The Causes of Severe Convective Outbreaks in Alberta. Part I: A Comparison of a Severe Outbreak with Two Nonsevere Events

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ABSTRACT

Mesoscale and synoptic-scale analyses were carried out for a severe convective outbreak and two nonsevere convective events in central Alberta. High-resolution upper-air and surface observations gathered during the Limestone Mountain Experiment (LIMEX-85) permitted a detailed diagnosis of the evolution of the atmosphere over the Alberta foothills. On the severe day, deep convection was triggered when upper-level cooling, associated with an advancing, synoptic-scale trough, occurred in phase with strong surface heating over the Alberta foothills from 0800 to 1200 local daylight time (LDT). The deep destabilization over the elevated topography acted to amplify the mountain-plain circulation and to generate mesoscale uplooking moisture transport. Concurrently, the surface synoptic pressure gradient gave rise to northeasterly winds that advected a tongue of moist plains air into the lower branch of the mountain-plain circulation. The plains moisture was thus permitted to reach the foothills in time to reinforce the initial convection and effectuate a secondary destabilization. On the nonsevere days, the absence of such joint meso-synoptic-scale uplooking moisture transport precluded the occurrence of severe convection.

1. Introduction

a. Background

Central Alberta\(^1\) is a region highly susceptible to severe summertime convection. Climatological statistics show that the area is affected by hail on an average of 61 days each summer (Wojtiw 1975), and between 10 and 20 tornadoes are reported annually (Bullas and Wallace 1988; see also Newark 1984). Although most of the hail comes from relatively weak single- and multicell storms, highly organized multi- of supercell storms develop three to five times per summer (Smith and Yau 1987) and produce widespread large hail and/or tornadoes. In particular, the Edmonton tornado of 31 July 1987 (F4 scale; Fujita 1981), which left 27 people dead and 300 injured and caused 250 million dollars of property damage (Bullas and Wallace 1988), demonstrated the vulnerability of the population to these organized outbreaks of severe convection and underscored the need for improved forecasting, public awareness, and warning dissemination of such events.

A number of conditions have generally been found to be associated with the occurrence of severe convection (see Fawbush et al. 1951; Newton 1963; Ninomiya 1971; Miller 1972; Browning 1986). They include the existence of the atmosphere of

1) a large amount of latent energy,
2) strong convective instability ($\partial \theta_v \partial z < 0$),
3) large vertical shear of the horizontal wind, and
4) a mechanism capable of releasing the latent energy.

Current forecasting of severe convection in Alberta basically consists of first identifying conditions 1, 2, and 3 using synoptic data, guidance from numerical weather prediction models, and satellite imagery. If the potential exists for severe weather, an attempt is made to determine condition 4 with the aid of experience, empirically based methods (e.g., Miller 1972), and/or computer-based expert systems (Bullas et al. 1990). Very short term forecasts of less than 6 h (nowcasts) rely heavily on observations from weather radar and satellites. In North America, these forecasts often take the form of severe weather watches or warnings.

Conditions 1 and 2 are usually satisfied by the observation of the so-called "loaded gun" sounding de-
Single-cell storms show light winds with little wind shear; those for multicell storms have moderate unidirectional shear; and those for supercell storms are distinguished by large shear concentrated in the lower levels of the troposphere. Observations such as these suggest that the vertical wind shear plays a role in determining the form that convective storms may take. Weisman and Klemp (1986) describe how the presence of vertical wind shear may increase the ability of gust fronts to trigger new convective cells and how it may interact with a cumulonimbus updraft to produce the enhanced, quasi-steady storm structure typical of supercell storms.

A common signature of the soundings in areas where severe convection eventually occurs is a layer of strong static stability located in the lower troposphere. This layer or capping lid prevents the formation of deep convection but allows the buildup of latent energy through an increase in boundary-layer heat and moisture contents (Fulks 1951; Newton 1963). A crucial question in forecasting severe convection is the timing and location of the breakdown of the capping lid, or in other words, when and where is condition 4 satisfied? In general, the lid can be removed by any combination of 1) differential temperature advection, 2) adiabatic lifting, 3) surface heating, 4) low-level moisture-flux convergence, and 5) evapotranspiration. The first two mechanisms act principally to modify the temperature structure of the lid, while the remaining three can be thought of as increasing the buoyancy of surface air parcels. As late as the 1960s, lid breakdown by surface heating alone was not thought to occur, and forecasters focused on identifying other mechanisms, such as the passage of a front, which could remove the lid by lifting (Newton 1963).

Several researchers have found that migrating upper-level jet streaks (wind maxima) and the associated divergence patterns can generate the lifting needed to remove the capping lid and trigger severe convection (e.g., Beebe and Bates 1955). Owing to the upward motion present there, areas located beneath either the right-rear entrance or left-front exit regions of jet streaks are considered to be favorable for the occurrence of severe convection. Following Uccellini and Johnson (1979), Carlson et al. (1983) proposed that jet streaks induced underrunning—a process whereby moist boundary-layer air flows out from underneath the capping lid—and were responsible for severe convective outbreaks on three separate occasions during SESAME (Severe Environmental Storms and Mesoscale Experiment). They showed from surface pressure tendency analyses that underrunning appeared to take place in locations where the low-level flow was aligned with the isallobaric wind. The strongest underrunning was induced by a migrating isallobaric couplet associated with an upper-level jet streak.

In a general sense, the goal of this study is to determine how the four conditions for the occurrence of
severe convection are usually met in Alberta. A brief review of previous studies in the next section will allow us to focus our goal considerably.

b. The Alberta problem

On days with severe convection in central Alberta, the morning soundings generally show a low-level tropospheric inversion or capping lid that initially inhibits the formation of deep convection (Fig. 2). Towering cumulus clouds first form over the foothills region\(^2\) in the early afternoon. They intensify rapidly into cumulonimbus, which move eastward.

Longley and Thompson (1965) attempted to isolate the important factors leading to severe convection and large hail in Alberta. They presented a series of mean 500- and 850-mb maps for major, minor, and no-hail days. Figure 3 shows that the mean 500- and 850-mb maps at 0600 LDT (1200 UTC) for days with major hail resemble those of a developing baroclinic wave. This finding led to the suggestion that the presence of a cyclone over southern Alberta is a feature characteristic of major hail days.

Strong (1986) presented evidence for vigorous dynamic forcing in the upper-level flow pattern on days of severe convection. He proposed that easterly boundary-layer flow due to surface cyclogenesis induces underrunning in the foothills. Strong argued that this underrunning acts to initiate deep convection in the area of the foothills where the lid has been most strongly weakened by large-scale ascent. Thyer (1981) investigated the potential role of the thermally induced mountain–plain circulation. Under clear sky conditions in the summer, an upslope wind regime is present in west-central Alberta because of the differential heating of topography (Longley 1968, 1969). This solenoidal circulation may act in concert with the synoptic-scale pressure gradient to give rise to severe convection. Thyer demonstrated the presence of both the upper

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\(^2\) This region is defined by the topographic ridge parallel to, and roughly 50 km to the east of, the Continental Divide (i.e., the border between Alberta and British Columbia). The mean elevation of the peaks in this ridge is around 1500 m.
Fig. 3. Mean (a) 500-mb and (b) 850-mb maps at 0600 LDT for days with major hail. Solid lines are heights in meters with a contour interval of 50 and 25 m for 500- and 850-mb maps, respectively. Dashed lines are isotherms with a contour interval of 2°C. Temperature to nearest tenth of a degree plotted to the left of height with a blank space for decimal point. Dewpoint temperature to nearest tenth of a degree plotted below temperature on 850-mb map (from Longley and Thompson 1965).

and lower branches of the mountain–plain circulation but failed to find any significant correlation between its magnitude and the occurrence of hail. Regardless of the source of the easterly component of the flow, its importance is documented by Smith and Yau (1987) who found, using a well-mixed boundary-layer model, that topographically induced upward motion is maximized in the Alberta foothills when the upshe flow has an easterly component.

Considering the comparable regional topographies, it is reasonable to expect that factors that give rise to severe convection in the High Plains of the United States (i.e., Colorado, Wyoming, and Montana) may be similar to those operating in central Alberta. It is therefore of value to consider Doswell’s (1980) synoptic composite, constructed mainly from an examination of synoptic maps (surface, 850-, 700-, and 500-mb levels) and tropopause–wind maxima charts on 30 severe weather days during June and July of 1979 over the High Plains. The major features in his composite include:

1) an upper-level trough upstream of the region where the convection is initiated,
2) southwesterly winds of at least 10 m s⁻¹ at 500 mb,
3) an easterly component of the surface flow arising from the synoptic pressure gradient, and
4) a low-level inversion.

These features are quite similar to those suggested by Strong (1986) and Longley and Thompson (1965). Although he did not have mesoscale observations to confirm it, Doswell felt that the increasing easterlies on days of severe weather observed by Modahl (1979) are the result of the surface synoptic circulation being enhanced by the diurnal upslope flow.

c. Statement of the problem

The evidence presented above suggests that the forcing provided by the synoptic environment is necessary, and perhaps sufficient, condition for the occurrence of severe convection in Alberta. Due to the coarse resolution of the observations, however, the forcing provided by the mesoscale environment, in particular that of the mountain–plain circulation, has not been investigated adequately. In this study, we will examine the interactions between the mesoscale and synoptic-scale prestorm environments over Alberta to provide answers to the following questions.

1) What role, if any, does the mountain–plain circulation play in the initiation of severe convection in Alberta?
2) What is the significance of the synoptic setting described by Longley and Thompson (1965) for major hail days? Is the existence of this setting a necessary
and sufficient condition for the initiation of severe convective outbreaks?

3) When, where, and how is the capping lid first weakened or removed on days of severe convective outbreaks in Alberta?

Our approach will be to carry out a detailed analysis of a high-resolution dataset collected in a mesoscale experiment in Alberta [Limestone Mountain Experiment 1985 (LIMEX-85)]. We shall focus on the evolution of the synoptic and mesoscale features, with special emphasis on three case days: 11 July 1985, a day of severe convection; 9 July 1985, a day of weak, isolated convection; and 17 July 1985, a day of widespread, nonsevere convection. Our results yielded a coherent picture of the interactions between the synoptic and mesoscale circulations that lead to severe convection.

Section 2 describes LIMEX-85, the database, and the method of analysis. The three case days are presented in sections 3 and 4 and discussed in section 5. In Part II of this study (Smith and Yau 1993), all 11 LIMEX-85 days are intercompared and a conceptual model of severe convective outbreaks based on this comparison is proposed.

2. Observational data and methods of analysis

LIMEX-85 was carried out during 8–23 July 1985. Its primary objective was to obtain accurate surface and upper-air data of high spatial and temporal resolution over the Alberta foothills on days exhibiting a capping lid. Figure 4 shows the LIMEX-85 observing network (see appendix A for station identifiers). Nine upper-air radiosonde and Airsonde stations provided soundings at 2-h intervals from 0800 to 1800 LDT. The surface network included 8 automated stations (CR-21 units manufactured by Campbell Scientific In-

3 The Airsonde (TS-3A ADAS) is an upper-air sounding system operating at frequency of 403 MHz. The signal strength of the system is weak at high altitudes, thus providing data only up to around 300 mb. Reliable data were, however, occasionally obtained beyond 150 mb. Warren–Knight optical theodolites were modified to provide upper-level winds in conjunction with the Airsonde ascents.
instrument Company), which recorded 5-min averages of temperature, relative humidity, and winds; 13 forestry service stations, which provided manual observations twice daily at 0830 and 1300 LDT; and the 33 regular surface-observation stations of the Atmospheric Environment Service of Canada (AES; see Fig. 1). A 10-cm, S-band radar located at Red Deer (AQF) was operated by the Alberta Research Council throughout the experiment. The digitized reflectivity data in plan position indicator (PPI) format, along with both polar-orbiting and geostationary satellite imagery, were used to determine the location and intensity of convection and to verify cloud observations reported by the manual observing stations.

In total, the full observational network was in operation on 11 days during LIMEX-85, and hail fell in the project area on nine of these days. Following the hail severity classification of Smith and Yau (1987; see appendix B), 5 days were classified as light hail days, 3 as moderate, and 1 (11 July 1985) as severe. On the remaining 2 days, only cumulus and towering cumulus clouds were present. A total of 449 soundings were released during the entire experiment.

Mesoscale and synoptic-scale analyses were carried out for each case day. The subjective synoptic-scale analyses of mean sea level (MSL) pressure, surface temperature, surface dewpoint temperature, and 2-h surface temperature and pressure tendencies cover all of Alberta. The data that went into the analyses include observations from the AES operational network and the surface measurements at the upper-air stations within the LIMEX-85 area. It should be pointed that the MSL pressure gradient in regions of elevated and/or sloping terrain is subject to considerable uncertainties. To check the accuracy of our MSL pressure analyses, a corresponding 850-mb height analysis was performed. It was found that the minima and maxima in MSL pressure were located within 30 km of the corresponding minima and maxima in 850-mb height. Since the actual magnitude of the MSL pressure gradient was not crucial to our arguments, no attempt was made to improve on its accuracy.

The objective mesoscale analyses cover an area depicted by the rectangle in Fig. 1. Horizontal cross sections of capping-lid strength, dewpoint temperature, pressure, temperature tendency, and convective available convective energy (CAPE) tendencies with

$$\text{CAPE} = R_d \int_{\psi_1}^{\psi_2} (T_{vp} - T_{ve}) d(\ln p)$$

where

$$R_d = \text{gas constant}$$
$$T_{vp} = \text{virtual temperature of a lifted parcel}$$

Fig. 5. AES operational 500-mb analysis at 1800 LDT 11 July 1985. Height and 500-1000-mb thickness contours (dam) are given by solid and dashed lines, respectively. Observations are plotted following conventional upper-level station model.

$$T_{ve} = \text{virtual temperature of environment}$$
$$p = \text{pressure}$$

were constructed at 1- or 2-h intervals for each of the 11 case days. The irregularly spaced data were first interpolated onto a regularly spaced (0.14°) 12 × 14 grid mesh using a version of the Barnes (1973) scheme. The Gaussian weighting function used decreases from unity at zero distance from the grid point to a value of 0.1 at the average distance of the four nearest stations. To determine the adequacy of the analysis scheme, some objectively analyzed fields were compared with the corresponding subjectively analyzed fields. Good agreement was found. Test analyses were also made using a coarser (0.28°) 6 × 7 grid mesh and the results were found not to deviate significantly from the fine-mesh analysis.

In all the analyses involving upper-air and sounding-derived variables, data from the eight upper-air stations were used. The distribution of stations is fairly uniform except in the southwest corner of the grid near WBA. Fortunately, our conclusions do not depend critically on upper-air data in this corner.

Data from all the surface stations shown in Fig. 4 were used in our surface analysis except after 1300 LDT, when forestry station data were no longer available. To construct vertical cross sections, data were interpolated linearly onto a grid with a horizontal and vertical grid lengths of 13.8 km and 10 mb, respectively.

3 Because many of the soundings released at station ABP were terminated below 500 mb, along with the fact that it was the only station located in rugged mountainous terrain, it was not included in the objective analyses.

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* Only data from 5 of the 13 stations were used in the analyses (see Fig. 4 for station locations).
Sounding-derived variables such as lifting and convective condensation levels (LCL and CCL), CAPE tendency, etc., were determined numerically using algorithms given in Stackpole (1967). Surface quantities required in these calculations were computed by averaging the variables in the lowest 30 mb of the sounding.

3. **A case study of a severe convective outbreak—11 July 1985**

   a. **Introduction**

   This day was marked by well-organized multicell storms that began moving out of the foothills around 1600 LDT. Radar reflectivities between 60 and 70 dBZ were present in these storms over the course of several hours. A total of 228 hail reports were received, and eight of these were for hail of golfball size (3.3–5.2 cm). Soundings were released every 2 h from 0800 to 2000 LDT.

   b. **Synoptic-scale analysis**

   The presence of a 500-mb trough upstream of central Alberta and a ridge downstream at 1800 LDT 11 July 1985 (Fig. 5) indicated favorable conditions for large-scale ascent over the region. The overall flow pattern resembled the Longley and Thompson (1965) mean 500-mb map shown in Fig. 3. Wind observations point to the presence of an upper-level jet streak with speeds of 25 m s\(^{-1}\) extending from south-central British Columbia through central Alberta on into western Manitoba. The southeastward displacement with time of the 558-dam 1000–500-mb thickness line (not shown) is consistent with a net cooling over central Alberta associated with the passage of the upper-level trough from 10–12 July.

   At the surface, a low pressure center in southeastern Alberta and a high toward the northwest dominated the MSL pressure field from 0800 to 1800 LDT (Fig. 6). During the day, the low deepened while the high

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**Fig. 6.** Subjectively analyzed MSL pressure and surface dewpoint temperature fields at 0800, 1200, and 1800 LDT 11 July 1985. Wind observations show station locations. Winds plotted using conventional speed scale. Contour interval is 0.5 mb and 2°C. Rectangle denotes LIMEX-85 observation area. The H's and L's designate local maxima and minima.
pushed southeastward. The resultant northeasterly flow channeled the moist plains air to the foothills along the northern edge of the LIMEX area. The influx of moisture is indicated by the westward migration of the 12°C dewpoint contour, which shows a well-defined moisture tongue extending from the Alberta–Saskatchewan border up to foothills northwest of the LIMEX area at 1800 LDT. The most intense midlevel storms were formed in the foothills and moved eastward along the axis of this moisture ridge where dewpoint temperatures exceeded 12°C.

c. Mesoscale analysis

1) WEAKENING OF CAPPING LID BY SURFACE HEATING

The mesoscale analysis will focus on the evolution of the capping-lid strength, defined as the difference between the potential temperature at the surface and the potential temperature at the CCL (see Fig. 2):

$$\Delta \theta = \theta_{CCL} - \theta_{SFC}.$$ 

A state of free convection is defined by $$\Delta \theta = 0.$$ A large positive $$\Delta \theta$$ indicates a strong inversion that inhibits deep convection. Because this definition is independent of the processes of formation, it can include lids formed by the advection of an elevated mixed layer (Farrell and Carlson 1989) or by other mechanisms such as radiation, subsidence, fronts, etc.

Time traces of $$\theta_{CCL}$$ and $$\theta_{SFC}$$ were plotted for each upper-air station on each case day. Figure 7 illustrates two examples with the corresponding cloud and weather observations (see appendix C for symbol definitions). In mid-July, sunrise occurs around 0500 LDT and sunset at 2100 LDT. The $$\theta_{SFC}$$ trace for LMW, a station in the foothills with an elevation of 1506 m, shows a rapid rise from 0800 to 1200 LDT and is almost constant thereafter. In contrast, $$\theta_{CCL}$$ decreases from 0800 to 1000 LDT and from 1200 to 1600 LDT. As will be shown later, this decrease was caused by an increase in low-level moisture, which lowered the CCL and thus the $$\theta_{CCL}$$. Similarly, a moisture decrease accounted for the increase in $$\theta_{CCL}$$ between 1000 and 1200 LDT.

The trace for ARM, a station located on the plains with an elevation of 988 m, displays a slower rate of surface potential temperature increase. Here $$\theta_{CCL}$$ remains quite uniform up to 1600 LDT. Towering cumulus clouds (TCU) were observed at LMW and cumulonimbus clouds (CB) at ARM at or after the time when $$\theta_{CCL} \approx \theta_{SFC}$$. It is evident that strong surface heating had caused a rapid erosion of the lid at LMW between 0800 and 1000 LDT. A similar weakening of the lid occurred over ARM but at a slower rate.

The evolution of the spatial distribution of capping-lid strength is displayed in Fig. 8. The locations of the upper-air stations are denoted by an asterisk. For the sake of clarity, only two of the surface stations (FMH and WBA) are plotted ( symbol). The foothills are oriented parallel to a line connecting ARL, LMW, AML, and FMH. Note that the greatest decrease in lid strength occurred over the foothills from 0800 to 1000
Fig. 8. Analyzed field of capping-lid strength at (a) 0800, (b) 1000, (c) 1200, and (d) 1400 LDT for 11 July 1985. Contour interval is 1 K.
Fig. 9. Analyzed fields of surface wind and (a) 2-h temperature tendency and (b) pressure tendency at 1000 LDT and boundary-layer depth at (c) 0800 and (d) 1000 LDT. Contour intervals are 1°C (2 h)^\(-1\), 0.2 mb (2 h)^\(-1\), and 100 m. For surface wind vectors, one east-west grid length is 10.6 m s\(^{-1}\).
Fig. 10. Analyzed field of surface wind and dewpoint temperature at (a) 1000, (b) 1200, and (c) 1400 LDT for 11 July 1985. Contour interval is 1°C. Wind vectors as in Fig. 9.
LDT and is followed by a slower rate of decrease out to 1400 LDT. Lid erosion was much more gradual over the plains. By 1400 LDT, a 3-K lid was still present above ACR and AEL. Observations of TCU at ARL, ACR, and AQF suggest that although the lid strength was positive everywhere except at AYC, the lid had been removed at several locations but on a scale too small to be resolved by the observing network. The lack of deep convection at AYC was due to the absence of a large amount of CAPE (300 J kg\(^{-1}\) at 1400 LDT).

Strong surface heating over the foothills (Fig. 9a) was the major cause of lid reduction. Heating was especially evident near LMW where the depth of the adiabatic boundary layer increased by 2 km (2 h\(^{-1}\)) during the same period (Figs. 9c,d). Regions of maximum pressure fall were nearly collocated with areas of max-
minimum temperature rise (Fig. 9b), a signal of the hydrostatic response of the atmosphere to surface heating.

2) **Underrunning and Low-level Moisture Transport**

In response to the pressure falls, surface winds over the plains veered appreciably, particularly from 1200 to 1400 LDT (Fig. 10). The resultant upslope flow associated with the lower branch of the mountain–plain circulation began transporting the moist air to higher elevations as early as 1200 LDT. By 1400 LDT, large dewpoint temperature increases (as much as 8°C in 2 h) were observed over much of the foothills as the moist air continued to flow upslope beneath the capping lid. An infrared satellite image taken at 1534 LDT displays the line of TCu (see arrow in Fig. 11) that formed along the foothills as a result of this underrunning process. For the most part, the strongest convective development seems to have been initiated over the foot-
Fig. 13. Analyzed fields of 500-mb wind and divergence at (a) 1400 LDT, (b) 500-mb, 2-h temperature tendency at 1200 LDT, (c) CAPE tendency below 300 mb at 1000, and (d) 1400 LDT for 11 July 1985. Contour intervals are 0.2°C (2 h)$^{-1}$, 5 × 10$^{-5}$ s$^{-1}$, and 100 J kg$^{-1}$ (2 h)$^{-1}$. For 500-mb wind vectors, one east–west grid length is 42.3 m s$^{-1}$. 
Fig. 14. Analyzed vertical cross sections of (a) 4-h temperature tendency, (b) CAPE tendency, (c) height tendency, (d) and $u$ component of the wind tendency ending at 1200 LDT 11 July 1985. Dashed contours show negative tendencies. Contour intervals are $1^\circ\text{C} (4 \text{ h})^{-1}$, 10 J kg$^{-1}$ (4 h)$^{-1}$, 5 m (4 h)$^{-1}$, and 4 m s$^{-1}$ (4 h)$^{-1}$. Shaded area is mean topographic cross section. Standard pressure levels shown in kilopascals. CAPE was calculated for each 10-mb increment based on the “lowest 30-mb” parcel used in the parcel ascent.

hills where the CAPE was largest (not shown). Figure 12 shows the location of the first radar echo (300°, 160-km range) at 1600 LDT, shortly after underrunning had taken place.

3) DIVERGENCE AND COOLING AT THE UPPER LEVELS

The development of the solenoidal mountain–plain circulation can also be inferred from the 500-mb divergence field (Fig. 13a). Over the foothills, a region of divergence is observed. The band of strong convergence to the east gives evidence of the return flow.

Upper-level cooling, acting in concert with low-level heating and the influx of moisture, destabilizes the atmosphere and renders it susceptible to convective overturning. The 500-mb temperature-tendency field at 1200 LDT (Fig. 13b) indicates negative values in the northern half of the grid—a feature consistent with the cooling observed in the operational 500-mb analyses.

The 2-h net CAPE tendency below 300 mb ending at 1000 LDT (Fig. 13c) depicts maximum positive tendencies directly over the foothills. The destabilization over this region is attributable to the combined presence of upper-level cooling and low-level heating. In contrast, the increase in CAPE from 1200 to 1400 LDT indicates a secondary destabilization arising from the influx of low-level moisture (Fig. 13d).

d. Vertical cross sections

To illustrate more clearly the development of the mountain–plain circulation, vertical cross sections of the 4-h tendencies of temperature, CAPE, height, and $u$ component of the wind ending at 1200 LDT are depicted in Fig. 14. The sections are in a plane perpendicular to the topography, running from LMW to AQF. Important features present over LMW are upper-level cooling and low-level warming (Fig. 14a), a maximum in CAPE tendency (Fig. 14b), low-level height.
falls due to surface heating, and upper-level falls linked to the advancing upper-level trough (Fig. 14c). The regions of maximum cooling above ACR and AQF in Fig. 14a were largely the effect of adiabatic mixing at the top of the boundary layer, which had extended up to 830 and 790 mb, respectively, at these two stations by 1800 LDT. Evidence of the solenoidal circulation induced by the localized destabilization is visible in Fig. 14d. The negative \( u \)-component tendencies above ACR (850 mb) demonstrate the accelerating upslope flow, while the positive tendencies, maximized around 600 mb, give an indication of upper-level return flow. The actual ACR winds at 850 mb changed from 7 m s\(^{-1}\), 300° at 0800 LDT to 4 m s\(^{-1}\), 45° at 1200 LDT. The 600-mb wind at ACR increased from 9 m s\(^{-1}\), 260° to 12 m s\(^{-1}\), 240° during the same period.

To clarify the presence of underrunning and the solenoidal circulation, vertical cross sections of anomalous equivalent potential temperature \( \theta_e \) and anomalous \( u \) component of the wind are presented in Fig. 15. If \( S \) is a scalar field, the anomaly field \( S_a \) is defined as

\[
S_a = S - \bar{S},
\]

where \( \bar{S} \) is the time-averaged (with respect to pressure) field for the period from 0800 to 1600 LDT. The upslope transport of anomalously high-\( \theta_e \) air beneath the capping lid is well illustrated in the 1200 and 1600 LDT cross sections where high-\( \theta_e \) air from the moist, capped boundary layer is transported by anomalous easterly winds to the narrow band of deepening convection over LMW. Upward transport of low-level easterly momentum by the ascending branch of the solenoidal circulation over the foothills is made evident by the negative anomalies above LMW at 1600 LDT.

Since the mountain–plain circulation and the synoptic-scale northeasterly flow both transported high-dewpoint air to the foothills, it is useful to examine the interactions between these two branches of the moisture transport. Note that the upslope flow induced by the mountain–plain circulation began in the morning over the foothills and moved northeastward during the day as the horizontal extent of the solenoid increased (Figs.
10 and 15). Concurrently, the northeasterly flow caused by the synoptic circulation began over eastern Alberta and moved southwestward as a result of ridging in eastern Alberta (Fig. 6). Before 1200 LDT, cooling aloft modified the upper-level stratification over the foothills so as to accelerate the destabilization already occurring there as a result of surface heating. The mountain–plain circulation transported moisture up from the western plains to the foothills. It was not, however, until approximately 1400 LDT, when the northeasterly synoptic surface flow had succeeded in advecting the moist air of the eastern plains into the lower branch of the mountain–plain circulation, that deep convection was triggered.

e. Vertical shear and jet streak

The strong upper-level flow in combination with the northeasterly upslope flow created conditions of strong vertical wind shear in central Alberta by 1400 LDT. At ARM, the 850–500-mb shear increased from $5.1 \times 10^{-2}$ s$^{-1}$ at 1200 LDT to $6.2 \times 10^{-3}$ s$^{-1}$ at 1400 LDT.

As pointed out earlier, the 500-mb operational analysis for 1800 LDT 11 July 1985 shows strong upper-level winds with speeds of 25 m s$^{-1}$ over central Alberta. A mesoscale isotach analysis at 1800 LDT (not shown) indicated a wind maximum of 28 m s$^{-1}$ near ARM, suggesting that the severe convection of 11 July 1985 may have formed on the north side of a migrating jet streak. This maximum was present only in the 1800 LDT isotach analysis, however, and it is difficult to determine whether it represents a true jet streak, a propagating gravity wave, or simply the result of an instrument error. Because of this uncertainty, it is not possible to determine whether the region just to the north of the LIMEX grid, where the strongest multicell storms occurred, was located beneath either the left-front exit or right-rear entrance regions of a jet streak.

Carlson et al.’s (1983) method of indirectly identifying the presence of a migrating upper-level jet streak from surface pressure tendencies has also been attempted. In our 2-h surface pressure tendency fields (Smith 1991), any well-defined, migrating couplet such as that shown in Fig. 6 of Carlson et al. (1983) was unidentifiable. The synoptic-scale pressure tendencies were, to a large extent, the result of low-level thermal forcing. Strong falls along the British Columbia–Alberta border at 1000 and 1200 LDT were aided by surface heating in the elevated topography, and the same applied to the region of maximum falls over southern Alberta. Thus, although the passage of an upper-level jet streak is probable over central Alberta on 11 July 1985, it had little or no effect on the surface pressure tendencies and did not induce underdumping in the manner put forth by Carlson et al. (1983). The importance of the jet streak (or simply the strong south-

westerly flow aloft) rested mainly in providing an environment of strong vertical wind shear into which the severe storms eventually developed.

4. Two nonsevere convective events

In the previous section, it was shown that the mountain–plain and synoptic circulations interacted to produce an outbreak of severe convection over Alberta on 11 July 1985. To affirm our findings, it is essential to establish that such interactions were either absent or suppressed on a nonsevere day. Furthermore, there remains the need to determine the relative importance of the meso- and synoptic scales of motion. To this end, the results from two nonsevere case days are presented.

An examination of the LIMEX dataset revealed that the nonsevere days were marked by two characteristic upper-level flow patterns. The first is the upstream-trough pattern similar to that on 11 July. The other is an upstream-ridge pattern. The following sections present the analyses on 9 and 17 July, when an upstream ridge and trough pattern occurred, respectively.

a. The 9 July case day—Weak, isolated convection

1) SYNOPTIC-SCALE ANALYSIS

A sharp 500-mb height ridge with its axis lying along the British Columbia–Alberta border was a conspicuous feature on 9 July 1985 (Fig. 16). The associated subsidence warming caused clear sky conditions over central Alberta where the northwesterly 500-mb winds were 15–20 m s$^{-1}$. The temperatures in the ridge were relatively warm ($-10^\circ$C at 500 mb over Edmonton).

[Fig. 16. Same as Fig. 5 but at 1800 LDT 9 July 1985.]
In response to the upper-level ridge, the MSL pressure field on 9 July (Fig. 17) differed considerably from that on the severe case day. While a northern high and a southern low dominated the pressure pattern on the latter day, a central high pressure center or ridge, and two low centers located to the north and southeast of the high were the main features on 9 July.

Locally within the LIMEX area, thermally induced upislope flow developed from 0800 to 1800 LDT 9 July. East of the LIMEX domain, however, the surface wind was directed away from the foothills rather than toward it as was the case on 11 July.

At 0800 LDT on both days, the highest dewpoints were found northeast of the LIMEX area. On 9 July, however, considerable drying occurred to the north of the LIMEX domain, where northwesterly surface flow swept the moist plains air eastward. Consistent with this drying, the infrared satellite image at 2106 LDT over Alberta was nearly devoid of cloud cover (not shown).

2) MESOSCALE ANALYSIS

Similar to the case of 11 July 1985, surface heating was responsible for a large part of the morning reduction in the capping-lid strength (Figs. 18a and 18b) with an accompanying large increase in the depth of the adiabatic boundary layer (not shown) over the foothills. At 1000 LDT, maximum surface temperature tendency was noted near LMW (Fig. 18c), where a local maximum of surface pressure falls also occurred (Fig. 18d).

The degree of destabilization from 0800 to 1000 LDT, signified by the CAPE tendency below 300 mb, was of the same order of magnitude on 9 and 11 July (Figs. 19a and 13c). The inference is that during this period, destabilization was primarily the result of strong surface heating. The synoptic environment was more stable on 9 July, however, because of the presence of the upper-level ridge. As a result, the net CAPE below 300 mb at 1000 LDT was much more negative on 9 July than on 11 July except in the region around LMW.
Fig. 18. Analyzed fields of capping-lid strength at (a) 0800 and (b) 1000 LDT 9 July 1985. (c) Surface wind and 2-h temperature tendency and (d) pressure tendency at 1000 LDT. Contour intervals are 1 K, 1°C (2 h)^{-1}, and 0.2 mb (2 h)^{-1}. For surface wind vectors, one east–west grid length is 10.6 m s^{-1}. 

Unauthenticated | Downloaded 10/23/23 05:50 AM UTC
Fig. 19. Analyzed fields of CAPE tendency below 300 mb at (a) 1000 LDT 9 July 1985 and (b) surface wind and dewpoint temperature at 1000 and (c) 1200 LDT. Contour intervals: 100 J kg\(^{-1}\) (2 h\(^{-1}\)) and 1°C.
Fig. 20. Vertical profiles of the difference in the \textit{u} component of the wind between 11 July and 9 July 1985 at ARM. Solid line is for 0800 LDT, solid line with plus sign for 1000 LDT, and dashed line for 1200 LDT. Standard pressure levels given in kilopascals.

Fig. 21. As in Fig. 5 but for 1800 LDT on 17 July 1985.

Figures 19b and 19c show that between 1000 and 1200 LDT, northeasterly upslope flow had developed west of the line joining ACR and AML, causing dewpoint temperatures to rise in this region. To the east of this line north to ARM, however, the surface winds were still characterized by a downslope component and significant drying had begun near ARL and ARM.

Warming in the upper levels from the approaching ridge was noted from 1000 to 1200 LDT. During this period, the 500-mb temperature rose by 0.2°C at both ARM and AQF (the sounding from ARL did not reach 500 mb). The corresponding 2-h temperature changes over ARM and AQF on 11 July 1985 were \(-1.1\)° and \(-0.6\)°, respectively. These short-term tendencies were consistent with the 12-h tendencies at Stony Plain, where a warming of 3.2°C at 500 mb from 0600 to 1800 LDT was observed on 9 July compared with a 3.0°C cooling on 11 July.

The 850–500-mb shear at 1200 LDT was \(2.9 \times 10^{-3}\) \(\text{s}^{-1}\) at ARM.\(^6\) To substantiate the claim that the mountain–plain circulation was amplified on 11 July 1985, the \textit{u} component of the wind on 9 July is subtracted from that on 11 July, and the differences is displayed in Fig. 20. At 1200 LDT, a layer of negative difference extends from the surface to around 650 mb, which demonstrates a significant enhancement of the easterly flow on 11 July. A larger westerly component is located near 475 mb at both 1000 and 1200 LDT, indicative of a stronger return flow on the severe day.

\textit{b. The 17 July case day—Widespread, moderate convection}

1) \textsc{synoptic-scale analysis}

A short-wave trough passed over central Alberta around 0700 LDT (Fig. 21). During the morning

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\(^6\) Because of the absence of deep convection, sounding releases at LMW, AYC, AEL, ACR, and AML were canceled after 1000 LDT. One additional sounding was released at ARM, AQF, and ARL at 1200 LDT. The 1800 LDT (0000 UTC) operational sounding at Stony Plain (WSE) was the only afternoon upper-air data available.
Fig. 23. Same as Fig. 18 but for 17 July 1985.
hours, the 500-mb winds over central Alberta were southwesterly at 5–10 m s\(^{-1}\). Associated with the trough passage, the 500-mb temperature over Stony Plain fell 1.8°C in 12 h. Although the upper-level flow was weaker on this day than on 11 July, it was similar to the upstream trough pattern associated with the occurrence of major hail described by Longley and Thompson (1965; see Fig. 3).

As on 11 July 1985, the MSL pressure pattern at 1200 LDT (Fig. 22) exhibited a high and a low located, respectively, over northwestern and southern Alberta. The resulting synoptic pressure gradient allowed for the development, later in the day, of northeasterly flow over the eastern plains. This northeasterly flow did not initiate severe convection, however.

2) **Mesoscale Analysis**

Figure 23 shows that cloudy conditions prevailed over all observing stations at 0800 and 1000 LDT. In contrast to the 9 and 11 July case days, the capping-lid strength was much weaker and was not subject to strong erosion over the foothills. The corresponding analyses of boundary-layer depth (not shown) also do not depict strong increases. These characteristics are related to the absence of strong surface heating over the foothills and is illustrated by the weak temperature tendency plotted in Fig. 23c. Consequently, the isallobaric gradient was weak (Fig. 23d) and did not favor the formation of thermally induced upshe flow at 1000 LDT. In fact, the 850-mb winds at ACR backed from 4 m s\(^{-1}\), 26° at 0800 LDT to 2 m s\(^{-1}\), 318° at 1200 LDT pointing to accelerating downslope flow.

Cloud shading also weakened the destabilization over the foothills. The CAPE-tendency field at 1000 LDT (Fig. 24) shows a local maximum of 400 J kg\(^{-1}\) (2 h)\(^{-1}\) near AML, far less than the values in excess of 2000 J kg\(^{-1}\) (2 h)\(^{-1}\) observed over the foothills on 11 and 9 July (Figs. 13c and 19a). Despite the lack of strong destabilization, CAPE values over the foothills at 1000 LDT were comparable to those on 9 and 11 July.

The damping effect of cloud shading can be estimated by comparing the actual CAPE below 300 mb over LMW with the CAPE that would have resulted if surface heating had taken place under clear sky conditions. The latter quantity can be approximated by applying the observed CAPE tendency ending at 1000 LDT 9 July [\(+2637\) J kg\(^{-1}\) (2 h)\(^{-1}\)] to the observed CAPE at 0800 LDT 17 July [\(−894\) J kg\(^{-1}\)]. The “potential” clear-sky CAPE at 1000 LDT 17 July turned out to be \(+1743\) J kg\(^{-1}\), more than four times the actual value of \(−558\) J kg\(^{-1}\)! In other words, without the reduction in surface heating due to cloud shading, the CAPE present above LMW on 17 July would have been in the range normally associated with high-energy hailstorms (Chisholm 1973).

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**Fig. 24.** Analyzed field of CAPE tendency below 300 mb at 1000 LDT 17 July 1985. Contour interval is 100 J kg\(^{-1}\) (2 h)\(^{-1}\).

<table>
<thead>
<tr>
<th>Intensity of convection</th>
<th>9 July Weak (no hail)</th>
<th>11 July Strong (severe hail)</th>
<th>17 July Moderate (moderate hail)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Synoptic setting</td>
<td>500-mb ridge upstream</td>
<td>500-mb trough upstream</td>
<td>500-mb trough upstream</td>
</tr>
<tr>
<td></td>
<td>500-mb warming and</td>
<td>500-mb cooling and</td>
<td>500-mb cooling and</td>
</tr>
<tr>
<td></td>
<td>stabilization</td>
<td>destabilization</td>
<td>destabilization</td>
</tr>
<tr>
<td></td>
<td>Surface northern low-central high-southern low</td>
<td>Surface northern high-southern low</td>
<td>Surface northern high-southern low</td>
</tr>
<tr>
<td></td>
<td>Surface synoptic downslope flow with moisture transported away from foothills</td>
<td>Surface synoptic upslope flow with moisture transported towards foothills</td>
<td>Surface synoptic upslope flow with moisture transported towards foothills</td>
</tr>
<tr>
<td>Stratification</td>
<td>Area-averaged* CAPE below 300 mb at 0800 LDT</td>
<td>Area-averaged CAPE below 300 mb at 0800 LDT</td>
<td>Area-averaged CAPE below 300 mb at 0800 LDT</td>
</tr>
<tr>
<td></td>
<td>−2613 J kg⁻¹</td>
<td>−1936 J kg⁻¹</td>
<td>−1240 J kg⁻¹</td>
</tr>
<tr>
<td>Mesoscale setting</td>
<td>Strong surface heating</td>
<td>Strong surface heating</td>
<td>Weak surface heating</td>
</tr>
<tr>
<td></td>
<td>Thermally induced upslope flow</td>
<td>Enhanced thermally induced upslope flow</td>
<td>Absence of thermally induced upslope flow</td>
</tr>
</tbody>
</table>

* Averaged over the LIMEX grid.

By 1200 LDT, cooling was occurring at 500 mb. Near the surface, the winds were very weak and continued to be directed downslope. Compared to the severe case day, the dewpoint temperatures were fairly high over the foothills, and the cloud and weather observations pointed to the existence of thunderstorms. The widespread, disorganized echo structure with weak reflectivity gradients shown in the radar PPI at 1500 LDT (Fig. 25) indicates, however, that these thunderstorms were not severe.

The weak winds aloft and the absence of low-level upslope flow at ARM resulted in a 850–500-mb shear value of $1.3 \times 10^{-3}$ s⁻¹ at 1200 LDT. The condition of weak vertical shear is consistent with the disorganized character of the convective storms depicted in the radar PPI.

5. Discussion

To summarize our findings concerning the synoptic-mesoscale interactions over Alberta, we list in Table 1 the important physical parameters on 9, 11, and 17 July 1985.

The 9 July analysis confirms the importance of the synoptic circulation. By the time thermally induced upslope flow had developed, northwesterly flow over the plains arising from the synoptic-scale pressure gradient had transported the plains moisture to the east. The relatively dry lower branch of the mountain–plain circulation was therefore unable to trigger moist convection in the foothills through underrunning.

The importance of the mountain–plain circulation is brought out by the 17 July case. The upstream trough aloft and the northeasterly flow at the surface provided the synoptic environment conducive to the formation of severe convection. The suppression of the mountain–plain circulation by early cloud cover resulted, however, in weak destabilization over the foothills and disorganized convective storms. The exact cause of the early cloud cover is difficult to determine. The axis of the 500-mb height trough was closer to the foothills on 17 July than on 11 July 1985. Consequently, by the morning of 17 July, synoptic-scale ascent may have already weakened the capping lid and destabilized the atmosphere to the point where convection was easily realized. In this sense, the improper phasing of dynamic forcing with diurnal heating may increase the horizontal extent of convection, yet lessen its severity.

The three case studies presented here, along with other LIMEX-85 case days, will be synthesized into a conceptual model of severe convective outbreaks in Alberta in Part II (Smith and Yau 1993).

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APPENDIX A

Station Identifiers

<table>
<thead>
<tr>
<th>Station</th>
<th>Elevation (m MSL)</th>
</tr>
</thead>
<tbody>
<tr>
<td>AQF—Red Deer</td>
<td>900</td>
</tr>
<tr>
<td>ARM—Rocky Mountain House</td>
<td>988</td>
</tr>
<tr>
<td>AYC—Calgary (University of Calgary)</td>
<td>1114</td>
</tr>
</tbody>
</table>
APPENDIX A (Continued)

<table>
<thead>
<tr>
<th>Elevation (m MSL)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LMW—Limestone Mountain West 1506</td>
</tr>
<tr>
<td>ABP—Bow Pass 1710</td>
</tr>
<tr>
<td>ACR—Caroline 1068</td>
</tr>
<tr>
<td>AEI—Elkton 1130</td>
</tr>
<tr>
<td>AML—Mountaineer Lodge 1441</td>
</tr>
<tr>
<td>ARL—Ram Lookout East 1294</td>
</tr>
<tr>
<td>WSE—Stony Plain 766</td>
</tr>
<tr>
<td>YEG—Edmonton International 723</td>
</tr>
<tr>
<td>WBA—Bannaf 1397</td>
</tr>
<tr>
<td>FMH—Mockingbird Hill 1902</td>
</tr>
<tr>
<td>FGH—Ghost 1433</td>
</tr>
<tr>
<td>FBB—Blue hill 1985</td>
</tr>
<tr>
<td>FBL—Baseline 1893</td>
</tr>
<tr>
<td>FBK—Burnstick 1227</td>
</tr>
<tr>
<td>FCW—Clearwater 1279</td>
</tr>
<tr>
<td>AEV—Evergreen 1009</td>
</tr>
<tr>
<td>AGL—Glennifer Lake 1012</td>
</tr>
<tr>
<td>APC—Prairie Creek 1406</td>
</tr>
<tr>
<td>ARF—Ram Falls 1618</td>
</tr>
<tr>
<td>ASU—Sundre West 1167</td>
</tr>
<tr>
<td>ATP—Tee-Pee Pole Creek 1380</td>
</tr>
<tr>
<td>AWV—Water Valley 1313</td>
</tr>
<tr>
<td>ACH—Cheddarville 1088</td>
</tr>
<tr>
<td>LMR—Limestone Mountain Ridge 2121</td>
</tr>
</tbody>
</table>

APPENDIX B

Hail-Severity Classification

The Alberta Hail Project (AHP) operated in central Alberta for 29 years (1957–85). Each day during the summer months, data on observed hailfalls (maximum hail size, time of the onset of hail, etc.) were collected. The major source of these data were hail cards sent in by farmers living within the project area and telephone surveys. Figure B1 displays the AHP experiment areas from which hail reports were received. Area A (35 484 km²) was in operation from 1957 until 1973, while area B (33 700 km²) operated from 1974 through 1985. Table B1 displays the potential number of observers within the project area for each year.

On any given day with hail, between 10% and 20% of the potential number would respond by mailing in a completed hail card, yielding an average observation density of 1 report per 16–32 km². Observers were distributed with fair uniformity within the intersection of areas A and B, where the farm density was highest. The number of farms decreased markedly to the extreme west, particularly near the foothills. For days when telephone surveys were conducted (usually days of severe convection), the density of observations increased to as high as 1 report per 3.2 km². As can be seen in Table B1, there was a steady decline in the number of potential observers from early to later years. This trend can be attributed to economic factors that resulted in fewer, albeit larger, farms. Considering the variability in observer response and the natural daily variability in convective activity, this decrease would not appear

![Figure B1. Areas of operations of the Alberta Hail Project. Area A was used from 1957 to 1973, while area B served from 1974 to 1985.](image)

<table>
<thead>
<tr>
<th>Year</th>
<th>Number of observers</th>
</tr>
</thead>
<tbody>
<tr>
<td>1974</td>
<td>22 800</td>
</tr>
<tr>
<td>1975</td>
<td>21 900</td>
</tr>
<tr>
<td>1976</td>
<td>21 900</td>
</tr>
<tr>
<td>1977</td>
<td>21 000</td>
</tr>
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<td>1979</td>
<td>20 000</td>
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<tr>
<td>1980</td>
<td>18 873</td>
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<td>18 499</td>
</tr>
<tr>
<td>1982</td>
<td>17 409</td>
</tr>
<tr>
<td>1983</td>
<td>19 464</td>
</tr>
<tr>
<td>1984</td>
<td>19 464</td>
</tr>
<tr>
<td>1985</td>
<td>19 464</td>
</tr>
</tbody>
</table>
to be important. Regardless of the potential number
of observers, the use of telephone surveys to verify ra-
der-observed storms ensured that only a very small
percentage of hail that reached the ground went un-
detected.

The total number of hail reports recorded each day
can be viewed as the outcome of a particular type and
strength of convection. Many hail reports represent, in
general, the outcome of organized, severe convection.
Few or no hail reports would be the outcome of weak
convection. Based on these considerations, Smith and
Yau (1987) proposed the following classification.

Severe hail (SH)  >150 hail reports
Moderate hail (MH)  51–150
Light hail (LH)  1–50
No hail (NH)  0

Although many hail reports can be generated by many
small storms, experience shows that larger, well-orga-
nized storms with long lifetimes produce the most ex-
tensive hail swaths and the greatest number of hail re-
ports. As a result, the total number of reports is posi-
tively correlated with maximum hail size. Therefore,
it is unnecessary to explicitly include reported hail size
in the severity classification. The choice of cutoffs in
the number of reports for the severity classes was based
on case studies (Smith 1986), which indicated that the
type of storms that produce more than 150 hail reports
are far more intense and have a greater degree of or-
ganization than those responsible for 50 or less reports.
It is quite difficult, however, to ascribe the storms that
result in 51–150 reports to some unique “moderate”
mode of convection. For example, 100 reports could
be generated by convection of near-severe intensity on
one day or widespread, weak convection on another.

APPENDIX C

Weather Phenomena and Symbols

<table>
<thead>
<tr>
<th>Specific phenomena</th>
<th>Symbol</th>
</tr>
</thead>
<tbody>
<tr>
<td>Thunderstorm</td>
<td>T, T+</td>
</tr>
<tr>
<td>Rain</td>
<td>R, R, R</td>
</tr>
<tr>
<td>Rain shower</td>
<td>RW, RW+</td>
</tr>
<tr>
<td>Drizzle</td>
<td>L, L, L+</td>
</tr>
<tr>
<td>Hail</td>
<td>A, A+</td>
</tr>
<tr>
<td>Fog</td>
<td>F</td>
</tr>
<tr>
<td>Haze</td>
<td>H</td>
</tr>
<tr>
<td>Smoke</td>
<td>K</td>
</tr>
<tr>
<td>Dust haze</td>
<td>D</td>
</tr>
<tr>
<td>Blowing dust</td>
<td>BD</td>
</tr>
<tr>
<td>Cumulus clouds</td>
<td>CU</td>
</tr>
<tr>
<td>Towering cumulus clouds</td>
<td>TCU</td>
</tr>
<tr>
<td>Cumulonimbus clouds</td>
<td>CB</td>
</tr>
</tbody>
</table>

Intensity of phenomena given by “-” very light;
“-” light; “” moderate; and “+” heavy.

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