%) was predicted throughout the entire 850-600 hPa column for the 24-hour period between 12Z on March 13 and 12Z on March 14. Mid-level relative humidity was also forecasted to be 90-100% as early as 6Z on March 13. Some QG forcing in the Appleton area was predicted at 6Z on March 13 and continued through 6Z on March 14, with only scattered weak forcing after that time. Based on this information alone, one would expect snow to start between 6Z and 12Z on March 13 and end between 6Z and 12Z on March 14. However, because the moisture at 6Z on March 13 was limited to levels above
Figure 3.27: Completed ingredients table for the 0Z March 13, 1997 ETA model run. See text for explanation.
700 hPa\textsuperscript{8}, precipitation generated above 700 hPa may have sublimated or evaporated before it reached the ground.

An investigation of the strength of forcing, atmospheric stability, and precipitation efficiency will facilitate a forecast for precipitation intensity. The ingredients table in Fig. 3.27 indicates that by 18Z on March 13 and through 0Z on March 14, moderate to strong QG forcing was predicted throughout the lower- to mid-troposphere. The 700-750 hPa ingredients map shows \( \nabla \cdot Q = -5 \times 10^{-15} \frac{Km^{-2} s^{-1}} \) close to Appleton at 0Z on March 14 (Fig. 3.28b). Also at 0Z on March 14, an area of negative \( PV_{es} \) in the 700-750 hPa layer over the forecast area is noted in the ingredients table and shown in Fig. 3.28b. The combination of forcing for ascent and conditional instability corresponds to a positive QPV value (Fig. 3.28d) for the 700-750 hPa layer in an area with 80-90\% relative humidity (Fig. 3.28d). Finally, the layer-average temperature over Appleton at 0Z on March 14 was -4 to -6\(^\circ\)C (Fig. 3.28c) at 700-750 hPa and -10 to -12\(^\circ\)C at 600-650 hPa, indicating that the atmosphere was not cold enough to expect maximum depositional ice crystal growth.

Because of the strong forcing, ample moisture, and isolated areas of instability, snowfall intensity during the 6-hour period between 18Z on March 13 and 0Z on March 14 is expected to be higher due to possible convection. Surface observations at Appleton indicate that light snow began at 9Z on March 13, and moderate snow fell during the 14 hours between 15Z on March 13 and 5Z on March 14. WSR-88D radar from Green Bay, Wisconsin, shows embedded regions of high reflectivities (30-40 dBz) characteristic of

\textsuperscript{8} Although the 700 hPa relative humidity over Appleton is 80\%, this area was close to the dry edge of a strong moisture gradient, and below 800 hPa the relative humidity was less than 70\%
Figure 3.28: As in Fig. 3.6 except for 700-750 hPa from the 24-hour forecast of the NCEP-ETA model valid at 0Z on March 14, 1997.
convection from 20Z on March 13 to 3Z on March 14. A convective cell just south of Appleton at 23:13Z on 13 March (Fig. 3.29) corresponds very well with the location of the QPV maximum in Fig. 3.28d.

Figure 3.29: WSR-88D radar reflectivity from Green Bay, Wisconsin 23:13Z at March 13, 1997. Echo intensity (in dBz) is shaded according to the legend at the bottom.

By 12Z on March 14, no QG forcing is indicated in the ingredients table at 850 hPa or 700 hPa for the Appleton forecast area, and only weak QG forcing is predicted at 600 hPa (see Fig. 3.30b). Coincident with the scattered weak forcing at 600 hPa, there is
negative $PV_{es}$ (Fig. 3.30b) throughout much of the state at this level. The co-location of negative $PV_{es}$ and negative $\nabla \cdot Q$ is reflected in a few 600-650 hPa QPV maxima in central Wisconsin (Fig. 3.30d). The combination of some forcing and instability indicates the potential for convection at this time. However, because the QPV maxima were located near a boundary of sufficient moisture (RH < 70% is only about 50 miles west of Appleton at 700-750 hPa and 600-650 hPa), isolated snow showers of short duration would be more likely than heavy thundersnow.

**Precipitation Type Forecast**

Although the observed *surface* temperatures in Appleton, Wisconsin, never rose above -4°C, the forecasted 850 hPa temperature rose from -6°C at 18Z on March 13 to approximately -2°C at 0Z on March 14 before falling sharply to -8°C by 12Z (see the ingredients table in Fig. 3.27). Because of the proximity of the 850 hPa 0°C isotherm to the Appleton forecast area at 0Z 14 March (Fig. 3.28a), a forecast for precipitation type at this time requires the examination of temperature and dew point profiles. Prior to 0Z and after 12Z on March 14, the “rain-snow line” was sufficiently south of the forecast area so that all precipitation could be expected to fall as snow. The model-predicted sounding for Appleton valid at 0Z on March 14 (Fig. 3.31) revealed an elevated warm layer with a maximum temperature of -2°C at about 800 hPa. Below this warm layer, the temperature dropped below -8°C between 900-950 hPa and warmed slightly to -6 °C near the surface. The dew point followed the temperature profile throughout the 1000-700 hPa layer, although it was typically 1-3°C less. Because the maximum
Figure 3.30: As in Fig. 3.6 except for the 600-650 hPa layer from the 36-hour forecast of the NCEP-ETA model valid at 12Z on March 14, 1997.
Figure 3.31: Model-predicted skew-T profiles for Appleton, Wisconsin, from the 24 hour forecast of the NCEP-ETA model valid at 0Z March 14, 1997. Temperature is the thick solid line and dew point is the dashed line. Isopleths of $\theta$ (light dashed) and $\theta_e$ (dotted), and wind barbs (m/s) are shown. Abscissa is temperature in °C and isotherms slope upward to the right. Ordinate is pressure in hPa.

Warm layer wet bulb temperature remained everywhere below freezing, this sounding indicates that the precipitation at 0Z on March 14 would be all snow at Appleton. Surface observations throughout the storm indicate that Appleton did in fact receive snow throughout the event. Between 2Z and 4Z on March 14 (Figs. 3.32b-d), however, a station 15 miles south of Appleton reported a brief changeover to rain. Because of the surface temperature between -3°C and -4°C at this station, Oshkosh, we suspect that the precipitation was not rain as reported by the Automated Surface Observing
Figure 3.32: Central Wisconsin surface observations at a) 1Z, b) 2Z, c) 3Z, and d) 4Z on March 14, 1997. Observations are plotted as in Fig. 3.9. The cities referred to in the text, Green Bay, Appleton, and Oshkosh, are labeled in (a).
System (ASOS) at Oshkosh, but was actually freezing rain\footnote{Most ASOS stations had freezing rain sensors by 1997, however a test in 1994-1995, prior to installation of these sensors on ASOS units, revealed that the freezing rain sensors agreed with observations only 66\% of the time, as discussed in http://www.nws.noaa.gov/asos/frezrain.htm (accessed 12/99).}. The transition from snow to freezing rain at Oshkosh is investigated further in section 3.3.4.

Based on the application of the IBM presented so far in this section, a forecast for precipitation duration, intensity, and type for the Appleton area predicts light snow to begin before 12Z on March 13, with little accumulation before daybreak. Precipitation should continue throughout the day, with moderate to heavy snow falling by 18Z. Thundersnow is a possibility in the afternoon and evening, primarily between 18Z on March 13 to 0Z on March 14. Some areas may experience a period of rain after 0Z on March 14 before changing back to all snow by 12Z. The precipitation will decrease in intensity by 6Z on March 14 with only light snow or rain continuing through 12Z on March 14, and scattered snow showers between 12Z and 18Z.

**Snowfall Accumulation Estimate**

The Garcia Method, in conjunction with an analysis of the ingredients, was used to estimate the snowfall accumulation in the Appleton forecast area between 12Z on March 13 and 12Z on March 14. Because this event was predicted to last 24 hours, two independent 12-hour Garcia forecasts were formulated: (1) 12Z on March 13 to 0Z on March 14 and (2) 0Z to 12Z on March 14. The average forecasted mixing ratio ($l$) at 700 hPa for the first period was 2.75-3 g kg\(^{-1}\) (initial $l = 2.5 - 3$ g kg\(^{-1}\) and 6-hour maximum $l = 3$ g kg\(^{-1}\)). During the second period, the average forecasted mixing ratio
at 700 hPa was 3-3.5 g kg$^{-1}$. A conventional Garcia Method forecast would double the sum of these average mixing ratios, giving a prediction of 12-13 inches of snow for the full 24-hour period. This is considerably less than the actual amount of snow that fell in the Appleton area. However, as described in Chapter 3.2.3, the conventional Garcia Method forecast must be modified based on a comparison of the ingredients in this storm with the “normal conditions” of the Garcia Method. In the latter half of the first 12 hours and early portions of the second 12 hours, heavy snow was expected due to the presence of strong forcing, high relative humidities, and some instability. Since the Garcia Method generally applies to snowfall of moderate intensity, the heavy snow forecasted in this case necessitates the use of a 4:1 ratio between snowfall and mixing ratio. This amounts to a 24-hour forecast for 24-25 inches of snow. This is in reasonable agreement with the maximum observed snowfall of 23.5 inches in Wautoma (see Fig. 3.18b).

This example illustrates the variety of issues involved in integrating the IBM with the Garcia Method for estimating snowfall accumulations. By accounting for the forecasted strong forcing and possible instability, an obtained. Note, however, that this forecast focused only on the final 24-hour portion of this event. Total storm accumulations for the 32-hour event were even greater for areas southwest of Appleton that received snowfall in the 8-hour period prior to 12Z on March 13 (including Wautoma, whose 32-hour accumulation was 28 inches).
3.3.4 Diagnosis of a Transition from Snow to Freezing Rain

Only 11 inches of snow were observed in Oshkosh, Wisconsin, located 15 miles south of Appleton, in the 24-hour period between 12Z on March 13 and 12Z on March 14. This accumulation is nearly half of the 21 inches reported in Appleton over the same time period (see Fig. 3.18b). The reduction in snowfall at Oshkosh can be largely attributed to a transition from snow to freezing rain that occurred between 2Z and 4Z in Oshkosh but not in Appleton (see Fig. 3.32). It is conceivable that this short period of freezing rain led to the observed variation in snowfall accumulation because while moderate freezing rain was falling in Oshkosh, moderate snow was rapidly adding to the existing snow cover in Appleton.

During the period of freezing rain, Oshkosh reported a surface temperature of -3 to -4°C, indicating that warmer air must have existed above the surface. The model-predicted temperature profile for Oshkosh valid at 0Z on March 14, 2 hours prior to the transition from snow to freezing rain, shows an elevated warm layer with a maximum warm layer temperature of -1°C at 800 hPa (Fig. 3.33). Hence, the atmosphere was everywhere below freezing at 0Z on March 14 when the precipitation was still snow.

As a first step toward diagnosing the transition to rain which occurred by 2Z on March 14, an assessment of how well the model-predicted sounding agreed with observations was performed. The forecasted temperature and dew point profiles valid at 0Z on March 14 for Green Bay, Wisconsin, the closest station to Oshkosh that launches radiosondes (Fig. 3.34a), were compared with the observed soundings at Green Bay
Figure 3.33: As in Fig. 3.31 except for Oshkosh, Wisconsin.

from the same time (Fig. 3.34b). The forecasted and observed soundings both show an elevated warm layer. The maximum warm layer temperature of the model-predicted temperature profile was -3.5°C at 800 hPa, and a cold layer was positioned close to the surface with a minimum temperature of -9°C. The dew point was 1-3 °C lower than the temperature through 700 hPa, and the air dries out slightly above that level. In contrast, the observed temperature and dew point profiles (Fig. 3.34b) were nearly coincident with each other, implying a more saturated atmosphere and a wet bulb temperature profile nearly equal to the air temperature profile. Furthermore, the observed maximum warm layer temperature was -2°C, 1.5°C warmer than predicted. The maxi-
Figure 3.34: a) As in Fig. 3.31 except for Green Bay, Wisconsin. b) Observed temperature and dew point soundings for Green Bay, Wisconsin. Data plotted and portrayed as in Fig. 3.31.
mum temperature occurred around the same level as in the forecast sounding.

Because of the similarity between the structure of the model-predicted Green Bay and Oshkosh soundings, and the proximity of the stations to each other (approximately 50 miles), an adjustment can reasonably be applied to the maximum warm layer temperature of the forecasted Oshkosh sounding, based on the difference between the Green Bay forecasted and observed maximum warm layer temperatures at the same time (+1.5°C). After this adjustment, the 0Z Oshkosh maximum warm layer temperature would be above freezing at 800 hPa, though only by +0.5°C. Based on the Czys et al. (1996) and Stewart and King (1987) techniques (see section 3.1.5) for determining the degree of melting in an above-freezing elevated warm layer, a maximum temperature of 0.5°C and warm layer depth of only 50-100 hPa corresponds to very little melting. Furthermore, the cold layer at 0Z beneath 800 hPa would have refrozen any precipitation that only melted partially. Thus, the 0Z Oshkosh model-predicted sounding after adjustment was still consistent with the surface observations of snow at Oshkosh at 0Z on March 14.

In order to extrapolate the subsequent evolution of the Oshkosh temperature profile, the thermal advection in the elevated warm layer was examined. Figure 3.35 shows a strong 800 hPa temperature gradient oriented nearly parallel to the 15 ms⁻¹ winds from the south, and warm advection of +2 °C hour⁻¹ in the Oshkosh area. Although this warming may have been partially offset by adiabatic cooling of the rising air, it is still likely that the 800 hPa temperature could have risen at least 3°C during the subsequent two hours. A 3°C rise in 800 hPa temperature above the adjusted model-predicted
warm layer temperature of +0.5°C would have led to a maximum elevated warm layer temperature of 3.5°C, warm enough for complete melting (Stewart and King, 1987). Although there was a deep cool layer near the surface, the temperature in this layer was not cold enough to initiate ice crystal formation, thus freezing rain would reach the ground.
Chapter 4

Conclusions and Future Work

This thesis presented an ingredients-based alternative to traditional rule-of-thumb techniques for forecasting mid-latitude winter season precipitation. The ingredients-based methodology (IBM) provides an effective framework for assessing the importance of five key ingredients in a winter precipitation event: quasi-geostrophic forcing for ascent, moisture, instability, precipitation efficiency, and temperature.

A comprehensive investigation of instability, the most often overlooked ingredient in winter season mid-latitude cyclones, was performed. Saturated equivalent potential vorticity ($PV_{es}$) and equivalent potential vorticity ($PV_e$) were employed as diagnostics for identifying regions of conditional instabilities (CI or CSI) and potential instabilities (PI or PSI), respectively. Through the analysis of numerical model data from actual midwestern snow events, the mechanisms that influence the evolution of $PV_{es}$ and $PV_e$ with time were explored. This analysis revealed that the evolution of $PV_{es}$ and $PV_e$ in a mid-latitude cyclone is dominated by horizontal advection, although there are some
regions where the saturation deficit term (SD), and the adiabatic generation term (AG) may contribute significantly to changes in $PV_{es}$ and $PV_e$, respectively. A comparison of the horizontal advection of $PV_e$ with the Eulerian change in $PV_e$ ($\Delta PV_e$) revealed substantial agreement in structure and location of these features. With a time interval of six hours, the magnitude of the $PV_e$ advection was two to three times the magnitude of $\Delta PV_e$. However, the magnitude of the advection terms approached that of the Eulerian $PV_e$ change as the time interval was reduced to one hour, indicating that the overestimation was related to storm motion.

A secondary, but potentially significant, contribution to the total Eulerian change in $PV_{es}$ was identified and called the saturation deficit term (SD). This term was shown to contribute to $\Delta PV_{es}$ where $l_s - l$ and horizontal derivatives of $l_s - l$ were large. This most commonly occurred behind a strong cold front in mature or decaying cyclones, where SD values of -0.4 to -0.8 PVU/day were typical, and a minimum SD of -1.6 was observed.

The adiabatic generation term (AG) in the $PV_e$ time tendency equation also made only a small contribution to the Eulerian change of $PV_e$. The most negative AG occurred primarily upshear of the occluded thermal ridge and along the cold front of mature or decaying cyclones, where isopleths of layer-averaged $\theta_e$ were rotated counter-clockwise with respect to contours of the isobaric thickness. Typical values of AG ranged from -1 to -2 PVU/day, and a maximum reduction of -3.8 PVU/day was observed near the triple point in the mature phase of a strong cyclone. In most situations, however, the location of the Eulerian change in $PV_e$ was not co-located with the time average
adiabatic generation of $PV_e$.

These preliminary findings about the time tendencies of $PV_e$ and $PV_{es}$ were not incorporated into the IBM; however, further investigation of the evolution of $PV_e$ and $PV_{es}$ may have utility for the IBM. An understanding of the mechanisms that transport and generate regions of instability in mid-latitude weather systems would enable forecasters to anticipate the evolution of regions of instability given initial conditions including the $PV_e$ or $PV_{es}$ field. For example, when numerical forecast model simulations are not verifying with observations, current analyses of winds, temperature and moisture parameters, and $PV_e$ or $PV_{es}$ could be used to infer the short term evolution of regions of instability. Additionally, one could predict the location or intensity of areas of instability at times between the standard 6-hour numerical forecast model time interval by anticipating the behavior of $PV_e$ or $PV_{es}$.

Following the investigation of the instability ingredient, the application of the IBM to forecasting winter season precipitation was detailed. Diagnostics for assessing each ingredient were presented. Q-vector convergence was employed to qualitatively assess QG forcing for ascent. $PV_{es}$ was used to identify regions of CI or CSI. Relative humidity and mixing ratio quantified moisture availability, and atmospheric temperature was used for both a determination of precipitation type and as a means to assess the efficiency ingredient. Isobaric ingredients maps were introduced to assist in the visualization of the ingredient parameters, and an ingredients table was used to organize the ingredients information from numerical model data for all forecast hours and in three isobaric layers (800-850 hPa, 700-750 hPa, 600-650 hPa). In some situations, the isobaric ingredients
maps and ingredients table analysis at the three standard pressure layers do not fully capture the relevant distribution of the ingredients. In these cases, a cross-sectional analysis provides a more complete picture of the vertical distribution of the ingredients. Two examples of isolated layers of instability which were not resolved in the standard isobaric layer ingredients maps but contributed to enhanced precipitation rates were presented.

The IBM can be used by itself to forecast precipitation duration, intensity, and type. It does not, however, independently provide an estimate of snowfall accumulation. This thesis presented an approach that combines the physical basis and flexibility of the IBM with the quantitative nature of a traditional forecast technique to make a preliminary prediction for snowfall accumulation. In this approach, the Garcia Method (Garcia, 1994) forecast for snowfall accumulation is considered the “normal conditions” prediction. Normal conditions for the Garcia Method correspond to moderate forcing for vertical motion, no instability, a 10:1 snow to liquid water ratio, and snow as the sole precipitation type. Thus, if the IBM indicates “abnormal conditions” (i.e., the potential for strong forcing, instability, or very cold surface temperatures), the GM-predicted snowfall accumulation should be increased. Garcia (1999) recommends a doubling of the inches of snowfall predicted by his original technique for heavy or convective snow events.

The utility of the IBM was illustrated for a case study of a poorly forecasted strong snow storm that affected Wisconsin on March 13-14, 1997. Together, the early onset of precipitation and the locally enhanced precipitation rates (as a result of convective
snow) led to storm total accumulations of up to 30 inches. When applied to this case, the IBM provided an accurate description of precipitation duration, intensity, and type. The IBM identified a mid-level instability coincident with ample moisture, QG forcing, and an air temperature that could support maximum depositional growth of ice crystals. These processes were likely involved in an unanticipated snow swath that fell prior to the predicted onset of precipitation. Later in the storm, significant clues to the potential for additional convective precipitation were identified using the IBM. Using the 4:1 ratio of snowfall to mixing ratio suggested by Garcia (1999) for convective snowfall events, an accurate estimate of the snowfall accumulations was obtained.

Because the ingredients diagnostics are tied closely to gridded numerical forecast model data, it is important to relate each ingredient to an observable quantity that can be monitored throughout the storm’s development. For example, regions of conditional instability could be identified in observed soundings and compared to forecasted values of the instability ingredient parameter, $PV_{cs}$, and upper air analyses of relative humidity could be used to verify the moisture ingredient. When compared with model-predicted ingredient parameters, observations of the actual magnitude and location of an ingredient enables the forecaster to anticipate modifications in the precipitation patterns if conditions do not evolve as forecasted by the model. Additionally, such observations may facilitate the decision to choose between different model scenarios as an event is unfolding. For example, if one model predicts less precipitation because of moisture limitations, observations of moisture parameters upstream of the forecast area could be compared with forecasted relative humidity to assist in determining the preferred
Furthermore, regular monitoring of IBM diagnostics and routine verification of winter precipitation forecasts prepared with this technique may prove useful in identifying weaknesses in the computation of the quantitative precipitation forecast by numerical forecast models. If events for which the instability or efficiency ingredient is important are consistently under-forecasted by a given numerical forecast model, this information would indicate that the convective parameterization scheme (for instability) or the microphysical parameterization (for efficiency) need further work.

The $\nabla \cdot Q$ diagnostic used to evaluate the forcing ingredient in the IBM presented in this thesis is arguably the weakest element. Use of this diagnostic excludes explicit consideration of the effects of ageostrophy on the redistribution of temperature and momentum and, thus, on the vertical motion forcing itself as suggested by Eliassen (1962). For this reason, it may prove useful to quantify the degree to which a given flow is well described by the QG assumptions. This could be accomplished by 1) computing the Rossby Number for the case, 2) comparing the model-predicted vertical velocity with that diagnosed for the QG-$\omega$ equation(equation 3.1), or 3) identifying synoptic patterns that are commonly associated with strong ageostrophic circulations.
Appendix A

The IBM in Operational Use

The following excerpts from forecast discussions by forecasters at the NWSFO in Dousman, Wisconsin were written in December 1999 and January 2000 following the introduction of the IBM analysis tools into the forecast office.

1. AREA FORECAST DISCUSSION
   NATIONAL WEATHER SERVICE MILWAUKEE/SULLIVAN WI
   330 AM CST WED DEC 15 1999

   **THE INGREDIENT ANALYSIS TOOL** indicates 2G/KG mixing ratios in the SE corner of WI with somewhat less to the west with only light QG forcing. No instability noted thus XPCTG 1-3 INCHES ACRS THE CWA. BUFKIT SOUNDINGS SHOW ALL SNOW WITH THIS EVENT.

   - Written by John Haase

2. AREA FORECAST DISCUSSION
   NATIONAL WEATHER SERVICE MILWAUKEE/SULLIVAN WI
   258 PM CST MON JAN 3 2000

   **WETZEL FCST INGREDIENTS MACROS ON AWIPS** show q vector convergence sustained over ern 2/3 of CWA thru 06Z and lingering in the east til 12Z. SPEC HUMIDITIES ARE IN THE 3- 4G/KG
RANGE. H8 TEMPS HANG IN THE 0°C RANGE OVER RACINE AND KENOSHA CNTYS...SO SNOW AMOUNTS MAY BE LIMITED BY MIX WITH RAIN. WL UP SNOW ACCUMULATIONS SLGTLY FOR IOWA...SAUK AND MQT CNTYS. CUR-
RENT BAND COULD LAY DOWN A FAST 1 TO 2 INCHES BEFORE 00Z...SO WL EXPAND THE RANGE SLGTLY.

– Written by Bob McMahon

3. AREA FORECAST DISCUSSION
NATIONAL WEATHER SERVICE MILWAUKEE/SULLIVAN WI
230 PM CST MON JAN 10 2000

ON WED...AVN GETTING EXCITED WITH NEXT SHORT WAVE WITH THIS ONE GENERATING DECENT ISENT LIFT AND STG WAA. USING THE FORE-
CAST INGREDIENT ANALYSIS TOOL SUGGESTS MDT FORCING WITH SPECIFIC HUMIDITIES INCRG TO OVR 2 G/KG. SOME INSTABILITY ALSO SHWG UP ACRS IA BY 12Z WED WHICH MAY TRANSLATE INTO SRN WI DURING THE DAY. WUD LIKE TO SEE ANOTHER MDL RUN. 0 H8 LN RMNS OVR NRN IL BUT ? REMAINS WHERE STGST LIFT AND HVST SNOWS OCCUR. MAY END UP N OF MKX CWA. FOR NOW WILL GO 50 POPS ON WED.

– Written by John Haase

4. AREA FORECAST DISCUSSION
NATIONAL WEATHER SERVICE MILWAUKEE/SULLIVAN WI
211 PM CST WED JAN 12 2000

NEW ETA MODEL PAINTING SNOWY PICTURE FOR TNGT AS IT DVLPS ANOTHER SURGE OF MOISTURE INTO THE MID LVLS. 290K ISENT PROGS SHOWING MIXING RATIOS INCRSG TO 2-3 G/KG AFT 00Z...AS UPR LVL IN-
STABILITY DVLPS AS RRQ OF JET MAX MOVES OVR SRN WI. THIS MOST APRNT BY LOOKING AT WETZEL INGREDIENTS AT 600 MB. MAX UVV VALUES OCCUR AT 00Z AND CONT INTO EVE HRS. LOOKS LIKE QUICK HITTER AS DYNAMICS SHIFT EWD BY 06Z. PSBL PD OF 3-4 HRS WHEN SNOWFALL WILL BE HEAVIEST.

– Written by Cris Garcia
5. AREA FORECAST DISCUSSION
NATIONAL WEATHER SERVICE MILWAUKEE/SULLIVAN WI
330 AM CST TUE JAN 18 2000

MDT Q-VECTOR CONVERGENCE AS SEEN ON THE FCST INGREDIENT ANALYSIS TOOL (FIAT). SPECIFIC HUMIDITIES ARE ARND 1.5 G/KG WITH TMPS PROFILE INDICATING A SNOW/WATER RATIO OF 15:1 OR 20:1. MDL QPF'S ALSO IN THE .10 TO .25 RANGE. SOME INSTABILITY HINTED IN SW WI. TIME SECTIONS SHOW THAT THE BULK OF LIFT/DEEP LAYERED MSTR IS WED AFTN UNTIL ABOUT 6Z THU. ALL THIS POINTS TO A 2-4 INCH SNOWFALL BY THE TIME IT ENDS LATER THURS NGT WITH POTENTIAL FOR 5 INCHES IN OUR WRN CWA. BULK OF THIS STILL IN XTND PART OF FCST SO WILL BEEF UP WORDING FOR THAT AND INCR POPS SOMEWHAT FM PRVS FCST.

– Written by John Haase
References


