Discrete Frontal Propagation over the Sierra–Cascade Mountains and Intermountain West

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(Manuscript received 30 September 2008, in final form 5 December 2008)

ABSTRACT

On 25 March 2006, a complex frontal system moved across the Sierra–Cascade Mountains and intensified rapidly over the Intermountain West where it produced one of the strongest cold-frontal passages observed in Salt Lake City, Utah, during the past 25 yr. Observational analyses and numerical simulations by the Weather Research and Forecast (WRF) Model illustrate that the frontal system propagated discretely across the Sierra–Cascade Mountains and western Nevada. This discrete propagation occurs in a synoptic environment that features a mobile upper-level cyclonic potential vorticity (PV) anomaly that is coupled initially with a landfalling Pacific cyclone and attendant occluded front. The eastward migration of the upper-level cyclonic PV anomaly ultimately encourages the development of a new surface-based cold front ahead of the landfalling occlusion as troughing, confluence, and convergence downstream of the Sierra Nevada intensify preexisting baroclinity over Nevada. Trajectories show that the new cold front represents a boundary between potentially warm air originating over the desert Southwest, some of which has been deflected around the south end of the high sierra, and potentially cool air that has traversed the sierra near and north of Lake Tahoe, some of which has been deflected around the north end of the high sierra. Although diabatic processes contribute to the frontal sharpening, they are not needed for the discrete propagation or rapid cold-frontal development. Forecasters should be vigilant for discrete frontal propagation in similar synoptic situations and recognize that moist convection or differential surface heating can contribute to but are not necessary for rapid Intermountain West frontogenesis.

1. Introduction

Rapidly developing cold fronts produce strong winds and dramatic temperature falls several times each year over the Intermountain West (Shafer and Steenburgh 2008). A particularly remarkable event occurred on 25 March 2006 when temperatures plunged 8°C in 15 min and 14°C in 2 h with a cold-frontal passage at the Salt Lake City International Airport (KSLC). The cold-frontal temperature change and accompanying 6-hPa pressure rise would rate as the fifth and third largest at KSLC in the 25-yr cold-front and pressure change climatologies presented by Shafer and Steenburgh (2008) and Koppel et al. (2000). The frontal system also moved from the northern California coast to KSLC in just 16 h [an apparent mean speed of \( \sim 60 \text{ km h}^{-1} \) \((16.7 \text{ m s}^{-1})\)] and intensified rapidly over Nevada where the temperature gradient increased from 3.5° to 13°C(100 km)\(^{-1}\) in less than 10 h. Over much of northern Utah, postfrontal wind gusts exceeded 35 m s\(^{-1}\).

The large apparent frontal system speed suggests that it may have moved discretely across the Sierra–Cascade Mountains and Intermountain West. Bryan and Fritsch (2000a) define discrete frontal propagation as “a process whereby a midlevel front moves through an area aloft without evidence of a front moving through the same area at the surface, while the surface frontal properties (e.g., pressure trough, wind shift, and thermal gradient) dissipate at one location and simultaneously develop at another location.” Recognition and understanding of discrete frontal propagation is important for
operational weather forecasting, particularly over the Intermountain West where cold fronts are frequently accompanied by severe winds and surface frontal analysis and prediction is complicated by the region’s complex topography (Shafer and Steenburgh 2008).

Several mechanisms may contribute to discrete frontal propagation. Charney and Fritsch (1999) describe how differential vertical motion acting on a prefrontal temperature inversion initiated discrete cold-frontal propagation over the southeast United States. Diabatic effects associated with precipitation amplified the development of the new surface front, but were not necessary for the discrete propagation. Hoskins et al. (1984) show that baroclinic development and frontogenesis in a nonuniform environment can lead to the formation of multiple surface fronts, a mechanism that Bryan and Fritsch (2000b) suggest can play a role in discrete frontal propagation. Some authors interpret the formation of a new surface front within a prefrontal trough or wind shift as discrete frontal propagation (Bryan and Fritsch 2000b; Schultz 2005). For example, Physick (1988) used idealized simulations to examine the evolution of a cold front approaching a coastline. In his simulations, the diurnal temperature rise along the coast produces a prefrontal surface trough that becomes frontogenetical and produces a new front, making it appear as though the original surface front accelerated rapidly toward the coastline. Hanstrum et al. (1990a,b) observed the formation of a new cold front within a prefrontal trough as the preexisting cold front undergoes frontolysis. Their climatology suggests that such an evolution is common during the summer months along the southern coast of Australia.

Bryan and Fritsch (2000a,b) present an example of discrete frontal propagation produced by prefrontal convection in the central United States. The original surface-based cold front stalls and weakens as it encounters a convectively induced cold pool, whereas the midlevel baroclinic zone moves continuously downstream. The midlevel baroclinic zone eventually couples with a new low-level cold front that forms on the downstream side of the cold pool. As a result, there is no surface-based cold-frontal passage within the cold-pool region. Daytime heating in the prefrontal boundary layer and cooling associated with thunderstorms initiates the development of the new front. Diabatic effects also appear to be responsible for a case of discrete frontal propagation examined by Kurz (1990) along the northern flank of the Alps.

Observational and theoretical studies show that frontal speed and intensity may be strongly affected by a mountain barrier (e.g., Mattocks and Bleck 1986; Blumen 1992; Egger and Hoinka 1992; Hoinka and Volkert 1992). In particular, Keuler et al. (1992) show that a surface front may move discretely across a mountain barrier with the original surface front remaining trapped on the windward slope while a new front forms in the lee. This discrete evolution results from nonlinear interactions between the frontal system and the mountain wave. Dickinson and Knight (1999) examine similar front–mountain interactions using primitive equation simulations of the Eady wave (Eady 1949). In simulations with taller and steeper mountains, the upper-level cyclonic potential vorticity (PV) anomaly moves relatively unimpeded over the barrier, whereas the surface front (defined using vorticity) is blocked. As it moves downstream, the upper-level cyclonic PV anomaly couples with a lee trough, which becomes the new surface front. For low, broad mountains, however, the degree of blocking is weak and the surface front moves continuously across the barrier. Frontal intensity also plays a role, with discrete evolution favored with weak fronts, which are more easily disrupted by the mountain-induced circulations. Schumacher et al. (1996) suggest that the strength of the upper-level PV anomaly also affects how much a surface-based front is retarded by orography.

The Intermountain West is a region where orographic and diabatic processes strongly influence frontal evolution, although knowledge of these local effects is just beginning to be elucidated (e.g., Horel and Gibson 1994; Steenburgh and Blazek 2001; Schultz and Trapp 2003; Shafer et al. 2006; Shafer and Steenburgh 2008). Here we document the discrete propagation of the 25 March 2006 frontal system across the Sierra–Cascade Mountains and Intermountain West and identify the processes responsible for the discrete propagation and the rapid formation of the new surface-based cold front, including the role of the region’s uniquely complex orography.

2. Data and methods

a. Observational datasets and analysis techniques

Our analysis uses upper-air analyses from the North American Mesoscale (NAM) Model, satellite and radar imagery, radiosonde observations, and manual surface analyses. At the time of the event, the NAM was run operationally at the National Centers for Environmental Prediction (NCEP) with 12-km horizontal grid spacing and 60 vertical levels, but obtained with 40-km horizontal and 25-hPa vertical grid spacing. The manual surface analyses use high-density surface observations from the MesoWest cooperative networks (Horel et al. 2002). These observations were quality controlled based on manual spatial and temporal consistency checks.
while performing the surface analysis, combined with an
inspection of MesoWest data quality ratings, which are
based on a comparison of observed values to an es-
timate based on multivariate linear regression (Splitt and
Horel 1998). Data quality was quite high, with only a
few observations not considered at each analysis time.
Although data from more than 2500 stations were used,
only observations provided by National Weather Service
(NWS) stations and/or Remote Automated Weather
Stations (RAWS) are plotted on the surface analyses
presented here to avoid clutter.

Large-scale surface analyses include a sea level pres-
sure analysis, which is appropriate over the Pacific Ocean
and regions upstream of the Sierra Nevada, whereas
regional surface analyses over northern California and the
Intermountain West include a 1500-m pressure
analysis. The latter derives from station observations
of altimeter setting following Steenburgh and Blazek
(2001). Frontal analyses use the conventional definition
of a front described by Bluestein (1986) and Keyser
(1986) wherein a front occupies the boundary on the
warm side of an elongated horizontal baroclinic zone, or
in the case of an occluded front, the axis of maximum
temperature. Important wind shifts or pressure troughs
without a sharp temperature transition are analyzed as
troughs. Unique analysis challenges and ambiguities pre-
sent in this case are discussed where appropriate in the
paper.

b. Numerical simulations

The observational analysis is supplemented with
two simulations by the Advanced Research Weather
Research and Forecast (WRF) Model, version 2.2
(Skamarock et al. 2005), which is based on a hydrostatic,
pressure-based, terrain-following \( \eta \) coordinate. The con-
trol run (CTL) is a full physics simulation. The “fake-dry”
run (FKDRY) examines the influence of moist-diabatic
processes on the discrete propagation and is identical to
CTL except that it does not include a cumulus parame-
terization or diabatic heating and cooling produced by the
explicit moisture scheme. FKDRY does, however, allow
simulated clouds and precipitation to interact with other
model physics packages, such as the radiation scheme.
Both simulations feature 36- and 12-km domains with 34
half-\( \eta \) levels in the vertical. Only output from the 12-km
domain is presented. Major physics packages include the
Rapid Radiative Transfer Model (RRTM) longwave
radiation scheme (Mlawer et al. 1997), the Dudhia
shortwave radiation scheme (Dudhia 1989), the Noah
land surface model (Chen and Dudhia 2001), the Mel-
or–Yamada–Janji\’c planetary boundary layer parame-
terization (Mellor and Yamada 1982; Janji\’c 2002), the
new Kain–Fritsch cumulus parameterization (CTL
only), which is a modified version of the scheme de-
scribed by Kain and Fritsch (1990, 1993), and Thompson

NAM Model analyses provide the cold-start atmo-
spheric and land surface (e.g., soil temperature, soil
moisture, snow cover) initial conditions at 0000 UTC 25
March 2006 and lateral boundary conditions through
the integration period. Some modifications were made,
however, to the NAM Model analyses and interpolation
algorithms used by the WRF, to produce an initial
analysis that more closely matched reality. The NAM
provides analyses for pressure levels below the NAM
surface based on an assumed lapse rate that is more
stable than dry adiabatic. The WRF typically uses this
subsurface data, along with the NAM surface analysis,
to interpolate vertically to the WRF grid. This intro-
duces, however, spurious surface-based stable layers
wherever the WRF terrain is lower than that of the
NAM analysis. Since a deep convective boundary layer
was present over the Intermountain West at the initial
time, we instead assumed well-mixed conditions at all
WRF levels below the lowest above-ground analysis
level provided by the NAM. This produced an initial
analysis that more closely matched observed sound-
ings.

Inspection of the NAM analysis also revealed excess-
ive snow cover, anomalously low ground temperatures,
and a spurious low-level cold pool over north-central
Nevada. The analysis in this region was adjusted by
removing snow cover from all but the highest elevation
regions, increasing the skin and first-layer ground tem-
pératures, and assuming well-mixed conditions below
750 hPa, consistent with satellite imagery, surface ob-
servations, and/or the Elko, Nevada (KEKO), sound-
ing. The resulting initial analysis more closely matched
reality, but had little impact on the simulation after
2–3 h of integration.

For figure clarity all horizontal scalar variable analy-
ses (e.g., geopotential height, temperature, potential
temperature, and frontogenesis) are smoothed using a
7-point cowbell spectral filter (Barnes et al. 1996). This
small grid increment enables one to identify the large-
scale signal without eliminating important mesoscale
terrain effects.

3. Observed evolution

The antecedent large-scale pattern featured a broad
upper-level trough centered off the west coast of North
America (near 140°W), with upper-level ridging down-
stream over the interior western United States (not
shown). In advance of the trough, a low-level baroclinic
zone extended from the eastern Pacific Ocean across northern California, northwest Nevada, and Oregon. On 24 March 2006, a shortwave trough and associated cyclonic PV anomaly entered the upstream side of the upper-level trough, initiating surface cyclogenesis over the eastern Pacific poleward of the low-level baroclinic zone and compaction of the upper-level trough, which became increasingly mobile (not shown).

By 0600 UTC 25 March, the upper-level trough axis and cyclonic PV anomaly (indicated by a dynamic tropopause pressure maximum) are upstream of the northern California coast (Fig. 1a). At 700 hPa, the low-level baroclinic zone extends from the eastern Pacific across northern California, northwest Nevada, and Oregon (Fig. 1b), but lacks a sharp temperature transition characteristic of a front at the surface. Consistent with cyclogenesis in a polar airstream (e.g., Reed 1979), the Pacific low center and attendant occluded front are found poleward of this low-level baroclinic zone (Fig. 1c), with a well-defined comma cloud evident in satellite imagery (Fig. 1d). Accompanying the occluded front is a weak “trough of warm air aloft” (trowal; Penner 1955; Galloway 1958; Martin 1998; Grim et al. 2007), which at 700 hPa extends southeast from the circulation center to the baroclinic zone. The surface front and accompanying trowal formed within the polar airstream rather than from a cold front overtaking a warm front as in the classical model. Similar frontal structures are sometimes referred to as an “instant occlusion” (Anderson et al. 1969; Locatelli et al. 1982; Schultz and Mass 1993), although in the present case the comma cloud and occluded front never merge with a frontal wave or cloud band. Surface features also include a weak trough upstream of the northern and central California coast, a trough in the lee of the Sierra Nevada, and a trough that extends northeastward across Oregon.

The surface occluded front passes KACV on the northern California coast (see Fig. 1d for location of this and other time series stations) at 1130 UTC, and is accompanied by a 4°C temperature fall, a broad pressure minimum, a wind speed maximum, a shift from southeasterly to south-southwesterly flow, and a frontal rainband (Figs. 2a and 3a). Based on radar imagery animation from the Eureka, California, radar (KBBX,
located near KACV), the occluded front appeared to be moving eastward at \( \sim 45 \text{ km h}^{-1} \) (12.5 m s\(^{-1}\)) prior to and during landfall.

The upper-level cyclonic PV anomaly, 700-hPa circulation center, and 700-hPa trough are approaching the Pacific coast by 1200 UTC 25 March (Fig. 4a,b). The surface low center and comma cloud are just upstream of the Pacific coast, with the surface occluded front progressing inland over northern California (Figs. 4c,d). Meanwhile, the broad 700-hPa baroclinic zone remains quasistationary over northern California, northwest Nevada, and southern Oregon. The 700-hPa flow in this region is nearly parallel to the isotherms, with orographic lift and large-scale ascent over the subsequent

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**Fig. 2.** Meteograms of temperature (solid), altimeter setting [in (a)–(e)], station pressure [in (f)] (dashed), and wind (full and half barb denote 5 and 2.5 m s\(^{-1}\), respectively) at (a) KACV, (b) KSIY, (c) KAAT, (d) KWMC, (e) KEKO, and (f) DPG17. Occluded or incipient cold-frontal passage is denoted by solid vertical line except in (b) where the frontal passage was diffuse. Trough A is identified by the dashed vertical line in (d). Barb density is reduced in (f) for clarity. See Fig. 1d for station locations.

**Fig. 3.** Lowest-elevation scan (0.5°) radar reflectivity composite at (a) 1200 UTC 25 Mar, (b) 1800 UTC 25 Mar, and (c) 0000 UTC 26 Mar. Reflectivity scale (dBZ) is at right of (a). The terrain is gray shaded.
6-h period producing higher relative humidity over the Sierra Nevada and Cascade Mountains of California (Fig. 4b). With the leading edge of the upper-level cyclonic PV anomaly pushing inland, a new low center is forming over south-central Oregon (Fig. 4c).

The occluded front weakens as it moves through southern Oregon and northern California. Many interior stations, such as KSIY, observe a broad pressure minimum, but temperature and wind changes are too small and variable to enable an unambiguous frontal analysis from surface data alone (Fig. 2b, see Fig. 4d for the location). Best estimates based on radar, satellite, and surface observations suggest the dissipating front moves across northern California at \( \frac{40}{2} \text{ km h}^{-1} \) (11.1 m s\(^{-1}\)), only slightly slower than the \( \frac{45}{2} \text{ km h}^{-1} \) (12.5 m s\(^{-1}\)) at landfall (Fig. 5a) and close to the speed of the leading edge of the upper-level cyclonic PV anomaly (Fig. 5c).

By 1800 UTC 25 March 2006, the upper-level trough and cyclonic PV anomaly are located over the northern California coastline (Fig. 6a). These features have compacted and become increasingly mobile and negatively tilted in the previous 6 h (cf. Figs. 4a and 6a). Meanwhile, the Pacific low center fills, while the sea level and 1500-m low pressure centers over southern Oregon deepen (Figs. 6c and 7a). Concurrently, the 700-hPa baroclinic zone over northwestern Nevada intensifies as the 700-hPa winds veer slightly and weak cold advection develops over northeastern California and northwestern Nevada (Fig. 6b). Convective clouds and weak, scattered radar echoes also develop over north-central Nevada (Figs. 3b and 6d).

The fate of the surface occlusion becomes increasingly unclear over northern California during this period. The 1800 UTC 25 March mesoscale analysis reveals a trough and wind shift (hereafter trough A) trailing the developing low center over extreme northeast California and southern Oregon (Fig. 7a). At KAAT in northeast California temperatures drop \( \frac{55}{2} \text{ C} \) and light rain and snow fall following the passage of trough A (Fig. 2c). Radar echoes in this region are weak and limited in area (Fig. 3b), but coverage in this region is very poor (e.g., Westrick et al. 1999). Given the data limitations, it is unclear if trough A represents the surface occlusion or a new trough that has developed from moist-diabatic processes. Using the timing of trough passage at KAAT and other stations yields an apparent speed of \( \frac{53}{2} \text{ km h}^{-1} \) (14.7 m s\(^{-1}\)) over northeastern California, somewhat faster than observed upstream (Fig. 5a). Based on this analysis we conclude that one cannot unambiguously trace the surface occlusion across northern California, but that its remnants are either coincident with or behind trough A.

![Figure 4](image-url)
By 1900 UTC 25 March another surface trough, labeled B, forms downstream over northern Nevada, well in advance of trough A (Fig. 7b). Frontogenesis occurs roughly coincident with trough B, which intensifies into a cold front as it progresses eastward across northern Nevada (Fig. 7c). As the incipient cold front passes through KWMC at 2015 UTC, it produces a temperature fall of 6°C in 1 h. After this time, trough A becomes indistinguishable.

By 0000 UTC 26 March the upper-level trough and cyclonic PV anomaly are over the California–Nevada border (Fig. 8a) and a fully developed Intermountain cyclone is centered over Idaho with the trailing cold front extending south-southwestward across central Nevada (Fig. 8c). Strong confluenex exists downstream of the Sierra Nevada over central Nevada at 700 hPa with the axis of dilatation aligned with the isotherms in the developing baroclinic zone (Fig. 8b), a configuration that is optimal for confluent frontogenesis (Martin 2006, 199–201) and has been implicated in previous studies of Intermountain frontogenesis (Shafer and Steenburgh 2008). Mushrooming cold cloud tops indicative of developing convection are visible along and behind the leading edge of the cold front with more limited cloud cover in the prefrontal environment (Fig. 8d). Radar imagery indicates scattered precipitation along and behind the cold front with some cellular reflectivity maxima exceeding 35 dBZ (Fig. 3c).

Differential daytime surface heating arising from inhomogeneous cloud cover and low-level diabatic cooling produced by postfrontal moist convection can be favorable for frontal intensification (Koch et al. 1995; Gallus and Segal 1999; Schultz and Trapp 2003; Segal et al. 2004). Soundings from KSLC and KEKO, the latter likely taken moments before frontal passage, show a deep convective boundary layer, indicative of strong prefrontal surface heating and favorable for strong diabatic cooling and the production of strong downdrafts (Fig. 9), as observed by Schultz and Trapp (2003) in another Intermountain cold-front event. Comparison of meteograms from KWMC, KEKO, and DPG17 reveals dramatic frontal sharpening as the front moves across Nevada and northwest Utah (cf. Figs. 2d–f) where pressure rises of 2–6 hPa h\(^{-1}\), abrupt wind shifts, and temperature falls of 10°–14°C h\(^{-1}\) accompany frontal passage. During this period the average speed of the front was 56–65 km h\(^{-1}\) (15.6–18.1 m s\(^{-1}\); Fig. 5a).

Figure 10 summarizes the discrete nature of the frontal propagation. As a surface-based occluded front makes landfall and moves inland across northern California, the downstream movement of the accompanying upper-level trough and cyclonic PV anomaly encourages cyclogenesis and the downstream development of a new surface-based cold front along a preexisting baroclinic zone over northern Nevada. This new cold front intensifies rapidly and becomes the dominant surface frontal feature as the occluded front weakens over the Sierra Nevada and western Nevada. Confluence to the lee of the Sierra Nevada, differential surface heating associated with inhomogeneous cloud cover, and diabatic cooling

![Fig. 5.](image-url)
associated with developing precipitation and convective downdrafts, appear to contribute to the frontogenesis. We now use WRF simulations to further examine the processes contributing to the discrete propagation and frontal development.

4. WRF diagnostic analysis

a. Large-scale evolution and validation

Consistent with the observed analysis, the CTL forecast valid at 1200 UTC 25 March 2006 features a 700-hPa circulation center, a trough of warm air aloft (trowal; identified with occluded front symbols), and a well-developed comma cloud just off the northern California coast (Fig. 11a). Farther east, a broad baroclinic zone extends from the eastern Pacific across northern California and northwest Nevada. At 850 hPa,1 a weak trough (labeled T1) and precipitation are embedded within this baroclinic zone along the California–Nevada border (Figs. 11b,c). Mesoscale ridging and the poleward deflection of the incident southwesterly flow occur windward of the Sierra Nevada, with troughing to the lee (Fig. 11b). The flow deflection is most pronounced around the high sierra (elevation >2250 m). Farther north, where the sierra is lower, southwesterly flow moves relatively unimpeded over Lake Tahoe and extends into western Nevada, whereas south of the sierra the flow accelerates across the Mohave Desert, turns cyclonically, and merges with southeasterly flow over southern Nevada. A wind shift (identified with a solid line in Figs. 11b,c) extends northeastward away from the Sierra Nevada where these two airstreams merge. The simulated windward ridge and lee trough compare well with observations, but the wind shift does not appear to have an observational analog (Fig. 4c). Instead, the observed surface winds over southern Nevada are light and variable, and there is no evidence of southerly flow penetrating across the sierra north of Lake Tahoe. These discrepancies may reflect the inability of the WRF to fully develop nocturnal cold pools within valleys and basins (e.g., Cheng and Steenburgh 2005) and fully resolve the crest height of the sierra at 12-km grid spacing.

1 The 850-hPa level is near or just below altitude of the basins and valleys of northern Nevada. The WRF 850-hPa analyses presented in this paper extrapolate (for geopotential height) or use the lowest half-$\eta$ level analysis (for wind and potential temperature) where the terrain is above 850 hPa.

FIG. 6. As in Fig. 1, but for 1800 UTC 25 Mar 2006.
vertically stacked occlusion and accompanying cyclonic wind shift (Fig. 11d). Two towers of locally high PV, presumably generated by diabatic processes, are found along and immediately behind the occlusion. These PV towers lie beneath the strong horizontal gradient in PV found at the leading edge of the upper-level cyclonic PV anomaly. Farther to the east, mountain waves are evident over and to the lee of the coastal range of California and the Sierra Nevada. A weak but broad baroclinic zone with locally high PV extends from Nevada westward and upward to the leading edge of the upper-level cyclonic PV anomaly. A nocturnal inversion lies over much of Nevada and Utah, but is weaker than observed.

The simulated occlusion weakens as it moves eastward with its remnants found over interior northern California at 1500 UTC 25 March (Figs. 12a,c). Because of the weakening thermodynamic signature and masking effect of mountain waves induced by the coast range, the frontal analysis at this and subsequent times is based primarily on the position of the accompanying PV tower and cyclonic wind shift. Farther downstream, trough T1 has migrated eastward and developed a positive tilt while a second trough, T2, develops in the lowland regions of northwestern Nevada. These troughs appear to be roughly analogous to observed troughs A and B, although their formation occurs earlier than observed and there are differences in location and orientation (cf. Figs. 12b,c and 7b). Simulated precipitation during this period falls near and poleward of the Nevada–Oregon border (Fig. 12c), consistent with radar reflectivity observations (not shown). Farther to the south, the flow, pressure, and thermodynamic patterns remain similar to 1200 UTC. Cross-section XY shows that the occluded front is now over the coast range, while farther downstream, some strengthening of the low-level baroclinic zone has occurred over western Nevada (Fig. 12d).

As the upper-level cyclonic PV anomaly progresses eastward, the Pacific low center weakens as troughing intensifies over Nevada (Fig. 13b), southeast of the observed 1500-m low center (Fig. 6b). By 1800 UTC the remnants of the occluded front are moving into the Sierra Nevada and Cascade Mountains of northern California (Figs. 13a–c). The T1 has dissipated while T2 has intensified, migrated eastward, and merged with the wind shift that extended across central Nevada, with the incipient cold front forming along the resulting confluence zone (Figs. 13b,c). The position of the trough and strength of the accompanying confluence at this time (1800 UTC) compares well with the 2100 UTC analysis (Fig. 7c). Precipitation continues both ahead of and behind the cold front near the Nevada border, as observed in radar reflectivity analyses (cf. Figs. 3b and 13c).
Cross-section XY reveals further intensification of both the cold frontal temperature gradient and accompanying cyclonic wind shift ahead of the decaying occlusion, destruction of the nocturnal surface-based stable layer, and the development of a convective boundary layer (Fig. 13d).

By 2100 UTC the discrete propagation is complete (Fig. 14). The Pacific low is now an open wave, the occlusion is no longer distinguishable, and the Nevada cold front is now the dominant surface front of the system (Figs. 14a,b). Precipitation is now entirely postfrontal, consistent with radar imagery (cf. Figs. 14c and 3b,c). Cross-section XY shows that the surface-based cold front is structurally continuous with deep tropospheric baroclinity that extends to the upper-level cyclonic PV anomaly (Fig. 14d). The convective boundary layer has deepened, particularly in the prefrontal environment, but is shallower than observed (cf. Figs. 9 and 14b). Prefrontal temperatures within the convective boundary layer are also 2°–3°C less than observed (not shown).

The analysis just described illustrates that the CTL simulation replicates the discrete frontal propagation in this case. As observed, the simulated occluded front weakens and decays over northern California as the upper-level cyclonic PV anomaly moves downstream and a new surface-based cold front forms along preexisting baroclinity over Nevada. The discrete propagation is well captured, albeit with the frontal development over Nevada occurring 2–3 h too early. The weak wind shift found over central Nevada early in the simulation does not have an observational analog, but does not appear essential for the frontal development. Underdevelopment of the simulated prefrontal boundary layer did not prevent an intense front from forming, but resulted in less frontal sharpening than observed.

b. Trajectory analysis

Three-dimensional trajectories from CTL confirm the discrete propagation, illustrate the fate of the postocclusion air mass, and help to reveal the processes contributing to the rapid frontal development over Nevada. The trajectory calculations follow Petterssen (1956, p. 27) and Seaman (1987), use three-dimensional grid-scale wind fields obtained from 15-min model output, and are constrained to remain on the lowest-half-$\eta$ level if they approach the model surface. To examine the volume of air encompassing the cold front at 2100 UTC, we examined thousands of trajectories, but for brevity present 9-h (1200–2100 UTC 25 March) trajectories that terminate along cross section XY at 850 hPa (or the lowest half-$\eta$ level if the terrain extends above 850 hPa) and 700 hPa.
There are three main families of trajectories that terminate at 850 hPa around the Nevada cold front. Trajectories identified with triangles at their origins begin behind the occluded front and terminate over northern California several hundred kilometers behind the incipient cold front (Fig. 15a). Net ascent (120–130 hPa) and a modest (~3 K) increase in potential temperature occurs along these trajectories, the latter consistent with daytime warming (e.g., trajectory 2; Figs. 15b,c).

Trajectories identified with squares at their origins begin in the broad preocclusion baroclinic zone that extends across northern California and northwestern Nevada (Fig. 15a, see Fig. 11 for 1200 UTC potential temperature analysis). These trajectories terminate behind the Nevada cold front, indicating that the frontal zone forms at least in part from frontogenesis within the preocclusion baroclinic zone. The westernmost of these trajectories are thrust over the Sierra Nevada, experiencing net vertical displacement (e.g., trajectory 7; Fig. 15b), whereas the vertical displacement of trajectories beginning over and downstream of the Sierra is smaller. All trajectories in this family experience warming from daytime heating, most by 5–9 K. For example, the potential temperature along trajectory 7 increases nearly 9 K after 1700 UTC (Fig. 15c).

Trajectories with crosses at their origins begin over the deserts of southern Nevada and form the prefrontal air mass at 2100 UTC (Fig. 15a). The potential temperature along these trajectories also increases (e.g., trajectory 12; Fig. 15c). Similar net change in potential temperature along trajectories 7 and 12 suggests that nearly all of the model surface frontogenesis along cross-section XY is due to the confluence and convergence of these trajectories and not differential diabatic heating, a hypothesis that is further supported in section 4c. Both trajectories begin with a potential temperature contrast of 10 K, experience similar diabatic warming, and end with a potential temperature contrast of 10 K. As noted previously, however, heating in the model prefrontal environment (i.e., along trajectory 12) appears to be underdone.


FIG. 10. Isochrones of the occluded front (solid with occlusion symbols), decaying occluded front (dashed with occlusion symbols), trough A (dashed), and the incipient cold front (solid with cold frontal symbols) between 1100 and 2300 UTC 25 Mar 2006. Topography is shaded.
Four trajectory families terminate at 700 hPa around the Nevada cold front (Fig. 16). As found at 850 hPa, one trajectory family (also marked with triangles at their origins) originates behind the occluded front (Fig. 16a). These trajectories experience net ascent and penetrate nearly to the California–Nevada border. Changes in potential temperature along these trajectories are small (e.g., trajectory 3; Fig. 16c).

The second trajectory family originates in the portion of the preocclusion baroclinic zone that extends over the eastern Pacific and, consistent with the trajectories that end at 850 hPa, is identified with squares. Collectively, trajectories 1–7 suggest that the postocclusion air mass has penetrated to near or just past the California–Nevada border, well rearward of the incipient cold front.

Trajectories in the third family, identified with diamonds, begin at low levels (below 900 hPa) farther to the east. These trajectories undergo ascent over the windward slopes of the sierra where they also experience diabatic warming within the orographic cloud (e.g., trajectory 9; Fig. 16b). Trajectory 10 is somewhat unique and originates within the blocked air mass windward of the Sierra Nevada (see Fig. 11c). The final family of trajectories, denoted with crosses, originates over the Mohave Desert of Nevada and California (trajectories 11 and 12) or upstream of the Sierra Nevada (trajectory 13). These trajectories experience net ascent and increase their potential temperature, the latter less than observed along near-surface trajectories. The entrainment of trajectory 11 into the frontal zone may reflect the accuracy limitations of the trajectory calculation or simply that the view of the cold front as a material surface is an overidealization, particularly at this level. Collectively, trajectories 10–13 show that the incipient cold front represents an important break between potentially warm air originating over the desert Southwest, some of which has been deflected around the south end of the high sierra, and potentially cool air that has traversed the sierra near and north of Lake Tahoe, some of which has been deflected around the north end of the high sierra.
FIG. 12. As in Fig. 11, but for 1500 UTC 25 Mar 2006.

FIG. 13. As in Fig. 11, but for 1800 UTC 25 Mar 2006.
These trajectories illustrate that the Nevada cold front is a new feature that has formed well in advance of the occluded front and postocclusion airmass. The cold front also represents a boundary between potentially warm air originating over the desert Southwest, some of which has been deflected around the south end of the high sierra, and potentially cool air of Pacific origin that has traversed the lower sierra crest near and north of Lake Tahoe or has been blocked and deflected northward around the high sierra. Changes in potential temperature along the trajectories that terminate at 850 hPa are relatively similar, suggesting that differential diabatic heating is not a major contributor to the model frontogenesis, at least along this cross section. This indicates that differential diabatic heating is not essential to the discrete frontal propagation, although it may have contributed to sharpening of the observed cold front since prefrontal warming was underdone in the model simulation. At 700 hPa, the most pronounced diabatic effect along this cross section is warming within the orographic cloud over the Sierra Nevada, which appears to have a frontolytic effect by warming the postfrontal air mass.

c. Frontogenesis diagnostics

Another perspective on the Nevada frontal development is provided by the surface frontogenetical function, defined following Petterssen (1936) and Miller (1948) as

$$F = \frac{d}{dt} |\nabla_{\eta} \theta|,$$

where

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

$$\nabla_{\eta} = i \frac{\partial}{\partial x_{\eta}} + j \frac{\partial}{\partial y_{\eta}},$$

the subscript $\eta$ denotes differentiation along the lowest half-$\eta$ surface, and $\eta$ is the $\eta$-coordinate vertical velocity. Following Miller (1948), Eq. (1) may be written as

$$F = F_W + F_T + F_D,$$

where

$$F_W = \frac{1}{|\nabla_{\eta} \theta|} \left[ \frac{\partial \theta}{\partial x} \left( \frac{\partial u}{\partial x} \frac{\partial \theta}{\partial y} + \frac{\partial \theta}{\partial y} \frac{\partial u}{\partial y} \right) + \frac{\partial \theta}{\partial y} \left( \frac{\partial u}{\partial x} \frac{\partial \theta}{\partial x} + \frac{\partial \theta}{\partial x} \frac{\partial u}{\partial x} \right) \right],$$

FIG. 14. As in Fig. 11, but for 2100 UTC 25 Mar 2006.
and the subscript $\eta$ has been dropped for convenience. Here $F_W$, $F_T$, and $F_D$ are the frontogenesis components produced by horizontal deformation and divergence (hereafter kinematic frontogenesis), tilting, and horizontal gradients in diabatic heating and cooling, respectively.

\[
F_T = -\frac{1}{\nabla_{\eta}\theta} \left[ \frac{\partial}{\partial \eta} \left( \frac{\partial \eta}{\partial x} \frac{\partial \theta}{\partial x} + \frac{\partial \eta}{\partial y} \frac{\partial \theta}{\partial y} \right) \right], \tag{6}
\]

\[
F_D = -\frac{1}{\nabla_{\eta}\theta} \left[ \frac{\partial}{\partial x} \frac{\partial \theta}{\partial x} \frac{\partial \theta}{\partial t} - \frac{\partial}{\partial y} \frac{\partial \theta}{\partial y} \frac{\partial \theta}{\partial t} \right], \tag{7}
\]

Although $F_T$ is nonzero because of the presence of a stable layer in the morning and a superadiabatic layer in the afternoon, it does not appear to contribute significantly to the observed frontal development and is not presented. We break $F_D$ into two components:

\[
F_D = F_M + F_{BL}, \tag{8}
\]

where $F_M$ is the diabatic frontogenesis produced by the WRF cloud microphysics and cumulus parameterizations (hereafter moist frontogenesis) and $F_{BL}$ is the diabatic frontogenesis produced by the boundary layer.
and radiation parameterizations (hereafter boundary layer frontogenesis). We calculate $F_M$ and $F_{BL}$ from temperature tendencies obtained directly from the WRF parameterizations.

At 1500 UTC kinematic frontogenesis maxima lie along T2 and the central Nevada wind shift that extends northeastward from the Sierra Nevada (Fig. 17a). Ahead of T2, a cold pool generated by precipitation near the Nevada–Idaho border (Fig. 12c) is flanked by moist frontogenesis maxima (Fig. 17b). Weaker and more localized moist frontogenesis maxima lie immediately behind T2 and along the central Nevada wind shift, but are largely opposed by boundary layer frontolysis (cf. Figs. 17b,c). Throughout the simulation, the moist and boundary layer frontogenesis components are frequently of opposite sign since cooling induced by precipitation leads to enhanced sensible heating over the (comparatively) warm model land surface.

By 1800 UTC, T2 and the central Nevada wind shift have merged with the incipient cold front forming along the resulting confluence zone. Strong kinematic frontogenesis is found within the southern portion of the frontal zone near the Sierra Nevada (Fig. 18a) where moist frontogenesis is negligible (Fig. 18b) and boundary layer frontogenesis is aligned along the cold front but is weak (Fig. 18c). The limited role of diabatic processes in the frontal development over this region is consistent with the trajectory analysis in the preceding section. The northernmost kinematic frontogenesis maximum is found where the cold-frontal zone intersects the baroclinity associated with the precipitation-induced cold pool (Fig. 18a). Precipitation behind the cold front has also produced a moist frontogenesis maximum in the cold air behind the front (Fig. 18b), which is partially opposed by boundary layer frontolysis (Fig. 18c). Overall, these diagnostics show that kinematic frontogenesis dominates the frontal development near the sierra at this time, with only weak reinforcement by differential boundary layer heating. To the north, kinematic and moist frontogenesis maxima are contributing, but are not in phase along the incipient cold front.

Kinematic frontogenesis intensifies and forms two strong maxima along the frontal zone by 2100 UTC (Fig. 19a). Consistent with the synoptic analysis presented in the preceding section (Fig. 14c), model (and observed) precipitation becomes primarily postfrontal during this period and produces a diabatic frontogenesis maximum that aligns with but slightly trails the cold front and kinematic frontogenesis maxima (Fig. 19b). Boundary layer frontolysis continues to partially oppose the moist frontogenesis (Fig. 19c). By 0000 UTC kinematic and diabatic frontogenesis have reached their
maximum intensity and are in phase with the frontal zone (Figs. 20a,b). These two effects easily overwhelm the frontolysis associated with diabatic heating in the cold postfrontal boundary layer (Fig. 20c). A strip of boundary layer frontogenesis lies along and just ahead of the cold front, as might be expected from differential boundary layer heating arising from inhomogeneous cloud cover. As noted previously, however, this effect is not well captured by the model and may have played a more significant role in the actual event.

The analysis described above is consistent with the view that kinematic processes initiate the cold-frontal development, with diabatic frontogenesis contributing to frontal sharpening later in the event as precipitation
intensifies along and behind the front. Boundary layer processes partially oppose the moist frontogenesis as precipitation cooled air is sensibly heated by the warm model surface. Although differential sensible heating from inhomogeneous cloud cover may have played a role in the actual event, underforecasting of prefrontal temperatures and boundary layer depths suggests that such effects were underdone in CTL and only contribute to model frontogenesis late in the simulation.

d. FKDRY

The frontogenesis diagnostics described above quantify the instantaneous and direct contributions of kinematic and diabatic processes to the observed frontal development. They do not account for indirect effects, such as enhancement of kinematic frontogenesis by diabatically induced circulations. The FKDRY simulation further confirms that diabatic processes are not necessary for the discrete propagation.

At 1500 UTC, FKDRY produces kinematic frontogenesis maxima roughly along and between trough T2 and the central Nevada wind shift, with the wind transition in the vicinity of the latter weaker than in CTL (cf. Figs. 17a and 21a). By 1800 UTC, a single frontogenesis maximum forms along the incipient frontal zone (Fig. 21b) and intensifies over the next 6 h (Figs. 21c,d). Near the sierra, the frontal temperature gradient is actually stronger in FKDRY relative to CTL, whereas farther to the north, it is weaker (cf. Figs. 20a and 21d). These results confirm that diabatic effects are not essential to the discontinuous propagation, but do contribute to frontal sharpening in areas with frontal and postfrontal precipitation.

The greater intensity of the FKDRY front near the Sierra results from the removal of upstream condensational heating within the orographic cloud. Without such condensational warming, surface potential temperatures are 1–2 K colder in FKDRY than CTL (cf. Figs. 20a and 21d). Thus, the moist-diabatic influences on the frontal evolution are both local and remote.

5. Conclusions

The analysis described above provides what we believe is the first documented example of discrete frontal propagation across the Sierra–Cascade Mountains and Intermountain West. Critical to the discrete propagation during this event is the large-scale pattern and its interaction with the regional topography. The antecedent conditions feature a mobile upper-level cyclonic PV anomaly that is initially coupled with a landfalling Pacific cyclone and attendant occluded front. As the upper-level cyclonic PV anomaly moves eastward, it encourages frontogenesis along preexisting baroclinity over northern California and northwest Nevada in advance of the landfalling occlusion. Low-level confluence and convergence, which appear to be terrain enhanced, concentrate this temperature gradient and produce the incipient cold front over Nevada. Ageostrophic circulations arising from geostrophic adjustment are likely essential for the
rapid rate of frontal contraction, as described by Hoskins and Bretherton (1972).

WRF trajectories confirm that the incipient cold front forms in the preocclusion air mass and represents a boundary between potentially warm air originating over the desert Southwest, some of which has been deflected around the south end of the high sierra, and potentially cool air that has traversed the sierra near and north of Lake Tahoe, some of which has been deflected around the north end of the high sierra. Despite the presence of a deep well-mixed boundary layer near the developing front, precipitation-induced diabatic cooling, as implicated in some discrete propagation events (e.g., Kurz 1990; Bryan and Fritsch 2000a,b), is not essential to the discrete propagation. Differential sensible heating arising from inhomogeneous cloud cover appears to contribute to the observed frontal sharpening, but also does not appear to be essential to the discrete propagation during this event.

The discrete propagation does bear some similarity to that produced in some idealized simulations of front–mountain interactions (e.g., Keuler et al. 1992; Dickinson and Knight 1999). In particular, in idealized simulations by Dickinson and Knight (1999), the upper-level cyclonic PV anomaly moves relatively unimpeded across the barrier while the initial surface front decelerates and stalls on the windward slope. The upper-level cyclonic PV anomaly then couples with a lee trough, which becomes the new surface front. The discrete propagation in the 25 March 2006 event is similar in that an upper-level cyclonic PV anomaly moved relatively unimpeded across the Sierra–Cascade ranges and encouraged the formation of a new surface-based cold front downstream and well ahead of the initial landfalling occlusion. In contrast, the initial occluded front did not appear to stall over the northern California topography, preocclusion baroclinity served as fuel for the new frontal development, and the new cold front did not form within a classical lee trough, which is typically aligned along a barrier (e.g., Palmén and Newton 1969; Steenburgh and Mass 1994). Instead, the new frontal development was initiated by troughing and confluence that were oriented nearly normal to the Sierra Nevada. Work to better understanding the flow–mountain interactions that led to the discrete propagation is the subject of ongoing research.

Only recently have new observations, data assimilation, and numerical model advances revealed the rapid
development and remarkable intensity of cold fronts over the Intermountain West (e.g., Shafer and Steenburgh 2008). This study illustrates that these fronts can form from preexisting baroclinity in advance of a landfalling Pacific frontal system and that kinematic processes are sufficient to rapidly produce an intense cold front over Nevada. Forecasters should remain vigilant for discrete frontal propagation in similar synoptic situations and recognize that moist convection or differential surface heating can contribute to but are not necessary for rapid Intermountain frontogenesis.

Acknowledgments. We thank Larry Dunn, John Horel, David Schultz, and two anonymous referees for providing reviews that significantly improved the manuscript. Use of the WRF is made possible by the National Center for Atmospheric Research, which is supported by the National Science Foundation. We gratefully acknowledge the provision of datasets, software, or computer time and services by the Unidata Program Center, the National Centers for Environmental Prediction, the University of Utah MesosWest program and contributors, the University of Utah Center for High Performance Computing, the Center for Ocean–Land–Atmosphere Studies, the Department of Atmospheric Science University of Wyoming, Sebastian Hoch, and Mark Stoelinga. This research is supported by the National Science Foundation under Grant ATM-0627937. Any opinions, findings, and conclusions or recommendations expressed in this material are those of the authors and do not necessarily reflect the views of the National Science Foundation.

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