The Eastern United States Side-Door Cold Front of 22 April 1987:  
A Case Study of an Intense Atmospheric Density Current

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ABSTRACT

A case study is presented of an unusual cold front that affected the east coast of the United States on 22 April 1987. Noteworthy aspects of this front are its genesis behind a preexisting back-door cold front, its propagation to the southwest, the shallow nature of the cold air associated with it (<600 m), and the absence of precipitation. This front has been termed a "side-door" cold front because of the significant differences between it and typical back-door cold fronts, such as its direction of propagation and vertical structure. Weather conditions in the mid-Atlantic states changed abruptly from partly cloudy skies, light winds, and 25°-30°C surface temperatures ahead of the front to a windy (gusts to 20 m s⁻¹), low overcast regime with temperatures 10°-15°C cooler behind the front.

Analysis of satellite imagery, surface data, and sounding data revealed that the front possessed density-current structure during a portion of its lifetime. Observed front-relative flow, as well as agreement between computed and observed frontal velocity, support this conclusion. Results from this study demonstrate that the cold front evolved without the blocking and channeling effects of a major mountain range, as is typically the case for similar events around the world. It is hypothesized that the front formed in response to differential surface heating and friction along the New England coast. Similarly, the differential heating across the coastline contributed to the intensification of the front as well as to the evolution of the density-current structure later in the life history of the front. Results of this study relevant to several numerical simulations of similar events are discussed.

1. Introduction

During the warm season, heat waves along the east coast of the United States are occasionally broken by cold fronts that propagate southward out of Canada to the east of the Appalachian Mountains. Fronts of this type are known as "back-door" cold fronts (Carr 1951; Bosart et al. 1973; Eichler and Shulman 1987). Bosart et al. (1973) constructed a back-door cold-front climatology and found that they are a warm-season phenomenon with highest (lowest) frequency in early autumn (spring). Typically, the east-to-west-oriented back-door cold front moves southward down the Atlantic coast while an anticyclone builds in behind it from the north. At 500 mb a short-wave trough moves eastward across northern Canada to the Atlantic coast. A similar situation began to unfold on 22 April 1987, as a back-door cold front moved southeastward from Quebec into New England. However, as this front stalled over the Gulf of Maine, frontogenesis occurred behind the front. This new front, which is termed a "side-door" cold front (because the primary motion of the cold air was westward), became the dominant front as it propagated southwestward into the mid-Atlantic states (Fig. 1a). As it moved through this area, wind gusts of 15-20 m s⁻¹ were experienced (Fig. 1b) with a rapid temperature fall from 25°-30°C to 10°C in roughly two hours.

As will be demonstrated, the cold air associated with the side-door cold front had the vertical structure of a density current [also known as gravity current; see Simpson (1987) for a comprehensive summary of density currents in the environment] during a portion of its lifetime. Density currents are created when two fluids of dissimilar density are placed adjacent to each other in the presence of gravity, such that the denser of the two fluids will flow in a shallow current beneath the less dense fluid. With the side-door cold front, the ocean provided a reservoir of cold water (Fig. 1b) that chilled the overlying air mass creating the dense fluid, while heated air over the coastal plain served as the less-dense fluid. Although these conditions occur quite frequently in spring, sea breezes are usually all that result. In this case, the larger-scale flow regime produced conditions that forced the marine air to penetrate westward several hundred kilometers inland.

Early theoretical work on the structure and propagation of density currents yielded mathematical and physical results that matched laboratory simulations quite closely (Benjamin 1968; Simpson 1987). The work of Benjamin (1968) has been applied to atmospheric events such as thunderstorm outflows (Charba
1974; Goff 1976; Wakimoto 1982), sea and land breezes (Clarke 1961; Schoenberger 1984), and the leading edge of synoptic-scale cold fronts (Carbone 1982; Hobbs and Persson 1982; Shapiro 1984; Seitter and Muench 1985; Bond and Fleagle 1985; Bond and Shapiro 1991). Density-current theory has also been applied to a class of shallow cold fronts that have undergone deformation upon encountering mountain barriers. A notable example of these events is the "southerly buster" of southeastern Australia (Baines 1980; Colquhoun et al. 1985). Similar phenomena have been observed in New Zealand (Steiner et al. 1987; Smith et al. 1991), the west coast of the United States (Dorman 1985; Mass et al. 1986; Mass and Albright 1987; Dorman 1987), near the Chilean Andes (Rutlant 1981), the Pyrenees (Hoinka and Heimann 1988), the Appalachians (Carr 1951; Bosart et al. 1973), and Central America (Parmenter 1970).

As will be discussed, the present case has significant similarities and differences to these other events. The side-door cold-front event of 22 April 1987 presents an excellent opportunity to study an intense, dry (no observed precipitation) atmospheric density current. The purpose of this study is to describe the life history of this cold front. It will be shown that the evolution of the front was crucially dependent upon the configuration of the larger-scale flow, and the density-current phase of the front developed as it crossed a differentially heated coastline.

Section 2 of this paper provides an overview of the methodology to this study, while section 3 outlines important features of the large- and small-scale flow at the surface and in the lower troposphere. Computation of vorticity, divergence, and frontogenesis are used to illustrate the evolution of the front in section 4, along with density-current calculations. A concluding discussion is contained in section 5.

2. Methodology

The development and temporal changes of the side-door cold front were evaluated by constructing manually analyzed mesoscale surface maps of pressure, temperature, dewpoint, wind direction, and wind speed for 0600, 1200, and 1800 UTC 22 April, as well as 0000 UTC 23 April 1987. High-resolution satellite images (1 km visible and 4 km infrared) were simultaneously analyzed and utilized in refining the analyses, particularly over the Atlantic Ocean where observations were relatively sparse.

To quantitatively assess the structural evolution of the flow patterns at the surface, the mesoscale analyses
were manually gridded on a $0.5^\circ \times 0.5^\circ$ latitude–longitude grid extending from $30^\circ$ to $44^\circ$N and $68^\circ$ to $82^\circ$W. The following diagnostic computations were performed:

\begin{align}
\text{relative vorticity: } \zeta &= k \cdot \nabla \times \mathbf{v} \\
\text{horizontal divergence: } \delta &= \nabla_h \cdot \mathbf{v} \\
\text{horizontal temperature advection: } A_T &= -\mathbf{v} \cdot \nabla T.
\end{align}

Surface frontogenesis, after Miller (1948), neglecting friction and diabatic heating and using observed winds is

\begin{equation}
\frac{d}{dt} \left| \nabla \theta \right| = \frac{1}{\left| \nabla \theta \right|} \left[ \frac{\partial \theta}{\partial x} \left( -\frac{\partial u}{\partial x} \frac{\partial \theta}{\partial x} - \frac{\partial v}{\partial y} \frac{\partial \theta}{\partial y} \right) \\
+ \frac{\partial \theta}{\partial y} \left( -\frac{\partial u}{\partial y} \frac{\partial \theta}{\partial x} - \frac{\partial v}{\partial y} \frac{\partial \theta}{\partial y} \right) \right].
\end{equation}

Geostrophic frontogenesis was also computed using geostrophic wind components derived from the sea level pressure field. Analyses of equivalent potential temperature were also generated from the surface mesoscale grid. To assess the large-scale flow evolution, National Meteorological Center (NMC) global gridded analyses of geopotential height and temperature on a $2.5^\circ$ latitude–longitude grid for the standard pressure levels were utilized. Synoptic-scale vertical motions were estimated qualitatively by overlaying the 1000–500 mb thickness field on the 500-mb vorticity field in accord with the concepts presented by Sutcliffe (1947) and Durran and Snellman (1987).

3. Synoptic and mesoscale overview

a. Upper-level features

At 0000 UTC 22 April 1987, roughly 12 h prior to the genesis of the side-door cold front, the main large-scale features at 500 mb (Fig. 2a) consisted of a stationary cutoff cyclone east of North Carolina, a large ridge extending from the Gulf of Mexico to the Great Lakes, and a short-wave trough to the east of Hudson Bay across Canada. Cutoff cyclones were also located over eastern Nebraska and northern Mexico. A similar

![Fig. 2. National Meteorological Center 0000 UTC 22 April 1987: (a) 500-mb height (solid, every 6 dam) and temperature (dashed, every 5°C); (b) 850-mb height (solid, every 3 dam) temperature (dashed every 5°C) and conventional surface frontal analysis; (c) 1000–500-mb thickness (solid, every 6 dam) and 500-mb absolute vorticity (dashed, every $2 \times 10^{-5}$ s$^{-1}$); (d) 850-mb horizontal temperature advection every 1°C (3 h)$^{-1}$, with negative values dashed.](image-url)
pattern also existed at 250 mb (not shown), with a zonal jet stream in Canada. At 850 mb (Fig. 2b) a plume of air whose temperature exceeded 10°C reached well into the Canadian maritimes east of the Hudson Bay trough, while upstream −10°C air was being advected southeastward. Accordingly, the 850-mb temperature advection map (Fig. 2d) features an area of warm-air advection to the east of Newfoundland and a region of strong cold-air advection [peak values near −6°C (3 h)−1] to the north of Maine that accompanied a surface cold front. Cyclonic-vorticity advection (CVA) by the thermal wind (Fig. 2c) implies ascent northeast of Labrador, over Iowa and Wisconsin, and north of Minnesota. In contrast, implied descent associated with the region of strong anticyclonic-vorticity advection (AVA) by the thermal wind north of the Great Lakes corresponds to the location of the surface anticyclone.

By 1200 UTC, the short wave at 500 mb had moved east to the Canadian maritimes (Fig. 3a). A region of AVA by the thermal wind (Fig. 3c) implies descent over Quebec, in the vicinity of the surface anticyclone. The main region of 850-mb cold-air advection (Fig. 3d) is located northeast of Newfoundland. As was the case at 0000 UTC, the prefrontal plume of warm air extends from the eastern United States to eastern Newfoundland (Fig. 3b). Over the Gulf of Maine, in the vicinity of the developing side-door cold front, there is weak cold-air advection.

The 500-mb short-wave trough axis was located east of Labrador at 0000 UTC, while an upstream ridge built up over the eastern United States (Fig. 4a). The 500-mb cutoff cyclone associated with the surface cyclone east of North Carolina remained quasi-stationary, with cold-air advection at 850 mb (Fig. 4d) decreasing in magnitude from earlier times and shifting to the east of the Canadian maritimes. In the region affected by the side-door cold front, there is near-zero temperature advection at 850 mb, reflecting the shallowness of the cold advection associated with the front. AVA by the thermal wind and the accompanied implied descent had split into two regions since 1200 UTC (Fig. 4c). One region was centered on Newfoundland, with the other located from New England to Nova Scotia. The latter region corresponds to the location of a new anticyclone that had developed over the Gulf of Maine. The combination of height rises behind the 500-mb trough and stationary cutoff east of North Carolina.

Fig. 3. As in Fig. 2 except for 1200 UTC 22 April 1987.
served to increase the surface pressure gradient over the ocean such that there were strong easterly winds over the ocean. This flow pattern was conducive to an inland penetration of relatively cold marine air. Details of the surface-flow evolution are now examined.

b. Surface features

The main surface pressure features at 0600 UTC (Fig. 5) consisted of a stagnant anticyclone over the mid-Atlantic states and a stationary cyclone offshore of North Carolina. Farther north, a cold front was aligned northeast to southwest from coastal Maine to western Pennsylvania. This back-door cold front moved rapidly east and south, and had already introduced much colder air into the northern regions where maximum temperatures of 25°–30°C on 21 April extended as far north as 50°N in eastern Canada ahead of the front. Behind the front, a 1030-mb anticyclone was located to the north of the Great Lakes.

The 0600 UTC surface analysis reveals that along the New England coast and offshore waters, air temperatures ahead of the front were approximately 5°–10°C colder than air immediately behind the front. This low-level cold air was a result of a southerly fetch of air that had a history of being cooled by contact with the relatively cold ocean waters (Fig. 1b).

Time series of surface observations centered on the time of frontal passage for stations in northern New England [Montpelier, Vermont (MPV), and Greenville, Maine (3B1)] and coastal New England [Portland, Maine (PWM), and Portsmouth, New Hampshire (PSM)] illustrate the effect of low-level cold air in advance of the front (Fig. 6). At 3B1 and MPV there was an abrupt pressure rise along with a sharp directional wind shift to northwest and an increase of wind speed from 3 to roughly 10 m s⁻¹. Along the coast, a similar pressure rise was observed; however, the winds switched from light southwesterly to light northeast. Three-hour temperature falls behind the front amounted to 13.8°C at MPV and 9.4°C at 3B1. Along the coast, both PSM and PWM indicate rising temperatures behind the front. After several hours, strengthening northeast winds advected colder air into these areas.

The postfrontal warm band observed at 0600 UTC (Fig. 5) persisted at 0900 UTC (Fig. 7), with the cold front having moved over the Gulf of Maine. A relatively homogeneous temperature field is evident both in the vicinity of and behind the cold front over the ocean.
However, a strong temperature gradient persisted along the coastline behind the front, with a warm tongue extending northward from central Massachusetts to eastern Maine. This warm tongue coincides with the postfrontal warming observed in the time-series analyses for PWM and PSM, and is likely due to previous adiabatic warming due to downslope flow east of the mountains. This effect would be most pronounced along the immediate coast, where the marine inversion was present.

The pressure field in Fig. 7 indicates that a weak inverted trough developed in the vicinity of the warm tongue along the New England coast. Over the Gulf of Maine, a strengthening surface pressure gradient accompanied northeast winds that increased to 10 m s⁻¹. By 1200 UTC (Fig. 8), the pressure gradient intensified over the Gulf of Maine such that strengthening northeast winds had advected the cold marine air southwestward. The net result was frontogenesis in the vicinity of the preexisting coastal baroclinic zone, which was accompanied by an amplification of the inverted trough. This new front will hereafter be termed the

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**Fig. 5.** Surface map for 0600 UTC 22 April 1987. Station observations are given with temperature (°C), dewpoint (°C), wind direction, and speed (half-barb = 2.5 m s⁻¹, full barb = 5 m s⁻¹). The "M" indicates missing or unreported data. Surface pressure is contoured every 1 mb with the leading "10" omitted.
by a steady temperature fall, as air with a trajectory from the Gulf of Maine reached the area.

The 1200 UTC Chatham, Massachusetts (CHH), sounding, launched at approximately 1110 UTC in between fronts (the side-door cold front moved through this location at approximately 1130 UTC) is given in Fig. 9. The sounding is essentially isothermal and moist from the surface to 850 mb, where it becomes drier with increasing lapse rate. Low-level winds are northwesterly, with a maximum speed of 10 m s\(^{-1}\) at 300 m. The lack of any substantial directional backing in low levels, even accounting for Ekman veering of the boundary-layer winds, indicates weak temperature advection, illustrating the decay of the back-door frontal boundary.

The southward propagation of the front during 1400–2100 UTC is clearly illustrated by visible satellite imagery (Fig. 10). At 1400 UTC (Fig. 10a) the side-door front is delineated by an arced band of stratus cloud extending from central Connecticut across eastern Long Island. An extensive area of stratus extends to the south along the coastal waters with the northern edge of the clouds associated with the offshore cyclone just visible at the bottom of the image. The motion of the front is illustrated by the westward motion of the arced cloud band in Figs. 10a–d.

From an analysis of hourly isochrones of position, the side-door cold front was found to propagate rapidly southwestward (17–21 m s\(^{-1}\)) down the coast from 1200 to 1800 UTC (Table 1). Rapid anticyclogenesis in the Gulf of Maine (cf. Figs. 8 and 11) produced a northeasterly low-level flow behind the front, favorable for its southwestward propagation. In addition, the cold air already in place over the relatively cold coastal waters southward to the Carolinas provided a continuously renewable source of cold air (Fig. 1b).

As the side-door cold front approached the New Jersey coast, one branch of the cold air continued channeling southwest along the coastal waters while another surged onshore. The southward moving portion of the flow formed a “nose” in the front that surged south along the coastal waters at 20 m s\(^{-1}\) (Table 1). The westward-moving portion of the cold front, as will be demonstrated, obtained a density-current structure as it crossed the coastline. The 1800 UTC surface map (Fig. 11) indicates that air temperatures immediately ahead of the front were 23°–24°C, with 9°–13°C air temperatures behind the front. The anticyclone in the Gulf of Maine developed over the past 6 h, maintaining a strong pressure gradient behind the side-door front. A westward-propagating inverted trough, which was a northward extension of the side-door cold front, was located between the new anticyclone and the preexisting anticyclone located over Quebec.

A prefrontal trough extended northward over the coastal plain from Virginia to eastern Pennsylvania, with high pressure over the mountains to the west. The 1800 UTC visible satellite image (Fig. 10c) indicates
that the side-door cold front was located at the leading edge of a deck of stratus cloud, as was the case for earlier times as well. This stratus deck extends eastward over the ocean south of New England in a region of east-northeasterly surface flow. The old back-door front was stationary in northern Pennsylvania at this time, with a cloud band farther to the north (Fig. 10).

The time series of observations at buoy 44009, located approximately 50 km east of the southern tip of New Jersey (Fig. 12), indicates there was little to no temperature contrast across the front when it passed this location. In addition, the observed pressure rise behind the front was slight, indicating that the depth of the cold air was roughly as deep as the marine boundary layer. Thus, the low-level baroclinity associated with the side-door cold front as it moved inland must have originated along the coast in response to the differential diurnal heating cycle along the coastline. This process probably contributed to the collapse in the horizontal scale of the frontal zone as it crossed the coast.

![Figure 7](image)

Passage of the side-door front in the mid-Atlantic states was marked by a gust-front line, a rapid pressure rise, 10°C temperature fall, and rapid influx of low stratus cloud. The sequence and magnitude of each of these changes varied with increasing distance from the coastline. For instance, at Atlantic City, New Jersey (ACY), these changes occurred almost simultaneously (Fig. 12). The temperature at ACY remained between 22°C and 23°C during the time of maximum heating due to an easterly gradient wind that was probably enhanced by a sea breeze. However, after the passage of the side-door cold front at 1730 UTC, the temperature dropped to 13.8°C by 1750 UTC.

When these observations of frontal passage over land are compared with the observations at buoy 44009, it is clear that the hydrostatic portion of the pressure jump behind the front was associated primarily with cooling in the boundary layer. Temperature falls of 10°C in 2 h behind the front in the Washington D.C. area (DCA, MTN, and IAD) (Fig. 12) are similar to those observed at ACY. However, the initial rate of fall is not as dra-
matic at the inland stations, with a steady temperature drop over a 2-h period. Similar to ACY, these stations experienced a sea-breeze effect (from Chesapeake Bay) prior to frontal passage. As the front pressed inland, satellite pictures (Figs. 10 and 14) and surface observations revealed a progressively increasing lag in the stratus deck behind the surface front. Since air in the lower troposphere over land was quite dry, it is possible the stratus dissolved and that it took longer to reform due to a longer land trajectory. In addition, it is likely that mixing behind the head of the density current would delay cloud formation due to entrainment of drier environmental air above the cold current.

The side-door density-current front continued to move westward at 13 m s\(^{-1}\) through Maryland, Virginia, and Pennsylvania during 1800–0300 UTC (Fig. 13). As the front moved inland, it was clearly defined on satellite imagery by a narrow line of cumulus clouds (Figs. 10d and 14). This observation is consistent with previous studies of large-scale atmospheric density currents, in which ascent over the head of the current was sufficient to produce a line of cloud. By closely analyzing this line of clouds on 1-km visible satellite images, it was found that the front was impeded by higher terrain in portions of northern New Jersey and Pennsylvania. In these areas, the terrain reaches ele-
Fig. 9. Skew T-logp format soundings for Chatham, Massachusetts, at 1200 UTC 22 April 1987. Sample moist adiabat (dotted), dry adiabats (thin solid), and temperature (thin solid) lines are given.

lations between 200 and 600 m, which implies that the cold air was less than 600 m deep. Soundings at ACY and Wallops Island, Virginia (WAL) (Fig. 15), also illustrate the shallow nature of the cold air. At 0000 UTC 23 April the only significant change in the soundings is confined below 900 mb. Comparison with the 1200 UTC soundings (dashed) above 900 mb indicates little change. At WAL, the depth of the cold air was 550 m, with a 12.5 m s⁻¹ low-level wind maximum at 265 m. ACY, farther back in the 900-m-deep cold air, reported a 13.5 m s⁻¹ wind maximum at 545 m. A prefrontal sounding at IAD at 0000 UTC (launched at 2310 UTC) is nearly identical to the 0000 UTC soundings at ACY and WAL above 850 mb (Fig. 16). At IAD, however, a deep mixed layer extended from the surface to 800 mb, where the marine inversion was located at ACY and WAL. By 0300 UTC 23 April the front reached its westernmost location along the foothills of the Appalachian Mountains. The front moved farther south into South Carolina overnight with the boundary ill defined, as a result of the large displacement from the source of cold air. The oceanic portion of the front became stationary offshore along the Gulf Stream boundary, in a region of easterly flow (Fig. 13).

4. Results

a. Gridded surface data computations

A quantitative diagnosis of the surface flow features observed in section 3 is now presented at 6-h intervals.

1) 0600 UTC 22 April 1987

At 0600 UTC (Fig. 17) a band of cold advection lies behind the northeast–southwest-oriented cold front from New Hampshire to Ohio, in contrast to a relatively homogeneous prefrontal environment. Cyclonic relative vorticity is focused in a narrow band along the front, with maximum values reaching $12 \times 10^{-5} \text{s}^{-1}$ over central New England and western Pennsylvania. The quasi-stationary cyclone east of Cape Hatteras is also reflected by an area of large cyclonic relative vorticity. The largest convergence values ($12 \times 10^{-5} \text{s}^{-1}$) are also located along the front, with an area of divergence exceeding $8 \times 10^{-5} \text{s}^{-1}$ located over coastal Massachusetts. A band of equivalent potential temperature $\theta_e$ values exceeding 308 K lies immediately in advance of the cold front, while over the coastal waters from Maine southward, $\theta_e$ values are less than 300 K. The gradient of $\theta_e$ along the coastline reflects the typical springtime land–sea contrast.

Both observed wind and geostrophic frontogenesis results at 0600 UTC (Figs. 18a and 18b) reflect values 2–4 K (100 km)⁻¹ (3 h)⁻¹ in northern Pennsylvania and southern New York. However, along the coast in Maine and New Hampshire there is geostrophic frontogenesis with values of 8 K (100 km)⁻¹ (3 h)⁻¹, while the observed wind frontogenesis has values near zero. The latter reflects the decay of the surface front as it reached the coast, with colder air in advance of the front over the ocean.

2) 1200 UTC 22 April 1987

By 1200 UTC, the observed wind frontogenesis results indicate a frontogenetic region along the developing side-door cold front in eastern Massachusetts (Fig. 18c). A similar situation exists for geostrophic frontogenesis, with a narrow frontogenetic band in Massachusetts embedded in a generally frontolytic region. The weakening remains of the back-door front are still reflected by a region of positive observed wind frontogenesis over southern Pennsylvania. The tongue of high-$\theta_e$ air located along the front at 0600 UTC was centered over Connecticut at 1200 UTC (Fig. 19d). This corresponds to a position behind the back-door front, and immediately ahead of the side-door front. As was the case at 0600 UTC, a gradient in $\theta_e$ was in place along the coast from eastern Massachusetts to Cape Hatteras.

Cold-air advection predominated from Pennsylvania to New England at 1200 UTC, with a pronounced maxima of −7°C (3 h)⁻¹ over coastal Massachusetts behind the side-door cold front. This reflects the advection of the relatively cold maritime air over the ocean into the relatively warm air behind the back-door cold front over land. The side-door cold-frontal boundary is also diagnosed to be convergent (Fig. 19b) with values of $8 \times 10^{-5} \text{s}^{-1}$, and it is associated with a narrow band of cyclonic vorticity (Fig. 19a). Behind the front there are large anticyclonic-vorticity values of order $-1 f$, in the region where rapid anticyclogenesis
Fig. 10. Visible satellite images for (a) 1400 UTC, (b) 1600 UTC, (c) 1800 UTC, and (d) 2000 UTC 22 April 1987. The black line indicates the location of the side-door cold front.
Table 1. Propagation speed of the side-door cold front to the nearest 1 m s\(^{-1}\).

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<th>Time (UTC)</th>
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took place during the following 6 h. Large cyclonic vorticity values east of Cape Hatteras are associated with the quasi-stationary cyclone.

3) 1800 UTC 22 APRIL 1987

As the side-door front made landfall in New Jersey at 1800 UTC, the baroclinic zone intensified and collapsed into a narrow zone, as reflected by a 30°C \( \theta_e \) gradient over 100 km in southern New Jersey (Fig. 20d). From a Lagrangian point of view, \( \theta_e \) can be changed only by sensible heating, so values near 330 K over land developed in response to the diurnal heating cycle, assuming this air remained near the surface. Sensible cooling from the sea surface also probably eliminated the \( \theta_e \) maximum that was over Connecticut at 1200 UTC, such that as in the 0600 and 1200 UTC results, values below 308 K extend southward along the coastal waters.

Another reflection of increased frontal baroclinity at 1800 UTC is the cold advection maximum along the New Jersey coast (Fig. 20c). Values of \(-16°C (3 h)^{-1}\) agree quite well with the observed temperature falls. Weaker cold advection extends southward along the coast to North Carolina. The vorticity field (Fig. 20a) indicates weak cyclonic-vorticity values near \(8 \times 10^{-1} \text{ s}^{-1}\) along the side-door front in western New Jersey and off the Delaware coast. A portion of the front in extreme southern New Jersey and Delaware is characterized by significant anticyclonic vorticity in the region of maximum \( \theta_e \) gradient (Fig. 20d). This anticyclonic relative vorticity was probably maintained by divergence (Fig. 20b). The same region was also quite frontolytic in the geostrophic sense (Fig. 21b), and marginally frontolytic with the observed winds (Fig. 21a). To the north, the side-door front is associated with convergence values of \(8 \times 10^{-4} \text{ s}^{-1}\). The front was also characterized by an increase in both observed wind and geostrophic frontogenesis since 1200 UTC, with values exceeding 14 K (100 km\(^{-1}\) (3 h\(^{-1}\) in the region of maximum \( \theta_e \) gradient. Although not evident in the geostrophic frontogenesis results, the back-door front (located along the New York–Pennsylvania border) is reflected in the observed wind frontogenesis results.

4) 0000 UTC 23 APRIL 1987

At 0000 UTC a band of \( \theta_e \) values exceeding 320 K preceded the front (Fig. 22d). The gradient of \( \theta_e \) at the front is less than at 1800 UTC, reflecting a modification of the shallow marine air mass by surface heating. Equivalent potential temperature values near 320 K are along the leading edge of the front, less than at 1800 UTC. Values of \( \theta_e \) over the coastal waters south of Long Island remained nearly constant in the range 292–296 K since 0600 UTC. Observed wind frontogenesis is maximized along the front from Pennsylvania into northern Virginia (Fig. 20c), but is much less than the frontogenesis observed along the coast at 1800 UTC. Geostrophic frontogenesis (Fig. 20d) reflects a similar trend. This modification is also reflected in the temperature advection results, which indicate maximum cold advection values of \(-8°C (3 h)^{-1}\) in central Maryland (Fig. 22c). Convergence at the front is weaker than observed at 1800 UTC. A significant majority of the front is characterized by small anticyclonic relative vorticity. This is a reflection of the highly ageostrophic nature of the wind flowing down the pressure gradient, resulting in little wind shift across the front.

b. Density-current considerations

The preceding results suggest that the onshore push of cooler air resembled a density current. This idea is explored further in this section. For a given frontal speed \( c \) and postfrontal fluid speed \( U_p \), a density current must exhibit positive front-relative flow, \( U_p - c > 0 \) (Smith and Reeder 1988). Table 1 gives the speed of the side-door cold front over both ocean and land. An approximately 120-km-wide section of the front for each time period was used to compute a representative frontal speed over the ocean along a line from eastern Massachusetts to the outer banks of North Carolina. A similar approach was used for the over-land portion of the front along a line from central New Jersey through the Washington D.C. area. While the front was over the ocean, from 1200 to 1800 UTC, \( c = 19 \pm 2 \text{ m s}^{-1}\), and while over land during 1700–0000 UTC, \( c = 13 \pm 1 \text{ m s}^{-1}\). Over the ocean, direct observations of the postfrontal wind speed are sparse, and do not allow for an estimation of front-relative flow. Over land, many stations reported surface winds sustained at 12 m s\(^{-1}\) with gusts of 15–20 m s\(^{-1}\) behind the front (Fig. 1b). However, as illustrated in the WAL and ACY soundings (launched several hours after the frontal passage) (Fig. 3), the core of the strongest winds (14 m s\(^{-1}\)) was several hundred meters above the ground. Thus, the peak winds observed at the surface behind the front probably provide a reasonable estimate of the sustained wind speed above the surface layer.
Exposed coastal locations also experienced somewhat higher sustained winds (e.g., BUZ3M reported sustained winds of 17 m s\(^{-1}\) behind the front for several hours). Hence, over land observed front-relative flow was approximately 3–7 m s\(^{-1}\). Given a frontal speed over the ocean of approximately the same magnitude as \(U_p\), this suggests that while the front was over the ocean it was not a density current.

Several versions of the empirical formula governing the propagation speed of density currents have been applied by various authors (e.g., Carbone 1982; Hobbs and Perras 1982). A form of the speed equation that includes the contribution of the ambient flow \(U\) given by Seitter (1986) is

\[
V = k \left( \frac{\Delta P}{\rho_w} \right)^{1/2} + 0.61U,
\]

where \(V\) is the velocity of the front, \(\Delta P\) is the hydrostatic surface-pressure difference across the front, and \(\rho_w\) is the density of the warm air. Seitter (1986) found a value of 0.79 for the Froude constant \(k\) by linearly fitting values from 20 density-current cases. At IAD,
in the warm air, the mean virtual temperature in the lowest 500 m is 297 K. Inspection of nine microbarograph traces (in New Jersey, Pennsylvania, Maryland, and Virginia) indicated a two-stage pressure rise, with the first stage lasting between 1 and 2 h. These stages corresponded to drops in temperature as well. Using a 2-h period after the frontal passage at these stations yielded a mean pressure rise of 2.8 ± 0.8 mb. This value was used for ΔP in (5), with the assumption that the pressure rise was primarily hydrostatic (Wakimoto 1982). Use of these data, along with a 2.9 m s⁻¹ prefrontal environmental wind in the direction of the front (nine station average over 6 h), results in a computed velocity of 13.9 ± 1.9 m s⁻¹. This compares quite well with the analyzed surface velocity of 13 m s⁻¹ from 1800 to 0000 UTC.

A hydrostatic estimate of the surface pressure rise behind the front (Schoenberger 1984) is

$$\Delta P = \frac{Pg \Delta Z \Delta T_v}{RT_{sw} T_{wc}},$$  \hspace{1cm} (6)

where ΔZ is the depth of the cold air, $T_{wc}$ the mean virtual temperature of the cold air, $T_{sw}$ the mean virtual temperature of the warm air, g gravity, and R the gas constant for dry air. From the 0000 UTC WAL sounding, ΔZ is found to be 550 ± 75 m, with a $T_{wc}$ of 284 K. Using an environmental pressure $P$ of 1015 mb in (6) yields a pressure rise of 2.9 ± 0.3 mb results. This agrees well with the observed mean value of 2.8 ± 0.8 mb from the microbarograph array. Thus, there appears to be good agreement between density-current theory and observed properties for this case.

5. Concluding discussion

A study of a shallow nonprecipitating cold front that affected portions of the east coast of the United States on 22 April 1987 has been presented. This event produced 10°C temperature falls in less than 2 h, with wind gusts to 20 m s⁻¹. In addition, sky conditions deteriorated from scattered cumulus clouds to a low stratus overcast with the frontal passage. This front has been termed a “side-door” cold front since the cold air surged westward from off the ocean. The side-door front developed behind a preexisting back-door cold front that had moved southeastward out of Canada. The abrupt transition in weather conditions was essentially unforecasted. Forecasts for the Washington D.C. area obtained from the Washington Post called for variable cloudiness with light winds and high temperatures of 24°–26°C. In reality, the high temperature was around 30°C, and gusty (15–20 m s⁻¹) winds were experienced late in the afternoon as temperatures fell to near 10°C. Events of this type have the potential to produce hazards to aviation due to increased low-level wind shear. In fact, an “aircraft mishap” reported at 2006 UTC at Dover Air Force Base may have occurred as a result of the passage of the side-door cold front around 1900 UTC. Thus, an understanding of this event would be important for all aviation interests along the heavily traveled corridor from Boston to New York City to Washington D.C.

The associated 500-mb flow pattern was found to be quite similar to the climatological mean for back-door cold fronts established by Bosart et al. (1973). A large ridge throughout the troposphere, from the Gulf of Mexico to the Great Lakes, maintained warm dry weather over land while a stationary cyclone east of North Carolina produced northerly winds along the cold coastal waters. The passage of a strong 500-mb short-wave trough in the westerlies north of New England triggered anticyclogenesis in its wake over the Gulf of Maine, and allowed the cold air to penetrate southwestward into a region of increasing easterly flow between the anticyclone and the stationary cyclone east of North Carolina. Thus, it appears that the evolution of the large-scale flow was crucial to the development and propagation of the side-door cold front. The rapid
development of a surface anticyclone over the Gulf of Maine in the wake of a midtropospheric transient short-wave trough contributed to the southwestward expansion of the cold air and to its ensuing inland penetration in the mid-Atlantic area.

The side-door front formed along the New England coast behind a preexisting back-door cold front. The low-level baroclinity associated with the back-door front underwent a significant alteration as it reached the coast. Since air over the cold ocean water was chilled to 4°C, air ahead of the front was initially 10°C colder than the continental air behind the back-door cold front. Consequently, a baroclinic zone remained fixed along the coastline, with a trough of low pressure aligned along the coastal plain coincident with a tongue of warm air after the back-door frontal passage. As the surface pressure rose over the Gulf of Maine, an increasing northeasterly flow produced enhanced convergence along the coastal trough/baroclinic zone resulting in frontogenesis and a southwestward propagation of the coastal baroclinic zone (the side-door cold front). The fact that this process resulted in the formation of a significant surface front in less than 4 h suggests that the semigeostrophic model of frontogenesis (e.g., Hoskins and Bretherton 1972) may be inadequate for this case. A temperature gradient was es-
FIG. 14. Visible satellite image (1-km resolution) for 2100 UTC 22 April 1987. The side-door cold front is located along the narrow cloud band that extends from the Chesapeake Bay north and east into northern New Jersey, highlighted by a black line.

FIG. 15. Skew $T$–log$p$ format soundings for (a) Atlantic City, New Jersey (ACY), and (b) Wallops Island, Virginia (WAL). Solid (dashed) lines represent soundings at 1200 UTC (0000 UTC) 22 (23) April 1987.
established due to differential heating across a coastline rather than by synoptic-scale deformation. In addition, the large degree of ageostrophy in the wind field for considerable distances behind the front throughout its life history reflected the absence of geostrophic balance in the cross-front pressure-gradient force.

Roughly 8 h after forming along the New England coast, the side-door cold front made landfall in New Jersey. Observations at a buoy 50 km offshore indicated little baroclinity across the front over the ocean with prefrontal air temperatures of 8.6°C and only a 1°C temperature fall with the passage of the front, and a 1 mb pressure rise. At the same time, over-land temperatures had reached 25°C–30°C, resulting in a 15°C–20°C temperature difference across the coastline. After landfall, the front was observed to have positive front-relative flow, indicative of a density current. In addition, a computed frontal speed of 13.9 m s⁻¹ from the density-current speed equation compared favorably with a mean observed speed of 13 m s⁻¹. These observations, in conjunction with surface mesoanalyses and satellite imagery indicated:

1) the side-door front was weakly baroclinic over the ocean to the south of Long Island, with the depth of the cold air being little more than the depth of the marine boundary layer;
2) a strong baroclinic zone was established well in advance of the front due to differential sensible heating across the coastline;
3) as the front reached the coast, its cross-front temperature gradient collapsed to a very narrow zone due to frontal convergence in the preexisting coastal baroclinic zone; and
4) the contraction of the frontal zone was associated with a density-current structure as the front surged inland.

These facts illustrate that the weakening of the cross-front temperature gradient over the ocean and its subsequent rapid intensification as the front reached the New Jersey coast are local effects that altered the frontal structure on very short time scales. So, although the large-scale flow was responsible for the general movement of the low-level cold air, local small-scale processes were responsible for the intensification of the front. Although frontal circulations were not evaluated for this case, the data suggest that frontogenesis occurred in a very shallow layer and was apparently not associated with a deep tropospheric secondary circulation, as is the case for balanced theories of frontogenesis based on deformation of the low-level thermal field by the large-scale geostrophic flow.

Prior to 1700 UTC 22 April, it is not possible to state conclusively whether the side-door cold front possessed a density-current structure, due to the lack of oceanic observations. However, a density-current structure may have been present over New England. With an observed frontal acceleration to near 20 m s⁻¹ over the oceanic waters south of New England it is unlikely that there was significantly positive front-relative flow, since the postfrontal flow was of the same magnitude. Beginning at 1800 UTC, one branch of the cold air continued southward along the coastal waters in a northeasterly flow regime to the west of a stationary offshore cyclone. A second branch of the cold air turned westward and crossed the New Jersey coast as a very narrow zone of intense baroclinity (qualitatively similar to 1200 UTC along the New England coast). The propagation speed of this portion of the front was reduced below the ambient speed of air behind the front, indicative of a density current. Increased surface friction over land, as well as surface sensible heat fluxes, may have played a crucial role in decelerating the shallow cold front below the velocity of the low-level current.

Soundings behind the front indicate that the cold air had a depth of less than 600 m. The shallow nature of the front was also revealed by its retarded motion as it reached hills with an elevation less than 600 m. Within this shallow cold-air mass, low-level wind speeds were on the order of 15–20 m s⁻¹, resulting in convergence along the front sufficient to support ascent to form a line of nonprecipitating cumulus clouds oriented along the front. This observation is consistent with other studies, where vigorous vertical motions along the head of density currents was sufficient to form a cloud band (e.g., Settet and Muench 1985). Surface
Fig. 17. Gridded surface data computations at 0600 UTC 22 April 1987 of (a) relative vorticity (every $4 \times 10^{-5}$ s$^{-1}$), (b) horizontal divergence (every $4 \times 10^{-5}$ s$^{-1}$), (c) horizontal temperature advection [every $1^\circ$C (3 h)$^{-1}$], and (d) equivalent potential temperature (every 4 K). All negative values are dashed.
FIG. 18. Observed wind [(a) and (c)] frontogenesis and geostrophic frontogenesis [(b) and (d)] at 0600 UTC [(a) and (b)] and 1200 UTC [(c) and (d)] 22 April 1987. Contours are every 2°C (100 km)⁻¹ (3 h)⁻¹, with negative values dashed.
Fig. 20. As in Fig. 17 except for 1800 UTC 22 April 1987.
Fig. 21. As in Fig. 18, except for 1800 UTC 22 April 1987 [(a) and (b)] and 0000 UTC 23 April [(c) and (d)] 1987.
FIG. 22. As in Fig. 17 except for 0000 UTC 23 April 1987.
observations of the rapid temperature falls, pressure rises, and strong winds with the passage of the event are also consistent with other observational studies of atmospheric density currents.

Using a two-dimensional model, Reeder (1986) showed that an oceanic front approaching a coastline propagated faster and was more intense when heating effects over land were included. Sea-breeze convergence ahead of the front accompanied a separate temperature gradient maximum. When the front reached the coast there was a contraction in the scale of the frontal zone, as a result of rapid frontogenesis in a region of enhanced convergence. Physic (1988) performed a similar study and found that a trough developed over land due to diurnal heating. Frontogenesis occurred at this trough producing a new front, which appeared to be an acceleration of the original offshore cold front. Over land, the front propagated slower, such that the front took on characteristics of an unsteady density current with positive front-relative flow. In the case studied here, the thermal gradient along the coastline during the afternoon of 22 April apparently did not result in the development of a new front, but intensified the crossfront thermal gradient of the existing side-door cold front as it reached the coast, which had weakened over the homogeneous oceanic PBL.

An apparently analogous event to the one under study is the "southerly buster" of Australia. In a modeling study, Baines (1980) concluded that the southerly buster formed due to a blocking mountain range oriented perpendicular to it. Another similar event is the southerly surge of marine air along the west coast of the United States (Dorman 1985; Mass et al. 1986; Mass and Albright 1987). Mass and Albright (1987) found that some of these surges are topographically trapped density currents formed by an alongshore pressure gradient. This alongshore pressure gradient creates ageostrophic downgradient flow along the coast as a result of synoptic-scale flow interacting with the coastal mountains. Thus, a significant difference between the side-door cold front and these events (including back-door cold fronts) is that the postfrontal downgradient flow is a result of isallobaric and frictional effects rather than physical blocking of the flow. Based on the observations of the behavior of this front over the ocean and land, it is hypothesized that differential heating and friction across the coastline contributed to a contraction in the horizontal scale of the crossfront temperature gradient as well as the speed of the front.

There have been very few studies of large-scale density-current cold fronts that are not associated with mountains, which block the postfrontal cold air resulting in the necessary unbalanced cross-front pressure-gradient force. Smith and Reeder (1988) concluded, "Except, possibly, where cold fronts become orographically trapped, there is little evidence to suggest that frontal speed is controlled by gravity current dynamics rather than by processes that operate on the frontal scale itself, where the ageostrophic crossfront circulation plays a central role." In the case at hand, it is not possible to state conclusively whether density-current dynamics completely determine the frontal speed. Although there is agreement between observed frontal velocity and the density-current velocity, this does not necessarily imply that density-current dynamics fully account for the speed of the front and that the importance of the large-scale flow to the movement of this front has been emphasized.

In a modeling study of a southerly buster event, Howells and Kuo (1988) found that the event was captured without topography in the model. They also conclude that "Analysis of the model results show that differential land-sea heating and friction between continental Australia and the Tasman Sea are capable of generating a realistic coastal southerly jet without the blocking and channeling effects of the Great Dividing Range." In a similar study, Garratt et al. (1989) also found that the frictional and diabatic effects in the vicinity of a coastline are sufficient to generate a front of this type. Although the shape of the coastline and other factors are quite different in this case, it appears the results of this study may provide observational evidence of the modeled results.

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