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A PROGRAM OF CHART ANALYSIS
(with some diagnostic and forecast implications)

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INTRODUCTION

The following discusses in compact form a minimum program of chart analysis recommended for all Central Region offices making forecasts. It brings out the main features of interest to most forecast problems and it will serve well in the briefing of the next forecast shift. It is aimed at WSFOs, but is appropriate for WSOs for whatever charts are posted for use. It treats Facsimile and Request/Reply (R/R) material, but not exhaustively. The analyses are intended to be functional, not decorative, and thus should be made primarily for the area of interest, which would be mainly in the upstream or potential upstream direction and for a distance appropriate to the area and time periods of responsibility. Special missions, like Fire Weather, may need additional analyses, as might any office to meet local needs or interests.

For additional information on facsimile charts, see Forecasting Handbook No. 1 - Facsimile Products. Copies of Technical Attachments referred to here are attached in numerical order as Appendix A. Additional TA's not referred to but of prime forecast interest are also attached, numbered in subject groups, as Appendix B. A list of additional forecast-related Tech. Memos is given as Appendix C, ordered geographically. The material in the Appendices was inserted so as to have a compact collection of the most useful material Central Region Scientific Services has produced to help meteorological understanding through analysis and diagnosis.

We realize that this material will not be appropriate to an office with AFOS equipment, because of the lack of hard copies. However, it will be years before all are without facsimile. Also, the analyses discussed here will most likely be needed in the AFOS era, probably with additional ones, but they will have to be done in a different way. How they will be done will the subject of a later Tech. Memo.

A. SURFACE CHARTS

a. Facsimile Chart (with data)

1. Color precipitation areas green, fog areas yellow, and fronts in the usual colors, with a colored dashed line for squall lines.

2. Analyze 3-hr pressure tendencies (usually use a 1 mb interval starting at 1 or 2 mb depending on the season and strength of movement of systems). After consideration of diurnal pressure tendencies (see Tech. Paper No. 1, publ. 1943, for synoptic times, and Supplement to TP #1, publ. 1945, for intermediate times for these tendencies), especially note in the cold air any decrease in rises or start of falls, as this portends rainfall, wave action, cyclogenesis at surface and aloft. For more on use of pressure tendencies, see 31 rules in Petterssen Wx. Anal. and Fcstng. 1940 edition, starting on p.398.

3. Analyze dewpoints at 5°F interval starting at 55° (50° in CO and WY and adjacent high areas). This is primarily for heavy rain (flash flood) and is not needed for lower values regardless of season.

4. Transfer outline of PVA area(s) from 50kPa analysis.

5. Note changes in all analyzed features in past 3-, 6-, 12- and 24-hours. Plot past positions of the Low center and front of interest. Yellow is a good color for this.

6. Try to explain cause of precipitation using factors related to upward vertical motion, which are:

i. PVA aloft (50 kPa)
ii. Warm advection (see 85 and 70 kPa chart or 100-50 kPa thickness)
iii. Cyclonic curvature of isobars (frictional convergence)
iv. Terrain or cold frontal (also arc cloud) lifting
v. Lake effects (cold air over warm water and frictional convergence near lee shore).
vi. NVA at surface
vii. Pure instability (rare in Central Region)

Notice that overrunning was not mentioned as a vertical motion producer. This is an imprecise term which is more precisely stated as warm advection. It is thus treated, but indirectly.

b. Local Check Chart

A locally-plotted limited-area surface map is an advantage at times, especially on difficult days when many available reports are not on the facsimile chart. Information in the remarks section of the hourly observation can be plotted for added detail. Analysis should be the same as the facsimile chart, plus 2 mb isobars and significant smaller scale features, especially discontinuity lines (lake breeze, pressure jumps, arc cloud, squall lines). Satellite and radar data can aid in this analysis. Speciality analyses, such as IFR areas for aviation can be best on this chart.

c. Surface Geostrophic Wind and Vorticity (Sangster) Chart (R/R)

1. Plot a few key winds shafts and barbs.

2. Note wind maxima and the LLJ axis, especially over the Great Plains in summer at 18Z, 21Z, and 00Z.

3. Plot several past positions of the key vorticity maximum on latest chart.

4. Note past changes in wind and the value of vorticity max.

5. See Tech. Proc. Bull. 111 for use of chart. The movement of weak surface features is usually best seen in the surface vorticity.

If dewpoints are high with southerly winds, be cautious of heavy rain where axis of strongest wind intersects a warm front (or other boundary such as an arc cloud) even though front is very weak (e.g. can be seen only by 85 kPa temp. grad). Winds blowing from low to high vorticity (indicates NVA) are favorable for upward motion. See also Tech. Att. 74-8, 74-11, 74-12, 75-14 (attached).

B. UPPER AIR CHARTS (with data)

a. <u>85 kPa Chart</u>

1. Outline and shade green the areas with $T-T_d$ of 5^o or less, as relating to existing or potential clouds and precipitation.

2. Color the 0^oC line.

3. Indicate axis of 85 kPa jet when appropriate (mainly summer).

4. Outline areas of warm advection (possible upward motion).

5. Note changes in past 12 hours.

6. Plot track of Low center for past 24 hours. Heavy snow, if any, mostly occurs just north of track (see Mon. Wea. Rev. May 1970 or CR Tech. Memo 12).

b. <u>70 kPa Chart</u>

1. Outline and shade green the areas with $T-T_d$ of 5^o or less. In thunderstorms, low moisture is good for severe (SELS) weather; high moisture good for flash floods.

2. Note temperature advection areas (warm adv. same as at 85 kPa; cold adv. in warm season may reduce stability and aid convection.

3. Note and draw in minor trough lines that seem to relate to weather but are not seen at 50 kPa.

c. <u>50 kPa Chart</u>

1. Analyze 12-hour height changes at 30m intervals. Shade rises in red and falls in blue for values of 30m or more in absolute value. Note changes from 12- and 24-hours ago as suggesting changes in strength or movement of short waves.

2. Color a key contour line, seasonally selected.

3. Draw in trough- and ridge lines (cross contours at point of maximum curvature), and add positions from 12- and 24-hrs. ago.

4. Note areas of temperature advection. Cold advection behind a trough line and/or warm advection behind a ridge line indicates amplification.

5. Compare key features with LFM, and possibly Barotropic, 12-hr prog from 12-hrs earlier data time. If 12-hr prog good, remaining progs likely to be good also. If strongest winds are upstream of trough line it indicates deepening; if downstream, it indicates a shortwave trough coming out.

Winds at this level are good steering winds for thunderstorms, therefore, weak winds are "best" for flash floods. Strong winds give more weather in the cold season and they favor severe (SELS) storms in the warm season.

d. <u>30 kPa Chart</u>

1. Locate and color axis of jetstream(s). Axis tend to be a high level cloud boundary, with clouds on south side.

2. Note max wind speed (see 50 kPa). Strongly diffluent areas favor surface Low development, weather, and severe storms.

3. Draw in trough line(s), and note difference from 50 kPa. Since the main divergence aloft is much nearer 30 kPa than 50 kPa, the troughline and divergence due to PVA at 30 kPa is really the more significant. Usually troughs are vertical, but if not, mentally adjust divergence (upward motion) due to PVA area accordingly.

e. <u>20 kPa Chart</u>

While this chart is seldom analyzed, you may wish to note the strength and movement of the warm pools because they associate with the 50 kPa troughs. The warmer the temperature in these pools the stronger the trough action tends to be at 50 kPa.

1. Shade in red temperatures warmer than -50^{o}C and note warmest temp.

2. Note change in past 24 hours, and change from most recent trough passage.

C. RAOBS

1. Plot soundings of local interest, especially from 12Z data.

2. In warm season, compute K index (K = 85 temp. - 50 temp + 85 dewpoint - 70 dewpoint depression). If low-level moisture tops out just below 85 kPa, discount K index, as it will be unrepresentatively low.

3. If no frontal passage since yesterday, from 12Z raob compute max. temp. using low-level area change from 12Z to 24Z yesterday with appropriate change for differing cloud condition, if any. If winds at a bit above surface are strong today or yesterday, adjust accordingly (see T.A. 77-3). If fresh snow on ground, see T.A. 75-5.

4. In cold season, if sfc temp. below freezing, and precipitation possible, see if a shallow region a bit above sfc has temp. above freezing, if so, think freezing rain (see T.A.s 71-21 and 73-25).

5. Note layers of high relative humidity and thus of cloud potential. Note them especially if between mandatory layers, since the fax charts will not show them. Clouds at temperatures warmer than freezing will rarely have rain created in them.

D. PROGNOSTIC CHARTS

a. <u>Barotropic 4-panel</u> Treat all panels the same except as indicated.

1. Color purple a key contour line (the same as on 50 kPa chart with data).

2. Draw in trough lines, crossing contours at point of maximum curvature.

3. Plot 0-, 12-, and 24-hr positions of vorticity max, and possibly trough lines, on 36-hr panel, and note trends.

4. Outline and shade in red the PVA area(s). While this may go from trough line to ridge line, and thus be mostly delineated by these lines, the main (strong) area frequently is considerably smaller, and should be so analyzed. Shifting the outline of these PVA areas to the appropriate panels of the subjective surface progs frequently helps understand the prog weather.

5. Compare panels for the same valid time with LFM from data time 12 hours earlier, and when available, with LFM from same data time. A good comparison increases confidence in correctness of charts. A bias in the barotropic progs is a tendency to have the south end of a trough move too slowly. To correct, start at the vorticity max and extend the trough approximately perpendicular to the mean flow, especially if it was this way initially.

b. <u>LFM</u>

1. Treat 50 kPa charts as for Barotropic and note differences.

2. Outline and shade in green relative humidity of 70% and more, and forecast precip. areas. These are highly correlated fields.

3. Color a 100-50 kPa line as selected seasonally (540 in cold season?).

4. Color red the plus 2 vertical velocity line and possibly shade VV > 2.

E. MISCELLANEOUS CHARTS

a. 4-Panel Composite Moisture Chart

Contains Lifted and K indices, precipitable water, mean relative humidity, freezing level. All relate to clouds and precipitation.

1. Low LI best for severe storms; shade less than zero, green.

2. High K best for general thunderstorms; color 30 line, red.

3. Shade green the precip. water over one inch.

4. Shade RH of 70% and more, green.

b. Subjective Surface Prog - PoP - QPF

1. Check these charts for consistency. Maker of sfc prog usually does not see the latest PoP chart, so they may be inconsistent. Judgment, not rules, resolves differences.

2. Fronts and weather on surface prog can be colored as on fax chart with data.

3. The 50% line of PoFP (prob. of frozen precip.) is the best snow-rain line we have and it is good. Color it red. Heavy snow, if any, is generally a bit north of this line.

F. REQUEST/REPLY (R/R)

a. PoP and PoF etc. - FOUS 12 (some on facsimile)

See Tech. Proc. Bull. 217. PoP1 is based on LFM and obs. PoP2 is on LFM and other models and should be better, but first 12 hours of PoP2 should be the same as PoP1 unless backup equations used for PoP2. Note trend of PoP and PoFP. Note 6-hr vs. concurrent 12-hr PoPs. 6-hr values should not be greater than 12-hr. Resolve conflict if so.

b. Max/Min Temps. - FOUS 22 (some on facsimile)

See Tech. Proc. Bull. 220. Has early and final guidance as for PoP, and again final should be better. Caution - These are based on calendar day max/min temps., which is OK for a day with normal diurnal temp. cycle. When fronts pass they may be invalid. If snow is on the ground, see TA 74-1.

c. Detailed Guidance Output - FOUS 60-77

See Tech. Proc. Bull. 173. This gives detailed information on many parameters from the numerical models in 6-hour increments. It is excellent for seeing trends and establishing timing of events.

A discussion of the anlysis of satellite and radar data was intentionally omitted because of their special nature. However, forecasters must free enough time to fully evaluate such data or the short period forecasts will unquestionably suffer, occasionally seriously. Information on the analysis and use of satellite data is contained in the "Handbook for Meteorological Use of Satellite Pictures", or in "Application of Meteorological Use of Satellite Pictures", Tech. Report NESC 51, and in the many GOES Tech. Attachments produced by the Central Region SSD. Information on the analysis and use of the radar data (for other than a radar operator) is in the NOAA publication, "Introduction to Weather Radar."

Much of the future for the human forecaster will probably lie in the use of satellite and radar data, as these observations, continuous in space, contain information not available elsewhere, and especially satellite obs. are not yet in numerical forecasts. They contain detail that must first be noticed, and then understood and blended in with other analyses. Details for short term forecasts will depend heavily on the analysis of satellite and radar data, and the short term forecast is the prime area for the human forecaster in the future.

APPENDIX A

REFERENCED TECHNICAL ATTACHMENTS

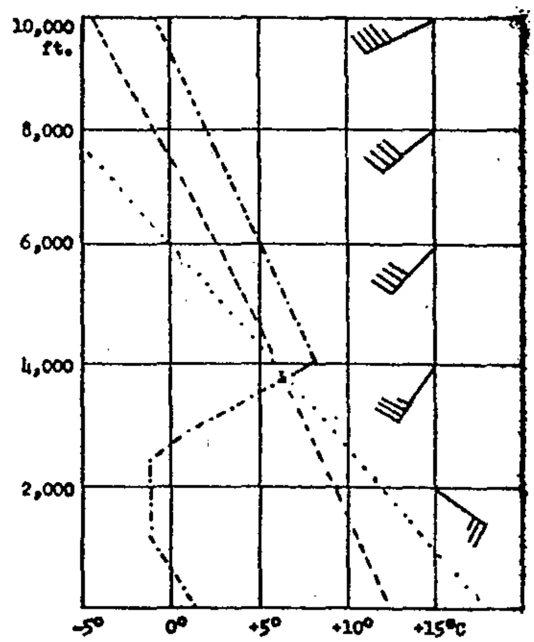
71-21	Forecasting Freezing Rain
73-25	Forecasting Ice Storms
74- 1	Bias of NMC Objective Surface Temperature Forecasts
74- 8	Nighttime Showers and the Sangster Chart
74-11	Check Sheet for Use of the Surface Geostrophic Wind (Sangster) Chart
74-12	The Flood Producing Rains in Missouri and Illinois of Mid-May 1974
75- 5	Maximum Temperature Over Snow
75-14	Meteorological Aspects of the Dakotas-Minnesota Floods of Early July 1975
77- 3	Raob Analysis - Mixing and Heating

Technical Attachment 71-21

Forecasting Freezing Rain

Freezing rain can be a very destructive event, but fortunately it is a rare event. It is rare mainly because it takes a set of circumstances within narrow limits to bring about freezing rain instead of rain, snow, sleet, or snow pellets.

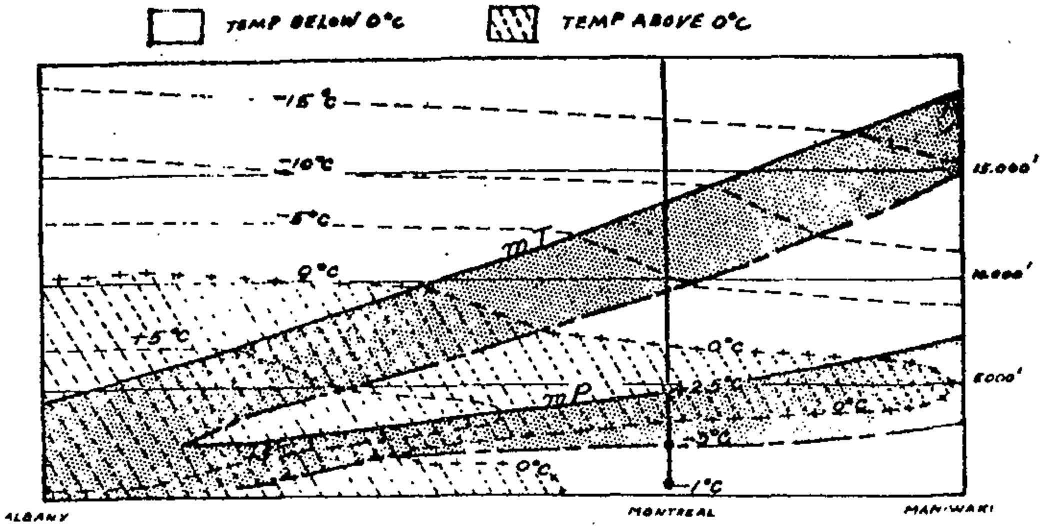
There is little question but that the precipitation involved when the snow-rain-freezing rain question arises is formed high enough in the atmosphere to start out as snow. Obviously, if it is to be rain, it must pass through a portion of the air that is warmer than freezing. Then if it is to freeze again, there must be a subfreezing layer below the above-freezing layer. This condition is depicted by Williams (1) as shown in Figure 1, and similarly by Mahaffy (2) and by Harlin (3).



Wind and temperature in lower levels for Huntsville, Ala., 1300 CST, 2 March 1960. The dash-dotted curve denotes the temperature lapse rate; the dashed curve, the pseudo adiabat; the dotted curve, the dry adiabat.

Figure 1

The condition depicted fits the usual warm frontal situation, as seen by Figure 2 (from Mahaffy). However, the lower subfreezing layer under an above-freezing layer restricts the number of occasions considerably, but this restriction is not enough, for, in addition, precipitation has to ba falling, and the thickness of the two lower layers has to be within reasonable limits. If the warm layer is too thin, complete melting of the snow will not occur,



Cross-section of the lower atmosphere from Albany, N. Y. to Maniwaki, P. Q., at 1900, 25 February 1961.

Figure 2

and snow pellets may form. If the cold layer is too shallow or too little below freezing, the rain may not be able to freeze again before hitting the ground, while if it is too thick, sleet will occur. The proper thickness of these layers would, of course, depend on the extent the temperature reaches above or below freezing in the appropriate layer. From the references given it appears that, for freezing rain, a warm layer about 5000 feet thick is proper, while a lower cold layer about half that value is reasonable.

There are several refinements to keep in mind to further help in the decision. It is a common winter condition that as a warm front approaches, the precipitation at the ground is initially snow, but turns to rain or freezing rain as the frontal inversion lowers in height as the front gets closer. One would generally think that once the change to rain had occurred that the threat of snow was over. This is not necessarily true. If the warm layer is not close to saturation and is only marginal in depth or the inversion nose is blunt so the temperature is just slightly above freezing, the heat given up by the air in melting the first snow falling through it may cool the layer to below the freezing point so the rain or freezing rain turns back to snow again. This is especially likely to happen if the warm frontal advance is associated with a small wave passing through and the wave crest is just passing, so the warm frontal advance upon the station has just stopped.

Another refinement is due to the heat island of the larger cities. At critical times this added heat is sufficient to have an advancing snow situation turn to rain over and/or downwind of the city, or to have freezing rain in the suburbs and just rain in and downwind of the city.

Also to be considered is the recent temperature condition. If there had been quite cold conditions which have quickly warmed quite a bit as the possible freezing-rain condition moves in, the coldness of the ground, trees, and telephone wires, and other objects, means that an icing condition can be more severe than had the ground and objects been at a temperature much closer to freezing.

A final point involves a small area with at least a few hundred feet difference in elevation within it. The sounding by Williams shown above (Figure 1) was for just such a situation, as it was in the Appalachian Mountain area. The sounding resulted in only rain at the location of the sounding, but, as noted by Williams, "light rain and sleet began at Huntsville Airport at 2216/1st. The sleet ended in fifteen minutes, but the rain continued throughout the following day with three and one-half inches accumulating."

"During the day on March 2nd, one could look at the surrounding hills and see a clear demarcation where ice was accumulating on higher elevations." Evidently, the warmer layer nearest the ground saved the city from an ice storm that at 1000 feet higher was serious.

Forecasting temperature conditions horizontally and vertically with the detail and accuracy necessary to make such sounding forecasts is probably not possible at this time. Instead, the problem takes on the character of "now-casting" in that one should examine the nearby soundings as soon as they are available, and if precipitation is expected to occur soon and the above diagrammed conditions exist in the vertical initially, a short-term weather forecast becomes reasonable on the assumption the sounding changes little. The sounding taken in support of air pollution work could be of considerable benefit for such a weather forecast.

References

(1) Williams, B. B., 1960: The 1960 Ice Storm In Northern Alabama. Weatherwise V. 13, No. 4, 196-199.

(2) Mahaffy, F. J., 1961: The Ice Storm of 25-26 February 1969 at Montreal. Weatherwise, V. 14, No. 6, 241-244.

(3) Harlin, B. W., 1952: The Great Southern Glaze Storm of 1951. Weatherwise, V. 5, No. 1, 10-13.

TECHNICAL ATTACHMENT 73-25

FORECASTING ICE STORMS

Have you ever forecast an ice storm? Put out an Ice Storm Watch or Warning? Probably not, as these are rare, but also forecasters seem reluctant to do this and usually go for the less desirable condition of a Traveler's Advisory. This is probably because they are uncertain of how to forecast them and uncertain as to just how bad a condition is actually occurring. This note will be concerned with how to forecast ice storms, applying the information of Tech. Attachment 71-21 to two recent bad ice storms, that of December 3-4 and November 20-21.

If freezing rain is a threat, and it is in almost every storm from November through February, and especially those with the 500 mb vorticity max coming out of the SW U.S., as the moisture is more abundant then, the first thing to do is to LOOK FOR FAVORABLE RAOB SOUNDINGS. Of course, if freezing rain is already occurring, this fact can be used too, but as will be seen later, an existing freezing rain situation can change quickly, so soundings must be used. Such a sounding is shown in T. A. 71-21. It must have a nose above freezing (usually around 5,000 ft. above ground) and a lowest layer below freezing. Figure 1 and 2 shows such soundings for 00Z December 4, 1973 for DDC and OMA.

Next, one must LOOK AT THE TWO LAYERS OF CONCERN, as discussed in T. A. 71-21*. If the warm nose is too small, complete melting may not occur and snow pellets (graupel) will occur. If the nose has too low a humidity, the evaporation of the melting snow and the heat required for melting (the former effect is 7½ times the latter for a given amount of water) may cool the layer to below freezing and cause freezing rain to change to snow. In Figure 1 the warm nose at DDC is small, but precipitation was not occurring at the time. One hour later the precipitation was freezing rain. But note that the nose is saturated, so it can probably continue to exist even though small. At OMA, Figure 2, the bottom cold layer is probably too large for freezing rain, and sleet would occur but there was no precipitation to confirm this. Both these points would suggest that the maximum ZR probability line, considering the temperature structure only, would be a bit east of both locations.

The next point is: CONSIDER HOW THESE SOUNDINGS AND THE FAVORABLE ZR AREA MIGHT CHANGE WITH TIME. The 850 mb chart (or 700 mb in the western Plains), is usually adequate for considering changes to the warm nose. In this case the flow was from the NNE, which was almost along the line between the two soundings in Figure 1 and 2, but slightly more northerly, especially at DDC, so with time some cooling could occur and the small warm nose at DDC would disappear over DDC. The small night cooling that was likely even under cloudy skies could also cool the lowest layer so that it might be too large for ZR, and for both these reasons, sleet or snow would be more likely a bit after observation time at DDC. Actually DDC had snow pellets mixed with ZR within a few hours and all snow at 0715Z.

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*Please correct T.A. 71-21 in this regard so that the first complete sentence on page 2 reads, "If the <u>cold</u> layer is too shallow or too little below freezing, the heat in the ground may prevent freezing rain, while if the layer is too thick or too cold, sleet will occur.

-2
T.A. 73-25 Cont.

Since the 850 mb winds showed little advection in the warm nose layer, and the 700 mb wind was light, the area east of DDC-OMA line favorable for ZR would be QUASI-STATIONARY and conducive to a sizable deposition of freezing rain. The most ice would, of course, occur in the ZR area where the AMOUNT OF PRECIPITATION IS LARGEST. To forecast this detail is difficult, but observing it by means of regular observations, or plotting manual d/radex data can be tried. Also, one should call out to spotters and/or other contacts to really know what is occurring.

WIND IS ALL IMPORTANT TO THE DAMAGE resulting from freezing rain, as tree limbs and power lines are very sensitive to wind when they are ice laden, and down limbs and lines add to traffic hazards. The one thing you should know for sure from observations you have is that if freezing rain is occurring with thunderstorms you have a serious problem, as the amount of precipitation is almost always great enough to create a major problem (incidentally, this is also the case when concerned with heavy snow instead of freezing rain). Therefore, WATCH FOR THUNDERSTORMS and the larger d/radex numbers.

Figure 3, from OMA at 12Z December 4, 1973 (12 hours after Figure 2) is an almost ideal freezing rain sounding. Incidentally, these soundings (except for Figure 5) are reduced copies of the computer plotted soundings at NSSFC, where both temperature and dewpoint curves are plotted except when there is saturation and then the overcast symbol is used to show the two curves are coincident. Freezing rain and ice pellets (sleet) was reported at OMA at sounding time and for several hours later, while one hour earlier there was only freezing rain. This means that the below-freezing lowest layer is close to the maximum size for freezing rain, because if it was larger, only sleet (ice pellets) would occur. If the air had not been saturated in this layer, it probably would have been too large for ZR as evaporation would have caused more cooling of the air and then of successive drops and only sleet would occur.

At the time of Figure 3, DDC was too cold for ZR, and TOP had cooled off to a good ZR sounding, but the surface temperature at TOP was right at freezing, so it would be the eastern edge of the ZR area. The 850 mb winds were from the northeast, so the likely ZR area would go from NE Kansas northeastward into Iowa. OMA didn't get much freezing rain, as the favorable temperature condition ended shortly after the precipitation started, but the persistence of this favorable condition being continuously advected into Kansas, where quite a bit of precipitation occurred, produced the worst ice storm in decades. Not only were the power lines down, but thousands of poles snapped off as well. While not all the facts are in as yet, we understand that besides the Kansas damage, at least powerline damage went across much of Iowa from SW to NE and touched the NW portion of Missouri and the SW portion of Nebraska. Quite a storm.

The November 20-21 storm was not as extensive and not quite as damaging, but it was very significant in eastern Nebraska, where power was out for days in some areas. The LBF sounding for 00Z Nov. 20 illustrates a point (see Figure 4). With a warm nose that small, it would require warm advection at that level to make it significant, however LBF had weak cold advection and the nose most likely disappeared soon. Also, note the low humidity in this warm nose. This is the type of situation in which the cooling effect of melting snow and evaporation would quickly eliminate that portion of the warm nose that is above freezing.

LBF did not have precipitation at 00Z but had liquid precipitation at 01Z and snow at 02Z and beyond, confirming the above.

The HON sounding at the same time as LBF above (00Z), was very similar to that of LBF (see Figure 5), but conditions changing the sounding were obviously different and the effect totally different. The small warm nose, like that at LBF, seems hardly large enough to last. But note that it is saturated, so evaporation cannot cool it. Also, at 850 and especially at 700 mb, there was warm advection, so the nose could enlarge with time. In the lowest layer it was initially above freezing, and rain was occurring, but with the cooling at night and some cold advection near the ground, the temperature quickly dropped below freezing, and the rain turned to freezing rain which lasted for several hours but with only a small amount of water equivalent. Since there hadn't been any appreciable time in the past few days with surface temperatures below freezing, the heat of the ground must have helped reduce the deposition of ice.

SUMMARY - Steps to an Ice Storm Warning

1. Based on evidence from the surface chart, select and plot key sounding(s). Usually 1 or 2 and rarely as many as 4 are needed for the area of a single state.

2. Note the presence and size of a warm nose at/about 5000 feet above ground, surface air temperature, and the size of the lowest sub-freezing layer.
   a. If the above freezing portion of the warm nose is small, it cannot have low humidity or cold advection if it is to survive. Use the 850 mb and/or 700 mb chart, or the turning of the winds with height to determine the advection in the layer with the warm nose. If the warm nose cools to below freezing, the rain will turn into snow pellets, then snow.

   b. The freezing rain will turn to sleet when the bottom layer gets too large or too cold. This/is brought about mainly by cold advection, but if the layer has low humidity and is fairly large, it can get too large by evaporational cooling. Diurnal temperature change even though small under the cloudy skies is also a factor.

3. Consider how the favorable ZR area will change position with time due to the movement of the pressure system and/or advection (these are not independent).

4. If freezing rain is occurring, it will give a large deposition and/or become damaging if:
   a. There is a lot of rain - watch for thunderstorms(plot d/radex) and note whether or not the ZR area should move much.

   b. There is considerable wind with the ice deposition.

   c. Wet snow comes later and adheres to the ice.

If the ground and other surface objects are well below freezing, ZR can occur even if the air temperature is at or just above freezing. Conversely, a marginal but significant freezing rain condition can occur when the ground is warm and the air temperature falls from above freezing to 5° or so below freezing. In this case above-ground objects, especially metal bridges, can cool to below the freezing point and ice up, while the ground heat keeps the ground ice free.

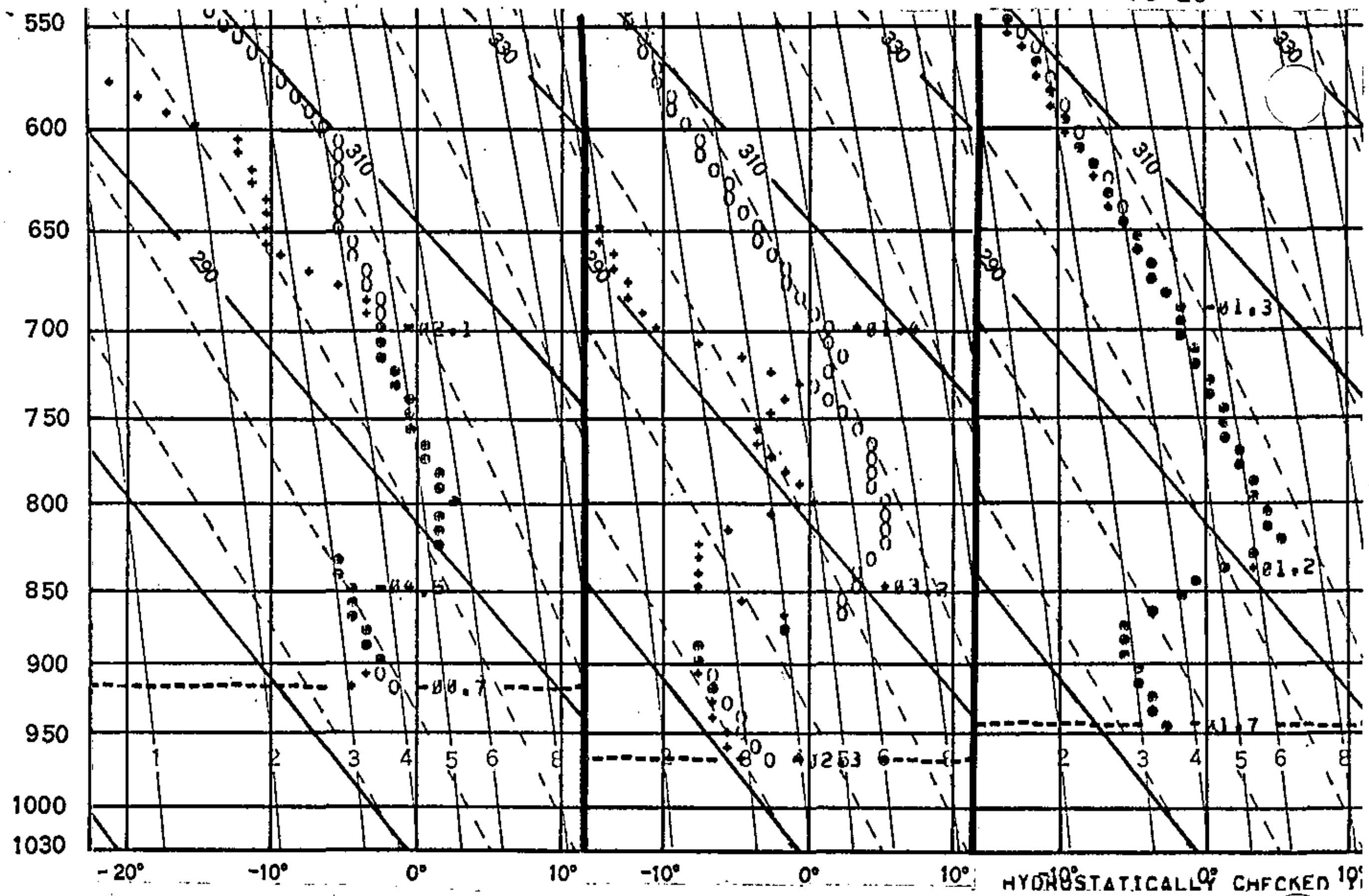


Fig. 1 DDC 00Z Dec. 4

Fig. 2 OMA 00Z Dec. 4

Fig. 3 OMA 12Z Dec

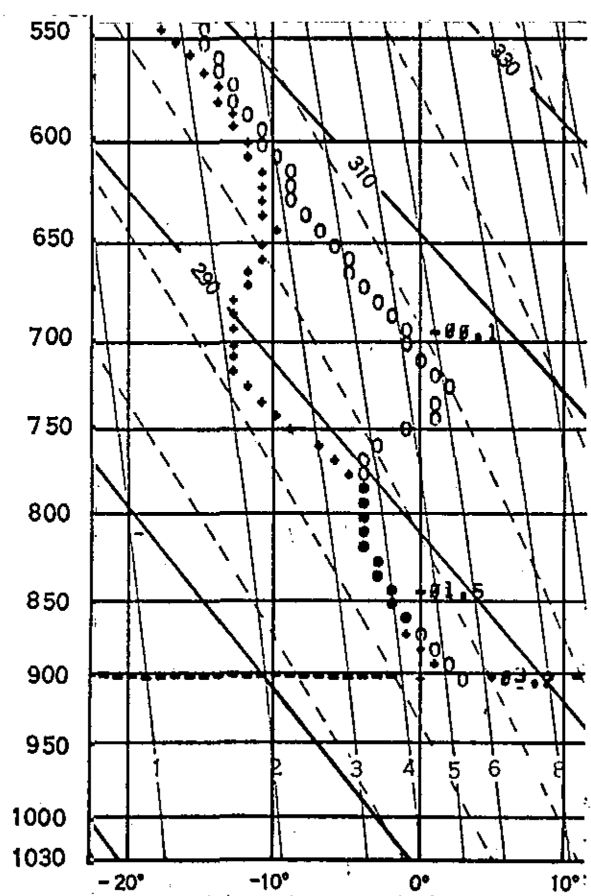


Fig. 4 LBF 00Z Nov. 20

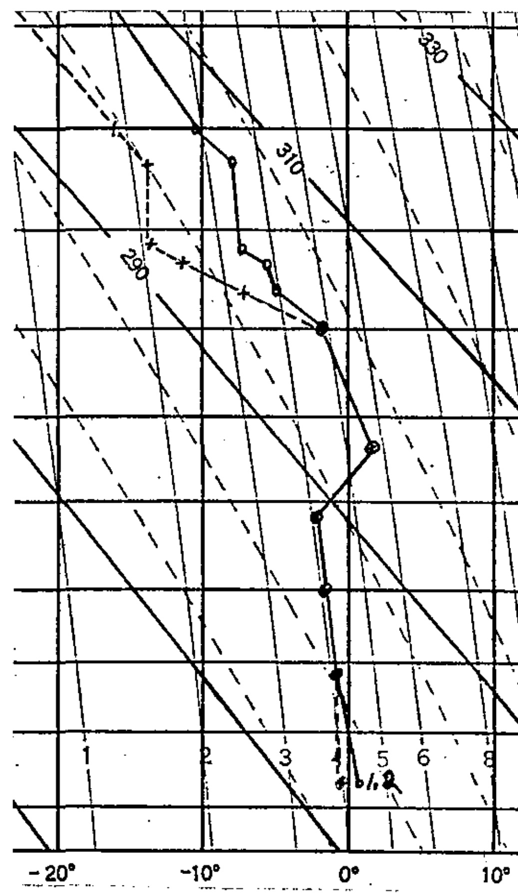


Fig. 5 HON 00Z Nov. 20

BIAS OF NMC OBJECTIVE SURFACE TEMPERATURE FORECASTS

John T. Curran, Jr., WSFO Omaha

The new NMC Objective Surface Temperature Forecasts have been generally accepted by the forecasters at WSFO Omaha as good. However with the onset of the snow cover over Nebraska, a large bias developed in the NMC maximum temperature forecasts. The effect of this bias is that maximum NMC temperature forecasts usually are too warm when snow cover is present.

The data used for this study included 24, 36, and 48 hour maximum temperature forecasts, when available, for OMA, GRI, LBF, BFF, and SUX from December 4, 1973 through January 15, 1974 when 1" or more of snow cover was present at the forecast point. A total of 353 forecast temperatures were used in the study.

The maximum for the next day (36 hour forecast) was used from 12Z data. The maximum today (24 hour forecast) and maximum tomorrow (48 hour forecast) was used from 00Z data. The forecasts were obtained from WMSC through the dedicated circuit.

Figure 1 is a scatter diagram of the NMC forecast temperatures vs. the observed temperatures. The diagonal line represents the line of perfect forecasts.

It can be seen from the scatter diagram (Fig 1) that a wide range of temperatures existed during the study with maxima occurring from below zero to above 60. Increasing deviation from the line of perfect forecasts becomes apparent with higher NMC forecast temperatures. Thus, one could conclude that the warmer the NMC maximum temperature forecast with snow cover present, the greater the amount of overforecasting.

Each point in the scatter diagram of Figure 2 represents the mean deviation of all the NMC forecasts for a particular temperature from the line of perfect forecasts. A curve was drawn to fit the scatter of points and represents the correction a forecaster should apply to the NMC forecasts to remove the bias. Again, this is for maximum temperatures only when snow cover is present.

The curve of Figure 2 indicates that the NMC maximum temperatures are forecast too high at all temperature ranges down to about 5 degrees above zero. The average error for forecasts from the mid teens through the 20s is about 7 degrees. The error steadily increases for forecasts above 32 degrees to the point where the average error is about 20 degrees when a high of 70 degrees is forecast.

In areas where extensive snow cover is the rule, rather than the exception as it is in Nebraska, the large NMC error is probably not present as this normal condition would be reflected indirectly through the developmental data.

Added by CRH SSD

Fred Ostby, Jr. of NSSFC also sent in some information on this subject. His short note dealt with three days in mid-January which had a sizable snow cover over Kansas City and none over Rapid City. His figures are as follows:

	RAPID CITY			KANSAS CITY		
	Fcst	Obs	Error	Fcst	Obs	Error
Jan 15	58	63	-5	71	42	+29
Jan 16	63	69	-6	70	43	+27
Jan 17	54	62	-8	64	43	+20

While we haven't looked at the bias for Rapid City for a larger sample, these figures and conditions, along with those of Curran, strongly suggest that the large overforecasting bias over the snow was due to presence of the snow. Ostby's maps of the bias on the three days given above suggest that in general the bias was small or negative in the Dakotas and at least the northern parts of Minnesota and Wisconsin, as well as the upper Michigan peninsula, allof which also had good snow cover. But this is the normal thing there and is thus allowed for in the MOS statistics. The large positive bias ran from Colorado ENE to at least lower Michigan and northern Indiana, while the snow free Kentucky had low bias values.

Technical Procedures Bulletin 2 discusses heat sources and sinks in the PE model. While some details have changed, the basics are unchanged. In essence, it says that there is extra cooling over snow when the sun angle is low (less than 10^{o}). However, their way of determining snow is a weekly look at satellite photos, so they tend to adjust for the permanent and semi-permanent snow cover, rather than the more transient cover. TPB 94 on the max-min PEATMOS temperature forecasts indicates that the forecaster must allow for the effects of a snow cover. While on the point it is worthwhile to refresh on all the points the forecaster must consider when using the MOS temperature forecast. TPB 94 says, ". . the MOS equations... do not account for mesoscale features such as sharp frontal zones, squall lines, and low stratus clouds." Also, ".... specific localized conditions will have to be considered subjectively by the forecaster.... These factors include land and sea breezes, mountain and valley winds, sea surface temperatures, snow cover, soil moisture, fog, thunderstorms, smoke and dust, and urban-suburban differences. When these conditions are important or abnormal, adjustments should be made...". Station moves must also be adjusted for.

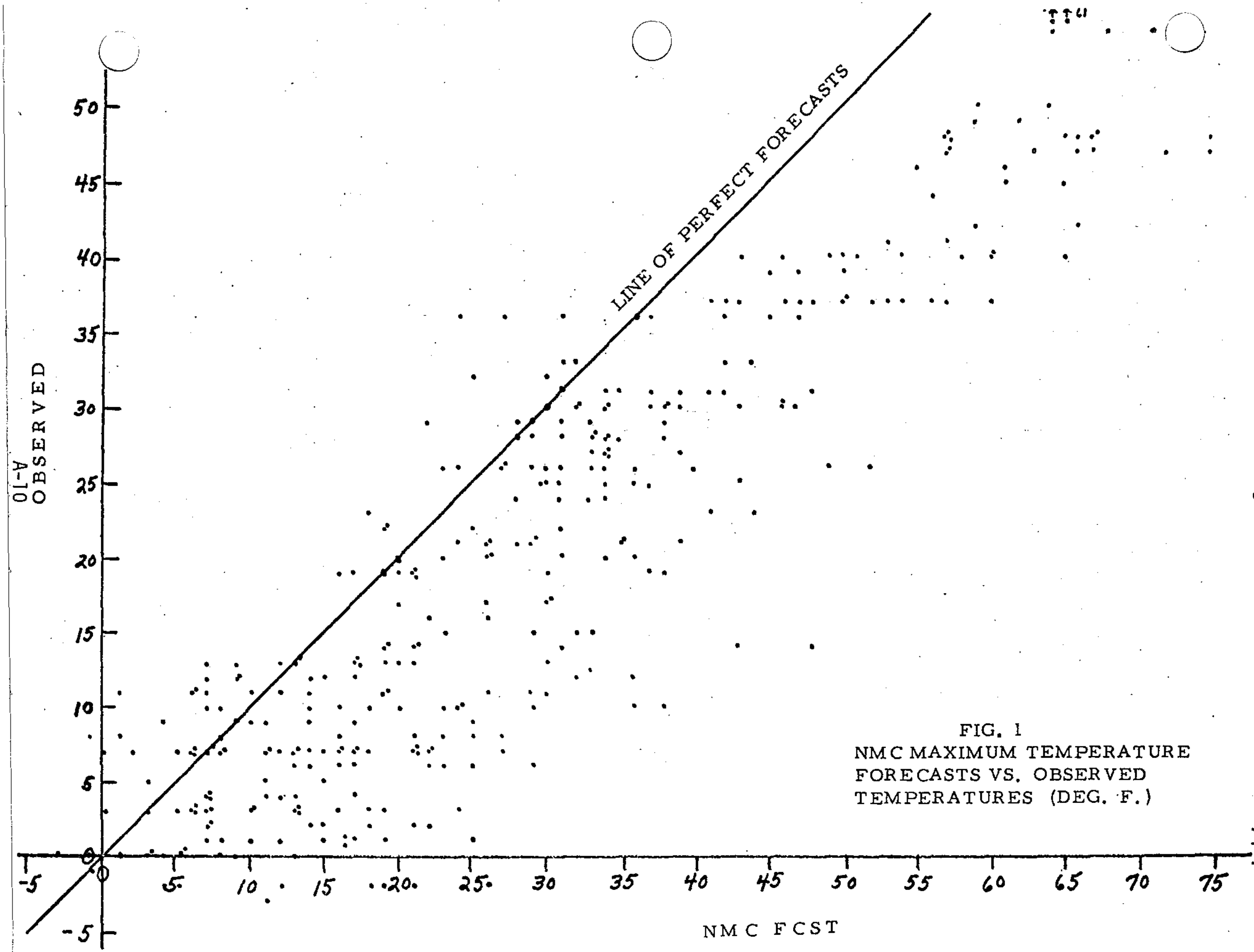
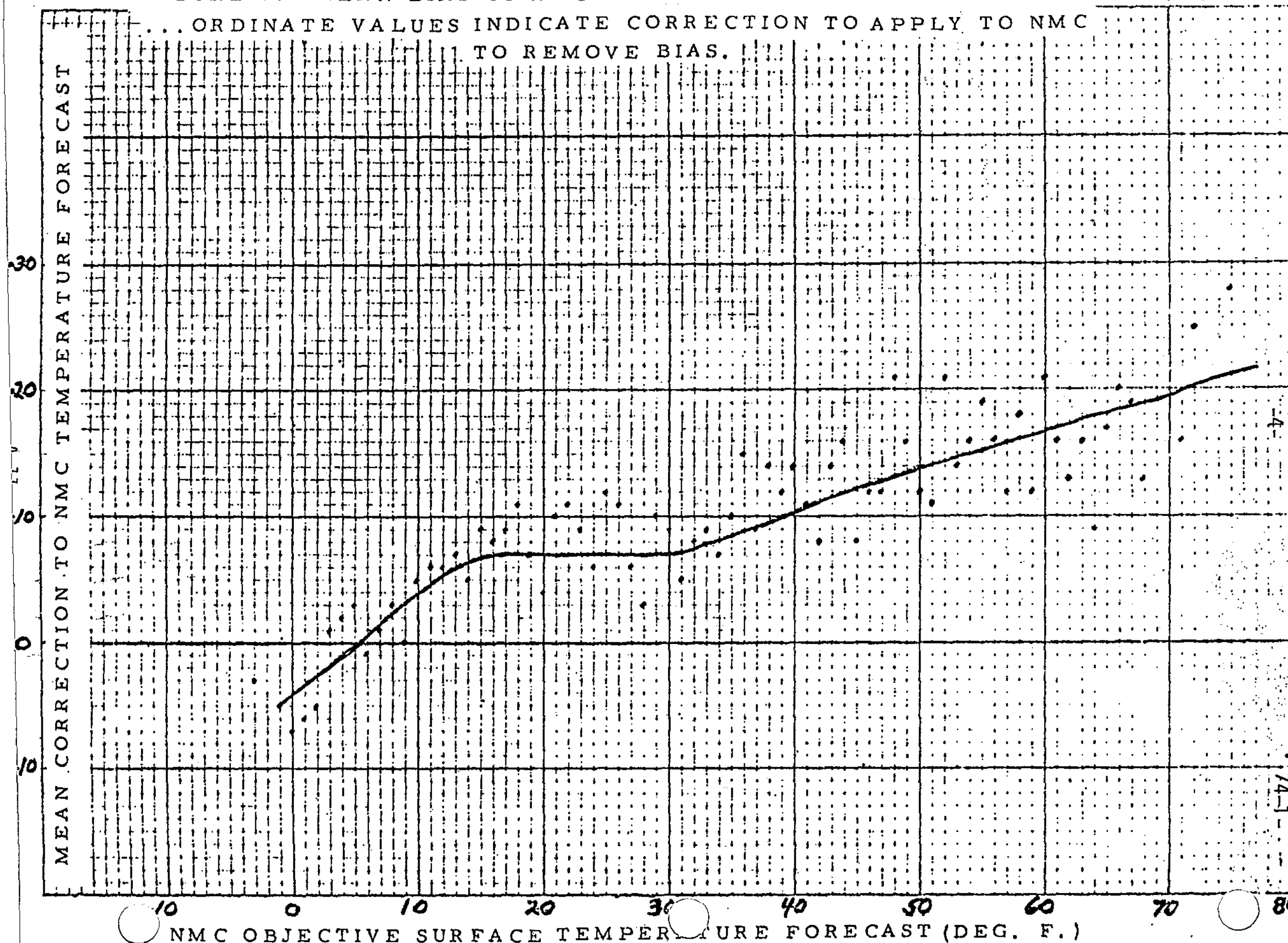


FIG. 1
NMC MAXIMUM TEMPERATURE FORECASTS VS. OBSERVED TEMPERATURES (DEG. F.)

FIGURE 2. MEAN BIAS OF NMC OBJECTIVE TEMPERATURE FORECASTS ... ORDINATE VALUES INDICATE CORRECTION TO APPLY TO NMC TO REMOVE BIAS.



74-1

TECHNICAL ATTACHMENT 74-8

NIGHTTIME SHOWERS AND THE SANGSTER CHART

The Sangster surface chart recently became available on request/reply for WSFOs. This T.A. is the first of several with which we hope to indicate ways of using this chart for short term forecasts. In my (Hughes) long forecasting experience, this chart is one of the most helpful for noting, understanding, and forecasting the smaller scale events. We have somewhat indicated this in sections 4 and 5 of Tech. Procedures Bulletin 111 dealing with this chart, but in these T.A.'s we plan to go beyond where the space limitations of the TPB forced us to leave off.

This first T.A. deals with a rain situation which appeared to be associated with a short wave trough, but which underwent nocturnal amplification and exhibited a pattern at the end of the night that was much more in line with the low-level jet hypothesis than with the expected pattern with a short wave trough.

The synoptic situation at 500 mb is shown in Figure 1a-1c, and is simply that of a common weak shortwave trough moving across the central Plains in a 24-hour period. The sea-level facsimile charts of Figures 2a-b show a slowly moving inverted trough in the west central Plains, with rather low moisture flowing northward across the central Plains (surface dewpoints in the upper 30s).

The Sangster surface chart for the latest time available for the afternoon forecast for "tonight" is shown in Figure 3, giving only the geostrophic winds as they would be on the teletype message. These winds are analyzed in 5 knot intervals, as may be profitable to do in the early stages of the use of the chart, as was suggested in the TPB. Let us look at how this information should be used, working only with the Sangster surface chart for 18Z on the 17th, the 12Z 500 mb chart on the 17th and the 500 mb progs for the next 24 hours. We will use the perfect progs of real data, to save space, as the actual PE progs were excellent for this time.

According to the TPB we should a) "Locate the maximum wind speed and determine its magnitude", the higher the better, with 40 knots uncommonly high, b) "From the wind speed maximum, go downstream (use wind directions for this) and note if an area of significant decrease in wind speed is found (still following the streamline through the wind speed maximum). The greater the decrease (per unit distance) the greater the probability of showers. The area most likely for showers is to the right (looking downwind) of the area of maximum decrease in wind speed."

In this case the wind speed was very high (48 knots at a grid point and perhaps slightly higher between grid points). The line one would follow, according to "b" above, is given in Figure 3, and it shows that practically all of the decrease of wind along the direction of the wind was in Kansas, and to the right of the line from the maximum wind would be in eastern Kansas.

In a quiet large-scale system as is typical of most of summer, eastern Kansas would be the prime area of concern for nighttime rainfall (possibly Western Missouri for the early "today" period), and it probably would have been heavy if the moisture content of the air had been up to summer averages, as the max wind was well above 40 knots and with a good wind speed decrease downstream. But this time the shortwave trough would be a factor and would tend to move the favorable area slightly south of due east as the trough advanced, so the threat area for showers tonight would be eastern Kansas, most of Missouri, except possibly extreme north, extreme northern Arkansas and northeast Oklahoma.

The RADU chart representing the latest radar data that would be available to the forecaster before finalizing his late afternoon forecast is shown in Figure 4a, along with the interpolated position of the 500 mb trough line. This precipitation broke out in western Kansas about 12Z, at the time the positive vorticity advection (PVA) area ahead of the upper trough was getting into the area (the precipitation did not start farther west perhaps because of the downslope effect of the upper level flow across the mountains).

Thus the precipitation seems to have been initiated by the upper trough, and had been advancing across Kansas with the upper trough. Its location in Kansas instead of north or south of there may well be related to the low-level jet as discussed above. However, a forecaster not familiar with the hypothesis indicated in the TPB and above, could easily take the precipitation at the speed and direction of the 500 mb trough, which would mean, according to Figure 1c, that it would have passed almost out of Missouri by 12Z April 18.

The RADU chart at 1235Z (Figure 4b) shows that this that this was clearly far from the case, in that the precipitation had not even cleared out of Kansas by that time. The holding back of the precipitation to a position well behind the upper trough was most likely due to the low-level jet model as discussed above. The point then is that the low-level jet effect can be strong enough, even when working against a weak upper-level trough, to be the major factor in the precipitation during the night and early hours after.

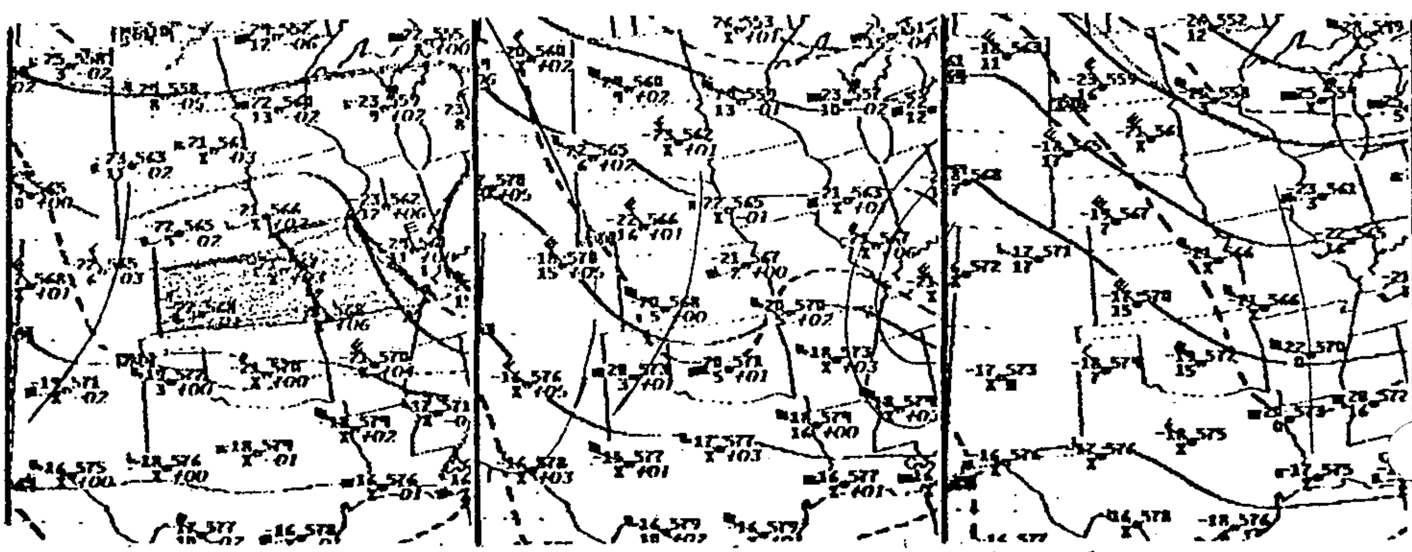


Figure 1a. 500 mb 12Z April 17, 1947. Figure 1b. 500 mb 00Z April 18, 1974 Figure 1c. 500 mb 12Z April 18, 1974

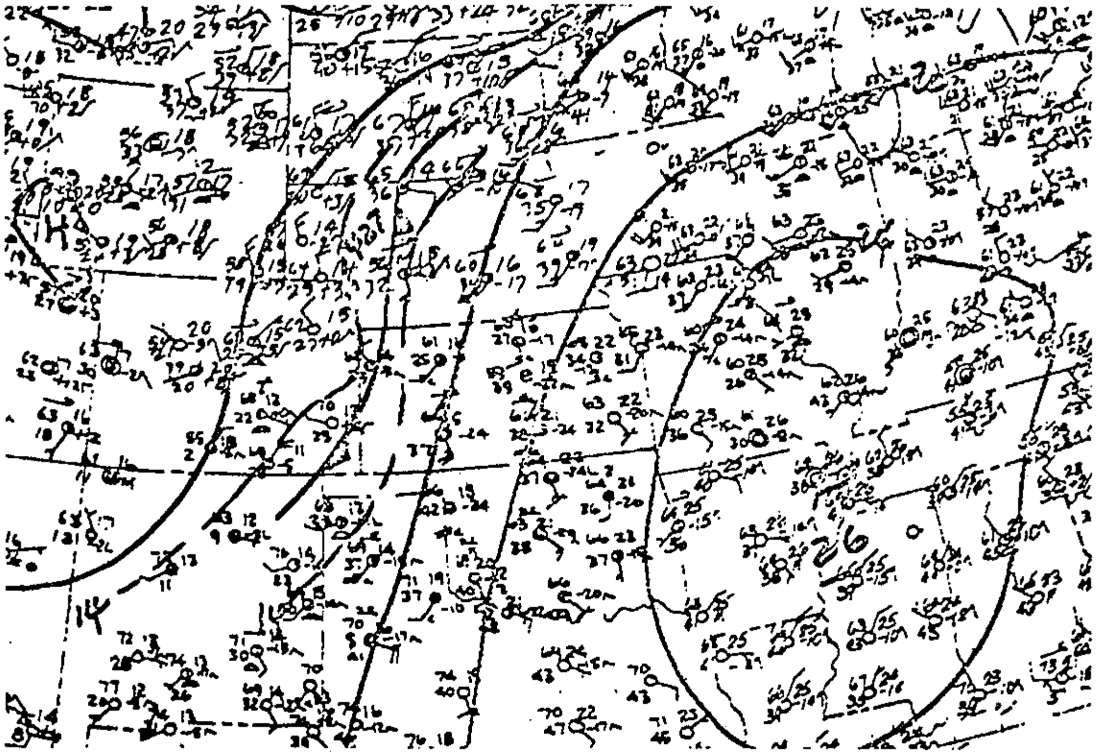


Figure 2a. Sea-level Chart
18Z April 17, 1974

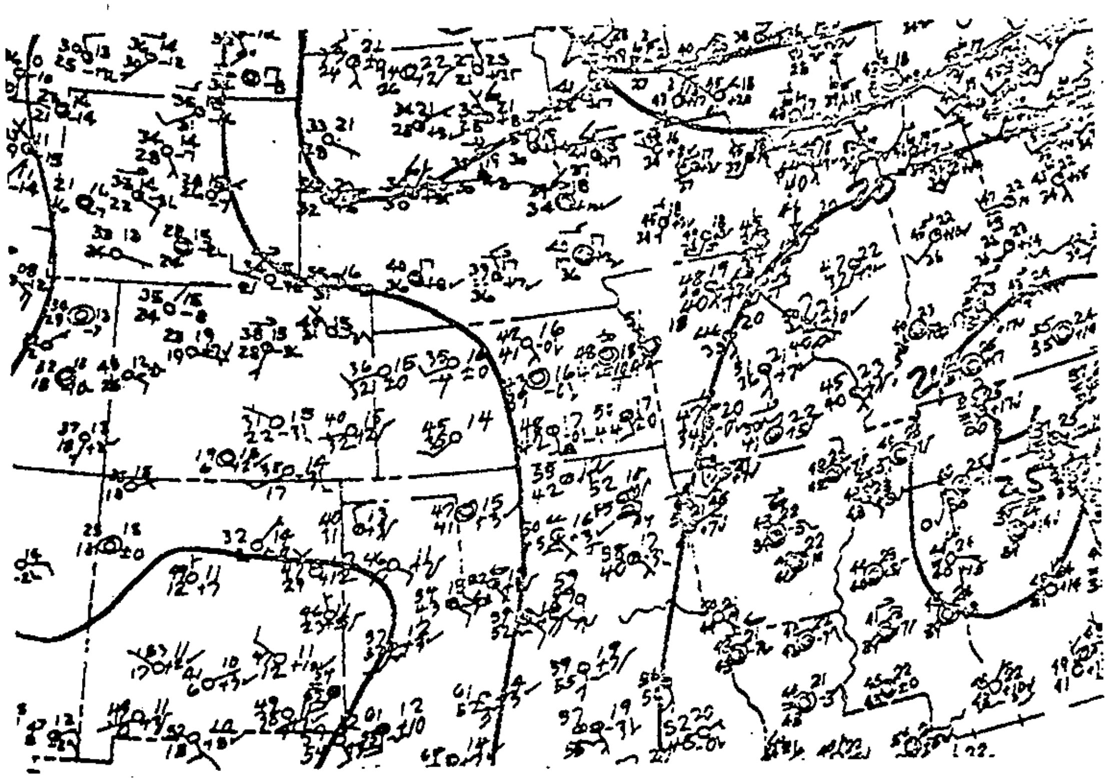


Figure 2b. Sea-level Chart
12Z April 18, 1974

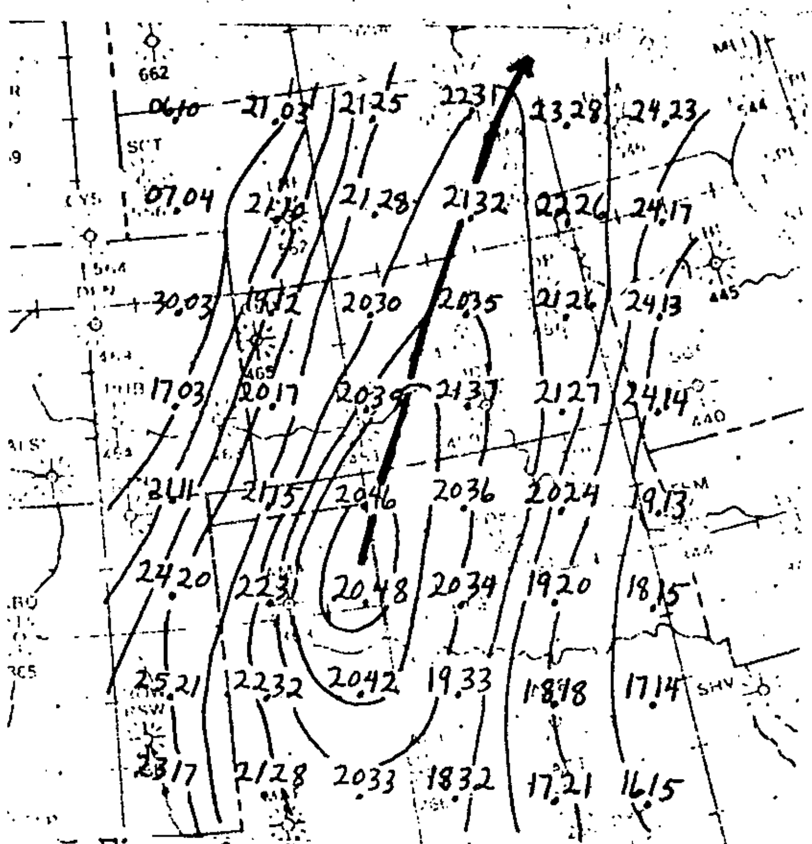


Figure 3. Surface Geostrophic Winds
18Z April 17, 1974

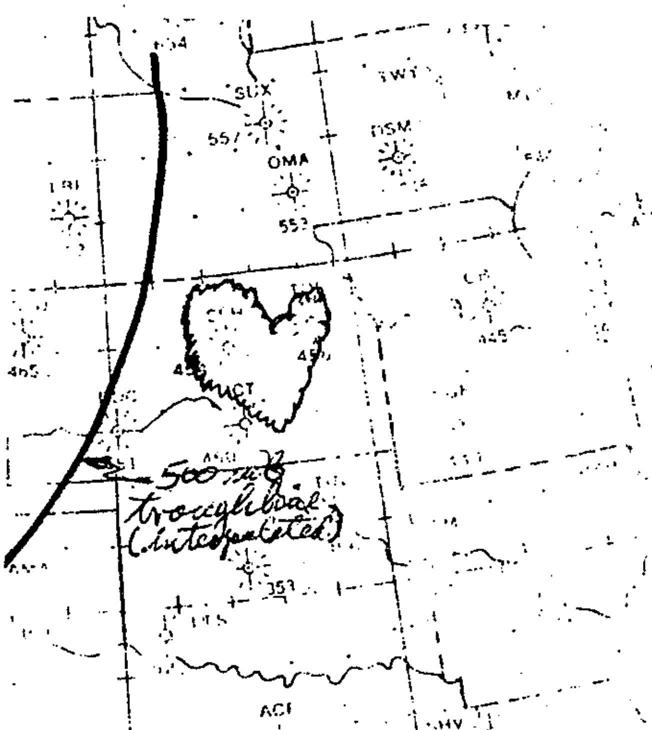


Figure 4a. RADU Chart
2035Z April 17, 1974

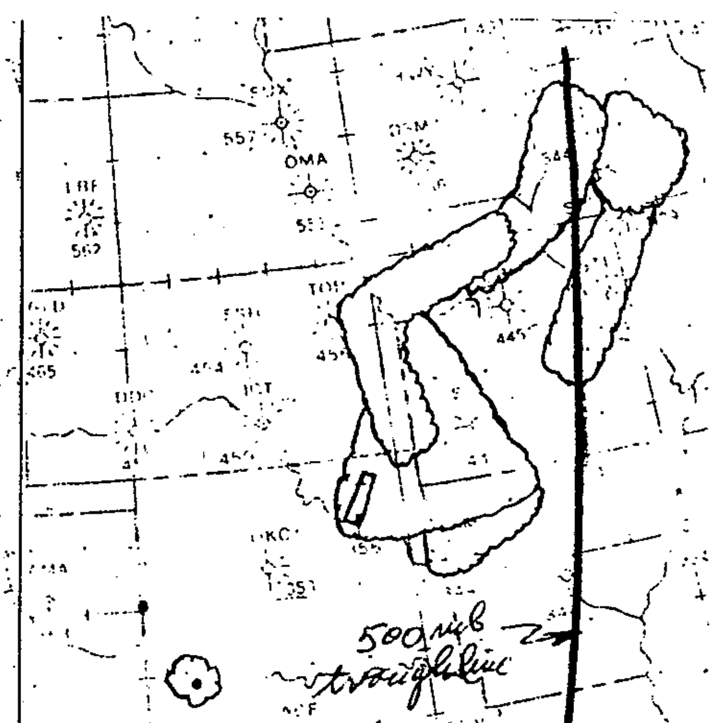


Figure 4b. RADU Chart
1235Z April 18, 1974

TECHNICAL ATTACHMENT 74-11

CHECK SHEET FOR USE OF THE SURFACE GEOSTROPHIC WIND (SANGSTER) CHART

Technical Procedures Bulletin No. 111 describes the Sangster chart and indicates its main uses. The check list below gets more explicit on how to use this chart (primarily warm season use and for the 4 PM forecast) in conjunction with other analyses and guidance material. The order indicated is that one is most likely to follow on shift, and does not necessarily indicate the importance of the individual items. The items are listed twice. The first with an explanation, the second time as an outline for forecast desk use. The surface vorticity part of the chart is given little comment, as this has more utility in the colder part of the year. The concern is mainly nocturnal convective activity.

1. Look at the larger scale upper-level (500 mb) flow, using the barotropic, LFM, and PE. The barotropic is first as it tends to handle smaller systems better and the forecaster experience level is higher. The LFM is next because of its smaller grid size. Note the direction of flow for a first approximation to the movement of convective systems. Note PVA areas, as these will tend to augment convective activity.

2. Look at the facsimile sea-level (surface) chart. This gives the larger scale surface picture. Note Lows that appear to be moving (use also the vorticity part of the Sangster chart to assist in determining the location and movement of surface Lows). The movement of Lows can change the low-level jet (LLJ) picture from that one expects from diurnal changes alone. The pressure change chart gives clues here as well.

Note also the location of warm or quasistationary fronts which, in conjunction with the LLJ, could produce overrunning. If a front is not shown explicitly, the upstream (use surface winds) edge of a large cloud shield can suffice.

3. Look at the 18Z Sangster chart (the maximum wind on this chart tends to be at the time of maximum temperature, and the minimum wind at minimum temperature time, so the 18Z chart can be significantly different from the 12Z chart as far as the LLJ is concerned). Note where the LLJ axis crosses the warm front (maximum overrunning), and also note the maximum wind in the LLJ and the location of areas with a significant decrease in wind along the direction of the surface wind.

4. Look at the 850 mb chart. Note the areas of warm advection from that chart and note the areas of low-level warm advection using the 850 mb temperature lines and the LLJ axis. Consider the low-level diurnal change in wind direction and speed, (see figure in TPB 111) to judge diurnal variation in overrunning. Use of the LLJ should narrow the most likely strong overrunning area. Compare the most favorable area thus obtained with that estimated

above from use of the front. The indications of the 850 mb chart should be the better estimate. Note also the dewpoint spread. (see next).

5. Look at the stability, precipitable water, and relative humidity chart(s). Note the dewpoint spread at 850 mb and 700 mb and plot a critical sounding or two to note whether small or large amounts of lift are needed for saturation. The speed decrease along the LLJ, when acting alone, i.e. without PVA aloft or warm advection, is not likely to produce precipitation with air dryer than average. Intense rains, such as with most flash flood threats, will probably not occur unless the moisture condition, at least in the low levels, is above average. The stability and moisture parameters, especially the moisture, can give sufficient information to say that no rain will occur or that it will be light at most, but rarely can they help refine the most likely area.

6. Look at the radar (RADU) chart. If existing convective areas are extrapolated to move into favorable precipitation areas of the LLJ, the existing areas will very likely increase in area and intensity during the night. The most likely portion of the area to enlarge is the upwind side using the surface wind, as this is the area where the LLJ can get lifted by the outrushing cool air of pre-existing thunderstorms.

7. Compare all of the above with conditions exactly 24 hours ago. More favorable conditions would be a stronger LLJ max wind, a larger decrease of wind along the LLJ per unit distance, stronger warm advection or stronger PVA aloft. The most important favorable item may be the relative movements or changes in magnitude of the favorable areas in the past 24 hours so that today the PVA aloft has moved to the location of the LLJ speed decrease or warm advection areas etc. through moisture, stability, max LLJ wind, local effects, and frontal positions.

1. 500 mb analysis and progs - direction and speed of flow, and PVA

2. Sea-level (surface) fax chart - moving Lows, fronts

3. Sangster chart - axis, max speed, and speed decrease in the LLJ

4. 850 mb chart - warm advection, moisture

5. Stability and Moisture chart. See also 850 mb and 700 mb moisture and soundings

6. Radar chart - existing systems to be augmented

Since WSOs do not have the Sangster chart available, they cannot evaluate item 3. However they should evaluate all the other items as routine, then contact their WSFO if they feel item 3 could be significant to them. Knowing WSOs do not have the Sangster chart, WSFOs should be alert to inform them when item 3 is significant for the WSO's area of responsibility.

TECHNICAL ATTACHMENT 74-12

THE FLOOD PRODUCING RAINS IN MISSOURI AND ILLINOIS OF MID MAY 1974

The upper air charts gave no appreciable clues to the events that occurred. The 500 mb chart had southwesterly flow averaging about 35 knots (slower speeds are more favorable for heavy rain), but the area receiving the rain was under the southern edge of the westerly belt and the vorticity advection (PVA) areas were very weak and came from questionable wiggles in the contours. The clues for the events thus had to come primarily from the surface chart.

There was a surface front in the area, and it was obvious from the precipitation of the night before and the surface dewpoints in the lower 70's that it would occur again, so the question became one of where the maximum probability would be for the coming night.[1] The Sangster chart, as discussed in Tech. Procedures Bulletin No. 111 and in CR Tech. Attachment 74-8 and 74-11 is excellent here, and better than the facsimile sea-level chart as it contains features of significance such as a diurnal cycle that is suppressed in the sea level chart. Also, giving the geostrophic wind explicitly makes for simpler evaluation.

TPB 111 gave two ways in which the Sangster chart could help find the most likely area for nocturnal shower activity. One involved a decrease in speed along the axis of the low-level jet (LLJ), and the other involved the area where the LLJ axis crossed a surface front. These two conditions appeared to figure strongly in this very heavy rain situation. Figure 1 shows the Sangster chart data for 18Z May 16, 1974, as is now available to WSFO's on request/reply teletype. The isotachs of geostrophic wind speed are dashed and labeled (in knots), and the geostrophic streamlines (actually stream function) are the solid lines. The double maximum in the wind speed is an unusual feature, and maximum wind speeds over 45 knots, as in even the weaker maximum, is higher than average. It is quite obvious from this chart that the maximum overrunning of the front would be in northern Illinois where the LLJ axis crosses the front, but the overrunning would extend southwestward along the northern side of the front well into Kansas. The most significant area of speed decrease along the streamlines and the overrunning is shaded, although the boundaries are not fully definable lines, especially to the north in the overrunning area.

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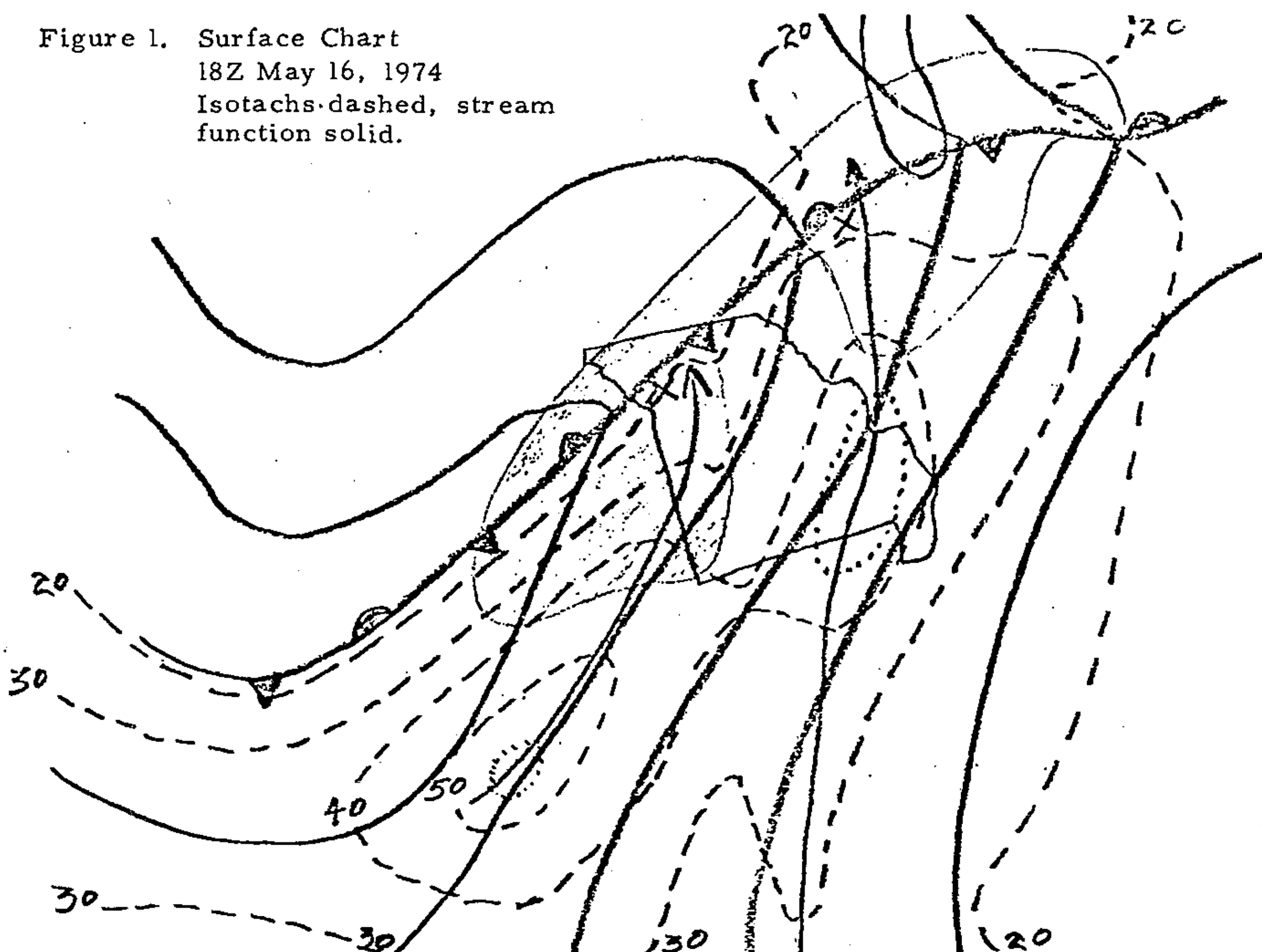
[1] You may recall from CR Tech. Att. 74-9 that except for the front range of the Rockies and southeast Missouri, the maximum frequency of heavy rain in the Central region west of the Mississippi river is almost exclusively in the night (00-12Z) period. An examination of about a dozen very heavy rain situations in the CR showed that all were at night.

The observed precipitation that night was roughly in a band almost 200 miles wide from about Concordia,Kansas through Rockford, Illinois into lower Michigan. There was some precipitation north of the shaded area, but it was almost all quite light. There were two definite maxima of precipitation with amounts over 4 inches, one near Kansas City and one in northern Illinois (both marked with an X). Because the max frequency of heavy rain is usually quite late in the Kansas City area (see T.A. 74-9), the rain went a bit into the "tomorrow" period. Note that the two precipitation maxima are closely associated with the two wind maxima and the jet axis from these maxima. It would fit the model given in TPB 111 a bit better if the max precipitation had been to the right of the jet axis instead of to the left, but the reason for this deviation is not understood.

The extreme southwest portion of the shaded area received no precipitation. This might have been expected if one looked at the sounding or the upper-level charts, as there was a strong inversion present, topping out about 850 mb, with the air above the inversion so dry that the dewpoint spread was given as an X on the facsimile charts.[2] On the other hand, the unshaded area in northeastern Missouri, etc., while definitely a minimum of precipitation between the two maxima, did receive precipitation. This too could be expected because the southwesterly flow aloft, acting as a steering current for convective cells which in themselves tend to be self perpetuating once formed, would move the cells from the generation area in western Missouri to the apparently less favorable area in northeast Missouri.

Figure 1. Surface Chart
18Z May 16, 1974
Isotachs dashed, stream function solid.



The next night heavy rain occurred again, and Figure 3 shows that situation as the latest data available to the forecaster before the late afternoon forecast (18Z May 17). The upper air support was still nil with southwesterly flow. While there is some decrease in wind along the LLJ, it is obvious in this case that strong overrunning is likely to occur in extreme north eastern Kansas and northward, and that this will be the dominant feature for the coming night's weather, with precipitation spreading northeastward in the upper flow.

The observed precipitation covered about the same area as the night before, with the following exceptions:

1. it extended farther north with considerably larger amounts farther north, most likely because overrunning was the stronger and speed decrease was weaker,

2. there was no maximum in northern Illinois, as the eastern wind max was no longer present,

3. there was a clear maximum, again more than 4 inches, in the Kansas City area, this time just to the right of the LLJ axis in the more normal location.

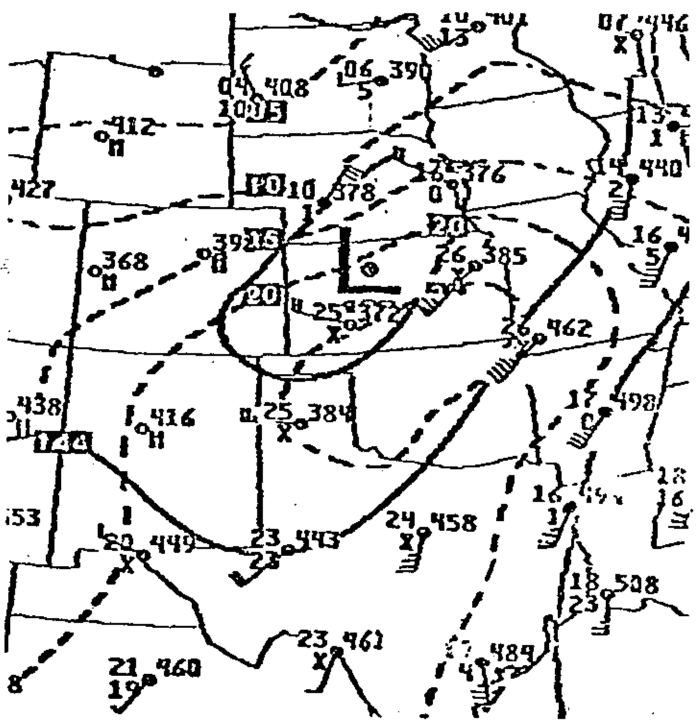


Figure 2a 850 mb chart
12Z May 16, 1974

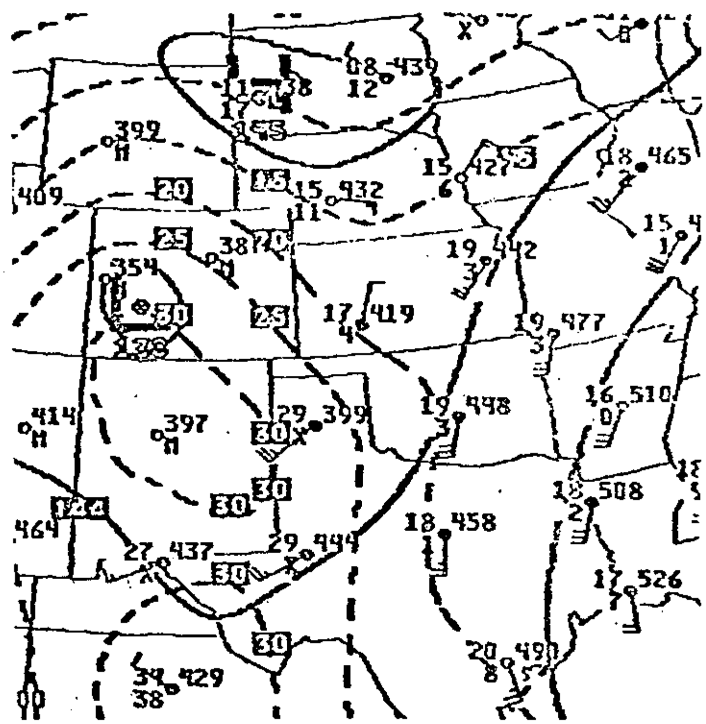


Figure 2b 850 mb chart
00Z May 17, 1974

2 The diurnal vertical movement of this inversion (higher in the daytime due to convection) gave odd looking changes to the 850 mb chart that some might have trouble understanding without having a plotted sounding. These involved major changes in temperature and in dewpoint spread that obviously could not occur due to horizontal advection, as, for example (see Fig. 2) the temperature dropped to 7°C in 12 hours at Topeka and Springfield in spite of warm advection.

The subjective progs from NMC did a fairly good job for these two nights, but they did not indicate the two maxima that one should expect the first night, nor did they pinpoint as well the maximum on the second night.

IN SUMMARY. On the Sangster chart analyze the isotachs and watch for speed decreases in the LLJ and note the location of the crossing of the front by the LLJ. This should help you gain confidence in some NMC QPF forecasts, add detail to other forecasts, and augment their efforts in some cases where they do not show precipitation.

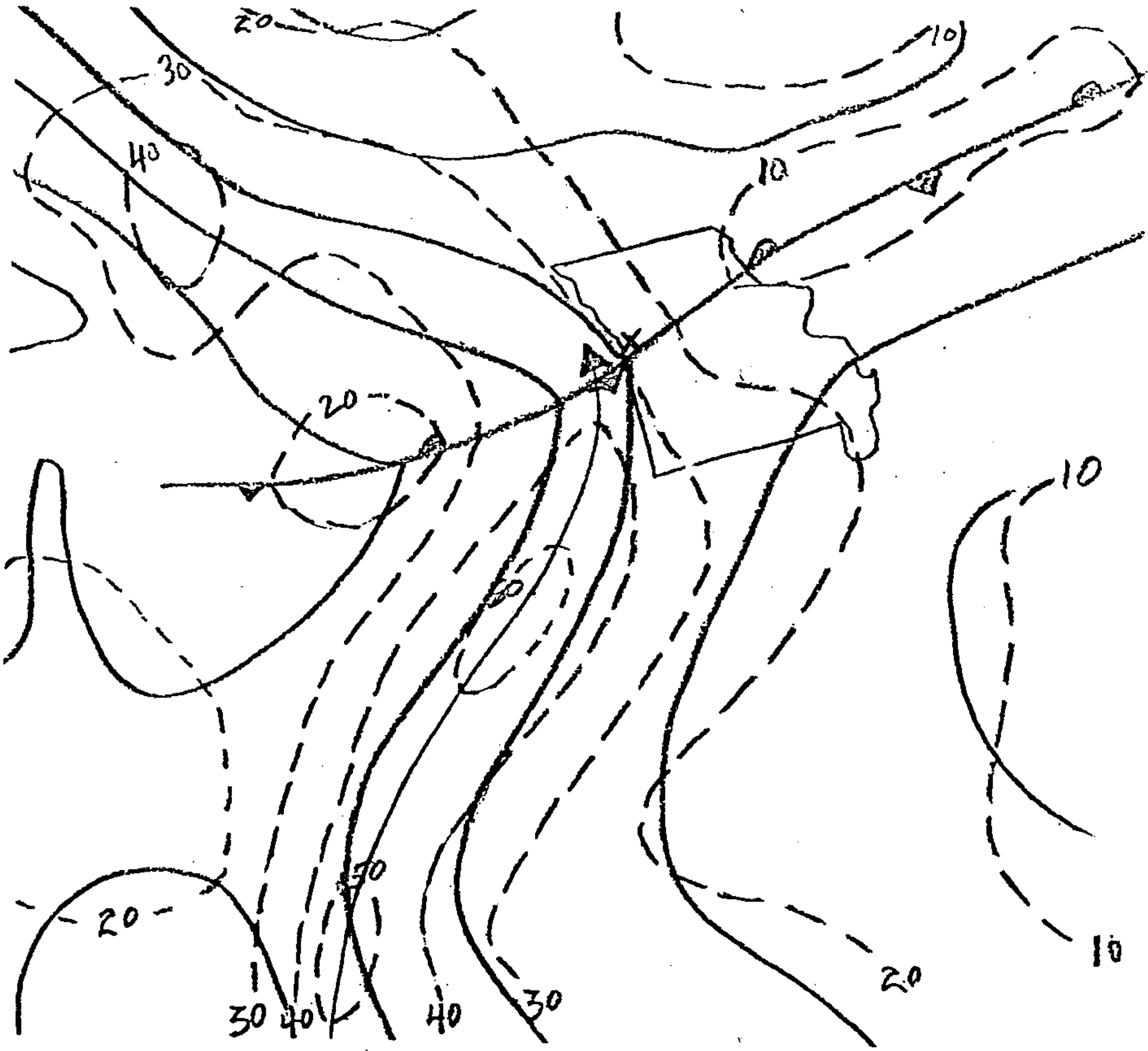


Figure 3. Surface Chart 18Z May 17, 1974

TECHNICAL ATTACHMENT 75-5

MAXIMUM TEMPERATURE OVER SNOW - CHEYENNE

The effect of snow cover of various depths on the maximum temperature has been treated for 10 years of data for Sioux Falls in Technical Attachment 73-5 of the Central Region. The present TA presents the same type of material for Cheyenne, based on 21 years of CYS data processed at CYS and recently sent to Scientific Services. This TA compares the two sets of data to examine the premise that higher temperatures are more likely to occur over snow at Cheyenne since at times diurnal heating is augmented by heating due to foehn (chinook) winds.

The table below gives the frequency distribution (in percent) for the airport at Cheyenne (Sioux Falls--taken directly from TA 73-5) for snow of various depths vs. max temperature, as well as the highest observed temperature for the various snow depths. Of course the much larger Cheyenne sample would have some effect, since extreme conditions are more likely in large samples. However, the sample size is fairly large in both cases and it would have more effect on the highest temperature observed than it would the percentages in the various categories.

	Snow Depth (inches) at 12Z				
Max. Temp. (°F)	1	2 or 3	4 pr 5	6	All
30-34	22(41)	33(30)	29(45)	40(43)	28(40)
35-39	24(29)	25(30)	28(33)	18(43)	24(34)
40-44	29(13)	21(30)	29(18)	18(7)	25(16)
45-49	15(7)	12(6)	11(2)	11(6)	14(5)
50-54	7(6)	6(2)	3(2)	7(1)	7(3)
55-59	3(1)	2(2)		4	2(1)
60-64	<1(2)	<1		2	1(1)
65-69	(1)	<1			<1(<1)
	305(189)*	243(118)*	79(126)*	45(183)*	672(616)*
Highest Temp. (°F)	62(69) Jan-Apr.	67(57) Apr.	53(51) Oct.	64(52) Apr.	

*Values are about double these of TA 73-5 in order to compensate for the smaller size of the sample in TA 73-5.

Note under the ALL category that there is a higher percentage of cases in the higher temperatures (≥40°) at CYS than at FSD. Since the figures are in percentage, this of course must give lower percentages at CYS in the colder temperatures. Also notice that with deep snow (at least 4 inches in depth) CYS had 6% of its cases with temperatures 55° or more, while FSD had none.

Thus it appears that the premise is correct, and the foehn effect is significant. This fact adds complication to the adjustment for snow cover of objective schemes for forecasting temperature such as those currently providing temperatures based on model output statistics. It would suggest that those east-of-the-Rockies locations with foehn effects be treated separately from those without such effects.

According to MIC Beebe at CYS, "In looking at a few of the unusually high temperatures with substantial amounts of snow (more than 4 inches), these situations are associated with westerly winds or light southerly winds. Also, the high temperatures did not (entirely) melt the snow to permit these high temperatures to occur". The westerly winds are a lot easier to understand than the light southerly winds, as they fit much better with the premise.

Using dates provided by Mr. Beebe on which these high temperatures occurred with deep snow, we looked at seven other cases, using the one-a-day Daily Weather Map information. The maximum temperature ranged from 43 to 52, so the cases were average and definitely not extreme. Using some judgment for what happened between chart times, all cases had 500 mb winds from nearly west at 30 knots or more over Cheyenne for at least part of the day. The surface winds varied quite a bit in strength and direction but were mainly westerly there as well as aloft. The most pronounced case of the group was January 14, 1973 with a 51° max temperature when the season would dictate the coldest conditions (the 52° case was in mid-March). In this January case, the 500 mb wind was about 290° at 50 knots, and the surface wind was also strong and from just north of west. These cases also tend to confirm that the foehn effects are significant, and suggest that the greater temperature anomalies are with the higher 500 mb wind speeds when the 500 mb wind is close to west.

The highest temperature observed in each category for snow depths over 1 inch, is at CYS rather than FSD, which also fits the premise. However, why CYS is so close to FSD for the 4-5 inch cases is not known. The month(s) of the highest temperature for CYS is given below the highest temperature value. It is interesting that April predominates.

Another point of interest is that two-thirds of the cases with snow 6 inches or more on the ground occurred during March and April, while 45% of the cases with 1 inch occurred in December and January. Note in the table above, that while CYS had somewhat more days with snow on the ground than FSD, that CYS had far fewer days with large amounts of snow on the ground. This is in spite of the fact that CYS averages 52 inches of snow a year while FSD averages only 40 inches. CYS is about 2° latitude south of FSD but is almost 5000 ft. higher in elevation. The difference in the frequency of larger snow depths is probably due to two other things, 1) the lesser moisture supply over CYS, and 2) the more rapid melting and evaporation of snow at CYS due to the foehn.

It would be interesting to continue the comparison by using data at a station fart south than CYS where the foehn should be stronger. Further treatment for the CYS data would be to stratify with respect to wind direction and perhaps speed.

TECHNICAL ATTACHMENT 75-14

METEOROLOGICAL ASPECTS OF THE DAKOTAS-MINNESOTA FLOODS OF EARLY JULY, 1975

Lawrence A. Hughes, Scientific Services, CRH

Floods from thunderstorm rains over so large an area and in rivers as large as the Red River of the North are unusual, especially in July. Such hydrometeorological events, of course, require unusual conditions to produce the necessary water as thunderstorm rainfall. Of the meteorological events that took place, none by themselves are unusual, but the persistence of the combination was unusual. The events included stability lower than usual for the time of year, high moisture, a stronger than usual low-level jet, warm advection in the low levels, slow movement of thunderstorm cells at times, and a perisistence produced by a stagnating flow pattern at both the surface and aloft. The initial thunderstorms started a chain reaction in which the strong winds of the low-level jet impinged on the front-like boundary of the pool of rain-cooled air from previous thunderstorms and the stagnation caused these events to recur day after day.

Troughs at 500 mb and their associated positive vorticity advection appeared to play a minor role. This is not unusual at this time of year, and if such troughs had played a larger role, the stagnation that was necessary for the repeated rainfall would not have existed, as the surface pressure systems with such troughs would have changed air masses with each passing trough. With upper-air systems weak, the low-level jet frequently plays a major role in mid-west weather systems, as it did in this case.

The main water of the flood was in the heavy rain of four consecutive days starting in late June. The main portion of these rains were in a band roughly from SE N. Dakota to NW Minnesota, with the greatest storm total of 20.6 inches reported about 25 miles SW of Fargo. However, the slow recession of rivers in this nearly level country allowed one or two rains in the following week to also contribute to the highest crest on the Red River downstream of Fargo. Here we will look primarily at the meteorological conditions of these four consecutive days of heavy rain.

The last major upper-level trough to cross the area before the pattern stagnated was on June 26, when a cold front and associated showers moved through. The rain amounts were significant enough to wet the ground well but were not large. The main rains started on June 28th when another cold front and associated showers moved into the area (see Figure 1). By this time though, the upper air pattern had stagnated enough that the moisture seen in the precipitable water on the facsimile chart had a tongue of high values from the Gulf of Mexico over the Plains to the Canadian border, with values over one inch in the Dakotas. The stability was low in this tongue, with Lifted-Index values as low as -5 and

K Index values as high as 39, in the Dakotas. The push of the cold front, combined with a center of low-level warm advection (Fig. 2), and even some positive vorticity advection (PVA) aloft with a 500 mb trough whose main PVA, though, was northwest of the area of interest, (see Fig. 3) set off the initial thunderstorm activity and heavy rains. An examination of the radar charts from RADU showed that early on the 28th, as the stagnation set in, the cell movement indicated on the charts dropped from around 35 knots to mostly 15-20 knots, which is well below average speed. The rest of the days also had mostly below average cell speeds.

The warm advection over the area of interest initially (Fig. 2) is not strong, especially by winter standards, but it is average for the time of the year. There is no question that the maximum warm advection is over the area of interest, but exactly which portion of the area it is over and exactly how strong it is cannot be determined readily or perhaps at all with the wide spacing of upper-air data.

The stagnation that set in as the initial heavy rains came allowed the low-level warm advection to remain over the area of interest. The 20°C temperature line at 850 mb in the area hardly moved during the next three days in spite of moderate winds nearly perpendicular to it, attesting to the upward vertical motion that occurred. The 500 mb charts and progs showed little if any PVA aloft during these three days. The stagnation caused the surface cold front to stop moving, and thus the front did not sweep out the moist unstable air ahead of it, and, at least equally important, it did not disturb the pool of rain-cooled air that was created by the initial thunderstorms.

This rain-cooled air was important as its boundary acted like a miniture cold front so that the winds of the low-level jet, which were a maximum in the area and stronger than average, were lifted by impinging on this cold front so as to start new thunderstorms. Also, the low-level jet had significant speed decreases along the streamlines over the area, and this is believed to also contribute to low-level convergence and lifting of the low-level air. These low-level jet effects can be seen from Fig. 4, from the Daily Weather Map which shows the stagnated front and the squall line generated by the earlier thunderstorms, and Fig. 5 which shows the streamlines and isotachs of the low-level geostrophic wind as analyzed from the winds on the so-called Sangster surface chart available on request/reply teletype. The 21Z time is used as being the nearest available chart to the maximum geostrophic wind condition.

An interesting point that can be seen only with satellite data, is that the main thunderstorms in the area on the 29th started very late in the afternoon in extreme north-central S. Dakota at the junction of the stationary front and the remnants of the squall line that are indicated earlier in the day on Fig. 4, and then quickly spread over much of N. Dakota. Earlier in the afternoon there was a very large arc cloud of the type discussed by Purdom (1973), that came out of the dissipating thunderstorm mass, but it reached its limit late in the day and was not readily visible when the main new thunderstorms broke out again. The same thing happened again the next day, although the initiation area was then a bit farther east.

Those somewhat familiar with the low-level jet in the mid-west may think of it as creating only nocturnal thunderstorms, as it is noted for this, but while the real winds of the jet do maximize at night, the winds are stronger along its axis than elsewhere at any time of the day and can be strong in the daytime as well. In this case the maximum geostrophic wind of over 50 knots was unusually high (around 40 knots is normal for 21Z), so this jet wind impinging on the discontinuity of the rain-cooled air could initiate thunderstorms at any time of the day, and rain at some time in this period did occur at almost all possible hours, although late afternoon and night-time hours predominated in the production of rain. This low-level jet pattern also persisted for several days with little change.

The shift of the heavy precipitation northeastward to northwest Minnesota on the last of the four days is interesting. Note that the upper air chart for the 2nd (the end of the precipitation period), as shown in Fig. 6, showed even less favorable conditions than earlier, as by then the southwesterly flow usually considered so favorable, had moved westward a bit to over Montana. This westward shift of upper-conditions and eastward shift of precipitation strongly suggest a non-upper air cause for the precipitation. The surface conditions seemed much the same (Fig. 7), but the low-level jet axis at 21Z on the 1st, (Fig. 8), had shifted eastward so as to miss N. Dakota entirely, thus strongly suggesting the jet axis position as the factor relating to the location of the heavy rain. The stopping of the heavy rain for a few days after the rain early on the 2nd also appeared to be well related to the low-level jet, as can be seen by Fig. 9. Notice that the winds over the area of interest are markedly different from earlier, with no jet axis and much lighter speeds. Incidentally, the high moisture and low stability were still present in the area, further pointing to the major role of the low-level jet.

One could easily take from the above that the upper-air patterns, especially through the mechanism of the PVA with troughs, were not a significant factor. From the evidence one can see on the facsimile 500 mb analysis of the raw data or as initialized in any of the numerical models this would be true, as they did not show troughs with sufficient size or continuity to be sure of their existence, except for that of the first day. While the upper-air network is adequate, and perhaps overly adequate for most large scale systems, it may not be adequate for the small synoptic systems that can be a factor in warm seasons rains.

An observing tool with excellent time and space resolution is the GOES satellite. When these pictures were examined, it appeared that upper-level systems did, at times, relate, but mostly they were not critical. Figs. 10a through g, GOES IR pictures, show this clearly for the most prominent upper air system of the 2nd through the 4th day.

Note at the initial time (10a) that the thunderstorms in the low-level jet had fired up considerably in the Dakotas and Minnesota, but the sky was clear to the west until reaching western Montana where the radar chart showed that the clouds contained showers. Six hours later (10b) the Dakotas-Minnesota thunderstorms had weakened some and moved little, while the clouds and associated showers in Montana had moved quite a bit. About twelve hours from the initial time (10c) only a small area of the thunderstorms over Minnesota was left, while the Montana clouds and showers had moved into N. Dakota.

By 18Z (10d) the cloud mass that had come from Montana was clearly dominant, and since it had moved into the area of the low-level jet with its favorable winds, moisture, and low stability and at a time of good surface heating, the showers changed to significant thunderstorms. The area continued to move, so 24 hours after the initial time (10e) the cloud mass and thunderstorms had moved into north eastern Minnesota. These thunderstorms had weakened by then, but note that a new area of thunderstorms had broken out north of Grand Forks, ND and west of the upper-air associated cloud system. This new area was associated with the low-level jet axis, which had moved eastward, and the low-level speed decrease which had moved northward from the earlier times (see Fig. 8). In the next six hours, see 10f and g, the low-level jet induced thunderstorms became the main system, while the area that had come from Montana weakened to no precipitation at all, according to radar reports.

Fig. 11 shows the precipitation gage trace from the only currently available recording station in the climatological network in northwest Minnesota (Thie Lake Refuge). From this and the satellite pictures in Fig. 10 one can see that the low-level jet, apparently working against (behind) the upper level system, produced the same amount of precipitation as the upper-level system working in and with the low-level jet.

The charts of the 500 mb initial condition at 00Z and 12Z on the 1st, and 00Z on the 2nd were looked at closely to see how they related to the satellite observed area moving from Montana. On the chart with data, there was no appreciable indication of a system at that level that would relate to the satellite pictures... Likewise on the barotropic series of initial condition charts. The LFM had some indication, but it was very weak. Oddly, the PE model had the best indication in spite of its loss in initialization. The reason for this about face in initializing smaller systems is not known, but this case does show how satellite pictures can assist the forecaster in selecting among numerical models, and how the pictures can augment upper-air analyses.

SUMMARY

A stagnating flow pattern at the surface set up moisture, stability, and low-level jet conditions that caused repeated thunderstorm activity over the same general area for four days in a row, thus creating most of the rain that caused the flooding, although some later rains also contributed. The location of th main heavy rain areas seemed well associated with the low-level jet axis, low-level jet speeds and speed gradient along streamlines. Upper-air features normally associated with precipitation appeared to play a minor but not insignificant role, and these features were best seen and followed via the GOES satellite pictures. The rain with the upper-air systems became more intense as the system came into the low-level jet region, although the high moisture and low stability of this region may be as much the cause as the low-level jet winds.

REFERENCE

Purdom, James F. W., 1973: Meso-highs and satellite imagery, MON. WEA. REV. 101, 180-181.

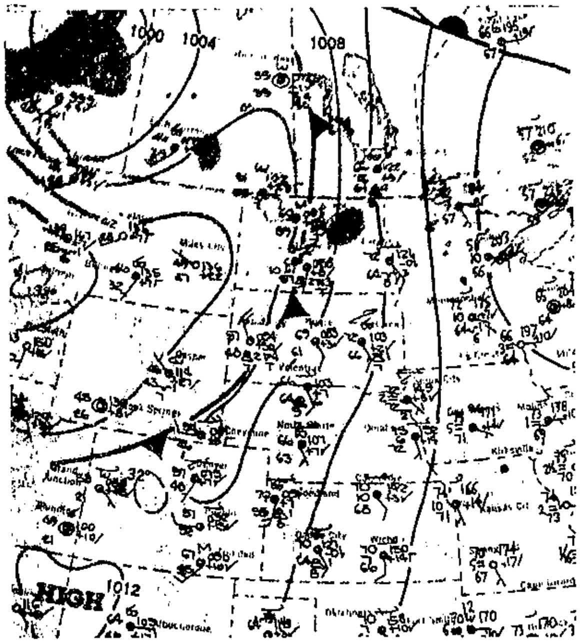


Fig. 1. Surface (sea-level) Chart
12Z June 28, 1975

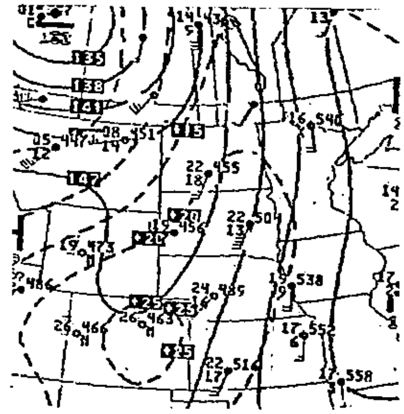


Fig. 2. Same as Fig. 1 but for
850 mb.

Fig. 3. Same as Fig. 1 but for 500 mb

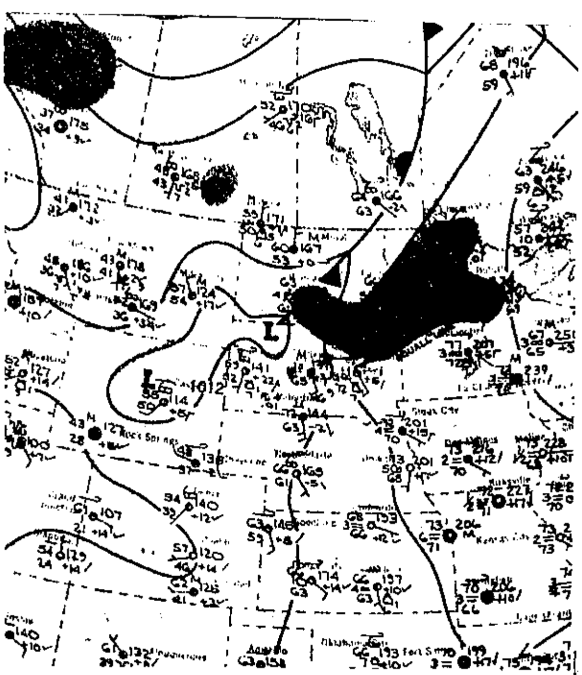


Fig. 4. Surface (sea-level Chart
12Z June 29, 1975

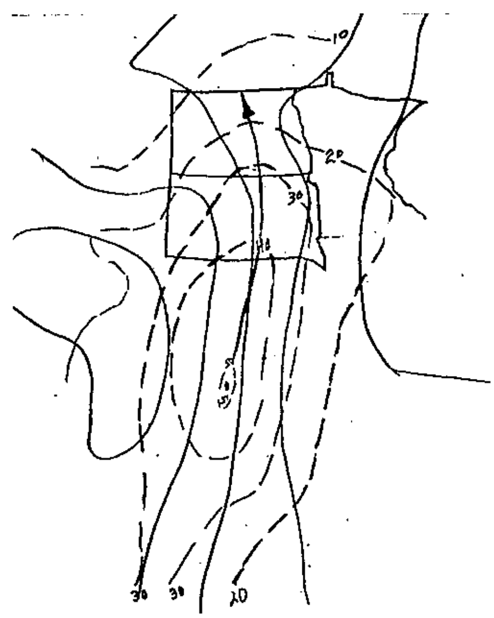
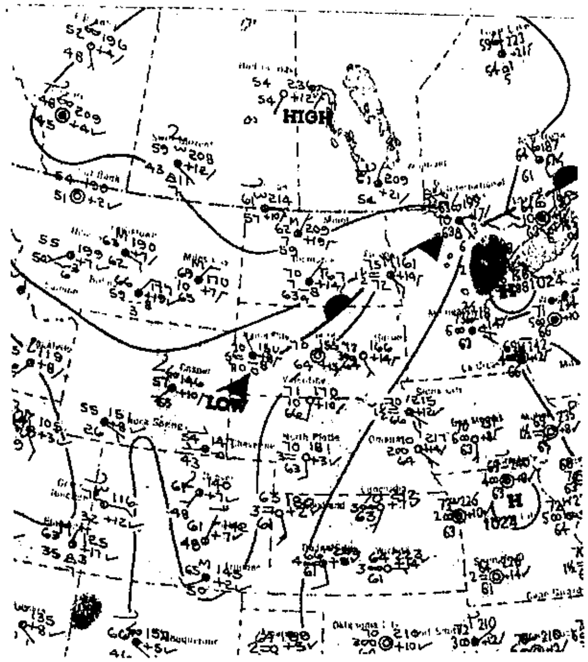


Fig. 5. Surface Geostrophic Wind
(from Sangster Chart).
Streamline solid; Isotach (kts)
dashed. 21Z June 29, 1975

Fig. 7. Surface (sea-level) Chart
12Z July 2, 1975



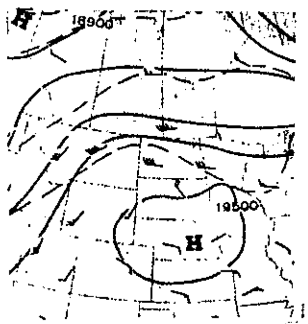


Fig. 6. 500 mb Chart
12Z July 2, 1975

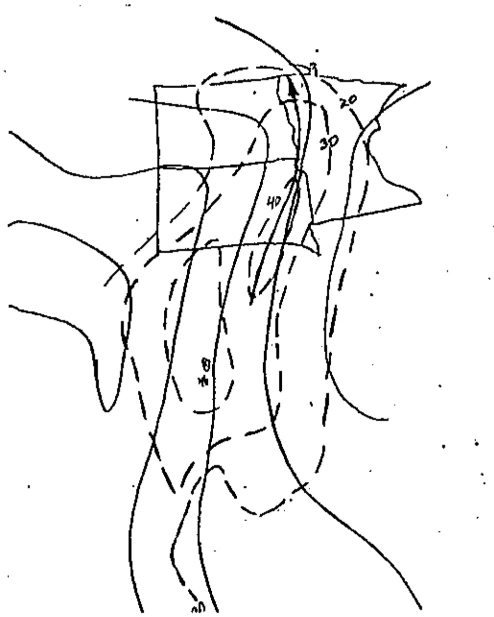


Fig. 8. Same as Fig. 5 but for
July 1, 1975

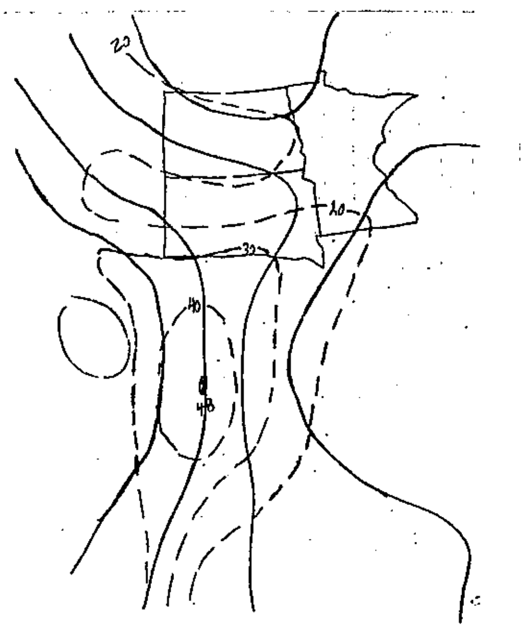


Fig. 9. Same as Fig. 5 but for
July 2, 1975

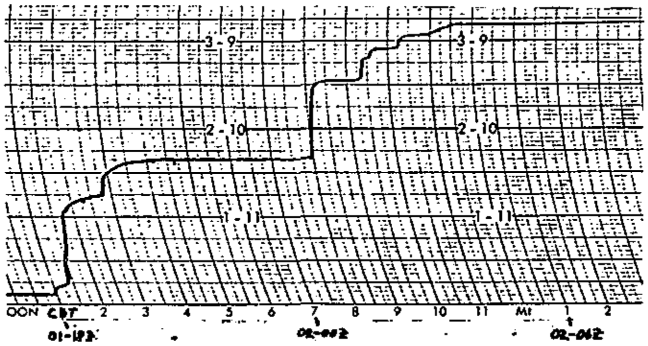


Fig. 11. Rain Gage Trace,
NW Minnesota,
July 1-2, 1975

CRH-SSD
February 1977
L.A. Hughes

TECHNICAL ATTACHMENT 77-3

RAOB ANALYSIS-MIXING AND HEATING

The 9th of December, 1976 was interesting because the maximum temperature of the preceding two days at Kansas City had been in the low 20s and yet there was some thought of breaking the high temperature record (68°F) on the 9th. This TA discusses forecasting temperature changes in the boundary layer in general using this case because such changes can be of significance in forecasting more than maximum temperatures. It looks at the effects of mixing and heating separately for the convenience and understanding, and then combines them.

Figure 1 shows the Topeka sounding for 12Z on the 9th (the dashed lines to be discussed later). Although the surface winds were light before dawn, the day should clearly be one with strong surface winds and therefore strong low-level mixing because the 12Z surface geostrophic wind from the Sangster chart was 40 kt. and increasing. Let us first consider the effect of thorough mixing in the boundary layer.

First, note that Figure 1 shows little mixing, as the lapse rate near the ground is not dry adiabatic. This lack of mixing with such a strong geostrophic wind is probably due to the fact that the geostrophic wind increased (actually doubled) during the night when clear skies and good radiation suppressed convection and therefore mixing. If the geostrophic wind had increased during daytime hours instead, heating of the ground would have produced convection and mixing, so the surface wind would have responded more quickly to the increased geostrophic wind and the higher winds shown in the raob. This is an important point to keep in mind when forecasting low-level winds.

Figure 2 shows how this sounding might be expected to change after considerable mixing[1] (again neglect the dashed line for now). How high the mixing would go must be decided by the forecaster. It obviously depends somewhat on the strength of the wind, as well as on the amount of heating. But it also depends on the strength of the low-level inversion or stable layer, as obviously it would be hard to move air upward because of the great stability with an inversion as strong as that of Figure 1. An estimate of the depth of mixing might be obtained from the soundings of a recent similar case[2].

The dry adiabat used should be selected so that areas B and C on Figure 2 (neglecting the dashed line) are equal, as this indicates that no heat has been added to or taken away from the air, i.e. only mixing took place. Note that the lowest part of the sounding can be <u>warmed up by mixing alone</u> to a surface temperature about 50°F, and that the upper part of the boundary layer <u>cooled off by mixing alone</u>. This cooling can be important to rain-snow or icing, and we will come back to it later.

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[1]Saucier's "Principles of Meteorological Analysis" discusses this in more detail starting on page 78.

[2]It is assumed that the applicable soundings will be plotted by all offices routinely.

We now consider the effects of heating. Here we need knowledge of the amount of energy that can be put into ground heating by the sun at this time of the year. The best way to get this is from prior knowledge of how much the sounding can be changed at this time of the year, i.e. how big an area of the low-level inversion can heating alone wipe out. Possible choices to be discussed are shown as dashed lines in Figure 1. The warmest of these (line 1), yielding a surface temperature of about 70°F would be typical in summer. It is a reasonable stopping point at that time, as any higher temperature would have to be achieved by heating a layer almost twice as deep, i.e. to about 770 mb.

It is not a reasonable line for December, however, as too much energy (area on the diagram) is needed to create it, even with full sunshine all day. The next lower line(line 2) yielding a surface temperature of around 60°F goes to a similar stopping point, where much more depth of the layer is also involved in additional heating. Whether that curve is still too high for the heating possible in December would depend on the station and the clouds, but it is probably too high for any Central Region location and curve 3 is probably more reasonable.[3]

Once the clear-sky area is selected, it would have to be adjusted downward for cloudiness, and downward for snow cover or moist ground, and upward or downward depending on advection of temperature. In this case the sky was clear except for the small effect of some cirrus. The ground was wet, with a trace of snow, but there was little or no temperature advection as there was little directional change of the wind with height in the raob (the 850 mb chart can also be used for this). Let us say then that the appropriate heating area is to the left of line 4 (area A), after reducing from line 3 because of clouds and damp ground.

Now we are ready to add the mixing and heating effects. This is done by moving the dry adiabatic line on Figure 2 to the right to the position of the dashed line, such that the area between the old and the new line is equal to the area A on Figure 1. If desired, a check on this shift can be made as shown in Figure 3, such that areas B' and C' are equal. Take my word for it, this task is more difficult to explain than to do, so don't be turned off by the explanation. Notice that the amount the dry adiabat in Figure 2 would move to the right is not a lot different with curve 4 vs. curve 3 from Figure 1, as so much more depth of the atmosphere is involved, so the adjustment for cloudiness and wet ground is not sensitive in this case.

Notice on Figure 1 that due to heating alone (area A--use line 4), a max temperature in the low or mid 40s might be expected, while from Figure 2, mixing alone would get it to around 50°F, and both combined would reach mid to upper 50s. Of course, neither of these really acts alone, so a max temperature in the low 40s would be unreasonable.

The observed 00Z sounding for Topeka, obviously near max temperature time because of the dry adiabatic lapse rate all the way to the ground, is shown in Figure 4, with the 12Z sounding dashed. The max temperature in Kansas City was also in the mid 50s. Note in the 00Z sounding the large amount of cooling that occurred around the 875 mb level. This cooling is proof that mixing played the major role in the temperature changes in the lowest levels, as there is no other way to cool this layer under the conditions that existed. Any time mixing plays a major role,

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[3] We are currently putting together plans for a simple project to determine the clear-sky no-advection energy area for a number of Central Region locations at various times of the year, and we will inform selected offices at a later date as to how to go about the project.

some cooling in the upper part of the mixing layer is going to occur. Of course, it is possible with a lot of surface heating to wipe out that cooling aloft, but there is not as much chance at this time of the year. This cooling showed up in an interesting way since the 850 mb chart showed that Topeka cooled off about 4°C during the day, which was unexpected and, for some, not explainable. Such cooling could be quite important at times, for instance if there is a cloud layer that is cooled to below freezing, creating an icing condition, or if there was freezing rain and the cooling brought the sounding all below freezing and changed the rain to snow. The effect of mixing should always be considered in freezing rain situations.

As a final point, note the warming in the layer in Figure 4 centered at about 770 mb. Notice that the lapse rate in the top part of the figure is dry adiabatic and that the inversion (or isothermal layer) has moved to a lower altitude. These are the symptoms of continuing subsidence.

In this example, we started with the mixing effect, but the result would have been the same if one had started with the heating effect. Of course, in nature they both operate together, usually. It may be best to start with the larger effect. If so, one would mostly start with mixing in winter and heating in summer.

SUMMARY

1. When estimating maximum surface temperature from a 12Z raob sounding, both heating and mixing must be considered, with heating usually more important in summer and mixing possibly more important in winter.

2. When considering heating, one must have a reasonable estimate of the maximum amount of energy (the area on the sounding) that the lowest part of the sounding can absorb at the time of year in question. This is then reduced for cloudiness and moist ground.

3. When mixing is a more prominent part of the sounding change than heating, cooling at the upper part of the boundary layer can result from mixing, even during the daytime.

4. Cooling in the upper part of the boundary layer produced by mixing could create an icing problem in the cooled layer, and/or could change a freezing rain situation to a snow situation.

5. An increase in the surface geostrophic (Sangster) wind at night will probably have a delayed effect on the observed surface wind, while a similar increase in the daytime will not have such a delayed effect.

6. A marked change in the boundary layer part of the 12Z sounding is likely to occur during the daytime, following surface geostrophic wind increases of quite a bit during the night.

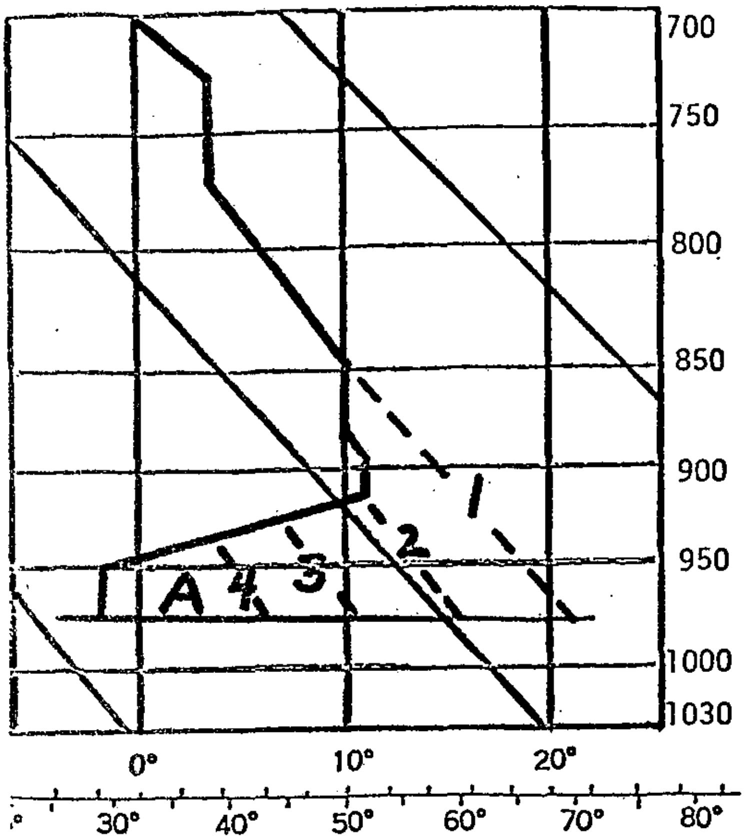

Figure 1. 12Z Dec. 9, 1976 (dashed lines--see text)

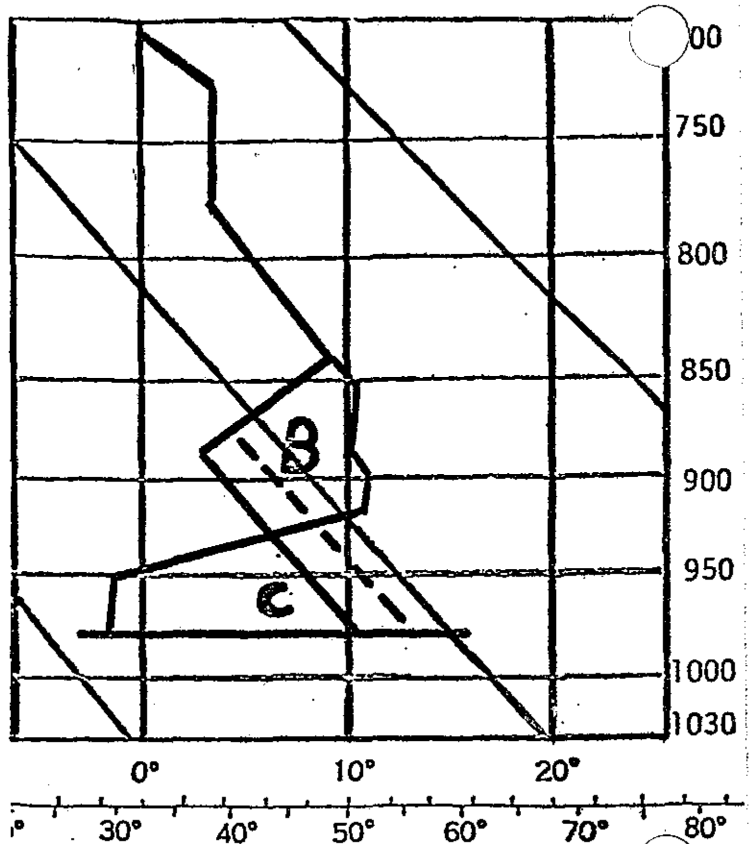

Figure 2. Fig. 1 sounding modified by mixing and heating

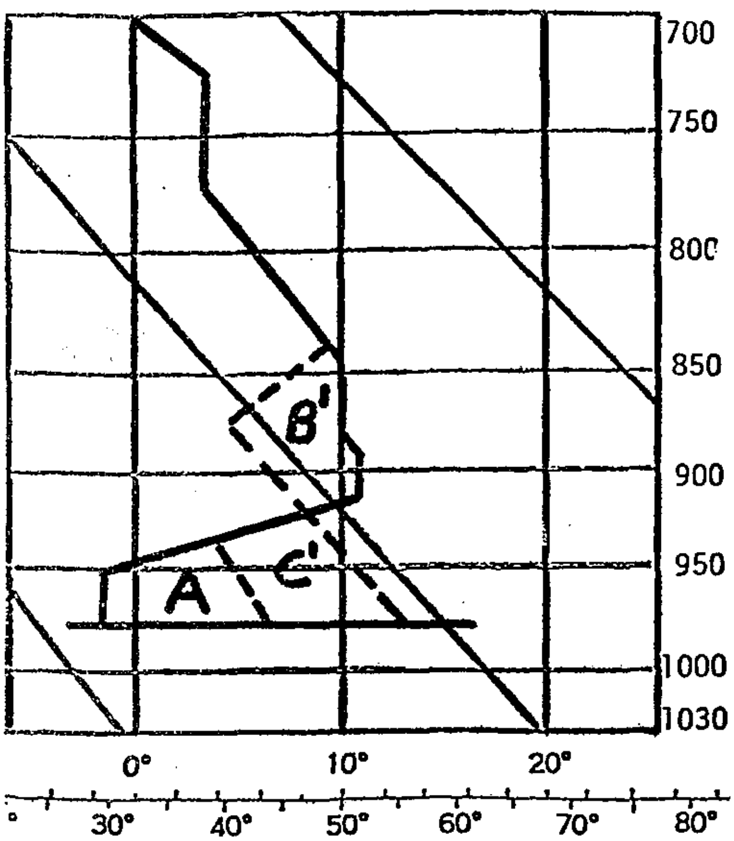

Figure 3. Fig. 1 sounding--a checking pro-dure

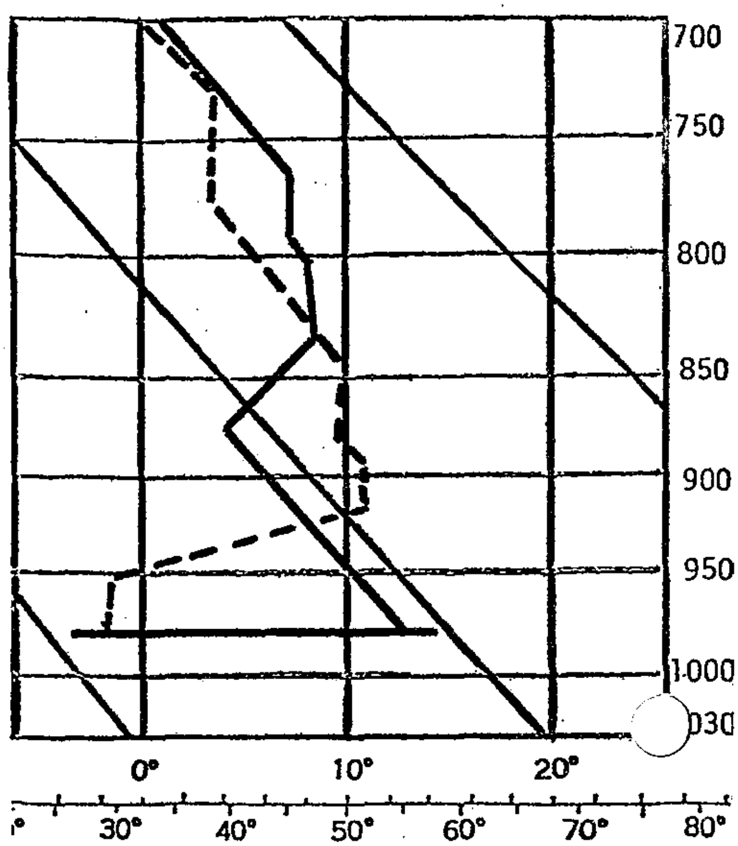

Figure 4. 00Z Dec.10, 1976, with 12Z Dec.9

APPENDIX B

NON-REFERENCED TECHNICAL ATTACHMENTS

1a	Nov. 1967	TA-B	On Lee Cyclogenesis
1b	75- 3		The Storm of January 10-11, 1975, "Why so Intense?"
1c	75- 9		Merger of Surface Vorticity Maxima Portends a Major Storm
1d	71-14		Hage Lows
1e	71-18		Some Rules for Frontal Wave Formation and Movement
1f	73- 2		Gulf Coast Waves
2	73- 4		Reverse Slope Systems
3	75- 4		Arctic Outbreaks
4a	Jan 1970	TA-A	Discussion of a Difficult Precipitation Forecast (Conclusions Only)
4b	74-17		Forecasting Drizzle
5	May 1968	TA-B	Boundary-Layer Wind Maxima and Nocturnal Thunderstorms
6a	71-13		Dew Forecasting
6b	73- 6		Forecasting (White) Frosts
7	73- 7		Short Term Water Level Changes on the Great Lakes
8a	77- 8		Prog K Values vs. Thunderstorm and Precipitation Frequency
8b	77-10		1951 Floods and Heavy Rains in Kansas
8c	77-11		Climatological Moisture Data
8d	77-17		Kansas City Flood of Sept. 12-13, 1977

CRH SSD
November 1967
LAH

TECHNICAL ATTACHMENT B

On Lee Cyclogenesis

Three major systems crossed the Rockies into the Central Region in the last eight days of the past October. While there were major differences among the systems as they moved well into the region, they all followed rather well the simple model discussed below. The steps in the model are as follows:

1. A front enters the region from the northwest with a moderate intensity sea-level Low in Wyoming or Montana.

2. The short wave trough at 500mb with this front shifts the winds over the Rockies from northwest (with only a small component perpendicular to the mountains) to west or WSW (almost perpendicular to the mountains) with the jet axis at 500mb crossing the Colorado Rockies.

3. The sea-level Low starts elongating southeastward into the warm air into southeast Colorado when the 500mb wind shifts to nearly perpendicular to the Colorado Range. The vorticity advection aloft at this time is near zero.

4. Lee cyclogenesis continues to occur in southeast Colorado, with the winds aloft remaining nearly perpendicular to the mountains, as the PVA area at 500mb becomes superimposed on the mountain effect.

5. The Low initially is very warm (unusually high 1000-500mb thickness) and dry (low humidity, little clouds and rain) due to the downslope lee effect, but the thickness decreases and the humidity, clouds, and rain increase as the PVA area initiates upward motion.

6. The sea-level Low moves away from the mountains with the PVA area aloft, frequently weakening or elongating as the PVA area and the mountain effect disassociate.

7. Subsequent events depend on the movement and intensity of the PVA area aloft.

Vorticity advection at 500mb (PVA) is usually the prime factor in sea-level cyclogenesis in the Petterssen development equation (1). However, when the 500mb jet axis crosses nearly perpendicular to the Colorado Rockies, the "lee Low" effect is so strong and so concentrated that it can overcome the effect of vorticity advection aloft, or at least strongly modify the vorticity advection indications. The lee effect is very strong because the mountains-to-plains height change is very large, and the mountains have sufficient lateral extent so the air goes over instead of around them. The effect is concentrated because with the jet axis over Colorado, both the wind and the average height of the mountains decreases north and south of Colorado.

Generally, cyclogenesis takes place with upward motion due to low-level convergence, and as a result much cloudiness and precipitation occur. However, with the "lee Low" the Low is formed as a result of a concentration of strong downward motion giving in effect a heat Low from subsiding air. Such a condition is obviously not conducive to much cloudiness or precipitation, and these systems were quite dry in their formative states as seen on the mean relative humidity charts. For those familiar with the Petterssen development equation, the term he discusses as a brake on development (the Laplacian of the product of vertical motion and stability) turns into an accelerator when downward motion ensues.

The center of the Low had unusually high thicknesses associated with them, as a consequence of this heating effect of downward motion; and later on, when the system was away from the mountain and the PVA became the dominant vertical motion term (producing upward motion), the thickness at the center of the Low was reduced without the usual means of occlusion.

When the PVA became dominant, the vertical motion was mainly upward and clouds and weather were initiated over a much larger area. The subsequent motion of the Low center was tied to the PVA area at 500mb. In the first two of these three cases, the upper trough moved rapidly through the Great Plains and pulled the Low away from the mountains, deforming and weakening it at first as the mountain effect diminished, then intensifying it in the Great Lakes region. The last Low went southeastward as the upper trough closed into a Low aloft over Texas. Since the upper trough stopped in that area, so did the sea-level Low.

Lee Lows sometimes are created in the Montana-Wyoming border area if the strong 500mb winds cross the mountains in that area from around the WSW. It is likely that the three main areas of cyclogenesis in western North America, as given by Petterssen (1, pp 267 and 269), are at least partly because of these lee effects. The three areas are Colorado, Nevada, and Alberta. The Alberta range is as effective a barrier for producing these Lows as the Colorado range, but the 500mb flow doesn't usually cross this area with the stronger winds found in 500mb troughs (the ridges have weaker winds) except in summer. In summer, the lee effect discussed above can be very prominent and it produces the very large and very intense summer Lows discussed by Hage (2).

In summary, we can say that lee cyclogenesis can be quite strong in the Colorado area (moderate in the Wyoming-Montana area) when the 500mb axis of strongest winds shifts from nearly parallel to the line of mountains to about perpendicular. The resulting Low is warm (high thickness 1000-500mb) and with low humidity. The Low moves away from the mountains mainly under the influence of PVA areas aloft (a pattern of warm-cold thickness advection can move it also), and usually stays with the PVA area. As the Low first leaves the mountains it usually changes into an irregular shape, since two creative mechanisms are acting in different but nearby places. After the mountain effect is gone, the Low may reintensify with the PVA area as the sole mechanism. This mechanism changes the Low into a cloudy, rainy one as upward motion becomes dominant.

1. Petterssen, Sverre, Weather Analysis and Forecasting, 2nd Ed., Vol. I McGraw-Hill, 1956.

2. Hage, K. D., On Summer Cyclogenesis in the Lee of the Rocky Mountains, Bull. of the AMS, January 1961.

CRH-SSD
February 1975
LAH

TECHNICAL ATTACHMENT 75-3

THE STORM OF JANUARY 10-11, 1975--WHY SO INTENSE?

The Petterssen Development Equation soon will have been discussed in detail at all WSFOs in the Central Region. The storm of January 10-11, 1975 is an excellent example of the use of this equation, not only for the final development, but for intermediate stages as well, and, as will be seen, these intermediate stages play a large role in the ultimate strength of the main storm.

As was brought out in the talk in the equation, the beginning of a cold season Great Plains cyclone is frequently with a lee Low that forms in Montana as the advancing 500 mb ridge brings the strongest wind at that level, nearly perpendicular to the Rockies in Montana (see Figures 1a and 1b). This surface Low quickly intensified, doubling its vorticity to a value of F (20×10^{-5} sec^{-1})* on the Sangster chart. However, with warm advection ahead of the Low and no cold advection behind it, the Low should pull away from the mountain and weaken. That this occurred is seen by Figure 2 as the surface chart 24 hours later. If a short wave trough had moved over this Low, it could have survived as a Low. However, the deepening of the 500 mb trough pulled the strength of the trough aloft away from this lee Low, which lost its identity as a Low and slowed its eastward progression as the warm advection weakened. However, it continued as a vorticity center--a factor of significance later, as will be seen.

If the arctic air had been able to come down, so that it could create cold advection west of the Low to compensate the warm advection to the east, the Low probably could have survived. But the arctic air did not yet have significant 500 mb flow over it, so it stayed in its source region (a general treatise on plunging arctic air is planned in another Technical Attachment to appear next time).

As was also stated in the model of these developments, the lee Low reforms where the first one moves away from the mountains. Since the strongest combination of wind and terrain had shifted the strongest lee effect to Colorado as a result of the deepening 500 mb trough, the lee Low formed there. It formed quickly, and quickly (by 06Z of the 9th), became quite strong (Sangster value of F+ or 23 units) because the winds at 500 mb were strong.

As the Colorado lee Low got warm advection ahead of it and the strongest westerlies at 500 mb moved on south of Colorado, this lee Low also moved eastward toward the warm advection and weakened, as still another lee Low formed in the panhandles region with a value of F (about 20 units) near 00Z on the 10th. The sea-level picture at this time is given in Figure 3, where the Montana lee Low has lost its identity, but its vorticity max still existed in the extreme eastern Dakotas; the Colorado lee Low was in northeast Kansas, and the main lee Low was now in the panhandles. The prologue was now over, and the stage set, as PVA was about to take over.

*See TPB No. 111

The strength of the lee Low is a significant factor in the ultimate intensity of the final surface Low. This is because the Petterssen equation is a measure of the amount of intensification, whereas the final intensity of a Low is a sum of its intensification and the initial intensity. Putting it another way, the Petterssen equation is a measure of the lower-level convergence, but the final amount of spin (vorticity) produced by the convergence depends on how much spin there is initially. This is fully analogous to the spinning ice-skater whose final spin speed depends on the amount of convergence of arms and legs and the amount of initial spin.

The vorticity advection, which should be and was the ultimate intensifier, was slow to organize and come out of the deepening upper trough, probably as there was no kicker upstream, i.e. no vorticity maximum upstream to shorten the upstream wavelength. The vorticity center went unusually far south, but by 00Z on the 10th the positive vorticity advection (PVA) at 500 mb had reached the panhandles sea-level Low (Figure 4). The PVA was at least moderate as judged from the following: 1. the trough was sharp, which tends to concentrate instead of spread out the PVA, 2. the 500 mb winds were moderate, about 65 knots, in the center of the PVA area, 3. the vorticity maximum was high, about 22 units on the LFM, which with its smaller grid size should give the best measure of vorticity.

Using 00Z data of the 10th, when the main show was about to start, there were three keys to events to come: 1. the surface and 500 mb vorticity centers had or soon would be at the longitude where they normally turn more northward (100°W), 2. the baroclinic and barotropic charts gave quite different 500 mb positions of the vorticity maximum on their 24-hour prog, with the PE and LFM over Ft. Worth, and the barotropic near Kansas City. But it has been stated many times before that the baroclinic is the best model as the upper trough deepens down in, and it was, while the barotropic is best when the trough is on the way out, and it was quite good this time.* 3. The strongest winds at 500 mb were still well upstream of the PVA area, but as the vorticity center takes on some northward component of movement, the strongest winds round the bottom of the trough and usually spread east of the trough. This strong increase in winds in the PVA area (there were over 100 knots winds upstream over Utah) would increase the PVA quite a bit, and this should take place around 12Z.

The surface vorticity increased mainly from about 06Z to 18Z on the 10th, when it increased from about 20 units to about 28 units. It continued to increase to 31 units in the 24 hours following the main development. While the strong PVA acting on strong initial vorticity started the development, and could have brought off a good storm by itself, several additional factors added to the development.

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*It is customary when the barotropic model takes a vorticity max northeastward correctly as it comes out of the southwest United States, for the trough to trail off southward way too much. The usual correction in these cases is to start the trough at the vorticity max and run it roughly SSEwd so it is about perpendicular to the 500 mb flow. This change concentrates the vorticity advection and moves it much farther along. Such a correction was applicable and correct here.

One was that as the PVA area aloft moved, it continued to pick up other vorticity to concentrate, as provided by the earlier lee Lows. This vorticity path for the developing Low acted much as an ionized path does lightning. The convergence produced in the low levels by the PVA aloft thus continued to bring in air with high vorticity, so the convergence was all the more effective. Another factor was the low stability in and around the developing Low. Low stability makes it easier for upward motion to come about, as there is little resistence to such motion because of buoyancy. At 0030Z, just before the main development started, the thunderstorms were all around the panhandles Low (see RADU chart, Figure 5, where the position of the surface Low is given by an X). But by 0630Z, when the development really got started, the thunderstorms were fairly well centered on the Low which was in extreme northeastern Oklahoma, with a strong line of thunderstorms close to the center of the Low. According to Tracton (Monthly Weather Review July 1973) and from Petterssen's equation, having the lowest stability centered on the Low will aid development.

The final factor in the intensity was the plunge of the arctic air, which by pushing into the west side of the Low, created stronger cold advection behind the Low than existed ahead of it, which further augments development. Thus, everything that could be favorable for development was favorable, and to a strong degree.

1. Strong 500 mb winds crossed the Rockies at various places, creating a series of strong lee Lows, all of which added to the final storm intensity (see 3. below).

2. The same strong 500 mb winds gave strong vorticity advection which eventually acted on the final and strong lee Low.

3. The strong vorticity advection continued to act on air which had high vorticity as a result of earlier lee Low activity over Montana and southward.

4. Stability was very low, and thunderstorms were centered on the Low.

5. The plunge of arctic air caused a symmetry of the thickness advection pattern around the developing Low such that cold advection predominated.

Figure 1a - Sea-level with 1000-500 mb Thickness, 12Z Jan. 8, 1975

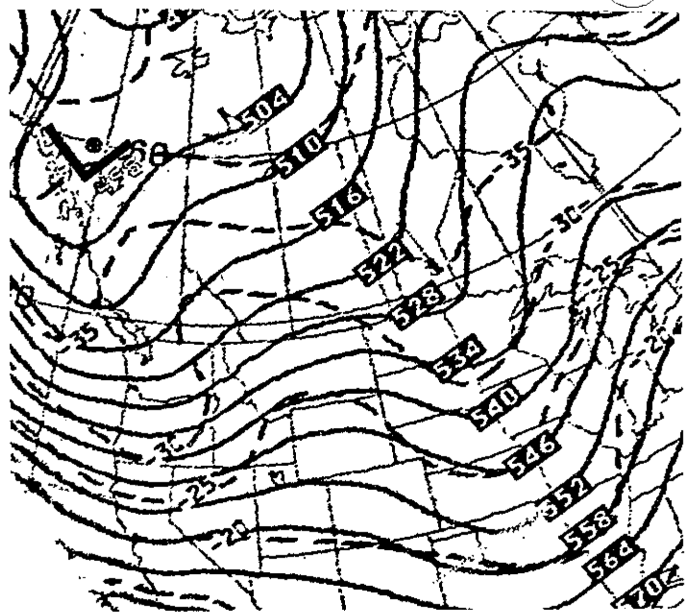


Figure 1b - 500 mb Contours and Isotherms, 12Z Jan. 8, 1975

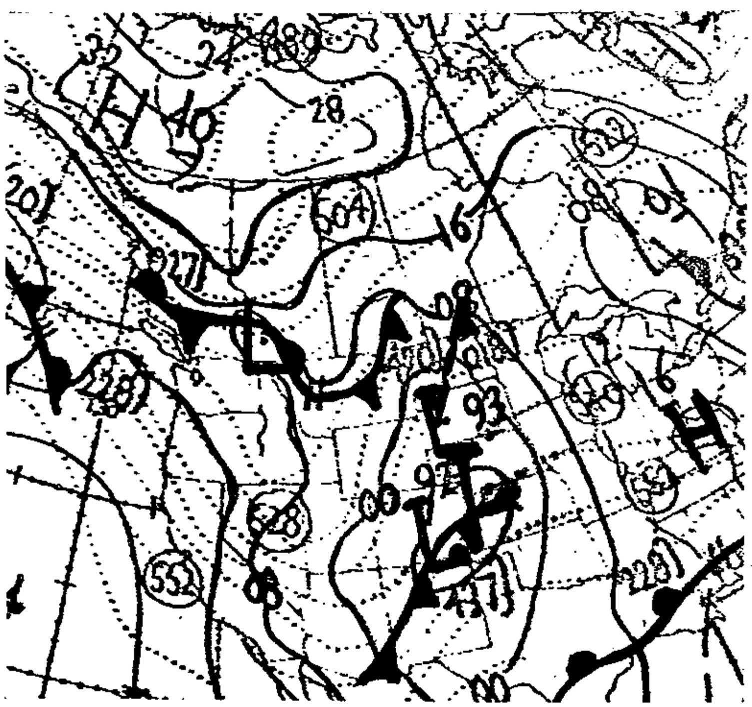


Figure 2. - Same as Figure 1 but 12Z Jan. 9, 1975

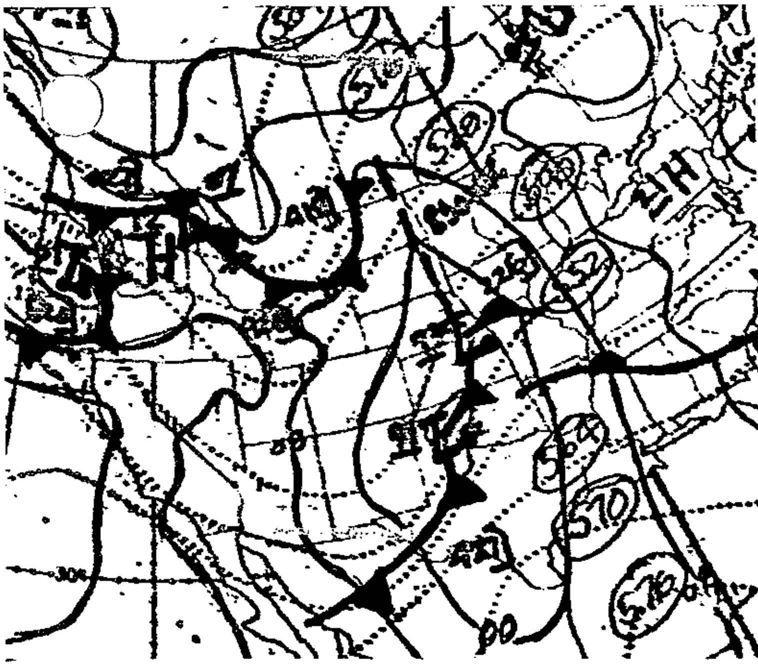


Figure 3. Same as Figure 1 but 00Z Jan. 10, 1975

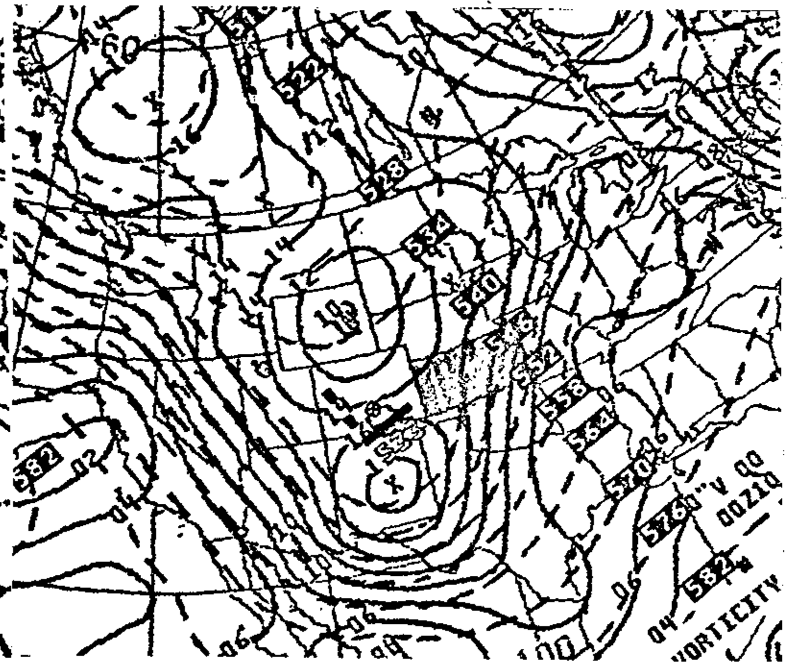


Figure 4. 500 Initial Vorticity and Contours, 00Z Jan. 10, 1975

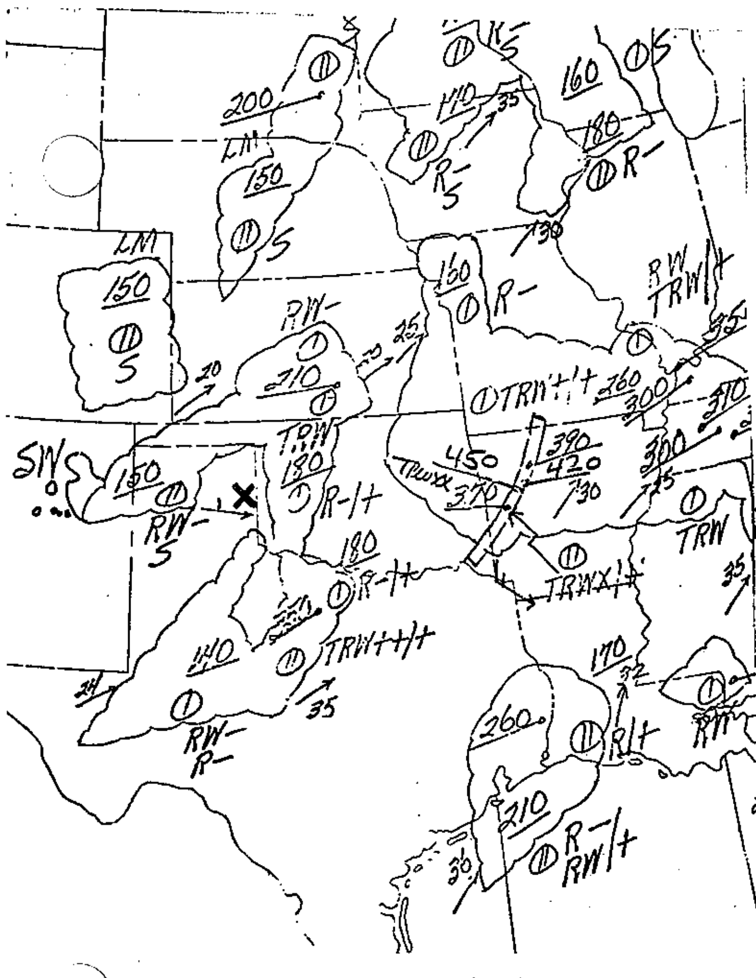


Figure 5. RADU Chart 0030Z Jan. 10, 1975

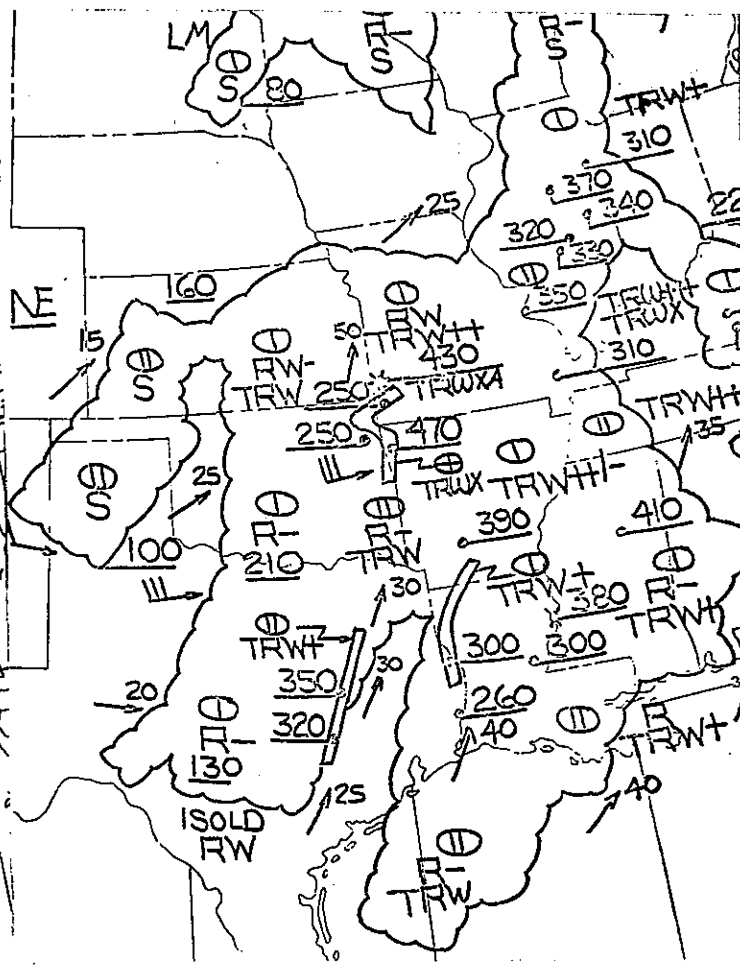


Figure 6. RADU Chart, 0630Z Jan. 10, 1975

TECHNICAL ATTACHMENT 75-9

MERGER OF SURFACE VORTICITY MAXIMA PORTENDS A MAJOR STORM

A recent series of the surface geostrophic wind and vorticity (Sangster) charts is interesting to view in relation to the development of a major surface low in the Mississippi Valley. At 18Z March 31, 1975 a surface geostrophic vorticity maximum was over extreme sourthern Wyoming (associated with an upper trough) and another maximum (associated with lee effects) was near Amarillo (see attached Figure). In the next 24 hours the Colorado maximum moved only slowly to southwestern Colorado. Meanwhile the lee maximum moved eastward to near Springfield, Missouri and became stationary there probably due to lack of upper-level vorticity advection or thermal advection.

Then the western maximum moved out of the mountains along a track nearly the same as the other vorticity max while the eastern maximum remained in southwest Missouri. 09Z April 2 is the last map to show two maxima, as their merger was completed at 12Z April 2.

By 18Z April 12, the single vorticity maximum had increased from an "E" value to an "F" value and moved eastward to near St. Louis. A further increase from "F" to "G" occurred as the vorticity max, now associated with a developing surface Low, moved to east of Indianapolis.

This merging of lesser vorticity maxima to form a major storm is probably fairly common, (see CR TA 75-3 on the January blizzard) and tracking of these maxima at critical times is a recommended forecast procedure.

But why did the surface Low development wait until the two vorticity centers had merged in Missouri, rather than taking place as a Colorado development? The Petterssen Development equation should provide an explanation. According to Petterssen*, "Cyclone Development at sea level occurs when and where an area of appreciable vorticity advection in the upper troposphere becomes superimposed upon a slowly moving or quasi-stationary front at sea level". The first vorticity center (produced by lee effects) did not have vorticity advection with it, and development did not occur. The second did, but the main PVA was north of the surface vorticity max and it moved over the surface ridge, when it first moved into the Plains, where development of a surface Low is not likely for several reasons:

1. The PVA tends to be counteracted by the Laplacian of the (cold) temperature advection (this comes directly from the Petterssen equation).

2. The great stability in the low levels may have forced any PVA-created convergence to be at the top of the cold dome instead of near the earth's surface. Since precipitation broke out in the cold ridge as the PVA advanced over it (amounts were light) some upward motion must have been occurring.

*Weather Analysis and Forecasting, 1956 - Volume 1, p. 337 (see also p. 334-339).

3. The surface vorticity is always low in a ridge, so even if there was some low-level surface convergence, there was little rotation to concentrate (remember the ice-skater analogy).

But when the PVA aloft reached the front, so it had no counterbalance, the development took place. The development was sudden and strong because:

1. The PVA was strong (see discussion below)

2. The Laplacian of cold advection was strong, and the speed of the 500 mb trough and PVA area was moderate, while the surface front was stationary, so the change from the advection counterbalancing the PVA was sudden and large.

3. There was plenty of low-level vorticity to concentrate, with the two vorticity maxima coming together.

4. The stability was low, although the Laplacian of the stability was not favorable.

The PVA aloft strengthened in the 12 hours preceeding surface development. This commonly occurs for the following reason. A stationary or slowly moving 500 mb trough generally has the characteristics of a long-wave trough (shortwave troughs can't remain stationary), and long-wave troughs have little PVA as the vorticity changes produced by the curvature of the flow are opposite to and about equal to those produced by changes in the earth's vorticity as one follows a fluid particle. But as a shortwave trough moves toward such a trough from the upstream direction it changes the character of the flow of the quasi-stationary trough by shortening its wavelength on the west side and increasing its curvature changes until eventually the long-wave trough becomes a large short-wave trough with good PVA and it moves faster. This effect is quite obvious on the 500 mb charts of 00Z and 12Z on April 3.

The forecaster making the early morning forecast on April 2 (upper air data from 00Z and surface data to 08Z or 09Z should think as follows:

1. The upper trough is coming out of the southwest faster than its previous speed (the numerical models show this).

2. Its vorticity advection should increase as it comes out (experience and numerical models show this).

3. The PVA area aloft will be over the surface ridge at first, so any development to occur should be well east of the last two big storms. The front and the quasi-stationary vorticity max at the surface in Missouri suggest this area for the beginning of development.

4. The movement of the second surface vorticity max on the Sangster chart

suggested merging of the two maxima at 12-15Z, and, since the second max was with the PVA area aloft, that the PVA area would be over the surface vorticity centers and the front at that time and therefore development should commence then and there.

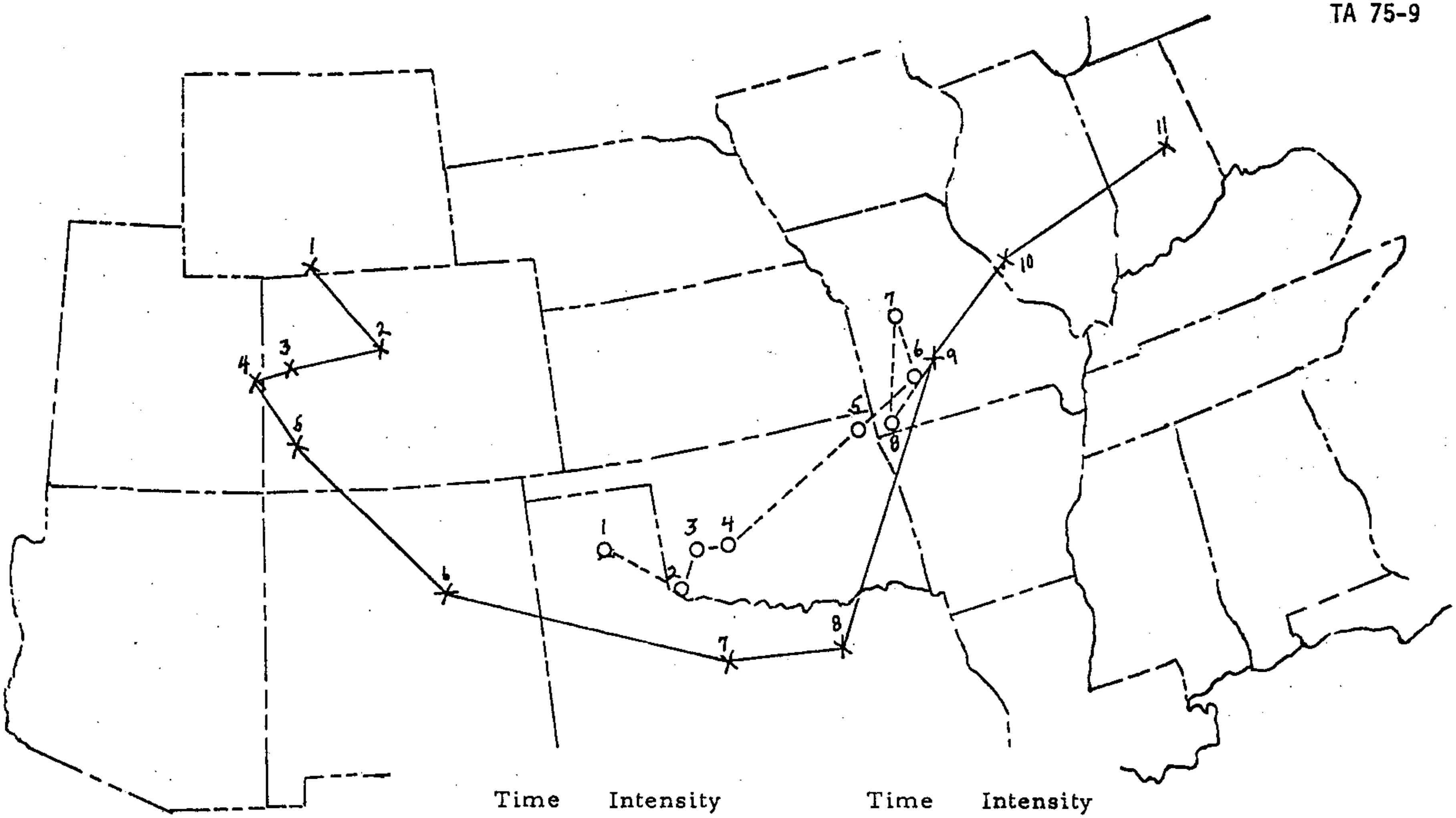
5. The 24-hour forecast PoFP value of 50% for 00Z April 3, considered the southern edge of any snow deposition (see CR Tech. Attachments 73-22 and 73-24), went from the southeast tip of Iowa eastnortheastward to northeast Ohio. The center of any heavy snow band can be considered as a bit north of this line.

6. The PE, LFM, and subjective forecasts of the surface all predicted development of a surface Low, with the subjective having the farthest north position and the PE the farthest south (some typical locked-in PE error?).

7. This would be a "Panhandle Hook" type of development (see CR TM 38) which tends to have higher than average water content, and the 00Z precipitable water was moderate and should increase a lot with the development. TM 38 says that in "early spring these storms can be particularly disastrous to Milwaukee and the Chicago metropolitan areas" due to the high water content of their snowfall.

8. In view of the above, the <u>least</u> alerting should be a Winter Storm Watch for this afternoon and tonight and a heavy snow warning in the PoFP band of 60-90% or so ahead of the developing Low for this afternoon and tonight would be very reasonable.

9. A careful weather watch, including the later movement of the surface vorticity center, should then be instituted.



	Time	Intensity X	Intensity O		Time	Intensity X	Intensity O
1	31/18Z	E	F	7	02/06Z	E	E
2	01/00Z	E	E	8	02/09Z	E	E
3	01/06Z	E+	E+	9	02/12Z	E	-
4	01/12Z	F	E+	10	02/18Z	F	-
5	01/18Z	E+	E+	11	03/00Z	G+	-
6	02/00Z	E	F				

TRACKS OF SURFACE ABSOLUTE GEOSTROPHIC VORTICITY MAXIMA - March-April 1975

CRH SSD
July 1971
LAH

Technical Attachment 71-14

Hage Lows

Hage lows are intense sea-level Lows forming in southern Canada in the summer (June-Aug.) (see Hage, AMS Bull., January 1961). This note indicates where and when to expect such developments, and what effect they will have on weather systems in the Central Region.

The sea-level development is brought about by the northeastward acceleration of a 500 mb Low from a position just off the west coast of mainland U.S. The sea-level Low develops just east of the Alberta range in the cold air north of any frontal system. Usually there is a pre-existing lee trough at sea-level in the area of expected development, and the development concentrates this circulation as the positive vorticity advection at 500 mb becomes superimposed on it.

All upper-level Lows that come inland around the Canadian border in summer have some sea-level development, but the more of the following criteria that are favorable, the greater the probability that intense development (sea-level pressure less than 990 mb) will occur. Favorable factors:

1. The minimum 500 mb temperature in the Low is less than -25°C and stays that cold as the Low comes inland.

2. 500 mb winds southeast of the 500 mb Low are 50 knots or more and increase as the Low moves inland.

3. A 500 mb trough in mid-Pacific is moving fast enough that in two to three days it has or is forecast to greatly reduce the distance between it and the Low off the west coast.

4. The 500 mb Low intensifies as it moves inland.

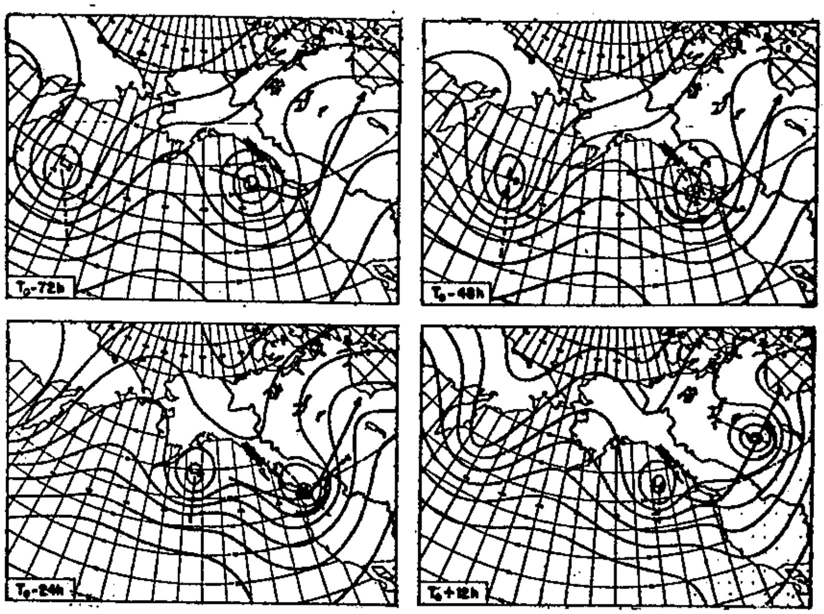
5. The axis of strongest 500 mb winds and the PVA aloft move across the highest part of the Alberta range (southern Alberta).

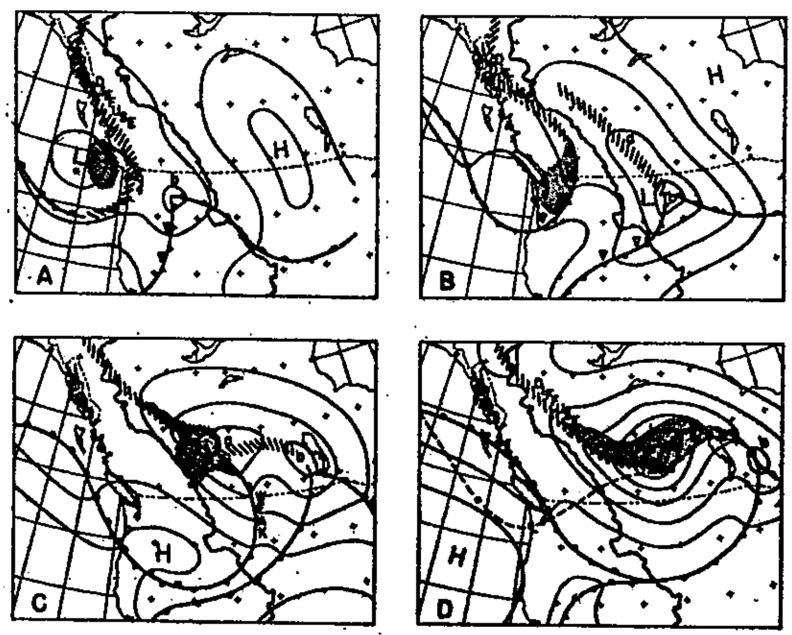
6. The sea-level lee trough in Alberta is pronounced before development.

The figures herein show schematically the events at the surface and aloft. Note that the frontal Low undergoes some development first but then weakens as the intense development occurs farther north and in the cold air. Don't be misled by the early frontal Low development, as this is not the main Low, and it will soon weaken if the PVA aloft will cross the Alberta range north of it.

The main effect in the Central Region is that the intense secondary development reduces the push of cold air behind the front so that one must delay the push, and associated precipitation, except perhaps in the Dakotas, until well after the intense development. In timing the development, the maximum intensification of the sea-level Low should occur when the center of PVA aloft is over or just east of the Alberta range.

One should note in predicting this event using numerical progs, that it is a characteristic bias of the barotropic model to move quasi-stationary 500 mb Lows inland from the west coast at a time when they are still remaining off the coast. The baroclinic model doesn't have this bias, and thus should be best at first. However, once the Low is moving inland, as can be anticipated when the diminishing distance upstream to the next upper trough indicates that acceleration inland is likely, then the barotropic may give a better speed of movement. Watch this point this summer.





Note - see case of July 7, 1971

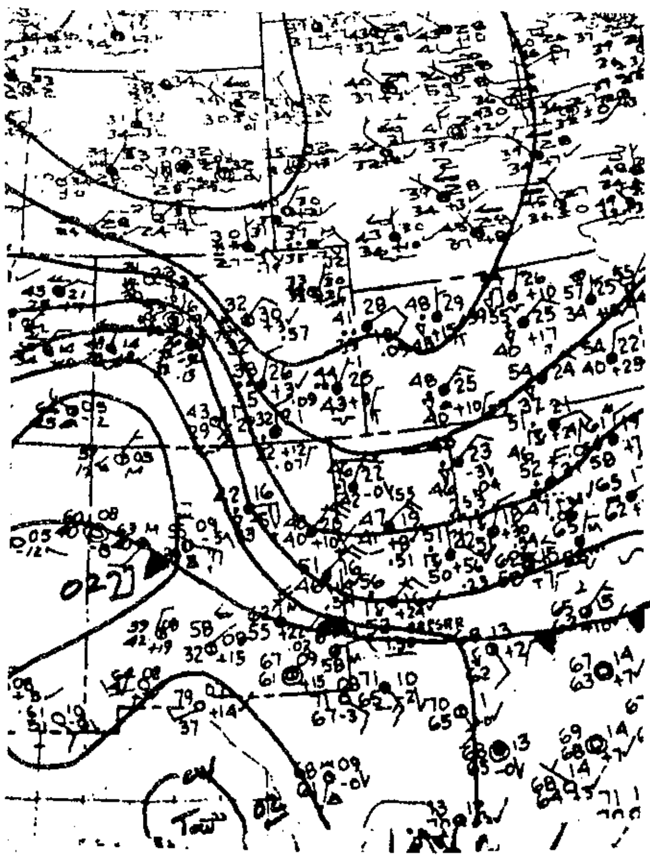
CRH SSD
October 1971
LAH

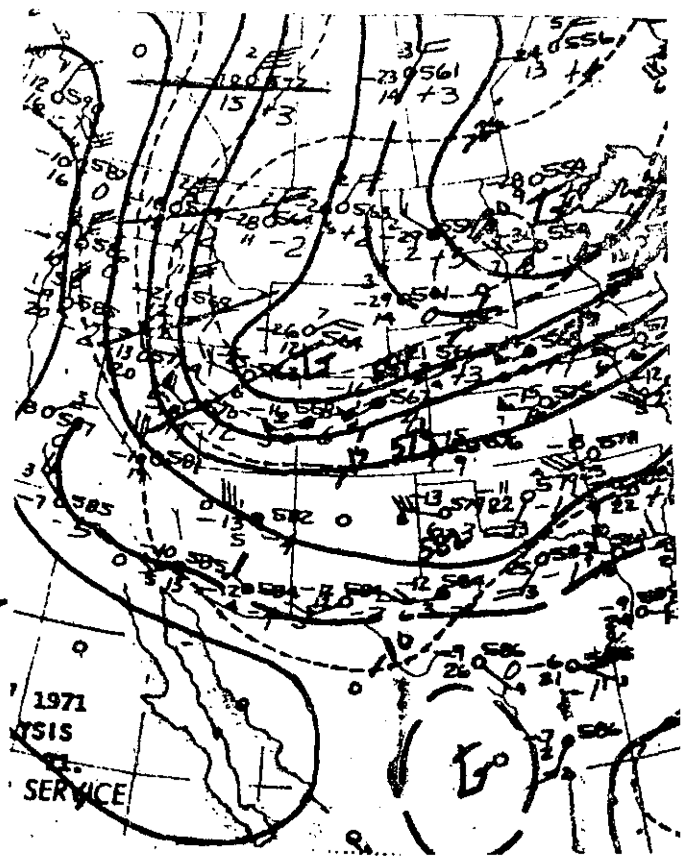
Technical Attachment 71-18

Some Rules for Frontal Wave Formation and Movement

In shifting to the cold season there is sometimes a clashing of gears instead of a smooth shift. Even those with a lot of experience still have trouble. The following discusses one of two similar situations that caused some clashing of gears by forecasters recently. They involved frontal wave development centered around September 18 and 22. We realize that case studies are hard to read, so we have treated only the earlier case, and we have kept the proof (reproductions of maps, etc.) meager, with the thought that those without great recall of events may want to look back through the facsimile maps. Only the initial maps are given to start off your thinking. Our intent is to bring certain points to the forefront for later use (they are listed at the end).

The facsimile surface analysis at 12Z September 17 (below) showed a slowly moving cold front and a southeastward push of a cold ridge into the Plains. The precipitation in the central Plains gave the usual problem of whether or not it was all caused by upslope low-level flow. At 500 mb there was a small sharp trough over Utah, and the past positions at 12-hour intervals show its recent movement (movement also borne out by 500 mb vorticity analyses).





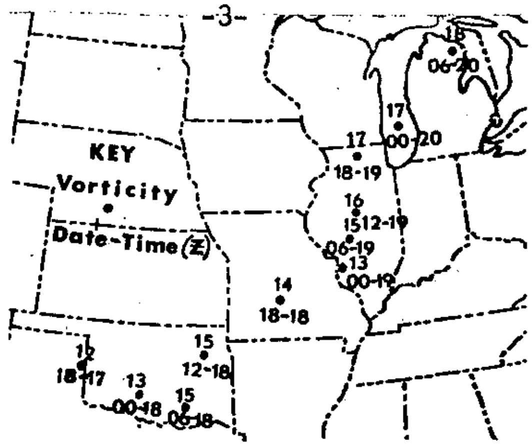
PROBLEM - What effect will the small upper trough over Utah have on the weather as it continues to rotate around the large scale trough and moves out over the Plains?

The answer to the question was clouded by the fact that the barotropic and baroclinic progs were quite unclear on the future of the small trough. This should not be surprising nor lead one to believe that the trough (or vorticity max) disappeared (they usually continue to rotate through the major trough). One must realize that: (1) the wavelength of the trough of concern was very small, but that it appeared prominent because it was near the center of the larger scale trough around which it was rotating. The same moving vorticity max would appear much less prominent near the center of a major ridge; (2) the PE and Barotropic progs cannot resolve small systems well, especially the baroclinic (hopefully the LFM will help here); and (3) the strongest winds at 500 mb over the Plains were well north of the surface position of the cold front, and over the surface ridge, so the small system could not create much of a frontal wave, at least at first. NMC subjective progs seemed to take no account of the small system, but the FOUS data gave a good hint of events, with precipitation forecast for Topeka in the 6-18 hour period after the above initial conditions.

The first good analysis clue to the coming sequence of events probably came in the surface vorticity chart of Sangster (available only at MKC). At 18Z (the next chart after the times shown above, since charts are made only every 6 hours), a completely new, but very well-defined SURFACE vorticity maximum appeared centered over the Texas panhandle. The clarity of this new event, in a climatologically favored region for surface cyclogenesis in response to the eastward movement of upper troughs, should tip one off that the small trough was probably still alive and active. The surface system was only on the threshold of detectability on the fax surface analysis.

The weakness of the surface system on the fax chart should be expected, because the upper system was moving out over the cold air and over initial surface anticyclonic relative vorticity (the system showed up better after it moved a bit east of the initial surface ridge line). There are two reasons why the Sangster chart showed the system so clearly. One, the surface vorticity accentuates small surface systems in the same way the 500 mb vorticity accentuates small 500 mb systems. Two, the 12-hour averaging and perhaps other aspects of the pressure reduction to sea-level tend to smooth out small systems, while there is no time averaging or significant pressure reduction over flat terrain in the Sangster chart. The surface system in the September 21-22 case was much clearer as the upper system came over the surface front and a frontal wave and surface Low were formed.

The surface system was very easily seen and followed on successive charts of surface vorticity, and it was a significant weather producer all the way (see track of surface vorticity maximum on next page). However, it wasn't until 00Z on the 19th (a day and a half from the initial map above) that a sea-level Low appeared on the fax map. The weather knew that the small system was alive and active, as could be seen easily after the fact from the fax 24-hour precipitation charts.



The above events are common in the cold season, and they occur in varying intensities. When the systems are weak, and especially when the upper system moves out over the cold air well north of the front, they are hard to follow via a model-of-the-day and thus are hard to forecast correctly. Let us list the points to keep in mind.

1. Short wavelength vorticity maxima at 500 mb are frequently poorly handled by the barotropic and baroclinic progs, and especially the latter, possibly because of the large grid size in the model. Hopefully, the LFM will do better.

SOLUTION - Extrapolate small systems by hand, and watch the LFM.

2. 500 mb vorticity maxima that move across the Rockies into the Plains are almost always of significance in the weather patterns (rain or overcast in their PVA areas), almost regardless of their initial strength.

SOLUTION - Beware of weak circulation systems that are hard to follow in progs. They can be significant weather producers.

3. If weak PVA aloft moves into the Plains well north of a surface front and over a sea-level ridge, there may be no visible manifestation in the sea-level isobars, but weather with it is likely, especially if with it initially.

SOLUTION - Use the surface vorticity chart (available only at MKC) and/or watch the clouds and weather under the 500 mb PVA area.

4. If weak PVA aloft moves into the Plains over or near a surface front, expect wave formation and movement with the PVA area and the increase and movement of associated weather.

SOLUTION - Relax a bit as the situation will be fairly clear.

5. If strong PVA aloft moves across the Rockies, major cyclogenesis will almost certainly occur in the lee under the upper PVA area.

SOLUTION - Relax a lot and use numerical model output as they handle this quite well, especially the barotropic for the position of the upper system, and the baroclinic surface chart. Beware of the "locked-in" error of the PE model.

6. If the surface vorticity system associated with an upper PVA area moves away from the Rockies with some southward component, the minimum latitude is almost always near the longtitude shown in the track given above—middle Oklahoma (a bit farther west for the northern Plains).

SOLUTION - Use climatology; it still is useful.

CRH SSD
January 1973
LAH

Technical Attachment 73-2

GULF COAST WAVES

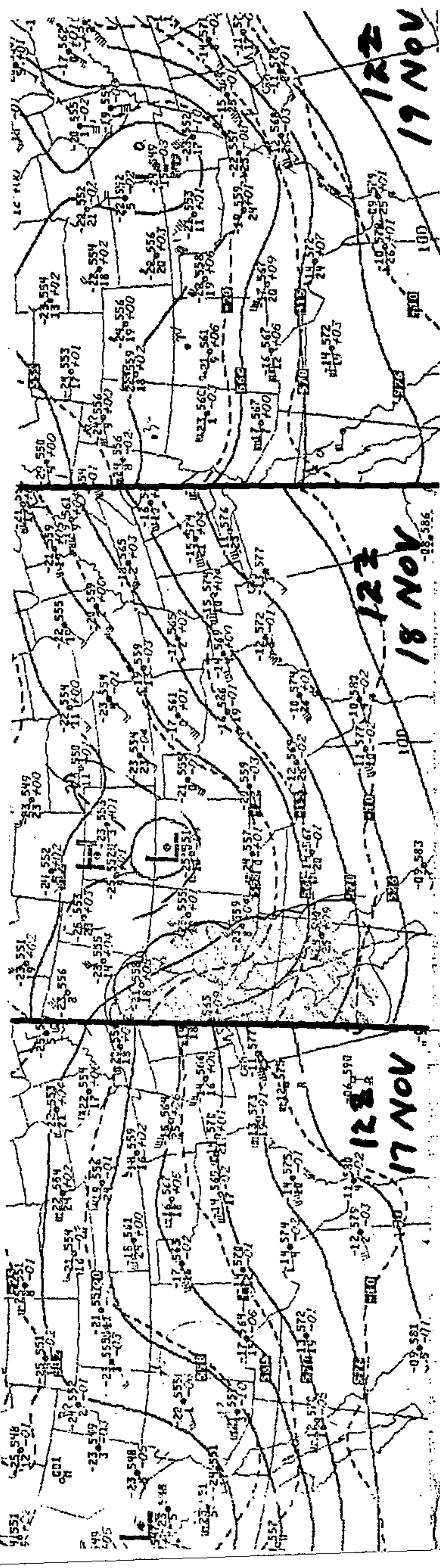
The waves of concern here are not in water, but on fronts, with the wave that formed November 17 and 18, 1972, as an example. The slow development on the 17th and early (local time) on the 18th (see maps attached) was associated with the heat flow into the cold air from the warm Gulf. The shape of the shore in the western Gulf is favorable for development by this heat addition according to a lesser used term in the Petterssen development equation. If the shoreline had been straight instead of curved, the early development would not have occurred. As long as there is cold air over the rather warm water of the western Gulf, and no upper air system moving across the area, the Gulf Low tends to slowly intensify and hold its position.

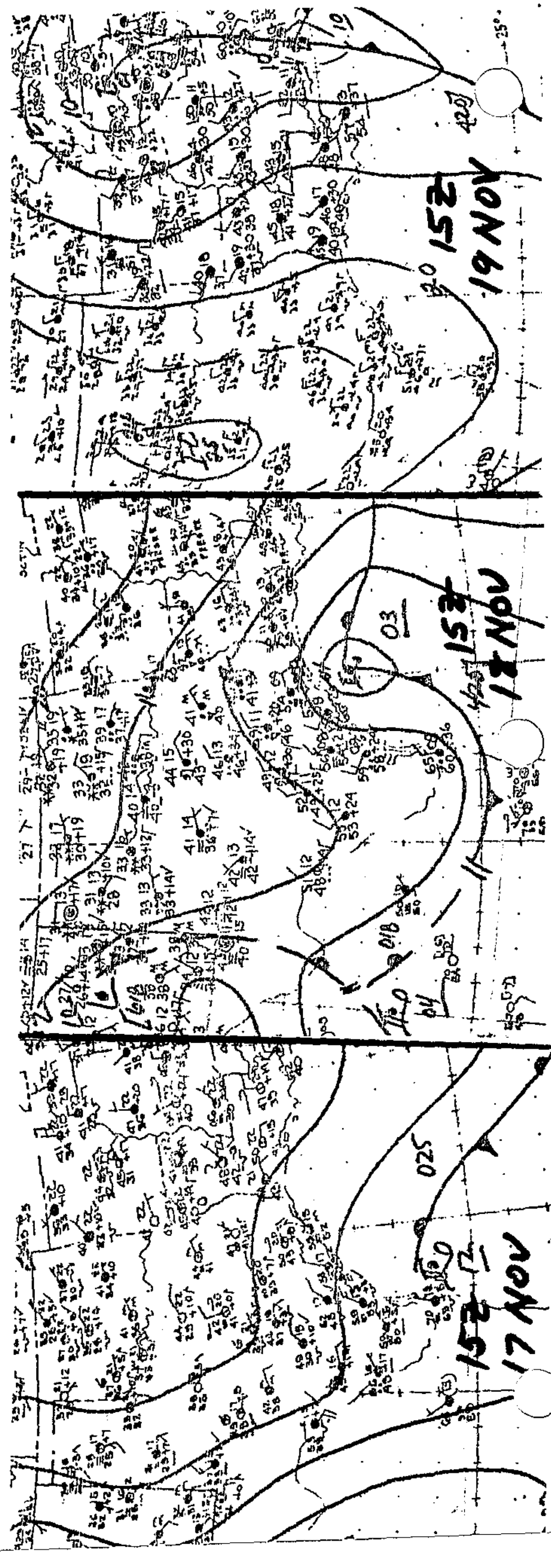
Similar developments occur in the Gulf of Alaska and near Hatteras. In the latter case, an old paper by Miller (JOM June 1946) showed that the developments were really initiated either side of Hatteras, rather than directly off Hatteras. From the Petterssen equation this could be because the shore has a favorable shape north and south of Hatteras, but an unfavorable one at Hatteras. The PE and LFM both have heat addition from water to air, so they could be expected to show at least some of this development and they did, with the LFM much the better surface prog as usual. A similar thing occurs in the Great Lakes area for the Lakes as a whole, but this is <u>not</u> considered in the PE-LFM equations, so the Great Lakes effects will have to be added by the forecaster.

The development late on the 18th and on the 19th, particularly the sudden northward jump of the Low (see maps) was due to PVA aloft, also in accordance with the Petterssen development equation. The development was also in accordance with the first example in the paper by Saucier on "Texas-West Gulf Cyclones", in the August 1949 Monthly Weather Review--a paper worth looking over in spite of its early date. Incidentally, the surface Low did not move continuously from New Mexico on the 18th to Tennessee on the 19th. Rather, the western Low died out and a new Low reformed in Louisiana, with anticyclonic flow over the southern Plains all the time. This <u>a</u>typical movement from the Rockies was probably a result of the presence of the Gulf wave. With warm air over the Gulf, the sea-level Low would probably have moved continuously across the southern Plains, probably through Oklahoma, as is more typical.

The importance of Gulf wave developments is that the wave tends to keep the surface low pressure system further south than otherwise, at least until it reaches the Mississippi River, thus for the Central Plains, affecting the surface winds, reducing the chance for precipitation especially of the larger amounts, but increasing the likelihood that precipitation would be snow.

In summary, the events are started by heat addition to the cold air by the warm water over the western Gulf because of the favorable shape of the shoreline. The PVA aloft moving into this area picks up the existing surface system in the Gulf rather than taking the main system across the southern Plains. The latitude of the PVA area as it reaches the eastern Texas border determines the extent of the northward elongation of the initial Low into the Gulf states and the northward extension of light snow into the Central Region.


12Z 19 NOV
12Z 18 NOV
12Z 17 NOV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CRH SSD
February 1973
LAH

Technical Attachment 73-4

REVERSE SLOPE SYSTEMS

An interesting situation occurred on an SSD visit early in this cold season in which a surface Low changed from a normal sloping system (upward to west) to a vertical or reverse sloping one. Several other cases have come to our attention since then, the latest being that of 00Z January 26, 1973, as the initial time. This type of system can give forecasters a hard time, as the first approximation is always for a normal slope. Let us look into the problem of the slope of pressure systems with height, so we can better judge the reasonableness of particular progs and not discard a good one or improperly adjust one which is suggesting the right answer.

It is required from hydrostatics that low pressure centers or troughs slope upward toward the coldest air. Thus, a 500 mb Low or trough is usually west or northwest of a surface Low because the circulation of the Lows brings cold air from northern latitudes in behind them. The amount of displacement from the surface Low to the Low or trough at some upper-level depends on two factors. One is the surface pressure gradient, such that the more intense the pressure (height) system the <u>less</u> the displacement. The other factor is the mean temperature gradient from the surface to the higher level, such that the stronger the temperature gradient, the <u>greater</u> the displacement.

There are several ways in which abnormal slopes can come about. Forecasters are well aware of the way which involves an old, well occluded, and previously intense surface system in which the cold air has gone completely around the Low so that there is little temperature gradient left, while the still moderate intensity of the system tends to hold it nearly vertical.

Another possibility involves an initially weak but normally positioned temperature field, but with the cold pool over the western half of the contiguous 48 states. As this pool moves downslope in reaching east of the Great Plains, the subsidence warming can reduce the cold pool sufficiently so as to eliminate it, thus causing the pressure system to become vertical. This may have been the case with the January situation mentioned above.

A third possibility is even more a Central Region problem. It involves a weak to moderate surface Low which is moving toward a strong cold ridge ahead of it. In such a case, the thermal gradient with the warm front ahead of the Low is stronger than that of the cold front with the Low.

A case like the third possibility occurred in early October during the SSD station visitation mentioned in the beginning. There was a surface Low over central South Dakota at 12Z on 2 October, and the thermal field is best shown by the 850 mb chart of Figure 1 (the 1000-500 mb thickness was missing for this time). Note the strong thermal trough over the eastern U. S. compared with the weaker one over Montana. A moderate 500 mb height trough over Montana was expected to pick up the Dakota surface Low and take it on eastward with it while, according to NMC subjective surface progs, maintaining a normal slope. In reality, the surface Low became nearly vertical with the upper system and even was a bit west of it after a while.

The positions of the surface low center and the 500 mb trough or Low center are given in Figure 2, along with the vorticity values of the surface system in units of 10^{-6} near the surface positions, as obtained from Sangster's surface chart program.

One interesting point for this situation is that the LFM handled the system quite well, and its surface prognoses was better than the NMC subjective one for this area and time in that the human forecaster maintained a normal slope and thus moved the surface Low too fast. In the January case mentioned above, the NMC forecaster correctly modified the LFM to remove the reverse slope and make the system vertical. Our experience suggests that the LFM should be watched with this type system, and when it forecasts a vertical or reverse slope, a nearly vertical system at the position of the 500 mb Low, as in the January case, is probably the best solution. The numerical model solutions are always hydrostatically consistent, but with weak surface systems and thermal fields, small errors in the models can cause forecasts of reverse slopes where the more usual essentially vertical conditions are verified.

Once this sequence of events is expected, additional related information becomes pertinent. As the upper system approaches the surface system, the PVA aloft tends to intensify the surface Low as it passes over it, but once past the surface Low, the Low tends to weaken again. This sequence can be seen in the surface vorticity values given by the positions of the Low center in Figure 2. Thus, knowledge of the relative motion of the systems at two levels also yields information about the intensity changes of the surface system.

CONCLUSIONS

1. When surface pressure systems are weak, and their associated thermal fields are also weak, the slope of the surface system with height can easily change to vertical or even to negative, i.e., the surface center west of the upper center.

2. Under the above conditions, the LFM may well give the best clues (the lack of resolution in the PE is such that the location of weak centers is frequently not given). Usually the system does not have a large negative displacement, but rather it can easily become vertical, in which case the surface Low should be very nearly under the prognostic position of the 500 mb Low (or vorticity center if no Low exists).

3. As the system becomes vertical, the surface system may intensify a bit as the PVA aloft comes over the surface Low, but the Low weakens again as the PVA aloft passes it.

4. Examination of the thermal field on the 1000-500 mb thickness or the 850 mb chart will indicate those cases where the thermal field is weak enough or of the right type to allow the system to become vertical. These are:

a. Old occluded surface Low with weak thermal conditions by Low.
b. Very weak thermal field.
c. Weak to moderate thermal field, with cold pool over the Plains moving eastward to lower elevation.
d. Weak to moderate thermal field, with stronger thermal trough to east than to west.

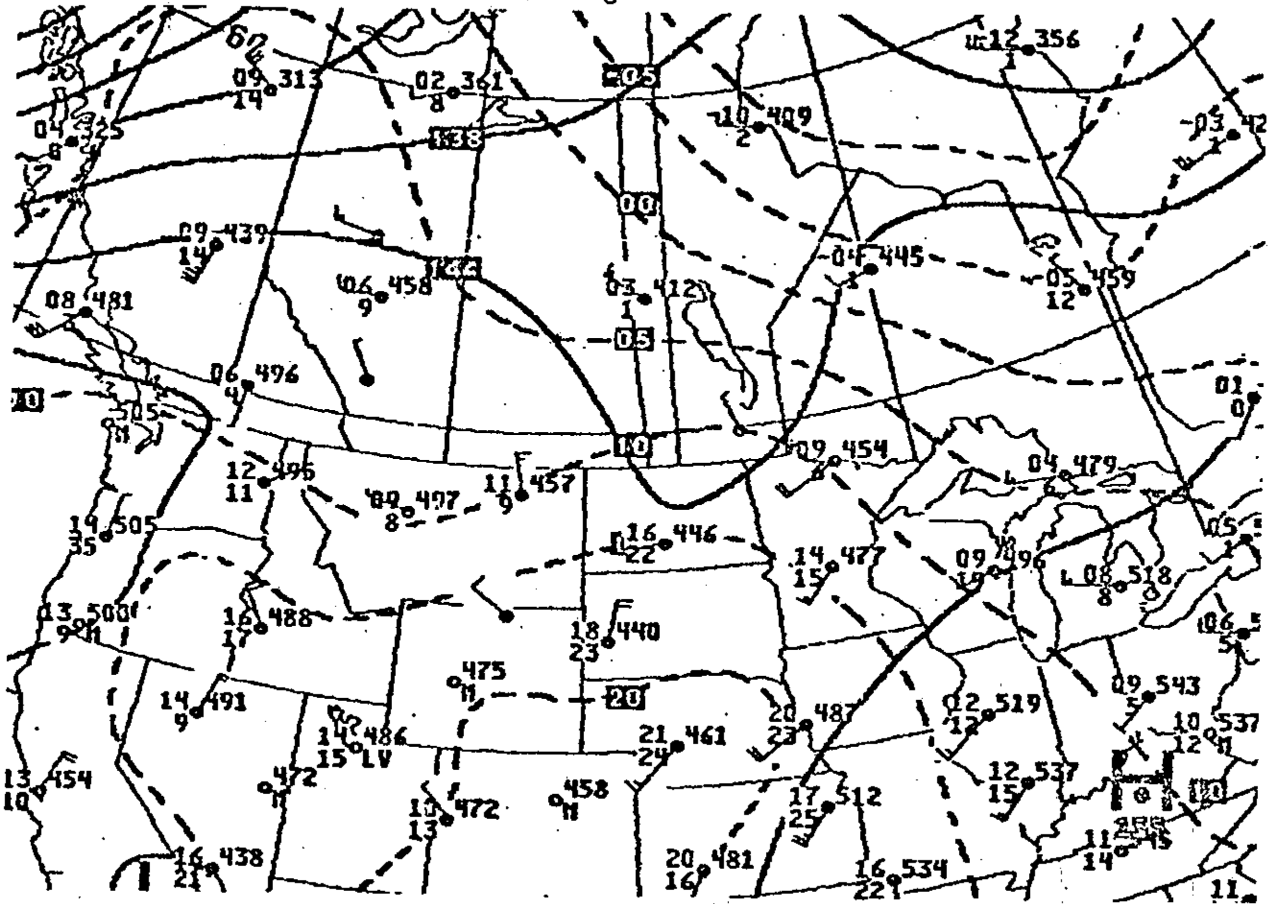

Figure 1 - 850 mb Chart for 12Z October 2, 1972.

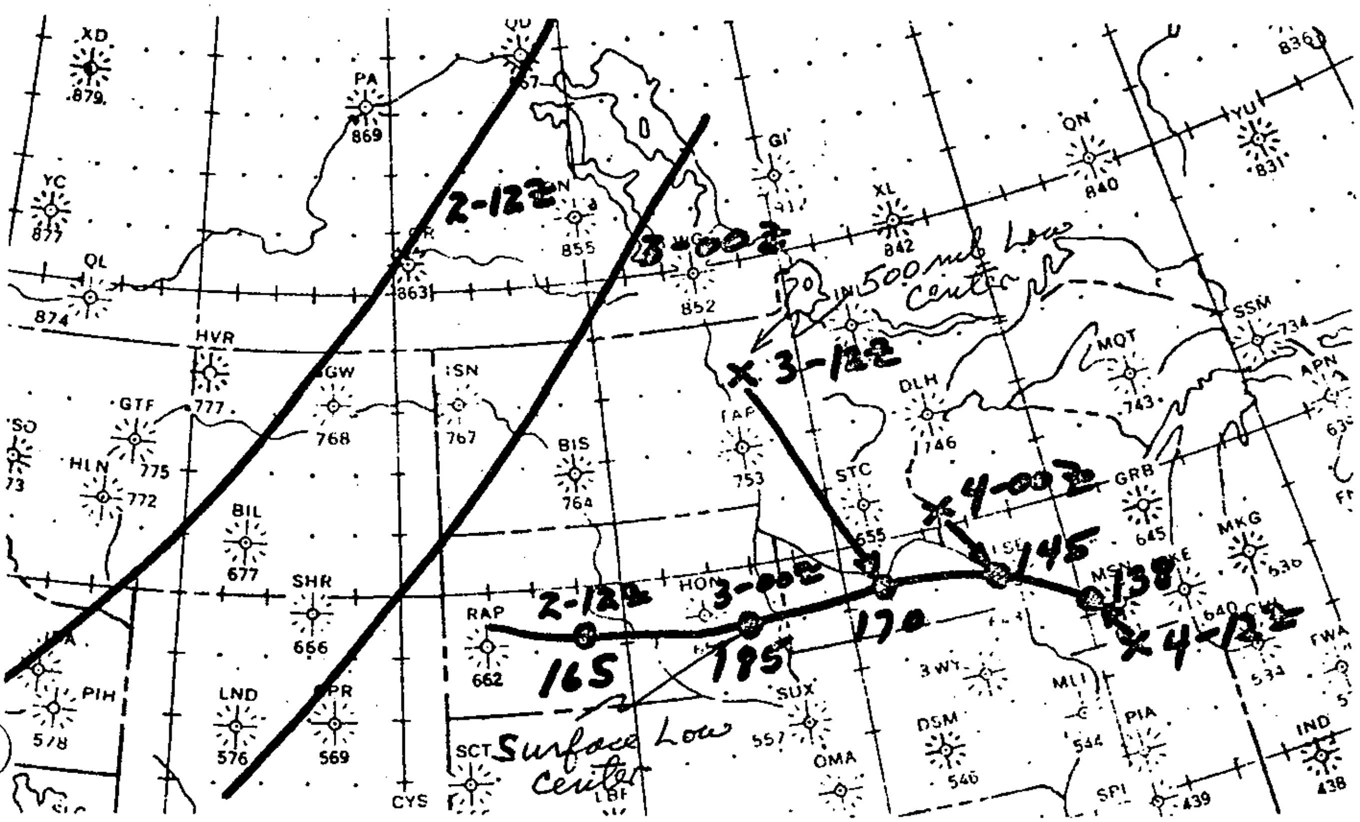

Figure 2 - Track of 500 mb Trough-Low and Surface Low

TECHNICAL ATTACHMENT 75-4

ARCTIC OUTBREAKS

The big blizzard of January 10-11, 1975, brought heavy snow, strong winds and very cold air together, with a resulting loss of life of people and livestock. In Minnesota it gave the lowest pressure on record. This paper will look at that storm in an unusual way---at the plunge of arctic air---rather than in the usual look at the cyclonic development and precipitation. However, why it was so intense was treated in Central Region Technical Attachment 75-3. The reason for this approach is that the strong push of arctic air was significant in creating the strong circulation of this system and its devastating effect on life and limb, and forecasters need a semi-objective way to determine such movement in order to adjust an incorrect numerical prog and know when to leave a good one alone. We examined a number of arctic pushes in the past, using the Daily Weather Map Series, and we found that this January case is an excellent example of the conditions for such pushes. This T.A. is more extensive than most, but there is quite a bit of useful forecasting information in it to justify a careful reading.

The arctic air began developing about January 3, when a nose of the Siberian High was squeezed off by a cold front and pushed into Alaska where it formed a 1026 mb center (Figure 1a). At that time the 500 mb pattern (Figure 1b) had a Low over Alaska so that the 500 mb flow over the surface High was meager, but eastward.

Three days later (the 6th--Figure 2a) the High was centered only a bit farther east. The 500 mb pattern was unchanged, (Figure 2b), and its relationship to the surface High was unchanged. The central pressure of the High had increased a little (4 mb), while the 500 mb height increased some (about 50 meters) over the sea-level High center. The sea-level Low in southern Alberta might be considered as bringing down the arctic air, as it was a fairly strong Low close to the arctic air, but, as the 500 mb flow was unfavorable, this did not occur, as three more days later the High was only slightly farther east (Figure 3a) and had spread only a bit farther south than three days earlier.

In this second 3-day period (to the 9th), the upper pattern started to change. The 500 mb Low that was over the surface High weakened, changed into a trough and started moving eastward more rapidly (see Figure 3b). This further raised the 500 mb heights over the surface High (about 120 meters) and the High further increased in central pressure by 10 mb to 1040 mb. One can see that building of sea-level Highs (increase in central pressure) tends to take place when the 500 mb heights increase over the High. This can be seen from simple hydrostatics, since if the mean temperature (1000-500 mb thickness) doesn't change, the 1000 mb height (sea-level pressure must change the same way as the 500 mb height) i.e., both go up and down together. While the surface vorticity was not examined, it probably decreased, possibly in response to NVA aloft as would be indicated by the Petterssen Development Equation.

Another difference brought on by the change at 500 mb was the beginning of a net flow of consequence at 500 mb over the High. With anticipated more rapid eastward movement of the upper trough, (as progged by numerical models), one should expect more and more upper-level flow over the surface High. This is very significant, as strong 500 mb flow over the arctic High is what brings out the entire High and empties the source region.

One day later, (January 10th), the High had moved little, but had increased 5 more mb in central pressure as the 500 mb height increased 60 m more over the center. The observed 500 mb flow was now strong enough over most of the High to expect movement of significance, and the expected increase in 500 mb ridging in the eastern Pacific would further increase this upper flow. For these reasons one should be reasonably sure at this time that the whole mass of arctic air would indeed finally move out of its source region. This point was stated in real time in the Kansas City map discussion, where this developing storm had been followed closely for the previous two days. Actually, it was an effort to bring down this large very cold High too early (January 7th) that triggered this study and gave the general key that was put to good use as discussed above. One day later the 500 mb flow over the High was very strong (Figure 4) and two days later the High was dominating the U.S. (contiguous 48 states) and the Canadian source region was empty, (see Figure 5). Note that a good portion of the High is over the Great Basin and that the 500 mb flow (see Figure 4) in Alberta crossed the Divide from the east side to the west side (see principle 4 below).

One might note that the 500 mb heights continued to increase over the sea-level High after the 10th, but the central pressure of the High did not increase. In our examination of many cases of surging Highs, it was noticed that when the High was moving, a further increase in central pressure did not occur, even though a considerable rise in 500 mb heights did occur. Why this is so lies in the hydrostatics used to compute 500 mb heights on the raob. When an arctic High is moving out of its source region it is being warmed from below. This will increase the 1000-500 mb thickness (the mean temperature of the column), and if the sea-level pressure is steady, the 500 mb height must rise as the thickness rises, i.e. the 500 mb height is the sum of the 1000 mb height, which changes almost exactly as the sea-level pressure changes, and the 1000-500 mb thickness.

A more complicated case of an arctic surge (actually two surges) occurred more recently (February 3-10, 1975). But before looking at this case, let us state a working hypothesis for the development and movement of arctic Highs (in looking at the Daily Weather Maps for cases, it was noted that the same principles seem to apply to mP Highs, as one might expect).

1. An arctic High in the Canadian source region will remain quasi-stationary as long as the 500 mb flow over it is weak.

2. If the High is elongated, and significant 500 mb flow comes over the eastern portion of the High, that portion will further elongate

and perhaps form a center which will then move as in 3. below.

3. The High center will move when 500 mb flow of significance comes over it. It will move in the direction of the 500 mb flow, and the stronger the 500 mb flow, the faster the High will move, although the relationship is not a simple one.

4. As a corollary to the above, if arctic air is to get into the Great Basin, one needs arctic air west of the Great Divide in Canada with significant 500 mb winds over the arctic air and blowing parallel to the coast south-southeastward), or arctic air east of the Divide, with the 500 mb flow crossing the divide from east side to the west side. Also, if the High is to stall over the U.S., the 500 mb flow over it must weaken considerably, which probably comes about by the upper Low moving over or near the surface High.

5. The central pressure of the High, when the High is quasi-stationary or slowly moving, will increase if the 500 mb height over the High center increases. Once the High starts to move or the east portion elongates significantly, the central pressure will no longer increase significantly, and probably will decrease.

6. Pushes of arctic air generally start with a sea-level Low passing south of the arctic High, the circulation of which draws down some arctic air. Whether it is a small amount, a medium amount, or the whole air mass depends on the 500 mb flow as given above. The whole air mass will come down only when the 500 mb flow over the High center becomes strong.

Let us apply this to the February 1975 case mentioned above. On January 31, 1975, a sizable arctic High (1041 mb) was again centered in the Yukon and an arm of it had pushed southeastward toward the Great Lakes under the west-northwest 500 mb flow (Figure 6a and 6b). The 500 mb flow over the sea-level High center was weak and toward the south, but the High can't move in that direction very well because of the coastal barrier blocking its movement, and destruction of the cold air over the much warmer ocean for that air passing this barrier, so it stayed quasi-stationary. Note the strong 500 mb ridge south to north over Alaska, as this becomes significant very soon.

Three days later (February 3), the sea-level Low or vorticity maximum over the Dakotas had the beginnings of an arctic push west of it (Figure 7a). Note in Figure 7b that the 500 mb ridge that had been over Alaska had rotated so as to bring strong northerly flow over the eastern arm of the Yukon sea-level High. This brought about a southward movement of this arm, the push of arctic air into Montana, and the creation of a separate High as a breakoff of the main High. The main High center had moderate northeasterly flow over it and because of the coastal barrier still retarding movement, should remain quasi-stationary.

Two more days later (the 5th--Figures 8a and 8b), the main High in the Yukon still had about the same 500 mb flow over it and it hadn't moved significantly. There was an increasing 500 mb flow to the southeast over the breakaway High center now just north of Montana so that center should move into the U. S.

Two more days later (the 7th) the breakaway High was now over east Texas, and the 500 mb flow over the main center in the Yukon had increased considerably (Figure 9a and 9b) and was forecast to increase quite a bit more. Thus one should expect, as in the January case discussed earlier, that the main High would move into the U. S. and again empty the source region. With 36 hours, the Yukon High was centered on the Nebraska-Dakota border (Figure 10) under a nearly easterly flow which then turned eastward to center it in Virginia in 36 more hours, (Figure 11).

The Central pressure of the Yukon High slowly and steadily rose from 1041 mb on the 31st to 1052 mb on the 3rd, as the 500 mb height over it increased and while the High was quasi-stationary and not expanding appreciably or sending out offspring High centers. But from the 3rd on, the central pressure mainly decreased as it sent out its first offspring center and later when it moved into the U. S., ending at 1029 mb when the High was centered over Virginia as can be seen from the set of figures starting with 6a.

As stated in the beginning, the purpose of looking at these cold surges is to give the forecaster a semi-objective way to determine what should occur so he can correct an incorrect numerical prog or leave a good one alone. The PE surface progs for these two cases, as judged by the 48-hour sea-level prog, were rather good on the movement of the Highs, except at times they moved the High out too soon. For example, the prog for the 3rd of February, when the breakaway High was forming, showed the whole High near the position of the center breaking away. Later, when the whole High did come out, the prog anticipated it 12 hours or so too soon. Similarly on the January push, the PE progs were quite good except that when the High was starting out (12Z on the 10th), the PE prog valid at that time had it well on its way. However, the later progs quickly corrected the error and were then good as the High moved to the east coast.

For those offices having a good file of the Daily Weather Map, additional examples can be obtained by looking at the cases around the following dates, all in 1972: January 3, 12, 18, 23; February 1, 21; March 17; April 2, 6. The January 3 case is quite interesting as the 500 mb flow is NNWly over western North America, bringing down the cold air, but the flow is clearly split, resulting in a push into both the Great Basin and the Great Plains, with a single High splitting into two two parts with one in each area as a result. The February 1 case is interesting in that the cold air flows mainly into the Great Basin as a result of the upper flow crossing the Divide from the east to the west side on both the 1st and 2nd of February, with the main flow into the Great Basin and not the Great Plains.

Forecasters should keep a set of principles listed earlier close to the forecast desk, and use them as a skeleton around which to build their experience. Modify or expand the rules as you see fit, although they are rather comprehensive as listed. Going back over the figures given with this T.A., by examining the 500 mb flow, the surface High interrelationship, without reading the text will help cement the idea in the mind.

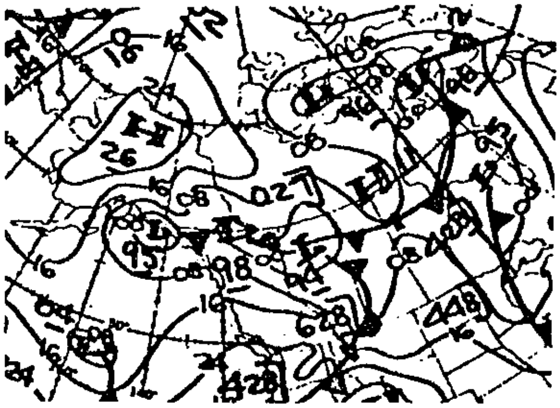

Figure 1a. Sea-Level Chart

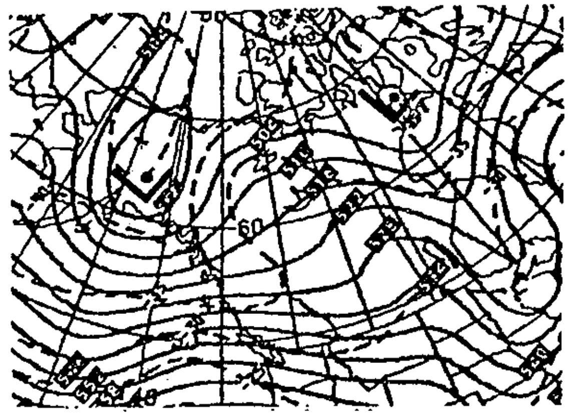

Figure 1b. 500 mb Chart

12Z January 3, 1975

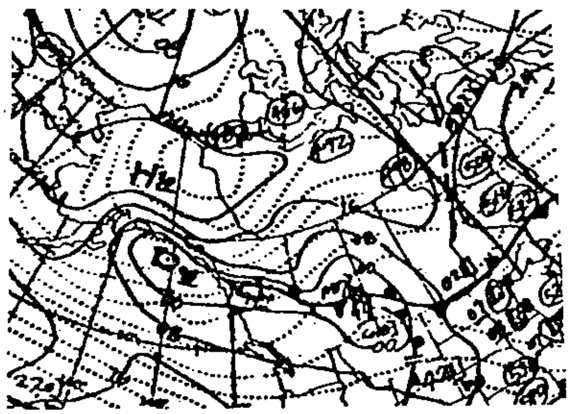

Figure 2a. Sea-Level Chart

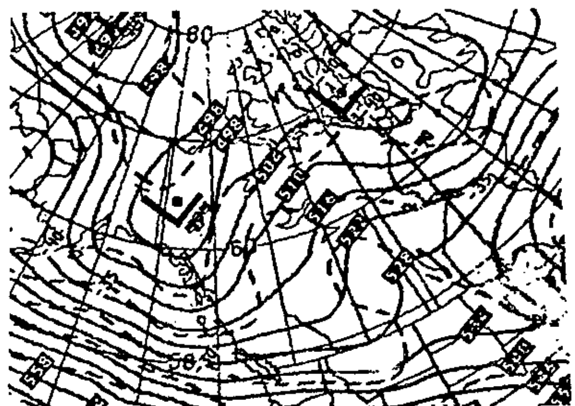

Figure 2b. 500 mb Chart

12Z January 6, 1975

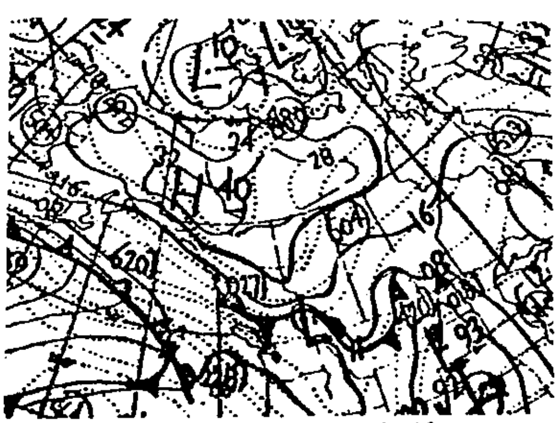

Figure 3a. Sea-Level Chart

Figure 3b. 500 mb Chart

12Z January 9, 1975

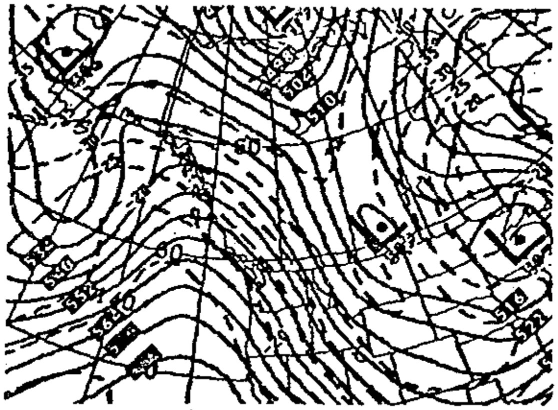

Figure 4. 500 mb Chart
12Z January 11, 1975

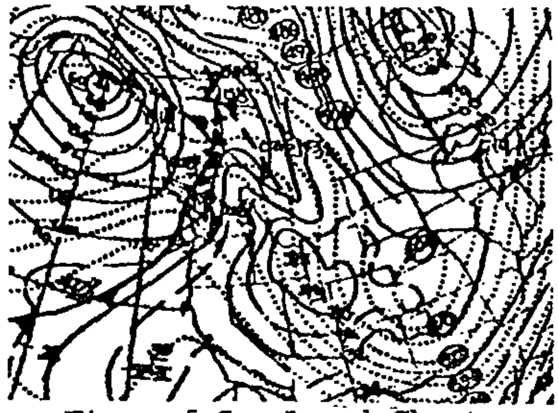

Figure 5 Sea-Level Chart
12Z January 12, 1975

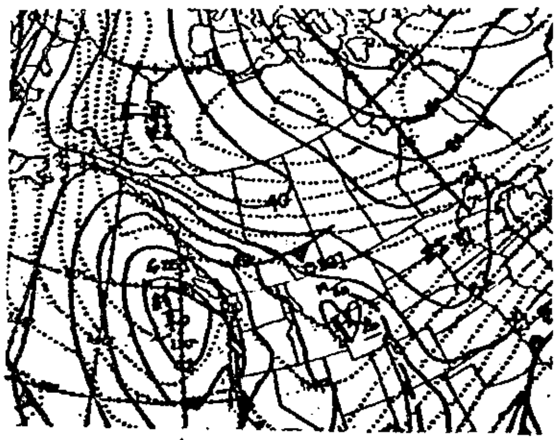

Figure 6a. Sea-Level Chart

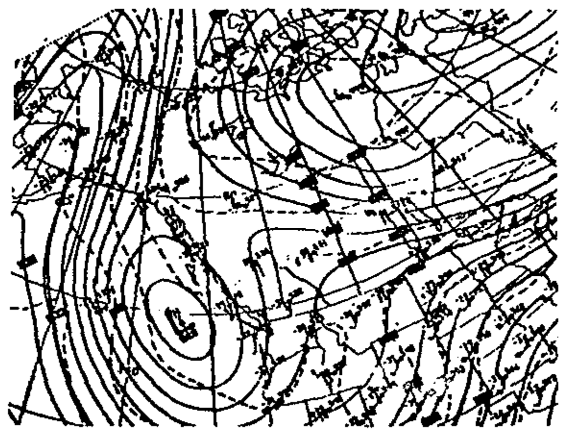

Figure 6b. 500 mb Chart

12Z January 31, 1975

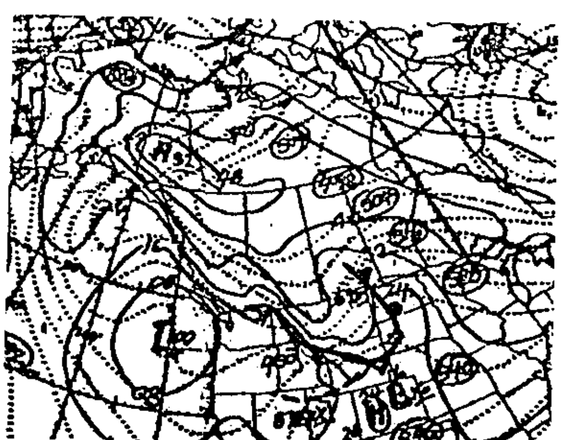

Figure 7a. Sea-Level Chart

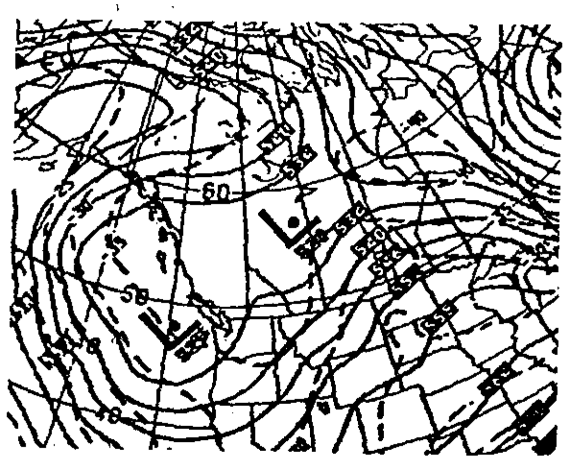

Figure 7b. 500 mb Chart

12Z February 3, 1975

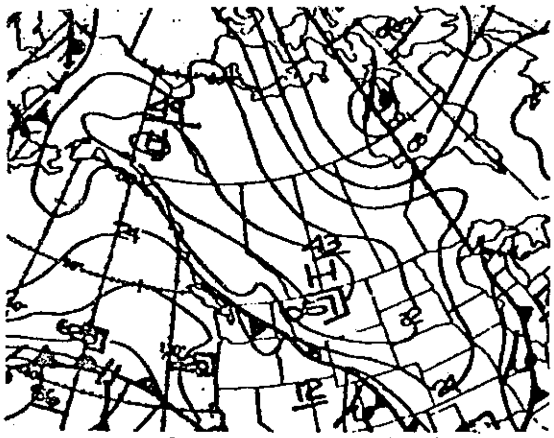

Figure 8a. Sea-Level Chart

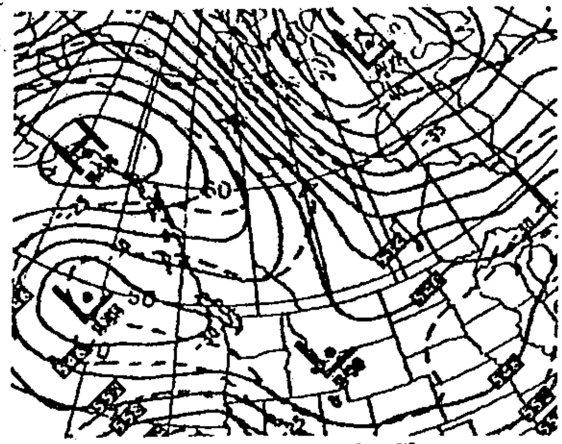

Figure 8b. 500 mb Chart

12Z February 5, 1975

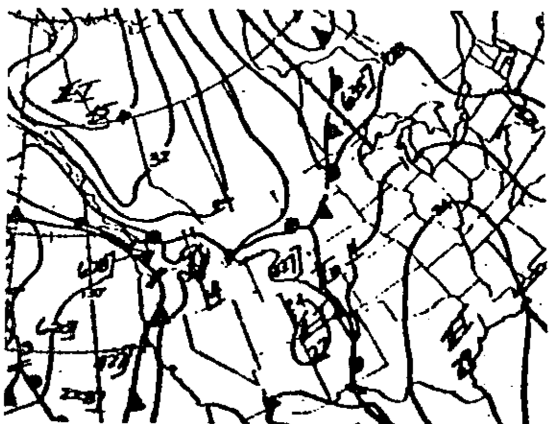

Figure 9a. Sea-Level Chart

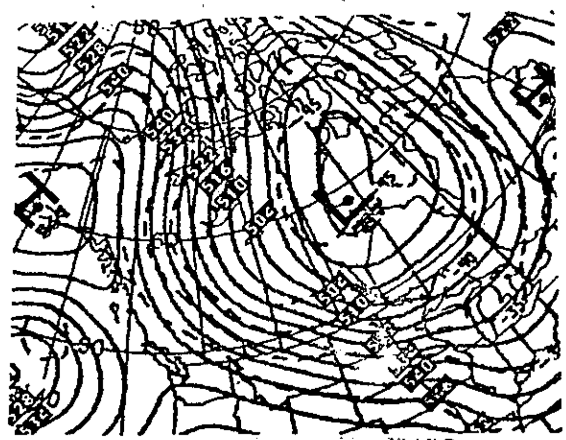

Figure 9b. 500 mb Chart

12Z February 7, 1975

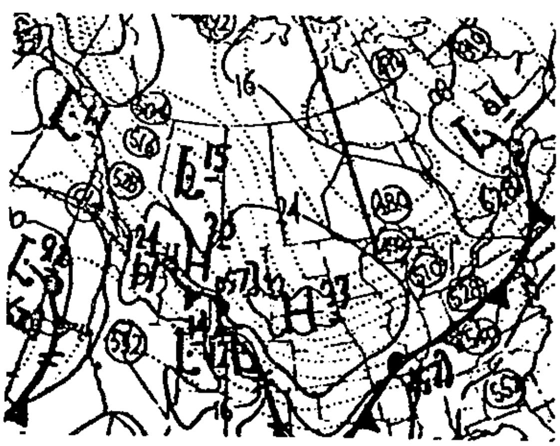

Figure 10. Sea-Level Chart
00Z February 9, 1975

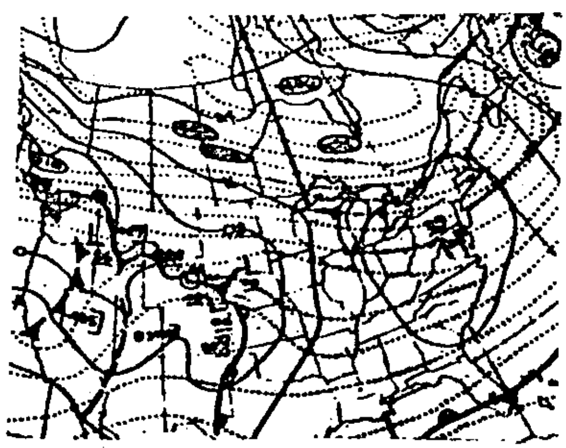

Figure 11. Sea-Level Chart
12Z February 10, 1975

Discussion of a Difficult Precipitation Forecast

CONCLUSIONS

1. Be cautious about discounting precipitation that is under a 500 mb level PVA area even though the precipitation is spotty and/or light. This is especially true if any one or more of the following things are occurring:

   a. The 500 mb trough is intensifying or is forecast to intensify.

   b. The PVA area is moving so as to join forces with other upward-motion producing mechanisms such as warm advection or cyclonically curved flow at the earth's surface.

   c. The PVA area is moving so as to intercept a high humidity area, especially if the high humidity level is within or just below the $-12^{\circ}C$ to $-18^{\circ}C$ temperature level.

A first approximation rule is that IF PRECIPITATION IS ASSOCIATED WITH A PVA AREA AT 500 MB AT THE INITIAL TIME, IT IS LIKELY TO REMAIN WITH THE PVA AREA AS THE PVA AREA MOVES. Probabilities below 10% are generally unwise in any area at the time of an existing or predicted PVA area aloft. If there is any precipitation in the PVA area at the initial time, even only traces, probabilities under 20% are unwise unless one is certain that the precipitation is not associated with the PVA area. Correct and more frequent use of the middle and high probability values can come from careful and thorough use of the MODEL OF THE DAY concept to evaluate the causes of the initial condition, after which the relative motions of significant features provide clues to changes in the model of the day.

FORECASTING DRIZZLE

With the cold season at hand, smaller scale convective systems are being replaced by larger scale advective systems. Likewise, the shower activity associated summer convective currents is giving way to a multitude of precipitation forms. One such is drizzle. Drizzle occurs when droplet growth is limited and drop size is small (in the range 100-500 μ). A look at the parameters which produce this phenomenon reveal that it can be forecast from the processes involved. Mainly these parameters are: 1) temperature in the lowest saturated layer, 2) low-level wind, 3) low cloud thickness, 4) cloud base height, and 5) upward vertical motion, and 6) low-level inversion.

The layer temperature is a prime consideration. The ice-crystal mechanism, (i.e. a mix of water droplets and ice crystals) must be absent or at least minimized or drop size is not limited, and rain will be produced. Mason[1] found that ice-crystals were rarely observed in air at temperatures $\geq$ -4°C, and the rate of crystal formation increased from -4°C to -12°C. A temperature -12°C is accepted as the point where the difference between the vapor pressure with respect to water and that with respect to ice is at a maximum. At this point ice-crystals grow very rapidly and the result is normally rain and/or snow. Due to this effect, we will consider that the low-level saturated layer must have a temperature of $\geq$ -4°C for drizzle production.

Special wind conditions are necessary, it would seem, to produce drizzle. Assuming no ice-crystals, the time required for the growth of cloud droplets to drizzle size would be too great to be realistic except in the case of very fine drizzle or heavy fog conditions. Therefore, it is believed that a degree of low level turbulence is necessary to bring about capture of the small droplets by the larger ones, thus creating droplets large enough for drizzle. We were unable to find mention of a wind speed threshold but from an examination of a number of drizzle cases at Kansas City, we will consider that surface observed winds must be at least 5 kts. for significant drizzle to be created, although an occasional observation $<$ 5 kts. may occur in a drizzle period.

Cloud base height and cloud thickness were investigated by Mason and Howarth[2]. The result, shown in Figure 1, is that drizzle occurs mainly with cloud bases up to 2,000 ft. and rarely with bases greater than 4,000 ft. Cloud thickness was mainly from 2,000 - 6,000 ft. and rarely greater than 10,000 ft. Drizzle occurs much more often than rain alone or rain and drizzle combined when

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[1] Mason, B. J. "Clouds, Rain and Rainmaking" pg. 62, Cambridge Univ. Press, 1962.

[2] Mason, B. J. and Howorth, B. P. "Some Characteristics of Stratiform Clouds over North Ireland in Relation to their Precipitation", Quarterly Journal of RMS Vol. 78, 1952, pg. 226.

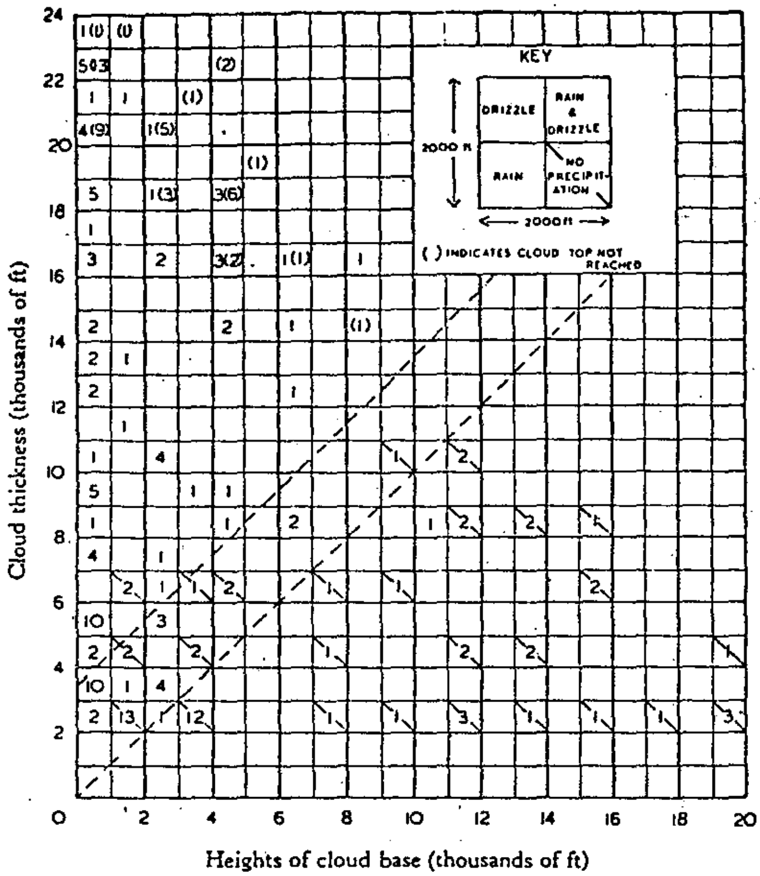

Figure 1. Cloud thickness and height of cloud base of stratiform clouds vs. type of precipitation. The number of cases in each interval of 2000 ft. is given in the square appropriate to the type of precipitation. (See Key) -- from Mason and Howorth.

cloud bases are less than 4,000 ft. AND cloud thickness is less than 10,000 ft., and this result is even without consideration of cloud temperature. Using data from Battan[3], a droplet falling from a cloud base at 2,000 ft. with a drop radius of 200μ would evaporate before reaching the ground (assumes 90% relative humidity from cloud base to ground), thus droplet evaporation is the probable reason for low frequency of drizzle with cloud bases $>$ 2,000 ft.

A statistical study of drizzle at Kansas City was compiled using the past fourteen years of surface observations (from Sept. through April, 1960-74, excluding Nov. - Dec. 1974). Of the 3603 observations recorded, only reports of drizzle or freezing drizzle were used. Drizzle combined with other forms of precipitation was omitted. The results are as follows:

	0-1000 ft.	1000-2000 ft.	2000+ ft.
Frequency (%) of drizzle by cloud base height	88	10	2

	Sept.	Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Apr.
Frequency (%) of drizzle by months	5	9	12	22	28	10	9	5

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[3] Battan, Louis J. "Cloud Physics and Cloud Seeding", pg. 24, Doubleday & Co., 1962.

These figures agree well with Mason and Howorth and show the strong tendency toward cloud bases below 1,000 ft. A mean base height for all cases was 625 ft. AGL, and the highest base observed was 4,000 ft. Looking at the frequency of drizzle by month, essentially half the occurrences were in the 2-month period of December and January, the two coldest months of the year, with quite low frequency at the end points of April and September. While these figures are for Kansas City, they should apply reasonably well to the region as a whole.

To look at the synoptic patterns associated with drizzle, 17 cases of observed drizzle were extracted from the past two years of the Daily Weather Map, and examined. At the surface, wide-spread drizzle occurs most often on the over-running side of a warm or stationary front, and with disorganized systems. The disorganization seems a necessary ingredient, since all 17 cases turned into rain producers when system organization took place. The 500 mb trough line was displaced far west of the drizzle area, producing weak PVA and weak upward vertical motion over the drizzle area. It appears, however, that some large scale upward vertical motion may be necessary for drizzle, and that frontal overruning is the main vertical motion producer, although terrain upslope is adequate*. Turning now to the temperature-dew-point profile, all 17 cases were examined using the computer-created adiabatic diagrams from NSSFC. It was found that a saturated low-level inversion layer was observed in all soundings, with the top of the inversion averaging about 100 mb above the surface. In all cases the top of the low-level saturation layer was between 850 mb to 800 mb, i.e. above the top of the inversion.

In summary, the conditions for drizzle are:

1. The cooler part of the year, and especially December and January.
2. Cloud base $\leq$ 4,000 ft. and especially $\leq$ 1,000 ft.
3. Cloud thickness $\leq$ 10,000 ft. and especially 2,000 - 6,000 ft.
4. Saturated low-level layer temperature $\geq -4^{\circ}$C.
5. Surface Wind $\geq$ 5 kts.

We also believe that some but not much upward motion is also necessary. This is usually with a front, but it may be due to terrain. In either case a saturated low-level inversion about 100 mb thick is present.

If any of the above items are not present or within the specified limits, the chance for significant drizzle is low.

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*In a recent case of "return flow" drizzle, a low overcast and fog covered a fairly large area of the Plains, but the wide-spread drizzle was where there was low-level upslope winds.

TECHNICAL ATTACHMENT B

BOUNDARY-LAYER WIND MAXIMA AND NOCTURNAL THUNDERSTORMS

Thetime of the year is here during which a phenomenon known as a boundary-layer wind maximum (hereafter referred to as a BLWM) plays an important role in determining the nocturnal thunderstorm weather of an area centered on eastern Nebraska and eastern Kansas. This phenomenon is not handled by any numerical model in operational use in the Weather Bureau, so forecasters are on their own in dealing with it. These maxima are still incompletely understood, but two possible physical causes are known. The relative importance of each remains to be determined, however, so this attachment is somewhat speculative in nature. A BLWM is characterized by a maximum in the vertical profile of wind speed (sometimes 70 knots or more) at 1500 feet or so above the ground. The term "low-level jet" is often applied to wind speed maxima in the horizontal, as well as in the vertical. The two often occur simultaneously, but wind maxima in the vertical are the prime concern of this attachment, so "BLWM" is used to refer to them rather than "low-level jet" to avoid confusion.

Blackadar (1) has proposed a simple explanation of the BLWM, which he concedes is oversimplified and does not completely explain it. During the day turbulence arising because of an unstable lapse rate causes a rather large frictional force to be exerted on the air in a fairly deep boundary layer. The frictional effect reduces the wind speed and causes a deflection of the actual wind to the left of the geostrophic wind. In the evening the air starts to become stabilized and so the frictional force is reduced, especially in the upper part of the daytime boundary layer. Now the pressure force is not balanced by the resultant of the frictional and Coriolis forces, so the air is accelerated. After a time the Coriolis force has increased due to the greater speed and the actual wind vector begins to rotate to the right as an inertial oscillation takes place. The theoretical period of this oscillation is 1/2 pendulum day or $12/\sin \phi$ hours, where ϕ is the latitude. At latitude 40 degrees, 1/2 pendulum day is about 18.7 hours. The wind speed would tend to reach a maximum at 5 or 6 pendulum hours (8 or 9 clock hours) after stabilization begins. This would put the maximum around 0200 CST.

This preceding discussion is based on the assumption that the <u>geostrophic</u> wind does not vary diurnally in the boundary layer -- not a very good assumption over sloping terrain as shown in (2). There have been studies which attempted to explain the BLWM on the basis of this phenomenon alone, without bringing the diurnal frictional variation into the picture, but these treatments have not been too convincing. The fact that the Amarillo-Oklahoma City-Dodge City triangle experiences the greatest frequency of the BLWM's (3) is suggestive that the diurnal geostrophic wind variation is at least enhancing the frictional effect.

The geostrophic wind effect on the BLWM is greater when there is a large diurnal temperature change, where there is the largest slope of the terrain, and where the wind is parallel to the terrain contours. The afternoon frictional effect can be regarded as roughly proportional to the square of the wind speed, so it increases rapidly as the wind speed increases. Thus, the diurnal variation in the low-level wind speed is likely to be a maximum when afternoon skies are sunny, and the sea-level geostrophic wind is at least moderate southerly. An important point to remember is that in summer the sea-level chart gives a good representation of the southerly surface geostrophic flow over the Great Plains only during the early morning hours (see Fig. 4 of (2)). By afternoon the south component can actually be 15 knots or so stronger. The reason for this is that errors are introduced by the plateau correction over sloping terrain and the temperature averaging in the reduction to sea level.

Aside from the importance of knowing of the existence of and being able to predict the BLWM for aircraft operations, the BLWM is of interest due to the convergence fields which are set up during the night because of differential effects. Various combinations of circumstances may lead to upward vertical motion fields and therefore thunderstorms if stability and moisture conditions are favorable.

It has long been known and recently documented again (4,5) that from May through September there is a maximum in precipitation frequency in the eastern Great Plains centered around eastern Nebraska in the 0000-0600 CST period, with a minimum similarly located but in the 1200-1800 CST period. It is physically reasonable and many people believe that the precipitation maximum and the BLWM are effect and cause. The problem from the forecasting standpoint is how to utilize the knowledge we have of the BLWM to produce better precipitation forecasts.

To a first approximation then, nocturnal precipitation will exist in the eastern Great Plains from May through September, with the maximum probability in eastern Nebraska. This last may well be because the average slope of the Plains decreases north of Oklahoma. But precipitation doesn't occur in eastern Nebraska every night, nor is the maximum probability in eastern Nebraska every night. What are the factors that raise and lower, as well as shift the maximum probability? We suggest you build experience this summer around the principles given above and the speculations given below, so that by the end of this summer you may have specifics in more detail for your own station.

The main precipitation inhibitor during this period may well be the moisture content of the air. The precipitation will associate closely with the low-level moist tongue, and if the low-level flow is from near due south, the precipitation may well maximize in the moist tongue around the Nebraska-Kansas border, even if the moisture tongue is in western Kansas-Nebraska.

But if the flow is from the southwest, the convergence due to the BLWM must overcome the low-level downslope effect of the terrain and the precipitation maximum will be in the moist tongue or high moisture area but displaced downwind to extreme eastern Kansas or extreme western Iowa. Under such conditions it may well start a bit later at night.

If moisture is too <u>low</u>, there will not be precipitation even if the BLWM is producing low-level convergence for several hours a day. If the stability is too <u>high</u> for thunderstorms (low moisture may be the cause), precipitation also will not occur. If there is a weak diurnal temperature cycle from northern Texas northward, due to cloudiness, the BLWM effect on precipitation will be lessened. Of course, if there is pronounced warm low-level advection or positive vorticity advection at 500mb, these effects are likely to be dominant and the BLWM effect hard to find.

The BLWM effect on precipitation can be augmented as well as diminished. Unusually high moisture or unusally low stability can aid the BLWM effect and also shift the most likely area of precipitation. Weak warm advection or weak PVA aloft can allow the BLWM effect to be dominant but augment it. Cold fronts passing across the Plains can aid the BLWM effect. Cold fronts passing through the eastern Nebraska and eastern Kansas areas in the middle of the night should have a better chance of precipitation or more widespread and heavier precipitation than if they pass through in the middle of the day. Thus, there is likely to be a variation in the precipitation with cold fronts moving through the Plains depending whether they pass through in phase or out of phase with the BLWM effect.

A prominent augmenting effect occurs when cloudiness suppresses daytime heating, and thus the BLWM effect, in Nebraska and/or Kansas, but more sunny conditions exist farther south. A common summer situation of this type occurs when a slow moving front lies east to west across the Great Plains, with fair skies and southerly winds south of the front and rather cloudy skies with a different wind direction to the north. Even though the front may be rather weak in its temperature gradient, the differential effect on the BLWM will usually bring about a large amount of nocturnal shower activity. If the upper flow is weak, the convective cells will have a slow movement, and heavy rain can easily result. The activity can be to the south of the front as well if there is a diffluence of the isobars or change in curvature to more cyclonic in this area.

If nocturnal activity is rather pronounced, it is rather easy to advect the thunderstorms into Missouri, Iowa, eastern Minnesota and western Wisconsin before they die out. If the upper level flow is light they might not advect as far but could easily give heavier rains in the areas they do cover. If the upper flow is quite strong, say 50 knots or more at 500mb, the activity may well advect into Illinois and eastern Wisconsin, and occasionally much farther, before dissipating. Thus, while the BLWM effect is primarily a Great Plains phenomenon, its overall effect can influence the forecasts for the whole Region except probably for Wyoming and western Colorado. It is also an effect that, as far as precipitation is concerned, is almost entirely in the Central Region.

REFERENCES:

1. Blackadar, A. K., 1957: Boundary layer wind maxima and their significance for the growth of nocturnal inversions. Bull. Amer. Meteor. Soc., 38, 283-290.

2. Sangster, W. E., 1967: Diurnal surface geostrophic wind variations over the Great Plains. Central Region Technical Memorandum 13, 16 pp.

3. Bonner, W. D., 1965: Statistical and kinematical properties of the low-level jet stream. Dept. of the Geophysical Sciences, The University of Chicago, Satellite and Mesometeorology Research Paper No. 38, 54 pp.

4. Scientific Services Division, 1966: Climatic frequency of precipitation of Central Region stations. Central Region Technical Memorandum 8, 51 pp.

5. Jorgenson, D. L., 1967: Climatological probabilities of precipitation for the conterminous United States. ESSA Technical Report WB-5, 60 pp.

CRH SSD
July 1971

Technical Attachment 71-13

Dew Forecasting

The formation of dew is of interest to agriculturalists and others. While we don't make forecasts of dew formation, knowledge of the phenomena is sufficient that we could. The following method of dew forecasting appeared in the Western Region's Technical Attachment 71-25, as written by Mr. Earl Bates, AAM at Corvallis, Oregon. Even though the conditions in Corvallis are quite different from most of those in the Central Region, it is believed that the principles given and even the forecasting graph are general enough that they will be applicable to CR locations.

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Dew is the condensation of atmospheric water vapor on plant surfaces and other objects near the ground due to temperatures dropping below the dew point during the night. The two major formation factors are strong radiational cooling during the night and available moisture near the ground level.

Synoptic weather patterns most conducive to dew formation are those associated with:

1. Essentially clear nights.
2. Little or no wind.
3. Moderately low dew points as reported from weather shelters.

(Rather high dew points with clear skies and light winds are usually associated with the formation of radiation fog rather than dew. Also, the wetting of plant surfaces due to rain and fog, associated with moist airmasses, is referred to as leaf wetness rather than dew.)

The amount of water vapor available for condensation at plant level depends on the moisture content of the air, including advective changes taking place, transpiration of moisture from plant tissue, and evaporation of moisture from the underlying soil. Evaporation of soil moisture has been indicated by Smith and Carpenter of South Carolina to be of prime importance in dew forecasting [1]. It is interesting to note that there is an inverse relationship between afternoon shelter relative-humidity observations and dew intensity after midnight; i.e., moderately low relative-humidities are highly correlated with dew formation [2].

A minimum of low-level diffusion is a vital factor in dew formation. The surface wind is used to evaluate the intensity of this diffusion. In the Mississippi Delta, using wind observations 40 feet above ground level, it was found that moderate dew was often associated with winds less than 3 mph, light dew with winds 5 - 6 mph, and no dew with winds 7 mph or more [2]. On the coastal plain, direction and strength of the pressure gradient are important. An onshore wind at evening is considered a 'no-dew' factor, because nighttime cloudiness usually accompanies onshore flow. Weak flow from the interior is favorable for dew formation.

Returning to moisture considerations, the local moisture source is of prime consideration when forecasting the intensity of dew formation expected. The main source of local moisture is the soil.

Forecasters in the Willamette Valley do not have a routine knowledge of the amount of available moisture in the soil, but they can obtain a useful estimate of soil moisture by noting the number of days since the last rain. In this area, irrigation of crops never lets soil moisture be depleted to the wilting point of plants. Within the root zone, soil moisture probably always equals or exceeds half an inch of water. Therefore, for the first three days after rainfall, an abundance of moisture is usually available in the soil, and heavy dews are likely if synoptic conditions are favorable. As time progresses beyond this point with no additional rain, the intensity of dew formation decreases since the available moisture decreases.

The accompanying graph for the Willamette Valley gives a joint relationship between (1) available soil moisture (i.e., number of days since rain) and (2) average wind speed after midnight, and formation of dew. (This graph can be used as a starting point for other locations, and then modified as local data become available.)

PROPOSED DEW FORECAST PROCEDURE

When the following criteria are forecast to exist from midnight to sunrise, use attached graph to estimate the dew intensity for 'tomorrow' morning.

1. Night sky cover .4 or less.
2. Weak pressure gradient with offshore component.
3. Relative humidity 85% or greater by midnight.
4. [Noncoastal areas - winds less than 7 mph.]

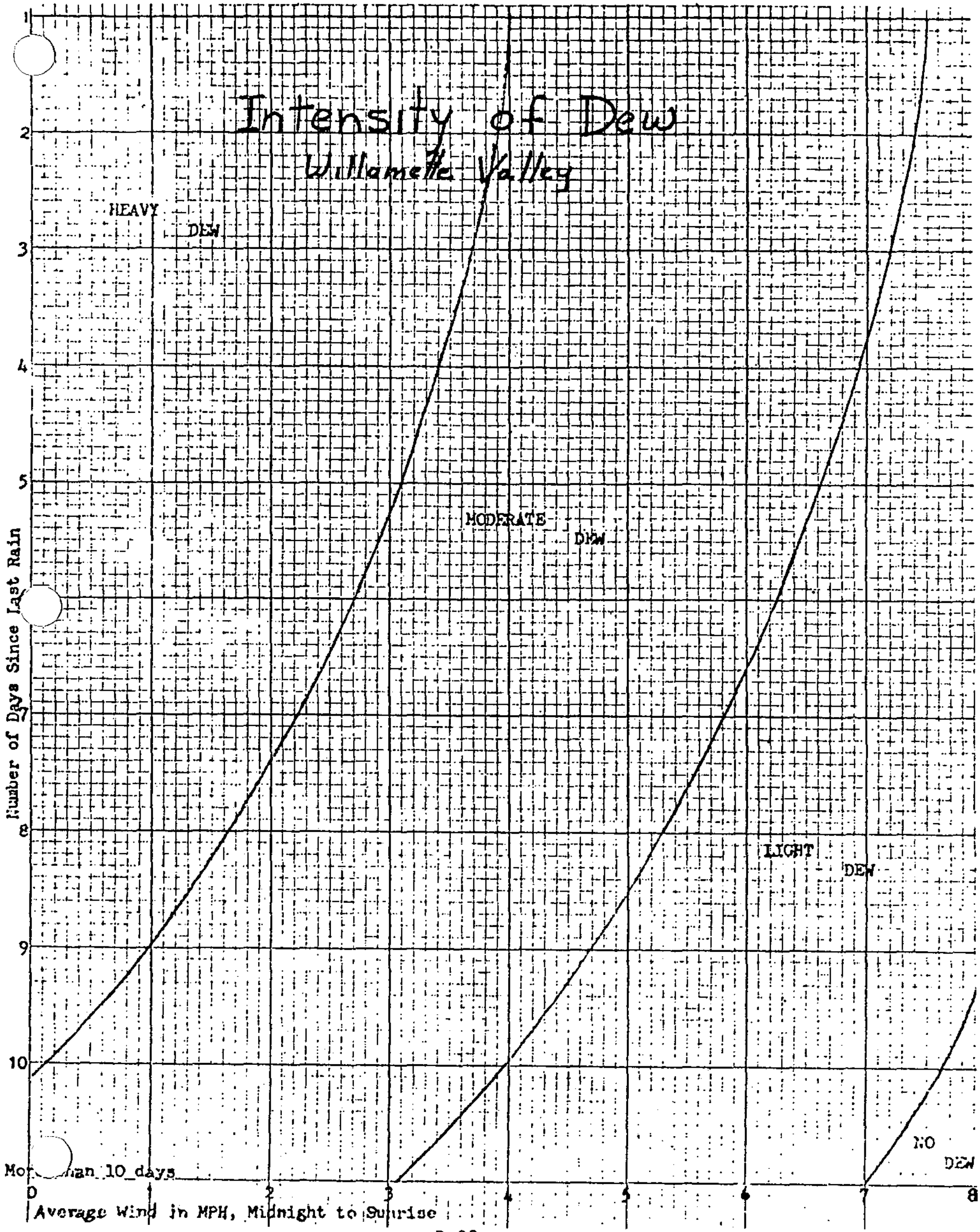
DEW DISSIPATION FORECAST RULES

1. Forecast the time when both relative humidity will be 85% or lower and wind speed will be 5 mph or greater.
2. If morning wind is not forecast to rise to 5 mph, forecast the time relative humidity will drop to 80%.

REFERENCES:

1. Smith and Carpenter, "Dew Forecasting Study for South Carolina," South Carolina Department of Agriculture, April 1966.
2. Riley, J. A., "Moisture in Cotton at Harvest Time in the Mississippi Delta", Monthly Weather Review, September 1961: p.345-345.

(Geiger's book, "Climate Near the Ground", p.110-111, is a good general reference.)


Intensity of Dew
Willamette Valley
HEAVY DEW
MODERATE DEW
LIGHT DEW
NO DEW
Number of Days Since Last Rain
1
2
3
4
5
7
8
9
10
More than 10 days
0
1
2
3
4
5
6
7
8
Average Wind in MPH, Midnight to Sunrise

Technical Attachment 73-6

FORECASTING (WHITE) FROSTS

This T.A. is not concerned with the whole spectrum of frost forecasting as relates to the agricultural program. It is limited mainly to the condition which produces sublimation on bridges and overpasses that contributes to a slipperly and unsafe road condition on the bridge surface. Such frosts will occur before and at times without frost on the ground, so the condition may not be readily noticed by motorists until too late, thus making it a major road hazard.

Forecasting white frost should be very similar to forecasting dew, although, of course, the temperature is also involved, as the skin layer near the ground or exposed object must get below the freezing point. The dew forecasting conditions essentially as stated in our Tech. Att. 71-13 (from WR T.A. 71-25) are:

1. Essentially clear sky ($\leq$3/10 cover) all night, except for high thin cirrus.

2. Little or no wind, especially after midnight.

3. Moderately <u>low</u> dewpoints at shelter height.

(Rather <u>high</u> dewpoints with clear skies and light winds are usually associated with fog rather than dew.)

Whether it will be dew or frost depends only on the dewpoint. If the dewpoint is less than 32°F at shelter height, frost will occur rather than dew. Of course, the temperature has to cool low enough to reach the dewpoint at the skin level of the earth or other objects, so the official shelter temperature must get fairly close to freezing, say less than 40°F, or frost cannot occur. Local variation of this critical temperature will occur, depending on whether the official site is in a pocket or on a ridge, so some experimentation and observation should be made so as to determine the critical temperature for each observing site. Bridges and overpasses tend to be near low places and thus near the extremes of minimum temperature.

One major factor making it easier to get frost deposition on bridges than on the ground is the lack of heat conduction from the large ground reservoir. In that sense, a massive stone or concrete bridge would frost slower and less often than a steel bridge, as the steel bridge would store less heat during the daytime, and conduct it to the surface more rapidly at night. Likewise, a frost is more likely when clearing after a cloudy day not only because the temperature can drop below freezing easier, but because there is less heat from direct sunshine stored in the structure. Similarly, the portion of a bridge or road shielded from the sun by a cliff will frost up more readily, and even the north side of a hill (road or bridge) will frost quicker.

Incidentally, when considering frost over soil, the moisture of the surface of the ground becomes of prime importance, according to our T.A. 71-13. Thus, a frozen and/or dry ground is much less likely to allow frost deposition than unfrozen moist ground.

We, thus, can add three things to the three listed on the previous page to get frosting on bridges:

4. Dewpoint at shelter height less than 32°F.
5. Minimum temperature at shelter height below about 40°F and within 8°F of dewpoint (but not so close as to make fog more likely).
6. Cloudy days (but clear nights) and steel bridges give maximum results.

If the conditions for frost are favorable, the degree is mainly dependent on the wind speed, especially after midnight. Using the criteria of T.A. 71-13, we would have heavy frost with the average wind after midnight as 0-4 mph, moderate with 4-7 mph, and light with 7-10 mph.

A related condition that could further increase the road hazard is that of snow-melt water on the bridges (or elsewhere) which freezes at night, whether or not frost forms. This is especially treacherous when, like frost, it occurs on bridges first or only occurs there.

CRH SSD
March 1973
LAH

Technical Attachment 73-7

SHORT TERM WATER LEVEL CHANGES ON THE GREAT LAKES

As is well known by now, the level of the Great Lakes is extremely high. Lakes Ontario, Erie and St. Clair are a little below to a little above the all-time record highs, and several feet above normal, while Michigan-Huron is only about a foot below record levels, but as much as three feet above the average levels of the past 10 years. Lake Superior, which is almost completely controlled, is only a bit above normal.

Upon this long term cycle, which of course results from variations in precipitation-evaporation, is the annual cycle with its low levels in the winter and the peak in the summer. The further north the lake, the later the high and low points, so that Lake Erie generally peaks in June and Superior in September. This variation is of course also caused by precipitation-evaporation variations, and especially the tie-up of precipitation into snow in the winter. The average magnitude of the annual cycle from low to high level averages about a foot on Superior and Michigan-Huron, and increases downstream to nearly two feet on Ontario.

Naturally with the water this high and the seasonal peak still ahead, many people are concerned with the short term changes in water level that can be quite large and therefore seriously aggravate an already serious problem. Precipitation, even if it is excessive, does not produce significant changes in lake levels over a period of one day or even a few days. Astronomical tides must exist in all bodies of water, but the Great Lakes are too small for this effect to be significant either. There are three main ways to make significant changes in the real or effective level of these lakes within a single day. These are:

1. WAVE ACTION, caused by the breaking of waves onshore.
2. SURGES, produced by the pressure jump and/or wind of a squall line.
3. SETUP, caused by the wind stress tipping the surface of the lake.

WAVE HEIGHTS, but not their onshore action, has been discussed by Hughes in Central Region Tech. Memo No. 21 entitled "Wind Waves on the Great Lakes". SURGES, especially in lower Lake Michigan, have been discussed in considerable detail in articles by Platzman, Irish, and Hughes in the Monthly Weather Review (May 1965). SETUP has been discussed by a number of people. Key among them is an article by Irish and Platzman in the Monthly Weather Review (Feb. 1962), which dealt with Lake Erie, and an unpublished paper by L. Bajorunas entitled "Water Level Disturbances in the Great Lakes and their Effect on Navigation", which was given at a conference in 1960 and, among other things, relates effects on the other Lakes to those on Erie for like meteorological conditions. This Tech. Attachment will touch on the key points of these papers as they relate to the high water problem. However, none of these efforts deal with Lake St. Clair because of its extreme shallowness, but some general comments will be made.

WAVE ACTION is the best known and the most common effect. Wave action of concern is mainly from the erosion caused by it, or the flooding that can be caused by waves splashing over a breakwater or through beach runup carrying beyond the first rise from the beach. There is probably no concern for these two factors unless lake levels are high. Appendix I in Tech. Memo 21 gives the wave height away from shore for various wind speeds, fetches, and wind durations. The actual fetch (distance the wind blows across the water), when approximately along the long axis of the lake, must be reduced considerable on all the lakes before entering the table. This, as discussed on page 8 of the Tech. Memo, is because the lakes are long and narrow and amounts to a fetch reduction of 40-50% on all but Superior.

An extreme case for sustained strong winds was the Armistice Day storm of 1940 as reported in the Monthly Weather Review for June 1941, where land winds were reported to average about 35 mph for about 7 hours (wind speeds over water most likely were 50-55 mph).* In this case the longest fetch of the lake is not fully utilized, as the wave height is limited by the duration of the strong wind, so that the max significant wave height would be about 20 feet. However, as noted on page 3 of the Tech. Memo, the most frequent waves would be about half the significant height, while the greatest wave would be almost twice the height.

Of course, wave action could be aggravated by setup (see below) as the waves would always be related to the height of the water, whether the lake surface is raised by setup or not. Fortunately, the bottom effects that allow setup to get large have the effect of reducing the peak waves, so there is some compensation between these two effects. Also, the strong winds needed to produce large waves occur in the colder part of the year when the lake level is lower. Wave action is also reduced in winter by both shore ice and by free floating ice, and in spring whenever warm air flows over the very cold water (the surface water usually starts warming rapidly about June 1).

SURGES of much consequence are fairly rare, even on lower Lake Michigan where they have been studied the most and probably are the most frequent, occurring there possibly several times a year for small (a foot or so) surges to one each 10 years or so for the large surges. The largest unofficially reported lake level change at the shore with these surges is a rise of 10 feet. However, since they are squall-line caused, they tend to come in summer when the lake level is high.

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*The conversion of land observed wind to over-water winds is a very complicated matter. The main factors are the type of anemometer exposure and its height, the distance of the anemometer from shore and the wind direction, the time of day, the air-water temperature difference, and the general stability of the lowest portion of the atmosphere. To simplify this, let us say we are interested only in the stronger winds and thus with good mixing in the low-levels. We will also eliminate that portion of the year with significant amounts of ice on the lake, and also those cases especially in spring where the air temperature is much greater than the water temperature. Then, taking land winds at Weather Service reporting offices and lake winds at the 65 foot anemometer height of boats, an increase of the land observed wind by about 40% would yield a reasonable lake wind. A significantly smaller percentage would be more reasonable if the land reporting point were only a mile or two from the water and an onshore wind existed.

These surges are brought on by the pressure jump and/or wind gusts of the squall-line, but to be significant the line must move with the speed directly related to the water depth. This speed according to Bajorunas is:

Superior	85 mph
Michigan	64 "
Huron	54 "
Erie	30 "
Ontario	63 "

Since the maximum of squall-line activity is southwest of the lakes, the frequency of squall-lines moving with the necessary speed (the direction of movement can also be critical) over any of the lakes is low enough that large water level changes due to this effect are uncommon on all of the Lakes. However, formal statistics are lacking.

The surge, when it does come, gives a rise of water that is fairly slow (5-10 minutes) and generally without significant breaking action such as with waves. Return surges generally of decreasing magnitude will usually occur at intervals of one to several hours, depending on the Lake in question and the direction of the squall-line movement.

SETUP occurs when winds blow along the long axis of a lake, causing a movement of water to the downwind end. Setup is best known on shallow Lake Erie. This is because bottom friction slowing the sub-surface return flow allows setup to occur, and the shallower the lake, the more the setup for given wind conditions. Irish and Platzman gave rather good statistics for the larger cases of this event on Lake Erie as follows:

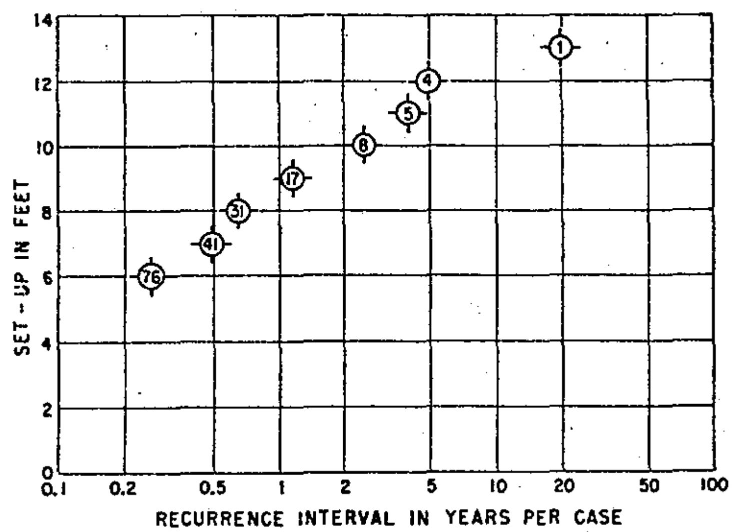

FIGURE 3.—Recurrence interval for which Buffalo-minus-Toledo set-up exceeded specified amounts. The circled numbers give the number of cases for which set-up exceeded 6, 7, 8, 9, 10, 11, 12, 13 foot during the 20-year period 1940–59.

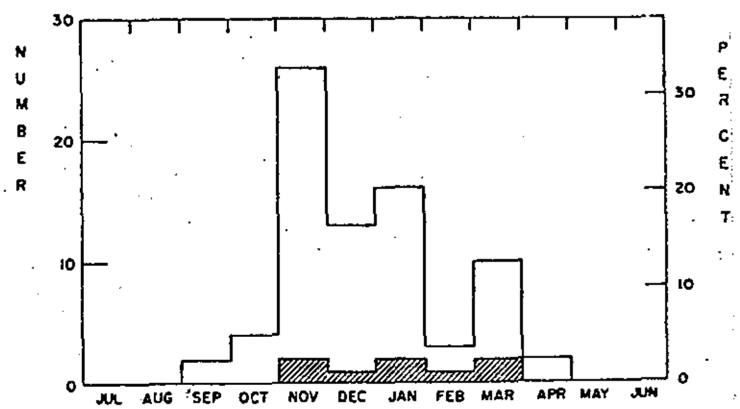

FIGURE 4.—Monthly distribution of 76 cases in which Buffalo-minus-Toledo set-up exceeded 6 feet during the 20-year period 1940–59. Hatching shows distribution of 8 cases in which set-up exceeded 10 feet.

These figures relate only to setup of more than six feet, where the setup is defined as the water level of Buffalo minus Toledo and thus apply to west wind induced setup. That produced by east winds was not studied, but they noted only three such cases in the 20-year period where the Buffalo stage was lower than that at Toledo by six feet or more. This averages to one such case in about seven years. This is fortunate, as high water at the west end of the Lake is much more troublesome than high water at the east end.

If the data of Figure 3 above are extrapolated downward to a setup of 2 feet, there would be around 20 such cases a year. While this averages one to two a month, the peak would still be in the cold season with some summer months getting none. The greater-than-6-foot figure, of course, converts to about 4 per year.

Bajorunas's figures for the relative frequency of setup on the other lakes adjusts for the greater depth and generally greater fetches of the other lakes so that, taking Erie as 100%, the relative frequency (in percent) for the other lakes would be:

Superior	17
Michigan	27
Huron	25
Ontario	18

Thus if a particular wind produced a maximum setup on Erie of 8 feet, such a wind would produce only 2 feet of setup on Huron. The relative frequency of setup-producing winds did not enter these figures, but a major difference among the Lakes is not likely.

Setup takes about 6 hours to become fully established, and a bit longer to drop back to normal once the wind diminishes. Additional high setup values recur as the lake tends to rock back and forth until friction or other winds damp the oscillation completely. Their return period is not nearly so regular between storm systems as with surges, but is mostly between 12 and 24 hours, depending on the way in which the wind speed and direction changes. Succeeding rises are always much less than the primary rise.

SUMMARY

WAVE ACTION

The wind speed and fetch are obviously important in creating wave action, so obviously butperhaps more important here when comparing the effect of the last storm and the one coming in, is the duration of the strong winds. This is for three reasons. First, because the wave height is usually more limited by duration of strong winds, than by fetch. Second, because setup will aggravate a wave condition, and it takes at least six hours to get to peak setup. Third, because the longer the wave action is great, the worse the erosion or flooding. Wave action is probably greatest in fall as the lake level is higher than in winter and spring, there is no reducing effect of ice, and little reducing effect of warm air over cold water.

SURGES (from squall-lines)

These occur in the warm season when lake levels are high. Large water level changes are rare, but can amount to 8 feet or so. They need not be associated with wave action and erosion. Generally squall-lines must be fast moving to get a significant surge. They are probably most common in Lake Michigan, but significant ones have occurred on all Lakes.

SETUP

Setup is clearly most common and of most concern in Lake Erie, especially with high water in western Erie, although this east wind condition is relatively rare. Setup is also quite significant in Saginaw Bay off Lake Huron. Any persistent wind will cause some setup of significance in these two locations, and water level changes of 4-5 feet at Buffalo with westerly winds occur yearly. Setup on the other Lakes is much less significant because of their great depth.

LAKE ST. CLAIR

Lake St. Clair is quite small compared to the Great Lakes and it is very much shallower (average normal depth about 10 feet). But the lake is 2-3 feet above normal, and this is a substantial percentage increase in the depth, so unusual conditions are produced not only by the high water level as on the Great Lakes, but by the change in depth as such. Waves cannot get very large because of the shallowness of the water, but here the increase in depth has had some effect in apparently allowing larger than average waves, and the effect of these waves is great in both erosion and flooding because of the flatness of the surrounding land. Surge induced water level changes are probably rare enough and small enough on such a shallow lake so as to be neglected.

Setup on the lake has the greatest potential for water level changes, because setup is greatest where the water level is least. Well documented cases of setup are lacking, however similar conditions exist in both Lake Okeechobee in Florida and Lake Ponchartrain in Louisiana. Both these lakes are very similar in depth,shape, and size to St. Clair. A devastating flood occurred at Okeechobee in September 1928 (see Monthly Weather Review for that month) when a hurricane caused setup of a reported 10-15 feet with sustained winds of well over a 100 mph. A similar situation occurred on Lake Ponchatrain in September 1965 (see Monthly Weather Review of February 1968--Goudeau and Conner) where a setup of 10 feet was well documented with winds again over 100 mph. Of course such winds are not believed possible over St. Clair, but this evidence clearly suggests that winds of 60 mph, which are rare but possible, could produce a setup of several feet on St. Clair. It is a rule of thumb in the New Orleans forecast office of the NWS* that the setup in Ponchartrain amounts to about 1 foot for each 10 mph of land observed wind. While this may be too high for the light winds, it is reasonable over a wide range of stronger winds in the experience of that office. The figure should be applicable to St. Clair if there is no ice in the lake, if the air temperature is <u>not</u> much warmer than the water temperature, and if the winds are moderate to strong.

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*Personal communication with Mr. Conner, MIC, New Orleans.

TECHNICAL ATTACHMENT 77-8

PROG K VALUES VS. THUNDERSTORM AND PRECIPITATION FREQUENCY

Based on work by
Perry Baker and Jerald L. Adams
WSFO Des Moines

The following table is based on the precipitation and thunderstorm occurrences in the 12-hour period FOLLOWING the valid time of the prog K index from the trajectory model, as reported in the FOUS 10 and 11 messages. These data are thus for the period 24-36 hours following the initial data time used by the numerical models, i.e. for the second period of the usual forecast.

The data sample was from three seasons as follows: May 13 - August 2, 1972; May 1 - July 25, 1973; April 1 - July 31, 1976. The precipitation events (measurable) were only at Des Moines airport, and thus the probabilities are point probabilities. The thunderstorm events were for a 50 nm radius around Des Moines and thus are areal probabilities. The thunderstorm determination involved use of the DSM regular surface observations and the DSM radar overlays. An observed thunderstorm, lightning, or a cb cloud reported in its vicinity would be a thunderstorm. From radar the determination was based on the radar operator's notations, the reported radar tops and the intensity (dB).

The table combines the day-valid and the night-valid 12-hour periods. The data were separated by periods, but the differences between them were considered to be no more than sampling variation, so only the combination is presented.

Prog "K"	Total Cases	Thunderstorm Occurrence	Precipitation Occurrence	Areal Coverage
≤ 0	126	5%	1%	20%
1-5	41	15%	5%	33%
6-10	65	29%	14%	48%
11-15	72	40%	22%	55%
16-20	68	47%	30%	63%
21-25	69	54%	31%	57%
26-30	77	77%	39%	51%
≥ 31	40	73%	68%	93%

There are a number of interesting points that come from this table. Note that the K index categories from 6 through 30 have essentially the same frequency in each, with a rapid drop off above 30, but evidently there are a number of "less-than-zero" cases contributing to the maximum there. Notice also that as the index gets above zero, the areal probability of thunderstorm (and thus precipitation) rises rapidly. Since there would reasonably be some showers without thunder in all K index ranges, the <u>areal</u> probability of <u>precipitation</u> would most likely rise ever faster than the thunderstorm occurrence percentages indicate.

The frequency of precipitation at a point obviously does not rise as rapidly, but a K index of 10 or more would indicate that some mention of precipitation should be in the body of the forecast. Notice that at K above 30 the precipitation frequency jumps rapidly and is nearly the same as the thunderstorm area probability, which leads to the next point.

Some time ago we devised the equation

$$P_p = C P_a,$$

where P_p is the point probability of precipitation, P_a is the areal probability, and C is the expected areal coverage if precipitation occurs. Since in the table the areal thunderstorm probability is P_a* and the point precipitation probability is P_p, we can get C from the proper ratio of these. These values are given in the right hand column. Note that the coverage is around 55% for K values all the way from 6 through 30, then it suddenly jumps markedly. This is saying that at a K of around 30 the character of the precipitation changes a lot, from a spotty shower condition at 30 and less to essentially a widespread precipitation at K of more than 30.

Getting back to the K values of 6 to 30, the fact that the areal coverage is essentially uniform says for these almost two-thirds of the cases, that when showers do occur they have about the same areal coverage, but at the lower K values they actually occur less often. Thus both P_a and P_p vary directly with K even though C is uniform with K in this range. This would say that <u>if</u> from some other means, such as satellite or radar data, as the time for the beginning of the period gets closer, one can be quite confident that showers will occur, i.e. the areal probability is near 100%, then the point probability to be used for the airport or any other point should be 50%. Another way to put it is that if the areal probability is easier to get a feeling for than the point probability, then one-half the areal probability is the best estimate for the point probability. This nearly uniform areal coverage is quite interesting.

It was a bit surprising that the night-valid period was so similar to the day-valid period that the two could be combined. However, DSM is definitely in the area with a nocturnal maximum of precipitation, as probably caused by the low-level jet. But for this reason one needs to be very cautious of using these statistics in climatologically different regions, especially since the present numerical models do not adequately model the LLJ.

The following is based not on actual numbers but on the feeling of the creators of the table from their work with the data. With a high K value, if the rain does not occur in the 12-hour period just after the time to which the K value pertains, it most likely would have occurred before the period started. Also, again with high K, it is more likely to have precipitation if the K value from the previous numerical run was lower than if it was higher than the latest; i.e. an upward trend of K is more favorable for precipitation than a downward one. Finally, 1976 was a dry season and the other two used were wet. However, 1976 did not give any appreciable bias, so the results may well be valid in a wide variety of conditions.

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*The areal probability of measurable precipitation would probably be a bit higher than shown, because some showers would probably occur without thunder. This would lower the areal coverage from that given in the table.

TECHNICAL ATTACHMENT 77-10

1951 FLOODS AND HEAVY RAINS IN KANSAS

Wayne E. Sangster, SSD

The floods at Rapid City and Big Thompson Canyon are fresh in our minds, but the floods of July 1951 in Kansas and Missouri are in the distant past. Not many forecasters are forecasting now who were forecasting then. I was a wheat farmer in south central Kansas (south of the brunt of the heavy rainfall) at the time, so I remember it well. Technical Paper No. 17 of the Weather Bureau on these floods was unearthed the other day while looking for something else in my office, and it stirred my interest. TP No. 17 stated "for some areas and durations the July storm was the greatest of record in the United States north of the 37° parallel." This may not still be true. Most of the Central Region lies north of the 37° parallel, which is the southern border of Kansas.

The rainfall causing these floods fell primarily at night (does that sound familiar?) over a three-day period. A tabulation from TP No. 17 of the frequency of hourly amounts of .50 inch (12.7 mm) or more for all Kansas stations from noon July 9 to noon July 13, 1951 was made with the following results (the precipitation shifted to extreme southern Kansas on the fourth day):

AM Hour ending LST

1	2	3	4	5	6	7	8	9	10	11	12
19	17	17	22	24	19	17	11	7	5	1	0

PM Hour ending LST

1	2	3	4	5	6	7	8	9	10	11	12
1	1	1	4	1	2	3	5	5	8	9	19

As can be seen 4 to 5AM had the highest frequency (24). No reported amounts of .50 inch or more occurred between 11 and noon. 87 percent of the heavy hourly amounts fell between 9 PM and 9 AM. Since this was a four-day period this statistic is significant.

The heaviest hourly amount at a recording gage was 2.60 inches at Ottawa (in east central Kansas) between 6 and 7 AM on July 11. The largest daily amount was 7.75 inches at Barnard (in north central Kansas) on July 11 (at official stations).

This was an obvious east-west stationary front--boundary layer wind maximum case, no stranger to experienced Plains forecasters (see maps from TP No. 17 at end of TA). The surface geostrophic wind and vorticity charts no doubt would be interesting for the case, if they could be produced easily.

To get a look at the monthly variation of very heavy rain, the frequencies of 5.00 inches (127 mm) or more in 24 hours for stations in Kansas from the beginning of record through 1946 by month were tabulated from "Climate of Kansas," published in 1948. They ranked as follows:

1.	September	66	6.	November	21
2.	June	58	7.	October	17
3.	August	53	8.	April	11
4.	July	44	9.	March	5
5.	May	27			

Had 1951 been included, July would no doubt have come up in the ranking. Five-inch rains have been reported in every month of the year except December, January, and February. June through September represents the big threat. The number one ranking of September may come as a surprise. April (the proverbial month of showers) is way down on the list, but November, usually considered a dry month is two steps higher. A tabulation for Dodge City, Wichita, and Kansas City reveals that the 24-hour climatological probability of .01 inch or more in 24 hours is a minimum in November, and a maximum in May. Yet May is only one notch above November in the ranking of 5-inch rains.

The lesson to be learned is--don't let down your guard against heavy rains in Kansas except in winter. Then you can be on the alert for heavy snow. Anybody want to transfer to San Diego?

This TA dealt with Kansas because the data were at hand. Other states could be studied if the data were readily available. Those interested might want to look at a paper by Dyck and Mattice, "A Study of Excessive Rainfall" in the October 1941 Monthly Weather Review. LFM's, VFM's, NGM's and numerical models yet to be invented will only partially solve the heavy rain problem. A good solution is probably going to require a mix of good dynamical models and sophisticated statistics (with feedback from statistics to the dynamics). Radar and satellites, of course, are good observational tools (as is a rain gage), which can be used to make short-term forecasts.

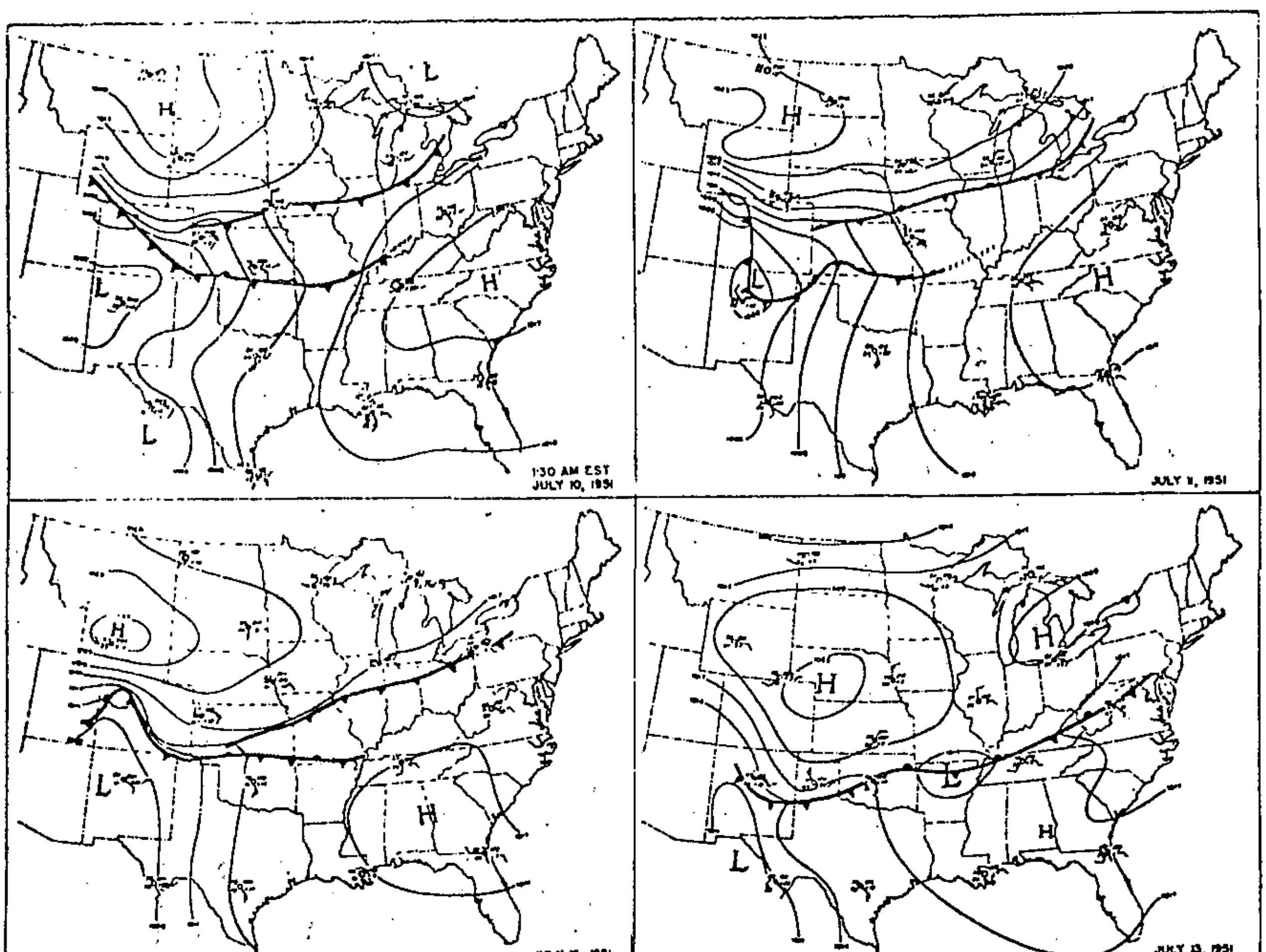


Fig. 28--Surface Weather Maps for Storm of 1951

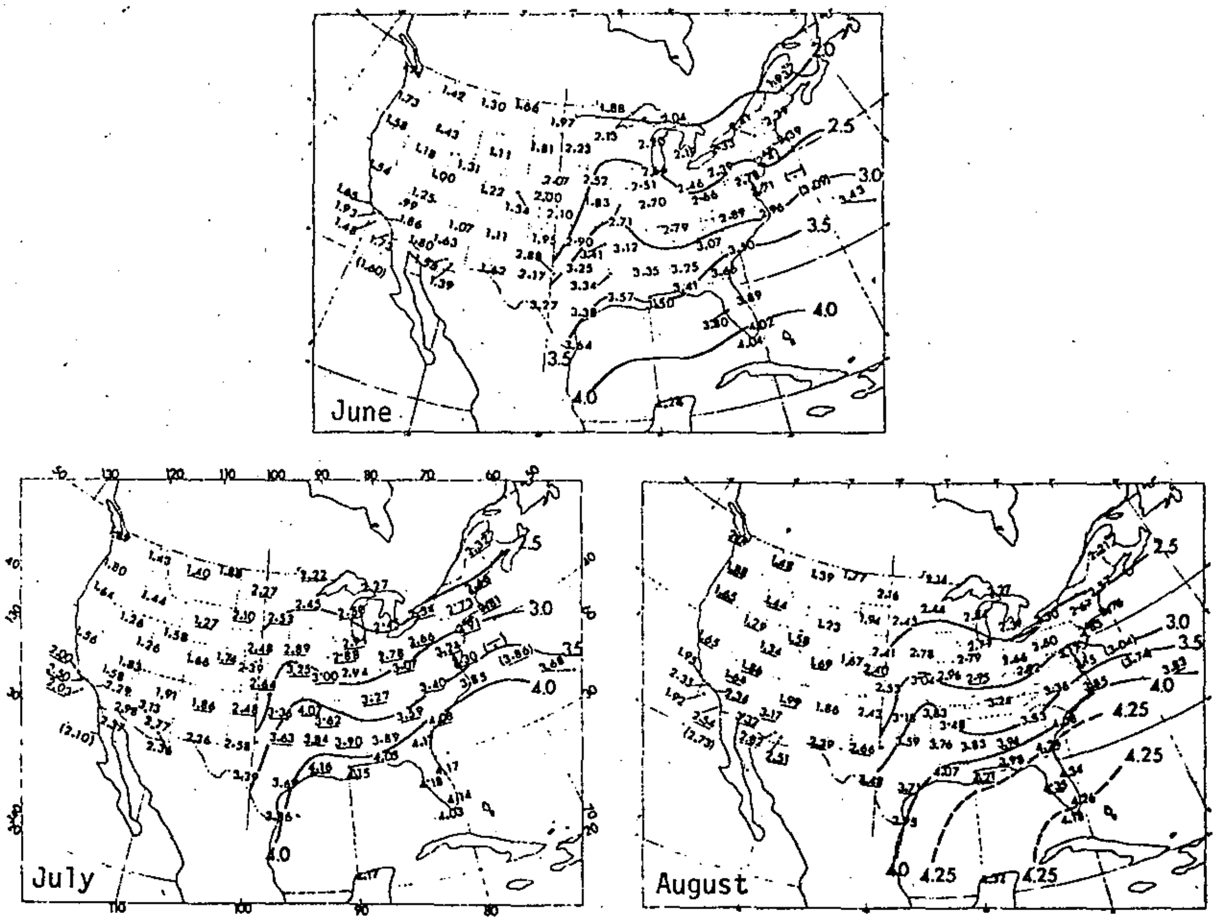
(Old-fashioned sea-level isobars.)

TECHNICAL ATTACHMENT 77-11

CLIMATOLOGICAL MOISTURE DATA
(for use in flash flood forecasting)

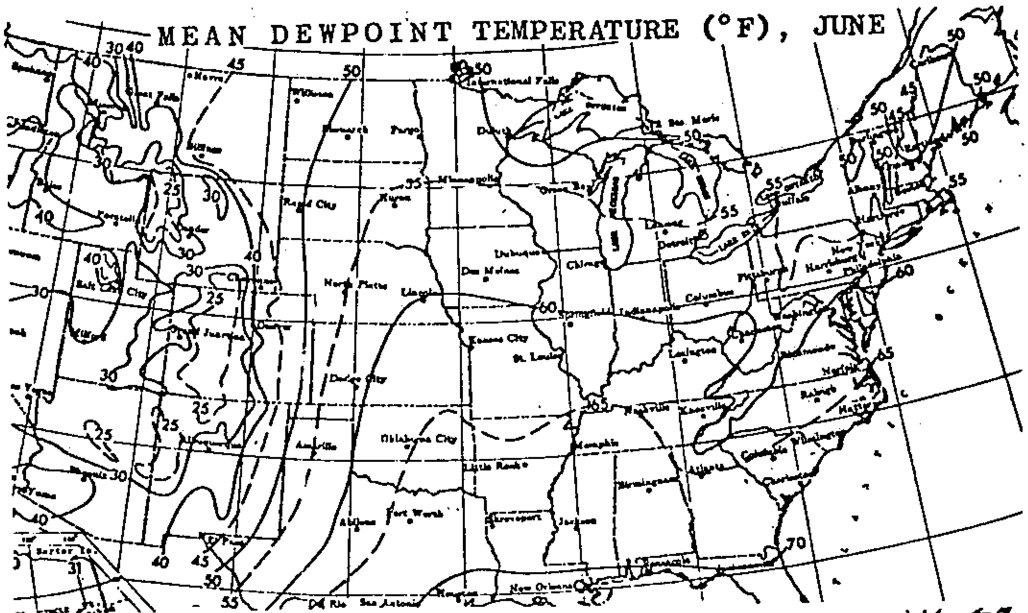
In our videotape comparing the Big Thompson Canyon and the Rapid City floods, it was noted that in both cases the surface dewpoints were well above normal, and also that moisture was high above the surface as well. To assist the forecaster in such an evaluation on a current basis, we present here the mean surface dewpoints and the mean surface-500 mb precipitable water for the months of June, July, and August. The dewpoint data (°F) are taken from the "Climatic Atlas of the United States", an ESSA and EDS publication of June 1968. The precipitable water data (cm) are taken from NOAA Technical Report NWS 20, "Precipitable Water Over the United States, Volume 1: Monthly Means." To convert the facsimile chart values from inches to cm, it will be adequate to multiply by 2.5 (actually 2.54).

The annual maximum in these two parameters tends to occur about mid-July. We present here only the three months centered on the maximum because monthly means drop off outside these months mainly because of averaging in more and more polar air conditions. In other months of the year one should consider that the moisture condition is high, and suitable for heavy rain, if it exceeds that of June or August, whichever is closer.

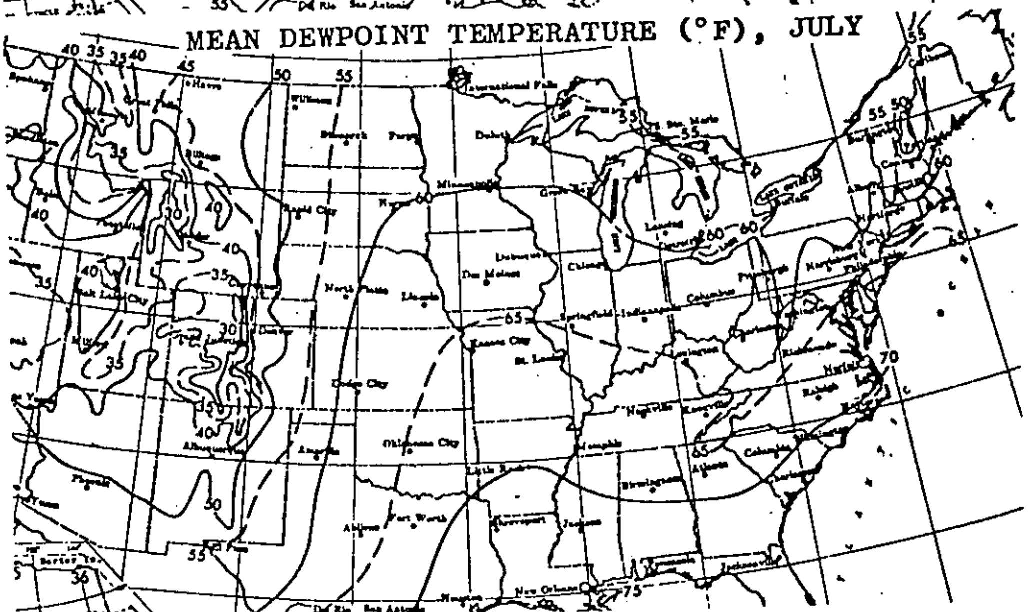


Mean Monthly Precipitable Water (cm), Surface to 500 mb.

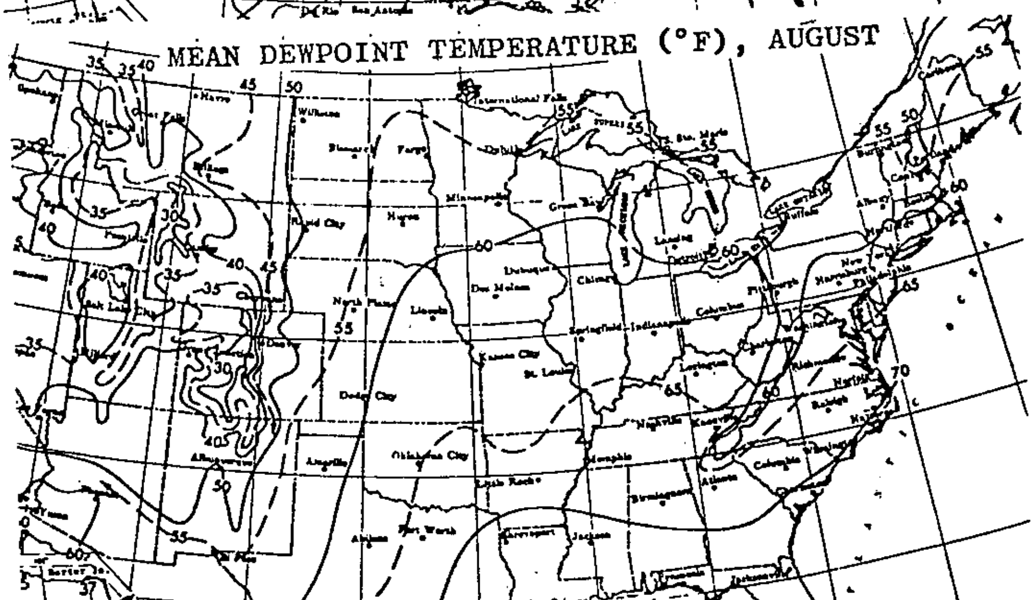
MEAN DEWPOINT TEMPERATURE (°F), JUNE



MEAN DEWPOINT TEMPERATURE (°F), JULY



MEAN DEWPOINT TEMPERATURE (°F), AUGUST



TECHNICAL ATTACHMENT 77-17

KANSAS CITY FLOOD OF SEPTEMBER 12-13, 1977

September 12, 1977, was very unusual in Kansas City, Missouri, because on that day there were two separate 6-hour rain events, each of which was easily equivalent to a once-in-a-100-year event, according to the values in Weather Bureau Technical Paper No. 40. While it is not rare in the Great Plains in the warm season for very heavy rain to occur on successive nights, it is extremely rare to have both events maximize at the same location, as happened at Kansas City. Not only were these two events in the same metropolitan area, but both maximized at the same point in the area (the center of the metropolitan area). This caused the second rain to fall on saturated soil and much paved surface and run off rapidly and fully, bringing about the main flooding.

Both events yielded about the same maximum rainfall observed at 6-7 inches in a 6-hour period. The first was midnight to 6AM, with the second from 6PM to midnight. The second event lasted more than 6 hours and gave the larger rain amounts. There was an observed maximum for the 30-36 hour storm period of almost 17 inches, but larger but unrecorded values probably occurred. There was sunshine in the afternoon between the two events, and there were sizable differences in the "cause" of the two events.

Both events had the same air mass, with the Composite chart on facsimile showing Lifted Indices below zero, K Indices at least in the mid-30s, mean relative humidities over 60% for the first event and over 70% for the second, and precipitable water well above normal. The surface dewpoints were also well above normal, and even above the normal of the month (July) when the normals are highest. All of these conditions are favorable for heavy rain from thunderstorms.

Figures 1 and 2 show the surface and 500 mb charts respectively for 12Z on the 12th (between the two events), as taken from the Daily Weather Maps series. The main features here are 1. the weak surface warm front, 2. the weak surface Low and cold front in Kansas forecast to move slowly toward Kansas City, and 3. the weak short wave trough aloft over Colorado also forecast to move toward Kansas City and provide lift by putting positive vorticity advection over Kansas City during the evening and into the night of the second event. The 500 mb winds were fairly high in both events, being around 40 knots.

The quantitative precipitation forecasts, both subjective and from the LFM, gave the general character of these heavy rains by having sizable amounts of precipitation in the Kansas City area and having precip. maxima near KC in a large scale sense. However, neither verified well enough in the smaller scale sense to truly indicate the magnitude of either event. Since the meso- or small-synoptic events were the key, let us look at these features.

Figure 3 shows the surface synoptic situation for the area at 18Z on the 11th (Sunday). These would be the basic data available to the forecaster for the afternoon forecasts for the event that occurred after midnight. The front and Low center positions were taken from the facsimile chart. The winds shown are the geostrophic winds from the so-called Sangster surface chart in the standard 4-digit format, as available to WSFOs on request-reply. The isotachs delineate the low-level jet axis. The maximum lifting due to warm advection (overrunning)

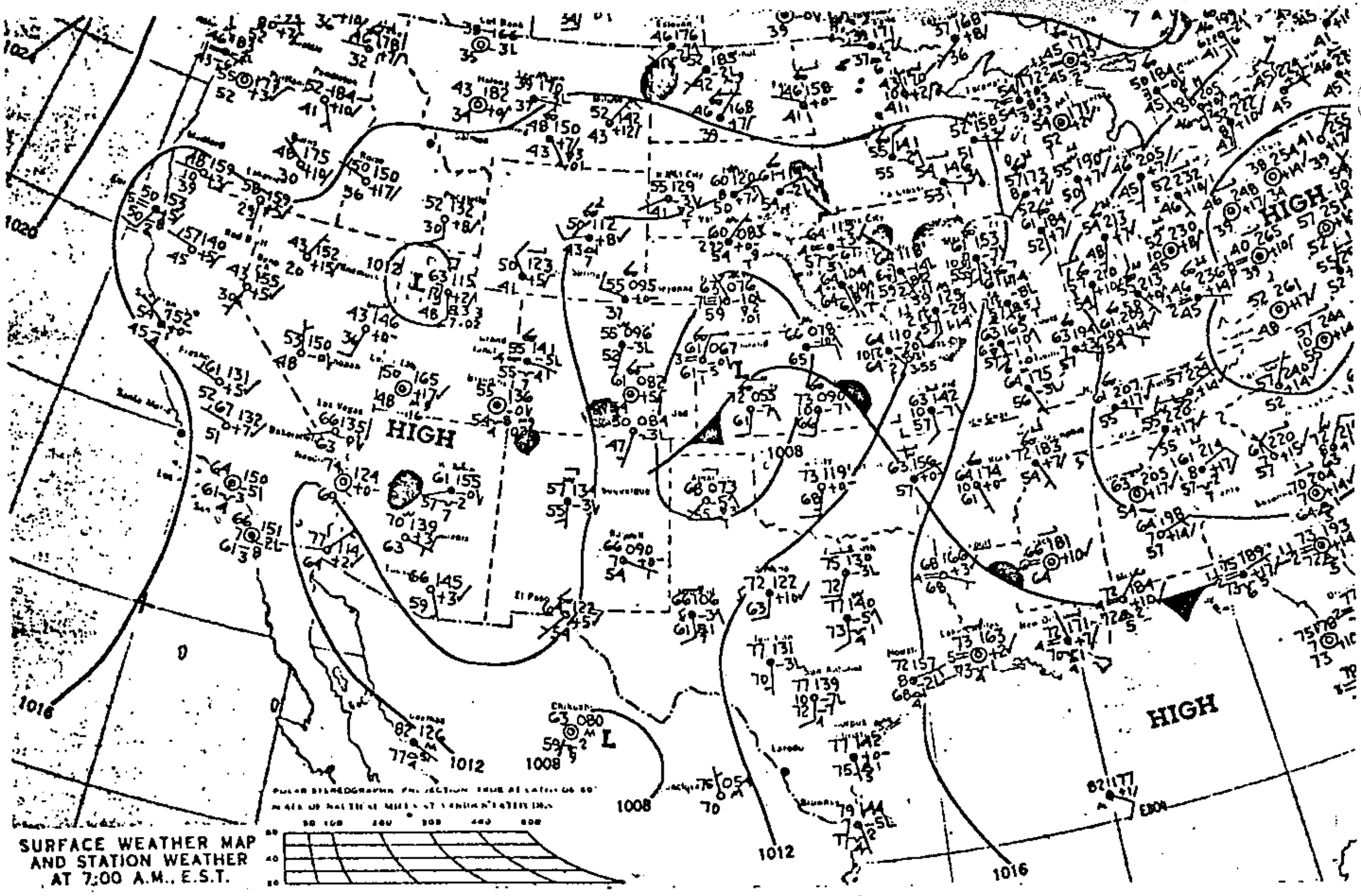


Figure 1. Surface Chart at 12Z September 12, 1977

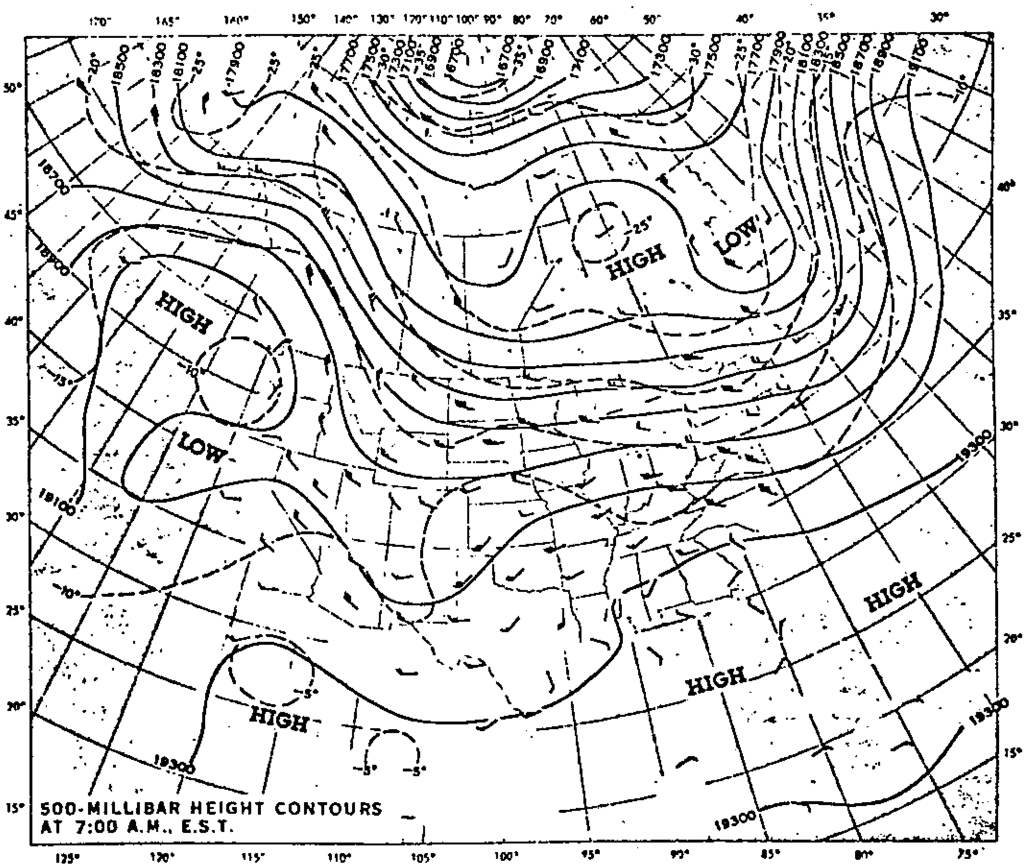


Figure 2. 500 Mb. Chart, 12Z Sept. 12, 1977

should be fairly close to this axis. Unfortunately, we don't have enough pibal winds to give this amount of detail to the real wind, but the real wind must adjust toward the geostrophic with time.

Shower activity initiated very late Sunday afternoon at points marked A & B on Figure 3. These merged to a stronger center at C a few hours later, and by midnight local time it was centered near Kansas City at point D. We know from knowledge of the real wind of the low-level jet (LLJ), that it increased well into the night, and that its direction changed to a more westerly component. Both these effects would increase the amount of warm advection over the warm front later in the night. Since the large scale effects were moving the geostrophic LLJ slowly eastward, it reached Kansas City about midnight local time at the beginning of the 12th, along with the rain.

The main rain swath of the first event lay from point C into Kansas City, parallel to the warm front, with the largest rain amounts occurring in Kansas City and where the LLJ would place the strongest warm advection during the night. The rain died off after daylight and the sun came out. The first event was thus mainly due to the LLJ impinging on a warm front.

The LLJ played a lesser role in the second event, as its axis continued to move eastward through the day, while taking on a more westward (downslope) component, so by 24 hours after Figure 3, it was across southern Missouri, where it was more parallel to the front and the arc cloud mentioned below.

The second event was more complex. As mentioned earlier, a short wave trough at 500 mb was moving in, giving PVA aloft and presumably associated upward motion over eastern Kansas and eventually western Missouri. In addition, the cold front was advancing across Kansas (see Figure 4 for the geostrophic winds, isotachs the front and the trough for 24 hours after Figure 3). The boundary of the rain cooled

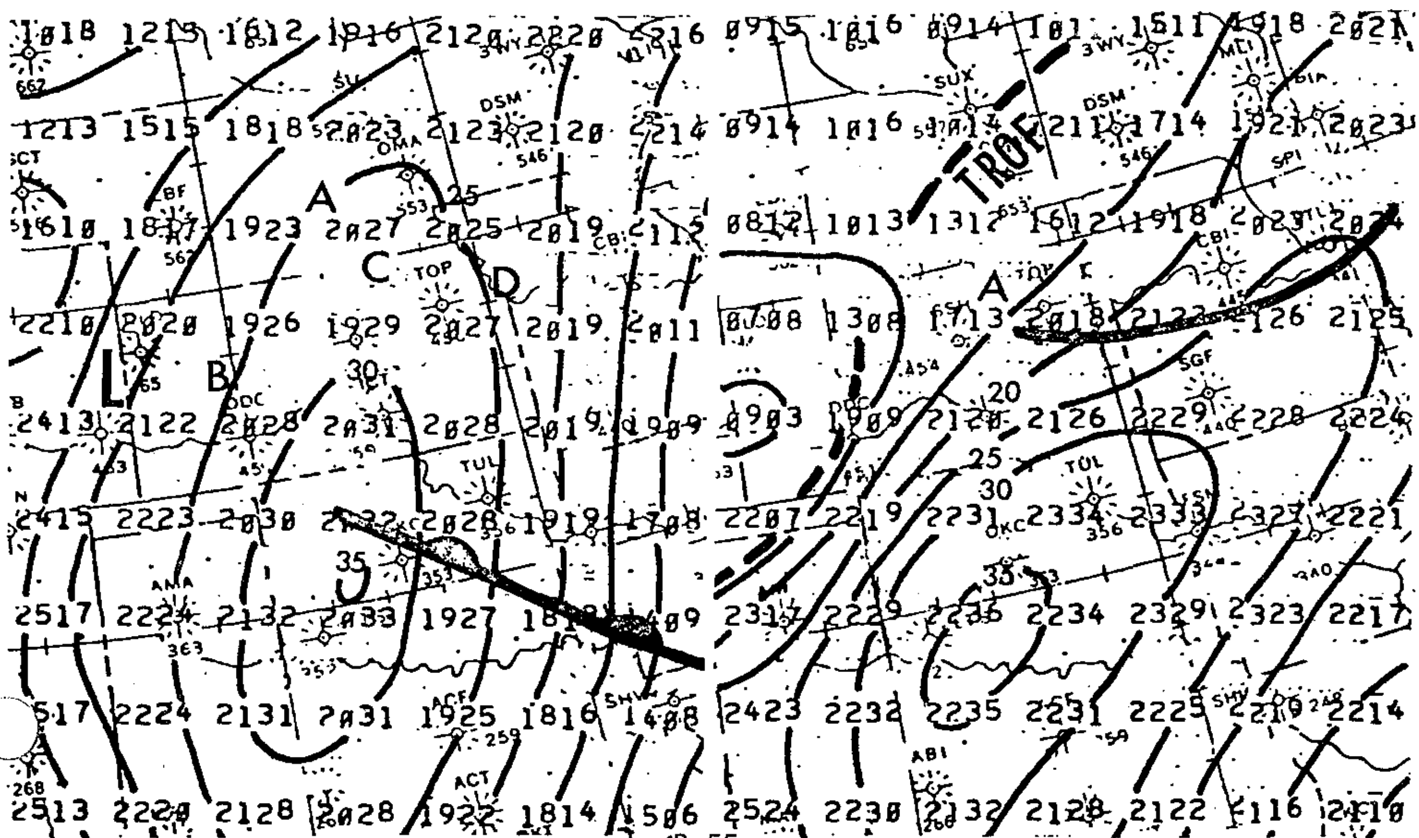


Figure 3. 18Z Sept. 11, 1977

Figure 4. 18Z Sept. 12, 1977

air from the first event was clearly delineated across central Missouri by the typical arc cloud seen in the visual satellite picture and given on Figure 4. The arc line was taking over in at least eastern Kansas as the boundary of significance instead of the warm front still carried by NMC but not by SELS. The activity initiated again late in the afternoon, but being closer to KC, (point A on Figure 4), the activity reached KC earlier. The initiation of showers was in the vicinity of the intersection of the arc cloud with the advancing cold front, and this activity moved along the arc line as the cold front advanced, while also breaking out along the front southwest of the initial activity. The heaviest rainfall remained along the arc line, as is usual, and the intense rainfall ceased at KC about the time the cold front and intersection passed the area. It continued on eastward for several hours, but decreased considerably late at night as the low-level flow became more parallel to the arc line.

We, in the Central Region, have heavily stressed both the LLJ and intersections as seen in satellite data. In this case both events were near intersections. The first near the intersection of the LLJ axis with the warm front--the maximum overrunning, and the intersection of an arc line with an advancing cold front. Note also that in the second event there was a merging of cells seen in the satellite pictures (A + B = C in Figure 4), which the NESS people have stressed in our satellite workshops as being associated with very heavy rain.

In the Great Plains it is not uncommon for a flood in which the low-level jet is a prime factor to be brought about by rains on more than one night. The flood in July 1951 in Kansas and that around July 1, 1975 in North Dakota are two that come easily to mind. Both had at least three consecutive nights of heavy rain. However, the 1977 KC flood was unusual even in that sense, since the heaviest rains both nights were at almost exactly the same location, and this is the point that made it so devastating. With drainage basins as small as those involved, a shift of the heaviest rain of either event of only a few miles (so they weren't coincident) would have considerably alleviated or even prevented the flood. Another detail of significance is that the axis of radar echoes in the second event lay WSW-ENE which is along the line of Brush Creek and the other small basins in the KC area, and along the line of radar echo movement. This allowed several thunderstorms to traverse the basins at speeds a bit higher than that generally associated with very large rainfall amounts, as the line moved slowly across the basins. Thus, the higher than usual winds aloft were not as critical as they would have been with different orientation of the line. These details are on a scale that cannot be predicted. This is the reason that all flash flood warnings cannot be expected to verify as well as this one, and the very serious extent of this flood could not be fully predicted.

A final point. An examination of the publication "The Climate of Kansas" which was reported in Central Region Technical Attachment 77-10 this past July, indicated that September is the month of maximum frequency of heavy rains of 5" or more in 24 hours from the start of record through 1946. The next highest month was June. September may be the peak because the summer moisture and instability are still around, while some large-scale events are starting to take place. This was certainly the case for Sept. 12, 1977 around Kansas City.

APPENDIX C

PRIME FORECAST-RELATED TECH. MEMOS

1.	TM CR 52	Cold Air Funnels
2.	TM CR 39	A Synoptic Climatology of Blizzards on the North-Central Plains of the United States
3a	TM CR 38	Snow Forecasting for Southeastern Wisconsin
3b	TM CR 30	An Aid in Forecasting Significant Lake Snows
3c	TM CR 48	Manual of Great Lakes Ice Forecasting
3d	TM CR 21	Wind Waves on the Great Lakes
4a	TM CR 2	A Study of Summer Showers Over the Colorado Mountains
4b	TM CR 3	Areal Shower Distribution-Mountain Vs. Valley Coverage
4c	TM CR 43	Summer Shower Probability in Colorado as Related to Altitude
4d	TM CR 57	Summer Radar Echo Distribution Around Limon, Colorado
4e	TM CR 61	An Updated Objective Forecast Technique for Colorado Downslope Winds
5	TM CR 58	Guidelines for Flash Flood and Small Tributary Flood Prediction