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The Frontal Hydraulic Head: A Micro-α Scale (≈1 km) Triggering Mechanism for Mesoconvective Weather Systems

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ABSTRACT

Measurements from the NOAA Boulder Atmospheric Observatory (BAO) 300 m tower, the National Center for Atmospheric Research (NCAR) Sabreliner aircraft, and the NOAA GOES-3 satellite, give evidence for the cross-front scale collapse of nonprecipitating surface cold-frontal zones to horizontal distances of ≈1 km or less. The leading edges of these fronts possess the characteristic structure of density current flows: an elevated hydraulic head followed by a turbulent wake. Vertical motions at the frontal heads exceed 5 m s⁻¹ at 300 m (AGL). The ascent at the frontal head may act as a (≈1 km-scale) triggering mechanism for the release of potential instability and the formation of intense squall-line mesoconvective.

1. Introduction

In reviewing the concepts and analyses of surface fronts, one finds that fronts near the ground were first considered zero-order discontinuities in density (temperature) and wind velocity, between air masses of different origin. The early observational studies (reviewed by Bergeron, 1959) documented the abrupt switch in wind direction and temperature drop during frontal passage. The accompanying theoretical treatments (e.g., Margules, 1906) used laws of hydrodynamics applied to “two-density” fluid models containing a sloping frontal interface in describing the motion, slope and vertical circulations of fronts. With the advent of kite and balloon-borne upper-air observations, it was discovered that fronts aloft were actually transition zones of finite width having characteristic vertical and horizontal scales of ≈1 km and 100 km respectively (Bjerknes and Palmén, 1937). At this stage in frontal history, the concept of zero-order discontinuities at fronts within synoptic-scale (≈1000 km) cyclones was discarded, and fronts were thereafter treated as sloping transition zones bounded by first-order discontinuities in temperature and velocity. However, it was recognized that sharp horizontal discontinuities could exist at the micro-α scale (0.2–2 km) such as at the outflow boundaries of convective precipitation systems, land–sea breeze convergences and orographically forced flows (Charba, 1974; Goff, 1976; Simpson et al., 1977; Matthews, 1981; Wakimoto, 1982; Carbone, 1982; and Hobbs and Persson, 1982).

It is the authors’ opinion that the shift from the discontinuous wedge (two-density) model to the continuous “zone” model was the result of combining synoptically spaced (≈400 km) upper-air observations with the meso-β scale spacing (≈100 km) of the hourly reporting surface stations, and the decline in the use of single-station (continuous record) analysis of frontal zones. Furthermore, with the expansion of synoptic observing networks and the application of the Norwegian cyclone-frontal models, fronts were treated as synoptic-scale (≈1000 km) in length whose narrow width resulted from the contractions of synoptic thermal gradients by the geostrophic deformations and their ageostrophic circulation response. The quasi- and semi-geostrophic equations which were later used to describe frontal contractions did not contain the nonhydrostatic physics which contribute to the final frontal scale collapse of the earlier wedge model or sharp discontinuities which had been observed with continuous temporal resolution observations.

Bergeron (1928) first described the role of synoptic-scale deformation fields in the contraction of temperature and velocity gradients into meso-β scale (≈100 km) frontal zones. Since then, the importance of geostrophic deformations and their coupled vertical secondary circulations have been examined in frontal studies by Petterssen (1956), Sawyer (1956), Eliassen (1959, 1962), Williams (1967), Hoskins and Bretherton (1972), and Shapiro (1981), among others. Eliassen (1959) recognized that secondary circulations act as a “self-sharpening” process to accelerate the
scale contraction of surface fronts. Williams (1967) and Hoskins and Bretherton (1972) simulated this process (ageostrophic contraction) through numerical and analytic solutions, respectively. In the absence of fine-scale turbulent motions, there are no limits to the scale to which frontal gradients may contract.
under the combined actions of geostrophic and ageostrophic motion. Inviscid, adiabatic frontogenesis produces infinitesimally narrow fronts within a finite time (Hoskins and Bretherton, 1972). The finite scale of frontal zones is a balance between frontogenetical synoptic-scale forcing and frontolytical turbulent-scale mixing. Further frontal contractions occur when moist convection breaks out within the ascending motion at the leading edge of fronts (e.g., Newton, 1950; Carbone, 1982), after the frontal dynamics are modified by the effects of water vapor phase changes and nonhydrostatic pressure changes. Surface boundary layer heat and momentum fluxes also contribute to vertical circulations about fronts (e.g., Keyser and Anthes, 1982; Shapiro, 1982; Koch, 1984). In summary, a wealth of information and considerable understanding has been acquired about the structure and processes which govern the evolution of surface fronts. One can say that surface fronts are presently considered one of the better understood and predictable of mesoscale atmospheric phenomena.

The present study was inspired by observations from the 300 m meteorological tower of the NOAA/Wave Propagation Laboratory (WPL) Boulder Atmospheric Observatory (BAO), which documented the micro-α scale contraction of a nonprecipitating surface cold front to a horizontal cross-front distance of less than 200 m (Shapiro, 1984). Vertical motions and cross-front convergence were found to be two orders of magnitude larger than previously observed for “typical” cold fronts, such as shown in Sanders (1955). Figure 1 shows the potential temperature and
cross-front wind speed, and Fig. 2 shows the vertical motion for this case.

In the present study, we provide additional documentation of this fine-scale structure. The first example is another case observed at the BAO tower. The second is a sharp cold front documented with the NCAR Sabreliner research aircraft over west Texas. In both cases, the width of the front was ~1 km near the ground and the leading edge of the front had an elevated “head” similar in form to that
Figure 5. Temperature (°C, solid lines) and wind speed (m s⁻¹, dotted lines) for 1950 to 2020 GMT 19 September 1983, measured at the 250 m level (top) and 50 m level of the BAO tower.

The observations for this study were taken with a variety of observing systems. Surface temperature and wind velocity were obtained from the NOAA Program for Regional Observing and Forecasting Services (PROFS) mesoscale network located east of the Rocky Mountains near Denver, Colorado. At the center of the network (30 km north of Denver), the BAO tower (Kaimal and Gaynor, 1983) provided fast-response (<1 s) measurements of horizontal velocity, vertical velocity and temperature at eight levels (10, 22, 50, 100, 150, 200, 250 and 300 m). The NOAA/WPL acoustic sounder (Brown and Hall, 1978) profiles (adjacent to the tower) were a source of additional information during frontal passage. The National Center for Atmospheric Research (NCAR) Sabreliner research aircraft provided 1 s measurements of horizontal wind velocity, vertical motion, pressure, temperature and moisture during frontal penetration over west Texas. Satellite visible imagery of narrow (rope-like) cloud lines and of mesoconvective cloud systems,

Figure 6. Acoustic sounder record of the frontal passage for 1950 to 2015 GMT 19 September 1983, measured at the BAO tower.
3. The surface cold front of 19–20 September 1983

On 19 September 1983, a sharp cold front moved southward through eastern Colorado in advance of the first major Canadian outbreak of late summer. A synoptic view at 2100 GMT 19 September (Fig. 3) showed the front oriented east-west across Colorado. Temperatures in advance of the front exceeded 30°C, while to the north they were near 0°C. A mesoscale view of the front was observed by the PROFS surface network (Fig. 4). The station temperatures (Fig. 4) were adiabatically adjusted to the surface pressure at the tower (a PROFS procedure) to remove topographic height variations from the resulting temperature analyses. This gave potential temperature analyses reduced to the ~850 mb pressure level. The front entered the network shortly after 1800 GMT (Fig. 4a) and passed rapidly southward (Figs. 4b, c and d) to the southern limit of the network by 2100 GMT with a speed of ~15 m s⁻¹. The westward movement of the front and cold air was impeded by the blocking action of the steep slope of the mountain lee. The sharpness of the temperature gradient and wind direction shift between Platteville (PTL) and Loveland (LVE), Colorado, at 1910 GMT (Fig. 4b) gave the first indication of the narrowness of the front. The discrepancy

were obtained from special processing of NOAA/GOES-5 satellite images with the University of Wisconsin, McIDAS System. Satellite infrared images were obtained with the NOAA/Environmental Research Group ground station receiver.
between the frontal analyses shown in Figs. 3 and 4d illustrates the value of regional networks in describing mesoscale frontal phenomena.

The BAO tower measurements revealed several dramatic characteristics of the front, which are shown in the temperature and wind speed traces from the 250 and 50 m levels of the tower (Fig. 5). The 50 m traces contained an 8°C temperature drop and a coincident 7 m s⁻¹ wind speed surge and cyclonic shear within 1 min as the front passed shortly after 2000 GMT. The frontal convergence and vorticity were coincident as was the case described in Shapiro (1984). Orlanski and Ross (1984) have suggested that frontal convergence and vorticity can decouple with the convergence leading the vorticity. For a frontal speed of 15 m s⁻¹ as taken from Fig. 4, the 1 min interval of frontal passage translates to a microscale frontal width of 900 m. The wind direction (not shown) shifted by 90° (315° to 045°) during passage. A second wind speed surge (from 10 up to 17 m s⁻¹) occurred between 2005 and 2008 GMT, 4 min after frontal passage at 50 m. Similar changes in wind and temperature were recorded at the 10, 22 and 100 m tower levels. The temperature traces at 150 m, and above, contained features not evident at lower levels. This is shown in the 250 m temperature trace (Fig. 5), where the front passed in two distinct temperature drops between 2000 and 2002 GMT, with a slight temperature rise between the two, after which the temperature rose by 3°C at 2005 GMT, and then cooled with turbulent undulations of 2°C amplitude.

The acoustic profiler record (Fig. 6) provided the necessary information for interpreting the unusual undulations in the frontal traces measured at the tower. Figure 6 represents the contouring of the maximum echo intensity which appeared on the acoustic profiler records. The acoustic profiler receives enhanced returns from elevated turbulent layers, especially those with strong refractive index discontinuities like those found at fronts, stable layers and elevated inversions. The acoustic record showed frontal passage at 2000 GMT and the rapid rise in altitude of the front into an elevated 550 m head. The frontal head was followed by a turbulent wake that lowered to below 150 m after which the front rose to above 300 m. This represents a remotely sensed record of the hydraulic head at the leading edge of the front. The acoustic record below 300 m is similar in structure with the observed oscillations in potential temperature measured at the tower and analyzed in Fig. 7.

The BAO tower and acoustic sounder observations were composited into an analysis of the front. The acoustic record was used to specify the vertical excursion of the frontal head above the 300 m limit of the tower. The resulting potential temperature analysis (Fig. 7) shows the arrival of the elevated frontal head 3 min after frontal passage, and the turbulent wake that followed in its lee. The leading edge of the front passed the tower in two distinct surges, with the major portion of the wind direction switch occurring within the first packet of isentropes. The analysis of the total horizontal wind speed showed surges to 18 m s⁻¹ at the 150 m level in the near-adiabatic layer behind and beneath the turbulent wake. This speed surge beneath the turbulent wake is outlined by the stippled 12–14 m s⁻¹ isentropic interval (Fig. 7). Note the in-phase relationship between the undulations in potential temperature and wind speed in the wake between 2005 and 2012 GMT (Fig. 7) as the vertical motions of trapped gravity waves and shear instability turbulence simultaneously raised and lowered the isopleths of temperature and velocity within the frontal layer. Wind speeds in the adiabatic layer beneath the frontal head averaged 9 m s⁻¹ between 2001 and 2005 GMT.

The vertical velocity measurements from the tower sonic anemometers contained strong ascending motions at the leading edge of the frontal head (Fig. 8), with the maximum upward motion of 4.5 m s⁻¹ at
Fig. 9. (a) NOAA/GOES-5 satellite visible image at 2100 GMT 19 September 1983. (b) NOAA/GOES-5 IR image at 0000 GMT 20 September 1983. Contour grey scale is from the enhancement (MB) curve of Clark (1983).
the 300 m level. The magnitude and distribution of this ascent is similar to that of the previously described front (Fig. 2). There were no observations of the vertical extent of the ascent plume nor its maximum value. The vertical motions following frontal head passage were quite chaotic (especially within the turbulent wake), but lesser in magnitude (<2.5 m s\(^{-1}\)), and were not included in the analysis of Fig. 8.

After Carbone (1982), we calculated the speed of the leading edge of the density current, \(V\), from the expression (von Kármán, 1940)

\[
V = \left( \frac{K^2 gh}{T_1 - T_2} \right)^{1/2}
\]

where \(T_1\) and \(T_2\) are the virtual temperatures of the lighter and denser air masses, respectively, \(g\) is gravity, \(h\) the depth of the layer behind the head and \(K^2 = 1.4\). For this case, \(h\) was determined from the acoustic record which showed the continued thickening of the post frontal cold boundary layer (beneath the frontal inversion) to \(\approx 500\) m at 20 min after frontal head passage (not shown in Fig. 6). For the present case, \(h = 500\) m, \(T_1 = 298\) K, \(T_2 = 290\) K and \(K^2 = 1.4\) and the resulting speed of the density current (frontal head) was \(14\) m s\(^{-1}\), in close agreement with the previously discussed frontal motion taken from Fig. 4. It should be noted that the leading edge of the density current and its hydraulic head propagated at a greater horizontal velocity than the wind component normal to the front beneath the head; i.e., \(15\) m s\(^{-1}\) versus \(9\) m s\(^{-1}\), respectively.

The satellite cloud observations over eastern Colorado did not exhibit sufficient structure to permit easy frontal identification from a single image. However, the animated McIDAS video loop with 30 min temporal resolution did show a narrow band of enhanced cloudiness imbedded with a broken cumulus layer which surged southward with the leading edge of the front over the PROFS mesonetwork (Fig. 4). This surging band was suspected to result from strong ascent (Fig. 8) at the frontal head. Figure 9a shows the NOAA/GOES-5 satellite visible image at 2100 GMT 19 September 1983. Of specific interest is the narrow “rope” cloud line that extended northeastward across northwestern Kansas at the leading edge of the front shown in Fig. 3. The transverse scale of the thin cloud (~3 km) is similar to that of the ascending plume at the leading edge of the frontal head shown in Figs. 2 and 8. At 0000 GMT, the rope cloud was also present in the satellite infrared imagery (Fig. 9b) and extended diagonally across northwestern Kansas. The cold (darker) upper-level cloud structure which extends northwestward from the rope cloud (Fig. 9b) is the outflow from the mesoconvective activity that erupted as the front intercepted the region of enhanced low-level moisture over eastern Nebraska and Iowa. The gray-scale contours for Fig. 9 are taken from the

Enhancement (MB) curve as described in Clark (1983). Figure 10 shows the 2035 GMT 19 September radar summary which outlines the areas of severe mesoconvective activity and the severe weather watch area.

From the available observations, it was not possible to provide conclusive evidence that the rope cloud (Fig. 9) was the signature of mesoscale ascent at the eastern extension of the frontal hydraulic head (Figs. 6–8). However, the barograph traces at surface stations within Nebraska, Colorado and Kansas showed strong pressure rise (surges) at time of frontal passage. An isochrone analysis of the time of surge passage for the 19 September 1983 front is shown in Fig. 11 along with some selected pressure traces. The pressure surge at the tower was coincident with the frontal passage. The position of the surge at 2100 and 0000 GMT (Fig. 11) coincides closely with the 2100 GMT and 0000 GMT position of the rope cloud in the satellite images (Fig. 9). It should be noted that standard National Weather Service (NWS) barographs have damping mechanisms which inhibit their ability to resolve hydraulic-head pressure perturbations whose time-scale of passage is tens of seconds or less.

On the following day, at 1800 GMT 20 September 1983, the front extended from Arkansas across Texas (Fig. 12) with a narrow rope cloud line of enhanced cumulus development at its leading edge (Fig. 13a). By 2000 GMT (Fig. 13b), the cloud line of Fig. 13 developed into a squall line with embedded cumulonimbus cellular “popcorn” structure. The above two-
FIG. 11. Frontal pressure surge, 19–20 September 1983. (a) Isochrones (hr) of pressure surge and plotted times (GMT) of surge onset at selected stations. (b) Barograph traces from numbered stations in (a). Three-hourly pressure tendencies [mb (3 h)] for the 3 h interval following surge onset.

FIG. 12. Surface streamlines (solid lines) and temperature (°C, dashed lines) at 1800 GMT 20 September 1983. Flags and barbs, as indicated in Fig. 3; rope cloud of Fig. 13, stippled line.
Fig. 13. NOAA/GOES-5 satellite visible images at (a) 1800 and (b) 2000 GMT 20 September 1983. Rope cloud of (a), same as indicated on Fig. 12.
each of the three flight levels (Fig. 15a) when the Sabreliner penetrated the front. Note that at the 857 mb flight level, the wind was northwesterly in the cold air beneath the head and then returned to southwesterly flow in the warmer air to the west.

The equivalent potential temperature analysis (which shows the combined effect of temperature and moisture discontinuities) and gust probe vertical velocity measurements (Fig. 15b) provided further documentation for the structure of the frontal head. Ascending vertical velocities exceeding 5 m s⁻¹ were measured just in advance of the frontal head, in agreement with the previously described cases (Sections 1 and 3). The leading edge of the front possessed the character of the hydraulic frontal head described in Section 3. The observations of the microscale frontal structure (Fig. 15) were taken just prior to the onset of severe mesoconvective activity. All flight legs of Fig. 15 were made below cloud base and were flown under Visual Flight Regulation (VFR) conditions without entering precipitation. Shortly after the Sabreliner’s landing at Abilene, tornadoses were reported in association with thunderstorms near the front. The extent of the mesoconvective that erupted as the front entered the Gulf of Mexico moisture of >11 g kg⁻¹ mixing ratio over central Texas is shown in the NWS radar summary for 0135 GMT 6 May 1982 in Fig. 16. Though this front exhibited microscale head structure, the satellite imagery did not contain a narrow rope cloud as a precursor to the mesoconvective development.

5. The surface cold front of 9–10 June 1984

There are numerous examples when visible satellite imagery shows the cloud signature associated with micro-α to meso-γ (0.2 to 20 km) ascent at the leading edge of surface fronts (e.g., Shapiro, 1982; Koch, 1984) prior to mesoconvective outbreaks. The narrowness of these clouds suggests that they are associated with fronts that have contracted down as narrow as the microscale. The rope cloud can form in the narrow ascending plume within the upward branch of meso-β secondary circulations (Koch, 1984), or be forced by hydraulic heads at the leading edge of frontal density currents (Figs. 2 and 8). The following example illustrates the development of mesoconvective activity along a pre-existing frontal rope cloud.

On the afternoon of 9 June 1984, a cold front to the lee of the Rocky Mountains was situated across eastern Nebraska, eastern Kansas and western Oklahoma (Fig. 17). By 1800 GMT, mesoconvective activity was occurring along the northern segment of the front over northeastern Kansas and southeastern Nebraska (Fig. 18a and 19a). Figure 18 shows the NOAA/GOES-5 visible imagery at hourly intervals between 1800 and 2100 GMT 9 June 1984. At 1800
GMT (Fig. 18a) a thin line of enhanced cumulus development extended from central Kansas to the Texas panhandle along the leading edge of the cold front. This rope cloud line marked the discontinuity between the cool, dry northwesterly flow to the northwest and the warmer moist southerly flow from off the Gulf of Mexico. Note the 40-km wavelength cloud lines embedded in the stratocumulus layer (Fig. 18a, central Oklahoma) in advance of the front. These cloud lines are probably the result of vertical circulations in the pre-frontal boundary layer, whose orientation is transverse to the boundary layer vertical wind shear vector. Reference to Fig. 17 shows that the rope cloud (Fig. 18b) is situated at the frontal wind shift between the two surface network stations at Wichita, Kansas (indicated by dashed circle). By 1900 GMT (Fig. 18b), the rope cloud began to broaden in response to the development of the cu-
mulus activity along the front. At 2000 GMT (Fig. 18c), the anvil outflow signature of cumulonimbus had erupted along the front in southern Kansas and northern Oklahoma. The final image (Fig. 18d) shows the fully developed cirrus outflow along the 500 km long squall line, with new convection breaking out at the southern remnant of the rope cloud. The NOAA/GOES-5 infrared images at 2 h intervals, from 1800 GMT 9 June to 0400 GMT 10 June 1984 are shown in Fig. 19. These images show the horizontal and vertical development of the mesoconvective system (MCS) cirrus anvil outflow. The visible and infrared satellite images of Figs. 18 and 19 suggest the presence of micro-α scale ascent as a precursor to the outbreak of the mesoconvective activity on this day. Unfortunately, there were no surface temperature or wind observations to document the coincidence of this rope cloud with frontal hydraulic head structure of the type described in Sections 1, 3 and 4.

6. Summary and discussion

The present study has shown additional documentation for the contraction of surface frontal gradients from the mesoscale down to the microscale, in the absence of convective precipitation. Of special notice are the observations of the elevated hydraulic head at the leading edge of the front and the several meter per second ascent in advance of the head as it propagates into the prefrontal warm air. The fronts described herein have the structural characteristics of classical density current flows cited in the introduction of this article.

It is interesting to note that we have come full circle in our interpretations, having put aside the

early near zero-order discontinuity frontal models only to find such structure almost a century later using “modern” instrumentation. The fast response of the platinum wire temperature sensors on the BAO tower recorded frontal passages in time intervals as small as 10 s, giving the 170 m scale for the front shown in Fig. 1. The quartz thermistors on the same tower, similar to those used for continuous temperature records at the PROFS mesonetwork stations and NWS stations, required up to 6 min to record the same temperature drop as the platinum wires. The near zero-order frontal discontinuity concepts were undoubtedly based upon the fast response of the human skin (and mercury thermometers) as researchers stood in the elements to directly sense the wind and temperature changes during frontal passages. The conceptual models of fronts have evolved in response to the sensitivity and spatial resolution of the observing networks. Sanders (1955) suspected that the surface cold front he observed in a meso-β scale (∼100 km) network was only “a few kilometers wide at the surface,” but because the spatial resolution and hourly reporting frequency of the network he was unable to resolve its true time or space scale.

Fronts and their associated vertical circulations have long been recognized as a mechanism for lifting moist air to its level of free convection and initiating the release of potential instability through deep mesoconvective cloud systems. An alternative hypothesis for triggering prefrontal thunderstorms was proposed.

1900 GMT 9 JUNE 1984

FIG. 17. Surface streamline analysis at 1900 GMT 9 June 1984. Rope cloud, (stippled) is from Fig. 18b. Flags and bars as in Fig. 3.
Fig. 18. NOAA/GOES-5 satellite visible cloud images at 1800 GMT, 1900 GMT, 2000 GMT and 2100 GMT 9 June 1984 as prepared on the University of Wisconsin McIDAS interactive meteorological graphics system.
in the theoretical treatments of Freeman (1948) and Abdullah (1949), with extensive observational documenta-
tion by Tepper (1950). These studies proposed that frontal and prefrontal squall lines were initiated by
hydraulic pressure waves or “jumps” formed by the accelerated motion of fronts. On occasion these pressure
jumps were thought to propagate out in advance of the triggering front. The documentation by Tepper (1950) was taken from an automatic 55-
station microbarograph, wind and temperature ob-
serving network with ~3 km spatial resolution of the observing stations. Newton (1950) questioned the
importance of the pressure jump triggering mecha-
nism, stating that the vertical excursions in trapped
surface moisture layers caused by the jumps and their
propagating gravity waves would not be sufficient to
raise air parcels to their level of free convection, and
hence produce squall line convection. Fujita (1955)
argued and presented evidence that the pressure jump
lines of Tepper (1950) were the result of, rather than
the cause of squall lines. We suggest that the limited
acceptance of nonhydrostatic, micro-α scale phenom-
ena as an initiation mechanism for mesoconvective
systems was due to the lack of documentation of the
vertical structure of the pressure jump lines. It is only
recently that instrumented towers and research aircraft
have provided some corroborating evidence for density
current (hydraulic head) characteristics of surface
cold fronts, thunderstorm outflows and sea-breeze
fronts. The present study has shown examples where
nonprecipitating fronts have assumed the characteristics of density-current flows. The large ascending
vertical motions (~5 m s⁻¹) at the frontal head are of
sufficient magnitude to lift potentially unstable air
through the trapping “lid” inversions (Carlson and
Ludlam, 1968; Carlson et al., 1983) which often cap
moist surface layers. The frontal head is capable of
releasing the potential instability of trapped moist
layers on time scales of minutes rather than hours as
required for the 20 cm s⁻¹ lifting of meso-β scale
(~100 km) frontal vertical circulations. It is not
unreasonable to consider the possibility that the frontal
head may separate itself from its parent front and
propagate out in advance of the front as a solitary
wave (Abdullah, 1949). The separation process could
produce prefrontal pressure waves (as in Tepper,
1950; Uccellini, 1975) and initiate precipitation sys-
tems in the prefrontal environment.

It remains for future research to 1) establish the
relationship between frontal deformations, their ver-
tical circulations, and surface boundary layer processes
in the scale collapse of frontal gradients and the
formation of frontal hydraulic heads, 2) derive the
theoretical treatments which describe the transition from geostrophic-momentum (semigeostrophic) dy-
namics over to nonhydrostatic, primitive equation
dynamics as frontal gradients transcend the meso-α
scale down to the micro-α scale and 3) carry out

further field experiments with research aircraft,
Doppler lidars, Doppler radars and fast response (~1
s) surface based instruments so as to clearly establish
the role and representativeness of the hydraulic frontal
heads, described herein, in initiating the formation of
narrow rope cloud lines and severe mesoconvective
weather systems.

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