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AIRMASS MODIFICATION BY VERTICAL MOTION:

DISCUSSION AND CASE STUDY

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1. INTRODUCTION

"Can the airmass become saturated in time to warrant introducing a threat of precipitation into the forecast for a given forecast period? And even if it does become sufficiently saturated, will the airmass be too stable for measurable precipitation?" These are basic questions that are commonly asked by forecasters when faced with an initial airmass that is "dry" and stable. And many times, regardless of the time of year, and despite painstaking thoroughness in analysis and meteorological reasoning, forecasters are repeatedly surprised at the ability of an airmass to modify rapidly in response to organized, large-scale vertical motion.

Such airmass modifications are focused in two areas:

1. Moisture content

2. Vertical lapse rate (in combination with divergence or convergence)

Processes leading to, and impacts of, vertical motion have been a popular subject among authors of research and technical papers for many decades (e.g., Sutcliffe, 1947; Draghici and Davies, 1978; Trenberth, 1978; Durran and Snellman, 1987; Dunn, 1988; and Rogash, 1992). And in the most recent decade, with the advent of video tape technology, this has also become a popular topic for training tapes. One of the first was produced by the National Weather Service (NWS) Western Region (1980). This tape illustrated just how quickly and drastically atmospheric stability can be modified while moisture increases through a deep tropospheric layer.

There is abundant documentation of cases where these processes resulted in significant meteorological "surprises." For example, the NWS Southern Region (1981) documented a case study over North Texas and Southern Oklahoma. In this situation, despite observed and forecasted low surface dewpoints, despite analyzed and projected low mean humidity in the lower troposphere, despite a very stable airmass, and a situation where wind speed actually ***decreased*** with height, significant thunderstorms developed. The case study showed that in just 12-h, mean relative humidity increased by more than 50%. In the upper portion of the middle troposphere (700-500 mb), it more than doubled. The lifted index changed from +10 to -2, the K index changed from -2 to +26, and precipitable water doubled. Brenner (1979) documented a comparable example of a dramatic early September event which occurred over the desert southwest. Rapid destabilization in the middle troposphere resulted in an unexpected significant thunderstorm event in what was supposed to have been a very "dry" airmass.

Understanding concepts, processes, and relationships associated with vertical motion is fundamental to the science of meteorology. This discussion is not the proper forum to bring the reader up-to-date on existing or new research in this arena. For this, the reader is encouraged to begin by referring to the references for research and literature on this subject. However, this discussion will focus on providing some added insight into effects vertical motion has on the moisture profile and the vertical temperature lapse rate.

2. SIGNIFICANCE OF VERTICAL MOTION

Vertical motion is likely the most important aspect in diagnosing current weather conditions and more importantly, forecasting possible changes. The normal state of large-scale hydrostatic balance is an environment where the relatively large vertical pressure gradient force is offset by gravitational acceleration. However, since pressure surfaces are closely packed in the vertical through the troposphere, a small vertical displacement can result in a large pressure change. (Holton [1988] describes various vertical velocity approximations. Two primary contributions to quasi-geostrophic vertical velocity isolated in the Omega Equation are differential vorticity advection and Laplacian of thickness [temperature] advection. The reader is referred to this reference for details and equations describing vertical velocity estimates.)

3. VERTICAL MOISTURE MODIFICATION BY VERTICAL MOTION

Given that our ability to quantitatively predict large-scale vertical motion is at present limited (based on data density, data quality, and model skill), the importance of understanding how upward motion can modify airmass moisture content becomes crucial.

Large-scale flow is essentially horizontal in nature, with vertical flow typically less than a tenth of one percent of the horizontal flow. However, as pointed out by Doswell (1982), since this upward motion may be sustained for long periods, it can have significant effects on the vertical distribution of moisture as well as saturation.

Upward motion, with a maximum at middle levels, generally has three primary effects on an existing moisture field, namely: (1) mixing ratio increases: (2) relative humidity increases; and (3) precipitable water increases.

1. Mixing Ratio Increase:

Fig. 1 is an illustration of the equation of conservation of water mass, which states:

Change Of Mixing Ratio At A Point = Horizontal Advection + Vertical Advection

Horizontal motion does not alter the mixing ratio since the airmass is initially assumed to be horizontally uniform (e.g., horizontal winds continually replace air at any point with air that has the same amount of moisture, resulting in no change). However, since mixing ratio is assumed to be highest at lowest levels and decreases with height, mixing ratio at all levels (except the surface) will increase as a result of vertical advection, at least until an air mass becomes saturated.

2. **Relative Humidity Increase:**

Relative humidity is defined as :

Observed mixing ratio divided by saturation mixing ratio

Saturation mixing ratio, a function of temperature, decreases with height for a stable lapse rate. During periods of upward motion, mixing ratios increase at a given level above the surface. As a result, relative humidity increases much more rapidly than other moisture parameters.

3. **Precipitable Water Increase:**

Fig. 2 displays how precipitable water may increase during upward motion. Continuity of mass implies upward motion is associated with low level convergence and upper level divergence. Because mixing ratios are higher in the lower atmosphere, water vapor that is converged inward at lower levels is not fully compensated by that which is diverged outward at higher altitudes. This results in a net increase in precipitable water.

Even slight, but persistent upward motion can alter moisture distributions substantially, particularly in the middle troposphere. An example is shown in Fig. 3. MacDonald (1975) also illustrated this concept by assuming a large, horizontally homogeneous airmass in which mixing ratios decreased with height. Given this initial environment, relative humidity at 500 mb of 30% undergoing upward motion of 1 microbar per second increased to 100% in a little over 24-h. Similarly, Doswell (1982) noted that maintaining vertical motion of only 5 centimeters per second for 24-h resulted in a net vertical lift of more than 4 km!

4. VERTICAL LAPSE RATE MODIFICATION BY VERTICAL MOTION

Large-scale upward motion can be a significant contributor to airmass destabilization. This mechanism is especially important with regard to severe convective storm development as it relates to the elimination of an inversion cap in an atmosphere that is already convectively unstable. Daytime heating from below may sometimes be sufficient to eliminate the inversion (Doswell, 1982). Doswell emphasizes that in most cases there is a clear relationship between severe thunderstorms and at least one source of upward motion (fronts, short-wave troughs, etc.). It is this correlation that strongly suggests that lifting eliminates the inversion.

Changes in stability will be amplified due to stretching (or compression) when layers undergo vertical motion. This is primarily true for layers of significant thickness. It is also important to emphasize that the more stable the layer is initially, the greater the potential ***change*** in stability will be due to lifting and stretching. This results from large-scale lifting driving the lapse rate toward moist adiabatic (note that the moist adiabatic lapse rate approaches the dry adiabatic lapse rate at high altitudes where moisture content is negligible). Therefore, a layer which is already stratified nearly adiabatic will undergo a lesser change, and a layer that is very stable initially will modify rapidly by lifting and stretching processes.

Changes in lapse rate can be inferred from the relationship between convergence/divergence, vertical motion (ascent/descent), and thickness (stretching/shrinking) (see *AWS*, 1969). In the lower atmosphere, effects on lapse rate from ascent and descent without convergence/divergence are that thickness will increase (stretch) for ascent and decrease (shrink) for descent. These are the same effects, respectively, that occur for convergence and divergence without vertical motion. The fact that these are the same for the lower atmosphere has a practical significance, since the earth's surface serves as a constraining boundary. This boundary forces convergence to accompany ascent, and divergence to accompany descent. Therefore, in the lower atmosphere, effects of convergence and ascent (stretching thickness) on lapse rate are to destabilize (assuming initially stable). Effects of divergence in combination with descent (shrinking) are to stabilize.

At upper levels, the relationship between convergence/divergence, vertical motion (ascent/descent), and thickness (stretching/ shrinking), and inferences with regard to change in lapse rate, can be identified with 12 combinations (*AWS*, 1969). These are shown in Table 1.

Category Number	Divergence	Vertical Motion	Thickness	Change In Lapse Rate (If Initially Stable)
1	Divergence	Descent	Shrinks	More Stable
2	Divergence	Ascent	Shrinks	More Stable
3	Divergence	Ascent	Stretches	Less Stable
4	Convergence	Descent	Shrinks	More Stable
5	Convergence	Descent	Stretches	Less Stable
6	Convergence	Ascent	Stretches	Less Stable
7	Divergence	None	Shrinks	More Stable
8	Convergence	None	Stretches	Less Stable
9	Divergence	Ascent	No Change	No Change
10	Convergence	Descent	No Change	No Change
11	None	Ascent	Stretches	Less Stable
12	None	Descent	Shrinks	More Stable

Table 1. Possible combinations of divergence, vertical motion, and thickness related to changes in lapse rate (from *AWS*, 1969).

Categories 7-12 are those where one parameter stays constant. Categories 2-5 are those that do not occur in the lower atmosphere where the earth's surface forms a fixed boundary.

5. DECEMBER 8-9, 1992, ANALYSIS AND DIAGNOSIS

The day shift on 8 December 1992 at WSFO Atlanta, Georgia, was confronted with an interesting forecast problem. The ATLFPCATL product based on the Limited Fine Mesh (LFM) 1200 UTC model forecasted a 60% Probability of Precipitation (POP12) for Atlanta (ATL) in the second forecast period (verifying 0000 UTC 10 December) and a 90% POP12 in the third period (verifying 1200 UTC 10 December). ATLFWCATL POP12 forecasts based on the Nested Grid Model (NGM) were 7% and 93%, respectively, for identical verification times.

The NGM 36-h forecast, valid 0000 UTC 10 December, of 500 mb heights and vorticity is shown in Fig. 4a; 700 mb height and mean relative humidity in Fig. 4b; and surface pressure and thickness in Fig. 4c. Just from these charts, one would not readily question the low NGM POP12 (7%) at this time, since the dynamics associated with the approaching short wave trough were forecast to occur west of Georgia.

The LFM 36-h forecast, not shown, moved the approaching trough faster, bringing the leading edge of dynamical forcing into Northwest Georgia and correspondingly increasing mean relative humidity at ATL to over 90%. This contributed to high POP12 (60%) at this time (the LFM POP06 for the same time was 70%).

Resolving this discrepancy, however, was considered to involve more than simply deciding which model's timing would be correct. The complicating factor with this particular situation was that the 1200 UTC 8 December sounding from Athens (Fig. 5) indicated a quite "dry" and stable airmass, especially below 500 mb. So additional issues became: (1) could the airmass become saturated over a deep layer in time to warrant introducing a precipitation threat into the forecast for the second period; and (2) even if it did become sufficiently saturated, would the airmass be too stable for measurable precipitation? For each issue, the outcome was highly dependent upon vertical motion.

Vertical velocities forecasted by the LFM and NGM from the 1200 UTC 8 December model initializations are provided in Table 2. The 00-h "observed" (model-analyzed) vertical velocities, obtained after the fact from subsequent NMCFRHT65 products, are shown for comparison, as are the corresponding POP12 forecasts from each model.

FORECAST HOUR AND VERIFICATION TIME (UTC)									
MODEL	00-h (8/1200)	06-h (8/1800)	12-h (9/0000)	18-h (9/0600)	24-h (9/1200)	30-h (9/1800)	**36-h (10/0000)**	42-h (10/0600)	48-h (10/1200)
LFM		-2.9	-1.2	-1.7	0.0	+1.1	**+7.0**	+3.2	+1.9
NGM		-1.5	-1.2	-0.4	+0.8	-0.6	**+2.4**	+12.3	+0.5
"OBSERVED" (00-h)	-0.8		-1.2		-0.7		**+1.2**		M
LFM POP12					0%		**60%**		90%
NGM POP12					0%		**7%**		93%

Table 2. LFM and NGM 700 mb vertical velocity forecasts (microbars per second) for ATL (AFOS Products ATLFRH65 and ATLFRHT65). Also shown are POP12 forecasts from each model (AFOS Products ATLFPCATL and ATLFWCATL, respectively). Model initialization time was 1200 UTC 8 December 1992. "Observed" (00-h) vertical velocities for 1200 UTC 8 December, 0000 UTC and 1200 UTC 9 December, and 0000 UTC 10 December are shown for comparison.

At the time of the 8 December 1200 UTC model initialization, 700 mb vertical velocity (VV) was model-analyzed as -0.8 microbars per second. Negative values of VV were forecast for the next 18-h (through 0600 UTC 9 December) by both the LFM and NGM. By 1200 UTC 9 December, VV values were forecast to be zero for the LFM and +0.8 microbars per second for the NGM. The models deviated in the sign of VV at 1800 UTC on 9 December, with the LFM forecasting VV of +1.1 microbars per second, and the NGM forecasting -0.6 microbars per second. This deviation no doubt reflected model storm motion timing differences (as mentioned above).

By 0000 UTC 10 December, the LFM VV forecast was significant (+7.0 microbars per second) while the NGM VV forecast was +2.4 microbars per second (followed by an impressive +12.3 microbars per second at 0600 UTC 10 December!).

Primary questions became: (1) Even if the LFM was accurate with its VV forecast, was this sufficient to saturate the airmass over a deep enough layer in time to warrant introducing a threat of precipitation into the forecast for the second period? and (2) Even if the airmass did become sufficiently saturated, would the airmass be too stable to produce measurable precipitation at ATL ***before*** 0000 UTC 10 December? As mentioned, the airmass over Northern Georgia was "dry" and stable (Fig. 5). Both the LFM and NGM forecasts indicated that VV would continue to be negative for another 18 hours, offering little hope for deep layer saturation resulting from upward motion before 0600 UTC 9 December. If valid, this meant that significant positive VV would probably be required in the 18-h period from 0600 UTC 9 December to 0000 UTC 10 December in order to produce sufficient airmass saturation and destabilization necessary for measurable precipitation before 0000 UTC 10 December.

However, from Table 2, significant positive VV was not forecast by either model until after 1800 UTC 9 December. The forecaster (author) reasoned that even if the forecasted values of VV did occur as predicted by the models after 0600 UTC 9 December, the significant positive VV would not be occurring long enough in that 6-h period from 1800 UTC 9 December to 0000 UTC 10 December to produce deep layer saturation ***and*** sufficient airmass destabilization. As a result, the forecast followed the reasoning that the major role of the forecasted large-scale upward motion, even at the rate of +7.0 microbars per second, in this crucial 6-h period would be to "prepare" the airmass for precipitation that might occur ***after*** 0000 UTC 10 December. As a result, a POP of 20% was issued for ATL for the second period, and a POP of 90% was issued for the third period.

The forecast was a good one. The only precipitation that occurred at ATL in the second period was a trace, from widespread virga, and for a very short time from 2220-2225 UTC 9 December. (I only wish I had the courage to have gone even closer to the NGM POP12 of 7%!) Measurable rain was first observed between 0200 UTC and 0300 UTC 10 December.

The question remains: Did VV serve to "prepare" the airmass by gradually saturating it over a deep layer, as well as eventually causing it to destabilize so that measurable precipitation could occur after 0000 UTC 10 December?

Table 2 shows that model analyzed VV at 1200 UTC 9 December was still negative (-0.7 microbars per second). While not exactly forecasted by the two models from the 1200 UTC 8 model runs, this was consistent with 0% POP12 from each. Model analyzed VV at 0000 UTC 10 December was only +1.2 microbars per second, far less than the +7.0 microbars per second forecasted by the LFM, and half that which was forecasted by the NGM. However, while insufficient to cause saturation ***and*** destabilization required for measurable precipitation prior to 0000 UTC 10 December, it was sufficient to "prepare" the airmass for eventual precipitation.

Figs. 6-8 and the following discussion vividly depict how airmass modification occurred through 12-h intervals and how moisture and lapse rate modification were primarily responsible for delaying precipitation until ***after*** 0000 UTC 10 December.

0000 UTC 9 December:

Changes in the AHN sounding from 1200 UTC on the 8th to 0000 UTC on the 9th can be summarized by: (1) moistening above 500 mb, and (2) temperature lapse rate destabilization between 490 mb and 800 mb (also Showalter, K and Total Totals indices showed a less stable atmosphere). These changes were consistent with horizontal vergences and vertical motions as depicted on the 0000 UTC 9 December sounding (Fig. 6); namely, convergence near 480mb with rising air above and sinking air below. Sinking and dry air extended from 490 mb down to near 880 mb where lateral divergence was inferred. This sinking also produced a nearly dry adiabatic lapse rate to 680 mb. Air in the lowest 100 mb was converging and rising dry adiabatically to 880 mb. The strong temperature inversion between 880 mb and 800 mb represented the intersection between sinking air above and rising air below.

1200 UTC 9 December:

Temperature and moisture sounding changes from 0000 UTC on the 9th to 1200 UTC on the 9th can be characterized by: (1) decent of the region of nearly saturated air with a moist adiabatic lapse rate from 480 mb to about 540 mb, where lateral convergence was inferred; (2) sinking and dry air between 550 mb and 900 mb, where lateral divergence was inferred, with continued destabilization of the lapse rate between 550 mb and 780 mb (consistent with air sinking at the dry adiabatic lapse rate); and (3) little change below 900 mb. Corresponding vergences and vertical motions associated with these changes are sketched on the 1200 UTC 9 December sounding (Fig. 7). Winds increased with height up to about 200 mb, along with slight positive vorticity advection (PVA). However, much of the PVA impact at 500 mb was likely diminished due to the absence of moisture between 550 mb and slightly above 900 mb in combination with the strong inversion. Clearly, rising was confined to layers above 540 mb, while sinking was evident beneath 540 mb (consistent with 700 mb sinking shown in Table 2).

0000 UTC 10 December:

Temperature and moisture sounding changes from 1200 UTC on the 9th to 0000 UTC on the 10th displayed the identical trend as the previous 12-h tendency; namely: (1) the branch of nearly saturated rising air with moist adiabatic lapse rate continues to descend from 540 mb to near 650 mb, where lateral convergence was inferred; (2) the region of sinking and dry air shrinks and was confined between 700 mb and 900 mb, where lateral divergence was inferred, with dry adiabatic lapse rates between 700 mb and 830 mb, and (3) little change occurred below 890 mb. Arrows corresponding to vertical motion, and vergences at this time are sketched on the 10 December 0000 UTC sounding (Fig. 8).

Obviously, the trend from 1200 UTC on the 8th through 0000 UTC on the 10th was for the rising branch of vertical motion over AHN to lower from near 480 mb to about 650 mb and the sinking branch of dry air to remain well defined, but become shallower and eventually dissipate. Convergence initially just above 500 mb and corresponding divergence aloft represented the rising branch of the circulation, which lowered through time. Simultaneously, convergence initially just above 500 mb and corresponding divergence near 900 mb represented the sinking branch of the circulation which also lowered through time. The rising branch was "priming" the air mass for rain; however, the sinking branch decoupled the lower atmosphere and its large moisture source and prevented measurable rainfall prior to 0000 UTC on the 10th. At and after 0000 UTC on the 10th, the sinking branch became very thin and eventually vanished, thus allowing measurable precipitation to begin.

8. SUMMARY

Assuming that basic assumptions hold, large-scale upward motion can substantially and rapidly modify airmass moisture content and vertical lapse rate. Upward motion effects on lapse rate can be inferred from the relationship between divergence/convergence, vertical motion (ascent/descent), and thickness (stretching/shrinking). Changes in stability will be amplified due to stretching (or compression) when layers move vertically. Typically, large-scale upward motion is a significant contributor to airmass destabilization.

It is quite important to realize that even when the atmosphere appears to be "dry" and stable, rapid modification of moisture and lapse rate are possible given sufficient magnitude and duration of upward motion. Advection of temperature and moisture should certainly be among the main considerations for forecasters contemplating development of a precipitation regime. However, important consideration must also be given to how moisture and lapse rate may be modified by vertical motion. As part of this consideration, forecasters need to strive to determine: (1) whether the airmass can become saturated over a deep layer in time to warrant introducing a precipitation threat into a given forecast period; and (2) even if it did become sufficiently saturated, would the airmass be too stable to produce measurable precipitation within that time frame?

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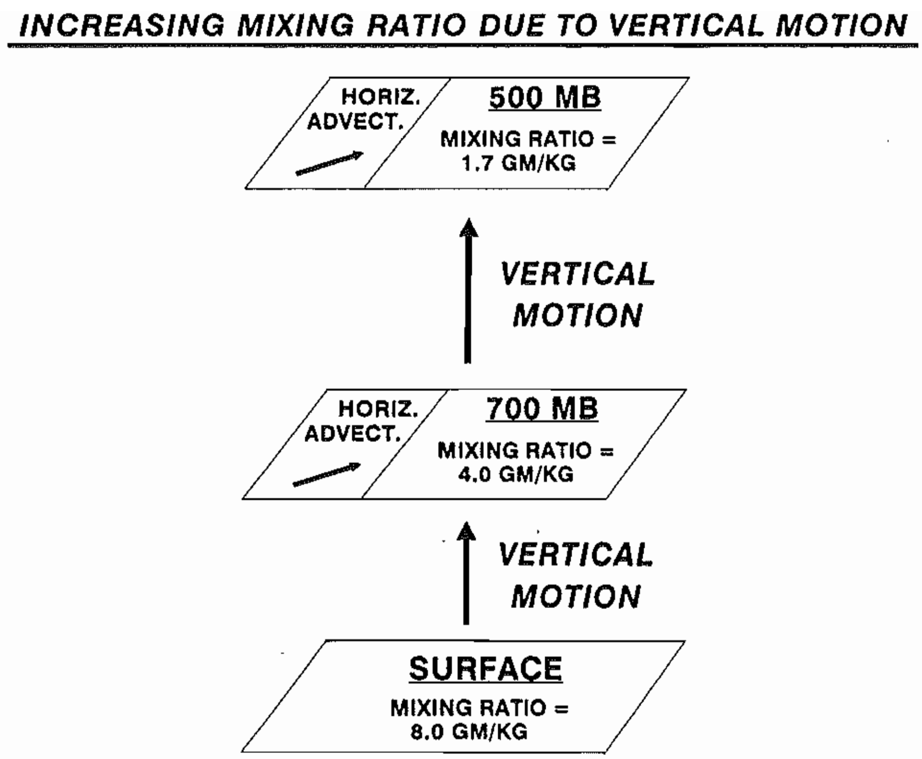

Figure 1 Increasing mixing ratio at all levels above the surface as a result of upward motion. (Assumes a horizontally uniform airmass and decreasing mixing ratios with height.)

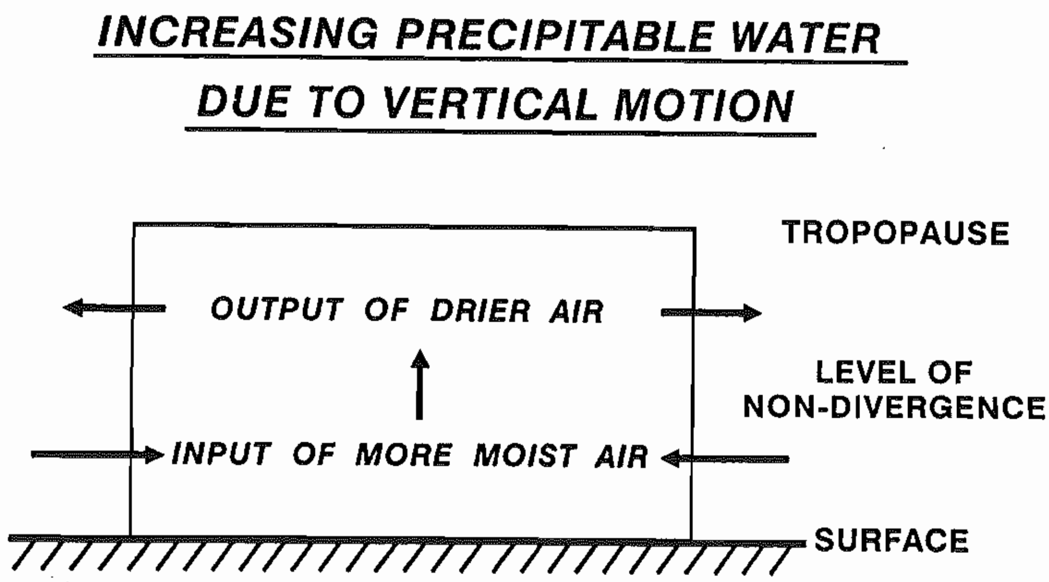

Figure 2 Increasing precipitable water as a result of upward motion. Moisture convergence is not compensated fully by moisture divergence.

VERTICAL TEMPERATURE AND MOISTURE PROFILE - 24-HR CHANGE

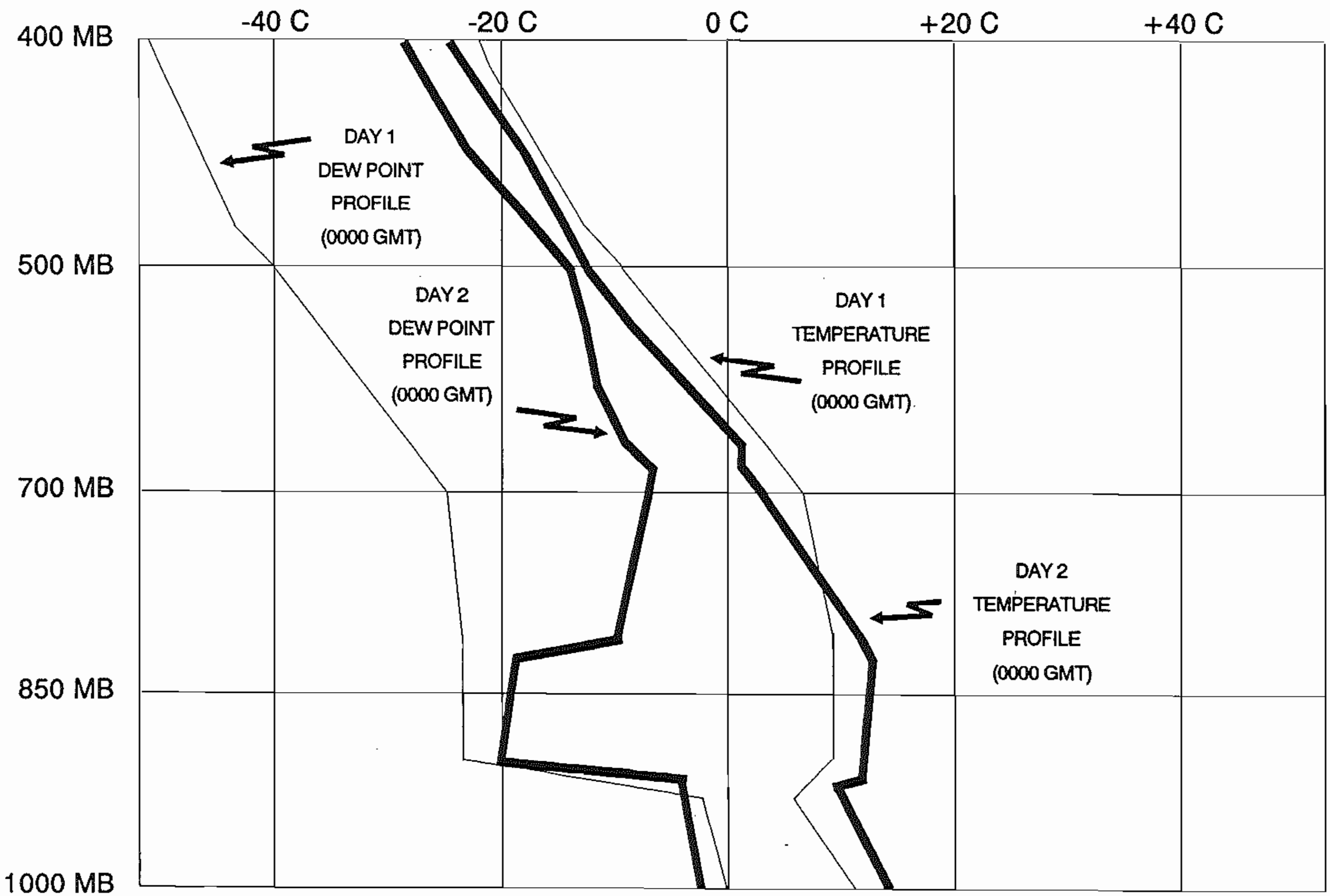

Figure 3 Illustration of vertical temperature and moisture modification over 24 hours due to upward motion.

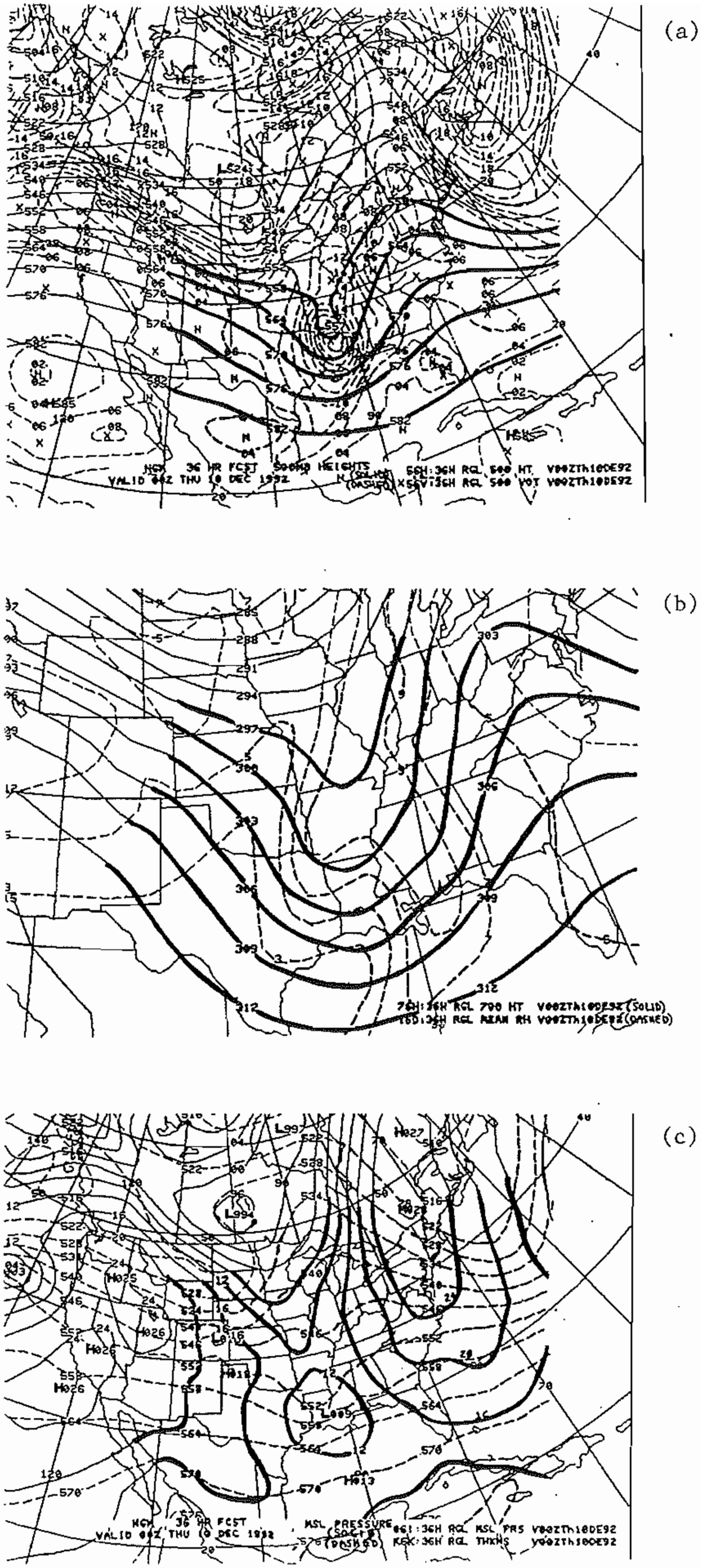

Figure 4. (a) NGM 36-hour forecast of 500mb heights (60 gpm intervals) and vorticity (dashed lines with 2 unit intervals). (b) NGM 36-hour forecast of 700mb heights (30 gpm intervals) and relative humidity (20% intervals). (c) NGM 36-hour forecast of mean sea-level pressure (4 mb intervals) and 1000-500mb thickness (dashed lines with 60 gpm intervals).

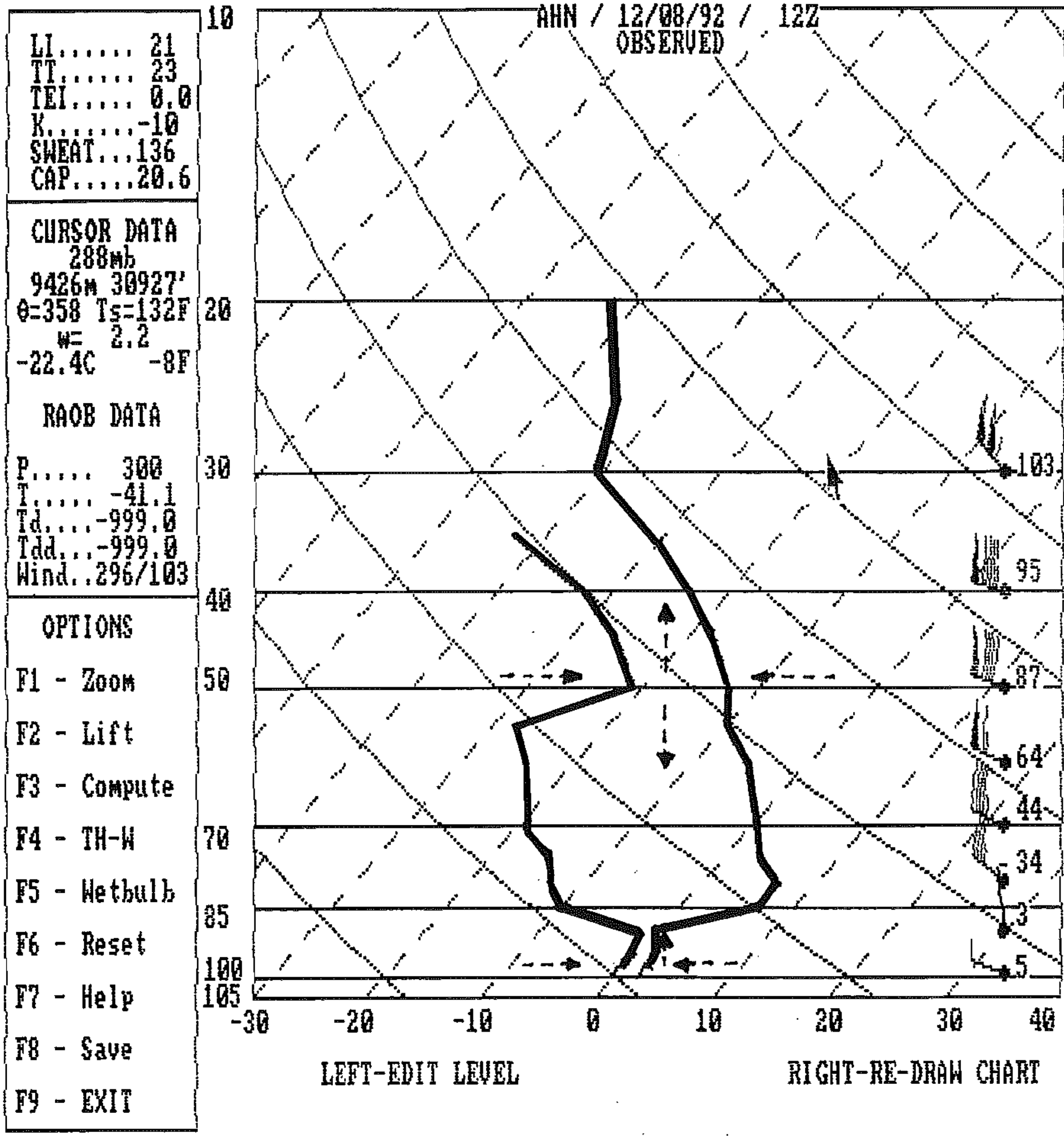

Figure 5. Radiosonde sounding at Athens, GA, (AHN) for 1200 UTC December 8, 1992. Dashed or solid arrows indicated inferred vergences and vertical motions. Right curve is temperature, left curve is dewpoint temperature. Winds at specific levels are displayed at far right.

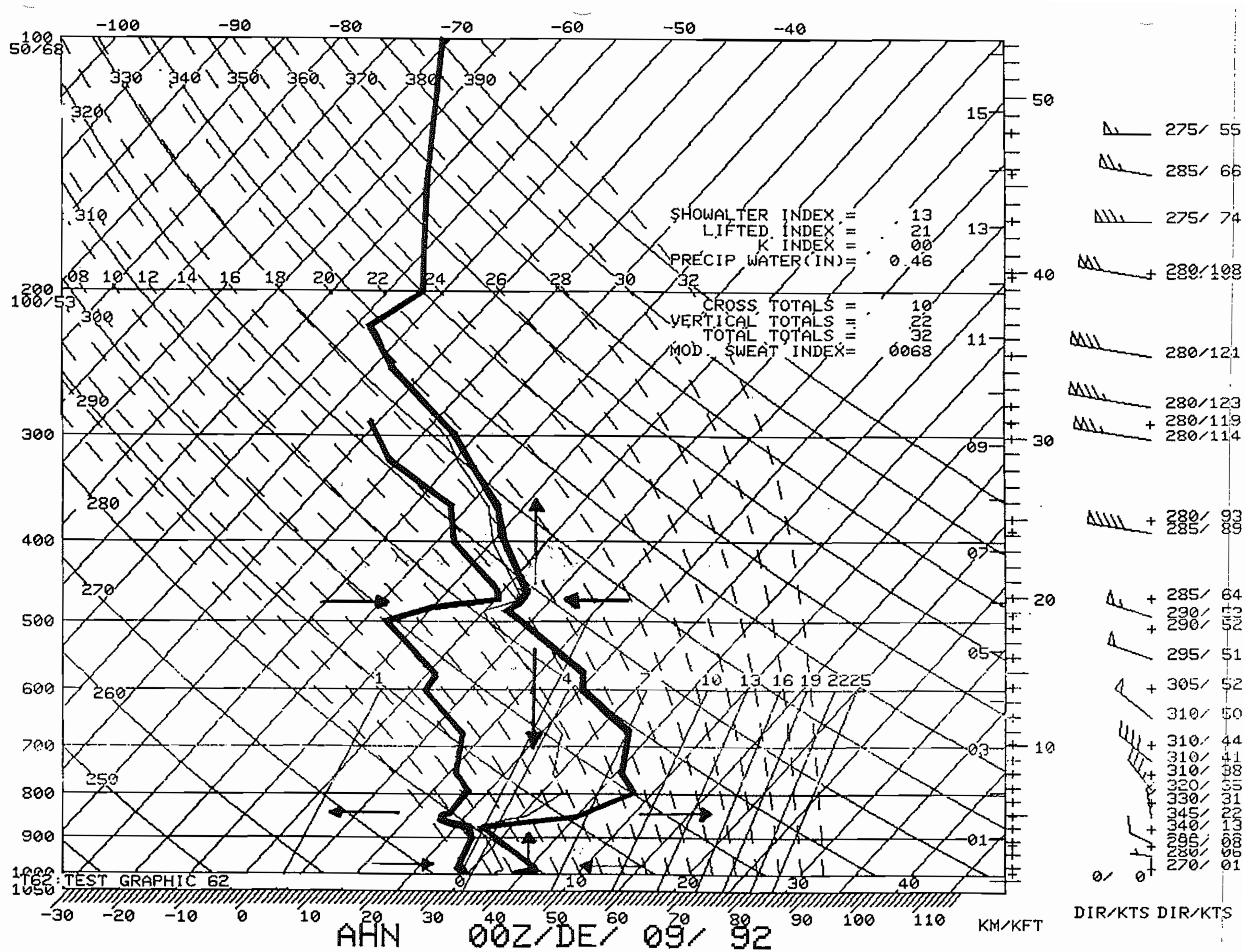

Figure 6. As in Fig. 5, except for 0000 UTC December 9, 1992. Stability indices and precipitable water are shown in the upper right of the diagram.

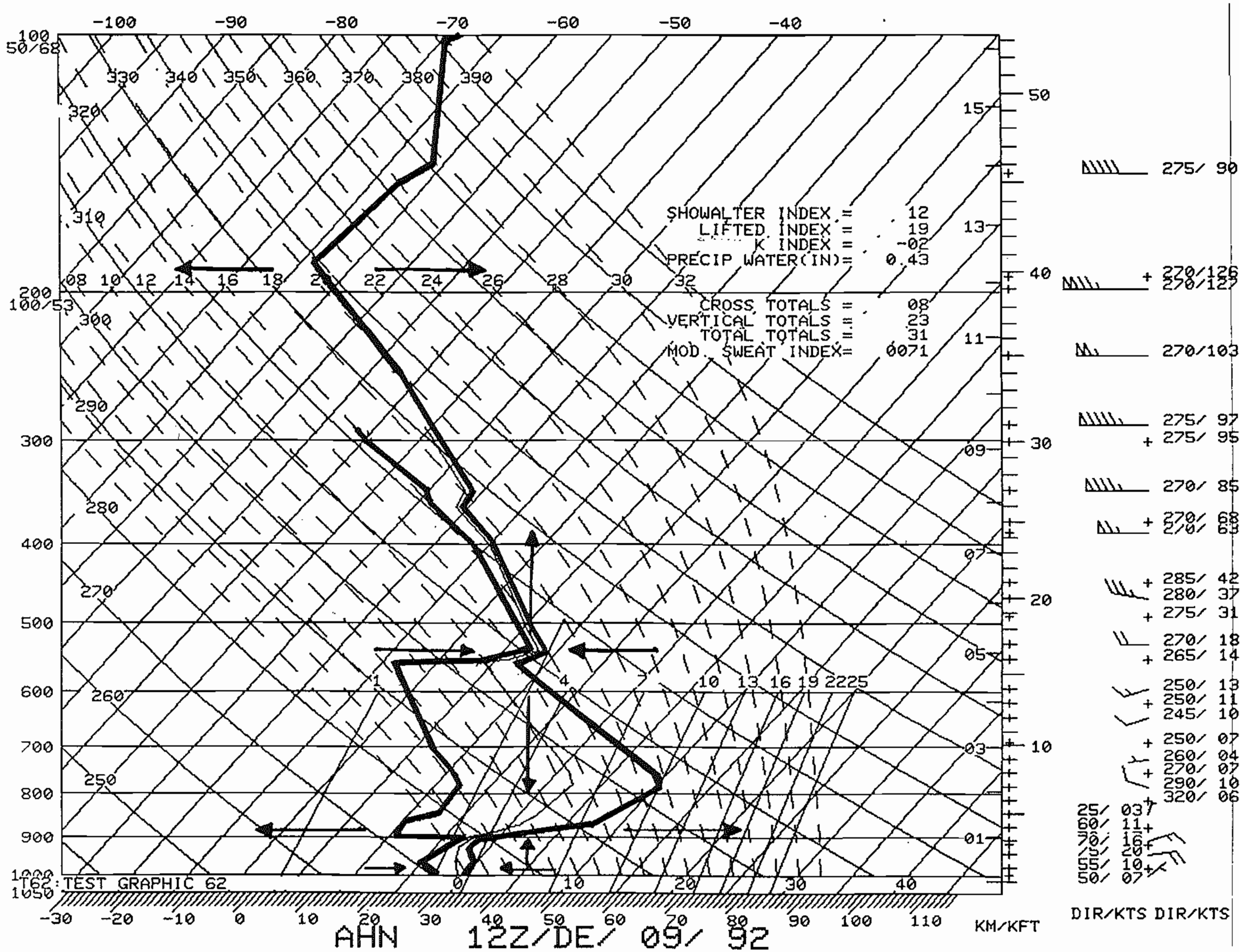

Figure 7. As in Figure 6, exc or 1200 UTC December 9, 1992.

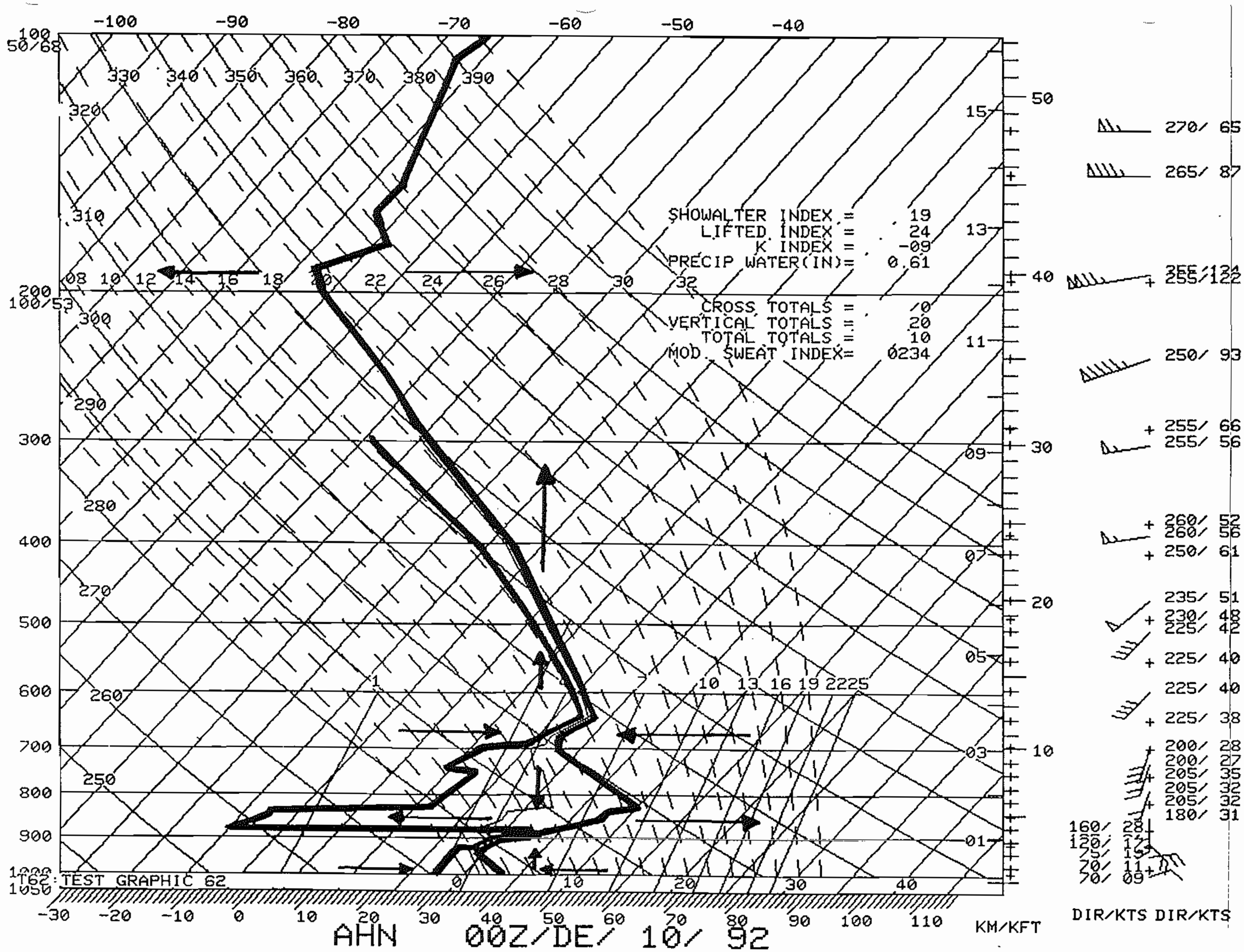

Figure 8. As in Fig. 6, except for 0000 UTC December 10, 1992.