Diabatically Driven Discrete Propagation of Surface Fronts: A Numerical Analysis

GEORGE H. BRYAN AND J. MICHAEL FRITSCH
Department of Meteorology, The Pennsylvania State University, University Park, Pennsylvania

(Manuscript received 2 July 1998, in final form 30 August 1999)

ABSTRACT

Discrete frontal propagation has been identified as a process whereby a surface front discontinuously moves forward, without evidence of frontal passage across a mesoscale region. Numerical simulations are employed to examine the upper-level evolution of a discrete frontal propagation event and to explore the processes that were responsible for the discrete movement.

Model results indicate that a frontal pressure trough was not able to penetrate through a deep surface-based layer of cool air created by a precipitating convective system several hundred kilometers in advance of the front. Meanwhile, a new low-level baroclinic zone formed well ahead of the front along the southern side of the cool layer. As the midlevel front moved continuously over the cool layer, a new low-level front developed in the new baroclinic zone and the original low-level front dissipated. At the surface, the simulated front did not pass through the cool layer.

Frontogenesis terms reveal that the prefrontal circulation that becomes the new frontal circulation initially forms directly from diabatic frontogenesis. Daytime heating in the prefrontal boundary layer and cooling from thunderstorms combine to create a thermal gradient and a mesoscale pressure perturbation. Winds turn in response to the altered pressure field and form a convergent boundary, resulting in kinematic frontogenesis. The boundary subsequently undergoes rapid intensification.

Sensitivity studies were conducted in which latent heating due to precipitation was withheld and the influence of clouds on the radiation scheme was ignored. In a simulation with both of these effects withheld, the original front passes continuously through the region, that is, there is no discrete propagation. Thus, diabatic processes associated with a large complex of thunderstorms were necessary to induce the discrete frontal propagation in this case. This conclusion contrasts with previous studies, where fronts were observed to propagate discretely in dry environments.

1. Introduction

Discrete surface frontal propagation is defined as discontinuous movement of a front at the surface. Bryan and Fritsch (2000) examined a case of discrete frontal propagation in a convective environment. In their case, a prefrontal convective system generated a large, deep cold dome just ahead of an approaching cold front. The low-level cold front was observed to stall and undergo frontolysis as it encountered the cold dome. Meanwhile, a trough formed on the downstream side of the cold dome and underwent rapid frontogenesis. Eventually, the original surface frontal trough disappeared, the mesoscale wind field adjusted to the new pressure pattern, and the prefrontal trough replaced the original front as the dominant frontal feature. Their analysis suggests that, although the low-level frontal properties (e.g., pressure trough, wind shift, and temperature gradient) were disrupted by the deep, stable cold dome created by the convection, the midlevel front propagated continuously.

This paper presents a numerical simulation of the case studied by Bryan and Fritsch (2000, hereafter referred to as BF). A number of aspects of this case warrant further study. In particular, calculations of surface frontogenetical tendencies in BF suggest that diabatic processes played an important role in initiating the prefrontal trough. Output from a numerical simulation can be used to confirm or refute this suggestion. Furthermore, it remains to be seen whether the diabatic processes were completely necessary in inducing the discrete frontal movement in this case. In a study of discrete frontal propagation in stable winter conditions, Charney and Fritsch (1999) found that cold fronts can propagate discretely without the need for precipitation processes. On the other hand, there is evidence that diabatic processes may be important in initiating frontogenesis by creating mass field changes and, eventually, wind field changes as the system adjusts toward geostrophic balance. The mesoscale wind shift lines and troughs may then undergo frontogenesis. For example, several studies (e.g., Purdom 1982; Pielke and Segal 1986; Segal...
et al. 1986) note that gradients in cloud cover can induce mesoscale circulations similar to sea and land breezes. Another diabatic process that can generate mesoscale boundaries is precipitation. Doswell (1982) showed that thunderstorm complexes can form intense outflow boundaries with properties similar to cold fronts. Ryan et al. (1989) presented a case where precipitation generated a new baroclinic zone several hundred kilometers in advance of an approaching cold front.

This paper examines the 5–6 September 1994 discrete frontal propagation event using a mesoscale-resolution simulation. The model output complements the observational analysis and adds insight that cannot be drawn from standard analyses. The role of physical processes are explored by examining output from the various model equations. Section 2 describes the numerical model configuration. Section 3 examines the simulated evolution of the event, and compares the results with observations. Section 4 presents the vertical structure and evolution of discrete frontal propagation. Analysis of frontogenesis terms and quantification of diabatic processes are presented in section 5. The results of several sensitivity experiments using the numerical model are included in section 6. A conceptual model for discrete frontal propagation is presented in section 7, followed by conclusions in section 8.

2. Model design

The fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5) is utilized for this investigation (Dudhia 1993; Grell et al. 1994). MM5 is a nonhydrostatic model with prognostic equations for temperature, horizontal, and vertical components of motion, pressure, water vapor mixing ratio, cloud water mixing ratio, and rain-water mixing ratio. Twenty-five sigma ($\sigma$) levels are used in this study with 11 levels below 700 hPa where boundary layer processes are important. Two model domains are used: a coarse mesh of $61 \times 73$ grid points with 60-km grid spacing, and a two-way interactive nested fine mesh domain of $64 \times 76$ grid points with 20 km grid spacing (Fig. 1). The location of the nested fine mesh domain was chosen such that information from the outer boundaries of domain 1 would not be advected into domain 2 during the model simulation. This is important for dry run experiments since the boundary conditions for domain 1 are observed conditions that include precipitation processes.

Model physics used in these simulations include the explicit bulk precipitation scheme with simple ice physics based on the studies of Lin et al. (1983), Rutledge and Hobbs (1983), Hsie et al. (1984), and Dudhia (1989). The planetary boundary layer (PBL) parameterization is based on the work of Blackadar (1976, 1979) and adapted for MM5 by Zhang and Anthes (1982) and Zhang and Fritsch (1986). The subgrid-scale convective parameterization used in the simulation is the scheme developed by Kain and Fritsch (1990). Of particular value to this study, the Kain–Fritsch scheme includes parameterized downdrafts necessary for simulating the convectively modified low levels.

The model initial conditions are derived from first-guess fields supplied by the National Centers for Environmental Prediction that are then objectively enhanced with observed surface and upper-air data (Manning and Haagenson 1992). At the beginning of the forecast, the four-dimensional data assimilation technique of Stauffer and Seaman (1994) is utilized to bring the dynamic and thermodynamic fields into closer agreement with the observations while still satisfying the governing hydrodynamical system.

The numerical simulation of the 5–6 September 1994 discrete frontal propagation event was initialized at 0000 UTC 5 September 1994 (0000/5) and run for 36 h, ending at 1200/6. For the first 12 h, four-dimensional data assimilation was applied. This simulation is hereafter referred to as the control simulation (CTL).

For cross-section analyses of model output, averages of several cross sections oriented perpendicular to the fronts are constructed. The averages eliminate high-frequency perturbations (such as those associated with moist parameterized convection at individual grid elements) but still retain the mesoscale structure and vertical circulations associated with the fronts. The methodology of the cross section averaging is described in the appendix.

3. Surface and upper-air evolution

Despite minor differences in the timing and magnitude of features, the model is able to capture the basic sequence of events in this case (Fig. 2). At the surface, a prefrontal mesohigh and trough form at about the correct time and location. Over time, the surface trough intensifies and moves southward while the original front slows, stalls, weakens, and dissipates. After the original
Fig. 2. Surface analysis from the fine mesh domain at (a) 1200 and (b) 1800 UTC 5 Sep and (c) 0000 and (d) 0600 UTC 6 Sep 1994. Sea level pressure is contoured every 1 hPa (solid) and temperature is contoured every 2°C (dashed).

front dissipates, large-scale pressure and wind fields adjust to the new frontal feature. Based on the fact that discrete frontal propagation is captured by the model, it is assumed that the basic processes are represented in the model output. Therefore, the model output can be used to draw conclusions about the mechanisms responsible for discrete frontal propagation.

Looking more carefully at the model output, conditions at 1200/5 capture the cold and warm fronts at the surface with good placement (cf. Fig. 2a with BF Fig. 4a). By 1800/5 a pronounced mesohigh and cold pool have developed ahead of the cold front (Fig. 2b). At this time, a prefrontal trough (PFT) is also analyzed. Winds have responded to the interfront pressure rises, creating confluence and thus frontogenetic tendencies along the PFT. It is also important to note that the model is able to simulate the warm pocket that develops behind the original front in central Kansas. Thus, the surface temperature gradient at the front is locally (and temporarily) reversed from what one would expect at a cold front.

At 0000/6, frontolysis is occurring at the original front (Fig. 2c). The front is still identifiable by a pressure trough and wind shift in southeastern Kansas and north-central Oklahoma, though no strong temperature gradient exists. The PFT at this time continues to intensify and move southward. By 0600/6, the pressure trough and convergence zone associated with the front are virtually nonexistent and identification of a surface front can no longer be justified (Fig. 2d). As the frontal trough disappears, the interfront mesohigh becomes incorporated into the synoptic-scale high pressure system to the north. Meanwhile, pressure and wind fields adjust to the new surface front. It is readily apparent in this sequence (Fig. 2) that there is a large area through which the original front does not propagate at the surface.

Simulated precipitation is shown in Fig. 3. The model generates a broad swath of precipitation in the interfront
region, in roughly the same area that radar echoes were observed (cf. Figs. 3 and 5 of BF). The model also reproduces several specific convective features; for example, a convective system is predicted behind the front in Kansas.

The evolution of conditions at upper levels also compares favorably with observations. At 850 hPa, a single frontal trough is apparent in the height and wind fields at 1200/5 (Fig. 4a). The thermal pattern, marked by a warm ribbon to the south and a strong temperature gradient to the north, also supports the existence of a single front at this level.

By 1800/5, a more complicated thermal pattern emerges at 850 hPa due to the introduction of a cold anomaly in the interfront region (Fig. 4b). Now there are two warm ribbons: one above the original front, and one above the developing prefrontal trough. A weak height ridge in the same location as the cold anomaly suggests that the cold dome extends farther above 850 hPa. Over most of the PFT, a second trough and wind shift do not develop at this level; the exception is along the far eastern end of the PFT. This trough is associated with a shortwave that moves rapidly through the northwest flow, mostly to the east of where the discrete propagation occurs. It may have been influential along the eastern end of the PFT, but a height trough (and associated wind shift) does not emerge over the PFT in Oklahoma or Texas. Furthermore, this trough is later observed to move ahead of and away from the PFT.
(new front), further suggesting that it was not critical in the development of the PFT. This evolution at 850-hPa contrasts with BF, which shows a double structure at 850 hPa as well as the surface. It will be shown later that a second upper-level trough does form in this simulation, but it does not initially reach 850 hPa. Thus, the developing new front in the simulation is not as deep as the observed system.

Six hours later at 0000/6, the cold anomaly at 850 hPa is not as well pronounced, as it begins to merge with the postfrontal temperature gradient (Fig. 4c). A warm pocket over eastern Kansas and northern Missouri is all that remains of the warm ribbon that was located along the original front. The shortwave trough at the eastern end of the new front continues to deepen. Later, this shortwave moves away from the PFT. During the next 12 h, the thermal field continues to adjust to the new frontal location (not shown).

Rapid changes also occur at 700 hPa. Like at lower levels, there is a single frontal structure at 1200/5 (Fig. 5a). However, there is already an indication of a cold anomaly just ahead of the front. By 1800/5, there is already a pronounced interfront cold anomaly (Fig. 5b), as seen at 850 hPa. Unlike at lower levels, locally higher heights do not develop in the interfront region. Instead, lower heights prevail, which is consistent with the observations (BF) and Zhang and Harvey (1995, a study of convection ahead of a cold front, which will be discussed later). Like at 850 hPa, deepening heights along the eastern end of the PFT can be seen through this sequence. A height trough and wind shift do not develop over most of the PFT at this level. Rather the original height trough of the front moves continuously through Oklahoma and western Texas (Fig. 5c).

Farther aloft at 500 hPa, it is difficult to identify a strong frontal structure (Fig. 6). This is consistent with observed warm season fronts in other studies (e.g., Sanders 1955; Ogura and Portis 1982; Reeder and Smith 1992). However, the strongest temperature gradient in the postfrontal region at 1200/5 (Fig. 6a) moves southward into the interfront region. This pattern is interrupted at 0000/6 by the appearance of a warm anomaly, apparently induced by the convection (Fig. 6b). In addition, a weak shortwave trough intensifies in the interfront region. The emergence of a shortwave trough during the period of discrete frontal propagation has implications for the conceptual model of this event. In a case study of convection ahead of a cold front, Zhang and Harvey (1995) show that a mesoscale convective system acted to intensify the midlevel height trough.

Fig. 5. Simulated fields at 700 hPa at (a) 1200 and (b) 1800 UTC 5 Sep and (c) 0000 UTC 6 Sep 1994: geopotential height (solid contours, 180 = 3180) every 15 m, and temperature (dashed) every 2°C [an intermediate contour of 7°C is included in (b) and (c)]. Positions of surface fronts are indicated in gray. A low-pass filter (Barnes 1964) was applied to the fields.
Fig. 6. Simulated fields at 500 hPa at (a) 1200 UTC 5 Sep and (b) 0000 UTC 6 Sep 1994: geopotential height (solid contours, 580 = 5580) every 20 m, and temperature (dashed) every 2°C. Positions of surface fronts are indicated in gray. A low-pass filter (Barnes 1964) was applied to the fields.

Fig. 7. Response curves used for two low-pass filters (B1 and B2) and the response curve used for the bandpass filter (BP). The BP is obtained by subtracting B2 from B1 and then normalizing the resulting curve to a maximum value of 1.0. See Maddox (1980) for additional details.

They further show that this convectively induced intensification acts to enhance synoptic-scale frontogenesis and cyclogenesis.

4. Vertical structure

a. Synoptic view

To highlight the vertical structure of the frontal trough, a bandpass filter was utilized. Following Maddox (1980), two low-pass filters are applied (Barnes 1964, 1973). One filter effectively eliminates wavelengths less than 150 km (B1 in Fig. 7); the second filter eliminates wavelengths less than 900 km (B2 in Fig. 7). The bandpass response curve is obtained by subtracting B2 from B1. The resulting response curve is then normalized to a maximum value of 1.0. The bandpass response curve used here (BP in Fig. 7) highlights features with a wavelength of about 400 km, that is, mesoscale wavelengths.

Vertical cross sections of bandpass geopotential height reveal that the frontal trough has a forward-sloping structure at 1300/5 (Fig. 8a), similar to that noted by BF in their analyses of the observations. Later, the midlevel trough amplifies and detaches from the surface trough (Fig. 8b). As this occurs, the PFT forms and intensifies. Notice that the trough that was observed over the far eastern end of the PFT at 850 and 700 hPa (Figs. 4 and 5) does not appear over the PFT in this cross section; this observation further supports the conclusion that the shortwave trough was not instrumental in the development of the PFT. Later, the midlevel trough connects with the PFT (Fig. 8c). Low heights associated with the original low-level front fill in and disappear. The resulting structure is a continuous height trough from the surface to midlevels, which is more tilted back than was noted earlier. These bandpass height analyses suggest that the front aloft propagates continuously while discrete frontal propagation occurs in low levels. This agrees with the observational study (BF), where a double-frontal structure was found only in the lower troposphere.

Large-scale cross sections of potential temperature further reveal the overall structure and evolution of the frontal features (Fig. 9). Early in the simulation, there
Fig. 8. Average cross sections of bandpass height (contour interval 2 m, zero contour excluded) at (a) 1300/5, (b) 2300/5, and (c) 1100/6. Data are from the coarse mesh domain. The location of the cross section is approximately from western Nebraska to southern Louisiana.
Fig. 9. Average cross sections of potential temperature (contours, K) and vertical velocity (shaded, representing values greater than 2 cm s$^{-1}$) from the coarse mesh domain at (a) 1200/5, (b) 1600/5, (c) 2200/5, and (d) 0000/6. The location of the cross sections is from South Dakota (left) to the Gulf of Mexico south of Louisiana (right).

b. Mesoscale structure

As noted in BF, a sudden deep-layer wind adjustment to the new low-level front occurs near the completion of discrete surface frontal propagation. This process can be seen in the model simulation also. Cross sections of wind speed normal to the fronts show that the original convergent pattern associated with one front (Fig. 10a) is replaced by a double structure (Fig. 10b). As the original low-level front dissipates, winds just ahead of the front turn to northerly (Fig. 10c), resulting in only one region of contrasting cross-frontal winds—along the PFT. In this same time period, vertical motion at the original low-level front undergoes a rapid transition from ascent to descent (Figs. 11b and 11c). The ascending circulation at the PFT intensifies throughout this time period. By 2300/5, the only appreciable region of ascent is associated with the new frontal circulation. Combining these vertical and horizontal motions together, the mesoscale flow surges downgradient and toward the PFT as the discrete frontal jump occurs. This sequence occurs as heights fall along the PFT and as heights rise along the original front; that is, the original frontal trough dissipates and becomes incorporated into the synoptic-scale high pressure to the north. As this occurs, low-level convergence and ascent aloft cease.

Further insight can be drawn by inspecting the potential temperature pattern together with the horizontal wind field. Along the PFT, frontogenetic tendencies are strong due to a collocation of temperature gradient and convergence (cf. Figs. 10b,c and 12b,c). In contrast, convergence along the original low-level front is not frontogenetic since virtually no favorable horizontal temperature gradient exists there. Thus, frontogenetical support for the original front is disrupted by the very cold (relative to temperatures behind the cold front) prefrontal environment generated by the persistent convection.

Another potential factor in frontolysis along the original front is the introduction of substantially cooler air just ahead of it. The presence of this cool air weakens the horizontal temperature gradient and thereby reduces the rate of frontogenesis. In addition, the model reveals that ascent is quickly replaced by descent at this time, so tilting frontogenesis ceases along the original front. Thus, there is no mechanism to support low-level frontogenesis and the surface front weakens. Hypothetically, if the original frontal trough was more intense, the frontal trough might not have been canceled by the cold pool–induced high pressure anomaly, and the front...
Fig. 10. Average cross sections of horizontal wind speed normal to the front ($u$, m s$^{-1}$) from the fine mesh domain: (a) 1200, (b) 1600, and (c) 2300 UTC 5 Sep 1994. Contour interval is 2 m s$^{-1}$.

Fig. 11. Average cross sections of vertical velocity ($w$, cm s$^{-1}$) from the fine mesh domain: (a) 1200, (b) 1600, and (c) 2300 UTC 5 Sep 1994. Contour interval is 2 cm s$^{-1}$, The zero contour is excluded.
would have propagated continuously through the cold pool. In fact, this is observed to occur: fronts can encounter low-level, convectively generated cold pools and pass through the area seemingly unaffected (e.g., Doswell 1982). The key difference in this case appears to be the depth and intensity of the prefrontal stable layer, and a shallow front that does not penetrate through the stable layer. Thus, the surface frontal trough weakens on one side of the cold dome and emerges on the other side.

5. Analysis of frontogenesis

To quantify the frontogenetical tendencies that were inferred from the thermal and wind fields, the frontogenesis equation (Miller 1948) is utilized:

\[ F = \frac{D}{Dt} \left| \nabla \theta \right|, \tag{1} \]

where \( \theta \) is potential temperature. For a cross section oriented perpendicular to the front (i.e., for the cross sections that are presented here), a two-dimensional frontogenesis equation can be used:

\[ F = \frac{D}{Dt} \frac{\partial \theta}{\partial x}, \tag{2} \]

which can be expanded to yield

\[ F = -\frac{\partial u}{\partial x} \frac{\partial \theta}{\partial x} - \frac{\partial w}{\partial x} \frac{\partial \theta}{\partial z} - \frac{\partial w}{\partial y} \frac{\partial \theta}{\partial z} + \frac{\partial q}{\partial x}. \tag{3} \]

where positive \( x \) points toward the warm air (to the right on the cross sections), positive \( y \) points into the cross section, positive \( z \) points upward, and \( q \) is the heating tendency due to diabatic processes. The first term quantifies horizontal confluence deformation, and will hereafter be referred to as \( F_1 \). The second term is horizontal shearing deformation. Since this term has been found to be negligible compared to the other terms in this case, it will not be presented. The third term is the tilting term (\( F_2 \)). The fourth term is the diabatic term (\( F_3 \)). Calculations of \( q \) were obtained directly from the numerical model. Temperature tendencies from the following effects were summed to yield \( q \): the explicit moisture scheme, the Kain–Fritsch convective parameterization, the Blackadar PBL parameterization, longwave and shortwave radiation, and turbulent diffusion. Fifteen-minute averages of these terms were computed, where the average comprises data from every time step during model integration.

At 1600 UTC, the prefrontal cold dome is beginning to emerge (Fig. 12b). The \( q \) field supports the hypothesis made earlier that diabatic cooling generated this feature (Fig. 13a). Analysis of the individual heating terms (not shown) confirms that evaporative cooling from the explicit moisture scheme and parameterized cooling from the Kain–Fritsch scheme are responsible for this cooling. Near the surface on either side of the cold dome, there exists warming from the PBL scheme, associated with turbulent mixing of sensible heat flux from the surface. The strong heating in the PBL behind the front agrees with observations in BF, and is the cause of the
reversal of temperature gradient there. In contrast, strong differential heating exists across the zone where the prefrontal trough forms. Later, at 2300 UTC, a deep layer of cooling continues in the interfront zone, although the PBL heating becomes negligible (Fig. 13b).

The frontogenesis terms at 1600/5 show two low-level maxima: one behind the original front, and one where the PFT develops (Fig. 14d). There is a deep layer of positive F3 about 300 km ahead of the front, where the PFT forms. Above the surface, F2 mostly cancels with F3, which is consistent with other studies (e.g., Ballentine 1980; Orlanski et al. 1985; Koch et al. 1995). However, near the surface where $w$ is necessarily small, F3 is not counteracted by tilting. Thus, diabatic frontogenesis is a strong contributor to the creation of the PFT. An aspect of frontogenesis that was inferred from other figures but quantified here is frontolysis at the original front caused by cooling ahead of the front and by warming behind the front (Fig. 14c). Also, as suggested earlier, confluence frontogenesis is negligible along the front (Fig. 14a), despite well-defined convergence there (Fig. 10a).

Over the next several hours, kinematic frontogenesis (F1 and F2) dramatically increase in magnitude and area along the prefrontal trough (Fig. 15). The terms F1, F2, and F3 all contribute to frontogenesis along the PFT. The corresponding plot of total frontogenesis shows an intense maximum near the surface. Near the original front, no significant values of frontogenesis exist. These data support the PFT as the dominant frontogenetic feature by this time.

Hourly plots of the frontogenesis terms near the PFT show that positive values of F3 appear along the PFT before the other two terms (Fig. 16). This time-height plot neglects the complete three-dimensional structure of frontogenesis, including the typical dipole structure of the tilting term [as shown in Figs. 14b and 15b, as well as other studies (e.g., Sanders 1955; Baldwin et al. 1984; Koch et al. 1995)]. Nevertheless, these data support the conclusion inferred from the thermal and wind evolution that diabatic processes initiated the creation of a new low-level thermal gradient. Then the pressure and therefore wind field adjusted to the new thermal pattern, initiating confluence deformation and tilting frontogenesis, which further intensified the thermal gradient. This model contrasts with the quasigeostrophic view of frontogenesis (e.g., Hoskins 1982; Bluestein 1986), where large-scale cyclogenetic processes impose horizontal deformation onto a preexisting low-level baroclinic zone. In the present case, there is no preexisting baroclinic zone—it is created by the diabatic processes before the usual frontogenetic processes can ensue.

These conclusions agree with several studies that show how sensible heating from PBL processes can enhance frontal structure (e.g., Keyser and Anthes 1982; Thorpe and Nash 1984; Shapiro 1983; Keyser 1986, section 10.3.1; Moore 1991; Blumen et al. 1996; Miller et al. 1996; Koch et al. 1997). A recent numerical study by Koch et al. (1995) is highly relevant to this work; they found that differential PBL heating enhanced frontogenesis when compared to numerical simulations of uniform PBL heating. The results of the present paper confirm that differential PBL processes are strongly frontogenetic. Furthermore, this paper shows that PBL processes can play a role in creating new ageostrophic circulations by creating horizontal temperatures gradients that did not previously exist.

6. Sensitivity studies

To see how sensitive discrete frontal propagation is to diabatic processes, additional simulations were conducted in which one or more physical parameters were withheld from the model solution. The first of these is a dry run simulation (hereafter referred to as DRY1) in which latent heating due to precipitation is neglected. Specifically, heating tendencies from the explicit mois-
ture scheme (which includes evaporation and melting of hydrometeors) and the convective parameterization are not considered. Moisture still exists in the simulation and is included in virtual temperature calculations and cloud-radiative effects. Thus, this simulation is not truly dry. To make sure that the presence of moisture does not affect the momentum fields, water loading effects are not considered.

Output from DRY1 clearly retains a signal of discrete propagation (Fig. 17a). The primary factor leading to discrete frontal propagation can be seen in the surface thermal field: a temperature gradient develops well ahead of the front. Analysis of the model output reveals that the presence of cloud water in conjunction with the radiation and PBL schemes is primarily responsible for creating the thermal gradient. Although latent heating due to phase changes is not considered, clouds are still created by the explicit and implicit moisture schemes. Other physical processes (such as radiation) interact with these clouds, so warming occurs in the prefrontal clear air while conditions under the clouds remain cool. Thus, there is an indirect effect of condensation/sublimation on the generation of a prefrontal boundary: cloud patterns induce differential solar heating, and thus diabatic frontogenesis. This result is similar to studies that have noted developing mesoscale circulations at cloud boundaries (e.g., Purdom 1982; Segal et al. 1986).

Based on these results, a second sensitivity experiment includes latent heating from phase changes, but neglects clouds in the radiation and PBL schemes. In effect, this is a clear sky radiation experiment (hereafter referred to as CLR). This simulation also contains a signal of discrete frontal propagation in the solution (Fig. 17b). Once again, a strong thermal gradient is created well ahead of the approaching front. This time the gradient is created primarily by the precipitation-induced cooling that occurs in the interfront region.

The final sensitivity experiment eliminates heating due to phase changes and imposes clear sky radiation conditions, that is, no clouds or precipitation. This experiment (hereafter referred to as DRY2) does not feature a discrete propagation of the front (Fig. 17c).
er, the original low-level front moves continuously through Oklahoma and Missouri.

From these sensitivity experiments, it is clear that diabatic processes were necessary for the discrete frontal propagation in this case. This is in contrast to the case studied by Charney and Fritsch (1999), which propagated discretely through adiabatic processes alone.

To highlight the magnitude and extent of diabatic processes, difference fields between the various experiments were calculated. By subtracting DRY2 from CTL, the effects that condensation and sublimation have on the model solution, both direct and indirect, are revealed. The interfront mesohigh and anticyclonic circulation are clearly evident in the difference fields (Fig. 18a). The mesohigh is coincident with a broad region of cooler temperatures throughout the interfront zone (Fig. 18b). As suggested earlier, the depth and strength of the prefrontal cold anomaly should play a role in discrete frontal propagation by determining whether frontogenetical tendencies can continue unabated at low levels. Cross sections of temperature difference taken through the center of the surface high pressure anomaly reveal that the primary cold anomaly (the interfront cold dome) extends to about 700 hPa. This depth is consistent with other studies (e.g., Lewis 1975; Kuo and Anthes 1984; Cotton et al. 1989; Ryan et al. 1989). Above 600 hPa, warm anomalies from latent heat release are apparent. The strongest forcing for the discrete frontal propagation, revealed by the highest perturbation values and strongest gradients, develops near the surface, but extends into the midtroposphere.

A cross section of the geopotential height difference between CTL and DRY2 shows the vertical extent of convective effects. The cross section at 1700/5 shows a typical convective signature, with high heights near the surface, low heights in midlevels, and high heights farther aloft in the upper troposphere (Fig. 19a). This plot also suggests that amplification of midlevel (700 and 500 hPa) height troughs over the interfront region is a direct result of convective warming above these levels [which agrees with Zhang and Harvey (1995) and Stensrud (1996)] and is consistent with the quasigeostrophic height tendency equation (Hirschberg and Fritsch 1993). Later, at 2300/5, the same pattern can be
Fig. 16. Time series of frontogenesis terms along the prefrontal trough. Here $F_1$, $F_2$, and $F_3$ are confluence frontogenesis, tilting frontogenesis, and diabatic frontogenesis, respectively; and $F$ is the sum of $F_1$, $F_2$, and $F_3$. Contour interval is $4 \times 10^{-9} \text{K m}^{-1} \text{s}^{-1}$.

seen. But, in addition, low heights are evident over a deep layer at the location of the PFT (Fig. 19b). This figure shows that a deep layer of height falls occur at the prefrontal trough during discrete frontal propagation. Also note that high height anomalies exist at low levels near the location of the original front, showing how the height trough associated with the original front fills in.

7. Conceptual model

An example of discrete frontal propagation induced by convection has been presented. Based on this analysis, discrete frontal propagation can be viewed as a process whereby a midlevel front moves through an area aloft without evidence of a front moving through the same area at the surface, while the surface frontal properties (e.g., pressure trough, wind shift, and thermal gradient) dissipate at one location and simultaneously develop at another location. Discrete movement in this case is initiated when the cold front encounters a deep cold dome generated by moist convection. The frontal trough did not penetrate through the deep stable layer. Rather, in effect the low-level front discretely moved into a new, more favorable position for development on the south side of the cold dome rather than continuously propagating through the unfavorable intervening area.

Based on observations and a numerical simulation of the event, a conceptual model is proposed for discrete frontal propagation in a convective environment. Since other cases of discrete frontal propagation will likely differ from this event in minor ways, this model only stresses the important processes involved and the basic structural evolution. The first step in the conceptual model is the creation of a deep ($>1 \text{ km}$) and long-lived ($>3 \text{ h}$) convectively induced cold dome ahead of an approaching cold front. Convection must be persistent, so that a deep, coherent mesoscale cold dome can be created and maintained by individual thunderstorms or thunderstorm complexes. The cold dome must be long-lived, since if it is not (i.e., if the cold pool and surface mesohigh quickly dissipate), the original surface trough
will simply slow and then move continuously. The cold anomaly must be deep/strong enough to induce a high pressure anomaly at the surface. Low-level winds turn in response to this newly formed mesohigh, creating convergence and, thus, a focused region for new frontogenesis.

The cold dome is created ahead of the approaching front. When the front encounters the cold dome, the cross-frontal temperature gradient is reduced, inhibiting horizontal confluence frontogenesis. Tilting frontogenesis is also reduced, since ascent above the front must lift higher static stability air than previously. The low-level frontal structure thus weakens.

A crucial element of discrete frontal propagation involves the creation of a new low-level frontogenetic trough. Initially, diabatic frontogenesis occurs along the boundary between the convectively created cold air and a prefrontal mixed boundary layer. As winds adjust to the new trough, confluence frontogenesis intensifies the thermal gradient. A vertical, thermally direct, ageostrophic circulation is created. Near the completion of discrete movement, the original frontal trough disappears and is incorporated into the postfrontal high pressure system. As the original frontal trough weakens, a mesoscale adjustment of the low-level wind field occurs.

8. Conclusions

A numerical simulation is used to investigate the vertical structure of discrete frontal propagation on 5–6 September 1994. The results confirm BF's observational analysis that the front aloft propagates continuously over a deep, convectively generated cold dome. A new baroclinic zone forms between the cold dome and a well-mixed planetary boundary layer. The new low-level front develops within this baroclinic zone while the original low-level front dissipates.

The important role of diabatic processes in discrete frontal propagation has been revealed in this study. A prefrontal trough that becomes a new frontal boundary was initiated by diabatic processes. Once a horizontal temperature gradient was created by differential heating, winds adjusted to the new pressure pattern and induced confluence and tilting frontogenesis. This is in contrast to other studies of the initiation of fronts, in which dynamic terms act to intensify preexisting baroclinic temperature gradients. In this case, diabatic processes create the initial gradient. Sensitivity experiments reveal that the front does not propagate discretely without the effects of precipitation and clouds. Differential diabatic processes are identified as strong contributors to frontogenesis here.

Further study into discrete frontal propagation should be carried out. Several dynamic considerations have been noted in this study, but have not been rigorously investigated. For example, the role of the depth and intensity of the prefrontal cold anomaly versus the intensity of the original frontal circulation may be a key
FIG. 19. Cross sections of geopotential height difference (contour interval 5 m) between CTL and DRY2 (i.e., CTL-DRY2) at (a) 1700 and (b) 2300 UTC 5 Sep 1994. The surface locations of the original front and the prefrontal trough in CTL are indicated by the arrows.

FIG. 18. Difference fields between CTL and DRY2 at 1800 UTC 5 Sep 1994: (a) sea level pressure (contour interval 0.25 hPa) and surface wind vectors (kt), (b) surface temperature (contour interval 1°C), and (c) cross section of temperature (contour interval 1°C) where the location of the cross section is indicated in (a). Negative contours are dashed. Frontal symbols denote the location of fronts in CTL.
in differentiating between discretely propagating fronts and continuously propagating fronts that encounter prefrontal thermal anomalies. Idealized variations in the depth, location, and intensity of prefrontal thermal anomalies, similar to the studies of Hoskins et al. (1984), might reveal the sensitivity of discrete frontal propagation to differences in the mesoscale environment. In addition to dynamical studies, a thorough climatological study of discrete frontal propagation in observations could be performed. Also, upstream discrete propagation (i.e., discrete backward movement of cold fronts) as well as discrete propagation of warm fronts should be explored.

Acknowledgments. We would like to thank Jay Charney for his assistance during this work, especially for his help with MM5 programs, data display, and his comments in interpreting the results. Valuable input was also added by Greg Forbes, Rob Rogers, Jack Kain, and Chuck Doswell. Reviews by Roger Smith and an anonymous reviewer strengthened the content and organization of this manuscript. This work was supported by NSF Grant ATM 92-22017. The first author was supported for one year by an American Meteorological Society Graduate Fellowship sponsored by the National Weather Service.

APPENDIX

Description of Cross-Section Averaging Technique

In cross section analyses of model output, averages of several cross sections oriented perpendicular to the fronts are constructed. The averages eliminate high-frequency perturbations (such as those associated with moist parameterized convection at individual grid elements) but still retain the mesoscale structure and vertical circulations associated with the fronts.

Cross sections are spatially normalized before averaging to account for variations in the separation between the original front and the new front. The normalization is achieved in the following manner. First, the surface locations of the front (F) and prefrontal trough (PFT) are subjectively determined. The center of the pressure trough was primarily used to define the locations of the front and PFT (see, e.g., Fig. 2). For normalization, the postfrontal region (the cold air mass behind the original cold front) is defined as the left part of the normalized cross section, the interfront region (the intervening zone between the original cold front and the PFT) is defined as the middle part of the cross section, and the pre-PFT region (the warm sector ahead of the PFT) is defined as the right part of the cross section. This methodology is illustrated in Fig. A1a. For each of the individual cross sections that comprise the average, values of selected variables (such as temperature, vertical velocity, etc.) are interpolated to equally spaced points across the three regions (Fig. A1b). The individual cross section fields are then added together and averaged. Since the averages are computed on sigma surfaces, the boundary layer structure is retained. An average terrain for the cross sections is computed from the mean terrain of the individual cross sections.

REFERENCES


Berson, F. A., D. G. Reid, and A. J. Troup, 1957: The summer cool...


Stensrud, D. J., 1996: Effects of persistent, midlatitude mesoscale
regions of convection on the large-scale environment during the warm season. *J. Atmos. Sci.*, 53, 3503–3527.


