THE DAMMING EFFECT OF THE SOUTHERN APPALACHIANS

by

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ABSTRACT

It is well-known that the Appalachian Mountains "funnel" or "trap" cold, stable air southward, particularly during the winter. The simple thermal-wind relation can be used to show that when stable air flows toward a mountain barrier in the Northern Hemisphere, its streamlines should be deflected to the left. This theory was tested on two cases of suspected damming over the southern Appalachians. Results were encouraging. In one case, a satellite loop showed the effect vividly.

Some guidelines were developed for the field forecaster to lessen erroneous forecasts of ceilings, visibilities, temperatures, and rain-snow lines. Also, it is shown how this damming effect is fundamental to the formation of East Coast mesoscale features of coastal warm fronts and "backdoor" cold fronts.

1. GEOGRAPHY

The Appalachian chain, like most other mountain systems, is not simply long, continuous upland stretching from Maine to northeast Alabama. Instead, it consists of regional groupings of mountain systems or ridges, each with differing local topography. For example, Fig. 1 shows the Green Mountains of Vermont and the Catskills of New York. The most ideal mountainous terrain for trying to understand broadscale, orographically induced wind dynamics is found along the eastern slopes of the Appalachians from Maryland southward -- the southern Appalachians.

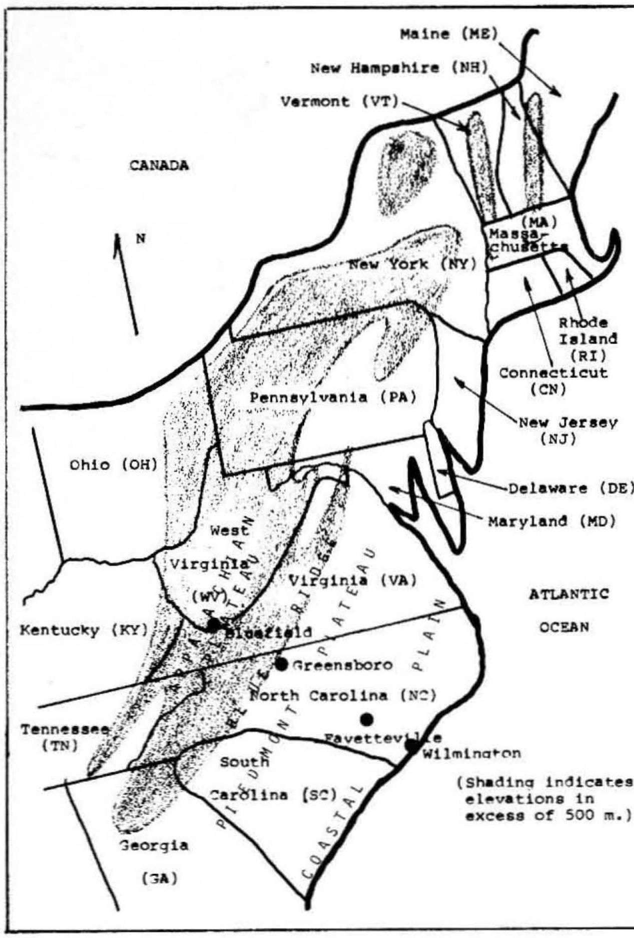

Figure 1. Eastern United States showing Appalachian chain of mountains. Shading indicates continuous elevations in excess of 500 m.

The easternmost ridge of the southern Appalachians is the Blue Ridge, which has continuous elevations in excess of 700m. The Blue Ridge is most prominent in Virginia. Westward of the Blue Ridge in Virginia is a

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valley whose floor is about 300m above sea-level. From the Blue Ridge westward through the southern Appalachians, the terrain rises to the Plateau portion where continuous elevations are at least 1000m, with the highest peaks in western North Carolina.

Fig. 2 shows a typical terrain cross-section of the southern Appalachians from Bluefield, WV southeastward to Wilmington, NC. The slope of the land is easily divided into two characteristic parts. From the coast landward to Greensboro, NC (GSO), the slope is 1/1000. This area is known as either the Coastal Plain or the Piedmont Plateau. From Greensboro uphill to the heart of the Appalachian Plateau, the slope is 7/1000.

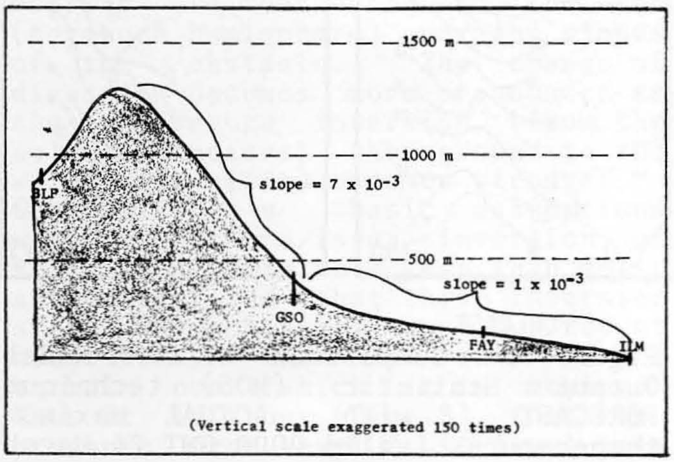


Figure 2. Typical terrain cross-section of the southern Appalachians.

2. THE PROBLEM

Veteran forecasters in the region know that often the eastern slopes of the Appalachians "funnel" or "trap" cold air southward, particularly during the winter season. Let us call this phenomenon "damming". Because of it, northward moving warm fronts are frequently held back east of the mountains. A recent example of this is shown in Figs. 3 and 4. Fig. 3 is our surface analysis using usual station plotting convention, except that temperatures and dewpoints are in degrees C. The signature or earmark that damming has occurred is a "nose" of high pressure in the sea-level pressure pattern over the eastern slopes. Fig. 4 is a

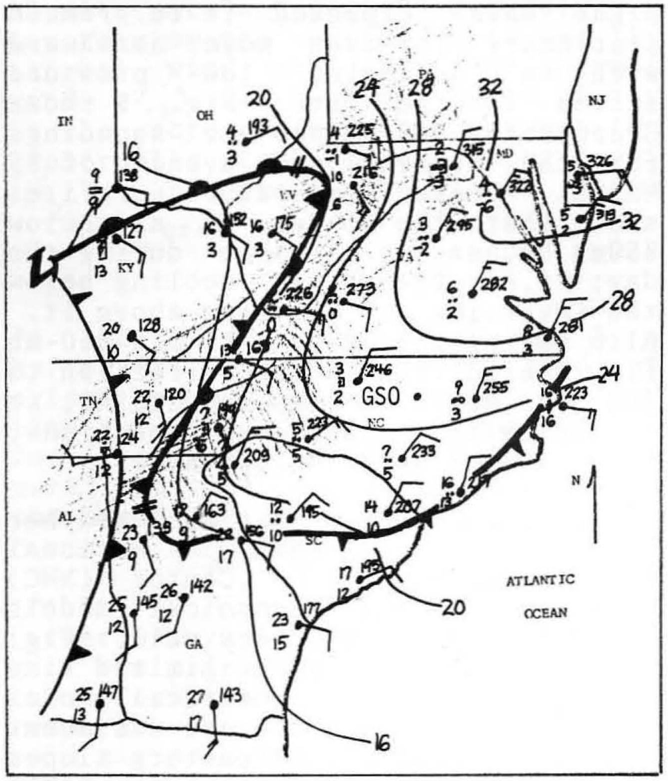


Figure 3. Sea-level pressure analysis for 1800 GMT 25 March 1978. Temperatures and dewpoints in degrees C and pressure in millibars (mb).

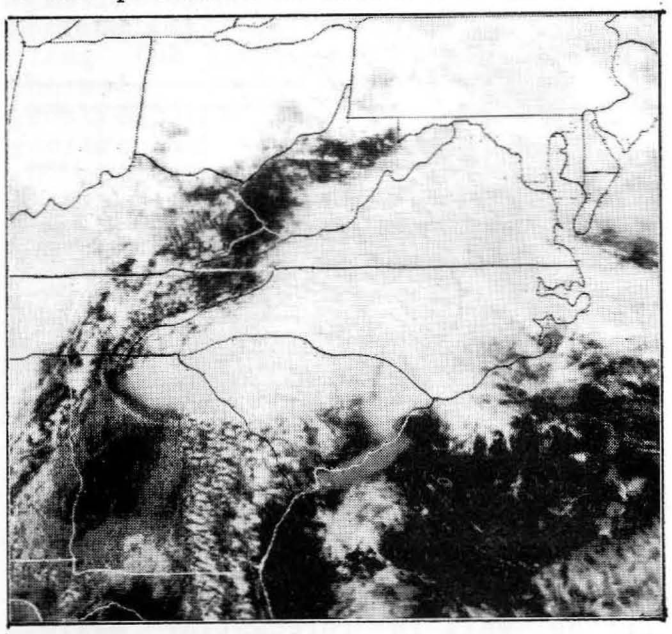


Figure 4. GOES Satellite photograph at 1831 GMT 25 March 1978.

rare satellite view of damming. It is rare because usually the stratus and stratocumulus cloud of the cold, dammed air is obscured by high- and middle-level cloudiness. Satellite movies of the day showed that east of the mountains the cool, stable air actually pushed southward when one might have expected it to remain stationary or even move northward with an advancing low- pressure system to the west. Fig. 5 shows Greensboro, NC upper-air soundings for the morning and evening of 25 March. These temperature profiles show that the layer of air below 850mb became more stable during the day; i.e., there was cooling below the inversion and warming above it. Also during the day, a 15 m/s 850-mb flow veered from a SSE direction to due S. West of the mountains, at synoptically analogous locations, temperatures were 10-15C warmer.

How well do today's National Weather Service (NWS) - National Meteorological Center (NMC) statistical and dynamical models handle damming? Not very well. Fig. 6 shows that the 12-h Limited Fine Mesh II (LFM II) numerical model sea-level pressure forecast was about 4 mb too low over the eastern slopes of the southern Appalachians. The reason for this poor forecast is probably not horizontal resolution, since the LFM II grid-spacing is smaller than the "nose" of high pressure representing the damming effect.

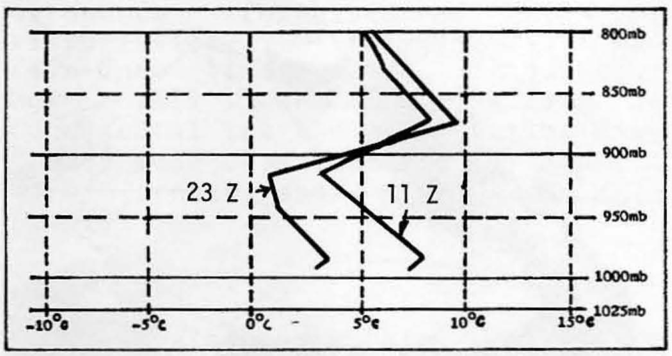
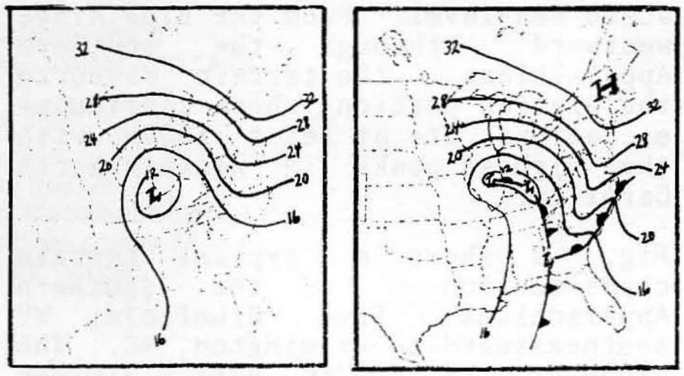


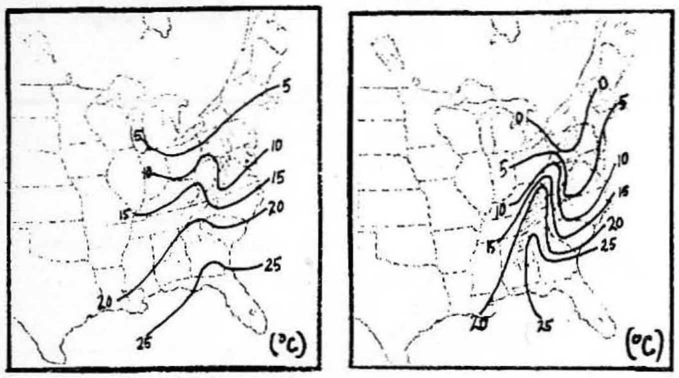
Figure 5. Upper-air soundings from Greensboro NC (GSO) for 25 March 1978.

Fig. 7 shows that even the Model Output Statistics (MOS) technique did not do well in forecasting maximum temperatures on 25 March. Over the eastern slopes, 24-h forecast maximum temperatures were 5-8C too warm -- a formidable error. Man-adjusted temperature forecasts from the various NWS Forecast Offices were not available for comparison. It can be seen, then, that the field forecaster still has the challenging mission of forecasting the meso-scale damming effect and its ramifications on the other meteorological parameters.



FORECAST ACTUAL

Figure 6. Comparison of 12-h LFM II FORECAST with ACTUAL sea-level pressure pattern (valid 0000 GMT 26 March 1978).



FORECAST ACTUAL

Figure 7. Comparison of 24-h Model Output Statistic (MOS) technique FORECAST with ACTUAL maximum temperatures (valid 0000 GMT 26 March 1978).

3. THEORY OF WHY DAMMING OCCURS

The flow of air over mountainous terrain in general has received considerable attention both theoretically and descriptively; e.g., mountain waves, foehn, etc. Near-neutral low-level (i.e., below 850 mb) stability is usually assumed or observed. Situations where statically stable air moves toward a mountain barrier have received far less attention. There is evidence that a mountain barrier should exert a kind of blocking action on air moving toward it, and that this effect should be proportional to its stability.

Baker (1971) searched exhaustively for, and qualitatively tested, previously proposed theories of the windward "nose" of high pressure and rejected all or part of each for one reason or another. In looking at a strong case of the high pressure ridge (26-28 April 1966), he came to the following observations. He rejected the at-first-glance hypothesis that the ridge is due simply to systematic errors in reducing pressure to sea-level over mountainous terrain. Rather, he concluded that the ridge "...is a real atmospheric phenomenon..." and that it merely is "reflecting changes in the inversion height aloft or, equivalently, changes in the depth of the cold air". Schwerdtfeger (1973) dynamically explained the unusual wind climatology of the area north of the Brooks Range in Alaska and concluded that "the streamlines of a shallow, stable airmass near the ground flowing toward a mountain barrier are deflected to the left (northern hemisphere) over the slopes of the obstacle. The change of direction becomes more pronounced as the temperature inversion (from the cold air nearest the ground to the warmer air aloft) becomes stronger." Schwerdtfeger's basic assumptions were that there is an inversion, or at least a lapse rate less than moist adiabatic; and that this inversion slopes in the same sense but not at necessarily the same rate as the terrain over which it is moving. Thus it is true (Fig. 8) that the pressure from level 2 to level 1 is inversely proportional to the mean virtual temperature of that layer. This implies an additional horizontal pressure gradient force at level 1 from B to A and, hence, an additional geostrophic wind component parallel to the obstacle (toward the viewer).

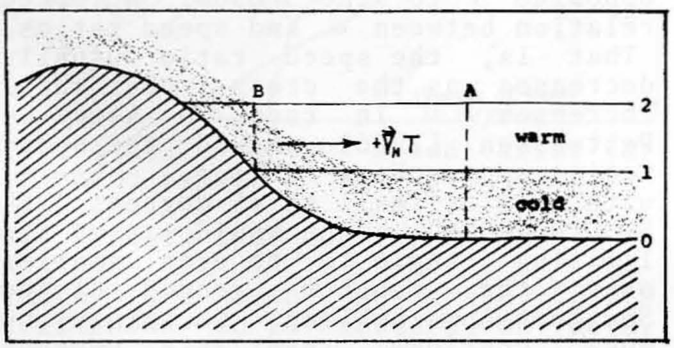


Figure 8. Schematic of stable air toward a mountain barrier.

An alternate and perhaps more elegant way to view the effect is to use the thermal wind concept. The thermal wind ($\vec{V}_{th}$) for a given layer of air is the vector difference between the geostrophic wind at some upper level ($\vec{V}_{gU}$) and that at a lower level ($\vec{V}_{gL}$) or:

$$\vec{V}_{th} = \vec{V}_{gU} - \vec{V}_{gL}$$

In the case shown in Fig. 8, a horizontal temperature gradient exists perpendicular to the contour lines of the terrain. Thus the $\vec{V}th$ for the layer between 1 and 2 parallels the obstacle, points into the paper and is given by:

$$\vec{V}_{th} = \frac{g}{f} \frac{\Delta T}{\bar{T}} (\vec{G} \times \vec{k})$$

(after Dalrymple et al., 1966) where g is the gravitational constant, f the Coriolis parameter, T the temperature difference between the cold air near the ground and warm air aloft, $\bar{T}$ the average temperature of warm and cold air, and $\vec{k}$ the unit vector upward. If then the geostrophic wind at some undisturbed level above the mountains is known, the surface wind can be predicted along the slope by subtracting the thermal wind from the upper geostrophic wind to obtain the lower geostrophic wind, and then applying an appropriate correction for the friction effect.

4. TESTING THE THEORY

Damming occurs quite frequently. To determine how often, and to test the theory, we examined all NMC 1200 GMT surface pressure analyses from a 2-1/2 year period from January 1974 to May 1976 for the visual clue to damming -- that "nose" of high pressure. Of a total of 18 cases found, 15 were during the cold season (November through March) and 3 were during the warm season.

For each case appropriate surface and upper air parameters were measured and recorded from NMC analyses. These included observed surface winds ($\vec{V}_o$) and geostrophic winds as suggested by the sea-level isobars ($\vec{V}_{gsL}$) by using a geostrophic wind scale. To test the proposed theory an undisturbed level above the mountains had to be assumed. We used

850mb because only scattered peaks in the southern Appalachians reach that level, and distortion of the 850-mb flow pattern and isotherms has never been noticed. Accordingly, 850-mb geostrophic winds ($\vec{V}_{g85}$) were measured again by using a wind scale. Also, soundings were plotted for all upper-air stations to determine the stability of the lower levels of air.

All parameters were then averaged to find the thermal wind caused by stable air flowing toward the mountain barrier. Greensboro NC was chosen as the test station because it is an upper-air station and is a rather typical slope location. The following average values for 16 cases for Greensboro yielded the vector solution to Eq. (2) shown in Fig. 9. The predicted surface wind ($\vec{V}_s$) is very close to the average observed surface wind ($\vec{V}_o$) especially when compared to the surface wind one might have deduced from the sea-level isobars, ($\vec{V}_{gsL}$).

$$\overline{\vec{V}}_{g85} = \text{SSW at } 12.5 \text{ m/sec}$$

$$\overline{\Delta T} = 6^{\circ}C$$

$$\overline{T} = 279^{\circ}K$$

$$|\vec{G}| = 7/1000$$

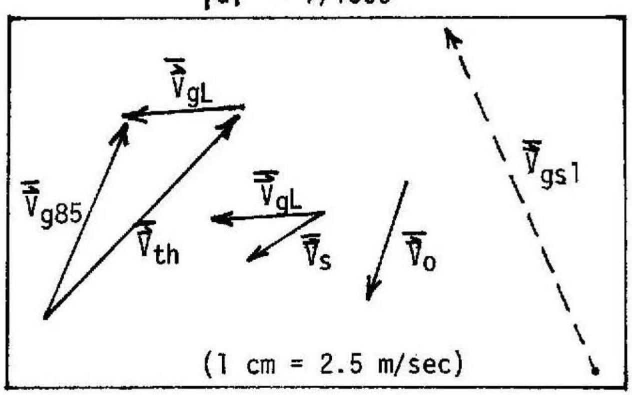


Figure 9. Vector solutions to Eqs. (1) and (2) using average values for Greensboro NC.

A nearly perfect ($\vec{V}_s$) would have been predicted if ($\vec{V}_{g85}$) were turned counter-clockwise and reduced in strength. As determined later, this would be the orientation of the 850-mb flow pattern during an earlier stage of damming. And since the "nose" of high pressure appears during the later part of the damming process, our method of selecting cases was accordingly biased.

To determine quantitatively how poorly the sea-level isobars represent the actual flow over the eastern slopes of the southern Appalachians when damming is occurring, average cross-isobar angles ($\bar{\alpha}$) between ($\vec{V}_s$) and ($\vec{V}_{gsL}$) were determined along with average speed ratios $(|\vec{V}_o| / |\vec{V}_{gsL}|)$. Plots of these are shown in Figs. 10 and 11. The number of cases shown in parentheses beside each station is usually less than the total 18, because often damming did not extend over the whole southern Appalachians; and, at some times, data were not available. No detailed discussion of friction or local effects is intended in this brief sketch, but the pattern shown in Figs. 10 and 11 warrant comment -- especially over the sloping terrain. Over the relatively flat Coastal Plain the correlation between $\bar{\alpha}$ and average speed ratios agrees reasonably well with previous studies. Haltiner and Martin (1957) show α ranges from around 10 degrees over smooth ground and/or with unstable conditions to 50 degrees over rough ground and/or with stable conditions. Inland of the Coastal Plain, over the Piedmont Plateau and the slopes, $\bar{\alpha}$ curiously increases to a 120 degrees maximum band over the steepest slopes. (Recall that this is an average value of between 12 and 18 cases.) In some cases individual α's exceeded 180 degrees; i.e., the actual wind was blowing exactly opposite to the direction suggested by the sea-level isobars. Also in individual situations, calm winds were observed with 20-30 m/s geostrophic gradients! Notice that the speed ratio remains around 0.3 or perhaps increases slightly over the slopes. This is opposite to the usual accepted relation between α and speed ratios. That is, the speed ratio usually decreases as the cross-isobar angle increases. In fact, a study by Petterssen (1956) showed the speed ratio approaching 0 (i.e., $|\vec{V}_o| \rightarrow 0$) with α's as small as 45 degrees OVER SMOOTH GROUND. One would expect the limiting angle to be even smaller over the rougher ground of the Southern Appalachians. Certainly one must conclude that this unusual pattern of increasing α's with increasing speed ratios is simply an indication that the sea-level pressure pattern does not represent the true flow when damming is happening.

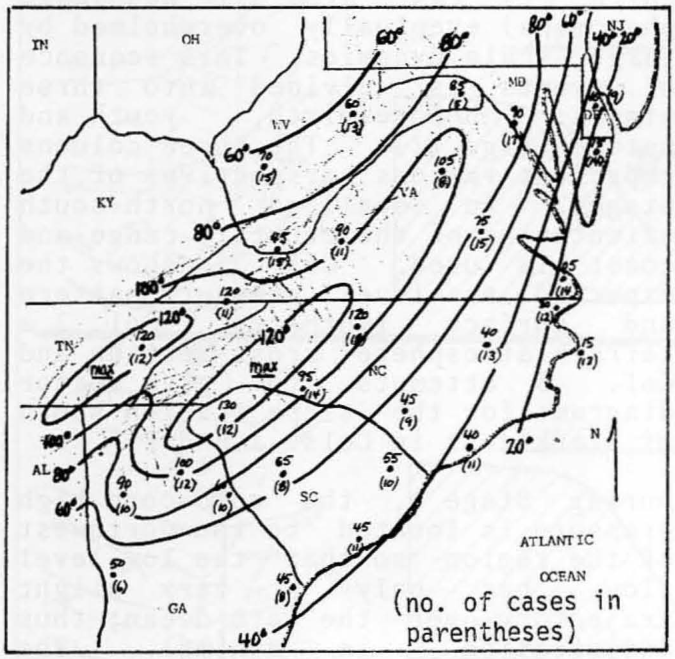


Figure 10. Average cross-isobar angles (α's) for selected cases, in degrees.

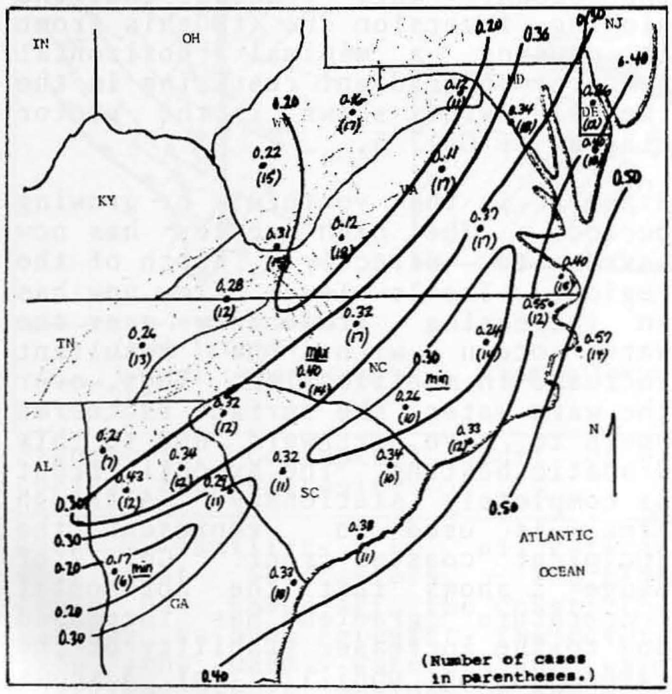


Figure 11. Average speed-ratios for selected cases.

5. RELATED EAST-COAST PHENOMENA

There are two mesoscale fronts well-known to forecasters on the U. S. east coast. They are, in part, due fundamentally to the damming effect of the Appalachians; they are the so-called "coastal" warm front and the "backdoor" cold front. This section describes each system and discusses the contribution of damming to each.

5.1 COASTAL FRONTS

A "coastal" front is a mesoscale pseudo-warm or stationary front which forms mainly during the winter months along certain parts of the east coast of the United States. Bosart et al (1972) studied those which form along the southeastern New England coast, and described them in the following way:

> "Temperature gradients of 5-10C over 5-10 km separating light north or northwest winds from strong easterly flow are common. The coastal thermal contrast may follow 6-12h after establishment of a cold anticyclone to the north of New England such that winds back to northerly... The frontal line is often the dividing line between frozen and non-frozen precipitation. Coastal fronts tend to occur on time scales of 6-12h in advance of the passage of the coastal low pressure center and create a serious forecast challenge."

Orography, coastal configuration, land-sea temperature contrast and friction all play roles in the formation of coastal fronts. The relative importance of each is still open to speculation.

We have noticed, and Bosart et al (1972) have observed, that coastal fronts also form along the Carolina coast (and to some extent along the Texas coast). This is not surprising since the geography of the two regions is so similar. Along both coastlines the land mass protrudes out into the Atlantic Ocean, terminating at Cape Cod in New England and Cape Hatteras in the Carolinas. Also, during the winter the ocean is warm relative to the air masses -- especially during the early part of the season. In the case of the Carolinas, the 15-20C Gulf Stream current is immediately offshore. The land-sea temperature contrast in New England is probably about the same, with the "cooler" warm ocean being offset by the less-modified and therefore colder air of the migrating Arctic high pressure systems.

To determine semi-quantitatively what role mountain-damming may play in coastal frontogenesis, we examined in detail the formation of such a front which formed on 5 February 1975 over the eastern Carolinas. The key observation of that study was that during the 24h preceding the detection of the coastal front, the 850-mb geostrophic wind at slope station Greensboro NC veered from ESE at 3 m/s to SSE at 15 m/s, while the surface winds over the sloping terrain remained NE at about 5 m/s. Keeping the theory of damming in mind, one way for this to happen would be for the stability of the uphill-moving air to increase. Indeed, Greensboro's upper air soundings showed that an inversion at 900mb increased in strength from 3C to 10C during the 24-h period. With all other things being equal, this means that the thermal wind due to the sloping stable air would have had to triple in magnitude to maintain the lower geostrophic (and thus surface) wind. This observation led us to develop a conceptual model of how mountain-damming contributes to coastal frontogenesis.

There is more evidence that the orography plays a major role. Bosart et al (1972) said that the New England coastal fronts form locally (i.e. do not advect in from the sea) and tend to stagnate on a Boston-Providence RI line. That line is parallel to the terrain contours. We noticed the same orientation over the Carolinas. This observation that COASTAL FRONTS PERSIST OR STAGNATE ON LINES MORE PARALLEL TO THE TERRAIN CONTOURS RATHER THAN TO THE COASTLINE seems to be convincing evidence that mountain-damming does play an important role in the formation of coastal fronts.

The result of this study along with those of Bosart et al (1972), Bosart (1975), and Baker (1971) were used to develop the conceptual model shown in Fig. 12. The formation and maintenance of the coastal front are seen as resulting from a positive-feedback sequence of events which is (as are all mesoscale phenomena) eventually overwhelmed by larger-scale dynamics. This sequence of events is divided into three stages: before-birth, youth and mature (Fig. 12). The three columns represent various perspectives of the stages. An idealized north-south orientation of the mountain range and coast is used. Col. 1 shows the expected sea-level pressure pattern and surface isotherms, Col. 2 a terrain-atmosphere cross-section and Col. 3 attempts to give vector diagrams for the slope station shown by black dots in Cols. 1 and 2.

During Stage 1, the cold-core high pressure is located to the northwest of the region so that the low-level flow has only a very slight trajectory over the warm ocean; thus modification is minimal. The synoptic-scale front has become stationary over land but is still moving southward as a cold front over the ocean. Col. 2 shows that the sloping inversion due to this front is causing a minimal horizontal temperature gradient resulting in the thermal wind shown in the vector diagram of Col. 3.

Stage 2 is the youthful or growing period. The high center has now moved to directly north of the region. The low-level flow now has an increasing trajectory over the warm ocean with the resultant increase in modification. Thus, over the warm water the surface isotherms begin to move northward due to this diabatic heating. The synoptic front is completely stationary. A trough line is used to represent the incipient coastal front. Col. 2 of Stage 2 shows that the horizontal temperature gradient has increased due to the increased stability of the layer moving uphill. Col. 3 shows that the lower geostrophic wind is being maintained by the stronger thermal wind balancing the veering and increasing upper geostrophic (850mb) wind. THIS CONTINUAL STRENGTHENING OF THE SLOPE-INDUCED THERMAL WIND DUE TO THE INCREASED STABILITY IS THE FUNDAMENTAL BASIS FOR THIS MODEL.

Finally, the sea-level pressure pattern of Stage 3 is one that should

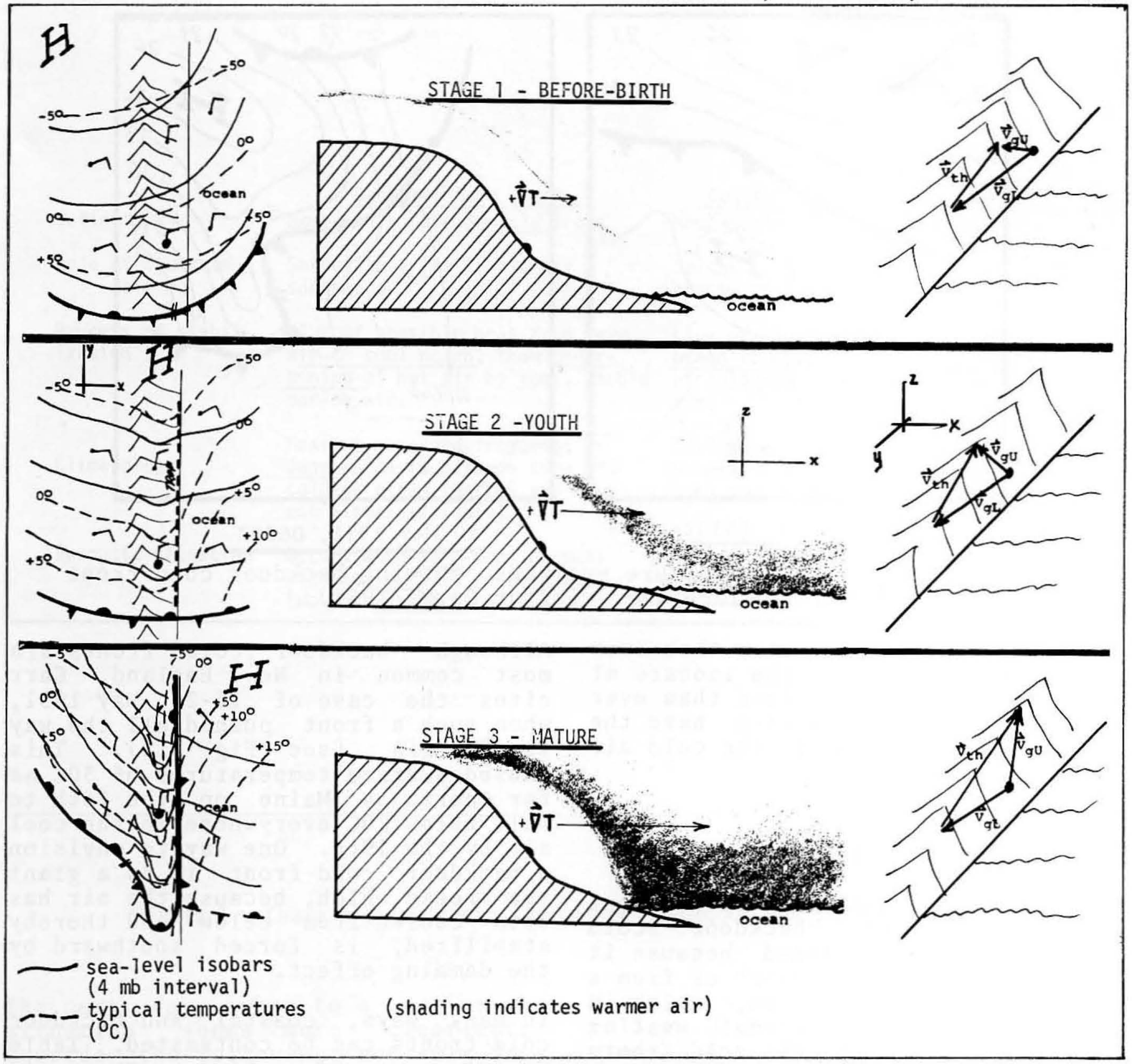


Figure 12. Conceptual model of how Appalachian damming contributes to coastal frontogenesis.

look familiar to all veteran east-coast forecasters. The high center has moved off the coast; the air has an even greater trajectory over the warm water; and warm maritime air is moving up and over the slope-induced circulation which is continually being strengthened. The "nose" or ridge of high pressure is now very conspicuous because of the damming and because the pressure has lowered more rapidly on either side of the eastern slopes. It has lowered seaward as the hydrostatic reflection of warm, maritime air replacing the colder, continental air. Over the mountain tops, the pressure has lowered because the synoptic front has become a northward-moving warm front, causing the depth of cold air to decrease, thus resulting in less pressure. In some cases on synoptic analyses, the synoptic warm front is joined to the coastal front as we have done. Really, though, the two fronts are very different. The meso-scale coastal front has a vertical-length scale closer to 10km.

Other physical principles play roles in the formation of coastal fronts. The most important of these is probably friction. Since friction

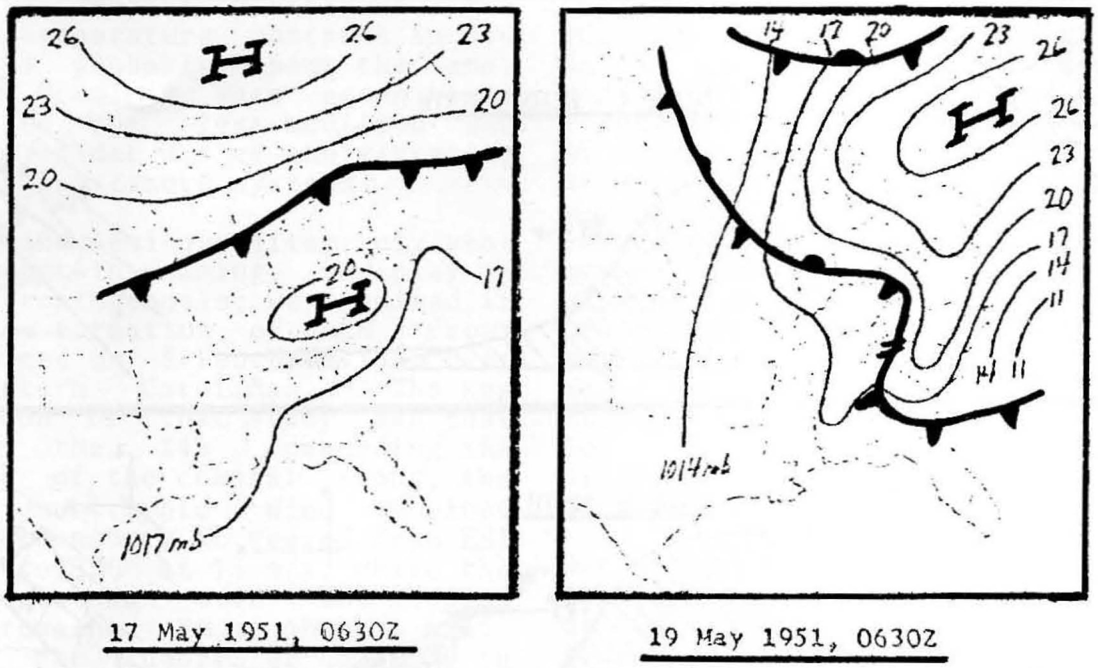


Figure 13. Sea-level pressure analyses showing backdoor cold front over the Southern Appalachians. (after Carr, 1951).

over land is greater than that over water, the winds cross the isobars at a greater angle over land than over water. This would also have the effect of maintaining the cold air southward.

5.2 BACKDOOR COLD FRONT

The other east-coast mesoscale front is the so-called "backdoor" cold front. It is so named because it sneaks in the "back door" or from a northeasterly direction, which is contrary to the normal west or northwest approach of cold fronts moving in the westerlies. Carr (1951) says:

> "The favored geographic regions for the occurrence of 'backdoor' cold fronts in the U.S. are the east slopes of the Rocky Mountains and east slopes of the Appalachian Mountains. In those regions cold air banks up against the mountains and is forced southward AT A WIDE ANGLE ACROSS THE ISOBARS (emphasis added) in the area. Thus, the cold air advances farther south than it would if the orographic effect were absent."

Although backdoor cold fronts are most common in New England, Carr cites the case of 16-20 May 1951, when such a front pushed all the way to Georgia (see Fig. 13). This caused surface temperatures of 30C as far north as Maine on the 16th to fall below 20C everywhere in the cool air by the 19th. One way to envision a backdoor cold front is as a giant sea breeze which, because the air has been cooled from below and thereby stabilized, is forced southward by the damming effect.

In many ways, coastal and backdoor cold fronts can be contrasted. Table 1 tries to do this. The main difference in the two fronts is the way in which the low-level flow is stabilized. Once this stabilization happens, the damming effect can start; and, depending on the season (which controls the land-sea temperature contrast), either a coastal or a backdoor cold front can result.

6. GUIDELINES FOR FORECASTERS

Before giving specific rules, let us review some "old-timers'" methods of dealing with the damming effect.

Veteran forecasters of mid-Atlantic weather often forecast "rain or snow", or "rain changing to snow" for

	Backdoor Cold Front	Coastal Front
Definition	Meso-scale, pseudo cold front	Meso-scale, pseudo warm front
Role of mountain-damming	Serves to drive stable air southward.	Serves to maintain stable air southward.
Process of stabil-ization	Flux of sensible heat from warm air to cool ocean, then under-mining of hot air by cool, stable marine air.	Flux of sensible heat from warm ocean to cold air, then overriding of cold continental air by marine air.
Climatology	Maximum observed frequency in June which is between time of coldest ocean (March) and warm-est airmasses (July).	Maximum observed frequency in December between times of warmest ocean (Septem-ber) and coldest airmasses (January).
Synoptic Situation	Occurs when warm-core Bermuda high to south is retrograding (moving westward).	Occurs when cold-core Arctic high to north is moving eastward.

Table 1. Chart showing contrasts between backdoor and coastal fronts.

the next day, when to a novice the situation seems more clear-cut. Consider the following: not only does the terrain rise landward in the mid-Atlantic region, but the slope of the cold air associated with damming and coastal fronts can be greater than the terrain slope. Apparently, then, these forecasters with 20-plus years of experience with the local area have a mental catalogue of weather situations that warns them to be ambiguous. More often than not, rightly so. In fact, some of the "old timers" have developed statistically-based forecast rules which can be found in the literature. For these mountain-effect problems, the rules are based on the central pressure of an anticyclone somewhere to the north, whose central pressure is a rough indicator of the cold air availability.

Here, then, are some guidelines which follow directly from the dynamics used in this study. The first question from a forecaster would probably be, "Just when should I expect damming to occur?" Two conditions must be fulfilled. The first is that the layer of air below 850mb must have a lapse rate less than moist adiabatic. Since this is often the case, the degree of damming is then directly proportional to the static stability if the second condition is fulfilled. That is, the lower layers of air must have some component of the flow directed uphill; i.e., perpendicular to the terrain contours.

Probably the most important things forecasters want to know is how to determine the space and/or time beginnings and endings of damming.

Since beginning this study, we have found that the 850mb pattern is an excellent semi-dynamical, semi-empirical predictor of these considerations. At the location and/or time that the 850mb flow has any component directed toward the terrain contours, some damming will occur. For example, in winter the 850mb flow usually veers (turns clockwise) with time as high pressure moves by to the north and/or low pressure to the south. Since the mountains are oriented NNE-SSW, damming usually starts with a NE wind and does not cease until the 850mb wind veers to SW. The northern and southern extremities of the damming, then, are where the 850mb wind has these directions. Remember that the degree or intensity of the effect will be directly proportional to the static stability.

Even if a forecaster has missed the start of the damming, he can redeem himself by using another result of this study. We have seen repeatedly that THE SEA-LEVEL ISOBARS CANNOT BE USED WITHOUT MODIFICATION to diagnose or predict the surface winds over the eastern slopes of the Appalachians. Our statistics showed that when damming is occurring, surface winds regularly blow 90-180 degrees different from, and at only 20-40 percent of, the speed suggested by the sea-level isobars. So, once the tell-tale "nose" of high pressure has appeared, wind forecasts can be adjusted.

Before the advent of numerical models of the atmosphere, forecasters were concerned with the weather for tomorrow. Now, with reasonably good forecasts of the location and intensity of synoptic-scale features for at least 36-48h ahead, forecasts of mesoscale damming and related coastal and backdoor cold fronts should be possible 2 or even 3 days in advance, rather than just for the next day.

6. CLOSING

Until "perfect" numerical models of the atmosphere and its interaction with orography and the oceans are developed, the forecaster still has a challenging mission. This mission is constantly changing, however. We hope that this study serves to focus forecaster interest and enthusiasm to where it should now be, to make the best of both "man and machine" to make as-near-perfect-as-possible forecasts.

7. ACKNOWLEDGMENTS

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REFERENCES

Baker, D.G., 1970: A Study of High Pressure Ridges to the east of the Appalachian Mountains. Ph.D. Thesis, Dept. of Meteorology, Massachusetts Institute of Technology, 127 pp.

Bosart, L.F., C.J. Vaudo and J.H. Helsdon Jr., 1972: Coastal Frontogenesis. J. Appl. Met., 11, pp. 1236-1258.

__________, 1975: New England Coastal Frontogenesis. Quart. Jour. Royal Met. Soc., 101, pp. 957-978.

Carr, J.A., 1951: The East Coast Backdoor Cold Front of May 16-20, 1951. Mon. Wea. Rev., 79, pp. 100-105.

Dalrymple, O., H.Lettau and S. Wollaston, 1966: South Pole Micrometeorology Program: Data Analysis. In "Studies in Antarctic Meteorology" (M.J. Rubin, ed). Amer. Geophys. Union Antarctic Research Series 9, pp. 13-58.

Haltiner, G.J. and F.L. Martin, 1957: Dynamical and Physical Meteorology. McGraw-Hill, New York, 470 pp.

Petterssen, S., 1956: Weather Analysis and Forecasting, Vol. 1. McGraw-Hill, New York, 428 pp.

Schwerdtfeger, W.E., 1973: "Mountain Barrier Effect on the Flow of Stable Air North of the Brooks Range". Climate of the Arctic, pp. 204-208.