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HETEROGENEOUS NUCLEATION AND ITS RELATIONSHIP TO PRECIPITATION TYPE

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Corrections:
Remove pages 1 - 12
Replace with the attached pages 1 - 12

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1. INTRODUCTION

An operational meteorologist is often faced with the difficult task of determining precipitation type. Occasionally, all atmospheric variables suggest that snow will occur yet freezing rain or rain occurs. In these situations, forecasters must go beyond numerical guidance, nomograms, thickness schemes and various "rules-of-thumb" in order to more accurately predict the precipitation type. A basic understanding of cloud microphysics is essential.

The purpose of this study is to present important elements of cloud microphysics that will be useful to the operational meteorologist in determining precipitation type. Synoptic-scale environments and vertical atmospheric structures of cases, where freezing precipitation occurred, will be examined. Furthermore, only cases in which the entire depth of the troposphere was below freezing are studied. The absences of lower tropospheric warm layers (above freezing) suggest that the primary atmospheric process that influenced precipitation type was heterogeneous nucleation rather than melting.

2. HETEROGENEOUS NUCLEATION

Many particles of micron and submicron size, in the earth's troposphere, attract water vapor. These particles are commonly referred to as condensation nuclei. The process by which water droplets form on nuclei from the vapor phase is called heterogeneous nucleation (Byers 1965). This is termed heterogenous since the water vapor and the nuclei have different physical and/or chemical structures.

Differences in the kinetic energy of the molecules for various phases of water exist, resulting in a free energy barrier. This free energy barrier called Gibbs Free Energy, must be overcome when water changes phases (vapor to liquid or liquid to ice). This phase transition process is also called nucleation (Rogers 1979).

Many types of condensation nuclei are present in the atmosphere. Some nuclei, such as haze particles, become wettable at relative humidities less than

100 percent. Wettability is simply a function of the nuclei size and chemical composition. As relative humidities approach 100 percent hygroscopic, or wettable nuclei, begin to serve as centers of condensation.

It can be shown through a simple manipulation of the Clausius-Clapeyron equation (replacing the latent heat of vaporization with the latent heat of sublimation) that the saturation vapor pressure over ice is less than the saturation pressure over water (Hess 1979). Thus, the atmosphere may be nearly saturated with respect to water but supersaturated with respect to ice. If ice crystals and supercooled water droplets coexist in a cloud then the ice crystals may grow at the expense of the water droplets. In other words, the ice crystals will grow by diffusion of the water vapor and the drops will evaporate to compensate for this. This process can also be termed heterogenous nucleation as the ice crystals are "forced" to become condensation nuclei due to the lower saturation vapor pressure over ice. This process is most efficient when temperatures are between -10°C and -15°C, as the saturation pressure difference between water and ice is a maximum in this temperature range (Neuberger 1967).

3. ICE NUCLEATION

In clouds where temperatures are below 0°C, supercooled water droplets may or may not freeze depending upon whether ice nuclei are present. Cloud chamber experiments have shown that pure water droplets do not freeze until a temperature of -40°C is reached. When ice crystals are present (hexagonal plates and dendritic structures), freezing occurs at -10°C to -17°C. Therefore, supercooling of water droplets down to -15°C or colder is not uncommon (Fletcher 1962).

A cloud is stable when it consists of many droplets of the same size. Droplet sizes change very little with time. If precipitation occurs, the droplet population is unstable and some drops grow at the expense of others. There are two mechanisms that destabilize a cloud: 1) collision and coalescence of water droplets and 2) the presence of ice crystals.

Collision and coalescence of water droplets in a cloud will result in a non-uniform droplet size distribution and the cloud environment will be unstable. The instability arises due to the variability in equilibrium vapor pressure over the droplet surfaces which is a function of droplet radius/curvature (Byers 1965). Larger droplets have lower equilibrium vapor pressure and will grow at the expense of the smaller droplets.

When ice crystals coexist with a substantial population of supercooled water droplets, the cloud environment is unstable due to the different saturation vapor pressures of water and ice. If the cloud is continually resupplied with ice crystals and the supercooled liquid water droplet population remains unchanged or decreases, then eventually all the liquid water droplets will be scavenged out of the cloud by diffusion onto the ice crystals, thus regaining cloud stability. The resulting precipitation will ideally become all snow, assuming steady state conditions and the absence of above freezing layers. Similarly, if the ice crystal supply is cut off, all the ice crystals may precipitate out of the cloud and only supercooled water droplets will remain, again establishing cloud stability.

4. SNOW VS FREEZING RAIN

Ice crystals usually appear in clouds when the temperature is at or below -15°C, therefore cloud temperatures can be as low as -10° to -15°C and the precipitation may still fall as rain or freezing rain. Frequently, a sounding is examined for an above 0°C warm layer to determine if freezing rain or rain is possible. If the sounding is well below 0°C at all levels, snow is usually predicted.

Figures 1 and 2 are examples of atmospheric temperature profiles that produced freezing drizzle or a mixture of light snow and freezing drizzle. In each case, the entire depth of the troposphere was at or below 0°C. For both cases, the low level moisture was in an atmospheric environment warmer than -15°C (too warm for ice crystal formation and growth) and bounded by relatively deep elevated dry layers.

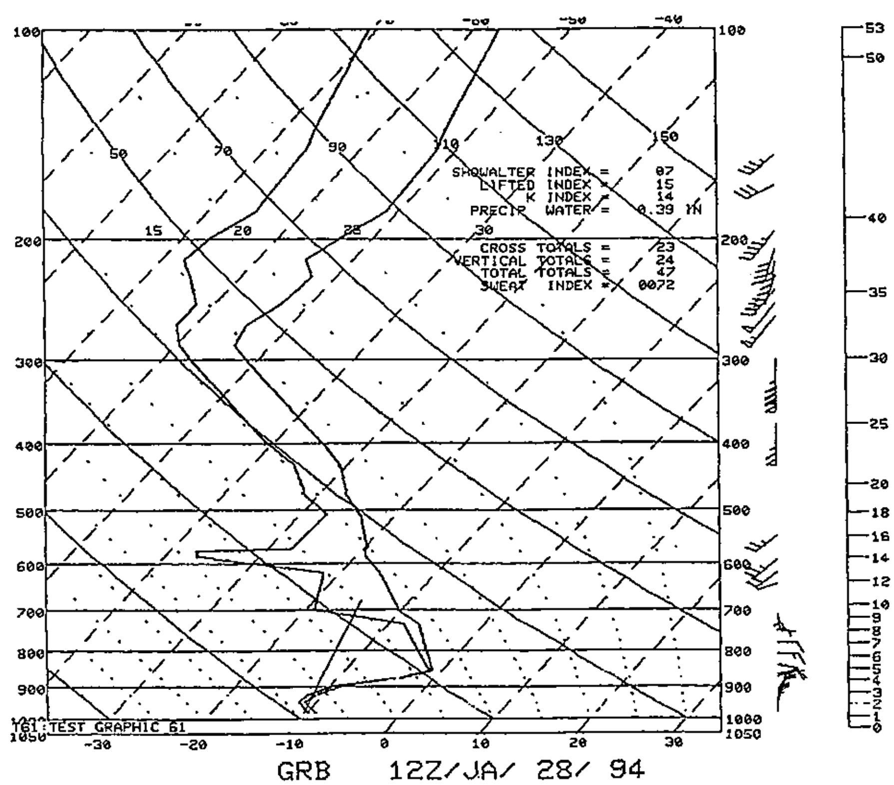

Figure 1. Green Bay sounding from 1200 UTC 28 January 1994.

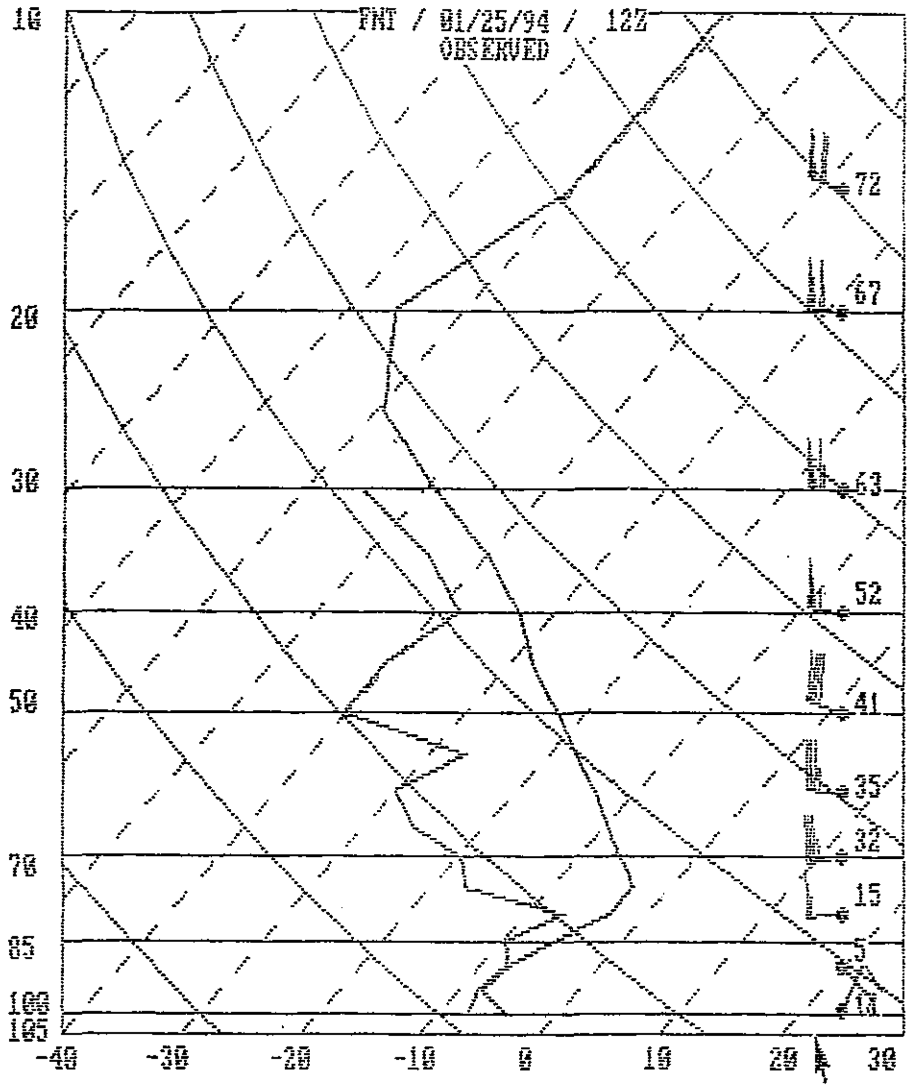

Figure 2. Flint, Michigan, sounding for 1200 UTC 25 January 1994.

The elevated deep dry layers appeared to play an important role in precipitation type by cutting off or limiting ice crystal seeding into the lower moist layers. In both cases, moist layers were present above 18,000 feet and temperatures were cold enough at this level to support ice crystals. However, the dry layers were substantial (more than 5000 feet deep and temperature/dew point spreads greater than or equal to 10°C). Ice crystal sublimation was taking place through a substantial depth of the troposphere and appeared to limit heterogeneous nucleation in the lower moist layers. The result was either a mixture of snow and freezing drizzle or just freezing drizzle at the surface. Quantitative dry layer parameters necessary for complete ice crystal sublimation are unknown and will vary with different synoptic environments. The purpose here is to notify operational forecasters the importance of significant dry layers, and the possibility of mixed or freezing precipitation when soundings show that the entire depth of the troposphere is below 0°C.

The role of elevated convection must also be considered when mid-tropospheric dry layers are present. If a lower tropospheric dynamical forcing mechanism for rising motion can together displace upwards a low-level moist layer and a mid-level dry layer, then elevated convection may occur. The latter is due to potential instability being released, especially if decreases of theta-e with height exceed 5°K in the dry layer (Moore 1992).

The elevated convection may erode the dry layer and reestablish/establish ice crystal seeding (heterogeneous ice nucleation) into the lower level moisture. If the entire troposphere is below freezing, then ideally the precipitation will change over to all snow.

It is not uncommon for convective snow to occur along a cold front, while at the same time, poleward of the front, deeper into the cold air freezing precipitation is observed. This can be the case during situations when there is discontinuous vertical saturation (in the cold air) and the low-level moisture does not extend upward to where temperatures are -15°C or colder.

5. SYNOPTIC SETTINGS

The most frequently observed synoptic-scale environments in which heterogeneous nucleation appears to be the primary atmospheric process influencing precipitation type are: 1) the mature cyclone where aggressive "dry slotting" has developed to the west of the cold front and 2) in arctic air masses where the cold air and moisture fields are shallow. An example of each case follows.

In the mature cyclone case, the aggressive mid-tropospheric drying results in discontinuous vertical atmospheric saturation, thus partially or completely separating the ice crystal source region from the lower tropospheric moist layer. As a result heterogeneous ice nucleation may decrease (or cease) in the lower moist layer and a mixture of light snow and freezing drizzle (or just freezing drizzle) is possible.

Figures 3 through 8 illustrate an example of the mature cyclone case over the lower Great Lakes on January 28, 1994. The 1200 UTC RAFS (Regional Analysis and Forecast System) predicted that thicknesses were at or below snow thresholds in the Milwaukee area at 1200 UTC (Figures 3, 6, and 7). Typically, snow thresholds for 1000-500-mb, 1000-850-mb, and 850-700-mb are 5400m, 1300m, and 1555m (McNulty 1991). Mid-level drying was also evident over all of eastern Wisconsin as 700-mb relative humidities had decreased to 50 percent or less by 1200 UTC (not shown). Milwaukee continued to report freezing drizzle and light snow until 1420 UTC although thicknesses continued to decrease (Figures 1, 3, and 8). Milwaukee reported just snow after 1420 UTC. It appeared as though mid-tropospheric moisture associated with a 500-mb cyclone (which was present throughout the depth of the troposphere) (not shown) moved into eastern Wisconsin after 1400 UTC and reestablished continuous vertical saturation and reinitiated ice crystal seeding into the lower level moisture.

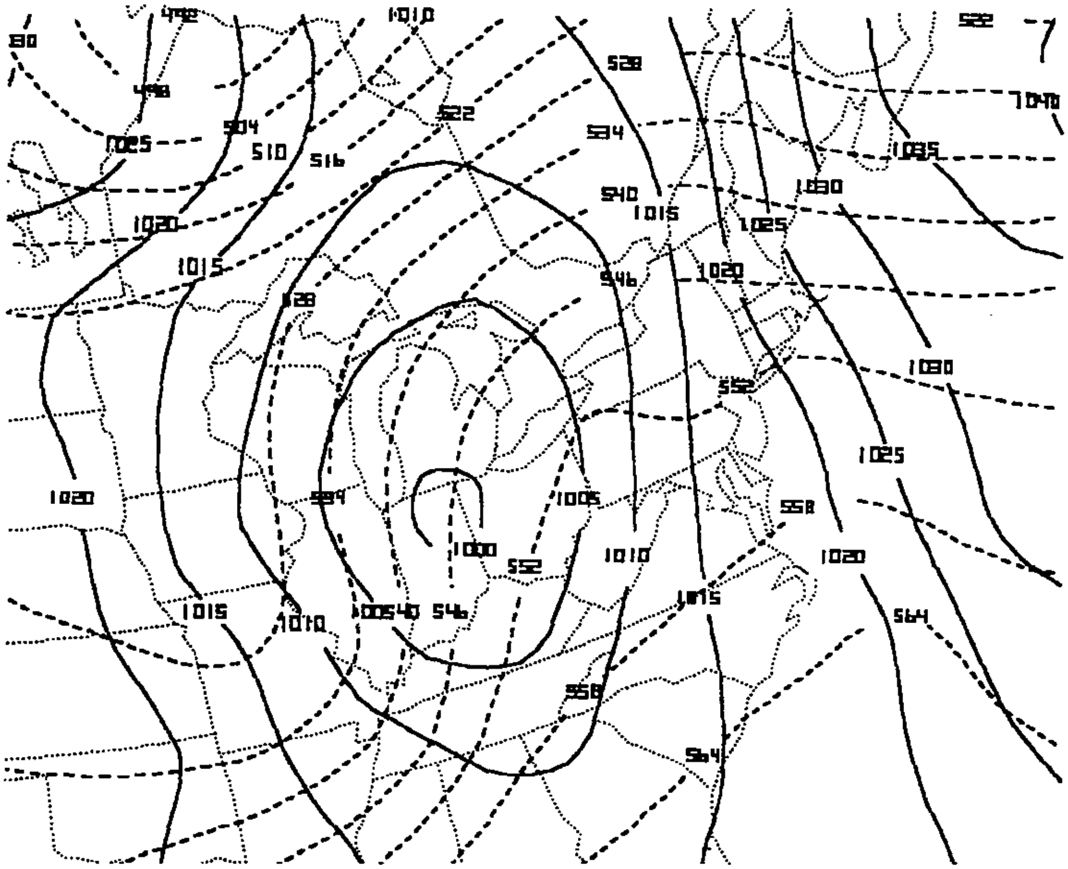

Figure 3. Mean sea-level pressure (mb) and 1000-500-mb thickness (dm) for 1200 UTC 28 January 1994 (RAFS 00 hour analysis).

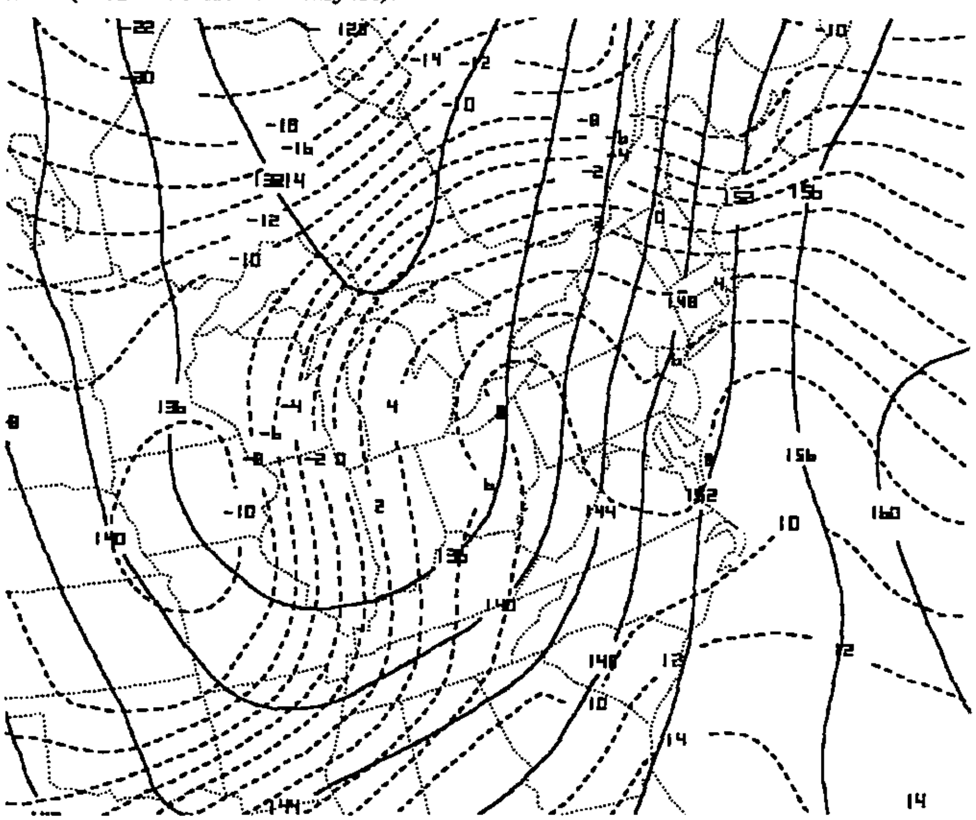

Figure 4. 850-mb heights (dm) and temperatures (°C) for 1200 UTC 28 January 1994 (RAFS 00 hour analysis).

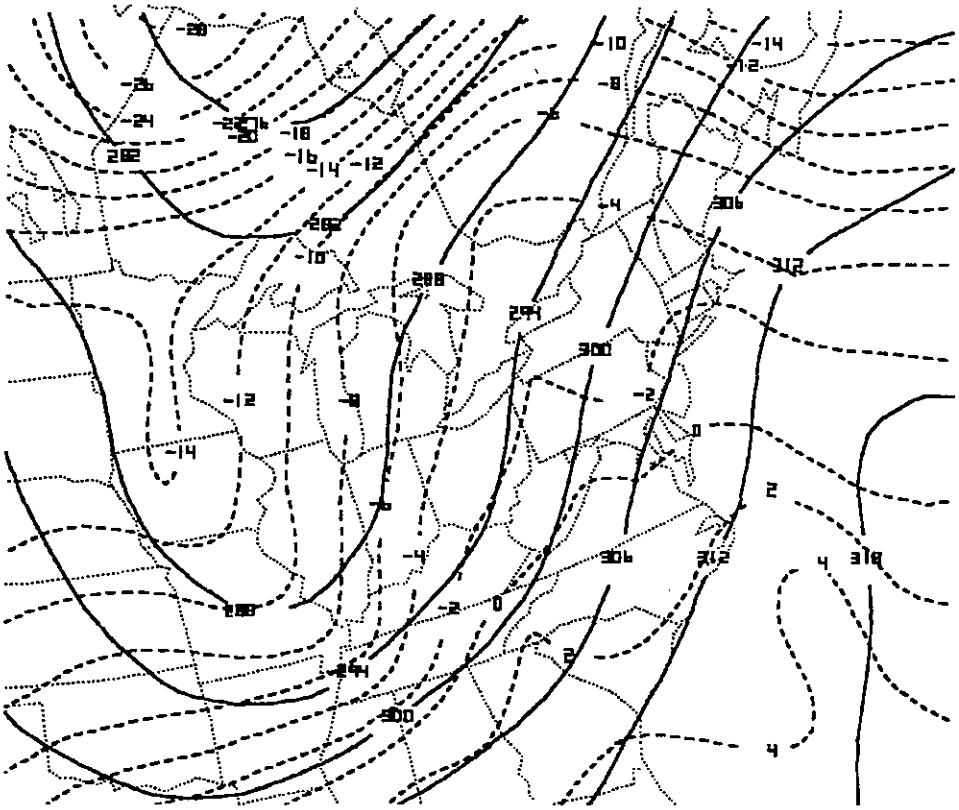

Figure 5. 700-mb heights (dm) and temperatures (°C) for 1200 UTC 28 January 1994 (RAFS 00 hour analysis).

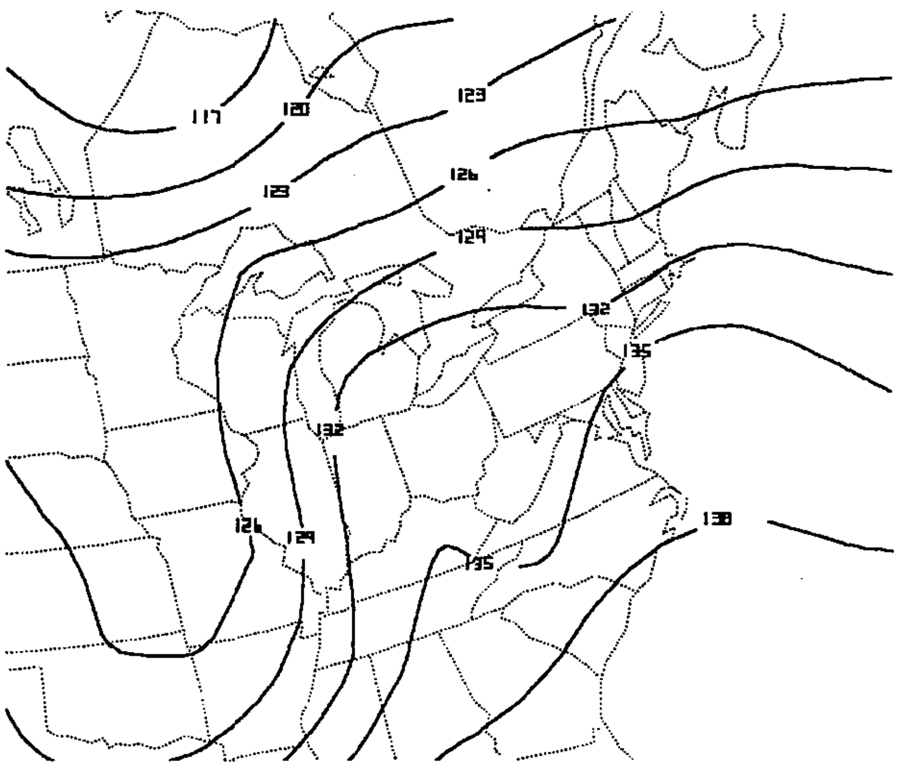

Figure 6. 1000-850-mb thickness (dm) for 1200 UTC 28 January 1994 (RAFS 00 hour analysis).

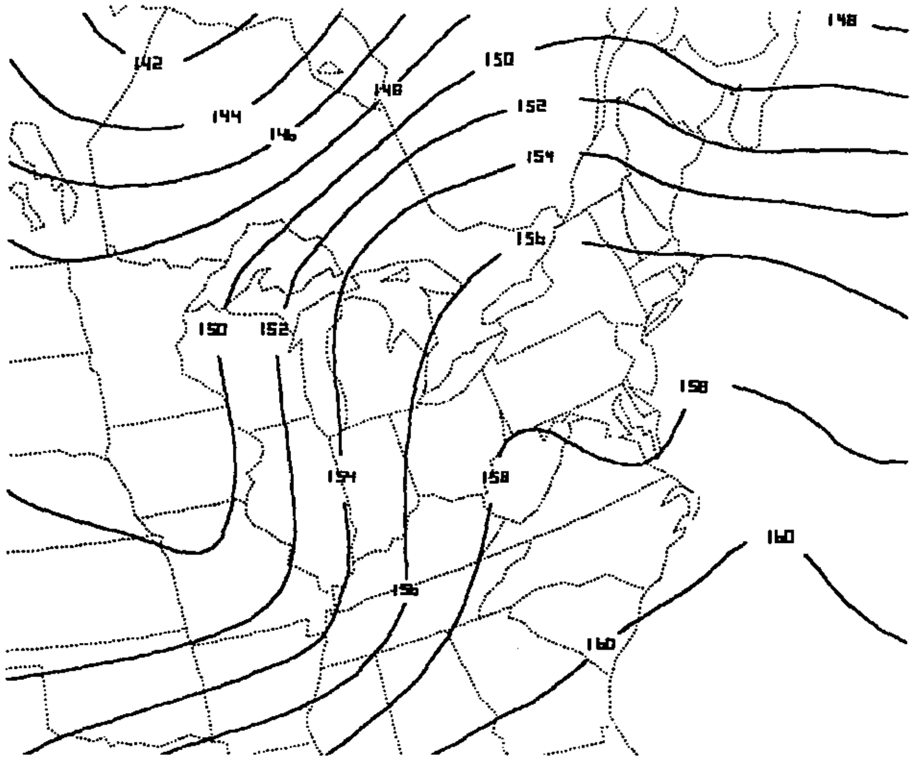

Figure 7. 850-700-mb thickness (dm) for 1200 UTC 28 January 1994 (RAFS 00 hour analysis).

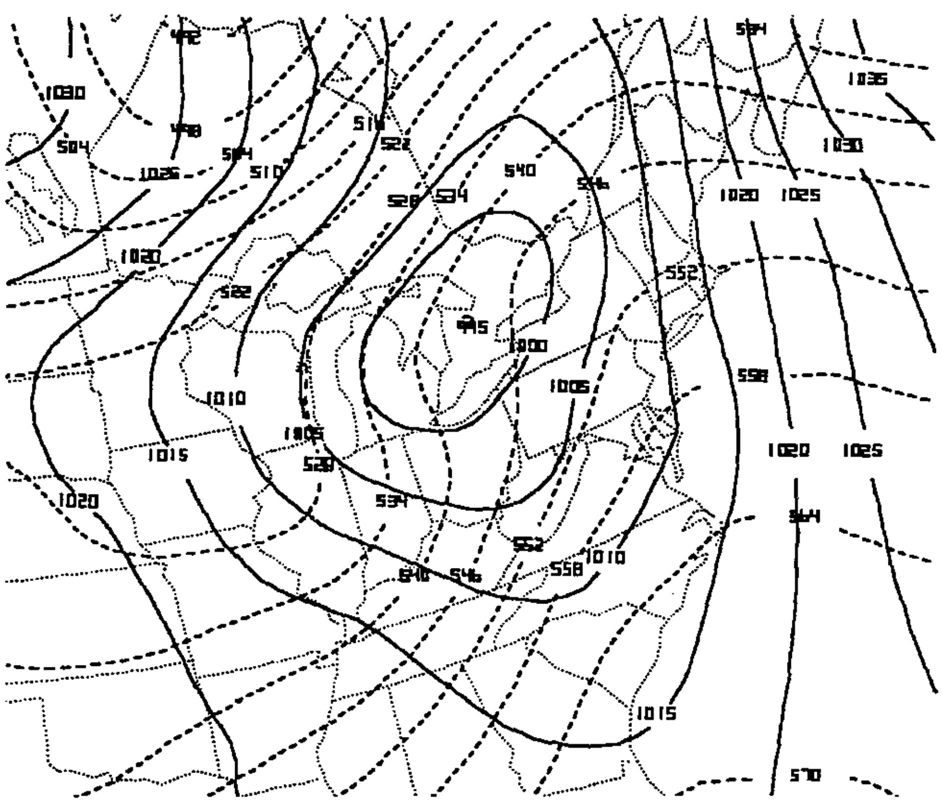

Figure 8. Mean sea-level pressure (mb) and 1000-500-mb thickness (dm) for 1800 UTC 28 January 1994 (RAFS 6 hour analysis).

Figures 9 through 16 depict a shallow arctic air mass case, over southern lower Michigan on January 25, 1994. The 1200 UTC RAFS surface and 850-mb analyses showed that the leading edge of arctic air was over the northern Ohio valley. The front was 100 to 150 miles south of southeastern lower Michigan, thus southeastern Michigan was well into the arctic air. From Figures 10 and 13, it is seen that the arctic air mass was relatively shallow as temperatures near the surface were as cold as or colder than the 850-mb temperatures. The moisture was also shallow in the arctic air mass as 700-mb relative humidities were between 20 and 40 percent over southern lower Michigan.

Several stations in southern lower Michigan reported freezing drizzle or a mixture of light snow and freezing drizzle during the morning of January 25, with Flint reporting a mixture of snow and freezing drizzle until 1440 UTC. Thicknesses were well below snow thresholds over all of southern Michigan at 1200 UTC (Figures 9, 11, and 12) thus, providing the forecaster with little insight that freezing precipitation was possible. Mid-tropospheric moisture increased over southern lower Michigan shortly after 1200 UTC in response to positive differential vorticity advection (cyclonic vorticity advection increasing with height) over southern Michigan in the 850-mb to 500-mb layer (not shown). Apparently the increasing mid-level moisture was substantial enough to reinforce ice crystal nucleation in the lower moist layer and as a result, the precipitation changed to all snow in the Flint area after 1440 UTC.

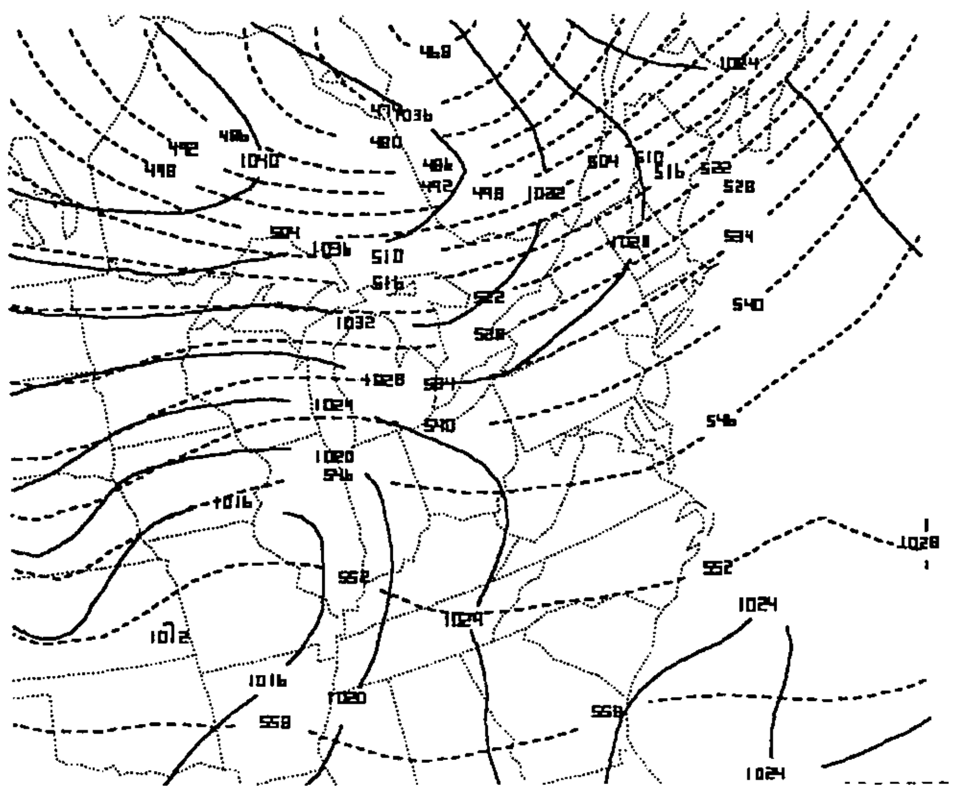

Figure 9. Mean sea-level pressure (mb) and 1000-500-mb thickness (dm) for 1200 UTC 25 January 1994 (RAFS 00 hour analysis).

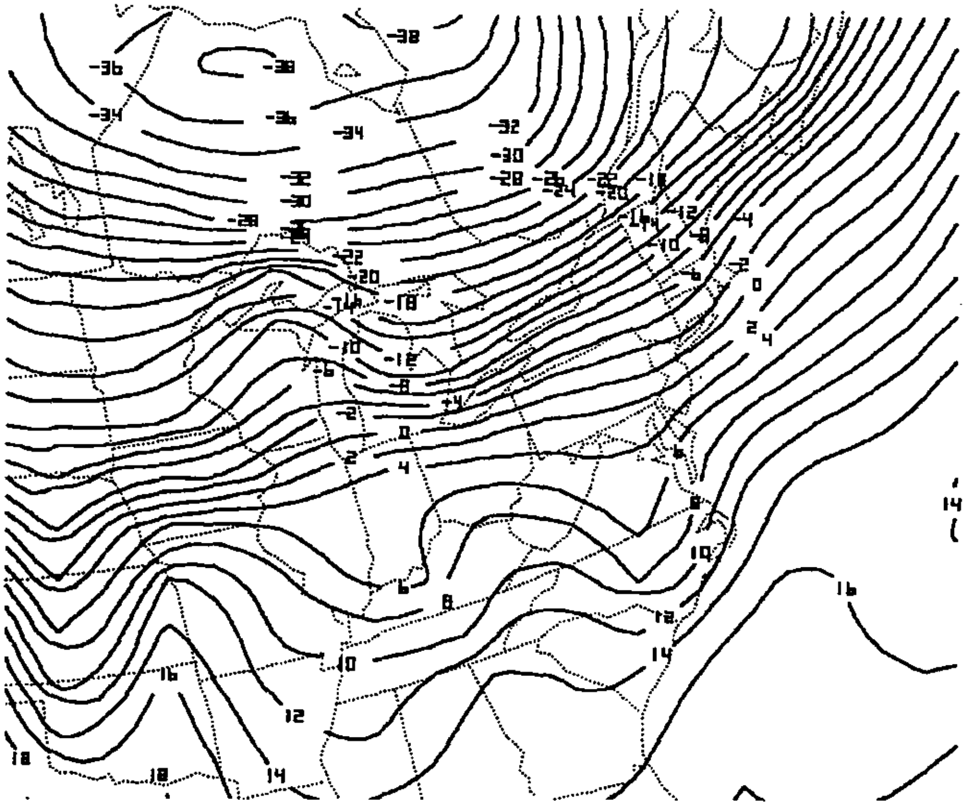

Figure 10. 1000-mb temperatures (°C) for 1200 UTC 25 January 1994 (RAFS 00 hour analysis).

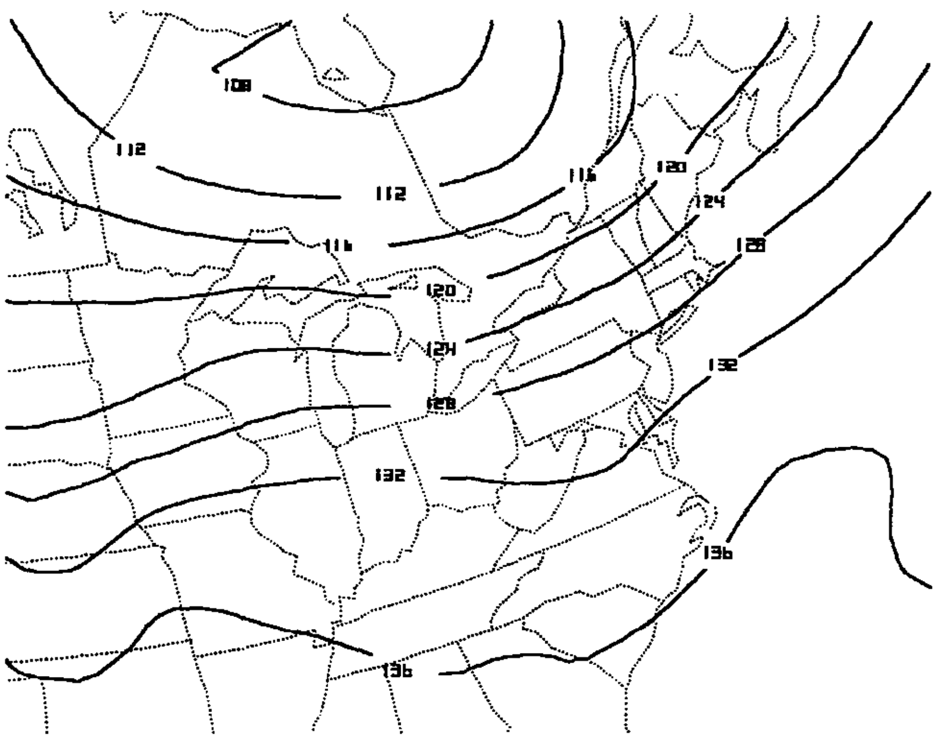

Figure 11. 1000-850-mb thickness (dm) for 1200 UTC 25 January 1994 (RAFS 00 hour analysis).

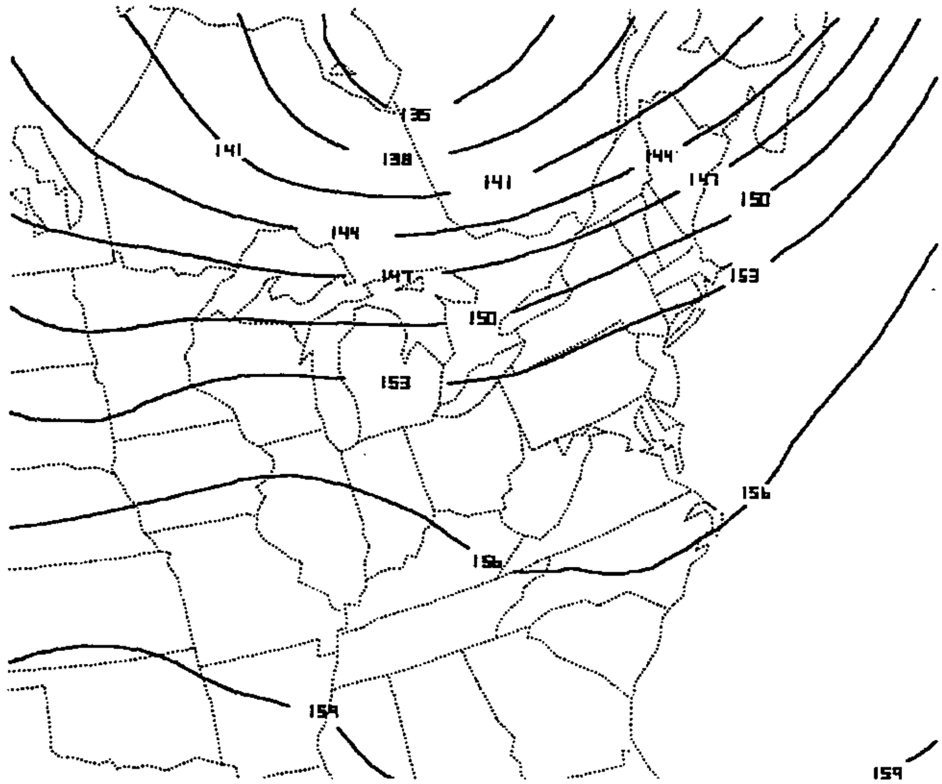

Figure 12. 850-700-mb thickness (dm) for 1200 UTC 25 January 1994 (RAFS 00 hour analysis).

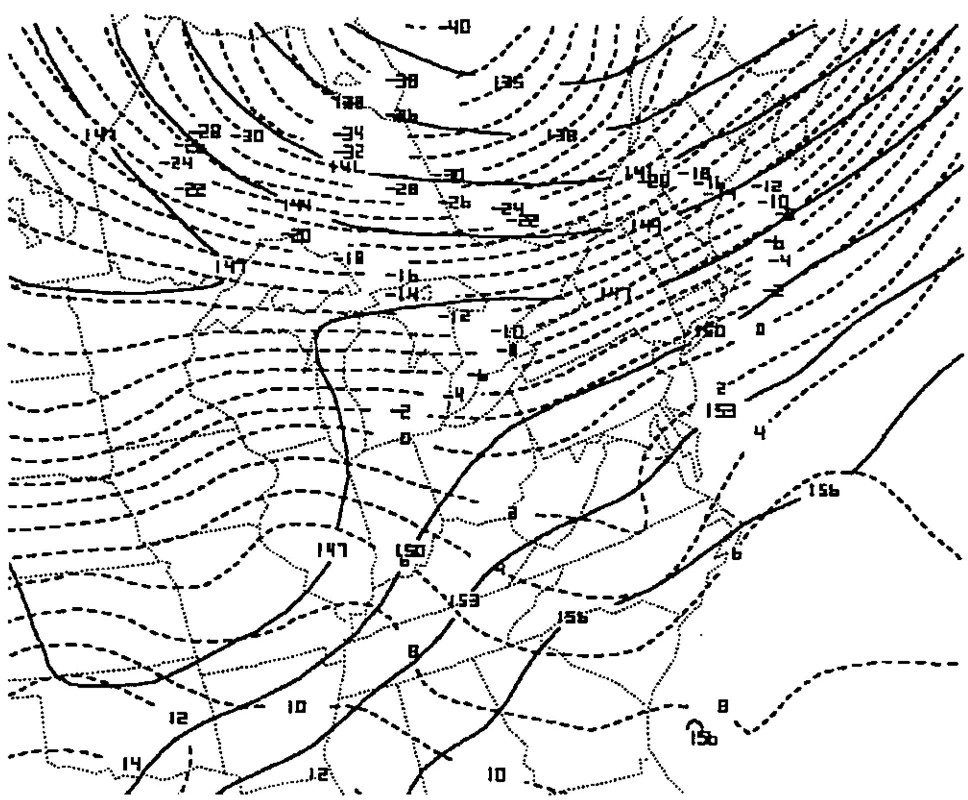

Figure 13. 850 mb heights (dm) and temperature (°C) for 1200 UTC 25 January 1994 (RAFS 00 hour analysis).

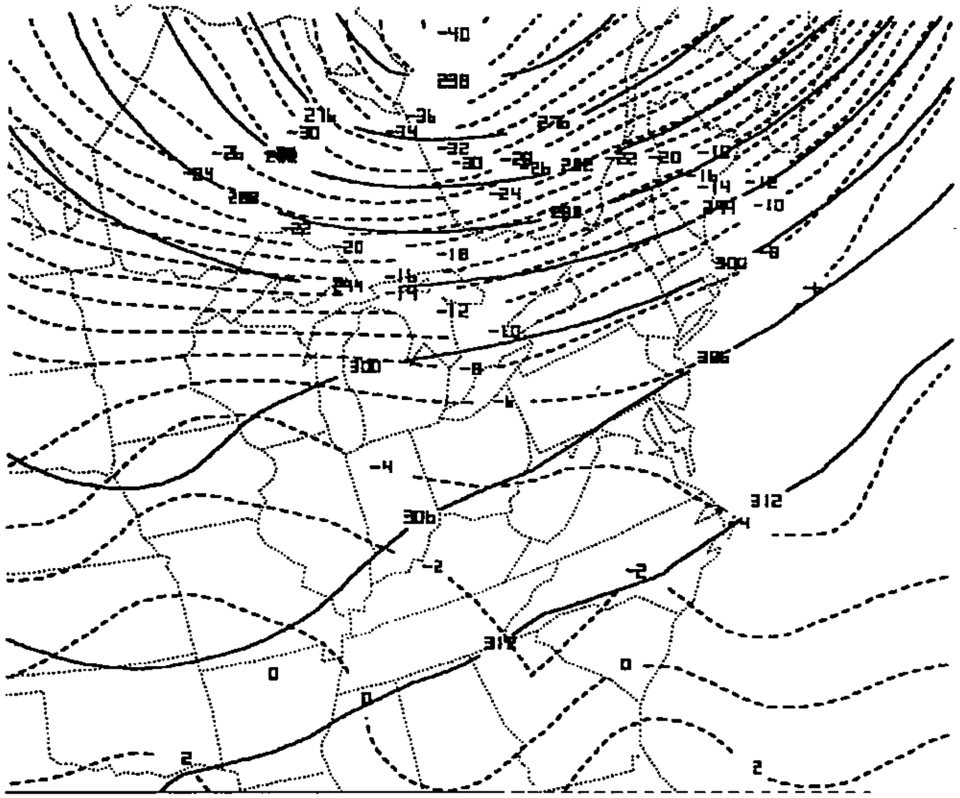

Figure 14. 700-mb heights (dm) and temperature (°C) for 1200 UTC, 25 January 1994 (RAFS 00 hour analysis).

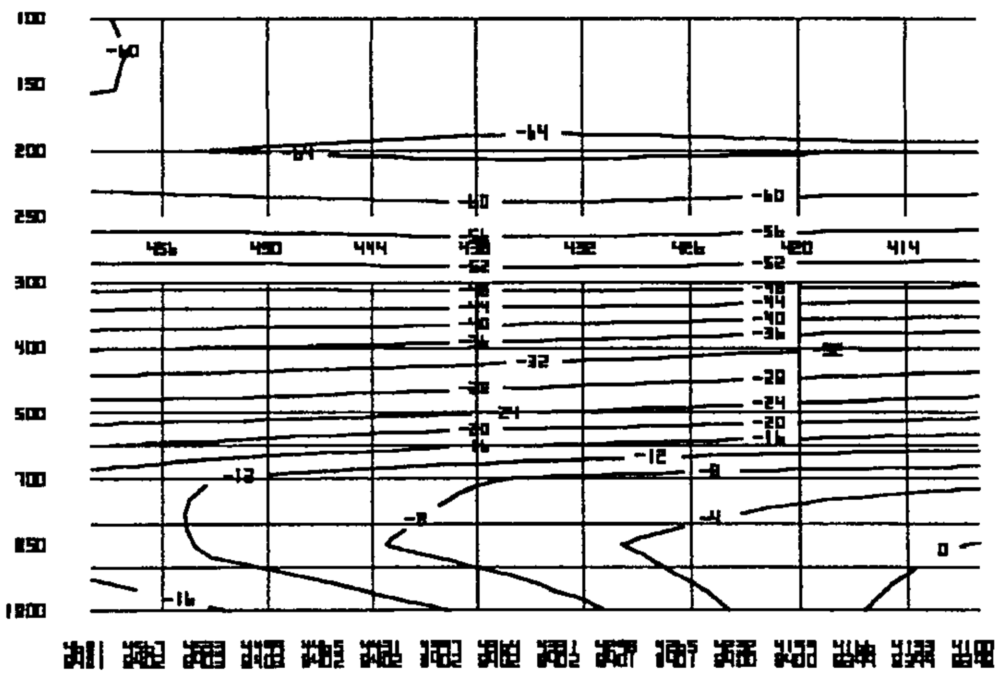

Figure 15. Temperature (°C) height section (Toledo-Flint-Alpena) for 1200 UTC 25 January 1994 (RAFS 00 hour analysis).