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THE IMPACT OF SUPERIMPOSED LOW-LEVEL JETS DURING THE 2003 PRESIDENTS' DAY WINTER STORM

Michael T. Kiefer[1], Michael L. Kaplan[2], and Yuh-Lang Lin[1]
[1]*North Carolina State University, Raleigh, North Carolina*
[2]*Desert Research Institute, Reno, Nevada*

1. Introduction

During 15-18 February 2003, almost exactly 24 years after the infamous 1979 Presidents' Day event (18-19 February 1979, hereafter PD79), another winter storm of great intensity impacted the eastern third of the United States. Snowfall amounts of greater than 50 cm were common from Virginia northward to extreme southern New England, with numerous reports of event snowfall exceeding 100 cm over northeastern West Virginia and western Maryland (hereafter, the *region of interest*). While both events produced excessive amounts of snow and ice over the Metropolitan areas of the mid-Atlantic and northeastern U.S. (See Bosart 1981 for snowfall analysis during the former), the extreme snowfall observed during the 2003 Presidents' Day event (PD03) occurred despite the absence of a rapidly deepening coastal cyclone, a feature noted to be critical to the heavy snowfall reported during PD79 (Uccellini et al. 1985). Within this paper, the development and subsequent interaction of two low-level jets (termed the continental and maritime, see Figure 1 for proposed conceptual model during height of event) will be considered in detail in a numerical modeling analysis of the forcing for heavy snowfall over the mid-Atlantic U.S. and in particular the region of interest.

The primary motivation of this study has been improving the understanding of winter storm dynamics, and subsequently our ability to adequately predict heavy snowfall in such events as PD03. While operational model forecasts of PD03 were generally reasonable, the extreme snowfall amounts observed within the region of interest were poorly forecast. One component present in PD03, the continental low-level jet directed from the Gulf of Mexico toward the region of interest in the 600-800 hPa layer (see Fig. 1), is absent in prior studies of winter storm precipitation. A preliminary assessment of other cases reveals a number of events (e.g. 19-20 January 1978, 4-5 December 2002, etc.) in which similar jets developed, yet little research has been done to understand the implications and origins of such features. It will be shown that the continental and maritime low-level jets just discussed contributed significantly to snowfall over the region of interest primarily through moisture transport and frontogenesis. Dynamical analysis in this study is performed primarily through the use of a mesoscale numerical model, the Non-Hydrostatic Mesoscale Atmospheric Simulation System (NHMASS) version 6.3 (Kaplan et al. 2005).

An observational summary of this event is presented in Section 2. Section 3 provides a brief description of the NHMASS model, while Section 4 describes principal results of this study, and Section 5 presents concluding remarks.

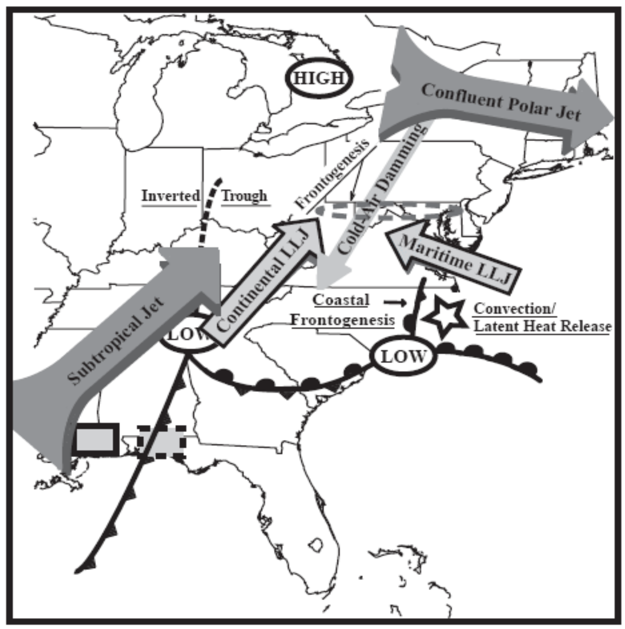

Fig. 1. A conceptual model of Presidents' Day 2003 winter storm. Regions over which moisture calculations in Table 1 are indicated by a solid box [16/0000 and 16/0600 calculations] and dashed box [16/1200 calculation] along the Gulf coast.

2. Observational Summary

The Presidents' Day 2003 winter storm can essentially be traced back in time to the merging of two upper-tropospheric jet streaks over the eastern United States. As early as 1200 UTC 14 February 2003, one can observe two airstreams: the subtropical jet (STJ) directed from the northeast Pacific Ocean across northern Mexico and toward the mid-Atlantic United States, and the polar jet (PJ) oriented from the Canadian plains across the Great Lakes region toward the mid-Atlantic U.S. (c.f. Figs. 1 and 2). A weak trough propagating along the poleward side of a western U.S. mid-level ridge (not shown) contributed to the formation of a lee cyclone seen at 850mb at 1200 UTC 14 February 2003 (Fig. 2b1), which subsequently was located on the Missouri/Kansas border at 1200 UTC 15 February 2003 (not shown), and over western Tennessee by 1200 UTC 16 February 2003 (Fig. 2b2).

At the surface, a low is noted at 1200 UTC 14 February 2003 over southwestern Kansas (Fig. 2c1) with a developing frontal boundary to the east. By 1200 UTC 16 February 2003 (Fig. 2c2), the surface low had moved toward central Alabama, and an anticyclone over northern Saskatchewan had moved to a position near Montreal, Quebec (CYUL) with strong cold-air damming (CAD) present in the lee of the Appalachians, as evidenced by the southward bulge of the isobars. Also seen at the time is a pair of inverted troughs on either side of the CAD region, the eastern one associated with coastal frontogenesis taking place and the western inverted trough consistent with previous studies of CAD events (Bell and Bosart 1988). A weak cyclone had also formed offshore of the North Carolina coast in the vicinity of the coastal front trough.

The kinked stationary frontal boundary over Tennessee apparent in Figure 2c2 was associated with a tongue of warm air transported northward by the continental low-level jet above 850 hPa (Fig. 3a). Elevated convection was present over northern Kentucky, southern Indiana, and southern Ohio just north of the surface warm air tongue. Also seen at that time is a band of light to moderate precipitation from northern West Virginia to southern New Jersey (Fig. 4). This band is parallel to an intensifying frontal zone aloft in the 600-800 hPa layer (not shown) indicated in Fig. 1 by the dashed ellipse across northern West Virginia and Virginia. It was within this broader band that the intense snowfall was reported across northeastern West Virginia and western Maryland 6-18 hours later.

Figure 5 depicts the observed trends of 24 hour accumulated precipitation between 1200 UTC 14 February 2003 and 1200 UTC 17 February 2003, indicating a dramatic shift of precipitation following 1200 UTC 16 February 2003 from the Tennessee River valley toward the mid-Atlantic U.S. What must be considered is what mechanism(s) are largely responsible for the large precipitation amounts evident in Fig. 5d across the mid-Atlantic U.S. and especially over the region of interest.

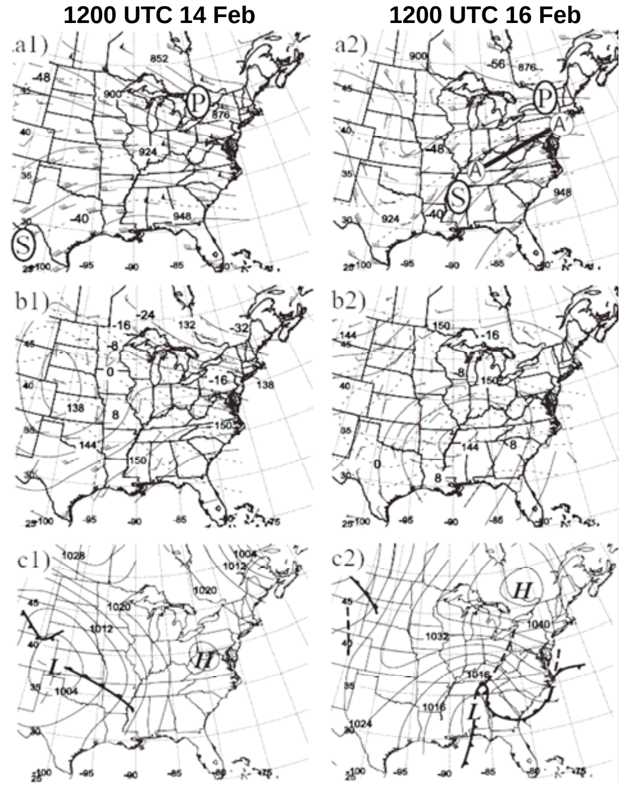

Fig. 2. NCEP/NCAR Reanalysis plots of a) 300 hPa, b) 850 hPa heights (dm), temperature (^{o}C), and winds (ms^{-1}), and c) Sea-level pressure (hPa) and surface frontal analysis, valid 1200 UTC 14 Feb 2003 (a1,b1,c1) and 1200 UTC 16 Feb 2003 (a2,b2,c2). Approximate positions of polar jet (P) and subtropical jet (S), determined from 300 hPa geostrophic wind maxima, noted in (a). Labels are omitted every other contour in all figures. Axis A-A' in (a2) for cross-section in Fig. 6.

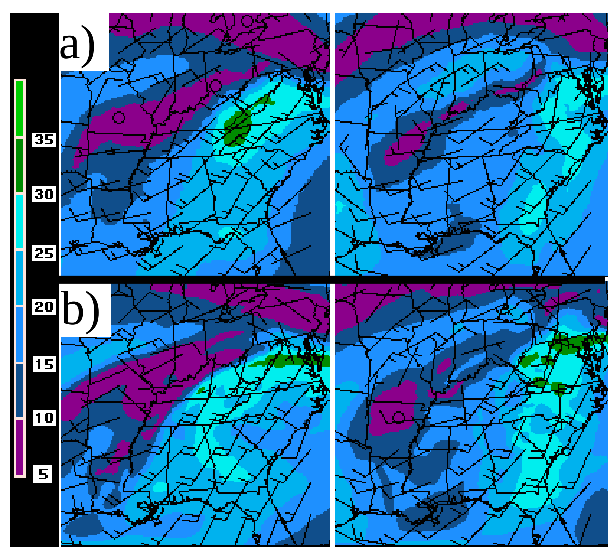

Fig. 3. 750 hPa heights [dm], isotachs [m/s, shaded], wind vectors [m/s] valid 0000 UTC [left two panels] and 1200 UTC [right two panels] 16 Feb 2003 for (a) NARR data and (b) 18 km NHMASS.

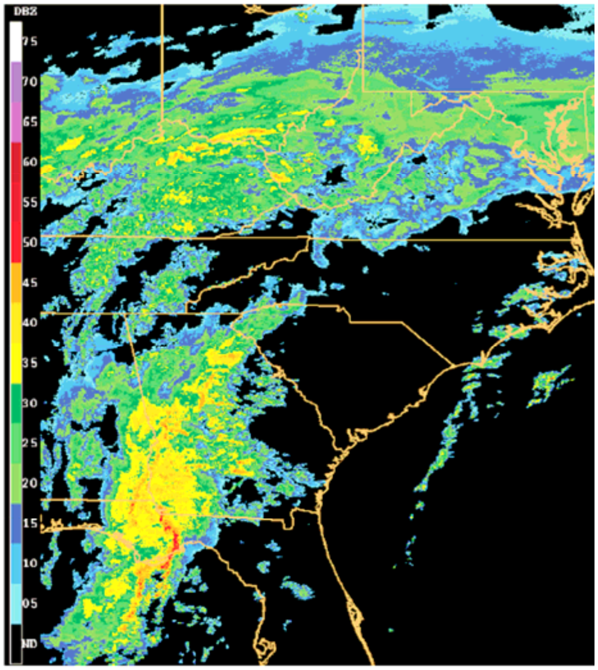

Fig. 4. NOWRAD 2-km base reflectivity valid 1200 UTC 16 Feb 2003.

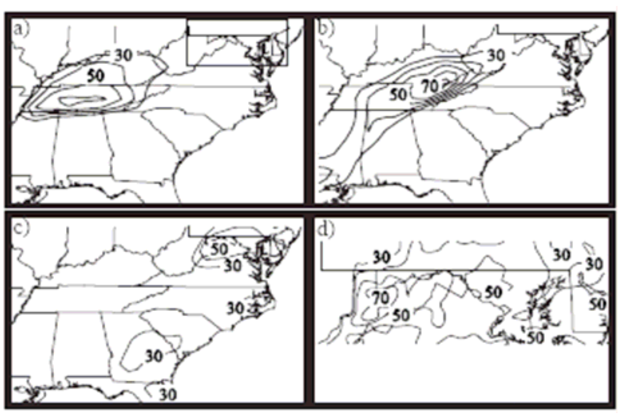

Fig. 5. 24 hour accumulated precipitation [mm, greater than 30 mm plotted, labeled every other contour] from CPC 1/8 degree rain-gauge analysis produced from objectively analyzed rain-gauge dataset (courtesy NCEP), valid a) 1200 UTC 15 Feb 2003, b) 1200 UTC 16 Feb 2003, c) and d) 1200 UTC 17 Feb 2003, with (d) inset image of (c). Inset region for (d) indicated in (a). Labels are omitted every other contour in all figures.

3. Model Description and Validation

a. Model Description

A mesoscale numerical model is used in this study to diagnose the multi-scale processes contributing to the excessive precipitation amounts reported in western Maryland. While observational data, for example NCEP/NCAR Reanalysis (Kalnay et al. 1996) and the North American Regional Reanalysis (NARR) (Mesinger 2004) data sets, wind profilers, and 2km NOWRAD Doppler radar data were analyzed when available, the resolution and/or temporal coverage of these sources of data made the use of a mesoscale numerical model essential for this study. The model chosen for this study is the NHMASS version 6.3 (Kaplan et al. 2005). One-way nesting was performed from 18 km horizontal resolution to 222 m resolution, with a separate 36km simulation performed for trajectories and validation of large scale patterns with NARR data. The 18 km and 36 km resolution runs were initialized with NCEP/NCAR Reanalysis data. For a complete summary of simulations performed and physics parametrization options utilized, see Kiefer (2005).

A number of sensitivity experiments were performed in order to isolate the impact of specific processes believed to be important in generating the excessive total precipitation observed across the region of interest, including a smoothed terrain simulation and a dry simulation. The former experiment will briefly be considered in this paper.

b. Model Validation

In order to lend credibility to the numerical modeling study of the dynamics leading to the generation of extreme amounts of snowfall, validation of the model synoptic fields and verification of the model precipitation was performed. Simulated 300 hPa and 850 hPa isobaric analyses and surface analyses, all from a coarse (36 km NHMASS) simulation valid 1200 UTC 16 February 2003, were compared to the previously discussed NARR datasets. At each of the levels considered, the simulation results compare favorably with the observed data (not shown).

4. Model Results

a. LLJ Interaction – Moisture Transport

The juxtapositioning of the two low-level jets, i.e., maritime and continental, is the focus of this and the following section. The impact of these two low-level jet/frontal systems on the precipitation development over the region of interest will now be considered. Both jets act to transport moisture toward the region of interest, evidenced by moisture transport calculations performed with both NHMASS and NARR data for the continental jet ($\rho\bar{q}\bar{V}$; see Uccellini et al.. 1984 for methodology) in Table 1. The values are comparable to that diagnosed for the PD79 low-level jet and for springtime low-level jets during convective scenarios (Uccellini et al. 1984). Additionally, the two jets transport air of greatly differing thermal characteristics into mutual proximity. The transport of warm air in the approximately 600-850 hPa layer over the dense cold air below 850 hPa associated with the maritime low-level jet will be considered in its implication for frontal lifting and closely-related frontogenesis.

b. LLJ Interaction - Frontal lifting / Frontogenesis

In order to begin this analysis of the impact of the superpositioning of the two jet/front systems, a southwest-northeast vertical cross-section of potential temperature and wind speed from the 6km NHMASS simulation (Fig. 6) proves invaluable. Immediately noticeable are dual shear zones and frontal inversions present over the northern Mid-Atlantic region, including western Maryland. Also apparent is the slope of isentropes north and east of the highest terrain, a direct result of the strong anticyclone over southern Quebec, wherein the depth of the cold air increases as one approaches the anticyclone (see Fig. 2c2). One result of the superpositioning of the two jets and their representative airmasses was the process wherein the generally 20-30 m s^{-1} continental jet progressed north and east over the dense cold-air damming airmass evident in Figure 6. The impact of the continental jet directed up the strongly sloped isentropes is apparent in a 6 km NHMASS simulated composite reflectivity image overlaid with isobars on the 300 K isentropic surface at 0000 UTC 17 February 2003 (Figure 7). One notices that the greatest composite reflectivity values are poleward of a frontal zone generally located between 690-730 hPa just south of the Pennsylvania-Maryland border. Also apparent is a largely meridional pressure gradient located just southwest of the 45 dBz composite reflectivity maximum. Considering the direction of the continental jet from the southwest (see Fig. 3), and the orientation of the continental jet approximately normal to the strongly sloped isentropic surfaces in the vertical cross-section in Figure 6, it appears that the vigorous frontal lifting of the continental jet is playing a role in producing the zonal band of heavy precipitation in the Mid-Atlantic U.S.

Observing the layer between approximately 750 and 850 hPa in Figure 6, it is apparent that the two low-level jets, the continental and maritime, are directed in a confluent manner in that layer, a critical observation in that such an interaction has great implications for frontogenesis. It should be noted that the following treatment of frontogenesis deviates from the recent work of Novak et al. (2004) in that this study considers a fairly shallow layer of frontogenesis produced through low-level jet interaction, whereas the study of Novak et al. (2004) looked at heavy snow banding resulting from (1) deformation zones northwest of surface cyclones and (2) deep layers of frontogenesis owing to mid- to upper-level confluent flow. Comparing the band of snowfall evident in Fig. 4 to the mid- to upper-level confluent flow across the eastern U.S. in Fig. 2a2, the broad band from southern Ohio to southern New Jersey appears consistent with Novak et al.'s non-banded case (see their Fig. 15b). What is being considered in this section is the impact of the two low-level jets, the continental and maritime, on frontogenesis across the mid-Atlantic U.S., largely below the level considered in Novak et al. To assess the various contributions to frontogenesis over the mid-Atlantic U.S., this study has utilized a 2-dimensional form of the Miller (1948) frontogenesis equation in height coordinates, defined as:

$$F=\frac{1}{\left|\nabla_{\sigma}\theta\right|}\left\{\begin{array}{l}\dfrac{\partial\theta}{\partial x}\left[-\left(\dfrac{\partial u}{\partial x}\dfrac{\partial\theta}{\partial x}\right)_1-\left(\dfrac{\partial v}{\partial x}\dfrac{\partial\theta}{\partial y}\right)_2-\left(\dfrac{\partial w}{\partial x}\dfrac{\partial\theta}{\partial z}\right)_3\right]\\ +\dfrac{\partial\theta}{\partial y}\left[-\left(\dfrac{\partial u}{\partial y}\dfrac{\partial\theta}{\partial x}\right)_4-\left(\dfrac{\partial v}{\partial y}\dfrac{\partial\theta}{\partial y}\right)_5-\left(\dfrac{\partial w}{\partial y}\dfrac{\partial\theta}{\partial z}\right)_6\right]\\ +\dfrac{1}{C_p}\left(\dfrac{P_o}{P}\right)^k\left[\left(\dfrac{\partial\theta}{\partial x}\dfrac{\partial}{\partial x}\left(\dfrac{\partial Q}{\partial t}\right)\right)_7+\left(\dfrac{\partial\theta}{\partial y}\dfrac{\partial}{\partial y}\left(\dfrac{\partial Q}{\partial t}\right)\right)_8\right]\end{array}\right\}. \quad (4)$$

Terms 1 and 5 are confluent deformation terms, while terms 2 and 4 represent shearing deformation. Terms 3 and 6 represent tilting effects in the x- and y-directions, respectively, while terms 7 and 8 are diabatic heating terms. Figures 8-9 represent (a) confluent deformation, (b) tilting effects, and (c) diabatic heating effects for 2000 UTC 16 February 2003 (Figure 8) and 2100 UTC 16 February 2003 (Figure 9) during the period of heaviest precipitation over the region of interest.

An analysis of the frontogenesis equation reveals that the result of two streams of air parcels approaching the region of interest from directions varying approximately 30-40 degrees (see wind vectors in Fig. 6 near center of cross-section in 700-800 hPa layer) is a sustained stripe of confluent deformation from western Maryland into northeastern West Virginia (Figs. 8-9a), along the Alleghany mountain range (not shown). Results from a smoothed terrain simulation indicate that the frontogenetical band ceases to exist when the terrain is strongly smoothed (not shown). The other band of confluent deformation, albeit weaker, appears to have been produced through speed convergence, as low-level parcels approaching the mid-Atlantic from the south and southwest, decelerated significantly (see Fig. 3).

In light of the previous discussion on frontal lift, horizontal frontogenesis should imply a greater slope of isentropic surfaces, stronger vertical velocities and, in the presence of a saturated atmosphere, greater quantities of precipitation. Given the fact that the greatest precipitation rates are on the poleward side of the frontal zone, the impact of secondary circulations due to horizontal deformation, opposing frontogenesis and thereby producing descending motion on the poleward side (Keyser and Shapiro 1986), appears negligible. The presence of the highest simulated total precipitation (e.g. Figure 7) east of the meridionally oriented 800 hPa baroclinic zone (with predominately southwesterly 600-800hPa flow) and north of the zonally oriented front (with generally south to southeast low- to mid-level flow), appears to implicate confluent deformation in the production of heavy snowfall over the region of interest.

The tilting and diabatic terms were noted to be important in repositioning the frontal zones, while the primary source of frontogenesis, establishing and maintaining heavy precipitation over the region of interest, was horizontal deformation (primarily confluent deformation). It is this mechanism which produces narrow regions of strongly sloping isentropes, thereby generating a band of strong vertical velocity and (in a saturated atmosphere) a band of heavy precipitation, with the diabatic heating (and tilting term to a lesser degree) then gaining importance and acting to modulate the fronts locally. The tilting term, while noted here to be important especially in the presence of complex terrain, is likely of greater importance further up into the middle troposphere, as tilting effects there are expected to dominate over horizontal deformation as temperature advections are weaker and vertical motions stronger (Miller 1948). Quite apparent is the tenuous relationship between the low-level jets, their frontal counterparts, and the positioning of mesoscale bands of precipitation across the region of interest, wherein slight variations in the intensity and position of the first two phenomena can have an enormous impact on the intensity and positioning of the heaviest snowfall.

c. LLJ Interaction - Additional Impacts

Before continuing further, a few additional significant impacts of the atmospheric structure described in Figure 1 will be discussed. First, an unbalanced subtropical jet exit region upstream of a mid-tropospheric ridge and north of a surface frontal boundary is a region of the atmosphere known to be conducive to the generation of inertia-gravity wave activity (Koch and Dorian 1988, among others). Indeed, such a synoptic setup did exist during PD03, and NHMASS simulations do indicate corresponding wave activity, although a limited observational analysis has been performed to verify this. Additionally, the thermal structure and wind-shear profiles resulting from the superpositioning of the two low-level jets produces a lower-atmosphere conducive for low-level wave-ducting (Lindzen and Tung 1976). Finally, vertically-propagating inertia-gravity waves amplifying and breaking in the upper-troposphere and lower-stratosphere are one known source of lower-stratospheric turbulence (Clark et al. 2000; Lane et al. 2003). Currently, work is being performed in order to improve prediction of lower-stratospheric turbulence produced, among other sources, by breaking inertia-gravity waves.

Time (UTC)	WD (dir)	WS (m/s)	Q g/kg	Moisture Transport ($x10^3$ $kgm^{-2}s^{-1}$)
16/0000	200 (200)	16.02 (17.39)	6.13 (6.63)	95.18 (112.30)
16/0600	200	20.36	7.47	149.21
16/1200	210 (210)	16.77 (15.07)	8.94 (8.18)	145.82 (120.11)

Table 1 Moisture Transport along axis of low-level jet averaged over 750 to 800 hPa layer for 18 km NHMASS simulation and NARR data (latter in parenthesis). Included are mean wind direction (WD, deg), wind speed (WS, ms^{-1}), and mixing ratio (q, gkg^{-1}). See Fig. 1 for area over which calculation was performed. See text for discussion.

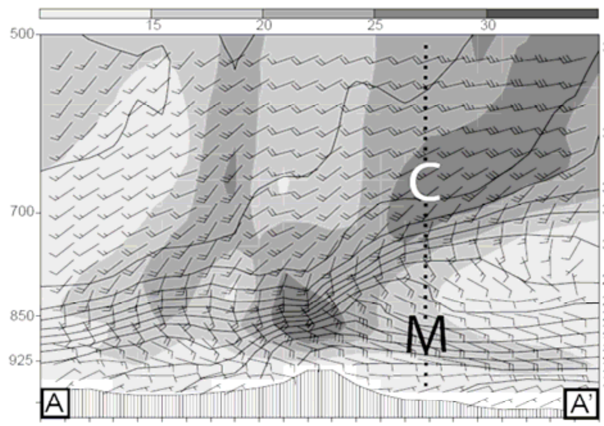

Fig. 6. 6km NHMASS vertical cross section of equivalent potential temperature [solid, K], total wind speed [shaded greater than 10 ms^{-1}] and total wind vectors [ms^{-1}], valid 0000 UTC 17 Feb 2003. Dashed vertical represents the location of western Maryland. Cross -section axis A-A' shown in Fig. 2a2. Labels "C" and "M" denote approximate positions of the continental and marine low-level jets, respectively, over western Maryland.

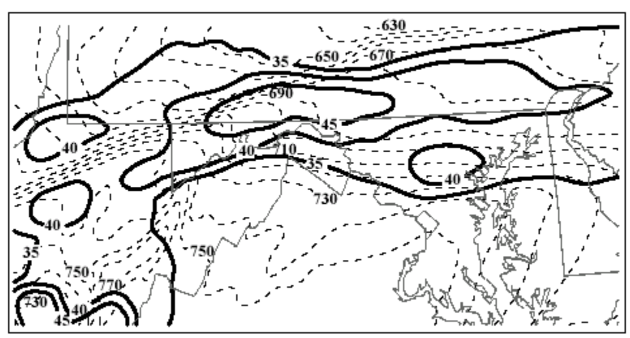

Fig. 7. 6km NHMASS Model Composite Reflectivity [solid, every 5 dBz, 35 dBz and greater] and pressure on the 300 K isentropic surface [hPa, every 10 hPa, alternate contours labeled] ending 0000 UTC 17 Feb 2003.

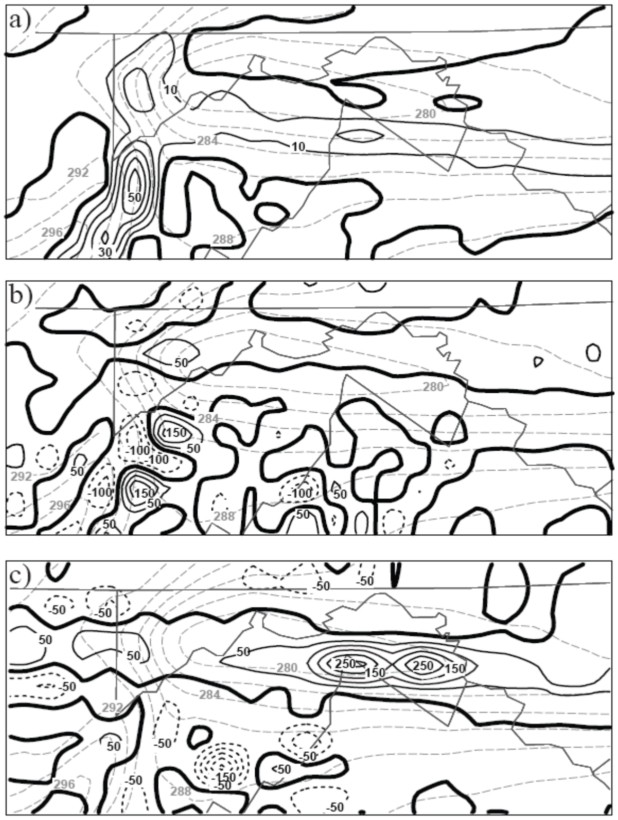

Fig. 8. 6 km NHMASS 800 hPa potential temperature (K, dashed gray lines) and frontogenetical forcing due to a) Confluent deformation (x 10^{-8}), b) tilting effects (x 10^{-8}), and c) diabatic heating (x 10^{-4}) [K/ms, solid-frontogenetical forcing, dashed-frontolytical forcing] all valid 2000 UTC 16 Feb 2003.

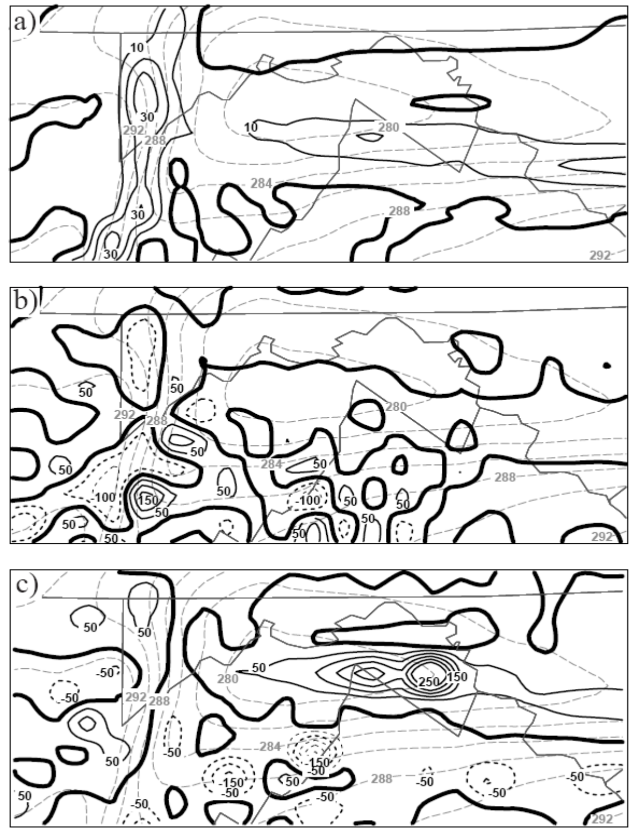

Fig. 9. As in Fig. 8, except valid 2100 UTC 16 Feb 2003.

5. Summary and Conclusions

This paper considered the impact of interaction of two jet/front systems on frontogenesis and heavy precipitation. It was shown that velocity convergence within the continental jet and confluence between the continental and maritime jets produced two bands of confluent deformation across the region of interest, one aligned with the primary terrain ridge in northeastern West Virginia and the second oriented zonally from near the first band to the Maryland shore. The positioning of a component of the continental jet normal to the two steepening frontal bands contributed to strong lift over the region of interest.

The juxtapositioning of the two low-level jet/front systems impacted not only the primary forcing for heavy precipitation over the region of interest, namely frontogenesis/warm-air advection and upper-level divergence, but also the secondary finer-scale mechanisms (such as terrain effects) that produced locally enhanced snowfall rates in the vicinity of complex terrain. These latter impacts will be considered in a future paper. Further implications of the synoptic and mesoscale atmospheric structure described in this paper were noted, including inertia-gravity wave generation and production of lower-stratospheric turbulence.

The critical point of this study is that the knowledge of the larger scale dynamic and thermodynamic structure of the atmosphere affords one the additional knowledge of the potential for fine-scale mechanisms that may lead to locally higher snowfall totals. Of note is the fact that the operational ETA model 24 hour forecast for 24 hour precipitation ending 1200 UTC 17 February 2003, encompassing the heaviest period of snowfall over the mid-Atlantic U.S., featured a precipitation maximum shifted significantly southeast of the region of interest. A forecasted continental jet weaker than that which was observed, and the coarse nature of the terrain dataset utilized by the ETA model raise questions as to how best an operational forecaster can interpret a model forecast, based on the dynamics considered within this study, and issue more accurate forecasts of snowfall in areas of complex terrain during similar winter storm scenarios. Much future work is required though before a new conceptual model can be utilized by the operational forecasting community. Additional winter storm case studies, particularly those focused in areas of small-scale complex terrain are required to evaluate the commonality of superimposed low-level jets during winter storms. With these additional efforts, the goal of improving operational forecasting of extreme snowfall during winter storms, and thereby improving the well-being of the public-at-large during such events, may be accomplished.

This study began with a number of reports of snowfall totals exceeding 100 cm across western Maryland and northeastern West Virginia during the 2003 Presidents' Day winter storm. It is believed that an assessment of the development and modification of low-level jet/front systems similar to the continental and maritime in this study will improve the timely and accurate prediction of such extreme snowfall, through improved understanding of the anticipated multi-scale processes capable of generating locally enhanced precipitation. While much more work is necessary to achieve the goal of improving operational forecasting of extreme snowfall in complex terrain, the author considers this endeavor as an important intermediate step between the previous studies on winter storm precipitation contributing to the conceptual model presented in Kocin and Uccellini (1990) and ongoing work toward a new conceptual model of winter storm precipitation incorporating the continental jet discussed here (of which Figure 1 is a prototype).

7. Acknowledgements

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