P1.11 AN EXAMPLE OF FORECASTING MESOSCALE BANDS IN AN OPERATIONAL ENVIRONMENT

Philip N. Schumacher*
National Weather Service, Sioux Falls South Dakota

1. INTRODUCTION

The forecasting of mesoscale bands of precipitation is one of the more difficult non-convective forecasts in the operational environment. In winter, these bands of precipitation can produce in excess of 30 cm of snow in 24 h. Gradients of snowfall on the order of 20 cm (100 km)$^{-1}$ mean small errors in location will result in large errors in snowfall totals at a particular location. While mesoscale model output of precipitation can provide guidance as to the amount and location of precipitation bands, the errors associated with these bands can be in excess of 50 km – even within 24 h of the forecast verification. Therefore, operational forecasters need to use other tools to help determine the most likely location for precipitation bands.

Previous research has shown that the development of mesoscale bands is the result of interaction of synoptic and mesoscale forcing mechanisms. Shapiro (1982) hypothesized that a favorable superposition of a front and upper jet could determine whether upright convection would develop. Loughe et al. (1994) used psi-vectors (Keyser et al., 1989) to show that the along wind component ("synoptic" scale) of the vertical motion was a significant contributor to the total vertical motion associated with major northeastern winter storms. Korner and Martin (2000) and Morgan (1999) used potential vorticity (PV) inversion to show the influence of upper level anomalies upon low level fronts. They were able to show that PV anomalies in the vicinity of fronts can enhance low level frontogenesis through increased convergence along portions of a frontal zone. Finally, Roebber et al. (2002) showed that location and strength of a PV anomaly was a critical part of determining the location of severe convection associated with the 3 May 1999 outbreak.

Therefore the understanding of synoptic scale and mesoscale dynamics, and the interactions of these scales, is critical to accurately forecast the location of mesoscale bands. This paper uses a case study of a snowband across eastern South Dakota to provide a methodology for forecasting significant mesoscale bands by examining the different scales of involved in the forecast. This will entail access to gridded data at different resolutions similar to that discussed by Roebber et al. (2002).

*Corresponding author address: Philip N. Schumacher, National Weather Service, 26 Weather Lane, Sioux Falls SD 57104; e-mail: phil.schumacher@noaa.gov

2. SYNOPTIC OVERVIEW

From 0000 UTC 14 March 2002 through 0600 UTC 15 March 2002, a mesoscale snowband developed across South Dakota and Minnesota. In excess of 30 cm of snow, with a maximum of 48 cm, was observed within this band from southwest South Dakota into southwest Minnesota (Fig. 1). The gradient in snowfall south of the band ranged from 2.5-5.0 cm (10 km)$^{-1}$. For one county, this meant a snowfall range of 25 cm from north to south.

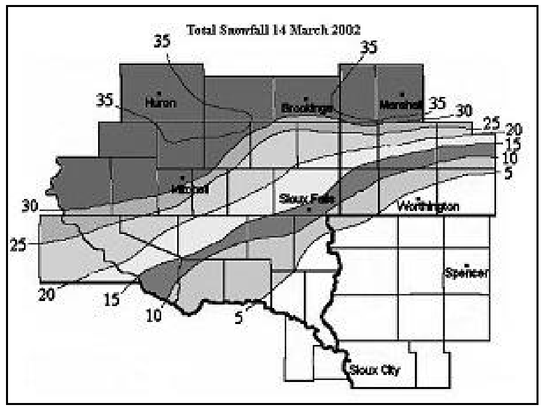

Fig 1. Total snowfall for the WFO Sioux Falls County Warning Area for 14 March 2002. Greater than 5 cm shaded. Contours every 5 cm.

At 0000 UTC 14 March, 700 hPa front was extended from the South Dakota and Nebraska border into southern Minnesota. A 300 hPa jet streak was located across northern Minnesota placing the front within the right entrance region of the jet. To the southwest, an upper level PV anomaly was moving from the central Rockies into western Nebraska. Light snow developed across western South Dakota prior to 0000 UTC 14 March. The snow spread along the mid-level front eastward into southern Minnesota by 1200 UTC 14 March. As the PV anomaly moved to the northeast, the band of precipitation narrowed and strengthened and became stationary from south central South Dakota into southwest Minnesota. As the PV anomaly moved across the area after 1800 UTC 14 March, the band of snow slowly moved southeast toward northwest Iowa and weakened as the upper level anomaly moved into Minnesota.

3. FORECAST METHODOLOGY

3.1. Synoptic scale analysis

A methodology similar to that assumed for "PV thinking" is applied. This assumes that the primary forcing for vertical motion is located near the tropopause and near the surface. For this paper, the surface will denote the three-dimensional frontal surface. As shown by Loughe et al. (1994), the synoptic scale lift is a significant contributor to frontal circulations. Surface forcing tends to be dominated by mesoscale forcing mechanisms. Therefore the synoptic scale forcing will be examined only above 500 hPa. Diagnostic studies using PV inversions apply a similar methodology to split the PV into multiple levels and examine the influence of each level (Morgan, 1999). While a PV inversion similar to Korner and Martin (2000) and Morgan (1999) would provide the best analysis for determining the influence of an upper level anomaly, the computing power to do a PV inversion in real time does not exist in the operational environment. Therefore, Q-vectors (Hoskins et al., 1978) in the 300 hPa to 500 hPa layer are used to examine the qualitative strength and evolution of the synoptic scale forcing. For National Weather Service forecasters, the Advanced Weather Interactive Processing System (AWIPS) includes a filter to remove small scale waves from the 80 km resolution grids

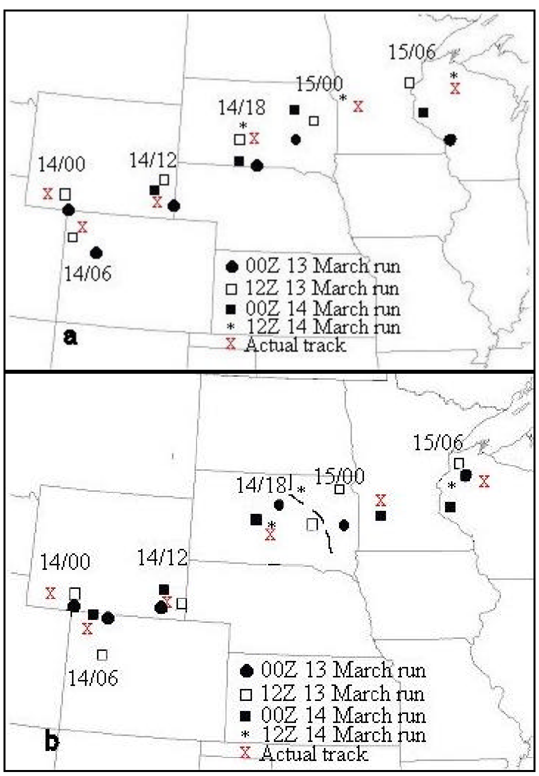

Fig. 2. Track of the 300 to 500 hPa Q-vector convergence maximum from the a) Global Forecast System model (GFS) and b) Eta model every 6 h from 0000 UTC 14 March – 0600 UTC 15 March 2002. The actual track was taken as the initialization of the GFS at the hour listed.

available from the Eta and GFS. The advantage of using a smoothed height field for Q-vectors is that it is relatively independent of vertical motion fields produced by the mesoscale models which are forced by mesoscale (or smaller) features within the model. Also, the divergence of Q is independent of the moisture forecast within the model.

For the 14 March 2002 event, Q-vectors are examined in both the GFS and Eta models. Both models showed remarkable agreement in the qualitative evolution of the Q-vector convergence. These models show the minimum divergence moving from southeast Wyoming across central South Dakota into northwest Wisconsin by 0600 UTC 15 March. The differences between models were negligible and both models were consistent 36 h prior to the event (0000 UTC 13 March). This area of Q-vector convergence was also located north of the precipitation maximum in both models at 0000 UTC 13 March. The run-to-run consistency and consistency between models gave forecasters confidence that the synoptic scale forcing was well forecast and would be across South Dakota.

3.2 Mesoscale forcing

As discussed by Schultz and Schumacher (1999), many mesoscale bands form as the result of frontogenetical forcing, especially in the absence of terrain. Therefore, the assumption is made that frontogenesis is the primary forcing mechanism. The difficulty in using frontogenesis, especially in winter, is that the frontal boundary slopes with height. So the maximum frontogenesis is not in the same location when moving higher into the atmosphere. This leads to the question as to where the lift and saturation will be maximized and result in the heaviest snow. To aid forecasters, an initial assumption is made that maximum lift will be located where there is a collocation between the frontogenesis maximum and upper level forcing. It is also assumed that lift initiated by frontogenesis above 600 hPa will be less likely to produce heavy precipitation since the amount of moisture available decreases markedly.

In practice, forecasters will load fields of Pettersen frontogenesis (Pettersen, 1956) from 850 hPa to 600 hPa and Q-vector divergence in the 300-500 hPa layer. Forecasters will then examine where the frontogenesis maximum is with respect to the Q-vector divergence minimum. The result is generally a 100 mb layer where the frontogenetical maximum approaches the Q-vector divergence minimum. Then forecasters will overlay lapse rates in the 700 to 500 mb layer (or around 600-500 hPa if frontogenesis is located above 700 mb) to provide an approximation of where stability between the front surface and upper level wave is minimized.

In the case from 14 March 2002, frontogenesis from 850 hPa was maximized from central Nebraska into central Iowa, while at 600 hPa it was maximized from south central South Dakota into east central Minnesota – a width in excess of 300 km. Overlaying the Q-vector divergence and mid-level lapse rates showed that the best interaction was likely to be across south central South Dakota into southern Minnesota. By examining all three parameters the 700 hPa frontogenesis was chosen (Fig. 3). This is the location where the frontogenesis approaches near the Q-vector divergence minimum and stability is decreasing.

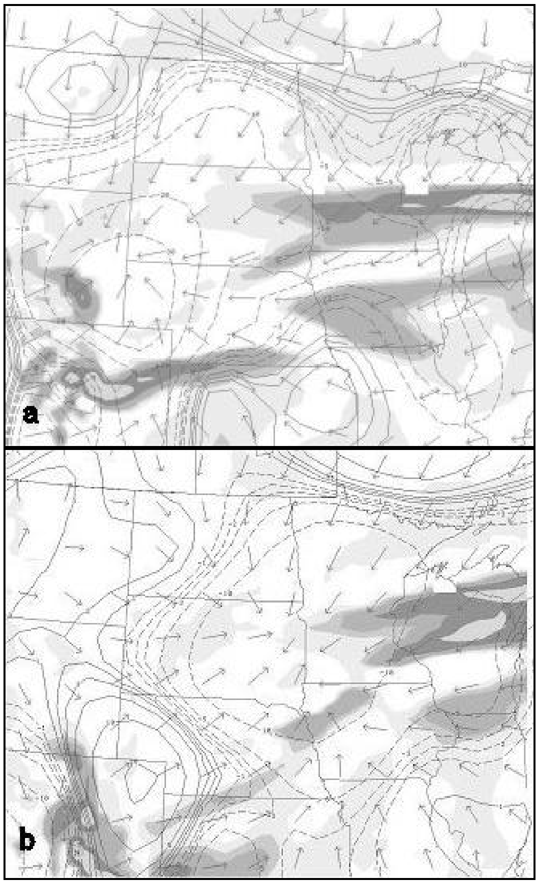

Fig. 3. 300 to 500 hPa Q-vectors, Q-vector divergence (< 0 dashed) and 700 hPa frontogenesis (> 0 shaded) from the 1200 UTC 13 March 2002 Eta model. A) 24 h forecast verifying at 1200 UTC 14 March. B) 36 h forecast verifying at 0000 UTC 15 March. Note: The Q-vectors are calculated from an 80 km smoothed height field and frontogenesis is calculated from the 40 km grid.

3.3 Stability and moisture

After determining where the best interaction between mesoscale and synoptic appears to be maximized the forecaster then examines the stability. Initially, the potential for upright convection is examined. When upright convective instability exists south of the best where the best scale interaction, the forecaster will adjust the location for the heaviest precipitation toward the area with upright convection instability. If upright convective instability does not exist, then conditional symmetric instability (CSI) is examined. Schultz and Schumacher (1999) noted that the response to frontogenesis is related to the CSI. In general, CSI is examined in a 50-100 hPa layer above the level where the mesoscale (frontogenesis) forcing is expected to interact best with the synoptic scale forcing. Areas with weak symmetric stability (< 0.25 PVU) or weak symmetric instability will result in a vigorous frontal ascent and increase the likelihood of interaction between the upper level wave and mesoscale front. If time permits, cross-sections perpendicular to the front are used to examine the stability by looking at saturated equivalent potential temperature changes with height above the frontal surface.

Finally, saturation is examined. Temperatures at or below -12°C are necessary for ice crystal formation and for heavy snow one needs saturation in a layer around -16°C to produce dendritic crystals. While moisture profiles are highly dependent upon where the model produces lift, examination of model soundings indicate whether it is saturated around these critical temperatures. Or, if the forecaster suspects lift may be maximized north of the model solution, one can make sure that temperatures will be at or below -12°C above the frontal surface which indicates ice crystal formation is possible.

For 14 March 2002, the potential for upright convection did not exist. An analysis of CSI between 600 and 500 hPa showed the largest instability was across Nebraska and southeast South Dakota (Fig. 4). The frontogenetical maximum was located along the northern gradient of this instability. An examination of moisture showed that the air was not saturated where the instability was greatest in east central Nebraska (not shown). Therefore the best vertical motion was likely to be along the stability gradient where the atmosphere became saturated. In general, the Eta saturated the atmosphere south of where the maximum interaction was expected. However, forecasts did show that if it did saturate where the dynamical forcing was located, the critical temperatures would be met and dendritic ice crystals were likely to form.

3.4 Precipitation

One goal of forecasters is to produce a quantitative precipitation forecasts and, when below freezing, snowfall forecasts. Even if the location of the model forecast is in question, forecasters can assume that if the model contains the dynamical forcing, it will likely get the nature of the event. This is similar to the assumption made by Roebber et al. (2002) when they discuss the use of high resolution models in convective forecasts. Therefore forecasters can examine the structure of the model precipitation forecast. If they observe a narrow band of precipitation, it provides them confidence that their diagnosis for banded precipitation is correct. They will also examine QPF amounts from the high resolution Eta to provide an estimate for maximum amounts within the band of snow. Other factors which can influence precipitation forecasts include the amount of time interaction between the upper wave and mesoscale boundary is maximized and the degree of instability. In a weakly symmetrically stable atmosphere with favorable interaction the forecaster can anticipate snowfall rates of 2.5 cm h^{-1}.

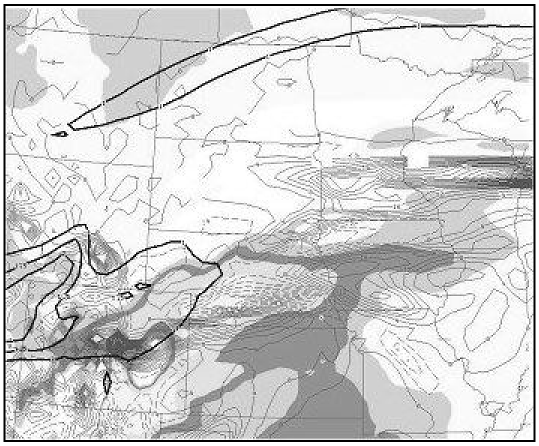

Fig. 4. Geostrophic equivalent potential vorticity (EPV) using saturated equivalent potential temperature in the 600 to 500 hPa layer (shaded < 0.25 PVU), 700 hPa frontongenesis (gray contours,) and 300-400 hPa potential vorticity (> 1.0 PVU, thick black contours). The 24 h forecast from the 1200 UTC 13 March 2002 Eta.

For 14 March 2002, the Eta model consistently showed a narrow band of precipitation. At 0000 UTC and 1200 UTC 13 March, the heaviest precipitation extended from south central South Dakota into northwest Iowa with a maximum amount approaching 4 cm (Fig. 5). By 0000 UTC 14 March, the maximum precipitation had moved along and north of interstate 90 – from Chamberlain South Dakota to Windom Minnesota. In all cases, this was too far south of where the forcing was maximized (and where precipitation verified).

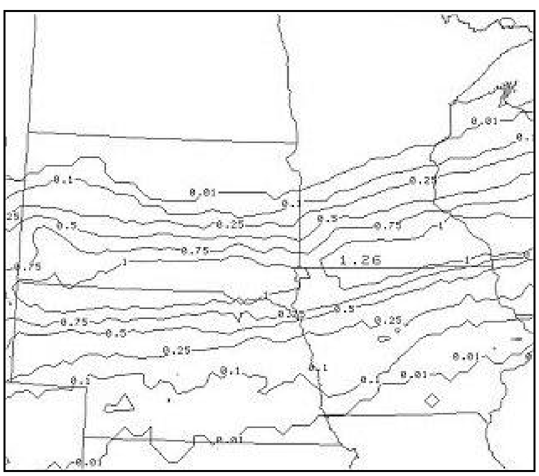

Fig. 5. 48 h QPF from the 1200 UTC 13 March 2002 Eta forecast. Contours are in inches.

However, amounts were considered fairly accurate and forecasts of 25 cm of snow were issued for portions of South Dakota and Minnesota based upon the 1200 UTC 13 March 2002 Eta forecast.

4. CONCLUDING REMARKS

The technique described here provides forecasters with a scientifically based tool to adjust model solutions for banded precipitation events. By examining the synoptic and mesoscale forcing as well as stability and moisture, the forecaster can determine which model solution is more likely to occur and provide forecasts with higher confidence at longer ranges. The National Weather Service office in Sioux Falls South Dakota has been using this technique for over one year. The case described above is only one of several cases of correct adjustments to precipitation location and amounts. In one case, a narrow band of light freezing rain was anticipated over portions of southeast South Dakota and southwest Minnesota even though the Eta QPF was zero. In another case winter storm watches and warnings were issued even though Eta QPF suggested snowfall amounts near zero where watches were issued. In both cases the public was given 18 to 30 hours lead time.

The technique requires forecasters to use Q-vectors to examine the qualitative strength and location of upper level forcing for vertical motion. With this knowledge forecasters can then examine the mesoscale forcing, usually frontogenesis, and stability to determine where the best interaction between the synoptic and mesoscale forcing will be. Then examination of high resolution QPF can provide confidence in the structure of the precipitation event and knowledge of the amount of rain or snow expected.

The implication of this research is that forecasters will continue to need access to grids at multiple resolutions (Roebber et al., 2002). Even when using a smoothed height field, grid resolutions below 80 km result in use of wave energy which is not of synoptic scale and violates the assumptions made to derive the quasi-geostrophic equations. In addition, examination of derived quantities like frontogenesis, generally need to be done on grid resolutions greater than 30 km or the amount of small scale noise results in difficulty in determining the signal in the model mass or moisture field. Finally, high resolution model output can be used to examine the structure of the precipitation band and expected amounts. As shown by Roebber et al. (2002), even if the forecast is in the wrong location, validation of the hypothesized structure can give forecasters greater confidence to issue watches and warnings when necessary.

Disclaimer: The views expressed here are those of the author and do not necessarily represent those of the National Weather Service.

Acknowledgements: I would like to thank the forecasters at the National Weather Service office in Sioux Falls for their discussion and input for this research. Without their hard work and dedication, the paper could not have been completed. I would also like to acknowledge Dr. Paul Roebber and discussions we had about the 3 May 1999 event that led me to pursue this research.

REFERENCES

Barnes, S. L., F. Caracena, A. Marroquin, 1996: Extracting synoptic-scale diagnostic information from mesoscale models: The Eta model, gravity waves, and quasigeostrophic diagnostics. *Bull. Amer. Meteor. Soc.*, **77**, 519-528.

Keyser, D., B. D. Schmidt, D. G. Duffy, 1989: A technique for representing three-dimensional vertical circulations in baroclinic disturbances. *Mon. Wea. Rev*., **117**, 2463-2494.

Hoskins, B. J., I. Draghici, and H. C. Davies, 1978: A new look at the ù-equation. *Quart. J. Roy. Meteor. Soc*., 104, 31-38.

Korner, S. O. and J. E. Martin, 2000: Piecewise frontogenesis from a potential vorticity perspective: Methodology and a case study. *Mon. Wea. Rev*., **128**, 1266-1288.

Loughe, A. F., C.-C. Lai, and D. Keyser, 1994: A technique for diagnosing three-dimensional ageostrophic circulations in baroclinic disturbances on limited-area domains. *Mon. Wea. Rev*., **123**, 1476-1504.

Morgan, M. C., 1999: Using piecewise potential vorticity inversion to dagnose frontogenesis. Part I: A partitioning of the Q vector applied to diagnosing surface frontogenesis and vertical motion. *Mon. Wea. Rev*., **127**, 2796-2821.

Pettersen, S., 1956: *Weather Analysis and Forecasting, Vol. 1, Motion and Motion Systems.* 2nd ed., McGraw-Hill, New York, 428 pp.

Roebber, P. J., D. M. Schultz, and R. Romero, 2002: Synoptic regulation of the 3 May 1999 tornado outbreak. *Wea. Forecasting*., **13**, 399-429.

Schultz, D. M. and P. N. Schumacher, 1999: The use and misuse of conditional symmetric instability, *Mon. Wea. Rev*., **127**, 2709-2732.

Shapiro, M., 1982: *Mesoscale Weather Systems of the Central United States*. CIRES, Univ. of Colo./NOAA, Boulder, CO, 78 pp.