Mesoscale Coastal Processes during GALE IOP 2

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(Manuscript received 7 April 1989, in final form 1 September 1989)

ABSTRACT

During Intensive Observation Period 2 of the Genesis of Atlantic Lows Experiment, a number of mesoscale phenomena were observed with special and conventional observing systems over the land and coastal waters. This study involved analysis of these data for the period 24–26 January 1986 in order to define the structure and dynamics of three features: the coastal front; a shallow cyclone that propagated along the coastal front, modifying it as it moved northward; and a low-level jet that formed in the strong coastal pressure-gradient field.

The coastal front formed in an existing pressure trough over the Gulf Stream as a result of both ageostrophic deformation and differential diabatic heating. There existed considerable variability in the frontal strength and position on both the mesoscale and mesoscale. The level of strongest frontogenesis was near the surface, with frontolysis calculated above 950 mb.

The marine atmospheric boundary layer (MABL) over the Gulf Stream was conducive to cyclone formation. Latent and sensible heat fluxes of up to 800 W m⁻² and 400 W m⁻², respectively, were calculated early in the study period, and a deep, moist conditionally unstable boundary layer was present. Calculation of the vorticity tendency associated with the sensible heating yielded a narrow band of positive values to the east of the coastline. As a weak midtropospheric wave reached this favorable region to the east of Florida, a shallow cyclone formed along the coastal front. As the cyclone tracked northeastward along the front, geostrophic deformation ahead of it strengthened the front while strong cold-air advection to its rear displaced the coastal front to the east, leaving behind a dry, stable MABL with low-level, cold-air advection and weak descent. As the cyclone moved northward along the front, conditionally unstable, moist, low-level air ahead was forced by the southeasterly flow to rise along the coastal front and its extension over the cold air near the coastline, causing enhanced precipitation.

A low-level northeasterly jet was also observed over the Carolinas, and formed as a result of the strong low-level pressure gradient created by the proximity of the cold continental air over land and the warm air of the Gulf Stream MABL near the coast. This jet, with a maximum near 960 mb, showed a diurnal variation of up to 20 m s⁻¹, which likely resulted from day/night variations in mixing at jet level, an inertial oscillation associated with the frictional decoupling of the low-level flow at sunset, and isallobaric accelerations.

1. Introduction

Cyclogenesis can result from a variety of distinct or cooperative dynamic mechanisms and complex scale interactions, where the existence of spatial maxima in the global distribution of cyclogenesis frequency (Petersen 1956) points to the importance of local forcing. Well-documented local factors include mountain ranges, coastal regions and sea surface temperature contrasts. Often, two or more of these effects act in concert along the East Coast of the United States. The coastal front, an intense mesoscale front that forms near the coastal baroclinic zone, is one phenomenon that may be an important ingredient in coastal cyclogenesis. Functioning as a source of low-level positive vorticity and moisture, the coastal front can focus processes operating in the low levels to initiate and enhance cyclogenesis. Kocin and Uccellini (1984) found that coastal fronts were present in 13 of 18 cases of major East Coast cyclogenesis and often are an important component in the cyclogenesis process. Though studied rather extensively, coastal frontogenesis and the interaction with the synoptic scale and mesoscale environments are still not completely understood, in part due to the lack of data on scales sufficient to resolve the important processes. The Genesis of Atlantic Lows Experiment (GALE), conducted from 15 January through 15 March 1986, used an array of special measurement platforms to provide a high temporal- and spatial-resolution dataset for the investigation of a variety of processes and phenomena associated with East Coast cyclogenesis. The objectives and an overview of the field project can be found in Dirks et al. (1988) and SethuRaman and Riordan (1988).

During the second Intensive Observation Period (IOP 2), which extended from 23 January to 29 January 1986, a variety of meteorological phenomena were observed, including intense coastal frontogenesis, a diurnally varying low-level jet (LLJ), and mesoscale cyclogenesis along the coastal front. In this study, me-

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soscale analyses and diagnostic calculations will be used to document the evolution, structure and interaction of these phenomena. This work serves as the foundation for ongoing research using the Penn State/National Center for Atmospheric Research (PSU/NCAR) mesoscale model to address questions pertaining to the dynamics of these coastal processes.

The coastal front, characterized by a cyclonic wind shift and a large thermal contrast, is a shallow (less than 500 m deep) boundary separating warm, moist oceanic air and cold, dry continental air. More frequent in the winter months, coastal fronts have been observed in many regions including New England (Bosart et al. 1972; Bosart 1975; Sanders 1983), the Carolinas (Bosart 1981; Bosart and Lin 1984; Riordan et al. 1985; Forbes et al. 1987; Stauffer and Warner 1987; Keshishian and Bosart 1987), the Gulf Coast of Texas (Bosart 1984) and the Black Sea coast (Draghi 1984). Often in the eastern United States the easterly flow of highly stable cold air associated with a surface anticyclone in the New England states, is blocked by the Appalachian Mountains and is funneled southward parallel to the mountains, enhancing the coastal baroclinic zone. This blocking of shallow cold air, or cold-air damming, is manifested as a persistent, narrow, inverted surface pressure ridge to the east of the Appalachians. The coastal front typically coincides with the eastern edge of the cold dome at the surface and slopes gradually westward with height, rising over the entrenched cold air. Thus, cold-air damming can play an important role in the frontogenesis and cyclogenesis processes along the East Coast, as documented by Baker (1970), Richwine (1980), Forbes et al. (1987), Stauffer and Warner (1987), and Bell and Bosart (1988).

Differential surface roughness, irregular orography, the land–sea temperature contrast, and coastal configuration have been hypothesized as important mechanisms contributing to coastal frontogenesis (Bosart et al. 1972; Bosart 1975). The importance of differential diabatic heating for coastal frontogenesis has been stressed in the numerical investigation of Ballentine (1980) and in the observational studies of Bosart (1981, 1984) and Bosart and Lin (1984). Geostrophic deformation, though frequently important in the early stages of synoptic scale frontogenesis, is seldom responsible for initiating coastal frontogenesis. In the case of the coastal front, ageostrophic processes are crucial for the initiation and eventual collapse of the front (Bosart 1975; Bosart and Lin 1984; Keshishian and Bosart 1987). However, in some special situations when low-level flow parallels the coastal front, and the anticyclone to the north is absent, geostrophic deformation may play a more substantial role. Specifically, as weak mesoscale disturbances propagate along the front, geostrophic deformation strengthens the preexisting baroclinic zone ahead of the weak cyclone, which then preconditions the low-level environment for later cyclogenesis events (Clark 1983; Bosart 1984; Keshishian and Bosart 1987). Since the baroclinic zone is tightened ahead of the cyclone, Clark (1983) refers to these disturbances as "zipper" lows. In addition to the low-level processes operating near the coast, the ascending branch of a thermally indirect circulation associated with the exit region of an upstream jet streak may intensify low-level warm advection in the Carolinas (Uccellini et al. 1984, 1987), potentially enhancing the coastal frontogenesis and cyclogenesis processes.

In this paper, diagnostic calculations and detailed mesoscale analyses are used to investigate coastal frontogenesis, mesoscale cyclogenesis along the coastal front, and a low-level jet observed during GALE IOP 2. The methodology is described in section 2, followed by a synoptic scale overview in section 3. Sections 4 and 5 contain a description of the mesoscale structure of the coastal front and the coastal cyclogenesis event. The structure and evolution of the low-level jet are described in section 6 and the conclusions are provided in section 7.

2. Data analysis procedures

A variety of National Weather Service (NWS) and GALE surface and upper-air data are used in this study. The GALE dataset is discussed in detail by Mercer and Kreitzberg (1986), Dirks et al. (1988) and Sethu-Raman and Riordan (1988). A description of the GALE instruments and the characteristic instrument errors can be found in GALE (1985). The upper-air sounding data are obtained from Cross-chain Loron Atmospheric Sounding System (CLASS) rawinsondes, miniradiosondes, special GMD and VIZ rawinsondes, Omega dropwindsondes, and the conventional (NWS) rawinsondes. The conventional data used in this study were obtained from sites operated by the NWS, the military, and the Federal Aviation Administration (FAA). In addition, NOAA buoys, NOAA instrumented platforms including Coastal Marine Aids to Navigation (C-MAN) towers, and ships-of-opportunity provided marine measurements. The special GALE surface data were obtained from 51 Portable-Automated-Mesonet (PAM) II sites, eight instrumented GALE research buoys, and two oceanographic research vessels (R/V) (the R/V Endeavor and the R/V Cape Hatteras). A map of the various surface and upper-air stations used in the study is shown in Fig. 1.

In order to allow the use of consistent winds from a variety of marine observing systems, wind speeds measured at platforms of known height have been adjusted to a reference height of 10 m by the North Carolina State University (NCSU). The adjustment method, described by Riordan (1987), makes use of a wind-speed dependent drag coefficient and the log-wind law. Because the original GALE buoy data showed too much high-frequency variability in the wind directions due to the instantaneous sampling and buoy motion, NCSU applied a 5-point time filter to these data (Riordan 1987).
For the analysis of mesoscale pressure perturbations, such as associated with the coastal front and small-scale cyclones, it is extremely important to obtain spatially consistent sea level pressure fields. In this study, a two-stage procedure was employed to check for bias errors in the pressure data. First, the PAM data were compared with the NOAA and FAA data to isolate measurement-system differences. Then individual sites were checked for error. To begin the process, the PAM surface pressures were reduced to sea level. For each PAM site, these 5-minute data were averaged for a 24-h period on the meteorologically quiescent day of 23 January 1986. These average PAM pressures were then corrected for biases by careful subjective comparison with objective and subjective analyses of similarly time-averaged NOAA and FAA sea level pressure observations for the same period. In the second step, hourly sea level pressure observations from PAM and all other stationary marine and land sources (with the exception of the NOAA buoys) were temporally averaged for the 120-h period from 0000 UTC 24 January through 0000 UTC 29 January 1986. A smooth objective analysis (Barnes 1973) of these averaged sea level pressures was then used to calculate a correction at each station necessary to provide consistency with this analysis. Several ships-of-opportunity considered crucial for the analysis were also corrected for systematic pressure errors by subjective comparison with nearby observations.

The upper-air data were error checked and interpolated to constant pressure surfaces at 10-mb intervals by the Gale Data Center. However, interpolated data were not used here when the original data levels were greater than 150 mb apart. The temperature, geopotential height, winds, and dewpoint depression were then objectively analyzed onto a 44-km grid mesh for times when both the GALE inner and regional areas were available. For a finer scale, 20-km grid mesh for the times when only the inner region observations were available. Isobaric analyses were produced every 20 mb from the surface to 100 mb using a modified Barnes (1973) objective analysis scheme, which employs a spatially varying weighting function that depends upon the data density. Upper-air time sections were also generated by use of the interpolated 10-mb soundings.

NWS radar films, radar mosaics, satellite imagery, and the GOES-6 GALE videotape from the University of Wisconsin were also used to help define specific mesoscale features. The sea surface temperature (SST) field used in this study was derived from the NOAA-9 infrared imagery. This high resolution 14-km analysis has a root-mean-square error of 0.5°-0.7°C and a bias error of less than ±0.5°C (Bill Pichel, personal communications, 1989). It is capable of resolving the strong gradients associated with the western edge of the Gulf Stream.

Surface, oceanic sensible and latent heat fluxes, \( H_s \) and \( H_L \), respectively, are diagnosed using the standard bulk aerodynamic formula

\[
H_S = \rho C_p C_H (T_{sea} - T_{air}) |V| \\
H_L = \rho L_e C_F (q_l - q) |V|,
\]  

where \( \rho \) is the air density, \( C_p \) is the specific heat of air at constant pressure, \( T \) is temperature, \( q \) is the specific
humidity, \( q_s \), is the saturation specific humidity at the SST, \( |V| \) is the wind speed, and \( L_e \) is the latent heat of vaporization. The exchange coefficients \( C_H \) and \( C_E \) are based on the formulations described by Frihe and Schmitt (1976) and applied by Riordan et al. (1985). The NOAA 14-km SST analysis was used for the heat-flux calculations.

An estimate of the local vorticity tendency due to sensible heating is based on Petterssen's development equation (Danard and Ellenton 1980) and is computed by:

\[
\frac{\partial \xi}{\partial t} = -\frac{R_d g \ln \left( \frac{p_0}{p_{LND}} \right)}{fC_p} \Delta p \nabla^2 H_S, \tag{2}
\]

where \( \xi \) represents the 1000-mb relative vorticity, \( p_0 \) is 1000 mb, \( p_{LND} \) is the pressure at the level of non-divergence estimated as 500 mb, \( R_d \) is the gas constant for dry air, \( f \) is the Coriolis parameter, \( g \) is the acceleration of gravity, \( H_S \) is the sensible heat flux from (1), and \( \Delta p \) is the depth of the heating. The sensible heating is assumed to be confined to a 150-mb deep layer, a depth estimated using soundings from coastal locations, research vessel soundings, and dropwindsondes. The 1000-mb height tendency associated with sensible heating can then be obtained from the vorticity tendency, assuming the vorticity is equal to the geostrophic vorticity (Bosart 1984).

To investigate the kinematics of the coastal frontogenesis, the gradient of equivalent potential temperature \( (\theta_e) \) is used. It reflects both the temperature and humidity of the air, and is a suitable measure of the coastal-front intensity. The frontogenesis function is defined as

\[
\frac{d}{dt} \left| \nabla \theta_e \right| = \left| \nabla \theta_e \right|^{-1} \left\{ - \left[ \left( \frac{\partial \theta_e}{\partial x} \right)^2 \frac{\partial u}{\partial x} + \left( \frac{\partial \theta_e}{\partial y} \right)^2 \frac{\partial v}{\partial y} \right] \right. \tag{A}
\]

\[
- \left[ \frac{\partial \theta_e}{\partial x} \frac{\partial \theta_e}{\partial y} \left( \frac{\partial v}{\partial x} + \frac{\partial u}{\partial y} \right) \right] \tag{B}
\]

\[
- \left[ \left( \frac{\partial \theta_e}{\partial x} \frac{\partial \theta_e}{\partial y} \frac{\partial \omega}{\partial x} + \left( \frac{\partial \theta_e}{\partial y} \frac{\partial \theta_e}{\partial p} \right) \frac{\partial \omega}{\partial y} \right] \tag{C}
\]

\[
+ \left[ \frac{\partial \theta_e}{\partial x} \frac{\partial}{\partial x} \left( \frac{d \theta_e}{dt} \right) + \frac{\partial \theta_e}{\partial y} \frac{\partial}{\partial y} \left( \frac{d \theta_e}{dt} \right) \right] \right\}, \tag{3}
\]

where \( \omega \) is the vertical velocity, and \( u \) and \( v \) are the horizontal wind components. Terms A and B represent the contribution of the deformation field to frontogenesis. Term C, the twisting term, is neglected at the surface because vertical motions are assumed to be small. Term D represents the frontogenesis due to horizontal variations in diabatic heating. The contribution of sensible heat fluxes and moisture fluxes to frontogenesis was estimated over the ocean based upon the vertical flux convergence calculated for a marine boundary layer of an assumed depth of 150 mb. The frontogenesis contribution from geostrophic deformation is found by substituting the components of the geostrophic wind into terms A and B in (3). Vertical velocities were calculated using a mesoscale omega equation similar to that used by Tarbell et al. (1981). The vertical velocity was assumed to vanish at the surface and 100 mb. Latent heat effects were incorporated in the omega equation using a moisture-corrected static stability, following Hirschberg (1989).

3. The synoptic scale setting

At the beginning of the study period, a massive surface anticyclone dominated the eastern one-third of the United States. Air flowing around the southern portion of this anticyclone, centered near southern Quebec at 0000 UTC 25 January (Fig. 2a), is modified by heat and moisture fluxes within the marine atmospheric boundary layer (MABL) as it approaches the southeast coast of the United States. At this time, an inverted sea level pressure ridge reflects the wedge of entrenched cold air that is positioned to the east of the Appalachian Mountains. Near the coastline, this ridge adjoins a trough that is associated with the coastal front, where the trough extends from Florida northward along the Carolina coastline and is nearly coincident with the shoreward side of the Gulf Stream. The coastal front is rather disorganized at this time, and several distinct mesoscale fronts are probably present. The wedge ridge remains established east of the Appalachian Mountains throughout 25 January (Figs. 2b and 2c), as the surface anticyclone moves eastward. The coastal front, now over 1000 km long, extends from Florida to New Jersey and is advancing westward into North Carolina with an intensification of the baroclinic zone. As the surface anticyclone continues to move offshore, warm moist air modified by the Gulf Stream continues to be advected toward the East Coast at low levels.

During 25 January, a mesoscale wave forms along the coastal front off the Georgia/Florida coast. Eventually this disturbance has a significant impact upon the coastal front evolution and subsequent disturbances that form along the coast during IOP 2. The cyclone, located just offshore of South Carolina by 0000 UTC 26 January (Fig. 2c), continues to track rapidly northwestward along the front, with heavy precipitation in advance of it and drier westerly flow to its rear. In the wake of the cyclone, the coastal front and associated baroclinic zone move eastward (Fig. 2d), perhaps delaying and disrupting ensuing cyclogenesis. A major purpose of this paper is to document not only the structure, evolution and mesoscale features associated with the coastal front, but also to describe the interaction of the front and this cyclone. Two other more powerful cyclones form along the trailing baroclinic
zone after an arctic air mass surges into the eastern United States.

The corresponding fields of 850-mb and 500-mb geopotential height, temperature, and wind are shown in Figs. 3 and 4, respectively. Throughout 25 January, pronounced warm advection occurs in the lower troposphere along the Carolina coast with prominent ageostrophic winds observed at 850 mb (Figs. 3a-c). An 850-mb wave is seen in the windfield and massfield along the coastline in the vicinity of the surface disturbance in the 0000 UTC and 1200 UTC 26 January analyses. This wave is most easily identified at 500 mb when it is located over Louisiana at 1200 UTC 25 January. As it propagates eastward it is difficult to follow, except in the height tendency field, but a weak trough is seen in the windfield over Georgia and South Carolina at 0000 UTC 26 January.

4. Mesoscale structure of the coastal front

The mesoscale conditions at the surface at 0000 UTC 25 January are illustrated in Fig. 5a. The stippled regions correspond to composites of the radar images from Cape Hatteras, NC, Wilmington, NC, Charleston, SC, Athens, GA, and Volens, VA. An inverted pressure ridge with the characteristic northerly ageostrophic surface flow, associated with the cold air dammed to the east of the Appalachians, is well defined at this time. This northerly ageostrophic flow funnels cold air southward to the east of the Appalachian Mountains, enhancing the low-level coastal baroclinic zone. The NOAA 14-km sea surface temperature analysis for 0000 UTC 25 January is shown in Fig. 6. The strong sea surface temperature gradient located along the western edge of the Gulf Stream is nearly coincident.
with the coastal atmospheric baroclinic zone at this time. Surface wind speeds are observed to be approximately 10 m s\(^{-1}\) greater over the shelf-water east of the Carolinas than over land. This may be a result of the lower surface stress over water, as well as the greater sea level pressure gradient to the east of the coastline, associated with the strong, coastal baroclinic zone. Also, soundings along the Carolina coast at 0000 UTC 25 January indicate wind speeds in excess of 10 m s\(^{-1}\) between 900 and 970 mb. Thus, stronger momentum mixing in the unstable boundary layer in the vicinity of the Gulf Stream may result in higher surface wind speeds as documented by Sweet et al. (1981). Farther offshore, a trough in the sea level pressure field is located east of the Carolinas. This weakens slightly to the south, before becoming more well defined again off the coast of Florida; a trend also supported by the low-level cloud structures in the satellite imagery. This trough is discernible from approximately 1400 UTC 24 January, though the determination of the actual time and location of the formation is difficult because of sparse oceanic surface data. A strong cyclonic wind shift and intensifying baroclinic zone, apparent since at least 1700 UTC 24 January, are associated with a portion of the pressure trough to the east of the Florida coast. In this trough, the southern extension of the surface coastal front is located near the western edge of the Gulf Stream. To the north, coastal frontogenesis has also been taking place for several hours prior to 0000 UTC 25 January. The hourly observations from the ship KEOC, from 1000 UTC 24 January through 0600 UTC 25 January, and the NOAA 14-km SST analysis are shown in Fig. 7. These particular ship observations provide insight into the initial stages of coastal frontogenesis to the east of the Carolinas. Note that the sea surface temperatures reported by KEOC are not used in the analysis because they erroneously reflect an uncharacteristically weak gradient across the north wall of the Gulf Stream. As KEOC traversed the gradient on the north side of the Gulf Stream between 1300 and 1400 UTC 24 January, the wind speed increased by 9 m s\(^{-1}\), the direction veered by 30 degrees.
and the air temperature and dewpoint temperature both increased. Another significant change occurred between 1600 and 1700 UTC, when the air temperature rose an additional 4°C and dewpoint temperature increased by 7°C. The air temperature and dewpoint temperature rises measured during this period when the ship moved through the large SST gradient, seem to be related to the differing airstream origins and the strong spatial difference in local sensible and latent heat fluxes. Surface streamline analyses (not shown) imply that the MABL air to the south of the north wall of the Gulf Stream originated to the east–northeast over warm water, whereas the MABL air on the cold side of the north wall originated over cooler surfaces to the northeast. Several hours later, as KEOC nears the gradient on the eastern side of the Gulf Stream, an additional 40 degree veer in the wind direction occurs between 2200 UTC 24 January and 0000 UTC 25 January. This is the first evidence of a significant cyclonic wind shear to the east of the Carolinas and may be an early indication that coastal frontogenesis is taking place in the Gulf Stream area. At 0000 UTC 25 January, wind directions both to the east and to the west of KEOC remain from the NNE (Fig. 5a), suggesting that the coastal front is forming locally over the Gulf Stream rather than advecting from inland locations or from farther out to sea. Also, a band of showers at 0000 UTC 25 January is located over the warm waters of the Gulf Stream and is possibly coincident with the surface coastal front. In summary, the initial coastal frontogenesis at 0000 UTC 25 January appears to be most prominent just to the east of the Carolina and Florida coasts, with a less defined frontal zone between these two areas. However, the analysis is hindered by sparse oceanic data in this area.

Between 0000 UTC and 0600 UTC 25 January (Fig.
5b), the northern portion of the coastal front begins to advance westward. By 0600 UTC, broad easterly flow exists on the warm side of the coastal front with a rather abrupt cyclonic wind shift across the front. The cold-air damming with the associated northerly flow continues to help maintain the shallow confluent baroclinic zone. The coastal front strengthens during this period, especially along its northern and southern portions. Along the central part of the front, east of South Carolina, the baroclinic zone remains poorly defined and weak. A line of precipitation located on the warm side of the northern portion of the coastal front, moves westward during this period. The coastal front does not appear to extend north of North Carolina at 0600 UTC, although the veering winds over the ocean north of Cape Hatteras during the previous six hours may be an indication that frontogenesis is occurring locally in that area. Along the South Carolina coastline, wind directions back in the several hours prior to 0600 UTC, while wind directions at offshore buoy and platform locations remain unchanged or veer only slightly. Along the Florida coast, the coastal front remains almost stationary, positioned near the western edge of the Gulf Stream. However, the wind directions have veered significantly on the warm side of the front during the 6-hour period.

Between 0600 UTC and 1200 UTC 25 January, the coastal front moves onshore in northern North Carolina and weakens as it stalls near the western side of the Pamlico Sound. To the north, the coastal front is becoming more defined, with an abrupt cyclonic wind shift across the front present by 1200 UTC (Fig. 5c). The coastal front near Florida remains offshore, anchored near the western edge of the Gulf Stream. At
0600 UTC, a line of radar echoes is associated with the coastal front off the North Carolina coast (Fig. 5b). Two hours later, another line of echoes forms inland on the cold side of the coastal front and extends from about Greensboro, NC to the northeast toward PAM station 43 (Fig. 1). This line of showers eventually weakens and merges with echoes positioned along the coast to produce a swath of precipitation across eastern North Carolina at 1200 UTC (Fig. 5c).

As the coastal front continues its slow westward movement in North Carolina from 1200 to 1500 UTC, frontolysis continues so that only a diffuse baroclinic zone remains. The erosion of the wedge of cold air entrenched to the east of the Appalachians also continues. However, 1531 UTC satellite imagery indicates a partial dissipation of the clouds in southeastern North Carolina so that, during the following three hours, the old, coastal front appears to consolidate with an enhanced baroclinic zone created by differential diabatic heating near the cloudy/partly cloudy boundary. Figure 8 shows the associated rapid increase in the temperature gradient between PAM stations 31 and 32, which are located 62 km apart on opposite sides of this intensifying segment of the coastal front. On the warm side of this front in northeastern North Carolina, a small pool of cool air exists, perhaps as a consequence of southeasterly flow at the surface being modified by the cool waters of the Pamlico Sound. This results in a secondary weak baroclinic zone near the coastline on the warm side of the main coastal front. Interestingly, the radar composites for 1200 UTC and 1800 UTC (see Figs. 5c and 5d) show that the precipitation is confined along the North Carolina coastline on the warm side of the main coastal front near this secondary baroclinic zone, despite the redevelopment of the front.
farther inland. North of this region, coastal frontogenesis continues to result in an intensifying baroclinic zone. During the period between 1500 UTC and 1800 UTC, as the northern portion of the coastal front undergoes redevelopment farther inland, the southern portion remains almost a separate feature. At 1800 UTC, the coastal front bends sharply in southeastern North Carolina, associated with a weak segment of the front connecting the two stronger portions.

Surface time sections shown in Fig. 9 illustrate the coastal frontogenesis process using hourly and half-hourly temperature, dewpoint temperature, and wind data. As the coastal front passes by the Diamond Shoals Light Ship (DSLN7) between 0130 UTC and 0200 UTC 25 January, the temperature increases approximately 4°C and an abrupt cyclonic wind shift occurs. After the coastal front passage, both the wind and temperature remain steady. The frontal passage at Cape Hatteras, NC, only 24 km to the west, occurs almost 5 h later, between 0600 UTC and 0700 UTC. It is interesting to note that the cyclonic wind shift occurs 1–2 hours before the temperature increase, indicating that the deformation zone and the maximum temperature gradient are not collocated at this time. For PAM station 45, located just to the south of Cape Hatteras, a frontal passage occurs just after 0700 UTC, with a strong veer in the wind direction and a temperature rise of several degrees Celsius. However, PAM station 47, 30 km to the north of Cape Hatteras, indicates no distinct frontal passage (not shown). Only a slow veering of the wind direction is evident. Stations offshore to the north, such as the Chesapeake Light Ship (CHLV2), also show no frontal passage and only a slow veering of the wind. PAM stations 46 and 48, just to the west and northwest, show a frontal passage several hours after the passage occurs at Cape Hatteras. The movement of the coastal front to the south of Cape Hatteras is illustrated by the time section for Cape Lookout (CLKN7), where a modest temperature rise occurs after 1400 UTC 25 January. However, just 44 km to the south, buoy 41007 reports a strong frontal passage just after 0600 UTC, about eight hours earlier than at Cape Lookout. This results in an abrupt rise in temperature and a veering in the wind direction to easterly. Interestingly, GALE buoy 2 (BY2), only 74 km to the west of buoy 41007, indicates a frontal passage nearly 10 hours later. The slow-moving character of the coastal front in this region is also exemplified by the neighboring GALE buoy 4, which remains on the cold side of the front through 1800 UTC. From these observations, considerable fine-scale structure in the coastal-front strength and position is distinctly apparent.

The vertical structure of the coastal front is illustrated by the cross sections of potential temperature, mixing ratio and wind shown in Fig. 10. Note that the availability of data from different measurement systems and locations at different times dictated the use of varied cross section axes in this figure. The first cross section (Fig. 10a) is for 0000 UTC 25 January and extends from Dayton, OH, to a dropwindsonde located 500 km to the southeast of Wilmington, NC; an orientation that is nearly perpendicular to the coastline. The cold air entrenched to the east of the Appalachians extends through a deep layer to about 870 mb. The developing coastal front is located between buoy 41007 and ship KEOC, and is characterized by steeply sloping isentropes extending westward over the cold dome. A secondary baroclinic zone positioned to the west of buoy 41007 near the shoreline is likely associated with the land–sea thermal contrast. The 200-mb deep boundary layer to the east of Wilmington is potentially unstable as a result of air flowing from the cool coastal waters to the north over the warm Gulf Stream. Near the coastline, strong easterly and southeasterly flow near the top of the cold dome is causing moist air and warm-air advection from maritime locations to the southeast.

Figure 10b shows a smaller scale vertical cross sec-

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**Fig. 9.** Time section of temperature (°C) and wind for 0000 UTC through 1800 UTC 25 January 1986 for the Diamond Shoals Light Ship (a); Cape Hatteras, NC (b); PAM station 45 (c); Cape Lookout, NC (d); and buoy 41007 (e). One full barb represents 5 m s$^{-1}$.
Fig. 10. Vertical cross section of potential temperature (K, solid lines), mixing ratio (g kg$^{-1}$, dashed lines) and wind from Dayton, OH to a dropsonde for 0000 UTC 25 January 1986 (a); from Asheville, NC to Cape Hatteras, NC for 0600 UTC 25 January 1986 (b); from Dayton, OH to the Research Vessel Cape Hatteras for 1200 UTC 25 January 1986 (c); from Asheville, NC to Cape Hatteras, NC for 1800 UTC 25 January 1986 (d). The isotherm interval for potential temperature is 1 K and the isohume interval is 0.5 g kg$^{-1}$. One full barb represents 5 m s$^{-1}$.

tion, entirely over land except for the Pamlico Sound, from Asheville, NC, to Cape Hatteras, NC, for 0600 UTC 25 January. At this time, the surface position of the coastal front is near Cape Hatteras located just to the east of PAM station 45. The depth of the cold air to the east of the mountains has decreased considerably over the last six hours, and as a result, the slope of the upper extension of the coastal front appears to have decreased. An additional region of baroclinity exists between Greenville and PAM station 41. This possibly results from the contrast in parcel trajectories that arrive from continental locations to the west and from maritime locations to the east; a situation implied by the shear in this region shown in Figs. 5b and 10b. At higher levels, a tongue of high-moisture-content air with specific humidity in excess of 5 g kg$^{-1}$, is seen rising up the isentropic surfaces within the frontal zone over the cold dome.

The vertical cross section for 1200 UTC 25 January (Fig. 10c) between Dayton, OH, and the R/V Cape Hatteras (RVC) shows an intense shallow coastal front positioned just offshore at the surface near GALE buoy 2. Within and just above the frontal inversion zone, easterly and southeasterly flow still produces a westward intrusion of moist air; but this process is confined to a more shallow layer than six hours before because the maritime fetch only exists near the surface, with drier air arriving above from the west and southwest. A second, weaker baroclinic zone is located near RVC in the vicinity of the western edge of the Gulf Stream.

After the front reintensifies in eastern North Carolina at 1800 UTC 25 January, it is shown by the vertical cross section from Asheville, NC, to Cape Hatteras (Fig. 10d) that an intense coastal front forms with a surface position near PAM station 37. Even though the baroclinic zone lies to the west of station 37, the wind shift
at the surface is to the east of it. The characteristic tongue of moisture and a shallow layer of potential instability (convective instability) slope westward above the cold dome. Some baroclinicity remains near the coastline to the east of PAM station 46. Precipitation at this time (Fig. 5d) is near the coastline where the stability in the lower troposphere is a relative minimum, and low-level moisture is a maximum.

The significant heat fluxes associated with the air-mass modification that occurred over the Gulf Stream played an important role in the coastal frontogenesis process. The sensible and latent heat fluxes for 1200 UTC 24 January, calculated using (1), are shown in Figs. 11a, b. The axes of maximum latent and sensible heat flux are coincident with the warmest waters of the Gulf Stream that parallel the coastline (Fig. 6). These heat-flux maxima are not only a result of large, air-sea temperature differences, but are also enhanced by the relatively high wind speeds located near the coast, associated with the strong pressure-gradient force in this region. At 0000 UTC 25 January, the sensible and latent heat fluxes shown in Figs. 11c, d are somewhat reduced because of warmer and more moist low-level air over the Gulf Stream and shelf waters as well as slightly weaker wind speeds. During this time period, the latent-heat flux is about twice as large as the sensible heat flux. Over the following 18 h, the sensible and latent heat fluxes decrease considerably, with maximum values at 1800 UTC 25 January that are about one-half of those observed at 0000 UTC 25 January. Thus, the heat-flux gradient is reduced also.

These heat fluxes computed for the early part of the GALE IOP 2 are not exceptionally large, however, they still are a potentially important component in the coastal dynamics. For example, the vorticity tendency due to sensible heating at 0000 UTC 25 January (Fig.

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**Fig. 11.** Surface sensible (a) and latent (b) heat fluxes (W m$^{-2}$) for 1200 UTC 24 January 1986 and surface sensible (c) and latent (d) heat fluxes (W m$^{-2}$) for 0000 UTC 25 January 1986. The isopleth interval is 50 W m$^{-2}$. 
12a) was computed using (2). The elliptical shape of the sensible heat-flux maximum yields a narrow band of positive vorticity tendency just to the east of the coastline. This corresponds to the 1000-mb, 6-h height falls shown in Fig. 12b, having a maximum of 63 m. The height-fall axis along the coastline is nearly coincident with the weak, observed sea level pressure trough in Fig. 5a. The development of this trough and the associated isallobaric wind, increases the surface convergence and may, in turn, act to increase the ageostrophic deformation within the frontal zone during the early stages of coastal frontogenesis. Additionally, these heat fluxes cause boundary layer destabilization which can contribute to precipitation near the coastal trough (e.g., see Fig. 5a). The convective heating and boundary layer convergence associated with this precipitation may also feed back to the development of the coastal trough, as shown by the numerical results of Ballentine (1980).

The frontogenesis function fields computed at the surface according to (3), for the 18-h period from 0000 UTC 25 January through 1800 UTC 25 January, are shown in Fig. 13. In the early stages of coastal frontogenesis, typified by the 0000 UTC fields, the diabatic contribution from differential sensible heat fluxes and moisture fluxes acts frontally in the vicinity of the warmest waters in the core of the Gulf Stream. In this region of diabatic frontolysis, the deformation field contributes frontogenetically. Closer to the coastline, over the shelf waters to the east of the Carolinas, a region of frontogenesis from diabatic forcing is offset by a strongly frontotyical deformation field. However, off the Florida and Georgia coastlines, both the deformation and diabatic terms contribute frontogenetically. The fact that both terms act in concert may explain why the coastal front is more intense here than to the north at this time. The resulting total frontogenesis at 0000 UTC indicates a maximum in the southern portion of the analysis domain, a secondary maximum over the Gulf Stream east of North Carolina, and weak frontolysis along the shelf waters and over some areas of the Gulf Stream.

At 0600 UTC 25 January, the deformation and diabatic terms (Fig. 13) now both generally act frontogenetically along the coastal waters from Florida northward to North Carolina. The two distinct regions of deformation related frontogenesis correspond to the two more well defined portions of the coastal front discussed above. The frontogenesis contribution from the differential surface fluxes has decreased over the past six hours, in part due to a weakening in the gradient of the fluxes. Six hours later at 1200 UTC 25 January, as the coastal front begins to move onshore in eastern North Carolina (Fig. 5c), the diabatic frontogenesis maximum remains off the coast, whereas the deformation frontogenesis and total frontogenesis areas have moved somewhat inland and are centered over the Pamlico Sound. The persistence of the maximum in the diabatic term off Cape Hatteras may be related to the heat-flux gradient caused by the proximity of the Gulf Stream and the cold Pamlico Sound waters. The frontogenesis maximum east of the Georgia and Florida coasts continues to remain stationary with positive contributions from both terms. Between the two maxima in total frontogenesis lies a zone that appears to have much weaker frontogenesis. Consistent with this is the observation that the coastal front does not move close to the coastline off northern South Carolina and southern North Carolina (Fig. 5c), where it seems to be anchored to the western edge of the Gulf Stream.

The coastal front becomes well-defined inland in eastern North Carolina (Fig. 5d) by 1800 UTC 25 January. However, the total frontogenesis (Fig. 13) may be significantly underestimated at this time because

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**Fig. 12.** The 1000-mb vorticity tendency due to sensible heating \(10^{-10} \text{s}^{-2}\); isopleth interval of \(20 \times 10^{-10} \text{s}^{-2}\) (a) and the 1000-mb height tendency due to sensible heating \([\text{m} (6 \text{ h})^{-1}]\); isopleth interval of \(10 \text{ m} (6 \text{ h})^{-1}\) (b) for 0000 UTC 25 January 1986.
Fig. 13. The surface frontogenesis due to deformation (left), surface frontogenesis due to differential diabatic heating (center), and the sum of the frontogenesis at the surface due to deformation and differential diabatic heating (right) [°C (100 km)^{-1} (6 h)^{-1}]. The isopleth interval is 5°C (100 km)^{-1} (6 h)^{-1} with the intermediate dashed isopleth interval of 2.5°C (100 km)^{-1} (6 h)^{-1}.
differential surface heat fluxes associated with the cloudiness gradient were not included in the computations. As noted in section 2, heat fluxes were only calculated over the ocean. A small area of frontogenesis remains near the Outer Banks of North Carolina, despite the coastal front consolidation that took place inland. The frontogenesis computations also suggest that the weak frontal zone between the north and south segments of the front is strengthening, whereas to the south along the Florida and Georgia coastlines, the frontogenesis appears to be weakening.

A comparison of the frontogenesis fields associated with the geostrophic wind (not shown) and observed wind deformation fields shows that the deformation associated with the ageostrophic part of the wind is dominant during the developing stage of the coastal front at 0000 UTC and 0600 UTC 25 January. The geostrophic deformation actually contributes frontally along the northern Florida and Georgia coastlines, whereas a frontogenesis maximum is present in the observed field. Also, the contribution by the geostrophic deformation to frontogenesis off the South Carolina and North Carolina coasts appears to be small at this time. However, the geostrophic deformation contributes more substantially to the frontogenesis at 1200 UTC and 1800 UTC 25 January, as the pressure trough associated with the coastal front becomes more defined. These results are in agreement with Bosart (1975), Bosart and Lin (1984), and Keshishian and Bosart (1987) in that the ageostrophic deformation appears to play an important role in the development of the coastal front.

The frontogenesis function projected onto the cross section plane from Asheville, NC, to Cape Hatteras, NC, is shown for 1200 UTC 25 January in Fig. 14. The observed frontal zone lies approximately between the two dashed lines. The strong computed frontogenesis is confined to a fairly shallow layer, within the observed frontal zone, that is only a few hundred meters deep. The frontogenesis has its maximum at the surface and rapidly diminishes with height, with only frontolysis present above 950 mb. The largest contribution to the frontogenesis is from the deformation term. The tilting term acts frontolytically in the upper portion of the frontal zone and thus contributes to the shallowness of the coastal front, in agreement with the numerical study of Ballentine (1980) and the observational study of Bosart (1981).

The ageostrophic wind and the vertical motion, computed as described in section 2, and projected onto the cross section plane, are depicted in Fig. 15 for a cross section from Asheville, NC, to Cape Hatteras, NC, for 1200 UTC 25 January. The base of the frontal inversion, coincident with the lower dashed line depicting the base of the frontal zone, is maintained by the warm moist easterly flow of air rising up into the zone. Low-level confluence is present near the surface within the frontal zone, and is associated with a weak thermally direct circulation. The associated updraft has a maximum vertical velocity of 2.6 μb s⁻¹ located near 880 mb on the cold side of the front. The vertical velocity maximum is about five times weaker than Bosart (1981) computed for a more intense coastal front. The maximum is positioned near the region of showers indicated by the radar composite for this time (Fig. 5c).
5. A weak coastal cyclone event

A small-scale, weak cyclone developed along the coastal front to the east of Florida, several hours prior to 1500 UTC 25 January. In this region, the coastal front had remained stationary over the preceding 24 h and was particularly intense, as discussed earlier. Also, the narrow, inverted sea level pressure trough associated with the coastal front in this area had been well defined for a considerable period of time, with the cyclonic shear across the front producing local maxima in the low-level vorticity and convergence. The enhanced baroclinic zone coupled with the pronounced axis of confluence along the coastal front resulted in strong frontogenetical forcing (see Fig. 13 for 1200 UTC). The heat fluxes in this area were not particularly large when the cyclone formed on 25 January, however, they were large for an extended period prior to the time of cyclogenesis. Note the tongue of large heat fluxes at 0000 UTC 25 January in Fig. 11 along the Georgia and Florida coasts in the region of subsequent cyclogenesis.

An early indication of the small-scale surface disturbance is evident in Fig. 16, which shows time series of wind observations from ship NLFI and buoy 8 (BY8) both located to the northeast of Florida (Fig. 1). The ship NLFI remained nearly stationary, approximately 30 km to the north of buoy 8, as the cyclone passed just to the west between 1500 UTC and 1800 UTC 25 January. As the disturbance approaches buoy 8 from the south, the wind backs between 1330 UTC and 1500 UTC, while the direction remains relatively constant at NLFI to the north. In the next hour, as the cyclone nears buoy 8, the wind direction veers considerably. The response of the surface wind direction at NLFI lags the wind fluctuations observed at buoy 8 by about an hour because the cyclone approaches from the south. The largest pressure falls occur between 1530 UTC and 1730 UTC at buoy 8 and an hour later at NLFI.

The 3-hourly positions of the surface cyclone and the coastal front to its rear are shown in Fig. 17. After moving northward along the coastal front off the Florida coast, the cyclone moves more northeastward along the Georgia and South Carolina coastlines immediately ahead of the cyclone, the coastal front remains nearly stationary, close to the maximum sea surface temperature gradient along the western edge of the Gulf Stream. This initial position of the coastal front ahead of the cyclone is plotted because it corresponds well to the frontal position ahead of the low at other times. The disturbance continues its northeastward movement along the coastal front and moves onshore at about 0400 UTC 26 January. A distinct cyclonic circulation is evident in eastern North Carolina at 0600 UTC. Though the central pressure of the cyclone decreases moderately during the time period studied, the perturbation pressure is just a few millibars.

The vertical extent and structure of the shallow cyclone is seen in time sections of profiles of temperature advection, relative vorticity and vertical motion shown in Fig. 18. These diagnostic quantities apply at a point just south of Greenville, NC, and just west of Beaufort, NC, where the cyclone passes at ~0600 UTC 26 January. As seen especially in the vorticity and vertical motion fields, the circulation is confined below about 700 mb. The vorticity maximum is located at approximately 60 mb above the surface with the vertical-motion maximum at about 100 mb above that level. The temperature advection pattern is asymmetric, with the cold air advection behind the cyclone being stronger than the warm air advection ahead of it.

The precipitation associated with the coastal front and cyclone are difficult to separate, but both seem to have significant contributions. Figure 19 shows a 3-hourly sequence of mosaics from the NWS radars for the period 0000–0900 UTC 26 January. An area of heavy rain showers is present along the coastal front in southeastern South Carolina just to the northeast of the surface cyclone, according to the NWS radar summary for 1935 UTC 25 January (not shown). The echo tops are less than 4 km at this time. Another area of heavier precipitation is located along the coast of North Carolina. In the following four hours, this area expands and intensifies considerably as the cyclone approaches from the south (Fig. 19a). This area of convection continues to move northeastward along the coastal front ahead of the surface cyclone. The region of precipitation expands and intensifies between 0000 UTC (Fig. 19a) and 0300 UTC 26 January (Fig. 19b). The

Fig. 16. Plot of temperature (°C), dewpoint temperature (°C), wind (m s⁻¹) and pressure (mb)
for the ship NLFI and buoy 8 from 1200 UTC through 2300 UTC 25 January 1986.
echo tops are now over 8 km as indicated by the 0535 UTC NWS radar summary. The 6-h accumulated precipitation for the period 0000-0600 UTC 26 January (not shown) indicates over 25 mm in eastern North Carolina associated with the area of convection. The convection advances to the east ahead of the coastal front at 0600 UTC (Fig. 19c) and is located near the coastline by 0900 UTC (Fig. 19d). By 1200 UTC, convection lies offshore and is organized in a banded structure along the coastal front.

The horizontal moisture flux from source regions over the Gulf Stream was significantly influenced by the passage of the weak cyclone. For example, Fig. 20 shows a vertical time section of the horizontal moisture flux \(q_v\) for Wilmington, NC, from 1800 UTC 25 January to 1500 UTC 26 January. As the coastal cyclone approaches Wilmington, the low-level southerly wind from over the Gulf Stream strengthens considerably. For example, the 850 mb wind speed doubles between 0000 UTC and 0600 UTC 26 January. The strengthening winds and advective moistening in the lower troposphere contribute to a doubling of the low-level moisture flux between 1800 UTC 25 January and 0600 UTC 26 January. After the cyclone passage at 0600 UTC, the horizontal advective flux decreases in the drier westerly flow at low levels. Other coastal locations such as Duck, NC, Greenville, NC, and Beau...

Fig. 17. Composite of the 3-hourly positions of the surface cyclone and the trailing coastal front from 1800 UTC 25 January 1986 through 1200 UTC 26 January 1986. The coastal front ahead of the surface low remained near its original position at 1800 UTC 25 January 1986, shown by the dashed line.

Fig. 18. Vertical time section of temperature advection \(10^{-5} \text{C s}^{-1}\) (a), relative vorticity \(10^{-5} \text{s}^{-1}\) (b), and vertical velocity \(0.1 \mu \text{b s}^{-1}\) (c) at 3-h intervals from 0000 UTC through 1200 UTC 26 January 1986 for a point just south of Greenville, NC and just west of Beaufort, NC. The isopleth interval for the temperature advection is \(3 \times 10^{-5} \text{C s}^{-1}\); for the relative vorticity it is \(1 \times 10^{-5} \text{s}^{-1}\), and for the vertical velocity it is \(0.5 \mu \text{b s}^{-1}\)
Fig. 19. Radar mosaics from the NWS radars for 0000 UTC 26 January 1986 (a), 0300 UTC 26 January 1986 (b), 0600 UTC 26 January 1986 (c), and 0900 UTC 26 January 1986 (d). The radar locations are Cape Hatteras, NC (H); Wilmington, NC (I); Charleston, NC (C); and Volens, VA (V). Heavier shading represents higher values of reflectivity.

fort, NC, show a similar increase in the horizontal moisture transport ahead of the cyclone. It is probable that this increased moisture flux contributed to the noted, enhanced rainfall ahead of the cyclone along the coastal front. That is, the strong low-level winds associated with a strengthening of the geopotential gradient in advance of the cyclone, transport Gulf Stream-modified air northward into the convective region ahead of the surface cyclone and provide the moisture needed to produce the copious precipitation amounts observed. The 6-hour accumulated precipitation totals were in excess of 30 mm in eastern North Carolina. A similar large increase in the moisture flux noted by Uccellini et al. (1984) aided the development of heavy snow in the Carolinas during the Presidents' Day storm. As the weak cyclone tracks northeastward along the
coastal front, a great deal of small-scale variability is present in the surface wind and temperature, as shown in Fig. 21 for PAM station 36. The coastal front at the surface is to the west of the station prior to 1800 UTC 25 January, and subsequently moves slowly toward the east, passing PAM station 36 between 0200 and 0230 UTC 26 January, as signified by the abrupt temperature drop and strong veering of wind. As the surface cyclone approaches, the coastal front moves westward in the easterly flow ahead of it and passes the station again at 0500 UTC, but going in the opposite direction. In the warm air to the east of the front, southerly winds are reestablished. However, several hours later, near 0730 UTC, the coastal front passes back across the station in the westerly flow to the rear of the cyclone.

There are other significant changes induced in the coastal zone by the passage of the cyclone. In addition to the coastal-front movement noted above, the strength of the front is also modified. Specifically, this weak cyclone has some characteristics similar to the zipper-low phenomenon discussed by Clark (1983), Bosart (1984) and Keshishian and Bosart (1987). That is, the surface low propagates northward along the coastal front, continually strengthening the baroclinic zone ahead of it and slightly to its rear. To illustrate this, Fig. 22 shows the evolution of the temperature gradient between two pairs of stations that bound the coastal front until just after the cyclone passage: Wilmington, NC, and PAM station 22; and PAM stations 41 and 37. For both station pairs, there is a marked increase in the temperature gradient in advance of the surface cyclone. This enhancement of the baroclinic zone appears to be a result of the geostrophic deformation to the north of the surface low, strengthening the preexisting baroclinic zone. Calculations for 0600 UTC 26 January (not shown) indicate significant contributions from geostrophic frontogenesis just ahead of the surface low and to the rear. The largest values in advance of the surface low appear to be located on the cold side of the front, as hypothesized by Keshishian and Bosart (1987). However, this cyclone differs from zipper lows which are associated with an absence of cold air advection and which tend to leave the preexisting baroclinic zone undisturbed in its wake (Keshishian and Bosart 1987).

The modification of the MABL behind the first cyclone appears to have had a significant impact upon the timing and location of the subsequent IOP 2 cyclogenesis events. As the midlevel long-wave trough approached from the west, the first weak surface wave was propagating along the coastal front and displacing the baroclinic zone to the east. Thus, the significant vorticity advection (not shown) associated with the intense long-wave trough approaching from the west encountered a more stable, drier, lower troposphere with low-level, cold advection and weak descent.
Fig. 23. Wind (m s\(^{-1}\)) on the 960 mb pressure surface for (a) 0000 UTC 24 January 1986, (b) 0600 UTC 24 January 1986, (c) 1200 UTC 24 January 1986, (d) 1800 UTC 24 January 1986, (e) 0000 UTC 25 January 1986, (f) 0600 UTC 25 January 1986, (g) 1200 UTC 25 January 1986. The isotach interval is 5 m s\(^{-1}\). One full barb represents 5 m s\(^{-1}\). The dashed
6. The coastal low-level jet

Low-level wind maxima or jets have received considerable attention in the scientific literature during the past three decades and have been documented for many geographic locations (Paegle and Rasch 1973; Wipperman 1973). Bonner's (1968) climatological study indicates that a large number of LLJs occur in the Great Plains of the United States, but secondary maxima exist along the Atlantic Coast and in particular along the coast of North Carolina (see Fig. 23a). A variety of mechanisms have been suggested as explanations for the development and evolution of LLJs. For example, Blackadar (1957), Buajittit and Blackadar (1957), and recently Parish et al. (1988) state that the supergeostrophic property of the nocturnal Great Plains LLJ is a result of an inertial oscillation with a period of $2\pi/f$, that is initiated a few hours after sunset as a result of the frictional decoupling that occurs at the top of the nocturnal inversion. In addition, LLJs have been associated with: the diurnal heating cycle over sloping terrain (Holton 1967; Bonner and Paegle 1970; McNider and Pielke 1981); the damming of stable air against mountains and the associated sloping inversion layers (Schwerdtfeger 1975; Forbes et al. 1987; Stauffer and Warner 1987; Bell and Bosart 1988); baroclinicity near coastal regions (Dickison and Neumann 1982; Zemba and Friese 1987); midlatitude fronts (Kreitzberg 1968; Browning and Pardoe 1973); isallobaric accelerations in the Great Plains (Djuric and Damiani 1980; Djuric 1981); mass adjustments forced by upper-level jet streaks (Uccellini and Johnson 1979; Uccellini 1980); and vertical displacement of parcels in baroclinc zones coupled with jet streak circulations (Uccellini et al. 1987).

During the GALE IOP 2 period, a persistent, diurnally varying LLJ was observed between 0000 UTC 24 January and 1200 UTC 25 January 1986 along the Carolina Coastal Plain. In Fig. 23 is shown a series of 6-hourly, subjective analyses of wind speed at 960 mb, the approximate level of the LLJ maximum. Between 0000 UTC and 1200 UTC 24 January, the northeasterly LLJ increases in amplitude and expands inland, with maximum values exceeding 20 m s$^{-1}$. Speed increases of as much as 10 m s$^{-1}$ occur during the first six hours of this period. By 1800 UTC (Fig. 23d), the LLJ maximum appears to have receded to the east, even though the data void over the ocean makes interpretation difficult. Relatively high wind speeds persist at Cape Hatteras, NC, and Beaufort, NC, whereas locations to the west and south experience significant decreases. Specifically, speed decreases of 10–15 m s$^{-1}$ are observed at Fayetteville, NC, Wilmington, NC, Greenville, NC, and Sumter, SC. Generally, the winds in this area remain northeasterly during the first 18 h of 24 January. During the period 0000–0600 UTC 25 January (Figs. 23e and 23f), the center of the LLJ moves slightly westward as the wind speeds increase at inland stations and decrease at the coast. Directions also veer to an east–southeasterly direction along the Coastal Plain in this 6-hour time period, as the synoptic scale ridge moves offshore and an inverted coastal trough begins to form parallel to the coastline. Between 0600 and 1200 UTC, the wind continues to veer along the coast at the 960-mb level, as the inverted trough strengthens and becomes well defined in the low levels.

The evolution of the LLJ is further illustrated by the vertical time sections of wind speed and direction from 0600 UTC 24 January to 1200 UTC 25 January for Fayetteville, NC (Fig. 24a) and from 0000 UTC 24 January to 1200 UTC 25 January for Beaufort, NC (Fig. 24b). The Fayetteville sounding data are missing for 0000 UTC 24 January. In general, the LLJ wind-speed maximum remains below the 950-mb level which is approximately 650 m above the surface. One feature shown earlier in the 960-mb analyses, that is also illustrated well in Fig. 24a, is the rapid decay in the intensity of the LLJ between 1200 and 1800 UTC 24 January at Fayetteville. The wind speed at the 970-mb level decreases by approximately 19 m s$^{-1}$ in the 6-hour period. Because it is unlikely that the 6-hour data are actually reflecting the maximum and minimum values of the LLJ speed, the real amplitude is certainly even greater than this. During the same time period and at the same pressure level, the wind speed decreases by only 5 m s$^{-1}$ at Beaufort (Fig. 24b).

The dynamic forcing of the LLJ can be revealed by first looking at the geostrophic wind field. The structure of the wedge-ridge during IOP 2—with the coldest air just to the east of the Appalachian Mountains and the warmer air near the coast—results in a low-level max-
imum in the horizontal pressure gradient force perpendicular to the coastline between 0000 UTC 24 January and 0600 UTC 25 January. Cross sections normal to the mountains from Dayton, OH, to Wilmington, NC (not shown), at 1200 UTC 24 January and 0000 UTC 25 January indicate that a northeasterly geostrophic wind-speed maximum is present below 900 mb between Greensboro, NC, and Wilmington, NC. An area-average geostrophic wind in this region was computed using observations located at the corners of a quadrilateral defined by Fayetteville, NC, Wilmington, NC, Beaufort, NC, and Greenville, NC. The pressure gradients were calculated along opposite sides of the quadrilateral and then averaged in order to obtain values for the geostrophic wind components that apply to the geographic area. Because of the shallowness of the baroclinic zone in this region, the geostrophic wind speed decreases with height in low levels, similar to the situation described by Dickison and Neumann (1982).

A time section of this geostrophic wind profile between the surface and 850 mb, shown in Fig. 25, illustrates that significant fluctuations in the geostrophic wind speed are present during the period 0600 UTC 24 January through 1200 UTC 25 January (missing Fayetteville data prevented the calculation of the geostrophic winds at 0000 UTC 24 January). Between 0600 UTC and 1200 UTC 24 January, the geostrophic wind increases by about 4 m s^-1 (30%) to a maximum value of 17 m s^-1 at 960 mb near the level of the LLJ. Near the surface, the increase is from about 14 to 21 m s^-1. This speed increase is predominantly a result of an increase in the geopotential gradient perpendicular to the coast. In the following 12 hours, from 1200 UTC 24 January to 0000 UTC 25 January, a 9 m s^-1 (53%) decrease in the average geostrophic wind speed is observed at 960 mb, and is accompanied by a slight veer in direction. During the following six hours, from 0000 UTC to 0600 UTC 25 January, the coastal baroclinic zone (front) begins to become more defined between the surface and about 900 mb as it moves onshore with the coastal trough. This strengthening baroclinic zone over land causes the geostrophic wind at this location to the west of the coastal trough to strengthen and become more northeasterly once again. After 0600 UTC, the coastal front is sufficiently to the west of the coastline so that the strongest baroclinic area is to the west of the area for which the geostrophic wind is calculated. This results in the significantly different geostrophic

Fig. 25. Vertical time section of geostrophic wind speed (m s^-1) for a point at the center of the quadrilateral defined by the observation locations at Fayetteville, NC, Wilmington, NC, Beaufort, NC, and Greenville, NC, from 0000 UTC 24 January through 1200 UTC 25 January 1986. The isotach interval is 2 m s^-1. One full barb represents 5 m s^-1.
wind profile seen at 1200 UTC 25 January. It is worth noting that another quadrilateral of stations (Greensboro, NC, Wilmington, NC, Greenville, NC, and Sumter, SC) was chosen, and similar variations in the geostrophic wind were observed to occur during the 30-h period. The oscillation of the geostrophic wind speed may result simply from nonperiodic changes in the mesoscale structure of the wedge ridge, but it is also reasonable to attribute it to periodic variations in the pressure gradient perpendicular to the coastline resulting from the diurnal heating cycle.

The observed wind profiles for Beaufort (Fig. 24b) and the geostrophic wind profiles computed for the area of the quadrilateral with Beaufort at one corner (Fig. 25) show interesting similarities and differences. Both have an oscillation in the speed at low levels, however, the maxima are 12 h apart for the wind but 18 h apart for the geostrophic wind. This period difference may be close to 0 h or 12 h, however, if there is significant aliasing produced by the lack of adequate temporal resolution in the 6-hour soundings. Also, the observed winds at Beaufort show a maximum near 960 mb, but the geostrophic winds have a maximum at the surface. Above 960 mb, both show decreasing speeds with height, but substantial ageostrophic components exist at all times. Actual ageostrophic forcing obviously explains some or most of the differences, but also, even small errors in massfield observations on the mesoscale may create large geostrophic wind errors.

The persistent existence of the above-described low-level, geostrophic wind maximum in the area (Fig. 25) implies that the overall forcing for the LLJ is geostrophic and results from the mesoscale baroclinic zone that prevails between the shallow cold air to the east of the Appalachian ridge and the warm air over the Atlantic coastal waters. However, significant ageostrophic forcing effects need to be explained. In Table 1 are shown the 960 mb observed, geostrophic, ageostrophic and isallobaric wind speeds and directions for the area of a quadrilateral defined by the soundings at Beaufort, NC, Wilmington, NC, Greenville, NC, and Fayetteville, NC. First note that the geostrophic flow is from the northeast quadrant, roughly parallel to the coast and to the observed wind in the LLJ. The ageostrophic wind speed shows a tendency for a nocturnal maximum and a daytime minimum. Its nocturnal direction ranges from northerly to east southeasterly, roughly in the same direction as the jet, whereas in the only midday sounding (1800 UTC 24 January) it is opposed to the jet. That is, the flow is generally supergeostrophic, except at 1800 UTC when it is subgeostrophic. The isallobaric wind speeds and directions were calculated from the temporal variation of the geostrophic winds (Fig. 25). Also shown are the vector differences between the isallobaric winds and the ageostrophic winds interpolated from the 6 h observation times. Generally, the isallobaric wind speeds are less than one-half the speeds of the ageostrophic components. However, even though the isallobaric wind is relatively weak, the isallobaric accelerations (to the right of the isallobaric wind) act in the direction of the LLJ for the nocturnal times (0900 UTC 24 January and 0300 UTC 25 January) when the LLJ winds are observed to be stronger, and oppose the LLJ during the daytime hours (1500 UTC and 2100 UTC 24 January) when the LLJ winds are observed to be weaker (Fig. 24).

Another ageostrophic mechanism involves the jet-level effect of the surface stress which varies diurnally. Specifically, the decrease in turbulent boundary layer mixing after sunset results in a reduction of the frictional stress and initiates an inertial oscillation with a period of approximately 21 h at 35°N, as the windfield adjusts to a new balance with the pressure gradient (Blackadar 1957). However, significant changes in the horizontal forces acting on a parcel can also occur because of the large mesoscale variations in atmospheric structure experienced by a parcel as it moves over rel-

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<td>18.3</td>
<td>11.1</td>
<td>8.7</td>
<td>7.2</td>
<td>3.4</td>
</tr>
<tr>
<td>00/25</td>
<td>11.9</td>
<td>5.1</td>
<td>8.4</td>
<td>10.9</td>
<td>170°</td>
</tr>
</tbody>
</table>

Table 1. The 960 mb observed, geostrophic and ageostrophic wind speed and direction based on sounding data from Beaufort, NC, Wilmington, NC, Greenville, NC, and Fayetteville, NC. Also shown are the 960 mb isallobaric wind speed and direction, and the vector difference between the isallobaric wind and the 960 mb ageostrophic wind interpolated in time. Directions have been rounded to the nearest ten degrees.
atively short distances. For example, parcels embedded in the northeasterly low-level flow near the North Carolina coastline originate in the relatively deep maritime boundary layer over the Gulf Stream near the coast. Kinematic trajectories calculated for the 278 K isentropic surface indicate that the parcels within the LLJ near Beaufort have a source region over warm water to the east of the coastline and rise from near the ocean surface to the jet level. Thus, over a short distance and time period, the parcels move from a maritime regime in which there is presumable vigorous, buoyantly driven mixing, to a regime over land characterized by a stable inversion layer over the shallow pool of cold air, and perhaps a nocturnal surface-based inversion as well. This may have provided the necessary frictional decoupling mechanism to initiate an inertial oscillation.

The effect of surface friction can also explain some of the differences in the vertical time sections of the observed winds for Fayetteville and Beaufort (Fig. 24). The dark solid lines in Fig. 24 represent the base of the inversion associated with the transition between the shallow layer of cold air advected from higher latitudes by the ageostrophic northerly flow in the wedge ridge, and the easterly flow of warm maritime air above. For the nocturnal hours from 0000 UTC to 1200 UTC 24 January, the inversion base was located at about 980 mb at both Fayetteville and Beaufort, just below the LLJ. Differences in airstream origin and local thermal gradients may result in more warm advection near the top of the inversion layer at locations along the coast and may explain why the static stability in the middle of this inversion at Beaufort was nearly twice that of the stability at Fayetteville during this time period. The most remarkable difference between the Fayetteville and Beaufort time sections is seen during the period 1200 UTC 24 January to 0000 UTC 25 January when the LLJ undergoes a dramatic breakdown at Fayetteville but remains relatively intact at Beaufort. Even though the maximum speeds at 1200 UTC 24 January only differed by a few meters per second at the two locations, the speed at Fayetteville decreases to 5 m s$^{-1}$ while the speed at Beaufort only decreases to 17 m s$^{-1}$ during the daytime period of 24 January. This difference appears to be, in part, a function of the relative strength and depth of the mixing vertical within the planetary boundary layer which, of course, depends upon the vertical wind shear and stability. In order to qualitatively estimate the depth of the turbulent mixing over land, the local Richardson number is calculated, where the shaded regions in Fig. 24 indicate where Ri is less than 0.25. Even though inadequate vertical resolution in sounding data may result in an underestimate of the shear and an overestimate of the calculated Richardson number (Uccellini et al. 1986), the shaded region does reflect the changes in the depth of the layer in which turbulence is generated at the two locations. Between 1200 UTC 24 January and 0000 UTC 25 January, the inversion base at Fayetteville lifts substantially, most likely due to the mixing processes associated with the diurnal heating cycle. As the mixed layer deepens, the LLJ momentum is mixed downward and dissipates near the surface. Time sections for other inland locations on the North Carolina Coastal Plain (Sumter, SC, Greenville, NC) show a similar daytime dissipation of the LLJ. During the same period at Beaufort, the existence of the larger temperature stratification means that the daytime turbulent layer is confined below the LLJ level. As a result, the LLJ momentum does not dissipate to the same degree. Interestingly, of the other coastal-sounding sites, only Cape Hatteras showed a similar daytime persistence of the LLJ. Wilmington, NC, and Charleston, SC, also along the coast but farther south, showed some dissipation of the LLJ, potentially because of their more continental wind fetch at low levels.

As the coastal front strengthens and moves onshore between 0000 UTC and 0600 UTC 25 January, the inversion associated with it and the eastern extension of the cold dome of air decreases in height (Fig. 24). Because of its greater proximity to the coastal front, the inversion base is several hundred meters lower at Beaufort at 0600 UTC. During this period, the existing LLJ reforms just above the descending inversion at Fayetteville, while at Beaufort, the level of maximum winds in the existing LLJ descends with the inversion.

In summary, the LLJ that occurred during IOP 2 in eastern North Carolina appears to be a complex interaction among an oscillating geostrophic forcing, resulting in isallobaric accelerations, an inertial oscillation, and boundary layer mixing processes. The geostrophic forcing, which weakens with height in the low levels and varies in strength with about a diurnal period, creates a favorable environment for the development of the diurnally oscillating LLJ. The LLJ wind speeds may have been further modified by an inertial oscillation that results from the air flow becoming decoupled from the surface-frictional stress after sunset. Variations in the local inversion strength may have modified the depth of the boundary layer mixing, thereby enabling the LLJ to persist at some locations in eastern North Carolina, such as Beaufort, and weaken substantially elsewhere, such as Fayetteville.

7. Summary and conclusions

In this study of coastal processes during GALE IOP 2, three physically related phenomena were investigated: the coastal front; a shallow cyclone that propagated along the coastal front, modifying it as it moved northward; and a low-level jet that formed in the strong coastal pressure-gradient field. Early in the study period, the coastal front was located over the Gulf Stream waters in the vicinity of an established surface pressure trough. This relatively static mesoscale trough presumably formed as a result of the sensible heating over the Gulf Stream, where the deformation associated with the early stages of coastal frontogenesis may have been produced by the isallobaric acceleration during the
coastal trough formation. Subsequently, the northern segment of the coastal front moved westward while the southern segment remained anchored to the western edge of the Gulf Stream. However, careful analysis of the frontal movements shows evidence of considerable fine-scale variation in its location and presumably, structure. Local mesoscale forcing, such as the documented example of differential solar heating along a cloudy/partially cloudy boundary, caused intermittent movement and a discontinuous structure to the feature.

Calculation of the frontogenesis function for $\theta$, showed that contributions early in the study period from deformation and differential surface fluxes tended to oppose each other near the surface along the North Carolina coast. There, the diabatic effect caused frontolysis over the Gulf Stream and frontogenesis over the shelf waters closer to the coastline, while the deformation had a contribution of opposite sign. However, both terms acted frontogenetically off the Georgia and Florida coasts. Intensification of the coastal front occurs later, as both terms act in concert over the shelf waters.

For the early part of the study period especially, the ageostrophic contribution to the frontogenesis for the deformation field is dominant. In fact, the ageostrophic deformation acted frontolytically off the Georgia and Florida coasts where the coastal front was observed to be strongest. The level of strongest total frontogenesis is at the surface, with frontolysis calculated above 950 mb. Above the coastal front and its inland extension over the cold air to the east of the Appalachians, there exists considerable advection of warm, moist air with source regions near the surface in the MABL over the Gulf Stream. Moderate surface sensible heat and latent heat fluxes are computed for the MABL over the Gulf Stream, with maximum values of $\sim 400$ W m$^{-2}$ and $\sim 800$ W m$^{-2}$, respectively, obtained early in the period. When the vorticity tendency from these fluxes is calculated, a narrow band of positive values is obtained just to the east of the coastline. This corresponds to a 1000-mb, 6-hour height-fall maximum of 63 m, indicating that these differential surface fluxes can contribute significantly to the coastal-trough formation.

A shallow cyclone formed along the coastal front to the east of Florida, in a region of the pressure trough where the MABL was characterized by positive vorticity, convergence, neutral stability, and large surface heat and moisture fluxes during the previous 24 h. For the entire study period, it remained a shallow phenomenon, with the perturbations in the vorticity and vertical motion patterns being confined to below 700 mb. As the cyclone tracked northeastward along the front, geostrophic deformation caused the front to strengthen. Also, the southeasterly flow ahead of the cyclone caused the westward transport of warm, convectively unstable moist MABL air from over the Gulf Stream, which encountered the cool stable air to the west of the coastal front and caused enhanced precipitation. Behind the cyclone, strong, cold air advection prevailed as the northwesterly flow in the MABL displaced the coastal front to the southeast. In its wake remained a dry, stable MABL with low-level cold advection and weak descent conditions not nearly as favorable for cyclogenesis.

As a result of the low-level pressure gradient caused by the proximity of the northerly flow of cold continental air over land and the warm air of the Gulf Stream MABL near the coast, a low-level northeasterly geostrophic wind existed. This forced a low-level jet in the windfield near the coast in North Carolina that showed a maximum near 960 mb and a large, approximately diurnal, variation with time. At locations with more continental wind fetches at low levels, the temporal variation in speed was as large as 20 m s$^{-1}$. Because of the observed nocturnal maximum and daytime minimum in the jet, it is hypothesized that frictional decoupling at sunset creates an inertial oscillation in the wind, with a nocturnal maximum, while boundary layer mixing during the day causes the high momentum jet-level air to be mixed downward and dissipated. Also, the diurnally varying low-level pressure gradient contributes to isallobaric accelerations that modulate the jet strength and direction.

The high resolution data from this field project has enabled mesoscale features to be defined and diagnostic calculations to be performed with greater accuracy and detail than possible using only the conventional database. However, additional work is needed to further examine the dynamics and interaction of the coastal LLJ, coastal front and mesoscale cyclone. An analysis of these processes is continuing using the Penn State/NCAR mesoscale model.

Acknowledgments. This research was funded by National Aeronautics and Space Administration Grant NAGW-994, Office of Naval Research Grant SFRGC No. N00014-86-K-0688, and NOAA Grant NA82AG00027. We thank Alfred K. Blackadar, Peter R. Bannon, Toby N. Carlson and Gregory S. Forbes for useful discussions and suggestions that improved the manuscript. William F. Roberts is also thanked for providing the radar-reflectivity mosaics used in this study. Paul A. Hirschberg collaborated in the development and implementation of the mesoscale omega equation. Computing support was provided by the National Center for Atmospheric Research which is funded by the National Science Foundation. The manuscript was capably typed by Joann Singer.

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