

# MID-TROPOSPHERIC DRY LAYERS AND THEIR RELATIONSHIP TO PRECIPITATION TYPE IN A SUB-FREEZING TROPOSPHERE

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## Abstract

*The effect of mid-tropospheric dry layers on precipitation type is examined for cases in which freezing (vs. frozen) precipitation fell when the entire depth of the troposphere was well below freezing. Aggressive mid-level dry slotting associated with a mature cyclone is examined. The depth of the mid-level dry layer and associated ice crystal sublimation, and subsequent disruption of ice nucleation and diffusional ice crystal growth in the lower tropospheric moist layer, is studied.*

*The lack of lower tropospheric warm layers (above freezing) suggest that ice crystal sublimation in the mid-level dry layer was the primary atmospheric process responsible for freezing, rather than frozen, precipitation. A shallow arctic air mass case is also presented, in which the lower tropospheric moist layer was confined to an environment warmer than  $-15^{\circ}\text{C}$  and bounded by a deep mid-tropospheric dry layer. A forecaster checklist is presented based on the study findings.*

## 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, rain, or a mixture of precipitation types occurs. In these situations, forecasters must go beyond numerical weather prediction 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 absence of lower tropospheric warm layers (above freezing) suggest that the primary atmospheric process that influenced precipitation type was ice crystal sublimation in mid tropospheric dry layers, which disrupted ice crystal seeding and diffusional ice crystal growth in the lower tropospheric moist layers.

## 2. Heterogeneous Nucleation

Many particles of micron and submicron size, in the 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 heterogeneous

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 phase (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 (or hygroscopic) 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 saturated, or 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 vapor will diffuse to the ice crystals and the ambient vapor pressure will decrease. The droplets must now evaporate as the vapor pressure decreases to less than the equilibrium value for the droplets (Byers 1974). This also results in a larger free energy barrier between the vapor and liquid droplet, as the free energy of the vapor decreases. In other words, the droplet surface tension force is now greater than the vapor pressure gradient force.

This process of diffusional growth is also most efficient when temperatures are between  $-10^{\circ}\text{C}$  and  $-15^{\circ}\text{C}$  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  $0^{\circ}\text{C}$ , supercooled liquid water droplets may or may not freeze depending upon whether suitable ice nuclei (or freezing nuclei) are present to initiate ice crystal formation. Cloud chamber experiments indicate that pure water droplets do not freeze, or complete the ice nucleation process, until a temperature of  $-40^{\circ}\text{C}$  is reached. This process is also known as homogeneous nucleation. Ice crystals usually appear in clouds when the temperature is at or below  $-15^{\circ}\text{C}$ , indicating that heterogeneous ice nucleation is occurring (Rogers 1979). Heterogeneous ice nucleation is less common at warmer temperatures since active ice nuclei populations are much smaller. Active ice nuclei populations can double with only a  $4^{\circ}\text{C}$  decrease in temperature. Therefore, supercool-

ing of water droplets down to  $-15^{\circ}\text{C}$  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, then 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 a 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 all the liquid water droplets will diffuse onto the ice crystals. As a result, the ice crystals may become large enough to fall and collisions with other ice crystals or liquid droplets may occur. Eventually the cloud will achieve stability by diffusion and/or scavenging out all of the liquid droplets. The resulting precipitation will ideally become all snow, or a mixture of snow and graupel, depending on the type of collisions (ice crystal with ice crystal or ice crystal with water droplet). Near steady state conditions and the absence of above freezing layers must also be assumed. 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  $-15^{\circ}\text{C}$ . Therefore, cloud temperatures can be as low as  $-10$  to  $-15^{\circ}\text{C}$  and the precipitation may still fall as rain or freezing rain. Frequently a sounding is examined for an above  $0^{\circ}\text{C}$  warm layer to determine if freezing rain or rain is possible. If the sounding is well below  $0^{\circ}\text{C}$  at all levels then snow is traditionally forecast.

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  $0^{\circ}\text{C}$ . For both cases, the low-level moisture was in an atmospheric environment warmer than  $-15^{\circ}\text{C}$  (too warm for efficient ice crystal formation and growth) and bounded by relatively deep elevated dry layers.

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 5,000 feet deep and temperature/dew point spreads greater than or equal to  $10^{\circ}\text{C}$ ). Ice crystal sublimation was taking place through a substantial depth of the troposphere and appeared to limit diffusional ice crystal growth 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

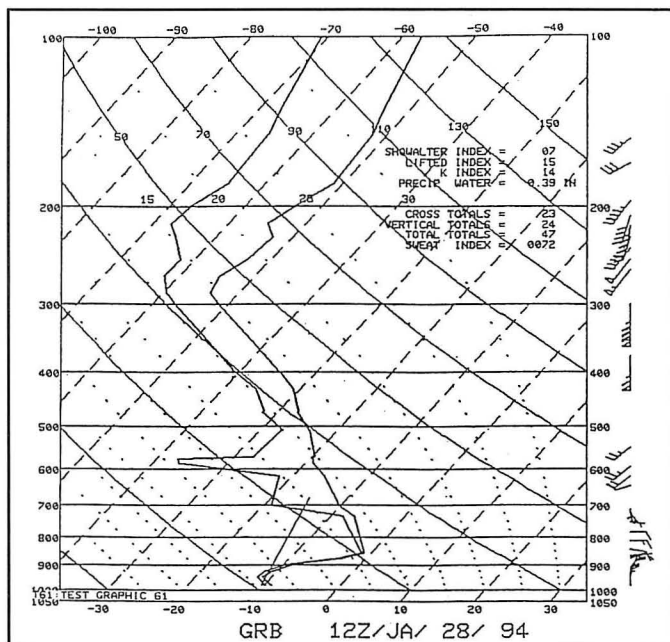


Fig. 1. Green Bay, Wisconsin sounding for 1200 UTC 28 January 1994.

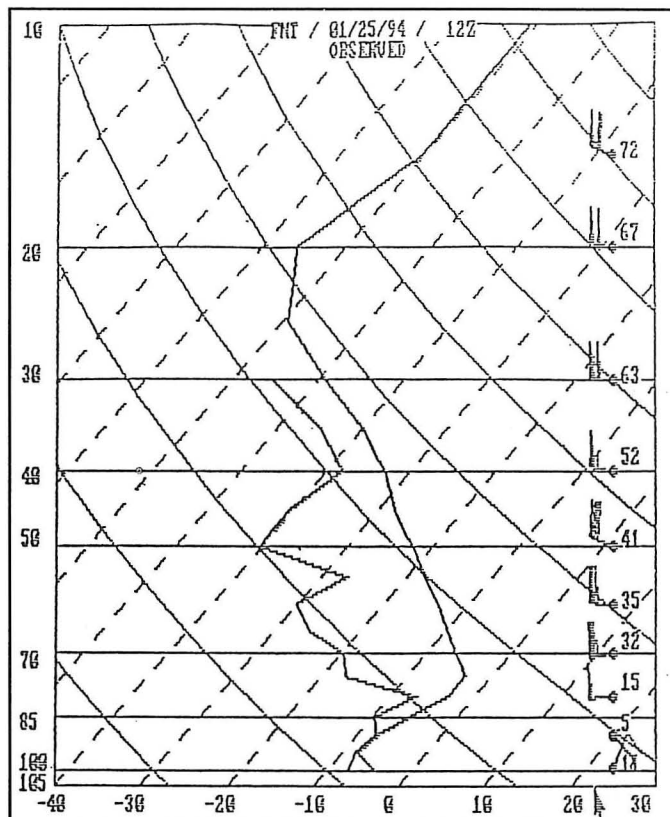


Fig. 2. Flint, Michigan sounding for 1200 UTC 25 January 1994.

crystal sublimation are unknown and will vary with different synoptic environments. The purpose here is to notify operational forecasters of 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-level 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°C in the dry layer (Moore 1992).

The elevated convection may erode the dry layer and reestablish/establish ice crystal seeding and diffusional ice crystal growth in 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 ice crystal sublimation in mid-tropospheric dry layers 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.

In the mature cyclone case, the aggressive mid-level 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, diffusional ice crystal growth 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 6 illustrate an example of the mature cyclone case over the lower Great Lakes on 28 January 1994. The 1200 UTC NWS RAFS (Regional Analysis and Forecast System) products predicted that thicknesses were at or below snow thresholds in the Milwaukee area at 1200 UTC (Figs. 3, 5 and 6). Typically, snow thresholds for 1000-500 mb, 1000-850 mb, and 850-700 mb are 5400 m, 1300 m and 1555 m (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. 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.

Figures 7-12 depict a shallow arctic air mass case over southern lower Michigan on 25 January 1994. The 1200 UTC RAFS surface and 850-mb analyses indicated that the leading edge of the arctic air was over the northern Ohio valley and the front was 100 to 150 miles south of southeastern lower Michigan. 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 25 January, with Flint reporting a mixture of snow and freezing drizzle until 1440 UTC. Thicknesses were

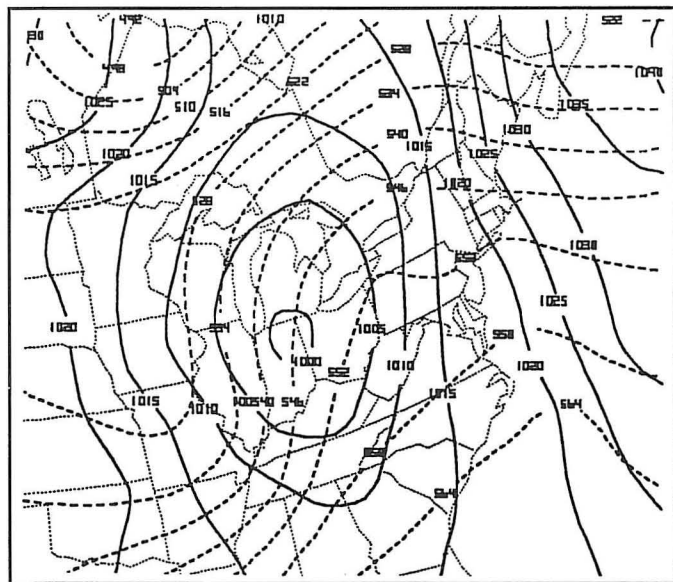


Fig. 3. Surface pressure (mb, solid) and 1000-500 mb thickness (dm, dashed) for 1200 UTC 28 January 1994 (RAFS 00-h analysis).

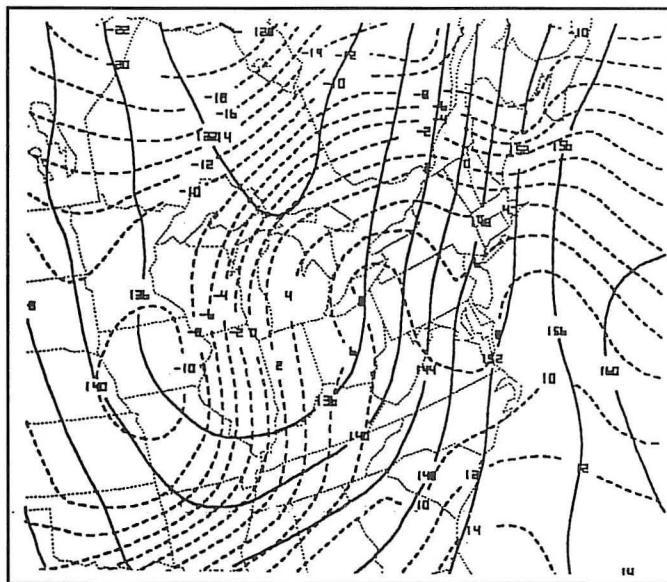


Fig. 4. 850-mb heights (dm, solid) and temperatures (C, dashed) for 1200 UTC 28 January 1994 (RAFS 00-h analysis).

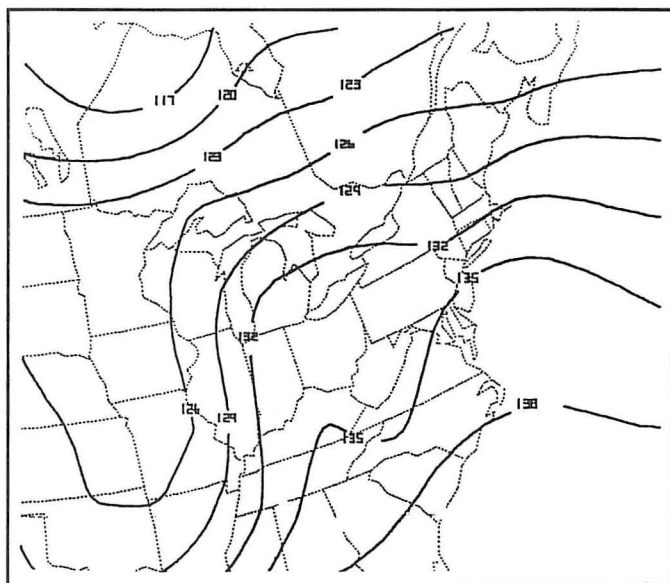


Fig. 5. 1000-850 mb thickness (dm) for 1200 UTC 28 January 1994 (RAFS 00-h analysis).

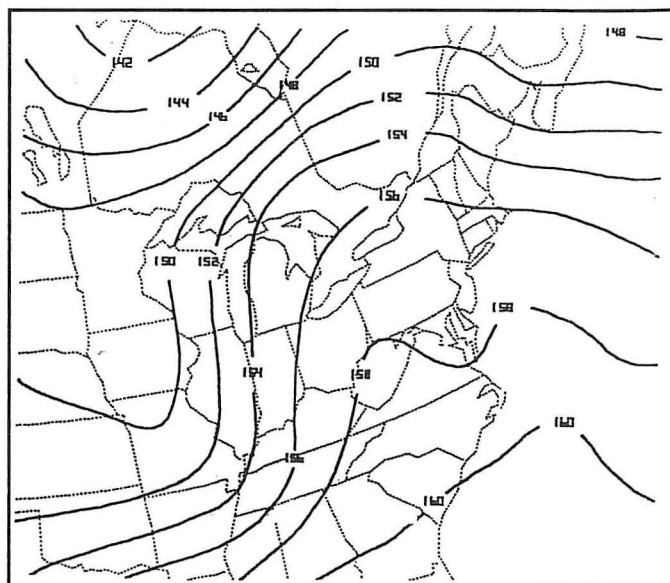


Fig. 6. 850-700 mb thickness (dm) for 1200 UTC 28 January 1994 (RAFS 00-h analysis).

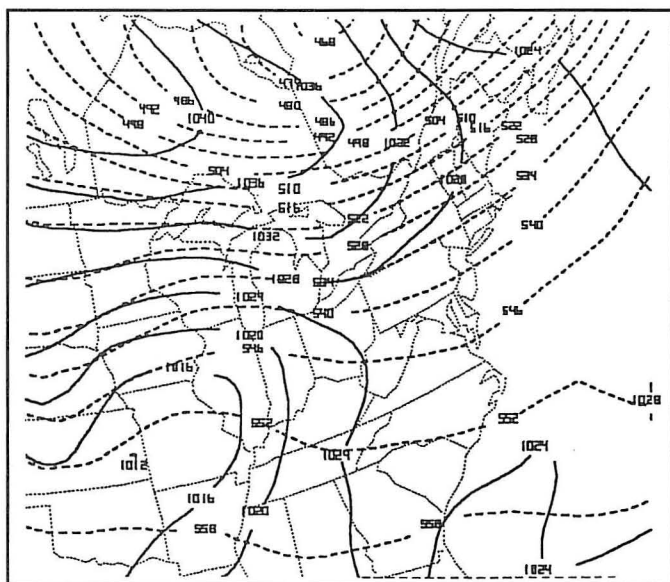


Fig. 7. Surface pressure (mb, solid) and 1000-500 mb thickness (dm, dashed) for 1200 UTC 25 January 1994 (RAFS 00-h analysis).

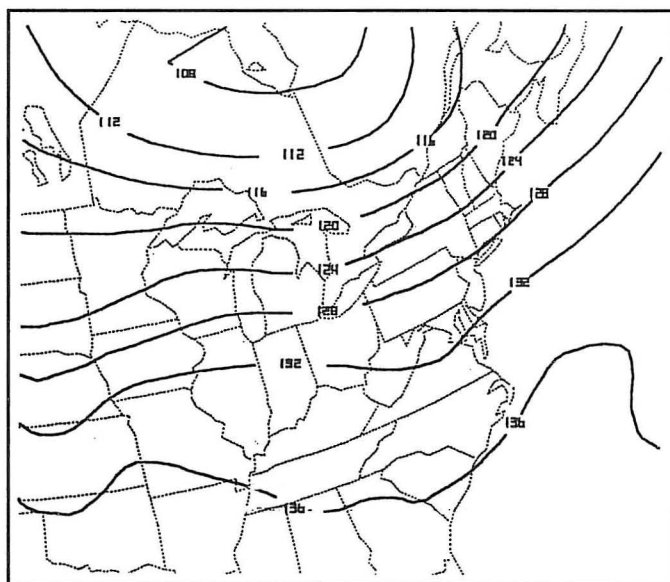


Fig. 8. 1000-850 mb thickness (dm) for 1200 UTC 25 January 1994 (RAFS 00-h analysis).

well below snow thresholds over all of southern Michigan at 1200 UTC (Figs. 7–9) 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 850-500 mb layer (not shown). Apparently the increasing mid-level moisture was substantial enough to reinforce ice crystal seeding in the lower moist layer and as a result, the precipitation changed to all snow in the Flint area after 1440 UTC.

The depth of the cold air and moisture associated with arctic air masses appears to play an important role in precipitation type. The forecaster should be alerted to the possibility of freezing precipitation in cases where the low-level moisture (generally below 800 mb) is in an environment warmer than  $-15^{\circ}\text{C}$  (but below  $0^{\circ}\text{C}$ ) and bounded by a dry layer above. It should be noted that the precipitation will often be light due to the discontinuous vertical atmospheric saturation. The forecaster must also be aware of mid-tropospheric moisture changes that may enhance or disrupt ice crystal seeding and diffusional growth in the lower moist layers. Using the NWS

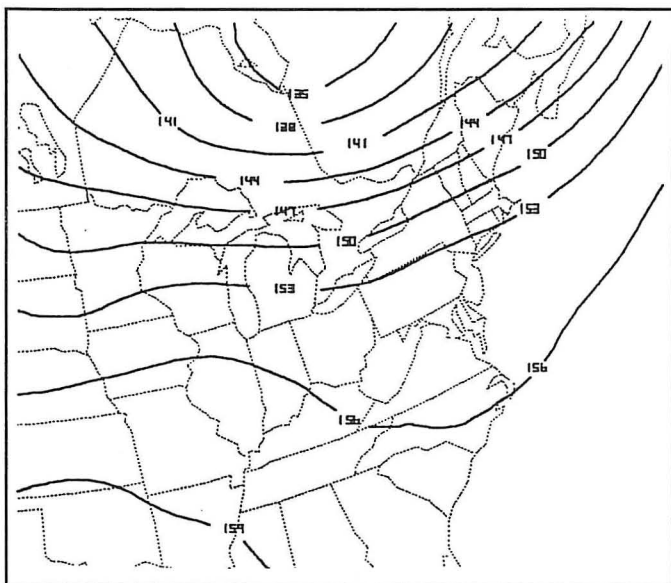


Fig. 9. 850-700 mb thickness (dm) for 1200 UTC 25 January 1994 (RAFS 00-h analysis).

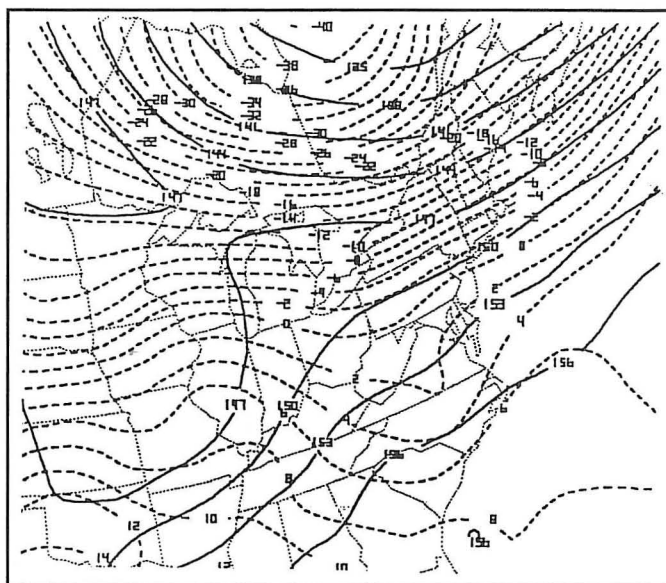


Fig. 10. 850-mb heights (dm, solid) and temperature (C, dashed) for 1200 UTC 25 January 1994 (RAFS 00-h analysis).

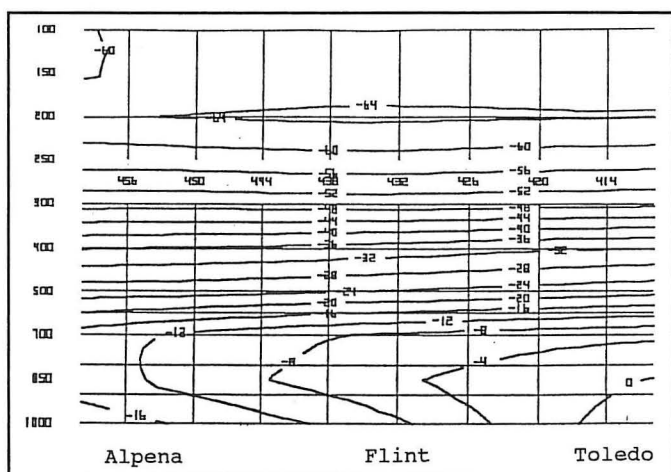


Fig. 11. Temperature (C) and height (mb) cross section (Alpena-Flint-Toledo) for 1200 UTC 25 January 1994 (RAFS 00-h analysis).

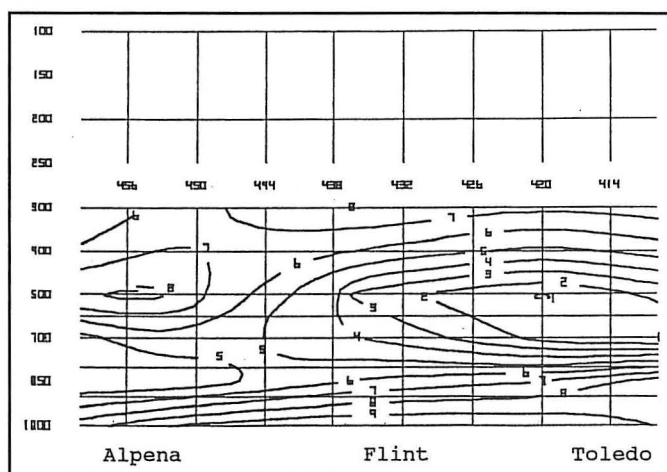


Fig. 12. Relative humidity ( $\times 10\%$ ) and height (mb) cross section (Alpena-Flint-Toledo) for 1200 UTC 25 January 1994 (RAFS 00-h analysis).

PCGRIDDS software package to generate height sections of moisture and temperature fields (Figs. 11 and 12) may provide the forecaster with valuable insight into the determination and timing of freezing precipitation or mixed precipitation associated with this process.

## 6. Forecaster Checklist

Several atmospheric parameters must be assessed by the forecaster to determine if ice crystal seeding and diffusional growth will occur or be disrupted. The following checklist provides both qualitative guidelines for assessing the possibility of freezing precipitation or a mixture of snow and freezing precipitation. The guidelines are only applicable for below freezing temperatures through the entire depth of the troposphere.

- Does a lower tropospheric moist layer extend upward to where temperatures are  $-15^{\circ}\text{C}$  or colder?  
If not, then freezing precipitation is possible.
- Is a mid-tropospheric dry layer present or forecast?  
If so, then the ice crystal seeding and diffusional growth may be disrupted and a mixture of snow and freezing precipitation or just freezing precipitation is possible.
- How deep is the mid-level dry layer?  
Mid-level dry layers deeper than 5,000 feet with temperature/dew point spreads greater than  $10^{\circ}\text{C}$  may be significant. The precipitation may completely change to freezing precipitation or a prolonged period of snow and freezing precipitation is possible.
- Is mid-tropospheric moisture increasing?  
If freezing precipitation is occurring and mid-level mois-

ture is increasing then the precipitation may change to all snow.

e. Be aware of elevated convection.

Elevated convection (due to the release of potential instability) may erode the mid-tropospheric dry layer and reestablish/establish ice crystal seeding and diffusional growth in the lower tropospheric moist layer, ideally changing the precipitation over to all snow.

## 7. Conclusions

The disruption of ice crystal seeding and diffusional growth is an important atmospheric process that is often overlooked. It can provide the forecaster with valuable insight into precipitation type when below freezing temperatures are present through the entire depth of the troposphere. The forecaster should be alerted to the possibility of freezing precipitation when the low-level moist layer is confined to an atmosphere warmer than approximately  $-15^{\circ}\text{C}$ , as this is the warmest temperature in which efficient ice crystal formation and growth occurs in clouds. Two synoptic environments in which ice crystal seeding and diffusional growth can be disrupted, in lower tropospheric moist layers, and thus influence precipitation type are: 1) aggressive "dry slotting" associated with mature cyclones and 2) arctic air masses where the cold air and moisture fields are shallow.

Mid-tropospheric dry layers appear to play an important role in the ice crystal seeding and diffusional growth process in lower tropospheric moist layers. Upper tropospheric moist layers cold enough to support heterogeneous ice nucleation, and subsequent ice crystal growth, may be partially or totally cut off from the lower level moisture as mid-level dry layers sublimate the ice crystals before they reach the lower level moisture. As a result, a mixture of snow and freezing precipitation may occur at the surface. The precipitation may completely change to freezing precipitation if the mid-level dry layer is substantial and temporally persistent. Dry layers deeper than 5,000 ft with temperature/dew point spreads greater than  $10^{\circ}\text{C}$  may be substantial and result in several hours of freezing precipitation. More work is needed to obtain quantitative spatial and temporal parameters to assess mid-tropospheric dry layers and the associated precipitation type. The result will be improved predictability.

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