

THE EASTERN MISSOURI SLEET STORM OF 29–30 DECEMBER 1990: A DIAGNOSTIC ANALYSIS

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Abstract

Those physical mechanisms responsible for an extended period of sleet over eastern Missouri during 29–30 December 1990 are investigated. The post-frontal precipitation pattern is shown to be the result of a direct thermal circulation forced by frontogenesis along an arctic front. This circulation was apparently further enhanced by vertical motions associated with the entrance region of an upper-level jet streak. 1000–850 mb and 850–500 mb “partial” thicknesses are shown to be useful in predicting sleet as opposed to freezing rain or snow. A Canadian technique using these partial thicknesses is described and applied to this unusual event.

1. Introduction

During the evening of 29 December 1990 an arctic cold front moved southeastward through the Mississippi Valley. As surface temperatures in St. Louis, Missouri dropped from 12.8°C (55°F) at 1800 UTC 29 December to –2.2°C (28°F) by 0000 UTC 30 December, precipitation changed from rain to freezing rain, and eventually to sleet (ice pellets). As seen in Fig. 1, sleet, moderate to heavy at times, continued from 0100 UTC until 1600 UTC on 30 December. Over this time period sleet accumulated to around 5.1 cm (2 in.) in the St. Louis Metropolitan area, crippling transportation and restricting even walking for over a week (NCDC 1990).

In this paper we examine those physical processes contributing to this extended period of post-cold front (anafront) sleet. In particular, we will focus our discussion on the development of a direct thermal circulation (DTC), which was likely the result of low-level frontogenetical forcing coupled with the secondary circulation in the entrance region of a strong upper-level jet (ULJ). Techniques for diagnosing this circulation and predicting the type of precipitation are discussed with the aim of anticipating these wintertime extended precipitation events.

2. The Arctic Front

At 0000 UTC 30 December 1990 an arctic cold front stretched from lower Michigan southwestward into central Texas. Temperatures ahead of the front were greater than 15.5°C (60°F) from Ohio southward, while behind the front cold arctic air of less than –23.3°C (–9.9°F) moved through the

Central Plains (Fig. 2a). Precipitation, mostly in the form of rain and freezing rain, was found along and behind the arctic boundary.

By 1200 UTC 30 December the arctic front had moved through Missouri to a position extending from Lake Erie southwestward to eastern Texas (Fig. 2b). The radar summary (Fig. 2b) reveals that a good deal of the precipitation was now post-frontal. This precipitation pattern is characteristic of an anafront type cold front, described by Moore and Smith (1989) as a cold front whose frontal speed exceeds that of the winds normal to the front aloft, thus generating relative upslope flow, post-frontal cloudiness and precipitation. In addition, anafront cold fronts typically are associated with a sudden temperature drop, sharp veering winds, high post-frontal relative humidity and slow clearing with frontal passage. At this time, St. Louis, Missouri was reporting heavy ice pellets and light snow, even though video integrator and processor (VIP) levels of only 2 (30–40 dBZ) were reported.

Arctic cold fronts are typically very shallow and this case was no exception. Figure 3, an isentropic cross section taken normal to the front at 1200 UTC 30 December from Bismarck, North Dakota to Slidell, Louisiana, reveals a highly compacted, sloping, frontal zone extending from around Topeka, Kansas (TOP) at 700 mb down to a surface position between Little Rock, Arkansas (LIT) and Jackson, Mississippi (JAN). A further indication of the shallowness of the arctic air can be seen in Fig. 4, a sounding taken at Paducah, Kentucky (PAH) at 1200 UTC 30 December. Note the well-defined inversion between 962 mb, where the temperature is –1.3°C, and 898 mb where the temperature is 14.0°C! The wind shift in the layer is equally dramatic with winds of 8 kts from 332° shifting to 17 kts from 224° at the top of the inversion.

3. Low-Level Frontogenesis

The surface temperature advection fields for 0000 and 1200 UTC 30 December (Fig. 5a–b) reveal diagnostic evidence of a positive/negative thermal advection couplet over a meso α scale distance (several hundred kilometers). At 0000 UTC warm air advection of greater than +0.4 K hr^{–1} across southern Illinois and Indiana is contrasted with cold air advection of less than –1.6 K hr^{–1} across south-central Missouri. By 1200 UTC this thermal advection gradient is weaker and has shifted slightly southeast over western Tennessee (Fig. 5b).

At 0000 UTC 30 December, the 850-mb temperature advection field reveals cold air advection dominating Missouri (Fig. 6a). However, by 1200 UTC a region of marked gradient in thermal advection develops across Missouri with

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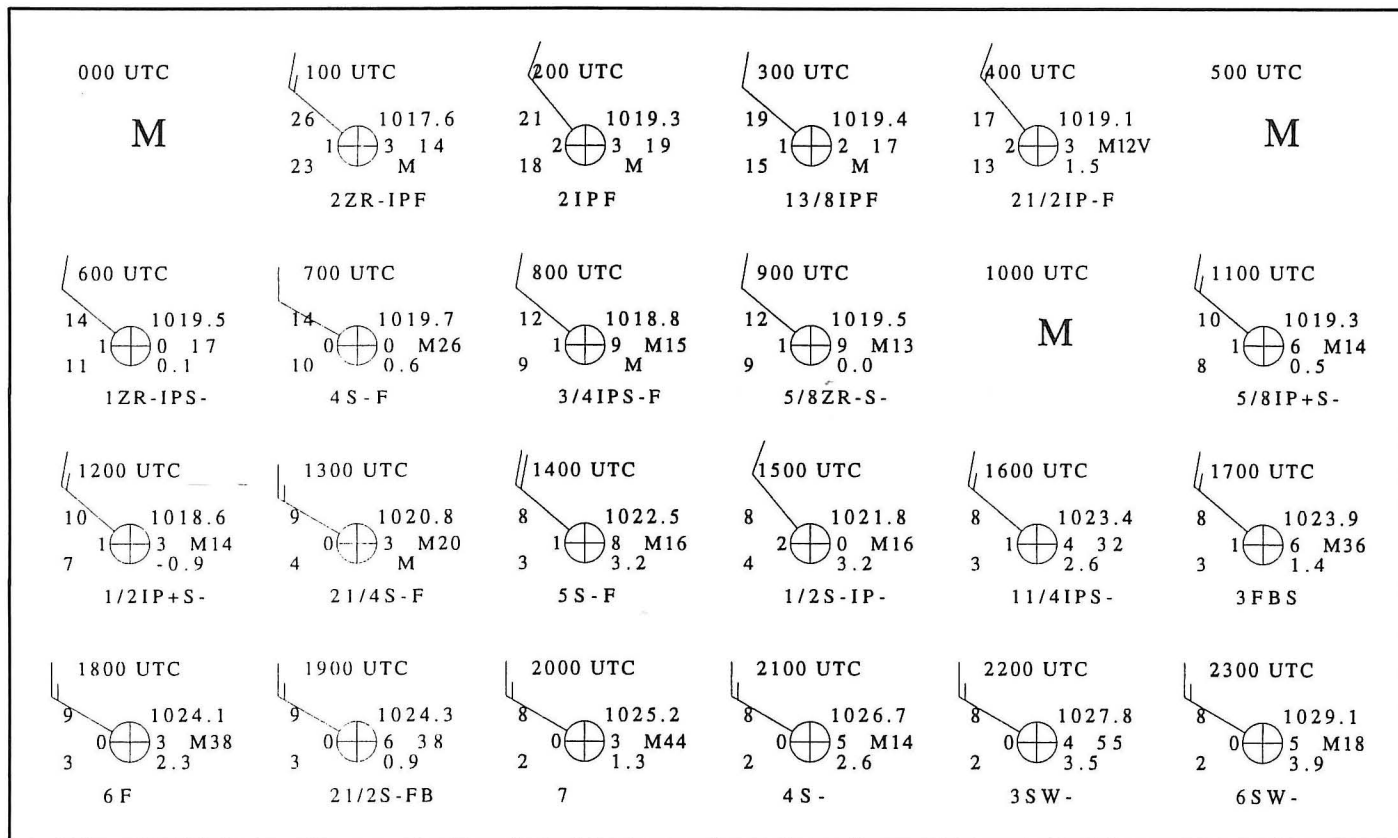


Fig. 1. Time series of station model reports from St. Louis, MO during 30 December 1990. Upper left: temperature ($^{\circ}\text{F}$); bottom left: dew point ($^{\circ}\text{F}$); upper right: mean sea level pressure (mb); middle left: tens digit of wind direction; middle right: units digit of wind speed (knots); bottom right: pressure tendency (3h, in mb); middle right outermost: ceiling in hundreds of feet; bottom middle: visibility in miles and weather. M means missing observation.

values of $+0.5 \text{ K hr}^{-1}$ juxtaposed with values of -2.5 K hr^{-1} (Fig. 6b). Note that the thermal advection gradient at 850 mb is much stronger and to the north of the surface thermal advection gradient (Fig. 5b) at this time.

The two-dimensional form of the frontogenetical function (Petterssen 1956) was computed using:

$$F = \frac{d}{dt} \left| \nabla \theta \right| = \frac{1}{2} \left| \nabla_H \theta \right| \left\{ D_H \cos 2\beta - \text{Div} \right\} \quad (1)$$

where D_H is the resultant deformation, Div is the horizontal divergence and β is the angle measured from the axis of dilatation to the potential isotherms. The resultant deformation is computed as the square root of the sum of the squares of the stretching and shearing deformation. The effect of horizontal deformation is to promote frontogenesis when the axis of dilation lies within 45° of the potential isotherms and to promote frontolysis when the axis of dilatation lies between 45° and 90° of the potential isotherms. Convergence acts frontogenetically while divergence acts frontolytically. At 0000 UTC 30 December the frontogenetical function (Fig. 7a) reveals values exceeding $20 \text{ K (100 km)}^{-1} (3 \text{ hr})^{-1}$ parallel to the thermal advection gradient region in Fig. 5a. The most intense regions appear in eastern Missouri and northern Texas. By 1200 UTC 30 December the frontogenetical forcing (Fig. 7b) has weakened considerably while shifting south-eastward, again reflecting the changes in surface temperature advection.

At 850 mb the two-dimensional, adiabatic form of Miller's (1948) frontogenesis equation in pressure coordinates was used to evaluate the frontogenetical function (see Moore and Blakley 1988). At 0000 UTC 30 December weak values of the frontogenetical function appear across Oklahoma (Fig. 8a). However, by 1200 UTC 30 December values greater than $30 \times 10^{-10} \text{ K m}^{-1} \text{ s}^{-1}$ are found across southern Illinois/southeastern Missouri, west Texas and New Mexico.

The convergence field associated with the arctic front in proximity to a very strong temperature contrast across the front undoubtedly increased the frontogenetical field over meso α scale distances (i.e., across the frontal zone). As discussed in Holton (1979), frontogenesis, acting over suitably long time periods, leads to a response in the momentum field in the form of a direct thermal circulation (DTC)—a secondary, ageostrophic vertical circulation created to rectify the imbalance between the thermal (thickness) field and the thermal wind (shear of the geostrophic wind).

Figures 9a–b reveal how this DTC develops over a 12 hr period (0000–1200 UTC 30 December) normal to the frontal zone. At 0000 UTC (Fig. 9a) one can see the initial stage of the DTC with a well-defined rising branch between points A and B in Missouri (see Fig. 7a), but no indications of a downward branch. However, a complete DTC cell is obvious by 1200 UTC (labeled “D” in Fig. 9b) with rising motion ahead of the cold front and sinking in the colder air. The expansion of the anafront (post-frontal) precipitation pattern

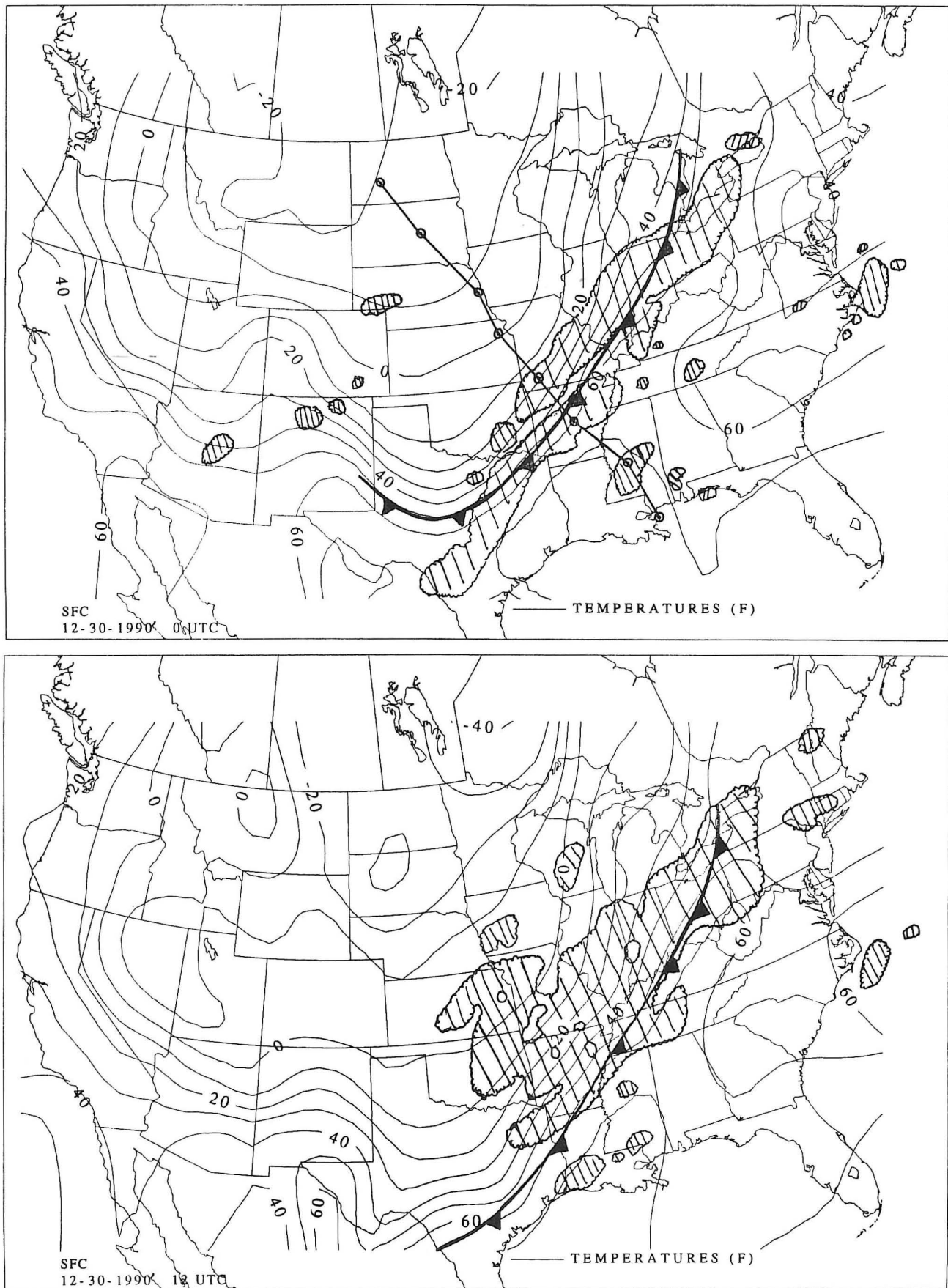


Fig. 2. (a) Isotherms (°F) at surface for 0000 UTC 30 December 1990. Precipitation region scalloped. Cold front designated with standard symbols. Line denotes cross section seen in Fig. 3. (b) Same as Fig. 2a, except for 1200 UTC 30 December 1990.

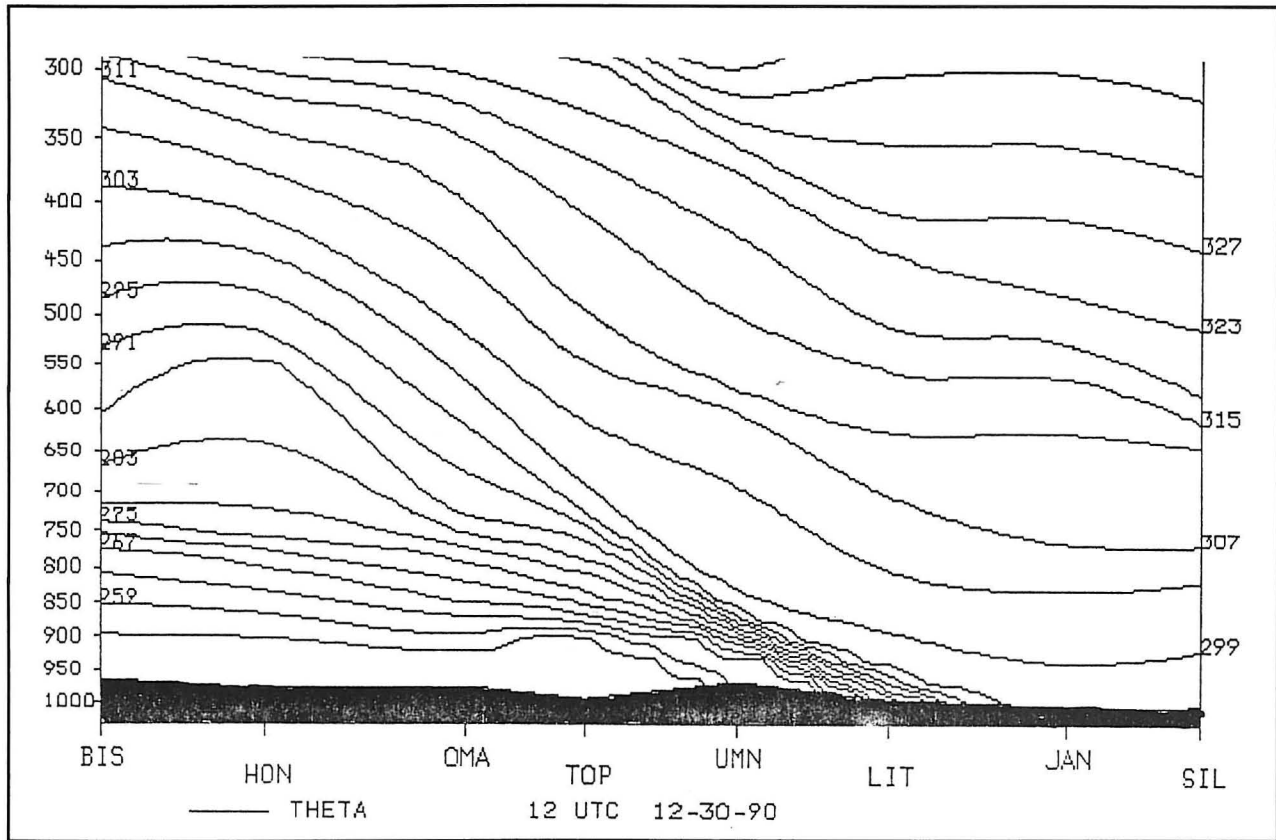


Fig. 3. Isentropic cross section along line shown in Fig. 2a at 1200 UTC 30 December 1990. Solid lines are isentropes. Rawinsonde stations listed across the x-axis are: BIS = Bismarck, North Dakota; HON = Huron, South Dakota; OMA = Omaha, Nebraska; TOP = Topeka, Kansas; UMN = Monett, Missouri; LIT = Little Rock, Arkansas; JAN = Jackson, Mississippi; and SIL = Slidell, Louisiana.

and the extension of the time scale of the precipitation are a direct result of the DTC created by the low-level frontogenetical forcing. However, as will next be discussed, the role of an ULJ in the development of the DTC needs to be considered as well.

4. Upper-Level Jet: Entrance Region Vertical Circulation

As the DTC seen in Fig. 9a–b extended to at least 400 mb, we investigated the possible influence of ULJ dynamics upon the vertical circulation. As discussed by Uccellini and Johnson (1979), an indirect thermal circulation (ITC) is typically found in the exit region of a straight-line ULJ, while a DTC is found in the entrance region of the ULJ, as the mass field adjusts to the momentum field. Figure 10 shows that at 1200 UTC 30 December a jet core exceeding 140 kts (70 m s^{-1}) was located over James Bay, Canada. Divergence is found on the equatorward side of this ULJ streak extending from eastern Oklahoma northeastward into Michigan with values greater than $2 \times 10^{-5} \text{ s}^{-1}$ over central Indiana. Convergence less than $-2 \times 10^{-5} \text{ s}^{-1}$ is found over the eastern half of Iowa.

The divergence/convergence fields associated with the entrance region of this strong straight-line jet streak being superimposed over the established DTC associated with the low-level frontogenetical field reinforced and deepened the DTC.

5. Predicting Sleet

Predicting mixed precipitation or extended periods of freezing rain or sleet can be difficult. Using the 1000–500 mb thickness together with the mean potential temperature of the lowest 100 mb near the ground can yield mixed signals. In this case, for instance, the 1000–500 mb thickness and mean potential temperature varied from 5530 gpm/270 K at 0000 UTC to 5460 gpm/270 K at 1200 UTC 30 December at St. Louis. Using the 5400 gpm as a cut-off value for liquid/solid precipitation must be reconciled with the cold boundary layer temperatures (usually values $< 280 \text{ K}$ indicate frozen precipitation).

Figure 11 taken from a U.S. Air Force training manual (USAF 1983) is a nomogram designed for mixed precipitation forecasting. It utilizes two thicknesses, 1000–850 mb and 850–700 mb. We have plotted these thickness values for both time periods. This approach works better than using the 1000–500 mb thickness alone, since it considers more shallow layers. Even so, at 0000 UTC it predicts rain, since the 850–700 mb layer is so warm. At 1200 UTC this nomogram predicts ice pellets/freezing rain mostly due to the cold 1000–850 mb thickness value. Numerical models must have good resolution of the lowest 200 mb to resolve arctic boundaries such as the one shown here.

Canadian meteorologists Cantin and Bachand (1992) discuss a technique for forecasting precipitation type based upon partial thicknesses (i.e., 850–700 mb and 1000–850 mb

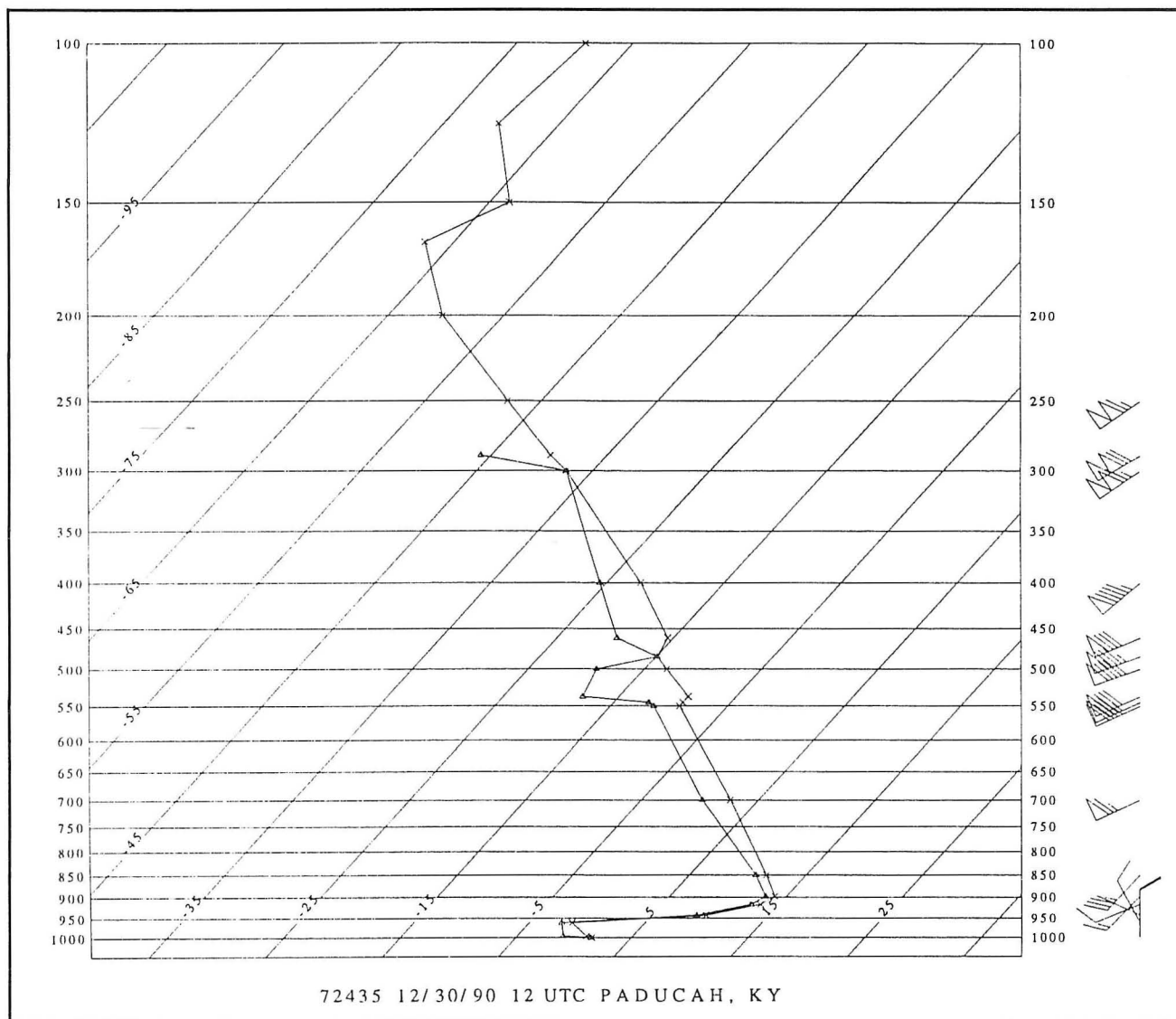


Fig. 4. Sounding for Paducah, Kentucky (PAH) for 1200 UTC 30 December 1990. Wind barbs on right hand side are in knots.

thicknesses) and a subjective evaluation of the vertical motion and low-level thermal advection magnitudes (see Table 1). The term “*overlapping*” refers to whether the 850-700 mb thickness (specifically, the 1540 gpm thickness line) overlaps or is north of the 1310 gpm 1000-850 mb thickness line. Overlapping in this fashion implies temperatures greater than 0°C in the 850-700 mb layer overlying temperatures less than 0°C in the 1000-850 mb layer. The latter situation implies a thermal stratification typical of sleet (ice pellets) and/or freezing rain. Two other important parameters to note are the vertical motion and low-level thermal advection. As a rule of thumb, strong upward vertical motion and/or strong low-level cold air advection favor solid precipitation (ice pellets/snow) over liquid precipitation (freezing rain/rain).

Conditions at 0000 UTC 30 December at St. Louis, Missouri included: 1000-850 mb thickness = 1335 gpm, 850-700 mb thickness = 1590 gpm with weak upward vertical motion (see Fig. 9a) and strong low-level cold advection (see Fig. 5a). The 1540 gpm thickness overlapped the 1310 gpm thickness well north and west of the St. Louis area. Under these

conditions, Table 1 suggests all rain, possibly changing to freezing rain as the 1000-850 mb thickness lowered. Conditions at 1200 UTC 30 December at St. Louis, Missouri included: 1000-850 mb thickness = 1300 gpm, 850-700 mb thickness = 1575 gpm, strong upward vertical motion (see Fig. 9b) and moderate low-level cold air advection (see Fig. 5b). The 1540 gpm thickness overlapped the 1310 gpm thickness over the St. Louis area. These conditions would suggest freezing rain and possibly ice pellets as the 1000-850 mb thickness decreased, according to the criteria in Table 1. This table appears to offer useful criteria for forecasting precipitation type in a complex thermal stratification setting.

The National Weather Service Techniques Development Laboratory (TDL) has developed statistical schemes for forecasting precipitation type. Bocchieri and Maglaras (1983) discussed a conditional probability of precipitation type (PoPT) model output statistics (MOS) scheme developed for the limited-area fine mesh model (LFM). It discriminated among snow or ice pellets, freezing rain or drizzle, and rain or mixed types. The probabilities are conditional, because

	THICKNESS	THICKNESS	PRECIPITATION TYPES	
	850-700	1000-850	Signif. up. V.V. and /or signif. low lvl. cold air advection	Weak up. V.V. and near zero low lvl. cold air advection
WITHOUT OVER-LAPPING	<154 <154 <154	<129 129-131 >131	S S (R #1) R	S (IP/ZL) R/S (R #1) R
WITH OVER-LAPPING	<154 <154 <154	<129 129-131 >131	S (IP #2) S/IP (ZR #2) (R #1) R	S (IP/ZR #2) IP/S (R #1) (ZR/ZL #3) R
	>154 >154 >154	<129 129-131 >131	IP (S #2) ZR (IP #2) R (often >132)	ZL/ZR/IP ZR/ZL (R/L) R

TABLE 1. Precipitation types according to partial thickness intervals and upward vertical velocity (evaluated subjectively). (): Types between parentheses are possible but not dominant except for those with one of the three specific following conditions: #1) warm sector or southerly winds; #2) close to the 154 dam; and #3) ahead of warm front and > 152 dam. X/Y: The two types (or more) are equally possible (Cantin and Bachand 1992).

the system assumes precipitation will occur. An updated form of the MOS precipitation type guidance for the nested-grid model (NGM) is described by Dallavalle et al. (1992). The NGM-MOS precipitation type scheme distinguishes among rain, freezing or ice pellets and snow. The most important numerical model predictors include: 850-mb temperature, boundary layer potential temperature, and 850-500 mb and 1000-850 mb thicknesses. The increased vertical resolution, especially in the lower troposphere, of the NGM versus the LFM provides a greater ability to discriminate among liquid and freezing type precipitation. NGM-MOS values for this case were unfortunately unavailable to compare to the previous methods.

6. Summary and Conclusions

An unusually long period of sleet accompanied the passage of an arctic cold front in east-central Missouri on 30 December 1990. It has been shown that this anafont type precipitation regime was the product of a direct thermal circulation due to the combined effects of a frontogenetical circulation and the vertical circulation in the entrance region of an ULJ streak. The net result of these superimposed vertical circulations was to reinforce the upward vertical motion behind the front and extend the time scale of the precipitation. Koch (1984) has described a mesoscale frontogenetical circulation that helped initiate a warm season squall line that is parallel to this work.

In addition, it has been shown that as the arctic frontal zone was very shallow, precipitation forecasting was tricky; utilizing schemes relying on large layer thicknesses (e.g., 1000-500 mb) would forecast rain. A U.S. Air Force nomogram for forecasting precipitation type offered some utility as it considered more shallow thickness values for input. However, criteria for forecasting precipitation type, discussed by Cantin and Bachand (1992), based on 1000-850

mb, and 850-700 mb thicknesses, and a subjective evaluation of the strength of the upward vertical motion and low-level thermal advection also proved useful for discriminating precipitation type.

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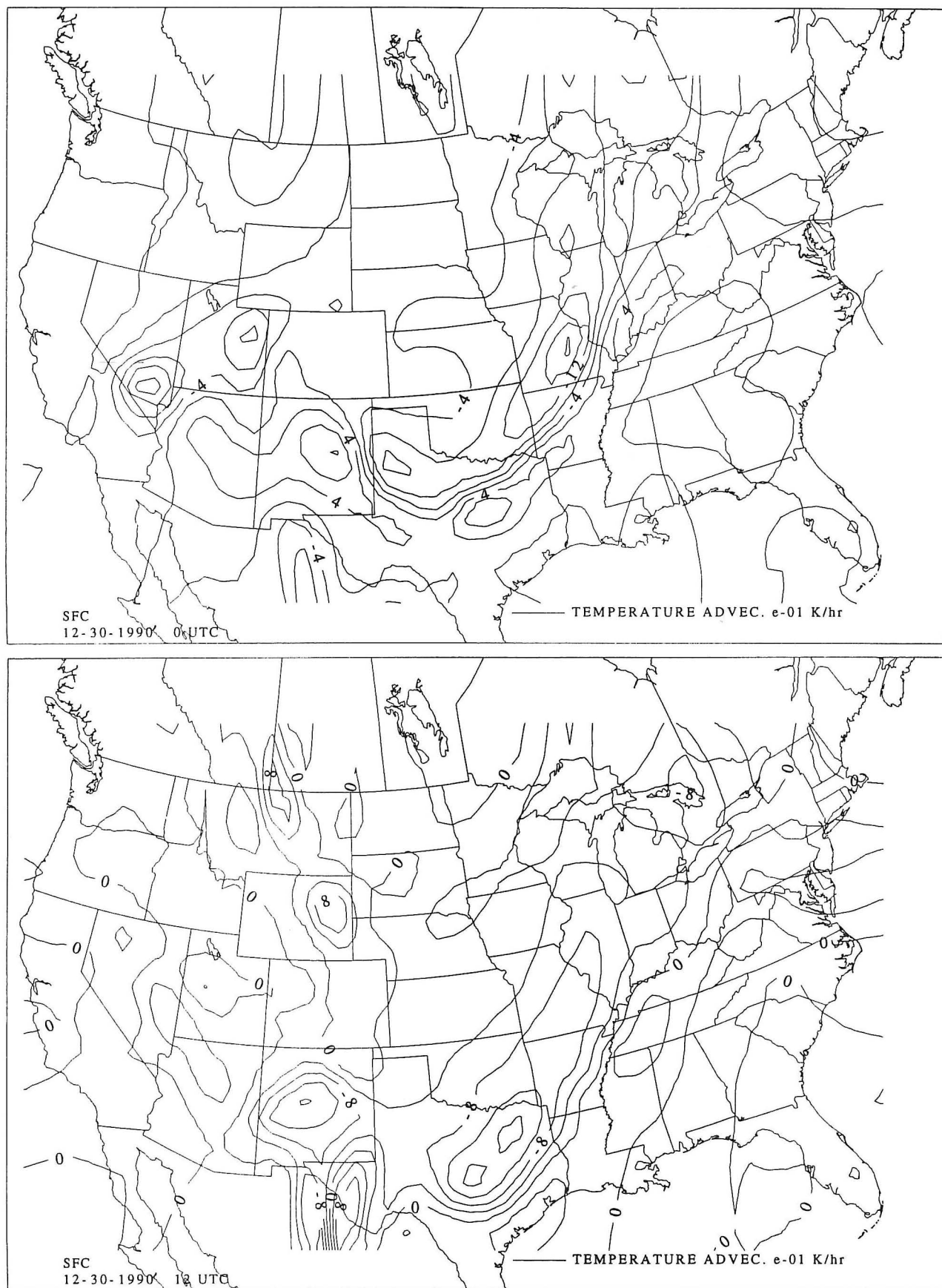


Fig. 5. (a) Surface temperature advection ($10^{-1} \text{ K hr}^{-1}$) for 0000 UTC 30 December 1990. (b) Same as Fig. 5a, except for 1200 UTC 30 December 1990.

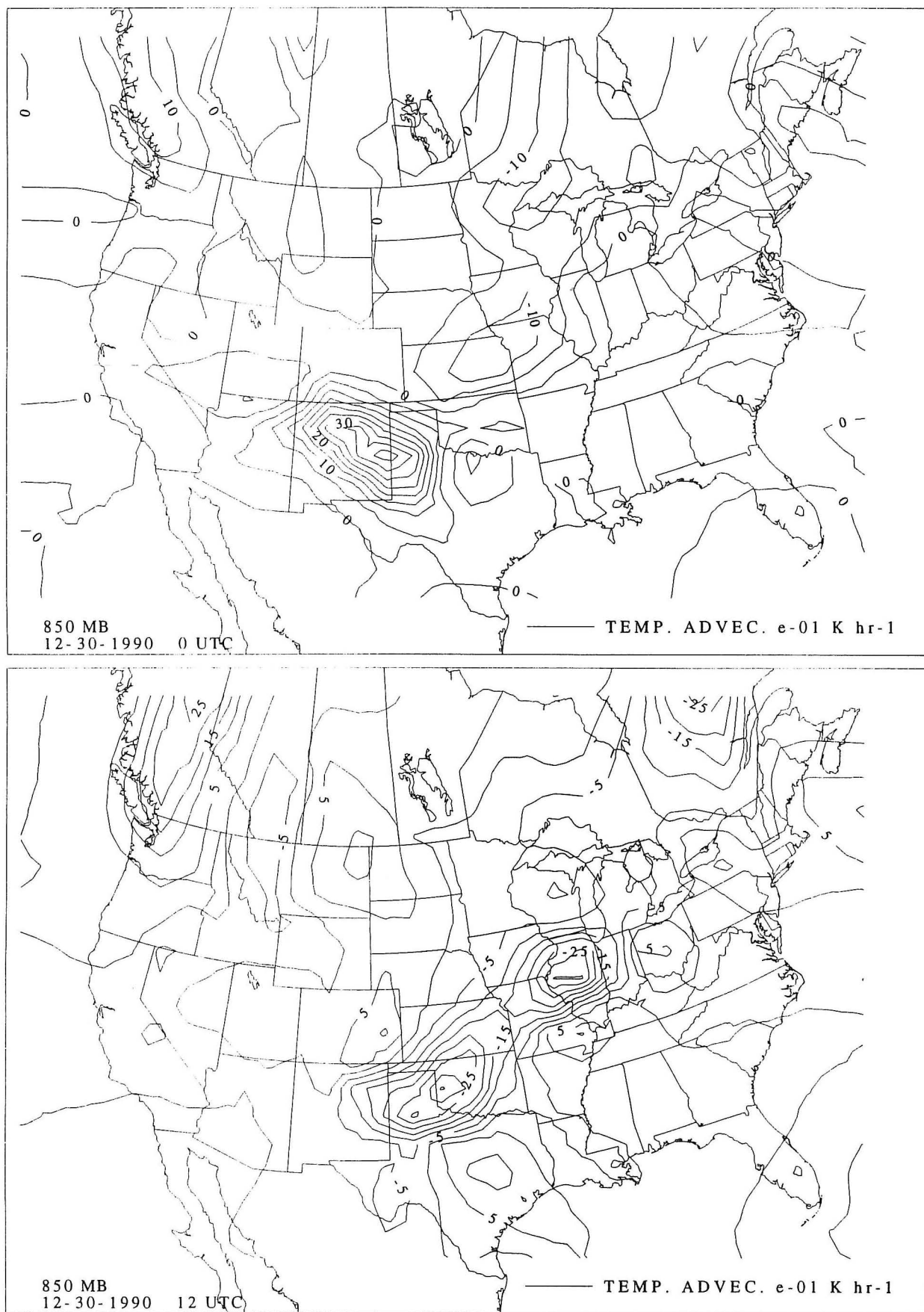


Fig. 6. (a) 850-mb temperature advection ($10^{-1} \text{ K hr}^{-1}$) for 0000 UTC 30 December 1990. (b) Same as Fig. 6a, except for 1200 UTC 30 December 1990.

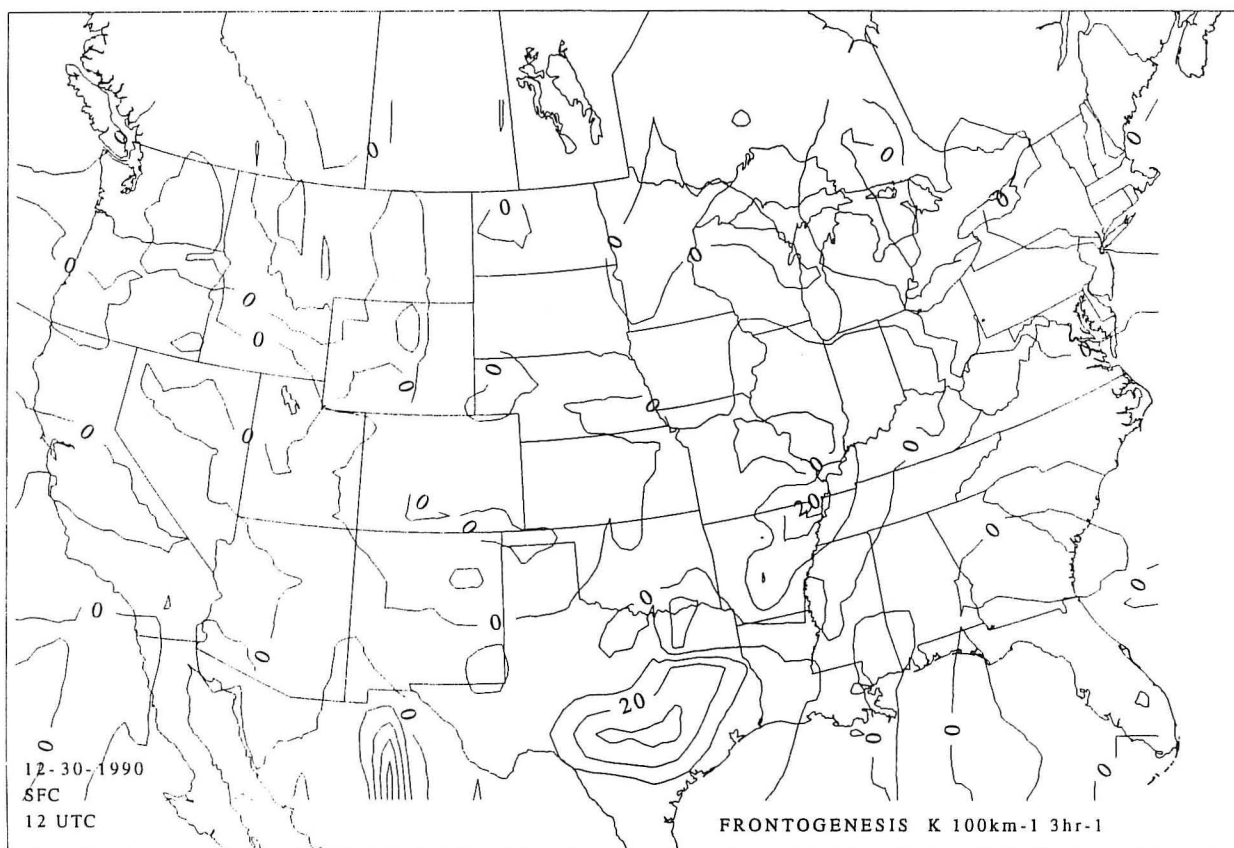
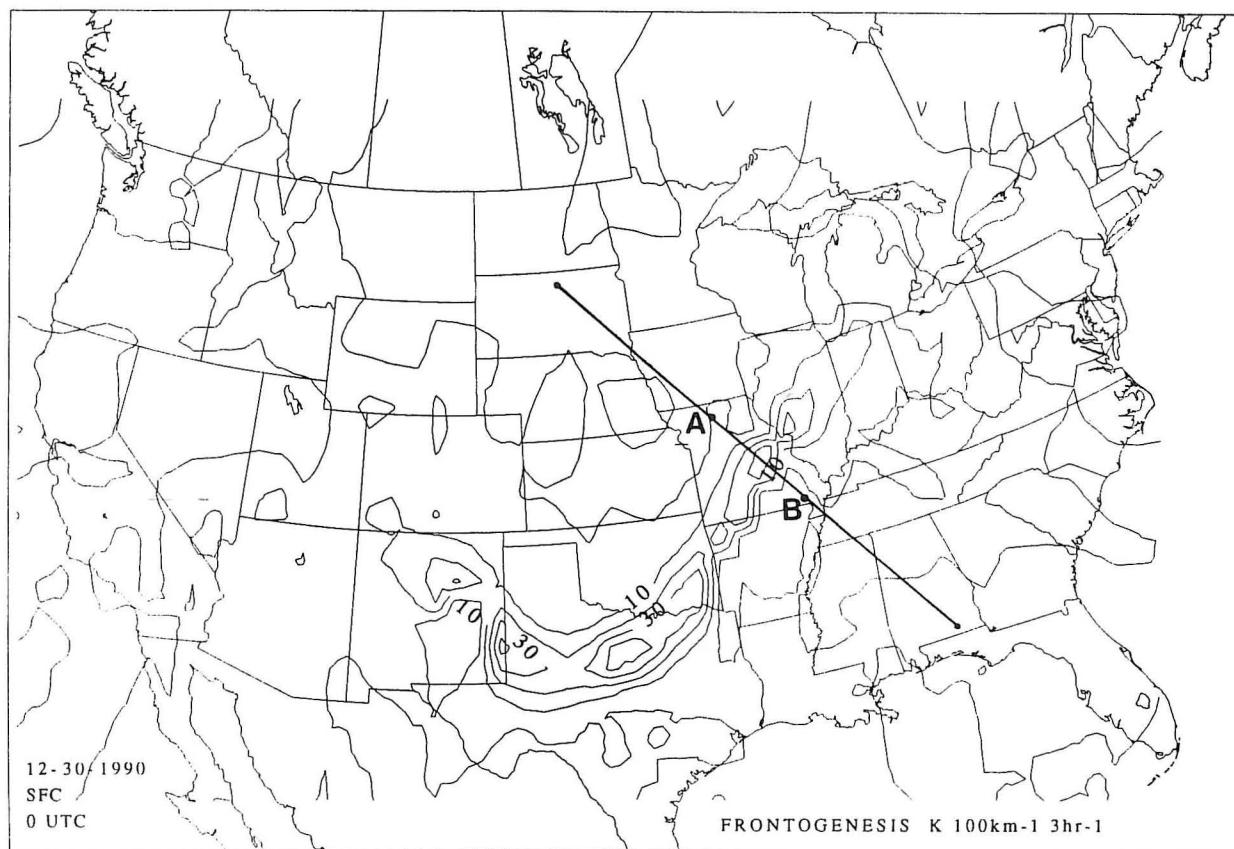


Fig. 7. (a) Surface frontogenesis [$K (100 \text{ km})^{-1} (3 \text{ hr})^{-1}$] for 0000 UTC 30 December 1990. (b) Same as Fig. 7a, except for 1200 UTC 30 December 1990. The line shown in Fig. 7a is the line along which the cross sections in Fig. 9a-b were taken.

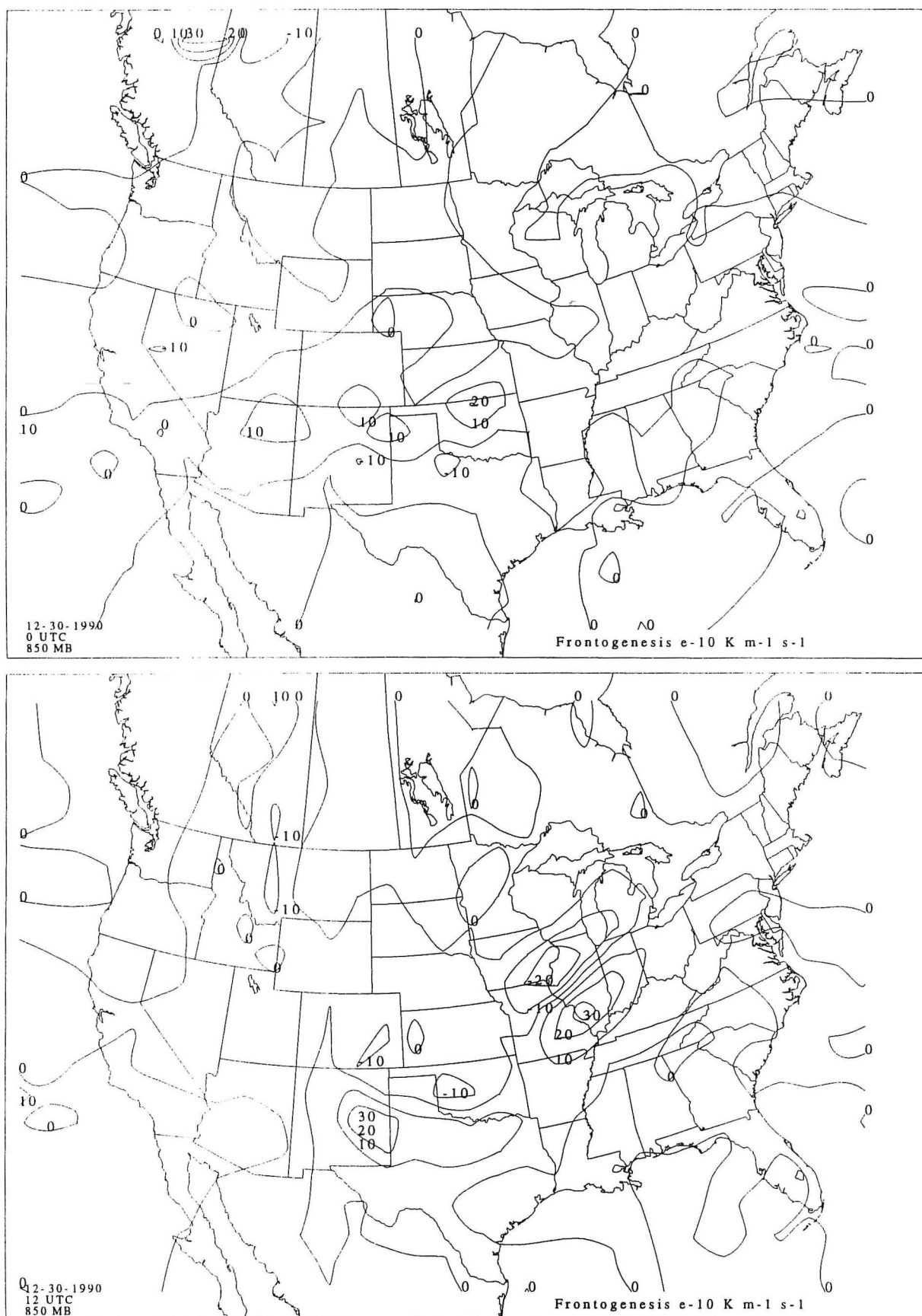


Fig. 8. (a) 850-mb frontogenesis ($10^{-10} \text{ K m}^{-1} \text{ s}^{-1}$) for 0000 UTC 30 December 1990. (b) Same as Fig. 8a, except for 1200 UTC 30 December 1990.

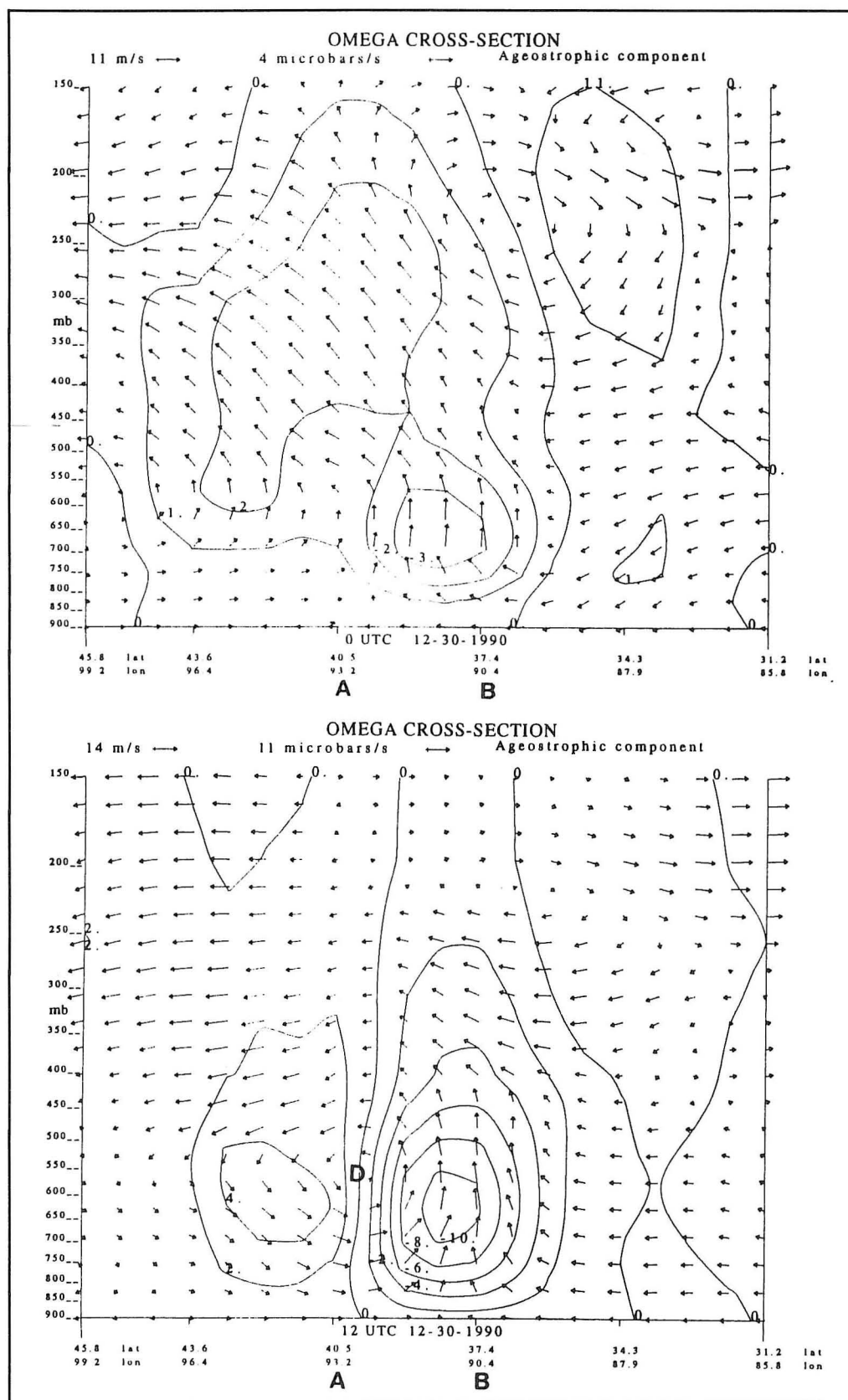


Fig. 9. Cross section along line shown in Fig. 7a depicting ageostrophic wind component tangent to the cross sectional plane (horizontal component of arrow) and vertical motion (vertical component of arrow) for (a) 0000 UTC and (b) 1200 UTC 30 December 1990. Points A and B, shown along the abscissa, are along the cross section shown in Fig. 7a. Solid lines are isopleths of vertical motion (ω) in $\mu\text{bars s}^{-1}$.

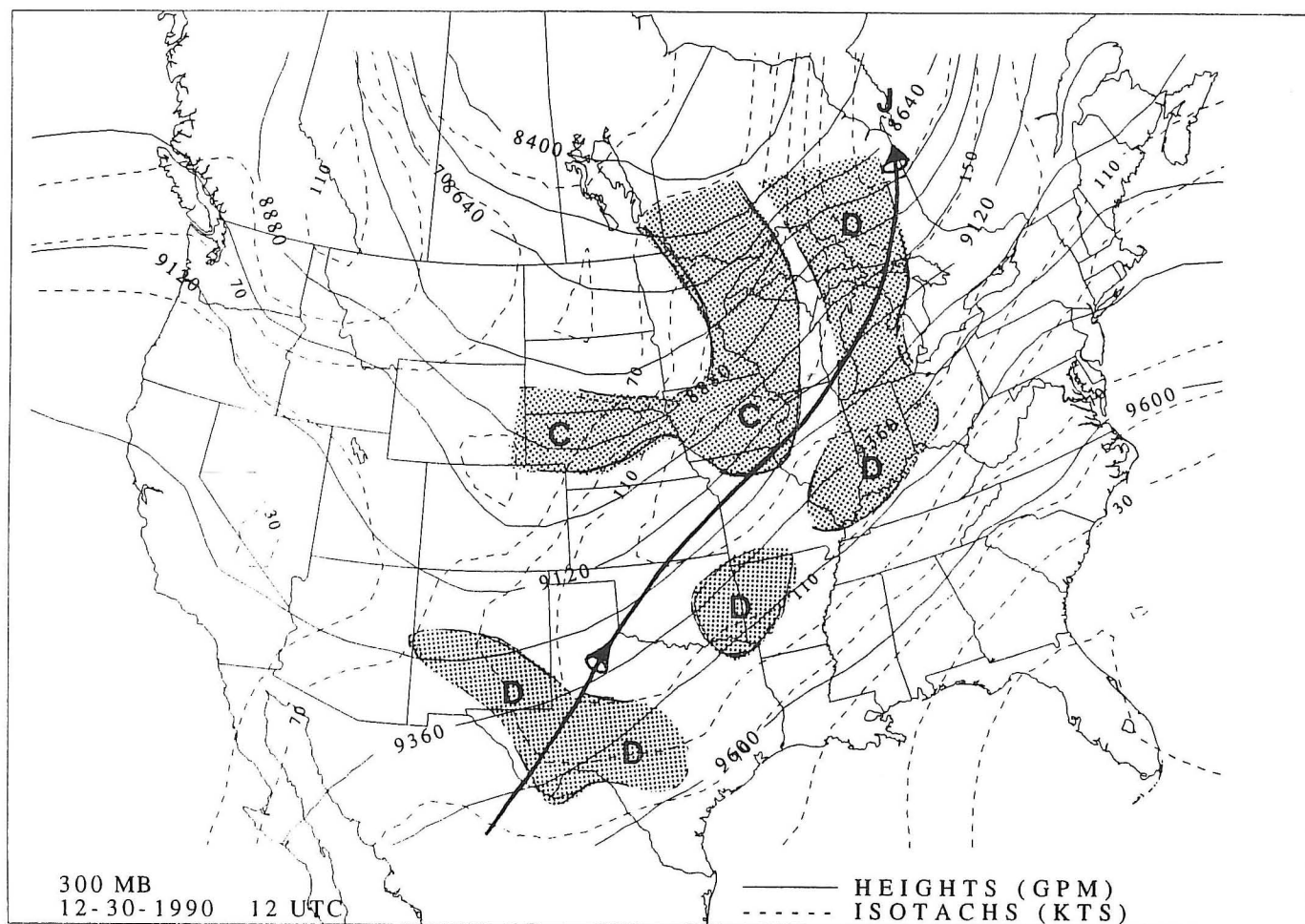


Fig. 10. 300-mb surface for 1200 UTC 30 December 1990. Heights (solid lines, gpm), isotachs (dashed lines, knots), thick solid line is jet axis, and divergence/convergence greater than $+1 \times 10^{-5} \text{ s}^{-1}$ or less than $-1 \times 10^{-5} \text{ s}^{-1}$ is shaded, D = divergence, C = convergence, stippled. Bold J is Jet core.

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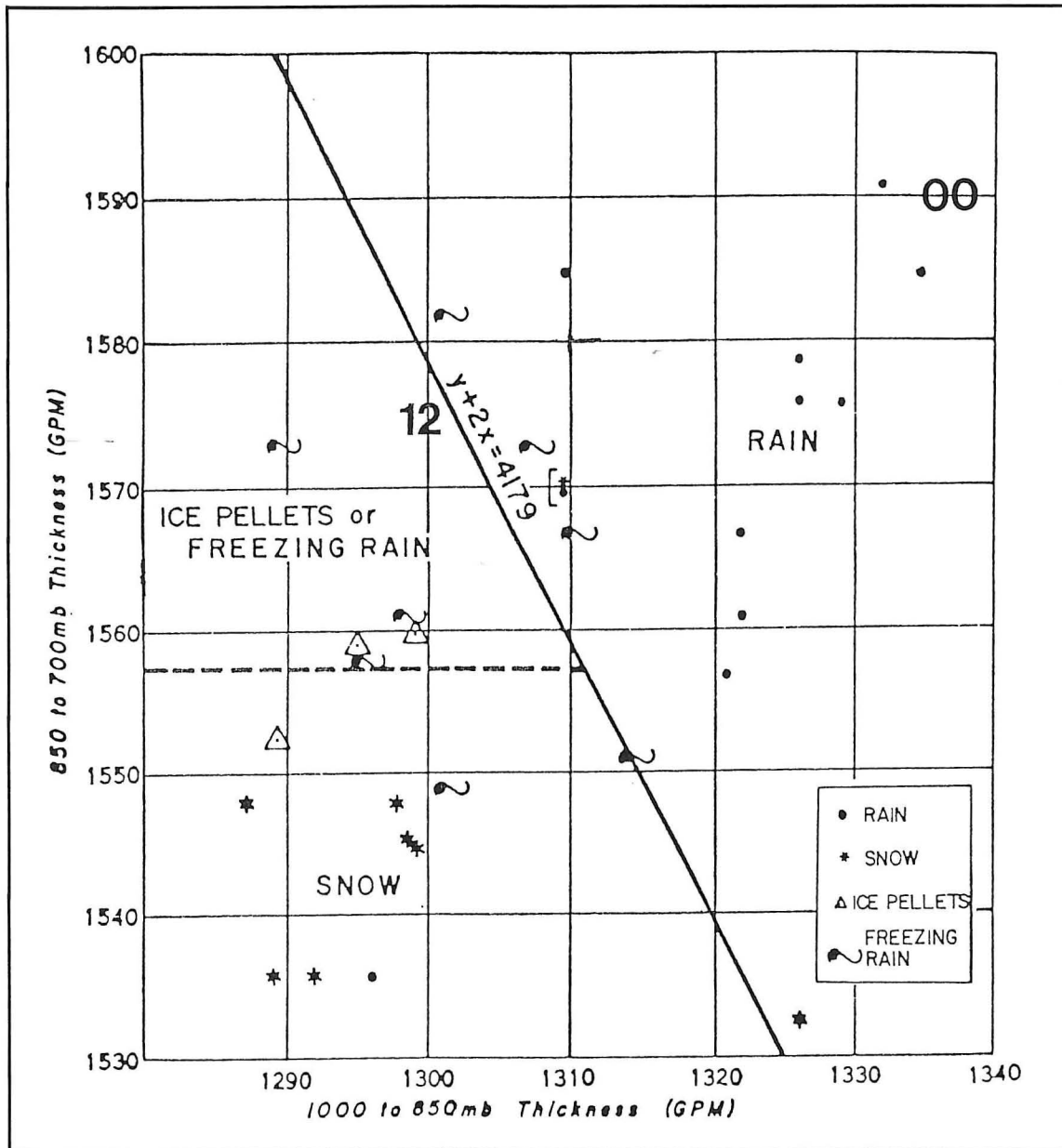


Fig. 11. Nomogram for forecasting precipitation type based on 1000-850 mb thickness and 850-700 mb thickness. (From USAF 1983 Extension Course Institute, Meteorological Technician Course (AFSC 25170), Volume 4.) Values at 0000 UTC 30 December 1990 labeled "00", values at 1200 UTC 30 December 1990 labeled "12".