Mesoscale Frontogenesis During Cold Season Cyclones Along The Southern New England Coastal Plain

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Abstract

As cold season cyclones propagate northeast off the New England coastline (sometimes referred to informally as "nor'easters) mesoscale processes can develop especially in eastern Southern New England. For example, the cyclonic rotation within the surface and mid level circulation results in a thermal boundary between more moderate mP (maritime polar) inflow from the Atlantic and a colder continental (cP, or continenta pola; cA or continental arctic) meridional flow over interior. Although the alignment of this thermal boundary or "coastal front" is in turn, dependent upon on the magnitude of the prevailing isallobaric field, the front also tends to align with the concave structure of the Southern New England eastern coastline. This is due in part to the boundary layer temperature gradient between relatively warm SST and colder interior landmass temperatures and associated "cold air damming" along the lee of the northern flank of the Appalachian mountain chain. This mesoscale feature results in a high impact on precipitation intensity and type especially during winter precipitation events through frontogenetic forcing and the associated increase in symmetric instability and isentropic lift especially along and immediately west of the frontal boundary; and confluence with the marine layer warming from mP air mass to the east of the front Within the eastern zone the warmer sensible and wet bulb temperatures within the marine layer can result in a transition in precipitation type from frozen to liquid, or if the precipitation type prevails i.e. snowfall there is a commensurate increase in precipitable water ratio.

This paper will investigate the dynamics of the coastal front and will cite several case studies of coastal cyclones in which this phenomenon was a high impact mesoscale forcing.

The northeast coast of the U.S. is subject to frequent cyclogenesis especially during the cold season. This high frequency is in part the result of the geographic structure from Cape Hatteras to Eastport Maine that tends toward cyclonic curvature and positive vorticity advection due to its downstream location within the prevailing middle latitude flow, and the differential thermal field between the cooler landmass of North America and the warmer Atlantic basin and associated Rossby wave forcing. Moreover, the juxtaposition of the relatively warm Gulf Stream in the wester North Atlantic basin with colder North Atlantic SST in the northern basin contributes to baroclinic forcing for developing coastal cyclones or so called "nor'easters" i.e. *Type C Cyclogenesis* (Sanders, 1959) The orographic lifting of the Appalachian mountain chain, in

Stretching" and spatial / temporal coupling of the aforementioned processes results in explosive cyclogenesis, which in turn contributes to heavy precipitation and high wind events especially along the coastal plain.

Categorization of Cold Season Cyclones along the U.S. East Coast.

Cold season coastal cyclones can be taxonomized into two archetypal categories according to the evolution of the surface cyclone along the U.S, east coast and associated upper atmospheric Rossby wave propagation and ageostrophic flows (Miller 1958).

Miller Type A

Sutcliffe ("self development") type cyclogenesis frequently occurs along or over the southeast U.S. coastal plain or the Gulf of Mexico (Figure A). Often referred to in operational meteorology (especially in the media) as "coastal huggers" due to the proximity of its NNE track along the coastline, the surface cyclone is coupled with the diffluent jet exit region of an eastward propagating high index Rossby wave. The deepening of the surface and mid level cyclones is associated with baroclinic forcing due to the deepening thermal gradient along the frontal boundary between the juxtaposed mT advection originating from the Gulf of Mexico / Gulf Stream (+ sensible and latent heat fluxes) and the cP/cA advection originating over the continent coupled with positive vorticity advection. The aforementioned processes, in turn results in symmetric instability and isentropic lift to the west of the frontal axis where the highest precipitation output occurs

Additionally, the high meridional track of the surface cyclone results in the alignment and negatively tilting of the 500 hPa mean trough axis; consequently the potential for cold air damming along the northeast coastal plain is diminished due to the westward shift of the negative isallobaric field commensurate with surface cyclone propagation along or near the coastal plain

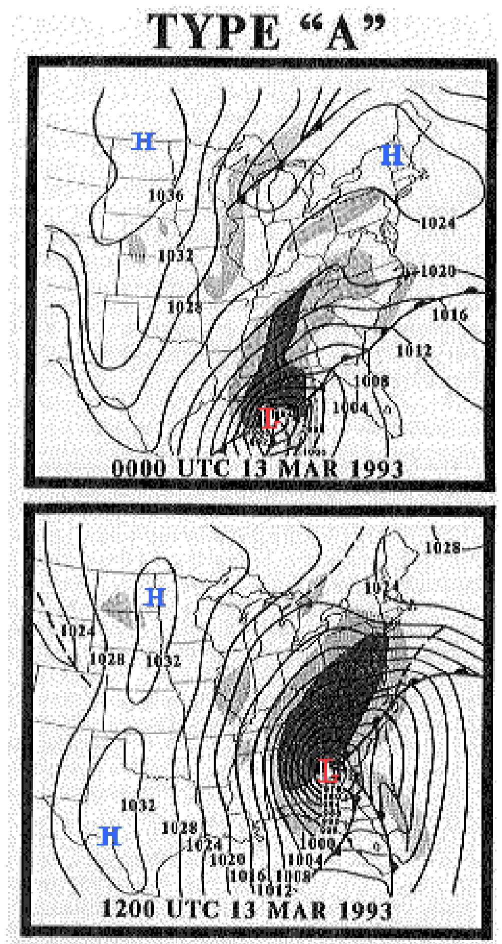

Figure 1: **Miller Type A (Sutcliffe) cyclogenesis along southeast U.S. coast dubbed as the "Super Storm Of 1993"**

Miller Type B

Miller Type B east coast cyclones also features Rossby wave induced cyclogenesis over the southeast U.S. However in this case the mean 500 hPa trough axis and associated vorticity advection is aligned further west therefore surface cyclogenesis tends to commence toward the central Gulf coast (Figure 2). Hence the resulting cyclonically curved mT meridional flow induces warm frontal genesis across the southeast. Commensurate with the aforementioned processes an inverted ridge develops downstream within the deepening isllabaric field along the east coastal plain. This inverted ridge is also due to a topographic and pressure gradient induced cP / cA meridional flow i.e "cold air damming" (CAD) along and east of the Appalachians. The result is secondary cyclogenesis as the vorticity advection becomes coupled with the baroclinic zone associated with the frontal boundary. Then as the 500hPa trough axis propagates toward the east, rotates counterclockwise becoming "negatively tilted" it decreases in wavelength as the attendant diffluent jet exit region coupled with sensible and latent heat fluxes induce rapid or even "explosive" intensification of the surface and mid level cyclones i.e. 850hPa, 700 hPa (Gyakum 1983, Kocin, Uccelini 1994)

Cold Air Damming

Cold air damming (CAD) typically occurs when a strong cP or cA anticyclone develops upstream over eastern Canada or the Maritimes resulting in an equatorial directed meridional flow along the U.S. east coastal plain. This flow is produced by the negative pressure gradient directed toward the center of low pressure downstream i.e. over the southeast U.S. and topographic forcing along the lee of the Appalachian Mountain chain. Hence CAD develops when the synoptic scale conditions $\partial p / \partial y > f$ (where $f = 2\Omega \sin \varphi$ and Ω represents the earth's rotation in rad/sec and φ represents latitude). CAD is signaled by an inverted ridge along the coastal plain contributing to the developing baroclinic zone or "ribbon" and coastal frontogenesis where cP/cA air masses become longitudinally juxtaposed with cP/mT, the latter of which originates from the Gulf of Mexico or Gulf Stream

CAD and Coastal Frontogenesis in Southern New England

Type B cyclones appear more favorable for coastal mesoscale frontogenesis in southern New England. This is due to CAD as warm marine advection is drawn into the developing surface and 850 hPa cyclonic circulation becoming juxtaposed with cP /cA air mass over the interior southern New England.. Frontogenesis is also determined by the location of surface cyclogenesis and axis of propagation of the 700 hPa and the 850 hPa cyclones. Typically surface frontogenesis occurs within the deformation zone as the result of horizontal shear across the thermal gradient. Commensurate with this process the thermal gradient associated with the 850 hPa cyclonic circulation develops an "S" shaped configuration as the associated low level jet streak becomes perpendicular to the Laplacian of the thermal field (Kocin and Ucellini, 1998).

The Geology of Southern New England: A Brief Overview and its Impact on Coastal Frontogenesis.

New England geology has been highly impacted by both plate tectonics and glaciation. For example five major orogenic events have occurred beginning with the Grenville orogeny during the Proterozoic eon at appx 1.2 x10^9 yr BP (Before Present) and concluded with the most recent orgenic event i.e, the "Allegheny at appx 2.4 x 10^8 yrs BP (i.e. the early Paleozoic era) Thereafter long term penaplanation has occurred resulting in the eroded remnants that now compose the northern branch of the Appalachian mountain chain in western New England and eastern New York (i.e. the Adirondacks, Berkshires , Green and White mountains). Artifacts of these orogenic events are also predominant further east as evidenced by the Connecticut Valley and the uplifted terrain of central Massachusetts and southern New Hampshire (i.e. the Monadnock region). The coastal plain, in turn, has also been impacted by tectonic movement and the interaction with both meso-continents e.g.the Avalon Terrain and macro-continents e.g Gondwanaland. For example the Boston Basin is the result of Pre-Cambrian sedimentary deposits (>5.5 x 10^8 yr BP) e.g "Roxbury conglomerate" and "Cambridge argillite" as a result of these processes.

Furthermore, most recent glacial maximum i.e. the "Wisconsin" the last of four maxima during the Pleistocene epoch beginning at 1.8 x 10^6 yr BP ending at appx 10^4 yr BP impacted New England with the equatorial movement four distinct divergent glacial lobes across its landscape along the equatorial directed terminus of the Laurentide ice sheet, (Maximum thicknesses ranged up to > 10^3 m across southern New England!) The aforementioned processes resulted in glacial deposition as evidenced by a series of NW-_SW trending drumlins (including the Boston Harbor islands,) recessional moraines that compose Cape Cod and the adjacent islands, and glacial scouring as evidenced by Stellwagon Bank in Cape Cod Bay.

East of the coastline along the western North Atlantic basin the equatorial directed Labrador current parallels the continental shelf north of Cape Cod that imports SST below the ambient mean (2 – 10c); and the Gulf Stream, the poleward directed western component of the North Atlantic gyre (hence North Atlantic current) parallels the continental shelf south of Cape Cod imports SST above the ambient mean (25c – 30c)

The aforementioned features therefore form a complex matrix of forcings that significantly contribute to the evolution of cold season coastal cyclones through the deformation and juxtaposition of contrasting air masses hence increasing thermal gradients across the coastal plain resulting in various mesoscale processes including coastal frontogenesis. In this sense, geologic process that have occurred many millennia in the past become feedback loops for present day meteorological processes!

Coastal Mesoscale Frontogenesis And Coastal Geography and Regional Topography.

During coastal cyclone events the concave structure of the Massachusetts coastline tends to form a thermal boundary between the warmer SST (mP) air mass and the colder (cP / cA) interior air mass. This is due in part to curvature of the cyclonic rotation and deformation along the thermal boundary in the frontogenetic zone. The inflection point for this deformation is dependent upon the track of the surface cyclone.

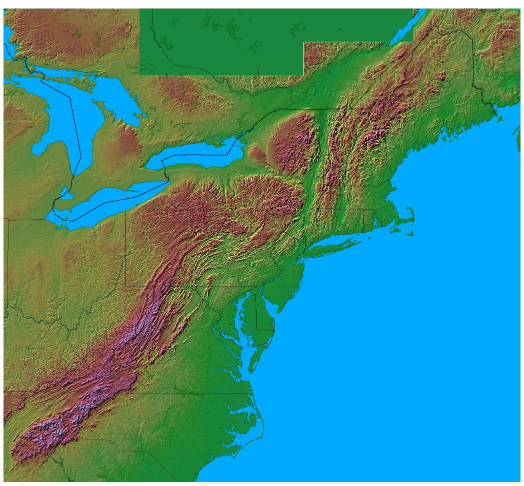

Figure 2: Appalachian Mountain Chain forces an equatorial directed meridional flow responsible for CAD (cold air damming) along the coastal plain during coastal cyclone events.

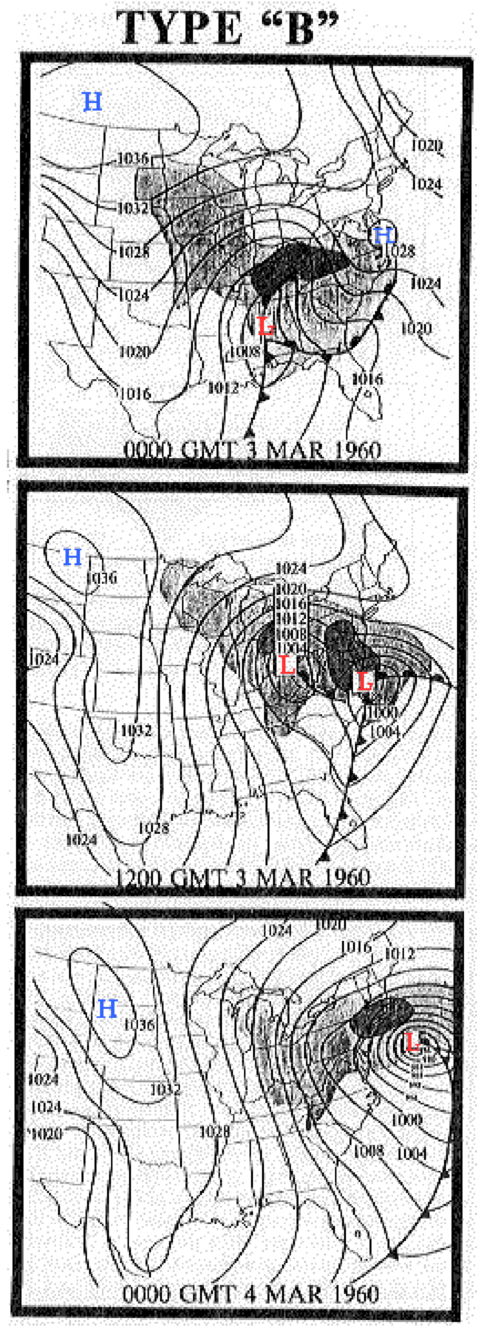

Figure 3

Mesoscale Frontogenesis:

Mesoscale frontogenesis during cold season coast al cyclones is most common and pronounced during the late fall –early winter period as this is when the SST – landmass temperature contrast is at its maximum. Typically coastal frontogenesis occurs where ΔT values are not uncommon across a 5-10 km length gradient with a duration period of 6 – 12 hours commensurate with the development of a cold anticyclone to the north. Frontolysis ensues when the surface cyclone propagates beyond the latitude quasi-perpendicular to the coastal front resulting in a shift in the cyclonic curvature of the airflow from an easterly (landmass directed) to a westerly (ocean directed) component (Bosart et al 1972).

Coastal frontogenesis is determined by the track of the surface, 850hPa, and 700 hPa cyclones and associate air mass advection patterns proximate to the coastal. The duration of the coastal front boundary can significantly impact precipitation intensity and type. To the east of the front i.e. "warm" advection" (ocean) side hydrometeor type, precipitable water to measured snowfall ratios depend upon a number of variables 1) the initial and ambient dry/wet bulb temperature profile of the boundary layer, 2) the omega, temperature, and moisture values within the vertical column especially through the 500 hPa layer 3) SST, wind speed / direction especially within the boundary layer.

To the west i.e. ""cold" advection (landmass) side of the front where vertical ascent and associated slantwise symmetric instability, transverse flow and EPA (equivalent potential vorticity) becomes coupled with mid level frontogenesis, precipitation output is often maximized. Typically within this zone high intensity hydrometeor rates e.g. low precipitable water to measure snowfall occur, or depending upon the boundary layer / vertical column thermal and dynamic environment sleet, ice pellets and freezing rain,

Type One: Interior.

Type one interior frontogenesis tends to correlate with Miller Type A cyclogenesis due the synoptic scale cyclonic flow juxtaposing more modified mP air masses (and in some cases mT) with cP/ cA air masses interior from the coastline. Frontogenesis in this case typically occurs along a poleward directed inverted trough as the developing surface and 850 hPa cyclones propagate poleward. This process can result in a significant contrast in surface – 850 hPa temperature across the frontal boundary and associated precipitation types as warm advection from the Atlantic basin is directed toward the Laplacian of the longitudinally aligned temperature gradient. In fact the 850 hPa thermal field develops in "S" shaped configuration in response to the cyclonic rotation and juxtaposition of air masses. In addition, the most significant precipitation occurs to the west of the frontal boundary due to isentropic lift and symmetric instability.

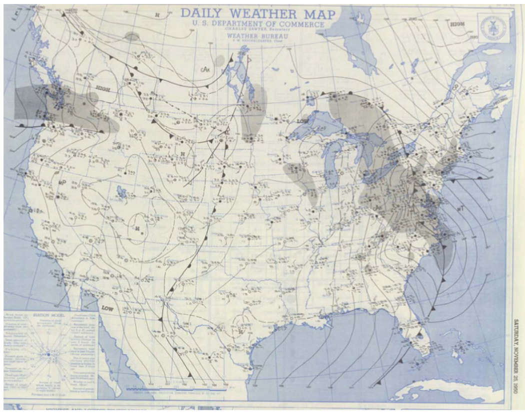

Figure ()The "Great Appalachian Storm" of Nov. 24-26 1950, A Miller Type A cyclogenesis that developed a complex juxtaposition of air masses and associated high Meso α fontogenesis and significant isentropic lift to the west of the frontal boundary.

Type 2: Coastal

Coastal frontogenesis is more consistent with Miller type B cyclogenesis due to the more pronounced ageostrophic flow along the U.S. east coast. In this case (as aforementioned) the cP/cA anticyclone located upstream over eastern Canada results in inverted ridging coupled with equatorial directed cold advection and CAD. Frontogenesis develops in response to the juxtaposition of cP/cA air masses over the interior and mP /mT over the Atlantic basin and over the southeastern U.S. As the surface / 850 hPa cyclone propagates poleward frontogenesistends toward recommencement especially in response to concave coastline geography (e.g. North Carolina and east coastal New England.) This process may be due to the SW – NE tilting and deformation of the temperature gradient within the boundary layer as the cyclonic flow encounters surface friction and differential sensible and latent fluxes between the land surface and marine environment (This thermal difference is enhanced if the land surface features a deep snow pack.)
The fluctuation and movement of this frontal boundary is determined by the isallobaric field commensurate with the axis of propagation of the surface, 850 hPa and 700 hPa cyclones.

Coastal frontogenesis: Impact on the Populace

This phenomenon, especially in an extreme state, can be an enigma to both the public and forecasters alike as it presents a meteorological "Tale of Two Cities" scenario. For example it is not uncommon (as in the case of the "Great December (11-12) Nor'easter of 1992", for a mostly "cold rain" event along the coastal plain east of the front e.g. Boston, while snowfall >80 cm west of the front (coupled with orographic enhancement) e.g. Worcester! Also the frontal boundary is subject to longitudinal fluctuation depending upon cyclonic curvature and associated juxtaposition and movement of contrasting air masses In a Type B (cold air damming)environment in which a cA/cP type anticyclone / inverted ridging is positioned to the north. strong cold advection west of the front can result in a tranverse temperature gradient of > 10c across a <20 km domain! This in turn has obvious impacts on hydrometeor type e.g. snowfall where precipitable water ratios can range from 5:1 within the “warm” zone to the east of the front, and up to 20:1 within the “cold” zone west of the front!

Summary

The New England is located upstream stream within the middle latitude geostrophic flow hence subject to high impact from eastward / northeastward propagating cyclones. This feature coupled its proximity to the Atlantic basin resulting in intensification of coastal cyclones. This effect is amplified by the juxtaposition of continental / marine type air masses which, in turn is influenced by geologic / geographic forcing. Coastal frontogenesis results from these processes impacting hydrometeor type / intensity and steep quasi - longitudinal thermal gradients. Coastal frontogenesis, in turn can be categorized into two basic types: Type 1: A higher meso α event associated with Miller type A cyclogenesis and develops along a poleward directed inverted trough axis. Type 2: A lower meso α event associated with Miller Type B cyclogenesis and develops along the quasi-longitudinal thermal gradient Often in closer proximity to the coastline.

Future Considerations

Mesoscale frontogenesis especially during cold season coastal cyclones can have significant impacts due to variability hydrometeor intensity, type. The spatial and temporal location / fluctuation of frontogenesis continues among the most daunting challenges in mesoscale meteorology. Hence similar to other mesoscale phenomena deeper research into the dynamics of mesoscale frontogenesis coupled with more refined numerical modeling is required for the increased accuracy of operational forecasts. This, in turn can better prepare the populace including public works and emergency management agencies.

Appendix

A) Cold Air Damming: CAD

A blocked flow is related to the ratio of inertia of the approaching air to the energy required to surmount the barrier. This dimensionless ratio is related to the Froude number

$$F\,r = \frac{U}{N\,H}$$

where U is the barrier-normal wind speed; H is the height of the barrier, and N the Brunt-Vaisala frequency is a measure of static stability

$$N^2 = \frac{\theta}{g} \, \frac{d\,\theta}{dz}$$

and the potential temperature Θ is defined by

$$\Theta = T\,(\,Po\,/\,P\,)^k$$

where K = 0.2854, or Poisson's constant

Frontogenesis function F (with potential temperature as the scalar field) is defined as the rate of change of the potential temperature gradient following the flow

$$F = \frac{d\,|\nabla_p\ \Theta|}{dt}$$

B) Equivalent Potential Vorticity

$EPV = -\eta \cdot \nabla \theta_e$

or

$EPV = -(\nabla \times \mathbf{u}) \cdot \nabla\theta_{es}$

∇ = 3D grad operator; u is the 3D velocity vector (set to the geostrophic wind); θ_{es} is sometime set to the equivalent potential temperature.

η = 3 dimensional vorticity vector ∇ is the del operator in x, y and p coordinates θ_e is the equivalent potential temp.

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