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ANTICIPATING DAMAGING FOEHN WINDSTORMS EAST OF THE CENTRAL APPALACHIANS

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1. INTRODUCTION

Lower tropospheric flow over mountain barriers can have a significant impact on downslope weather. Studies over the last two decades have focused on mountain waves and resultant lee slope high winds, but most of the assessments for North America have been documented over the western most part of the continent. The Santa Ana winds (Burroughs 1987; Lessard 1988) and Sundowner winds (Blier 1998; Ryan and Burch 1992) of southern California, the Taku Winds (Coleman and Dierking 1992; Dierking 1998) near Juneau, Alaska, and the Chinook Winds (Oard 1993; Nkemdirim 1986; and many others) of the Front Range of the Rockies have all been directly correlated to mountain waves. Large changes in surface temperature and relative humidity have also been noted with these events, but it is the strong, gusty surface winds that often lead to damage of personal property and danger to public safety.

For eastern North America, specifically the Appalachian Mountains, these downslope "foehn" wind events are also observed. Gaffin (2002) showed a case over eastern Tennessee where a mountain wave likely produced large changes in observed surface temperature and humidity, but only a modest increase in wind speed. Brady and Waldstreicher (2001) discussed the influences of mountain waves on precipitation shadows in the Wyoming Valley of northeast Pennsylvania. However, no documented cases of severe downslope windstorms in the Appalachians were found. While they seem to be less common than their western North American counterpart due to the lower terrain profile of the Appalachians, these severe downslope windstorms can occur. For the central Appalachians, predominate westerly flow favors the development of these wind events on the eastern slopes.

This paper will show an eastern North American case where wind, temperature, and relative humidity all responded significantly to the presence of an amplified mountain wave. This case will also be compared to other central Appalachian high wind events from the 2002-2003 cool season, most of which exhibited little or no warming on the lee side of the mountains, with the intent to establish pattern recognition. In doing so, it is our hope to increase the awareness of severe foehn winds in the eastern U.S., and provide forecasters with synoptic patterns, as well as thermodynamic and wind profiles, that favor the occurrence of these strong downslope winds.

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2. DATA

For this study we focused on the 2002-2003 cool season (December - February) and identified six high wind events, all of which were confined to the eastern slopes of the central Appalachians (Fig 1). An event was defined if non-thunderstorm wind damage or a measured severe wind gust (≥ 26 m s^{-1}) was observed in three or more counties. The six events are listed in Table 1. Events 3 and 4 (8 and 9 January 2003 respectively) both displayed distinct foehn characteristics: a significant warming and drying on the lee side of a mountain ridge. They are considered two distinct events since there was a significant separation between the periods where damaging winds were reported. Wind damage associated with these two events covered a number of counties, but was rather spotty in nature. The other events in the table were not associated with significant lee side warming, likely due to the fact that strong cold advection in the lower troposphere was masking any effects of adiabatic warming. Wind damage with a couple of these cold advection events was widespread in many of the counties reporting damage. A synoptic overview of all the events will be provided in the next section.

Surface and upper air data were collected for all the events from a combination of internal National Weather Service datasets, as well as on-line data sources. Observational data and model initial analyses were used to assess synoptic patterns, temporal trends in surface parameters, and vertical thermodynamic and wind profiles. The Blacksburg, VA (KRNK in Fig. 1) sounding location lies in the heart of the central Appalachians and is therefore a representative location for examining vertical profiles associated with these events. However we also examined soundings from Dulles VA (KIAD) and Greensboro NC (KGSO) just downstream (to the east) of

Event (Date)	Time of Occurrence	No. Counties with reported damage	Max observed wind gust ($m\ s^{-1}$)
1 (14 Dec ‘02)	12 UTC - 19 UTC	3	27 (54 kts)
2 (25 Dec ‘02)	16 UTC - 23 UTC	25	31 (61 kts)
3 (8 Jan ‘03)	06 UTC - 15 UTC	15	29 (57 kts)
4 (9 Jan ‘03)	05 UTC - 10 UTC	13	32 (63 kts)
5 (24 Jan ‘03)	00 UTC - 08 UTC	14	44 (87 kts)
6 (23 Feb ‘03)	06 UTC - 20 UTC	26	26 (52 kts)

Table 1. Non thunderstorm wind events are listed chronologically for the central Appalachians during the 2002-2003 cool season. The peak wind gust in Event 5 was estimated based on damage. Gusts for the remaining events came from actual measurements. The number of counties affected are preliminary and subject to change when Storm Data is published.

the Appalachians. No upstream soundings were examined, since they either were too far upstream or too far north to be considered representative of the area of study.

In addition, satellite data were used in some cases to confirm the presence of mountain waves. Radar data were not particularly useful in these events since environments were relatively dry with little in the way of reflectivity data to provide adequate wind observations.

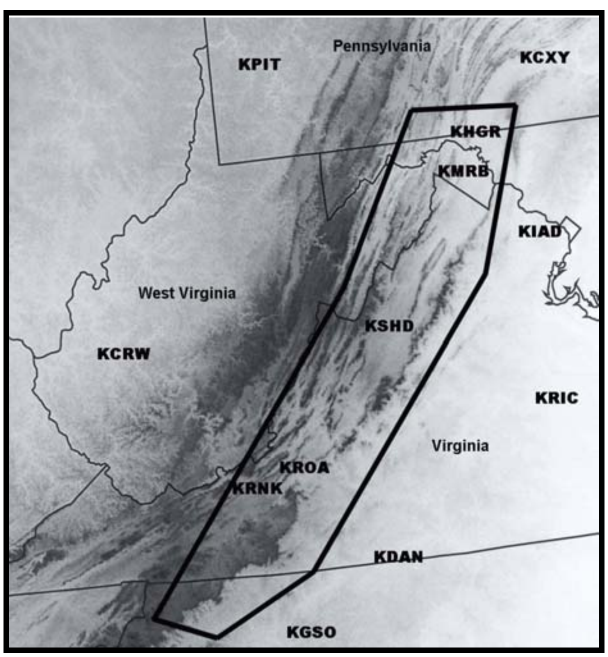

Fig.1. Location of central Appalachian windstorms examined.

3. THE COLD AIR ADVECTION EVENTS

Figure 2 shows a series of upper air and surface charts for each of the four cold air advection events in the study, at the closest synoptic time to the height of the high wind. The 300mb height and wind fields are in the left-hand column, 850mb heights, temperatures, and winds are in the middle, and isobars of mean sea level pressure as well as 1000-500mb thicknesses are shown in the right-hand column. In all four cases, a deep large scale trough is in place at 300 mb over the eastern United States, with the jet stream to the south or southeast of the Appalachians. Cold air advection, often very strong, is apparent at 850mb with the low centered north or northeast of the central Appalachians. At the surface, a deepening cyclone is generally just to the northeast of the central Appalachians, except in Event 5 (Fig. 2c) where it is already well offshore. As a result, dramatic surface pressure rises occur as the surface cyclone moves out of the area. These pressure rises, primarily on the west side of the Appalachian mountains, ranged from 6 to 18 mb in 12 hrs for these four cases. In contrast the surface pressure tendencies in Event 3 and 4 were nearly neutral or only slightly rising.

Since significant synoptic scale pressure rises are commonly observed to be associated with widespread high wind events in the Appalachians and forecasters have relied on this signature for some time to anticipate wind events, Event 3 and 4 are of particular interest since they were not accompanied by the significant pressure rises and strong cold advection.

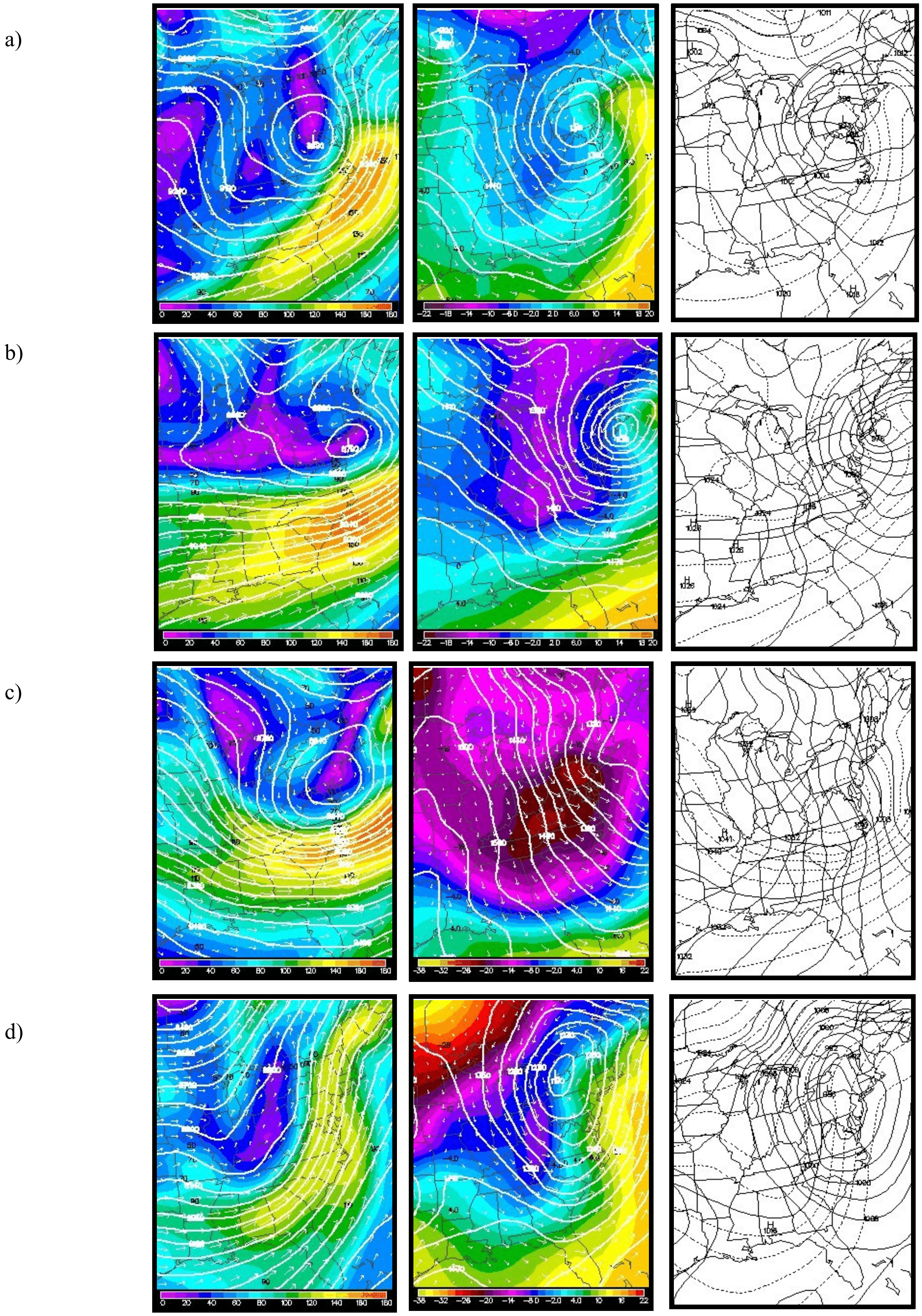

Fig. 2. Eta initial analyses of 300mb heights, wind vectors and image field of isotachs (left column); 850 mb heights, wind vectors, and image field of temperature (center column); and mean sea level pressure (solid) and 1000-500mb thickness (dashed) (right column) for (a) 12 UTC 14 Dec 2002 - Event 1, (b) 00 UTC 26 Dec 2002 - Event 2, (c) 00 UTC 24 Jan 2003 - Event 5, and (d) 12 UTC 23 Feb 2003 - Event 6. Eta panels courtesy of Unisys.

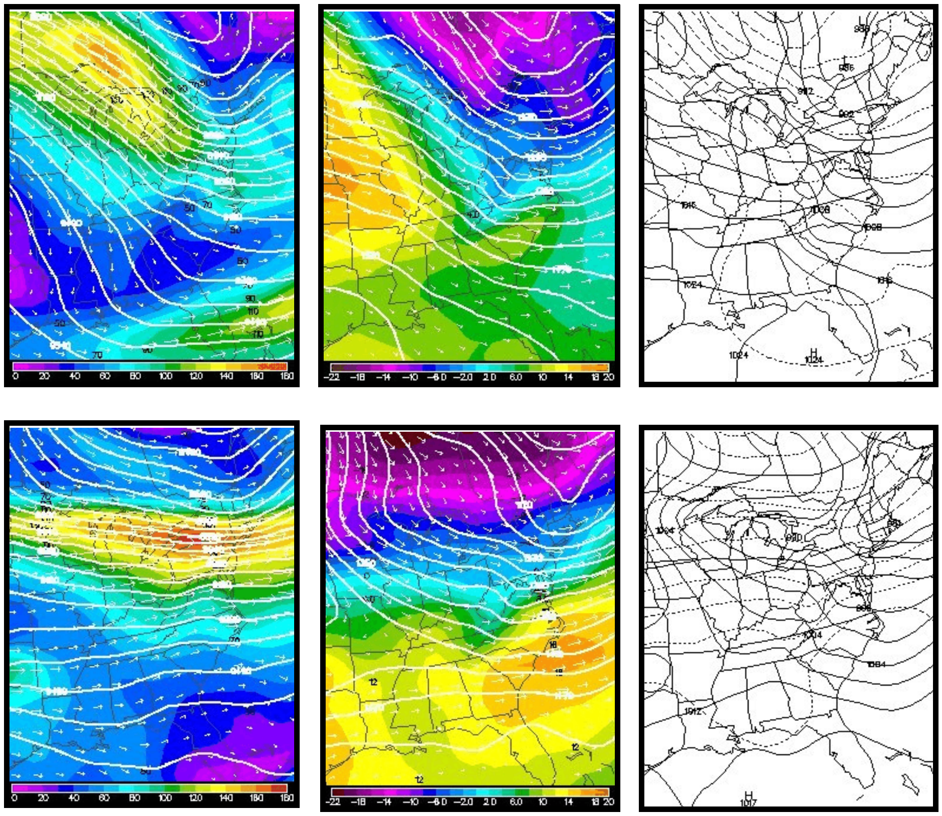

Fig. 3. Same as Fig. 2, but for (a) 12 UTC 8 Jan 2003 - Event 3, and (b) 12 UTC 9 Jan 2003 - Event 4.

4. THE 8-9 JANUARY 2003 FOEHN EVENTS

4.1 Synoptic Overview

Figure 3 shows a similar series of charts for the 8-9 January 2003 cases (Events 3 and 4) as was shown in Fig. 2 for the cold advection cases. By contrast, the jet stream is to the north of the central Appalachians with a less amplified large scale trough at 300 mb over the eastern U.S. There is warm advection pushing into the western Appalachian region at 850 mb (seen in Kentucky and West Virginia in the center column of Fig 3a and 3b, and the surface cyclone or surface trough is over New England and extending westward over the Great Lakes. At mid levels (not shown), two short waves passed near or just north of the central Appalachian region, one on the morning of 8 January, the second during the early morning of 9 January. Both waves were associated with 50 m s^{-1} mid level jet streaks, and corresponding 850 mb winds of 25 to 30 m s^{-1}.

In addition, substantial warm air advection was most notable within a 200 mb layer from 850 to 650 mb, where a 15 ^{0}C rise in temperature was observed between 00 UTC 7 January and 00 UTC 8 January. The 48 hour (00 UTC 7 January to 00 UTC 9 January) net change for the same layer was +25 to +30 ^{0}C.

4.2 Foehn Signatures at the Surface

During these events, observed surface temperatures on the immediate lee side of the Appalachian mountains either warmed during the overnight hours, or remained nearly steady (remaining relatively warmer than upstream or downstream observations). Meteograms for two representative locations for 9 January (Fig. 4) show that temperatures warmed 5 ^{0}C (9 ^{0}F) between 04 UTC and 06 UTC at Roanoke, Virginia (KROA), and 12 ^{0}C (20 ^{0}F) at Martinsburg, West Virginia (KMRB). The locations of KROA and KMRB are shown in Fig.1. A drop in surface dewpoint was noted at KROA, while dewpoints remained steady or rose slightly at KMRB; this corresponds to a 30 to 50% reduction in relative humidity at KROA and KMRB respectively. Wind speeds increased from the west at both stations shortly after the significant temperature rises began, with gusts to 14 m s^{-1} (28 kts) at KROA and 22 m s^{-1} (43 kts) at KMRB. In the cold advection cases, nocturnal temperature rises were not observed.

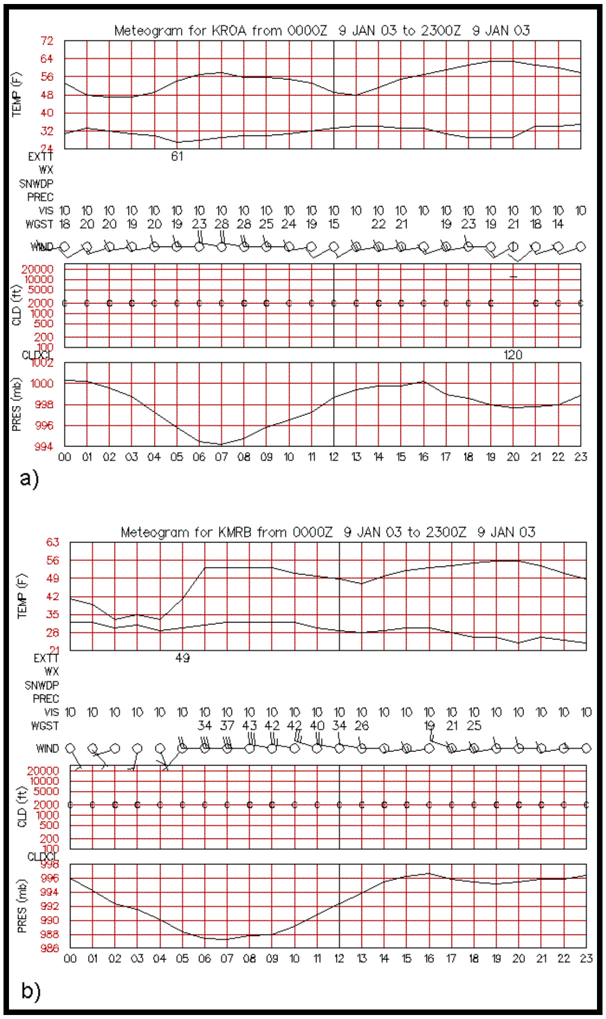

Fig. 4. Meteograms from 00 -23 UTC for 9 Jan 2003 for a) Roanoke, VA (KROA) and b)Martinsburg, WV (KMRB). Locations of sites shown in Fig. 1. Most recent time is on the right. Temperature and dewpoint on top graph given in ^{0}F, wind speed in knots, pressure in mb. EXTT is midnight to midnight max temperature. Courtesy of Plymouth State College.

It should also be noted that while these lee side observation sites showed local pressure rises after the winds increased (seen at the bottom of Fig 4a and 4b), locations upstream of the Appalachians did not experience significant pressure rises before or during the high winds events of 8 or 9 January, which is in sharp contrast to the four cold advection events.

4.3 Satellite Signature

Cloud patterns from the infrared satellite imagery, especially the early morning of 9 January (as seen at 1030 UTC in Fig. 5), showed strong evidence for the presence of an amplified mountain wave. Only one wave crest is visible at cirrus level where there was apparently enough moisture to produce clouds. In animations the western edge of these clouds was observed to remain stationary, giving the impression of streaming or "burning" cirrus since it appeared to be originating from a fixed source, similar to smoke from a line of fires. In addition, a loop of water vapor imagery (not shown) indicated areas along and just downwind of the highest terrain (and just upstream of the wave crest) where drying from subsidence was maximized. This subsidence signature was present the morning of 8 January, across western Virginia and northwest North Carolina, but by the morning of 9 January began gradually shifting north into northern Virginia and east-central Pennsylvania (by 12 UTC).

The wave-induced cirrus in IR imagery could also be seen shifting poleward almost 4 degrees latitude during the early morning hours of 9 January as an upper level jet streak passed to the north. Damage reports on the morning of 8 January were mainly from near Roanoke, Virginia (KROA) southward to northwest North Carolina. On the morning of 9 January a much larger area of wind damage occurred, from just south of Roanoke, northeastward along the east slopes of the Appalachians to near Martinsburg, West Virginia (KMRB). In fact, when satellite imagery was overlaid with the actual damage reports, it was found that the most significant wind, temperature and humidity changes occurred just upstream of the resultant wave cloud and just to the lee of the mountain barrier.

5. DISCUSSION

5.1 Mountain Wave Theory

Topography for the central Appalachians features a southwest to northeast continuous mountain range with average orographic relief just under 1000 m. The eastern slopes are typically steep with gentle rising slopes on the western aspect. This west-to-east profile (shown in Fig. 6) fits empirical criteria established by Queney et al (1960), and theoretical support by Smith (1977), and Lilly and Klemp (1979), that mountain barriers with steep lee slopes and gentle windward slopes are the most effective generators of large-amplitude mountains waves. Although vertically challenged compared to the mountainous terrain of western North America, the Appalachian terrain profiles can support the development of large-amplitude waves and associated downslope windstorms if environmental conditions are favorable.

Previous studies of atmospheric flows over mountain barriers have identified several dynamic mechanisms to explain downslope windstorms. Colman and Dierking (1992) summarized three important conditions favorable

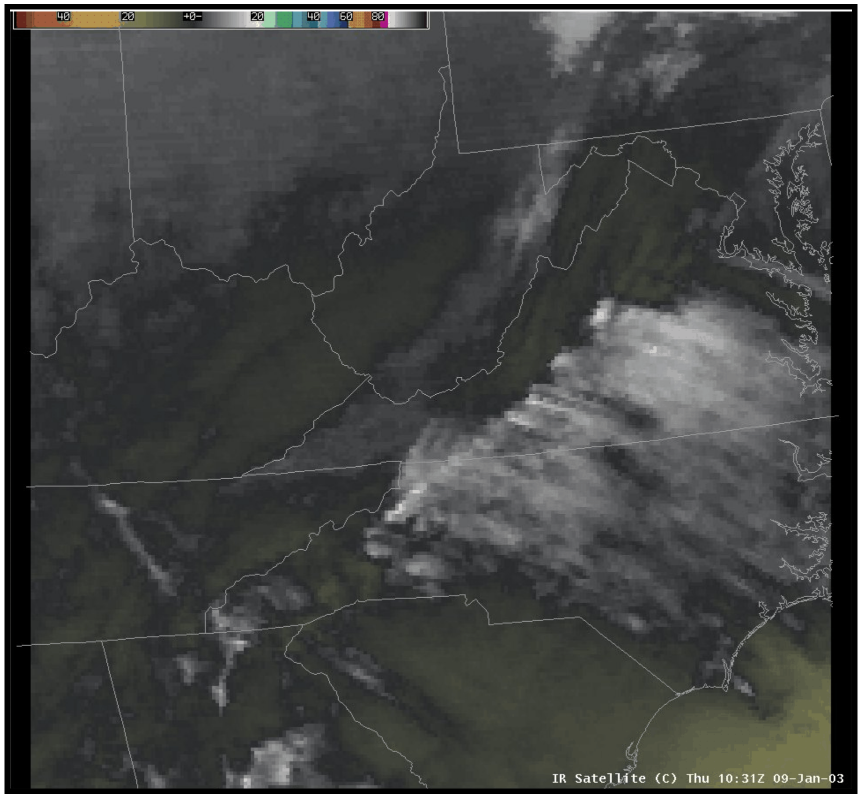

Fig. 5. IR satellite image at 1030 UTC 9 Jan 2003.

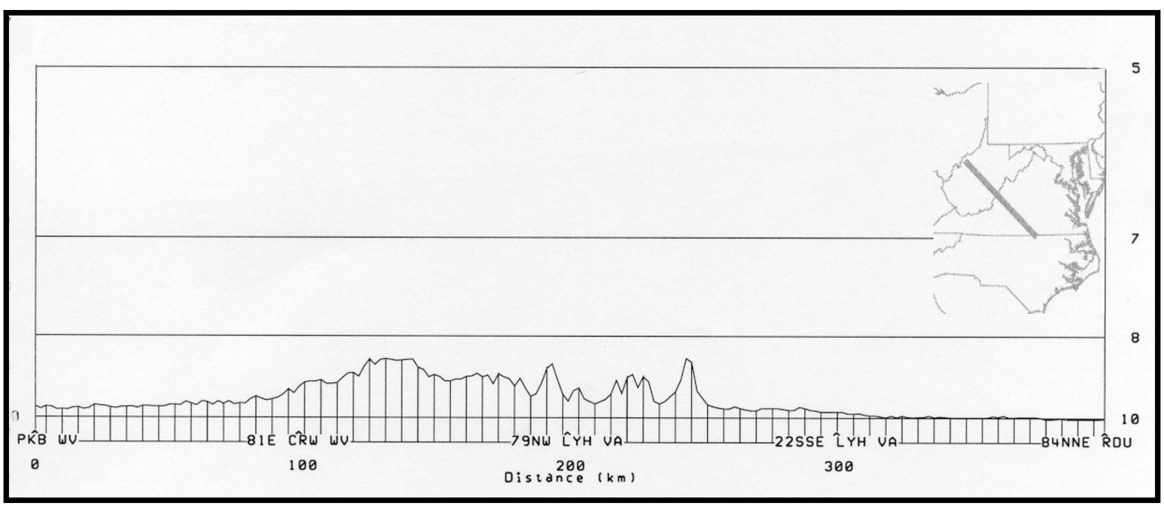

Fig. 6. A representative cross-section taken normal to the central Appalachians. Heights given on y-axis are x10^2 mb. View is to the northeast; the cross-section is taken from northwest (left margin) to southeast (right margin).

for strong downslope winds, which others have also highlighted: 1) an inversion (stable layer) at or just above ridge top, 2) strong cross-barrier flow near ridge-top (typically 15 to 20 m s^{-1} within 30 degrees of perpendicular), and 3) cross-barrier flow decreasing with height to zero (defined as a critical level). Lilly and Klemp (1979) used non-linear theory to suggested the critical level helps to reflect and amplify the wave. Whether a pre-existing critical level is necessary to amplify a mountain wave, or a "self-induced" critical level created by the wave itself can cause further amplification (Peltier and Clark, 1979) is a topic of debate. Finally, Klemp and Lilly (1975) emphasized the need for a relatively deep layer of unstable air in the mid troposphere, above the stable layer.

5.2 Vertical Structure on 9 January 2003

Soundings from Blacksburg VA (KRNK) from the mornings of 8 January and 9 January 2003 (near the times of two separate high wind events) showed vertical structures similar to that described above. Figure 7, from 1200 UTC 9 January, indicates the presence of strong cross-barrier (northwest) winds of 35 to 40 ms^{-1} (70 to 80 kts) near and just above a stable layer. The base of the stable layer is at approximately 1500 m above the ground level, which would place it approximately 800 to 1000 m above the higher ridges in the central Appalachians. A deep layer of instability was also observed above the stable layer, with a very high tropopause for mid winter. There was no evidence, at least not from the observed sounding data, of a critical level.

An examination of the Eta model initial analysis cross-sections of model vertical motion, temperature, and wind profiles (model resolution is 12 km in the horizontal, but displayed on a 40 km grid and vertically in 25 mb increments) did not show any indication of a critical level at 12 UTC 9 January 2003 (Fig. 8). However, Fig. 8 does show that the Eta model resolved the mesoscale subsidence immediately downwind of the Blue Ridge (eastern-most ridge in Figs. 1 and 6), which tilted upward toward the west where significant synoptic scale subsidence was occurring. The Eta model also resolved the deep upward motion just downstream of the Blue Ridge maximized at mid and upper levels, and tilting back upstream with height. This deep ascent was coincident with the wave cloud formation observed in Fig. 5.

It is interesting to compare this Eta cross-section with a figure from Lilly and Zipser (1972), created from aircraft observations during a Boulder, CO downslope windstorm on 11 January 1972 (Fig. 9). Vertical motion and stability implied by the isentropes indicate not only the vertical extent of the upward motion in the lee of the mountains but a sense of the tilt in the upstream direction with height. Strong low-level downslope flow also occurred immediately downstream of the Continental Divide in the Boulder cross-section per Lilly and Zipser (1972). This is similar to the structures of the downward vertical velocities in Fig. 8 just downstream of the Blue Ridge. Therefore, it appears that the Eta model can resolve, to some degree, aspects of large amplitude mountain waves in the lee of the Appalachians.

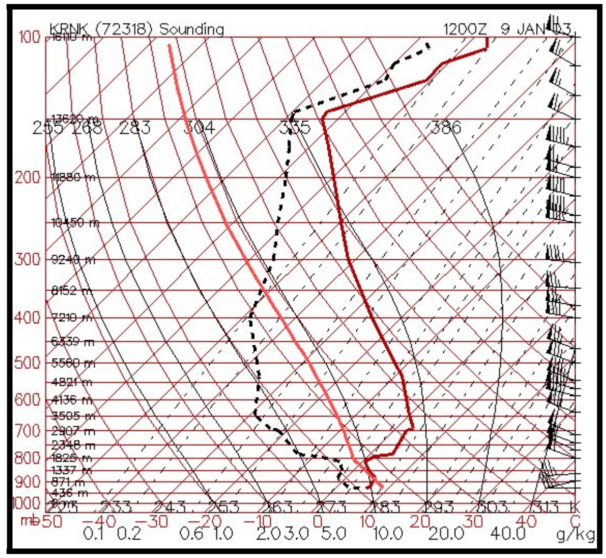

Fig. 7. Blacksburg, VA (KRNK) rawinsonde for 12 UTC 9 Jan 2003. Temperatures are in ^{0}C and winds are in knots. Location of KRNK is shown in Fig. 1. Courtesy of Plymouth State College.

5.3 Comparison of Soundings from Other Events

Soundings for the four cold air advection high wind events for the 2002-2003 cool season also indicated strong cross-barrier flow with wind speeds ranging from 20 to 35 m s^{-1} (40 to 70 kts). Inversions were present near or just above ridge top level in all four events. However for 14 Dec. 2002 (Event 1), where damage was limited to three counties and mainly in the highest elevations, the inversion was somewhat higher than the others (roughly 1500 m above the highest ridges), and a little weaker as well. In addition, all of the cases, except for 24 Jan. 2003 (Event 5), indicated the presence of a deep layer of instability just above the stable layer. The 24 Jan. 2003 sounding (shown in Fig. 10) also indicated a possible critical level (near 4 km) as the winds became northerly and weakened with height, thus decreasing the cross-barrier flow. Given the orientation of the Appalachians, the wind would need to be more northeasterly for the cross-barrier flow to go completely to zero.

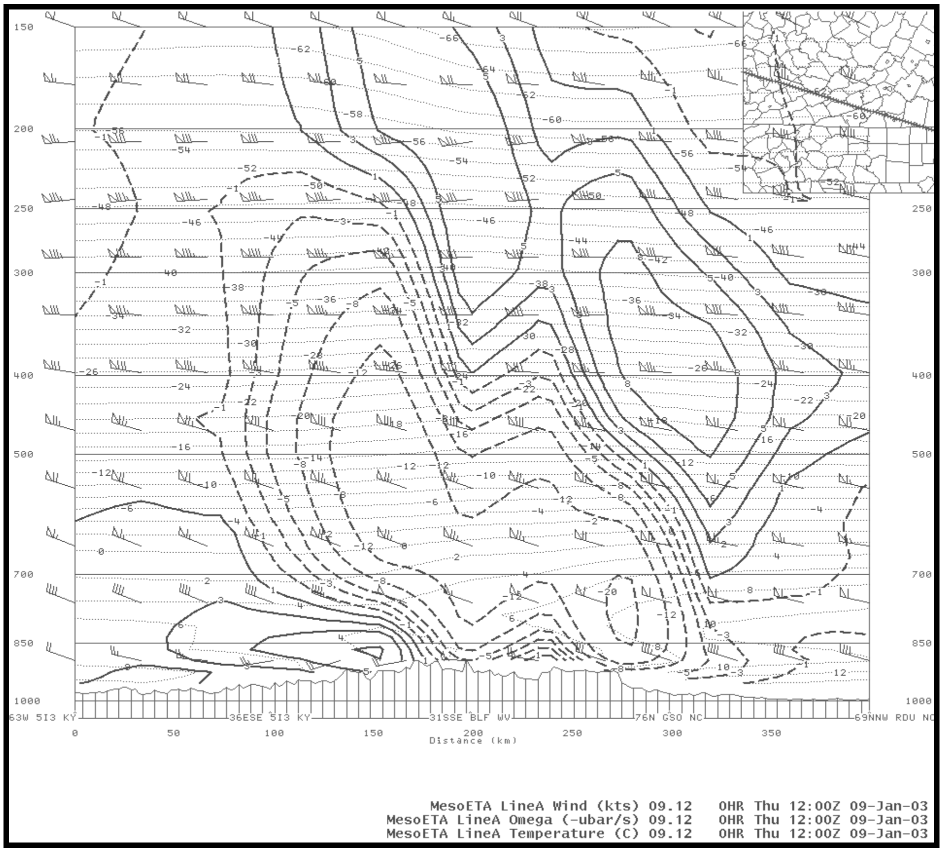

Fig. 8. 12km Eta model initial analysis (viewed at 40km horizontal and 25mb vertical resolution) cross-section for 12 UTC 9 Jan 2003. Left side (northwest corner) is near Jackson, KY, right side (southeast corner) is just east of Danville, VA; cross-section length is 400 km with the orientation shown in upper right. Dark solid (dashed) contours are model upward (downward) vertical motion respectively, in -µbar/s. Thin dotted lines are temperature in ºC. Wind barbs are shown in knots. Terrain is shown as hatched area at bottom. Heights are in mb.

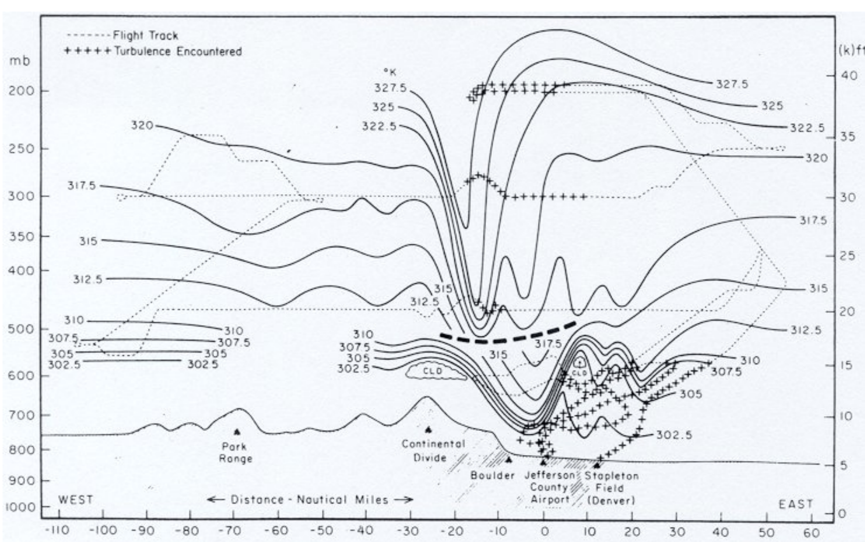

Fig. 9. Cross-section of the potential temperature field (solid lines) observed in a very strong mountain wave over Boulder, CO, on 11 January 1972. Heavy dashed line separates observations taken at different times; the dotted lines show the aircraft flight tracks; the crosses indicate regions of turbulence. (From Lilly and Zipser, 1972).

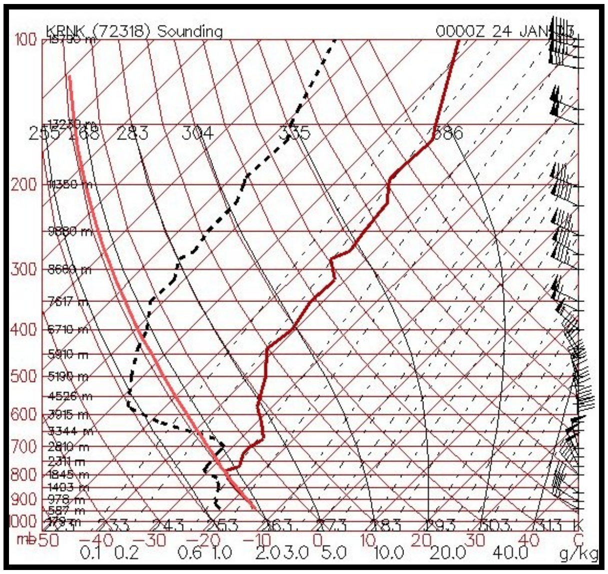

Fig. 10. Same as Fig. 7, but for 00 UTC 24 Jan 2003.

A difficulty with respect to the cold advection cases was caused by the rapidly changing environments during those events. The degree to which vertical snapshots once every 12 hours are truly representative of the atmosphere at the time of the highest surface winds is questionable. Nevertheless, the similarities with the two foehn events in terms of the strong cross-barrier flow, stable layer near ridge top, and in most cases the unstable layer above the inversion are important to note.

The implications are that the vertical structures in the cold air advection cases were generally supportive of amplified mountain wave production just as the two foehn events were, but not always as clearly. Since two of these cold advection cases feature very widespread wind damage that included ridge tops as well as lee slopes and even locations well to the east of the mountains, it is unclear whether or to what extent an amplified mountain wave was contributing to the wind damage. More precise locations of wind damage or estimates of wind speeds could aid in this determination.

6. CONCLUSIONS

The orientation of the central Appalachians suggests the eastern slopes will experience the highest frequency of severe downslope winds given the climatologically favored westerly flow. For downslope wind damage to take place, our findings from the 2002-2003 cold season confirm the importance for cross barrier flow to exceed 20 ms^{-1}, with an inversion near or just above ridge top. A critical level may or may not be observed, but its presence would favor deeper amplification of a mountain wave according to theory.

Deepening surface cyclones and their associated pressure gradients tend to produce very favorable conditions for strong downslope winds when the central Appalachian mountains are in the southwest quadrant of the cyclone. The net synoptic scale subsidence, in addition to any subsidence from the presence of a mountain wave, tends to result in enhanced wind damage along the higher ridges and just downwind of the mountain barrier. Strong gradient winds may be observed farther downstream, but evidence shows the greatest damage typically occurs closer to the mountains where the potential for mountain wave contributions exist.

In the absence of strong cold advection and surface pressure rises in the wake of an exiting surface cyclone (such as the true foehn or warm advection events presented here), synoptic signatures leading to damaging lee slope winds may be more subtle. The vertical structure prerequisites for the development of a mountain wave still apply. However, in these cases the role of a high amplitude vertically propagating mountain wave is probably more important. The presence of a single wave cloud (e.g. "burning cirrus") in the satellite imagery, enhanced subsidence seen on water vapor imagery near the eastern side of the mountains, and sudden rises of surface temperature, accompanying lowering humidity, and increasing winds immediately in the lee of the mountains are signals of a possible amplified mountain wave. Synoptic scale subsidence above the inversion as well as strong warm air advection in that layer may also be acting to strengthen the inversion, possibly contributing to the amplification process.

The 8 and 9 January 2003 events are atypical of the majority of central Appalachian windstorms in that they were not associated with strong cold air advection and significant pressure rises behind a rapidly deepening cyclone. To better anticipate severe downslope windstorms in the central Appalachians it is important for forecasters to understand the similarities between all these events in terms of mountain wave meteorology, as well as recognize the differences in synoptic patterns and surface changes. True foehn events will likely not be associated with strong cold advection and significant surface pressure rises on the west slopes of the Appalachians, but can still produce severe downslope winds with significant contributions from an amplified mountain wave.

7. NEED FOR ADDITIONAL STUDY

The 2002-2003 season study was limited to one season and did not include events where wind damage was not observed. Conditions which favor enhanced downslope winds, although not necessarily severe, occur quite often just in the lee of the central Appalachians with noticeable fluctuations of temperature, wind, and humidity. This poses a number of forecast challenges, especially to aviation and fire management. The analysis of "null" events, as well as additional severe wind events, should help lead to the development of some specific thresholds for forecasting severe downslope windstorms, as well as better pattern recognition.

Analysis and prediction of these and similar events with a high resolution mesoscale model is also an important next step. We are expecting to eventually set up the workstation Eta model, and will run it at the lowest resolution available in hopes of better resolving these and other types of orographic-related phenomena.

8. ACKNOWLEDGMENTS

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