The Orographic Modulation of Pre-Warm-Front Precipitation in Southern New England

RICHARD E. PASSARELLI, JR., AND HANNAH BOEHME

Massachusetts Institute of Technology, Department of Meteorology and Physical Oceanography, Cambridge, MA 02139

(Manuscript received 8 June 1982, in final form 20 January 1983)

ABSTRACT

Topographic forcing over the hills and small mountains of southern New England plays an important role in determining the distribution of pre-warm-front precipitation from winter cyclones. Upslope regions receive 20–60% more precipitation than do nearby downslope or coastal regions. Both the intensity and duration of precipitation contribute to the positive upslope anomalies. The magnitude of the upslope anomalies, the details of the precipitation intensity distributions at proximal upslope and downslope gauges, and the results of simple models indicate that precipitation scavenging in orographic clouds can explain the orographic enhancement. Also, the existence of a positive precipitation anomaly over the coastal plain suggests that frictional convergence may be generating weak, but persistent vertical motions.

1. Introduction

Understanding how topography influences the local distribution of precipitation is an important step towards formulating improved quantitative precipitation forecasting techniques. Even a cursory glance at a climatological atlas shows that mountains have a strong effect on the long-term average distribution of precipitation. Unfortunately, this information is of little use in assessing how a particular storm event will be influenced by topography or in providing insight into the physics of orographic effects. In this paper we examine the magnitude and nature of orographic effects that influence pre-warmfrontal precipitation over the southern New England region of the United States.

Heavy cold-season precipitation in southern New England usually falls in advance of a surface warm front. During these situations the atmosphere is typically very stable at low-levels and the surface winds are easterly. The terrain features (300–500 m high) are oriented in north–south ranges, normal to the low-level easterly flow. While the orientation of the terrain with respect to the pre-warmfrontal winds is favorable for orographic precipitation, the modest relief and low-level stability are not. Indeed, Browning and Harrold (1969) characterized pre-warmfrontal orographic effects as "altogether negligible" based on a very detailed case study in the British Isles. These authors and others (e.g., Douglas and Glasspoole, 1947; Sawyer, 1956; Browning *et al.*, 1974) point out that orographic effects over Britain are strongest during the southwesterly flow ahead of a cold front where the atmosphere is typically moist and conditionally unstable. Elliot and Shaffer's (1962) study of orographic precipitation over the western mountains of the United States shows that large orographic precipitation rates are favored by a conditionally unstable atmosphere. In contrast Wilson and Atwater (1972) found significant orographic anomalies in storm total precipitation amounts over the small terrain features in Connecticut during easterly flow precipitation events.

While the stability of the atmosphere during pre-warmfrontal precipitation events does not favor large orographic precipitation rates, it is certainly true that the associated easterly flow and large-scale precipitation are persistent (12–24 hrs). Since precipitation rates are typically small (1–2 mm h^{-1}), it is possible that a weak orographic effect, integrated over the duration of a storm, could impact the spatial distribution of precipitation.

In this study, hourly raingauge data are used to examine the magnitude and nature of orographic effects over southern New England during pre-warm-front periods. The results show that stable orographic lifting over the hills and small mountains of southern New England plays an important role in determining the distribution of pre-warmfrontal precipitation. The most striking anomaly is that upslope regions receive 20–60% more precipitation than adjacent downslope or coastal regions. Both the intensity and duration of precipitation contribute to the upslope anomalies. Several mechanisms for the upslope enhancement are discussed, and it is concluded that the scavenging of orographically forced or enhanced cloud by the large-scale precipitation is probably the dominant mechanism (e.g., Bergeron, 1949; Sawyer, 1952). This hypothesis is supported by the results of a comparison of proximal upslope and downslope rain gauge data and by the results from some simple models of orographic precipitation. These results also show that

© 1983 American Meteorological Society

Unauthenticated | Downloaded 10/03/23 11:28 PM UTC

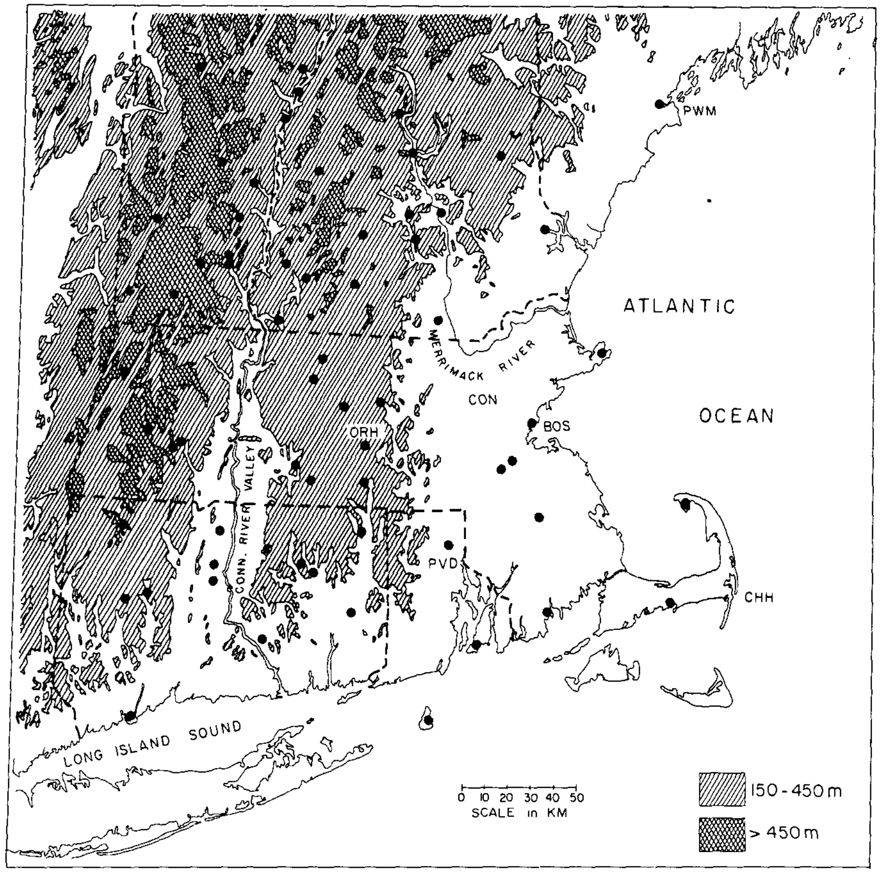

FIG. 1. Terrain of southern New England and the locations of the raingauges.

even thin (~500 m) orographic clouds can account for the observed upslope anomalies. The analyses also suggest that frictionally-induced convergence may act to enhance precipitation over the relatively flat coastal plain.

2. The terrain in southern New England

Fig. 1 shows the topography in the study region. There are two north–south oriented regions of higher terrain. The eastern hills are bounded by the coastal plain to the east and the Connecticut River Valley to the west. The western mountains are bounded by the Connecticut and Hudson River Valleys. Also, note the Merrimack River Valley.

Locations of the 66 rain gauges used in this study are indicated by the black dots. These gauges are maintained by various government agencies and hourly data are compiled by NOAA and reported in the New England Hourly Precipitation Data publication.[1] Of the 66 gauges, 34 have 0.01 inch resolution, 20 have 0.1 inch resolution and 12 were a mixture of these two resolutions for the storms that were studied.

Fig. 2 is an east–west cross-section of average terrain elevations for the latitude strip between 42 and 43°N—nominally the Commonwealth of Massachusetts. From the east, the coastal plain rises gradually from sea level to ~80 m, the eastern hills rise to ~300 m, and the western mountains rise to ~550 m. The Connecticut and Hudson River Valley drop to ~100 m. Approximate values of the terrain slope (in %) are indicated for the various upslope and downslope regions.

3. Storm selection criteria

We have selected twelve cyclones that produced sustained periods of easterly flow and widespread precipitation over the study area. Before selection of the storms, Bosart's list (personal communication, 1980) of coastal front events was consulted so that none of these would be included. The New England coastal front is known to enhance precipitation and is characterized by strong low-level confluence and temperature contrast. Surface synoptic analyses were made for each storm to determine the type of precipitation (rain or snow), to examine the temporal

[1] Available through the National Climatic Center, Asheville, NC 28801.

Unauthenticated | Downloaded 10/03/23 11:28 PM UTC

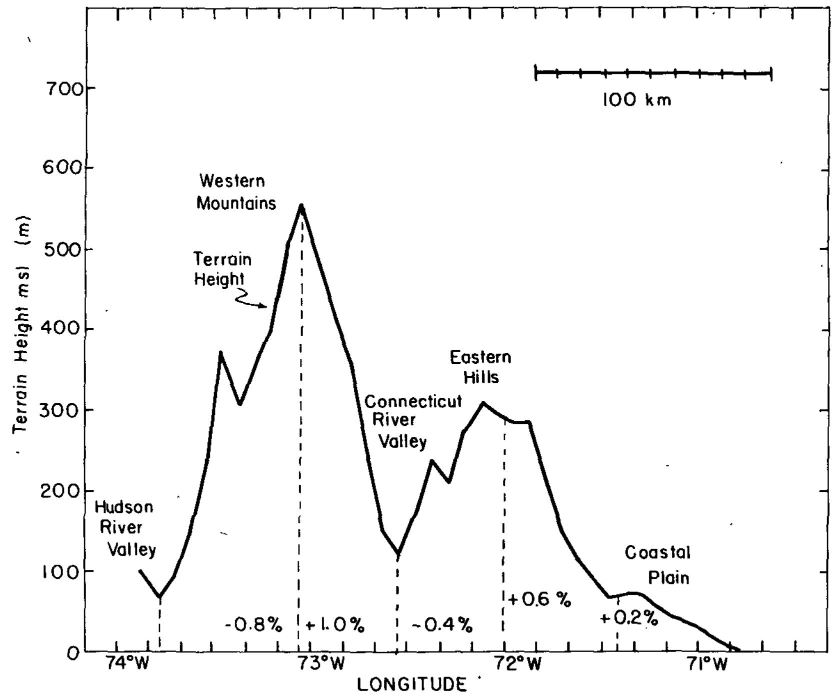

FIG. 2. Cross-section of the terrain averaged over the state of Massachusetts showing the various regions that are discussed in the text. Average values of the east-to-west terrain slope (in percent) are given for each region.

behaviour of the surface winds and to double check for the presence of a coastal front. The storms included in this study are predominantly rain because of the difficulty in obtaining accurate measurements of equivalent snow depth.

The hourly precipitation data are analyzed only during the period of surface easterly flow as determined from the hourly wind reports at four stations within 100 km of Boston, Massachusetts (BOS, ORH, CON, PVD). The beginning of a storm period is defined as when the station average wind direction is between 60 and 120° and there is rain reported in any one of the 66 gauges. The end of a storm period is defined when the wind shifts out of this range of directions or when precipitation ceases at all gauges. The predominant surface wind direction is south of east for the defined storm periods. The storm dates, times and durations are given in Table 1. The average storm duration is 25 hours. The storm tracks, corresponding to the periods given in Table 1, are shown in Fig. 3 and are typical of storms that produce moderate or heavy cold-season precipitation over New England.

4. Precipitation analyses

For each of the 66 gauges and each of the twelve storms, the total precipitation and the number of hours during which precipitation actually fall were determined for the storm period. The precipitation total averaged over all gauges is given in Table 1 for each storm. Since the storms are of different durations, it is convenient to characterize the intensity of precipitation by defining a normalized storm total equal to the storm total at a gauge, divided by the defined storm duration. The normalized storm total averaged over all gauges is given in Table 1 for each storm. The average storm total is 27 mm, representing ~3% of the total annual precipitation for the region illustrating that these are significant precipitation events.

The area-average precipitation rate was computed for each storm by dividing the total precipitation at each gauge by the number of hours during which precipitation actually fell and then averaging over all gauges. Only the more numerous 0.01 inch resolution gauges were used. (This is discussed later.) Table 1 gives the area-average precipitation rates and the area-average duration of precipitation for each storm. The precipitation rates are greater than the normalized storm totals since precipitation was rarely continuous at any gauge during the entire storm period. Table 1 also gives the area-average fractional duration of precipitation (the duration of precipitation at a particular station, divided by the storm duration). Averaged over all storms and all stations, precipitation fell ~65% of the time.

The area-average storm properties given in Table 1 show that these storms are significant precipitation events due to their duration (~15 h of precipitation)

Unauthenticated | Downloaded 10/03/23 11:28 PM UTC

TABLE 1. Area-average precipitation summary.

Date	Time (GMT)	Storm duration (h)	Storm total precipitation (mm)	Normalized storm total precipitation (mm h^{-1})	Precipitation* rate (mm h^{-1})	Duration of precipitation* (h)	Fractional duration of precipitation* (h)
29–30 Nov 1971	1900–0500	11	24	2.2	2.9	8.4	0.76
Nov 30–1 Dec 1972	2100–1000	12	19	1.6	1.9	9.8	0.82
26–27 Apr 1973	0300–2000	41	25	0.6	1.3	20.9	0.51
14 Dec 1973	0200–1400	13	13	1.0	1.7	7.9	0.61
Mar 31–Apr 1, 1973	0400–0200	17	34	2.0	2.4	12.9	0.76
25–26 Apr 1976	0700–1300	31	25	0.8	1.4	19.8	0.64
22–23 Mar 1977	1500–0600	16	27	1.7	2.3	11.0	0.69
5 Apr 1977	0000–2000	21	25	1.2	1.8	14.9	0.71
23–25 Apr 1977	1100–0900	47	52	1.1	1.5	35.3	0.75
9–10 May 1977	0200–0200	25	28	1.1	1.8	13.8	0.55
7–9 Nov 1977	0700–0700	49	29	0.6	1.6	21.6	0.44
26 Nov 1977	0200–1600	15	17	1.1	1.8	7.7	0.51
Average over all storms		25	27	1.3	1.9	15.3	0.65

* Based on 0.01″ gauges.

not their precipitation rate (1–2 mm h^{-1}). At 0°C, a vertical velocity of 10 cm s^{-1} in a saturated atmosphere corresponds to an upward moisture flux of 2 mm h^{-1} which gives a measure of the pre-warmfrontal vertical air motion. For a terrain slope of 1% and a barrier-normal windspeed of 10 m s^{-1}, the orographically-forced lifting will be 10 cm s^{-1} at the surface which is comparable to the pre-warmfrontal upward motion. Thus one might expect orographic lifting to play a significant role in determining the distribution of precipitation.

Analyses of single storm events suggest that topography does influence the spatial distribution of precipitation, but there is considerable variability that is not clearly related to the topography. This variability is not surprising since the pre-warmfrontal region of this type of storm usually contains mesoscale bands or areas of heavier precipitation. The specific trajectories and life histories of these features undoubtedly contribute to the spatial variability of precipitation from a single storm. Hence the distribution of precipitation from a single storm has two components: orographic features (meaning those features tied to a geographic location) and non-orographic features (meaning those features that are spatially transient).

By averaging raingauge data over many similar storm events, one can generate an ensemble-average spatial distribution of precipitation for a particular storm type. Ideally, non-orographic features will make no contribution to the spatial variability of the ensemble-average pattern since they are (by definition) randomly distributed over the study area. The variability of the ensemble-average pattern is due only to effects that are somehow linked to specific geographic locations. Thus ensemble averaging over many similar precipitation events is appropriate to the study of orographic precipitation features.

Several quantities are available for ensemble averaging. Fig. 4 shows contours of the ensemble-average precipitation total, i.e., the precipitation total at a gauge averaged over all twelve storms. There is almost a factor of three variation in this quantity and three dominant features: a broad maximum over the upslope of the eastern hills extending eastward onto the coastal plain, a minimum over the Connecticut River Valley downslope region, and a maximum over the western mountain upslope. There is also a weak minimum associated with the Merrimack River Valley downslope and an apparent north–south trend (more precipitation to the south). We note that these patterns are very smooth and easily contoured.

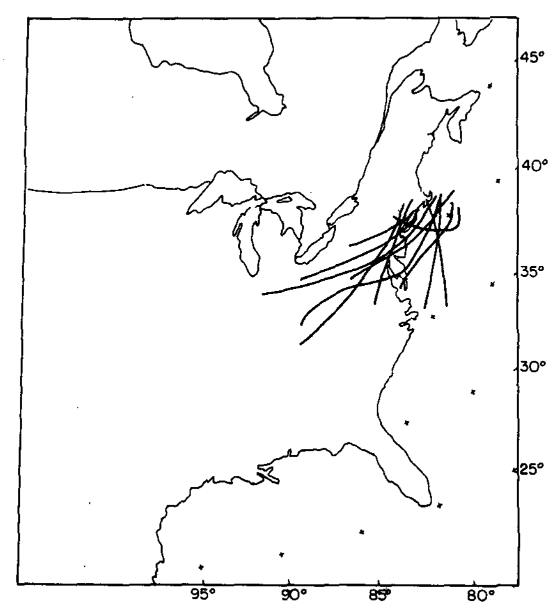

FIG. 3. Cyclone tracks for the defined storm periods.

Unauthenticated | Downloaded 10/03/23 11:28 PM UTC

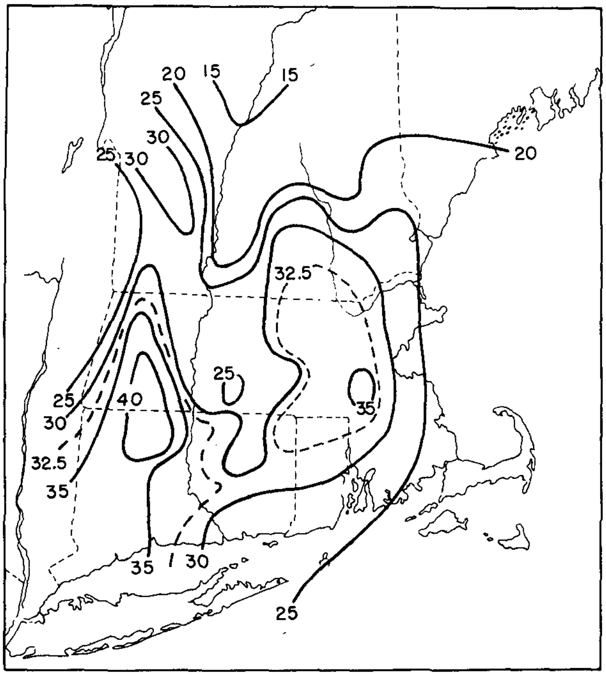

FIG. 4. Ensemble-average precipitation totals in mm.

For better illustration of the correspondence of the precipitation anomalies to the terrain, ensemble-average precipitation totals for stations in 10′ longitude strips were averaged together for the latitude strip 42–43°N. These are shown in Fig. 5 along with the corresponding terrain elevation. The three features are clearly tied to the terrain slope except for the broad maximum over the eastern upslope which extends over the relatively flat coastal plain.

The ensemble-average precipitation totals clearly show that upslope gauges receive 20–60% more precipitation than downslope or coastal gauges to the east. It is interesting that the precipitation totals over the Connecticut Valley downslope are approximately equal to the totals at the coast to the east. This suggests that upslope regions act to enhance total precipitation while downslope regions have little net effect.

Further, the duration of precipitation can be studied by examining the fraction of time that a gauge receives precipitation during a defined storm period. This quantity was computed for each station, for each storm, and then averaged for each station over all storms. The difference between the two gauge resolutions (0.1 and 0.01 inch) creates a problem. Since precipitation rates were usually less than 0.1 inch per hour, the 0.1 inch gauges report fewer hours of precipitation. The more numerous 0.01 inch gauges more accurately reflect the actual duration of precipitation. The computation of the ensemble-average fractional duration of precipitation was done separately for each gauge type and then the 0.1 inch gauge results were multiplied by an adjustment factor (~2) so that the average duration, over all stations, for the two gauge types was identical.

Fig. 6 shows contours of the ensemble average fractional duration of precipitation (expressed in percent). The contour analysis was performed for the 0.01 inch gauges first. The adjusted 0.1 inch gauges were added later but did not produce any substantial changes in the pattern. Averaged over all stations, precipitation fell 65% of the time. Regions where the precipitation duration is above average are shaded. Again the correlation with terrain slope is clear. The duration of precipitation is 20–40% longer over upslope regions as compared to stations along the eastern shore. The precipitation duration over the Connecticut River Valley downslope is the same or greater than over the eastern shore.

The precipitation rate for each storm and each station is computed by dividing the total precipitation by the number of hours of precipitation. This was done separately for both gauge types and the 0.1 inch gauges were adjusted to the 0.01 inch values using the same adjustment method that ws applied in the previous analysis. Fig. 7 shows contours of the ensemble-average precipitation rate in mm h^{-1}. The average rate is 1.9 mm h^{-1}, and areas of above-average precipitation rate are shaded. The same basic anomaly pattern is present, but there is a strong maximum along the southern coast. The eastern maximum is rather broad, including the eastern mountain upslope and the coastal plain.

5. Discussion of orographic mechanisms

The analyses thus far show a clear orographic effect on the windward slopes of the hills and small moun-

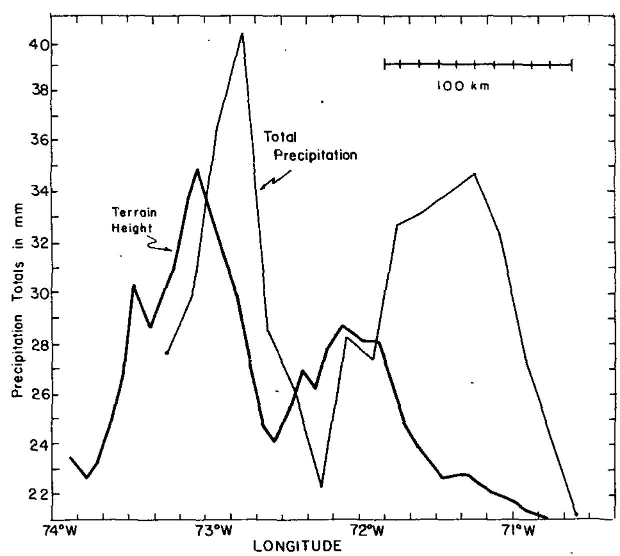

FIG. 5. Average precipitation total as a function of longitude for Massachusetts. The terrain elevation from Fig. 2 is shown for reference.

Unauthenticated | Downloaded 10/03/23 11:28 PM UTC

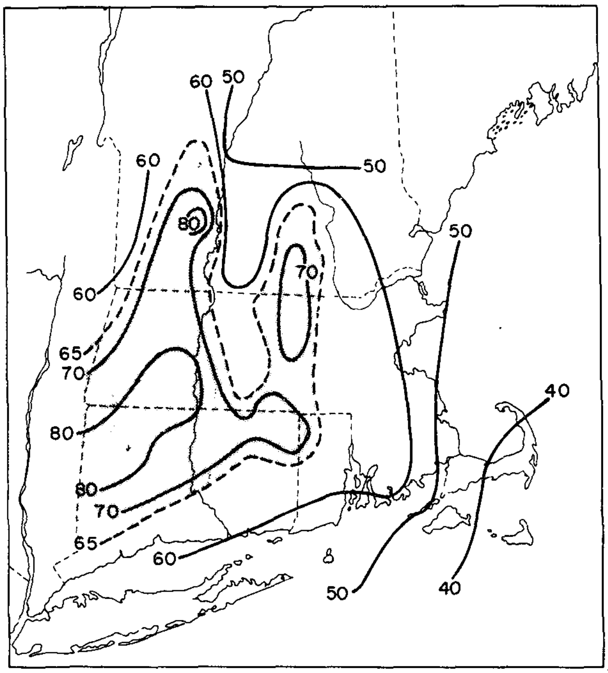

FIG. 6. Ensemble-average fractional duration of precipitation in percent. The fractional duration is taken with respect to the defined storm duration.

tains. Both intensity and duration of precipitation appear to contribute to the greater upslope precipitation amounts. In this section we will examine various mechanisms of orographic precipitation and discuss them in light of the observations.

Before considering mechanisms that may contribute to the orographic precipitation anomalies it is important to understand the "typical" storm environment. Soundings at Albany, Portland and Chatham were examined for every case and these show that the atmosphere is typically very stable below the warmfrontal inversion and slightly stable (with respect to moist adiabatic ascent) aloft. Saturated or near saturated conditions prevail within the lowest 6 km. The atmosphere is highly sheared at low levels, with easterly winds at the surface going to southwesterly winds at ~2 km. Of course the details vary from storm-to-storm, but these general thermodynamic and kinematic conditions are common to all the storms selected for study.

The low-level stability of the environment is not conducive to the triggering of large convective clouds by forced lifting. Heavy rain gushes, lightning and other signatures of deep convection are rare during these storms Recently Smith (1981) hypothesized that blocking of the low-level flow by the terrain, coupled with cold advection aloft, could lead to convective instability over and upwind of a mountain barrier. However, since the pre-warmfrontal region is characterized by warm advection aloft, it is unlikely that this mechanism operates during these storms.

Forced lifting over the hills and small mountains in the stable, sheared environment will produce gravity waves and, since the environment is near saturation, the gravity waves will produce clouds. It is doubtful that lee waves are contributing to the ensemble-average precipitation anomalies since variations in environmental conditions will likely lead to a melange of amplitudes, phases and wavelengths. At any instant, lee-wave vertical motion may contribute to positive and negative precipitation anomalies, but these contributions will be transient in space and time and, hence, incoherent in the ensemble average.

The regions of forced upslope motion will be spatially coherent from storm-to-storm and these upslope regions coincide with positive precipitation anomalies. Bergeron (1949; 1965) postulated that shallow lifting over rather modest terrain features leads to shallow clouds and that it is the scavenging of these clouds by the large scale precipitation that causes the enhancement of precipitation on windward slopes. On the basis of the data presented here it is not possible to either confirm or reject the hypothesis that low level scavenging in shallow orographic clouds is responsible for the observed precipitation anomalies, but the mechanism can explain various aspects of the observations. The remainder of this section is devoted to examining this mechanism in greater detail.

First, the precipitation anomalies are in close coincidence with terrain features (e.g., Fig. 5). This

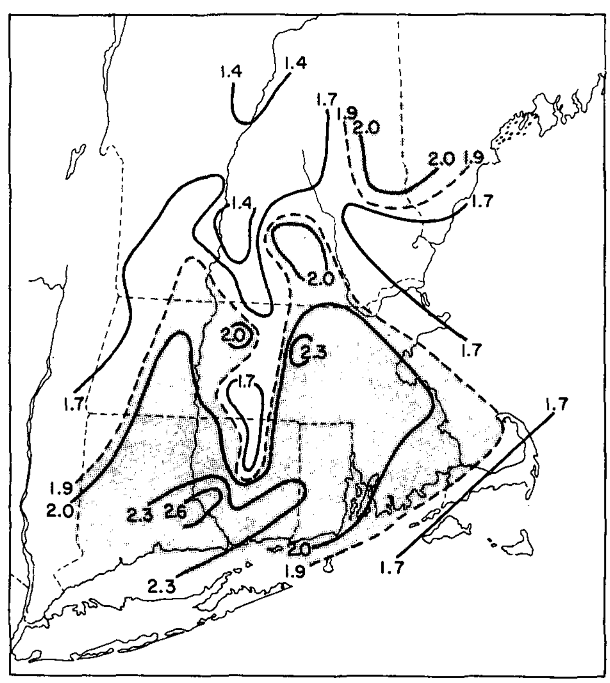

FIG. 7. Ensemble-average precipitation rate (mm h^{-1}), i.e., the total precipitation divided by the number of hours during which precipitation actually fell.

Unauthenticated | Downloaded 10/03/23 11:28 PM UTC

might not be the case if the orographic enhancement were occurring at great altitude since the precipitation particles could be advected away from the region of enhancement. For example, an ice crystal at 6 km, falling at 1 m s^{-1} to 1 km and then melting to rain and falling at 6 m s^{-1} requires 1.4 hours, during which time it could travel a considerable distance. The storm-to-storm variations in the wind profile would smear the resulting anomaly pattern. Hence, the close correspondence between terrain features and the precipitation anomalies suggests low-level enhancement.

If low-level enhancement is occurring over upslope regions then one would expect the upslope precipitation rates, on average, to be slightly greater than the downslope precipitation rates. On the other hand, if deep convection were occurring over upslope regions, one might observe that the upslope anomalies were due to short periods of intense precipitation. Thus the character of precipitation over upslope and downslope regions is an indicator of how the precipitation was formed. The spatial variability of precipitation precludes an hour-by-hour comparison of upslope and downslope precipitation records, but one would expect similar statistical properties of the precipitation at proximal gauges. For example, the frequency distribution of hourly precipitation amounts should be similar at proximal gauges.

To compare the characteristics of upslope versus downslope precipitation, two gauges in the Connecticut Valley, 30 km apart, were selected. The upslope gauge is on the western side and the downslope gauge is on the eastern side of the valley. The gauges have 0.01 inch resolution and data were available at both sites for eight of the twelve storms. Over the eight storms, the upslope station received 205 mm of precipitation in 125 h while the downslope station received 113 mm in 94 h. The average upslope precipitation rate is 1.6 mm h^{-1} while the average downslope rate is 1.2 mm h^{-1} (total precipitation divided by the number of hours of precipitation).

Fig. 8 shows the hourly distribution of hourly precipitation amounts for both stations. The hourly precipitation amounts are most often 1–2 mm h^{-1} and hourly precipitation amounts greater than 4 mm h^{-1} were seldom received at either gauge. This illustrates that intense rainfall associated with convection is not the cause of the anomalies. The upslope hourly amounts are shifted toward higher precipitation rates by 0.5 mm h^{-1}. In other words, if one added 0.5 mm h^{-1} to the downslope hourly amounts, the result would be similar to the observed upslope distribution. This is what one would expect if the large-scale precipitation were scavenging the moisture in orographically enhanced clouds. Of course part of the difference could be due to the low-level evaporation of the downslope precipitation.

The gauge comparison illustrates that orographic enhancement is small (~0.5 mm h^{-1}), but that the

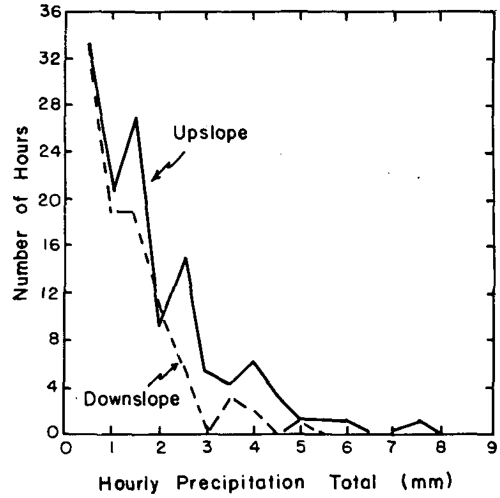

FIG. 8. Precipitation rate spectra for two proximal upslope and downslope stations for eight storm events. See text for details.

net effect can be quite substantial since the precipitation rates for these storms are only 1–2 mm h^{-1}. In the remainder of this section we will address two questions via simple models. First, can the lifting over the terrain supply 0.5 mm h^{-1} of additional moisture, and second, is the efficiency of precipitation scavenging adequate to convert the orographic cloud to precipitation in the available time?

Previous modeling studies show that precipitation scavenging of low-level orographic clouds can easily account for a 0.5 mm h^{-1} enhancement of the large-scale precipitation (e.g., Storebø, 1975; Bader and Roach, 1977). These studies employed numerical models of condensation and accretion and an assumed airflow over a hilly terrain. Here, we shall adopt a more simple analytical approach to answer the two questions that were posed above.

Assume that saturated air is lifted moist adiabatically at a characteristic orographic vertical velocity w_0 and that this lifting occurs over a depth H. The magnitude of the orographic enhancement of the precipitation rate ΔR_0 can be estimated by assuming that all condensate is converted to precipitation, i.e.,

$$\Delta R_0 \sim w_0 a H, \qquad (1)$$

where $a \equiv \overline{d\rho_{vs}/dz}$ is the moist adiabatic change in the saturation vapor density ρ_{vs} with height, averaged over the layer H. An upper-bound estimate of the characteristic vertical velocity can be obtained from

$$w_0 \approx US, \qquad (2)$$

where U is the windspeed normal to the barrier and S the terrain slope. For U = 10 m s^{-1} and S = 1%, w_0 = 10 cm s^{-1}. Assuming that the depth of orographic lifting H is 500 m (the nominal terrain height) and that near 0°C $a \sim 2$ g m^{-3} km^{-1}, it yields R_0

Unauthenticated | Downloaded 10/03/23 11:28 PM UTC

~ 0.4 mm h^{-1}. This is of the right order to explain the observed orographic enhancement rates, but it implies that the orographic precipitation process is either very efficient or that lifting occurs over a greater depth.

A simple model of precipitation falling through a moist-adiabatically rising parcel can be used to examine the parameters that govern the efficiency of orographic precipitation scavenging. Consider a simple model of vapor and cloud water in a rising parcel. We shall assume that a monodisperse precipitation falls through the parcel at a precipitation rate

$$R = \frac{\pi}{6} D^3 \rho V_T N, \qquad (3)$$

where we have assumed spherically symmetric particles of diameter D, fallspeed V_T, constant bulk density ρ and concentration N. These particles will remove cloud droplets by accretion at a rate

$$\frac{\pi}{4} D^2 V_T N X_c, \qquad (4)$$

where X_c is the cloud water content, and we have neglected the diameter and fallspeed of the cloud droplets and assumed perfect collection. For a shallow cloud we assume that cloud water in the rising parcel is generated at a rate wa, where w is the updraft speed and a is the average height derivative of the saturation vapor density for moist adiabatic ascent. The governing equation for the cloud water content of the parcel is

$$\frac{dX_c}{dt} = \frac{wdX_c}{dz} = wa - \frac{\pi}{4} D^2 V_T N X_c, \qquad (5)$$

where $z = wt$ is the height of the parcel. For $X_c(0) = 0$, one can show that

$$X_c(z) = \frac{2wa\rho D}{3R}\left[1 - \exp\left(\frac{-3Rz}{2\rho Dw}\right)\right]. \qquad (6)$$

The model describes the behavior of the cloud water content as a parcel travels up a mountain barrier. The water content increases to a limiting value, i.e.,

$$\lim_{z\to\infty} X_c \equiv X_{c,\infty} = \frac{2wa\rho D}{3R}, \qquad (7)$$

since the scavenging rate is proportional to the cloud-water content. Initially, X_c increases until a balance between generation and depletion is achieved. A small value of $X_{c,\infty}$ implies efficient removal of precipitation. If the generation rate wa is large, then $X_{c,\infty}$ is large. Thus a very steep slope that leads to large w is not as conducive to efficient orographic enhancement as a more gentle slope that rises to the same height. For a given precipitation rate, small, low-density particles (such as snowflakes) favor efficient scavenging (small $X_{c,\infty}$). When the precipitation rate R is larger, there are more precipitation particles in this monodisperse model, hence more efficient scavenging.

The e^{-1} height for the approach to equilibrium is

$$h \equiv \frac{2\rho Dw}{3R}. \qquad (8)$$

A small value of h means a rapid approach to $X_{c,\infty}$ and implies efficient scavenging. The dependence of h on the various physical parameters is similar to the dependence of X_c. Efficient orographic precipitation scavenging is favored by a gentle incline, large precipitation rate and small, low-density precipitation particles. The dependence of the scavenging efficiency on the various physical parameters as implied by (7) and (8) is in accord with the Bader and Roach (1977) study that employed a more detailed model.

To evaluate the model we select values consistent with light snow falling through an orographic cloud at 0°C:

$$w = 10 \text{ cm s}^{-1}$$
$$a = 2 \text{ g m}^{-3} \text{ km}^{-1}$$
$$\rho = 0.1 \text{ g cm}^{-3}$$
$$R = 1 \text{ mm h}^{-1}$$
$$D = 0.1 \text{ cm}.$$

The e-folding height is $h = 25$ m and the asymptotic value is $X_{c,\infty} = 0.05$ g m^{-3} of cloud water. For comparison, the cloud water content after 0.5 km of lift with no depletion would be 1 g m^{-3}. If instead of snow, raindrops were doing the scavenging then (melting the snow in the previous example) $\rho = 1.0$ and $D \sim 0.05$ (cgs) and we obtain an e-folding height of 125 m and an equilibrium water content of 0.25 g m^{-3}. Hence, snow can scavenge the available cloud water much more efficiently than rain, but in either case the efficiency is high.

The model illustrates that precipitation scavenging can be very efficient for gently sloping terrain. For the mountain barriers in New England one would expect near complete scavenging of orographic cloud. Hence the $\sim$0.5 mm h^{-1} that is required to account for the observed orographic enhancement can be supplied by scavenging in shallow clouds. For comparison, Bader and Roach (1977) found that a 1.5 km deep cloud formed over a 400 m hill can increase the precipitation rate by 1 or 2 mm h^{-1}.

Before concluding, it is interesting to discuss the coastal plain precipitation. The greater precipitation over the coastal plain (as compared to shoreline and island locations) has several possible explanations. It could be an artifact due to the typically greater windspeeds at the coast which may lead to a less efficient raingauge catch. Ideally this should not be a problem since raingauges are supposed to be sited in sheltered areas whereas wind measurements are supposed to

Unauthenticated | Downloaded 10/03/23 11:28 PM UTC

be made in exposed areas. It is also possible that precipitation over the coastal plain is being enhanced by lifting generated by frictional convergence (e.g., Bergeron, 1949). The winds at the shoreline are stronger than winds at inland stations so we know that frictional convergence must occur. The magnitude of the vertical velocity generated by frictional convergence can be estimated from

$$w \approx \frac{H}{L} \Delta u,$$

where H is the depth of the atmosphere over which the windspeed is reduced by Δu and L is the horizontal length scale. For $\Delta u \sim 1$ m s^{-1}, $L = 10$ km and $H = 1$ km, $w \approx 10$ cm s^{-1} at the top of the layer H. The slope of the coastal plain is $\sim$0.2% so that for a speed of 10 m s^{-1} the upslope vertical velocity is 2 cm s^{-1}. Thus, the combination of weak upslope motion and frictional convergence could produce vertical air motions that would account for the excess precipitation on the coastal plain.

6. Conclusions

The small terrain features in southern New England play an important role in shaping the distribution of pre-warmfrontal precipitation. Upslope regions receive 20 to 60% more precipitation than adjacent downslope or shoreline locations. The enhancement rates are only $\sim$0.5 mm h^{-1}, but the large scale precipitation rates are only 1–2 mm h^{-1}. Thus over the course of a storm, the weak orographic enhancement leads to significant precipitation anomalies.

Of the possible mechanisms discussed here, the low-level scavenging of orographically-forced cloud by the large-scale precipitation best explains the observations. The limited scope of the data do not permit one to say that this is the sole mechanism for enhancement. However, it is certain that orographically forced clouds will form in saturated air that is lifted over mountain barriers and that precipitation from aloft will fall through these clouds and scavenge the available moisture (either by riming and/or direct deposition for snow, or by accretion for rain). The modeling illustrates that since precipitation scavenging is a rather efficient process, even thin orographic clouds ($\sim$500 m) can account for the observed anomalies. Hence the case for low-level scavenging is a strong one, but a more comprehensive set of observations would be required to verify that this is the dominant mechanism.

These analyses reiterate the importance of natural seeding in the release of precipitation from shallow, persistant cloud systems and lead one to speculate whether these storms would be suitable for artificial seeding to enhance precipitation. However, upslope precipitation falls 70–80% of the time implying abundant natural seeding so that little would be gained by artificial seeding. A more suitable period for artificial seeding could be during northwesterly flow after the passage of a cyclone, when non-precipitating stratocumulus clouds are fairly common.

Acknowledgments. The authors thank James Franklin for assistance with computer programming and Liz Manzi and Isabelle Kole for the preparation of this manuscript. This research was sponsored by the National Science Foundation under Grants NSF-G 77-27634-ATM and ATM-8013711.

REFERENCES

Bader, M. J., and W. T. Roach, 1977: Orographic rainfall in warm sectors of depressions. *Quart. J. Roy. Meteor. Soc.*, **103**, 269–280.

Bergeron, T., 1949: The problem of artificial control of rainfall on the globe. II: The coastal orographic maxima of precipitation in autumn and winter. *Tellus*, **1**, 15–32.

——, 1965: On the low-level redistribution of atmospheric water caused by orography. *Suppl. Proc. Int. Conf. on Cloud Physics*, Tokyo and Sapporo, 96–100.

Browning, K. A., and T. W. Harrold, 1969: Air motion and precipitation growth in a wave depression. *Quart. J. Roy. Meteor. Soc.*, **100**, 309–330.

——, F. F. Hill and C. W. Pardoe, 1974: Structure and mechanism of precipitation and the effect of orography in a wintertime warm sector. *Quart. J. Roy. Meteor. Soc.*, **100**, 309–330.

Douglas, C. K. M., and J. Glasspoole, 1947: Heavy orographic rainfall in the British Isles. *Quart. J. Roy. Meteor. Soc.*, **73**, 11–42.

Elliot, R. D., and R. W. Shaffer, 1962: The development of quantitative relationships between orographic precipitation and air mass parameter for use in forecasting and cloud seeding evaluation. *J. Appl. Meteor.*, **1**, 218–228.

Sawyer, J. S., 1952: A study of the rainfall of two synoptic situations. *Quart. J. Roy. Meteor. Soc.*, **78**, 231–246.

——, 1956: The physical and dynamical problems of orographic rain. *Weather*, **11**, 375–381.

Smith, R. B., 1981: A differential advective model of orographic rain. *Mon. Wea. Rev.*, **110**, 306–309.

Storebø, P. B., 1976: Small scale topographic influences on precipitation. *Tellus*, **28**, 45–59.

Wilson, J. W., and M. A. Atwater, 1972: Storm rainfall variability over Connecticut. *J. Geophys. Res.*, **77**, 3950–3956.

Unauthenticated | Downloaded 10/03/23 11:28 PM UTC