The Interactive Role of Subsynoptic Scale Jet Streak and Planetary Boundary Layer Processes in Organizing an Isolated Convective Complex

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ABSTRACT

Mesoscale model simulations with and without diurnal planetary boundary layer heat flux are compared to a detailed surface analysis for a case of an isolated tornadic convective complex development. The case study, 3–4 June 1980, is of particular interest because of the development of several destructive tornadic storms within the Grand Island, Nebraska metropolitan area during a period of relatively weak synoptic scale forcing. This type of case presents an opportunity for the mesoscale numerical simulation of the subtle interactions between an upper tropospheric jet stream and surface diabatic heating. Model simulations run with and without diurnal surface sensible heating show marked differences in processes both within and above the planetary boundary layer (PBL). The results of the simulations indicate that the evolution of the subsynoptic scale low pressure system and its accompanying low level jet streak, areas of moisture convergence, and regions of convective instability are influenced by the interaction of the deep surface-heated PBL with a weak synoptic scale jet streak. The model simulations show that the distribution and evolution of tropospheric velocity divergence cannot be realistically decoupled from the thickness changes caused by PBL heating in this case of relatively weak dynamic forcing. Modifications in the simulated velocity divergence and low level warm advection caused by PBL heating led to a more realistic pattern of pressure falls, low level jet formation, and a significant reduction of the lifted index values near the region of observed convection. Comparisons with observations, however, also indicate that the modeling system still requires: 1) enhanced soil moisture information in the data base utilized for its PBL parameterization to achieve the proper amplitude and distribution of surface sensible heat flux and 2) the proper parameterization of convective scale processes such as latent heating to completely capture the evolution of the subsynoptic scale low pressure system into a mesoscale low pressure system. The most significant implication of these modeling results is that previous dynamical models of upper and lower tropospheric coupling during the pre-storm environment should include consideration of the effects of diurnal surface sensible heating upon a pre-existing jet streak.

1. Introduction

For many years investigators have developed theories linking the onset and intensification of severe local storm systems to upper and lower tropospheric jet streams. These studies span a period in excess of 30 years from the pioneering work of Fawbush et al. (1951), Fawbush and Miller (1953, 1954), Beebe and Bates (1955), House (1958), Lee and Galway (1956, 1958) and Newton (1963) to contemporary studies such as that by Uccellini and Johnson (1979). These investigators have related upper tropospheric velocity divergence, vertical wind shear, and dynamic destabilization to the development of severe convective storms.

Recently, Maddox and Doswell (1982) have noted that well-defined classical jet streak dynamics are not apparent in all cases of intense convective complex development. They point to the critical role played by lower tropospheric warm advection in organizing the upward motions necessary to release convective instability. Many investigators have found a relationship between regions of warm surface temperatures or low-level warm advection and well-organized convective storms often accompanied by heavy rain or tornadoes (e.g., Whiting, 1957; Darkow et al., 1958; Matsumoto et al., 1971; Bosart and Carr, 1978; Ninomiya, 1971; and Maddox et al., 1979, 1980). Others including Whiting and Bailey (1957), Tegtmeier (1974) and Moller (1980) have found tornadic storms to be most numerous in the northeast quadrant of subsynoptic surface low pressure systems which are often regions of significant low-level warm advection.

The emphasis of the studies noted above has been to study the influence of larger scale dynamic processes upon the development of severe local storms. Recent mesoscale modeling studies by Chang et al. (1981) and Maddox et al. (1981) have begun to deal with the problem of the influence of the convective storm systems upon the large-scale environment by modeling the parameterized convective scale latent heat release and the subsequent perturbation of the jet stream flow. However, questions still remain concerning the
relative importance of heat flux within the planetary boundary layer in causing pressure changes which ultimately modify jet stream velocity fields prior to the onset of deep convective complex development.

The primary purpose of this paper is to examine to what extent deep, well-mixed, and well-heated boundary layers play a role in perturbing a weak jet streak in proximity to the development of an isolated but intense convective complex associated with the Grand Island, Nebraska tornado outbreak of 3–4 June 1980. This case is particularly important because of the relatively weak upper-level wind velocity patterns observed prior to convective complex development, making the severe convection a very difficult event to forecast (Maddox and Doswell, 1982). This paper will relate rawinsonde and surface analyses to mesoscale model simulations with and without diurnal surface sensible heating in an effort to determine the significance of the dynamics prior to the development of severe convective storms.

A brief description of the case is presented in Section 2 which emphasizes 3-hourly surface analyses, radar and satellite data. In Section 3, we discuss the results of the numerical experiments comparing differences in the runs with and without diurnal surface sensible heating and relating them to the observations. In Section 4 a discussion of the dynamical processes which are responsible for these simulation differences is presented. The significance of these differences in terms of their effect upon the preconvective environment is discussed in Section 5. A summary of the results is presented in Section 6.

2. Case study: 3–4 June 1980

a. Satellite and radar data

On 4 June 1980 Grand Island, Nebraska (GRI) was devastated by a major tornado outbreak. The GOES satellite imagery and NWS radar summaries (Figs. 1 and 2) indicate that before 0000 GMT 4 June, convection is generally isolated and weak throughout the plains states. By 0000 GMT, however, there is a cell or cluster of cells near GRI (Fig. 1b). This feature is first depicted on radar at 0035 GMT. The NWS radar summaries and GOES satellite imagery indicate that this area of convection, with clouds approaching 20 km in depth, continues to increase in size with its cirrus shield covering most of eastern Nebraska by the 0200 GMT time period. Severe weather reports (Fig. 3) indicate that this convection is producing several tornadoes in Grand Island between 0200 and 0500 GMT.

b. Hourly surface data analysis

The Barnes (1964) objective analysis scheme utilized to analyze the hourly Airways surface data is the same as that used to analyze the MASS model initial data fields and is described in Kaplan et al. (1982). The 1200, 1800, 2100, and 0000 GMT 3–4 June 1980 altimeter setting, surface temperature and surface
Fig. 2. National Weather Service radar summaries for (a) 2335, (b) 0035, and (c) 0135 GMT 3–4 June 1980.

dewpoint analyses derived from hourly surface observations are depicted in Figs. 4, 5 and 6 primarily for Kansas and Nebraska. The dominant feature in the altimeter setting analyses at 1200 GMT 3 June (Fig. 4a) is the trough of low pressure extending from western South Dakota toward southwestern Kansas. The surface temperature analyses (Fig. 5a) for 1200 GMT indicates a weak thermal ridge extending along and east of the trough axis in the central Great Plains. The surface dewpoint analysis for 1200 GMT (Fig. 6a) reveals the existence of a distinct dry line which extends from the Wyoming–South Dakota border southward. Except near the dryline feature, the 1200 GMT wind observations indicate generally light, uniform surface winds from the south or southeast (Fig. 4a).

During the period between 1200 and 1800 GMT, distinct subsynoptic features in the altimeter setting, temperature, and dewpoint analyses begin to evolve (Figs. 4b, 5b, 6b). By 1800 GMT the minimum in the altimeter setting field expands to cover much of northwestern Kansas. This low pressure system has a trough extending eastward toward north-central Kansas. As this trough propagates northeastward, the isobars begin to shift their orientation across southern and western Kansas from a north–south to an east–west direction.

The surface temperature analysis for 1800 GMT (Fig. 5b) indicates that significant warming has oc-

Fig. 3. Severe weather reports for the period 1800 GMT 3 June to 1200 GMT 4 June 1980 (Maddox and Doswell, 1982).
curred across most of eastern Colorado and Kansas. A thermal ridge extends northward from northern Kansas toward southern South Dakota. The maximum surface temperature at 1800 GMT is seen in western and central Kansas (Fig. 5b). It is also interesting that these stations include and are generally downstream from Dodge City, Kansas (DDC) where a very warm (25°C) 1200 GMT 85 kPa temperature observation was recorded. In spite of well-organized surface thermodynamic features, winds still remain uniform in velocity and direction across Nebraska and Kansas (Fig. 4b).

During this same 6-hour period between 1200 and 1800 GMT (Figs. 6a and b) the surface dewpoint analysis indicates a well-developed tongue of higher dewpoints oriented from the southeast toward the northwest across Nebraska and Kansas. This feature appears to be the result of moisture transport northwestward into central Nebraska.

At 2100 GMT a distinct subsynoptic low in northwest Kansas continues propagating toward the northeast (Fig. 4c). Accompanying the evolution of this system is an increase in the southerly winds in northcentral Kansas in the region where the isobars are becoming progressively oriented in an east–west direction. With the exception of extreme southwestern Kansas, this region also has the highest surface temperatures (Fig. 5c) with several stations exceeding
33°C. Associated with this continued heating in central Kansas is the development of a distinct east–west oriented thermal boundary near the Kansas–Nebraska border. Surface temperatures in central Nebraska rise less rapidly than in central Kansas. By 2100 GMT the central Nebraska surface temperatures average nearly 5°C lower than those in central Kansas. During this period the temperature gradient also increases between eastern and western Nebraska. The juxtaposition of the 6-hour surface temperature changes between 1500 and 2100 GMT relative to the 9-hour mean sea level pressure changes between 1500 and 0000 GMT are depicted in Fig. 7. This indicates that the maximum pressure falls are occurring in the gradient of the temperature change region where the warm advection is maximized.

By 2100 GMT the moist air has advected northward ahead of this surface low in Kansas with the 20°C isotherm reaching into central Nebraska (Fig. 6c). An eastward bulge in the dry line has also formed by 2100 GMT in western Kansas, further indicating the juxtaposition of relatively dry and moist air masses. The preferential formation of tornadic storms near the dryline bulge is a phenomenon observed by Tegtmeier (1974) and McCarthy and Koch (1982).

By 0000 GMT the subsynoptic low is slightly northwest of Hill City, Kansas (HLC) with relatively little change in the overall shape of the pressure pattern since 2100 GMT (Fig. 4d). There is a continued shift in the surface winds towards the south, particularly in north-central Kansas, while the winds in central Nebraska maintain a southeasterly direction (Fig. 4d). Surface winds have shifted slightly more to the east in eastern Nebraska and to the southwest at HLC as compared to the 2100 GMT analyses (Fig. 4c). The north–south surface temperature gradient,
which had become well-developed by 2100 GMT near the Nebraska–Kansas border has maintained its structure through 0000 GMT with the 30°C isotherm moving into the region just southwest of GRI (Fig. 5d). The dewpoint analyses indicate slightly drier air moving into southwestern Nebraska, while slightly higher dewpoint values continue to move into north-central Nebraska. The observed 0000 GMT 50 kPa lifted index reflects this low level transport of moisture (Fig. 8a). It is at this time that GRI reports a cumulonimbus west-northwest of the station moving eastward and by 0100 GMT thunder is reported at that station.

c. Upper air observations

The NWS rawinsonde observations valid at 1200 3 June and 0000 GMT 4 June were analyzed utilizing a Barnes analysis technique in a manner similar to that described in Kaplan et al. (1982). On close examination of the 30, 50 and 85 kPa charts (Figs. 8b–g) for 1200 3 June and 0000 GMT 4 June, there is little indication of many of the intense well-organized features often present prior to tornado outbreaks as have been described by Newton (1963), Danielsen (1974), Uccellini and Johnson (1979), Hoxit and Chappell (1975), Fujita et al. (1970), Fujita and Forbes (1974), and Zack (1981). Maddox and Doswell (1982) have pointed out that, as in many other cases, the Grand Island outbreak occurred during relatively weak middle to upper tropospheric flow patterns exhibiting little positive vorticity advection. At 1200 GMT there are two identifiable jet streaks over the country as indicated on the 30 kPa analysis (Fig. 8b). One jet is located over the northern tier of states extending from northern California to the northern
Rockies and a second extending from Mexico to northeast Oklahoma and then eastward to the Ohio Valley region. The 50 kPa surface at 1200 GMT again reflects the basic trough–ridge–trough system across the United States (Fig. 8d) with relatively weak winds in the south central part of the country. The 85 kPa surface (Fig. 8f) depicts a band of 10–15 m s\(^{-1}\) southerly winds from northern Oklahoma to South Dakota as the most relevant feature. This is also a general region of warm air advection at 1200 GMT as emphasized by Maddox and Doswell (1982).

By 0000 GMT the basic upper air pattern remains relatively the same as before. The ridge line at 30 and 50 kPa has shifted slightly eastward reflecting significant height rises across the north central Great Plains (Fig. 8e). The position of the jet has shown little change (Fig. 8c). Therefore, by 0000 GMT, eastern and central Nebraska lies within the left exit region of the weak jet streak centered over New Mexico. The 85 kPa analysis indicates that the flow over Nebraska and South Dakota has remained 10–15 m s\(^{-1}\) from the south with a noticeable tightening of the height gradients in this region (Fig. 8g), coinciding with the area in which the surface subsynoptic low developed and the moisture and temperature advections were increased (previous subsection).

The changes noted above are subtle, but important. This is a rather weak and nonamplifying system. It is not at all like the classical pattern of an amplifying baroclinically unstable wave which typically precedes the synoptic scale cyclogenesis and which leads to severe storm formation (Newton, 1963; Danielsen, 1974). In an effort to reconcile this dilemma, Maddox and Doswell (1982) emphasized the importance of low-level warm advection as a mechanism for generating low-level ascending motion which releases the convective instability. However, while there is a positive Laplacian configuration in the 85 kPa warm advection pattern over Nebraska and Kansas by 0000 GMT, a large portion of the plains states is also being subjected to a similar pattern during this time period (Fig. 8g). Nevertheless, the development of the severe convective storm system occurred over only a very isolated part of the region experiencing a significant Laplacian of warm advection, a region which is also near the exit region of the upper-level jet as defined on the 30 kPa surface (Fig. 8c).

It is possible that the increased lower tropospheric upward motion below the exit region of the mid- to upper level jet (Uccellini and Johnson, 1979) is acting to enhance the environment for the development of convection over southern Nebraska. An important signature of this process would be mid- to upper level divergence in the left exit region of the jet (Beebe and Bates, 1955; McNulty, 1978). As 30 kPa heights rise 60–90 m in the mid- and upper Mississippi River Valley between 1200 and 0000 GMT, the winds tend to adjust by exhibiting a more dominant westerly component at Topeka, Kansas (TOP) and Omaha, Nebraska (OMA). This increase in the westerlies across eastern Kansas and eastern Nebraska contributes to the development of weak mid- to upper tropospheric velocity divergence to the west of these stations by 0000 GMT (Fig. 9). The synoptic scale velocity divergence analysis is from Schaefer and Doswell, 1982.) An analysis of the 0000 GMT synoptic scale velocity divergence (Fig. 9a) indicates weak divergence at 50 kPa over eastern Nebraska and over southeastern Nebraska and eastern Kansas at 30 kPa (Fig. 9b). The divergence pattern is consistent with the positive vorticity advection inferred from the 50 kPa vorticity analysis (Fig. 9c). It should be noted, however, that the relatively basic model of jet streak adjustments conceptualized by Uccellini and Johnson

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**Fig. 7.** Observed altimeter setting change analysis for the period from 1300 to 0000 GMT 3–4 June 1980 and observed surface temperature change analysis for the period from 1300 to 2100 GMT 3 June 1980. Solid lines represent pressure change in kPa. Dashed lines represent temperature change in °C.
(1979) cannot define the totality of the mass–momentum adjustments which result in this divergence pattern. For example, the effect of curvature near the axis of the upper level ridge has to be considered as is done in House (1961), Kaplan and Paine (1977), and Uccellini et al. (1984).

A second possible link between the massive organized tornado outbreaks and the isolated variety, of which the Grand Island case study is an example, may be a well-mixed boundary layer upwind of the region of intense storms. Features similar to this have been emphasized by Hoxit and Chappell (1975), Danielsen (1974), Fujita et al. (1970), and Zack (1981) in the famous 3–4 April 1974, Palm Sunday and Red River Valley outbreaks. Carlson et al. (1983), Carlson and Ludlum (1968), Anthes et al. (1982),
Benjamin (1983) and Lanicci and Carlson (1983), all discuss the existence of well-mixed boundary layers behind the dryline during severe weather outbreaks. This type of structure is apparent in the 1200 GMT sounding at DDC (Fig. 10a) which is much warmer in the layer between 80 and 90 kPa than is the sounding at North Platte, Nebraska (LBF), for example (Fig. 10b). However, at 50 kPa the temperature at DDC is not very different than the other stations which all report temperatures $\sim -10^\circ$C. The region north of DDC was seen as the region of observed maximum surface temperature increases (Fig. 7). The deep, very dry, nearly adiabatic layer above the surface-based morning inversion should be a favorable precursor to the formation of a well-mixed deep PBL. This type of PBL could tend to enhance a pre-existing thermal boundary by allowing a very deep mixing of surface sensible heat. This could result in the amplification of spatial variations in the vertical mixing of the surface sensible heat. The region of deep mixing may be much more sharply demarcated than can be inferred from the synoptic 85 kPa analysis because
of its inherent mesoscale structure. Given the fact that these features exist, a question arises as to whether or not they modify the background synoptic scale jet streak dynamics and contribute to changes in the preconvective environment that aided in the development of severe convection in southeastern Nebraska. This issue will be studied with the numerical experiments in the following section.

d. Summary of the observations from the case study

The observed surface and rawinsonde data have indicated the following significant features prior to the onset of convective complex development near Grand Island, Nebraska:

1) a well-organized subsynoptic scale surface low developed in western Kansas and moved toward GRI prior to the onset of convection;
   2) a well-organized gradient in the surface temperature developed in proximity to the region of lowest surface pressure;
   3) surface wind adjustments appear to have been linked to the evolving subsynoptic surface pressure field;
4) a surface dewpoint maximum in the Grand Island area developed along the leading edge of the subsynoptic surface low pressure system which was linked to the increased advections to the east of the surface low in Kansas;
5) a weak synoptic scale jet streak, approaching

**FIG. 9.** (a) 50 and (b) 30 kPa observed synoptic scale velocity divergence analyses for 0000 GMT 4 June 1980. Solid lines are contours of velocity divergence and dashed lines are contours of velocity convergence in units of $10^{-3}$ s$^{-1}$. (From J. T. Schaefer and C. A. Doswell, III, NOAA, National Weather Service, Techniques Development Unit, NSSFC).

**FIG. 9c.** 50 kPa observed synoptic scale absolute vorticity analysis for 0000 GMT 4 June 1980. Solid lines are isovorts in units of $10^{-6}$ s$^{-1}$. 
scale rawinsonde data pose an interesting challenge to mesoscale modelers. Given the data limitations, it appears that the only means to resolve the relative importance of, and interrelationships between, jet streak dynamics and boundary layer sensible heating is through numerical experiments. In this section, numerical experiments are described which are designed to determine: 1) if this case can be properly simulated with conventional observations at synoptic time and space scales, 2) to what extent PBL and middle to upper tropospheric features interact, and 3) to what extent this interaction establishes an environment conducive for severe convective storm development.

Two numerical simulations for the Grand Island case are described. Experiment I consists of running a mesoscale numerical model for 18 hours initialized from a synoptic-scale data base valid at 1200 GMT 3 June 1980 and therefore allow the PBL to respond to the diurnal surface heating cycle. Experiment II consists of running the same model with the same data, except that no diurnal surface sensible or surface latent heat flux will be allowed. These two simulations 1) test the model sensitivity to diurnal surface sensible heat flux variations and determine how the subsequent mass changes modify the deeper tropospheric flow patterns and 2) determine the extent to which the environment preceding intense convection is modified by the diurnal PBL heating.

a. The numerical model

Specific details on the modeling system have been described by Kaplan et al. (1982) and Wong et al. (1983). Therefore, the model used in these experiments (MASS 3.0) is only briefly discussed in this section and described in Table 1. It should be emphasized that in an effort to simplify the determination of its sensitivity to surface sensible heating, the cumulus parameterization scheme in MASS 3.0 is turned off. MASS 3.0 is designed for subsynoptic (meso α) scale sensitivity for time periods of 12 to 36 hours with its 58 km horizontal and 14 layer vertical resolution covering the North American continent and environs. The model also emphasizes relatively comprehensive PBL physics employing a generalized similarity PBL which is sensitive to time-dependent changes in the static stability, mean PBL wind, PBL height, as well as to spatial variations in albedo, ground temperature, soil moisture and surface roughness. This scheme employs a comprehensive surface energy and moisture balance which enhances its ability to model diabatic interactions between the earth's surface and free atmosphere for a variety of stability and flow regimes. Crucial to the proper simulation of the surface energy balance is the initialization of the soil moisture. For these simulations a uniform 60% soil saturation ratio was employed. Future simulations will allow more

3. Numerical simulations of the Grand Island tornado outbreak

The relatively weak synoptic scale forcing for the Grand Island case and lack of detailed subsynoptic

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FIG. 10. Observed soundings depicted in emagram format for (a) Dodge City, Kansas, and (b) North Platte, Nebraska for 1200 GMT 3 June 1980. Full wind barb = 5 m s⁻¹.
TABLE 1. Basic characteristics of MASS 3.0 (June ’83).

1. A matrix of 128 × 96 grid points with ~58 km horizontal grid mesh covering most of North America and the adjacent waters.
2. 14 vertical layers in a sigma-p coordinate system.
3. Nested grid capability to ~14 km.
4. Euler-backward time marching.
5. Fourth-order accurate space differencing.
6. Generalized similarity theory planetary boundary layer with a surface temperature and moisture budget as well as time dependent equations for the PBL height.
7. Modified Fritsch-Chappell cumulus parameterization scheme.
8. Stable latent heating.
10. Time dependent boundary conditions.
11. Static initialization scheme.
12. LFM analysis, mandatory and significant level rawinsonde, and hourly surface data sets used to initialize the model.
13. Meso-mesh terrain, soil characteristics, sea surface temperature, ground wetness, albedo, and vegetative cover.

realistic variations in soil moisture based upon antecedent precipitation. The model’s atmospheric dependent variables are initialized from a data base valid at 1200 GMT 3 June 1980. The data base is comprised of the NWS Limited Fine Mesh (LFM) analysis, significant level rawinsonde data and hourly surface data. For the sake of brevity, the simulation made without surface sensible heating is termed adiabatic and with surface sensible heating diabatic. Depicted difference fields refer to the quantity (diabatic simulation values-adiabatic simulation values). While the simulations produced favorable forecasts over the entire United States and accounted for the severe weather in the central Great Plains, only the geographical region in proximity to the Grand Island, Nebraska tornado outbreak is displayed.

b. Basic comparison of mesoscale model simulations with and without diurnal surface heating

At 1800 GMT the adiabatic and diabatic simulations of the altimeter setting fields are similar (Figs. 11a and b). Both simulations show a subsynoptic low pressure center just northwest of Goodland, Kansas (GLD) with the diabatic version’s altimeter setting values being slightly lower. A comparison with the observations (Fig. 4b) indicates that the simulated position of the subsynoptic low is displaced slightly to the northwest of its observed position near GLD. The diabatic simulation at 1800 GMT represents the lower altimeter setting value of 101.0 kPa. With a 0.1 kPa pressure difference between the adiabatic and diabatic simulations, the gradients of the isobars around each subsynoptic low do not differ significantly. The differences in the lowest altimeter setting value are, however, beginning to reflect the effects of the 2–4°C temperature increases in the air column below the 85 kPa level in the diabatic run with 100–50 kPa thickness already 10–30 m higher in western Nebraska and Kansas when compared to the adiabatic run. By 1800 GMT the 85 kPa wind fields also reflect subtle differences between the two model runs (Figs. 12a and b). Even though the 85 kPa velocity maximum extends from north central Nebraska toward South Dakota in both experiments, the flow across northwestern Kansas has a slightly stronger southeasterly component in the diabatic version than in the adiabatic version. At this time the only observational comparison possible could be inferred from Fig. 4b where the surface winds in western Kansas and Nebraska have a weak easterly component which agrees best with the diabatic simulation.

By 0000 GMT substantial differences develop in the altimeter setting fields between the two model simulations (Figs. 11c and d). In the adiabatic simulation altimeter setting values over southern Nebraska actually rise approximately 0.1 kPa between 1800 and 0000 GMT while the altimeter setting values fall nearly 0.2 kPa in the diabatic simulation. The diabatic simulation better represents the surface pressure tendency where the observed altimeter setting falls actually exceed 0.3 kPa between GLD and GRI between 1500 and 0000 GMT (Fig. 7). In the adiabatic simulation there is virtually no pressure gradient between southwestern Nebraska and southwestern Kansas by 0000 GMT. However, the pressure gradient in the diabatic simulation is more than twice what it was at 1800 GMT (Figs. 11b and d). This trend continues through 0300 GMT.

Significant differences in the effect of the pressure gradient force on the low-level wind fields should be expected between the two model runs. The 0000 GMT simulated 85 kPa wind fields (Figs. 12c and d) reflect the effects of the different pressure gradient force within each simulation. The diabatic simulation produces an 85 kPa wind maximum in excess of 15 m s⁻¹ near DDC with an extension northward into south central South Dakota compared to a velocity value near 10 m s⁻¹ in the diabatic run in this region. The 0000 GMT 85 kPa observations indicate that the diabatic simulation velocity maximum near DDC is more accurate than the diabatic simulation which underforecasts this feature in favor of an unrealistically high velocity maximum in northeastern Nebraska and southeastern South Dakota.

Figures 13a–b contain the simulations of 85 kPa heights and temperatures at 0000 GMT for comparisons with Fig. 8g. The adiabatic simulation (Fig. 13a) is much cooler than both the observations and the diabatic simulation. This is particularly apparent from western South Dakota to western Kansas and eastern Colorado. While the diabatic simulation seems somewhat warmer than the synoptic scale analysis in this region, a closer examination indicates that this simulation may be rather accurate. If one assumes that the observed vertical lapse rate of temperature
between the surface and 85 kPa at DDC is representative of the region encompassed by DDC, LBF and Denver, Colorado (DEN), the observed surface temperatures which range from 31 to 34°C yield, when this DDC lapse rate of temperature is applied, an 85 kPa temperature range from ~26°C in northwestern Kansas to ~28°C over southeastern Colorado. It is also possible that the lack of subsynoptic 0000 GMT 85 kPa data is limiting the observed analysis. Zack (1981) was able to construct lower tropospheric temperature cross sections (utilizing 3-hourly SESAME data for the 10 April 1979 tornado outbreak) which indicate that, in the well-mixed boundary layer over the high plains, 85 kPa heating rates as simulated for this case by the diabatic version can realistically occur. The improved simulation of 85 kPa temperatures in the diabatic simulation has resulted in a much better simulation of 85 kPa heights. The diabatic simulation of 85 kPa heights much more accurately reflects the 0000 GMT observed height gradients across Kansas and Nebraska (Fig. 8g) than does the adiabatic run.

A comparison of 0000 GMT model dewpoint fields at the first sigma level (σ = 0.96) (Figs. 14a and b) further emphasizes these aforementioned differences between the diabatic and adiabatic simulations. In the diabatic simulation a pronounced tongue of higher dewpoints has been organized across north-central Kansas and east-central Nebraska while in the adiabatic simulation this feature is less well-defined. When compared with the observations, this moist tongue in the diabatic simulation accurately replicates the pattern of low level dewpoints, particularly between GRI, TOP and Wichita, Kansas (ICT) (Fig.
The differing pattern and positions of the dewpoint maxima in the adiabatic and diabatic simulations have been significantly influenced by the differing positions and intensities of the simulated subsynoptic 85 kPa wind maxima at 0000 GMT (Figs. 12c and d). The accurate positioning of the southerly wind maximum in central Kansas in the diabatic simulation is responsible for the significant transport of high dewpoint values northward into south-central Nebraska by 0000 GMT. A similar moist tongue and surface wind maximum are evident in the surface analyses (Figs. 4d and 6d).

A comparison between the two model simulations indicates a gradually increasing disparity in time between the adiabatic and diabatic versions' 50 kPa heights in the western plains (Figs. 15a–b) which can be related to the existence or absence of PBL heating in the lower troposphere. At 2100 GMT the differences (diabatic–adiabatic) in height patterns are trivial. However, by 0000 GMT height differences between the two simulations (Fig. 15c) are typically on the order of 10 m in the region between LBF and northeastern Wyoming. More importantly, the height increases in the diabatic simulation are not uniform across the model domain. There is a difference in the height increase between OMA and LBF, as there is virtually no increase at OMA and approximately a 10 m increase at LBF. This variation in the height field in the diabatic simulation is, in part, responsible for the ageostrophic winds over western Kansas and
Nebraska which modify the 50 kPa absolute vorticity pattern with the appearance of a tighter gradient of vorticity between north-central Kansas and northwestern Nebraska by 0000 GMT (Fig. 15b). By 0300 GMT the trend continues with maximum 50 kPa height differences between the simulations increasing to approximately 15 m in northwestern Nebraska, southwestern South Dakota and northeastern Colorado. Accompanying this height perturbation is a further tightening of the vorticity gradients across central and northwestern Nebraska (not shown). This 50 kPa height perturbation is in direct proximity to the region of maximum 85 kPa warming (Fig. 13b).

The 30 kPa wind and height fields are depicted in Figs. 16a and b for the two simulations for the 0000 GMT time period. By 0000 GMT the height perturbation which was maximized in western and central Nebraska at 50 kPa in the diabatic simulation is now also apparent at 30 kPa. The relatively sharp northward extension of the 9480 m isoline into South Dakota (which is also observed) (Fig. 16b) is much better resolved in the diabatic simulation than it is in the adiabatic simulation (Fig. 16a). This perturbation in the 30 kPa height field has resulted in an enhanced gradient in the 30 kPa wind velocity field northwest of GRI. This feature is particularly interesting because it has resulted in a velocity divergence increase near GRI and velocity convergence increase over northwestern Nebraska (Fig. 17c). The 0000 GMT 30 kPa velocity and velocity divergence difference fields (Figs. 16c and 17c) clearly indicate that the gradient of velocity divergence over central and

![Diagram](image_url)

**Fig. 13.** MASS adiabatic (a) and diabatic (b) simulations of 85 kPa temperature and height valid for 0000 GMT 4 June 1980. Temperatures are dashed in °C and heights are solid in meters.

![Diagram](image_url)

**Fig. 14.** MASS adiabatic (a) and diabatic (b) simulations of model level 1 (σ ~ 0.96) dewpoint valid for 0000 GMT 4 June 1980 in °C.
Fig. 15. MASS adiabatic (a) and diabatic (b) simulations of 50 kPa height and absolute vorticity valid for 0000 GMT 3–4 June 1980. Solid lines are contours of height in meters and dashed lines are isovorts in units of $10^{-2}$ s$^{-1}$. (c) Difference between MASS diabatic and adiabatic simulations of 50 kPa height in meters and MASS ageostrophic 50 kPa wind vectors and isochronals from the diabatic simulation valid for 0000 GMT 4 June 1980 in m s$^{-1}$ (vectors at alternate rows and columns only).

Fig. 16. MASS adiabatic (a) and diabatic (b) simulations of 30 kPa height, wind vectors, and isochronals valid for 0000 GMT 4 June 1980. Solid lines are contours of height in meters and dashed lines are isochronals in m s$^{-1}$ (vectors at alternate rows and columns only). (c) Difference between MASS diabatic and adiabatic simulations of 30 kPa wind speed for 0000 GMT 4 June 1980 in m s$^{-1}$. 
Comparisons of 0000 GMT rawinsonde observations to grid point 30 kPa simulated velocities reveal that there is, in general, reasonably good agreement between the diabatic MASS-simulated winds and heights and the observations. This is further reinforced by the fact that the pattern in the diabatic simulation of the 30 kPa velocity divergence field (Fig. 17b) over Nebraska is apparent in the observed 30 kPa analysis in Fig. 9b. This feature is in direct proximity to the convective complex which affects Grand Island for the next several hours. This feature highlights the differences between model simulations as it is a direct integrated response to the PBL heating and associated low-level thickness differences between both simulations. In the next section we will continue the discussion by investigating possible dynamical processes which may be responsible for the differences in the two simulations.

4. Discussion of important dynamical processes as revealed by model simulations

As was discussed earlier, it is apparent that the maximum observed mean sea level pressure falls in northwestern Kansas and southwestern Nebraska between 1500 and 0000 GMT occur just to the north and northeast of the maximum surface temperature increases between 1500 and 2100 GMT (Fig. 7). In a somewhat similar manner, the MASS diabatic and adiabatic simulations of altimeter setting tendencies are controlled by the differences in 85 kPa temperature patterns in this region (Figs. 11 and 13). Specifically, altimeter setting tends to be displaced to the north and east of the maximum of heating on the 85 kPa surface in the region including northeastern Colorado, northwestern Kansas and western Nebraska (Figs. 11 and 13). In this section of the paper we will describe the MASS model simulation of the diabatically-induced 85 kPa temperature gradient across western Nebraska and Kansas.

There are three fundamental model-simulated dynamical processes which are responsible for the development of this diabatically-induced 85 kPa temperature gradient across western Nebraska and Kansas, i.e., 1) the development of a variable depth planetary boundary layer, 2) the horizontally variable vertical flux of sensible heat within the PBL and its effects upon frontogenesis and 3) mesoscale patterns of pressure which are induced by the differential warming and their subsequent perturbation of the wind field.

The first of these model-simulated processes concerns the development of a strongly sloping planetary boundary layer over western Nebraska and Kansas. Figure 19a indicates that west of GRI the top of the model-simulated PBL extends above 85 kPa by 2100 GMT. The existence of this higher PBL lid in western Nebraska in the diabatic simulation is much stronger than it is in the adiabatic simulation. These differences continue to increase through 0300 GMT.
 FIG. 18. Cross section location for model diagnostic studies. B—represents beginning point, E—ending point, L—location nearest LBF, G—location nearest GRI, C—position indicator representative of 1/50th of total distance along the cross section between B and E.

Nebraska is a response to the higher surface elevation in this region where there is decreasing model-simulated static stability and increasing model-simulated PBL wind velocity.

The second process leading to a larger 85 kPa temperature gradient over western Nebraska and Kansas evolves from the horizontally nonuniform sensible heating within the sloping PBL. It is clear from Eq. 1 that such a spatial inhomogeneity in the depth of sensible heating can act to produce a non-uniform temperature Laplacian on an isobaric surface, i.e.,

\[
\nabla^2 \left( \frac{\partial T}{\partial t} \right)_{85 \text{ kPa}} = \nabla^2 \left( -m \mathbf{v} \cdot \nabla T - \omega \frac{\partial T}{\partial P} + \frac{\alpha}{C_p} \omega \right) + T_{\text{flux}} + T_{\text{cond/evap}} + \frac{\partial T}{\partial t_{\text{conv}}},
\]

where \( m \) is the map scale factor for the polar stereographic map projection, \( T_{\text{flux}} \) the sensible heating term, \( (\partial T/\partial t)_{\text{conv}} \) the convective heating term and \( T_{\text{cond/evap}} \) the heating (cooling) associated with stable condensation (evaporation). The 85 kPa heating which occurs between the 1800 and 2100 GMT period is not uniform over western Kansas and Nebraska, being nearly twice as large just west of LBF as opposed to just west of GRI, hence, a diabatically-induced temperature gradient field can evolve (Fig. 19a). This region where there is a simulated gradient of temperature change at 85 kPa is similar in location to the observed surface temperature change depicted in Fig. 7. The simulated 85 kPa temperature boundary apparent in Fig. 13b between GRI and the region west of LBF is largely the result of this differential heating process and subsequent differential horizontal advective processes. In addition to producing differences in the horizontal gradient of temperature the heating has produced a vertically varying temperature gradient as is indicated by the adiabatic layer west of LBF in the cross section depicted in Fig. 19b (note Fig. 18 for the location of this cross section). It is interesting to note that the simulated cross-sectional structure of the isentropes resulting from this differential heating (Figs. 20b–c) is similar to that observed by McCarthy and Koch (1982) and Carlson et al. (1983) for case studies of severe storm development in proximity to dry lines.

FIG. 19. (a) MASS diabatic simulation of the pressure at the top of the planetary boundary layer valid for 2100 GMT 3 June 1980 and 85 kPa temperature change valid for the period from 1800 to 2100 GMT 3 June 1980. Dashed lines are pressure in kPa × 10 and solid lines are temperature change in °C. (b) Cross section of MASS diabatic simulation of potential temperature (solid in K) and mass flux divergence in gm⁻¹ s⁻³ × 10⁻² valid for 2100 GMT 3 June 1980. Solid lines are mass flux divergence and dashed lines are mass flux convergence. (For location note Fig. 18).
The third model-simulated process concerns changes in the distribution of mass over western Nebraska and Kansas in response to the aforementioned thermodynamical processes. Note the development, in the cross section depicted in Fig. 19b, of a mesoscale area of mass flux divergence ($\nabla \cdot \pi v > 0$) near LBF between the 2 and 4 km levels by 2100 GMT (Fig. 19b). This feature lies within the significant horizontal gradient of temperature on the 85 kPa surface between GRI and the region west of LBF and above the downfolded 310 K isentrope. This mass flux divergence is in response to two physical processes: 1) horizontal warm air advection ($\nabla \cdot \pi v > 0$) north and east of the maximum 85 kPa sensible heating near GLD and 2) mass outflow caused by an increased eastward-directed pressure gradient force ($-m(\partial \phi / \partial x) > 0$) in the $u$ momentum equation in response to thickness increases above the region of maximum sensible heating. The result of this diabatically-induced area of mass flux divergence is to force negative values in the two terms which comprise the pressure tendency equation in $\sigma_p$ coordinates, i.e.,

$$\frac{\partial \pi}{\partial t} = \int_0^\infty \nabla \cdot \pi v d\sigma_p$$

$$= \int_0^\infty v \cdot \nabla \pi d\sigma_p + \int_0^\infty \pi \nabla \cdot v d\sigma_p, \quad (2)$$

where $\pi = P_{surface} - P_{top}$. The fundamental consequence of this process is to produce height falls within the lower troposphere (<2 km) to the north and east of the region of maximum sensible heating between 1800 and 2100 GMT as is depicted in Fig. 21a. This is consistent with quasi-geostrophic theory in which vertically differential warm air advection resulting from the sensible heating distribution is forcing low level height falls. Concurrently heights are rising above 2 km in response to the heating-induced thickness increases from western Kansas northward as is depicted in Fig. 21b. These changes in the gradient of the height field force the development of an isallobaric wind which varies in sign between the 85 and 70 kPa surfaces. Note, as is depicted in Figure 21, how the region encompassing Kansas and Ne-
braska is covered by east to northeast ageostrophic wind vectors at the 85 kPa level while at 70 kPa there are westerly ageostrophic wind vectors over northcentral Nebraska. It is interesting to note that this simulated pattern of low level ageostrophic winds is similar to that simulated by Lanicci and Carlson (1983) and to that observed by McCarthy and Koch (1982) and by Carlson et al. (1983) for case studies of severe storm development. These ageostrophic flows are associated with mass flux divergence \( \nabla \cdot \tau v > 0 \) predominating between the 2 and 4 km levels where \(-m(\partial \phi / \partial x) > 0\) while mass flux convergence \( \nabla \cdot \tau v < 0 \) predominates below 2 km where \(-m(\partial \phi / \partial x) < 0\) near LBF (Fig. 19b). This lower tropospheric feature clearly dominates the magnitude of \( \nabla \cdot \tau v \) along the cross section depicted in Fig. 19b at 2100 GMT. It should be noted that this ageostrophic flow is dominantly due to the isallobaric wind.

The development and evolution of the diabatically-induced circulation, as is simulated by MASS over central and western Nebraska, is depicted in Figs. 20a–c for 1800, 2100, and 0000 GMT. [This circulation is nonexistent in the adiabatic simulation (not shown).] During the 3-hour period between 1800 and 2100 GMT there is a significant increase in the northerly and westerly wind component near LBF between the 2 and 4 km levels while the southerly and easterly wind component gradually increases southeast of LBF below 2 km. By 0000 GMT, the lower tropospheric circulation west of LBF has contributed to the increased vertical velocity near GRI. This is the result of the convergence associated with the increasing low-level southeasterly jet near GRI. The convergence was induced by the changing pressure gradient force and the resultant isallobaric wind. The changing pressure gradient force arising from the height rises (falls) to the east (west) at 85 kPa will produce a significant northwesterly ageostrophic wind. The isallobaric–ageostrophic wind will be strongest at the location of greatest \( \nabla \cdot (\partial \phi / \partial t) \)—in northeastern Nebraska, weaker over central Nebraska, where the gradient weakens, and again stronger over southwestern Kansas (Fig. 21a). This pattern results in isallobaric convergence near GRI and at the western border of Nebraska resulting in lower tropospheric upward motion. This area of upward motion near GRI lies within the left front exit region of the upper-level jet. The intensity of this ascending motion near GRI has also been enhanced by the upper-tropospheric velocity divergence which was amplified, in part, by boundary layer heating. The heating caused height rises which modified the velocity divergence at the 30 kPa level over central Nebraska (Figs. 16 and 17). Most importantly, by forcing such a strong variation of ageostrophic winds between the 85 and 70 kPa levels, the diabatically-induced circulation over western Nebraska and Kansas is extremely effective in producing the differential advections and low-level vertical motions conducive to the development of convection near Grand Island, Nebraska after 0000 GMT. The process of establishing an environment conducive to convective development will be addressed in the next section.

5. The use of the diabatic model simulation for the prediction of severe convective storms

We will now relate the diabatically-induced circulation patterns in the aforementioned simulation to fields often used as indicators of a favorable environment for convection.

The diabatic simulation's cross sections of relative humidity, mixing ratio and equivalent potential tem-
perature indicate that two responses to the diabatically-induced circulation are playing a key role in organizing the environment for convective complex development between LBF and OMA by 0000 GMT (Fig. 22). The first response involves the development of the low-level jet primarily below 3 km over central Kansas between 1800 and 2100 GMT which propagates northward into southcentral and southeastern Nebraska by the 0000–0300 GMT time period (Figs. 12 and 22). The accelerations which comprise this low-level jet are a response to the heating and pressure falls in western and central Nebraska. This jet is responsible for the rapid northwestward transport of moisture into the region between LBF and GRI during the 1800–0000 GMT time period (Fig. 22). As can be seen in Figs. 22a and b, the mixing ratio nearly doubles in magnitude below 3 km west of GRI between 1800 and 0000 GMT. The equivalent potential temperature values, responding to this moisture transport and the diurnal PBL heating, increase by more than 10 K just west of GRI near the 2 km level during this time period (Figs. 22c and d). The velocity convergence associated with this low-level jet is also responsible for producing significant mesoscale values of upward motion. The simulated upward motion maximum near GRI exists over a very shallow layer just below the 2 km level during the 0000 to 0300 GMT time period (Fig. 21c). Note the rapid increase in the relative humidity values near the 2 km level between LBF and GRI in response to both the moisture transport and upward motion during the 1800–0000 GMT time period as is depicted in Figs. 22c–d. This is the time period just before observed convection develops west of GRI.

The second process in the diabatic model simulation which will contribute to the development of convection near GRI can be seen in Fig. 22. The feature consists of an area of relatively dry air near the 4 km level which moves from northwest of LBF at 2100 GMT southeastward to just west of GRI at 0000 GMT (Figs. 21 and 22). The effect of this dry air advection, which is enhanced by the eastward-directed isallobaric wind, is to promote the southeastward transport of relatively dry air (Fig. 22). This relatively dry air moves from west of LBF at 1800 GMT to west of GRI by 0000 GMT (Figs. 22c and d). This dry air advection acts to enhance the potential instability as a minimum in the equivalent potential temperature field develops near the 4 km level by 0000 GMT (Fig. 22f). This minimum eventually phases with the low-level moisture maximum near 1 km which is the result of the northwestward moisture transport associated with the low-level jet which is evident just west of GRI by 0000 GMT (Fig. 22). These processes lead to a maximization of the vertical gradient of the equivalent potential temperature near GRI between 0000 and 0300 GMT (Figs. 22f and g). This differential moisture transport is similar to the elevated mixed layer phenomena described in Carlson and Ludlum (1968) and Carlson et al. (1983). This phenomena occurs when the moist air flows under dry air which is being advected away from a region of elevated terrain. The elevated mixed layer is often observed ahead of the dryline itself.

The effect of this low-level moisture advection on the 50 kPa lifted index field is quite dramatic as can be seen in Fig. 23. Between 1800 and 0000 GMT the combined action of surface heating and low-level moisture transport act to decrease the LI values dramatically in the region between LBF and GRI. The simulated LI value of −11°C at 0000 GMT is just west of the observed initiation of convection described earlier (Fig. 1). Thus, the environment, as simulated by the diabatic model, is being enhanced for convection very close to its observed location of development, i.e., the instability is being maximized and a mechanism for its release established close to where the observed convective complex develops.

It should be noted that the MASS model simulation of 50 kPa lifted index and 70 kPa vertical motion has been diagnosed to be a reliable indicator for mesoscale convective development in a large number of case studies exhibiting a wide range of synoptic variability (Koch, 1983). This model evaluation study indicated that mesoscale processes similar to those described in this paper for the Grand Island case study are commonly produced by this model and frequently relate well to convection.

6. Summary

Surface analyses and numerical simulation sensitivity studies are compared for a case of isolated tornadic convective complex development (3–4 June, 1980). Several tornadic storms occurred over the limited geographical region in and around Grand Island, Nebraska during a relatively weak synoptic scale flow pattern.

The case study observational analyses reveal the following: 1) the spatial distribution of surface heating and subsynoptic pressure falls are highly correlated, which is consistent with the findings of Whiting and Bailey (1957), Tegtmeier (1974), and Moller (1980) and 2) the evolution of observed surface winds over subsynoptic time scales appears to be linked to the changing gradients of the observed surface pressure 3–6 hours prior to the severe convection.

The mesoscale model simulations with diurnal heating better represent these observed changes in the surface fields on subsynoptic time and space scales than simulations without diurnal surface heating. Furthermore, many of the often-observed characteristic features associated with the environment prior to and during convective complex development (subsynoptic low pressure centers, low-level jet streams,
isolated zones of significant upward motion and rapid destabilization) are more realistically simulated with the inclusion of diurnal surface heating in this case study than without such heating. A very significant finding was that the middle-to-upper tropospheric subsynoptic patterns of velocity divergence and absolute vorticity often associated with jet streaks were influenced by the inclusion of diurnal surface heating and the subsequent development of a deep well-mixed planetary boundary layer. The sensible heating acted to enhance the middle-to-upper level divergence at and below the jet streak level and the isallobaric forcing of the low-level jet. Thus the diabatic heating in the boundary layer adds another component to the complex interaction between the lower and upper troposphere which can significantly influence the pre-convective environment (House, 1961; Kaplan and Paine, 1977; and Uccellini and Johnson, 1979).
There are several fundamental implications of these findings. The most significant finding concerns the fact that mesoscale modelers employing simple PBL representations are likely limiting their ability to properly represent important subsynoptic dynamical adjustments prior to and during convective complex development. A second significant finding is that although the existing adiabatic-dynamic models of upper and lower tropospheric jet streak coupling provide a background signal for the pre-storm environment, the effects of surface heating within deep well-mixed planetary boundary layers could act to significantly enhance the development of the low-level jet and influence the evolution of the entire jet streak system. Finally, based upon the results of these simulation experiments, there still are significant deficiencies in the model utilized in this case study: 1) MASS 3.0, with the cumulus parameterization scheme turned off, failed to concentrate the magnitude of pressure falls at a sufficiently short length scale when compared to the observed evolution of a mesoscale low pressure/convective complex system and 2) MASS 3.0 evidently underpredicted some of the observed mesoscale structure in the PBL features possibly due
to insufficient information concerning soil moisture variations in the surface data base which was employed. Future model experiments, which will employ a nested grid version of MASS 3.0, will be designed to study the role of convective latent heat and soil moisture variations in modifying the aforementioned circulation patterns.

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