A Review of Cold Fronts with Prefrontal Troughs and Wind Shifts

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ABSTRACT

The conceptual model of a classical surface-based cold front consists of a sharp temperature decrease coincident with a pressure trough and a distinct wind shift at the surface. 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). Although many authors have recognized these prefrontal features previously, a review of the responsible mechanisms has not been performed to date. This paper presents such a review. Ten disparate mechanisms with different frontal structures have been identified from the previous literature. These mechanisms include those external to the front (i.e., those not directly associated with the cold front itself): synoptic-scale forcing, interaction with lee troughs/drylines, interaction with fronts in the mid- and upper troposphere, and frontogenesis associated with inhomogeneities in the prefrontal air. Mechanisms internal to the front (i.e., those directly associated with the structure and dynamics of the front) include the following: surface friction, frontogenesis acting on alongfront temperature gradients, moist processes, descent of air, ascent of air at the front, and generation of prefrontal bores/gravity waves. Given the gaps in our knowledge of the structure, evolution, and dynamics of surface cold fronts, this paper closes with an admonition for improving the links between theory, observations, and modeling to advance understanding and develop better conceptual models of cold fronts, with the goal of improving both scientific understanding and operational forecasting.

1. Introduction

Conceptual models of surface-based cold fronts presented in textbooks (e.g., Wallace and Hobbs 1977, sections 3.2 and 3.3; Carlson 1991, p. 343; Bluestein 1993, p. 246) generally feature a discontinuity (or near discontinuity) in temperature, a simultaneous wind shift, and coincident pressure trough with cold-frontal passage. These characteristics of classical cold fronts are predicted from zero- and first-order discontinuity theory (e.g., Petterssen 1933; Godson 1951; Saucier 1955, p. 109; Bluestein 1993, 240–248). Not all cold fronts, however, possess the classic cold-frontal passage featuring the simultaneity of the temperature decrease and wind shift. In other instances, separate structures may occur in close proximity together (e.g., a dryline in advance of a surface cold front), complicating interpretation.

In fact, these characteristics were noted in some of the early Norwegian literature on cold fronts. For example, Bjerknes (1919) analyzed a *forerunner*, or a prefrontal wind shift, in an early schematic of the Norwegian cyclone model, although this feature was dropped in later versions because the feature lacked generality. [See the discussion in Friedman (1989, 128–134) for more information on the short-lived forerunner.] A different type of cold-frontal structure, the *double cold front*, was discussed by Bjerknes (1926, 1930):

> The case which was demonstrated in my lecture showed a cold front starting as a well-defined line of discontinuity, and within the range of the map changing into a double cold front. The foremost of the two cold fronts showed only a very small drop of temperature and veer of wind, whereas the second, which followed not more than 30 miles behind, was well defined both in respect to temperature and wind. Nev-

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> ertheless, the rain accompanied the first of the cold fronts, and the second did not give any precipitation whatever. (Bjerknes 1926, p. 32)

Like the forerunner, support for the double cold front as part of the Norwegian cyclone model quickly waned. Although the forerunner and double cold front concepts did not find a more permanent place in the Norwegian cyclone model, the Norwegians had observed frontal structures that did not fit their conceptual model. Such nonclassical structures were not limited to the Norwegians, however. As shown in this article, published results ranging from the 1920s to the present using observational analysis, theoretical approaches, and numerical simulations reveal many fronts possessing a variety of nonclassical structures that require explanation.

Understanding the processes that control the structure and evolution of fronts is essential for the accuracy of weather forecasts for several reasons. First, Sanders (1967, 1983, 1999a) has argued that 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 as dissipative processes (e.g., mixing) dominate. Second, in some cases, the prefrontal wind shift or trough may develop a temperature gradient and thus become frontogenetical, eventually leading to clouds and precipitation. Third, given the right environmental conditions, 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) suggested that prefrontal troughs were responsible for convective initiation in southeast Australia because 50% of thunderstorms 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 in these situations has the potential to improve forecast performance.

I restrict attention in this paper to prefrontal features associated with surface-based cold fronts for five reasons. First, the conceptual model of surface cold fronts emphasizes the simultaneous occurrence of the temperature drop, wind shift, and pressure minimum. If structures differ from this conceptual model, then learning more about the structure and evolution of cold fronts would improve our understanding. Second, the surface signatures associated with cold-frontal passages tend to be sharper and better defined than for other types of surface fronts. Third, cold fronts are more commonly discussed in the literature compared to other types of surface fronts (e.g., warm, occluded), as noted by Keyser (1986). Thus, there is more literature to draw upon for a review article on these nonclassical frontal structures. Fourth, to the knowledge of this author, little, if anything, has been written about nonsimultaneity in upper-level fronts. Finally, substantial evidence exists in the literature for the occurrence of prefrontal features associated with cold fronts, and more than one hypothesis has been suggested for their origin. Despite such evidence, a comprehensive review of these mechanisms, which would allow for categorization of these events, as well as a more rigorous testing of these proposed hypotheses, has not been compiled.

The purpose of this paper is to compile a review of these mechanisms and highlight possible research directions to resolve many of these issues with such prefrontal features. Section 2 of this paper presents a suite of characteristics of prefrontal features observed in the literature. This section demonstrates the variety that these features have and provides some insight into the difficulty that researchers and forecasters have had in trying to understand these features. Section 3 assembles various mechanisms that have been proposed for the occurrence of prefrontal troughs and wind shifts. These mechanisms group into 10 categories, illustrating the diverse processes that produce these structures. Although these 10 mechanisms constitute one classification scheme, other self-consistent taxonomies are possible, as noted by one of the reviewers of this manuscript. Section 4 synthesizes the results from the various mechanisms and proposes connections between theory, observations, and diagnosis that are needed to more fully understand these phenomena. Through better understanding, improved conceptual models of cold fronts can be developed. Section 5 summarizes the results of this paper.

2. Characteristics of prefrontal features

To illustrate some of the variety of prefrontal troughs, their characteristics are discussed briefly below.

- *Horizontal scale.* Some of these proposed mechanisms for the occurrence of prefrontal troughs and wind shifts (discussed more fully in section 3) operate on the synoptic scale (through quasigeostrophic principles) and some operate on the mesoscale. Even smaller-scale fronts not associated with extratropical cyclones are also known to possess troughs in the warm air, such as thunderstorm gust fronts (e.g., Charba 1974; Goff 1976; Mahoney 1988) and sea breezes (e.g., Geisler and Bretherton 1969; Alpert and Rabinovich-Hadar 2003).

- *Relative intensities.* Prefrontal wind shifts can be abrupt, or they can occur over several hours. Prefrontal pressure troughs can be sharp or subtle. After their formation, prefrontal features sometimes develop frontal characteristics. Sometimes they become the dominant front, leading to frontolysis of the original front (e.g., Hanstrum et al. 1990b; Charney and Fritsch 1999; Bryan and Fritsch 2000a,b). Other times, they maintain their intensity, yielding two or more cold fronts (e.g., Browning et al. 1997). A third possibility is that the prefrontal feature remains a trough or a wind shift and never develops frontal properties (e.g., Hutchinson and Bluestein 1998).
- *Fronts and airstream boundaries.* A front represents the boundary between two air masses, featuring a wind shift and a temperature gradient. Fronts are a subset of airstream boundaries. An airstream boundary represents the boundary between two air streams, also known as conveyor belts (e.g., Carlson 1980) or coherent ensembles of trajectories (e.g., Wernli and Davies 1997; Wernli 1997). A prefrontal trough is one example of an airstream boundary, where a wind shift is present, but the temperature gradient, if present, is not considered strong enough to qualify as frontal. Sanders (1999a) refers to such nonfrontal wind shifts as baroclinic troughs. Whereas diagnostics to evaluate frontogenesis are well known and popular (e.g., Petterssen 1936; Keyser et al. 1988), diagnostics to evaluate airstream boundaries are less widely used (e.g., Petterssen 1940, 252–256; Petterssen 1956, p. 38; Cohen and Kreitzberg 1997; Cohen and Schultz 2005).
- *Relative speeds.* After the formation of a prefrontal feature, sometimes the cold front catches up and merges with it (e.g., Shapiro 1982; Locatelli et al. 1989, 2002a; Bluestein 2005, manuscript submitted to Amer. Meteor. Soc. *Meteor. Monogr.*, hereafter BLU; Bluestein 1993, p. 162; Keshishian et al. 1994; Colle and Mass 1995; Browning et al. 1997; Hutchinson and Bluestein 1998; Neiman et al. 1998; Neiman and Wakimoto 1999; Parsons et al. 2000; Stoelinga et al. 2000; Rose et al. 2002). Other times, the prefrontal feature develops and travels faster than the cold front, thereby increasing their separation (e.g., Sanders 1999b; Locatelli et al. 2002b).
- *Separation between prefrontal feature and front.* The temporal separation between the prefrontal feature and the front can range anywhere from minutes to hours. Given typical frontal speeds (order 10 m s^{-1}), these prefrontal features could lie anywhere from one kilometer to hundreds of kilometers ahead of the temperature gradient, respectively. Sometimes multiple features of different scales appear during the same event, suggesting that different processes are responsible for each feature, or that the same process acted more than once. For example, the cold front analyzed by Seitter and Muench (1985) possessed two prefrontal features. The first feature was associated with the rope cloud, an increase in wind speed, and a veering of the wind direction. The second prefrontal feature arrived 5–10 min later when the temperature fell about 4°C. Two hours later, however, what Seitter and Muench (1985) termed the "original front" passed with a further shift in wind direction of 20° and temperature decrease of several degrees. Another example in which the rope cloud is about 250 km ahead of the largest temperature decrease was observed by Dorian et al. (1988, their Fig. 17).

Regardless of the characteristics of the prefrontal feature or its proposed mechanism for formation, if these features are hydrostatically balanced (which we believe they likely are for the majority of events discussed in the literature—one possible nonhydrostatic mechanism is discussed in section 3h), surface pressure changes or wind shifts must be accompanied by temperature changes within the overlying air column. For example, if a surface pressure trough is to exist, then the average temperature in the atmospheric column over the surface trough must be warmer than surrounding locations. Thus, the proposed mechanisms in this paper must include supporting evidence that a warmer column of air overlies the surface pressure trough. In addition, the sharpness of the surface trough must also match the sharpness of the vertically integrated thermal structure aloft.

3. Mechanisms for prefrontal troughs and wind shifts

Sanders and Doswell (1995, p. 510) said, "It often appears, however, that one or more wind shifts precede the zone of temperature contrast in cold fronts . . . The origins of such lines are not typically well known and they may arise from more than one source." Indeed, numerous diverse explanations have been proposed for such prefrontal troughs and wind shifts, as is shown in this review.

These mechanisms can be separated into processes external to the front and those internal to the front. I define *external processes* as those mechanisms for nonclassical frontal structures that are associated with processes not directly associated with the cold front itself, such as those with the environment of the front. On the other hand, *internal processes* are those associated with the cold front itself. Because such processes may not be mutually exclusive, it may be possible that some mechanisms overlap, as discussed further in section 4a. Those mechanisms external to the front include

(a) synoptic-scale forcing
(b) interaction with lee troughs and drylines
(c) interaction with fronts in the mid- and upper troposphere
(d) frontogenesis associated with inhomogeneities in the prefrontal air

Those internal to the front include

(e) surface friction
(f) frontogenesis acting on alongfront temperature gradients
(g) moist processes
(h) descent of air
(i) ascent of air at the front
(j) generation of prefrontal bores and gravity waves

The remainder of this section explores these possible explanations in the order listed above.

a. Synoptic-scale forcing

One way that fronts can develop prefrontal troughs or wind shifts is through the influence of a synoptic-scale system. Consider a case of a mid- to upper-level short-wave trough in advance of a surface front. A prefrontal surface pressure trough may be induced by surface pressure falls in advance of the short-wave trough aloft. Adjustment processes may result in the winds responding to this pressure trough, producing a prefrontal wind shift coincident (or nearly so) with the pressure trough. Thus, the synoptic-scale feature aloft can produce a prefrontal trough and wind shift. Topography may aid in this separation and is discussed further in section 3b. Because the scale of quasigeostrophic processes is quite large, the reflection of surface pressure falls ahead of surface front is likely a broad trough, at least initially.

Idealized frontal models are also capable of producing prefrontal troughs due to synoptic-scale forcing. As reviewed by Smith and Reeder (1988) and Snyder et al. (1993), two idealized frontogenesis models exist: the confluence-induced and horizontal-shear-induced Eady (1949) wave models, as formulated by Hoskins and Bretherton (1972) and Williams (1967), respectively. Confluence-induced models are discussed in this section, whereas horizontal-shear-induced models are discussed in section 3f.

At the time of frontal collapse in a confluence-induced model, the vorticity, surface convergence, and temperature gradient become collocated along the dilatation axis, although before that time these features may not necessarily be collocated (e.g., Davies and Müller 1988; Smith and Reeder 1988). Thus, except for transitory structures, the confluence-induced model as traditionally formulated does not produce prefrontal features.

In such a traditionally formulated confluence-induced model, Smith and Reeder (1988) stated that frontal motion could only occur in response to the secondary circulation itself. In other words, frontal motion was not very realistic. Cunningham and Keyser (1999) demonstrated that extending the confluence-induced model by adding a translating dilatation axis (i.e., the line of zero cross-front basic-state wind speed in a field of pure stretching deformation) could produce more realistic frontal motion. They showed that the motion of the front was ultimately controlled by the movement of the translating dilatation axis. For a translating dilatation axis moving from cold to warm air, the dilatation axis preceded the merged surface baroclinicity and vorticity axes by several hundred kilometers (Cunningham and Keyser 1999, their Fig. 2b). The implication of the results of Cunningham and Keyser (1999) is that frontal motion can be controlled by the large-scale flow in which the front is embedded.

Cunningham and Keyser (1999) showed that when the speed of the basic-state dilatation axis is large, the confluent asymptote of the total wind field (where the sign of the cross-front velocity component changes, the location of which is related to the translating dilatation axis and its speed of movement) may be in the warm air ahead of the maximum baroclinicity or vorticity (see Fig. 1 for an example of what this might look like on a horizontal map of streamlines). Thus, prefrontal wind shifts, when they occur, may occur at the confluent asymptote and/or the axis of maximum vorticity (where the sign of the alongfront velocity component changes). Furthermore, there may be initial periods where the surface baroclinicity and vorticity maxima are not coincident, as in Davies and Müller (1988). Cunningham and Keyser (1999) did not address the surface pressure field, so whether their wind shifts are associated with prefrontal troughs is not known. The Cunningham and

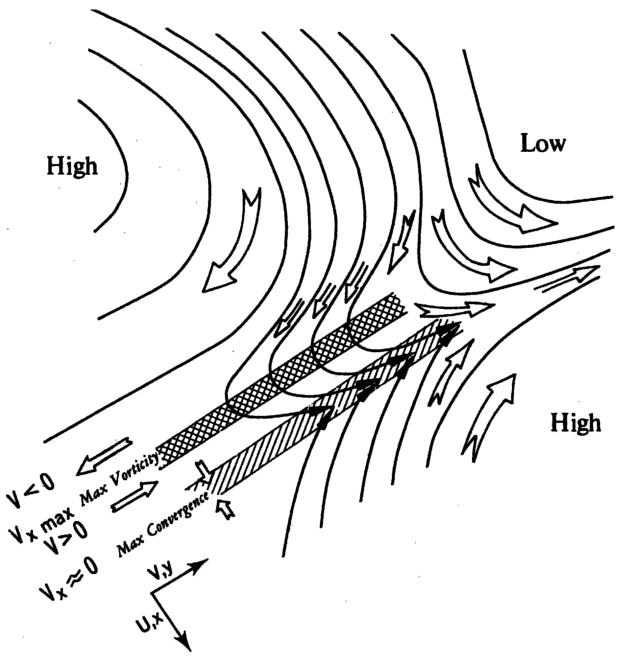

FIG. 1. Schematic of proposed streamline pattern near the surface within a cold-frontal zone with the cold air mass at the top of the figure. A coordinate system relative to the front is shown where x is the cross-front coordinate and y is the alongfront coordinate. The line of maximum vorticity is indicated by cross hatching. The line of maximum convergence is shown by hatching. Vorticity is zero along the maximum convergence line if vorticity and convergence are 90° out of phase (Orlanski and Ross 1984, their Fig. 24).

Keyser (1999) mechanism may exist outside of idealized frontal model simulations, but this mechanism has not been applied to observations of cold fronts.

Observations of fronts in large-scale flow environments resembling pure deformation zones have been documented in at least three instances: Rabin et al. (1987) examined a stationary front over the southern United States, Ostdiek and Blumen (1995) studied a moving cold front over the central plains of the United States, and Wakimoto and Cai (2002) analyzed a moving front over the North Atlantic Ocean. To the ability of the observations from these studies to detect nonsimultaneity, these latter two traveling fronts appeared to possess simultaneity while moving forward into the warmer air. Thus, these studies do not provide a rigorous test for the Cunningham and Keyser (1999) mechanism.

Another location where fronts resembling the confluence model occur frequently is near Australia. The *meridional front* (Civilian Staff 1945; Troup 1956; Taljaard 1972) derives its name from being a north–south-oriented front, sandwiched within a hyperbolic deformation zone created between two mobile anticyclones over the Southern Ocean. Many of these fronts are not true cold fronts (Troup 1956) and may simply represent deformation zones, however. Observations of such fronts before they arrive onshore do not exist, unfortunately. Thus, supporting observations have not arisen and this mechanism remains untested observationally.

b. *Interaction with lee troughs and drylines*

Fronts interacting with topography may cause a classical front to develop nonclassical characteristics. In idealized simulations of fronts moving over mesoscale mountains, Keuler et al. (1992) and Dickinson and Knight (1999) found that an initially simultaneous front produced by the Eady baroclinic wave became altered by its interaction with the mountain such that the surface temperature gradient associated with the front was up to 400 km behind the surface vorticity maximum. Illustrated schematically, the Eady wave approached the mountain (Fig. 2a), and the separation between the temperature gradient and the surface vorticity was due to the blocking of the original front on the upstream side of the mountain and the development of new vorticity in the lee of the mountain in the presence of lower-level warm advection (Figs. 2b,c). Once the surface cold air and the upper-level portion of the wave moved over the ridge, the vorticity maximum (Fig. 2d) and the temperature gradient recombined and moved eastward.

In the lee of the Rocky Mountains, Hutchinson and Bluestein (1998) showed that as many as 60% of the cold fronts in the central United States were associated with prefrontal wind shifts (Fig. 3). For example, Sanders (1967, 1983) showed a front over Oklahoma using high-resolution surface observations where the wind shift preceded the temperature drop by 1–16 min. A composite analysis of 1-min time series analyses at stations where frontal passage occurred showed that the maximum convergence and vorticity lay ahead of the temperature gradient, when composited across all the observing sites relative to the passage of the vorticity maximum (Sanders 1967, 1983). Other examples of fronts with prefrontal features of this type have been presented by Locatelli et al. (1989), Colle and Mass (1995), Schultz (2004), and Schultz and Roebber (2005, manuscript submitted to Amer. Meteor. Soc. *Meteor. Monogr.*, hereafter SR), Hutchinson and Bluestein (1998) and Schultz (2004) attributed these prefrontal wind shifts to drylines (e.g., Schaefer 1986) or lee troughs.

As illustrated schematically in Fig. 4a, westerly flow across the Rocky Mountains leads to the formation of a lee trough. An equatorward-moving cold/arctic front and prefrontal warm advection causes the movement of

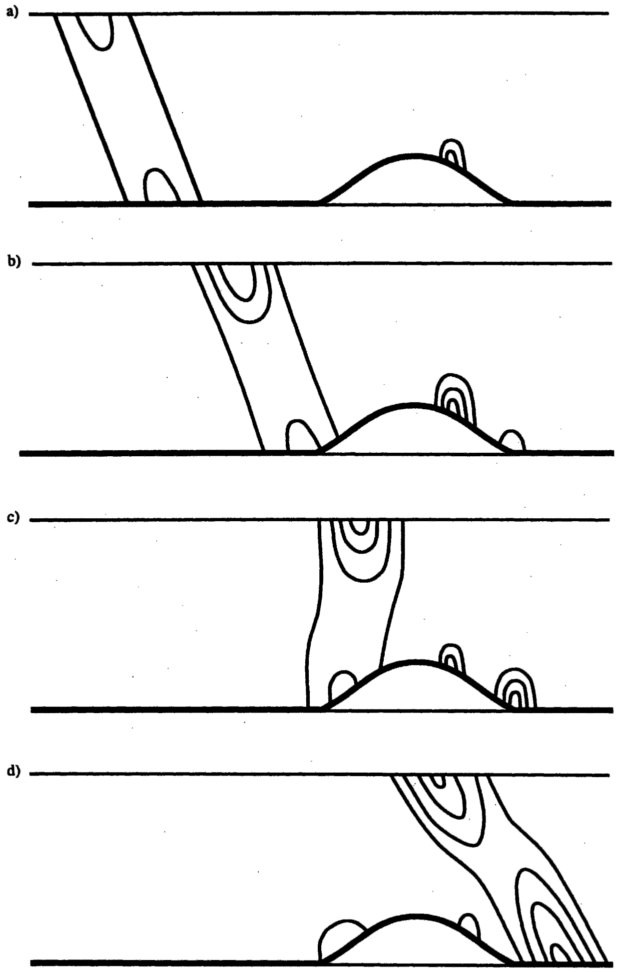

FIG. 2. Idealized schematic of a front moving over a mountain (Dickinson and Knight 1999, their Fig. 14). (a) The approach of a coupled upper- and lower-level trough toward a mountain; (b) the low-level blocking of the front along the windward slope, and the development of the lee trough and secondary trough along the lee slope; (c) the separation of the upper-level and lower-level frontal waves; and (d) the coupling of the upper-level frontal wave with the secondary trough in the lee of the mountain. Isopleths represent relative vorticity. Time between panels is 6–8 h, depending on the strength of the frontal circulation and the size of the mountain (M. Dickinson 2004, personal communication).

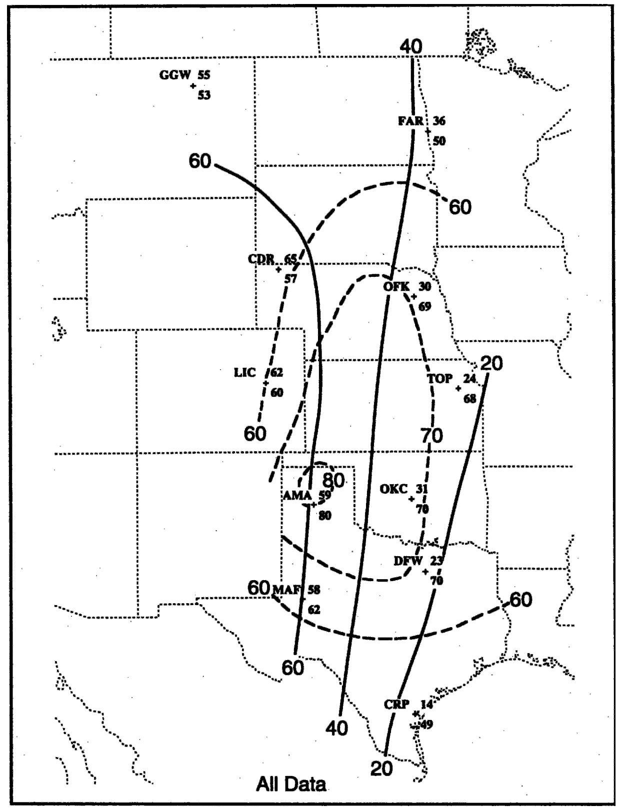

FIG. 3. A 2-yr climatology of prefrontal wind shifts between 1 May 1993 and 30 Apr 1995. Station model shows (upper left) station identifier, (upper right) percentage of cold fronts that were preceded by wind shifts, and (lower right) number of frontal passages during the sample. Percentage of fronts preceded by prefrontal wind shifts (solid lines every 20%) and the number of fronts (dashed lines every 10) (Hutchinson and Bluestein 1998, their Fig. 7).

the lee trough away from the mountains (Fig. 4b). This warm advection implies that trough development occurs ahead of the surface cold front. Such low-level warm advection may be due to secondary circulations associated with a 500-hPa short-wave trough (e.g., Uccellini 1980). Schultz (2004) presented an alternative explanation whereby the departure of the lee trough from the mountains was due, not to the warm advection, but to forcing for surface pressure falls by a mobile upper-level short-wave trough. Eventually, the faster-moving cold front catches up to the lee trough, becoming one feature at the surface (Fig. 4c), as discussed by numerous authors (e.g., Shapiro 1982; Locatelli et al. 1989, 2002a; BLU; Bluestein 1993, p. 162; Keshishian et al. 1994; Colle and Mass 1995; Hutchinson and Bluestein 1998; Neiman et al. 1998; Neiman and Wakimoto 1999; Parsons et al. 2000; Stoelinga et al. 2000; Rose et al. 2002). Thus, the structures and evolution of the conceptual model in Fig. 4, which are based on observational analysis, mimics that seen in the idealized model simulations of Keuler et al. (1992) and Dickinson and Knight (1999).

Prefrontal descent caused by topographic downslope flow (i.e., a foehn) can also lead to a prefrontal trough. For example, Heimann (1990) showed numerical simulations in which warming of 5 K occurred in the prefrontal lower troposphere. Such prefrontal warming increased the horizontal temperature difference across the front [also noted for a different case by Schultz and Trapp (2003, their Fig. 7b)], reduced the prefrontal sur-

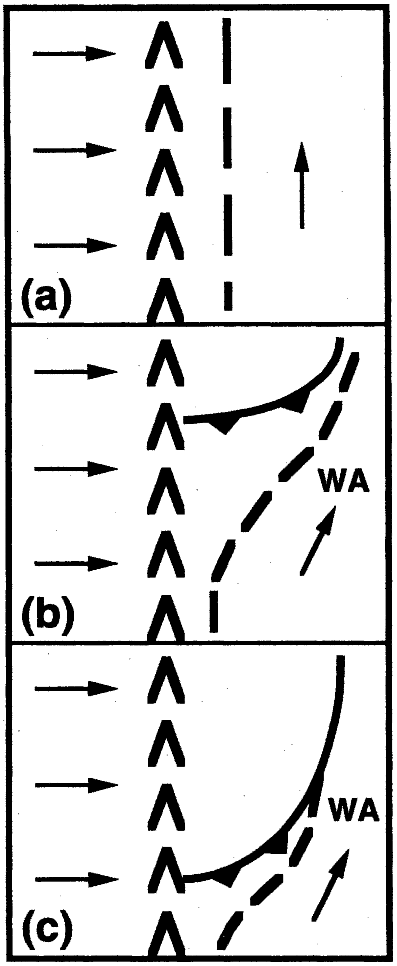

FIG. 4. Idealized evolution of a prefrontal wind shift (dashed line). (a) Initially a wind shift associated with a lee trough or dryline forms when low-level flow (arrows) is perpendicular to the mountains. (b) The wind shift propagates away from the mountains when the low-level flow veers with time and warm advection (indicated by WA) is occurring to its east. (c) Eventually, the front catches up to the wind shift (Hutchinson and Bluestein 1998, their Fig. 24).

face pressure, and produced a forward-tilting frontal structure (Heimann 1990).

Observationally, fronts moving through the western United States may not necessarily fit the classical model of fronts (e.g., Williams 1972). For example, Schultz and Doswell (2000, p. 162) examined a case of a cyclone moving from the Pacific Northwest and redeveloping in the lee of the Alberta Rocky Mountains. They showed that, although a surface pressure trough could be followed through the western United States, it could not be characterized as a front. Schultz and Doswell (2000, section 2c) argued that low-level baroclinicity does not move through the western United States as easily as upper-level features, allowing for the pressure trough associated with the upper-level wave to separate from any preexisting surface cold front. An example of such a mechanism may be the apparent separation between the surface trough and the baroclinicity in Sanders (1999b), where the forcing for surface pressure falls associated with the 500-hPa trough lay ahead of the surface cold front (cf. Figs. 1a, b, c and 4a, c, e in Sanders 1999b, respectively), although it is likely that diabatic processes cannot be divorced from this case (e.g., Hoffman 1995).

Drylines in the south-central United States can also serve as prefrontal wind shifts. The dryline represents a quasiclimatological boundary between the southerly moist air mass originating over the Gulf of Mexico and the westerly or southwesterly dry, well-mixed air mass originating over New Mexico and west Texas. Advancing fronts from the west or north can approach and interact with the dryline, giving the appearance of a prefrontal wind shift. Merger between drylines and fronts often results in convective initiation and severe weather (e.g., Koch and McCarthy 1982; Ogura et al. 1982; Shapiro 1982; Schaefer 1986; Neiman et al. 1998; Koch and Clark 1999; Neiman and Wakimoto 1999; Parsons et al. 2000; Stoelinga et al. 2000; Rose et al. 2003; BLU).

c. Interaction with fronts in the mid- and upper troposphere

The interaction between a surface-based cold front and an upper-level system to produce a nonclassical frontal structure can manifest itself in other ways as well. Here, interaction could mean the merger or simply collocation of these two features. During the 12–14 March 1993 Superstorm, Schultz et al. (1997) observed a prefrontal trough and wind shift ahead of the temperature drop associated with the cold front in eastern Mexico. Schultz and Steenburgh (1999) showed that these prefrontal features were a result of a midtropospheric front interacting with the surface cold front to produce a forward-tilting structure with multiple cloud bands. The cold advection aloft preceding that at the surface resulted in the warmest tropospheric mean temperatures at the leading edge of the midtropospheric front. Thus, a surface pressure trough preceded the surface cold front, with the surface wind shift responding to the trough.

Locatelli et al. (1995, 1997) suggested that weak surface pressure troughs or an inflection in the surface pressure traces occurred ahead of surface fronts owing to cold advection aloft, similar to the instability lines of Fulks (1951) and House (1959). Further observational evidence was provided by Neiman et al.

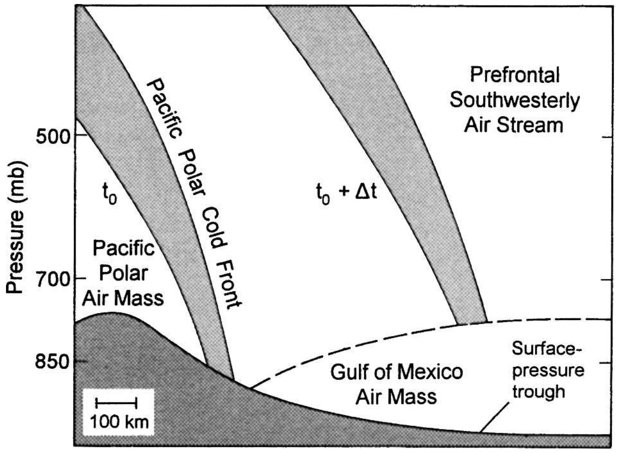

FIG. 5. Schematic of the decoupling of the Pacific cold front from the surface by the Gulf of Mexico air mass. The Pacific front is shown at two times (t_0 and $t_0 + \Delta t$, where Δt is about 18 h) relative to the stationary Gulf of Mexico air mass. Top of the shallow Gulf of Mexico air mass (dashed line) and intersection of the dashed line with the ground indicates the surface position of the dryline. West is to the left (Neiman et al. 1998, their Fig. 27).

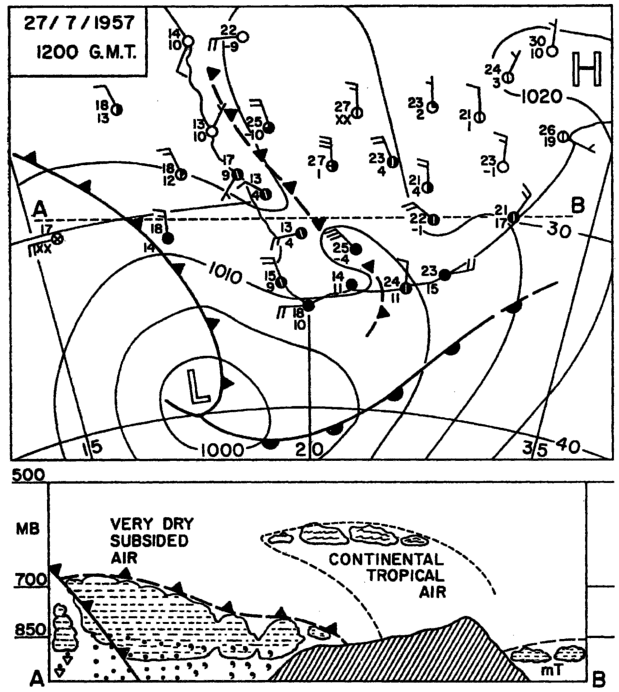

FIG. 6. Example of a "leader" or "intercell" front over the western part of the South African plateau in the winter in advance of the polar cold front. Section AB is along 30°S (Taljaard 1972, his Fig. 8.11).

(1998)[1] and Neiman and Wakimoto (1999) over the central United States, showing that surface pressure troughs were indicative of the location of a previous surface-based front having moved aloft (Fig. 5). Stoelinga et al. (2003) showed that the convergence associated with the moving surface pressure field associated with a cold front aloft could initiate and maintain a convective line. In Europe and the eastern United States, structures such as split fronts (e.g., Browning 1990; Koch 2001; and references therein) may show similar behavior in surface weather patterns, although this has not been demonstrated. Thus, the interaction between frontal zones at the surface and aloft may give the appearance of a nonsimultaneous cold front at the surface.

d. *Frontogenesis associated with inhomogeneities in the prefrontal air*

Several authors have proposed mechanisms involving inhomogeneities in the prefrontal air that lead to the formation of a prefrontal trough and wind shift. These mechanisms include (i) preexisting frontal features ahead of the surface front and (ii) inhomogeneities in the prefrontal environment that develop frontal characteristics due to synoptic or frontal forcing.

An example of the first mechanism is the *leader or intercell front* in South Africa (e.g., Taljaard 1972), which appears to be an onshore-moving sea breeze ahead of a cold front (Fig. 6). Perhaps most intriguing is the second mechanism by which a constant forcing applied to a region with horizontal variations in static stability develops multiple fronts. Hoskins et al. (1984) showed that the addition of a surface-based warm anomaly in the warm air can cause a second front to develop ahead of the primary front (Fig. 7). They performed idealized two-dimensional semigeostrophic simulations of growing baroclinic waves. One featured a surface temperature profile monotonically decreasing to the north (their "smooth temperature profile"); the other featured the same temperature profile, but with a warm anomaly in the warm prefrontal air (their case

[1] Neiman et al. (1998) analyzed the same case as Locatelli et al. (1995), arriving at a slightly different conclusion. Whereas Locatelli et al. (1995, p. 2648) found that there was no horizontal potential temperature gradient at the surface with the Pacific cold front before the merger with the dryline, Neiman et al. (1998, 2532–2533) found a surface frontal signature. In regard to other cases, however, Hobbs et al. (1996, p. 1173) state, "As this upper-level baroclinic zone, or CFA [cold front aloft], moves over the Rockies, it may be associated with a front at the surface. However, as it moves down the eastern slope of the Rockies and out over the Great Plains, adiabatic warming associated with the low-level downslope flow tends to erode the baroclinic zone at the surface." Later papers by Hobbs and collaborators emphasize the formation of the CFA as resulting from a surface Pacific cold front occluding with the lee trough, rather than from low-level frontolysis due to downslope adiabatic warming.

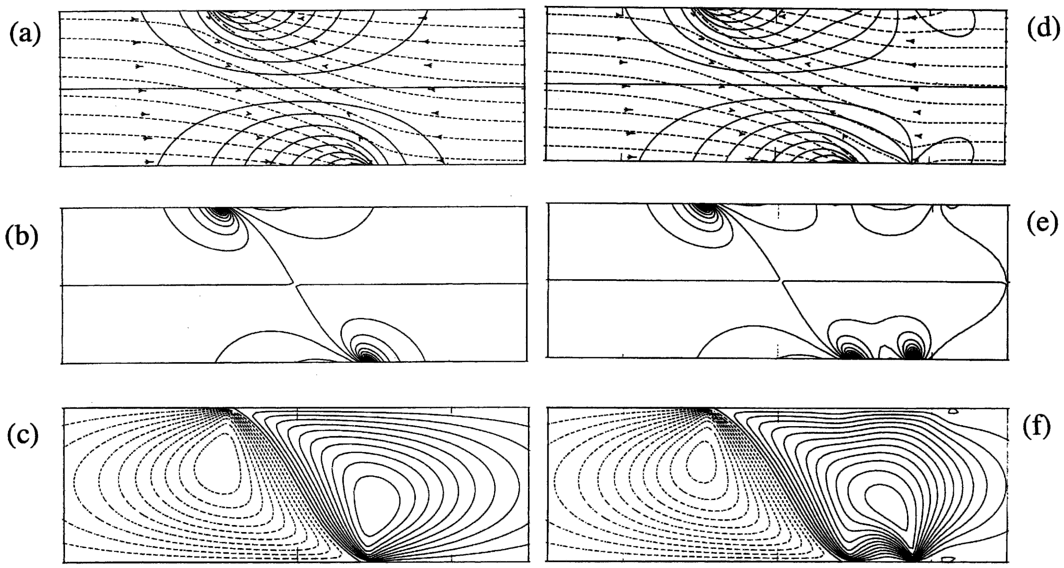

FIG. 7. Idealized frontal simulations: (a)–(c) the smooth-profile solution and (d)–(f) case IV, a solution with a warm anomaly of maximum 4°C added to the warm air, as per Fig. 3 in Hoskins et al. (1984). (a) Alongfront wind speed (solid lines every 5 m s^{-1}), potential temperature (dashed lines every 4°C), and the cross-frontal streamlines (arrows). (b) Relative vorticity (solid lines every $0.2f$); the zero contour passes along the middepth of the section. (c) Vertical velocity (every 0.1 cm s^{-1}; solid positive and dashed negative). (d), (e), and (f) Same as (a), (b), and (c), respectively, except for case IV (Hoskins et al. 1984, their Figs. 1, 4d, 5d, and 6d).

IV). In the smooth-profile simulation, a single region of strong surface temperature gradient, single surface vorticity maximum, and single midtropospheric vertical motion maximum form (Figs. 7a,b,c, respectively). In contrast, in case IV, the warm anomaly results in the production of a second front in the warm air ahead of the original front (Figs. 7d–f). Specifically, a second region of strong thermal gradient and second surface vorticity maximum forms, as well as a shift in the midtropospheric vertical motion maximum over the second front (Figs. 7d,e,f, respectively). This second front has a temperature gradient, vorticity maximum, and vertical motion as large as, or larger than, those of the primary front (Figs. 7d,e,f, respectively) and of the front in the case without the warm anomaly (Figs. 7a,b,c, respectively).

In a different idealized model configuration testing the same process, Reeder et al. (1991) showed that the formation of this second front is a reflection of the model's enhanced response to the frontogenetic forcing in the presence of weaker static stability induced by the warm anomaly. Observationally, such warm anomalies may occur owing to intense surface heating over continents ahead of cold fronts, as in the prefrontal troughs formed by heat lows or downsloping in the lee of orography over Australia (e.g., Fandry and Leslie 1984; Physick 1988; Kepert and Smith 1992; Skinner and Leslie 1999; Kraus et al. 2000; Preissler et al. 2002). The Spanish plume (Morris 1986; van Delden 1998) may also be a similar feature over western Europe. Despite these similarities, this mechanism remains undiagnosed using real data.

Whereas the previous four mechanisms dealt with external explanations for the prefrontal features, these next subsections address those situations for which the dynamics related to the front itself (internally) result in the pressure trough or wind shift being out ahead of the cold front.

e. Surface friction

Many early cross sections of cold fronts were drawn with a nose aloft protruding forward above the surface position of the front (e.g., Kobayasi 1923; Giblett 1927; Flower 1931; Brunt 1934, 344–345), rather than the rearward-sloping surface depicted by the Norwegians. This nose was later documented with high-resolution, temporal-resolution data from an instrumented tower by Shapiro (1984) (Fig. 8).

Brunt (1934) argued that such a structure would imply a prefrontal surface pressure trough because the leading edge of the cold advection arrives aloft first. Taylor et al. (1993) observed a 25-km separation between the wind direction and the temperature gradient, and they suggested that the overhanging cold air may have been responsible.

Such prefrontal structures have also been detected in

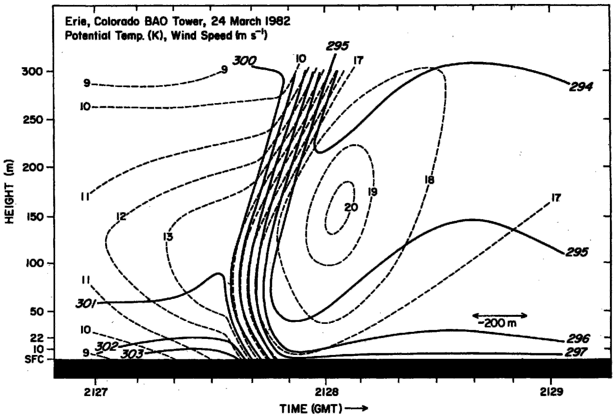

FIG. 8. Time section of the front-normal wind component (m s^{-1}; dashed lines) and potential temperature (K; solid lines) for 2127–2129 UTC 24 Mar 1982 cold front observed by the Boulder Atmospheric Observatory tower in Erie, CO (Shapiro 1984, his Fig. 3).

observations (e.g., Lawson 1971; Idso et al. 1972; Charba 1974; Goff 1976; Mahoney 1988) and numerical simulations (e.g., Mitchell and Hovermale 1977) of thunderstorm gust fronts, which have been related to density current dynamics. Such an overhanging nose structure has been attributed to surface friction (e.g., Simpson 1972; Mitchell and Hovermale 1977) and would also appear to be relevant to cold fronts acting like density currents. Indeed, Simpson (1972) attributed the overhanging nose to the no-slip lower-boundary condition, allowing the overrunning of lighter prefrontal air by the denser postfrontal air. This produces a thin layer of prefrontal fluid that is transported underneath the cold air. The lobe-and-cleft structure at the head of a density current is a result of the release of convective instability of this unstable stratification. Furthermore, Droegemeier and Wilhelmson (1987) showed that the pressure increase preceding the cold air at the surface was dynamically induced by the collision of the cold air and the prefrontal air.

f. Frontogenesis acting on alongfront temperature gradients

The horizontal-shear-induced model of frontogenesis was reviewed by Smith and Reeder (1988). Based on earlier work (e.g., Hsie et al. 1984; Reeder 1986; Reeder and Smith 1986, 1987), Smith and Reeder (1988) found that alongfront warm advection leads to surface pressure falls ahead of the frontal zone. Whereas the alongfront warm advection induces height falls that can propagate eastward, causing the prefrontal pressure trough and the wind shift, the isotherms constituting the frontal zone are transported by the wind at the advective wind speed. Thus, the possibility exists that propagation of the pressure trough could occur relative to the temperature gradient (Fig. 9), implying a separation between the pressure trough and the temperature gradient. Idealized fronts in such cases have the maximum surface relative vorticity, maximum surface convergence, and minimum surface pressure coincident, but they lie ahead of the maximum horizontal temperature gradient by order 100 km.

Sanders (1999a) offered a similar argument. He re-

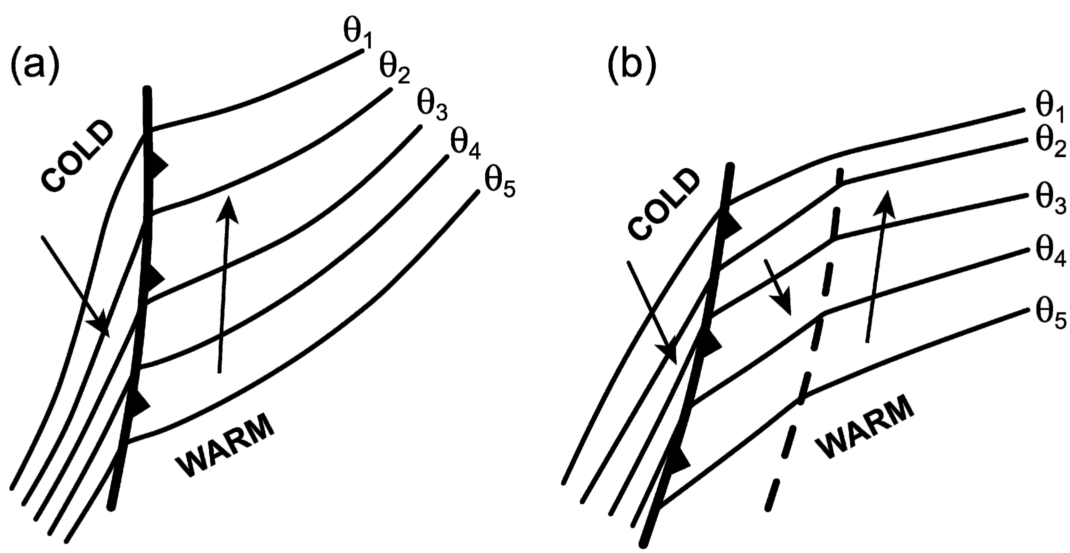

FIG. 9. Evolution of a front with an alongfront temperature gradient: (a) temperature gradient and wind shift line are coincident, (b) alongfront warm advection causes troughing, and associated wind shift line moves eastward faster than cold advection, producing a prefrontal trough.

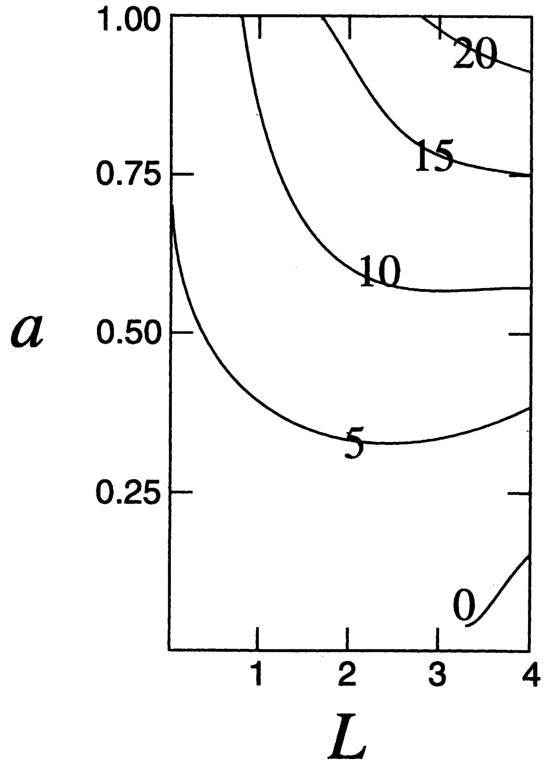

FIG. 10. Eastward propagation speed (m s^{-1}) of a surface trough as a function of wavelength, L (in 1000 km), and meridional temperature gradient, a [°C (100 km)$^{-1}$] (Sanders 1999a, his Fig. 4).

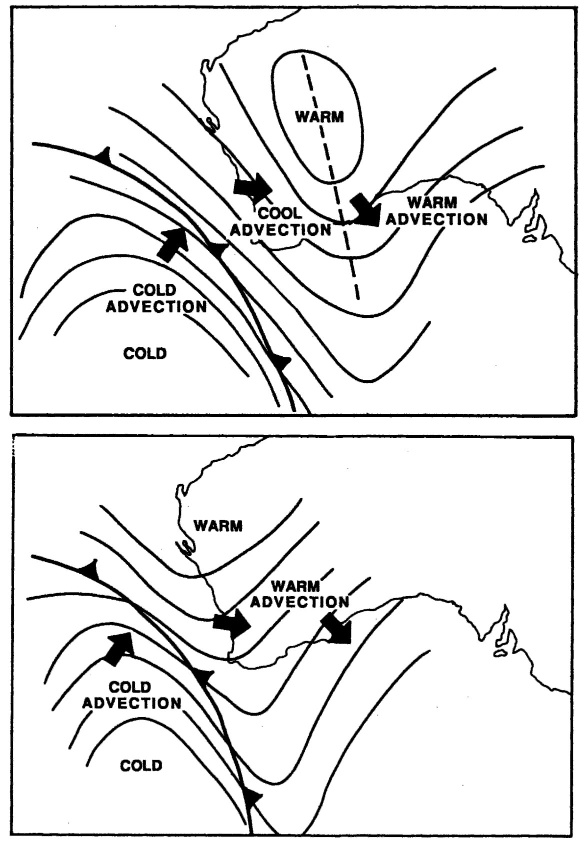

FIG. 11. Schematic thermal advection pattern for a cold front approaching western Australia in (a) spring and (b) midwinter. Solid lines represent surface isotherms. Arrows indicate wind flow associated with the front/trough system. The dashed line in (a) represents a discontinuity in thermal advection (Hanstrum et al. 1990a, their Fig. 8).

lied on the Sanders (1971) analytic model of baroclinic disturbances, which showed that propagation of the pressure trough relative to the earth could be 5–20 m s^{-1} (Fig. 10) for even modest values of the alongfront temperature gradient [<1 °C (100 km)$^{-1}$]. Sanders (1999a) argued that, because the isentropes move at the advective speed of the postfrontal wind, separation between the more rapidly moving pressure trough from the isentropes (front) is possible. In addition to the lee-trough mechanism (section 3b), the alongfront warm-advection mechanism suggests another way that prefrontal warm advection can lead to prefrontal features.

This mechanism has been proposed to explain the presence of prefrontal troughs, but has not been quantitatively tested using observations. For example, the west coast of Australia is known for its frequent occurrence of fronts with alongfront warm advection (e.g., Fig. 11), although the alongfront warm advection hypothesis has not been tested. Over the southwest United States, Sanders (1999b) found no agreement between the quantitative prediction from his theory and the observations (10 m s^{-1} computed versus 17.2 m s^{-1} observed). The applicability of this theory for this event, however, may be limited due to hypothesized diabatic effects on the frontogenesis (Hoffman 1995).

Possibly a similar mechanism for prefrontal features in horizontal-shear-induced frontogenesis models was proposed by Ross and Orlanski (1982) and investigated in more detail by Orlanski and Ross (1984) using a hydrostatic primitive equation model for a real atmospheric case (see also Emanuel 1985a). They found that the ageostrophic vorticity term in the surface divergence-tendency equation resulted in a negative feedback to limit intensification of the front (defined in these studies as the maximum of surface relative vorticity). This negative feedback consisted of descent occurring over the leading edge of front, effectively shifting the convergence maximum to the warm side of the frontal zone by up to 400 km (Fig. 1), although this effect was not found by Levy and Bretherton (1987). In a linear, two-layer, dry model, Orlanski and Ross (1984) found that vorticity and divergence within the surface front oscillated about an equilibrium state, with a periodicity close to the inertial period. Garner (1989a,b), on the other hand, concluded that this periodicity implied an inertial oscillation and that the energy of these oscillations was too small to produce the observed features. In fact, the unbalanced initial conditions used by Orlanski and Ross (1984) may have produced the inertial oscillations, which resulted in the separation. Thus, the relevance of this work, especially to the real atmosphere, is unknown.

Reeder (1986, p. 143) showed in his idealized numerical model simulations that convergence preceded the temperature gradient and vorticity by about 40 km. Two-dimensional shear simulations of idealized fronts by Orlanski and Ross (1977), Gidel (1978), Reeder and Smith (1986, 1987, 1988), and Keuler et al. (1992) showed the same behavior. Whereas Keuler et al.'s (1992) simulations featured a near-constant separation of about 200 km between the maximum temperature gradient and the surface convergence maximum, Reeder and Smith's (1986) simulations had a decrease in the separation of these features during frontal collapse. Reeder and Smith (1987) noted that, in a three-dimensional simulation of a nonsimultaneous Australian cold front, the surface convergence led the vorticity maximum by 95 km, followed in 158 km by the temperature gradient. In this case, the temperature gradient was in a region of horizontal divergence, leading to weakening. Gall et al. (1987) showed that the offset described by Orlanski and Ross (1984) was present in their model simulations, although it did not lead to a weakening of the front because there was still substantial overlap between the surface convergence and vorticity. In fact, the front continued to strengthen even with the separation because the separation was not large enough to produce divergence at the location of the vorticity maximum. Ultimately, Gall et al. (1987) concluded the limited vertical resolution of their model led to bounds on the frontogenesis.

Consequently, observational evidence for the Orlanski and Ross (1984) mechanism has not been readily forthcoming. Levy (1989) examined snapshots of four cold fronts over the ocean from satellite-derived wind data and found that the convergence maximum preceded the vorticity maximum by about 50 km, unless the front was in pure confluence, in which case the two were coincident. Whether the nonsimultaneity of these fronts was due to the Orlanski and Ross (1984) mechanism could not be examined owing to the lack of temporal resolution in the data on the frontal evolution. Nevertheless, these observations support the frontal structure described by the mechanism, if not the dynamics of the mechanism themselves.

The temperature field may be transported by inertial oscillations present in the frontal zone, which may undergo frontogenesis by the larger-scale flow and result in multiple frontal zones (Blumen 1997). Observational evidence for inertial oscillations was presented by Ostdiek and Blumen (1995, 1997). It is intriguing to speculate that the periodicity of this oscillation between coincident and separated vorticity and convergence is related to the inertial period, although this hypothesis has not been explored observationally.

In summary, there is much that remains to be learned about the role of alongfront thermal advection. Although theoretical and modeling studies suggest the plausibility of this mechanism, observations to date have not produced quantitative agreement.

g. *Moist processes*

Moist processes can also lead to prefrontal features through a variety of ways. First, Emanuel (1985b) argued that in the limit of small moist symmetric stability saturated ascent could occur 50–200 km ahead of the maximum in frontogenetic forcing in an idealized two-dimensional diagnostic model (Fig. 12). As Emanuel (1985b) showed, this situation was frontolytical at the front and frontogenetical ahead of the front, thereby leading to more rapid propagation of the front than if the front were dry, a result confirmed in a real-data numerical model simulation by Reeves and Lackmann (2004). Verifying these results in a prognostic idealized model has not been performed, however.

Second, several authors have argued that subcloud evaporation of falling precipitation under clouds ahead of the surface front can produce prefrontal pressure troughs and wind shifts (e.g., Sawyer 1946; Fujita 1959). The formation of prefrontal features in such dry subcloud environments in Australia has been demonstrated by Ryan et al. (1989). These frontal structures have also occurred in other arid regions of the world, such as China (e.g., Mitsuta et al. 1995; Takemi 1999) and the western United States (e.g., Schultz and Trapp 2003). A schematic of the prefrontal structure in Schultz and Trapp (2003) is shown in Fig. 13. In this case, sublimation/evaporation of hydrometeors leads to cooling aloft ahead of the surface front. This cooling aloft leads to an increase in surface pressure just before the surface front, resulting in the lowest surface pressure occurring minutes to hours ahead of the surface front (Fig. 13).

Third, in a case of discontinuous surface frontal propagation over the central United States, Bryan and Fritsch (2000a,b) found that prefrontal thunderstorms had laid down a surface cold pool through which the advancing cold front was unable to penetrate down to the surface. As the cold front moved overtop the cold pool, a prefrontal trough on the downwind side of the cold pool developed, subsequently resulting in surface

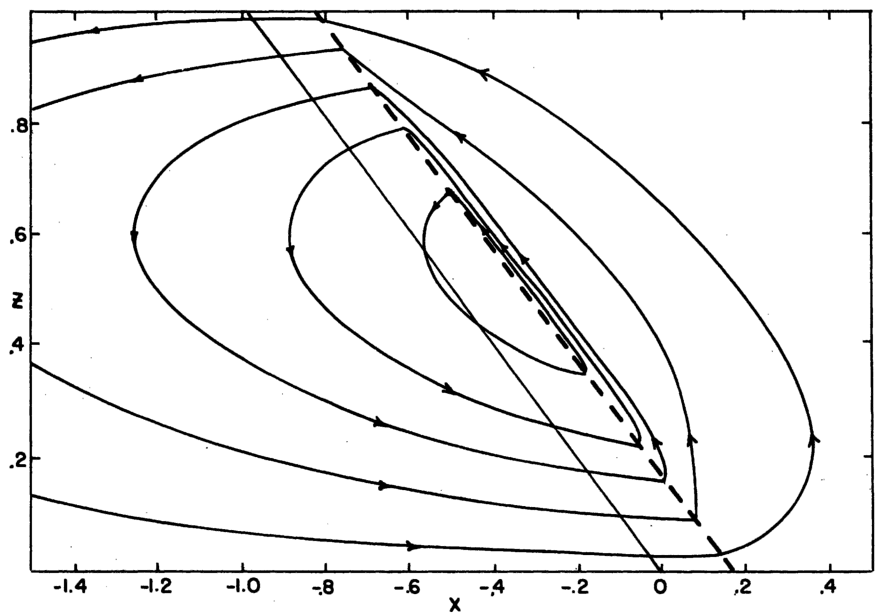

FIG. 12. Streamlines of the cross-front circulation in physical space from an idealized two-dimensional semigeostrophic model. The minimum value of the dimensionless streamfunction is −1.769; contours are 0.1, 0.3, 0.5, 0.7, and 0.9 times the minimum value; and the heavy dashed line denotes the surface $X = L$ (line of maximum frontogenetical forcing) (Emanuel 1985b, his Fig. 5).

frontogenesis. Eventually this prefrontal trough became the dominant front as the original front weakened. The mechanism bears some similarities to that discussed in section 3i of Charney and Fritsch (1999).

h. *Descent of air*

Another way that has been proposed to generate a front with multiple boundaries is to cause warming by adiabatic descent in the postfrontal air. For example, Read (1925), Bjerknes (1926, 1930), Gold (1935, p. 119), and Berry et al. (1945, p. 649) suggested that the postfrontal descent of air from midlevels could be responsible for the weakening of a cold front and the formation of a double cold front by adiabatic warming in the descending air. This mechanism has not been observationally verified, however.

More recently, Hoxit et al. (1976) and Rutledge (1989) argued that prefrontal descent can lead to surface pressure troughs and wind shifts ahead of precipitating fronts, without the weakening of the original front. Given this perspective, a surprising number of cross sections through some observed and modeled fronts, both precipitating and nonprecipitating, show regions of midlevel prefrontal descent 200–500 km ahead of the surface cold front (Table 1; Fig. 14). The magnitudes of this descent range from 3 to 70 mb h^{-1}—the large range is likely a result of the differing resolutions of the observing and modeling systems used in the cited studies. These papers in Table 1 generally do not address the cause of this prefrontal descent. [Hsie et al. (1984) attribute this prefrontal descent to evaporation.] It could be that this descent is due to the nonhydrostatic effect of the ascent plume associated with deep moist convection at the leading edge of the cold front reaching its level of neutral buoyancy and sinking, but this has not been evaluated. In the case of Chen and Bishop (1999) and other similar models with a rigid lid (not shown), the subsidence may be a result of the frontal updraft hitting the lid in the model and being forced to subside. Nevertheless, descent of this magnitude may be a way to generate prefrontal troughs and wind shifts.

i. *Ascent of air at the front*

The ascent at the leading edge of a cold front results in adiabatic cooling, which may alter the frontal structure (e.g., Ross and Orlanski 1982, p. 319; Mass and Schultz 1993). Indeed, weakening the temperature gradient across the cold front to return the front to thermal wind balance is one purpose of the secondary circulation associated with the front. Consider a front with an unsaturated 1 m s^{-1} updraft ingesting a prefrontal air mass with a lapse rate of 6.5°C km^{-1}. Lifting the air 1 km in 16.7 min would result in dry adiabatic cooling of

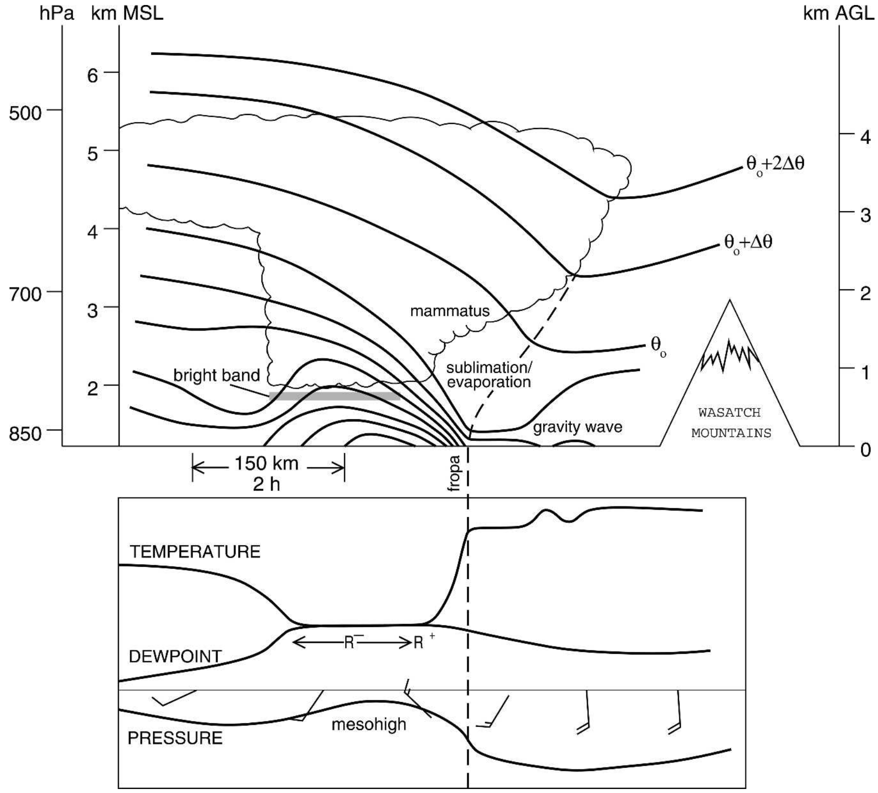

FIG. 13. Conceptual model of the cold front from 14 to 15 Feb 2000. (top) Schematic of cloud (scalloped lines) and potential temperature (thick solid lines); fropa = surface frontal passage. (bottom) Time series of temperature, dewpoint, sensible weather, winds, and pressure at the surface (Schultz and Trapp 2003, their Fig. 23).

3.3°C relative to the nonascending environmental air. Such cooling would be even greater for more stable lapse rates. For example, Taljaard et al. (1961, p. 38) noted the strong prefrontal cooling that can occur in the presence of a prefrontal inversion. Therefore, the effect of the adiabatic cooling owing to ascent can be considerable. This effect is mitigated by horizontal warm advection and the release of latent heat in the updraft (e.g., Bond and Fleagle 1985), negating the cooling entirely during saturated ascent if the prefrontal lapse rate equals the moist adiabatic lapse rate.

Another result of this lower- to midtropospheric adiabatic cooling is that the front may develop a forward tilt (e.g., Taylor et al. 1993; Mass and Schultz 1993; Steenburgh and Mass 1994; Locatelli et al. 1995). Because ascent is occurring at the leading edge of the front, adiabatic cooling would be greatest in the lower to midtroposphere right above the front. Thus, the cooling would arrive aloft before the cooling at the surface. For example, Steenburgh and Mass (1994) attributed their forward-tilting front to adiabatic cooling of the rising air at the leading edge of the cold front. Figure 15 shows the forward tilt of the leading edge of the thermal gradient from the surface (LCA) to 800 hPa, a distance of 150 km. Supporting their argument is that the gradient in relative humidity from moist cold air (presumably due to ascent) to dry prefrontal air is collocated with the forward tilt (Fig. 15). Other observational and modeling studies of forward-tilting cold fronts have been discussed by Schultz and Steenburgh (1999), Parker (1999), and Stoelinga et al. (2002).

If this lifting acts on a surface-based cold anomaly, then a prefrontal feature could be formed. Charney and Fritsch (1999) showed that a surface-based cold anomaly capped by an inversion could be tilted by the advancing ascent associated with a front (Figs. 16a,b).

TABLE 1. Studies of cold fronts with strong prefrontal descent.

Authors	Their figures
Hobbs et al. (1980)	Fig. 14
Ogura and Portis (1982)	Fig. 19
Hsie et al. (1984)	Fig. 3a
Schultz et al. (1997)	Figs. 8a,c
Thompson and Williams (1997)	Fig. 12d
Chen and Bishop (1999)	Figs. 1, 2

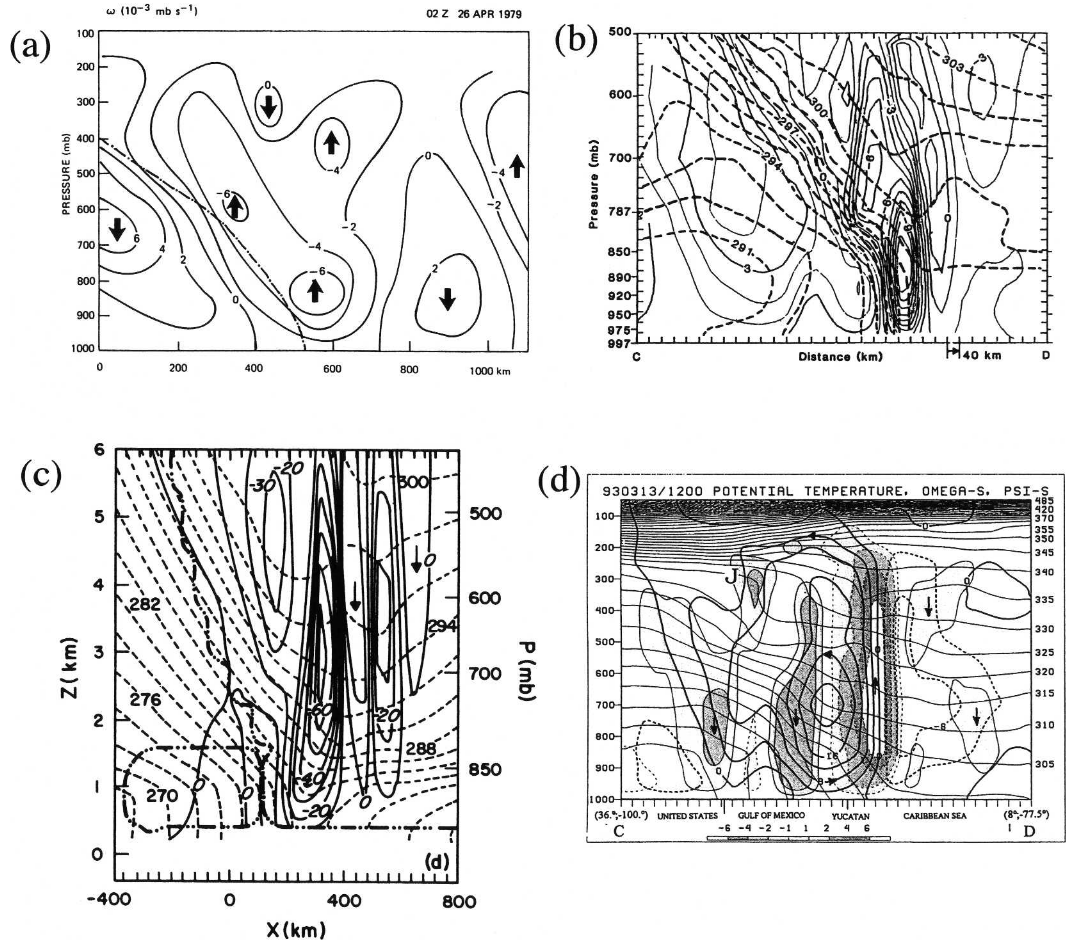

FIG. 14. Fronts with prefrontal descent: (a) vertical velocity (solid lines every 2×10^{-3} mb s^{-1}); dashed–dotted line represents axis of maximum vorticity (Ogura and Portis 1982, their Fig. 19). (b) Vertical velocity (solid lines in μb s^{-1}) and potential temperature (dashed lines every 1 K) (Thompson and Williams 1997, their Fig. 12d). (c) Vertical velocity (solid lines every 10 mb h^{-1}) and potential temperature (dashed lines every 2 K) (Hsie et al. 1984, their Fig. 4d). (d) Potential temperature (thin solid lines every 5 K), streamfunction for the divergent circulation (thick solid lines every 4×10^4 Pa m s^{-1}), and vertical velocity (0.1 Pa s^{-1}; shaded) (Schultz et al. 1997, their Fig. 8c).

They showed that frontogenesis by tilting, as well as precipitation-induced diabatic cooling (Fig. 16c), led to the development of a prefrontal horizontal temperature gradient. Secondary circulations associated with the frontogenesis resulted, further intensifying the prefrontal temperature gradient at the expense of the original front (Figs. 16c,d). Eventually, the prefrontal temperature gradient became the dominant feature, supplanting the original front (Fig. 16d). Thus, the surface front was said to have propagated discretely over 600 km in about 12 h.

j. Generation of prefrontal bores and gravity waves

There is abundant literature on the formation of prefrontal bores and gravity waves generated on stable layers in the prefrontal air. Just a sampling of the wide variety of those studies is described in this section; a more thorough review of bores and gravity waves as they relate to squall lines can be found in Locatelli et al. (2002b, 1646–1647).

To put bores and gravity waves in a broader context, Haertel et al. (2001) argue for a spectrum with gravity currents and gravity waves at opposite ends. Gravity currents represent the advection of cold air due to density differences between cold and warm air. After the passage of a gravity current, surface pressure increases abruptly. Gravity waves, on the other hand, are propagating waves in which buoyancy is the restoring force, resulting in only a temporary increase in surface pressure after their passage. Bores, which are partly advective and partly propagative, would lie in the middle of this spectrum. Traditionally, a bore is a solitary wave,

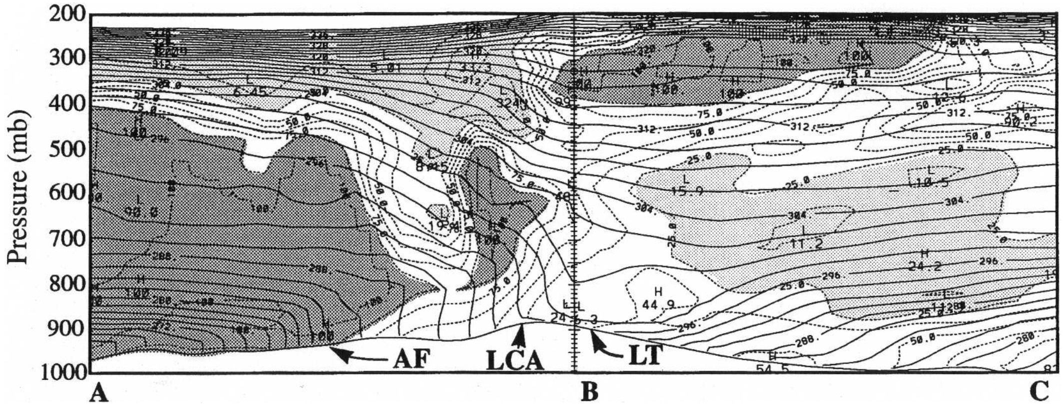

FIG. 15. Cross section of potential temperature (solid every 1 K) and relative humidity (dashed every 12.5%, light shading below 25%, and dark shading more than 87.5%). LCA represents the leading edge of the cold advection at the surface, AF represents the surface location of the arctic front, and LT represents surface location of lee trough (Steenburgh and Mass 1994, their Fig. 16b).

and its structure and dynamics are analogous to a hydraulic jump that propagates in advance of a cold front, characterized by sustained increases in surface pressure, the height of the stable layer, and the wind component in the direction of the bore motion.

The first study credited with applying these ideas to cold fronts was Tepper (1950), who proposed that accelerating fronts could produce pressure jumps, features now interpreted as bores and gravity waves. Tepper (1950) noted that such pressure jumps might initiate convection in the warm sector. Haase and Smith (1984), Karyampudi et al. (1995), Koch and Clark (1999), and Locatelli et al. (2002b) have shown cases where a prefrontal wind shift or pressure trough occurred as a result of a bore moving ahead of a cold front along a prefrontal stable layer. Australian fronts moving into prefrontal stable layers sometimes produce bores known as morning glories (e.g., Reeder and Smith 1992, 1998; Smith et al. 1995; Deslandes et al. 1999).

The strength of these bores and gravity waves engenders some debate about their relative importance to observed weather. Ley and Peltier (1978) and Levy and Bretherton (1987) argued that gravity waves emitted during frontal collapse can produce a convergence maximum 75–125 km ahead of the surface cold front. In contrast, Garner (1989b) argued that such waves were too weak to produce as large a disturbance in the surface wind field as has been observed. Observationally, the impact of bores and gravity waves on producing deep, moist convection can vary. For example, Koch and Clark (1999) showed a squall line over Oklahoma where the lifting by a prefrontal bore was insufficient by itself to initiate convection—the combined lifting from the bore and a gravity current was required to initiate the convection. In contrast, Locatelli et al. (2002b) showed that a bore during the 3–4 April 1974 Super Tornado Outbreak was sufficient to initiate convection.

4. Discussion

In this section, we synthesize and expand on the results from this review. The first goal of this section is to highlight connections and contrasts between the various mechanisms in section 4a. In section 4b, research opportunities to improve understanding of prefrontal troughs and their associated cold fronts are presented. This goal is accomplished most effectively by closing some of the gaps in understanding that separate theoretical, observational, modeling, and diagnostic research approaches. The last goal is to look toward the future of improved understanding and forecasting of prefrontal troughs in section 4c through developing closer relationships between the research- and operational-meteorology communities.

a. Synthesis

The 10 mechanisms for prefrontal troughs and wind shifts that were presented in section 3 can be simplified even further by classifying the nature of the process that produces the prefrontal pressure trough or wind shift. As was noted earlier, a trough at the surface exists because the overlying atmospheric column is warmer than the adjacent locations. To warm the column, we consider the terms in the thermodynamic tendency equation (e.g., Bluestein 1992, p. 197): warm advection, descent, and diabatic heating. The overlying column can be warmed through warm advection, which is what occurs with the inhomogeneities mechanism (section

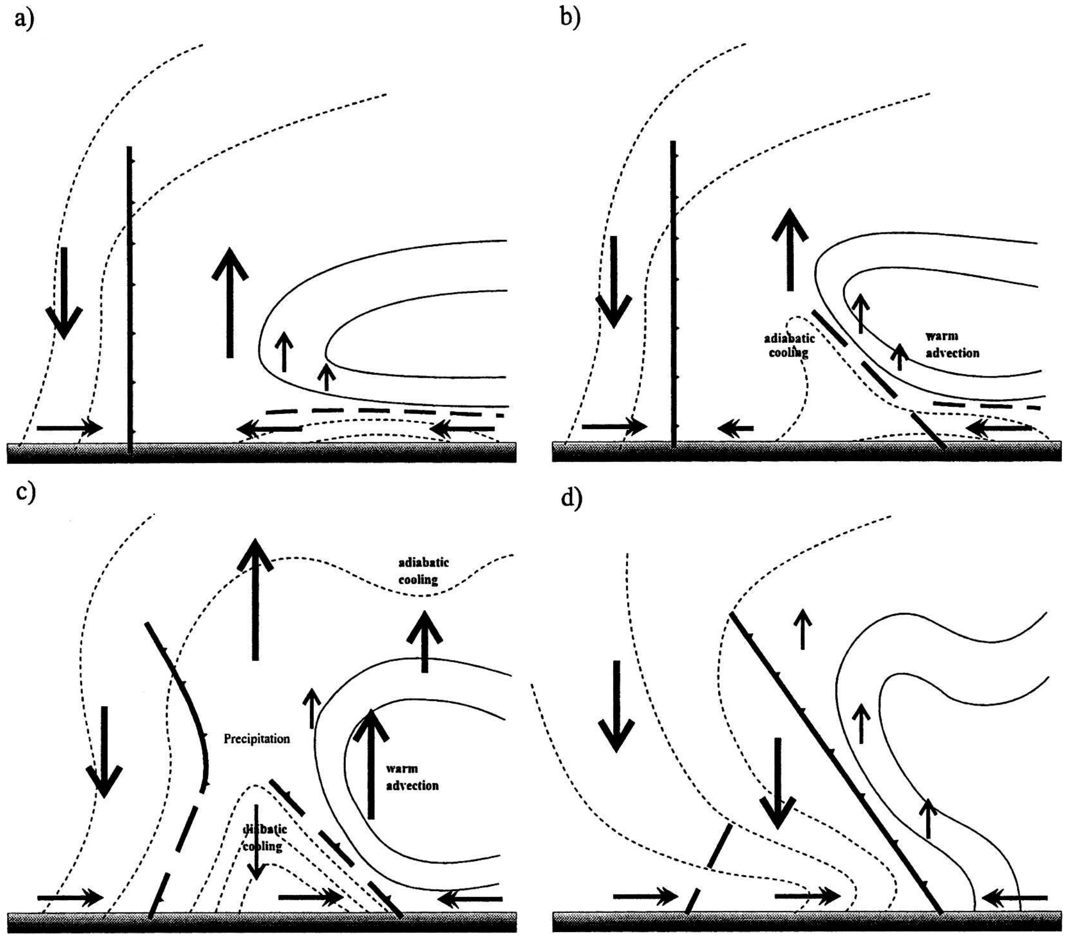

FIG. 16. Schematic representation of the lower-tropospheric two-dimensional structures associated with discrete frontal propagation. Temperature anomalies (solid positive, dashed negative) (Charney and Fritsch 1999, their Fig. 12).

3d) and the alongfront temperature advection mechanism (section 3f). The column can undergo descent, which is what occurs with the lee troughing mechanism (section 3b) and the descent mechanism (section 3h). Finally, warming the column can occur by diabatic processes, which has not been documented previously.

Alternatively, the trough can occur because adjacent surface pressure rises cause the resulting minimum locally. The synoptic-scale processes mechanism (section 3a) and upper-level fronts mechanism (section 3c) are a result of cold advection and ascent, producing column cooling. Cooling can also occur by ascent (section 3i). Mechanisms that rely on diabatic cooling include subcloud evaporation (section 3g) and the formation of the prefrontal stable layer discussed by Charney and Fritsch (1999) (section 3i).

Further similarities between these 10 mechanisms exist. One, in particular, is the presence of surface stable layers in the prefrontal air. In Charney and Fritsch (1999), this stable layer is tilted in the vertical to form a secondary baroclinic zone. In contrast, the cold front discussed by Bryan and Fritsch (2000a,b) does not tilt the stable layer, but moves over it, resulting in a prefrontal trough on its downstream side. Such a scenario is similar to frontogenesis occurring in the lee of mountains (e.g., Dickinson and Knight 1999) and over stable layers in the central United States (e.g., Neiman et al. 1998; Neiman and Wakimoto 1999).

As this review has demonstrated, Australian meteorologists have long been aware of the importance of prefrontal troughs associated with many of their cold fronts. Indeed, prefrontal troughs have been a significant forecasting problem for them, spawning heavy rains, convective storms, dust storms, dry gusty winds favoring wildfires, and even extratropical cyclones. That Australia is surrounded by oceans with sparse observations of these frontal systems before they come ashore limits understanding. For these reasons, Australians have concentrated on an active research program of observational, theoretical, modeling, and applied research. Their interest in prefrontal troughs has contributed to the identification of 4 of the 10 mechanisms cited in this review, as reviewed in the next paragraph.

First, variations in the prefrontal stability can produce prefrontal features (e.g., Reeder et al. 1991), as discussed in section 3d. Second, idealized simulations of the horizontal-shear-induced frontogenesis model with alongfront temperature gradients that resemble fronts in and around Australia (e.g., Reeder 1986; Reeder and Smith 1986, 1987) also produce a separation between the wind shift and the thermal gradient (section 3f). Third, subcloud evaporation of falling hydrometeors can lead to a prefrontal trough (e.g., Ryan et al. 1989), as discussed in section 3g. Finally, the morning glory cloud band is related to a bore moving along a prefrontal stable layer (section 3j). Thus, given the variety of different mechanisms and the frequency of frontal passages possessing such structures in and near Australia, that Australians have taken a prominent lead in research on prefrontal troughs is not surprising.

b. *Making connections within the research community*

Table 2 shows the mechanisms discussed in section 3 for the formation of prefrontal troughs and wind shifts. In addition, more mechanisms not listed in Table 2 may exist, either inadvertently omitted from this list or presently undiscovered. Table 2 also lists whether these mechanisms were identified from theoretical or observationally oriented research, showing that both theory and observations have been active in advancing our understanding of cold-frontal structure and dynamics, but both have contributed to a complete understanding of these mechanisms for only some of the mechanisms. That many of the mechanisms in Table 2 have only been identified from either theory or observations, but not both, suggests further research opportunities. In support of this admonition, others have also called for further research on frontal structures and dynamics, in general (e.g., Keyser 1986; Keyser and Pecnick 1987; National Research Council 1998, 82–83 and 175–177; Schultz 2005, manuscript submitted to Amer. Meteor. Soc. *Meteor. Monogr.*).

This review highlights several important directions that are worthy of exploration. Theoretical approaches that have been studied, but have not been observationally applied, include observational evaluation of the Cunningham and Keyser (1999) mechanism in deformation-induced fronts (section 3a), confirmation of the role of alongfront temperature gradients in producing prefrontal troughs (section 3f), and diagnosing the adiabatic cooling at the leading edge of a front (section 3i). This final study is important because of the potential role that adiabatic cooling might play in the removal of stable layers inhibiting deep, moist convection, especially over the central United States.

TABLE 2. Mechanisms for prefrontal troughs/wind shifts in the literature and whether they are primarily theoretically or observationally supported.

External to the front	Support
Synoptic-scale forcing	Theoretically and observationally
Interaction with lee troughs and drylines	Theoretically and observationally
Interaction with fronts in the mid- and upper troposphere	Observationally
Inhomogeneities in the prefrontal air	Theoretically
Internal to the front	**Support**
Surface friction	Theoretically and observationally
Alongfront temperature gradients	Theoretically
Moist processes	Theoretically and observationally
Descent of air	Observationally
Ascent of air at the front	Observationally
Generation of bores and gravity waves	Theoretically and observationally

Mechanisms that have been studied from an observational perspective, but require insight from more theoretical approaches, include a generalization beyond deformation- and shear-induced background flows to more realistic synoptic flows and their effect on frontal structure and evolution (section 3a), a theoretical foundation for the interaction between surface fronts and upper-level fronts (section 3c), an idealized modeling approach combining realistic synoptic flows with horizontally inhomogeneous prefrontal stability (section 3d), and understanding mechanisms producing prefrontal descent and their roles in prefrontal troughs (section 3h). Currently, we have only speculated on this last point—evidence to support this relationship does not exist.

Finally, very few of the above mechanisms have had frontal diagnostics, such as those described by Keyser et al. (1988) and Keyser (1999), applied to them. These mechanisms have been rarely related to dynamic frameworks, such as potential vorticity, quasigeostrophic, and semigeostrophic approaches.

c. *Making connections between research and operations*

The dichotomy between theory and observations is not the only chasm that our science must bridge. The

National Research Council (2000) terms the chasm between research and operations "The Valley of Death." Others have also noted the reality of this chasm (e.g., AMS 1952; Doswell et al. 1981; Doswell 1986; Stokes 1997; National Research Council 2003; Serafin et al. 2002). Understanding and recognizing these prefrontal features is of more than just academic interest for improving our understanding of cold fronts. For example, such prefrontal troughs may be one mechanism by which prefrontal squall lines (e.g., Fulks 1951; Newton 1950, 1963) may form. Forecasters at the National Oceanic and Atmospheric Administration (NOAA)/Storm Prediction Center and its predecessor organizations often have focused on these prefrontal features ("boundaries") as locations for the initiation of convection in a conditionally unstable atmosphere (e.g., House 1959; Sanders and Doswell 1995).

A second example is the summertime prefrontal troughs that pose crucial forecasting issues during the Australian wildfire season (e.g., Hanstrum et al. 1990a). The importance of the prefrontal troughs to fire-weather forecasting is related to their high prefrontal temperatures, strong dry winds, possible thunderstorms, and rapidly shifting wind direction. An unfortunate example was the Ash Wednesday bush fires of 16 February 1983 when 75 people were killed, 2545 buildings were destroyed, and over 390 000 ha were burned by 10 separate fires, exacerbated by the winds associated with a prefrontal trough (Bureau of Meteorology 1984; Mills 2005; Garratt 1988).

The literature is rife with alternative structures and evolutions of cold fronts that are often observed by operational forecasters and analysts, but have not been placed in a dynamical context. There are opportunities to expand the knowledge reviewed in this paper into the operational sector. Ultimately, this argument leads to the inevitable conclusion that forecaster training and manual analysis of the data are important to improved understanding of the atmosphere. Intuitive forecasters [i.e., forecasters who construct their conceptual understanding on the basis of dynamic visual images, as defined by Pliske et al. (2004)] are good at incorporating a variety of information into the hypothesis-formation and hypothesis-testing stages of forecasting (e.g., Roebber et al. 2004). Providing improved conceptual models of cold-frontal processes and dynamics leads to improved forecasting skill for intuitive forecasters. Consequently, operationally oriented research and effective forecaster education, along with an emphasis on weather-analysis skills, are required for the best forecasters to excel in their talents (e.g., Doswell et al. 1981; Bosart 2003; Doswell 2004).

5. Summary

This paper reviewed a number of different mechanisms for prefrontal troughs and wind shifts, illustrating the tremendous variety of physical processes acting in cold fronts, including some that may not be widely appreciated among the meteorological community. Ten different mechanisms were identified that have been discussed in the literature for the formation of prefrontal troughs and wind shifts (section 3). These mechanisms were classified into those that are external to the front and those internal to the front. Those mechanisms external to the front are associated with the environment of the front and with processes outside of the frontal circulations. Those mechanisms internal to the front are those processes associated with the cold front, its structure and circulation. Those 10 mechanisms are as follows:

- *Synoptic-scale forcing.* Synoptic-scale forcing in the form of mid- and upper-level short-wave troughs or translating axes of dilatation can produce prefrontal troughs due to surface pressure falls.
- *Interaction with lee troughs and drylines.* Prefrontal troughs and wind shifts may form in the lee of the mountains in the form of lee troughs and drylines as fronts traverse topography, separating the pressure trough/wind shift from the temperature gradient.
- *Interaction with fronts in the mid- and upper troposphere.* The merger or collocation of fronts in the mid- and upper levels with a surface cold front can produce a prefrontal pressure trough at the surface.
- *Frontogenesis associated with inhomogeneities in the prefrontal air*. Two different ways that inhomogeneities in the prefrontal air can lead to prefrontal features include preexisting frontal features ahead of the surface front and inhomogeneities in the prefrontal environment that develop frontal characteristics due to synoptic or frontal forcing.
- *Surface friction.* Because friction slows down the near-surface wind speed relative to that above the surface, cold advection is delayed at the surface, allowing for the development of a nose of cold advection overhanging the surface position of the front.
- *Frontogenesis acting on alongfront temperature gradients*. In the presence of alongfront thermal gradients, the surface pressure falls may propagate faster than the advective speed of the isotherms, thereby leading to a separation between the front and the prefrontal trough.
- *Moist processes*. The interaction between the dry dynamics of cold fronts and moisture can lead to prefrontal features in several ways. First, some idealized

models of frontogenesis in the presence of small moist symmetric stability produce the ascent 50–200 km ahead of the surface front. Second, subcloud evaporation of falling precipitation under forward-tilting clouds can lead to hydrostatic pressure rises, resulting in the lowest surface pressure being ahead of the front. Third, prefrontal stable layers laid down by previous convection can affect the movement and formation of cold fronts, leading to the development of prefrontal features.

- *Descent of air.* Prefrontal troughs can be formed by the adiabatic warming associated with prefrontal descent of air.
- *Ascent of air at the front.* In contrast to the previous mechanism, the ascent plume along a cold front can create significant adiabatic cooling and subsequent pressure rises at the surface, resulting in the formation of prefrontal troughs and wind shifts adjacent to these pressure rises. In addition, preexisting prefrontal stable layers can be tilted upright and develop frontal characteristics.
- *Generation of prefrontal bores and gravity waves.* Bores and gravity waves advancing on stable layers ahead of cold fronts can result in prefrontal troughs and wind shifts.

Many cold fronts behave similarly to extant conceptual models (e.g., wedge model, zero- and first-order discontinuities, Sawyer–Eliassen paradigm of secondary circulations). Yet, prefrontal troughs and wind shifts often occur and, when they do, deviate from the standard models of cold fronts. These anomalies from the standard models of cold fronts test the generality of our theories and conceptual models of the structure and dynamics of cold fronts. Some studies, although they possessed nonclassical features, have been interpreted as classical examples of cold fronts. For example, the cold front of Sanders (1955), long held up as a classical example of a cold front, recently has been shown to have a nonclassical prefrontal wind shift associated with it (Schultz 2004; SR).

Finally, although these 10 mechanisms have been either observed, theorized, or modeled using real or idealized initial conditions, complete multifaceted explanations for many of these mechanisms have not been forthcoming. This paper hopes to start a dialog on reconnecting theory, observation, and diagnosis to improve understanding of cold fronts and develop improved conceptual models.

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