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OF OROGRAPHICALLY ENHANCED RAIN

IN BRITAIN: A REVIEW

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ABSTRACT

The orographic enhancement of rain over hills of small or medium size can be very large when there is a strong and almost saturated low-level flow. Much of the orographic growth occurs in the lowest one or two kilometres and the rainfall maxima are therefore tied closely to the hills. Orographic rain is difficult to measure by means of raingauge networks because the horizontal gradients of rainfall tend to be large; orographic rain is also difficult to measure accurately by means of radar because of the tendency of the radar beam to be above the shallow zone of heavy rain.

The observational evidence reviewed in this paper supports the widespread applicability of a conceptual model by Bergeron, according to which small droplets in a low-level orographic "feeder cloud" located mainly below the $0^{\circ}C$ level are washed out by raindrops falling from a pre-existing "seeder cloud". A numerical treatment of the washout process by Bader and Roach provides a sound basis for deriving dynamically simple models for predicting the distribution of orographic rain. Such models can be driven by the output from large-scale dynamical models capable of predicting the background conditions in the absence of detailed orography. A test of one such combination of models shows good skill in predicting 24-h rainfall accumulations over hills. In order to predict the detailed time variation in orographic rainfall for the period 0 to 6 h ahead it is necessary to supplement the numerical models by radar observations of precipitation patterns upwind of the forecast area.

CONTENTS

1 INTRODUCTION

The underlying terrain influences the distribution of rain through forced uplift and also through thermal and frictional effects. The thermal effects are manifested as the triggering of convective showers by elevated heat sources and by organised mesoscale circulations. Frictional effects lead to enhanced rain through local boundary layer convergence brought about by differences in friction such as at coastal boundaries. When we speak of orographic (or oreigenic) rain, however, we generally have in mind rain that is due to the forced lifting of a moist airflow over a range of hills. Although more than one of the above effects sometimes occur together, the intention in this paper is to concentrate on those situations where the dominant factor causing enhancement of rain is indeed forced uplift due to hills. Orographic rain defined in this way is of great practical importance since in many countries it accounts for a large proportion of the water supply. In Britain orographic rain does not have the high intensity that is characteristic of showers and thunderstorms but by falling fairly steadily at rates of typically 2 to 8 mm h^{-1} for periods of many hours it nevertheless often gives rise to large daily totals.

Our purpose in this article is to review existing knowledge and to present some new results concerning the effects of hills of modest dimensions such as those encountered in the British Isles. The paper is geographically biased toward Wales and southwest England because this is where many of the observational studies have been made; however, the results are thought to apply to the hilly western areas throughout the British Isles. The paper deals with situations in which the low-level flow is warmer than $0^{\circ}C$ and in which rain rather than snow falls in the lowest kilometre or two. We shall be concerned with the structure, mechanism and prediction of orographic precipitation, but not with its augmentation by artificial means. Indeed, the type of orographic cloud considered in this paper - with a substantial amount of lower cloud warmer than $0^{\circ}C$ and natural seeding occurring from upper cloud layers - is not, according to Sax et al (1975), the kind of cloud for which there is a well-tested cloud seeding rationale.

Orographic rain is not an easy subject to study because of the complex form of most hilly regions and the difficulties of accurately measuring the rain in these areas. Progress has been understandably slow. However, as we shall show, recent observational and theoretical studies have brought us to the stage where a fairly clear picture of the mechanism of orographic rain is emerging that is very much in line with a conceptual model put forward by the late Tor Bergeron.

2 STRUCTURE OF OROGRAPHIC RAIN

a Plan distribution

There is often a close correspondence between the recorded mean annual rainfall and the height of the terrain around the raingauge (Bleasdale and Chan 1972). The increase in rainfall with height is particularly pronounced in parts of the British Isles that are directly exposed to airflows from the sea when the depth of the moist layer is far greater than the height of the hills. Many hilly parts of Wales, for instance, receive 3 or 4 times the annual rainfall of 700 mm which characterises large parts of central England. However, the average figures represent the integrated effect of rainfall generated by different mechanisms with different wind directions and synoptic situations, and what is sometimes not realised is just how large the orographic effect can be in individual cases. An example of a heavy orographic fall is given in Fig 1, which shows the distribution of rain in relation to topography in south Wales and southwest England during a period of prolonged rain ahead of a quasi-stationary cold front. Fig 1(a) was derived from daily raingauge totals on two successive rainfall days but the orographic rain was confined to an approximately 24-h period (see Fig 1(c)) during which the synoptic situation was almost constant with a continuous flow of moist warm air from the SSW. Figs 1(a) and 1(b) show that the total rain over parts of the south Wales hills about 500 m high exceeded 100 mm. Although Fig 1(a) shows that unsheltered coastal regions 40 km upwind had totals of typically 20 mm, by making allowances for showers which occurred after 18 GMT on 30 January, it can be seen that the coastal sites actually received only about 10 mm during the main period of orographic rain (see the autographic record for station P in Fig 1(c)). Other coastal locations such as those in SE Wales which were in the lee of the hills of SW England had far less rain even than this.

Data from several other cases of sustained orographic rain in south Wales are presented in a simplified form in Fig 2. This figure shows time-integrated rainfall profiles across the hills along the direction of travel of the rain areas. The cases have been selected because the direction of travel of the rain areas was approximately constant (at 250°) throughout the period of rainfall in each case. All but one of the situations were characterised by strong low-level winds. For these cases the peak rainfall did not occur in the region of maximum upslope motion where the rate of condensation would have been greatest; rather, the cloud water was advected up the hill to give the maximum rainfall marginally to the lee of the crest. The only case where the maximum rainfall occurred upwind of the hill crest (29 July 1967) was an occasion when the low-level winds were weak and there was heavy background rainfall (cf Sec 4(c)). Another feature of the data depicted in Fig 2 is that most of the cases were characterised by near-neutral stability with occasional weak potential instability especially above 800 mb. The 16 October 1967 case was more unstable than usual (Browning and Harrold 1969) and this was associated with the carry-over of rather more rain to the lee than on most other occasions, presumably because of the orographic triggering or intensification of convection (Browning et al 1974).

Orographic effects can also be detected for hills far smaller than those discussed above. Bergeron (1960), in Project Pluvius, operated a very dense network of raingauges in a lowland area surrounding Uppsala, in Sweden. He found that the fine scale distribution of daily rainfall was closely related to the distribution of hills only a few tens of metres high. The circumstances of his studies were not unlike those often encountered in Britain. Fig 3, from Bergeron (1967), shows that orographic effects were observable even in the monthly totals resulting from several occasions of widespread non-convective rain. Note in particular that Lunsen Hill (L), although only 50 m above msl, received 26 mm, twice as much as Lake Ekoln (E) at sea level only a few kilometres away. Of course, on occasions with travelling convective showers, effects such as this tend to be obliterated. Monthly totals for 16 other months at a time of year when convection is negligible confirm that the Lunsen hill rainfall anomaly is a regular occurrence (Table 1). Similar results have been reported by Wilson and Atwater (1972) for stratiform rain in Connecticut, USA. They show that variations in rainfall up to a factor of two frequently occur within a few tens of kilometres over terrain varying in height by only about 150 m. They, too, find that the pattern of orographic rainfall is highly repetitive from storm to storm.

b Vertical structure

Bergeron (1965) argued that the close correspondence between the rainfall maxima and the terrain height implies that most of the orographic rain originates at very low levels - otherwise the wind would carry it farther away from the hill crest. For the modest orographic effects near Uppsala he inferred that the enhancement occurred mainly below 500 m. Actual measurements of the 3-D pattern of orographic precipitation intensity have been obtained recently by the Meteorological Office Radar Research Laboratory using a quantitative weather radar scanning at a sequence of elevations. The radar was sited on the south coast of Wales so that it was possible to observe the change in the vertical structure of the precipitation as it travelled from the sea to the hills. Data obtained during the passage of frontal systems confirm Bergeron's contention that much of the orographic precipitation originates at low levels; however, with the hills in south Wales rising to 600 m, the enhancement is not restricted to quite such a low level as inferred by Bergeron. In the situations analysed so far, which are all cases of deep mainly stratiform precipitation, the bulk of the enhancement takes place in the lowest 1 or 2 km. An example is shown in Fig 4. It depicts the mean pattern of rainfall over a 5-h period within a vertical section oriented along the direction of travel of individual features of the precipitation pattern. The data were contaminated by a radar 'bright band' due to melting snow at an altitude of about 2 km. Although the intensity of the bright band was unusually weak in this case, it nevertheless causes the precipitation intensity near 2 km to be overestimated in Fig 4. Thus the rate of decrease of rainfall intensity with height between altitudes of 1 and 2 km would have been even greater than indicated in Fig 4.

The fact that most of the orographic enhancement above hills a few hundred metres high occurs in the lowest 1 or 2 km does not imply that the enhancement at higher levels is negligible. Small but significant effects are indeed detectable by radar in the middle troposphere. As an example, Fig 5 shows the variation in the height of radar-detectable precipitation over a 2-h period along a path from north Devon to south Wales which is oriented parallel to the wind direction of 200°. These data are for the same occasion which produced the heavy falls of rain depicted in Fig 1. The top of the detectable precipitation was mainly between 2 and 4 km over the hills of north Devon; it dropped to 0 to 3 km over the sea, and then rose to 4 to 5 km over the hills of south Wales.

c Implications for measuring orographic rain

The measurement of rain is not a topic we wish to discuss in detail in this review; however, it is appropriate to mention here that the great spatial variability of orographic rain described above makes it difficult to obtain accurate and spatially representative measurements. Chang (1977) suggests that for a mountainous region such as the Appalachians in West Virginia a gauge spacing of about 10 km is required for an accurate description of even the annual rainfall. The problem of measuring short duration rainfall patterns is far greater. Fig 1, for example, shows horizontal changes in the effectively 24-h rainfall totals amounting to 50 mm within 10 km on the flanks of hills. It is impossible to map such fields with the existing distribution of autographic raingauges whose spacing in remote hilly areas in Britain is typically 10 to 30 km. The average spacing of 3 to 10 km for 24-h gauges is more nearly adequate but it is only occasionally that a homogeneous rainfall type dominates the daily rainfall totals. In general, therefore, the true magnitude of the horizontal gradients in orographic rainfall intensity that occur in the very short term tends to be blurred in the daily totals.

Radar is a tool that is increasingly coming to be viewed as a means of obtaining representative estimates of areal rainfall, especially in situations where the data are required in real time for forecasting and flood warning purposes. A project has recently been completed (Dee Project 1977) in which it was demonstrated that, even though radar is difficult to use in hilly areas (because of echoes from the ground etc), the accuracy of radar in such areas is nevertheless adequate for most hydrological purposes. Radar was also shown to be more cost effective than a network of telemetering raingauges provided the area over which the measurements are required exceeds 3000 km^2. However, it is important to realise that the high accuracies achieved in this project were due in part to the measurement of rainfall being restricted to within 50 km of the radar. The presence of strong vertical gradients of orographic rainfall intensity, as exemplified in Fig 4, implies that at much larger ranges the radar may significantly underestimate the surface rainfall intensity in hilly regions. Consider, for example, a radar with a narrow 1° beam centred at $\frac{1}{2}^{\circ}$ elevation. Assume an optimum radar site in which the volume above a region of hills at a range of, say, 75 km can be observed with a horizon of 0° without the hills being so obtrusive as to produce extensive ground echoes. At 75 km the beam will extend from 400 to 1600 m above the height of the radar, over which interval the rainfall intensity may decrease by a factor of 2. In this case the mean reflectivity measured by the radar would underestimate the surface rainfall by 25%. At longer ranges or with a broader beam the errors would increase further. These errors are of course additional to those due to interception of the beam with the melting level which is increasingly likely to occur at the longer ranges.

3 FACTORS INFLUENCING OROGRAPHIC ENHANCEMENT OF RAIN

The heaviest orographically enhanced rainfall in the hilly western parts of Britain occurs in association with frontal systems travelling **from a generally westerly direction.** The principal factors influencing the orographic rain are the form of the airflow induced by the hills, the magnitude of the relative humidity, the strength of the wind, the wet-bulb temperature, convective instability, and the presence and nature of pre-existing precipitation. These factors will now be considered in turn from an observational point of view:

a The pattern of vertical air motion

The height and shape of a hill obviously influences the magnitude and location of the induced vertical air motion, the rate of condensation and the resulting orographic enhancement of the rain. Unfortunately not much is known about the detailed form of the airflow over natural hills of complex shape. Broadly speaking the magnitude of the vertical motion usually tends to decrease with height through the troposphere.

Sawyer (1956) has discussed the importance of the shape of the vertical profile of the wind and the static stability for determining the vertical extent of significant upward motion. However, since most orographic rain is generated in the lowest 2 km, the extent to which the upward motion penetrates into the middle troposphere is not a crucial issue, and for many purposes it may be sufficient to assume that the airflow follows the topography in the all-important lowest layer. Actually, for an isolated steep-sided hill of small horizontal extent, the air will tend to flow around it especially if the air is statically stable. But, for a smooth-shaped hill of considerable extent at right angles to the flow, the assumption that the broad-scale flow follows the shape of the hill without change of direction is a useful approximation.

In the case of a mass of hills deeply dissected by valleys, Pedgley (1970) points out that the main airflow does not follow every detail of the local topography. Where a valley faces into the wind we would expect to find a funnelling effect, with strong winds at the head of the valley. But where a valley lies across the wind, it will be sheltered, the broad-scale flow going from ridge to ridge, leaving an eddy within the valley. With a pattern of linked valleys having various orientations, as in Snowdonia (Fig 6a), the air is able to flow from one valley to the next and its direction may show little obvious relation to the direction of the broad-scale flow aloft. To a considerable extent, the valleys are filled with rather stagnant air that may be taken as part of an overall 'mountain dome'. As a result, gauge locations in the centre of the smooth envelope surrounding the mass of hills will receive orographic falls characteristic of the higher parts of the hills even though the gauges themselves may be situated deep in the valleys. Examples of such sites in Britain are Cwm Dyli at 94 m in the heart of Snowdonia (Site 11 in Fig 6) and to a lesser extent Treherbert at

183 m in the midst of the Blaenau Morgannwg (Nash and Browning 1977). Fig 6(b), from Pedgley's paper, shows the profile of rainfall across Snowdonia in a warm sector with a moist WSW'ly airflow. Most of the raingauge sites whose data are plotted here were in valleys; nevertheless the rainfall maximum is seen to be related fairly well to the pattern of smoothed topography as shown by the full curve in Fig 6(c).

b Relative humidity at low levels

Since most of the orographically enhanced rain originates in the lowest one or two kilometres, the magnitude of the relative humidity at these levels is most important. Bergeron (1965) reported that in situations with notable orographic effects the low-level air is almost saturated and even small hills are liable to be shrouded in scud or pannus cloud. Such conditions are most likely when the air is being moistened by falling rain and more especially when there is low-level convergence. This convergence might be in the form of boundary layer convergence in frontal regions with cyclonic shear (Browning et al 1975) or at the leading edge of low-level jets (Nash and Browning 1977). The effect of such factors is, however, readily negated by local subsidence in the lee of hills and so, as shown by Fig 7, there is a strong relationship between large orographic falls in south Wales and the occurrence of southwest winds incident from the open sea. The existence of a saturated low-level flow when the winds in south Wales are from the southwest is also favoured by the decrease of sea surface temperature with distance downwind along the trajectory of the approaching air.

c Strength of the wind at low levels

For a saturated flow crossing a given orographic barrier the rate of condensation at the low levels from which most of the orographic rain originates can be expected to be proportional to strength of the low-level winds. The importance of strong moist flows for producing heavy orographic rain in western Britain was noted as long ago as 1947 by Douglas and Glasspoole. The same point has been stressed by Saha and Bavadekar (1977) in regard to the southwest monsoons in the Western Ghats of India. An investigation of some of the heaviest 24-h falls of rain over the hills of south Wales has shown that 18 out of 20 cases were associated not only with strong winds (usually stronger than 30 ms^{-1} according to Table 2) but also with definite low-level jets with a maximum wind speed at around 900 m (Nash and Browning 1977). The low-level jets occurred in the warm sectors of depressions, often just ahead of the cold front, and they extended some way ahead of the position of the surface warm front. Low-level jets such as these have been studied observationally by Browning and Pardoe (1973); they are associated with what Harrold (1973) has referred to as conveyor belts. A notable example of a strong low-level jet associated with a conveyor belt is shown in Fig 8. The low-level jet system depicted in Fig 8 remained stationary for 24 hours and was responsible for the very heavy orographic fall of rain portrayed earlier in Fig 1.

d Wet-bulb potential temperature

Since the rate of condensation during orographic uplift increases with wet-bulb potential temperature (θ_w) it might have been expected that the value of θ_w would be a major factor in determining the intensity of orographic rain. However, when the cases in Table 2 were categorised according to the mean θ_w in the lowest 2 km of the warm-sector conveyor belt, it was found that the majority of cases (12) occurred between November and March when θ_w was in the range from 8 to 12°C, whereas only 6 cases occurred between June and October when θ_w was in the range 14 to 18°C. Thus, although the rate of condensation during orographic uplift increases with θ_w, this does not appear to be a dominant factor in determining the magnitude of orographic effects. On the contrary, Table 2 shows that the seasonal distribution of heavy 24-h rainfalls in south Wales reached a maximum in December in association with the situations of strongest low-level windspeeds when θ_w was relatively low (although still high for the time of year). This is in line with Bleasdale and Chan's (1972) finding that the rainfall-terrain height relationship for the British Isles as a whole is intensified in the months November-February at the time of the highest frequency of fast-moving westerly rain systems.

e Convective instability

The importance of convective instability for orographic rain in conveyor belt situations is not entirely clear. Certainly deep convection from the surface is not a feature of most situations of heavy orographic rainfall. In many cases there is a highly stable warm frontal layer in the lower troposphere. On the other hand, it is not uncommon in frontal regions for there to be a layer of shallow convection aloft (eg Browning et al 1969, 1973, 1974; Kreitzberg and Brown 1970; Atkinson and Smithson 1972, 1974, 1978; Harrold 1973; Hobbs et al 1975, 1978). This is usually revealed in radiosonde ascents as being associated with weak potential instability; often the lapse rate aloft is found to be close to saturated adiabatic. However, to some extent the failure to detect marked instability is due to the poor response of the humidity element in radiosondes and also to the fact that the instability is in the process of being released by large scale or mesoscale ascent. There is evidence (eg Elliott and Shaffer 1962, Elliott and Hovind 1964) that the instability aloft tends to be more pronounced on occasions of heavy orographic rain and such instability can lead to the area of heavy rain extending farther to the lee of the hills (cf Sec 2a). What is not yet clear is whether the convection aloft plays a major causative role in enhancing the rain over the highest part of the hills. Most of the orographic rain is generated in the low-level cloud and so if the middle-level convection does play an important role it would be as a result of more effective seeding of the low-level cloud.

f Pre-existing rain

Rain may form either through the initial growth of ice crystals which subsequently melt during fall or through the growth of large droplets in a water cloud by coalescence with smaller droplets. Ludlam (1956) gave the time taken for precipitation to develop by the ice-crystal process as roughly half an hour and by the coalescence process as one hour. As Sawyer (1956) pointed out, these periods are long compared with the time taken for a strong airflow to cross a hill of small or moderate dimensions. Thus it is difficult to see how orographic rain in the British Isles could simply be the result of cloud particles growing from scratch. However, as stressed by Pedgley (1970), on most occasions of orographic rain in Britain there is already extensive layer cloud producing some precipitation before the airstream reaches the hills. Apparently, therefore, the effect of the hills is to intensify existing rain areas rather than to produce major new areas of rain.

This view is supported by recent radar observations of areas of rain travelling from the sea over the hills of south Wales. Fig 9, for example, shows time records of the intensity of the same group of warm frontal rain areas crossing, first a line SS over the sea and, then, a line HH over the hills 60 km downwind. There is a close correspondence in area of regions of heavy rain over the hills (Fig 9b) with regions of moderate rain approaching from the sea (Fig 9a). Indeed, it can be seen that on this occasion there was seldom any heavy rain over the hills when the pre-existing rainfall rate over the sea was less than $\frac{1}{4}$ mm h^{-1}.

An apparent exception to this rule is the so-called anomalous orographic rain on the Hawaiian Island of Oahu, reported by Woodcock (1975), for which there was negligible rain observed at the surface upwind of the hills. Although

Woodcock considered that this was due to exceptionally rapid growth of droplets, we suspect that it represents a marginal kind of situation in which the clouds upwind of the hills contain drizzle-sized droplets which evaporate before reaching the ground. Of course, when larger raindrops already exist, the evaporation is only partial; for example, in the low-lying areas around Uppsala during widespread frontal rain, Andersson (1964) suggests there is commonly a decrease of about 10% in the precipitation amounts owing to evaporation in the 500 m layer below cloud base.

Another kind of situation in which orographic rain may occur in the absence of pre-existing surface rain is when convection is triggered aloft just upwind of the hills. In this circumstance the seeding particles may begin reaching the ground only in the immediate vicinity of the hills. Browning et al (1974) suggest that middle-level potential instability can be triggered as far as 50 to 100 km upwind of the hills. Ascent beginning far upwind of the hills has been predicted by Eliassen and Rekustad (1971) in a theoretical study of mesoscale mountain waves. The main effect of this mechanism, however, would appear to be to increase the extent of areas where middle-level convection is already active rather than to generate entirely new rain areas.

For precipitation to be formed independently of the hills there needs to be some dynamically induced large-scale or meso-scale ascent. Baroclinic ascent is preferable to buoyant convective ascent from the surface because the rapid vertical mixing implied by the latter often tends to rob the lower layers of its all-important high relative humidity. In general it is the baroclinic ascent of a deep moist layer that is most likely to produce the required pre-existing rain. Our radar observations in south Wales (eg Fig 4) show that the pre-existing precipitation is usually detectable up to heights of 4 to 6 km. This is consistent with Holgate's (1973) finding that, although most of the orographic enhancement occurs at low levels, heavy orographic rain nevertheless is usually accompanied by a moist layer of considerable depth. This is not to say that orographic rain never occurs with a fairly shallow moist layer: Browning et al (1975), for example, have observed heavy orographic rain when there was a very dry capping layer at 3 km, whereas on Oahu moderate orographic rain has been observed with the top of the moist layer as low as 2.1 km (Woodcock 1975).

Fig 9 suggested that on a given occasion there may be a good correlation between the pre-existing, or background, rainfall intensity and the orographically enhanced rainfall intensity. Nash and Browning (1977), however, examined many cases in south Wales and found that the overall correlation was not as good as they had hoped. One factor likely to have obscured the expected relationship would have been variations in the depth and relative humidity of the moist low-level flow from one case to another. Another factor which probably obscured the relationship is the small-scale variability in the rainfall intensity, as exemplified in Fig 9. This makes it difficult to make valid comparisons between individual coastal and hill raingauges as was attempted in the study by Nash and Browning. Detailed studies of other sets of data such as those in Fig 9, now in progress, tend to support the view that for a given situation there may be a reasonable correlation between orographic rain and background rainfall intensity. Such a correlation is of course necessary if we are to attempt to make short-term predictions of orographic rain using rainfall information obtained at places upwind of the hills (cf Sec 5).

4 THE MECHANISM OF OROGRAPHIC RAIN

a The washout model

The observational evidence summarised in Sec 3 is consistent with the simple model of orographic rain proposed by Bergeron (1950, 1965) as portrayed in Fig 10. The important components of this model are:

(1) A so-called feeder cloud, capping the hill, associated with

forced uplift of moist low-level air over the hill. Most of this cloud is composed of very small cloud droplets warmer than $0^{o}C$. Because of the short residence time of the droplets in the feeder cloud they do not have enough time to grow into raindrops and fall out in their own right.

(2) A so-called seeder cloud existing independently of the hills and producing precipitation which serves the dual role of:

(i) further moistening the low-level air upwind of the hills by evaporating below the base of the seeder cloud, and

(ii) washing out the cloud droplets in the feeder cloud over the hills.

The seeder cloud may not be entirely distinct from the feeder cloud as shown in Fig 10. Indeed, even if the seeder cloud were separated from the feeder cloud as in Fig 10, there would still tend to be some orographically stimulated increase in the cloud water content and some associated precipitation growth in the seeder cloud itself. Nevertheless, it is conceptually convenient to consider the seeder and feeder clouds as being separate entities.

The precipitation distribution can be described in terms of:

P_0, the intensity of the seeder precipitation entering the top of the feeder cloud,

P_1, the intensity of the surface rain upwind of the hills, and

P_2, the intensity of the surface rain in the orographic maximum near the hill crest.

The true orographic enhancement is given by $P_2 - P_0$. What is normally measured is the apparent orographic increment $P_2 - P_1$, which is the net effect of low-level evaporation (less often low-level growth) over the low ground and the generally much larger low-level growth due to washout of some of the cloud droplets in the feeder cloud.

b Quantitative aspects of the washout model

Hobbs et al (1973), in calculating the growth of precipitation in an orographic cloud, neglected the depletion of the cloud water content. Hobbs et al were, however, dealing with low precipitation intensities associated with slow-falling ice particles; in the case of raindrops falling in significant concentrations it is not justifiable to neglect this depletion. As pointed out by Sawyer (1956) and more recently Browning et al (1975), the orographic enhancement of rain is therefore dependent not only on how rapidly the hydrometeors from the seeder cloud can grow by sweeping out the cloud droplets in the feeder cloud but also on the rate at which the low-level flow can replenish the liquid water content of the feeder cloud through fresh condensation. This combination of factors is represented schematically in Fig 11. Following Bader and Roach (1977), the elements in this block diagram can be represented by the following equations:

The vertical air velocity of the low-level air, assuming it follows the smoothed shape of the hill, is given by

$$w = u \tan \alpha \qquad \text{..... (1)}$$

where u is the component of the wind normal to the range of hills and α is the hill slope in this direction.

The rate of condensation in the form of small cloud droplets within the feeder cloud, assuming an initially saturated airstream, is given by

$$C = -w \frac{\partial \rho_s}{\partial z} \qquad \text{..... (2)}$$

where $\frac{\partial \rho_s}{\partial z}$ is the variation of saturated vapour density with height and for a given θ_w can be specified approximately as a linear function of height.

The rate of accretion of the small cloud droplets by raindrops entering from above is given by

$$A = \pi \int_0^\infty N_r E_r V_r r^2 \, q \, dr \qquad \ldots\ldots (3)$$

where $N_r dr$ is the concentration of seeder raindrops of radius between r and r+dr, V_r is their terminal fallspeed, E_r is the collection efficiency of the raindrops with respect to the small cloud droplets, and q is the liquid water associated with the small cloud droplets.

The continuity equation for the liquid water content associated with the small cloud droplets within the feeder cloud may be written

$$\frac{Dq}{Dt} = C - A \qquad \ldots\ldots (4)$$

where $\frac{Dq}{Dt}$ is the rate of change of q following the air motion. It is assumed that gravitational settling and direct deposition of cloud droplets can be neglected, and that the whole system is in a two-dimensional steady state, so that $\frac{D}{Dt} = u\frac{\partial}{\partial x}$ where u is the air velocity along a trajectory running normal to the hill range at a constant height above the terrain. Thus the liquid water content at a given height above the terrain varies with distance x according to the equation

$$q = \frac{1}{u} \int_0^x (C - A) \, dx \qquad \ldots\ldots (5)$$

Bader and Roach carried out computations in an x-z matrix within the feeder cloud with vertical and horizontal grid lengths of 100 m and 2 km, respectively. The procedure adopted was to assume that the seeder raindrops at the top of the matrix were characterised by a Best drop size distribution (Best 1950). For a prescribed seeding rate P_0, this determined N_r as a function of r. The radius of the cloud droplets in the feeder cloud was taken to be 10 μm and E_r was obtained as a function of r in accordance with values given by Mason (1971). Assuming a saturated flow of depth d and taking prescribed values for θ_w and u(z), they were then able to compute q from eq (5) as a function of x along the top row of the x-z matrix. Their next step was to assume that the number flux of seeder drops remained constant during their descent through the feeder cloud, ie the raindrops were assumed neither to break up nor to coalesce with other raindrops. For simplicity they also assumed that the seeder drops fell vertically*. (See Footnote). They were thus able to compute the new radius of the seeder drops in each initial size category after they had descended to the next level of the matrix, using the equation

$$r(z) = r(z_t) - \frac{1}{4\rho}\int_z^{z_t} E_r \, q \, dz \qquad \ldots\ldots (6)$$

where ρ is the water density. They applied this procedure iteratively until they reached the bottom of the matrix, where they were able to derive the surface rainfall rate P as a function of x from the newly evolved drop size distribution, using the equation

$$P_x = \frac{4}{3}\pi\rho\int_0^{\infty} N_r V_r \, r^3 \, dr \qquad \ldots\ldots (7)$$

Examples of the kind of results derived by Bader and Roach are shown in Fig 12. Fig 12(a) shows the distribution of C, A and q within a vertical section through the feeder cloud formed over a 400 m high hill. Conditions are as specified in the figure legend. Note that, although the condensation rate (C) is a maximum in the upslope part of the hill, the rate of washout (A) in the presence of modest seeding rates is not large enough when there is a strong wind to prevent the water content (q) of the feeder cloud from building up to a maximum fairly near the crest of the hill. The result is that the rate of washout itself and hence the surface precipitation intensity both tend to reach a maximum quite near the hill crest (Fig 12(b)). The orographic enhancement is seen to become larger when the wind is stronger and when the seeding rate (P_O) increases. This is consistent with the calculations

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FOOTNOTE

* This is a useful first approximation if the seeder particles are fairly large raindrops. For example, a raindrop with a 5 ms^{-1} terminal fallspeed would be displaced horizontally only 6 km in falling through the lowest $1\frac{1}{2}$ km in the presence of a 20 ms^{-1} wind. However, in the case of smaller drops (or snowflakes) the displacement could be so great that the particles might be carried out of the lee side of the feeder cloud and suffer evaporation before reaching the ground.

of Storebø (1976) and is in line with the observational evidence presented earlier in this review. Other factors such as the composition, size and concentration of the condensation nuclii entering the feeder cloud have been shown by Storebø to have no significant influence on the precipitation pattern.

c Efficiency of washout

Following Sawyer (1956), Elliott and Hovind (1964), Browning et al (1975) and Bader and Roach (1977), the efficiency of the washout process can be defined as

$$E_w = \frac{P_T}{C_T} \qquad \text{..... (8)}$$

where P_T is the total mass of the orographic rainfall increment reaching the surface per unit time per unit width perpendicular to the airflow, integrated over the total distance from the foot of the hill (x = o) to the hill crest ($x = x_c$), ie

$$P_T = \rho \int_o^{x_c} (P_x - P_o)\, dx \qquad \text{..... (9)}$$

and C_T is the total mass of water condensed per unit time per unit width perpendicular to the airflow, integrated over the depth ($z_t - z_b$) of the feeder cloud and over the entire distance from the foot of the hill (x = o) to the hill crest ($x = x_c$), ie

$$C_T = \int_o^{x_c} \int_{z_b}^{z_t} C\, dz\, dx \qquad \text{..... (10)}$$

Bader and Roach have calculated the dependence of E_w and P_T on the wind speed u and the seeding rainfall rate P_o. They show (Fig 13) that a given orographic rainfall increment can be produced by either (i) a high P_o and a low value of u, in which case E_w is high, or (ii) a low P_o and a high u, in which case E_w is low. When the wind is strong, the resulting high rate of condensation gives rise to a large cloud water content; however, a given parcel of cloud crosses the hill so quickly that there is insufficient time for rain entering at a modest rate to sweep out more than a small fraction of it.

Fig 13 is useful for illustrating the principle of washout efficiency but it must be borne in mind that it is based on the assumption that the flow is initially saturated. Lack of saturation in the low-level flow will diminish the liquid water content of the feeder cloud and hence also P_T; it also will lead to some evaporation at low levels upwind of the hills so that the background rainfall intensity actually measured at the surface (P_1) will be unrepresentative of the true rate of seeding (P_o) at the top of the seeder cloud. Another limitation of Fig 13 is that it assumes a particular drop size distribution (see Sec 4b). Rainfall intensity is proportional to the mean droplet volume whereas rate of washout is proportional to mean droplet area. Thus, for a given rainfall intensity, smaller droplets would be more efficient than large ones at sweeping out the cloud water in the feeder cloud. On the other hand, as pointed out by Pedgley (1970), if the particles are too small (or in the form of snow), far from falling vertically as assumed by Bader and Roach, they may get carried almost horizontally out of the cloud before they can grow much.

5 FORECASTING OROGRAPHIC RAIN IN BRITAIN

a Synoptic climatology

Ever since the early studies of Douglas and Glasspoole (1947) it has been recognised that prolonged heavy rain over the hills of western Britain is most likely to occur within the warm sectors of depressions, especially when there is a strong moist flow from between west and south-southwest ahead of a slow-moving or quasi-stationary cold front. Heavy orographic rainfall can also be associated with warm fronts so long as there is a strong moist flow at low levels (eg Nash and

Browning 1977). Slight instability aloft often accompanies orographic enhancement of rain (Elliott and Shaffer 1962, Elliott and Hovind 1964). Thunderstorms occasionally produce large falls in the absence of a strong low-level flow but they tend to be rather localised and the orographic enhancement associated with them is quite small.

Typical rainfall sequences during the passage of a frontal system are shown in Fig 14. The figure shows that in the warm sector, for a distance of order 100 km ahead of the cold front, the hills are substantially wetter (full line) than nearby low-lying areas upwind of the hills (dashed line). Although rain on this occasion continued behind the cold front, there is seen to have been an abrupt end to orographic enhancement at the cold front. This was due partly to the decrease in wind speed at low levels and partly to the presence of drier air descending beneath the cold front.

Synoptic considerations such as these, when further qualified by the need for the moist layer to be fairly deep, form the basis for much of the present-day forecasting of heavy orographic rain in semi-quantitative terms (eg Holgate 1973). The accuracy achievable with these methods has been assessed by Nicholass (1975) using a dense network of rain-gauges in the hilly Upper Dee region of north Wales. He evaluated forecasts of rainfall accumulations for periods up to a day ahead in an area of about 250 km^2. With the forecast rainfall total over a 21-h period divided into just 3 broad categories (5-15, 15-30 and >30 mm), he found that the correct category was predicted about half the time using Holgate's rules. Nicholass also investigated the accuracy of some other synoptic forecasting procedures (Benwell 1967; Lowndes 1968, 1969) but found that these were on the whole not quite as successful as the procedures suggested by Holgate.

The detailed distribution of orographic rainfall is highly dependent on the surface wind direction and the synoptic type. Nicholass and Harrold (1975) have studied this dependence using a dense network of autographic raingauges in the hilly Upper Dee region. The total area of this region was 1000 km^2 and they divided it into 16 river subcatchments each with an area of typically 60 km^2. They classified the surface wind into 3 categories of speed and 4 of direction, and they considered 6 synoptic types (pre-warm-front, warm sector, post-cold-front, occlusion, cyclonic rain not associated with a well-defined frontal system, showers not associated with any of the other types). Periods of rainfall were defined during which these categories remained constant - each such rainfall event was typically several hours long. For each rainfall event they calculated the ratio between the rainfall in individual subcatchments (R_s) and the corresponding rainfall ($\overline{R}$) averaged over the entire 1000 km^2 area. Nicholass and Harrold built up a set of climatological statistics for this ratio using data acquired over a period of 2 years. They found that the mean value of $\left(R_s/\overline{R}\right)$ varied between about $\frac{1}{2}$ and 2 according to the subcatchment, the surface wind and the synoptic type. In situations of widespread rain the standard error of each ratio was found to be as little as 20 to 50% (and this remained the case even when the climatological ratios were applied over periods of only 1 h). The good predictability of $\left(R_s/\overline{R}\right)$ has important implications for forecasting rainfall amounts over small subcatchments as required for hydrological purposes. It suggests that once a set of mesoscale climatological statistics has been derived, then the accuracy of forecasts on this scale should depend primarily on the accuracy with which the rainfall amounts over the larger 1000 km^2 area can be forecast. However, there are situations when the application of such climatological statistics would break down; these would include occasions when stationary mesoscale rainbands are located over the subcatchments. (Harrold 1973; Browning and Bryant 1975).

Although Nicholass and Harrolds' mesoscale climatological study was for one particular hilly area, it is reasonable to expect that analogous climatological relationships should apply in other hilly areas. Unfortunately, the data needed to determine such relationships cannot easily be obtained, since areal rainfall measurements are required for short periods of constant synoptic type. A sufficiently dense network of autographic raingauges covering all areas of interest would be far too expensive to install and operate. Thus Nicholass and Harrold proposed that the climatological relationships for an extensive region should be derived either from a network of weather radars or from a numerical model capable of representing the effects of topography on rainfall (cf Sec 5b). A limited comparison of the values of $\left(R_s/\overline{R}\right)$ predicted by one particular fine-scale numerical model (Collier 1975) with the observed values in the Upper Dee area showed that the model did indeed produce values which tended to adjust the rainfall in the correct sense.

b Numerical prediction methods

Numerical prediction models currently in operational use do not properly resolve the orography; nor do they represent the physics of the orographic rain process. Even a model such as the fine-mesh version of the Meteorological Office 10-level model (Burridge and Gadd 1977) has a resolution (100 km) which is one or two orders of magnitude greater than the scale of the orographic features. Thus the predicted rainfall in the hillier parts of western Britain is often a considerable under-estimate. This is borne out by Fig 15(a), which shows an example of the predictions of the 10-level model for a hilly area of north Wales during 14 consecutive wet days with a southwesterly airflow. The predicted rainfall totals in Fig 15(a), which are for 24-h periods beginning at 0900 (based upon the preceding midnight analysis), are seen to be generally less than the corresponding rainfall totals as measured by the raingauge network.

Primitive equation prognostic models with spatial resolution of a few kilometres have been proposed for the prediction of orographic rainfall (eg Colton 1976) but such models are expensive to run. An alternative approach is to use the output from a standard coarse-resolution numerical prognostic model as an input to a separate, numerically simpler diagnostic model which has a grid size small enough to resolve the topography adequately. Kuhn and Quiby (1976) have used this approach to predict the detailed pattern of vertical velocity over the Swiss Alps, from which the rainfall intensity was inferred by means of a regression technique. Another such model by Collier (1975) has been used with some success to estimate the detailed pattern of rainfall accumulations in the hills of north Wales using an analysis of radiosonde observations as the input data. However, as Collier (1977) pointed out, his model does not contain the microphysics of the orographic rainfall process and it is valid only in the special circumstance when there is near 100% efficiency of washout (E_w). As discussed in Sec 4(c), the value of E_w depends critically on the background precipitation intensity and the wind speed at low levels. By running Collier's model for a hilly area upwind of the area of primary interest and comparing the predicted rainfall there with the actual rainfall measurable by radar, Harrold (1975) was able to derive E_w factors for different parts of precipitation systems. Then, by assuming that these efficiency factors remained the same when the different parts of the system being tracked by radar reached the area of interest, Harrold was able to obtain a much better forecast for that area than with Collier's unmodified model. The disadvantage of this approach of course is that it is limited to those occasions when there is a suitable region of high ground upwind of the forecast area. Thus Bell (1977) has derived an improved model which incorporates a microphysical scheme for explicitly predicting E_w whilst at the same time greatly simplifying the formulation for deriving local vertical velocity. The model, which has a horizontal resolution of $3\frac{1}{3}$ km and is based on a scheme by Jonas (1976), uses the equations of Bader and Roach (1977) to calculate E_w. The model also allows for the effects of evaporation and the horizontal drift of precipitation.

Bell has run his model using the output from the fine-mesh version of the 10-level model as input. He derived the vertical air motion simply from the sum of the large scale ascent predicted by the 10-level model and the topographically-induced component calculated assuming that the flow follows the terrain at the surface but decreases to zero at 500 mb. A more elaborate treatment of vertical motion such as that adopted by Collier was not used because the accuracy of the forecast input parameters was considered to be the dominating factor in determining the accuracy of the fine-scale model. Some indication of the accuracy of the predictions of daily rainfall over hilly terrain using Bell's model is given in Fig 15(b). This shows a considerable improvement over the results from the 10-level model alone (Fig 15a), with the predicted values of daily rainfall being much closer to those actually observed. Bell points out that his model does have a general tendency to overpredict the rainfall to the lee of hills, but he suggests that this can be overcome by taking proper account of the modification to the large-scale humidity field in the 10-level model due to the local removal of water by orographic rain.

c The use of data from a weather radar network

The good accuracy obtained in numerical model studies by Bell and others (eg Fig 15b) is applicable to forecasts of 24-h rainfall accumulations. This kind of numerical model is unlikely to resolve the detailed time variations in rainfall intensity, however, since it is driven by synoptic-scale input data and is not capable of predicting the occurrence of localised storms or mesoscale precipitation areas. The same would probably apply even to more sophisticated primitive-equation

mesoscale models such as that of Tapp and White (1976), at least in the case of the strong disturbed airstreams that are responsible for the situations of heavy orographic rain. Another approach toward more detailed forecasts might be to use a mesoscale numerical-dynamical model with observational inputs on the mesoscale.
Nash and Browning (1977) have studied the mesoscale patterns of wind and precipitation associated with large orographic falls of rain, mainly in and near the warm sectors of depressions, and they find that the wind and rainfall have a complicated distribution on the mesoscale in addition to the features imposed by orography. It thus seems that it will be very difficult to obtain sufficiently detailed observational data on the mesoscale far enough upwind to serve as the input to a full mesoscale primitive-equation model.

A more pragmatic approach to detailed prediction over short periods is to exploit the so-called Nowcasting method in which the actual distribution of precipitation is observed a limited distance upwind of the forecast area and individual mesoscale precipitation areas are tracked as they travel toward the target zone. The principal tool needed in this approach is an automated weather radar network with a system of communications which enables a map of the actual precipitation distribution to be received at the forecast centre in real time (Taylor and Browning 1974). Frequent cloud observations from a geosynchronous satellite are also likely to be valuable for the identification of likely precipitation areas in regions beyond the radar coverage. Since persistence of mesoscale features of the precipitation pattern plays so important a part in this kind of forecasting it might have been expected that this approach would break down in the presence of large orographic effects. However a preliminary study by Hill and Browning (1979) of a strongly orographic situation associated with warm sector rain in the Welsh hills offers some encouragement. Despite the very large modulation of rainfall intensity by orography, it was found that individual mesoscale precipitation areas tens of kilometres across could be tracked for 6 h over distances of 600 km, over land and sea, from the west coast of Ireland to southeast England.

The kind of procedure likely to be needed to achieve time-resolved quantitative precipitation forecasts up to several hours ahead is as follows:

(i) Replay a sequency of hourly radar (and satellite) plan pictures on a colour TV monitor to give a subjective impression of the movement of rain areas toward the target zone.

(ii) Derive simple objective forecasts of the movement of the spatially smoothed rainfall patterns over a 6-h period by means of a computer pattern-matching technique*. Many such extrapolation techniques exist: they have been reviewed recently by Collier (1978).

(iii) Replay the sequence of objectively forecast rainfall patterns on the TV monitor as a continuation of the sequence in (i) above, and modify the forecast patterns if necessary by iterative man-computer interaction so as to (a) produce subjectively more acceptable extrapolations of existing trends and (b) to take into account any development expected from other considerations.

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* **FOOTNOTE**

In the case of an area of precipitation which extends over a region of high land, it may be desirable, before determining its motion by extrapolation, to "disenhance" the precipitation intensity over the hills by means of the inverse application of the field of topographic enhancement factors mentioned in (iv).

(iv) Apply a computer-archived spatially-distributed field of topographic enhancement factors to the forecast precipitation patterns. The relevant field of climatological factors may be arrived at by selecting the appropriate category of surface wind and synoptic type (cf Sec 5a) and by using the washout efficiency E_w predicted by a fine-scale numerical model (cf Sec 5b). Occasionally the value of E_w changes abruptly, as for example at sharp cold fronts; radar data can sometimes be used to identify such a region from the characteristic echo pattern associated with it (James and Browning 1978).

The forecast product derived in this way within about 30 min of receipt of the latest radar data would be in a digital format suitable for immediate dissemination to users for display on a colour TV or, in the case of Water Authorities, for feeding into an objective hydrological model for the prediction of stream flow (Browning, Bussell and Cole 1977).

d A combined approach to forecasting orographic rain

The various forecasting techniques outlined above should be regarded as complementary. The numerical methods described in Sec 5b are likely to be the main source of the forecast for periods from 6 to 36 h ahead, although they may benefit from empirical refinements based upon a mesoscale precipitation climatology (Sec 5a) built up over a long period using data from a radar network. The radar network, supplemented by satellite data, is likely to provide the main source of data for more detailed time-resolved forecasts over the period 0 to 6 h ahead and in some cases up to 12 h ahead. However, radar data must be supplemented by estimates of the expected development and of orographic enhancement derived from numerical models and the climatological archive. The overall forecast system is depicted in the block diagram in Fig 16.

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TABLE 1 Some monthly totals of rainfall (mm) in the Uppsala area of Sweden for the years 1959 to 1965, restricted to periods with negligible convection. R_H and R_L represent the rainfall on the wettest part of Lunsen Hill and the driest part of Lake Ekoln, respectively. The overall value of $\frac{R_H-R_L}{R_L}$ for the 16 months tabulated is 0.50 (From Bergeron, 1967).

	September			October			November		
	R_H	R_L	$\frac{R_H-R_L}{R_L}$	R_H	R_L	$\frac{R_H-R_L}{R_L}$	R_H	R_L	$\frac{R_H-R_L}{R_L}$
1959	23	16	0.5	70	50	0.4	-	-	-
1960	45	30	0.50	55	35	0.57	-	-	-
1961	58	38	0.53	50	35	0.43	30	19	0.58
1962	66	44	0.50	35	25	0.40	25	13	0.95
1963	57	33	0.73	65	45	0.45	-	-	-
1964	67	49	0.27	75	55	0.36	-	-	-
1965	115	75	0.53	30	15	1.00	-	-	-
Σ	431	285	0.51	380	260	0.46	55	32	0.72

TABLE 2 List of the twenty occasions between January 1960 and December 1974 when the rainfall over a 24-h period exceeded 85 mm somewhere in the south Wales hills (within the circle shown in Fig 7). The 24-h period during which the rainfall exceeded 85 mm overlapped two so-called rainfall days on the 6 occasions denoted by asterisks. (From Nash and Browning 1977).

Rainfall day(s)	Dominant synoptic type	Maximum windspeed in low-level jet at 900 m (ms^{-1})	Maximum Rainfall in hills of south Wales (mm in 24 hours)	Representative coastal rainfall upwind of south Wales hills (mm in 24 hours)
3 December 1960	Warm sector/Warm front	40	150	30
11 September 1962	Warm front	30	90	50
17 November 1963	Warm front/cold sector trough	35	93	50
12 December 1964	Warm sector/Warm front	37	115	25
29/30 December 1964*	Warm sector	35	120	55
8/9 December 1965*	Warm sector	40	140	40
17/18 December 1965*	Warm sector/Warm front	31	220	50
27 February 1967	Warm front/Warm sector	40	110	25
29 July 1967	Warm front/Warm sector	25	105	40
16 October 1967	Warm front/Warm sector	35	135	40
23 March 1968	Warm sector/Warm front	37	110	15
11 August 1969	Thunderstorms		95	40
18 June 1971	Warm front/Warm sector	30	88	45
9 August 1971	Warm front/Warm sector	15	85	82
18/19 October 1971*	Warm sector/Warm front	35	110	20
15 February 1972	Cold sector Trough		90	20
4/5 December 1972*	Warm sector/Warm front	36	105	45
5 August 1973	Warm sector/Warm front	34	100	40
4 January 1974	Warm front/Warm sector	43	120	10
29/30 January 1974*	Warm front/Warm sector	35	140	20

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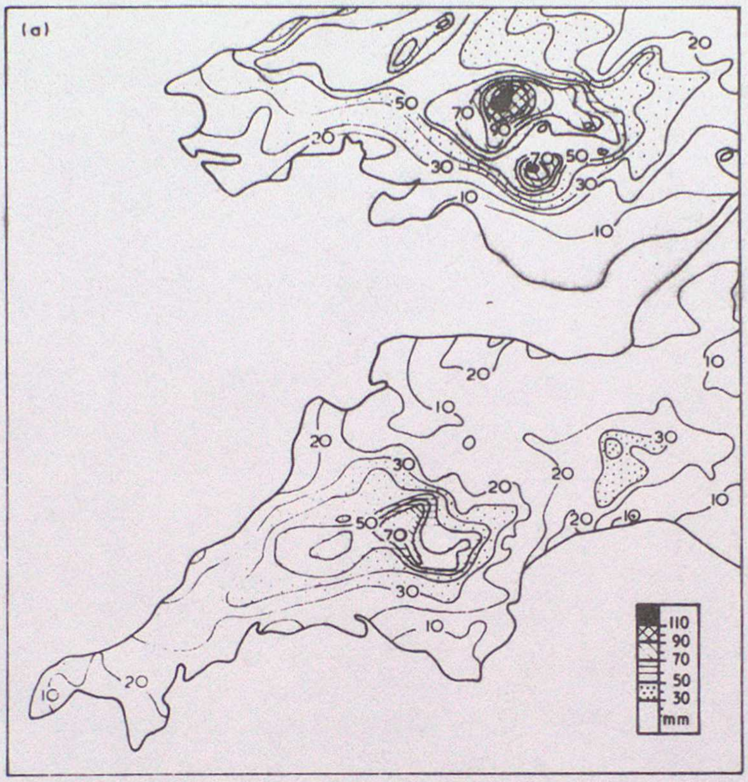
(a)



(b)

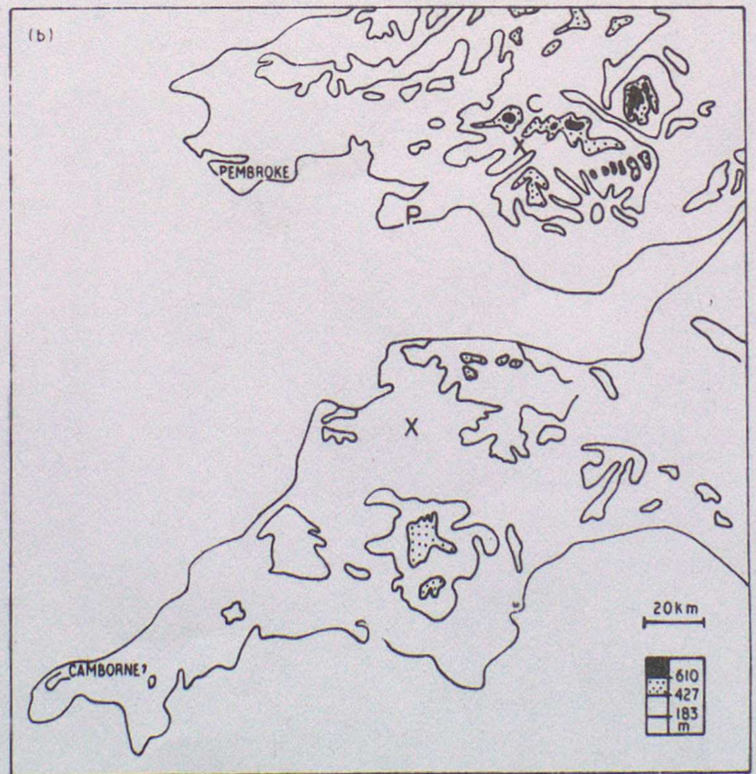
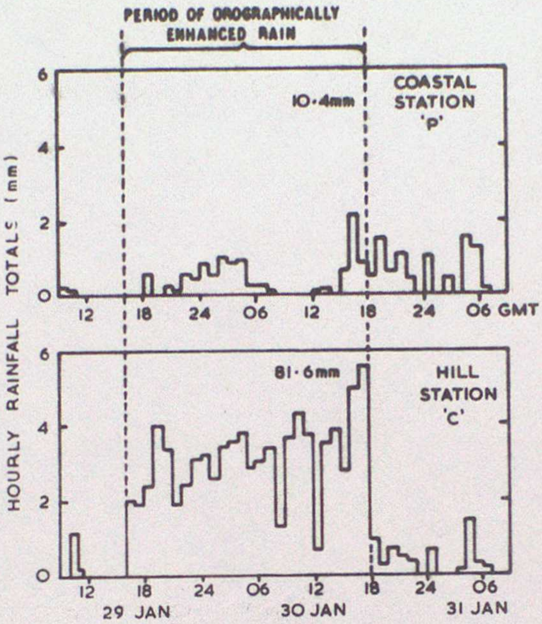


Fig 1



(c)

Fig 1 (a) Rainfall distribution in south Wales and southwest England on 29 and 30 January 1974. Isohyets are at 10 mm intervals.

(b) Topography of south Wales and southwest England, and locations of places mentioned in text.

(c) Histograms of hourly rainfall amounts for rainfall days 29 and 30 January 1974 at an unsheltered coastal station (P) and a hill site (C) in south Wales.

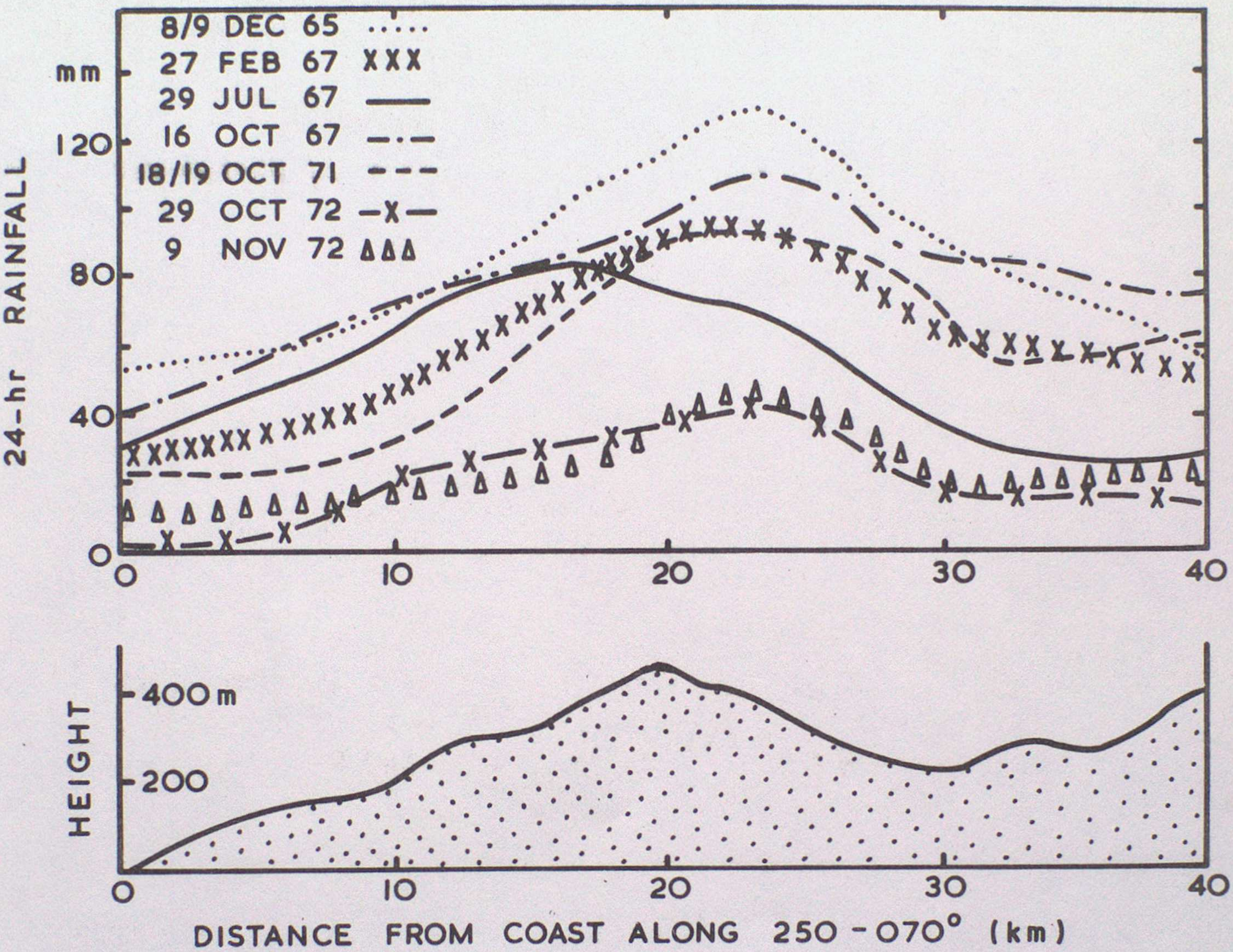


Fig 2 Distribution of 24-h rainfall within sections 10 km wide across the hills in south Wales known as the Blaenau Morgannwg for 7 occasions of prolonged orographic rainfall. In each case the winds were persistently from about 250° and a profile of the hill along this direction is shown at the foot of the diagram. (After Nash and Browning, 1977).

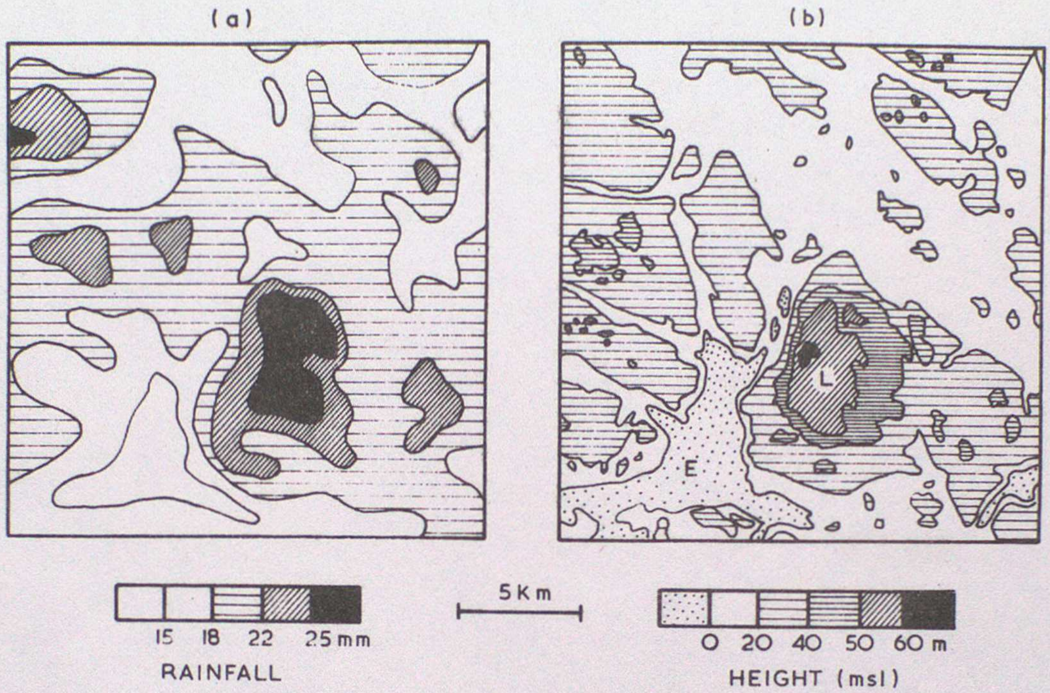


Fig 3 (a) Distribution of total rainfall in the Uppsala area of Sweden for the month of November 1962.

(b) Topography of the area surrounding Uppsala, showing the locations of Lunsen Hill (L) and Lake Ekoln (E) as mentioned in the text. (After Bergeron, 1967).

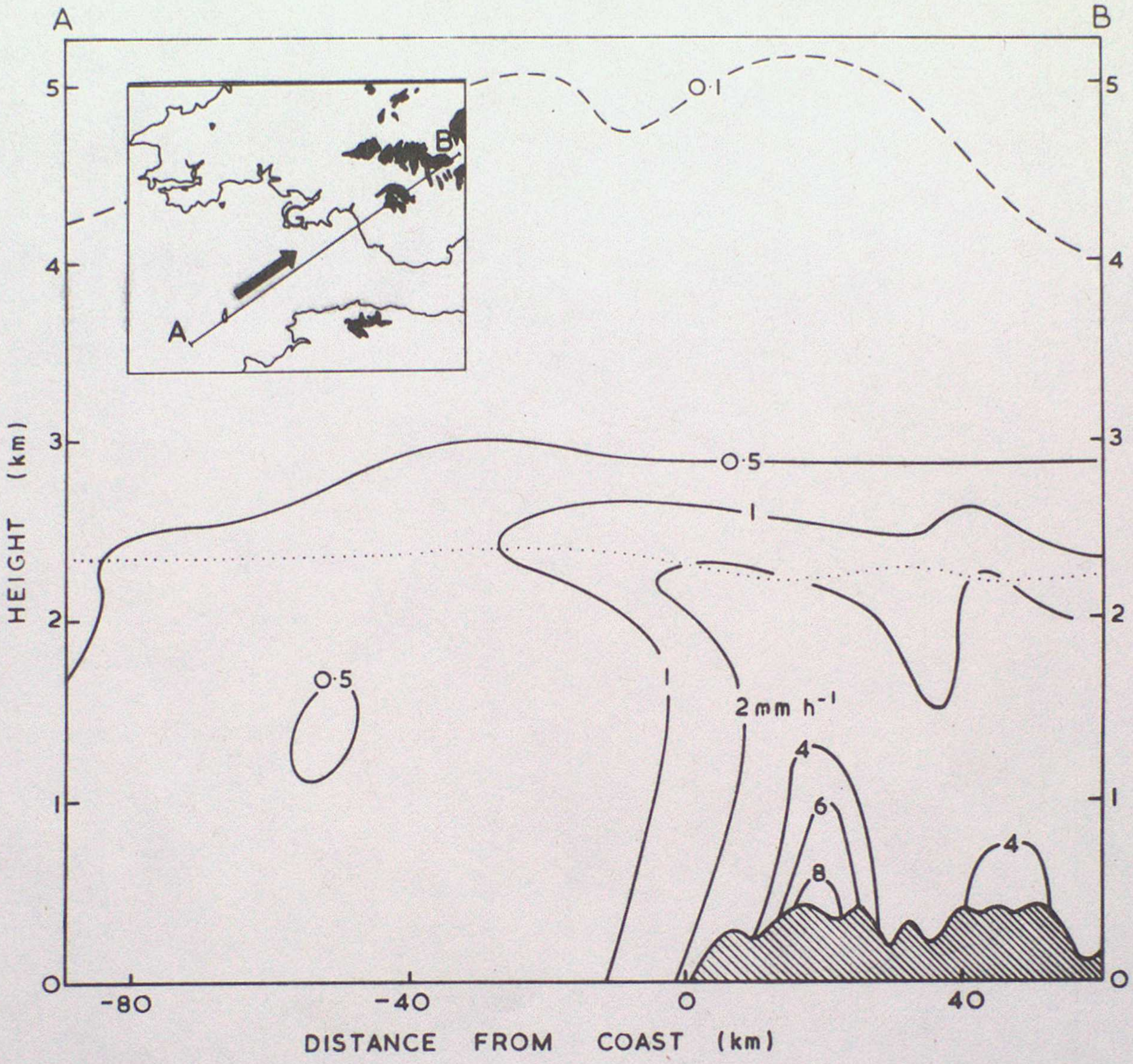


Fig 4   Time integrated pattern of radar echo intensity within a vertical section (AB) along the direction of travel of rain areas as they travelled from the sea over the hills of south Wales during a 5-h period of warm-sector rain on 28 November 1976. Isopleths are labelled in terms of rainfall intensity (mm h^{-1}), which is strictly meaningful only below the melting layer (dotted line just above 2 km). The dashed contour represents the upper limit of detectable precipitation. The inset shows the orientation of the section AB in relation to the coastline and hills (> 400 m) of south Wales. The raingauge-calibrated 10 cm radar used to obtain these data was situated at G. Analysis by J C V Pezzey and F F Hill, Meteorological Office Radar Research Laboratory.

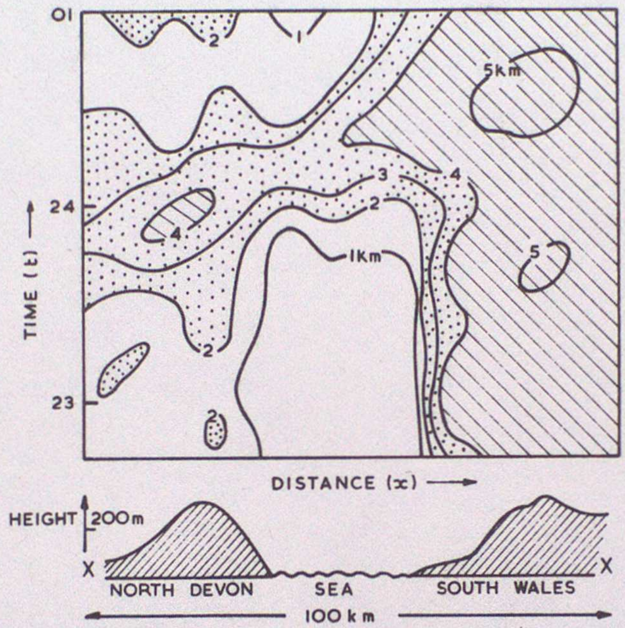


Fig 5 Height of the radar echo plotted in an x-t format, where t is time during the night of 29/30 January 1974 and x is distance parallel to the wind along a path from north Devon to south Wales (X to X in Fig 1b). The measurements have been obtained within a constant range interval of 80 to 100 km from a radar sited near Pembroke. Terrain height is plotted at the foot of the diagram.

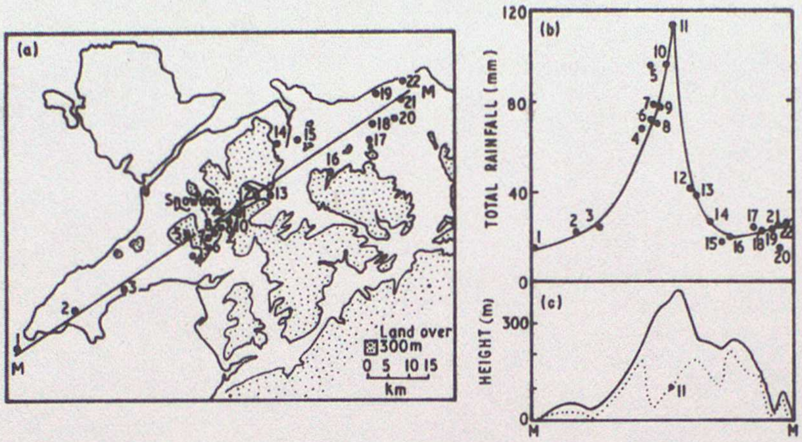


Fig 6 (a) Map showing the topography of Snowdonia and the locations of 22 raingauges whose data are plotted in (b).

(b) Distribution of rainfall (in mm) along the line MM in (a) which fell mainly in the period 1530, 26 June to 1220, 27 June 1966.

(c) Altitude of the raingauge sites. The height of gauges is indicated by the dotted curve. The full curve is a profile obtained by taking the average height within 3 km of each raingauge site. (After Pedgley, 1970).

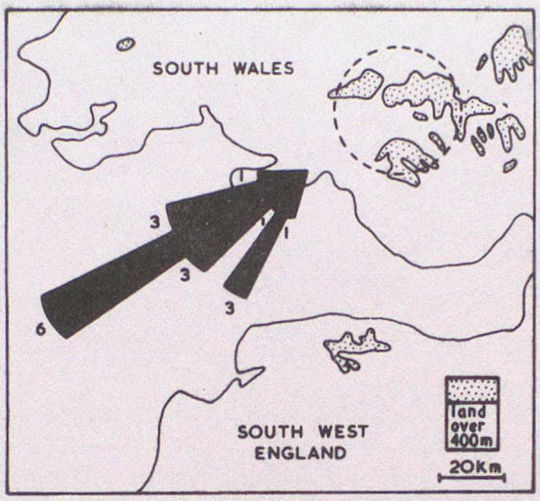


Fig 7 Wind rose giving the directions of the 900-mb wind associated with 18 of the 20 heaviest falls of orographic rain occurring during the period 1960 to 1974 in the parts of the south Wales hills shown encircled. The 18 cases each gave 24-h totals of more than 85 mm; they were all warm sector and/or warm frontal situations, with 900-mb winds mainly in excess of 30 ms^{-1}. The 2 cases not shown were cases of lighter winds not associated with warm sectors or warm fronts. (After Nash and Browning 1977).

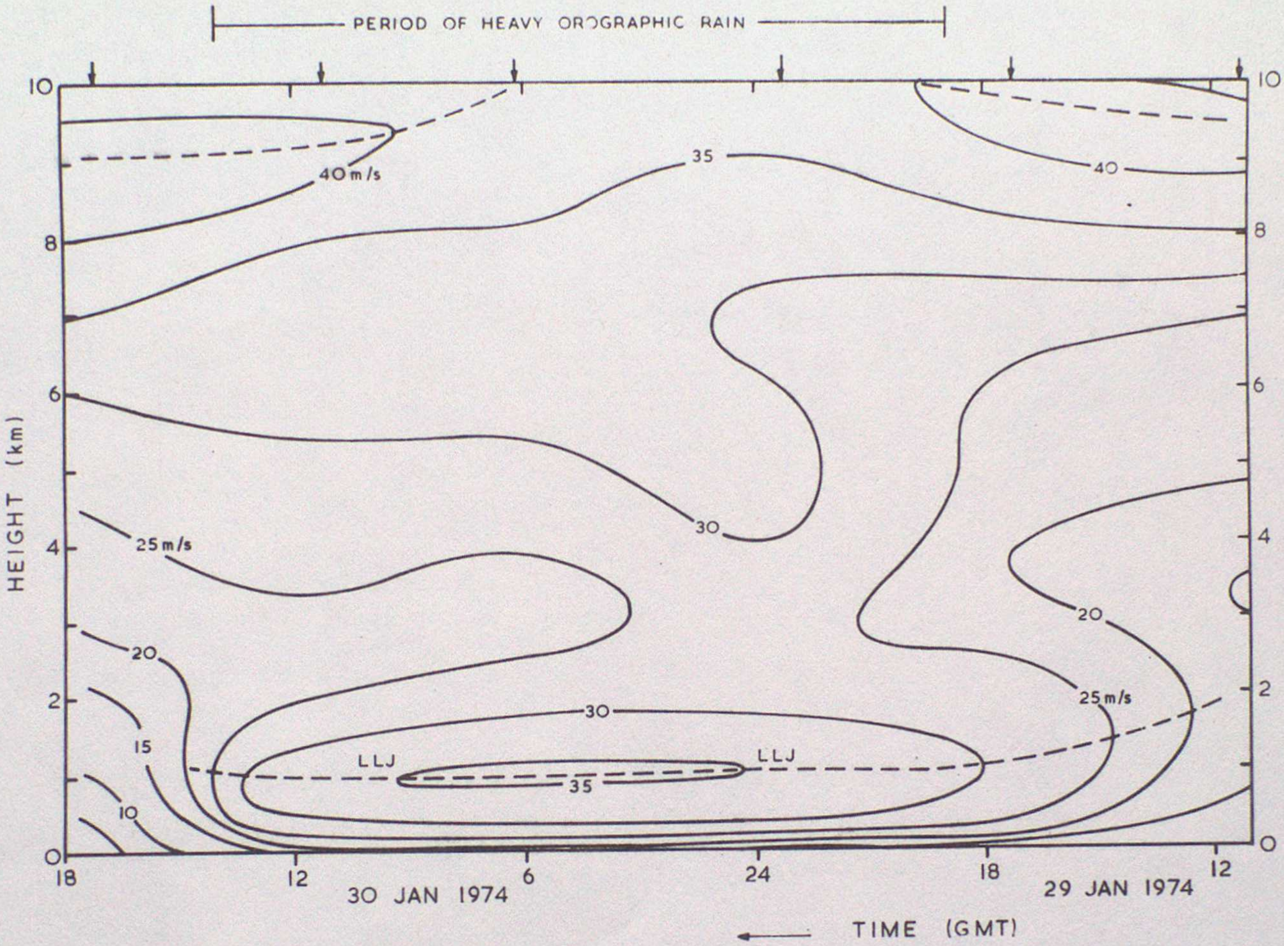


Fig 8 Time-height section of wind speed showing the strong low-level jet which persisted throughout the period of orographic rain that produced the large falls depicted in Fig 1(a). The times of the 6 soundings used to construct this diagram are shown by arrows at the top of the figure. The wind soundings were made from Camborne, the location of which is indicated in Fig 1(b).

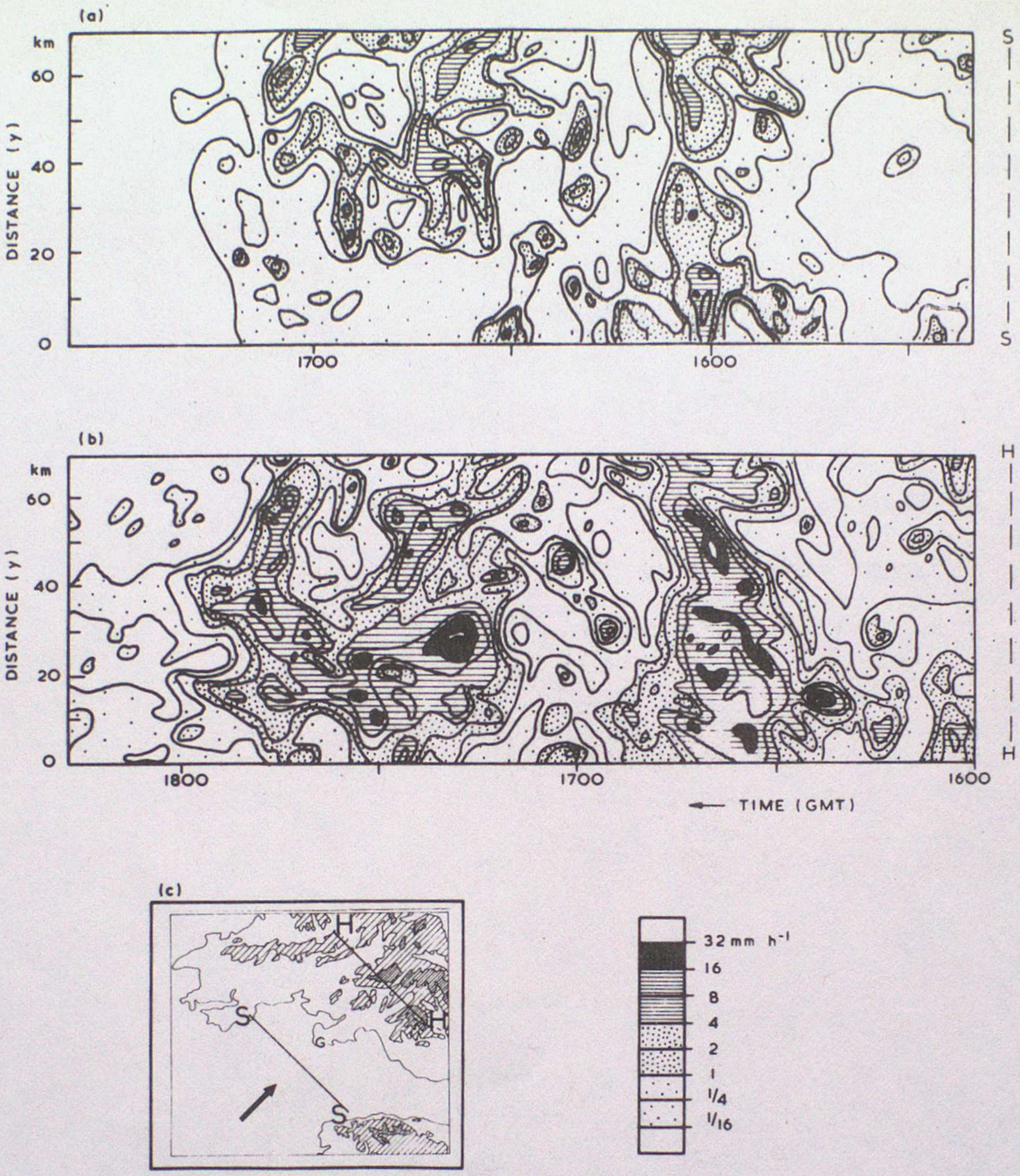


Fig 9 Rainfall intensity at an altitude of 800 m above msl associated with a warm front on 25 January 1977, plotted in a t-y format where t is time and y is distance normal to the direction of travel of small rain areas along (a) the line SS over the sea and (b) the line HH downwind over the hills of south Wales. The locations of lines SS and HH are indicated in (c). Land above 200 and 400 m, respectively, is shaded lightly and heavily in (c). Knowing the speed of travel of the rain areas, the time axes in (a) and (b) have been drawn so as to correspond to a distance scale comparable to the y axis. The two t-y diagrams are displaced laterally so that given rain areas are located in approximately the same relative positions in (a) and (b). Areas of rain which crossed HH before 1700 can be identified 40 min earlier crossing SS at the same position along the y axis; however, because of a slight change in the direction of travel of rain areas, those which crossed HH after 1700 crossed SS about 10 km farther northwest toward larger y. These diagrams have been derived using a raingauge-calibrated 10 cm radar on the Gower peninsula (G) between SS and HH. Analysis by F F Hill, Meteorological Office Radar Research Laboratory.

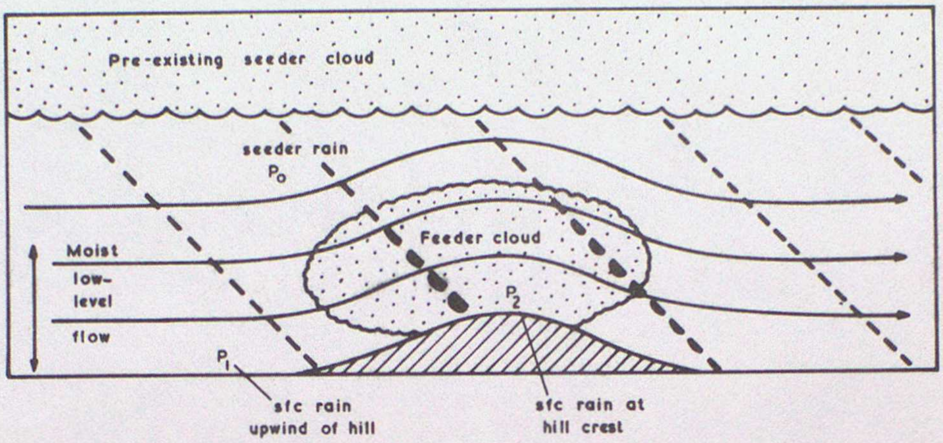


Fig 10 Conceptual model illustrating the orographic enhancement of rain (After Bergeron 1965).

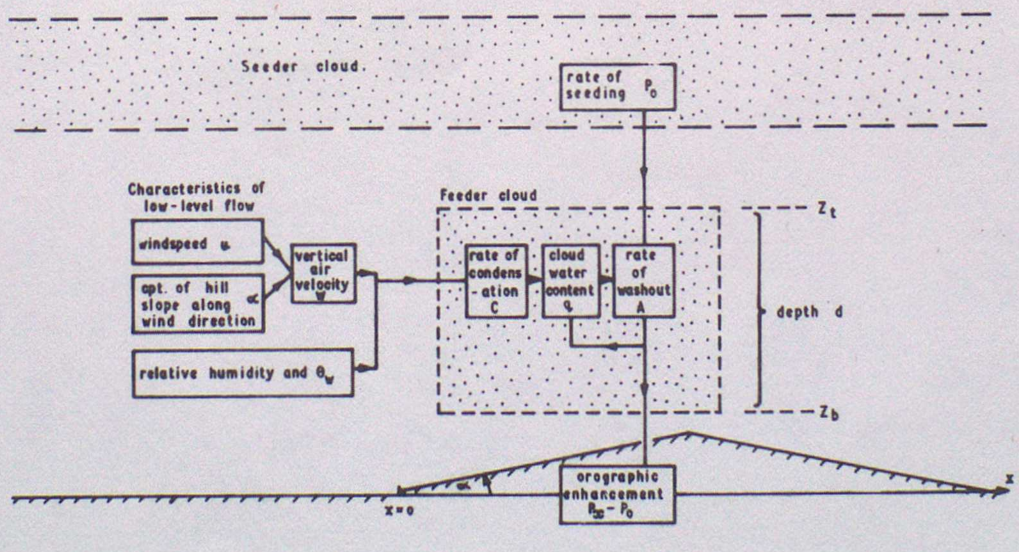


Fig 11 Schematic diagram illustrating the dependence of orographic enhancement on the characteristics of the low-level flow and the rate of seeding from above.

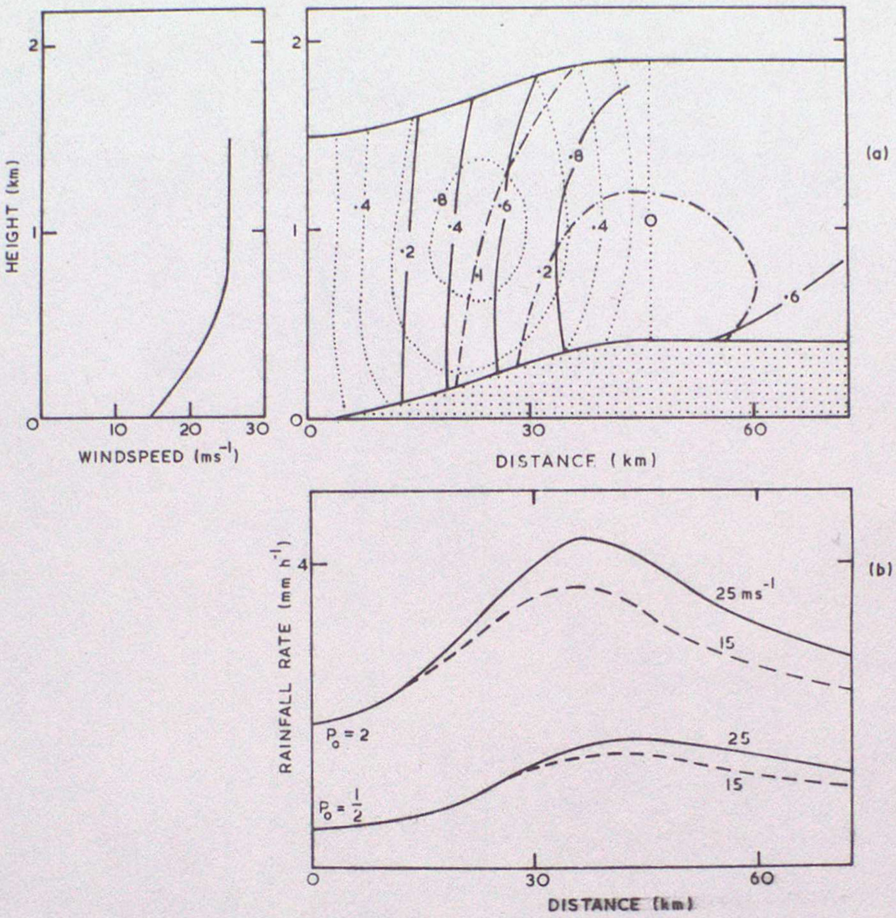


Fig 12 (a) Computed distribution of condensation rate, C, (......) and washout rate, A, (-.-.-.-) in mg $m^{-3}s^{-1}$, and liquid water content, q, (———) in g m^{-3} in a vertical section through an orographic feeder cloud oriented along the wind direction. The calculations assume a saturated airflow adjacent to the ground of depth d = 1.5 km, with $\theta_w = 10^oC$, seeded by rain from above at a rate P_o of 0.5 mm h^{-1}. The wind profile is as shown on the left, the maximum speed above the friction layer being 25 ms^{-1}. The assumed hill profile is shown stippled, the peak height being 400 m.

(b) Computed surface rainfall rate P as a function of distance x across the hill depicted in (a). The solid curves, for values of P_o of approximately 0.5 and 2 mm h^{-1}, are for conditions otherwise as specified for (a). The dashed curves show the change in P which results from decreasing u to 60% of the magnitude assumed in (a). From Bader and Roach (1977)

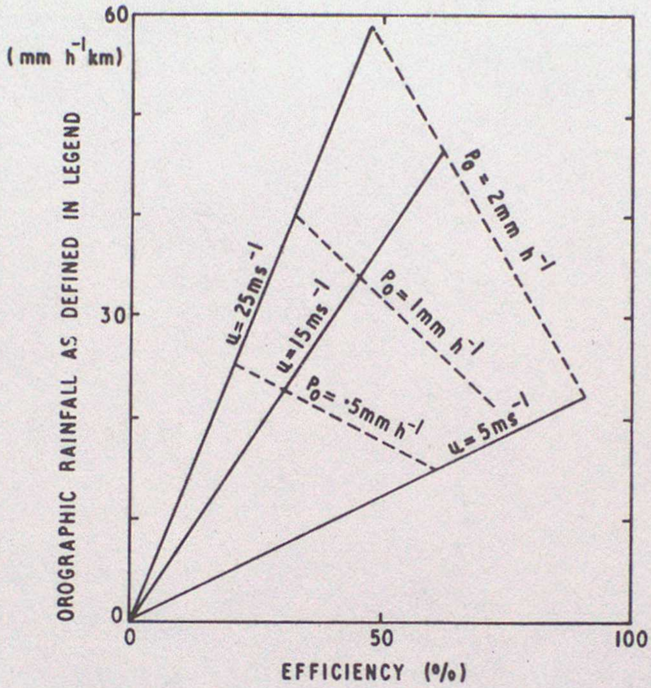


Fig 13 Orographic component of the rainfall intensity integrated from the bottom of the hill to the crest $\left(\int_o^{x_c}\left(P_x - P_o\right)dx\right)$ for the hill depicted in Fig 12, as a function of the efficiency of washout E_w, for different values of windspeed u and seeding rainfall rate P_o. The calculations assume a saturated airflow adjacent to the ground of depth 1.5 km, with $\theta_w = 10^{\circ}C$, and winds which decrease close to the ground as shown on the left of Fig 12. (From Bader and Roach 1977)

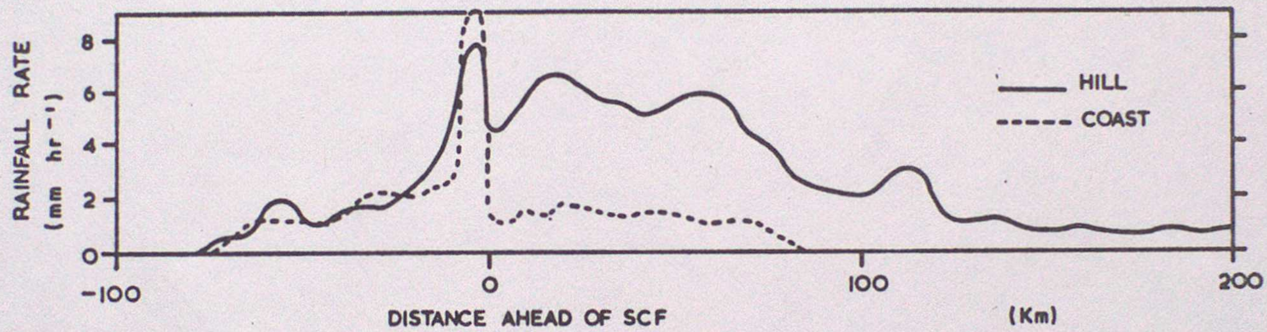


Fig 14 Time records of the average surface rainfall intensity at a number of hill sites (full line) and nearby coastal sites (dashed line) in south Wales during the passage of a cold front on 9 November 1972. The time axis is labelled in terms of distance normal to the surface cold front. (From Browning, Pardoe and Hill 1975).

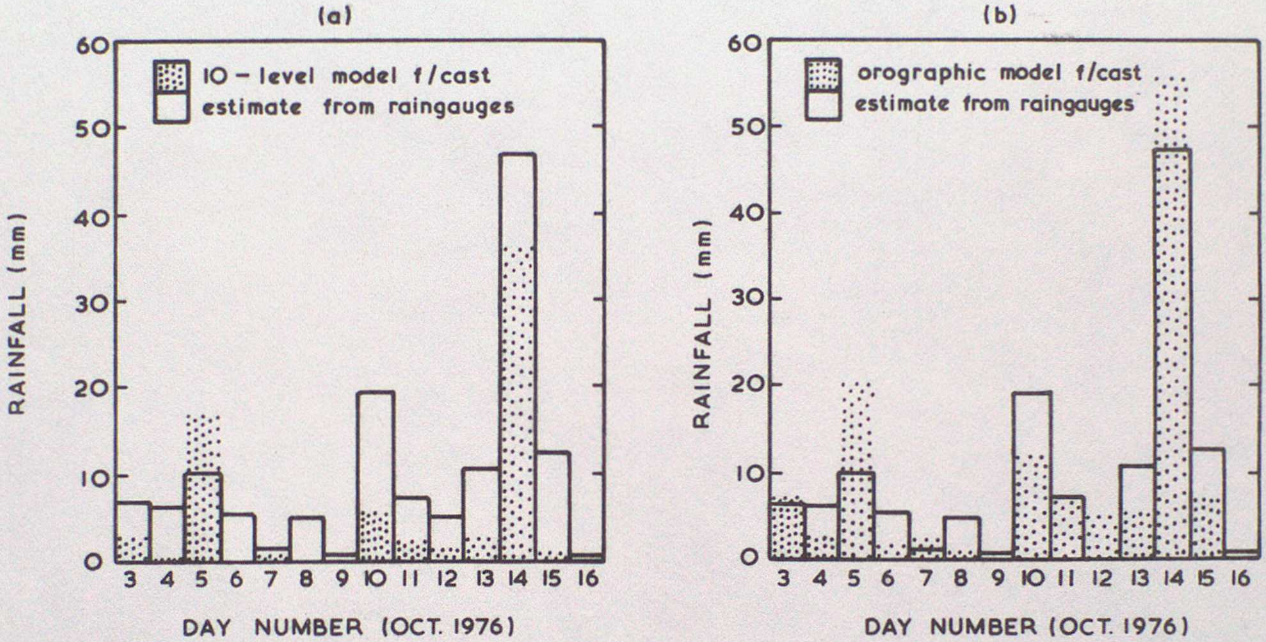


Fig 15 Predicted daily rainfall totals for the period 3-16 October 1976 for a 50 x 50 km area in north Wales derived using (a) the fine mesh version of the Meteorological Office 10-level numerical model alone and (b) an orographic model driven by the 10-level model. Observed rainfall is shown in both (a) and (b) for comparison. (After Bell 1978)

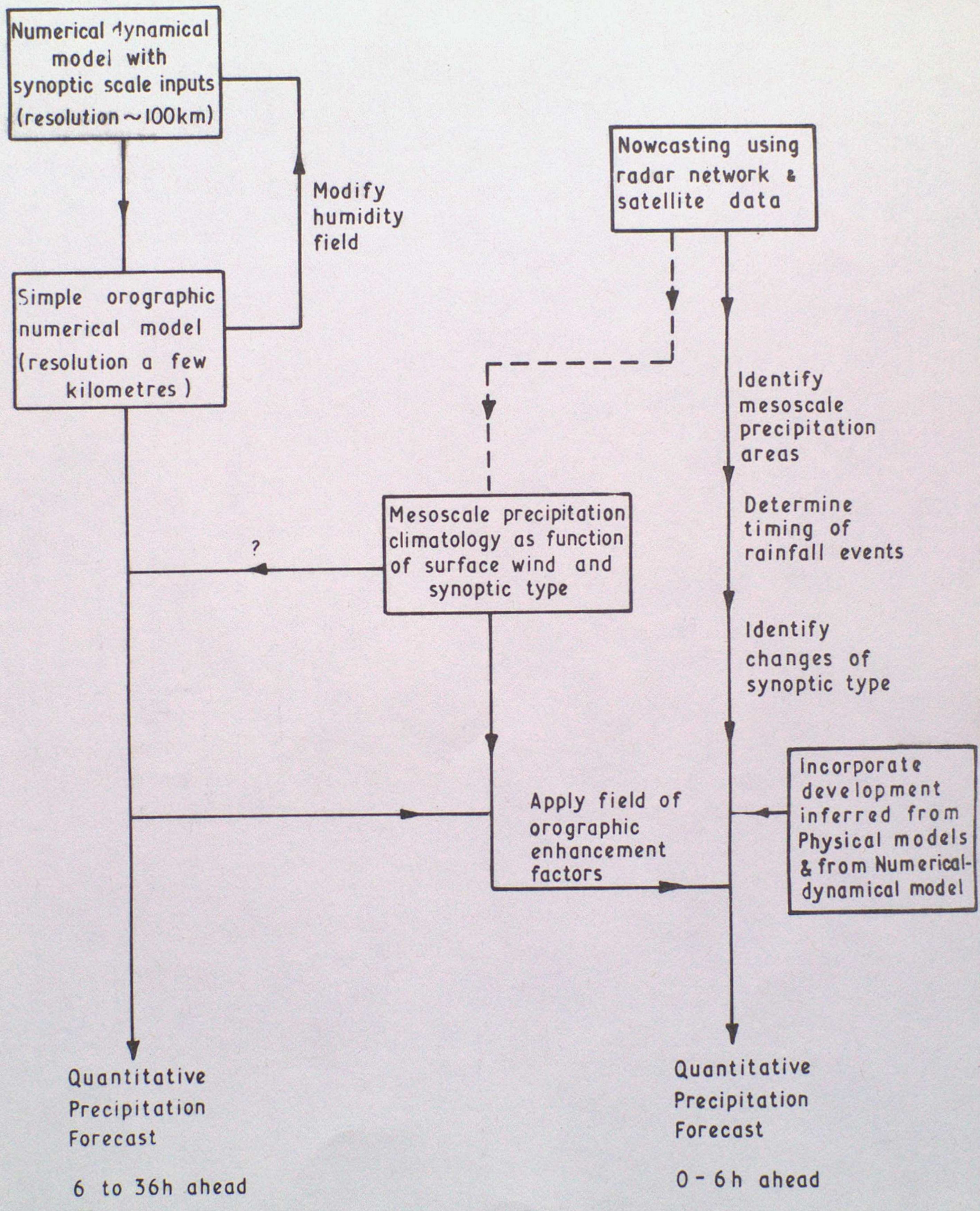


Fig 16 Scheme for forecasting orographic rain.