Western Region Technical Attachment
No. 90-16
May 1, 1990

A CASE STUDY ON THE SIGNIFICANT ATMOSPHERIC COOLING WHICH RESULTED IN HEAVY SNOWFALL OVER PORTIONS OF THE MIDDLE ATLANTIC REGION ON JANUARY 8, 1990

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[Editor's Note: The following is a reproduction of an Eastern Region Technical Attachment. The article does an excellent job of quantifying the contributions of adiabatic and diabatic processes to significant low-level cooling of the atmosphere. This particular case resulted in an unexpected snowfall.]

EASTERN REGION TECHNICAL ATTACHMENT
NO. 90-4B
APRIL, 17, 1990

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INTRODUCTION

An unexpected significant snowfall occurred on January 8, 1990, over portions of the Middle Atlantic states. Snowfall amounts of 4 to 8 inches were reported in a narrow band which extended from central West Virginia northeast through western Maryland, and into southeastern Pennsylvania and northwestern New Jersey. Little or no significant precipitation fell north and west of this area. Further east and south closer to the coast, nearly 1 inch of precipitation in liquid equivalent fell, but surface temperatures in this area remained high enough to prevent significant snowfall accumulation.

Freezing levels in the heavy snow area were initially above 8,000 feet, boundary and surface temperatures were above freezing, 1000-500 millibar thickness values were well above 540 decameters, and 1000-700 millibar thickness values were well above 285 decameters.

This case study will examine how significant cooling of the lower atmosphere occurred during the time of precipitation due to a significant increase in vertical velocity, the change of state involved in the melting process, and dry-air intrusion.

SURFACE

A low pressure system formed south of the panhandle of Florida on Sunday, January 7, 1990, and moved northeast into central South Carolina by 1200 UTC Monday morning. A large area of overrunning precipitation in the form of rain moved slowly northeast during Sunday night, and had spread as far north as Beckley, West Virginia and Richmond, Virginia by 1200 UTC.

During the remainder of the morning hours, the overrunning precipitation continued to spread north through most of West Virginia, Virginia, Maryland, Delaware and southern New Jersey. The rain mixed with sleet and snow over much of the Mid-Atlantic region, with precipitation predominantly in the form of snow well inland. The snow continued to spread north into southeastern Pennsylvania, northern New Jersey and southern New England during the afternoon and evening hours as the low pressure system continued to move northeast and rapidly intensified off of the Delmarva and New Jersey coasts.

Precipitation ended in West Virginia and Virginia by late Monday afternoon as the surface system began to move away from the area. Snow or rain ended over the remainder of the Mid-Atlantic and southern New England coastal states by or shortly after midnight.

SOUNDINGS and UPPER-AIR

The 1200 UTC soundings at Atlantic City (ACY) and Washington (IAD) indicated shallow low-level inversions from the surface up to around 980 millibars. Temperatures at the surface were near 0°C, but rose sharply to around +6°C at the top of the inversion layer. The Pittsburgh (PIT) sounding revealed a very weak inversion or nearly isothermal layer up to 800 millibars. Freezing levels at 1200 UTC for ACY, IAD and PIT were all around 8,500 feet (Figures 1-3). Temperatures at 850 millibars over the area which received significant snowfall during the day initially were between +1°C and +2°C (Figure 4). The, 1000-700 millibar thickness values in this area were between 286 and 288 decameters - well above the 280 to 285 decameter range considered favorable for heavy snow events (Figure 5).

By 0000 UTC, however, 850 millibar temperatures over much of the Mid-Atlantic region had cooled as much as 4°C through a combination of dynamic low-level atmospheric lifting, and thermodynamic evaporative and state-change processes (Figure 6). At this time, 1000-700 millibar thicknesses in the area where heavy snow occurred had fallen to between 284 and 285 decameters (Figure 7), and most of the lower atmosphere was now near or slightly below freezing (Figures 8-10).

DYNAMICS and THERMODYNAMICS

The dynamic and thermodynamic processes which helped significantly cool the lower atmosphere over the Middle Atlantic states are shown mathematically in the Appendix.

The diabatic process of absorption and loss of heat includes phase-changes and evaporation. It is interesting to note that evaporation is a strong factor which considerably cools large layers of the atmosphere by making that layer isothermal when it was once close to the dry-adiabatic lapse rate (Haltiner and Martin, 1957). Vertical motion cools a vertical profile more effectively when the lapse rate is not close to the dry-adiabatic lapse rate (see Equation 5 in the Appendix). Vertical motion increases the cooling effect as the lapse rate decreases, with the greatest cooling potential when the lapse rate is negative (indicating the presence of an inversion). Of course, an inversion suppresses vertical motion, so dynamic forcing is required to produce this effect. The low-level inversions which were indicated at 1200 UTC at both ACY and IAD (Figure 1) therefore helped to enhance the cooling effect induced by vertical motion.

Strong positive vertical velocity associated with the storm moved northeast through the heavy snow area during the day. The RAFS FOUS (FRHT61-62) for the Mid-Atlantic region indicated that 700 millibar vertical velocities in excess of 6 microbars sec^{-1} would move through the area by Monday evening. The lower atmosphere cooled in response to the positive increase in vertical velocity. Although the lower atmosphere could cool at the moist adiabatic rate of $6° C\ km^{-1}$, this cooling process was offset somewhat by diabatic processes such as low-level warm advection. The RAFS FOUS T1 and T3 forecasts indicated that temperatures below 850 millibars would cool from around +4°C at 1200 UTC, to 0 or +1°C by 0000 UTC.

Evaporational cooling on the northern fringe of the storm was also a likely factor in low level atmospheric cooling. The latent heat of vaporization is equal to approximately 590 gram calories. Therefore, a total of 590 calories are required to change a gram of water in the liquid state to the vapor state. The atmosphere was initially quite dry below 600 millibars at both ACY and IAD before precipitation began, and the atmosphere remained fairly dry throughout the entire day at PIT, which remained north of the precipitation shield (Figures 3 and 10). The lower atmosphere initially cooled over much of the Mid-

Atlantic region as mid-level precipitation evaporated when it fell into low levels which were still quite dry. It is likely that some low-level evaporational cooling continued on the northern fringe of the storm as drier air was entrained south on north to northeast winds.

Another process which helped in low-level atmospheric cooling was the state-change involved in the melting process. The latent heat of fusion is about 80 gram calories. Therefore, a total of 80 calories of energy is required from the atmosphere to change a gram of water from a solid to a liquid state. The soundings over much of the Middle Atlantic region on January 8th indicated that freezing levels were around 8500 feet during the morning, but gradually lowered to near the surface during the day in the heavy snow area (Figures 1-3 and 8-10). Some additional cooling of the lower atmosphere was therefore occurring as snow melted through this layer. Although both processes were involved, the thermodynamic cooling due to evaporation was significantly stronger than the cooling which resulted from the state-change involved in the melting of snow.

ADDITIONAL CONSIDERATIONS

All of the processes mentioned in the previous section cooled the lower atmosphere enough to allow snow to accumulate on the northern fringe of the storm system. It is interesting to examine the dynamics that forced large areas of vertical motion, which in turn produced copious amounts of precipitation. Large-scale vertical motions can be implied by examining the frontogenesis process and the Omega Equation.

A 300 millibar jet of greater than 110 knots stretched from southern Texas northeast to the Delmarva coast. Gulf moisture at high levels and Atlantic moisture at low levels of the atmosphere were entrained into the storm circulation. Strong divergence at 300 millibars moved northeast over the Mid-Atlantic region during the day - enhancing low-level convergence and implying strong upward vertical motion (Figure 11).

Very strong positive isothermal vorticity advection (PIVA) was also occurring during the day, with the heaviest snow area confined to a narrow band where 1000-500 millibar thickness values were between 544 and 546 decameters (Figures 12 and 13). PIVA is an attempt to combine differential vorticity advection with the height term and the thickness advection term of the Omega equation into a graphical presentation by advecting vorticity in a layer by the thermal wind of that layer (Chaston, 1989). The Omega equation, when simplified, indicates that upward vertical velocity normally associated with cloud-cover and precipitation is a function of both increasing positive differential vorticity advection with height (PVA) and warm advection. Since upward vertical motion due to PVA (advection by the geostrophic wind which blows parallel to the height field) can be offset by cold-air advection, a PIVA overlay (PVA advected on a thermal wind which blows parallel to the 1000-500 millibar thickness field) examines areas where both functions of the Omega equation are working together to produce strong positive vertical velocities. The use of this overlay (when PVA is increasing with height) has proven to be a good way to estimate the dynamic and thermodynamic effects caused by vertical motion, and is useful in better defining areas where significant precipitation may occur.

Perhaps one of the best ways of inferring where upward vertical motion is at a maximum is by examining Divergence of Q (Div Q) fields. This quasi-geostrophic diagnostic can be thought of as a complete solution of the Omega Equation. Areas where simultaneous advection of both temperature and increasing vorticity with height are represented by negative values, and imply upward vertical motion. Positive areas (diverging Q fields) imply downward vertical motion and subsidence. Q-vector convergence was evident at both 500 and 700 millibars over the Middle Atlantic region during the day (Figures 14 and 15).

Q-vectors represent the rate of change of the horizontal potential temperature gradient which develops in an air parcel moving with the geostrophic wind assuming no vertical velocity. Although this is not always true, the Q-vector is useful because it gives an approximation of the ageostrophic horizontal wind in the lower branch of the circulation which develops in order to maintain the thermal wind balance in a developing synoptic disturbance (Chaston, 1989). Although not shown, the Q-vectors at both 500 and 700 millibars over much of the Middle Atlantic Region during the day were crossing the isotherms from cold to warm air. This tightening of the thermal gradient suggested frontogenesis was occurring, with available potential energy being converted into kinetic energy.

Additional information concerning Q-vectors and heavy snowfall can be obtained by examining Western Region Technical Attachments Nos. 90-07 and 90-08.

SUMMARY AND CONCLUSIONS

Significant low-level atmospheric cooling in the Mid-Atlantic region occurred on January 8th, due to a combination of dynamic and thermodynamic processes. The diabatic local temperature change with respect to time is a function of phase/volume changes, horizontal advection, vertical motion and lapse-rate. The negative contribution of these functions to local temperature resulted in an unexpected heavy snowfall for a small area from West Virginia northeast into northern New Jersey. Although some of the guidance data available to the forecaster prior to the event indicated that some low-level cooling would occur during the day, the magnitude of this cooling was underestimated, especially by lower resolution models such as the LFM.

Strong vertical velocities, low-level convergence and upper-level divergence, atlantic and gulf moisture availability, strong positive isothermal vorticity advection and upslope conditions along the eastern ridges of the Appalachian mountains helped to generate copious precipitation amounts over much of the Mid-Atlantic states.

Heavy snowfall is possible under normally unfavorable conditions when strong PIVA moves into a region causing excessive vertical velocities, where 1000-500 millibar thickness values are equal to or less than 546 decameters, and where the T1 and T3 values of the RAFS FOUS are forecasted to be or fall to within a couple degrees of freezing. Sustained northeast winds to the north of a surface low pressure system may aid in upslope over higher terrain east of the Appalachians, and advected drier air at low levels of the atmosphere just north of the precipitation shield are additional factors which might help in better defining the potential for and general location of where heavy snow may fall.

ACKNOWLEDGEMENTS

The author wishes to thank Chet Henricksen (MIC - WSFO Philadelphia), Dean Gulezian (DMIC - WSFO Philadelphia), and Gary Carter (SSD - ERH) for their invaluable assistance and mathematical addition to this text.

REFERENCES

Haltiner, G. J., and F. L. Martin, 1957: *Dynamical and Physical Meteorology*, McGraw-Hill Book Company, Inc., pp. 206-208.

Chaston, P., 1989: *Graphical Guidance*, NWSTC, Kansas City, Mo., pp. 105 and 107.

APPENDIX

EQUATION 1: $\frac{dT}{dt} = \frac{\partial T}{\partial t} + \mathbb{V} \cdot \nabla T + \omega \frac{\partial T}{\partial z}$ $\quad - \frac{\partial T}{\partial z} = \gamma$

where $\frac{dT}{dt}$ is the total temperature change with respect to time,

$\frac{\partial T}{\partial t}$ is the local rate of temperature change,

$\mathbb{V} \cdot \nabla T$ is horizontal advection in x, y (or u, v) coordinates,

$\omega \frac{\partial T}{\partial z}$ is the product of vertical motion in z (or w) coordinates and lapse rate.

Equation 1 can be re-written as follows:

$$\frac{\partial T}{\partial t} = \frac{dT}{dt} - \mathbb{V} \cdot \nabla T - \omega \frac{\partial T}{\partial z}$$

and shows that the local change in temperature is equal to the total temperature change with respect to time minus the horizontal advection and vertical motion/lapse rate terms.

EQUATION 2: $\frac{dH}{dt} = C_p \frac{dT}{dt} - \alpha \frac{dP}{dt}$ $\quad \alpha = 1/\rho$ (DENSITY)

The First Law of Thermodynamics states that total heat exchange as a function of time $\frac{dH}{dt}$

is equal to the specific heat at constant pressure multiplied by the total temperature change with respect to time

$$C_p \frac{dT}{dt}$$

minus the product of specific volume and the total change in pressure with respect to time.

EQUATION 3: $\frac{dP}{dt} = \frac{\partial P}{\partial t} + \mathbb{V} \cdot \nabla P + \omega \frac{\partial P}{\partial z}$

Where $\frac{dP}{dt}$ is the total change of pressure with respect to time,

$\frac{\partial P}{\partial t}$ is the local rate of pressure change,

$\mathbb{V} \cdot \nabla P$ is the horizontal advection of pressure,

$\omega \frac{\partial P}{\partial z}$ is the product of vertical motion and pressure change with height.

The Hydrostatic Equation $\frac{\partial P}{\partial z} = -\rho g$ where g=gravity and the First Law of Thermodynamics can then be substituted into Equation 3 to yield an equation which may be simplified:

$$C_p \frac{dT}{dt} = \frac{dH}{dt} + \alpha \frac{\partial P}{\partial t} + \alpha \mathbb{V} \cdot \nabla P - g\,\omega$$

Since the contributions of local pressure changes and cross-isobaric flow are small (except on very small scale motion - such as in thunderstorms), the pressure terms of

$$\frac{\partial P}{\partial t} \quad \text{and} \quad \mathbb{V} \cdot \nabla P$$

in this equation are approximately zero, and can therefore be excluded. The equation in its simplified form is given below:

EQUATION 4: $\frac{dT}{dt} \cong \frac{1}{c_p}\frac{dH}{dt} - \gamma_d\,\omega$ $\quad \gamma_d = g/c_p$ DRY ADIABATIC LAPSE RATE

This simplified equation may be replaced for Equation 1 to yield a final equation:

EQUATION 5: $\frac{\partial T}{\partial t} = \frac{1}{c_p}\frac{dH}{dt} - (\gamma_d - \gamma)\,\omega - \mathbb{V} \cdot \nabla T$

This equation shows that the local temperature change with respect to time $\frac{\partial T}{\partial t}$

is dependent on diabatic absorption or loss of heat

$$\frac{1}{c_p}\frac{dH}{dt}$$

the product of the vertical velocity and the lapse rate subtracted from the dry-adiabatic lapse rate

$$(\gamma_d - \gamma)\,\omega$$

and the horizontal advection of temperature.

$$\mathbb{V} \cdot \nabla T$$

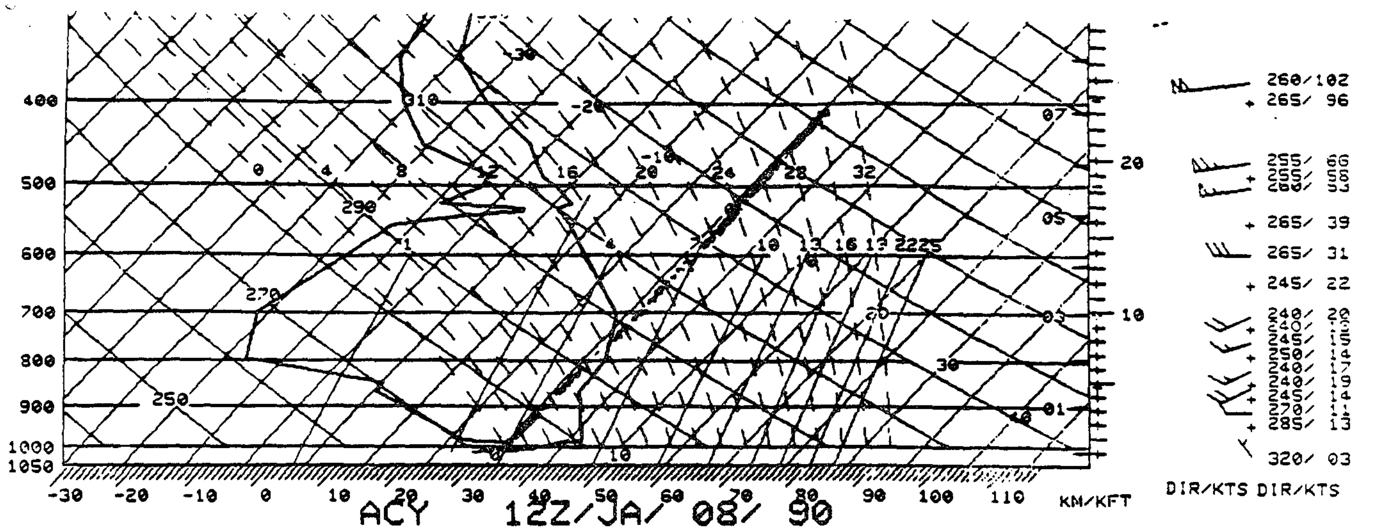

FIGURE 1 Atlantic City Sounding at 1200 UTC January 8, 1990

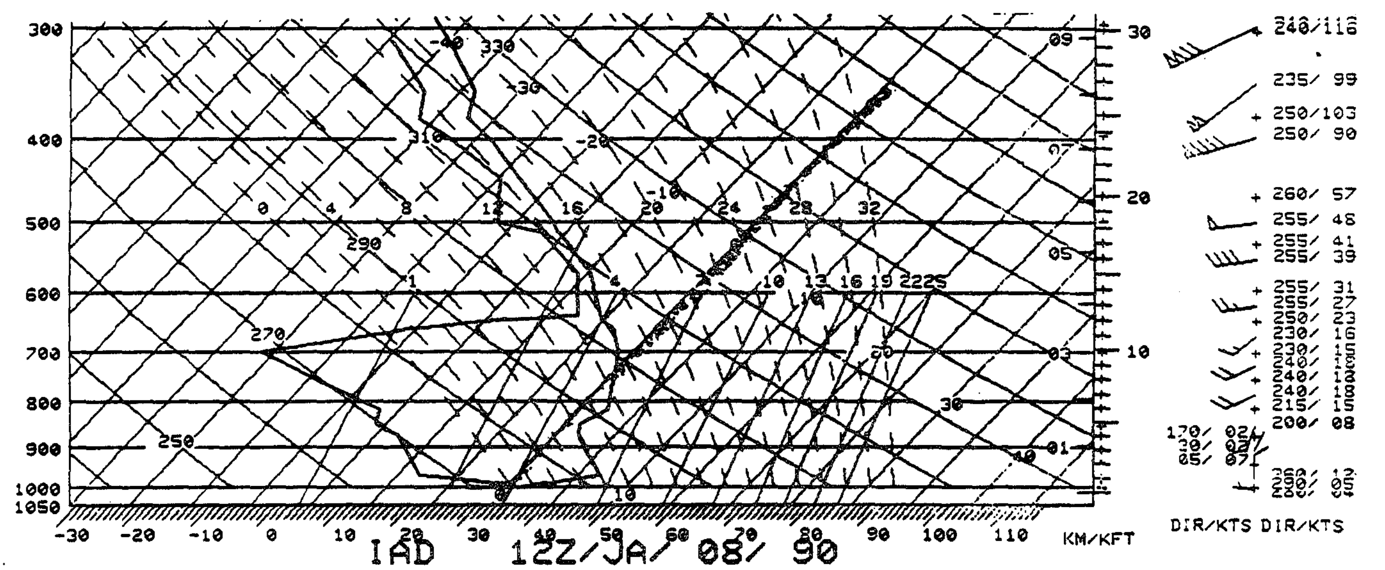

FIGURE 2 Washington Sounding at 1200 UTC January 8, 1990

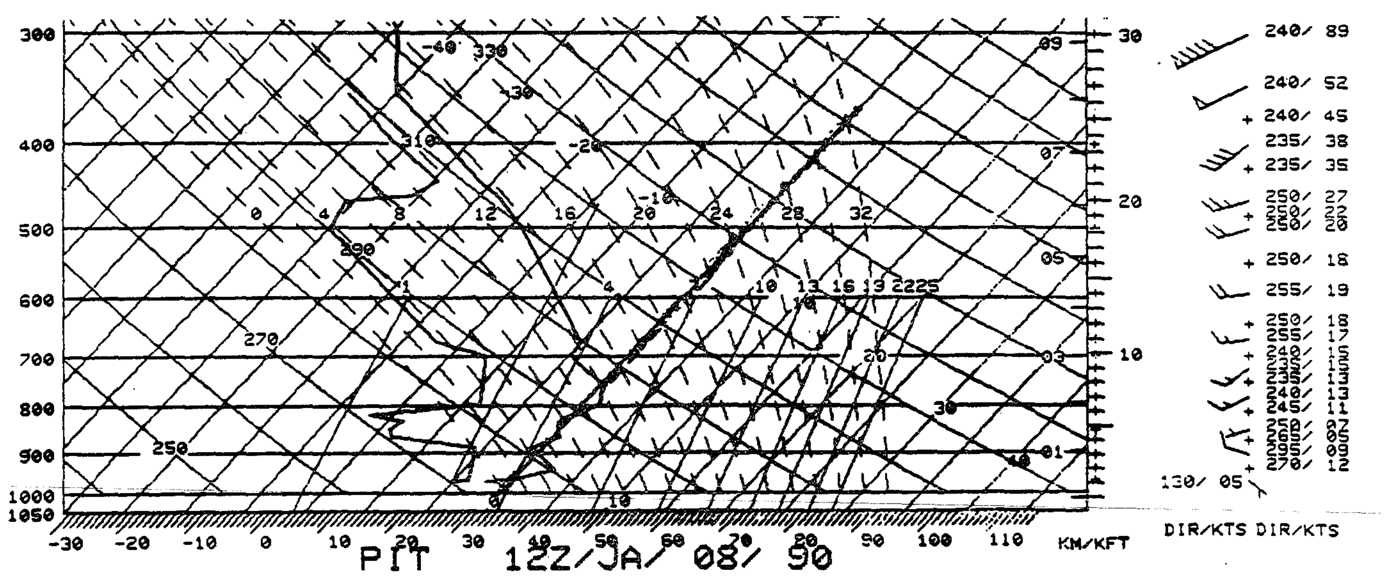

FIGURE 3 Pittsburgh Sounding at 1200 UTC January 8, 1990

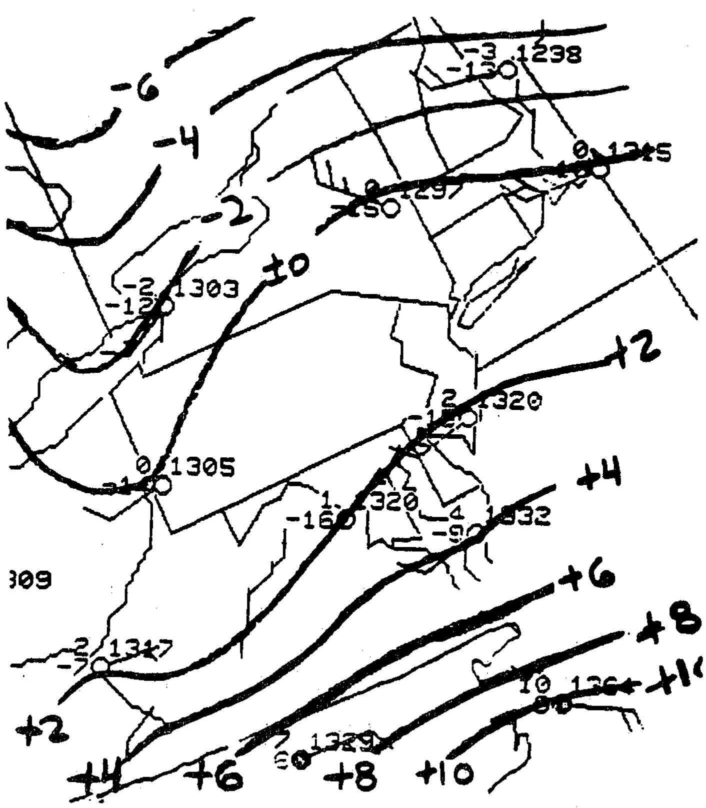

FIGURE 4 850 MB Temperatures
1200 UTC January 8, 1990

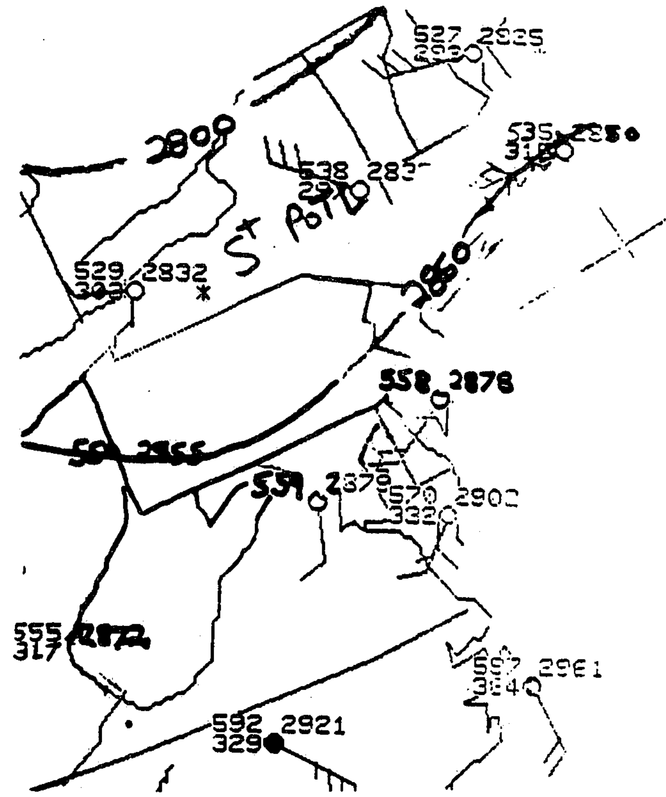

FIGURE 5 1000-700 MB Thickness (M)
1200 UTC January 8, 1990

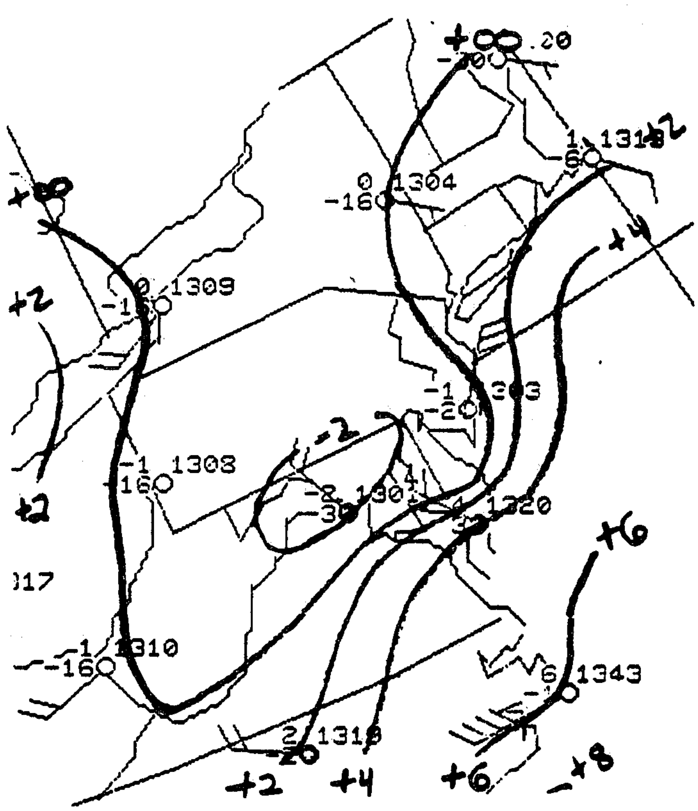

FIGURE 6 850 MB Temperatures
0000 UTC January 9, 1990

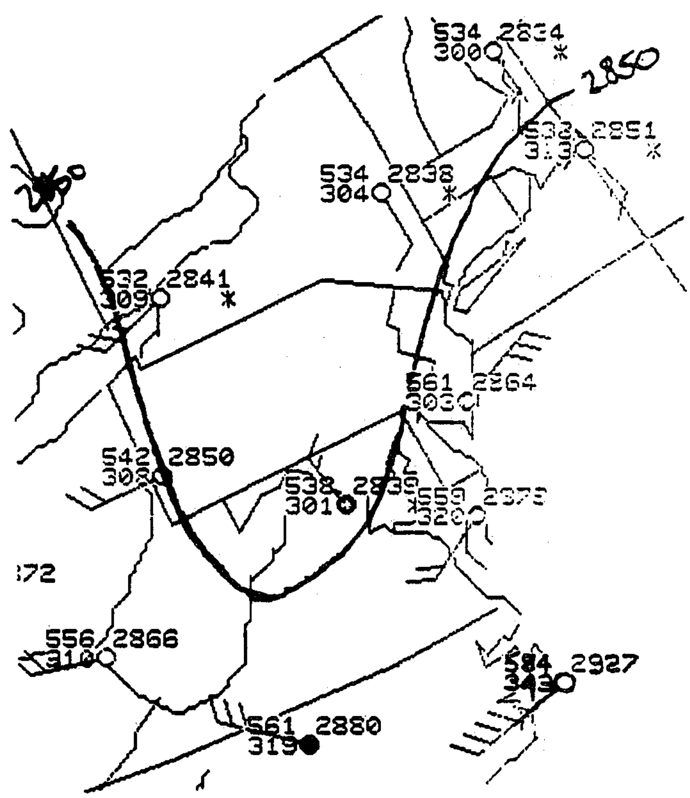

FIGURE 7 1000-700 MB Thickness (M)
0000 UTC January 9, 1990

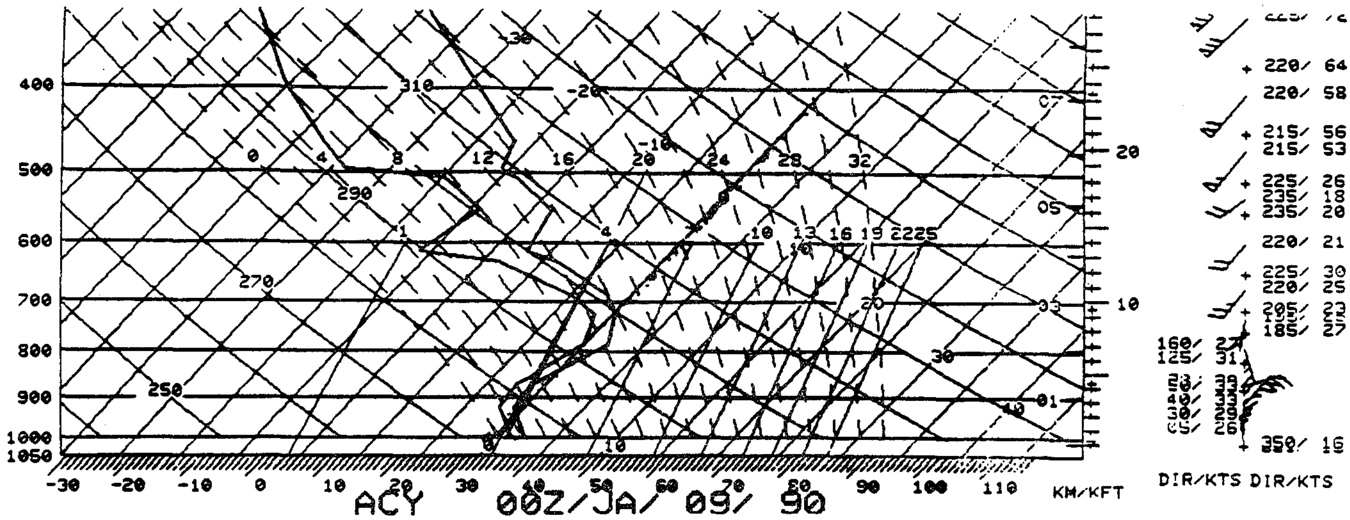

FIGURE 8 Atlantic City Sounding at 0000 UTC January 9, 1990

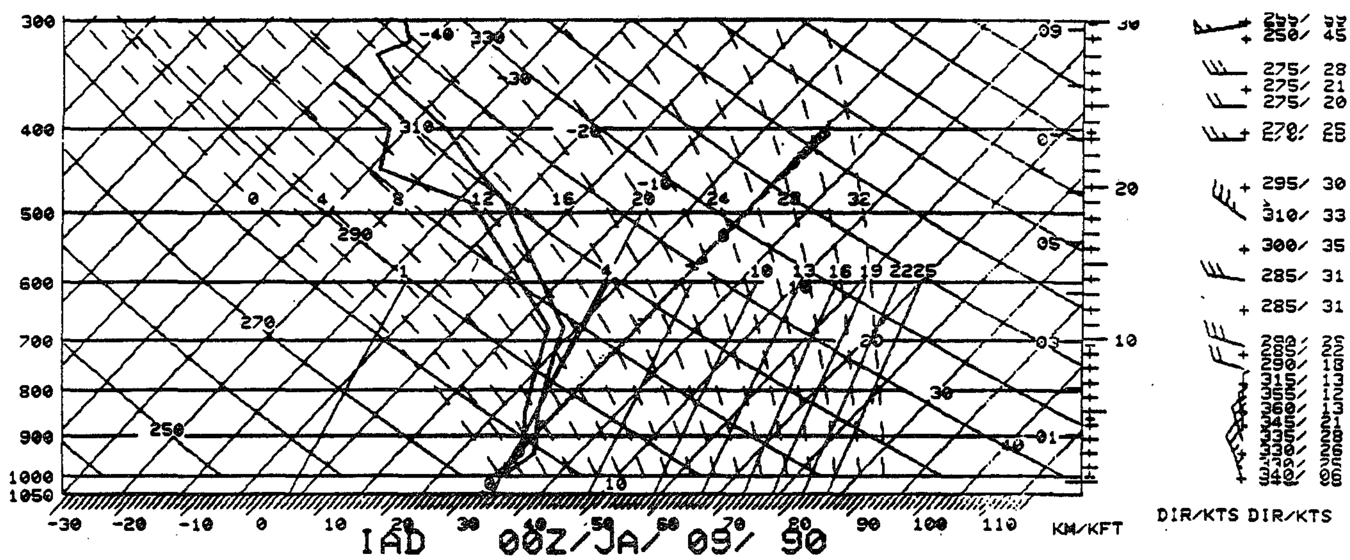

FIGURE 9 Washington Sounding at 0000 UTC January 9, 1990

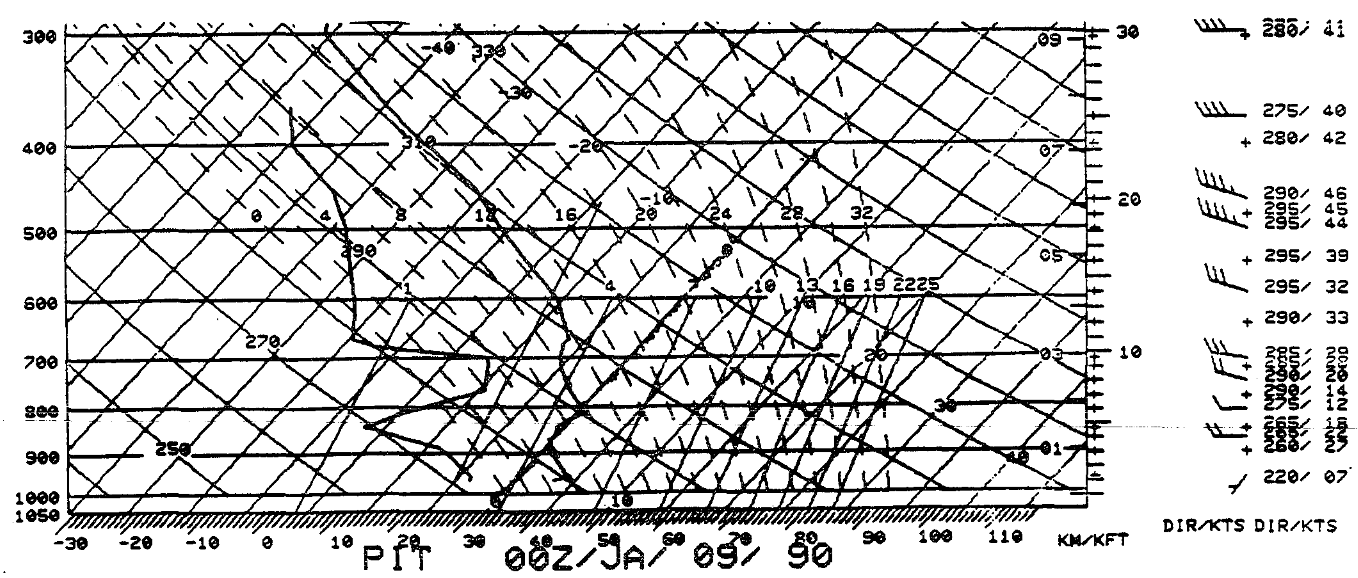

FIGURE 10 Pittsburgh Sounding at 0000 UTC January 9, 1990

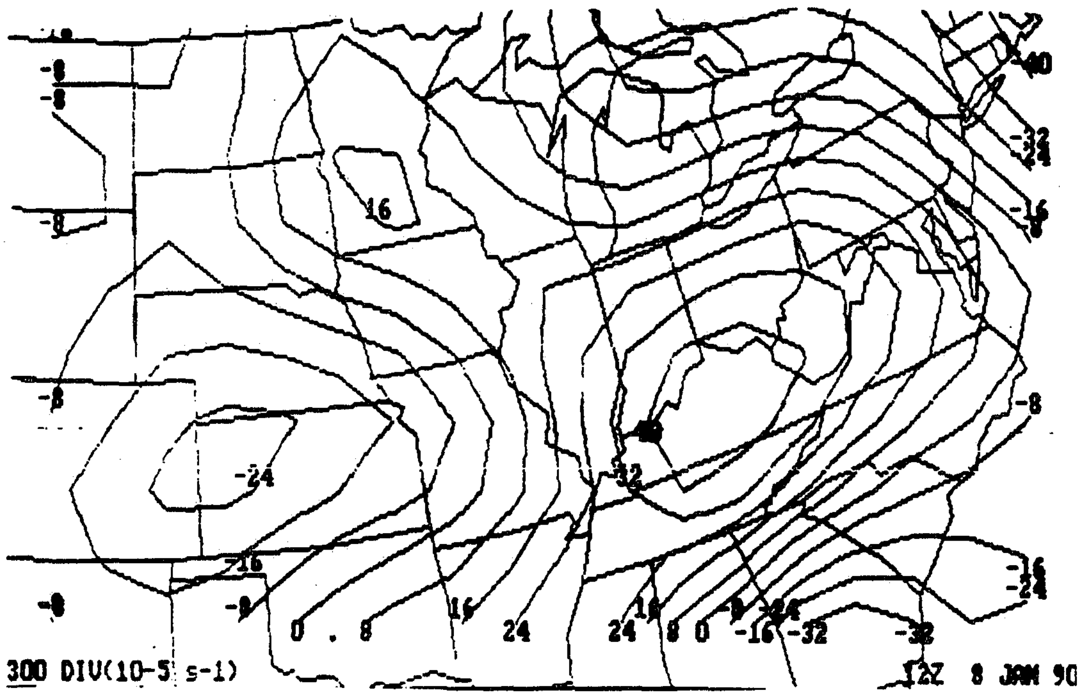

FIGURE 11 300 MB DIV (10 -5 s -1) at 1200 UTC January 8, 1990

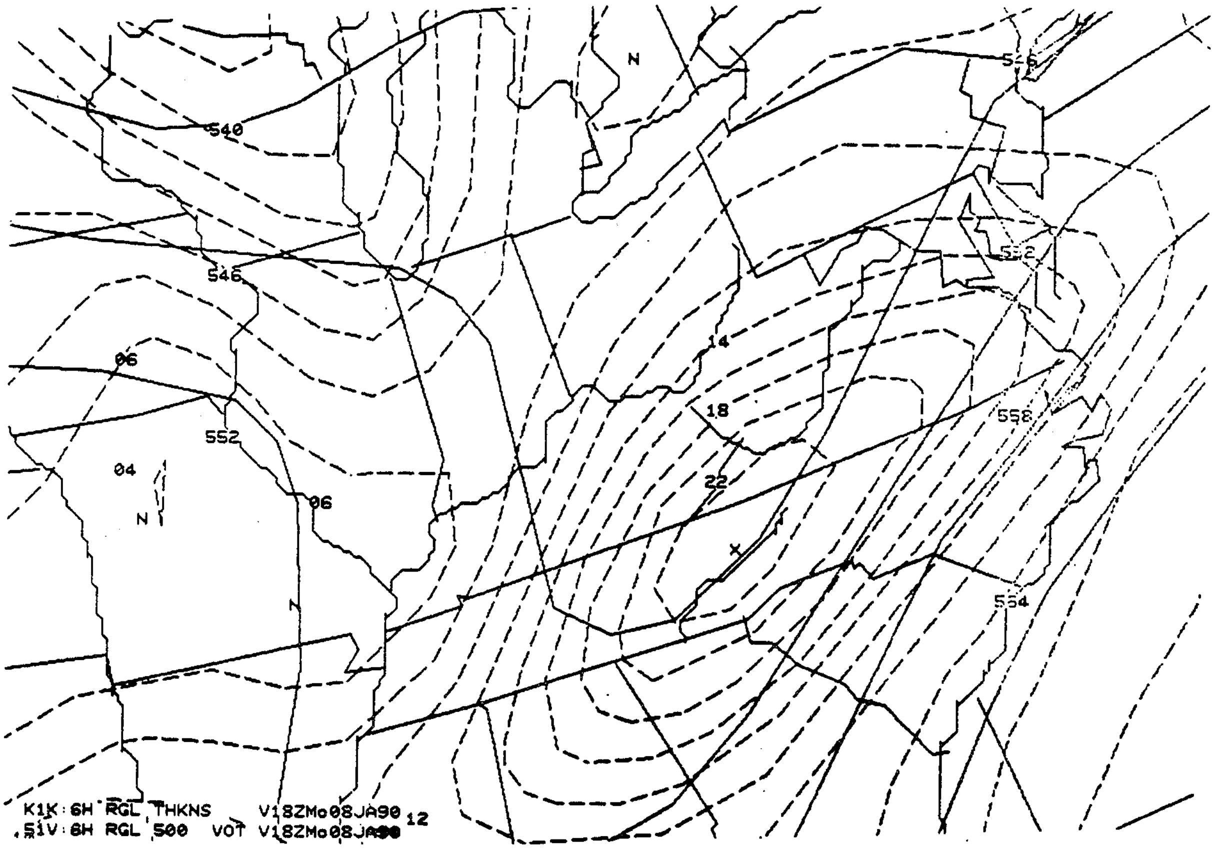

FIGURE 12 500 MB PIVA Overlay valid at 1800 UTC January 8, 1990

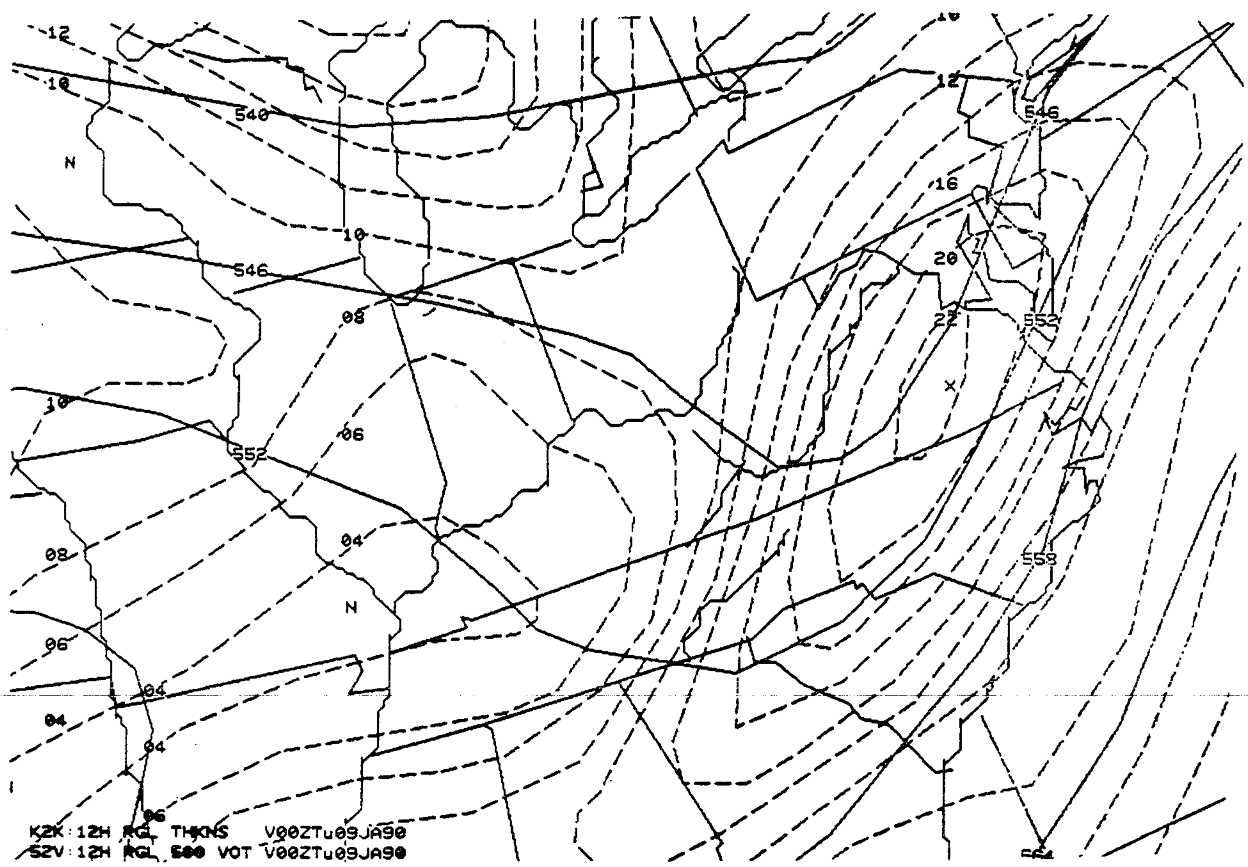

FIGURE 13 500 MB PIVA Overlay valid at 0000 UTC January 9, 1990

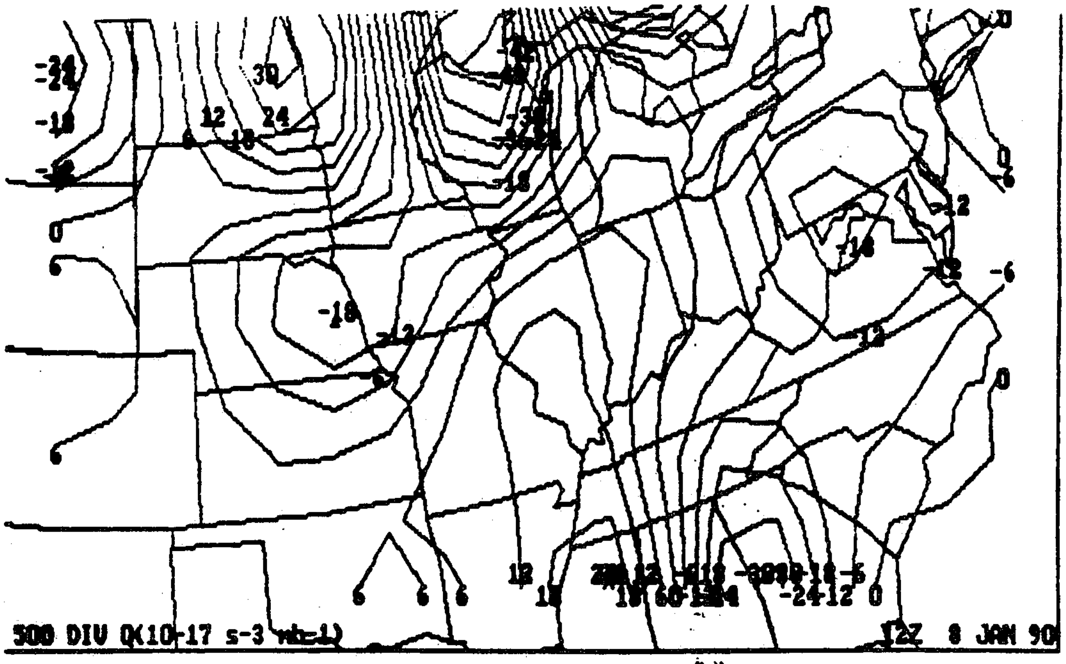

FIGURE 14   500 MB DIV Q (4x10 -17 M Kg -1 s -1) at 1200 UTC
January 8, 1990

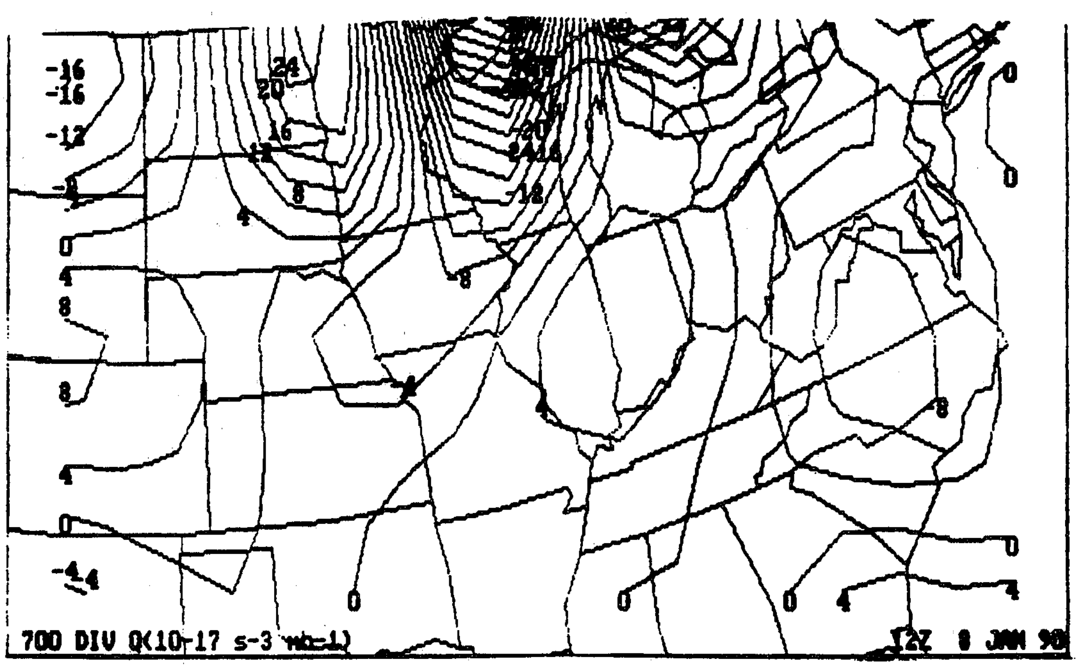

FIGURE 15   700 MB DIV Q (4x10 -17 M Kg -1 s -1) at 1200 UTC
January 8, 1990