SYNOPTIC CLIMATOLOGY OF NORTHWEST FLOW SNOWFALL IN THE SOUTHERN APPALACHIANS

Lester Baker Perry

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Approved by
Advisor: Charles E. Konrad, II
Reader: Peter J. Robinson
Reader: Thomas Whitmore
Reader: Thomas W. Schmidlin
Reader: Walter Martin

ABSTRACT

LESTER BAKER PERRY: Synoptic Climatology of Northwest Flow Snowfall
in the Southern Appalachians
(Under the direction of Charles E. Konrad, II)

Snowfall in association with low-level winds out of the northwest is a common occurrence at higher elevations and along windward slopes in the Southern Appalachian Mountains. These northwest flow snow (NWFS) events typically have low temperatures and considerable blowing and drifting snow. Due to the high degree of spatial variability of snowfall and limited ability of numerical models to predict these events, forecasting NWFS remains a challenge. This dissertation analyzes the synoptic climatology of NWFS events in the Southern Appalachians for the period 1950 to 2000. Hourly observations from first-order stations, daily snowfall data from cooperative observer stations, and National Center for Environmental Prediction (NCEP) reanalysis data are utilized to identify NWFS events, defined here as snow events with 850 hPa northwest flow (270 to 360 degrees) at the hour of greatest snow extent. Atmospheric fields of temperature, wind, moisture, and associated variables are analyzed for heavy and light snowfalls separately by calculating composite field values and constructing composite plots of the synoptic patterns. The NOAA Hysplit Trajectory Tool is used to calculate 72-hour antecedent upstream air trajectories, and composite trajectories are mapped in a geographic information system (GIS). The sample of events in the trajectory analysis is limited to those with synoptic-scale subsidence, a frequent occurrence with NWFS. Analyses of vertical soundings are coupled with NCEP data to determine the synoptic characteristics associated with different air trajectories. Results indicate that NWFS accounts for as much as 56 percent of mean annual snowfall along the higher elevation windward slopes. Heavy NWFS events are tied to higher values of synoptic-scale ascent and relative humidity in the lower troposphere, as well as lower 500 hPa heights and longer event durations. Additionally, upstream air trajectories with a Great Lakes connection have higher composite mean areal and maximum point snowfall totals along the higher elevation windward slopes than other northwest trajectories. Little Great Lakes influence is noted at lower elevations and on leeward slopes.

To Patience, Holden, and Chase

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CHAPTER I
INTRODUCTION

1.1. Research Context

Northwest flow snow (NWFS) events are responsible for nearly 50 percent of average annual snowfall totals along the windward slopes and higher elevations of the Southern Appalachian Mountain region. The low temperatures and considerable blowing and drifting of snow, coupled with the significant spatial variability of snowfall, substantially increase the societal impacts. In some cases, snowfall can be quite heavy, as evidenced by the widespread totals of 51 to 76 cm (20 to 30 in) along the windward slopes and higher elevations of eastern Tennessee and western North Carolina during an event on 18-20 December 2003. Significant late-season NWFS events also occurred during 2-6 April 1987 (up to 152 cm or 60 in) and 6-9 May 1992 (up to 102 cm or 40 in) (Fishel and Businger 1993, Sabones and Keeter 1989, NWS 1987). More commonly, however, NWFS results in lighter accumulations and occurs in association with mid-level synoptic-scale subsidence and moisture limited to below 700 hPa, with topography and convection providing the necessary forcing (Fig. 1.1). The general synoptic environment, therefore, displays certain characteristics that are associated with lake-effect snowfall (LES) in the Great Lakes region—for example, a shallow moist layer that is present beneath a capping inversion (e.g. Niziol et al. 1995, Lackmann 2001). One important difference is the role of topography in controlling the spatial patterns of NWFS, maximizing snowfall on the northwest (windward) slopes and leading to pronounced shadowing on southeast (leeward) slopes (Perry and Konrad 2006).

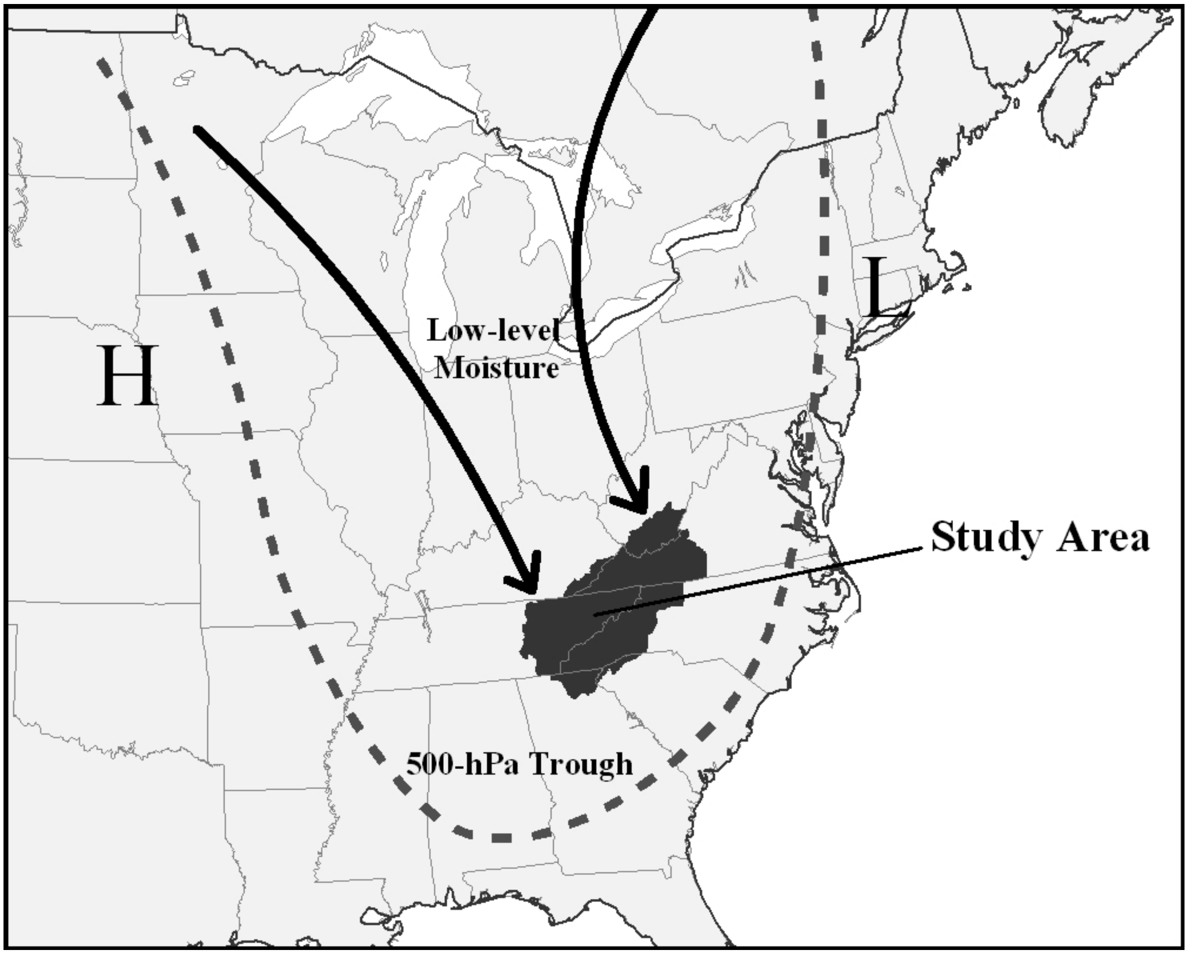

Figure 1.1. Location of study area and typical synoptic pattern for NWFS.

In this dissertation, NWFS events are identified on the basis of low-level northwest flow from a 50-year database of snowfall events constructed from daily cooperative observer data in the Southern Appalachians. The dissertation utilizes a synoptic perspective in the analysis of NWFS, explicitly linking the synoptic-scale (sub-continental) circulation patterns and NWFS events over a 50-year period. Through analyses of different synoptic fields (e.g. temperature, moisture, vertical velocity), comparisons are made between NWFS events of varying intensity in order to more fully understand the differences in synoptic fields between ordinary (i.e. light) and extraordinary (i.e. heavy) events. The NOAA Hysplit Trajectory Tool is also used to calculate 72-hour antecedent upstream air trajectories, and composite trajectories are mapped in a geographic information system (GIS) for a sample of events characterized by synoptic-scale subsidence. Analyses of vertical soundings are coupled with NCEP reanalysis data to determine the synoptic characteristics associated with each trajectory class.

1.2. Background

A synoptic climatological analysis of Northwest Flow Snowfall (NWFS) in the Southern Appalachians must be informed by several bodies of literature. This section provides background on the atmospheric processes associated with snowfall development, synoptic patterns of snowfall with special emphasis on the eastern U.S., and orographic and mesoscale processes and patterns of snowfall. Lastly, the chapter concludes with an examination of two major snowy events/periods in the Southern Appalachians in which NWFS contributed partly or entirely to snowfall totals. These two case studies were chosen to help illustrate some of the synoptic and spatial patterns associated with NWFS in the Southern Appalachians.

1.2.1. Atmospheric Processes of Snowfall Development

Snow, like other forms of precipitation, develops within a range of synoptic, mesoscale, and cloud environments. This section focuses specifically on the atmospheric processes of snowfall development and is organized in the following manner. First, a review of the cloud microphysical processes important for snowfall development is offered. It is followed by an overview of the vertical profiles of temperature and moisture associated with different synoptic patterns of snowfall. The final part of this section discusses the different mechanisms through which air may be lifted sufficiently for condensation, deposition, and associated snowfall development to occur.

1.2.1a. Cloud Microphysical Processes

Ice crystals grow to form snowflakes in three major ways: deposition, riming (or accretion), and clumping (or aggregation). Deposition involves the growth of ice crystals through water vapor diffusion directly to the ice nuclei and is more prevalent at colder temperatures. Riming occurs when water droplets collide and freeze directly onto an ice crystal, sometimes producing a heavily-rimed snowflake called graupel. Clumping, or the collision of ice and snow crystal with other crystals, occurs most frequently with temperatures at or just below 0º C and can produce very large snowflakes (Pruppacher and Klett 1997). Riming at temperatures near freezing can also contribute to ice crystal growth via collision-coalescence because rimed particles have liquid on the surface that helps to cement them together and they also have a higher likelihood of collisions due to different fall rates associated with a range of ice crystal sizes (Byers 1965a).

Snowflakes come in many different shapes and sizes, which are a direct result of the different cloud microphysical environments responsible for the growth of the ice crystals. Cloud temperature, through its influence on the concentration of liquid water droplets versus ice crystals, is the most fundamental influence on snowflake size and shape. Large bodies of pure liquid water become ice below 0 ˚C; however, much smaller droplets of water—such as those found in clouds—actually freeze at temperatures considerably below 0 ˚C. Therefore, at temperatures just below 0 ˚C, liquid water droplets significantly outnumber ice crystals. As cloud temperatures approach -15 ˚C, abundance of both supercooled liquid water and ice crystals occurs. Below approximately -20 ˚C, ice crystals predominate, and by -40 ˚C, clouds are comprised almost exclusively of ice crystals (Rogers and Yau 1989, Ahrens 1991, Pruppacher and Klett 1997).

The presence of sufficient ice nuclei is essential for ice crystal development to commence (Schemenauer et al. 1981). In the absence of atmospheric ice nuclei particles, supercooled liquid water may not freeze spontaneously and form ice crystals until -40 ˚C. However, terrestrial particles such as clay (kaolinite in particular) and even extraterrestrial particles such as meteorite dust and fragments provide a surface through which supercooled liquid water vapor may become ice via deposition and/or direct freezing (Byers 1965b, Droessler 1965, Ahrens 1991). Ice nuclei appear to be most numerous over land and at lower elevations, indicating the continental origin of the clay particles in particular (Byers 1965b, Rogers and Yau 1989). Pure ice nuclei, or those formed through the freezing of pure water at very low temperatures, may also lead to the development of ice crystals. Due to the extremely low temperatures (approximately -40 ˚C) needed for their development, however, their presence is generally limited to very cold air, typically found in the middle and upper troposphere (Ahrens 1991).

Ice crystal growth is also strongly influenced by the degree of supersaturation of water compared to ice, which is directly related to the cloud temperature. The saturation vapor pressure over water is greater than the saturation vapor pressure over ice. At 0 ˚C, this difference is almost negligible, whereas at colder temperatures the difference in the saturation vapor pressure over water and the saturation vapor pressure over ice becomes greater and increases steadily. At temperatures close to and just below 0 ˚C, clouds consist primarily of supercooled liquid water droplets and any ice is of extremely small quantities. At around -15 ˚C, significant quantities of both supercooled liquid water and ice crystals occur, with a higher saturation vapor pressure immediately surrounding the supercooled liquid water droplets and lower saturation vapor pressure surrounding the ice crystal. The lower saturation vapor pressure over the ice crystals creates a vapor pressure gradient, drawing water vapor away from the supercooled liquid water droplets and resulting in ice crystal growth via deposition. This vapor pressure gradient is evident as long as supercooled liquid water droplets exist. However, as temperatures drop below -20 ˚C, the concentration of supercooled liquid water diminishes rapidly, and hence the saturation vapor pressure over ice predominates, leading to slower ice crystal growth rates (Rogers and Yau 1989, Ahrens 1991, Pruppacher and Klett 1999). Therefore, as numerous field and laboratory studies indicate, ice crystal growth is most rapid between approximately -14 ˚C and -17 ˚C (Ryan et al. 1976, Auer and White 1982, Ahrens 1991, Pruppacher and Klett 1997, Fukuta and Takahashi 1999), suggesting this temperature range is ideal for heavy snowfall development.

Ice crystal structure and size is also strongly dependent on temperature. At 0 ˚C to -4 ˚C, for example, plates are most common, whereas from -4 ˚C to -10 ˚C prisms and needles predominate. From -10 ˚C to -20 ˚C thick plates and dendrites occur most frequently, and below -20 ˚C hollow columns and sheaths dominate (Byers 1965a, Schemenauer et al. 1981, Pruppacher and Klett 1997). The dendritic ice crystal range coincides with the maximum ice crystal growth rate observed between -14 ˚C and -17 ˚C. Therefore, dendrites comprise the majority of ice crystals associated with heavy snowfall. Numerous studies have reported that a secondary growth maximum occurs between -5 ˚C and -7 ˚C in association with the lower density and longer in-cloud residence times of columnar crystals and needles. Rising motions enable the lower density ice crystals to stay lofted in the cloud for a longer period, allowing more time for ice crystal growth. Additionally, the lower density columnar crystals and needles have a low fall velocity which translates into a greater in-cloud residence time. The minimum growth occurs around -10 ˚C as the isometric ice crystals are relatively dense and therefore have a high fall velocity. This leads to a shorter in-cloud residence time and diminished ice crystal growth via deposition (Byers 1965a, Fukuta and Takahashi 1999).

The vertical extent or depth of favorable conditions for ice crystal growth within a cloud is also important, as deeper clouds typically translate into a longer residence time and enhanced ice crystal growth via water vapor diffusion (Byers 1965a). Additionally, heavy snowfall tends to occur when the favorable conditions for ice crystal growth (cloud temperatures of -14 ˚C to -17 ˚C and supersaturation of water relative to ice) coincide with the area of maximum vertical ascent. Rising air motions are often maximized in a zone referred to as the level of maximum lifting, with zones of convergence below and divergence above. Approximately 70 percent of a sample of 75 synoptic-scale winter cyclones in the United States during 1980-1981 exhibited a level of maximum lifting at 600 hPa with corresponding temperatures between -13 ˚C and -15 ˚C (Auer and White 1982). For orographic snowfall in the western U.S., Auer and White (1982) found that the maximum vertical velocities and level of maximum lifting appeared to occur just at or slightly above mountaintop level. Thus, the level of maximum lifting for pure orographic snowfall in the Southern Appalachian Mountains may range from 1,500 m to 2,500 m, or approximately 850 to 750 mb, as the mountaintops range from 1,200 to 2,000 m. Heavy NWFS orographic snowfalls in the Southern Appalachians may also be characterized by the level of maximum lifting (850 to 750 mb) coinciding with maximum ice crystal growth temperatures (-14 ˚C to -17 ˚C).

Other factors important for heavy snowfall are a continuous supply of both water vapor and ice nuclei. Both of these fluxes should be high enough to offset losses through precipitation out of the cloud. The moisture flux is especially important, as abundant ice nuclei are typically present at colder temperatures (Schemenaur et al. 1981). The ratio of recorded precipitation versus total water vapor available to the system, or precipitation efficiency, also greatly influences the intensity and duration of snowfall events. Precipitation efficiency is generally quite high with heavy snowfall (approaching 80%), as synoptic-scale snowstorms are very efficient in delivering the condensed and sublimated water vapor to the ground as snowfall (Schemenaur et al. 1981, Auer and White 1982). Relative humidity values are typically quite high in association with the synoptic-scale environment of major winter storms, leading to very little evaporation or sublimation of falling snowflakes. When significant amounts of dry air are in place in the lower troposphere, as in situations of cold air wedging along the eastern slopes of the Appalachian Mountains, substantial loss of water vapor through melting/evaporation and sublimation may occur, reducing the precipitation efficiency. Precipitation efficiency of lake-effect snowfall and northwest flow orographic snowfall in the eastern United States may also be reduced at times, particularly at lower elevations, due to entrainment of dry air within banded snow squalls and substantial loss via melting/evaporation and sublimation of falling snowflakes.

The cloud microphysical environment in the lower troposphere can also greatly enhance ice crystal size and hence snowfall amounts. The presence of a saturated layer with temperatures below 0 ˚C within which supercooled liquid water droplets abound leads to rapid growth of seeder ice crystals falling from the middle troposphere via riming. For example, inversions associated with stable layers of cold air and high supersaturation of water relative to ice in the lower troposphere are known to be feeder regions of rapid snow crystal growth (Power et al. 1964, Wesley and Pielke 1990). This seeder-feeder process has also been found to significantly enhance precipitation in situations of orographic lifting, especially in the case of snowfall. The dendritic crystals, in particular, are very efficient scavengers of supercooled liquid water from the low-level feeder clouds produced by orographic ascent of moist air (Choularton and Perry 1986, Dore et al. 1990, Dore et al. 1992, Borys et al. 2000). Orographic snowfall and orographic enhancement of snowfall will be discussed in greater detail in section 1.2.3 on Orographic and Mesoscale Processes and Patterns.

1.2.1b. Lifting mechanisms

Snowfall is produced when sustained lifting of air in the lower and middle troposphere leads to condensation/deposition and ice crystal growth in the cloud layer. Large scale condensation occurs when sizeable masses of air are cooled to their dewpoint. Air can be cooled to its dewpoint through a variety of processes, including radiational cooling in the nighttime atmosphere, advection of moist air across a cold surface, or forced ascent and adiabatic cooling. Rising air motions resulting in adiabatic cooling, in reality, are the only processes which can generate sufficient sustained condensation through a thick enough atmospheric layer to realize anything more than snow flurries or drizzle. Surface convergence, frontal lifting, convectional lifting, and orographic lifting are generally recognized as the major lifting mechanisms associated with precipitation formation and are discussed in greater detail below (Ahrens 1991).

Surface convergence, or its corollary, upper level divergence, encourages dynamic lifting of air and is therefore an important lifting mechanism. As surface or lower tropospheric winds converge, air molecules pile up on one another and are forced upward. Conversely, as winds diverge in the middle and upper troposphere, lower tropospheric air is forced upward, resulting in adiabatic cooling, condensation, and if other conditions are favorable, precipitation development. These two processes often work together in a well developed synoptic-scale cyclone, with rising air motions dynamically forced by both surface convergence in association with a surface low and upper level divergence in conjunction with an exiting jet stream. Jet streaks, or areas of stronger wind speeds embedded within a larger scale jet stream, also provide an important mechanism for upper level divergence. These are typically located immediately downstream of the 500 hPa trough axis, where both curvature and shear vorticity become more positive (i.e. cyclonic). The acceleration of wind speeds creates an area of upper level divergence and a localized vorticity maximum that can significantly enhance snowfall (Uccelini and Kocin 1987, Harman 1991).

Frontal lifting of air occurs when air masses with different thermal properties meet. Since warm air is less dense and more buoyant than cold air, the two air masses do not readily mix. Strong lift can accompany a cold front, leading to a moderate to intense band of precipitation just preceding or immediately accompanying the front. Lifting associated with lower tropospheric warm advection is not as dramatic, and usually follows surfaces of equal potential temperature referred to as isentropes. Isentropic lifting can be an important lifting mechanism for prolonged snowfall that can result in heavy accumulations (Kocin and

Uccelini 1990). In fact, one of the heaviest snowfalls ever recorded in Greer, SC, was in association with a prolonged period of isentropic lift (Moyer 2001).

Convectional lifting results largely from surface heating, although cooling of the middle and upper troposphere can also contribute to its development. Insolation from a late winter or early spring sun is an important source of convective lifting leading to snowfall. Solar energy rapidly warms the ground and the boundary layer warms rather quickly as heat is transferred vertically via conduction, mechanical mixing, and convection. Ice-free or otherwise relatively warm bodies of water can also play important roles in heating air at the surface, as in the case of lake-effect snowfall to the lee of the Great Lakes. Convectional lifting is aided by high levels of instability resulting from steep lapse rates in the lower and/or middle troposphere. In an absolutely unstable atmosphere, once air particles are given a nudge upward (or downward), they will continue to rise (descend) until the equilibrium level is reached. Assuming sufficient water vapor is present and cloud temperatures are cold enough, localized heavy snowfall can result, accompanied by thunder and lightning. In fact, thundersnow occasionally develops on the leeward shores of the Great Lakes, where convectional lifting helps to generate intense lake-effect snow squalls (Market et al. 2002).

Orographic lifting occurs anytime a topographic barrier forces air to rise. When the air is sufficiently moist and the topographic relief is great enough, orographic lifting can lead to condensation of water vapor and precipitation formation. Orographic snowfall, therefore, is a common occurrence in many mountain regions of the world, including in the western U.S. and even in the Appalachian Mountains of the eastern U.S. However, orographic lifting is typically not solely responsible for precipitation that falls in mountain regions (Critchfield 1983). It often occurs in conjunction with synoptic-scale lifting, and hence it is perhaps more appropriate to refer to it as an orographic enhancement of snowfall. Regardless of whether the snowfall is exclusively orographic or not, orographic lifting can provide a trigger for snowfall development, in addition to providing a layer of supersaturated liquid water droplets near the surface. Falling snowflakes generated by seeder clouds higher in the atmosphere fall through this orographically enhanced supersaturated layer and grow rapidly through riming. Even small hills can lead to significant enhancement of snowfall through the seeder-feeder process (Dore et al. 1992). Orographic snowfall and orographic enhancement of snowfall will receive greater attention in subsequent sections.

1.2.1c. Vertical profiles of temperature and moisture

Adiabatic and psuedoadiabatic effects play very important roles in determining the vertical profiles of temperature and moisture associated with snowfall, as they do with other types of precipitation. An adiabatic process refers to cooling or warming without the addition/loss of heat energy. Rising or sinking of unsaturated air parcels, therefore, produces adiabatic cooling (warming) as air moves into a lower (higher) pressure environment and expands (contracts). Once a rising air parcel reaches saturation, however, the process becomes pseudo-adiabatic, because further rising and cooling results in condensation and the release of latent heat energy, which partially offsets the rate of cooling (Ahrens 1991). Rising air motions, therefore, will produce cooling and lead to a cooler atmosphere as elevation increases. Lapse rates are greatest in unsaturated air and lowest in saturated air at warmer temperatures, due to the release of abundant latent heat due to higher water vapor content at warmer temperatures. This suggests that when dry air is present in the Southern Appalachians, the temperature differences produced solely by adiabatic effects between the valleys and mountaintops can be quite dramatic, sometimes on the order of 15 ˚C. In fact, this occurs frequently in conjunction with low-level northwest flow, as orographic lifting condenses out most of the available water vapor on the windward slopes, leaving dry air to warm at the dry adiabatic lapse rate as the air parcels descend the leeward (southeastern) slopes of the foothills.

Diabatic (or non-adiabatic) effects refer to the change of temperature with the addition or loss of heat energy. In wintertime situations, the evaporation of rain or the sublimation of snow takes heat out of the atmosphere (i.e. converts sensible energy to latent energy), yielding a pronounced cooling effect. Surface temperatures may be well above freezing before the onset of precipitation, but if dewpoint temperatures, and especially wet-bulb temperatures, are below freezing, significant cooling as a result of evaporation and/or sublimation may result, changing the rain to snow (or sleet or freezing rain) at the surface. Additionally, the melting of falling snow also produces a diabatic effect by taking heat out of the atmosphere and producing cooling. In fact, it is interesting to note that 0 ˚C appears most often on vertical soundings in precipitation settings due to the diabatic cooling that results as snow melts (Djuric 1994). The melting of snow can also play an important, but often overlooked, role in cooling the surface if the precipitation rate is moderate to heavy. In fact, diabatic cooling due to melting led to a surprise early October snowstorm that dumped up to two feet of snow in eastern New York, including 15 cm (6 in) in Albany. As much as two-thirds of the observed cooling in the lower troposphere during this event was associated with melting (LaPenta 1989). Diabatic effects also played important roles in lowering the snow levels in the major late season Southern Appalachian snowstorms of May 1992 (Sabones and Keeter 1989) and most likely in April 1987.

Ice crystal growth and subsequent snowfall development can occur in a variety of synoptic and mesoscale environments. However, as mentioned previously, heavy snowfall usually results when a deep moist layer with temperatures between -14 ˚C to -17 ˚C coincides with the zone of strongest rising motions. In synoptic-scale cyclones of the central and eastern United States, this level of maximum lifting is typically found between 550 and 600 hPa (Auer and White 1982). In addition, a cold wedge is often found at the surface, ensuring that surface temperatures are sufficiently cold for snow to accumulate. A supersaturated layer of water with respect to ice often occurs at the top of the cold wedge, generally around 900 mb, leading to enhanced ice crystal growth via riming. Warm advection predominates in the layer above the wedge, creating a warm nose aloft and resulting in enhanced lift along isentropic (equal potential temperature) surfaces (Bell and Bosart 1988). Although warm advection is desirable to enhance isentropic lift in the layer above the wedge, if temperatures in the warm nose rise above 0 ˚C, snowflakes may melt partially or completely, reaching the surface as sleet, freezing rain, or a mixture thereof. In fact, strong warm advection in the lower troposphere is more likely to result in mixed precipitation, whereas weak to moderate warm advection is often associated with heavy snow (Konrad and Perry 2004). In many cases diabatic cooling due to melting of snow can reduce or negate the advective warming, providing an isothermal layer at 0º C and creating a significant forecast challenge.

Vertical profiles of temperature and moisture in association with easterly upslope snows in the Front Range of the Rocky Mountains show interesting patterns that bear some resemblance to the synoptic-scale cyclone environment discussed above. One synoptic pattern in this region features a cold wedge below 800 hPa extending along the eastern slopes in association with weak northwest flow (layer 1), a strong easterly flow with deep moisture from 800 hPa to 570 m (layer 2), and strong southerly flow with moderate amounts of moisture above 570 hPa (layer 3). Ice crystals form as a result of isentropic lift in the feeder region of layer 3, fall through a moisture laden layer 2, where temperatures are in the ideal range for dendritic crystal growth (-12 ˚C to -16 ˚C), and then fall through a supersaturated layer at the top of the low-level inversion where rapid growth through riming occurs (Wesley and Pielke 1990). This particular situation further illustrates how heavy snowfall results when 1) the level of maximum lifting coincides with the maximum ice crystal growth rate and 2) a supersaturated layer at the top of a low-level cold wedge leads to enhanced ice crystal growth via riming.

Orographic snowfall and orographically enhanced snowfall in the mountains of the western U.S., Japan, and Scotland also show the importance of a supersaturated feeder layer close to the surface where snowflakes grow rapidly by riming (Choularton and Perry 1986, Borys et al. 2000, Kusunoki et al. 2004). However, this supersaturated layer on the western slopes of these mountains is produced primarily by orographic lifting of moist air, whereas in the other situations the supersaturated layer is in conjunction with cold wedging and a low-level inversion. Heavy northwest flow orographic snowfalls in the Southern Appalachians often occur in a conditionally unstable lower troposphere with high values of relative humidity in the lower and middle troposphere. Since the level of maximum lifting here and elsewhere is often found near or just above the mountaintop level (Auer and White 1982)—ranging between 850 to 650 hPa depending on the mountain range, temperatures must be colder for the maximum growth rate of ice crystals to occur at a lower constant height. In other words, for the level of maximum lifting to coincide with the dendritic growth range (-14 ˚C to -17 ˚C), temperatures at or just above the mountaintop level must be colder than in situations of heavy synoptic-scale snowfall, where the level of maximum lifting is found higher in the troposphere. This suggests that surface temperatures may also be considerably colder with heavy northwest flow orographic snowfall than with heavy snowfall resulting from synoptic-scale lift.

Lake-effect snows in the central and eastern U.S. often exhibit an even lower level of maximum lifting, with a capping inversion occurring above cold arctic air (Lunstedt 1993, Niziol et al. 1995, Lackmann 2001). Convectional or convergent lifting of air is confined to a much thinner section of the troposphere due to the low capping inversion, and even colder temperatures throughout the atmospheric column are necessary for the level of maximum lifting to coincide with the maximum ice crystal growth rate. Lapse rates are generally very steep in lake-effect situations and in many cases the lower troposphere is absolutely unstable. In fact, lake-effect snow is probable when the temperature difference between the lake surface and that at 850 hPa is 13 ˚C or greater in the absence of synoptic-scale forcing (10 ˚C with synoptic-scale forcing), according to the Buffalo National Weather Service Forecast Office (Niziol et al. 1995). Synoptic-scale forcing, such as cyclonic vorticity advection downstream of a 500 hPa vorticity maximum, apparently aids the convectional and convergent lift already present, allowing lake-effect snow to develop in a more stable lower troposphere.

1.2.2. Synoptic Patterns of Snowfall

Snowfall in the eastern U.S. and across the Southern Appalachian Mountains often occurs in connection with several distinctive synoptic circulation patterns. These patterns are tied to increased cyclogenesis in the Gulf of Mexico and along the East Coast of the U.S. This section focuses on the synoptic patterns of snowfall by first discussing winter cyclone tracks in the eastern U.S., then identifying favorable synoptic patterns for snowfall in the Southern Appalachians, and finally by making some observations on the inter-annual variability of snowfall.

1.2.2a. Winter cyclones in the eastern U.S.

Cyclogenesis rates during the late fall through early spring display maxima to the lee of the Canadian Rockies in Alberta, to the lee of the U.S. Rocky Mountains in Colorado, in the Texas/Gulf Coast area, and along the eastern coast of the U.S. (Reitan 1974, Zishka and Smith 1980, Whittaker and Horn 1982, Nielsen and Dole 1992). The Alberta and Colorado cyclogenesis maxima are associated with the short-wave troughing and cyclonic vorticity as air parcels are forced downward on the lee of the Rocky Mountains. The cyclogenesis maxima along the Texas/Gulf Coast and the eastern coast of the U.S., however, primarily result from the thermal contrasts between the cold land and warm water, which produce a zone of enhanced baroclinicity (Harman 1991). Winter cyclones that develop in these areas often move along favored storm tracks conducive to snow in the eastern U.S. and Southern Appalachians. These cyclogenesis maxima are tied to four preferred cyclone tracks, which are discussed in greater detail below.

Despite the obvious differences in cyclone tracks, available moisture, and temperature, it is possible to make some generalizations about snowfall-producing cyclones in the eastern U.S. For example, the mean distance between the track of the vorticity max and the location of heaviest snowfall is approximately 282 km (175 miles) to the left. Similarly, the favored location for heavy snow on average occurs approximately the same distance to the left of the surface cyclone track (Goree and Younkin 1966). This preferred location is far enough away from the center of the cyclone and strong warm advection, yet close enough to the area of maximum ascent associated with the vorticity maximum and isentropic lift. This balance between lower tropospheric warm advection and near 0 ˚C temperatures is critical in determining where heavy snow occurs (Browne and Younkin 1970). Strong warm advection provides enhanced lift along an isentropic surface, but if temperatures in the lower troposphere (850 hPa for example) become too warm (i.e. too much warm advection), precipitation at the surface will mix with or change to sleet, freezing rain, and/or rain. For a surface cyclone tracking northeastward along the East Coast, therefore, the preferred location for heavy snow is approximately 282 km (175 miles) to the northwest (or inland) of the center's track, which is the approximate distance from the southeastern U.S. coast to the Blue Ridge of the Southern Appalachians.

Alberta clippers typically track southeast from Alberta into south-central Canada and north-central U.S., and then eastward into the northeastern U.S. Light to moderate snow often accompanies these cyclones to the north of their track, with lighter precipitation amounts to the south (Hutchinson 1995). As their name suggests, Alberta clippers usually move quickly, which keeps snowfall totals rather light. Since these systems often occur in conjunction with cold continental air masses, moisture is certainly at a premium, which may also reduce snowfall totals. Additionally, the cloud temperatures at the level of maximum lifting may be considerably below the temperature range (-14 ˚C to -17 ˚C) of maximum ice crystal growth rate, yielding lower snowfall totals. When a deep long wave trough persists across the eastern U.S., Alberta clippers will often track much farther to the south than is usually the case, bringing snows to the Southern Appalachians and, in some instances, intensifying off the mid-Atlantic coast (Kocin and Uccelini 1990). The northwestern slopes of the southern and central Appalachian Mountains may receive significant snowfalls in this scenario, due to orographic enhancement. To the lee of the Southern Appalachians in the North Carolina and Virginia foothills, however, little if any snow accumulations occur with Alberta Clippers.

Colorado lows tend to track east or northeast from the southern Rockies and then continue northeast to the west of the Appalachian Mountains. In the late fall and early winter, Colorado lows can become very strong, indicative of their common name, “Witches of November,” in the Great Lakes (Zielinski 2002). Such a storm track is less favorable for snowfall across the eastern U.S. in general, with the exception of the Great Lakes and New England, as areas to the south and east of the storm track lie in the warm sector. Lower tropospheric warm advection to the south and east of the surface cyclone quickly changes any snow to sleet, freezing rain, and rain. Many Colorado lows redevelop off the New England coast and may provide a period of wrap-around and northwest flow on the back side, but generally accumulations remain light. Therefore, Colorado lows rarely bring appreciable snowfall to the Southern Appalachians or East Coast.

The Texas/Gulf Coast lows are important snow producers for the Deep South and, in fact, for much of the East Coast of the U.S. (Kocin and Uccelini 1990, Suckling 1991, Mote et al. 1997). These systems can also be prolific snow producers in the Southern

Appalachians, bringing abundant moisture on initial south and southeast flow, and then ending with northwest flow associated with wraparound upslope snowfall. One of the strongest winter storms to affect the eastern U.S., the Blizzard of ’93, was an example of a Texas/Gulf Coast low. Miller (1946) originally categorized this type of cyclogenesis and associated storm track “Type A”, which is why these systems are commonly referred to as Miller Type A cyclones. They are characterized by the development of a surface cyclone along a frontal boundary in the Gulf of Mexico separating cold continental air from warmer and more moist Gulf or Atlantic air. The surface low tracks northeastward out of the Gulf of Mexico, paralleling the Atlantic coastline in the southeastern U.S. Further development may occur off the North Carolina coast, in the vicinity of Cape Hatteras, and the cyclone may become a nor’easter as it moves off to the northeast. This storm track places interior locations of the southeastern U.S. and mid-Atlantic in the favored location for heavy snowfall.

Nor’easters are strong cyclones that undergo cyclogenesis off the mid-Atlantic coast generally north of Cape Hatteras, NC, and track northeast paralleling the coast. The name nor’easter results from the strong northeasterly winds that impact coastal areas to the north and northwest of the storm’s track as a result of the cyclonic circulation. Their impacts are greatest across the northeastern U.S., particularly the mid-Atlantic and northeastern coastal areas where heavy precipitation (rain and/or snow), high winds, and heavy surf can be more destructive than a hurricane (Davis et al. 1993, Zielinski 2002). Nor’easters, as mentioned previously, may form as a result of further intensification of a Texas/Gulf Coast low (Miller Type A). Nor’easters may also develop from Texas/Gulf Coast lows that track west of the Appalachians, Colorado lows, or even Alberta clippers. In these situations, also known as Miller Type B, a secondary area of low pressure develops and rapidly intensifies off the southeast or mid-Atlantic coast and moves northeastward along the coast. Frontogenetical forcing in conjunction with a low-level wedge of cold air to the east of the Appalachians plays a major role in this secondary development (Kocin and Uccelini 1990). Of the 20 major nor'easters Kocin and Uccelini (1990) examined between 1955 and 1985, ten are classified as Miller Type A and ten as Miller Type B, indicating that both types of storm tracks and patterns of cyclogenesis can lead to major nor'easters.

1.2.2b. Favorable synoptic patterns for snowfall in the Southern Appalachians

In the Southern Appalachians, Texas/Gulf Coast lows are one of the most favorable storm tracks for heavy snowfall (Gurka et al. 1995). In particular, Miller Type A cyclones that originate in the Gulf of Mexico and move northeastward through the central and eastern Carolinas place much of the Southern Appalachians in the zone of preferred heavy snow (Knappenberger and Michaels 1993, Goree and Younkin 1966). As the surface cyclone approaches, temperatures in the lower troposphere are often close to or just below 0 ˚C with strong isentropic lift associated with warm advection leading to the development of heavy snowfall. As the low tracks through the Carolinas, further enhancement of snowfall can result from upper level divergence associated with jet streaks and cyclonic vorticity advection. Cloud temperatures in the level of maximum lifting are most likely in the range of rapid ice crystal growth, as suggested by Auer and White (1982). The southwest to northeast orientation of the Southern Appalachians also aids in orographic enhancement of snowfall totals on the southeastern slopes in association with southeasterly and southerly low and mid-level flow ahead of the surface cyclone. Likewise, wrap-around northwest flow on the back side is also enhanced by orographic effects. Consequently, some of the most memorable snowstorms in the Southern Appalachians, including the Blizzard of ‘93, were Miller Type A cyclones that originally developed in the Gulf of Mexico (Gurka et al. 1995).

Texas/Gulf Coast lows of the Miller Type B variety can also lead to heavy snow along the southeastern slopes of the Southern Appalachians, as well as significant accumulations of sleet and freezing rain (e.g. Gay and Davis 1993). In this synoptic pattern, the surface cyclone moves north or northeastward out of the Gulf of Mexico but tracks west of the Appalachians into the Ohio River Valley, before secondary cyclogenesis occurs off the mid-Atlantic Coast (Kocin and Uccelini 1990, Gurka et al. 1995). A strong anticyclone over the northeastern U.S. produces cold air wedging in the lower troposphere in the Southern Appalachians, while strong warm advection occurs above the wedge in a warm nose. The strong anticyclone often continues to support the cold wedge through cold advection at the surface, even as warm advection above the wedge in the warm nose strengthens. In this scenario, precipitation associated with isentropic lift often begins as snow over much of the region, but then changes to mixed precipitation or all rain as the warm advection overwhelms the cold air aloft and diabatic cooling effects associated with melting snow. An exception to this general pattern occurs along the southeastern slopes and adjacent Piedmont where the topography helps to anchor the cold air in place and serve as a hedge to the warm advection, allowing frozen precipitation to persist considerably longer than in areas farther to the southwest, west, and even northwest. Isentropic lift is also enhanced by the locally steeper surface of the cold wedge due to upsloping over the Blue Ridge escarpment (Bell and Bosart 1988), producing enhanced snowfall in this region. It is also likely that significant ice crystal growth occurs as a result of riming in the supersaturated layer at the top of the cold wedge and near the mountaintops (Power et al. 1964, Choularton and Perry 1986, Wesley and Pielke

1990). The resulting enhancement of precipitation may also lead to locally stronger diabatic cooling, further counteracting the ongoing warm advection and allowing heavier snow to persist longer.

Late season 500-hPa cutoff lows can bring substantial snowfall totals to higher elevations and occasionally even to valley locations. Therefore, they are a third important synoptic pattern conducive to heavy snowfall in the Southern Appalachians. Deep 500-hPa troughs often become “cutoff” from the main 500-hPa flow in the spring months, leading to the development of 500-hPa cutoff lows (Sabones and Keeter 1989, Gurka et al. 1995). At the center of the 500-hPa low, 850-hPa temperatures are below 0º C, enhancing the probability that any precipitation will fall as snow. Spokes of vorticity rotating around the system and low-level upslope flow provide sources of lift, whereas the quasi-stationary or slow movement of these systems allows the snow to persist over the region for days at a time. In some situations, the cold air associated with the 500-hPa lows actually moves into the region from the south as the cold pool migrates slowly north or northeastward, as was the case with the April 12, 1988, event that produced up to 56 cm (18 in) of snowfall in the North Carolina Mountains (Sabones and Keeter 1989). In this situation, a deep 500-hPa trough became cutoff in the lower Mississippi Valley and drifted slowly northeastward. Due to the heavy precipitation and strong lift that often occurs with these systems, the dynamic cooling and diabatic cooling (i.e. tied to melting snow) no doubt contribute to lowering snow levels to the valley bottoms. Two memorable recent events that brought record and near-record snowfall totals to the Southern Appalachians include April 2-5, 1987, when 152 cm (60 in) of snow fell at Newfound Gap on the NC/TN border (single storm record for NC) in the Great Smoky Mountains, and May 6-8, 1992, when 102 cm (40 in) and possibly as much as 152 cm (60 in) fell in the vicinity of Mt. Pisgah, North Carolina (Sabones and Keeter 1989, Fishel and Businger 1993). Diabatic cooling played a major role in lower snow levels in the May 1992 event, as this author remembers well the alternating periods of light rain and very heavy snow that accumulated to a depth of several inches in Waynesville, NC, and vicinity (792 m or 2600 ft). Prolonged low-level northwest flow on the back side of both of these systems contributed appreciably to the heavy snowfall totals as well.

Alberta clippers can be important snow producers in the winter months across the Southern Appalachians, although snowfall totals are typically much less than with Texas/Gulf Coast lows and 500-hPa cutoff lows. Moisture from both the Gulf of Mexico and the Atlantic Ocean essentially remains cut off with a strong west to northwest flow, meaning that the moisture content is significantly reduced. In addition, the systems are often fast moving. Conversely, surface temperatures are often sufficiently cold, allowing whatever snow that falls to accumulate, rather than melt. Snow densities are also much lower due to the colder temperatures throughout the atmospheric column and reduced melting and compaction. Alberta clippers can also provide a period of low-level northwest flow on the back side, bringing additional snowfall to the northwestern slopes via orographic lifting. Therefore, successive Alberta clippers in a prolonged period of northwest flow can produce a significant amount of snowfall, although the liquid water content may be much lower than with systems that have access to copious Gulf and Atlantic moisture.

Low-level northwest flow can produce light to moderate, and occasionally heavy, snowfall in portions of the Southern Appalachians. Northwest flow snowfall (NWFS) events are common occurrences on windward slopes and at higher elevations of the Southern Appalachian Mountains, but comparatively rare on lower elevation leeward slopes. NWFS often develops behind a strong cold front heralding in a deep 500-hPa trough over the eastern U.S. Abundant moisture is present in the lower troposphere to the lee of the Great Lakes, but a capping inversion predominates around 850 mb, characterized by subsidence above. Therefore, the available lift for NWFS is often confined to the lower troposphere below this capping inversion, leading to a much shallower layer of potential lift than observed with other synoptic patterns. Upslope flow and vorticity maxima rotating around the base of the long wave trough can, however, cause the height of the capping inversion to rise, providing a deeper layer through which lift can occur in the lower and middle troposphere (Niziol 1989, Lackmann 2001). On windward slopes in the Southern Appalachians, NWFS snowfall may account for between 25 and as much as 50 percent of the total annual snowfall (Schmidlin 1992, Perry and Konrad 2006). Moisture for many of these events is of Great Lakes origin (Schmidlin 1992), suggesting some commonality with synoptic patterns associated with lake-effect snowfall in snowbelt locations to the southeast of Lake Michigan and Lake Erie. Although most events are relatively light, prolonged periods of northwest flow associated with abundant low and mid-level moisture and some synoptic-scale forcing can produce heavy snowfall, as was the case December 18-20, 2003, when up to 76 cm (30 in) fell in the North Carolina Mountains.

Other types of synoptic patterns can certainly produce snowfall in the Southern Appalachians when the right ingredients come together, but these events occur much less frequently than those discussed above. Prolonged isentropic lift in the absence of an existing or deepening surface cyclone can occasionally bring heavy snowfall to portions of the region. In fact, the single storm snowfall record of 30 cm (12 in) in Greer, SC, occurred on January 7, 1988, in association with just such a pattern. In this case, a weak surface low did eventually develop in the Gulf of Mexico, but tracked well to the south across the Florida panhandle (Moyer 2001). Additional synoptic patterns that may give rise to heavy snow are certainly possible given the right ingredients, but it is difficult to comment further given their relative absence in the recent climatological record.

1.2.2c. Inter-annual variability of snowfall

Much inter-annual variability exists in the snowfall totals across the Southern Appalachians, which can be related to variations in hemispheric-scale circulation patterns. These circulation patterns directly influence temperature patterns and storm tracks across the eastern U.S. and Southern Appalachians, and therefore have an important influence on monthly and annual snowfall totals. Perhaps the greatest direct influence on temperature and snowfall patterns is the mid-level circulation pattern across the North American continent (Namias 1960). The Pacific-North American (PNA) teleconnection pattern is often used to characterize the middle-tropospheric circulation, with the positive phase characterized by enhanced troughing over the eastern U.S. The negative phase, on the other hand, is associated with a more zonal flow across North America (Konrad 1998). During PNA positive periods, long wave troughing prevails across the eastern U.S., producing extended periods of low temperatures (Konrad 1998) and an active Texas/Gulf Coast storm track (Namias 1960, Knappenberger and Michaels 1993). Additionally, precipitation totals and the number of precipitation events are generally below normal across the southeastern U.S., but precipitation event totals increase (Henderson and Robinson 1994), presumably due to stronger surface cyclones. Although precipitation totals are below normal, snowfall is actually above normal due to colder maximum temperatures on precipitation days (Serreze et al. 1998). Miller Type A cyclones along the Texas/Gulf Coast storm track also appear to be more numerous, particularly in the mid-Atlantic region. During PNA negative periods, zonal flow predominates, leading to mild temperatures and a more active storm track west of the Appalachians in the Ohio River Valley (Knappenberger and Michaels 1993). Precipitation totals may average above normal (Henderson and Robinson 1994), but snowfall is below normal as much of the eastern U.S. remains in the warm sector of the Ohio River Valley storm track (Knappenberger and Michaels 1993).

The North Atlantic Oscillation (NAO) is another teleconnection that influences the inter-annual variability of snowfall across the eastern U.S., especially in the Appalachian Mountains and New England (Hartley 1998, Hartley and Keables 1998, Hartley 1999). The NAO oscillates between positive and negative phases, with the positive phase characterized by above normal sea level pressure south of approximately 55° N latitude and below normal sea level pressure generally north of the Arctic Circle. The strong pressure gradient enhances the low-level westerly wind belt and also results in southerly flow across the eastern U.S., producing above normal temperatures and a shifting of Atlantic storm activity northeastward into Newfoundland (Hurrell et al. 2003). The negative phase is characterized by higher than normal sea level pressures across the Arctic—especially in the vicinity of Greenland and Iceland, and lower sea level pressures south of 55º N. The blocking high pressure in the North Atlantic allows a deep meridional 500-hPa flow to develop over eastern North America, producing frequent intrusions of Arctic air and an active storm track. As much as 50 percent of the annual variability of snowfall along the northwestern slopes of the Southern Appalachians is explained by the NAO (Hartley 1999). The negative phase is associated with above normal snowfall and below normal temperatures, whereas the positive phase is tied to below normal snowfall and above normal temperatures. Some intra-regional variation exists, however, as the relationship between the NAO phase and snowfall is much weaker on the eastern slopes of the Appalachians and farther south (Hartley 1999). These results suggest that NAO negative winters exhibit more frequent and stronger periods of NWFS in the region.

Although the PNA pattern and NAO are the most significant teleconnections influencing winter climate variability in the middle and high latitudes of the Northern Hemisphere (Hurrell et al. 2003), the El Niño-Southern Oscillation (ENSO) teleconnection also has an influence on winter precipitation patterns in the southeastern U.S. The ENSO warm-phase, or El Niño, is associated with positive sea surface temperature (SST) anomalies in the eastern tropical Pacific extending to western South America. The positive SST anomalies destabilize the lower troposphere, resulting in widespread convection and the release of latent heat. This perturbation in the tropical Pacific is manifested across the Gulf of Mexico through stronger westerly flow aloft, which translates into a more active subtropical jet stream and associated Texas/Gulf Coast storm track (Trenberth 1991), extending north along the East Coast of the U.S. (Hirsch et al. 2001). Therefore, above normal precipitation occurs across portions of the southeastern U.S. in association with the warm phase. However, during the ENSO warm-phase winters, a split flow in the 500-hPa pattern often occurs, leaving the necessary cold air well to the north and relegating the heavy snowfalls to the higher elevations of the Southern Appalachians (Smith and O’Brien 2001). In the ENSO warm-phase winter of 1997-98, for example, snowfall totaled 450 cm (177 in) at the coop station on Mt. Mitchell, NC (station elevation 1,902 m or 6,240 ft), almost double their average annual snowfall. Boone, NC (1,024 m or 3,360 ft), on the other hand, reported only

84 cm (33 in) of snowfall, considerably below the 114-cm (45-in) long term average annual snowfall. Precipitation, however, was well above normal, as 1,270 mm (50 in, and just shy of the yearly average) fell in Boone during the first four months of 1998 (NCDC 2002). Weak ENSO warm-phase (El Niño) winters, however, are not tied to relatively warm conditions, thus more snow may occur. Therefore, the strength of the particular event plays an important role as well. The ENSO cold-phase (La Niña), characterized by negative SST anomalies in the tropical Pacific extending to the west coast of South America, has a somewhat opposite effect in the Southern Appalachians, as temperatures are slightly above normal leading to more rain events than snow (Smith and O'Brien 2001).

1.2.3. Orographic and Mesoscale Processes and Patterns of Snowfall

Precipitation processes and patterns are quite complex in mountain regions due to the high variability of topographic parameters such as elevation, slope, and aspect over very short horizontal distances. Mountains also act as barriers to the movement of air, allowing lower tropospheric cold air to stay wedged in place during certain synoptic patterns, providing a favored location for frontogenesis. Topography may also lead to localized areas of enhanced lift downwind due to wave motion and/or convergence. The relatively warm waters of unfrozen lakes in the middle latitudes can also contribute to enhanced snowfall at the mesoscale, particularly in elevated terrain in close proximity to the leeward shores. This is especially the case immediately downwind of the Great Lakes, where prolific snowfall totals can occur. This section focuses on snowfall produced or enhanced by mesoscale topographic influences and lake-effect snow. In situations of NWFS in the Southern Appalachians, the Great Lakes are often upwind, suggesting that lake-effect snow processes may combine with orographic effects to produce or enhance snowfall. This Great Lakes connection is especially apparent in northern sections, hence the importance of discussing the lake-effect snow regime in the context of NWFS in the Southern Appalachians.

1.2.3a. Orographic snowfall

Mountains provide a lifting mechanism for snowfall development and/or enhancement when moist low-level winds are forced upward by the topography. Through this orographic lifting, air rises and cools adiabatically, leading to large-scale condensation and supersaturation of liquid water versus ice. This process is generally more common along a continuous mountain range rather than an isolated peak, where air may move around the mountain more easily than up the windward slopes. Furthermore, orographic lifting is more effective at generating snowfall over a larger area when the orientation of the mountain range is roughly perpendicular to the low-level wind direction. Since the Appalachians trend southwest to northeast, a northwest wind is very close to perpendicular to the large-scale topographic orientation and maximizes the topographic influence. A southeast wind also provides ideal conditions for orographic lift over a large area, and under the right conditions, southeast flow can produce very large snowfalls due to the upwind proximity of the eastern Gulf of Mexico and the Atlantic Ocean. However, southeast flow snowfall is much less frequent than NWFS as periods of southeast flow often occur in a lower troposphere that is too warm for snow.

Snowfall may either be enhanced or produced exclusively via orographic lifting. Orographic enhancement of snowfall is more common and occurs anytime snowfall produced by synoptic-scale lift affects mountain regions. The synoptic-scale ascent produces large-scale condensation and deposition in the middle troposphere; ice crystals grow primarily through vapor diffusion in the layer of maximum lifting. These seeder ice crystals fall through the cloud and may continue growing via vapor diffusion, riming, and/or aggregation. Along windward slopes, and particularly at higher elevations, low-level orographic lift produces condensation and a supersaturated environment (often in the form of a cap cloud) leading to rapid enhancement of the falling seeder ice crystals through riming. The seeder-feeder effect contributes to the significantly greater precipitation that results over higher terrain; it also serves to deplete the mountaintop clouds of supercooled liquid water and ice and hence reduce the amount of low-level moisture at locations downwind. At lower elevations on leeward slopes, therefore, snowfall may quite often still occur, although what snow makes it to the ground may be very light as this air will have a lower relative humidity and promote greater sublimation/evaporation (Barry 1992, Dore et al. 1992, Whiteman 2000).

Orographic snowfall results when the topography is the primary or sole source of lift in the lower troposphere, leading to snowfall in mountain locations only. This type of snowfall often occurs in a conditionally unstable atmosphere in the Southern Appalachians. The forced ascent eventually provides just enough sensible heating through the release of latent heat above the lifting condensation level for air parcels to become unstable and continue rising on their own until the equilibrium level is reached, at which point the air parcels are no longer buoyant. In this situation, the topography is forcing the lift *and* acting to enhance the snowfall; the seeder-feeder effect produces rapid growth of ice crystals at the mountaintop level and actively scavenges moisture from the mountaintop cap clouds, leading to drier low-level air and the absence of lift on leeward slopes. Much of the NWFS of the

Southern Appalachians occurs in the absence of synoptic-scale lifting mechanisms, and therefore the necessary lift is provided solely by the terrain. Preliminary work indicates that approximately 75 percent of events are tied to synoptic scale subsidence. When synoptic-scale ascent does occur, it is often insufficient to generate snowfall on its own, but rather acts to moisten the middle troposphere and lift the capping inversion (e.g. Lackmann 2001), allowing orographically lifted parcels of air to rise to a much higher level than would otherwise be the case. In the rare events in which significant synoptic-scale lift occurs in conjunction with a prolonged period of low-level northwest flow, prolific snowfall totals result, as evidenced by the 152 cm (60 in) of reported snowfall at Newfound Gap, TN/NC, during the April 2-5, 1987 event (NWS 1987).

The spatial patterns of orographically enhanced snowfall and orographic snowfall are quite complex, with accumulations varying dramatically over very short distances. The general pattern, however, is for snowfall rates and accumulations to be maximized on windward slopes and minimized on leeward slopes. Higher elevation windward slopes generally receive more snowfall than lower elevation windward slopes, although on mountains with significant relief, such as Mt. Rainier in the Cascade Mountains, the snowfall maximum apparently occurs in the middle elevations (Reifsnyder 1980, Barry 1992). As air parcels continue to be forced upward, the available moisture diminishes due to ongoing condensation/deposition at lower elevations. Slope is also a factor; field measurements and simulated models of orographic enhancement of snowfall in the South Wales hills of the U.K. suggest that long hills (i.e. ones with gentler slopes) tend to promote ice crystal enhancement through vapor diffusion, whereas shorter hills (i.e. steeper slopes) tend to have much greater amounts of supersaturated liquid water leading to ice crystal enhancement via riming. The shorter hills are more susceptible to wind drift effects with higher wind speeds, which may displace the heaviest snowfall slightly downwind of the crest (Choularton and Perry 1986). Such an observation is consistent with precipitation data from Snake Mountain (1,701 m or 5,581 ft.), a knife-edged ridge in northwestern North Carolina, where mean annual precipitation actually decreased with elevation, presumably due to wind drift effects (Smallshaw 1953). The wind drift associated with snow would even be expected to be much greater than for rain, due to slower fall speeds as a result of lower density and larger surface area of falling snowflakes.

The seeder-feeder mechanism is characterized by greater precipitation efficiency of orographically enhanced snowfall compared with orographic rainfall. This apparently results from the larger surface area and slower fall speeds of snowflakes compared with rain drops, yielding significantly greater ice crystal growth via riming in the supersaturated environment that prevails near the mountaintops. Consequently, the shadowing produced as a result of orographically enhanced snowfall is much greater, leading to an even more significant decrease in precipitation amount and low-level moisture as air parcels descend the leeward slopes (Choularton and Perry 1986). In situations of pure orographic snowfall, therefore, descending air is more likely to warm at the dry adiabatic lapse rate for a longer period than in orographic rainfall, leading to an even greater range of temperatures between low elevation leeward slopes and high elevation windward slopes. In fact, pronounced temperature differences routinely occur between the higher elevation windward slopes and lower elevation leeward slopes in the periods of NWFS in the Southern Appalachians. Sensible heating also plays a role due to the mostly sunny or clear skies that often predominate along the leeward slopes, leading to maximum solar energy receipt.

1.2.3b. Mesoscale topographic influences

Numerous mesoscale topographic influences also interact with synoptic-scale patterns to enhance snowfall. Cold air wedging, terrain-induced convergence zones, lee waves, and differential heating are four influences discussed in greater detail below. Cold air wedging frequently occurs on the eastern slopes of the Appalachian Mountains in conjunction with a cold surface anticyclone over the northeastern U.S. The mountains act as a hedge against the cold dense air and encourage advancement to the south rather than the west. As a 500-hPa trough develops in the southern Great Plains or lower Mississippi River Valley, warm air is advected northeastward and isentropic lift results as this warm air is forced up and over the cold surface wedge. Precipitation results and diabatic cooling as a result of evaporation or sublimation of falling precipitation into a very dry airmass helps to lower surface temperatures considerably in some instances. Further cooling occurs when the cold surface air in the wedge is lifted orographically with northeasterly or easterly winds along the eastern and southeastern slopes of the Appalachian Mountains (Bell and Bosart 1988). Along the Blue Ridge Mountains of North Carolina and Virginia, the diabatic and adiabatic effects may cool surface temperatures well below freezing, while producing a supersaturated environment at the top of the cold wedge just above the mountaintops. The slope associated with the precipitation-producing isentropic surface is locally steeper in these areas due to the lifting of the entire cold wedge, leading to heavier precipitation and a higher probability that diabatic cooling due to melting of snow may negate or minimize the warm advection occurring above the cold wedge. Furthermore, heavy riming of snow likely occurs in the supersaturated layer at the top of the cold wedge, leading to significant enhancement of snowfall. This, in turn, may explain why snow may persist much longer and accumulate to much greater depths than in locations only 26 km (16 miles) to the west and northwest.

Localized convergence zones may also form as a result of topographic influences, particularly when a large mountain or mountain complex does not form a complete barrier to the prevailing wind. In addition to being forced up the windward slopes, air parcels may also be forced around the topographic barrier. On the lee side, the split currents of air may come back together, producing a localized convergence zone and an area of pronounced lift (Whiteman 2000). If sufficient low-level moisture is in place, terrain-induced convergence zones may produce snow showers in some locations while other areas in close proximity may enjoy sunshine or only partly cloudy skies. The Puget Sound Convergence Zone in western Washington is an example of a localized convergence zone that can develop under the right conditions, sometimes producing rapid lifting and thunder snow showers along the western slopes of the Cascade Mountains, with as much as 10 to 20 inches (25-50 cm) of snow (Renner 1992). Although it is doubtful that convergence zones of this magnitude exist in the Southern Appalachians, smaller-scale areas of enhanced lift downwind of major mountain complexes may help to intensify snowfall under certain circumstances. Most of the mountain ridges, however, are oriented southwest to northeast, providing limited opportunity for split flow to develop.

Channeling is also an example of terrain-induced convergence. Winds may be channeled or funneled up valleys and drainage basins when the low-level wind direction is nearly parallel to the orientation of the valley (Whiteman 2000). Enhanced lift may result in the upper reaches of the valley, and when coupled with existing orographic lift, may produce locally heavier snowfalls. Some valleys and drainage basins in the Great Smoky Mountains and Unaka Mountains of the Southern Appalachians have a northwest to southeast orientation, with the elevation increasing to the southeast. The Pigeon, French Broad, Toe/Nolichuky, Doe, and Watauga River basins are all oriented more or less in this fashion, suggesting that additional enhancement of snowfall may occur in association with NWFS when the wind direction is ideal with respect to the valley orientation.

Upslope convergence due to differential heating of mountaintops or higher elevation ridges can also lead to the enhancement or formation of snow showers (Whiteman 2000). Higher elevations of mountains are often warmer than the free atmosphere at the same altitude in the late spring through early fall especially, producing lower surface pressures and initiating rising air motions up the mountain slopes. In a cold, moist, and conditionally unstable or absolutely unstable lower troposphere, upslope convergence may lead to the development of scattered snow showers. This process, however, is more common in association with afternoon summertime shower and thunderstorm development in mountains.

Lee waves are a final type of mesoscale topographic influence that can enhance snowfall totals. Strong moisture-laden winds that intersect a mountain range within 30 degrees of perpendicular can often form pronounced banding of precipitation on the leeward side. This appears to be particularly true if a stable layer associated with a low-level inversion exists in the lower troposphere (below mountaintop level) and a conditionally unstable atmosphere exists above the low-level inversion. Such a case has been documented in the Tennessee Valley of the Southern Appalachians, as moist conditionally unstable air propagated northward and downwind from the Great Smoky Mountains resulting in mesoscale bands of heavy snow (Gaffin et al. 2003).

1.2.3c. Lake-effect snow

As continental polar or continental arctic air masses cross the relatively warm waters of the Great Lakes in the late fall and winter, the cold dry air mass warms through sensible heating and evaporates abundant water vapor from the lake surface. This addition of sensible and latent heat helps to destabilize the otherwise stable air mass and leads to the formation of mesoscale bands of snow—often quite heavy—that extend to the leeward shores and some distance downwind of the lake shore itself. The change from a smooth lake surface to a rougher surface with rolling hills and vegetation also leads to frictional convergence and enhanced lift at or close to the leeward shores. The snowfall is further enhanced by orographic lifting in elevated areas in close proximity to the leeward shores, such as the Tug Hill plateau on the east side of Lake Ontario, where more than 180 inches of snow falls in an average year. The heaviest lake-effect snowfalls generally occur in the early winter, when the air-water temperature difference is the greatest. Later in the winter season this difference diminishes as many of the Great Lakes develop significant ice cover, greatly reducing the sensible heat and moisture fluxes (Niziol 1989, Niziol et al. 1995).

Lake-effect snowfall typically develops in situations when the difference between lake surface temperature and the air temperature at 850 hPa is maximized. Significant lake-effect snowfall may occur when the lake-850 hPa temperature is greater than 13º C without synoptic-scale lift, or 10 ˚C with the extra lift. The height of the capping inversion associated with the continental polar or arctic air mass is another important controlling factor on the evolution of lake-effect snowfall. In most cases, the capping inversion is found at a rather low-level—often 700 hPa or lower, signifying that the vertical extent of snow-producing clouds and ice crystal growth are confined to the lower troposphere (Niziol et al. 1995).

Convection associated with sensible and latent heat fluxes can locally lift the base of the capping inversion, as can cyclonic vorticity advection tied to 500 hPa vorticity maxima that often rotate through the long wave troughs that dominate the synoptic patterns associated with lake-effect snowfall. These synoptic-scale weather features, in particular, can "pre-condition" the troposphere for extreme snowfalls, such as the 178 cm (70 in) of snow that fell over a two-day period on the Tug Hill plateau in New York State (Niziol 1989). Some evidence also suggests that heavy lake-effect snowfall is associated with a warm and saturated layer in the middle troposphere that is advected into the region from the Gulf Stream waters of the Atlantic Ocean (Liu and Moore 2002). Such conditions support heavy snowfall by contributing a thicker moist layer and allowing the moist layer to coincide with temperatures in the dendritic temperature range.

Lake-effect snow bands may affect locations considerably downwind from the lake shores, extending to the mountains of West Virginia and even as far east as Long Island, NY (Niziol 1995). A combination of lake-effect snowfall and orographic lift of Great Lakes moisture may contribute as much as 25-30 percent to average annual snowfall totals in the northern mountains of West Virginia. In addition, snowfall at Snowshoe, WV, during periods of north-northwest or northwest flow is highly correlated with snowfall in the Lake Erie snowbelt locations, further suggesting downwind effects in elevated terrain nearly 500 km (310 miles) from the leeward shores (Schmidlin 1992). This downwind influence is clearly of Great Lakes influence, as continuous cloud bands can often be seen in satellite imagery well to the south along the western slopes of the Appalachian Mountains. The key variable, however, in producing snowfall well downwind of the Great Lakes is elevation and exposure, which together maximize the available low-level moisture via orographic lifting. The snowfall that extends well downwind of Lake Erie, in particular, to the northern mountains of West Virginia, is more than likely not pure lake-effect snow, but rather orographic snowfall whose low-level moisture source is quite often, if not always, the Great Lakes.

Some NWFS events in the Southern Appalachians are likely characterized by some moisture influx from the Great Lakes under favorable air trajectories, particularly the western Great Lakes of Superior and Michigan. Although low-level moisture from the eastern lakes of Huron and Erie may play an important role in NWFS farther north in the northern mountains of West Virginia, only northerly flow could conceivably transport low-level moisture farther southwest to the Southern Appalachians. Northerly flow is rare and is far from perpendicular to the southwest to northeast orientation of the mountains. Just how important this low-level moisture is remains to be seen, however, as previous air trajectory analyses for selected NWFS events in the Southern Appalachians have proved inconclusive (Lee 2002). Significant NWFS in the Southern Appalachians more than likely occurs in an atmospheric column that has been pre-conditioned by synoptic-scale lift, raising the height of the capping inversion and moistening the middle troposphere.

1.2.4. Memorable Snowy Periods in the Southern Appalachians

The Southern Appalachians have experienced some major snowstorms and prolonged snowy periods during the past fifty years. The April 1987, May 1992, and March 1993 events certainly stand out in terms of snowfall totals (approximately 152 cm or 60 in at the highest elevations), and the single storm impacts of the Superstorm of ’93 are nearly unprecedented in the region. However, this section focuses on the prolonged snowy period of February/March 1960 and the December 2003 snowstorm to further illustrate the synoptic and spatial patterns of snowfall across the region. Snowfall during the February/March 1960 period occurred in association with an active storm track and alternating periods of NWFS. The December 2003 event represents a prolonged period of NWFS that produced snowfall totals in excess of 61 cm (24 in) in the absence of a well developed surface cyclone and strong synoptic-scale ascent.

1.2.4a. February/March 1960

February 13, 1960, began an historic period of snow and cold across the Southern Appalachians that persisted for five weeks through the end of the March. A synoptic-scale cyclone tracked just south and east, placing most of the region in the favored area for heavy snowfall and dumping up to 70 cm (28 in) of snow, including 48 cm (19 in) in Knoxville, TN. A significant period of wrap-around NWFS followed, building the snowpack at higher elevations and on northwestern slopes. By March 26, 211 cm (83 in) of snowfall had accumulated in Boone, North Carolina, with many mountain locations in northwestern North Carolina receiving in excess of 152 cm (60 in) (Fig. 1.2). Continuous snow cover persisted through the end of March across much of region, extending into early April across the highest elevations. The maximum snow depth reported on the ground was 144 cm (44 in) on March 13-14 in Boone (NCDC 2002). Snow drifts evidently reached extreme depths, with some reports from Flat Springs, North Carolina, indicating that the drifts buried mature apple trees and persisted well into May and June in some hollows (Hicks 2002).

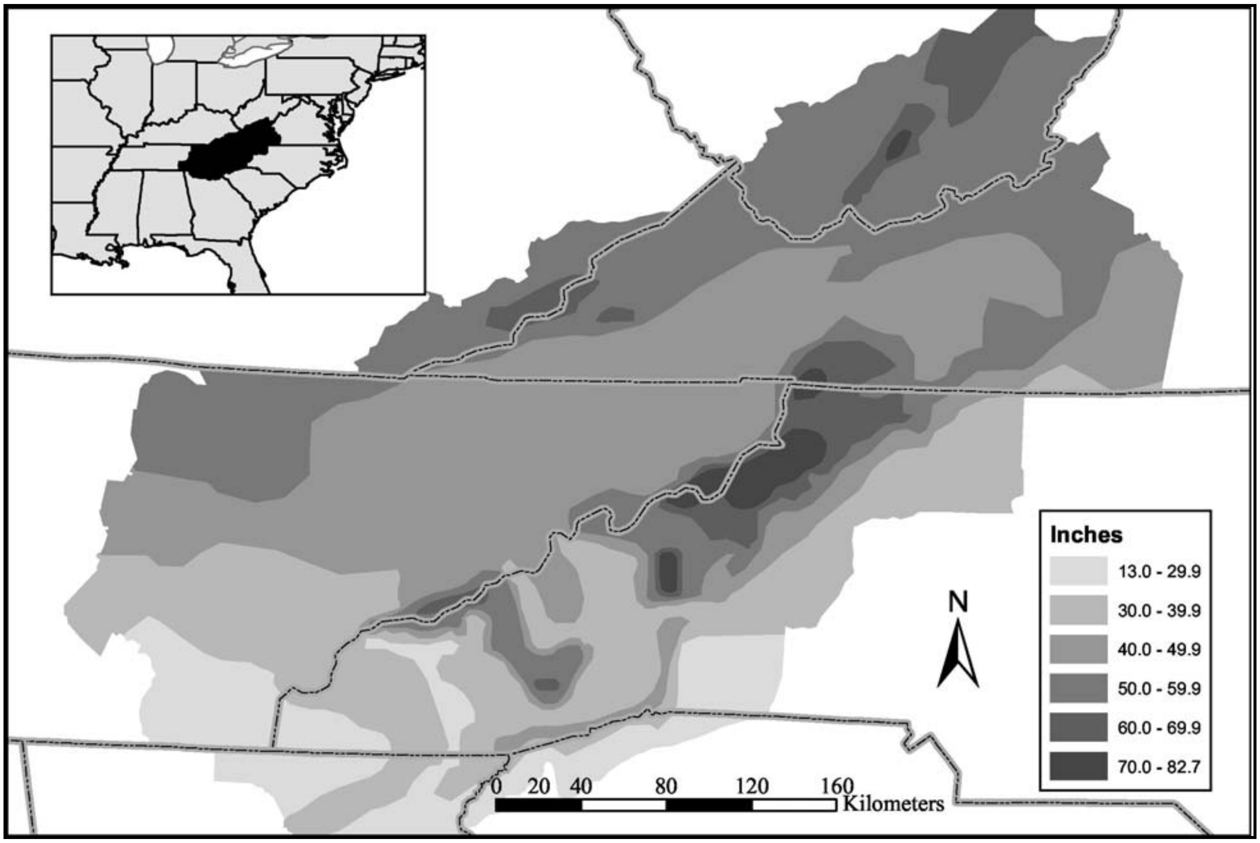

Figure 1.2. Total Snowfall for February/March 1960.

March alone broke all temperature and snowfall records on a monthly basis for most locations across the Southern Appalachians. Average monthly temperatures for almost the entire region were over 6 ˚C (11˚ F) below normal for the month (Fig. 1.3). Cold, arctic high pressure dominated the synoptic pattern in the interior of the United States, whereas an active Texas-Gulf Coast storm track persisted across the Atlantic Coast (Hardie 1960, Ludlum 1960a, Ludlum 1960b). Therefore, the region was under the alternating influence of both of these patterns the entire month. The first major synoptic-scale cyclone of the month came on March 2, a Wednesday, and was followed by other cyclones on the 9th and the 16th, also Wednesdays (Winston-Salem Journal 1960). In the higher elevations and along northwestern slopes, snow was almost a daily occurrence and wreaked havoc, particularly in northwestern

North Carolina, where the Red Cross and National Guard were called on to airlift food and other supplies into the region (Hardie 1960; Minor 1960; Watauga Democrat 1960). The 57 inches of snow that fell in Boone, North Carolina, during the month set a new state record for monthly snowfall. The almost daily snowfall in northwestern North Carolina was due to the passage of frequent synoptic-scale disturbances and persistent northwest flow in association with deep long-wave troughing across the eastern U.S.

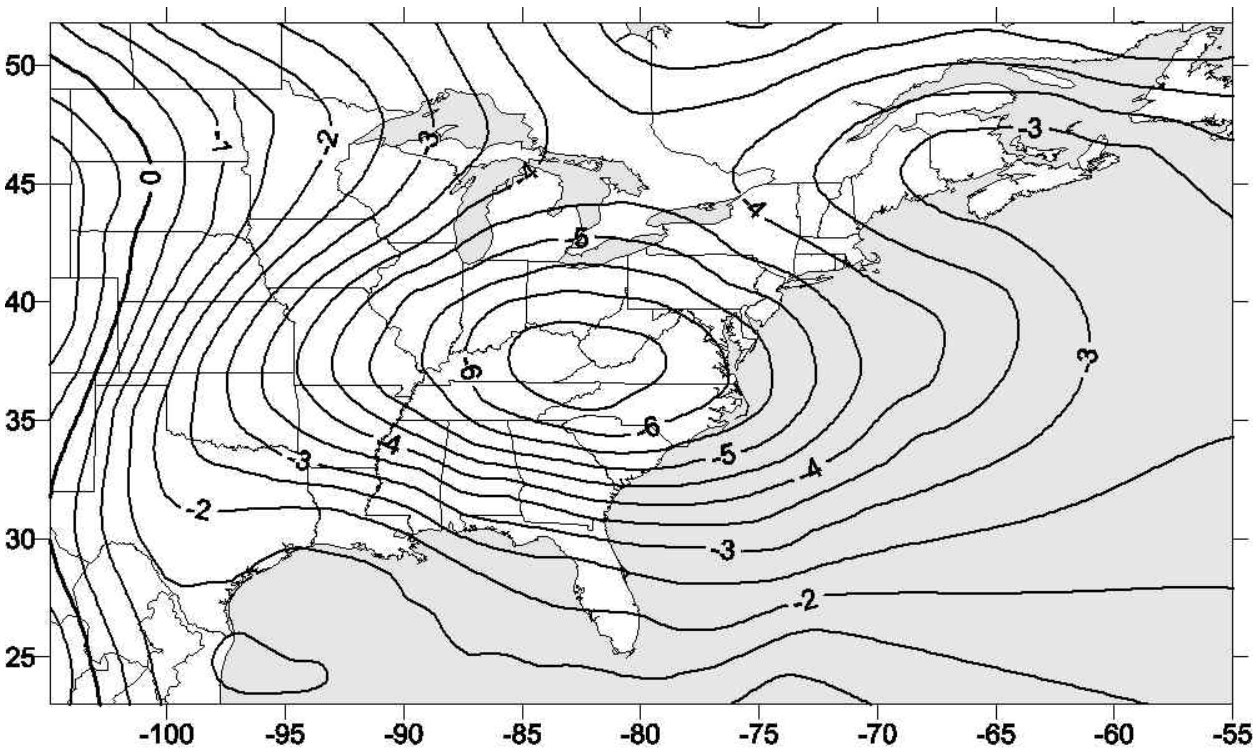

Figure 1.3. Mean 850 hPa temperature departures from normal, March 1960.

1.2.4b. December 18-20, 2003

A recent heavy snowfall event also helps to highlight some of the spatial and synoptic patterns associated with snowfall in the Southern Appalachians. During the three-day period December 18-20, 2003, as much as 76 cm (30 in) of snow fell on higher elevation northwest slopes in conjunction with a prolonged period of NWFS. An Alberta Clipper passed through the region on December 18, producing synoptic-scale ascent and light to moderate snowfall across the region. The surface cyclone was also actually preceded by a period of light snow associated with isentropic lift well ahead of the main period of snowfall. The synoptic-scale ascent not only contributed to the snowfall totals, but perhaps more importantly, it appears to have conditioned the synoptic environment by raising the height of the capping inversion and moistening the middle troposphere. The 00Z 18 Dec 2003 sounding for Blacksburg, Virginia, for example, indicated the base of a stable capping inversion at approximately 825 mb, whereas 24 hours later after the passage of the Alberta Clipper, the height of the capping inversion had risen to 550 hPa and the lower and middle troposphere had moistened considerably (Fig. 1.4). A prolonged period of low-level northwest flow followed, allowing orographic lift to extract the abundant moisture in a conditionally unstable lower troposphere.

Snowfall totals associated with this very heavy NWFS event varied predictably according to elevation and exposure (Fig. 1.5). The heaviest snow in the North Carolina Mountains occurred at higher elevations near the Tennessee border, with a distinct gradient in snowfall totals from the northwest to the southeast. Portions of the southern Blue Ridge in Transylvania and Henderson counties received less than 2.5 cm (1 in) of total accumulation, whereas locations at the same elevation 32 km (20 miles) or so upwind received 30 cm (12 in) or more. The foothills region on the southeastern slopes of the Southern Appalachians also received little, if any, accumulation from this event, highlighting the high precipitation efficiency of the seeder-feeder mechanism in the elevated terrain and the effective scouring of low-level moisture. Downsloping winds and adiabatic warming also aided in inhibiting further snowfall development, and the dry air sublimated much of the snow that may have spilled over from the northwestern slopes.

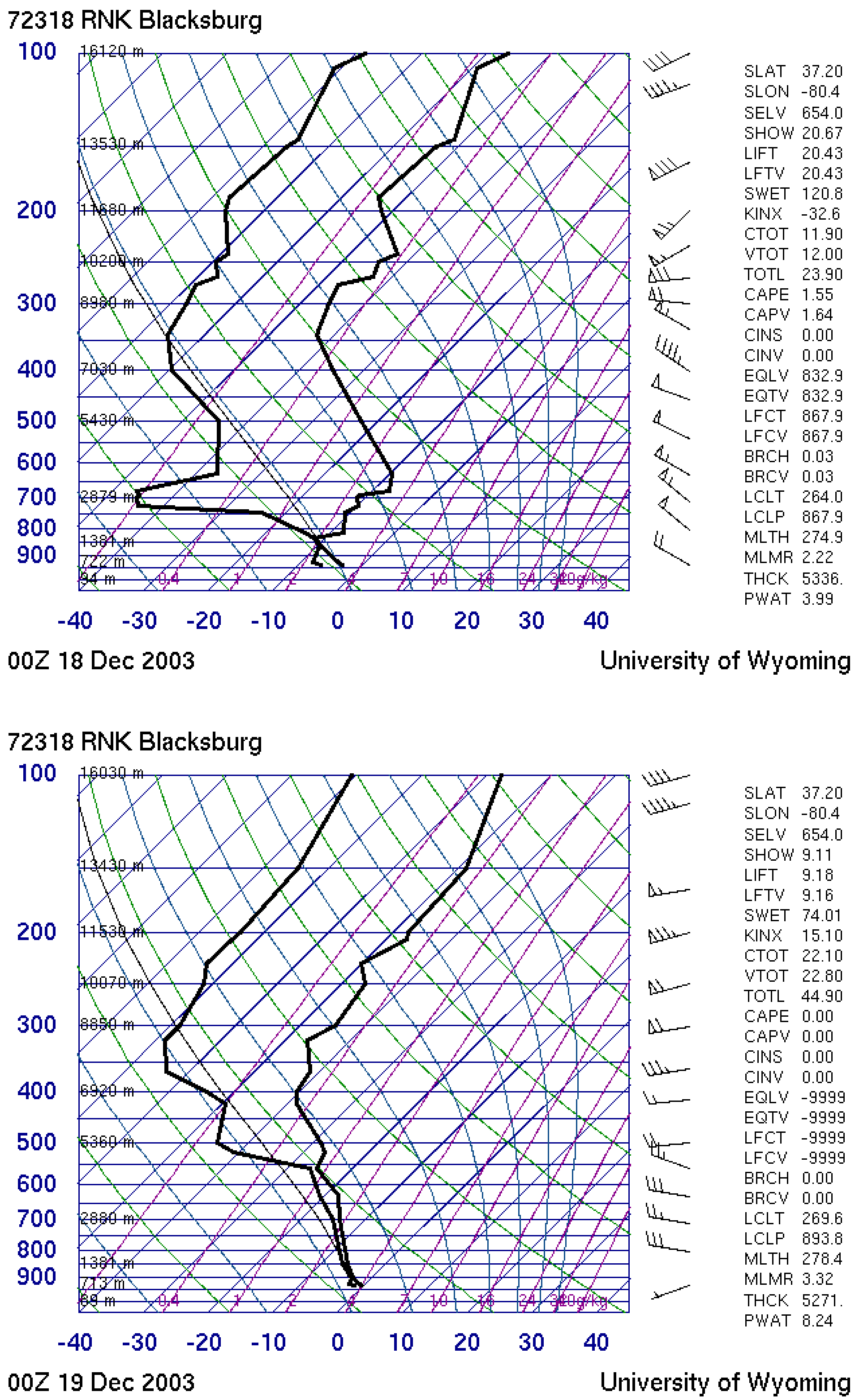

Figure 1.4. Vertical profiles of moisture and temperature for Blacksburg, Virginia, on 00Z 18 Dec 2003 (top) and 00Z 19 Dec 2003 (bottom).

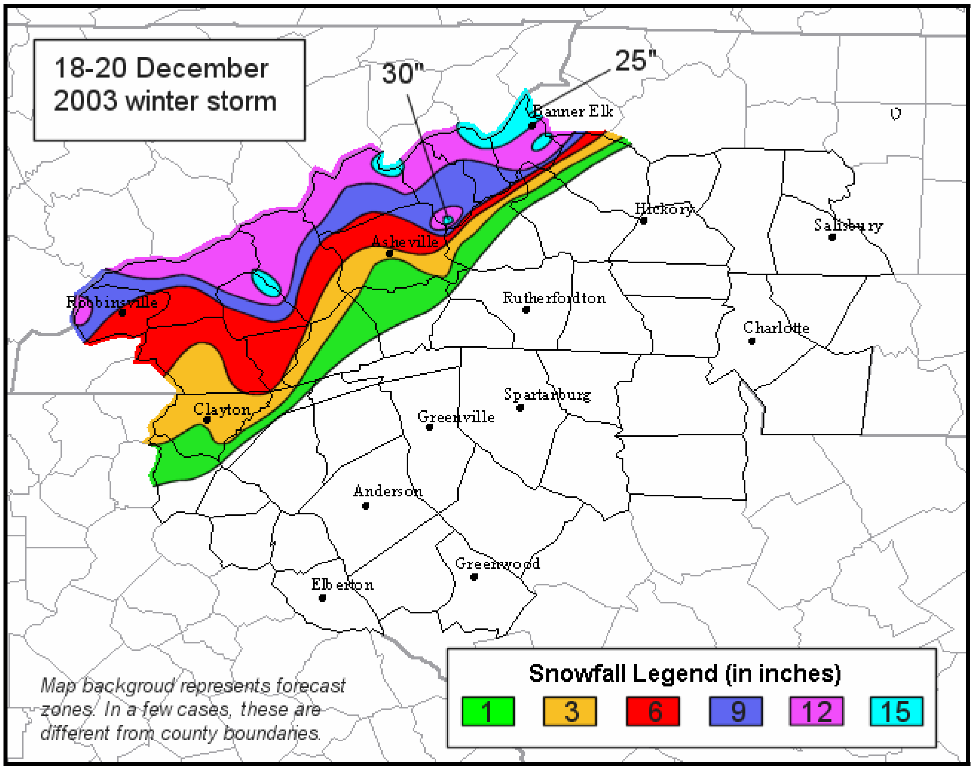

Figure 1.5. Total Snowfall for December 18-20, 2003, NWFS Event in the Greer, SC, County Warning and Forecast Area (http://www.erh.noaa.gov/gsp/localdat/December_18-20.htm).

1.3. Research Design

1.3.1. Research Questions

The following research questions serve to guide this dissertation:

a) What are the monthly and annual patterns of NWFS across the Southern Appalachians and how do these patterns vary between events tied to synoptic-scale ascent and synoptic-scale subsidence?

b) How do the synoptic patterns and parameters vary between NWFS events that produce light versus heavy snow? How do antecedent upstream air trajectories and vertical profiles of moisture vary between light versus heavy events? Why does measurable snowfall occur in some periods of low-level northwest flow, but not in others?

c) How do the synoptic patterns and parameters vary between events characterized by significant differences in snowfall by elevation and those with less difference by elevation? How do they vary between events characterized by significant “spillover” snowfall extending into leeward areas and those that are largely confined to the windward slopes?

d) What are the most frequent air trajectories associated with NWFS across the Southern Appalachians? How do the synoptic patterns, synoptic field values, vertical profiles of moisture and temperature, and snowfall totals vary among different air trajectories? What influence do air trajectories with a Great Lakes connection have on snowfall patterns across the Southern Appalachians?

1.3.2 Research Objectives

The research questions above will be addressed with the following objectives in mind:

a) Develop a 50-year database of NWFS events and summarize data by snow regions.

b) Calculate and map monthly and seasonal patterns of snowfall, including mean snowfall, mean NWFS, percent of NWFS tied to synoptic-scale subsidence, and mean NWFS accumulation by event.

c) Compare composite synoptic fields for NWFS events stratified by snowfall intensity, differences in snowfall by elevation, spillover effects, and antecedent upstream air trajectories. Develop composite synoptic plots and compare across different event types.

d) Complete an antecedent upstream air trajectory analysis for a sample of NWFS events and classify events based on air trajectory.

e) Undertake an analysis of the vertical profiles of temperature and moisture for a sample of NWFS events.

f) Compare composite air trajectories, snowfall totals, synoptic fields, and vertical profiles of moisture and temperature across different trajectory classes, paying particular attention to the influence of the Great Lakes.

CHAPTER II
DATA AND METHODS

2.1. Introduction

This chapter describes the data and methods used to develop a synoptic climatology of NWFS in the Southern Appalachians. It begins with a brief description of the topographic characteristics of the study area, paying particular attention to the topographic forcing that occurs in periods of low-level northwest flow. The chapter continues with a discussion of the source of the snowfall data and the procedures used to define snowfall events, delineate snow regions, and identify NWFS events. The methods used to extract the synoptic fields are also discussed. The final section of this chapter focuses on the methods used for the various analyses, beginning with the construction of a general climatology of NWFS, and continuing with a comparison of events on the basis of snowfall intensity, elevation, and spillover. Finally, the trajectory and sounding analyses are described.

2.2. Study Area: Southern Appalachians

The study area is situated in the Southern Appalachians (Fig. 2.1), a region that stretches from northern Georgia to southern West Virginia and from the Cumberland Plateau to the Blue Ridge foothills in the east. The study area was delineated so as to include the major topographic features of the Southern Appalachians as well as areas immediately upwind or downwind. Therefore, it does not necessarily correspond to cultural (e.g. Raitz and

Ulack 1984) or political delineations (e.g. ARC 2005). The study area trends southwest to northeast, capturing the general orientation of the topography. The highest peaks and associated greatest relief are found in the mountains of Tennessee and North Carolina. Elevations above 1,829 m (6,000 ft), in fact, are confined to the Great Smoky Mountains (NC/TN), Balsam Mountains (NC), Black Mountains (NC), and the Unaka Mountains (NC/TN), whereas 1,500-m (4,921 ft) elevations extend into southwestern Virginia and northern Georgia. Elevated terrain extends northeastward along the spine of the Appalachians through Virginia and West Virginia, but elevations remain below 1,500 m (4,921 ft). The Appalachian Plateau in Tennessee, eastern Kentucky, and extreme southwestern Virginia is characterized by maximum elevations above 1,000 m (3,280 ft), but less than 1,500 m (4,921 ft). The Blue Ridge of North Carolina and Virginia rises abruptly from the adjacent foothills, but elevations generally range between 1,000 and 1,500 m (3,280 to 4,921 ft). The Tennessee Valley is an expansive low elevation area in eastern Tennessee between the Cumberland Plateau and the North Carolina border. The New River Valley, somewhat higher in elevation than the Tennessee Valley, stretches across portions of southwestern Virginia.

The pronounced topographic relief and general southwest to northeast orientation of the Southern Appalachians results in ideal conditions for orographic lifting during periods of low-level northwest flow. Under such synoptic patterns, windward and leeward slopes are well defined (Fig. 2.2), particularly in the vicinity of the higher terrain in close proximity to the crest of the Appalachians. The significant topographic relief and steep northwesterly slopes of the Great Smoky and Unaka Mountains along the NC/TN border forces low-level air parcels to rise abruptly, thereby maximizing NWFS activity at the High Peaks in these areas (Perry and Konrad 2006). Areas immediately downwind of the Great Smoky and

Unaka Mountains experience downslope flow, resulting in a warming and drying of the low-level air mass and limiting snowfall. Farther downwind, however, air parcels descend the leeward slopes of the Blue Ridge, resulting in strong downslope flow and often leading to dissipation of clouds and snow. Similar patterns are evident across West Virginia and Virginia, where upslope flow is maximized across the higher elevations of southeast West Virginia and downslope flow is concentrated in the New River Valley and the southeastern slopes of the northern Blue Ridge.

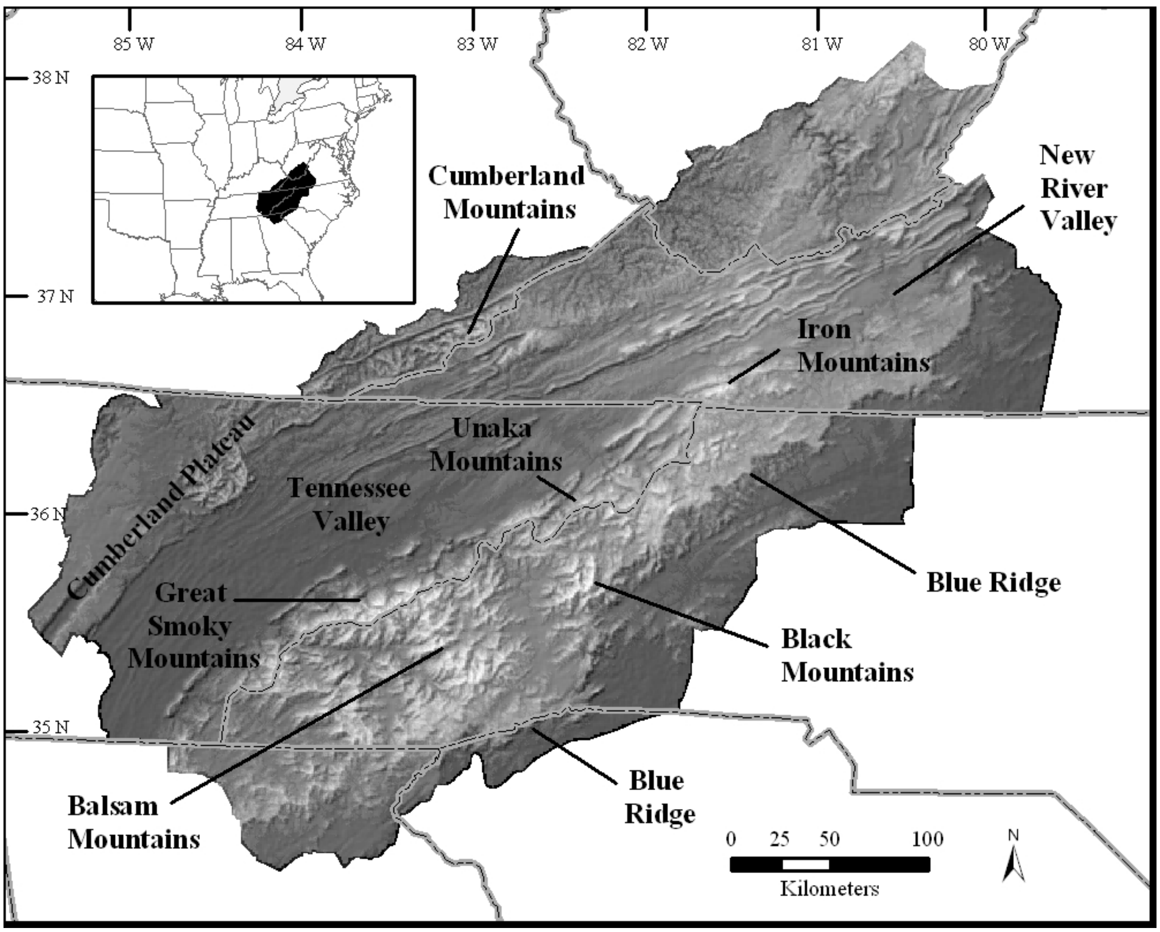

Figure 2.1. Location of study area and important topographic features (Fenneman 1938).

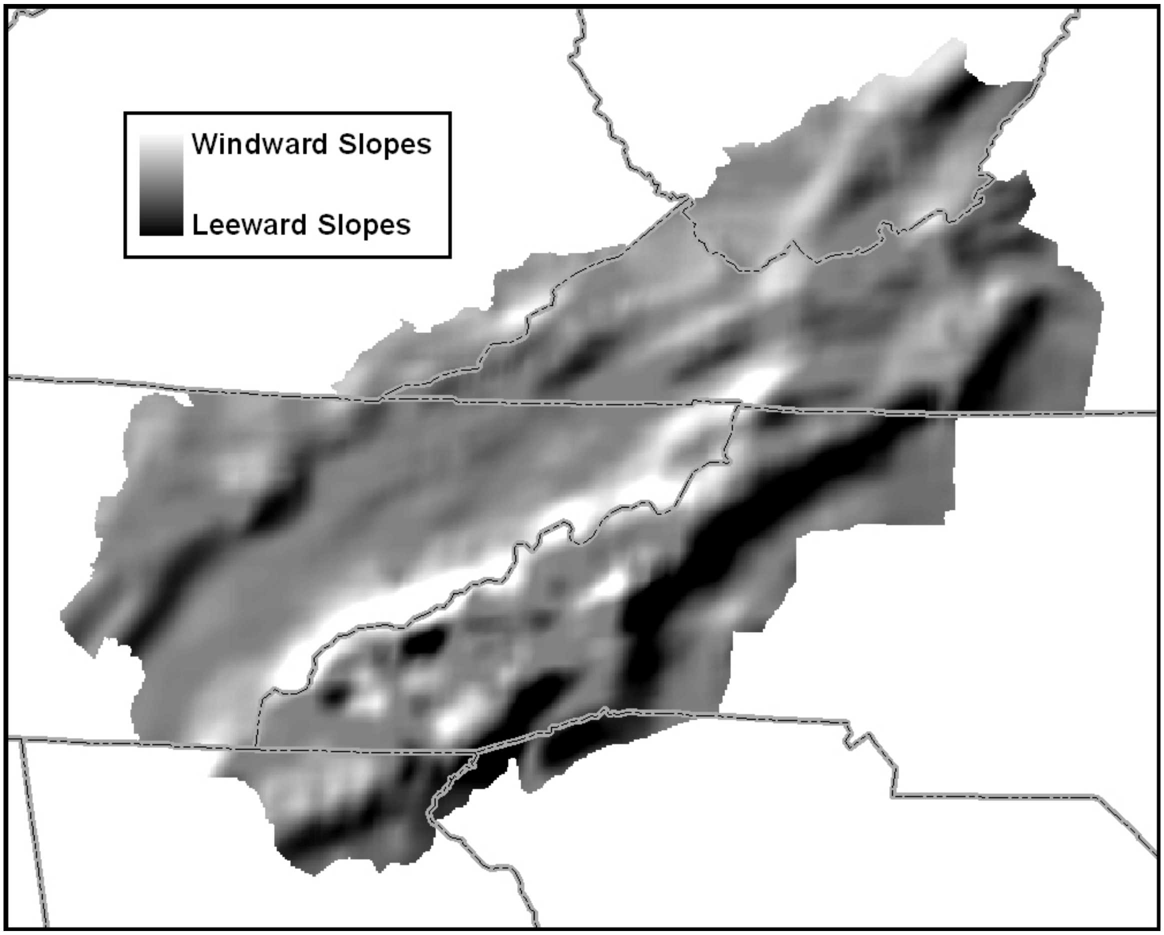

Figure 2.2. Locations of windward and leeward slopes during periods of northwest flow.

2.3. Snowfall Data and Synoptic Fields

2.3.1. Snowfall Data and Event Definition

2.3.1a Daily snowfall data

This study parsed daily snowfall data to estimate storm snowfall totals associated with specific synoptic disturbances. Daily snowfall records for the period from 1950 to 2000 served as the source of the snowfall data. These data came from the National Climatic Data Center's Cooperative Summary of the Day CD-Rom (NCDC 2002), which consists of daily observations of minimum and maximum temperature, precipitation, snowfall, and snow depth. This study used 121 cooperative observer stations in the Southern Appalachians, which represented all of the stations in the region with a minimum of 10 years of snowfall data.

Since most coop observers are volunteers, the quality of the data can sometimes be a concern, particularly with snowfall and snow depth measurement (Robinson 1989). However, even when observers follow standardized procedures for measuring snowfall, variation may still exist within the same area due to inter-observer variability (Doesken and Judson 1997, Doesken and Leffler 2000). The observation times of the coop data also introduce some inconsistency for snowfall measurement, as some stations report at 7 AM, others at 5 PM, and a handful at 12 AM. In situations where snowfall events occurred less than 24 hours apart, a small portion of a coop station's event total snowfall may be attributed to the prior or subsequent event due to different reporting periods. It would be helpful to have fully standardized and higher quality coop data, but unfortunately these are the only snowfall data available at sufficient spatial resolution to fulfill the aims of this research project.

2.3.1b. Event definition using first-order stations

Some periods of snowfall may be connected with a single synoptic-scale disturbance and therefore span only one day, whereas others may result from multiple disturbances or a prolonged period of upslope flow and thus have durations ranging from two to as many as five days. The daily snowfall data by themselves are informative, but are not tied explicitly to synoptic-scale snowfall events. Therefore, it was necessary to calculate event snowfall totals by summing the daily snowfall data through the duration of each event. A total of 1,641 snowfall were identified across the region for the 50-year period, or an average of 32 events per snow season.

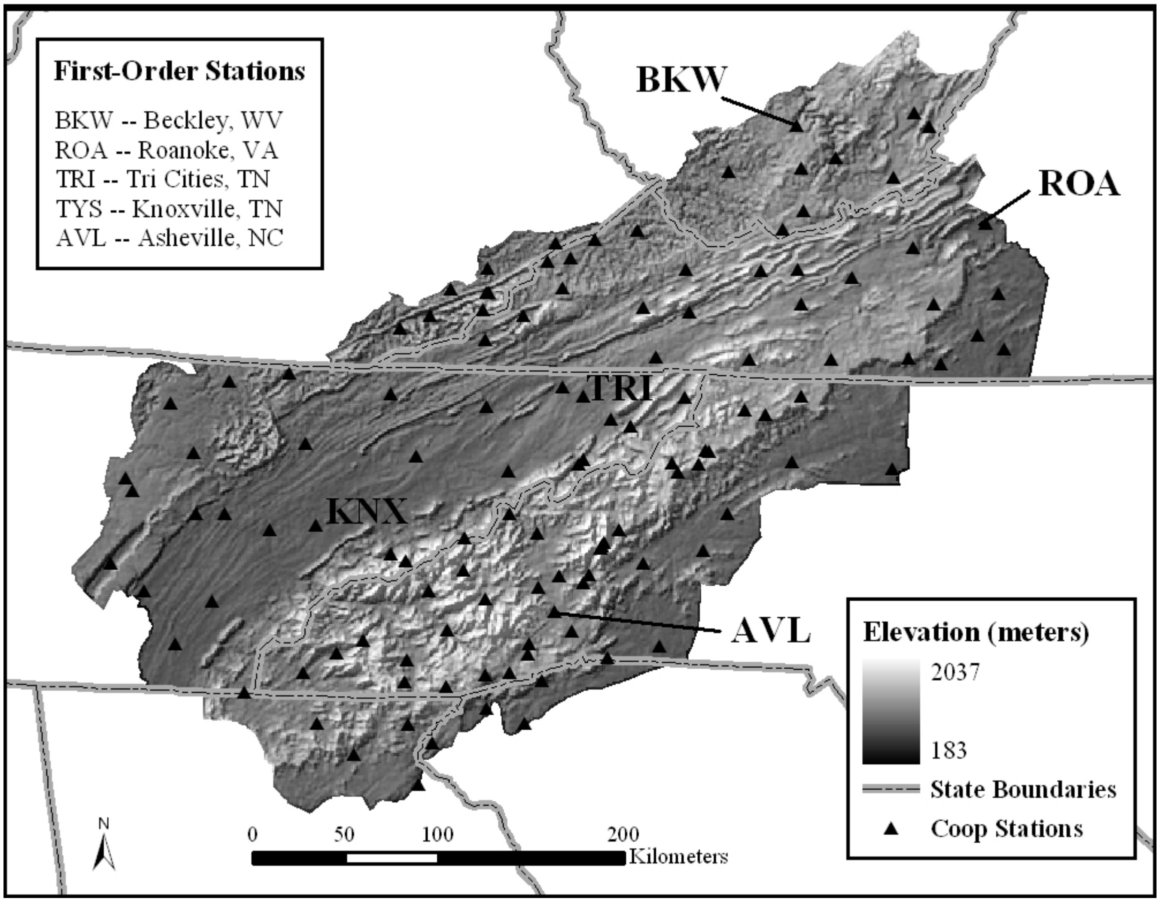

Figure 2.3. Topography of the Southern Appalachians with weather stations used in the study.

Each snowfall event was identified by the occurrence of measurable (≥ 0.1 in) snowfall at one or more coop stations across the region. If snowfall was reported, hourly weather conditions (i.e. precipitation occurrence and type), precipitation intensity, and temperature data were acquired for all available first-order stations in the region (Fig. 2.3): Knoxville, TN, Tri-Cities, TN, Beckley, WV, Asheville, NC, and Roanoke, VA. The beginning time of the snowfall event was defined as the hour snow was first reported at any of the first-order stations, the maturation time the hour in which the spatial extent of snow was the greatest (Fig. 2.4 – i.e. reported at the most number of first-order stations in the same hour), and the ending time the hour snow no longer was reported at any of the first order stations.

First-order stations

Maturation Hour = Greatest Spatial Extent of Snow

CITY	3-23	4- 0	4- 1	4- 2	4- 3	4- 4	4- 5	4- 6	4- 7	4- 8	4- 9	4-10	4-11	4-12	4-13	4-14
knx									S	S	S	S				
avl			R							S				S	S	S
tri	R	R				S	S	S	R	S	S	S				S
bkw	R		R	S	S	S	S	S	S	S	S	S	S	S	S	S
knx	47.	46.	45.	42.	41.	40.	39.	39.	38.	38.	37.	37.	39.	39.	40.	39.
avl	44.	47.	41.	42.	40.	39.	35.	34.	35.	33.	35.	36.	39.	37.	37.	36.
tri	45.	43.	42.	39.	37.	36.	35.	34.	35.	34.	35.	35.	37.	35.	35.	35.
bkw	36.	36.	36.	33.	31.	30.	30.	30.	29.	29.	29.	29.	29.	29.	28.	28.

Date	WISE 1 SE	WYTHEVIL	BECKLEY	BLUEFIELI	BLUESTOI	FLAT TOP
19871204	3.5	0	3	2	0	1.5
19871205	1.1	0	1	0	0	2.5
19871204	**4.6**	**0**	**4**	**2**	**0**	**4**

Coop stations

Two-day event total

Figure 2.4. Identification of event maturation hour and calculation of event snowfall totals. Shown are hourly patterns of precipitation type (S = snow, R = rain) and temperature for first-order stations (top) and daily snowfall totals and event snowfall totals for selected coop stations (bottom).

An event remained active as long as precipitation was reported at any first-order station other than Beckley, WV. In Beckley, snowfall often continued for several days due to orographic effects, when in fact different synoptic-scale disturbances in close succession were responsible for the prolonged snowfall. Since the other first-order stations are lower in elevation and considerably less exposed to northwest flow than Beckley, the orographic effects are much less pronounced. Therefore, event durations are shorter and snowfall can be more easily tied to the individual synoptic-scale disturbances. Therefore, it is possible to break up these prolonged periods of upslope snows in Beckley using data from Tri-Cities and Knoxville. The event total snowfall is thus calculated by summing the daily coop station snowfall data for each day the snowfall event remains active (Fig. 2.4).

A minor deficiency of the first-order time series data is the absence of a high elevation station. Beckley, at 763 m (2,504 ft) and Asheville, at 683 m (2,117 ft) are the highest stations; however, roughly 21 percent of the coop stations exceed these elevations and 11 percent exceed 1000 m (3,280 ft). In some events, NWFS will continue for a longer period in the higher peaks; therefore, the calculated event ending times may be somewhat earlier (and hence event durations somewhat shorter) than is actually the case since NWFS may continue for a longer period along higher elevation windward slopes. In other cases, snowfall may occur at the High Peaks while Asheville and Beckley receive only rain. In fact, approximately 20 percent of the snowfall events are not connected with any hourly snowfall reports at the first order stations. In these cases, the beginning, maturation, and ending times were estimated based on the hourly observations of temperature and precipitation at the first order stations. In a few instances, no hourly precipitation was reported at any of the first-order stations, in which case the beginning, maturation, and ending times were estimated using the daily precipitation data and hourly cloud cover and temperature data.

2.3.1c. Development of snow regions

Due to the large areal extent and significant topographic diversity within the Southern Appalachians, the study area was divided into 14 snow regions. The primary purpose of these groupings was to facilitate intra-regional comparisons. Coop stations were grouped together into snow regions based on similarities in snowfall patterns, elevation, and topography (Table 2.1 and Fig. 2.5). In most cases these regions correspond to zone groupings that the National Weather Service (NWS) uses for forecast products. However, it is important to note that the High Peaks Snow Region (Region 14) is not contiguous, but rather consists of all locations in the study area with elevations greater than 1,200 m (4,000 ft). These areas are largely limited to the mountains of eastern Tennessee and western North Carolina. Recent work (Perry and Konrad 2006) has shown that higher elevation stations (e.g. Mt. Mitchell, NC) in close proximity to valley stations (e.g. Asheville, NC) average as much as 75 cm (1,700 percent) more NWFS on an annual basis, suggesting dramatic differences in snowfall climatologies. Mean and maximum event snowfall totals for those stations reporting snow were calculated for each snow region. The use of snow regions also helps to minimize the deleterious effects of missing or inaccurate data at individual coop stations by placing greater emphasis on regional rather than the local or point scale patterns. Though useful for the purposes of this study, it is important to recognize that considerable variability of snowfall still occurs within the snow regions, particularly in those with greater topographic relief.

Table 2.1. Snow regions and cooperative observer stations.

	Snow Region	Coop Stations
1.	Southern Tennessee Valley	Athens, Cleveland, Dayton, Kingston, Knoxville, Lenoir City, Pikeville, Rockwood
2.	Southwest Mountains	Blairsville, Andrews, Murphy, Copperhill
3.	Southern Foothills	Clayton, Helen, Toccoa, Forest City, Tryon, Long Creek, Pickens, Salem
4.	Great Smoky Mountains	Cataloochee, Hot Springs, Marshall, Oconoluftee, Waterville, Waynesville, Gatlinburg
5.	Southern Blue Ridge	Asheville, Black Mountain, Brevard, Coweeta, Cullowhee, Fletcher, Franklin, Highlands, Lake Toxaway, Pisgah Forest, Rosman, Caesars Head
6.	Southern Plateau	Allardt, Crossville, Newcomb, Norris, Oneida, Tazewell
7.	Northern Tennessee Valley	Bristol, Elizabethton, Erwin, Greeneville, Kingsport, Rogersville, Abingdon
8.	High Country	Banner Elk, Blowing Rock, Boone, Celo, Jefferson, Transou, Trout Dale, Mountain City
9.	Central Foothills	Lenoir, Marion, Morganton, North Wilkesboro, Yadkinville
10.	New River Valley	Sparta, Blacksburg, Bland, Floyd, Galax, Meadows of Dan, Pulaski, Wytheville
11.	Northern Foothills	Martinsville, Philpott Dam, Roanoke, Rocky Mount, Stuart
12.	Central Plateau	Baxter, Blackmont, Burdine, Closplint, Cumberland Gap, Pine Mountain, Skyline, Big Stone Gap, Grundy, John Flanagan, North Fork Reservoir, Pennington Gap, Wise
13.	Northern Plateau	Burkes Garden, Beckley, Bluefield, Bluestone Lake, Flat Top, Lewisburg, Pineville, Princeton, Union, White Sulphur Springs
14.	High Peaks (Above 4000 ft)	Grandfather Mountain, Mt. Mitchell, Swannanoa, Mt. Leconte

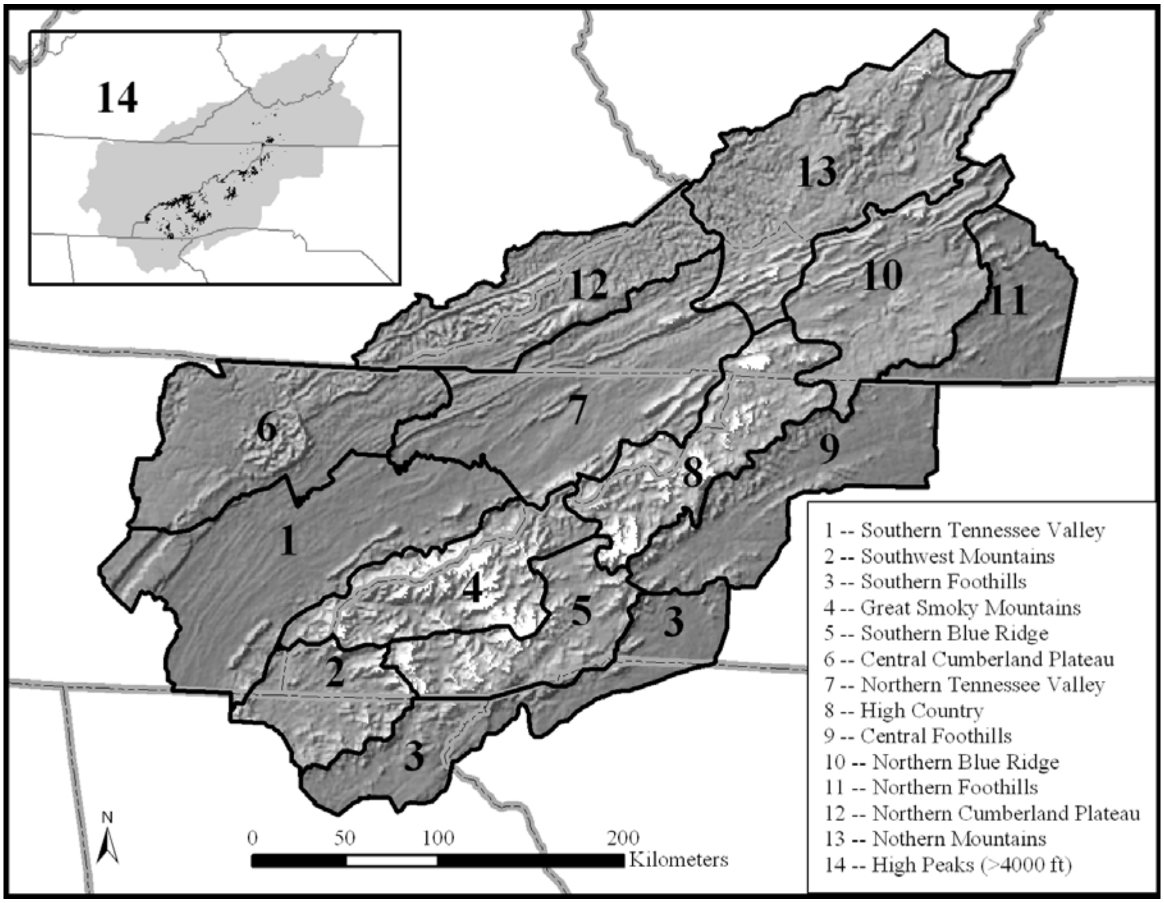

Figure 2.5. Snow regions used in the study. The High Peaks (Region 14) are shaded white on the large map and black on the inset map.

2.3.1d. Identification of NWFS events

NWFS events were defined by the occurrence of a northwest (270 to 360 degrees) low-level (850 hPa) wind at event maturation (hour at which the spatial extent of snow was greatest) (Fig. 2.6). Values for 850-hPa wind direction at the mean center of each snowfall event were calculated from NCEP reanalysis data (discussed in next section) using Synoptic Climatology Suite (Konrad and Meaux 2002) and written to a database file. NWFS events were then selected using the “filter” option in Microsoft Excel. Wind direction at the 850-hPa level (found at approximately 1,450 m (4,757 ft) in periods of northwest flow) is a good indication of the mean wind direction between 1,000 and 2,000 m (3,208 and 6,562 ft), where much of the orographic enhancement likely occurs. Additionally, the 850-hPa level is a standard height at which a variety of atmospheric data are routinely collected and archived by several governmental agencies and represents a good source of long-term wind direction data. A total of 859 NWFS events were identified across the Southern Appalachians during the period 1950 to 2000, or an average of 17 per year.

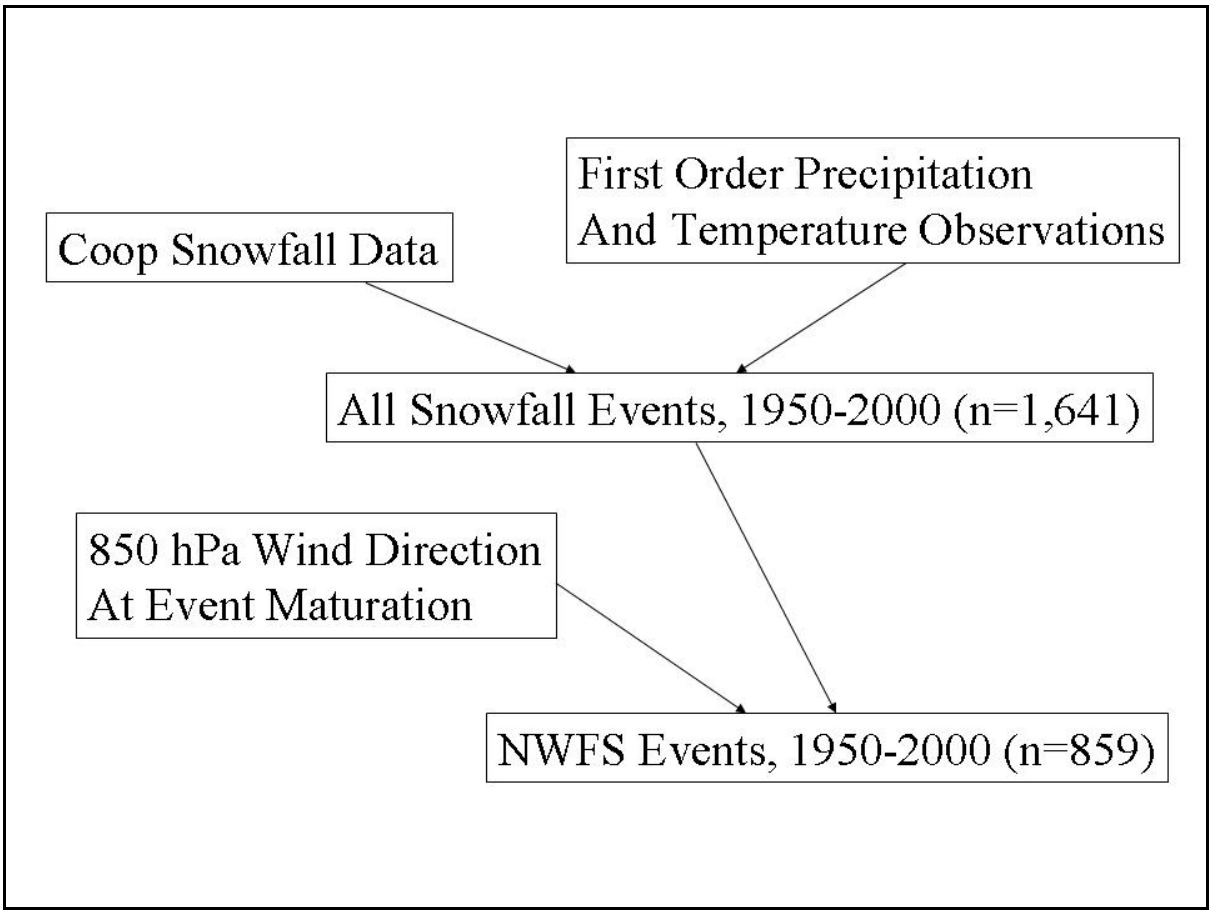

Figure 2.6. Identification of NWFS events.

The identification of NWFS events using the automated approach and NCEP reanalysis dataset in the manner described above has many advantages. First of all, it provides a consistent and objective method of identifying NWFS events based on their defining characteristic: low-level (850 hPa) northwest flow. Secondly, values for the 850 hPa wind direction are extracted at the time in which the spatial extent of snowfall (event maturation) is greatest across the region. Lastly, this automated approach is useful for exceptionally long time series of data, such as the one utilized in this study, where a manual synoptic typing using weather map analysis is unwieldy. One caveat of this methodology is that some of the major snowstorms that affect the Southern Appalachians may be classified as NWFS, even though they are associated with Gulf Lows of the Miller Type A variety (Gurka et al. 1995) with a prolonged period of NWFS on the back side of the surface cyclone. For example, using this methodology, the Blizzard of '93 exhibited an 850-hPa northwest flow at event maturation, but a substantial amount of the storm total snowfall was also tied to low-level south and southeast flow prior to backing of winds to the northwest. Therefore, the amount of snowfall occurring in conjunction with low-level northwest flow may be over reported. Due to the limited temporal resolution (i.e. 24-hr totals only) of the coop data, it was unfortunately impossible to assign snowfall accumulations to specific low-level wind directions in these events. Nonetheless, even in the extreme snowstorms such as the Blizzard of '93, the bulk of the total snowfall at higher elevations and windward slopes occurred in conjunction with low-level northwest flow (e.g. Goodge and Hammer 1993).

2.3.2. Synoptic Fields

Table 2.2 lists 16 synoptic fields that were extracted from the National Center for Environmental Prediction (NCEP) reanalysis dataset (Kalnay et al. 1996) and used to characterize the synoptic environment associated with the study events. These 2.5 x 2.5 degree gridded data were generated from a data assimilation routine that incorporated available surface and upper-air observations (radiosonde and aircraft) from around the world and therefore provide unmatched continuity over a long period of time. The gridded data were used to obtain the u and v components of 850 hPa wind direction. These data were then spatially interpolated onto a 1776 km grid (197 km resolution), with the center grid corresponding to the mean center (weighted by coop snowfall totals) of each snowfall event. Using the 0000 and 1200 UTC gridded synoptic fields, a temporal interpolation was undertaken to estimate field values during the event maturation time. This interpolation procedure was carried out for the remaining synoptic fields. This study used an inverse distance technique to carry out all spatial and temporal interpolations. These fields together provide a good characterization of the synoptic environment (e.g. moisture, circulation, stability, thermal structure, etc.) associated with NWFS.

Table 2.2. Synoptic fields derived from the NCEP reanalysis dataset used in this study.

Synoptic Variables	Units
1000 hPa Height	Meters
500 hPa Height	Meters
850 hPa Mixing Ratio	g H_20/kg dry air
850 hPa Relative Humidity	Percent
850 hPa Wind Direction	Degrees
850 hPa Wind Speed	ms^{-1}
850 hPa Temperature	ºC
850 hPa Thermal Advection	ºC/12 hour
1000-500 hPa Mean Relative Humidity	Percent
700 hPa Vertical Velocity	hPa/hour
850-500 hPa Lapse Rate	ºC
Precipitable Water	Inches
500 hPa Relative Humidity	Percent
850 hPa Theta-E	ºC
500 hPa Vorticity	$1*10^{-5}s^{-1}$
500 hPa Vorticity Advection	$1*10^{-5}s^{-1}$/12 hour

Although the NCEP reanalysis data are unmatched at the synoptic-scale and provide the only available long-term source of data, several minor limitations are worth discussing. The limited spatial resolution (2.5 degrees by 2.5 degrees) is one concern, as mesoscale vorticity maxima and other sub-synoptic-scale features are not readily discernible. Data are also limited to five mandatory levels: 1000, 850, 700, 500, and 200 hPa, when greater vertical resolution, particularly in the lower troposphere, would be helpful. The moist layer, which contributes to NWFS in the Southern Appalachians, is typically confined to the lower troposphere, with the top of the moist layer rarely extending above 600 hPa. Therefore, greater vertical resolution would aid in the assessment of synoptic patterns of humidity, moisture, and temperature within the critical layer of snowfall development. This is also true with some of the other synoptic fields, such as lapse rate, which is only calculated between 850 and 500 hPa. This measure therefore does not adequately capture the finer scale conditional instability in the synoptic environment below 800 hPa characteristic of most NWFS events. Greater temporal resolution of the NCEP reanalysis data would also be desirable, as the temporal interpolation scheme using the inverse distance technique is likely to introduce some error for events in which an abrupt change in the synoptic pattern occurs in conjunction with event maturation times four to six hours before of after the 0/12 Z data. Vertical temperature and moisture profiles will be calculated from atmospheric sounding data to provide a more detailed perspective (see Section 4f below).

2.4. Analyses

2.4.1. Climatology of NWFS

A general climatology of NWFS in the Southern Appalachians was constructed by snow region using the snowfall data from 1950 to 2000. Mean annual snowfall was calculated for each snow region by summing the mean snowfall of those stations reporting snowfall for each event and then dividing by 50 years. Likewise, average annual NWFS by snow region was calculated in a similar fashion, allowing the percent of average annual snowfall attributed to NWFS to be readily calculated. A negative bias in these calculations may exist, however, as coop stations are largely positioned in valleys as opposed to ridge tops. The NWFS events were stratified into two groups according to the polarity of the synoptic-scale vertical motions at 700 hPa (e.g. events associated with synoptic-scale ascent vs. subsidence). Events were stratified into these two groups in order to differentiate between NWFS events that are primarily due to synoptic-scale disturbances versus those in which orographic forcing predominates. It is well known that a large percentage of NWFS events occur in the absence of synoptic-scale ascent across the Southern Appalachians, but it is unclear how differences in the topographic characteristics of snow regions may influence the frequency or accumulations of NWFS events connected with the two primary types of forcing. Therefore, the percent of NWFS events in each region connected to a) synoptic-scale ascent and b) synoptic-scale subsidence were calculated. Mean annual snowfall was also calculated for each group and compared across snow regions.

The mean number of NWFS events by year and month for each snow region was also calculated and the results mapped. A NWFS event, by definition, occurred if at least one coop station in a snow region reported measurable snowfall (> 0.1 in, or 0.25 cm). Lastly, the mean snowfall accumulation associated with each NWFS event was calculated by totaling the snowfall for the 50-year period and then dividing by the number of events. This was done for the entire snow season as well as each month. The purpose of this analysis was to ascertain if any significant differences in mean event snowfall totals existed across the study area.

2.4.2. Intensity of Snowfall

Comparisons were also made between light and heavy NWFS events. Light events may only result in minimal to moderate impacts, whereas the impacts may be considerably greater with heavy events. Therefore, it is helpful to compare the synoptic fields and map patterns between light and heavy events to better understand how and why heavier snowfall occurs in some events, but not in others. Many previous precipitation studies have focused only on heavy events (e.g. Maddox et al. 1979, Mote et al. 1997), when a comparison of the exceptional (heavy) events with the ordinary (light) events is perhaps more useful from an analytical and forecasting standpoint. The snow regions chosen for further analysis were the higher elevation windward slopes: Great Smoky Mountains (Region 4), High Country (Region 8), Central Plateau (Region 12), Northern Plateau (Region 13), and High Elevations (Region 14) (Fig. 2.7). In the other snow regions, the number of NWFS events is significantly lower, as are the typical snowfall accumulations and associated impacts.

The Fields Developer and Climatology Explorer modules of the Synoptic Climatological Suite (Konrad and Meaux 2002) allowed for NWFS events to be stratified by intensity. Light events were defined as the bottom three quartiles of the mean snowfall of those cooperative observer stations reporting snowfall in the respective snow regions, and heavy events were defined as the top quartile. The heavy events represent the tail of the histogram (Fig. 2.8). In the High Country (region 8), all of these events met the current criteria for an advisory and over half for a warning, with an areal mean snowfall of at least 5.6 cm (2.2 in) (NWS 2006). Composite mean values for each of the 15 synoptic fields centered over each event were compared between the two groups and the statistical significance evaluated using a two-sample difference of means t-test. Composite plots of selected synoptic fields were also developed for both the light and heavy events to better illustrate the synoptic patterns. In addition, composite map patterns of the mean difference between heavy and light events were developed for those synoptic fields that displayed a significant ($p < 0.05$) difference.

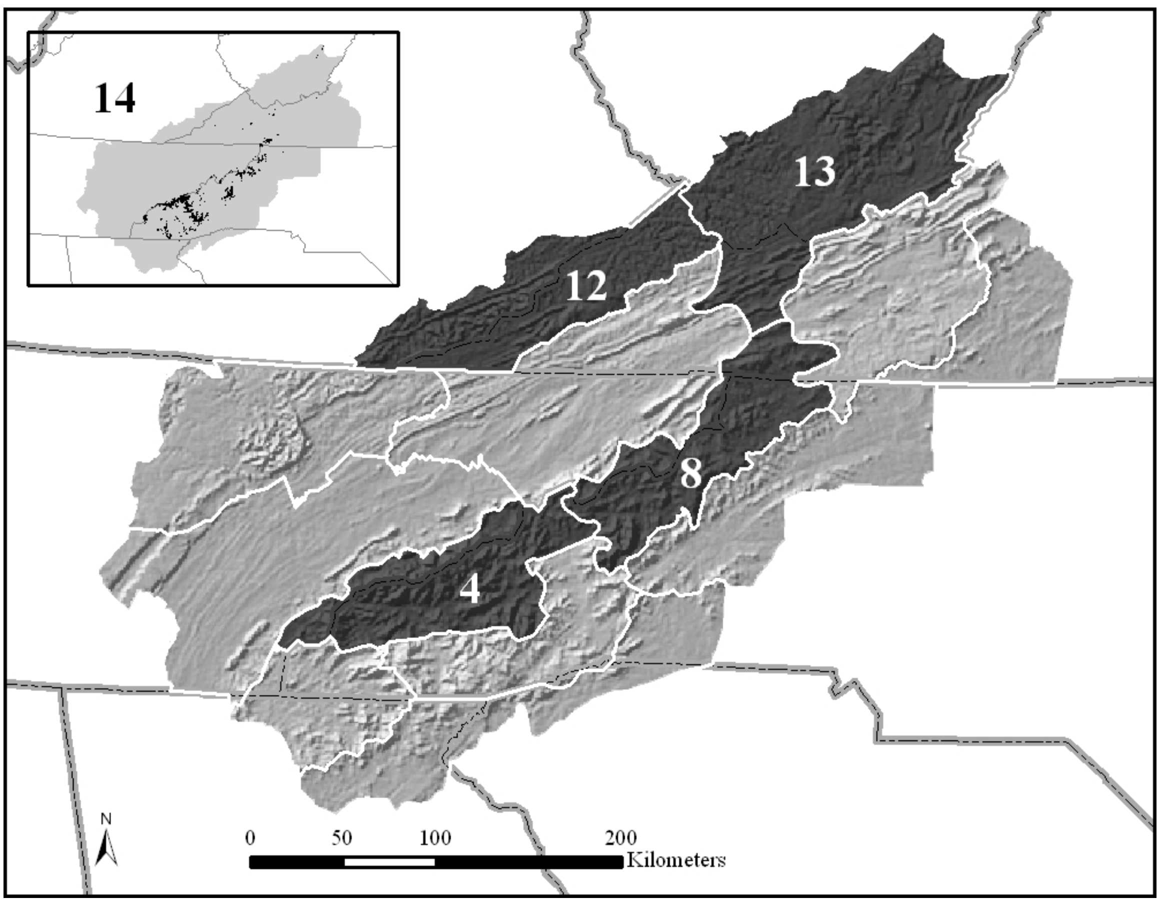

Figure 2.7. Snow regions analyzed in light vs. heavy comparison.

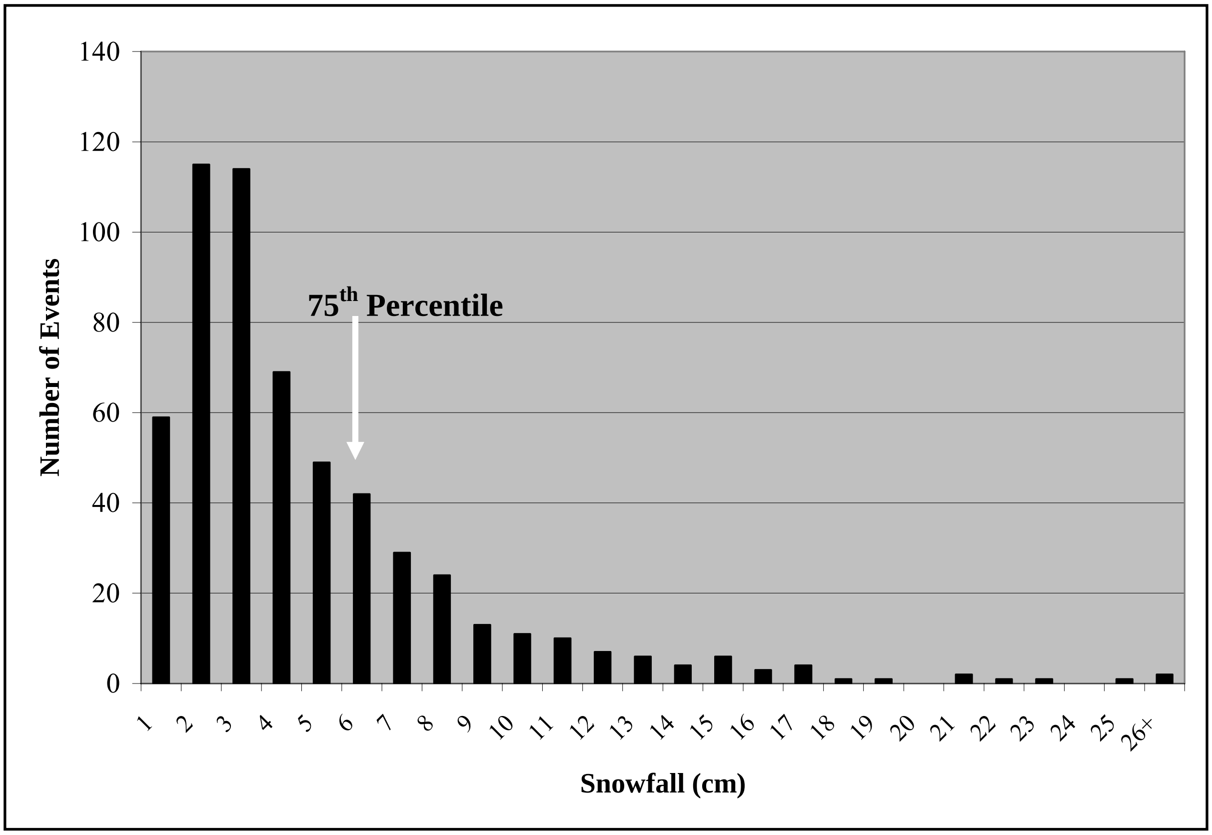

Figure 2.8. Frequency of NWFS events by snowfall amount for the High Country.

In addition to light versus heavy comparisons for the entire sample of NWFS events in each of the five snow regions, this study also evaluated snowfall intensity for events grouped by synoptic-scale vertical motions. Snowfall accumulations are typically greater and more widespread (e.g. more likely to occur at lower elevations and along windward slope) in the presence of synoptic-scale ascent, whereas forcing is primarily orographic in situations of synoptic-scale subsidence, with lighter accumulations that are limited to the higher elevations and windward slopes. Therefore, light and heavy events were compared for just those events that exhibited synoptic-scale ascent at event maturation in order to evaluate the differences in synoptic fields and composite map patterns. Likewise, snowfall intensity was evaluated for those events with synoptic-scale subsidence at event maturation. Composite plots were developed for selected fields to better illustrate the synoptic patterns associated with each type of event and to highlight the differences in map patterns between light and heavy events.

Not all episodes of low-level northwest flow during the cold season result in accumulating snowfall. Many times, only snow flurries or rime ice occurs, presumably due to limited moisture. In order to better understand the synoptic patterns associated with these “snow-free” episodes of low-level northwest flow, comparisons were made between NWFS events in which light accumulations occurred and episodes of low-level northwest flow with no measurable snow. The period 1988 to 2000 was chosen for analysis due to the availability of both of the high elevation stations of Mt. Leconte and Mt. Mitchell. The light NWFS sample consisted of the bottom quartile of the mean snowfall for events with measurable snowfall reported in the High Peaks (Region 14), which yielded a total of 33 events for the 12-year period. The other sample was constructed by using NCEP reanalysis data and the Synoptic Climatology Suite (Konrad and Meaux 2002) to extract values for 850 hPa wind direction, 850 hPa temperature, and 850 hPa relative humidity for the 0 Z and 12 Z maptimes for every day during the snow season (October 1 through May 31) between 1988 and 2000. A sample of 33 episodes of snow-free low-level northwest flow were identified that met the following criteria: 1) northwest 850 hPa wind direction (270 to 360 degrees), 2) 850 hPa temperature less than 0˚ C, and 3) 850 hPa relative humidity greater than 50 percent. Clearly, some episodes of low-level northwest flow are just too warm for snow, so it was important to limit the sample based on temperature. Likewise, snow will not develop in the absence of moisture, necessitating an additional criterion.

2.4.3. Influence of Elevation

Some NWFS events are exclusively confined to the High Peaks (Region 14), whereas others produce widespread snowfall across all elevations. In some situations this is a product of temperature, as adiabatic cooling associated with upslope flow produces cooler temperatures at higher elevations. In other events, however, temperatures are plenty cold enough across the region, but moisture is apparently a more significant limitation. At the High Peaks, the limited moisture may be maximized due to enhanced upslope flow as a result of greater topographic relief and higher relative humidity. Choularton and Perry (1986) suggest that precipitation efficiency and associated scavenging of low-level moisture is significantly greater when temperatures in the moist layer are within the dendritic temperature growth range. However, it remains unclear which processes are most significant in producing heavier snowfall in the High Peaks when compared with the remainder of the study area.

To evaluate the influence of elevation on NWFS, a ratio of the mean snowfall for the High Peaks to the remaining snow regions was developed. Using this approach, a high ratio indicates NWFS events in which the High Peaks received exceptionally more snow than lower elevations. The highest ratios signify events in which snow was observed at the High Peaks, but no snow fell at the remaining lower elevations snow regions. The Fields Developer and Climatology Explorer modules of Synoptic Suite were used to compare the top quartile of this ratio (large differences in snowfall accumulations) to the bottom three quartiles. Composite values of the 16 synoptic fields were compared between the two groups and composite map patterns developed for selected fields. In addition, composite plots of the mean difference in the synoptic fields between the groups were developed to further assess the differences in synoptic patterns. A second analysis was also carried out to evaluate the influence of elevation, but only included events in which snow was reported in the lower elevation snow regions of the Great Smoky Mountains (Region 4) and the Southern Blue Ridge (Region 5), which are perforated by most of the High Peaks region. As in the previous analysis, comparisons were made between the top quartile and bottom three quartiles of the ratio of the mean snowfall for the High Peaks to the remaining snow regions.

2.4.4. Spillover Effects

Determining which NWFS events have the potential to produce more widespread snowfall accumulations downwind from the favored upslope locations remains a particularly acute challenge for forecasters. NWFS occasionally exhibit accumulating snowfall that "spills over" the higher windward slopes to the Blue Ridge and even into the Foothills. In the overwhelming majority of the events, however, locations immediately downstream of the Blue Ridge receive little or no measurable snowfall due to the shadowing and downsloping that develops. Although the synoptic patterns that favor spillover snowfall are not well understood, stronger synoptic-scale forcing is likely a contributing factor. Other processes, such as low-level instability and lower upstream precipitation efficiency may also play a role.

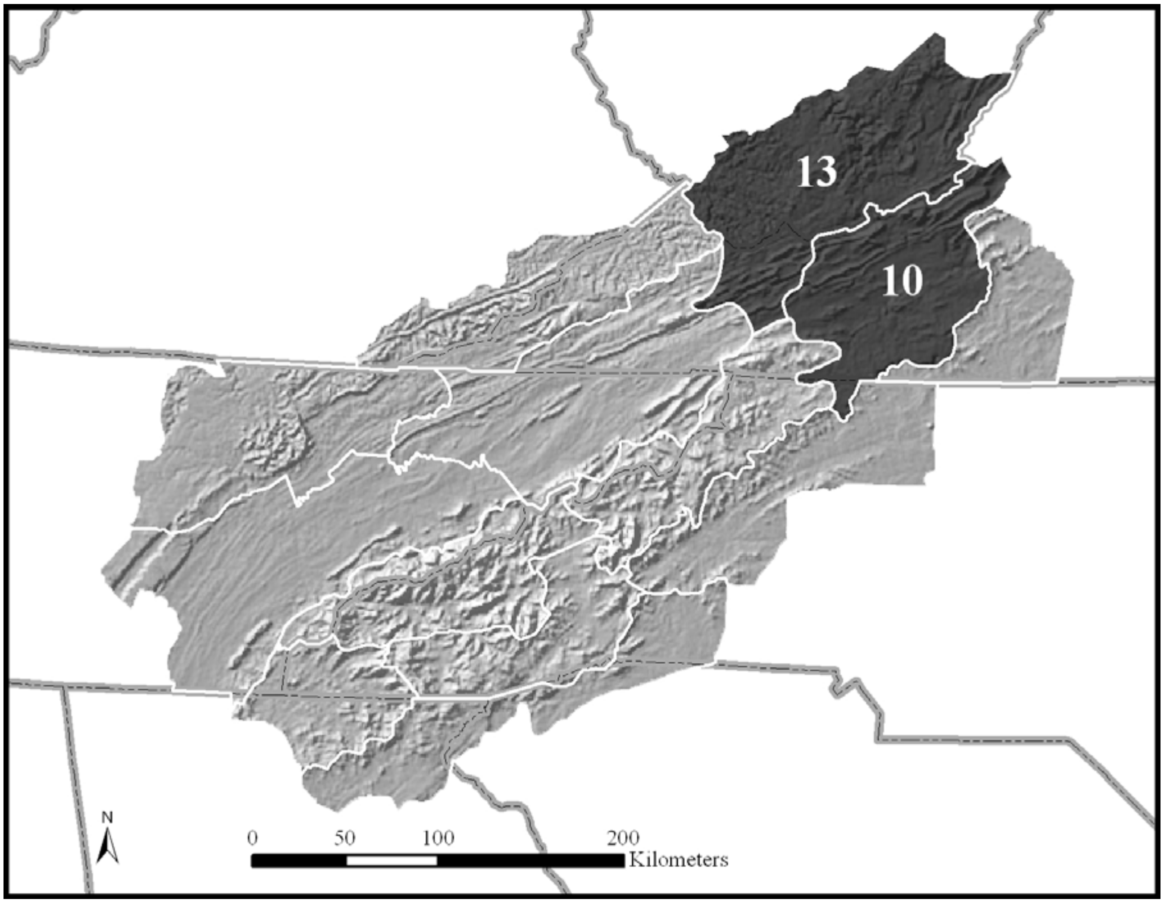

Figure 2.9. Snow regions used for spillover vs. no spillover comparison.

Two snow regions were selected to assess the spillover effects (Fig. 2.9). The windward slope snow region of the Northern Plateau (Region 13) was compared with the New River Valley (Region 10). This pair was selected due to their adjacency and similar elevations, so as to minimize the confounding effects of elevation and latitude in evaluating differences in snowfall accumulations. The ratio of mean snowfall between the snow regions (Region 13:10) was used to compare the spillover effects. A low value of this ratio indicates greater spillover (e.g. higher snowfall totals in the New River Valley compared to other events), whereas a high value translates into less spillover, with snowfall largely limited to the windward slopes. Comparisons in synoptic fields and map patterns were made between events in the bottom quartile (exceptional spillover) and top three quartiles (limited spillover) of the mean snowfall ratio.

2.4.5. Trajectory Analyses

In order to assess the influence of different air trajectories on snowfall patterns across the Southern Appalachians, a backward air trajectory analysis was undertaken for a sample of NWFS events. Previous work by Schmidlin (1992) found that the Great Lakes significantly enhanced NWFS in the northern mountains of West Virginia, leaving open the possibility of a Great Lakes influence on snowfall patterns farther south. A backward air trajectory consists of planimetric (two-dimensional) and vertical plots of the hourly values of latitude, longitude, and height (in hPa) for an air trajectory ending at a specified point and time. The NOAA Hysplit Trajectory Tool (http://www.arl.noaa.gov/ready/hysplit4.html) was utilized to calculate 72-hour backward air trajectories for air parcels at 1450 m, or approximately the 850 hPa level. The Hysplit Trajectory Tool uses u and v components of the horizontal wind, temperature, height, and pressure at different levels of the atmosphere to calculate the backward trajectories (Draxler and Hess 1998).

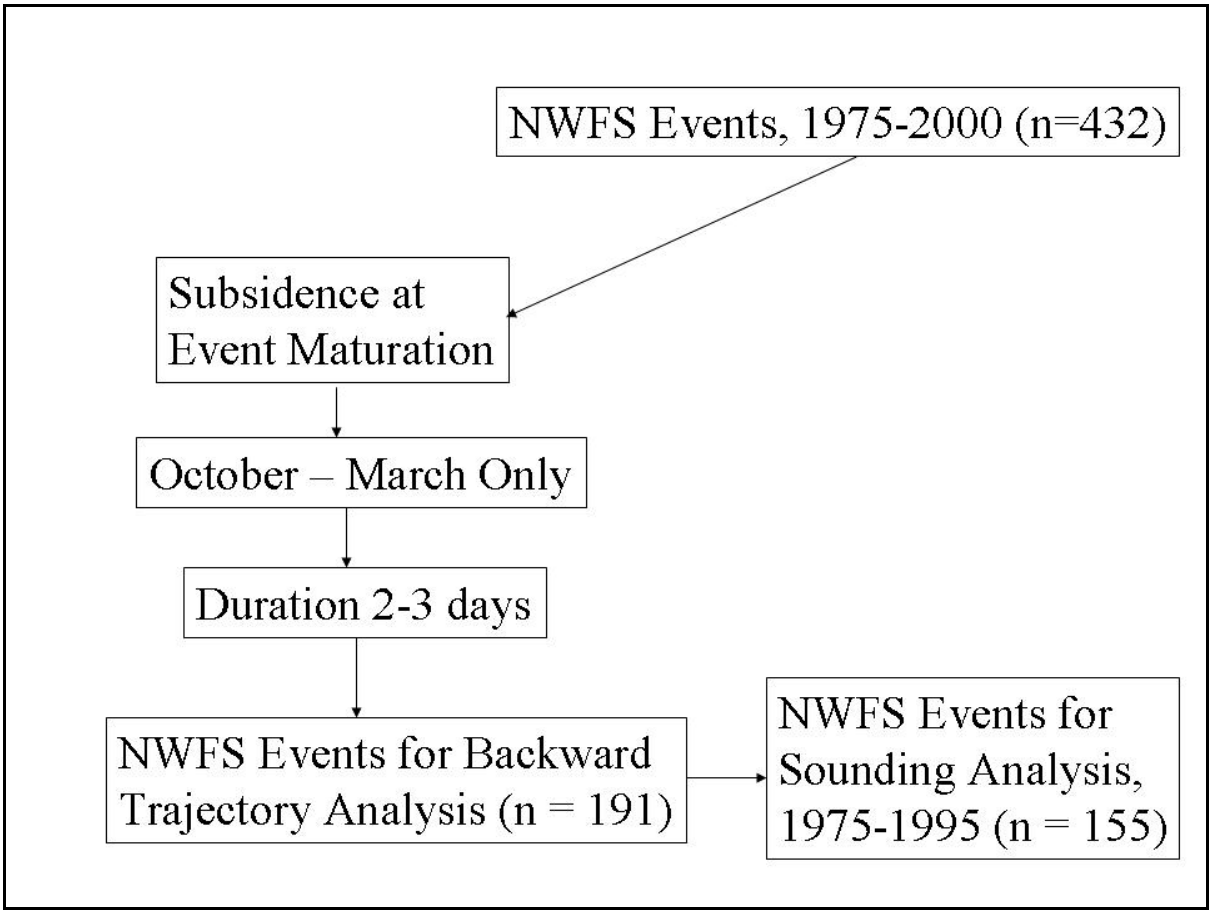

Figure 2.10. Selection of sample of NWFS events for the trajectory analysis.

The original sample of 859 NWFS events for the period 1950 to 2000 was narrowed for this analysis by identifying only those events between 1975 and 2000 with 1) synoptic-scale sinking motions (e.g. 700 hPa vertical velocity > 0) at event maturation and 2) durations between two and three days during the period (Fig. 2.10). A sample encompassing the entire period of analysis was not pursued so that trajectories with a Great Lakes connection could be tied to daily ice cover data in digital form available only after 1973. Events with rising motions were eliminated because a portion of the precipitation may be tied to synoptic features (i.e. approaching troughs) and may have confounded possible orographic and Great Lakes' influences. Exceptionally short and long events were also eliminated from the sample in order to control for the influences of event duration (e.g. exceptionally heavy snowfall totals resulting from an extended period of light snowfall). Also, 40 events in the months of April and May were excluded due to the complicating effects of diurnal patterns of convection, which are present in many of these events. Although low-level convection is an important lifting mechanism for NWFS throughout the year, its role is particularly significant and more closely related to diurnal patterns of surface heating in the spring. Therefore, it was necessary to exclude those events. The remaining 191 NWFS events served as the sample for the backward air trajectory analyses.

The weighted spatial mean or centroid of each event and the event maturation time were used as the starting point and time for the backward air trajectory analyses. A visual analysis of the computed trajectory allowed for classification of each six-hour segment by counting the number of segments that fell within each quadrangle from a trajectory classification grid (Fig. 2.11). The Great Lakes (grids 3.1 and 3.2) were roughly divided into the western lakes of Superior and Michigan and the eastern lakes of Huron and Erie. The Lake Ontario region was not explicitly delineated in this study due to the relatively infrequent occurrence of north-northeasterly flow from this region. Since the western and eastern Great Lakes sub-regions were the same spatial dimensions, it was possible to make direct comparisons between the two. The limited spatial resolution of the NCEP reanalysis data used to compute the backward air trajectories also supported using a more generalized classification grid, and not focusing on the precise lake-water boundary. Additionally, the concept of the lake-aggregate effect as recognized by Sousounis and Mann (2000), suggested that the combined influence of the Great Lakes moistens the lower troposphere and lowers surface pressures across the Great Lakes region, and not just over or immediately downwind of individual lakes.

The hourly latitude, longitude, and height (hPa) coordinates associated with each event backward air trajectory were used to develop composite trajectories for each region in the trajectory classification scheme. Additional composite trajectories were developed for T-12 (12 hours prior to event maturation) and T+12 (12 hours following event maturation). For each event with a GLC ($\geq$ 6 hours of the trajectory in trajectory grids 3.1 or 3.2) the Great Lakes Ice Atlas DVD (Assel 2003) was consulted to determine the percent ice cover for the western and eastern Great Lakes. The NCEP Reanalysis data (Kalnay et al. 1996) were also used to develop composite plots of 1000 hPa height for each trajectory class. Composite mean and maximum snowfall totals, composite backward air trajectories for each region in the Southern Appalachians, and the values of various synoptic fields were also calculated for each trajectory class and comparisons were made across trajectory classes.

Specific comparisons were made between northwesterly trajectories (e.g. those that extend backwards to the northwest, as opposed to the southwest or northeast) that exhibited a Great Lakes Connection (GLC) versus those that did not. Additionally, events with a GLC in which either the High Country (Region 8) or Northern Plateau (Region 13) received significant snowfall relative to the other region were investigated further to determine if part of the enhanced snowfall could be attributed to favorable air trajectories. Snowfall totals were compared between the High Country (Region 8) and Northern Plateau (Region 13) to identify events in which snowfall in one of these regions may have been enhanced due to more favorable lake trajectories. The top quartile of the ratio of the mean snowfall between the High Country and Northern Plateau (Region 8: Region 13) served to identify events with a GLC in which the High Country received greater snowfall, while the bottom quartile served to identify events in which the Northern Plateau received the heavier snowfall. Interestingly, the bottom quartile of this ratio was associated no measurable snowfall across the High Country.

2.4.6. Vertical Analyses of Moisture and Temperature

In order to provide more detailed information on the vertical characteristics of temperature and moisture associated with NWFS, an analysis of vertical soundings for Huntington, WV (HTS) (Fig. 2.11) was undertaken for a smaller sample of 155 of the original 191 events during the period 1975-1995. This detailed vertical analysis of sounding data complemented the fields obtained from the NCEP reanalysis data, which were quite useful for synoptic-scale analyses. HTS was chosen since it was the closest sounding station upwind of the study area and therefore provided the best assessment of the vertical profiles during periods of northwest flow. The full sample of 191 events from the trajectory analysis could not be used since the National Weather Service moved the HTS upper air station to Blacksburg (RNK), Virginia, in late 1995. The sounding time closest to event maturation served as the maturation, or T0 sounding, whereas the soundings 12 hours prior to and after the T0 sounding served as T-12 and T+12. Since the sounding data were only available every twelve hours except in special situations, the sounding data may be off as much as six hours from the calculated event maturation time. The sounding data may not be representative of the synoptic environment associated with NWFS in southern portions of the study area, particularly when the flow has more of a southwesterly or westerly component. Nonetheless, HTS represents the best available station for the Southern Appalachians as a whole.

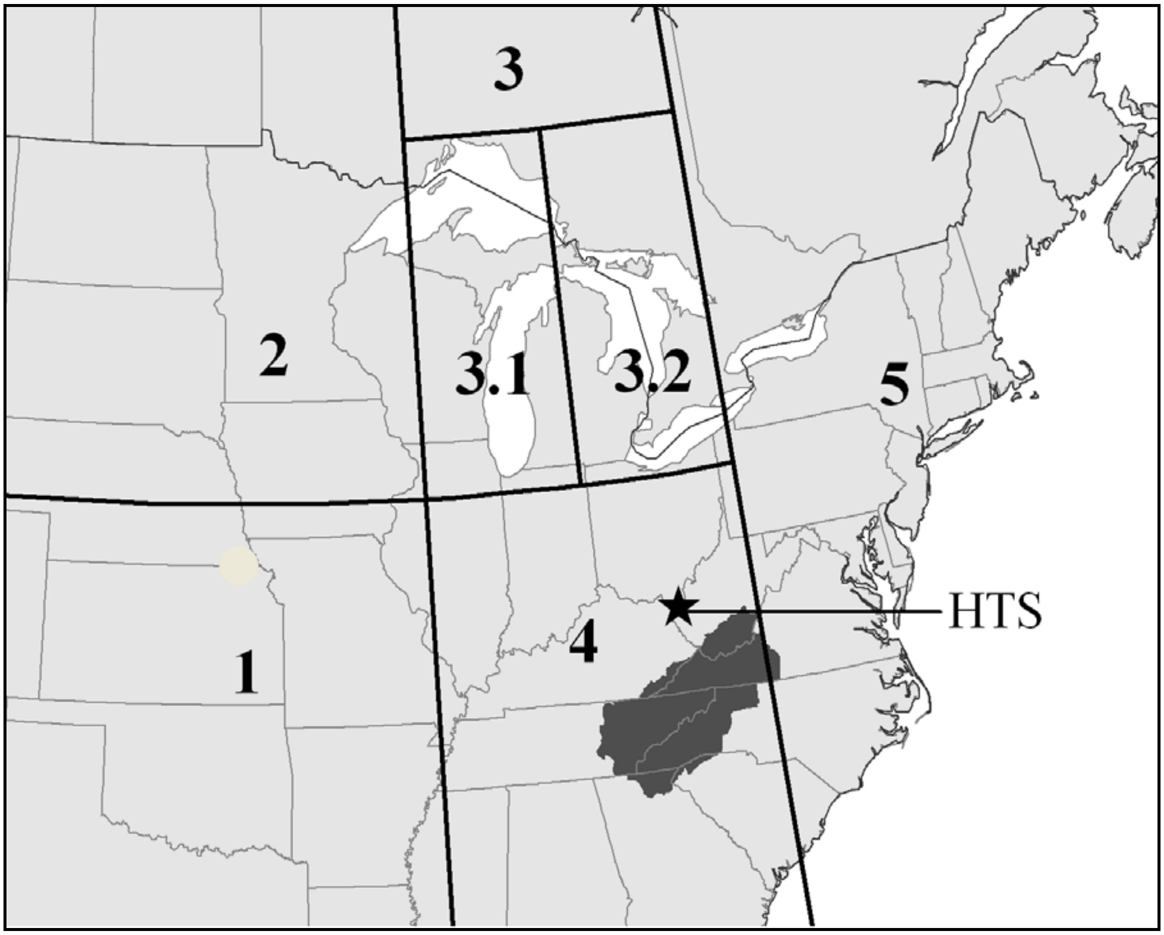

Figure 2.11. Trajectory classification grid and location of Huntington (HTS).

Table 2.3 summarizes the sounding variables that were calculated from the analysis. The moist layer was defined as that portion of the lower troposphere with mean relative humidity greater than or equal to 80 percent. The mean temperature of the moist layer was calculated by averaging the temperatures at the lower and upper heights of the moist layer. The presence or absence of moderately cold temperatures (-14˚ to -17˚ C) in the moist layer was also noted as ice crystal growth is maximized in the dendritic range of -14˚ to -17˚ C (Ryan et al. 1976, Pruppacher and Klett 1997, Fukuta and Takahashi 1999); furthermore, heavy snowfall is often associated with moist layer temperatures within the dendritic temperature range (Auer and White 1982). Additionally, forecasters at the Greer, SC, NWS Office have found that heavy NWFS events are more likely to occur in the North Carolina mountains when moist layer temperatures coincide with the dendritic temperature range (Lee 2005). Values for relative humidity, temperature, and height were taken from the University of Wyoming archived text soundings (http://weather.uwyo.edu/upperair/sounding.html).

Table 2.3. Sounding variables analyzed in this study. Note that these variables were calculated at maturation (T0), T-12, and T+12.

Sounding Variable	Units
Thickness of Moist Layer	hPa
Mean Temperature of Moist Layer	ºC
Top of Moist Layer	hPa
Dendritic Temperature Range	-14 to -17 ºC (0 = No, 1 = Yes)
Lapse Rate Above Moist Layer	ºC/next 100 hPa

The sounding variables calculated in this study provided a detailed profile of temperature and moisture characteristics unavailable from the NCEP reanalysis data, and therefore served to complement the synoptic-scale perspective taken thus far. Therefore, comparisons of the sounding variables were also made on the basis of trajectory class and snowfall intensity. For the air trajectories, sounding variables were compared across all trajectory classes. Additionally, comparisons were made between trajectories with a Great Lakes connection and those without one. The level of statistical significance between the mean values associated with each trajectory class was assessed using a two-sample difference of means t-test. Likewise, sounding variables were compared between light and heavy events for five snow regions on the windward slopes or at higher elevations **(**Fig. 2.7**)**. This comparison used the same approach taken in the analysis of snowfall intensity for the entire sample of NWFS events, with light events defined at the bottom three quartiles of the mean snowfall of those stations reporting snow in each snow region, and the heavy events the top quartile.

CHAPTER III
RESULTS AND DISCUSSION

3.1. Introduction

This chapter begins with a general discussion of the snowfall climatology of the Southern Appalachians. In this section and those that follow, emphasis is placed on the spatial variations of snowfall and synoptic fields by snow region within the Southern Appalachians. The third section focuses on differences in synoptic patterns and synoptic field values by snowfall intensity (i.e. between light and heavy events). Analyses are carried out for all of the NWFS events, those characterized by synoptic-scale ascent or synoptic-scale subsidence at event maturation, and a sample of events with information on upstream antecedent air trajectories and vertical profiles of moisture and temperature. The fourth section discusses the influence of elevation on NWFS, paying particular attention to the comparisons between events with a greater difference in snowfall by elevation and those with less of a difference. In the fifth section of this chapter, comparisons are made between events that "spill over" from the Northern Plateau (Region 13) into the New River Valley (Region 10) and those that are largely confined to the Northern Plateau. These spillover events are a particular challenge to forecasters due to the infrequency of accumulating NWFS along leeward slopes and other areas well downwind from windward slopes. The sixth, and final section, discusses the results of the antecedent upstream air trajectory analysis, with specific reference to trajectory classification, development of composite trajectory plots, mapping of composite snowfall totals by trajectory class, and comparisons of composite sounding and synoptic fields by trajectory class. The section concludes with an overview of the influence of the Great Lakes on NWFS events that display a northwesterly antecedent trajectory (i.e. a trajectory that extends to the northwest).

3.2. NWFS Climatology

Mean annual snowfall totals for the period 1950 to 2000 range from less than 40 cm (16 in) across the southern, lower elevation snow regions, to nearly 160 cm (63 in) in the High Peaks (Region 14) (Fig. 3.1). However, a shorter, 15-year climatology (1991-2005) using all available data from Mt. Leconte (1,979 m or 6,493 ft), yields a mean annual snowfall total of 262 cm (103 in). Much of the mountainous terrain above 600 m (1,968 ft) averages between 50 and 127 cm (20 to 50 in). The percent of mean annual snowfall tied to NWFS ranges from 25 to 30 percent in the Southern, Central, and Northern Foothills (Regions 3, 9, 11) to as much as 56 percent in the Central Plateau (Region 12) (Fig. 3.1). In general, the highest percentages are found along the higher, but not highest, elevation windward slopes, with the lowest percentages along the lower elevation leeward slopes. Interestingly, the percent of mean annual snowfall tied to NWFS at the High Peaks (Region 14) is somewhat less (48 percent) than the Central and Northern Plateau (56 and 52 percent). This reflects the importance of southeast and southwest flow in contributing significantly to snowfall on the High Peaks, which lowers the percentage tied to NWFS. Farther north and west (e.g. Central and Northern Plateau), snowfall in association with southerly and southeasterly flow is less significant due to shadowing effects from higher terrain to the south and east.

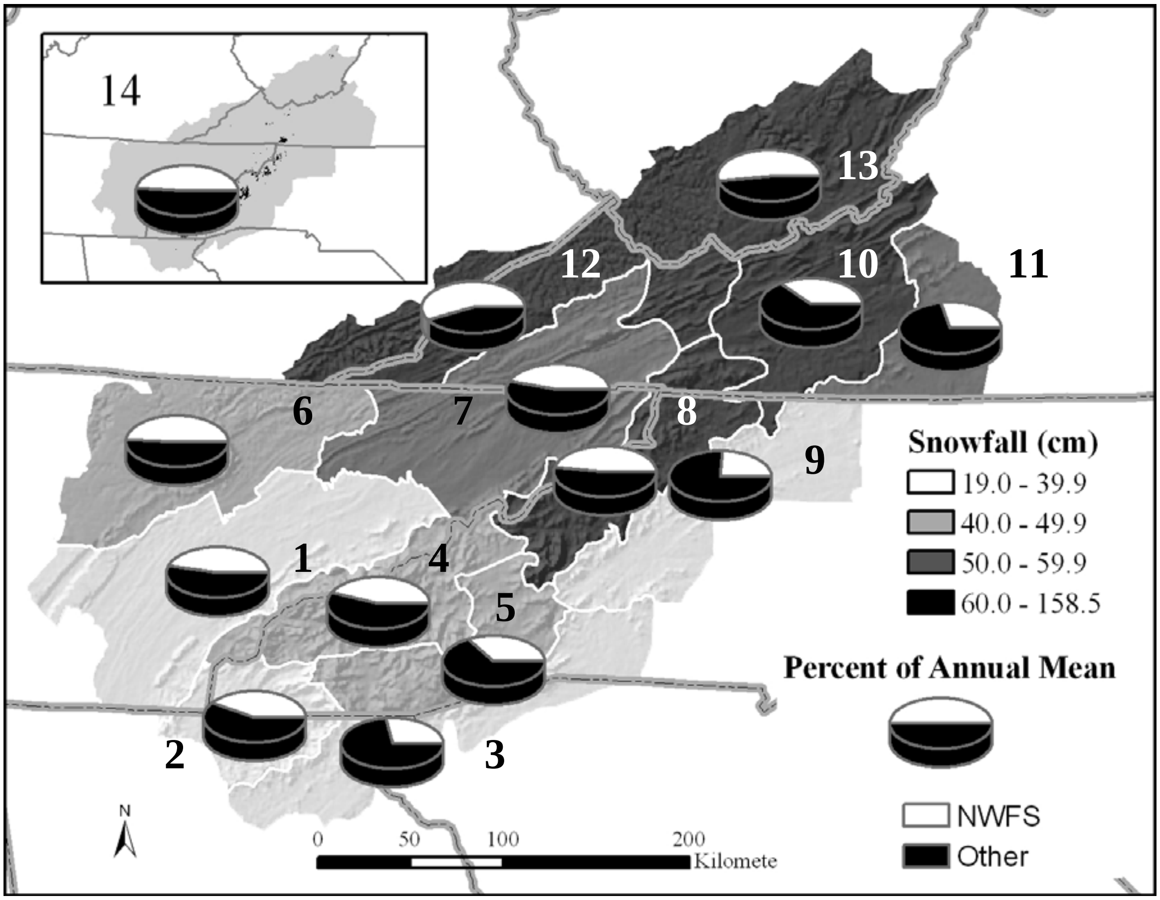

Figure 3.1. Mean annual snowfall and percent of mean annual snow tied to low-level northwest flow by snow region.

The spatial patterns of mean annual NWFS are similar to overall mean snowfall, although a more pronounced longitudinal gradient is evident (Fig. 3.2). The western snow regions generally receive more mean annual NWFS than the eastern snow regions at the same latitude. The Southern and Central Foothills (Regions 3 and 9) average very little NWFS, typically less than 10 cm (4 in) per year. The higher elevation windward slopes (Regions 8, 12, 13, and 14), on the other hand, receive considerably more, averaging 51 to 76 cm (20 to 30 in) per year. The percent of mean annual NWFS tied to 700 hPa synoptic-scale subsidence also varies considerably across the study area (Fig. 3.2), with the higher elevation windward slopes exhibiting the highest percentages and the lower elevation leeward slopes (e.g. Regions 3, 9, and 11) displaying the lowest. In the Northern Plateau (Region 13), approximately 61 percent of the mean annual NWFS is tied to synoptic-scale subsidence. These results indicate the importance of topography in controlling the spatial patterns of NWFS. Orographic lifting is of fundamental importance in periods of low-level northwest flow due to the nearly perpendicular nature of the flow to the southwest-to-northeast trending topography. When synoptic-scale forcing is lacking, topography often provides enough of a lifting mechanism for snowfall development at higher elevations and along windward slopes. At lower elevations and along leeward slopes, however, accumulating snowfall does not develop as frequently due to an absence of orographic lifting or downslope flow.

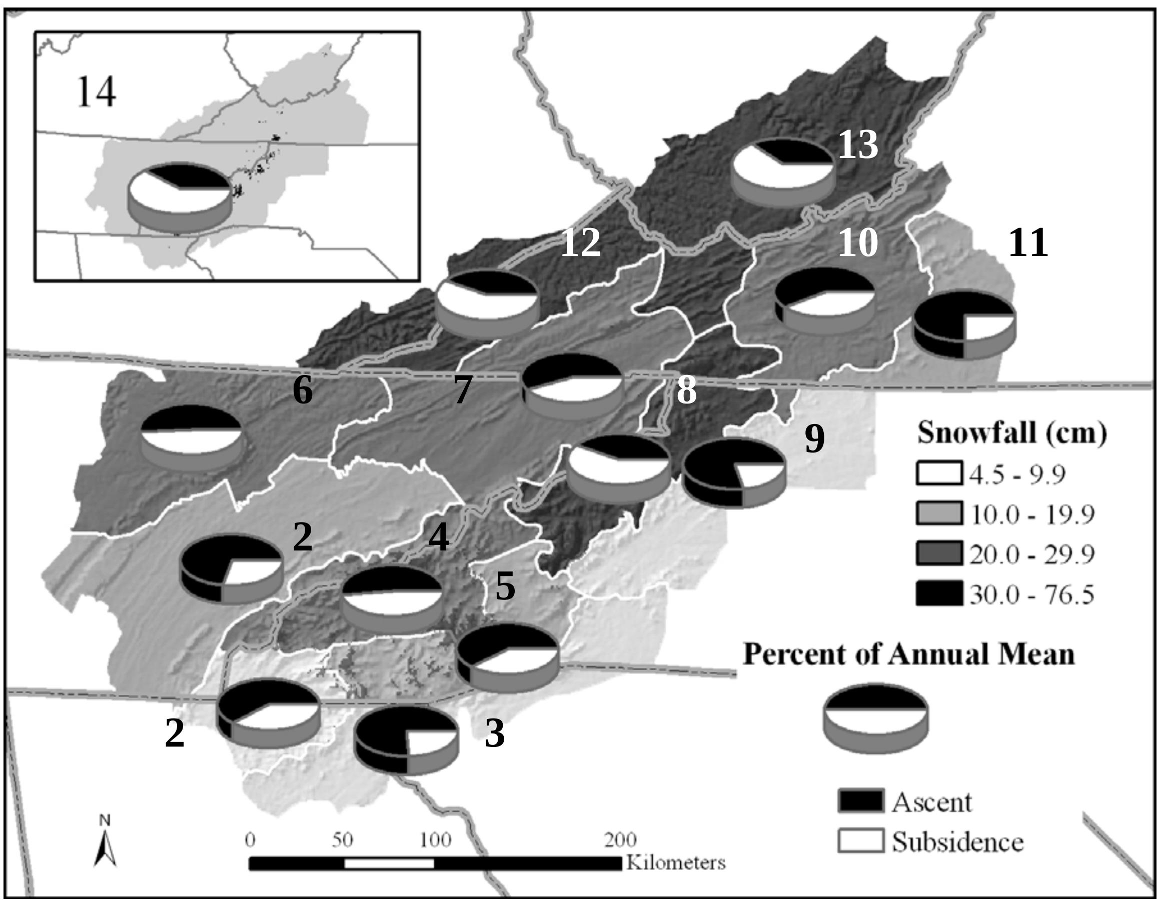

Figure 3.2. Mean annual NWFS and percent tied to synoptic-scale subsidence and ascent by snow region.

The mean number of NWFS events per year is highly correlated with mean annual NWFS (Fig. 3.3), with the higher elevation windward slopes averaging between 8 and 15 events per year. Likewise, the Southern and Central Foothills (Regions 3 and 9) average only approximately one event per year. The monthly pattern of total NWFS events (Fig. 3.3) shows a clear maximum in all snow regions during January, the coldest month of the year. The months of December, February, and March see relatively equal numbers of events across most snow regions. The vast majority of the events occur between December and March, although it is not uncommon see NWFS at the higher elevations in the transition months of October, November, April, and May. The highest mean number of NWFS events is found in the Northern Plateau (Region 13), where an average of one event every ten days is observed between the months of December and March. The Northern Plateau also averages approximately one event in November and April, and NWFS can be expected one out of three years in October and even less in May.

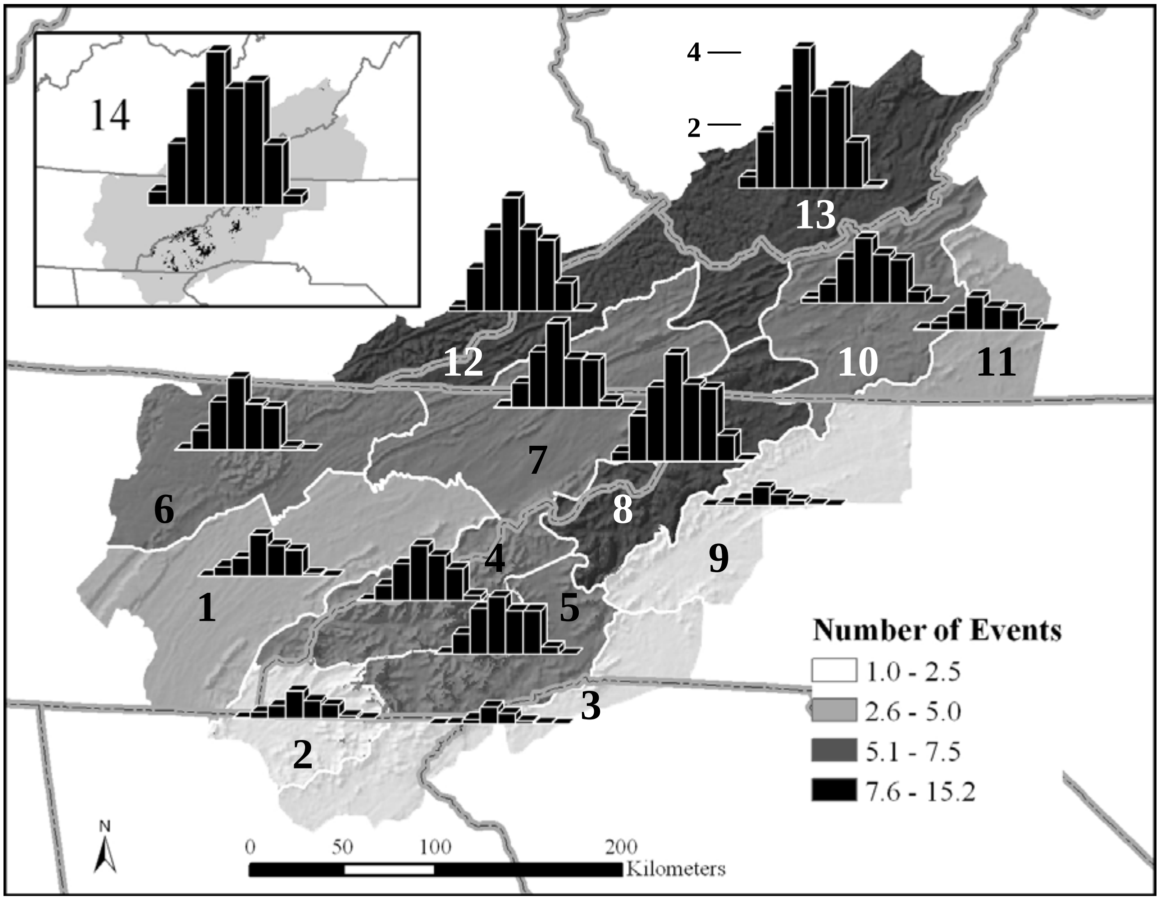

Figure 3.3. Mean number of NWFS events by year (shaded) and month (bar graph, October to May, left to right).

A different measure of NWFS is the mean accumulation by event (Fig. 3.4), which shows a very different pattern than mean annual NWFS or mean annual NWFS events. The highest mean event NWFS values are found in the Central and Northern Foothills (Regions 9 and 11), while the Southern Blue Ridge (Region 5) experiences the lowest mean event snowfall. Although NWFS events are infrequent in the Foothills' snow regions, when they do occur, it is due to strong synoptic-scale forcing which results in heavier snowfall. In fact, six of a sample of 12 events in which mean snowfall was greater than 7.6 cm (3.0 in) were tied to closed surface cyclones (eight of the Miller Type A variety). Even though the Southern Blue Ridge averages a substantially higher number of NWFS events, most of the events are exceedingly light. This apparently results from having a moderately high mean elevation of 1,012 m (2,570 ft), which is sufficient for the development of snowfall due to the orographic lifting, even in the absence of synoptic-scale ascent. However, in periods of low-level northwest flow, moisture is precipitated out on the windward slopes and higher elevations of the Great Smoky Mountains and Balsam Mountains, which are located immediately to the northwest. Some of the High Peaks in the Southern Appalachians are located in these two ranges, with several peaks over 2,000 m (6,561 ft). Therefore, the Southern Blue Ridge is significantly shadowed despite its relatively high elevation. Heavy NWFS can certainly occur in the Southern Blue Ridge, but the majority of the events are characterized by much lighter accumulations, thereby contributing to the lowest mean event NWFS total in the entire study area.

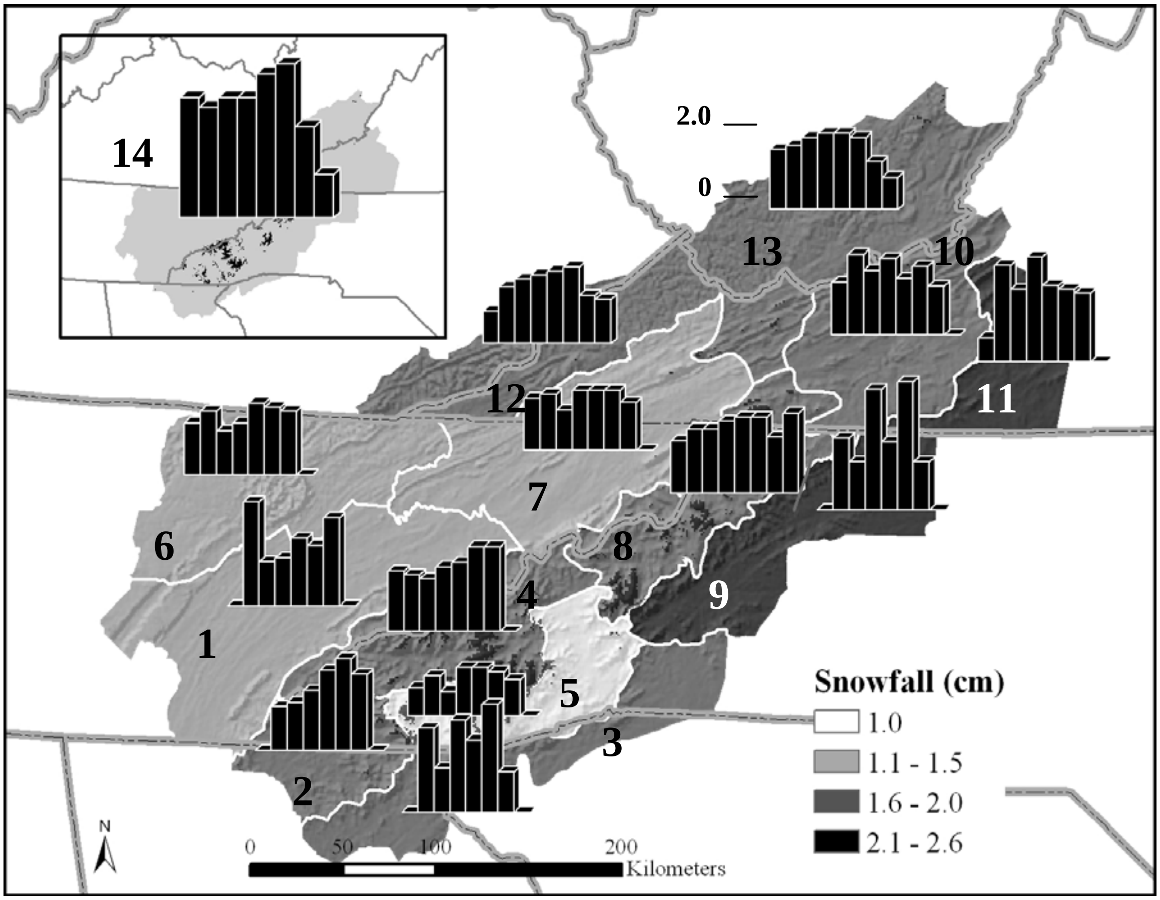

Figure 3.4. Mean NWFS accumulation by event overall (shaded) and by month (bar graph, October to May, left to right).

The mean NWFS accumulation by event varies somewhat by month (Fig. 3.4), particularly at the lower elevations and leeward slopes where the small sample size introduces greater variability. The highest mean event snowfall totals generally occur during the months of January, February, and March. Interestingly, some of the highest mean event snowfall totals at the higher elevations and windward slopes occur during March. This is particularly the case for the High Peaks (Region 14), where an average of 8.1 cm (3.2 in) falls in each event, approximately 2.5 cm (1.0 in) greater than the average for the other months. One possible explanation for this pattern is that lower tropospheric instability is quite high in the spring due to the higher sun angle. This, in turn, results in heavier snow, particularly higher up where elevated heat sources increase the low-level instability. In March, ground and surface temperatures at the High Peaks are also still consistently cold enough to readily accumulate snow, but by April and May this is no longer the case.

3.3. Snowfall Intensity

3.3.1. All Events

In order to develop a better understanding of the differences between ordinary and exceptional NWFS events, composite synoptic fields and map patterns were compared across two groups of events for the five snow regions representing windward slopes and higher elevations (Regions 4, 8, 12, 13, 14 – Fig. 2.7). The ordinary, or light, events consisted of the bottom three quartiles of the mean snowfall in each of the snow regions, whereas the exceptional, or heavy, events were defined as the top quartile. Table 3.1 summarizes the composite synoptic fields for the High Country (Region 8). Rather than include tables for each of the other snow regions, the general patterns are discussed relative to the values presented in Table 3.1 for the High Country. Event duration, lower and middle tropospheric relative humidity, 500 hPa height, 700 hPa vertical velocity, and 200 hPa divergence display highly significant differences (relative difference > |0.50|) between light and heavy events in the High Country and across the other snow regions. The values for 1000 hPa height (highly correlated with surface pressure) directly over the mean center of each event are also significantly different between light and heavy events across the different snow regions, with even greater differences noted along the South Carolina coast (Fig. 3.5). The heavy events are tied to a stronger surface cyclone, most likely of the Miller Type "A" variety, displaced farther to the southwest. As a result, lower mean heights are observed across the entire southeastern U.S.

Heavy events are tied to long durations across all regions. Heavy events are also consistently tied to higher values of lower and middle tropospheric relative humidity in all regions. The 1000 to 500 hPa mean relative humidity is approximately 10 percent higher in the heavy events in the five snow regions analyzed, yielding a relative difference of 0.72 for the High Country (Region 8) and 0.84 in the Northern Plateau (Region 13). The composite map patterns also show significantly higher 1000 to 500 hPa mean relative humidity across the southeastern U.S. for the heavy events, with the greatest differences found over central North and South Carolina (Fig. 3.6). Relative humidity at 500 hPa also displays highly significant differences, with the heavy events approximately 10 percent higher. Interestingly, little difference in the moisture fields of precipitable water or 850 hPa mixing ratio are evident, signifying very little difference in the amount of water vapor between light and heavy events. Rather, lower tropospheric temperatures are slightly colder, yielding a lower dewpoint depression and hence higher relative humidity.

Heavy events are also characterized by significantly lower 500 hPa heights across all snow regions. In the High Country (Region 8), the relative difference is -0.54, and this value climbs to -0.84 in the High Peaks. Mid and upper-level synoptic-scale dynamics also play an important role in differentiating between light and heavy events. Heavy events are tied to much lower values of 700 hPa vertical velocity (i.e. weaker synoptic-scale subsidence) and higher values of 200 hPa divergence. The composite map patterns of 700 hPa vertical velocity are particularly striking, as strong mid-level synoptic-scale subsidence is evident to the west in the Mississippi River Valley and synoptic-scale ascent is present off the southeastern U.S. coast for both the light and heavy events (Fig. 3.7). However, the heavy events are tied to much stronger synoptic-scale rising motions (e.g. more negative values), with the greatest mean differences just to the east of the study area in central Virginia and North Carolina. Events with lower values of 700 hPa vertical velocity over and just to the east of the Southern Appalachians favor heavier snowfall due to the combination of orographic lift and synoptic-scale ascent. Conversely, lighter events are characterized by stronger synoptic-scale subsidence; thus lift is limited to orographic and convective mechanisms.

In the High Country (Region 8) and Northern Plateau (Region 13), 850 hPa thermal advection also displays highly significant differences between light and heavy events. Light events are tied to much lower values of thermal advection (i.e. strong cold air advection), whereas heavy events exhibit more modest values (i.e. weak cold air advection). Cold air advection often occurs in association with subsidence, and therefore periods of strong cold air advection, in particular, are tied to synoptic-scale subsidence (Table 3.1), limiting snowfall. It is rather striking, however, that the relative differences for 850 hPa thermal advection are much greater for the High Country and Northern Plateau. This may be explained in two ways. First, substantial elevated terrain is needed to provide sufficient topographic forcing in the absence of strong synoptic-scale support. In many cases, these events are dominated by strong cold air advection in association with synoptic-scale subsidence. At the lower elevations, orographic effects are insufficient to produce accumulating snowfall, whereas at the higher elevations light amounts are observed. A second explanation is that in some events rain may change briefly to snow as strong cold air advection cools the higher elevations just enough behind a departing synoptic-scale disturbance. In this case, lower elevations remain too warm and are unable to accumulate snow. Examination of a sample of ten NWFS events with strong cold air advection (less than -15 ˚C/12 hr) confirms these explanations. Six of the events were tied to very strong cold fronts in the absence of a surface cyclone or trough, followed by low-level northwest flow and no discernible synoptic-scale disturbance. The other four were associated with low-level northwest flow in the wake of a strong synoptic-scale disturbance, suggesting that just enough overlap occurred between the departing synoptic-scale ascent and the arriving cold air in the higher elevations.

In the High Peaks (Region 14), heavy events are tied to 850 hPa relative humidity values 8.1 percent higher (relative difference of 0.60) than in the light events. The high relative difference of this field stands in contrast to other snow regions, where the relative differences are considerably less. Only in the High Country (Region 8), is the relative difference somewhat comparable (0.47). This finding may relate to the importance of higher 850 hPa relative humidity in generating supersaturated conditions in the mountaintop feeder clouds in association with the seeder-feeder effect. Likewise, events with lower 850 hPa relative humidity are less likely to develop a significant mountaintop feeder cloud, which may reduce the orographic enhancement at the higher elevations.

Table 3.1. Mean synoptic field values and differences for light versus heavy NWFS events in the High Country (Region 8). Duration variables include Days (the number of days snowfall is reported at coop stations), Beginning-Maturation (hours), Maturation-Ending (hours), and Beginning-Ending (hours). Note that stars indicate that the differences are statistically significant at the $p < 0.05$ level.

High Country (Region 8)	Light	Heavy	Abs_Dif	Rel_Dif	T-Score	
All Events	*n=431*	*n=143*				
Days	2.1	2.8	0.7	0.85	9.54	*
Beginning-Ending	25	40	14	0.88	8.86	*
1000-500 hPa Mean Relative Humidity	58.7	67.6	8.9	0.72	7.48	*
850 hPa Thermal Advection	-5.9	-3.5	2.4	0.57	6.87	*
Maturation-Ending	14	23	9	0.72	6.76	*
200 hPa Divergence	2.6	3.7	1.1	0.62	6.53	*
Beginning-Maturation	11	17	6	0.64	6.04	*
500 hPa Relative Humidity	34.5	44.1	9.6	0.56	5.51	*
850 hPa Relative Humidity	75.5	81.6	6.1	0.47	5.46	*
500 hPa Height	5429	5385	-45	-0.54	-5.41	*
700 hPa Vertical Velocity	12.3	1.4	-10.9	-0.57	-5.09	*
1000 hPa Height	129	106	-24	-0.45	-4.71	*
850 hPa Wind Direction	301.7	297.1	-4.6	-0.24	-2.90	*
500 hPa Vorticity	4.3	4.8	0.6	0.25	2.64	*
1000-500 hPa Thickness	5300	5279	-21	-0.26	-2.57	*
850 hPa Wind Speed	12.9	11.9	-1.0	-0.25	-2.55	*
850-500 hPa Temperature Change	18.4	19.1	0.7	0.18	1.93	*
Precipitable Water	0.31	0.33	0.02	0.19	1.83	*
850 hPa Temperature	-6.8	-7.5	-0.7	-0.16	-1.58	
850 hPa Theta-e	11.8	11.2	-0.6	-0.08	-0.82	
850 hPa Mixing Ratio	2.2	2.3	0.1	0.07	0.75	
500 hPa Vorticity Advection	1.2	1.3	0.1	0.02	0.18	

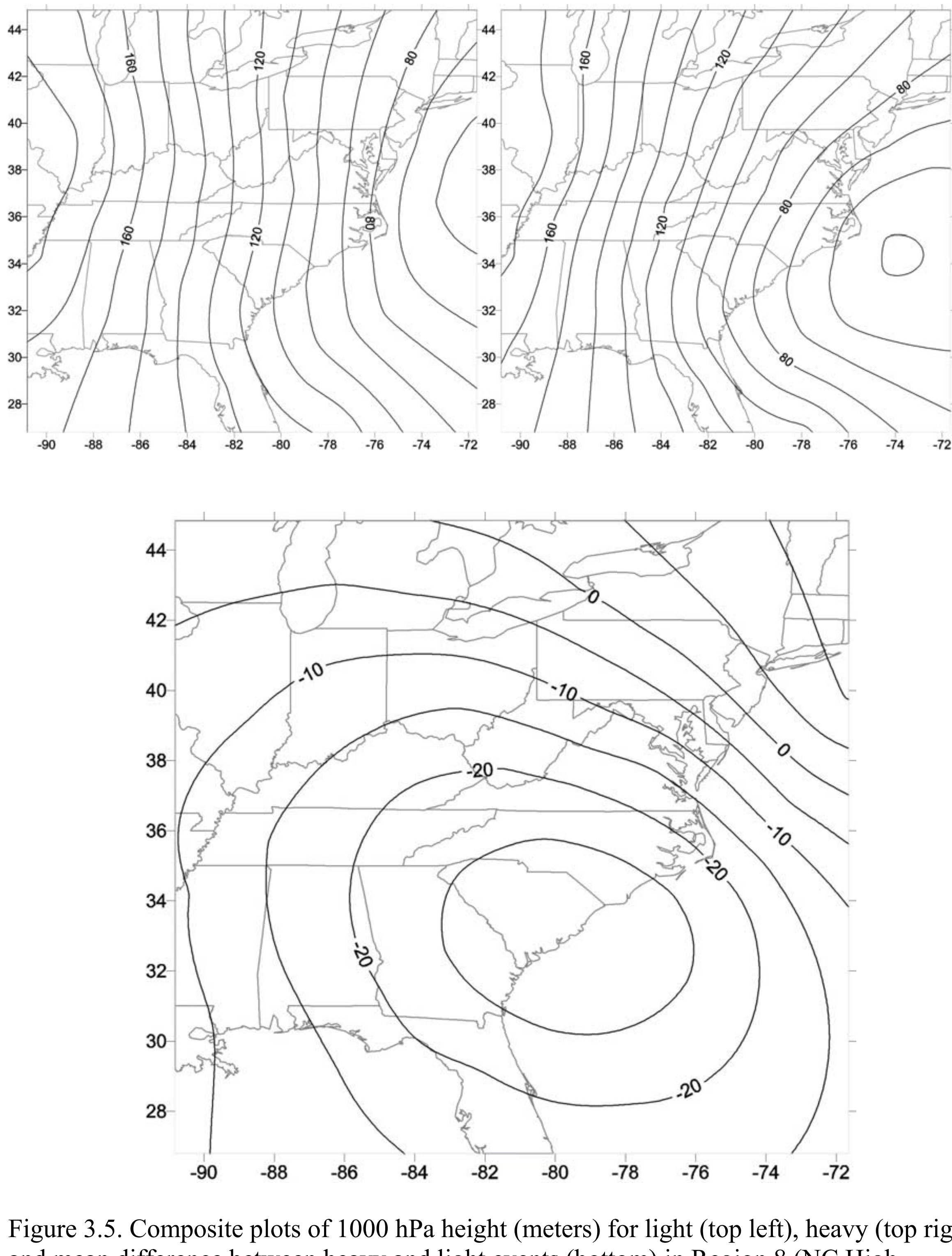

Figure 3.5. Composite plots of 1000 hPa height (meters) for light (top left), heavy (top right), and mean difference between heavy and light events (bottom) in Region 8 (NC High Country).

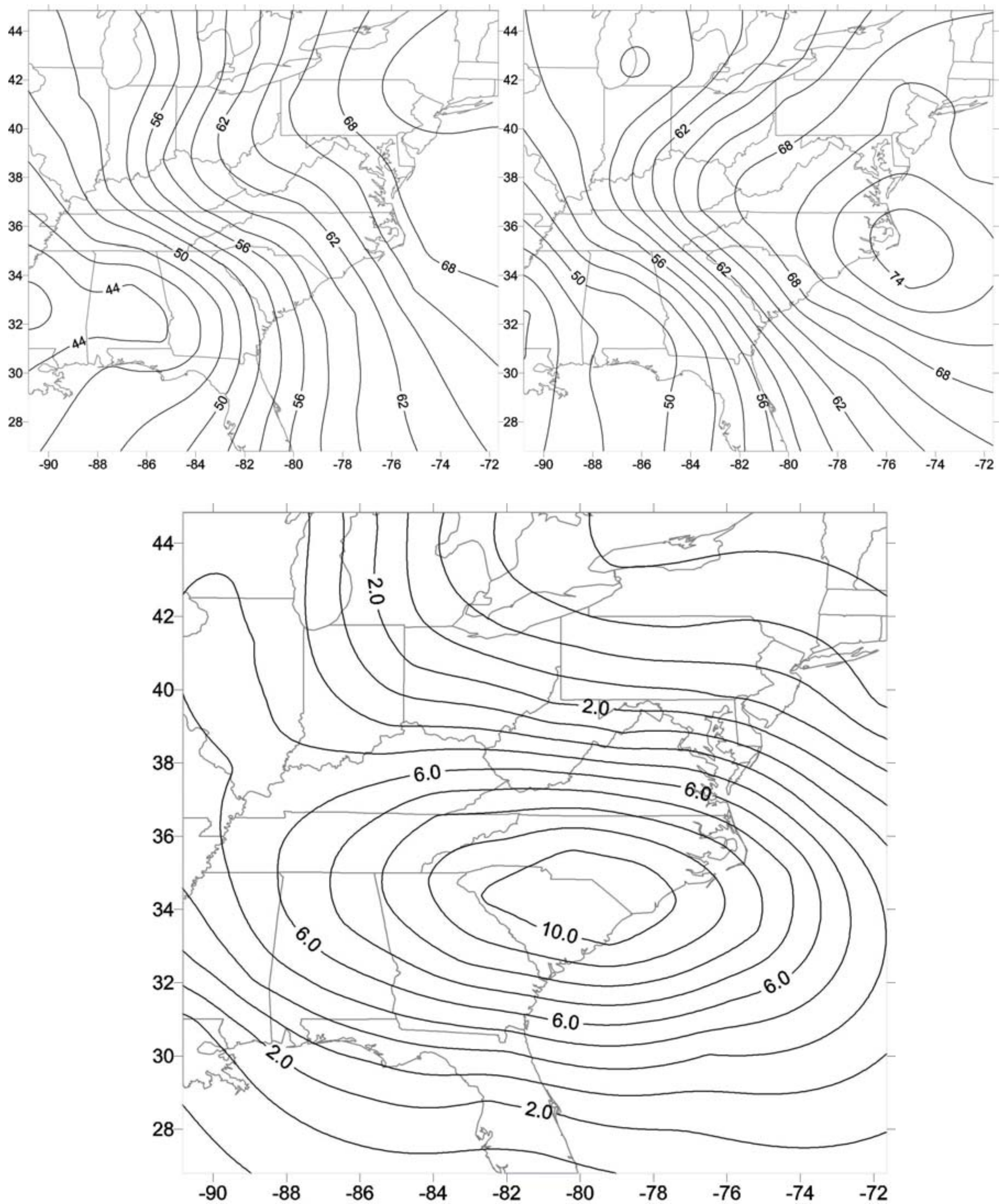

Figure 3.6. Same as Fig. 3.5, except for 1000 to 500 hPa mean relative humidity.

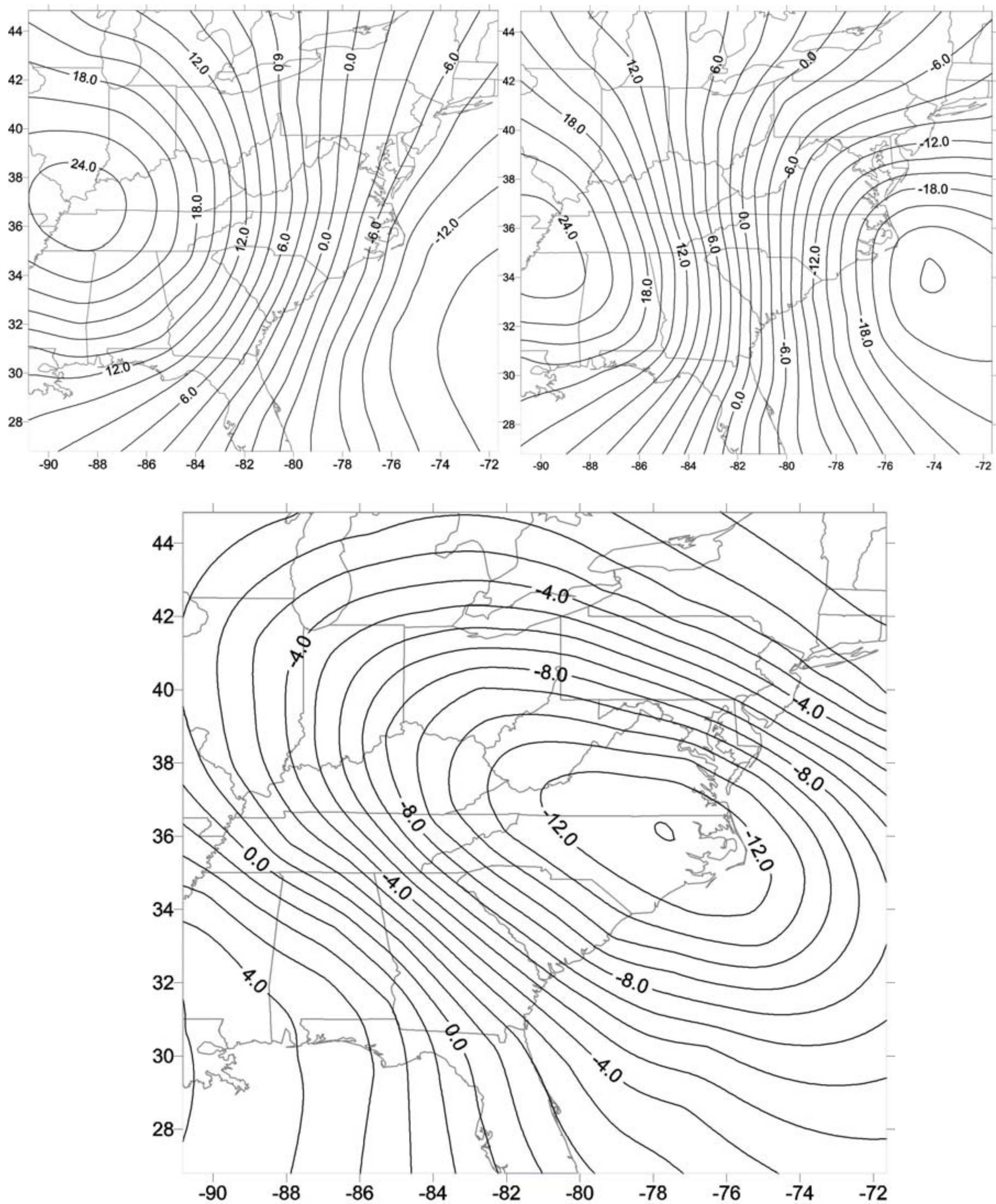

Figure 3.7. Same as Fig. 3.5, except for 700 hPa vertical velocity.

3.3.2. Synoptic-Scale Ascent

Due to the important role of synoptic-scale vertical velocity in influencing snowfall intensity (see section 3.3.1), light and heavy events were also compared just for those events exhibiting synoptic-scale ascent (e.g. 700 hPa vertical velocity < 0) at event maturation in each of the five snow regions representing the windward slopes and higher elevations (Regions 4, 8, 12, 13, 14). The purpose of this analysis was to better understand the differences between light and heavy NWFS events in the presence of synoptic-scale ascent. The composite 1000 hPa height pattern (i.e. surface pressure pattern) clearly shows that heavy events are tied to a deeper surface cyclone off of the Carolina coast (Fig. 3.8). The greatest differences in 1000 hPa height are found off of the South Carolina and Georgia coasts, and extend across the entire southeastern U.S. The location of the surface low indicates an East Coast storm track for both the light and heavy events based on the composite maps. However, the heavy events are not only associated with a stronger (deeper) cyclone, but also one that tracks closer to the Southern Appalachians, such that the regions of low-level northwest flow and synoptic-scale rising motions show greater superpositioning. A cyclone track closer to the Southern Appalachians also places more of the region within the favored location for heavy snowfall relative to the surface cyclone track, as identified by Goree and Younkin (1966).

Event duration, 1000 to 500 hPa mean relative humidity, 500 hPa relative humidity, 850 hPa thermal advection, 500 hPa height, and 700 hPa vertical velocity all exhibit highly significant differences (relative difference > |0.50|) between light and heavy events with synoptic-scale ascent (Table 3.2). Heavy events are almost twice as long from the beginning of the event to the end compared with light events in the High Country (Region 8), producing a relative difference of greater than 1.00 in four out of the five snow regions. Lower tropospheric relative humidity is also highly significant, with heavier events averaging nearly 9 percent higher 1000 to 500 hPa mean relative humidity. All snow regions exhibit much higher values of relative humidity at 500 hPa for the heavy events. Differences are greatest in the High Country and Central Plateau, where values are nearly 15 percent higher (relative differences of 0.80). The composite map patterns are particularly revealing, indicating higher 500 hPa relative humidity across the entire southeastern and mid-Atlantic portions of the U.S. with the heavy events, and a pronounced maximum over the center of the study area (Fig. 3.9). This higher 500 hPa relative humidity is tied to 500 hPa air trajectories extending to the Pacific Ocean in a sub-sample of seven heavy events, suggesting that some of the increase in mid-level moisture may be of Pacific origin.

Measures of mid- and upper-level dynamics remain important in differentiating between light and heavy events across all five snow regions, even when those events exhibiting synoptic-scale subsidence are eliminated from the sample. Heavy events are tied to more negative values of 700 hPa vertical velocity (i.e. stronger rising motions) and higher values of 200 hPa divergence. In fact, the absolute value of the relative difference for 700 hPa vertical velocity between light and heavy events in this sample (-0.71) is somewhat higher than in the overall sample (-0.57). Lower 500 hPa heights are also tied to heavy events in all snow regions. Lastly, 850 hPa thermal advection shows highly significant differences in all snow regions except the Great Smoky Mountains. Examination of a sample of seven events indicates that some are tied to a very strong frontal passage in the absence of a strong synoptic-scale disturbance, while others are associated with the passage of a strong synoptic-scale cyclone through the Ohio Valley. Interestingly, previous work (e.g. Knappenberger and

Michaels 1993) has found that an Ohio Valley storm track in general does not favor heavy snowfall in the Appalachians or mid-Atlantic regions. This may result from most of the precipitation falling as rain in the warm sector before strong cold air advection commences and produces a period of NWFS.

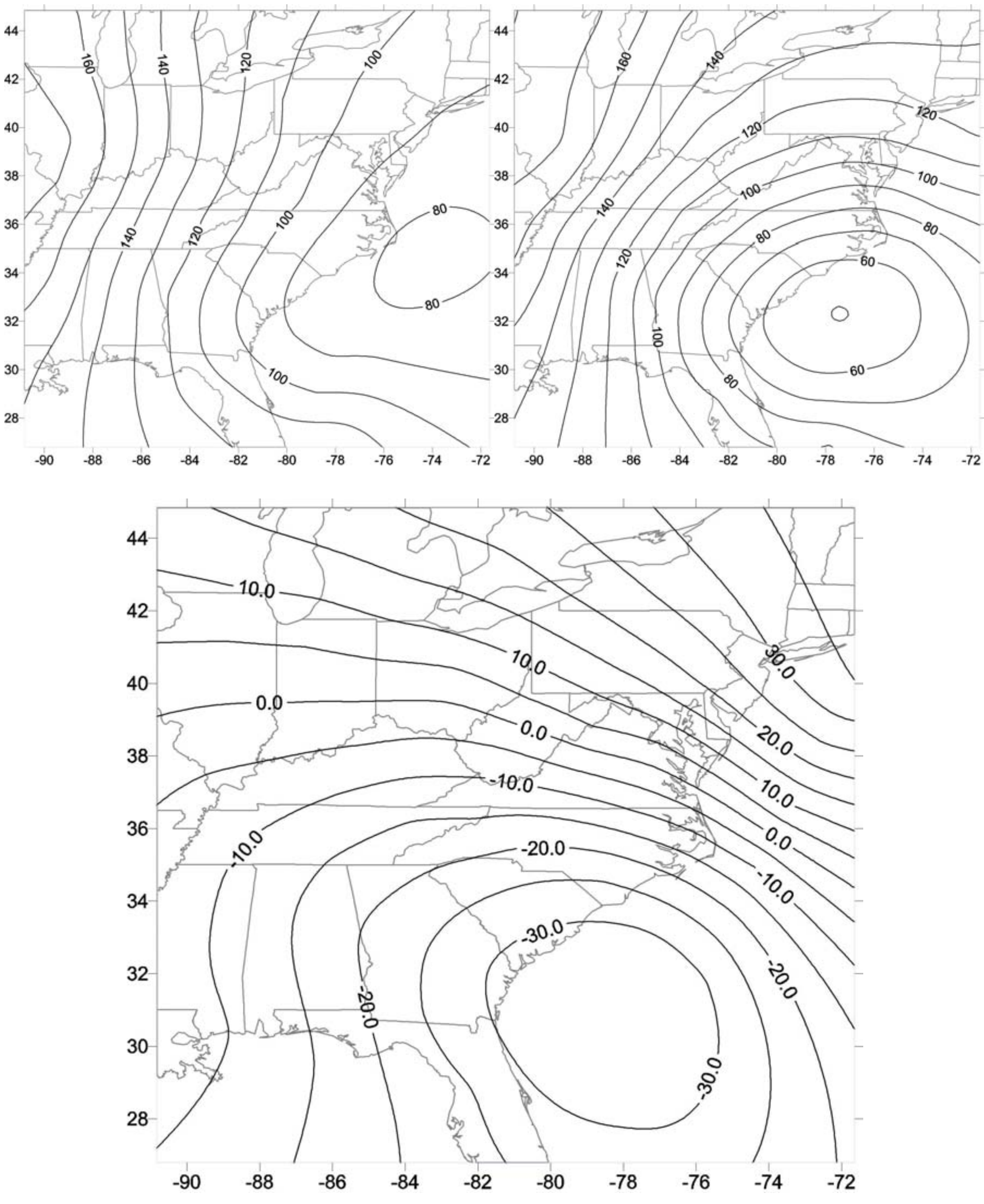

Figure 3.8. Same as Fig. 3.5, except for 1000 hPa height for events with synoptic-scale ascent.

Table 3.2. Same as Table 3.1, except for events in which synoptic-scale ascent is observed.

High Country (Region 8)	Light	Heavy	Abs_Dif	Rel_Dif	T-Score	
Synoptic-Scale Ascent	*n=120*	*n=39*				
Beginning-Ending	24	44	20	1.22	6.93	*
Days	2.2	3.0	0.8	0.92	5.69	*
Beginning-Maturation	9	17	8	1.01	4.75	*
1000-500 hPa Mean Relative Humidity	69.0	77.5	8.5	0.76	4.73	*
850 hPa Thermal Advection	-5.0	-1.8	3.2	0.7	4.69	*
500 hPa Relative Humidity	46.6	61.0	14.5	0.8	4.68	*
Maturation-Ending	15	26	12	0.9	4.66	*
200 hPa Divergence	3.3	4.4	1.1	0.55	3.12	*
500 hPa Height	5451	5403	-47	-0.54	-3.11	*
700 hPa Vertical Velocity	-11.9	-21.5	-9.7	-0.71	-2.88	*
850 hPa Relative Humidity	79.4	84.7	5.3	0.43	2.76	*
1000-500 hPa Thickness	5339	5308	-31	-0.4	-2.24	*
500 hPa Vorticity	3.3	4.2	0.9	0.36	2.17	*
850 hPa Temperature	-4.5	-6.1	-1.5	-0.37	-2.02	*
850 hPa Wind Speed	10.7	9.5	-1.2	-0.29	-1.64	
1000 hPa Height	111	95	-16	-0.32	-1.59	
850 hPa Theta-e	15.6	13.7	-1.9	-0.29	-1.54	
850 hPa Mixing Ratio	2.8	2.7	-0.1	-0.12	-0.64	
500 hPa Vorticity Advection	4.6	4.3	-0.3	-0.08	-0.45	
850-500 hPa Temperature Change	19.1	19.4	0.3	0.07	0.37	
Precipitable Water	0.42	0.41	-0.01	-0.04	-0.23	
850 hPa Wind Direction	299.0	298.8	-0.2	-0.01	-0.05	

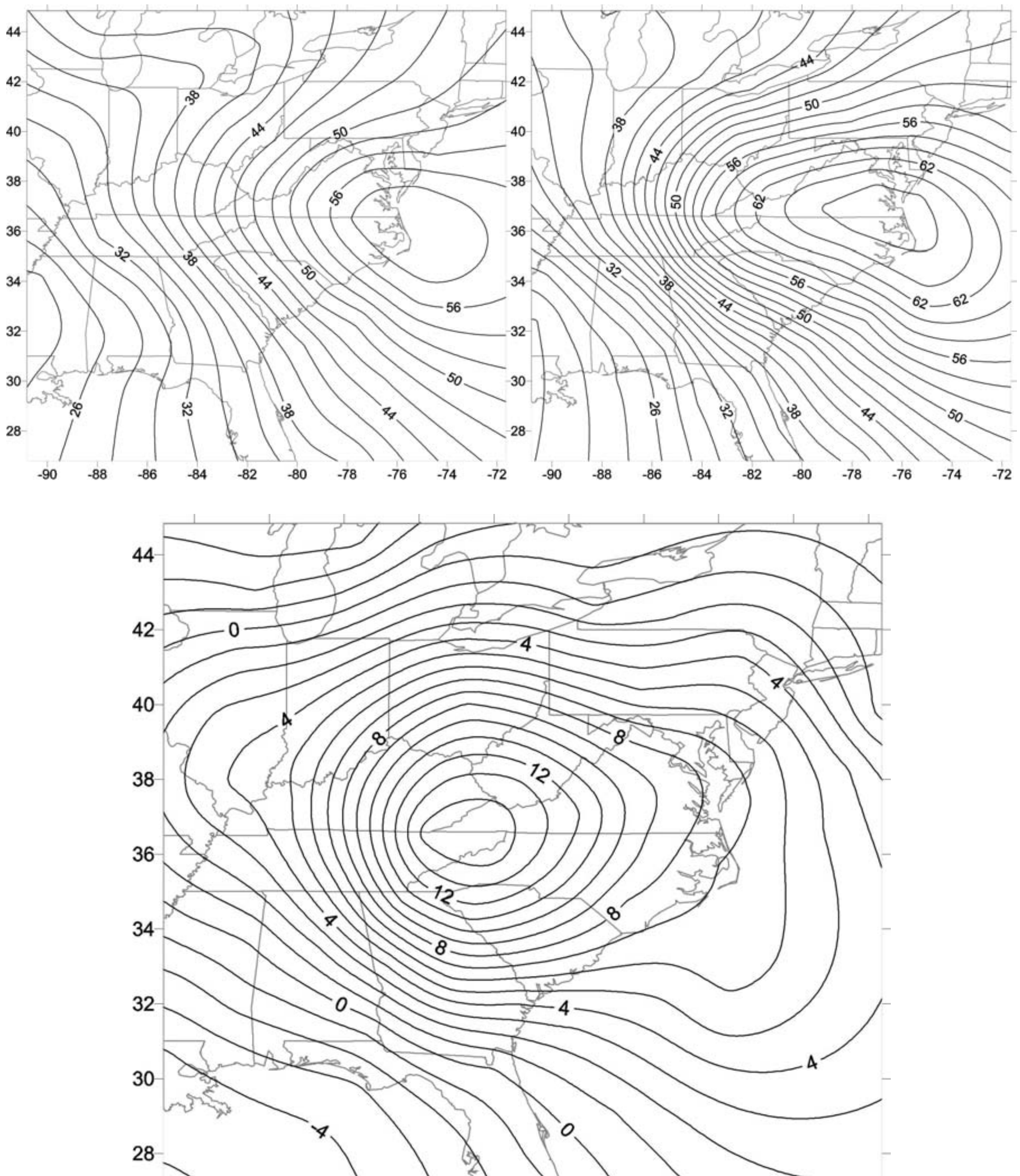

Figure 3.9. Same as Fig. 3.5, except for 500 hPa relative humidity for events with synoptic-scale ascent.

3.3.3. Synoptic-Scale Subsidence

In this analysis, comparisons were made between light and heavy snowfall for only those events exhibiting synoptic-scale subsidence at event maturation (700 hPa vertical velocity > 0). The majority of the NWFS events along the windward slopes and at higher elevations fall into this category, and therefore a comparison between light and heavy events was undertaken to better understand the synoptic patterns associated with exceptional (heavy) events. The surface pressure pattern associated with light and heavy events is almost identical (Fig. 3.10). A surface cyclone is evident off the mid-Atlantic coast, while a surface anticyclone lies to the west along the Mississippi River. The only difference between the two composite maps is the magnitude of the values over the mid-Atlantic, where heights are approximately 20 m lower, owing to a slightly stronger surface cyclone. These differences stand in contrast to those observed in section 3.3.2 (synoptic-scale ascent), in which the stronger composite surface cyclone was displaced farther to the south.

Event duration, 1000 to 500 hPa mean relative humidity, and 500 hPa height once again stand out as being highly significant (relative difference > |0.50|) between light and heavy events in the High Country (Region 8) (Table 3.3). Although the difference in 1000 to 500 hPa mean relative humidity is not quite as great as with the synoptic-scale ascent sample (6 percent or a relative difference of 0.56), the composite map patterns are very noteworthy (Fig. 3.11). In the light and heavy events, high relative humidity is found across the northeastern U.S. in conjunction with the surface cyclone, with progressively lower values extending to the southwest. In the heavy events, however, a tongue of higher relative humidity extends much farther to the west into the Ohio River Valley. The greatest differences in the composite map patterns are directly over and immediately to the northwest of the Southern Appalachians. The heavy events, therefore, are tied to significantly higher values of lower tropospheric relative humidity just upwind of the study area. As in the previous analyses (sections 3.3.1 and 3.3.2), there is little difference in the amount of water vapor between the light and heavy events; rather, the heavy events are tied to colder lower tropospheric temperatures that result in higher relative humidity.

The heavy events are also tied to substantially lower 1000 to 500 hPa thicknesses (relative difference -0.57) in the High Country. Temperatures at 850 hPa are also 2.1 ˚C colder (-9.3 ˚C, relative difference of -0.48), suggesting that moist layer temperatures in the level of maximum lifting between 800 and 700 hPa are more likely to be within the dendritic growth range of -14 ˚C to -17 ˚C. These results are therefore consistent with Auer and White (1982), in which they noted that heavy snowfall is typically tied to moist layer temperatures in the dendritic growth range. The other windward slopes and higher elevation snow regions display a similar pattern, with the exception of the Great Smoky Mountains (Region 4), where the only fields with relative differences greater than |0.50| are event duration, 1000 to 500 hPa mean relative humidity, and 500 hPa relative humidity. Portions of this region are considerably more shadowed than the other snow regions, so that may help to explain some of the differences, as might its more southerly latitude.

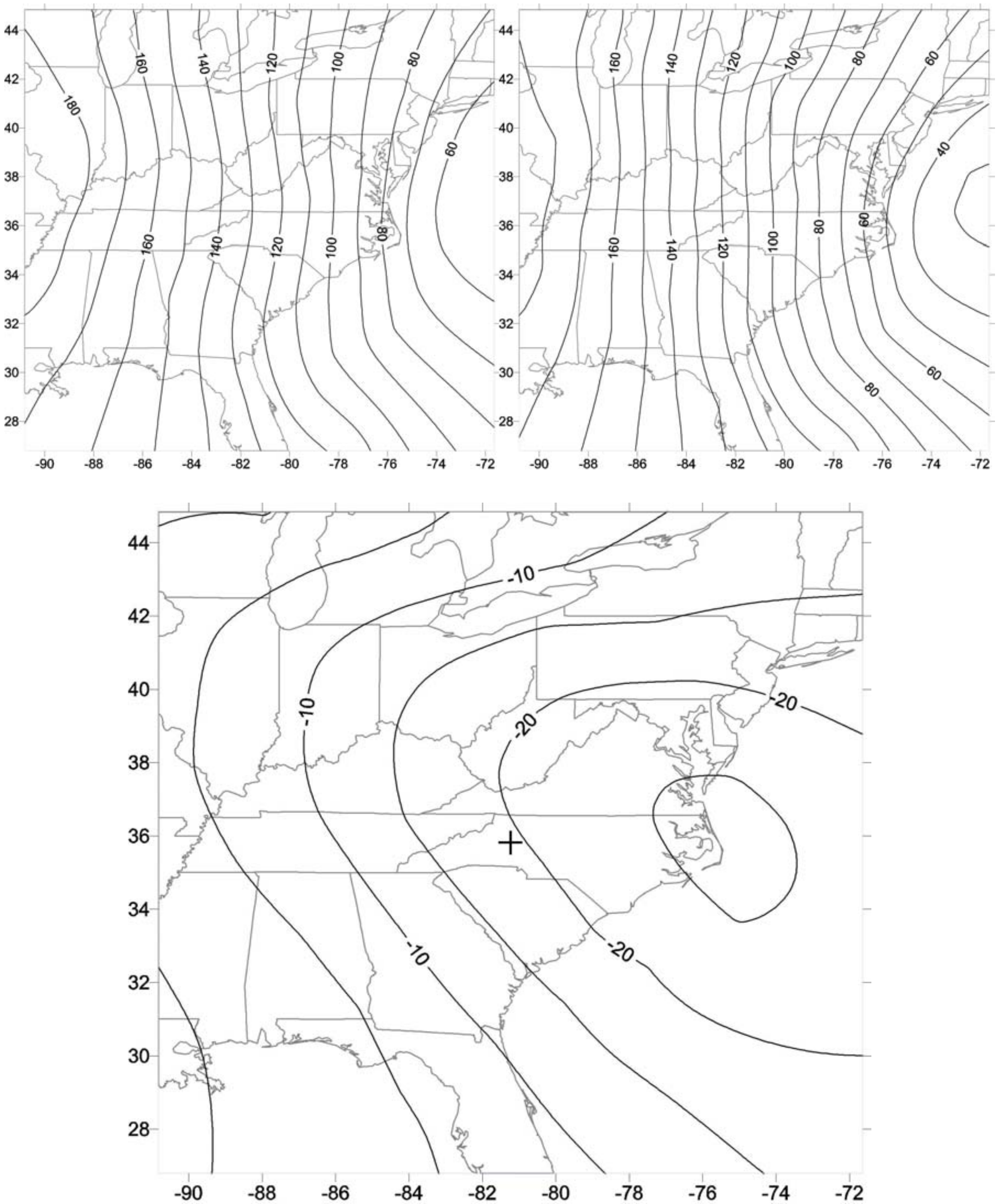

Figure 3.10. Same as Fig. 3.5, except for 1000 hPa height for events with synoptic-scale subsisdence.

Table 3.3. Same as Table 3.1, except for events in which synoptic-scale subsidence is observed.

High Country (Region 8)	Light	Heavy	Abs_Dif	Rel_Dif	T-Score	
Synoptic-Scale Subsidence	*n=313*	*n=102*				
Beginning-Ending	25	41	15	0.95	7.97	*
Days	2.1	2.8	0.7	0.87	7.87	*
500 hPa Height	5426	5363	-63	-0.8	-7.53	*
Maturation-Ending	14	23	9	0.82	6.57	*
Beginning-Maturation	12	17	6	0.65	5.1	*
1000-500 hPa Thickness	5292	5247	-45	-0.57	-5	*
1000-500 hPa Mean Relative Humidity	55.6	61.3	5.8	0.56	4.95	*
850 hPa Temperature	-7.3	-9.3	-2.1	-0.48	-4.19	*
850 hPa Relative Humidity	74.2	79.8	5.6	0.43	4.14	*
200 hPa Divergence	2.5	3.2	0.7	0.47	4.11	*
850 hPa Thermal Advection	-6.1	-4.4	1.8	0.43	3.95	*
500 hPa Vorticity	4.5	5.4	0.9	0.42	3.85	*
500 hPa Vorticity Advection	0.4	-1.1	-1.5	-0.43	-3.83	*
850 hPa Theta-e	10.9	8.3	-2.6	-0.41	-3.57	*
1000 hPa Height	134	116	-18	-0.36	-3.2	*
850 hPa Wind Direction	302.6	296.9	-5.8	-0.3	-3.15	*
500 hPa Relative Humidity	30.8	35.0	4.2	0.31	2.75	*
850 hPa Mixing Ratio	2.1	1.9	-0.2	-0.21	-1.92	*
Precipitable Water	0.28	0.26	-0.02	-0.17	-1.47	
850 hPa Wind Speed	13.4	13.8	0.4	0.1	0.89	
850-500 hPa Temperature Change	18.3	18.6	0.3	0.07	0.57	
700 hPa Vertical Velocity	18.7	18.8	0.2	0.02	0.14	

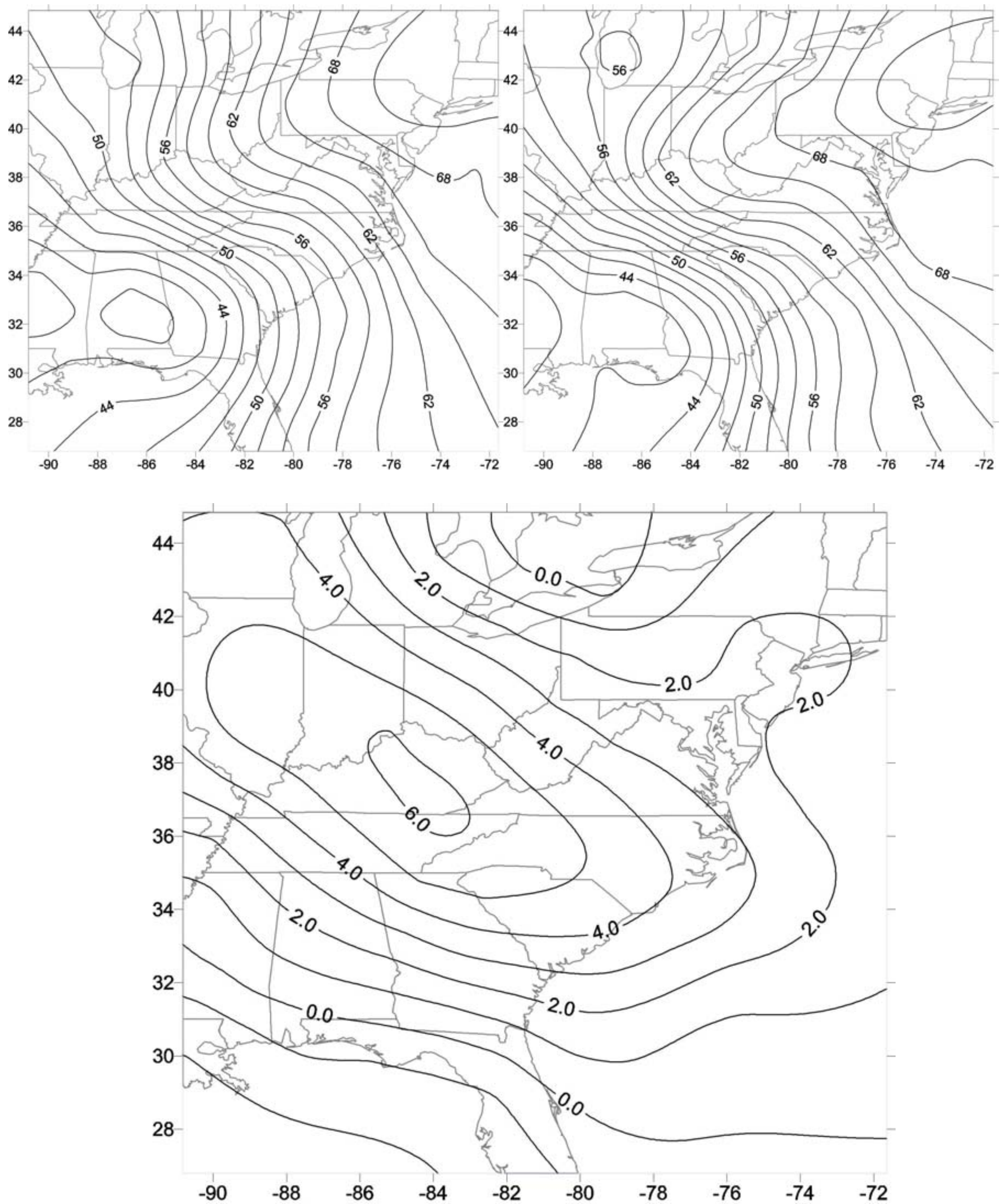

Figure 3.11. Same as Fig. 3.5, except for 1000 to 500 hPa mean relative humidity for events with synoptic-scale subsidence.

3.3.4. Synoptic-Scale Subsidence: Trajectory and Sounding Sample

This section compares antecedent upstream air trajectories and composite vertical profiles of moisture and temperature between light and heavy events. Data from the trajectory and sounding analyses used in this section, therefore, complement the synoptic-scale NCEP reanalysis data and provide a more detailed perspective on the synoptic climatology of NWFS. Trajectories were analyzed only at event maturation (T0), whereas vertical profiles of moisture and temperature were analyzed at three time periods: twelve hours prior to event maturation (T-12), event maturation (T0), and twelve hours after event maturation (T+12).

Many of the sounding variables at event maturation display highly significant differences between the light and heavy events (Table 3.4). The moist layer is nearly 100 hPa thicker in the heavy events, providing a deeper layer through which snowfall development can occur. Likewise, the height of the top of the moist layer is 61 hPa higher (645 hPa) for the heavy events at event maturation (T0). In the heavy events, however, the moist layer is considerably thicker, providing greater in-cloud residence times for developing snow crystals. It is noteworthy that these results are consistent with those of Auer and White (1982), in which they found the level of maximum lifting in heavy orographic snowfall to be at or slightly above mountaintop level.

Approximately 93 percent of the heavy events are characterized by moist layer temperatures within the dendritic temperature growth range (-14 ˚C to -17 ˚C), compared to only 66 percent of the light events at event maturation. Moist layer temperatures are also more likely to be in the dendritic temperature range in the heavy events at 12 hours before and after event maturation. Previous work has found that ice crystal growth is maximized in this temperature range (Fukuta and Takahashi 1999), so this finding is not unexpected. The mean temperature of the moist layer at event maturation, however, shows very little variation between light and heavy events, confirming the presence of a thicker moist layer in the heavy events that extends higher into the troposphere, where temperatures in the dendritic growth range are more likely to be present. Again, these results are consistent with those reported by Auer and White (1982), where they found that moist layer temperatures within the dendritic growth range are often tied to heavy orographic snowfall.

The lapse rate in the next 100 hPa above the moist layer also varies substantially, with much higher values at event maturation for the heavy events. The greater stability in this layer in the light events is often associated with a capping inversion just above the top of the moist layer that precludes additional vertical development of any mesoscale bands. In general, the significance of differences in sounding variables from T-12 or T+12 is considerably lower, suggesting that the vertical profiles of moisture and temperature at event maturation are most critical in differentiating between light and heavy events. Of the trajectory variables, only the percent of events with a Great Lakes Connection (GLC) shows a modest, though not significant ($p < 0.05$), increase between the light and heavy events. The number of hours air parcels spent in the vicinity of the Great Lakes is almost identical for the two event types, indicating that composite air trajectories for heavy events do not necessarily spend more time over the Great Lakes. The results of additional analyses comparing snowfall totals by trajectory class will be discussed in section 3.6 below.

Table 3.4. Mean sounding field values and differences for light versus heavy NWFS with synoptic-scale subsidence in the High Country (Region 8).

High Country (Region 8)	Light	Heavy	Abs_Dif	Rel_Dif	T-Score	
Trajectory and Sounding	*n=82*	*n=27*				
Thickness of Moist Layer (T0)	203	301	98.93	0.76	3.98	*
Dendritic Temperature Range (T0)	66%	93%	27%	0.60	3.68	*
Lapse Rate Above Moist Layer (T0)	-4.9	-7.8	-2.9	-0.69	-3.68	*
Dendritic Temperature Range (T-12)	29%	67%	37%	0.77	3.61	*
Top of Moist Layer (T0)	706	645	-61.03	-0.60	-2.81	*
Mean Temperature of Moist Layer (T-12)	-7.1	-11.6	-4.55	-0.55	-2.61	*
Thickness of Moist Layer (T-12)	180	259	79.29	0.51	2.22	*
Dendritic Temperature Range (T+12)	49%	70%	22%	0.43	2.08	*
Mean Temperature of Moist Layer (T+12)	-11.4	-13.6	-2.21	-0.40	-1.91	*
Great Lakes Connection	45%	59%	14%	0.28	1.29	
Thickness of Moist Layer (T+12)	115	142	26.29	0.26	1.13	
Top of Moist Layer (T+12)	783	760	-23.1	-0.20	-1.11	
Lapse Rate Above Moist Layer (T+12)	-1.3	-2.0	-0.67	-0.16	-0.71	
Mean Temperature of Moist Layer (T0)	-11.0	-11.6	-0.59	-0.12	-0.60	
Great Lake Hours	10.5	11.6	1.09	0.07	0.35	
Lapse Rate Above Moist Layer (T-12)	-5.6	-6.0	-0.33	-0.07	-0.32	
Top of Moist Layer (T-12)	636	626	-9.74	-0.05	-0.28	

3.3.5. Northwest Flow but No Snow

Periods of cold low-level northwest flow, characterized by 850 hPa temperatures below 0 ˚C, are common throughout the snow season. As discussed in the previous sections, accumulating snowfall is a frequent occurrence across the windward slopes and higher elevations. In some of these periods of northwest flow, however, accumulating snowfall does not occur. At issue is why accumulating snowfall occurs in some periods of northwest flow, but not in others. In this analysis, synoptic fields and composite map patterns are compared between periods of northwest flow in which no snow was reported and those in which light accumulations occurred. A sample of 66 events, with 33 representing episodes of northwest flow but no snow, and the other 33 representing NWFS events in the bottom quartile of mean snowfall, was selected for further analysis (see section 2.4.2 for more details).

Composite plots of 1000 hPa height show important differences between periods of no snow and light NWFS (Fig. 3.12). In the snow-free "events", lower heights are evident across the northeastern U.S., and the surface cyclone is displaced farther to the north. A surface anticyclone is anchored across the lower Mississippi River Valley, ridging into the Southern Appalachians. As a result, the composite map patterns reveal a more pronounced anticyclonic flow in the snow-free events. In the light NWFS events, a surface cyclone is present off the mid-Atlantic coast, while the surface anticyclone is in a similar position west of the Mississippi River. Heights across the Southern Appalachians and the entire southeastern U.S. are somewhat higher in the snow-free events, with the greatest difference off of the North Carolina coast. The 500 hPa height values are also significantly higher in the snow-free events, by 76 m (relative difference of -1.00), and are indicative of the higher 1000 hPa heights and warmer lower tropospheric temperatures.

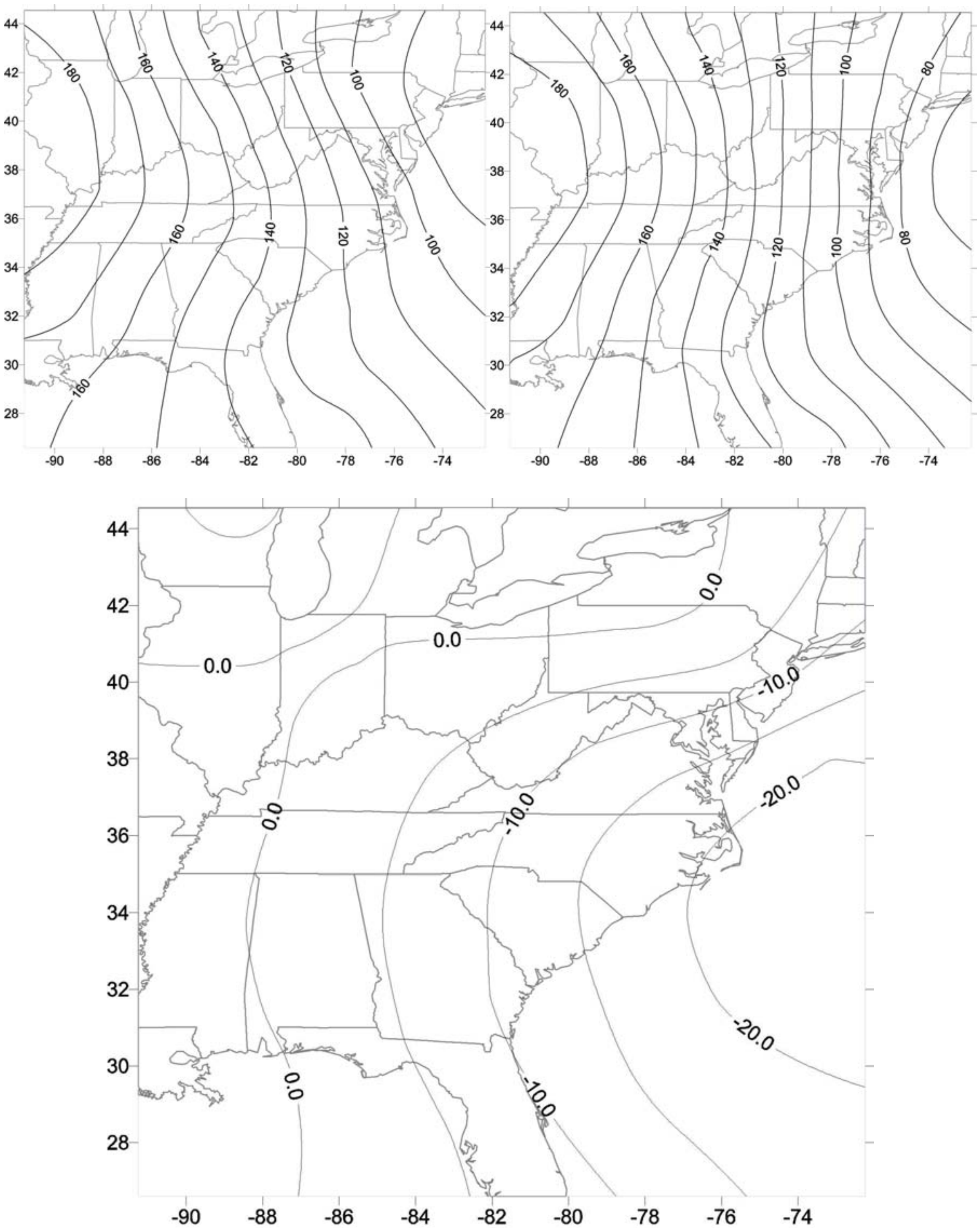

Figure 3.12. Composite plots of 1000 hPa height for no snow (top left), light snow (top right), and light snow minus no snow (bottom).

Table 3.5. Mean synoptic field values and differences during periods of low-level northwest flow in which no snow and light snow was reported.

High Country (Region 8)	No Snow	Snow	Abs_Dif	Rel_Dif	T-Score	
No Snow vs. Snow	*n=33*	*n=33*				
850 hPa Temperature	-1.8	-5.5	-3.7	-1.01	-4.76	*
500 hPa Height	5555	5478	-76	-1	-4.67	*
1000-500 hPa Thickness	5405	5337	-68	-0.98	-4.58	*
850 hPa Theta-e	18.7	13.3	-5.4	-0.85	-3.84	*
500 hPa Vorticity Advection	3.1	0.4	-2.8	-0.66	-2.83	*
500 hPa Vorticity	2.3	3.6	1.3	0.56	2.37	*
850 hPa Mixing Ratio	2.9	2.3	-0.6	-0.55	-2.31	*
Precipitable Water	0.38	0.31	-0.07	-0.48	-2.01	*
200 hPa Divergence	1.6	2.3	0.7	0.47	1.97	*
500 hPa Relative Humidity	25.7	30.7	5.0	0.31	1.28	
850 hPa Wind Direction	295.0	306.6	11.7	0.28	1.16	
700 hPa Vertical Velocity	14.0	18.2	4.1	0.26	1.05	
1000-500 hPa Mean Relative Humidity	51.2	53.5	2.3	0.21	0.85	
1000 hPa Height	150	141	-8	-0.2	-0.81	
850 hPa Wind Speed	12.3	13.0	0.6	0.16	0.66	
850-500 hPa Temperature Change	18.3	17.8	-0.5	-0.16	-0.64	
850 hPa Relative Humidity	69.6	71.4	1.8	0.13	0.52	
850 hPa Thermal Advection	-6.5	-6.9	-0.4	-0.1	-0.42	

Snow-free events are also tied to higher values of 1000 to 500 hPa thickness and 850 hPa temperature (relative differences of approximately -1) (Table 3.5). The composite mean 850 hPa temperature is only slightly below 0 ˚C in the snow-free events, while it is -5.5 ˚C for the light events. Composite map patterns for 850 hPa temperature are similar between the two groups of events, but the greatest temperature differences are found over and just to the south of the Southern Appalachians (Fig. 3.13**)**. Although surface temperatures at the High Peaks (Region 14) are clearly cold enough for snow, it is the moist layer temperatures that appear to be the limiting factor. Previous work in cloud microphysics has shown that liquid water droplets significantly outnumber ice crystals at temperatures just below 0˚C, and that cloud temperatures need to be considerably below freezing in order for snowfall to develop (Pruppacher and Klett 1997). When temperatures are marginal (i.e. only slightly below 0 ˚C), few ice nuclei are present and drizzle, rather than snow, is likely to develop. Lower values of 500 hPa vorticity and 200 hPa divergence are also tied to snow-free events. Therefore, not only is the lower troposphere too warm for snowfall development, but mid- and upper-level dynamics are also lacking. Interestingly, 700 hPa vertical velocity, 1000 to 500 hPa mean relative humidity, 850 hPa relative humidity, and 500 hPa relative humidity showed very little difference. Due to the nearly equal values of lower tropospheric relative humidity, it is possible to conclude that moist layer temperatures are too warm for ice crystal growth in the snow-free events.

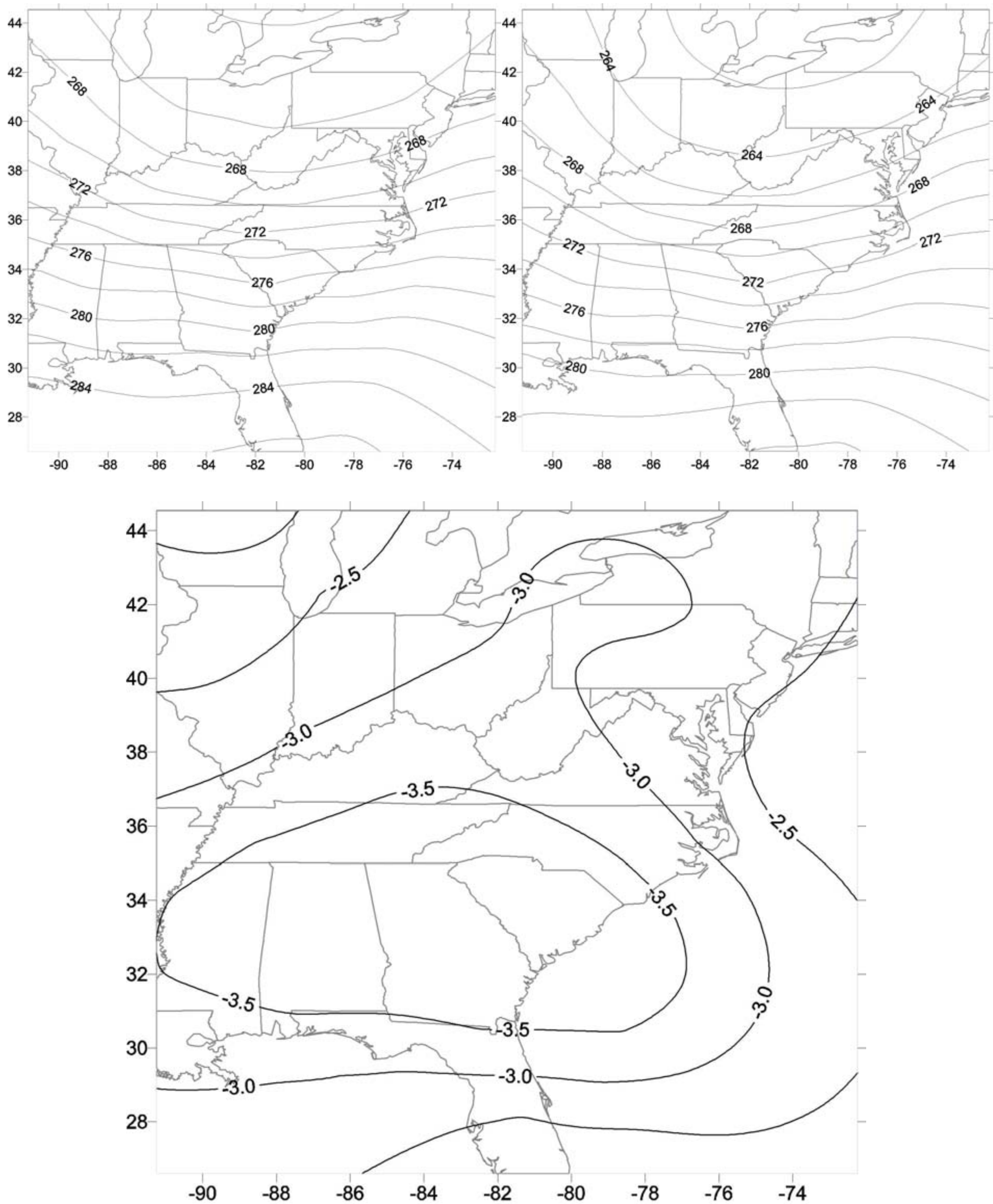

Figure 3.13. Same as Fig. 3.12, except for 850 hPa temperature (degrees K).

3.4. Influence of Elevation

Higher elevations typically receive heavier NWFS snowfall than lower elevations at the same latitude. However, in some cases, only the higher elevations receive snow. In others, they receive markedly more snow than lower elevations when snowfall is widespread at all elevations. Many of these instances are undoubtedly due to low-elevation surface temperatures that are well above 0 ˚C (Perry and Konrad 2006), but in some cases other factors may be play a role as well. In this section, the influence of elevation on snowfall totals is analyzed by comparing the ratio of mean snowfall between the High Peaks (Region 14) and all of the remaining snow regions (Regions 1 – 13). This is done for all events in which measurable snowfall was observed at the High Peaks as well as for only those events in which snowfall was relatively more widespread at all elevations.

3.4.1. Snowfall at High Peaks

This sub-section compares events in the top and bottom three quartiles of the ratio of mean snowfall between the High Peaks (Region 14) and the remaining snow regions for all events in which measurable snowfall was observed in the High Peaks. Events that display the greatest differences in snowfall by elevation are tied to warmer lower tropospheric temperatures (Table 3.6). In particular, 850 hPa temperature and 1000 to 500 hPa thickness are considerably higher (relative differences of 1.02 and 1.04) when compared with events that show less difference in snowfall be elevation. Values for 850 hPa theta-e, 850 hPa mixing ratio, and precipitable water are also significantly higher for those events with greater differences in snowfall by elevation. Interestingly, however, values of 1000 to 500 hPa mean relative humidity and 850 hPa relative humidity are decidedly lower, indicating that the greater moisture content of the lower troposphere is somewhat offset by less efficient use of available water vapor. Events with greater differences in snowfall by elevation are also tied to greater instability between 850 and 500 hPa, and shorter event durations.

Subtle, but important, differences between the two groups of events are evident in the composite plots of 1000 hPa height (Fig. 3.14). The events characterized by lower differences in snowfall by elevation (i.e. colder events) are tied to a surface cyclone just off the North Carolina coast, while the composite surface cyclone is off of the New England coast for those events with greater differences in snowfall by elevation (i.e. warmer events). The 1000 hPa height values show minimal differences between the two event types across the southeastern U.S., but are substantially lower over the northeastern U.S for the events with greater differences in elevation. In both cases, however, the surface anticyclone is positioned west of the Mississippi River. These subtle differences in the composite 1000 hPa height pattern translate into very important differences in lower tropospheric temperature between the two groups. A more southerly cyclone (i.e. off of the North Carolina coast) is tied to colder temperatures, whereas temperatures are considerably warmer in association with a more northerly cyclone. This pattern is therefore similar to the comparison of light versus heavy events associated with synoptic-scale ascent, where a stronger surface cyclone was displaced farther to south for the heavy events.

The warm temperatures, abundant lower tropospheric water vapor, but short durations suggest that many of the events with greater differences in snowfall by elevation may occur as the lower troposphere cools from the top down as a synoptic-scale disturbance exits the area. In these situations, it is a race to see how much overlap develops between the lower tropospheric cold advection and the departing synoptic-scale dynamics. Rain may change to snow much sooner at the higher elevations than the lower elevations, thereby significantly limiting snowfall accumulations at lower elevations. Examination of a sample of ten events that displayed the greatest differences of snowfall by elevation confirm these observations. Half of the sample is tied to recently departed cold fronts, troughs, or surface cyclones. In the other events, no synoptic-scale disturbance is noted, suggesting that these events are only differentiated from the bottom quartile events by the warmer lower tropospheric temperatures.

Table 3.6. Mean synoptic field values and differences for events in which snow occurred at all elevations and those in which the High Peaks (Region 14) received markedly more snow.

Influence of Elevation	All Elevs	High Elevs	Abs_Dif	Rel_Dif	T-Score	
(Ratio 14:1-13)	*n=431*	*n=143*				
850 hPa Temperature	-8.1	-2.9	5.2	1.04	11.25	*
850 hPa Theta-e	10.2	17.9	7.8	1.02	10.71	*
1000-500 hPa Thickness	5274	5358	83	0.93	9.93	*
850 hPa Mixing Ratio	2.1	3.0	0.9	0.9	8.64	*
500 hPa Height	5397	5469	73	0.81	8.55	*
Beginning-Ending	31	20	-11	-0.69	-8.09	*
850-500 hPa Temperature Change	18.4	21.1	2.6	0.67	7.68	*
Maturation-Ending	18	10	-8	-0.63	-7.65	*
Precipitable Water	0.3	0.41	0.11	0.78	7.54	*
Days	2.4	1.8	-0.6	-0.65	-7.44	*
Beginning-Maturation	13	10	-4	-0.43	-5.12	*
1000-500 hPa Mean Relative Humidity	63.5	59.9	-3.6	-0.3	-3.28	*
850 hPa Relative Humidity	80.1	76.0	-4.1	-0.32	-3.14	*
1000 hPa Height	122	111	-11	-0.22	-2.41	*
500 hPa Vorticity	4.6	4.2	-0.4	-0.19	-1.99	*
850 hPa Thermal Advection	-4.6	-4.2	0.4	0.1	1.1	
850 hPa Wind Speed	12.3	12.0	-0.4	-0.09	-0.87	
200 hPa Divergence	3.2	3.1	-0.1	-0.08	-0.84	
500 hPa Relative Humidity	39.5	39.1	-0.4	-0.02	-0.24	
850 hPa Wind Direction	301.7	301.4	-0.2	-0.01	-0.13	
700 hPa Vertical Velocity	8.5	8.3	-0.2	-0.01	-0.09	
500 hPa Vorticity Advection	0.9	0.9	0.0	0	0.06	

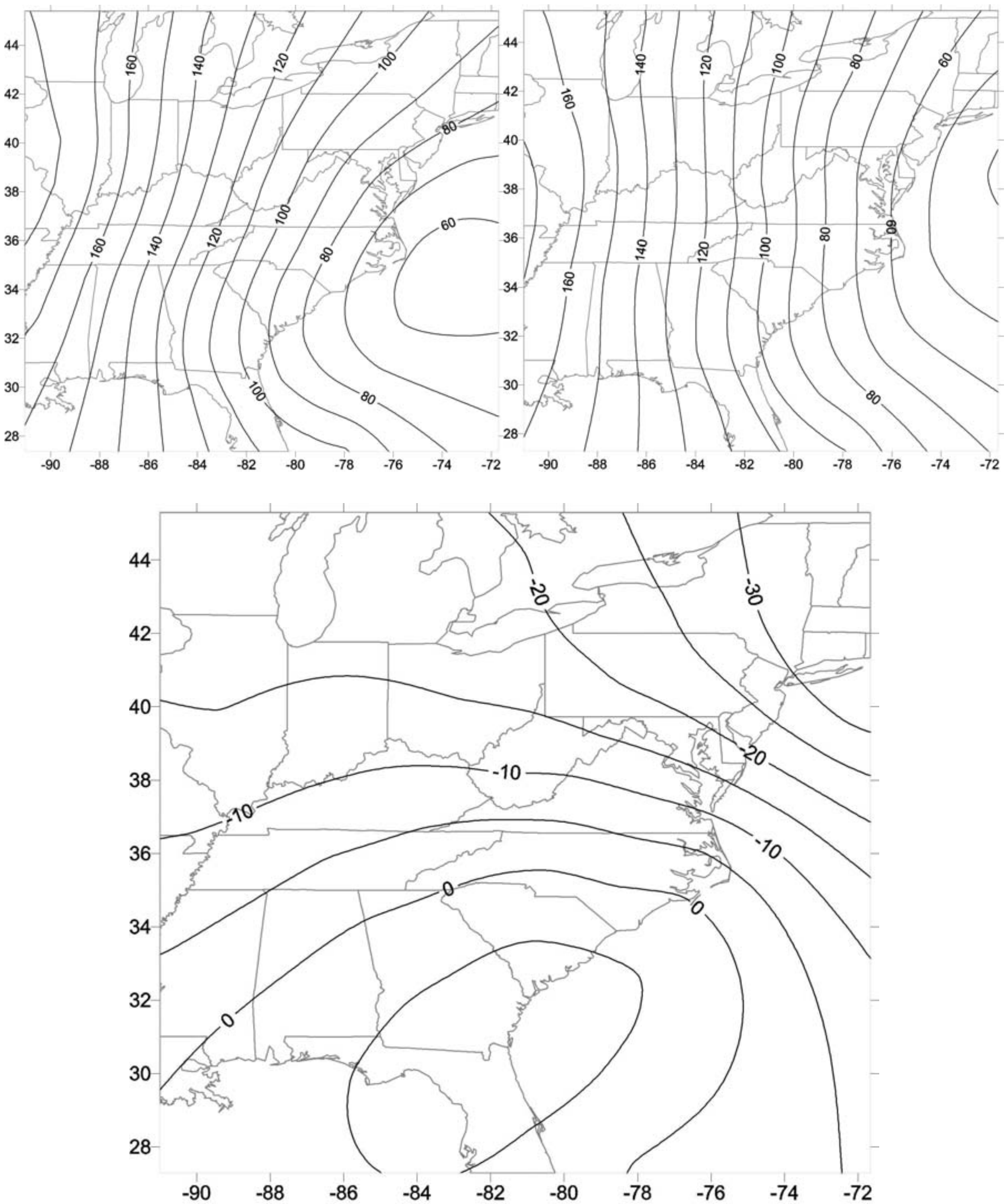

Figure 3.14. Composite plots of 1000 hPa height for events with less difference in snowfall by elevation (top left), those with greater differences (top right), and mean difference between the two (bottom).

3.3.2. Widespread at All Elevations

In order to further assess the influence of elevation on temperature, a second analysis examined just those events in which snowfall was more widespread at lower elevations. Only those events with measurable snowfall in the lower elevation snow regions of the Great Smoky Mountains (Region 4) and the Southern Blue Ridge (Region 5) were included in this analysis. This sample, therefore, represents events in which snow was relatively widespread across a wide range of elevations in western North Carolina and extreme eastern Tennessee. This was done to control for the influence of warm surface temperatures in limiting snowfall accumulations at lower elevations, which were found to exhibit some of the greatest differences between the two event types in the previous sub-section. As was done previously, comparisons were made between the top quartile and bottom three quartiles of the ratio of mean snowfall between the High Peaks and the remaining snow regions.

Composite map patterns for 1000 hPa height for this analysis are very similar to those discussed in the previous section (Fig. 3.14), and are therefore not included as a separate figure. The events with greater differences by elevation are associated with a composite surface cyclone off of the New England coast, and the greatest mean differences between the two event types are found over the northeastern U.S. However, slightly higher 1000 hPa heights are noted across the southeastern U.S. for events with greater differences by elevation. This pattern displays some similarities, therefore, with the composite 1000 hPa height pattern for events characterized by synoptic-scale subsidence. In that analysis, very little difference in the position of the composite surface cyclone off of the northeastern U.S. coast was noted between the light and heavy events, but heavy events were tied to a stronger composite cyclone (i.e. lower 1000 hPa heights).

As with the previous analysis, events with greater differences in snowfall by elevation are tied to lower values of 1000 to 500 hPa mean relative humidity and 850 hPa relative humidity (Table 3.7). Additionally, 500 hPa relative humidity is lower, while mid- and upper-level dynamics are also weaker. The composite map patterns for 700 hPa vertical velocity indicate stronger mid-level subsidence across the southeastern and mid-Atlantic U.S. (Fig. 3.15), with the greatest mean differences found directly over and to the east of the study area. Only the 850 to 500 hPa temperature change is appreciably higher, leading to two possible conclusions. First, the greater instability can be tied to a mid-level cold pool in association with the base of the 500 hPa trough. Secondly, sustained and continuous orographic forcing at the High Peaks plays an important role in snowfall development due to greater lower tropospheric instability. In these situations, sustained orographic lifting eventually results in a lower rate of cooling once saturation is acheived (i.e. due to the release of latent heat from condensation and deposition), allowing air parcels to become warmer than the surrounding environment. Further orographic lifting may result in the release of conditional instability and the development of low-topped convection.

Table 3.7. Same as Table 3.6, except for events that were widespread at all elevations.

Influence of Elevation	All Elevs	High Elevs	Abs_Dif	Rel_Dif	T-Score	
(Ratio 14:1-13), Widespread	*n=159*	*n=53*				
500 hPa Vorticity Advection	1.8	-0.8	-2.7	-0.57	-4.5	*
1000-500 hPa Mean Relative Humidity	68.3	62.7	-5.5	-0.47	-3.42	*
700 hPa Vertical Velocity	-0.7	8.8	9.5	0.45	3.33	*
500 hPa Relative Humidity	46.3	39.2	-7.1	-0.39	-2.65	*
850-500 hPa Temperature Change	18.3	19.9	1.6	0.4	2.47	*
200 hPa Divergence	3.9	3.3	-0.6	-0.33	-2.02	*
Beginning-Maturation	14	17	2	0.25	1.51	
850 hPa Wind Speed	11.5	12.2	0.7	0.16	1.08	
Precipitable Water	0.31	0.29	-0.02	-0.16	-1.04	
850 hPa Wind Direction	301.0	298.4	-2.6	-0.14	-0.99	
850 hPa Temperature	-8.6	-7.9	0.7	0.16	0.97	
Maturation-Ending	21	19	-2	-0.16	-0.96	
850 hPa Theta-e	9.5	10.4	0.9	0.14	0.87	
850 hPa Mixing Ratio	2.1	2.1	0.1	0.08	0.52	
1000 hPa Height	112	109	-3	-0.07	-0.47	
500 hPa Height	5376	5371	-5	-0.06	-0.4	
Days	2.6	2.6	0.0	-0.04	-0.26	
500 hPa Vorticity	4.7	4.8	0.1	0.03	0.19	
850 hPa Thermal Advection	-3.8	-3.9	-0.1	-0.02	-0.15	
1000-500 hPa Thickness	5263	5262	-2	-0.02	-0.14	
Beginning-Ending	36	36	0	0.02	0.09	
850 hPa Relative Humidity	80.9	80.8	-0.1	-0.01	-0.04	

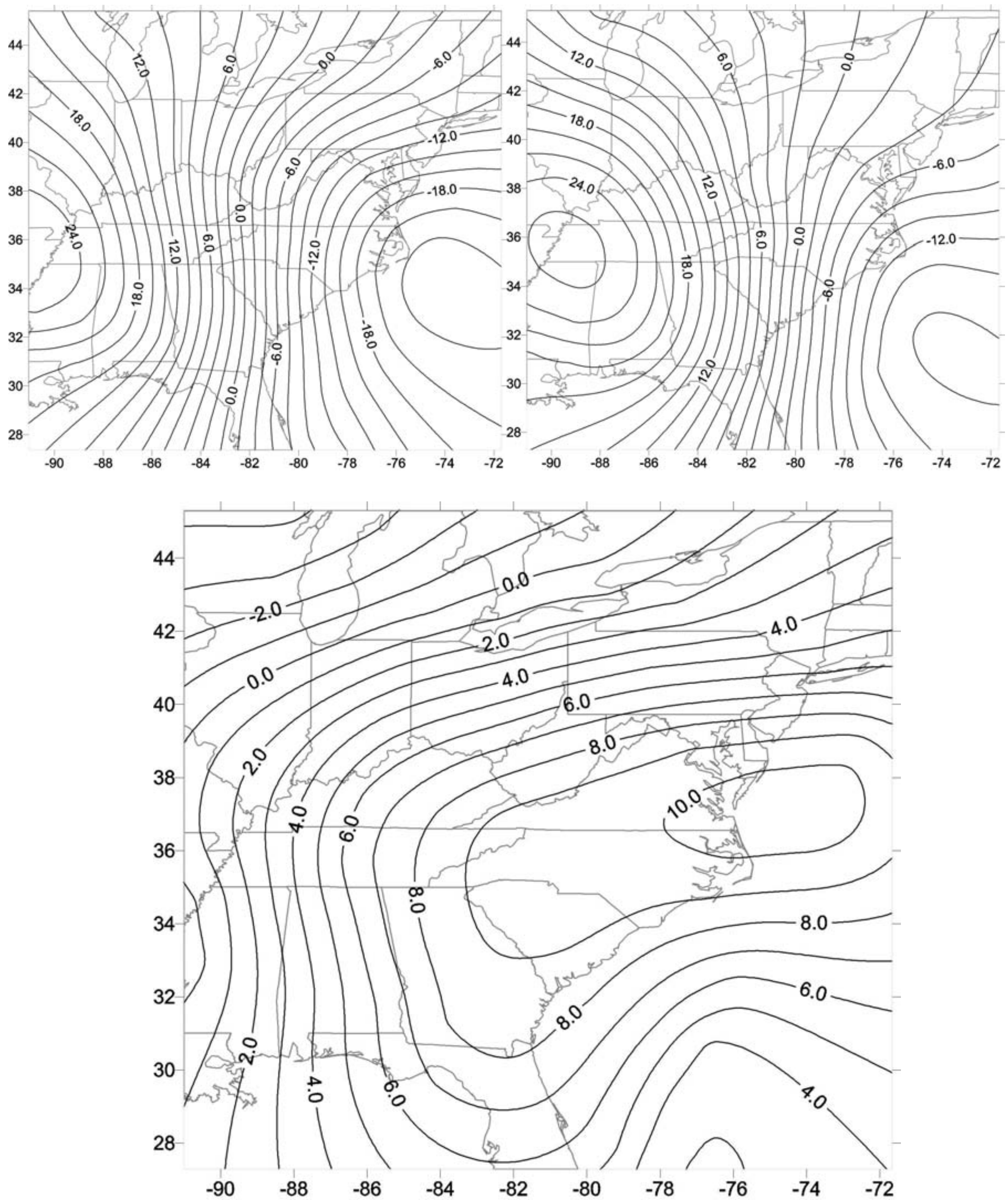

Figure 3.15. Same as Fig. 3.14, except for 700 hPa vertical velocity.

3.5. Spillover Effects

NWFS events that produce substantial amounts of snowfall in areas well downwind of the windward slopes and higher elevations (e.g. spillover snowfall) are particularly problematic for forecasters and the public, since such events are so infrequent and therefore not anticipated. They are infrequent due to the strong downslope flow that develops in periods of low-level northwest flow, which typically results in warming and drying of low-level air parcels. Spillover events can catch forecasters and the public by surprise, as evidenced by the unprecedented impacts in the Raleigh, NC, area caused by light spillover snowfall (NWS 2005). Although this event did not occur within the Southern Appalachian study area, it effectively illustrates the forecast challenges and societal impacts. In this section, spillover effects during periods of NWFS are analyzed by comparing the bottom quartile and the top three quartiles of the ratio of mean snowfall between the Northern Plateau (Region 13) and the New River Valley (Region 10). The bottom quartile of this ratio represents events in which significant spillover was noted, whereas the top three quartiles correspond to events with little or no spillover. Although spillover snowfall can also occur further downwind, into the Central and Northern Foothills (Regions 9 and 11), these events are exceedingly rare and the sample size is not sufficient for a synoptic climatological analysis.

Events with significant spillover into the New River Valley are tied to a surface cyclone displaced farther to the south than in events with little or no spillover (Fig. 3.16). Composite 1000 hPa heights are also somewhat lower across the southeastern U.S., with the greatest differences noted over Georgia and South Carolina. Very little difference in the position of the surface anticyclone to the west is evident, however. The more southerly location of a stronger composite surface cyclone for the spillover events is very similar to the composite patterns for the heavy events with synoptic-scale ascent and the widespread events at all elevations. Therefore, it is apparent that a stronger surface cyclone off of the Carolina coast is tied not only to heavy synoptic-scale ascent snowfall, but also widespread snowfall at all elevations and spillover snowfall.

These similarities are supported by composite patterns and composite synoptic field values for mid- and upper-level dynamics. Spillover events are tied to lower values of 700 hPa vertical velocity, indicating weak to near neutral synoptic-scale subsidence (Table 3.8). The composite map patterns for 700 hPa vertical velocity are particularly revealing, showing much weaker subsidence over the Southern Appalachians and just to east for spillover events (Fig. 3.17). However, the Northern Foothills (Region 13) is under the influence of synoptic-scale ascent at event maturation in the composite pattern. The 200 hPa divergence also shows significantly higher values, indicating the importance of mid- and upper-level dynamics in producing spillover snowfall. Lower and middle tropospheric relative humidity is also higher in the spillover events, while event durations are somewhat longer. The elevation and spillover analyses together highlight the importance of synoptic-scale forcing in producing accumulating snowfall at lower elevations and along leeward slopes. Not only do they share many similarities, but they also display many characteristics of the heavy synoptic-scale ascent events. Therefore, it is possible to conclude that heavy NWFS in the presence of synoptic-scale ascent *typically* results in widespread snowfall at all elevations and along both windward and leeward slopes.

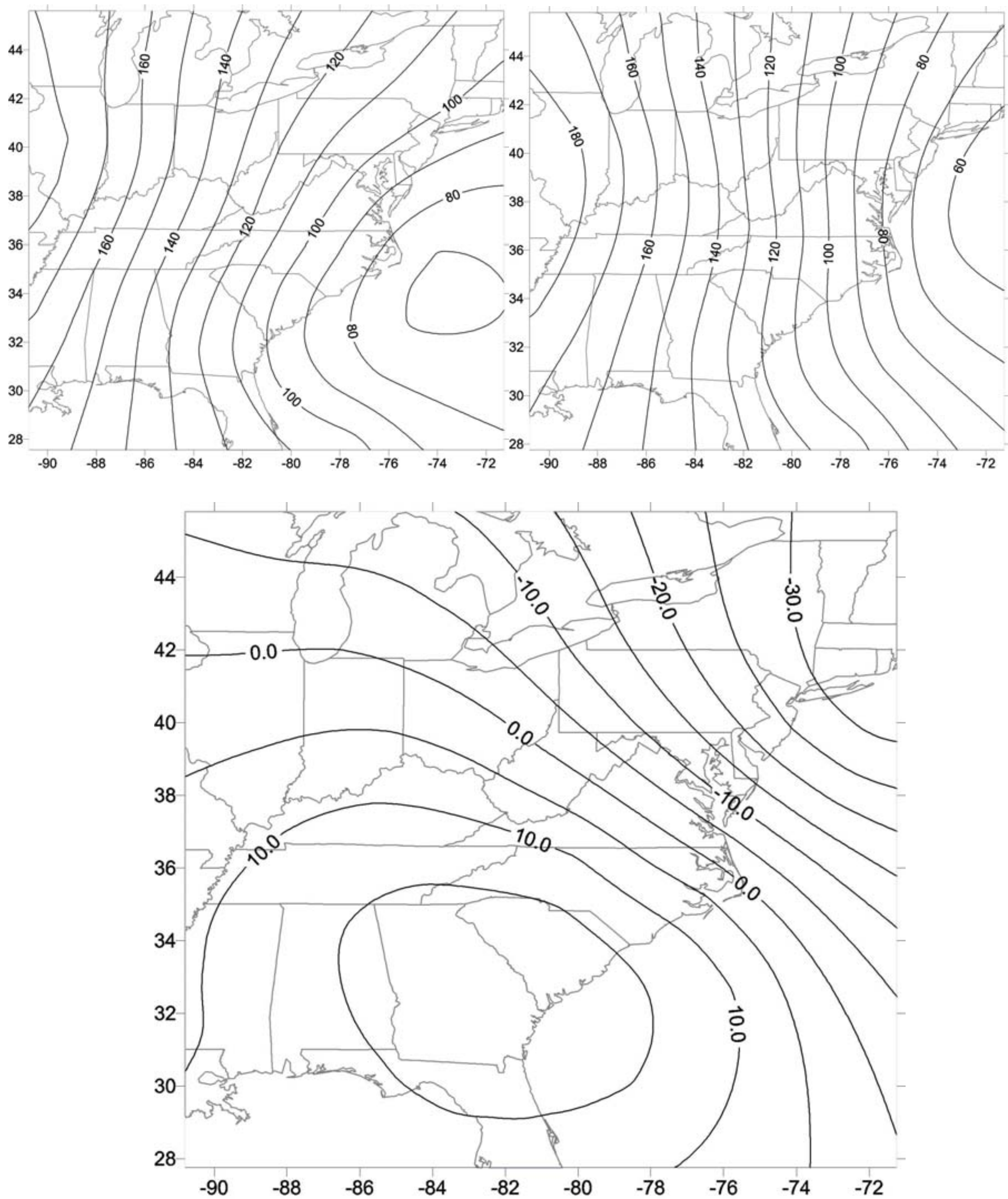

Figure 3.16. Composite plots of 1000 hPa height for events with significant spillover (top left), those with little or no spillover (top right), and mean difference between the two (bottom). Spillover was assessed by evaluating the ratio of mean snowfall between the Northern Plateau (Region 13) and New River Valley (Region 10).

Table 3.8. Mean synoptic field values and differences for events with significant spillover and those with little or no spillover.

Spillover (Ratio 13:10)	Abundant	Limited	Abs_Dif	Rel_Dif	T-Score	
	n=190	*n=570*				
700 hPa Vertical Velocity	1.7	13.7	12.0	0.66	7.32	*
1000-500 hPa Mean Relative Humidity	66.1	58.8	-7.3	-0.62	-6.96	*
Days	2.4	2.0	-0.4	-0.44	-5.31	*
Beginning-Ending	31	25	-7	-0.42	-5.1	*
500 hPa Relative Humidity	42.3	34.3	-8.0	-0.47	-5.07	*
850 hPa Wind Speed	11.3	13.1	1.8	0.45	5.03	*
200 hPa Divergence	3.3	2.7	-0.7	-0.37	-4.66	*
850 hPa Relative Humidity	80.9	76.8	-4.1	-0.32	-4.1	*
850 hPa Thermal Advection	-4.2	-5.6	-1.5	-0.35	-4.1	*
Beginning-Maturation	14	11	-3	-0.35	-3.99	*
Maturation-Ending	18	14	-4	-0.32	-3.89	*
500 hPa Height	5396	5417	21	0.26	3.11	*
500 hPa Vorticity Advection	2.0	0.8	-1.2	-0.26	-3.1	*
1000 hPa Height	121	131	10	0.19	2.3	*
Precipitable Water	0.31	0.29	-0.02	-0.19	-2.18	*
500 hPa Vorticity	4.6	4.3	-0.3	-0.14	-1.8	*
1000-500 hPa Thickness	5275	5286	11	0.13	1.63	
850 hPa Temperature	-8.0	-7.5	0.5	0.11	1.32	
850 hPa Theta-e	10.4	10.8	0.4	0.06	0.76	
850 hPa Mixing Ratio	2.2	2.1	0.0	-0.04	-0.51	
850 hPa Wind Direction	303.3	302.6	-0.7	-0.04	-0.46	
850-500 hPa Temperature Change	18.4	18.3	-0.1	-0.01	-0.15	

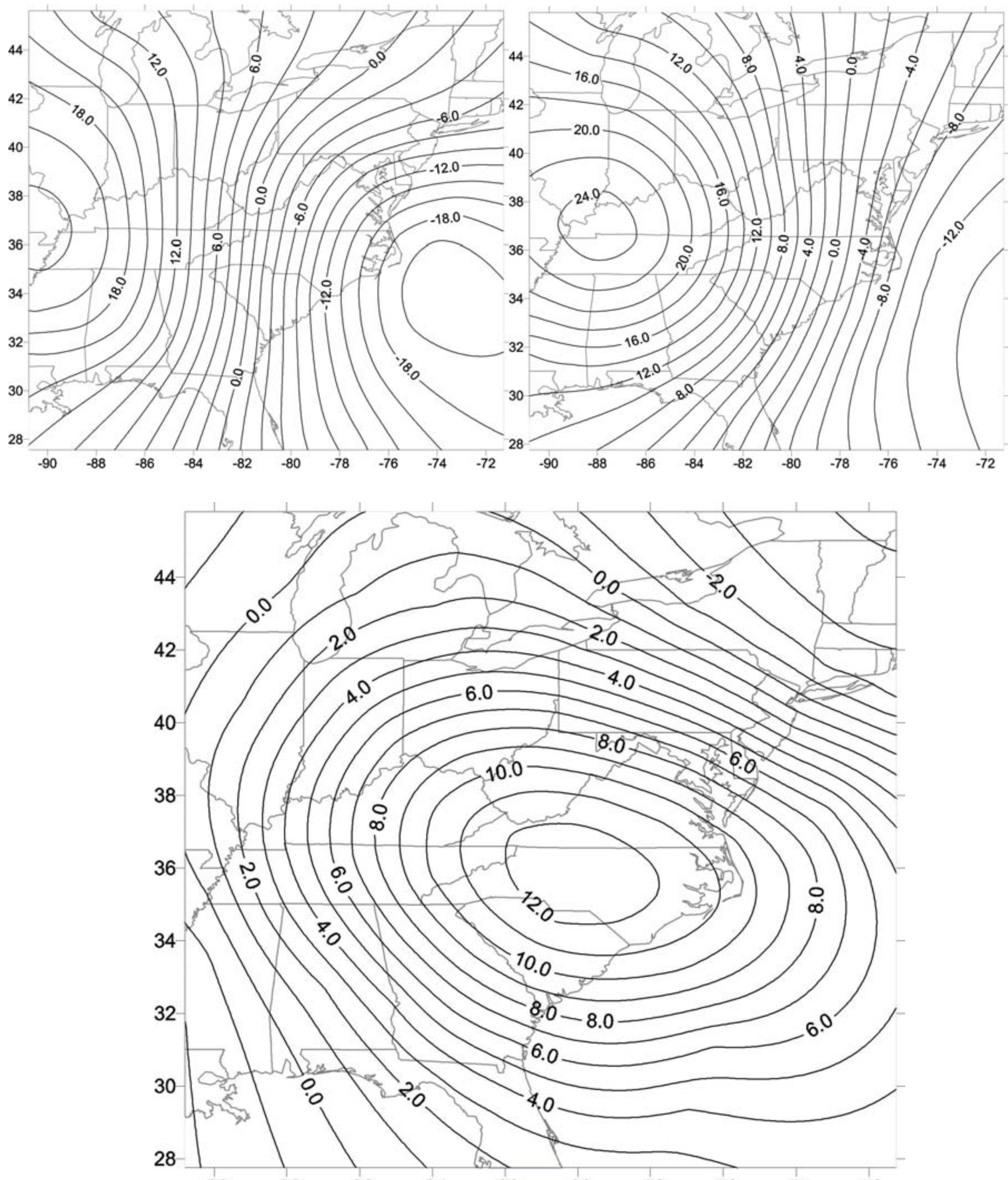

Figure 3.17. Same as Fig. 3.16, except for 700 hPa vertical velocity.

3.6. Trajectory Analysis

An antecedent upstream air trajectory analysis was undertaken in order to develop a more detailed understanding of the major air trajectories tied to NWFS. Previous work (e.g. Schmidlin 1992) has suggested that many of the NWFS events in the northern mountains of West Virginia are tied to air trajectories that extend downwind from the eastern Great Lakes, particularly Lake Erie. However, the frequency and influence of trajectories with a Great Lakes Connection (GLC) farther south in the Appalachians has not been well understood. A perusal of satellite and radar imagery for selected lake-effect snow events occasionally shows a clear connection with NWFS in the Southern Appalachians. This part of the analysis classifies antecedent upstream (backward) air trajectories using a classification scheme presented in section 2.4.5. In addition, it compares composite snowfall totals, synoptic fields, and vertical profiles of moisture and temperature across trajectories. Lastly, the influence of the Great Lakes on snowfall totals is assessed by comparing selected trajectories and analyzing the spatial patterns of snowfall across the Southern Appalachians.

3.6.1. Trajectory Classification

Table 3.9 summarizes the results of the trajectory classification. The most frequent air trajectory associated with NWFS in the Southern Appalachians is Trajectory Class 3.1 (30.9%), followed closely by Class 2 (26.7%). Interestingly, almost half (47.1%) of the 191 NWFS events analyzed in this study exhibit a GLC (Trajectory Class 3) and approximately three-quarters (73.8%) of the events display a northwesterly trajectory (Trajectory Class 2 or 3). Besides sub-classes 3.2 and 3.3 of Trajectory Class 3, the most infrequent trajectories are Trajectory Class 1 and 4, which are characterized by a more westerly trajectory, as opposed to northwesterly, at event maturation. The remaining events did not meet any of the trajectory classification requirements and were therefore grouped together in Trajectory Class 5.

Table 3.9. Frequencies of each backward air trajectory class.

Class	Frequency	Percent	Description
1	15	7.9%	≥36 hrs in Region 1, No GLC
2	51	26.7%	≥36 hrs in Region 2, No GLC
3.1	59	30.9%	≥6 hrs in Western Great Lakes
3.2	10	5.2%	≥6 hrs in Eastern Great Lakes
3.3	21	11.0%	≥6 hrs in both W and E Great Lakes
4	16	8.4%	≥36 hrs in Region 4, No GLC, <36 hrs in either 1 or 2
5	19	9.9%	Remainder
	191		**Total**

The temporal pattern of trajectory classes by month (Fig. 3.18) indicates significant variability within and among the classes. Most of the Trajectory Class 1 events were observed during the coldest months of the winter (i.e. December, January, and February), when westerly trajectories are associated with air sufficiently cold for snowfall to occur. Trajectory Class 3.1, on the other hand, displayed a relatively equal monthly distribution as the other trajectory classes were largely limited to the months of December, January, February, and March. The overwhelming majority of events in the month of November were associated with GLC trajectories, suggesting the importance of northerly flow in advecting air into the study region that is cold enough for snow to occur. The lake-air temperature contrast, which is greatest in the late fall and early winter when the sensible and latent heat fluxes are maximized (Niziol et al. 1995), may also contribute somewhat to the observed monthly patterns for Trajectory Class 3.1. The very low number of February events for Trajectory Class 2 is noteworthy, as is the spike of events for Trajectory Class 3.3 in the same month. Both of these aberrations likely result from small sample sizes rather than monthly variations in the synoptic environment.

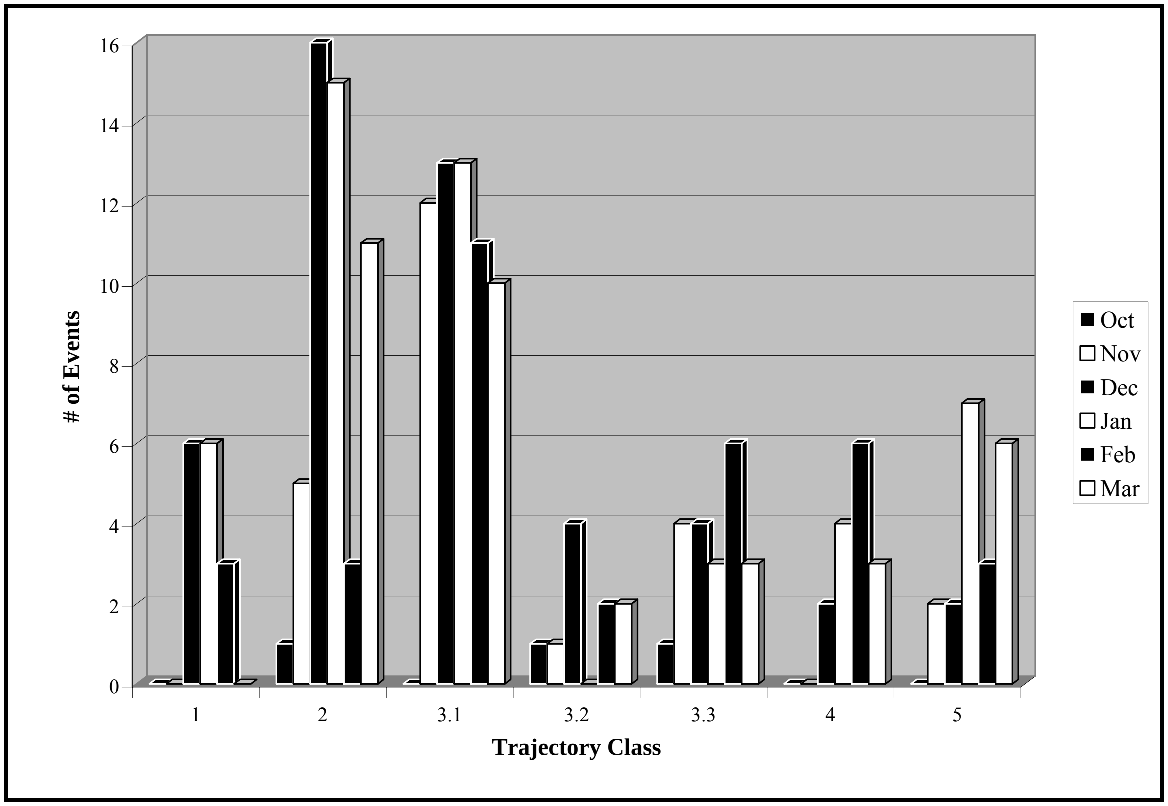

Figure 3.18. Trajectory class by month.

3.6.2. Composite Trajectory Plots

Northwesterly trajectories, or Trajectory Class 2 and 3, are discussed first, followed by the more westerly trajectories of Trajectory Class 1 and 4. Composite trajectory plots (Fig. 3.19) for Trajectory Class 2 show a backward air trajectory that extends to the northwest, but well to the west of the western Great Lakes. Subsidence is noted in the vicinity of the Ohio River Valley, with the air parcels descending from 800 hPa to just below 850 hPa, before ascending as a result of orographic forcing. Trajectory Class 3.1 is very similar to Class 2, but in this case part of the trajectory at T0 and T+12 crosses the extreme southern portion of Lake Michigan. Subsidence is also evident from 800 to 850 hPa over the 72-hour period, although not quite as abrupt in the vicinity of the Ohio Valley as with Class 2. Air parcels in association with Trajectory Class 3.2 spend a considerable amount of time in the vicinity of the Great Lakes, with the backward air trajectory centered over the eastern Great Lakes at TO. Air parcels are also near the surface for an extended period of time, with modest lift occurring as a result of the topography. A veering of the trajectory to a more northerly component is evident between T-12 and T0 for all three composite trajectories, with less consistency noted between T0 and T+12, particularly in Trajectory Class 3.2, where backing is noted.

The position of the composite cyclone is related to the orientation of the air trajectories (Fig. 3.19). Cyclones positioned farther south (e.g. off the mid-Atlantic coast) favor a more northerly low-level airflow with a GLC while cyclones positioned farther north off the New England coast favor a more westerly trajectory that is positioned west of the Great Lakes. Interestingly, in several previous analyses, composite surface cyclones displaced farther to the south were tied to heavy snowfall that was more widespread even at lower elevations and along leeward slopes. The position of the surface cyclone may also relate to a deeper 500-hPa trough and associated lower 500 hPa heights, which also characterize heavy events in previous analyses.

Trajectory Class 1 (Fig. 3.20) extends west from the study area in association with strong subsidence and backing of winds as air parcels descend from 650 to 850 hPa over the 72-hour period. The composite 1000 hPa height field reveals a strong anticyclone located to the west, with a departing surface cyclone to the east. Many of the 72-hour backward air trajectories extend as far west as the Pacific Ocean, suggesting that moisture for many of these events is of Pacific origin. A veering of the flow from southwest to northwest is evident from the temporal trajectory analysis, consistent with the temporal pattern for Trajectory Class 2. Trajectory Class 4 (Fig. 3.20) occurs in situations of weak southwest flow at T-12 that becomes northwest near event maturation, and then increases and becomes a more northwesterly trajectory as the event approaches T+12. The vertical trajectory plot clearly shows that air parcels are lifted from the surface in this case, which stands in contrast to the pronounced subsidence evident in Trajectory Class 1. The composite 1000 hPa height pattern indicates a cyclone to the east and southeast off the NC/SC coast, which is evidently far enough away to preclude synoptic-scale lift, but close enough to result in northwest flow at 850 hPa. It is important to recognize that the more westerly trajectories characterized by classes 1 and 4 are infrequent, together representing just 16.3 percent of the sample.

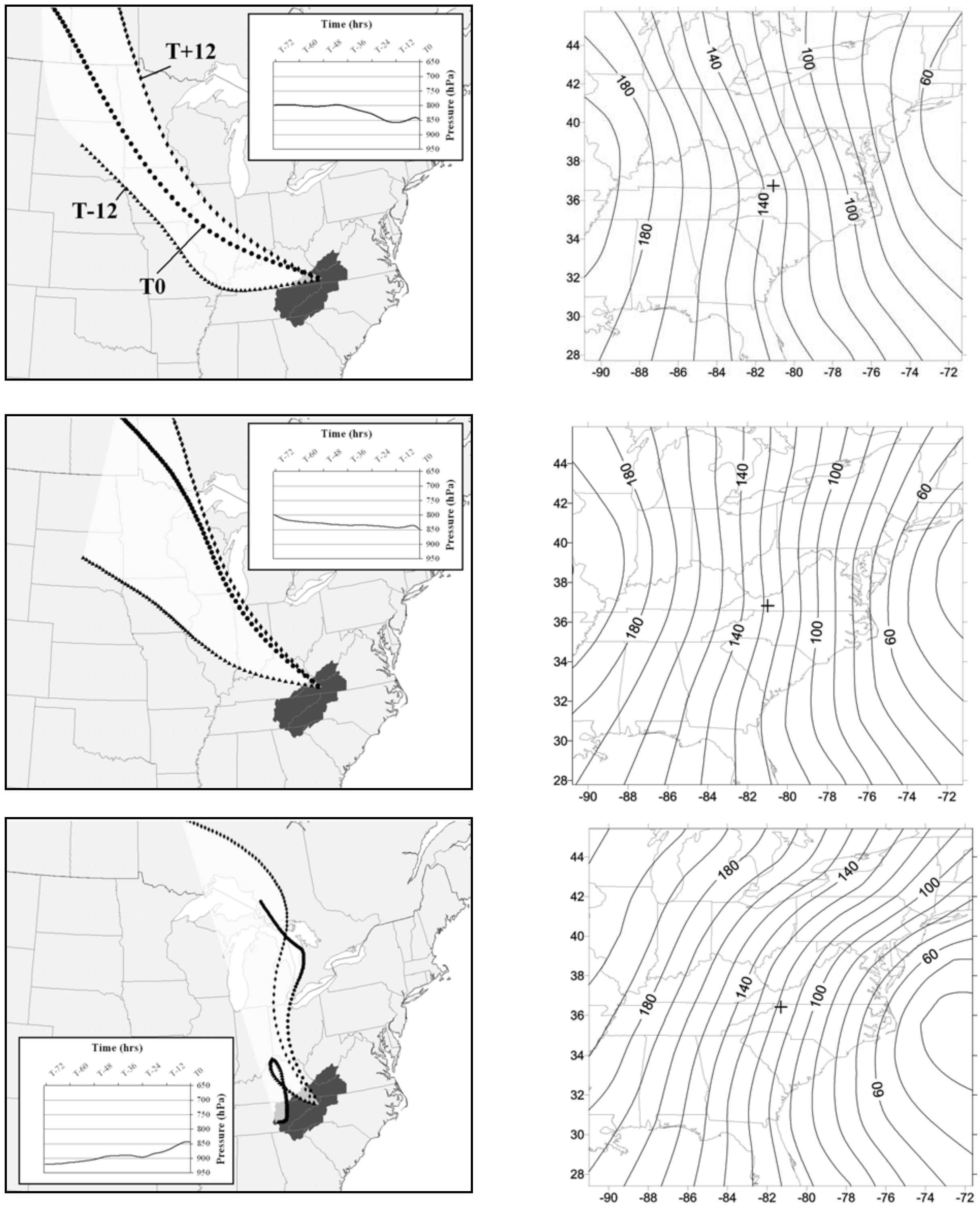

Figure 3.19. Composite trajectories at T-12, T0, and T+12 and surface pressure pattern at T0 (right) for Trajectory Class 2 (top), 3.1 (middle), and 3.2 (bottom).

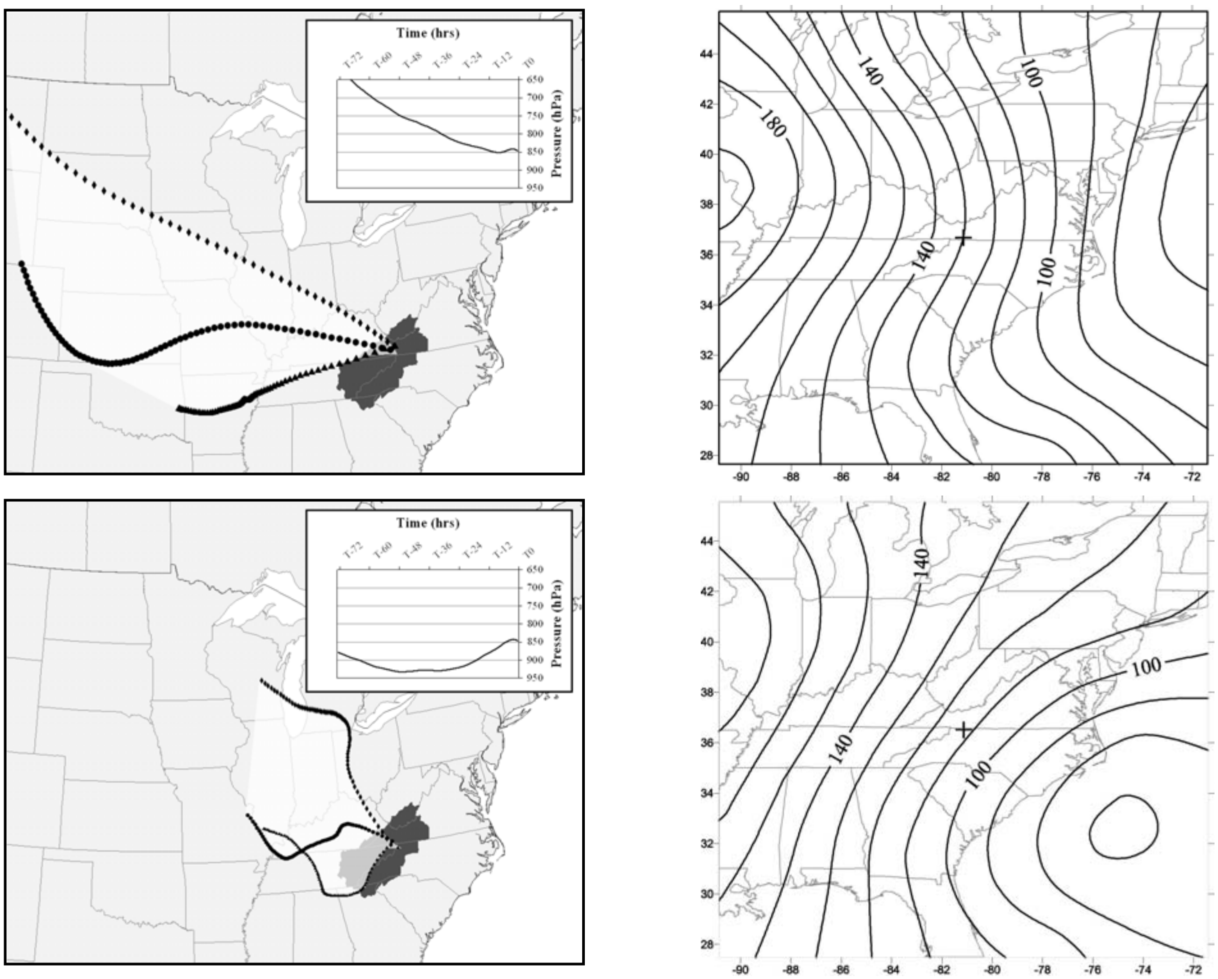

Figure 3.20. Same as Fig. 3.19, except for Trajectory Class 2 (top) and 4 (bottom).

3.6.3. *Composite Snowfall Totals*

Snowfall totals exhibit substantial variability among the different trajectory classes and across the study area (Fig. 3.21). The High Country (Regions 8), Central and Northern Plateau (Region 12 and Region 13), and High Peaks (Region 14) received the greatest amounts of snowfall across the different trajectories. The lower elevation regions, and particularly those on the leeward slopes (Southern, Central, and Northern Foothills, or Regions 3, 9, and 11), received no measurable snowfall across most of the trajectories. These results are consistent with those reported by Perry and Konrad (2006), in which the greatest snowfall totals for NWFS events occurred along the windward slopes and at higher elevations. The lowest composite mean snowfall totals are found in association with Trajectory Class 1 and 2 across all snow regions, whereas the highest snowfall totals generally associated with Trajectory Class 3.2, 4, and in some cases 5. The low composite mean and maximum snowfall totals in association with Trajectory Class 1, in particular, may be tied to the strong subsidence along a westerly trajectory. Trajectory Class 4 stands out as producing the highest snowfall totals across all but one of the snow regions, suggesting the presence of more abundant moisture and/or higher relative humidity. Additionally, as evidenced from previous analyses, heavy snow at all elevations and along leeward slopes is tied to a surface cyclone off of the Carolina Coast.

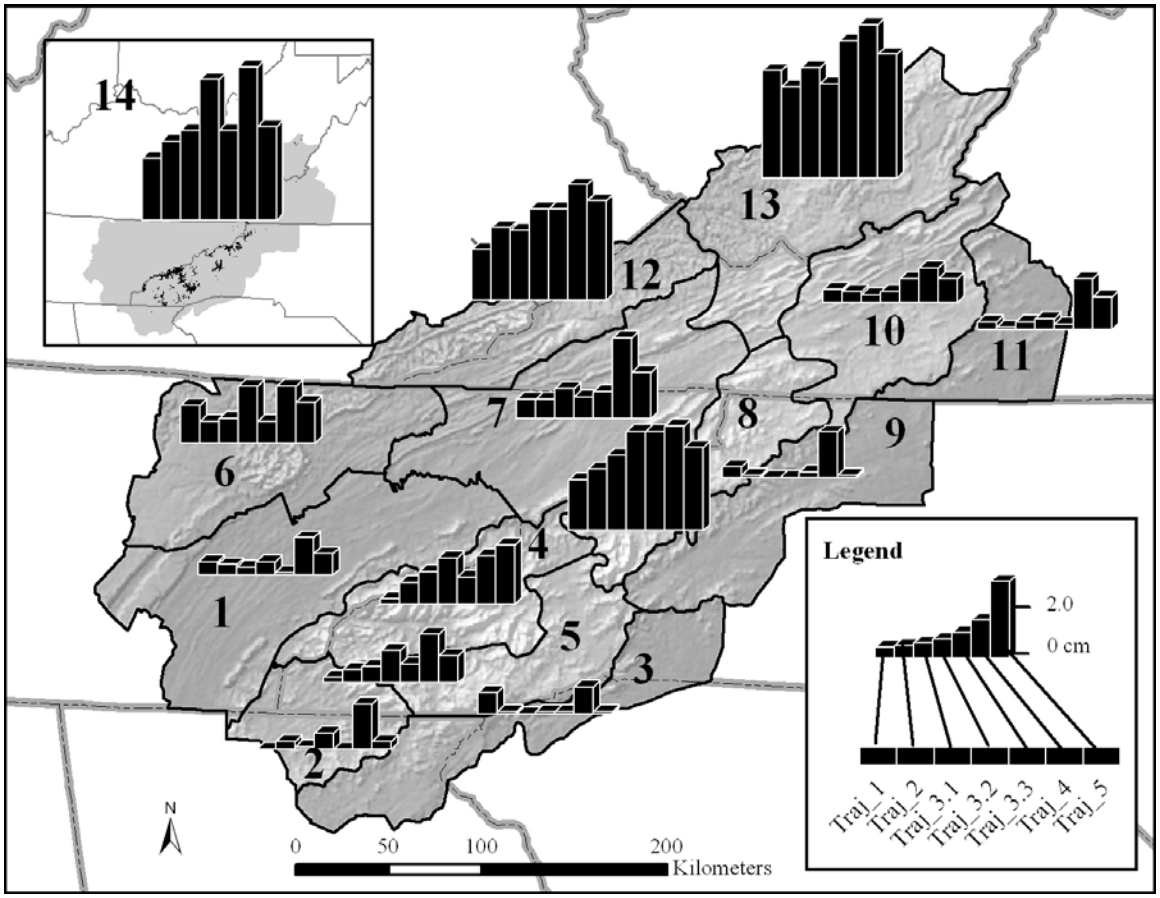

Figure 3.21. Composite mean snowfall totals by trajectory class and snow region.

The Southern, Central, and Northern Foothills (Regions 3, 9, and 11) exhibit the lowest composite mean snowfall totals across most of the trajectory classes, although the totals are substantially higher for Trajectory Class 4 and 5. It should be noted, however, that synoptic-scale lift in association with a coastal low extended just far enough to the west to bring significant snowfall to these areas in a small number of events (3 out of the 16 in the sample). Interestingly, some of the lowest composite mean snowfall totals across the different trajectory classes are also found in the New River Valley of Southwestern Virginia (Region 10). Even though the mean elevation is relatively high (686 m), this snow region is heavily shadowed in periods of northwest flow by the higher terrain of the Northern Plateau (Region 13) to the west and northwest. Therefore, it is relatively uncommon for accumulating snowfall to occur in the absence of synoptic-scale ascent in periods of low-level northwest flow. A general latitudinal gradient, with the exception of the New River Valley (Region 10) in composite mean snowfall is evident across the different trajectory classes as northern snow regions at the same elevation and exposure display higher totals (e.g. Southern Plateau (Region 6) vs. Central Plateau (Region 12)). This latitudinal gradient likely continues northward to the Central Appalachian Mountains of northern West Virginia, western Maryland, and Pennsylvania.

3.6.4. Composite Sounding and Synoptic Fields

Results from the sounding analysis and the NCEP reanalysis data provide some additional insight into the observed snowfall patterns by trajectory class (Table 3.10). The lower snowfall totals observed in Trajectory Class 1 and 2 in particular can be tied to a "thinner" moist layer, lower values of relative humidity and moisture at 850 hPa, and higher values of cold air advection. Previous work (e.g. Auer and White 1982) has demonstrated the importance of a deep moist layer in producing heavy orographic snowfall in the western U.S. Perry and Konrad (2004) also found that the heavy NWFS events in the Southern Appalachians were tied to higher values of lower tropospheric relative humidity, so the results of this study are consistent with previous work. The higher values of 850 hPa cold air advection for Trajectory Class 1 and 2 also imply stronger subsidence, and therefore a less favorable synoptic environment for snowfall in the lower troposphere. The 500 hPa vorticity, though still positive (cyclonic), is lower in Trajectory Class 1 and 2, while the remaining trajectory classes are tied to higher absolute values of 500 hPa cyclonic vorticity.

Several of the sounding and synoptic variables in association with the higher snowfall totals of Trajectory Class 3.2 are quite distinctive from the other trajectory classes. Temperatures were in the dendritic temperature range for only approximately one-third (38%) of the events, and the mean temperature of the moist layer was also considerably greater than the other trajectory classes. In addition, the 850 hPa mixing ratio was exceptionally high (2.66 g/kg), as were 500 hPa relative humidity, precipitable water, 1000-500 hPa mean relative humidity, and 850 hPa theta-e. The warmth of the moist layer is particularly striking, since previous work has demonstrated that ice crystal growth is most rapid and efficient in the dendritic temperature range of -14 ˚C to -17 ˚C (Ryan et al. 1976, Pruppacher and Klett 1997, Fukuta and Takahashi 1999); furthermore, Auer and White (1982) have tied heavy snowfall to moist layer temperatures within the dendritic growth range. These results indicate that heavier NWFS can occur—albeit infrequently—with moist layer temperatures outside of the dendritic growth range. However, a deep moist layer with exceptionally high values of lower tropospheric moisture appears to be needed to make up for the reduced precipitation efficiency. The synoptic characteristics associated with this trajectory class show some commonalities with events that exhibit greater differences in snowfall by elevation (i.e. markedly more snow at higher elevation), particularly in the exceptionally high water vapor content and warm temperatures.

Trajectory Class 4 is also associated with a considerably thicker moist layer, much higher values of humidity and moisture at 850 hPa and 500 hPa, and lower values of cold air advection when compared to the remaining trajectory classes. Interestingly, however, the 850 hPa wind speed for Trajectory Class 4 is more than 4 m s^{-1} below the speeds for the other classes. This reduction in 850 hPa wind speed is associated with a less pronounced pressure gradient, apparently due to a weaker surface anticyclone to the west (Fig. 3.20). The weak flow is evident in the composite trajectory plots (Fig. 3.20), with the 72-hour backward trajectories extending only slightly west of the Mississippi River. In this trajectory class, which once again occurs relatively infrequently (n=16, or 8.4% of sample) compared to the others, abundant and deep low-level moisture is lifted orographically, resulting in heavier snowfall. Therefore, significant NWFS can also result across the region when air masses with high values of relative humidity and moisture throughout the lower troposphere are lifted orographically, albeit weakly compared with other NWFS events.

Table 3.10. Mean sounding and synoptic field values by each trajectory class.

Field	Traj_1 n = 13	Traj_2 n = 40	Traj_3.1 n = 48	Traj_3.2 n = 8	Traj_3.3 n = 17	Traj_4 n = 11	Traj_5 n = 18
Top of Moist Layer	703	688	689	604	634	590	584
Dendritic Temp Range	0.54	0.65	0.69	0.38	0.65	0.64	0.5
Depth of Moist Layer	203	140	183	238	237	272	188
Avg Temp of Moist Layer	-9.3	-11.5	-11.7	-6.6	-8.6	-8.5	-8.3
Lapse Rate Above Moist Layer	-4.0	-3.7	-4.5	-5.4	-6.2	-5.8	-4.6
1000 hPa Height	142	146	135	114	110	122	117
500 hPa Height	5453	5392	5373	5459	5405	5437	5413
850 hPa Mixing Ratio	1.86	1.72	1.75	2.66	2.28	2.45	2.38
850 hPa Relative Humidity	70.4	73.1	77.9	82.5	81.5	82.0	82.8
850 hPa Wind Direction	306	300	306	309	301	309	296
850 hPa Wind Speed	13.2	14.2	13.8	14.3	14.5	8.7	12.7
850 hPa Temperature	-8.1	-9.8	-9.9	-5.5	-6.9	-6.3	-7.4
850 hPa Thermal Advection	-8.9	-7.0	-6.2	-2.7	-5.2	-2.9	-3.6
1000-500 hPa Mean Relative Humidity	56.1	54.4	55.7	64.8	60.7	62.7	66.0
700 hPa Vertical Velocity	11.7	20.0	23.6	16.4	16.3	17.4	13.9
850-500 hPa Temp Change	14.9	17.4	17.9	17.3	19.2	19.3	18.3
Precipitable Water	0.28	0.22	0.22	0.37	0.30	0.33	0.33
500 hPa Relative Humidity	27.3	29.4	28.7	37.4	34.2	30.8	35.8
850 hPa Theta-E	9.4	7.3	7.2	14.3	11.8	12.9	11.6
500 hPa Vorticity	2.6	4.4	5.2	4.5	6.0	4.4	4.1

3.6.5. Influence of the Great Lakes

Comparison of the northwesterly trajectories of Class 2 (no GLC) with classes 3.1, 3.2, and 3.3 (GLC) reveals that trajectories with a GLC are associated with higher composite mean and maximum snowfall totals across windward slopes and higher elevations of the study area, with the influence the greatest in the High Country and Northern Plateau (regions 8 and 13) (Fig. 3.22). In both of these regions, the composite mean and maximum snowfall totals are significantly greater ($p < 0.05$) for events with a GLC, with the composite max snowfall 1.8 to 2.1 cm greater (150 to 135 percent). However, the composite max snowfall may be a better indicator of local enhancement as a result of favorable trajectories than the composite mean. Modest increases are noted in several of the other snow regions, particularly those with portions along the windward slopes. At the lower elevations and along the leeward slopes, very little if any snowfall occurs with either trajectory, highlighting the importance of topographic controls on the spatial patterns of snowfall for events with a northwesterly trajectory.

The increase in snowfall totals in portions of the study area is tied to higher values of moisture and humidity, particularly in the low levels (Table 3.11). The difference is most pronounced and statistically significant ($p < 0.05$) with the sounding variables of thickness of the moist layer and lapse rate above the moist layer. The synoptic variables of precipitable water, 850 hPa relative humidity, 1000-500 hPa mean relative humidity, and 850 hPa mixing ratio also display significant ($p < 0.05$) differences, suggesting a more moist and saturated lower troposphere for events with a GLC. Significant differences in 500 hPa vorticity are also evident. Events without a GLC typically are accompanied by lower values of cyclonic vorticity, whereas events with a GLC display higher values of cyclonic vorticity. In GLC events, vorticity maxima at 500 hPa rotating through the base of the 500 hPa trough may help to raise the height of the low-level capping inversion, as suggested by Lackmann (2001) and Niziol (1989). Our sample, however, was restricted to events in which subsidence was observed at 700 hPa (e.g. no approaching troughs to provide synoptic-scale ascent). Interestingly, no significant differences in 700 hPa vertical velocity, 850 hPa temperature, or percent of events in the dendritic temperature range were noted.

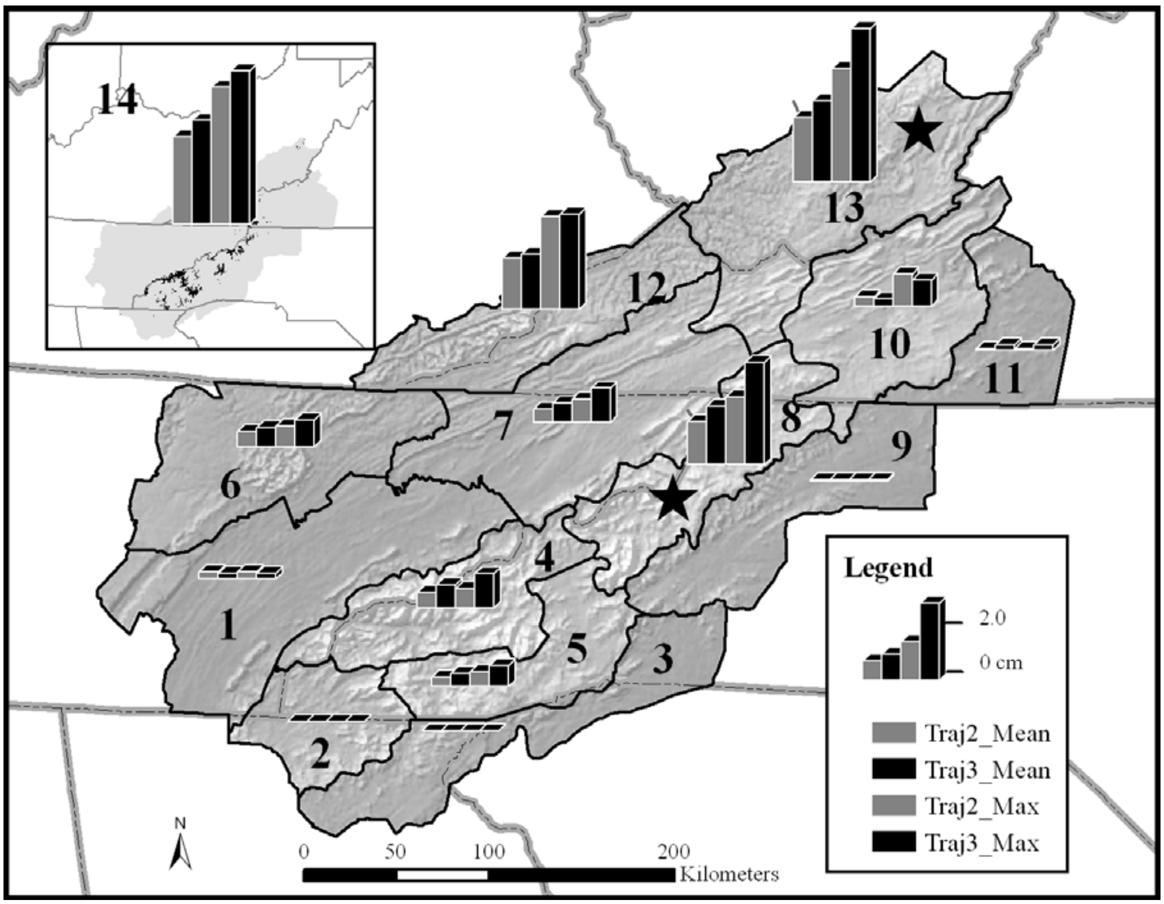

Figure 3.22. Composite mean and maximum snowfall totals between Trajectory Class 2 (no GLC) and Trajectory Class 3 (GLC) by snow region. Star denotes significance at $p < 0.05$.

Table 3.11. Mean sounding and synoptic field values and differences for Trajectory Class 2 versus Class 3.

Variable	Traj_2	Traj_3	Abs_Dif	T-Score	
Thickness of Moist Layer	140	202	61.61	2.61	*
Precipitable Water	0.22	0.26	0.04	2.34	*
500 hPa Vorticity	4.44	5.35	0.9	2.34	*
850 hPa Relative Humidity	73.1	79.3	6.14	2.32	*
1000 hPa Height	146	127	-19.59	-2.06	*
1000-500 hPa Mean Relative Humidity	54.4	57.9	3.43	1.96	*
850 hPa Mixing Ratio	1.7	2.0	0.26	1.87	*
850 hPa Thermal Advection	-7.0	-5.6	1.46	1.8	*
Lapse Rate Above Moist Layer	-3.7	-5.0	-1.34	-1.76	*
850 hPa Wind Direction	300.0	305.0	5.02	1.45	
850 hPa Theta-e	7.4	8.9	1.54	1.23	
850 hPa Temperature	-9.8	-8.8	0.99	1.22	
Avg Temp of Moist Layer	-11.5	-10.4	1.14	1.04	
700 hPa Vertical Velocity	20.0	21.1	1.11	0.55	
Top of Moist Layer	688	667	-21.51	-0.53	
850 hPa Wind Speed	14.2	14.0	-0.18	-0.26	
500 hPa Height	5392	5390	-1.95	-0.13	
Dendritic Temp Range	65	64	-0.01	-0.07	

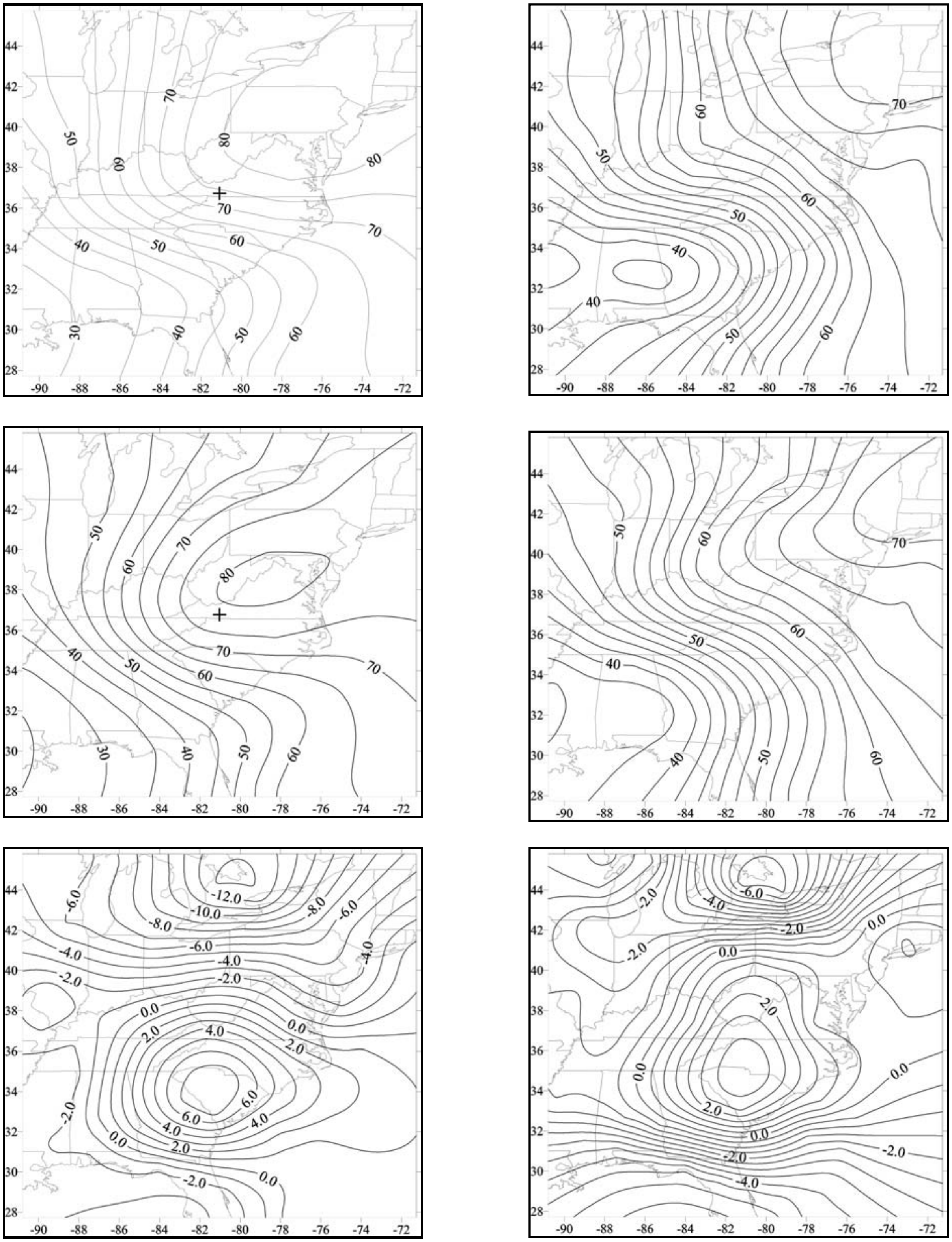

Figure 3.23. Composite plots of 850 hPa relative humidity (left) and 1000-500 hPa mean relative humidity (right) for Trajectory Class 2 (top panels), Trajectory Class 3 (middle panels), and mean difference between Class 3 and 2 (bottom panels).

The composite plots of 850 hPa relative humidity and 1000-500 hPa mean relative humidity for Trajectory Class 2 and 3 (Fig. 3.23) further highlight the significant differences in lower tropospheric relative humidity between the two trajectory classes. In both classes, the highest values of relative humidity are generally found to the northeast, but a distinct ridge of higher values extends southwest towards the Ohio Valley in association with Trajectory Class 3. The greatest mean differences between the two fields for both trajectory classes are located over western North and South Carolina, as higher values of lower tropospheric relative humidity in these areas in particular are tied to trajectories with a GLC. These composite plots also show that lower tropospheric relative humidity is higher across a large area downwind of the Great Lakes for trajectories with a GLC. The composite plots are also very similar to the composite 1000 to 500 hPa mean relative humidity plots for light and heavy events in section 3.3.3, in which a tongue of higher relative humidity extended from the maximum in the northeastern U.S. to the southwest in the Ohio Valley in the heavy events. Therefore, higher values of lower tropospheric relative humidity immediately upwind of the Southern Appalachians is consistently tied to heavier snowfall across the different analyses.

Interestingly, although higher values of lower tropospheric moisture and humidity across the Southern Appalachians occur in association with Trajectory Class 3, significant (at $p < 0.05$) increases in composite mean and maximum snowfall only occur in the High Country and Northern Plateau (regions 8 and 13). Mean elevation in these two snow regions is considerably higher than the others, with the exception of the High Peaks (snow region 14 or areas >1,220 m). Therefore, the increase in snowfall totals appears to be somewhat elevation-dependent, suggesting the important role orographic processes play in extracting lower tropospheric moisture. The greater regional topographic relief along the windward slopes of the High Country and Northern Plateau translates into a longer period of orographic lifting, as shown in modeled experiments in the South Wales hills of the U.K. (Choularton and Perry 1986). In addition, in a conditionally unstable lower troposphere, the orographic ascent along the windward higher elevation slopes may provide sufficient forcing for convection to develop, further extracting lower tropospheric moisture. Even though an increase in composite mean and maximum snowfall totals is evident for events with a GLC in the High Peaks (region 14), it is not so great as that seen in the High Country and Northern Plateau (regions 8 and 13). This may be related to the more efficient scavenging of low-level moisture at the High Peaks in Trajectory Class 2 (i.e. no GLC) when compared to stations at slightly lower elevations. No significant topographic features are present upwind of the study area in association with northwesterly trajectories, so it is safe to assume that no significant reduction in moisture due to topographic forcing occurs along either of the trajectories.

Additional findings related to the influence of the Great Lakes indicate that no significant difference ($p < 0.05$) exists in composite mean and max snowfall totals between events with percent ice cover in the top quartile (> 25 percent) and those in the bottom three quartiles (< 25 percent). This finding may be explained by the very small number of events with a GLC in which percent ice cover was greater than 75 percent ($n = 3$ or 3.3%) or even 50 percent ($n = 10$ or 11.1%). In these cases, much of the ice cover existed on Lake Superior and northern portions of Lake Michigan, leading to a much less pronounced effect on sensible and latent heat fluxes for trajectories crossing the southern end of Lake Michigan. Although extensive ice cover greater than 75 percent should greatly limit the sensible and latent heat fluxes and hence the lake-aggregate effect, it is difficult to use such a small sample of events to draw conclusions due to the complexity and diversity of necessary ingredients for NWFS events.

3.6.6. Comparative Air Trajectories

The spatial patterns of snowfall across the Southern Appalachians in association with different trajectories are best illustrated by comparing air trajectories for two samples of events: a) ones in which the highest totals occurred in the High Country (region 8), and b) ones in which the highest totals occurred in the Northern Plateau (region 13). These regions were chosen for comparative purposes due to the significant increase ($p < 0.05$) in composite mean and maximum snowfall totals for events with a northwesterly trajectory and a GLC. When considerably more snow fell in the High Country (top quartile of snow region 8:13 ratio), composite air trajectories for this region appeared to coincide approximately with the long axis of Lake Michigan (Fig. 3.24), whereas farther north, air trajectories were less favorable for lake interaction (e.g. across lower Michigan). Snowfall was observed along the windward slopes and higher elevations throughout the study area, with the greatest amounts centered over the High Country. For events in which the highest totals occurred in the Northern Plateau (bottom quartile of snow region 8:13 ratio), air trajectories were most favorable for lake interaction in that respective snow region (Fig. 3.24). To the south, only trace amounts of snow fell in the higher elevations—with no snowfall occurring elsewhere—in association with air trajectories well west of the western Great Lakes.

Measures of lower tropospheric relative humidity represent some of the most significant differences in synoptic fields between the bottom quartile and top quartile of the ratio of mean snowfall between the two snow regions (Table 3.12). These same fields also showed the greatest differences between Trajectory Class 2 (no GLC) and Trajectory Class 3.1, 3.2, and 3.3 (GLC) and are consistent with a moistening of the lower troposphere as a result of sensible and latent heating from the lake surfaces. Air trajectories, however, certainly are not the only factor influencing the spatial patterns of snowfall in these cases, as indicated by the significant differences in 1000 and 500 hPa heights, 500 hPa vorticity, and 850 hPa thermal advection. In particular, NWFS events in the bottom quartile (i.e. no snow in the High Country or region 8) of events with a GLC are tied to higher 500 and 1000 hPa heights, and lower values of 500 hPa vorticity. In these events, not only is the lower troposphere much drier, but the 500 hPa trough is displaced farther to the north. In addition, cold air advection at 850 hPa is significantly greater, which is tied to stronger subsidence in the lower troposphere. Therefore, the characteristics of the air mass being advected into the High Country (region 8) are not conducive for snowfall development, and may be partly explained by an air trajectory that is less favorable for interaction with the southern portion of Lake Michigan.

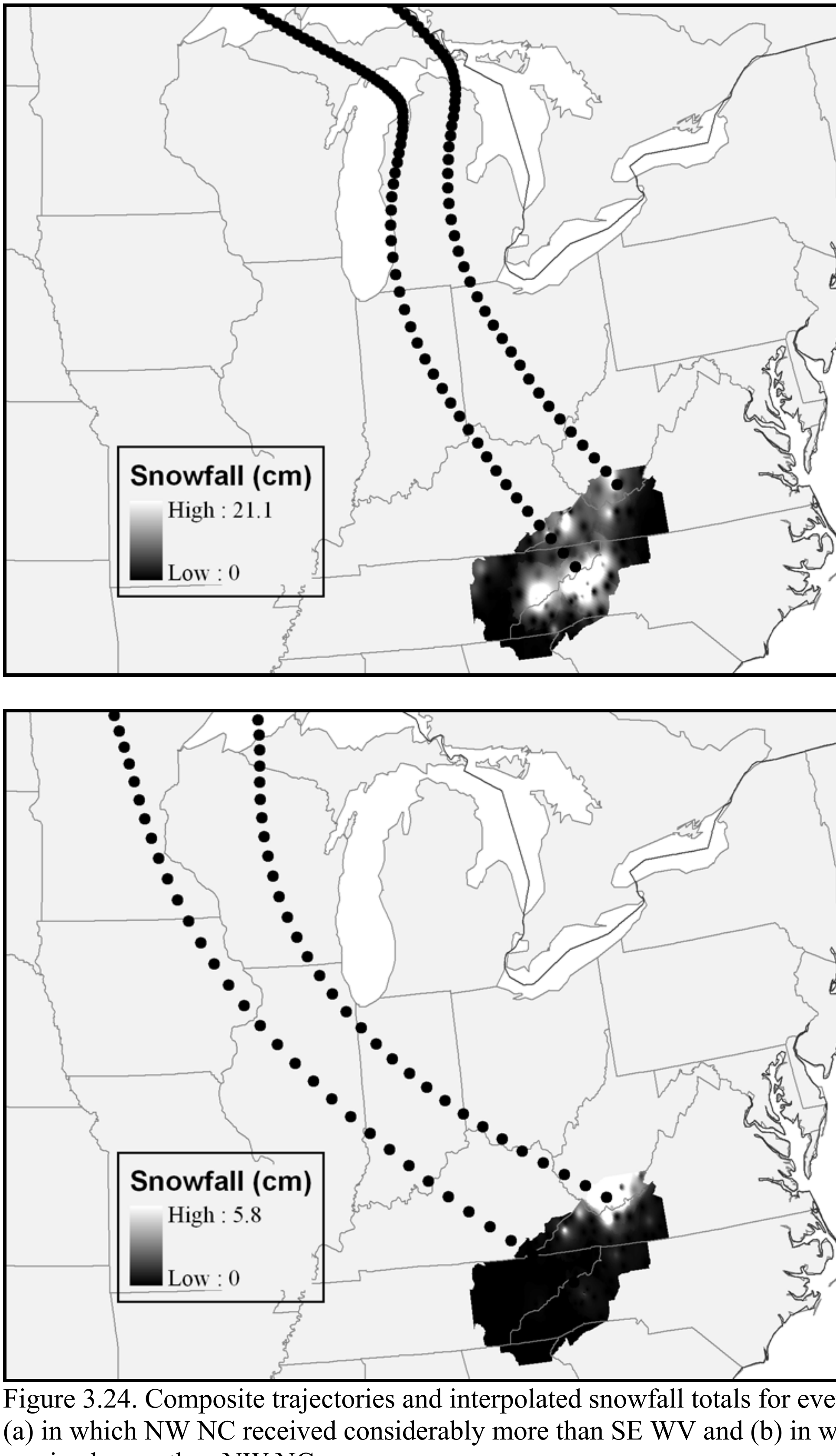

Figure 3.24. Composite trajectories and interpolated snowfall totals for events with a GLC (a) in which NW NC received considerably more than SE WV and (b) in which SE WV received more than NW NC.

Table 3.12. Mean synoptic field values over the High Country (region 8) for events with a GLC in the bottom vs. top quartile of the mean snowfall ratio between the High Country and Northern Plateau (region 13).

	Bottom Quartile	Top Quartile	Abs_Dif	T-Score	
1000-500 hPa Mean Relative Humidity	45.4	57.9	12.5	4.08	*
500 hPa Height	5517	5395	-122	-3.81	*
500 hPa Vorticity	3.03	5.43	2.41	3.4	*
850 hPa Thermal Advection	-8.0	-4.6	3.4	3.18	*
1000 hPa Height	176	131	-45	-2.99	*
850 hPa Relative Humidity	63.0	76.0	13.1	2.89	*
850 hPa Temperature	-5.5	-9.0	-3.5	-2.46	*
850 hPa Theta-e	12.3	8.6	-3.8	-1.98	*
500 hPa Relative Humidity	22.5	29.8	7.3	1.93	*
700 hPa Vertical Velocity	27.0	21.1	-5.9	-1.51	
850 hPa Mixing Ratio	1.92	1.86	-0.06	-0.31	

CHAPTER IV
SUMMARY AND CONCLUSIONS

4.1. Summary

This dissertation analyzed the synoptic climatology of Northwest Flow Snowfall (NWFS) in the Southern Appalachians using a combination of snowfall data and synoptic-scale atmospheric data for the period 1950 to 2000. A general snowfall climatology was constructed from these data, with particular emphasis placed on the intra-regional patterns across the Southern Appalachians. Ordinary (i.e. light) events were compared with exceptional (i.e. heavy) events in order to better understand the synoptic patterns and synoptic field values associated with heavy NWFS. Additionally, events with greater differences in snowfall by elevation (i.e. markedly more snow at higher elevations) were compared with those that were more widespread at all elevations. Events that displayed spillover snowfall well downwind to leeward slopes were also compared with those events in which snowfall was largely confined to the higher elevation windward slopes. Lastly, an analysis of antecedent upstream air trajectories and vertical profiles of moisture and temperature provided a more detailed perspective on NWFS events, with particular emphasis on the influence of the Great Lakes. The following sub-sections provide a brief summary of the major findings of this dissertation.

4.1.1. General Characteristics

NWFS events occur frequently in the Southern Appalachians from late fall through spring and account for over 50 percent of mean annual snowfall in parts of the region. Mean annual NWFS totals range from less than 10 cm (4 in) on lower elevation leeward slopes to as much as 77 cm (30 in) on higher elevation windward slopes, indicating a significant topographic influence on the spatial patterns of snowfall. The overwhelming majority of mean annual snowfall totals at lower elevations are tied to synoptic-scale ascent, whereas this percentage is considerably lower at higher elevations due to the high frequency of orographic snowfall in periods of synoptic-scale subsidence. The frequency of NWFS events shows a distinct maximum in January across the entire region, although accumulations per event are typically highest in March.

4.1.2. Snowfall Intensity

A comparison of NWFS events on the basis of snowfall intensity revealed that heavy events were distinguished from light events by the following:

a. Colder temperatures (at 850 hPa and as evidenced by lower 1000 to 500 hPa thicknesses) and higher relative humidity (at 850 hPa, 500 hPa, and 1000 to 500 hPa). The colder temperatures are responsible for the higher relative humidity since no significant differences in precipitable water or 850 hPa mixing ratio are evident. Colder temperatures and higher relative humidity also suggest that snowfall development is more efficient. Moist layer temperatures are more likely to be in the dendritic temperature range (-14 to -17˚ C), where ice crystal growth is most efficient.

Additionally, sublimation of falling snowflakes is reduced when low-level relative humidity (i.e. at 850 hPa) is higher.

b. Thicker moist layer, the top of which is found at a higher level (i.e. lower pressure) in the troposphere. Therefore, not only are temperatures in the moist layer more likely to be within the dendritic growth range, but a thicker moist layer provides a longer period of ice crystal growth and allows for greater vertical development of any orographically-induced convection.

c. Longer event durations, which can be tied to a longer period of upslope flow. Larger slower moving synoptic-scale disturbances are associated with prolonged periods of low-level northwest flow, contributing to higher snowfall totals. Likewise, a deep and persistent 500 hPa trough in the absence of a synoptic-scale disturbance with air trajectories extending downwind from the Great Lakes may also result in long event durations and an extended period of accumulating snowfall.

d. Relatively weaker synoptic-scale sinking motions at 700 hPa, which reflect a synoptic environment more conducive for orographic snowfall when compared with the light events (i.e. stronger 700 hPa subsidence). Weaker synoptic-scale subsidence aids in orographic and convectional lifting by lessening the mid-level cap. The heaviest snowfalls are tied to strong synoptic-scale ascent and large or slow moving systems, resulting in longer event durations.

e. Lower 500 hPa heights and higher values of 500 hPa vorticity, which are indicative of colder temperatures and deeper mid- and upper-level troughing. Moist layer temperatures in the dendritic temperature range and the greater thickness of the moist layer are associated with lower 500 hPa heights and higher 500 hPa vorticity. The 500 hPa vorticity advection in association with a vorticity maximum may help to raise the height of the low-level capping inversion, as suggested by Niziol et al. (1995) and Lackmann (2001) in periods of heavy lake-effect snow to the lee of the Great Lakes.

f. Stronger upper-level (200 hPa) divergence, resulting in relatively weaker values of synoptic-scale subsidence that promote a less capped environment that encourages heavier orographic snowfall.

g. Lower surface pressures (i.e. 1000 hPa heights), with a composite surface cyclone positioned farther to the south and off of the North Carolina coast. Such a position favors more northerly trajectories that extend downwind from the eastern Great Lakes of Huron and Erie.

h. Weaker low-level (850 hPa) cold air advection, which correlates with the presence of colder temperatures as a NWFS event commences (i.e. strong cold advection implies air that is too warm to support snow during the beginning stages of the even).

4.1.3. Influence of Elevation

4.1.3a. Snowfall at High Peaks

Events with the greatest increases in snowfall by elevation were distinguished by the following:

a. Warmer temperatures (at 850 hPa, higher 1000 to 500 hPa thicknesses), suggesting boundary layer temperatures across the region too warm for significant snowfall, except in the High Peaks (Region 14).

b. Higher water vapor content (precipitable water, 850 hPa mixing ratio), but slightly lower relative humidity (at 850 hPa and 1000 to 500 hPa), which can be tied to warmer temperatures (i.e. warm air has a greater capacity to hold water vapor than cold air) but less efficient use of available water vapor for snowfall development. Lower 850 hPa relative humidity results in greater sublimation of falling snowflakes, thereby reducing the precipitation efficiency. Likewise, lower 1000 to 500 hPa mean relative humidity is tied to greater entrainment of dry air and less efficient use of water vapor.

c. Higher 500 hPa heights, which are indicative of warmer temperatures and less pronounced mid- and upper-level troughing.

d. Shorter event durations, which result from a brief window of overlap between arriving sub-freezing air and departing moisture. Relatively warm boundary layer temperatures and short event durations largely limit snowfall accumulations to the higher elevations.

e. Greater instability between 850 and 500 hPa, indicating a synoptic environment more conducive for greater vertical development of orographically-induced rising motions.

4.1.3b. Widespread at All Elevations

Events with greater differences in snowfall by elevation in which widespread snowfall occurred at all elevations were distinguished by the following:

a. Lower relative humidity (from 1000 to 500 hPa and at 500 hPa) indicating reduced precipitation efficiency. Interestingly, very little difference is noted at 850 hPa, suggesting the importance of upslope flow in maintaining high relative humidity despite lower mean values throughout the lower troposphere.

b. Stronger synoptic-scale subsidence, suggesting that orographic lifting is the primary mechanism responsible for heavier snowfall in the High Peaks relative to the lower elevations. The upslope flow is not sufficient to produce significant snowfall at lower elevations and on leeward slopes, but on higher elevation windward slopes amounts may be significantly greater.
c. Greater instability between 850 and 500 hPa, again indicating a synoptic environment more conducive to the development of orographic snowfall.

4.1.4. Spillover Effects

Pronounced spillover snowfall into the leeward slopes occurred in the presence of:

a. Weaker to near neutral synoptic-scale sinking motions at 700 hPa and greater upper-level divergence (at 200 hPa), which can be tied to stronger mid- and upper-level dynamics along the leeward slopes and into the adjacent Piedmont. Synoptic-scale dynamics are needed in order to counteract the downslope flow and produce measurable snowfall along the leeward slopes.
b. Higher relative humidity (from 1000 to 500 hPa, at 850 hPa, at 500 hPa), resulting in efficient use of water vapor within the moist layer and limited sublimation of falling snowflakes below 850 hPa.
c. Longer event durations, which can be tied to a synoptic-scale disturbance tracking more closely to the region, resulting in a period of synoptic support and a prolonged period of orographic snowfall on the higher elevation windward slopes.
d. Weaker cold air advection, suggesting temperatures cold enough for snow when the event commences. Antecedent temperatures are particularly important at lower

elevation leeward slopes, where warm boundary layer and ground temperatures can quickly melt any spillover snowfall.

4.1.5 Trajectory Analysis

NWFS events analyzed in this dissertation displayed considerable variability in 72-hour backward air trajectories. In fact, over a quarter of the sample of events did not exhibit northwesterly trajectories. The remaining events displayed northwesterly trajectories, with almost half (47.1%) of all NWFS events in the sample interacting with portions of the Great Lakes region for a minimum of six hours during the analyzed 72-hour backward air trajectories. Trajectories passing over the western Great Lakes of Superior and Michigan were most prevalent, with the southern section of Lake Michigan seeing the highest frequency. Trajectories with higher composite mean and maximum snowfall totals are characterized by air parcels lifted from near the surface in weak to moderate cold air advection, a deep moist layer, and high values of 850 hPa relative humidity. In contrast, trajectories with lighter totals are tied more strongly to subsidence, stronger cold air advection, a thinner moist layer, and considerably lower values of 850 hPa relative humidity. These results are consistent with the results from analyses of snowfall intensity, the influence of elevation, and spillover effects. Heavy events, widespread snowfall at all elevations, and spillover events were all tied to high values of lower tropospheric relative humidity and weak cold air advection.

Northwesterly trajectories with a Great Lakes connection were also associated with higher values of moisture and humidity in the lower troposphere and enhanced snowfall totals at higher elevation windward slopes. This is particularly the case when comparisons are made between northwesterly trajectories, or when Trajectory Class 2 is compared with Trajectory Class 3.1, 3.2, and 3.3. An increase in composite mean and maximum snowfall totals is evident across the Southern Appalachians, with significant ($p < 0.05$) differences for the High Country of North Carolina (Region 8) and the Northern Plateau (Region 13) of southeastern West Virginia. This increase in snowfall totals is also associated with higher values of lower tropospheric water vapor and relative humidity, suggesting that the Great Lakes may act to enhance snowfall totals considerable distances downwind by moistening the lower troposphere. The exact air trajectories in association with events that display a Great Lakes connection also appear to influence the spatial patterns of snowfall across the Southern Appalachians. In particular, the heaviest snowfall occured in the High Country when composite trajectories extended southeastward from the long axis of Lake Michigan into this respective snow region, whereas the heaviest snowfall shifted to the Northern Plateau when upstream antecedent trajectories with a Great Lakes interaction were limited to extreme northern portions of the study area.

4.2. Conclusions

Several important conclusions about NWFS in the Southern Appalachians can be drawn from the results of this dissertation. Together, these analyses highlight the importance of the following characteristics in leading to the development of significant NWFS:

a. Topography, as snowfall mainly occurs on higher elevation windward slopes, with accumulations rarely observed on lower elevation leeward slopes. Orographic lifting allows snowfall to develop in the presence of synoptic-scale subsidence and enhances snowfall in association with synoptic-scale ascent.

b. Higher values of lower tropospheric relative humidity, which increase the efficiency of ice crystal growth and minimize sublimation of falling snowflakes. A deeper moist layer extending to a higher height (i.e. lower height and colder temperature) also aids in snowfall development by providing a longer in-cloud residence period and increasing the probability of moist layer temperatures falling within the dendritic growth range.
c. Stronger mid- and upper-level dynamics, which are tied to heavier and more widespread snow at all elevations and pronounced spillover snowfall on leeward slopes.
d. Longer event durations, which result in an extended period of accumulating snowfall.
e. Lower 500 hPa heights, which can be tied to a deeper 500 hPa trough and colder temperatures.
f. Antecedent upstream air trajectories with a Great Lakes connection, which are particularly important for snowfall enhancement when synoptic-scale subsidence is strong. Such trajectories are tied to higher snowfall totals along the higher elevation windward slopes, presumably due to the extraction of low-level moisture partly of lake origin via orographic processes.

The results of this dissertation have contributed to a better understanding of the NWFS events from a forecasting perspective. Three findings, in particular, are of significant relevance to the forecast process:

a. The importance of lower tropospheric relative humidity in differentiating between heavy and light NWFS events. A significant NWFS event is more likely when an air mass with high values of lower tropospheric relative humidity is advected into the region.

b. Moist layer temperatures within the dendritic growth range. Nearly the entire heavy NWFS events sample analyzed in this dissertation exhibited moist layer temperatures between -14 and -17 ˚C.

c. The importance of air trajectories in influencing snowfall patterns. Results indicate that trajectories with a Great Lakes Connection result in higher mean areal and maximum point snowfall totals along higher elevation windward slopes than other northwest trajectories.

4.3. Unanswered Questions and Future Work

This research raises many new questions about NWFS in the Southern Appalachians:

4.3.1. Vorticity Maxima

Perhaps one of the more vexing unanswered questions relates to the role of sub-synoptic scale vorticity maxima in enhancing snowfall in the presence of synoptic-scale subsidence. Due to the modest spatial resolution of the NCEP reanalysis dataset, it was not possible to evaluate the presence or significance of these features in this dissertation. However, forecast discussions from the National Weather Service Weather Forecast Offices frequently reference these vorticity maxima and suggest the cyclonic vorticity advection associated with them play an important role in producing NWFS in conjunction with orographic lifting. The cyclonic vorticity advection may not play a significant role in snowfall development per se, but may serve to lift the height of the mid-level capping inversion and aid in thickening the moist layer.

4.3.2. Low-level Wind Direction

Although the results of this dissertation found the mean 850-hPa wind direction to be approximately 300 degrees across all analyses, the importance of exact wind direction within the northwest quadrant (270 to 360 degrees) remains unclear. Due to the general southwest-to-northeast orientation of the Southern Appalachians, winds with a more westerly (i.e. closer to 270 degrees) or more northerly (i.e. closer to 360 degrees) low-level component are considerably less than perpendicular, with local and regional exceptions. Therefore, future research investigating the snowfall patterns and intensities tied to more specific wind directions (i.e. 270 to 285 degrees) would aid in answering some of these questions. Additionally, the extent and frequency to which west-southwest or even southwest low-level flow can contribute to orographic snowfall in the Southern Appalachians deserves further scrutiny.

4.3.3. ***Cloud Microphysical Environment***

Due to limitations in the temporal resolution of vertical soundings (i.e. limited to 0 Z and 12 Z) and the density of radiosonde (upper-air) stations, additional research is needed to develop a better understanding of the cloud microphysical processes associated with NWFS on the higher elevation windward slopes. Huntington (HTS), WV, though perfectly upstream of northern portions of the study area (i.e. Northern Plateau and New River Valley), is considerably distant from other regions of the Southern Appalachians. Therefore, the uncertainty introduced by the limited temporal resolution and distance should be viewed with caution. Further research utilizing field studies and coordinated radiosonde releases along the higher elevation windward slopes for a sizeable sample of NWFS events would significantly improve our understanding of the vertical profiles of moisture and temperature within the moist layer. The use of vertically-pointing radar systems and disdrometers could also provide important information on the depth of the moist layer, abundance of water versus ice, and precipitation particle diameter and fall speed.

4.3.4. Trajectory Analysis

Although this dissertation has found that air trajectories with a Great Lakes connection aid in enhancing snowfall totals under certain circumstances, the importance of the duration of favorable air trajectories remains unclear. The trajectory classification scheme presented in this study evaluated antecedent upstream air trajectories at event maturation, which is only a snapshot in time. Over the course of a two to three day NWFS event, a wide range of trajectories is evident, signifying that a trajectory without a Great Lakes connection at the event maturation may develop one towards the end of the event (e.g. Trajectory Class 4 at T+12). Additionally, the actual source of the low-level moisture associated with NWFS remains poorly understood. In some cases (e.g. Trajectory Class 2), moisture appears to be of Pacific origin, whereas in others it can be inferred to be partly of Great Lakes origin. This study has focused solely on those NWFS events in which synoptic-scale subsidence was observed at event maturation, and therefore intentionally excluded those NWFS events with synoptic-scale ascent. A question remains, therefore, as to the degree of enhancement, if any, which may occur in NWFS events with synoptic-scale ascent and a Great Lakes connection. The modest (~2.5˚) resolution of the NCEP reanalysis data may also be a limitation in the study, particularly in resolving moisture and instability plumes that may extend downwind of the individual Great Lakes and into the Southern Appalachians.

Further research incorporating the higher resolution North American Regional Reanalysis (NARR) dataset might prove beneficial in resolving some of the mesoscale features associated with NWFS, such as the role of 500 hPa vorticity maxima and moisture and instability plumes in lifting the height of the low-level capping inversion. Additionally, case study analyses and high resolution modeling of selected NWFS events could also provide valuable insight into the mesoscale orographic and convectional processes associated with NWFS events. In a high resolution numerical model such as the WRF, the influence of the Great Lakes could be turned “off” by completely freezing the lakes over in a simulated manner. Likewise, the orographic influence could also be turned “off” by removing the Southern Appalachians (Hollaway and Lackmann 2005).

4.3.4. Synoptic Classification

The need for a comprehensive synoptic classification of NWFS events remains, as little is known about the frequency and variability in synoptic patterns. Further work building upon the preliminary classification of Lee (2004) would improve the conceptual understanding of NWFS. A comprehensive synoptic classification would also allow for comparisons of snowfall and synoptic field values to be made between ordinary and exceptional events within a specific synoptic class. For example, snowfall events tied to a surface cyclone tracking west of the Southern Appalachians (i.e. an Ohio Valley storm track) can be analyzed by snowfall intensity in order to better understand the synoptic characteristics associated with exceptional (i.e. heavy) snowfall. Likewise, snowfall patterns and synoptic field values could also be compared across different synoptic classes.

4.3.5. Spatial Patterns of Snowfall

Although it is clear that topography strongly controls the spatial patterns of snowfall in periods of low-level northwest flow, important questions remain as to the importance of exact low-level wind direction (section 4.3.2) and synoptic class (section 4.3.4). Various spatial interpolation techniques, such as GIS-based regression modeling (e.g. Perry and Konrad 2006) could be used to evaluate the finer-scale spatial patterns of snowfall, complementing the use of larger-scale snow regions in this dissertation. Satellite imagery analyses also represent a potentially useful approach to understanding the spatial patterns of snowfall for events with no antecedent snow cover, so long as clear skies and cold temperatures allow imagery to be acquired soon after the ending of an event. In either case, maps of composite snowfall totals could be constructed by synoptic class and compared across different classes. Such an approach would provide forecasters with an understanding of the composite snowfall amounts and patterns tied to specific synoptic patterns.

4.3.6. Influence of Hemispheric Teleconnections

The 50-year snowfall dataset utilized in this dissertation provides an as of yet unexplored opportunity to evaluate the importance of hemispheric teleconnections in influencing NWFS in the Southern Appalachians. El Niño-Southern Oscillation (ENSO), the Pacific-North America (PNA) pattern, and the North Atlantic Oscillation (NAO) are three that previous work has shown to have important influences on winter climate patterns across the eastern U.S., including the Southern Appalachians. No studies, however, have addressed the relationship between NWFS and different hemispheric teleconnections. Developing a better understanding of these relationships can provide a better assessment of the expected snowfall patterns for medium (i.e. 6-14 days) and long-range (i.e. greater than 15 days) forecasts.

A related question involves the influence of global climate change on snowfall patterns in the Southern Appalachians. A warmer winter climate in eastern North America would likely limit the meager snowfall totals at lower elevations across the Southern Appalachians, whereas it is plausible that higher elevations may see an increase in NWFS due to higher water vapor content at warmer temperatures. Additionally, the enhancement due to the Great Lakes may increase as the lake surface warms and ice cover decreases, allowing greater sensible and latent heat fluxes. Future research is needed to test these hypotheses.

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