Cold Fronts With and Without Preefrontal Wind Shifts in the Central United States

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ABSTRACT

Time series of cold fronts from stations in the central United States possess incredible variety. For example, time series of some cold fronts exhibit a sharp temperature decrease coincident with a pressure trough and a distinct wind shift. Other time series exhibit a prefrontal trough and wind shift that precedes the temperature decrease associated with the front by several hours. In early March 2003, two cold fronts passed through Oklahoma City, Oklahoma (OKC), representing each of the above scenarios. The cold front on 4 March was characterized by a coincident sharp wind shift, pressure trough, and a strong temperature decrease of 10°C in 2 minutes. On the other hand, the cold-frontal passage on 8 March was characterized by a prefrontal wind shift occurring over a 7-h period before the temperature decrease of 10°C in 2 h. Twelve hours before frontal passage at OKC, both fronts had the same magnitude of the horizontal potential temperature gradient and Petterssen frontogenesis. By the time of frontal passage at OKC, the magnitude of the horizontal potential temperature gradient for the 4 March front was double that of the 8 March front and the frontogenesis was nearly four times as great. The simultaneity of the surface horizontal potential temperature gradient, and deformation and convergence maxima (coincident with the wind shift) was primarily responsible for the greater strength of the cold front in OKC on 4 March compared to that on 8 March. Whether a prefrontal wind shift occurred was determined by the timing and location of cyclogenesis in the central United States. On 4 March, a cyclone was adjacent to the slope of the Rocky Mountains and developed on the cold front as it moved through Oklahoma, permitting greater frontogenesis and resulting in a cold-frontal passage at OKC with a simultaneous temperature decrease and wind shift. On 8 March, the cyclone moved eastward through Oklahoma before the arrival of the cold front, resulting in a prefrontal wind shift associated with the northerlies behind the cyclone, followed by the frontal passage. A two-year climatology of cold-frontal passages at
OKC supports the two cases above, indicating that the timing and location of cyclogenesis was responsible for these two different cold-frontal structures. These results imply that, for situations resembling those of this study, the prefrontal trough is not directly associated with the cold front, but is caused by external processes related to the lee troughing.
1 Introduction

The conceptual model of a classical cold front consists of a sharp temperature decrease coincident with a pressure trough and a distinct wind shift. Many cold fronts, however, do not conform to this model—time series at a single surface station may possess a pressure trough and wind shift in the warm air preceding the cold front (hereafter called a prefrontal trough and prefrontal wind shift, respectively). Schultz (2004a) reviewed mechanisms by which cold fronts may exhibit prefrontal troughs and wind shifts. One of those mechanisms was related to lee troughs, such as occur downstream of the Rocky Mountains. During the cool season in the central United States, equatorward-moving fronts occur frequently, and Hutchinson and Bluestein (1998) found that as many as 60% of these cold fronts were associated with prefrontal wind shifts (Fig. 1).

Hutchinson and Bluestein (1998) schematically illustrated the development of such a cold front with a prefrontal wind shift in Fig. 2. Initially, westerly flow across the Rocky Mountains leads to the formation of a lee trough (Fig. 2a). An equatorward-moving cold front and prefrontal warm advection causes the movement or departure of the lee trough away from the mountains (Fig. 2b). This warm advection implies that troughing occurs ahead of the surface cold front. Eventually, the faster-moving cold front catches up to the lee trough, becoming one feature at the surface (Fig. 2c). In contrast, Palmén and Newton (1969), Steenburgh and Mass (1994), and Schultz and Doswell (2000) provide an alternate explanation for departure of the cyclone from the lee slopes. They showed that upper-level vorticity advection from a mobile shortwave trough was responsible for the lee trough departing the lee slopes. Regardless, that prefrontal wind shifts are associated with lee cyclones helps to explain the maximum in prefrontal troughs located along the front range of the Rocky Mountains (Fig. 1), where lee cyclogenesis is common (e.g., Petterssen 1956, 267–268; Zishka and Smith 1980, their Fig. 2). Thus, the prefrontal trough is not
directly associated with the cold front, but is caused by external processes related to the lee troughing.

If Fig. 2 represents the schematic pattern for cold fronts with lee troughs acting as prefrontal troughs, then what is the pattern for cold fronts without prefrontal troughs? Hutchinson and Bluestein (1998) did not address this question. Therefore, the purpose of this paper is to contrast the conditions that favor cold fronts in the central United States that exhibit prefrontal wind shifts from those that do not.

This question is more than just academic, as weather forecasts may depend on understanding these processes. First, as Sanders (1967, 1999a) has argued, the relationship between the wind shift and temperature gradient determines the future strength of the cold front. Cold fronts in which the wind shifts are coincident with the temperature gradient imply frontogenesis or a strengthening temperature gradient over time. In contrast, cold fronts in which the wind shifts are not coincident with the temperature gradient imply frontolysis or a weakening temperature gradient over time. Second, in some cases, the prefrontal wind shift or trough may develop a temperature gradient and thus become frontogenetical. Third, convective initiation and severe weather may develop along the surface convergence associated with the wind shift, cold front, or both (e.g., House 1959; Sanders and Doswell 1995). For example, Ryan and Wilson (1985) estimate that 50% of all thunderstorms in southeast Australia are prefrontal. Finally, terminal area forecasts (TAFs) for aviation and wildfire forecasts are crucially dependent on accurately forecasting the timing of the wind shift associated with an advancing cold front. Understanding the reasons for such wind shifts may help improve forecast performance for such situations.

Section 2 of this paper contrasts two cold fronts that passed through Oklahoma City, Oklahoma (OKC) only four days apart in early March 2003. The front on 4 March exhibited a classic cold-frontal passage at OKC with a simultaneous temperature drop, pressure trough,
and wind shift. For the front on 8 March, however, neither the wind shift nor pressure minima were coincident with the largest temperature decrease, and the wind shift took 7 h to complete. Although both fronts possessed a comparable magnitude of the horizontal surface potential temperature gradient 12 h before frontal passage at OKC, by the time of frontal passage at OKC, the 4 March front had become one to two orders of magnitude stronger than the 8 March front. Both fronts passed OKC at roughly the same time of day (about 2300 UTC or 1700 LST) with prefrontal clear-to-partly-cloudy skies and postfrontal overcast skies, eliminating significant cross-frontal differences in diurnal surface heating between the two fronts as a possible explanation. Neither front produced any precipitation at OKC, suggesting moist processes were not important. To explain the differences in strength between these two fronts, Petterssen (1936) frontogenesis diagnostics are employed in section 2. In section 3, a two-year climatology of cold fronts at OKC compiled by Hutchinson and Bluestein (1998) is revisited. Composite analyses of these events help distinguish between synoptic conditions favoring cold fronts with and without prefrontal wind shifts. Section 4 concludes this paper.

2 Case studies of cold fronts with and without prefrontal wind shifts

A classic cold-frontal passage occurred at 2306 UTC 4 March 2003 at OKC (Fig. 3a). The temperature dropped 6.2°C in the first minute, 10.0°C in the first two minutes, and almost 28°C in 10 hours. Given that the average speed of the front in the 12 hours from 1200 UTC 4 March to 0000 UTC 5 March was estimated to be 11.6 m s⁻¹, the time-to-space-converted temperature gradient for the first minute was estimated to have been 8.9°C km⁻¹. This is an order of magnitude greater than the front observed by Schultz and Trapp (2003) in northern Utah, which had an estimated horizontal temperature gradient of 0.6°C km⁻¹. [The value
published in Schultz and Trapp (2003, p. 2227) is incorrect.] In four minutes, the wind shifted nearly $100^\circ$ from southwest to northwest (Fig. 3a). The pressure fell steadily with the approach of the front, reached a minimum when the front passed, then rose rapidly with the arrival of the postfrontal cold air (Fig. 3a).

In contrast, the situation on 8 March exhibited a very different time series (Fig. 3b): neither the wind shift (about 1200–1900 UTC) nor pressure minima (about 0900 and 2000 UTC) were coincident with the most rapid temperature decrease (2300 UTC). The maximum temperature was reached at 2000 UTC, coincident with relative minima in pressure and dewpoint temperature. Following this maximum in temperature, the temperature decreased about 5°C before the arrival of the sharp temperature decrease at 2300 UTC. Although the total temperature drop (28.8°C) was comparable to that on 4 March, the initial rate of decrease with the front occurred much more slowly—2 h to fall about 10°C. Given the 12-h average speed of 13.9 m s$^{-1}$, the temperature gradient at OKC was estimated to have been 0.6°C km$^{-1}$, an order of magnitude smaller than on 4 March. The pressure trough was not as sharp as on 4 March and the wind shift from southeasterly to northerly took 7 h to complete (Fig. 3b). At 1700 UTC, within the period the wind shifted, a southwesterly to northwesterly prefrontal wind shift signaled the brief arrival of a warmer, drier air mass at OKC. [Figure 4d shows the localized extent of the warmer air at the surface over south-central Oklahoma.] Thus, the time series on 8 March indicated a prefrontal wind shift and pressure trough ahead of the largest temperature gradient, representing the front.

2a Synoptic patterns

What was the cause of these differences between the two frontal passages at OKC? To answer this question, surface maps roughly 12 h before and at the time of frontal passage at OKC for the two cases are examined (Fig. 4). The analysis is the National Centers for
Environmental Prediction’s Rapid Update Cycle (RUC; Benjamin et al. 2004) initial analysis with 20-km horizontal grid spacing, archived at a reduced 40-km grid spacing. Although the RUC analysis is relatively coarse, it reproduced the general structure and evolution of these two fronts adequately for illustrating general synoptic features and computing diagnostics on scales larger than, and including, the mesoscale. For the 4 March front, cyclogenesis was occurring in the lee of the Rocky Mountains at the same time the front was approaching from the north (Fig. 4a). By the time the front passed OKC, the cyclone was located along the warm side of the front, albeit in two separate centers 4 and 7 hPa weaker than 12 h earlier (Fig. 4b). For the 8 March front, the cyclone had already departed from the lee slopes for the Great Lakes region well before the cold front arrived in OKC (Figs. 4c,d). Surface winds slowly veered over 7 h from southeasterly to northerly (Fig. 3b) as the surface cyclone departed the lee slopes. Thus, the timing and location of the cyclone for each event appeared to play a significant role in the character of the frontal passages at OKC.

On 4 March, having the cyclone coincident with the arrival of the cold front in Oklahoma meant that the warmest air equatorward of the cyclone could be juxtaposed with the cold air behind the front, maximizing the temperature drop across the front. Also, the deformation field associated with the surface cyclone (e.g., as shown by the idealized cyclones of Figs. 13f and 14f in Schultz et al. 1998) helped keep the temperature gradient and wind shift coincident (Fig. 3a), as in a classical front. In contrast, on 8 March, with the cyclone departed from Oklahoma, the winds had already begun to veer, coming from the north, even before the cold front had arrived. The prefrontal northerlies resulted in weaker convergence and frontogenetical forcing along the front. In addition, with the cyclone well to the east, deformation associated with the cyclone that could have contributed toward frontogenesis was absent. More on deformation, convergence, and frontogenesis is discussed shortly.

The 500-hPa pattern for 1200 UTC 4 March featured confluent flow over the central
United States (Fig. 5a). By the time of frontal passage at OKC, absolute-vorticity maxima within the southern stream were moving over Oklahoma and Texas (Fig. 5b), likely associated with the surface cyclogenesis (Fig. 4b). At 1200 UTC 8 March, however, the southern stream was weaker and absolute vorticity maxima over Nebraska and Kansas (Fig. 5c) were likely associated with the surface cyclogenesis that was departing the lee slopes (Fig. 4c), the same mechanism discussed by Palmén and Newton (1969), Steenburgh and Mass (1994), and Schultz and Doswell (2000) for other examples of lee cyclogenesis. By the time of frontal passage at OKC around 0000 UTC 9 March, the shortwave trough in the northern stream had moved into the Great Lakes area (Fig. 5d), along with the surface cyclone (Fig. 4d).

2b Frontogenesis

Although the magnitude of the surface horizontal potential temperature gradients in the RUC analyses for the two fronts were comparable at 11–13 K (100 km)$^{-1}$ 12 h before frontal passage at OKC, by the time the fronts passed OKC, the maximum magnitude of the potential temperature gradient had more than doubled to over 28 K (100 km)$^{-1}$ for the 4 March front, but remained nearly unchanged for the 8 March front (not shown). To explain the difference in the strength of the fronts, the expression for Petterssen (1936) frontogenesis $F$ is used.

$$F = \frac{d}{dt} |\nabla_H \theta|,$$  \hspace{1cm} (1)

where

$$\frac{d}{dt} = \frac{\partial}{\partial t} + u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y},$$

$$\mathbf{V}_H = u \mathbf{i} + v \mathbf{j},$$

$$\nabla_H = i \frac{\partial}{\partial x} + j \frac{\partial}{\partial y}.$$  

Petterssen (1936) showed that (1) could be written

$$F = \frac{1}{2} |\nabla_H \theta| (E \cos 2\beta - \nabla_H \cdot \mathbf{V}_H),$$  \hspace{1cm} (2)
where \( E \) is the resultant deformation and \( \beta \) is the local angle between an isentrope and the axis of dilatation measured in a counterclockwise direction. In this paper, variations in surface elevation are neglected so that (2) can be applied at the Earth's surface.

Twelve hours before frontal passage at OKC, the values of surface frontogenesis for the 4 and 8 March fronts were comparable at 9 and 10 K (3 h\(^{-1}\) (100 km\(^{-1}\)), respectively (not shown). By the time these two fronts passed OKC twelve hours later, the values of surface frontogenesis increased to 60 and 16 K (3 h\(^{-1}\) (100 km\(^{-1}\)), respectively, with the value for the 4 March front nearly four times as large as that on 8 March (not shown). Such large changes in the magnitude of the surface horizontal potential temperature gradient corresponding to these incredible rates of frontogenesis were not observed, even within the 40-km horizontal grid spacing of the RUC analysis. It is generally accepted that, following the front, this large rate of surface frontogenesis is offset by mixing within the frontal zone since the front is not a material surface (e.g., Sanders 1955, 1999b), as discussed in Schultz (2004b).

Owing to the strong dependence of \( F \) on the magnitude of the horizontal potential temperature gradient in (2), an expression for the kinematic effect of bringing isentropes together, independent of the thermal gradient, is desired. Frontogenesis normalized by the magnitude of the horizontal potential temperature gradient, \( \tilde{F} \), is defined as:

\[
\tilde{F} = \frac{F}{|\nabla_H \theta|},
\]

\[
\tilde{F} = \frac{d}{dt} \ln |\nabla_H \theta|,
\]

which can be expressed as the sum of the deformation and the convergence terms:

\[
\tilde{F} = \frac{1}{2} (E \cos 2\beta - \nabla_H \cdot \mathbf{V}_H).
\]

Normalized frontogenesis for the two fronts (Fig. 6) shows that the differences in the wind field across the front had dramatic consequences for the strength of the fronts. Normalized
frontogenesis was relatively small 12 h before frontal passage at OKC for both fronts (cf. Figs. 6a,c), but, by the time of frontal passage at OKC, $\mathcal{F}$ was at least three times greater along the 4 March front compared to the 8 March front (cf. Figs. 6b,d). Both the deformation and convergence terms in (5) were contributing equally to the normalized frontogenesis at the time of the 4 March and 8 March frontal passages at OKC, although the two terms were each weaker for the 8 March front than their counterparts for the 4 March front (not shown). The explanation for this difference in the magnitude of normalized frontogenesis is the location of the surface cyclone. On 4 March, the deformation and convergence associated with the front were aided by the deformation and convergence associated with the cyclones moving along the front (Fig. 4b). Sustained frontogenesis on the synoptic scale concentrated the temperature gradient, leading to strong frontogenesis on the mesoscale, in the manner discussed by Hoskins and Bretherton (1972). On 8 March, in the absence of the surface cyclogenesis along the front, less synoptic-scale deformation and convergence were present, so the scale of the front never contracted and its intensity did not increase substantially. There was also a difference in the spatial relationships between the normalized frontogenesis and the magnitude of the surface potential temperature gradient. At 0000 UTC 5 March, the largest normalized frontogenesis was coincident with the largest magnitude of the surface potential temperature gradient (Fig. 6b), indicating the potential for strong frontogenesis. In contrast, at 0000 UTC 9 March, the largest normalized frontogenesis was occurring at the leading edge of the potential temperature gradient over Oklahoma (Fig. 6d), indicating a less-than-ideal configuration for strong frontogenesis. In Indiana and Illinois, closer to the deformation associated with the cyclone, the normalized frontogenesis was larger than that over Oklahoma (Fig. 6d) due to the scale contraction effect discussed above, illustrating the importance of the cyclone towards providing a favorable frontogenetical environment. This case also indicates that the occurrence of prefrontal wind shifts may not be consistent along
the length of the front.

This explanation for the comparative strength of the two fronts was previously discussed by Sanders (1967, 1983, 1999a). He hypothesized that, when the temperature gradient and the wind shift (i.e., representing a region of maximum convergence, deformation, and vorticity) coincide, frontogenesis would be large (Fig. 7a). The collocation of the normalized frontogenesis with the temperature gradient maintained the front against turbulence (e.g., Sanders 1999a), but when the wind shift separated from the temperature gradient, frontogenesis decreased or was even frontolytical (Fig. 7b).

Furthermore, Sanders (1955, 1967) showed that the collocation of the surface vorticity maximum associated with the front and the vertical motion maximum above the surface convergence maximum strengthened the surface vorticity at the front due to vortex stretching. In contrast, Orlanski and Ross (1984) and Sanders (1999a,b) showed that the nonsimultaneity of these quantities weakened the front over time. These results were previously discussed by Petterssen (1936, 20-21) in the context of idealized models of fronts in different flow fields. Given the importance of the relationship between surface convergence, deformation, and horizontal potential temperature gradient, plots of the relationship between these kinematic quantities are constructed to evaluate the above hypotheses.

In the 4 March front, the convergence maxima were along the leading edge of the front at 1200 UTC 4 March (Fig. 8a), but by 0000 UTC 5 March, with the surface cyclone traveling along the front, much of the magnitude of the surface horizontal potential temperature gradient was within the region of convergence (Fig. 8b). In the 8 March front, weak convergence was along the leading edge of the front at 1200 UTC 8 March (Fig. 8c). By 0000 UTC 9 March, the vorticity and convergence associated with the lee trough preceded the thermal gradient by 40–160 km, with an even greater separation in southwestern Oklahoma (Fig. 8d). The convergence and vorticity were at the leading edge of the cold front, if not farther into
the warm air. Divergence was occurring within the frontal zone, a frontolytical effect by (2). These diagnostics demonstrate why the simultaneity of the temperature gradient and wind shift in the 4 March front, associated with the cyclogenesis along the front, resulted in a substantially stronger front than the 8 March front (cf. Figs. 4b,d), in agreement with the Sanders (1999a) schematic (Fig. 7).

2c Discussion

Curiously, even when the fronts were at their strongest, the axes of dilatation were not parallel to the isentropes within the front (Figs. 8b,d), as would be expected for maximum frontogenesis. Bishop (1996) explained that, since the vorticity along the front would be associated with cyclonic rotation, a nonrotating front reaches a balance between the vorticity along the front acting to rotate the isentropes cyclonically and the deformation along the front acting to rotate the isentropes anticyclonically. This effect is apparent in these two cases (Figs. 8b,d).

What happened to these fronts after they passed through OKC? Despite the strength of the 4 March front at OKC, it stalled in southern Texas within 18 hours, never reaching the Mexican border (not shown). By the time the front passed through Austin, Texas, however, a prefrontal trough preceded the temperature drop by about 3 h. In contrast, earlier in South Dakota (not shown), time series at surface stations of the 8 March front exhibited the largest temperature drop and the wind shift simultaneous. With the development of the lee trough ahead of the front as the front moved equatorward, however, these time series exhibited nonsimultaneity (Fig. 3b). Thus, either external or internal factors to cold fronts may affect whether the temperature gradient remains coincident with the most significant wind shift/pressure trough, supporting Sanders’ (1999b) claim that many strong fronts are short-lived phenomena.
Before leaving this discussion of case studies, it is worthwhile to reexamine previous examples of fronts described in the literature. A surprising result, perhaps, is that even some famous fronts may be similar to the situation on 8 March. For example, the front studied by Sanders (1955) occurred in such a situation. A time series from Fort Sill, Oklahoma (FSI) showed a wind shift from southwesterly to northerly around 2100 UTC 17 April 1953 (Fig. 9). Surface maps (not shown) confirm that this wind shift is associated with a lee trough. Three hours later, the frontal passage was characterized by the temperature dropping 11.1°C in 1 h, the wind speed increasing to 10–22.5 m s$^{-1}$ followed by 5 h of blowing dust, and a slight veering of the wind direction to the north-northeast. Other previously published, strong cold fronts in the central United States also show a prefrontal trough and wind shift ahead of these cold-frontal passages (e.g., Brandes and Rabin 1991; Mecikalski and Tilley 1992; Colle and Mass 1995).

Other explanations have been offered in the past for these types of prefrontal features, as reviewed by Schultz (2004a). For example, some of these prefrontal features may be bores. If this were true, then warming would be expected to accompany the prefrontal feature. Since modest cooling is observed in the 8 March and many other cases, it is unlikely that prefrontal bores are responsible in these situations. Another potential explanation for prefrontal features are solitary gravity waves. For solitary gravity waves, however, the winds at a particular point would change from southerly (prefrontal), to northerly (gravity wave passage), followed by southerly (post gravity wave), and finally northerly (postfrontal). Such a trend is not observed either. The timing of the fronts and prefrontal wind shifts do not show a relationship with time of day, either, suggesting that stability-based explanations (e.g., Reeder et al. 1991; Sanders and Kessler 1999) are not a likely explanation either.
3 Climatology of cold fronts with and without prefrontal wind shifts

In the previous section, case studies of cold fronts with and without prefrontal wind shifts in the central United States were compared, concluding that the timing and location of the departure of a cyclone from the lee of the Rockies was responsible for the distinction between these two situations. To test this hypothesis with a larger dataset, a climatology of cold fronts with and without prefrontal wind shifts is desired. The dates and times of the frontal passages and prefrontal wind shifts (if any occurred) from the two-year climatology in Hutchinson and Bluestein (1998) were obtained (T. Hutchinson 2003, personal communication). Time \( t = 0 \) was defined as the 6-h time closest to, but before, frontal passage at OKC (for cases with no prefrontal wind shifts) or wind shift at OKC (for cases with prefrontal wind shifts). These times, as well as \( t = -24 \) h, \( t = -12 \) h, and \( t = +12 \) h were composited at the NOAA–CIRES Climate Diagnostics Center web site (http://www.cdc.noaa.gov) using the NCEP/NCAR Reanalysis dataset (Kalnay et al. 1996).\(^1\) Fifty fronts without prefrontal wind shifts (NOPRE) and 20 fronts with prefrontal wind shifts (PRE) were composited. An average time of 5 h (ranging from 2 to 13 h) occurred between prefrontal wind shift and frontal passage in PRE.

The time-lagged mean and anomaly 500-hPa geopotential height fields for NOPRE and PRE are shown in Fig. 10 and Fig. 11, respectively. NOPRE is characterized by a sharp trough over Utah at \(-24\) h that moved slowly eastward to Kansas, Colorado at 0 h, and Nebraska by \(+12\) h (Figs. 10 and 11). A ridge over the central and eastern United States preceded the trough. Large-scale geostrophic confluence was indicated by the two branches

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\(^1\)Because anomalies can only be calculated from the daily average fields at the web site, the 500-hPa anomalies shown in Fig. 11 are calculated from the mean of the daily averages over 1968–1996. As such, the anomaly fields in Fig. 11 will be smoother than anomalies would be if they were calculated from the 6-h data in Fig. 10.
of the jet stream at 0 h: one branch of northwesterly flow over Montana, Wyoming, and the Dakotas and the other of southwesterly flow over Texas, Oklahoma, Kansas, and Missouri (Fig. 10). The NOPRE composite in Fig. 10 at −12 and 0 h represents the corresponding times for the 4 March 2003 cold front in Figs. 5a,b, respectively. In contrast, PRE was characterized by a more mobile 500-hPa pattern farther north, with a sharp trough developing in northwesterly flow over western Canada at −24 h, which moved through the northern United States into the western Great Lakes area by +12 h (Figs. 10 and 11). The PRE composite in Fig. 10 at −12 and 0 h represents the corresponding times for the 8 March 2003 cold front in Figs. 5c,d, respectively.

These differences at 500 hPa were reflected in the sea level pressure field (Fig. 12). In NOPRE, the westerly flow and 500-hPa shortwave trough over Utah produced a cyclone at the Colorado–Kansas border at −24 h (Fig. 12). As the shortwave trough moved eastward, the surface cyclone drifted slowly southeastward by 0 h. By +12 h, the surface cyclone weakened, developing a broad trough from Wisconsin to Texas, with a surface low in eastern Mexico. This evolution was similar to that of the 4 March cold front (Figs. 4a,b). In PRE, a surface cyclone in eastern Montana at −24 h was attributed to the upstream 500-hPa shortwave trough in northwesterly flow (cf. Figs. 10 and 12). This surface cyclone moved quickly southeastward, elongated, and developed a closed low over Michigan at +12 h (Fig. 12), similar to the 8 March cold front (Figs. 4c,d). In both cases, lee troughs formed, but where the trough/cyclone was deepest depended on the location of the forcing for height falls by the shortwave troughs in the westerlies. [That the differences between the sea level pressure of NOPRE and PRE at 0 h (Fig. 12) were quite small was mostly due to the coarse resolution of the NCEP/NCAR Reanalysis (2.5° latitude × 2.5° longitude, approximately 280 km × 280 km).] Thus, the lee cyclone is important by providing a favorable frontogenetical environment and producing strong northerlies after its departure, which advance the front
These composites help to explain the differences in the large-scale flow that determine whether a surface frontal passage at OKC is associated with a prefrontal wind shift or not. The differing paths of the 500-hPa shortwave troughs (Figs. 10 and 11) suggests that the location of the forcing for surface cyclogenesis is significant (Fig. 12). Whereas the 500-hPa forcing in NOPRE is slower moving and closer to Oklahoma, helping to focus the surface deformation near the cold front, in PRE, the rapidly moving forcing, and the resulting surface cyclone, is farther poleward. Large-scale confluence in the 500-hPa pattern over the central United States in NOPRE suggests a large-scale control on keeping the front simultaneous compared to PRE.

4 Conclusion

This paper explored the similarities and differences between cold fronts in the central United States that exhibited prefrontal troughs and those that did not. Analysis of this problem entailed both case studies and a composite analysis based on a two-year climatology of cold-frontal passages in OKC. For the case studies, two cold fronts in the central United States from early March 2003 were examined. The 4 March cold front had a classical frontal structure featuring simultaneous surface horizontal temperature gradient, wind shift, and pressure trough. The 8 March cold front exhibited a wind shift and pressure trough in the warm air ahead of the largest temperature gradient. The reason for this difference was the timing of the departure of the cyclone from the lee slopes of the mountains, associated with a mobile shortwave trough aloft. If the cyclone departed at the same time as, or later than, the arrival of the cold front from the north, then no prefrontal wind shift occurred. If the cyclone departed ahead of the cold front, then a prefrontal wind shift was observed. Twelve hours before frontal passage at OKC, both the 4 March and 8 March fronts had the same
magnitude of the horizontal potential temperature gradient and frontogenesis. By the time of frontal passage at OKC, the magnitude of the horizontal potential temperature gradient for the 4 March front was double that of the 8 March front and the frontogenesis was nearly four times larger, consistent with arguments by Sanders (1967, 1999a) and others.

To examine a larger dataset, a climatology of cold-frontal passages in OKC previously constructed by Hutchinson and Bluestein (1998) was examined. Composite analyses were constructed, supporting and extending the results from the case studies. Specifically, if the cyclone remained along the slope of the Rocky Mountains when the cold front arrived in OKC, then a cold front without a prefrontal wind shift would be observed in OKC (NOPRE). Cyclogenesis along the cold front was forced by a slow-moving 500-hPa shortwave trough. On the other hand, if the forcing for cyclogenesis from the shortwave trough aloft preceded the cold frontal passage, then a prefrontal wind shift and trough would be formed (PRE). Thus, the timing and location of the cyclogenesis relative to the cold front controlled the resulting evolution.

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Figure captions

Figure 1. A two-year climatology of prefrontal wind shifts between 1 May 1993 and 30 April 1995. Station model shows station identifier (upper left), percentage of cold fronts that were preceded by wind shifts (upper right), and number of frontal passages (lower right) during the sample. Percentage of fronts preceded by prefrontal wind shifts (solid lines every 20%) and the number of fronts (dashed lines every 10) are contoured. (From Hutchinson and Bluestein 1998, Fig. 7.)

Figure 2. Idealized evolution of a prefrontal wind shift (dashed line). Initially (a) a wind shift associated with a lee trough forms when low-level flow (arrows) is perpendicular to the mountains. The wind shift moves away from the mountains (b) when the low-level flow veers and warm advection (indicated by WA) is occurring to its east. Eventually, (c) the front catches up to the wind shift. (From Hutchinson and Bluestein 1998, Fig. 24.)

Figure 3. Frontal passages on 4–6 and 8–10 March 2003 from the 1-minute data at the Automated Surface Observing System (ASOS) at Oklahoma City, OK (OKC). UTC=CST+6 h.

Figure 4. Mean sea level pressure (thick solid lines every 2 hPa), 2-m potential temperature (thin solid lines every 2 K), and 10-m winds (one pennant, full barb, and half barb denote 25, 5, and 2.5 m s\(^{-1}\), respectively) from the NCEP Rapid Update Cycle: (a) 1200 UTC 4 March 2003, (b) 0000 UTC 5 March 2003, (c) 1200 UTC 8 March 2003, and (d) 0000 UTC 9 March 2003.

Figure 5. The 500-hPa absolute vorticity of the total horizontal wind (10\(^{-5}\) s\(^{-1}\), shaded according to scale) and geopotential height (solid lines every 6 dam) from the NCEP Rapid
Update Cycle: (a) 1200 UTC 4 March 2003, (b) 0000 UTC 5 March 2003, (c) 1200 UTC 8 March 2003, and (d) 0000 UTC 9 March 2003.

Figure 6. Normalized frontogenesis ($10^{-5}$ s$^{-1}$, shaded according to scale), 2-m potential temperature (thin solid lines every 2 K), and 10-m winds (one pennant, full barb, and half barb denote 25, 5, and 2.5 m s$^{-1}$, respectively) from the NCEP Rapid Update Cycle: (a) 1200 UTC 4 March 2003, (b) 0000 UTC 5 March 2003, (c) 1200 UTC 8 March 2003, and (d) 0000 UTC 9 March 2003.

Figure 7. Evolution of a front: (a) newly formed steering line, and (b) wind shift moving away from weakening temperature gradient. (From Sanders 1999a, Fig. 3).

Figure 8. Relative vorticity of the total horizontal wind at 10 m AGL ($10^{-5}$ s$^{-1}$, shaded according to scale), convergence of 10-m winds ($10^{-5}$ s$^{-1}$, solid black lines for positive and zero values, dashed black lines for negative values), 2-m potential temperature (solid green lines every 2 K), and axes of dilatation of total horizontal wind [$10^{-5}$ s$^{-1}$, blue line segments scaled according to legend; separation between displayed axes of dilatation is 40 km (every grid point)] from the NCEP Rapid Update Cycle: (a) 1200 UTC 4 March 2003, (b) 0000 UTC 5 March 2003, (c) 1200 UTC 8 March 2003, and (d) 0000 UTC 9 March 2003.

Figure 9. Time series of the frontal passage at Fort Sill, Oklahoma, from the case studied by Sanders (1955).

Figure 10. Composite 500-hPa geopotential height (solid lines every 60 m) for fronts without prefrontal wind shift (NOPRE; left column) and fronts with prefrontal wind shift (PRE;
right column) at Oklahoma City, Oklahoma (OKC). Zero time is defined as the 6-h time closest to, but before, frontal passage (for cases with no prefrontal wind shifts) or wind shift (for cases with prefrontal wind shifts) at OKC. Top to bottom: 24 h and 12 h before front/wind shift passage at OKC, at front/wind shift passage, and 12 h after front/wind shift passage. Composite maps were provided by NOAA–CIRES Climate Diagnostics Center, Boulder Colorado from their web site (http://www.cdc.noaa.gov).

Figure 11. Composite 500-hPa geopotential height anomalies (solid lines every 20 m) for fronts without prefrontal wind shift (NOPRE; left column) and fronts with prefrontal wind shift (PRE; right column) at Oklahoma City, Oklahoma (OKC). Anomalies are defined from the mean of daily average fields over 1968–1996. Zero time is defined as the day of frontal passage (for cases with no prefrontal wind shifts) or wind shift (for cases with prefrontal wind shifts) at OKC. Top: day before front/wind shift passage at OKC; bottom: day of front/wind shift passage at OKC. Composite maps were provided by NOAA–CIRES Climate Diagnostics Center, Boulder Colorado from their web site (http://www.cdc.noaa.gov).

Figure 12. Same as Fig. 10, except for composite sea level pressure (solid lines every 1 hPa).
Figure 1: A two-year climatology of prefrontal wind shifts between 1 May 1993 and 30 April 1995. Station model shows station identifier (upper left), percentage of cold fronts that were preceded by wind shifts (upper right), and number of frontal passages (lower right) during the sample. Percentage of fronts preceded by prefrontal wind shifts (solid lines every 20%) and the number of fronts (dashed lines every 10) are contoured. (From Hutchinson and Bluestein 1998, Fig. 7.)
Figure 2: Idealized evolution of a prefrontal wind shift (dashed line). Initially (a) a wind shift associated with a lee trough forms when low-level flow (arrows) is perpendicular to the mountains. The wind shift moves away from the mountains (b) when the low-level flow veers and warm advection (indicated by WA) is occurring to its east. Eventually, (c) the front catches up to the wind shift. (From Hutchinson and Bluestein 1998, Fig. 24.)
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