Coastal Frontogenesis¹

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ABSTRACT

Detailed synoptic analyses are made of mesoscale frontogenesis situations along the southeastern New England coast. Temperature gradients of 5–10°C over 5–10 km along a line separating light northerly or northwesterly flow from stronger easterly flow are common. The coastal thermal contrast may follow 6–12 hr after the establishment of a cold anticyclone to the north of New England and persist for 12 hr. Dissipation takes place when an offshore cyclonic circulation reaches the latitude of southern New England such that coastal winds back to northerly. The whole process is called coastal frontogenesis.

A persistent tendency for coastal fronts to stagnate along a Boston to Providence line is noted. Precipitation appears to be enhanced along and just on the cold side of the frontal zone. Arguments are presented in favor of the importance of surface friction, orography, coastal configuration, and land-sea thermal contrast on the life cycle of coastal fronts. An inspection of eight years of data suggests that coastal frontogenesis is a maximum in late fall and early winter when land-sea temperature contrasts are greatest. Evidence is presented in one case of a coastal front, acting as a channel for a developing secondary wave disturbance, with latent heat release associated with the precipitation maximum proving to be important for development.

1. Introduction

The present paper is designed to document the existence of unusual mesoscale frontal features that form in New England during the winter season.² We have chosen to call these features coastal fronts since the effects of orography, coastal configuration, land-sea temperature contrast and friction evidently play major roles in their formation, intensification and dissipation. The present investigation grew out of observations of remarkable situations in southeastern New England of 10°C temperature contrasts over distances of 5–10 km on a length scale of 100 km. Such coastal fronts are characterized by cyclonic wind shifts and have many characteristics of warm fronts. The frontal boundary is often the dividing line between frozen and non-frozen precipitation. Coastal fronts tend to occur on time scales of 12 hr in advance of the passage of a coastal low pressure center and create a serious forecast challenge.

A key synoptic feature in coastal front formation is

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the presence of a cold anticyclone over northern New England resulting in a flow of northerly air along the coast. This anticyclone usually takes the form of a pressure ridge east of the mountains. The coastal front forms as the boundary between the northerly flow and the development of an easterly maritime air flow off of the adjacent Atlantic. It should be emphasized that such features usually form well north of where the surface warm front is found in association with the developing low pressure wave to the south. Similar frontal features have been observed in the Carolinas and to some extent the upper Texas coast. Documentation of these fronts will be reported elsewhere. It is important to note, however, that coastal fronts appear to exhibit marked geographical preference for New England and the Carolinas along the east coast.

A detailed description of the coastal front feature will bring into sharper focus the essential dynamics involved as well as reveal some significant forecasting clues. The approach here will be to look at case studies of different synoptic situations in which a coastal front played a significant role in the weather pattern over New England. The cases were selected because, collectively, they document the essential features that are likely to arise in coastal front situations. Fig. 1 shows the location of key observing stations and the topography of the region.

2. Case of 4 December 1968

Fig. 2 shows a series of surface and 500-mb maps for the period 3–5 December 1968 during which time a

FIG. 1. Locations of key places mentioned in the text. Elevations over 2000 ft and between 1000 and 2000 ft are indicated by heavy and light stippling, respectively.

major trough in the westerlies intensified and moved eastward across the United States. The surface patterns accompanying this development were chaotic but featured well-defined low pressure centers persisting on both sides of the Appalachians. Pronounced mesoscale variations in the temperature, pressure, wind and precipitation fields were observed over most of the eastern United States. Bosart (1973) has discussed mesoscale aspects of the precipitation field over the 48-hr period while Bosart and Cussen (1973) have reported on the extensive gravity wave activity in the cold air over the Carolinas and Virginia. In particular, note the period from 1200 (all times GMT) 4 December through 0000 on 5 December. A tendency for a coastal secondary is already evident by 1200. This feature

tended to evolve northeastward during the next 12 hr as a warm front wave. It is this warm front wave that we wish to examine further.

Detailed surface maps for southern New England are presented in Figs. 3 and 4. Careful analysis of all available data has been used in these and the following figures. Surface reports are taken from WBAN 10 forms obtained from the National Climatic Center in Asheville, N. C., in order to ensure time consistency where possible. Station pressures were calibrated under weak anticyclonic conditions in order to determine any systematic errors. Pressure averaging was used as an additional check. Reports from Coast Guard vessels and ships of opportunity were incorporated into the analyses when available. In most cases, ship winds

FIG. 2. Surface and 500-mb maps for 3-5 December 1968. Surface isobars every 8 mb; dashed every 4 mb where necessary. 500 mb maps include heights (solid) at 6 dekameter intervals and temperatures (dashed) at 5°C intervals. Conventional plotting and analysis intervals.

proved to be considerably more useful than pressure readings.

At 1200 on 4 December 1968 (not shown), a broad SE geostrophic flow prevailed over New England with a convergence region just south of Long Island. Ahead of it, moderate precipitation was beginning to break

out. Judged by observed wind shifts, the convergence region reached the vicinity of southern Long Island 3 hr later. Persistent light rain in the southerly air flow, in comparison with the ending of precipitation following most warm front passages, is to be noticed. An enhancement of the anticyclonic bulge across central New

FIG. 3. Detailed surface maps for southern New England at 1500 and 1800 GMT 4 December 1968. Isobars (solid) are drawn at 2-mb intervals; isotherms (dashed) at 5F intervals. Dotted line indicates coastal front location. Surface winds in knots, cloud cover and present weather indicated at stations where available.

By 1800, surface troughing is very much in evidence over southeastern New England. Winds have veered over Cape Cod and southeastern Massachusetts and backed over interior sections. Connecticut River Valley winds depart 90° counterclockwise from the geostrophic direction. Coastal winds are much closer to the geostrophic direction. The surface trough across Long Island has continued to drift very slowly northward. Again, continuous precipitation is in evidence in the southerly airstream in what otherwise might be thought of as the warm sector. A detailed analysis of the 1800 surface map over a larger region (not shown) provides convincing evidence that a more classical warm front is located further south near 39°N . The ESE winds at Nantucket light ship and at another vessel south of Block Island (BID) suggest that any surface warm front is still located further south. Yet clearly a tendency is being established for a surface convergence zone to form from northeastern Massachusetts toward eastern Long Island.

Over Cape Cod, Nantucket and Martha's Vineyard, surface winds continued to veer such that by 2100 a well-defined surface convergence zone is present over southeastern New England. Notice the calm wind at South Weymouth (NZW). This station is *practically* located on the convergence zone. Less than an hour later the wind picked up from the south at NZW. Another interesting feature is the north-south ridge developing over western New England ahead of a surface trough extending southward from central New York. The latter feature is connected with the original west-of-the-Appalachian primary low and the semblance of a secondary low centered southwest of New York City. The tendency of the winds to veer easterly again is evident at New York City. At the same time there is now evidence of a cyclonic circulation evolving on the convergence zone over extreme eastern Long Island, resulting in three separate cyclonic circulation centers over the New England-New York-New Jersey region. Observed winds and temperatures, however, indicate that warm frontogenesis apparently is taking place across southeastern New England.

The end result of the above process is seen at 0000 at which time a definite cyclonic circulation is found over southeastern Massachusetts embedded in a region of relative warmth. The heavy rain at Nantucket is rather local. Within the hour that station reported a wind shift to SSW with the passage of the surface warm front. Close scrutiny of the hourly maps drawn for this period clearly suggests that this cyclonic circulation evolved out of the coastal convergence region. A 3-hr deepening of 5 mb is noted. The north-south ridge line noted at 2100 GMT has moved eastward along with a cold trough in the isotherm pattern. In all probability this mesoscale ridge is associated with cold air trapped east of the mountains under the influence of an anticyclone to the north. A separate cyclonic circulation in the

FIG. 5. Six-hour precipitation in inches ending 1800 GMT 4 December 1968 and 0000 GMT 5 December 1968

New York City area represents the final wave moving northeastwards.

Fig. 5 shows 6-hr precipitation totals ending at 1800 and 0000, composited from the cooperative precipitation network, in addition to regularly reporting stations. A tongue of heavy precipitation from New York City toward Boston is evident. Amounts are generally light over Cape Cod, Nantucket and Martha's Vineyard. Very heavy precipitation is noted in southern Connecticut where the low-level convergence region appeared to be the strongest. It would also appear that the coastal convergence region described previously separates distinct relatively wet and dry areas. Such behavior can typically be found in association with warm fronts.

3. Case of 15 November 1970

Fig. 6 shows portions of the surface and 500-mb maps for 1200 on 15 November and 0000 on 16 November 1970. A weak north-south low pressure trough, with centers over Pennsylvania and southeastern Virginia, moved northeastward. The southern center gradually

FIG. 6. Surface and 500-maps for 1200 GMT 15 November 1970 and 0000 GMT 16 November 1970. All analyzed quantities are identical to Fig. 2.

became stronger during this period. The offshore coastal development is complicated by an elongated NE-SW surface trough having, evidently, a double center. Twelve hour deepening was an unimpressive 2-3 mb

on the synoptic scale. An extensive confluent trough was moving northeastward and weakening slightly at 500 mb. Meanwhile, an interesting coastal front formed and had a definite bearing on the weather

FIG. 7. Detailed surface maps for southern New England at 1500 and 2100 GMT 15 November 1970. All analyzed quantities are identical to Fig. 3 except isobars drawn at 1-mb intervals.

patterns during this 12-hr period, Fig. 7 presents detailed surface maps for southern New England at 1500 and 2100 while a larger scale map at 1800 is shown in Fig. 8.

At 1200 (not shown) the surface isobar pattern was similar to the previous December case and displayed a pressure ridge in the cold air east of the mountains. By

FIG. 8. Surface map of New England for 1800 GMT 15 November 1970.
All analyzed quantities are identical to Fig. 7.

1500 the wind field suggests veering is taking place from New Bedford (EWB) to the south and east while backing is occurring to the north and west such that the beginnings of a surface convergence zone are seen. Inspection of the hourly pressure change maps (not shown) reveals that pressures over coastal southeastern New England are falling faster than elsewhere during this period. Water temperatures in the vicinity of Nantucket and Martha's Vineyard ranged from 10–15°C, which suggests that advection of warmer air landward

by the onshore flow is being reflected in larger coastal pressure falls.

A pronounced surface convergence zone is evident across southeastern Massachusetts by 1800 (Fig. 8) in a similar position to the December 1968 case. The anticyclonic isobaric bulge located in the isotherm trough just inland from the coast is a well-defined feature along with continued backing of the winds on the cold side of the pressure trough. Offshore troughing is evident into the Gulf of Maine. Coast Guard stations along the

Maine coast corroborated the existence of a northward extension of the coastal front. There is no evidence in Fig. 8 or from available ship data at 1800 to suggest that the warm front near 38N at 1200 (Fig. 6) moved rapidly northward to lie along the coast 6 hr later and then retreated seaward in the following 6 hr. The absence of upper air data at this time prevents putting this conclusion on a more firm footing.

A curious feature is the incidence of cross-isobar flow toward higher pressure on the warm side of the convergence zone. Naturally, pressure values were viewed with some suspicion. Eighteen-hour pressure averages over New England were carried out for the day in question. The averaging period was determined by data availability from Hyannis and New Bedford. No errors in excess of ± 0.3 mb were detected. The Nantucket barogram trace revealed a sharp, steady fall of 5 mb between 1500 and 1700 with considerably smaller falls before and after this period. No evidence for any local gravity wave activity is found in the trace. We are thus forced to accept our analysis with some reluctance. Perhaps we are witnessing a weakening in the horizontal pressure gradient with the corresponding establishment of a cross-isobar flow toward higher pressure as the air decelerates.

A key question is whether the coastal front forms locally or advects in from the ocean. Evidence presented so far suggests the former conclusion. An inspection of all reported winds from Nantucket (ACK) to New Bedford (EWB) failed to disclose any systematic westward propagation of wind shifts from NE to SW. Rather, the winds over Cape Cod, Nantucket and Martha's Vineyard seem to veer simultaneously. In addition, hourly isallobaric analyses at 1600, 1700 and 1800 show the development of a region of maximum pressure falls in the vicinity of the coastal front located on the 1800 map. In view of the backing geostrophic wind direction over Cape Cod resulting from the northeastward extension of the surface trough, a forecaster would expect a weakening and backing of the SE flow. This indeed happens by 2100, as can be seen in Fig. 7. Consequently, the coastal front in this case is rather a short-lived phenomenon existing on a time scale of 6 hr. Wind records suggest the wind shift line propagated rapidly seaward in excess of 30 kt. Note that the wind at Nantucket light vessel (493 in Fig. 7) has veered to the SW. This places the surface warm front near 41N at 2100 with a weak cyclonic circulation somewhere east of Nantucket. The developing northerly circulation to the rear of this disturbance has helped dissipate the coastal front. No evidence for the coastal front feature was shown on the National Meteorological Center 1800 GMT surface analysis transmitted over national facsimile.

Fig. 9 shows the 6-hr precipitation totals ending 1800 derived from all available data. There appears to be a suppression of rainfall over extreme southeastern

FIG. 9. Six-hour precipitation in inches ending 1800 GMT 15 November 1970.

New England on the warm side of the surface convergence zone, as in the December 1968 case. The heaviest rainfall amounts occurred in the 3-hr period ending 1800. Little additional accumulation occurred thereafter. Presumably the period of coastal frontogenesis is associated with an enhanced thermally direct circulation that is buoyed by latent heat release, although the limited horizontal motion scale suggests that rainfall amounts should taper off inland immediately away from the coast. The configuration in Fig. 9 lends considerable support to this argument.

4. Case of 22–24 December 1970

Fig. 10 shows the surface and 500-mb maps for 0000 GMT 23 December through 1200 GMT 24 December 1970. The sequence of events in New England involved the passage of a weak trough aloft at 0000 on the 23rd followed by a weak ridge and then a much more potent trough during the day on the 24th. At the surface the weak trough was reflected in a small low over central Pennsylvania and New York and the development of a secondary low south of Long Island is indicated at 0000 on the 23rd. A cold anticyclone was centered north of New England. Simultaneously, with the secondary storm development, a coastal front formed in very nearly the same position as that described in the November 1970 case. Dissipation of the front occurred after the coastal low passed east of Nantucket as winds backed to northerly. This was followed by a renewed southeastward extension of the cold anticyclone such that by 0000 on the 24th a pronounced southwestward extension of the cold ridge east of the Appalachians is visible. With an occluded low over Michigan, a deep mid-tropospheric vortex moving eastward and cold air banked east of the mountains, conditions favoring coastal redevelopment are very good. In the following 12 hr rapid coastal intensification set in with the Long Island low center moving NNE, passing just west of

Nantucket and brushing outer Cape Cod after 1500. A very well defined coastal front was noticed 12–18 hr prior to the low passage.

Detailed maps for the coastal front on the 23rd are not presented because this front is overshadowed by a more dramatic front on the 24th. However, Fig. 11 does show a time section of hourly weather sequences at three Massachusetts stations aligned nearly normal to the frontal zone. The distance between stations is approximately 26 km. Both Boston and South Weymouth show distinct wind shifts from northerly to easterly around 2000 on the 22nd. The temperature rise at Boston is 10F in 1 hr while the temperature rise at South Weymouth is much more gradual, reflecting

that station's greater distance from the ocean. Official weather observations at Boston are taken at Logan Airport which juts into the harbor and is well exposed to maritime influences. Evidently the air arriving from the ocean has not had a very long overwater trajectory since air temperatures are much below the 5–6C sea surface temperatures. Most likely the air just offshore is of recent cold continental origin, a theory supported by the light N and NW flow prevailing earlier in the day. Note that inland at Bedford continuous light to heavy snow fell throughout the period and the temperature remained in the teens. Mixed sleet, ice pellets, and freezing drizzle at Boston and South Weymouth during the period of easterly flow suggest that the lower

FIG. 10. Surface and 500-mb maps for 0000 GMT 23 December 1970 through 1200 GMT 24 December 1970. All analyzed quantities are identical to Fig. 2.

FIG. 10. (continued)

troposphere is still fairly stable with a slight upward increase of temperature. Such behavior is expected since the true surface warm front is located to the south.

An outstanding example of a coastal front in New England is shown in Figs. 12 and 13. The front formed around 1800 on the 23rd when light northerly winds over Cape Cod, Nantucket and Martha's Vineyard began to veer to NE and finally E as a cold anticyclone moved eastward north of New England. A pre-existing temperature gradient of 5–10F is noted between the Boston area and outer Cape Cod at this time. Such gradients under northerly flow are not unusual in early winter because of the thermal lag in sea-water temperatures. By 0000 (Fig. 12) the coastal front running from just west of Boston light ship to just east of New

Bedford and then south-southwestward is well defined. Pronounced ageostrophic flow can be noted west of the coastal front. Precipitation is quite widespread although the heaviest amounts are occurring inland from the coastal front. Snow is reported on both sides of the front with the exception of the coastal waters south of New England where a changeover to rain has taken place. This is consistent with the observation that the easterly flow does not represent the advection of a maritime air mass from more southerly latitudes. The air is probably of recent continental origin, and it is gradually being warmed from below by contact with the sea surface. Snow is occurring over Cape Cod. Temperatures are between 35 and 40F so the air mass must be cold aloft.

Fig. 14 represents a north-south cross section of

FIG. 11. Time section of hourly weather at Bedford, Boston and South Weymouth, Mass., from 1800 GMT 22 December 1970 through 0300 GMT 23 December 1970. Conventional plotting model. Wind speeds in knots.

potential temperature and observed winds. The potential temperatures were computed from a modified version of a program given by Duquet (1964). Thermal wind constraints, based on vertical wind shear directions and magnitudes, were used to render the analysis as consistent as possible. The gross features of the prevailing north-south temperature gradient, having a westerly jet of 75 m sec^{-1} centered near Portland, are seen. The bulk of the lower tropospheric temperature gradient is concentrated between Portland and Chatham, or across the coastal front. In view of the surface observations, the lower tropospheric north-south temperature gradient is probably stronger than indicated in this figure. Note the stable layer centered between 750 and 800 mb at Chatham. It weakens and slopes upward to 750 mb at Portland while extending downward to near the surface at ship *Hotel*. This is the true synoptic-scale warm front inversion. The surface wind at *Hotel* at this time is 150° at 16 kt. Within an hour the wind shifted to SSW as the surface warm front passed. Below the warm front inversion at Chatham the air is much more unstable than comparable air at Portland. Evidently sensible heat transfer from the sea surface under the easterly flow regime is serving to destabilize the air mass.

Returning to Figs. 12 and 13 it can be seen that by 0600 the coastal front has moved westward to lie from

west of Block Island (BID) across Narragansett Bay to Boston and northeastern Massachusetts. A definite trough in the isobar pattern is associated with the coastal front. Again the surface ridge is most pronounced east of the mountains. Freezing rain and drizzle in the Hudson River Valley are occurring beneath advection of warmer air aloft associated with an advancing occluded front across central New York and Pennsylvania.

Little additional westward movement of the coastal front is seen between 0600 and 0900, although there is a sharpening of the temperature gradient. Rain is falling at Boston and South Weymouth; very light rain and drizzle are falling further south. The temperature contrast across the Boston metropolitan area is from 15–20F. Freezing rain and drizzle now occupy a larger area of southern New England as warm air continues to come in aloft ahead of the occluded front approaching the Hudson River Valley. By 1200 the coastal front is beginning to retreat seaward as the synoptic-scale cyclonic circulation continues to intensify. However, the surface trough in which the coastal front is embedded is still well defined. Winds at Boston and South Weymouth are backing to northeasterly at 1200 although the full impact of the temperature drop is yet to be felt. The temperature gradient is especially intense north of Boston with moderate rain and 36F at Newburyport Coast Guard station versus 13F and moderate snow at Portsmouth (PSM). Also moderate snow and 9F at Portland (PWM) is seen versus moderate snow and 33F at Portland Light Ship (393). The northward advance of the rapidly deepening coastal secondary low pressure is reflected in the surface warm front passage at Nantucket Light Ship (493) and a wind shift to SSW.

Figs. 15 and 16 depict time sections of hourly weather at Bedford, Boston and South Weymouth from 2100 on 23 December through 1800 on 24 December in an effort to better reveal the strong temperature and weather contrasts. The coastal front arrived with authority between 0300 and 0400 at Boston and South Weymouth with temperature rises of 14F and 10F, respectively. Throughout the time period the coastal front never reached as far inland as Bedford. Note the tendency for relatively heavy precipitation up until a few hours prior to the arrival of the coastal front. East of the front, snow gradually changed to sleet and then all rain as the lowest layers of the atmosphere became warmer. At the same time the rain became heavier as the easterly flow began to weaken and back around to northeast just prior to the retreat of the coastal front. This suggests that horizontal confluence and, by inference, horizontal convergence, is a major frontogenesis mechanism since part of the temperature drop occurs simultaneously with the wind direction and speed change. Evidently the effect is strong enough near the front to result in enhanced precipitation. The extreme shallowness of the cold air behind the retreating coastal

FIG. 12. Detailed surface maps for southern New England for 0000 and 0600 GMT 24 December 1970. All analyzed quantities are identical to Fig. 2.

front is indicated by the Boston report of ZL— with a 21F temperature at 1500.

Finally, Fig. 17 presents 6-hr precipitation patterns

for selected time intervals for the December 1970 cases. The first coastal front achieved its greatest intensity near 0000 on the 23rd. The patterns are not that

FIG. 13. Same as Fig. 12 except for 0900 and 1200 GMT 24 December 1970.

revealing but a dry Cape Cod relative to northeastern Massachusetts is suggested. The latter area is in the region of strongest overrunning. Meanwhile, the strong second coastal front case is characterized by another

precipitation maximum across northeastern Massachusetts. Again relative dryness is noted over Cape Cod and the adjacent coastal waters. In particular, the axis of maximum precipitation for each 6-hr period appears

FIG. 14. A north-south cross section of potential temperature (solid, 4K intervals) and observed winds (m sec^{-1}) from Caribou, Maine to ship Hotel on 0000 GMT 24 December 1970.

to be centered just inland from the mean coastal front position. Agreement is not as conclusive for the 6-hr period ending 1200 on the 24th. However, the rainfall maximum just prior to the retreat of the coastal front

has been noted in connection with Figs. 15 and 16. The quarter-inch isopleth for this same period over western New England is associated with a brief period of very heavy snowfall that occurred just in advance of passage of the occluded front from the west.

FIG. 15. Same as Fig. 11 except for 2100 GMT 23 December 1970 through 0700 GMT 24 December 1970.

FIG. 16. Same as Fig. 11 except for 0800 through 1800 GMT 24 December 1970.

FIG. 17. Same as Fig. 9 except for selected 6-hr intervals on 23 and 24 December 1970.

5. Discussion

In our presentation so far a number of factors have arisen which are important in any theoretical attack on the problem. These include the geography of New England, surface friction and orography, latent heat release, and ocean-land temperature differences. Un-

fortunately, most conclusions must remain speculative because the data quality and spacing do not readily permit quantitative computations that would be meaningful. This is especially true with regard to upper air observations, as the scale of the coastal front is considerably less than the spacing between radiosonde

stations. It is hoped that results presented in this paper will stimulate a more detailed observational and theoretical attack on the problem.

In an effort to understand the dynamics applicable to the coastal front, we observe that Petterssen's (1956) development equation for geostrophic balance may be written as

$$\frac{\partial \eta_{\theta l}}{\partial t} = -\mathbf{V} \cdot \nabla \eta_{\theta u} - \frac{R}{f} \nabla^2 \times \left\{ \int_{P_1}^{P_2} \left[-\mathbf{V} \cdot \nabla T + \omega(\Gamma_d - \Gamma) + \frac{1}{C_p} \frac{dq}{dt} \right] d \ln p \right\}. \quad (1)$$

Here η_{θ} refers to the absolute geostrophic vorticity. Subscripts l and u refer to the lower (essentially surface) and upper pressure levels respectively. The dry adiabatic and environmental lapse rates are given by Γ_d and Γ , while $\omega \equiv dp/dt$ and dq/dt refers to diabatic heating or cooling. The remainder of the symbols have their usual meteorological meaning.

In the interpretation of (1) it has proved useful to think of atmospheric development as commencing when upper tropospheric vorticity advection becomes superimposed over a surface region where suitable moisture and temperature contrasts, in addition to convergence, are present. The Laplacian of thickness tendency effects then becomes more important as the thermal configuration of the storm evolves. These effects then partially serve to determine the ultimate intensity and direction of the storm motion. Of course, atmospheric friction plays a key role as a brake on development.

The derivation of (1) does not properly incorporate latent heat release into the diabatic term so that quantitative computations involving this term are to be viewed with suspicion. A further difficulty is the restriction of precipitation rates to several hundredths of an inch per hour on the basis of quasi-geostrophic theory, discussed by Phillips (1963). Qualitatively, a region of maximum diabatic heating through the Laplacian term on the right side of (1) will result in an increase of surface geostrophic relative vorticity. In all cases described previously, there was a tendency for the axis of maximum precipitation to be found along and toward the cold side of the coastal convergence zone. As can be seen from Fig. 1, orographic influences cannot be the sole basis for this precipitation maximum. Hence the enhanced latent heat release exerts a cyclogenetic tendency through (1) along the coastal convergence zone.

Additional qualitative support comes from the quasi-geostrophic omega equation (see, e.g., Danard, 1964) given by

$$\left(\nabla^2 + \frac{f_0^2}{\sigma} \frac{\partial^2}{\partial p^2} \right) \omega = -\frac{f_0}{\sigma} \frac{\partial}{\partial p} (-\mathbf{V} \cdot \nabla \eta) + \frac{1}{\sigma} \nabla^2 \left(-\mathbf{V} \cdot \nabla \frac{\partial \Phi}{\partial p} \right) - \frac{R}{C_p \sigma p} \frac{dq}{dt}. \quad (2)$$

FIG. 18. Map showing locations where coastal front activity is often observed.

Here $\sigma \equiv (-\alpha/\theta)(\partial\theta/\partial p)$, where θ is potential temperature, α the specific volume, f_0 a constant value of the Coriolis parameter, and Φ the geopotential. The other symbols have their usual meteorological meaning or are described in connection with (1).

According to (2), upward motion is proportional to the Laplacian of the thermal advection as well as diabatic heating. This is consistent with our finding that the axis of maximum precipitation is located on the cold side of the coastal convergence zone or within the region of maximum lower tropospheric warm advection. In the December 1968 case considerable precipitation fell (Fig. 5) along the coastal convergence zone prior to 1800 on the 4th and before any sizable upward increase of cyclonic vorticity advection took place aloft (Fig. 2). Consequently, given that the surface convergence zone formed across southeastern New England (Figs. 3 and 4), it is tempting to speculate that latent heat release enhanced the development of a cyclonic circulation along this zone. This zone then serves as a channel for the developing secondary wave. Caution must be exercised, however, because the horizontal length scale of the coastal front approaches the lower limit of quasi-geostrophic theory. In addition, the vertical distribution of diabatic heating is unknown.

The geography and topography of the northeastern United States (Fig. 1) is such that cold air outflow from Canadian anticyclones passing to the north of the region is funneled southward east of the Appalachians much more readily than west of these mountains. A synoptic characteristic is the establishment of a surface pressure ridge east of the Appalachians within the cold air. An outstanding characteristic of these ridges, besides their persistence, is that the ridge axis is found east of the mountain range itself. Baker (1970) investigated occurrences of these cold high pressure ridges. He deduced that the ridge axis conformed closely to the line of greatest low-level cold air depth and that the blocking effects of the mountains served to pile up cold air to the east. A similar pattern on a smaller scale is observed in all three coastal front cases described previously. It is especially noticeable in the case of 24 December 1970 and can be seen in Figs. 10, 12 and 13.

Consequently, synoptic-scale orographic factors aid in the establishment of a northerly gradient along the essentially north-south aligned New England coast with a cold anticyclone centered to the north or north-west. Under such a northerly flow regime the land is to the right facing downstream. Cross-isobar flow toward lower pressure, due to surface friction and orographic factors, is greater over land than sea in such a case. This leads to coastal convergence and upward motion near the shoreline. Surface vorticity would be generated and thermal contrast along the coast established in the face of a prolonged northerly flow. However, in repeated observations over five years the first author has noticed a great tendency for coastal front structures to form in the New England and Carolina areas or at locations A and B in Fig. 18. In addition, there is some suggestion of coastal front formation along the upper Texas coast at location C. Coastal fronts appear to be much rarer between locations A and B although there have been cases where Carolina coastal fronts appear to extend northeastward to Chesapeake Bay. It is suggested that the shape of the coastline under the influence of an offshore cyclonic circulation is an important parameter. The idea has been employed on a much larger scale by Carpenter (1945) in a discussion of cyclogenetic regions along the Pacific coast of North America. In such situations the configuration of the coastline changes from cyclonic to anticyclonic from south to north across points A and B. Across point C the change is from cyclonic to uniform but the south-to-north change is still in the anticyclonic sense.

According to (1) the effect of the Laplacian of diabatic heating and cooling must be to produce cyclonic and anticyclonic geostrophic relative vorticity south and north of points A, B and C, respectively. Likewise from (2), low-level ascent and subsidence would be enhanced south and north of the same points. The net effect would be an anticyclonic bulge in the surface isobars over eastern New England and eastern North Carolina. In association with the mountain barrier to the west

such anticyclonic development would serve to keep shallow cold air banked up and trapped east of the Appalachians. This in turn would probably enhance the coastal convergence associated with the outflow from a cold anticyclone. The effect would probably be most pronounced just south of the anticyclonic bulge where the onshore geostrophic component would be enhanced.

Detailed inspection of the surface weather records, in view of a partial overlap between the temperature change and wind shear through the frontal zone, certainly suggests that horizontal confluence is involved. Large land-sea thermal contrast should enhance the thermally direct circulation through the twisting term which then contributes negatively to frontogenesis such that the coastal front rapidly loses intensity upward from the surface. Sanders (1955) has described similar circulations in case studies of intense surface fronts. Evidence has already been presented for the shallowness of the front in Fig. 14. In addition, the first author has observed many coastal front situations first hand. His observations suggest that temperature changes of 5–10°C over 2–3 km in intense coastal front situations are not unreasonable. The extreme shallowness of these features is further suggested by his personal observations of ESE flow at cloud base near 300 m coupled with a NNW surface flow.

Frictional effects would further suggest a shallow front through horizontal divergence above the top of the Ekman layer. Eliassen (1959, 1962) presented theoretical arguments which suggested that frictionally aided fronts are inherently shallow features while Clarke (1961) and Ball (1960) have suggested that friction plays a major role in some types of front formation, like the sea breeze front. It should also be noted that there is a greater tendency for cold anticyclones north of New England to be situated over land in December and further eastward near the coast in March and April which is a result of a tendency for a reversal of the land-sea temperature gradient during this period. This suggests that northerly flow regimes, together with a correspondingly greater depth of cold air backed up east of the mountains, are more likely in December than March, all other factors being equal. Surface friction would then allow a more persistent NNW flow in pressure ridge situations in December and a resulting larger thermal contrast near the coast.

Any intensification of the frontal thermally direct circulation is also likely to enhance precipitation within and just on the cold side of the frontal zone. Results from our case studies tentatively confirm this speculation but the station density is not adequate to resolve the question. In view of the findings for the December 1968 case, in which a wave disturbance moved along the coastal convergence zone which in turn was also an axis of maximum precipitation, it would be interesting to speculate whether coastal lows show any greater tendency to hug the coast in late fall and early winter than in late winter and early spring. The overall land-sea

FIG. 19. Geostrophic total deformation at 1000 mb times 10^{-5} sec^{-1} for 0000 GMT 24 December 1970. Line segments indicate deformation axes.

baroclinic zone would appear to be further offshore in March than December, for example. We are unaware of any relevant observational studies relating to this point with the possible exception of a study by Klein (1957). His results do not confirm our speculations. However, the 5° latitude-longitude technique employed by Klein would tend to obscure any differences on smaller scales.

In view of the influence on frontogenesis theories of the confluence mechanism of Namais and Clapp (1949), an effort was made to compute the total horizontal deformation for the case of 24 December 1970 at 0000

GMT. Computations were made using a grid spacing one-third the size of the current operational NMC forecast grid, or a distance of about 1° of latitude. Figs. 19–22 show the geostrophic and observed deformations, observed relative vorticity, and observed divergence at this time. The geostrophic calculation refers to the 1000-mb surface whereas the other fields were derived from isotach-isogon analyses of surface winds.

Comparison with Fig. 10 shows that the maximum of geostrophic deformation is concentrated along the frontal positions in keeping with classical ideas. In

FIG. 20. Total deformation computed from observed surface winds times 10^{-5} sec^{-1} for 0000 GMT 24 December 1970.

FIG. 21. Horizontal divergence computed from observed surface winds times 10^{-6} sec^{-1} for 0000 GMT 24 December 1970.

addition, a distinct maximum is found east of the Appalachians in the cold air in the region occupied by the pressure bulge. A separate minimum is located off of the North Atlantic coast and just east of the coastal front position. The observed deformation in Fig. 20 is somewhat *noisier* than Fig. 19 due to the observed winds but the large-scale geostrophic features are much in evidence. A major discrepancy is seen over the north-eastern United States. The maximum east of the Appalachians in Fig. 19 is replaced in Fig. 20 by a minimum centered over the Hudson Valley. Of most

importance is the distinct maximum running from coastal Maine to extreme eastern Long Island in Fig. 20. This position closely approximates the location of the coastal front at this time. Evidently the ageostrophic effect of surface friction, together with other terms in the equation of motion such as the acceleration term, are playing a significant modifying role in the total deformation field.

The large-scale synoptic features are also noted in the divergence and vorticity fields in Figs. 21 and 22. Coastal convergence within the cold air is quite pro-

FIG. 22. Vertical component of relative vorticity computed from observed surface winds times 10^{-6} sec^{-1} for 0000 GMT 24 December 1970.

nounced in comparison with the synoptic-scale circulations. Divergence is noted along and just to the east of the Appalachians over Pennsylvania, New York and New England. The vorticity pattern is somewhat noisy although a distinct maximum is found in southeastern Massachusetts. Again the influence of the mountains is seen across northern New England. The anticyclonic maximum just north of New York City resulted from a stronger easterly flow to the south over the ocean coupled with a southeasterly flow across eastern Pennsylvania. Undoubtedly the various small-scale features would stand out even more clearly with a finer mesh, yet we still see the influence of the coastal front on this scale.

A climatological investigation of the frequency of coastal front occurrences in New England was carried out from eight years of data from January 1964 through December 1971 using the months of September through May. The data source was the microfilm compilations of daily surface maps routinely transmitted over national facsimile and now available from the National Climatic Center in Asheville, N. C. The work of Baker (1970) and our own observations of pressure ridge situations east of the Appalachians proved useful in sharpening the selection process. We looked at all cases with a high centered over or to the north of New England and in this sense our criteria was subjective. Table 1 presents our findings.

TABLE 1. Number of New England coastal front cases (1964-71).

September 3	December 19	March 12
October 4	January 16	April 4
November 16	February 17	May 4

It should be emphasized that the monthly numbers are probably on the low side. Note that the coastal front situations described previously were not analyzed on surface maps sent over national facsimile. However, their presence could be inferred from observations of pressure ridges together with judicious inspection of the plotted observations. The resulting uncertainties suggest that results should be treated with caution and that any statistical tests would be suspect. With this in mind the results do suggest the tendency for a late fall and early winter maximum. We believe that such a finding is physically reasonable in view of the rather large land-sea temperature contrast at this time of the year.

In an effort to stimulate a theoretical attack on this problem, we will try to summarize what appear to be the essential elements. First, New England coastal fronts exist on time scales of 6-12 hr prior to passage of a coastal cyclonic disturbance. Coastal fronts tend to stagnate along a line from Boston to Providence to Block Island, R. I. It would appear that coastal fronts are initiated in almost all cases with a pressure ridge east of the Appalachians and a developing cyclonic

circulation further to the southwest. Dissipation usually takes place when the surface low reaches the latitude of New England as coastal winds back to northerly. Coastal fronts are usually not observed with the very strong offshore lows since presumably the strong cyclonic circulation over land serves to overwhelm local gradients.

Finally, our results suggest that friction, orography, coastal configuration, and land-ocean thermal contrast are all important factors, although the nod appears to go to friction and its related influences. The thermal contrast, while important, is probably not the generating mechanism. We appeal to the recent case of 1 May 1972 (not shown) to support our argument. In that case a coastal front stagnated along the coast for nearly 24 hr. Outer Cape Cod, Nantucket and Martha's Vineyard remained in a southwesterly airflow. The temperature contrast across the front was minimal because the southwest flow had a cold overwater trajectory. Likewise, changes in the pressure gradient across the front were hard to detect although the wind shift was pronounced.

Consequently, one would not expect the frontal models of Hoskins and Bretherton (1972) to be very effective in the present situation. A major requirement in that paper and many others on frontogenesis is that wind components parallel to the front satisfy geostrophic balance. We accept this assumption as a first-order approximation for fronts on the synoptic scale. However, we observe surface wind directions in excess of 90° counterclockwise to the geostrophic wind direction within the cold air at stations where local terrain effects are not suspect. Sanders (1955) has commented on the violation of this criteria for shallow, intense surface fronts. He computed front-parallel vertical wind shears three to five times smaller than the corresponding geostrophic shears. Evidently surface friction plays a major modifying role in surface frontogenesis.

6. Forecasting considerations

This paper would not be complete without some discussion of the forecasting problem that coastal fronts present to the forecaster in the field. Clearly, these fronts may cause considerable "tearing of hair" in forecast offices when thermal and precipitation (type and amount) gradients are as large as in the December 1970 case. Our results suggest that coastal fronts have a high affinity for a Boston-Providence axis. During winter this axis is likely to be near the crucial rain-snow line. On occasion ice storm conditions may arise northwest of this axis. In the Boston area the first author has observed that the line between freezing and non-freezing precipitation may persist for several hours between the Cambridge and Belmont-Malden regions. The enhancement of precipitation along and just northwest of the mean coastal front position should be noted, especially in any flood potential situations. Our results also tenta-

tively suggest that coastal fronts are a special forecasting headache in late fall and early winter.

We would also like to enter a plea for the incorporation of small-scale features like the coastal front into the routine synoptic analyses transmitted over national facsimile. The mission of the National Meteorological Center is to analyze and forecast the motions of synoptic-scale systems as seen on the daily weather maps. They have been very successful in their mission. However, as a matter of policy they are not trying to bring out mesoscale features such as the coastal front. We believe that we have pointed out the obvious importance of a particular type of mesoscale feature for forecasting. We hope that detailed regional analyses will be available in the future which will reflect the increased importance of such mesoscale systems.

7. Suggestions for further research

Further research should be directed toward both observational and theoretical attacks on the problem. We feel that the former could be accomplished realistically through the use of instruments to record continuously temperature, relative humidity, pressure, and wind direction and speed at selected New England locations, including the auxiliary Coast Guard stations, as a mini field program for a portion of the early winter season. Special attention should be devoted to obtaining low-level soundings through coastal fronts in an effort to resolve more satisfactorily the three-dimensional structure. Only then can meaningful quantitative computations be made in terms of frontal origin, maturity, dissipation and motion. Observations over the adjacent ocean are also of vital importance. We feel that we can predict a sizable percentage of coastal front occurrences 12–24 hr in advance in order to ensure the feasibility of low-level sounding attempts. Work is currently in progress on several Carolina front cases and this will be reported later.

On the theoretical side, the coastal front phenomenon lends itself to numerical modeling in which friction, orography, coastal configuration, and land-sea temperature contrast are important parameters. The problem should also be attacked in connection with the related phenomenon of cold surface pressure ridges east of the Appalachians. The latter appear to be very conducive to coastal secondary cyclogenesis. Finally, there is the curious problem of apparent cross-isobar flow toward higher pressure on the warm side of the coastal front. This has been especially noticed at the lightships off the Carolina coast under a SE flow regime. There is the

suggestion of such behavior at 1800 GMT on 15 November 1970 over Cape Cod. Similar effects in the Gulf of Maine are suggested for this case although the wind observations are probably not of the quality of the regularly reporting lightships off the Atlantic coast.

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