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STRUCTURE OF THE LOWER ATMOSPHERE ASSOCIATED WITH HEAVY FALLS OF OROGRAPHIC RAIN IN SOUTH WALES

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SUMMARY

In this paper we analyse for a 14-year period all the falls of rain in the hills of South Wales which exceeded 85 mm in 24 hours. Eighteen of the twenty cases of heavy rain which met this criterion were associated with warm sector/warm frontal conveyor belts, and all but one of these were characterized by moist low-level jets in which the winds at 900m were between 25 and 43 m s^{-1} (predominantly from the southwest). Maximum winds in these jets were close to geostrophic.

The existence of a strong moist flow in the lowest 2 or 3 km is important because it leads to a high rate of generation of liquid water within low level orographic cloud forming over the hills. Since most of the orographic increment in rainfall probably originates from this low-level cloud, it is sometimes referred to as a feeder cloud. For an initially saturated airstream of given velocity, the orographic increment depends on the efficiency with which the small cloud droplets in the feeder cloud are washed out by precipitation particles (seeder particles) entering it from above. In a recent article [redacted] Bader and Roach have suggested that it may be possible to estimate the washout efficiency from the intensity of the surface precipitation upwind of the hills. The present study to some extent supports this view since it indicates a threshold of rainfall intensity on the coast upwind of the hills below which a high washout efficiency was not achieved. Thus, when the coastal rainfall rate was less than 0.5 mm h^{-1}, the orographic increment in the wettest parts of the hills was always less than 4 mm h^{-1}. There was, however, no obvious dependence of the orographic increment on the coastal rainfall intensity

as the latter increased above 0.5 mm h^{-1}. The lack of any closer relationship is attributed mainly to growth or evaporation of precipitation at low levels upwind of the hills and also to triggering of potential instability by the topography, as a result of which the surface rainfall rate at the coast is not a reliable measure of the rate at which seeding particles from above enter the orographic feeder cloud.

It is widely held that the heaviest orographic rain in the western parts of the British Isles occurs in warm sectors, usually within 150 to 300 km of a surface front. The present study confirms this, but at the same time it demonstrates that a warm front that is preceded by a strong low-level jet can also be associated with very large orographic enhancement. Situations in which heavy rain was observed at orographic sites more than 150 km from the closest surface front were associated mainly with the leading edges of low-level jets embedded within the warm sector. The depth of moist air in warm sectors is highly variable, often over short distances, and by no means all parts of the low-level jets were characterised by orographic rain even in the exceptionally wet situations studied in this paper. Generally the most favoured locations for orographic rain are where the mesoscale circulations associated with the jets lead to low-level convergence.

1 INTRODUCTION

For many years it has been recognised that prolonged heavy rain can occur over the hills in Western Britain within the warm sectors of depressions if there are strong moist flows at the surface from between W and SSW ahead of a slow-moving or quasistationary cold front. Rainfall of high intensity can also be associated with the passage of a warm frontal zone but in this case the orographic increment (mountain - lowland) in rainfall is generally held to be of smaller magnitude (eg, Douglas and Glasspoole, 1947). These conditions, when further qualified by the requirement for a deep moist layer within the warm sector, provide the basis for practical forecasting of heavy orographic rain in semi-quantitative terms (eg Holgate, 1973).

Recently several studies have been carried out at Malvern in which radar observations have been combined with frequent rawinsonde ascents to investigate the detailed vertical structure of frontal rain systems. A belt of strong winds with a local maximum in the vertical between 0.9 and 1.5 km has been identified within warm sectors, usually immediately in advance of the cold front(Browning and Pardoe, 1973). Such low-level jets were associated with tongues of air of higher wet bulb potential temperature (θ_w) than their immediate environment. Harrold (1973) found similar moist warm tongues, with widths ranging from 100 to 1000 km, and he referred to them as conveyor belts. As is shown later in this paper, such conveyor belts are synoptic-scale features which can contain several mesoscale low-level jets.

Four case studies of the interaction of frontal systems with the hills of south Wales were reported by Browning, Pardoe and Hill (1975), hereafter referred to as BPH (1975). They showed that the presence of a low-level jet leads to the generation of intense orographic precipitation on many occasions, but only when certain other conditions are fulfilled. It is necessary, first, that the low-level flow be close to saturation so that ascent over hills results in the production of a 'feeder' cloud of small droplets. And secondly, there must be sufficient precipitation-size hydrometeors, sometimes called 'seeders', falling through the feeder cloud to wash out a significant proportion of the small droplets to the surface (cf Bergeron, 1965).

The rate of condensation in the feeder cloud is, to a first approximation, related linearly to the strength of the incident low-level flow. In the case of a narrow range of hills, however, a strong low-level flow implies a short lifetime for individual droplets within the feeder cloud, and so it becomes unlikely that microphysical interactions can generate an adequate supply of seeder particles from within the orographic feeder cloud itself. This is why it is necessary that seeder hydrometeors should be generated independantly of the orographic feeder cloud if the orographic rainfall potential of the high rates of condensation is to be realised. The presence of rainfall upwind of the hills may be taken as an indication of the availability of such seeder hydrometeors.

In this paper, we present the results from analyses of a large number of the wettest cases over the hills of south Wales during a period of approximately 14 years, with the intention of providing a representative sample of the mesoscale conditions associated with heavy orographic precipitation. Frequent rawinsonde ascents were not available in the majority of these cases and so detailed mesoscale information was generally limited to levels close to the surface. An emphasis on the structure of the atmosphere at low levels is reasonable because it is the mesoscale circulations at low levels that control the generation and distribution of moisture in the layers from which most of the orographic enhancement originates.

2 NATURE OF THE DATA

We have investigated the heaviest falls of rain from December 1960 until January 1974 in the area of south Wales bounded by the dashed line in Fig 1, using the criterion that more than 85 mm of rain should have fallen in any 24-hour period.

(a) The rainfall data

Data from the dense network of 24-h raingauges in south Wales indicate that large orographic increments are often generated on the Blaenau Morgannwg (the hills of central Glamorgan) and also on exposed areas of the neighbouring chain of hills oriented east-west which includes the Brecon Beacons, (Fig 1). The peak rainfall is often similar in the two areas, the latter hills being farther inland but of greater height. In this study we concentrate much of our attention on the Blaenau Morgannwg, since these are exposed directly to the sea when the wind is in the sector between W and SSW. Within this sector there is little variation with wind direction in the overall upslope gradient of the hills. Autographic raingauges at Treherbert Park and nearby Treorchy, in the heart of the Blaenau Morgannwg, receive very high orographic rainfall although they are situated in a deep valley. Data from the network of 24-h gauges showed that, when the low-level winds were from the open sea, the rainfall at these two sites was nearly always within 10% of the orographic maximum in the Blaenau Morgannwg. Thus, measurements from these two gauges have been used to form the basis of the intercomparison between rainfall systems in the present study.

In order to elucidate any relationship between the mesoscale pattern of low-level wind and orographic rainfall enhancement (sec 3(e)) the analysis of rainfall for each weather system has also been extended beyond the Blaenau Morgannwg to cover a much larger area. Large orographic falls are limited to exposed hills in the coastal regions of western Britain in the systems we have considered, with the result that the measurements of orographic rain are available essentially in only one dimension (y) across the country. We therefore compare the fields of wind and rain using a y-t format, in which data from orographic sites are plotted along a y-axis oriented at right angles to the overall synoptic scale motion of the weather system. Wind fields have been extracted at one hour intervals from x-y analyses of low-level winds, while the individual rainfall records from orographic sites (suitably time-shifted to compensate for the off-axis location of the sites) have been inserted directly into the y-t format. Variations from site to site in measured rainfall intensity (owing to differences in orographic increment caused by the differences in local topography) have been compensated by normalising rainfall intensity to that which would have been realised if that part of the system had passed over hills resembling those near Treherbert. The adjustment factor at a site was estimated from a number of comparisons between the rainfall at that site and that at Treherbert when atmospheric conditions were similar at both sites. The intensities, once compensated, are referred to as adjusted rainfall intensities throughout the remainder of this paper. Interpolation of the rainfall fields between sites with large orographic enhancement in the y-t diagrams has been based on the rainfall records from sites where orographic effects are less pronounced. In these latter regions the values of adjusted rainfall intensity represent a potential which was never attained in reality.

(b) The low-level wind data

Although in a limited number of cases sequential ascents and radar observations were available from Castlemartin (Dyfed) or from Defford or Pershore (Worcestershire), on most occasions the information on upper air structure was limited to the soundings of the operational rawinsonde network. The spacing of this network over the British Isles is such that it is impossible in most cases to define the mesoscale structure of the flow at upper levels with any precision. However, at a level of 900m, near the top of the friction layer, it has been demonstrated that information from sequences of hourly surface wind observations at unsheltered sites can be used to interpolate between the upper air observations within warm sectors and ahead of warm fronts, except where the main frontal surfaces intersect the friction layer (Nash, unpublished). Continuity allows fluctuations at individual stations originating from smaller scales of motion to be smoothed out, and final estimates of the magnitude of V_{900} at intermediate positions have been found to have a probable error of about $\pm$ 2 ms^{-1}. Surface observations at Mumbles (see Fig 1) play a major part in the estimation of the magnitude of V_{900} incident on the Blaenau Morgannwg. The coupling ratio of the magnitude of V_{900} to the surface wind speed at Mumbles is about 1.7 when the winds are incident from the open sea. This is a typical value for a coastal site whereas unsheltered inland stations generally have values between 2 and 2.5.

3 GENERAL CHARACTERISTICS OF THE CASES OF HEAVY OROGRAPHIC RAINFALL

(a) Synoptic classification and seasonal distribution

A list of the 20 cases investigated in detail is presented in Table 1. Dominant synoptic type, maximum 900m wind associated with the main period of heavy rain and an indication of the orographic rainfall from the 24-h gauge totals are indicated for each case. On 18 occasions the heavy rain was associated with warm sector/warm frontal conveyor belts, all but one with strong 900m winds in the range from 25 to 43 $m\ s^{-1}$. The two remaining cases, which are not treated in detail in this paper, were the result of, on one occasion, organised convection associated with a trough in a strong cold air flow 150 km from the centre of an occluded depression and, on the other, of a series of intense thunderstorms which developed in a strong baroclinic zone oriented north-south over Wales in a region of strong cyclonic shear but weak low-level flow.

The selection criterion acted to exclude many cases of intense but short-lived rainfall associated with fast-moving circulations of limited spatial extent. In only 2 of the 18 conveyor belt cases was the duration of the warm sector over the

Blaenau Morgannwg less than four hours. Rain within warm sectors contributed more than half the 24-h total over the hills in the majority (13) of the conveyor belts. In the remaining 5 cases the rain ahead of the surface warm front predominated.

The seasonal distribution of the heaviest daily rain over south Wales was found to reach a maximum in December (Table 1). When the cases are categorised according to the mean wet bulb potential temperature (θ_w) in the lowest 2 km of the warm-sector conveyor belt, it is found that the 12 cases which occurred between November and March had θ_w in the range from 8 to 12°C, whilst the remaining 6 between June and October had θ_w in the range from 14 to 18°C. Although the rate of condensation during orographic uplift increases as the θ_w of the moist air increases, this does not have a predominant effect on rainfall because the strongest low-level flows were associated with the winter-time situations of lower θ_w.

(b) Direction of the low-level flow

Figure 2 shows the distribution of the wind direction at 900m associated with the generation of heavy orographic rain over south Wales from warm sector/warm frontal conveyor belts. Histograms are presented for two categories of rainfall: the first (Fig 2(a)) for the 18 cases considered in this study and the second (Fig 2(b)) for 51 more cases from the same period in which rain within a rainfall day was between 50 and 85 mm. The Blaenau Morgannwg and central Brecon Beacons are directly exposed to the open sea when the surface winds are in the sector between 220 and 250°. The peak at 240° in the histogram of the heaviest rainfall corresponds to direct incidence from the open sea for the entire depth of the low-level flow. When the wind at 900m is from 220° the winds close to the surface approach the Blaenau Morgannwg from across southwest England, (see arrow in Fig 1) and consequently the relative humidity of the low-level flow will tend to be less than in an equivalent weather system incident from 240°. Thus, although system flows from 220° are as common as those from 240° in the lower rainfall category (Fig 2(b)), those from 240° dominate the heaviest rainfall category (Fig 2(a)). When flows are from the sector between 180° and 210°, the hills of central south Wales are sheltered to varying degrees by the hills of southwest England, particularly Exmoor and so the maximum orographic effect then shifts to the western parts of the Brecon Beacons.

(c) Vertical structure of the low-level winds

In 15 of the 18 conveyor belt situations that were investigated in detail, individual soundings existed in which a wind maximum was observed at or below 800 mb and above which the wind speed decreased by at least 2 ms^{-1} within 2 km. On the other three occasions, a low-level maximum was still observed but the speed decrement was only 0.5 ms^{-1}, which is within the limits of uncertainty in the measurements. Nevertheless it seems probable that in every conveyor belt considered in this paper there were large regions characterised by a significant low-level wind maximum. Ninety per cent of the observations of low-level maxima were at or below 850 mb, the magnitude of the maximum ranging from 15 to 43 ms^{-1}.

In an attempt to assess the departure from the geostrophic value of the magnitude of the winds in the low-level jet maxima, estimates of the wind speed at 900m have been plotted in Fig 3 against the surface geostrophic wind speed derived from the surface pressure field on a grid length of 100 km across the Bristol Channel. The estimates of V_{900} are considered to be subject to an uncertainty of only $\pm$ 1 ms^{-1} since they incorporate direct measurements from Camborne. The uncertainty in the estimate of the surface geostrophic wind resulting from difficulties in defining the surface pressure field on a grid length of 100 km can be as large as 10%. In most of the cases studied here the geostrophic winds at 900m will have been marginally less than the surface geostrophic winds and so this reinforces the impression given by Fig 3 that the winds at 900m are on average slightly supergeostrophic. By contrast the low-level jets found in association with bands of heavy rain in Japan during the Baiu season are often substantially supergeostrophic above 900m (Matsumoto, 1972). Evidently, there are significant differences in the origins and circulations of the low-level maximum in this study from many of those observed in the Baiu season in Japan.

On one of the occasions studied (30 January, 1974) the speed and orientation of the weather system made it possible to use data from the midday operational soundings augmented by additional pilot balloon ascents at $\pm$ 6 h to define the vertical wind structure with much greater precision than in the other cases. This situation is now described briefly to illustrate the kind of weather system with which this paper is concerned. The analyses of the low-level wind field and 1000-700 mb thickness are presented in Figs 4(a) and 4(b), respectively, whilst a vertical section constructed along the line AB in Fig 4(b) at right angles to the axis of the low-level jet is presented in Fig 5. The low-level jet maximum in advance of the front was 35 ms^{-1} above which there was a speed decrement of about 10 ms^{-1}. This decrement, one of the largest found in this study, was produced by strong warm-frontal baroclinicity east of the surface cold front (Fig 4(b)). The transverse circulation represented by the bold arrows in Fig 5 intensified this baroclinic zone by advecting colder drier air from continental Europe within the backed flow at low levels. Close to the surface cold front warm moist air from the boundary layer ascended on the cyclonically sheared flank of the low-level jet and then continued rising as slantwise ascent in the region of veered flow centred at about 800 mb. Heavy orographic rainfall was observed over the hills of southwest England and Wales at this time, with one rainfall zone extending from the cold front 150 km into the warm sector in association with the main low-level jet in Fig 5, and a second zone associated with an earlier low-level jet, the tail of which gives rise to a second weaker low-level maximum extending from 220 to 270 km in advance of the cold front.

(d) Vertical structure of humidity

The humidity structure within warm sector/warm frontal conveyor belts has large spatial variability. This is illustrated by the sequence of rawinsonde ascents through the system of 16 October 1967 described by Browning and Harrold (1969), a system which is also one of those selected for the present study. On this day the depth of the layer with dewpoint depression less than 3°C in the conveyor belt which fed the active warm front was less than 1.5 km about 100 km behind the front but increased to more than 5 km at the front itself. In at least four other situations the moist low-level flows feeding into a region of heavy rainfall were initially less than 2 km deep. On several other occasions, too, shallow moist layers were observed in warm sectors within 100 km of the region of heavy rainfall. However, when humidity soundings were obtained actually within a region of heavy orographic rain, in every such case the depth of the moist layer was greater than 3 km and in more than half of these cases it was greater than 4.5 km. The humidity structure at low-levels is strongly influenced by mesoscale regions of low-level convergence and divergence. Of course, saturation does not necessarily exist throughout a zone of low-level convergence; rather its extent depends on the initial relative humidity of the air entering this zone, on the magnitude of the convergence and on the strength of the low-level flow relative to the motion of the convergent zone. Unfortunately, the great spatial variability in the depth of the moist layer makes it impossible to interpolate between routine humidity soundings at upper levels with any confidence and so no attempt has been made to produce statistics of the depth of the moist layer over south Wales associated with high orographic rainfall.

(e) Distribution of heavy orographic rain in relation to the mesoscale pattern of low-level winds.

The analyses of low-level wind fields in the 18 conveyor belts of the present study have indicated that the heavy orographic rain is associated with regions of convergence in the vicinity of the low-level wind maxima which can be divided into two main categories:-

(1) C-type, in which convergence within the boundary layer is encountered where there is strong cyclonic wind shear bounding the left flank of the region of strongest winds at low levels.

(2) N-type, in which convergence within the low-level flow is encountered near the jet nose, ie, ahead of the region of strongest winds at low-levels, and where the convergent zone may not be linked to regions of strongly cyclonic wind shear in the boundary layer.

These types of situation may be encountered in association with conveyor belts both within warm sectors and where the conveyor belt flow begins to ascend above the warm frontal zone. Table 2 shows the frequency of occurrence of the different categories of circulation on occasions of sustained heavy rainfall at Treherbert. (Strong low-level baroclinicity is used here to denote 1000-700 mb geostrophic wind shear greater than 7 ms^{-1}). Not quite all the situations could be classified into the above simple categories. A few (referred to as C_o in Table 2) were the result of localised areas of rainfall associated with a strong low-level flow near a cold front at which there was no strong cyclonic shear. The remainder (referred to as W in Table 2) were situations in which heavy rainfall was generated by intense warm frontal rainbands where the 900m winds was less than 20 ms^{-1} and orographic enhancement was negligible.

We now present a few examples of the distribution of heavy orographic rain in relation to the pattern of low-level winds.

(i) Examples in which the heavy orographic rain was associated with warm frontal zones.

Warm frontal zones, either just ahead or just behind their analysed surface position, were found to be one of the commonest locations for heavy rain over the Blaenau Morgannwg. Table 2 shows that strongly baroclinic N-type circulations were alone responsible for almost half the heavy rain and the majority of these circulations were observed within 100 km of a warm front. The three highest sustained rainfall intensities associated with the conveyor belts in the present study were also observed in association with N-type circulations in regions of strong low-level baroclinicity close to warm fronts. Fig 6 depicts one of these three cases, 16 October 1967. Figs 6(a) and 6(b) show the low-level wind structure at 16 GMT and thickness pattern at 12 GMT as the system was crossing England. Figs 6(c) and 6(d) show the corresponding low-level wind structure and adjusted rainfall intensity at orographic sites in the y-t format as explained in Sec 2(b) during the passage of the system across the line Y_1Y_2 in Fig 6(b). A zone of N-type convergence at the surface extended from the surface warm front back into the warm sector ahead of the strongest winds at low-levels as did the region of strong low-level baroclinicity, whilst at 1.5 km the convergence extended 180 km ahead of the surface warm front. Orographic enhancement was large throughout much of the strongly baroclinic zone but it was negligible in regions more than 100 km ahead of the surface warm front where the precipitation grown during baroclinic ascent fell into weak flows (<15 ms^{-1}) at levels below 750 m.

Close to the surface warm front at Treherbert the adjusted rainfall intensity reached 17 mm h^{-1} (Fig 6d). The limited spatial extent of this area is consistent with the view that it originated from middle-level convection perhaps stimulated by orographic ascent (eg Browning et al 1974). In fact convection aloft was observed by radar close to this region. Similar localised areas of high orographic rainfall occurred on several other occasions but only on one occasion was it not associated with a surface front: this occurred on 23 March 1968 (not illustrated) within a region of otherwise moderate orographic rainfall over the Blaenau Morgannwg more than 300 km from the nearest surface front.

In the previous example the low-level flow ahead of the warm front was rather weak and orographic enhancement far ahead of the surface front was correspondingly small. Figs 7(a) and (b), however, show that on 27 February 1967 strong low-level winds extended a long way ahead of the warm front. A C-type zone of low-level convergence was associated with the cyclonic shear within the boundary layer just ahead of the surface warm front. Also precipitation falling from above the warm frontal zone helped maintain the moisture supply to the jet core within regions of low-level divergence farther ahead of the surface warm front. As a result orographic rainfall in excess of 6 mm h^{-1} was maintained at Treherbert for more than 4 hours ahead of the warm front (Fig 7(d)).

(ii) Examples in which the heavy orographic rain occurred within warm sectors in regions far from surface fronts.

High rainfall rates at orographic sites are not confined to regions of strong low-level baroclinicity close to surface fronts. Take the case of 17 December 1965, for example. Figs 8(a) and (b), respectively, show the analysis of the low-level wind field in x-y format at 16 GMT on that occasion and the thickness pattern at 12 GMT. The y-t presentations, shown in Figs 8(c) and (d), lie wholly within the warm sector in a region of weak low-level baroclinicity. An N-type zone of convergence was associated with rainfall over the Blaenau Morgannwg with a typical intensity of 5 mm h^{-1}. Most of the rain fell more than 275 km from the nearest surface front.

(iii) An example in which heavy orographic rain occurred along a cold front.

Rain heavier than 10 mm h^{-1} at orographic sites was reported in BPH (1975) in association with C-type circulations just ahead of cold fronts on 29 October 1972 and 9 November 1972. Neither of these cases met the selection criterion of the present study. Instead the occasion of 5/6 August 1973 has been chosen to portray this kind of situation. Fig 9 shows that to the north of Treherbert on this occasion there was a band of very heavy rainfall along the cold front associated with a C-type convergence zone.

(iv) An example in which a low-level jet was not associated with much heavy orographic rain

Although low-level jets in excess of 25 ms^{-1} were observed in almost all the case studies, heavy orographic rain was not always observed across the whole of the jet core. As noted before it is necessary in addition for the low-level flow to be saturated and for seeding particles to be available. An example in which these conditions were not fulfilled is illustrated in Fig 10. Here the main ascent was ahead of the low-level jet and there was no low-level convergence behind it, ahead of the cold front. The low-level jet was fed by initially unsaturated air and, despite it being the strongest jet observed in this study (jet maximum 43 m s^{-1}), there was scarcely any heavy rain directly associated with it.

(v) Examples of the distribution of heavy rainfall associated with complex conveyor belts.

Most of the above examples were parts of systems in which the conveyor belt was characterised by more than one low-level jet. To illustrate the complexity of the rainfall systems which have been considered here, two complete systems are included in Figs 11 and 12. In the first, Fig 11, most of the heavy rain fell at distances in excess of 100 km from the surface fronts. This applied particularly in central Wales where continuous heavy rain fell from 500 km in advance of the cold front to the front itself, in association with the N-type convergence zones which preceded the two main low-level jets. Along the northern portion of the cold front surface observations indicated negligible convergence and this may have contributed to the low orographic enhancement ahead of the front in this region. In Fig 12 an extremely complex rainfall distribution was observed in association with multiple low-level jets. The initial warm front was characterised by a circulation similar to that in Fig 7 whilst the succession of low-level jets located on the warm side of the quasistationary front maintained a region of almost continuous N-type convergence within the warm sector close to the front. Much of the heavy orographic rainfall within the warm sector was located within 100 km of this front except that associated with the N-type convergence zone which preceded the extensive area of strong winds ahead of the final cold front.

(f) Location of heavy orographic rain with respect to surface fronts.

The distribution of heavy rainfall at Treherbert with respect to the surface fronts, in Fig 13(a), shows that the most likely location of heavy orographic rain is within the warm sector close to the surface front; it can, however, occur as far as 300 km from the surface front. This is in agreement with the previous observations of Douglas and Glasspoole (1947). According to Fig 13(b) heavy rain associated specifically with C-type circulation is most likely to occur close to a surface front. Heavy rain events are more frequently associated with N-type situations, however, and these are prone to occur deep within the warm sector as well as within 100 km of the surface fronts.

4 OROGRAPHIC ENHANCEMENT OF RAINFALL OVER THE HILLS AROUND TREHERBERT: COMPARISON OF OBSERVATIONS WITH THEORY

(a) The dependence of orographic enhancement on the coastal rainfall rate and on the low-level wind speed.

In the examples in Sec 3(e), the orographic increment of rainfall over the Blaenau Morgannwg ranged from negligible to as large as 11 mm h^{-1}. A simple view might be that, if all the enhancement occurred within the low-level flow, then the rainfall intensity over the hills would be directly proportional to the rate of generation of liquid water at low-levels, ie, to the strength of the low-level flow. This makes the assumption, however, that a mechanism exists by which the small droplets generated within the orographic feeder cloud are washed out to the surface with equal efficiency in each case. The numerical model of Collier (1975), which was developed for use in forecasting orographic rainfall over complex surface topography, incorporates such an assumption. On the other hand Bader and Roach (1977) have modelled washout of varying efficiency by considering the microphysics of the coalescence between the droplets in the orographic feeder cloud and the larger drops that fall from levels above the feeder cloud. Their results suggested that orographic rainfall in conveyor belts might be predicted to a reasonable first approximation provided the coastal rainfall rates upwind of the hills are used as a measure of the efficiency of seeding of the low-level feeder cloud.

Our observational results are compared with the theoretical Bader-Roach predictions in Fig 14. The solid curves show the orographic increments predicted by the model when a 3 km layer of initially just saturated air of $\theta_w = 12^{\circ}C$ flows parallel to a hill of similar dimensions to the Blaenau Morgannwg. These curves suggest that, for coastal rainfall rate less than 2 mm h^{-1} and low-level winds in excess of 20 ms^{-1}, orographic increments (Treherbert minus coastal rainfall) should be relatively insensitive to the magnitude of the low-level flow and that variations in the coastal rainfall rate should be the dominating influence. Unfortunately, as we shall see, our observations bear this out only to a limited extent.

The individual point values plotted in Fig 14 represent the observed orographic increments ΔR sustained over 2-h periods over the Blaenau Morgannwg. These are subject to uncertainties of $\pm$ 2.5 ms^{-1} in V_{900} and $\pm$ 0.3 mm h^{-1} in orographic increment. The 2-h averaging period was adopted to minimise the influence of short-term fluctuations owing to small-scale convection. All values have been normilised to a θ_w of 12°C, although the resulting adjustments were always smaller than 20%. A few of the values in Fig 14 are bracketed to show that the Blaenau Morgannwg in those cases were in the lee of the hills of southwest England; otherwise all values refer to wind directions for which the Blaenau Morgannwg were fully exposed to the open sea. The orographic increments could be accurately estimated only for situations after 1966 because suitable autographic gauge records from coastal sites were not available at earlier dates. In order to increase the sample, data from about ten other occasions, including those analysed in BPH (1975), have been incorporated in Fig 14.

It is apparent from Fig 14 that situations with observed surface coastal rainfall intensity R_o in the range 0.1 to 0.5 mm h^{-1} (Category 0) generally have much smaller values of ΔR ($\leqslant$4 mm h^{-1}) than those with higher R_o. There is, however, no clear evidence of an increase in ΔR as R_o increases further. (The situations in which the flow was incident on the Blaenau Morgannwg after passing over the hills of southwest England (bracketed values in Fig 14) were associated with unusually large ΔR and these will be considered later).

A linear regression of ΔR on V_{900} for all the (unbracketed) observations with R_o > 0.5 mm h^{-1} and $V_{900} \geqslant$ 20 m s^{-1} yields the relationship:

$$\Delta R = 0.12\ V_{900} + 1.9 \qquad \ldots\ldots\ldots\ldots (1)$$

where ΔR is measured in mm h^{-1} and V_{900} in ms^{-1}. The correlation coefficient is 0.63. Deviations of the observed ΔR from this regression curve are plotted in Fig 15 for different categories of R_o. These indicate that the magnitude of ΔR, instead of increasing as we might have expected from Bader and Roach's calculations, actually decreased slightly as R_o increased above 2 mm h^{-1}.

We can summarize these findings as follows:

(i) No sustained orographic enhancement occurs without pre-existing surface rain at the coast; a surface coastal rainfall rate in excess of about 0.5 mm h^{-1} is required if high orographic enhancement is to be produced.

(ii) If the low-level wind and coastal rainfall exceed certain threshold values ($V_{900} \approx 20$ ms^{-1}, $R_o \approx 0.5$ mm h^{-1}), the estimate of the orographic increment can be improved slightly if some dependence on the low-level wind speed is incorporated.

(iii) When the surface coastal rainfall rate exceeds 0.5 mm h^{-1}, no further significant improvement in the estimate of orographic increment can be gained by using the surface coastal rainfall rate as an indicator.

(b) Other factors influencing the orographic enhancement of rainfall.

We have shown that the variance in the orographic increment cannot be accounted for solely in terms of the surface rainfall intensity at the coast and the low-level wind velocity. At least three other factors are important:

(i) Evaporation or condensation within the low-level flow

The results of Bader and Roach's calculations can be applied only if the rate of seeding at the top of the orographic feeder cloud is equal to the coastal rainfall rate at the surface. Unfortunately this situation does not pertain when, as a result of descent in the lee of the hills of southwest England or merely advection of dry air, there is significant evaporation at low levels upwind of the Blaenau Morgannwg. Clearly a higher input of precipitation particles from aloft is required to keep the low-level flow moist and to maintain a given surface rainfall rate at the coast when there is evaporation in the layers closest to the surface than when there is condensation which is generating cloud at low levels over the sea. Thus, <u>for a given surface rainfall rate (R_o) at the coast</u>, the implied heavier seeding from aloft when there is low-level evaporation may lead to a larger orographic increment (ΔR) in such a situation. Several situations of unusually large ΔR

occurred when the low-level flow had first passed over the hills of southwest England, (see bracketed values in Fig 14). These situations were also associated with strong baroclinic ascent, the precipitation from which may have helped to regenerate an almost saturated low-level flow by the time it reached the Blaenau Morgannwg. Situations suspected of producing large ΔR when there was significant low-level evaporation were mainly of type C. By contrast, many of the N-type situations studied here are believed to have been associated with condensation in the lowest layers. In these cases ΔR was generally close to or slightly lower than the values given by eq (1). All the situations with $R_o > 3$ mm h^{-1} in Fig 15 were of this kind.

So far we have considered situations in which the air at low-levels may have been dry enough to decrease the surface rainfall rate R_o at the coast yet not dry enough to prevent an orographic feeder cloud from forming at all. Assuming the airflow to be parallel to the surface topography, hardly any orographic feeder cloud can form over the Blaenau Morgannwg if the relative humidity at low levels is less than 80% and in these circumstances there can be little orographic increment. Some of the low values of orographic enhancement in Fig 15 with R_o between 0.5 and 3 mm h^{-1} were associated with regions well ahead of surface warm fronts where despite precipitation descending from aloft there was a very dry layer persisting within the low-level flow. Thus the difference in the surface rainfall intensity over the hills from that upwind over the coast is seen to depend rather critically on the moisture balance at low-levels upwind of the hills: there are circumstances in which evaporation at low levels may suppress the surface rainfall rate only at the coast but there are of course also circumstances in which the humidity at low levels is sufficiently low to inhibit the formation of a feeder cloud and hence to suppress the orographic rain as well. What is needed to distinguish between these two situations is some measure upwind of the hills of both the precipitation intensity aloft (say at 3 km) and of the relative humidity at low levels. We shall return to this problem again in Section 5.

(ii) Triggering of potential instability at middle-levels by orographic lifting.
A small raindrop with a fallspeed of 3 ms^{-1} initially at 3 km will reach the surface about 25 km downwind in a strong low-level flow. Ice crystals from higher levels will fall more slowly and will be carried even farther downstream. Thus, measurements of rainfall rates at the coast will give little indication of the presence of orographically triggered middle-level convection unless as sometimes happens it is initiated a considerable distance out to sea (cf Browning et al, 1974). In other words the low-level flow over the hills may be efficiently seeded by convection without any of the seeding particles reaching the surface at the coast. When this happens the orographic enhancement will be much larger than might have been expected on the basis of the coastal rainfall rate. Two cases when this occurred are incorporated in Fig 15; the first was responsible for the largest orographic increment plotted in Category 1 and the second (on 16 October 1967; Fig 6) was responsible for the largest increment plotted in Category 4+.

(iii) Variations in the degree of orographic uplift.
Bader and Roach (1977) assumed that the air trajectories follow the outline of the hills. In practice, however, the vertical extent and magnitude of the vertical motions induced over the hills will vary with the static stability and with the magnitude and vertical structure of the wind (eg, Sawyer 1956). The fact that there is little orographic enhancement ahead of many surface warm fronts, even in regions of fairly strong moist low-level flow, may also be due to the strong shear and veer of the wind at very low levels, as a result of which the winds in the layers closest to the surface may be much weaker and from a direction unsuitable for orographic lifting on the seaward slopes of the hills (surface winds are occasionally backed by 60^{o} with respect to those at 900m ahead of the surface warm front). Some of the relatively low orographic increments plotted in Fig 15 in Categories 2 and 3 were observed in such circumstances.

(c) Location of the orographic rainfall maximum
When the wind direction at low levels is approximately constant throughout the passage of a rain system, the location of the region of maximum orographic rainfall

on the Blaenau Morgannwg can be estimated with some precision from 24-h raingauge totals. Fig 16(a) shows the rainfall distribution in sections across the Blaenau Morgannwg for 7 such cases (5 from the present studies; 2 from BPH, 1975). These cases have been selected to represent as nearly as possible the complete range of conditions producing heavy orographic rainfall. The values of rainfall plotted here are averages across a band 10 km wide, the sections being taken along the mean direction of the 700 mb wind, (this direction is a good approximation to the direction of travel of mesoscale precipitation areas, often the dominant source of seeding of the low-level feeder cloud). The maximum orographic rainfall was observed close to the crest of the hills (Fig 16b) for these and most other warm sector and warm frontal conveyor belts incident directly from the open sea. Altogether there were only two exceptions to this rule: one was on 9 August 1971 (not shown in Fig 16) and the other was on 29 July 1967. In these cases the peak rainfall was located upwind from the hill top. Now, these two cases also happen to be the only situations associated with weak low-level flows ($V_{900} < 20\ ms^{-1}$); at the same time the coastal rainfall was also heavy. This suggests that the anomalous rainfall distribution on these occasions was related to the high efficiency and rapidity of the washout. In most of the cases, however, Fig 16 suggests that liquid water was being advected up the hill within the feeder cloud, as in the Bader-Roach model, to produce a maximum liquid water content in the cloud close to the hill top and a pronounced peak in orographic enhancement near that region.

5 SHORT PERIOD FORECASTING OF HEAVY RAINFALL OVER EXPOSED OROGRAPHIC SITES

To obtain quantitative forecasts of the rainfall from warm frontal/warm sector conveyor belts for exposed sites such as the Blaenau Morgannwg it is necessary to predict:

(i) the extent and intensity of the background (ie, non-orographic) rainfall associated with the baroclinic ascent within the synoptic scale weather system, and

(ii) the occurrence and movement of mesoscale low-level jet circulations and the patterns of low-level wind velocity and humidity.

The frequent data from a geosynchronous satellite combined with predictions from a synoptic scale model (eg Bushby and Timpson 1967) should permit adequate forecasts of the intensity and motion of the synoptic scale system. However, as can be seen for example in Fig 12, the mesoscale features are often very complex, and it seems unlikely that inputs to a full primitive-equation numerical-dynamical model can be obtained on the mesoscale, at least in the near future. We therefore believe that the most practical approach for short period forecasting will be to wait until these systems come within range of land-based radars and then observe their mesoscale structure directly as they approach the forecast area. Radars can be used to identify many of the main ingredients of orographic rain, such as:

- the intensity of the rain while it is over the sea,
- the velocity of the mesoscale precipitation areas likely to provide the seeding particles for the orographic rain,
- the velocity of the low-level wind within the areas of precipitation, assuming the radar is designed to measure the Doppler shift.

To obtain useful warnings, however, the radars must be well sited. Fig 17 illustrates how, from a technical point of view, two radars might ideally be located for the purpose of forecasting for Wales and southwest England. Specially designed pulsed Doppler radars, with good horizons towards the southwest, could identify the existence and motion of areas of precipitation and also identify the low-level wind velocity over large areas where there is precipitation. With the radar configuration shown in Fig 17 this could be achieved up to 400 km upwind of the south Wales hills. This would give about 6 hours warning of an approaching system, assuming the average velocity of the systems investigated here. Such measurements would provide much of the information required to fulfil items (i) and (ii) above. These observations, and others to be discussed shortly, could then be used as inputs to simple fine-scale numerical models developed along lines similar to those proposed by Collier (1975, 1977) and Jonas (1976) to predict the orographic enhancement over specific areas of complex topography.

Models such as those of Collier and Jonas provide detailed fields of vertical motion over the hills. It is then necessary to derive estimates of the rate of condensation in the low-level feeder cloud and also the rate of washout, ie, the microphysical efficiency of the conversion of cloud droplets to rain. Collier (1975) makes some very simple assumptions concerning the cloud physics; Jonas (1976) treats the cloud physics in some detail. To a first approximation the washout efficiency can be estimated from the surface rainfall intensity upwind; however, in Sec 4 we showed that the prediction of the orographic enhancement using coastal rainfall rate is subject to considerable error because the surface rainfall is not a reliable indicator of the rate of seeding from aloft. It is possible that observations with a narrow-beam radar scanning at different elevations may enable one to assess the precipitation intensity at an appropriate altitude (3 km, say). Moreover, the vertical gradient of radar reflectivity between 3 km and lower levels will also provide an indication of precipitation growth and evaporation and this can give a crude indication of the relative humidity at low levels and hence of the amount of condensation in the feeder cloud. Thus it is possible in principle for a radar scanning upwind of the hills at more than one elevation angle to go some way toward providing the additional information required as inputs to the fine-scale numerical prediction models. The efficacy of these ideas has yet to be demonstrated in practice, however, and further studies would have to be carried out to test them.

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TABLE 1 Catalogue of days with heavy rainfall

Rainfall day(s)	Dominant synoptic type	Maximum windspeed in low-level jet at 900m (ms^{-1})	Maximum Rainfall in hills of south Wales (mm in 24 hours)	Representative coastal rainfall upwind of south Wales (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

* The 24-hour period during which the rainfall exceeded 85 mm overlapped two so-called 'rainfall days' on these occasions.

TABLE 2

	Classification of the low level circulation		No. of cases	Duration of rainfall (h)	Total rainfall(mm)
Principal classification	Strongly baroclinic cases	C type	7	24	153
		N type	21	76	523
	Weakly baroclinic cases	C type	9	39	195
		N type	8	42	230
Other cases		C_o	3	6	36
		W	3	7	43

Table 2 Classification of mesoscale circulations associated with situations of strong low-level wind which produced more than 2 hours of rainfall at greater than 4 mmh^{-1} at Treherbert. See text for definition of terms. Note that more than one type of low-level circulation can occur in association with the rainfall for a given synoptic situation.

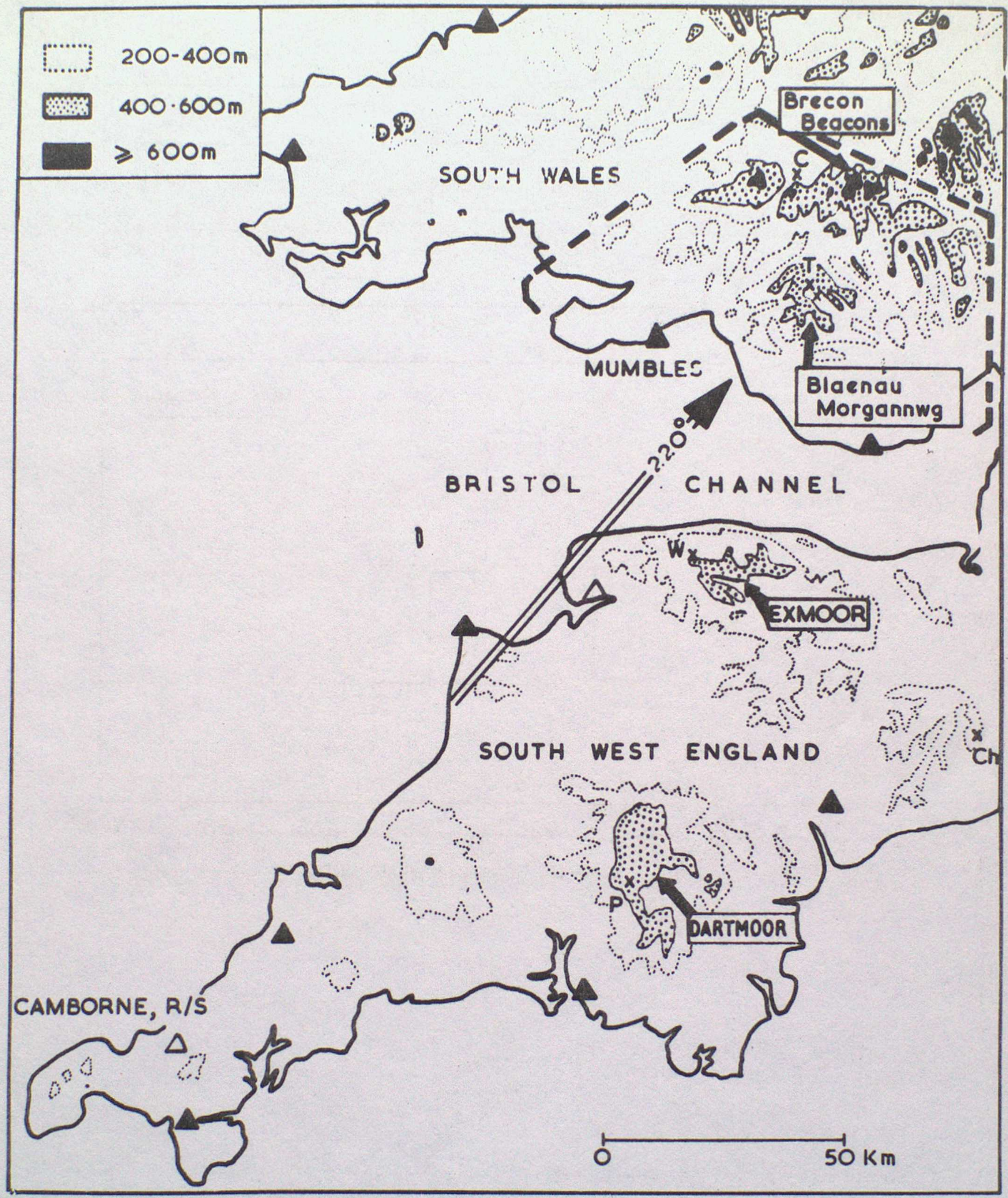


Fig 1 Topography of part of south Wales and southwest England. Dashed lines demarcate the region where the rainfall selection criterion was applied. Orographic raingauge sites which feature prominently in this study are indicated by crosses. They are: T, (Treherbert Park and Treorchy (to the SE)), C (Upper Cray Reservoir); D (Ddolwyn Bridge); Ch (Chard); P (Princetown); W (Woolhanger). The solid triangles represent locations of hourly wind observations.

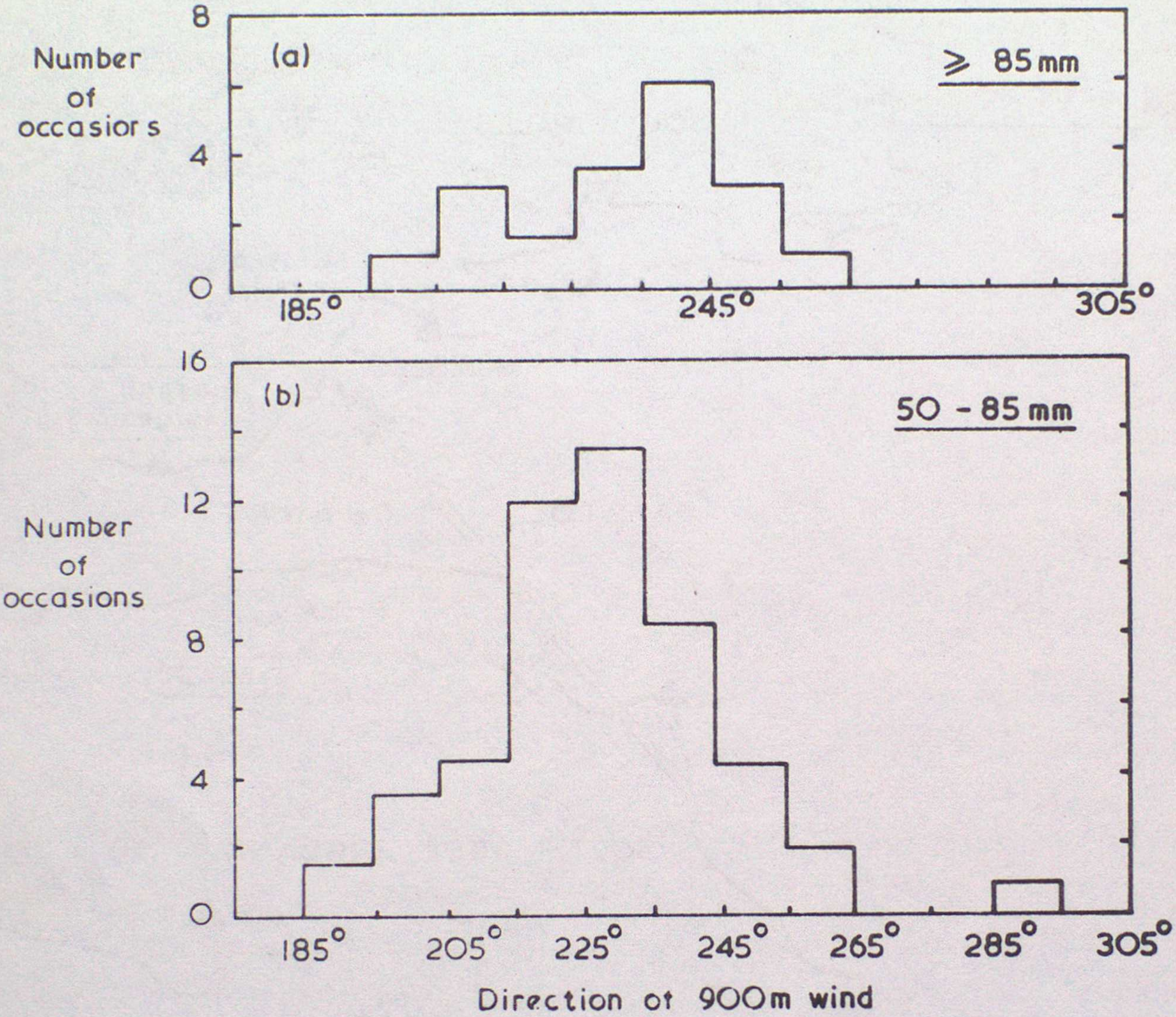


Fig 2 Direction of the 900 m wind associated with heavy orographic precipitation in warm sector/warm frontal conveyor belts over south Wales (a) for the 18 cases analysed in this paper with 24-h orographic rainfall exceeding 85 mm and (b) for 51 further cases with 24-h orographic totals in excess of 50 mm.

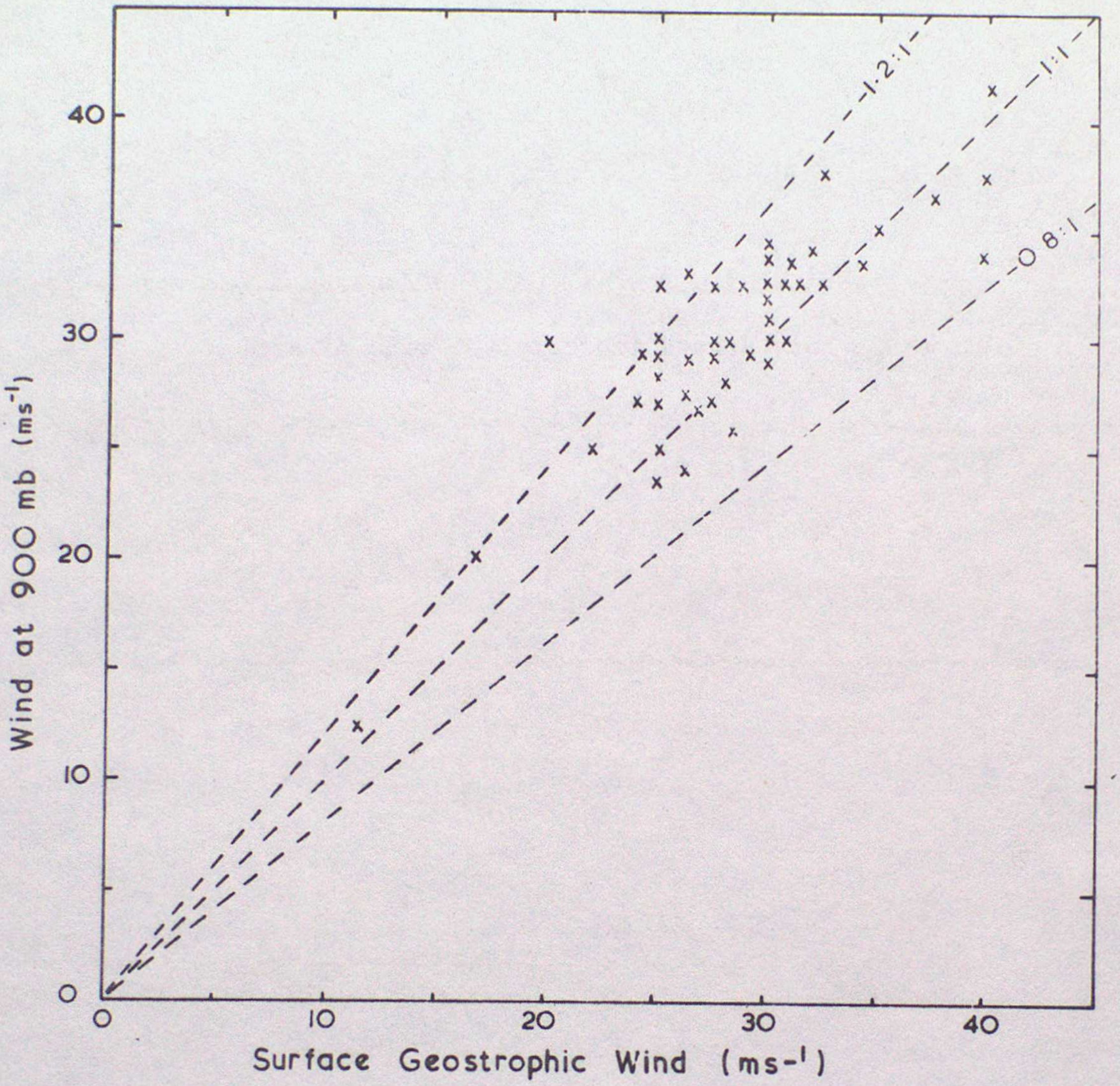


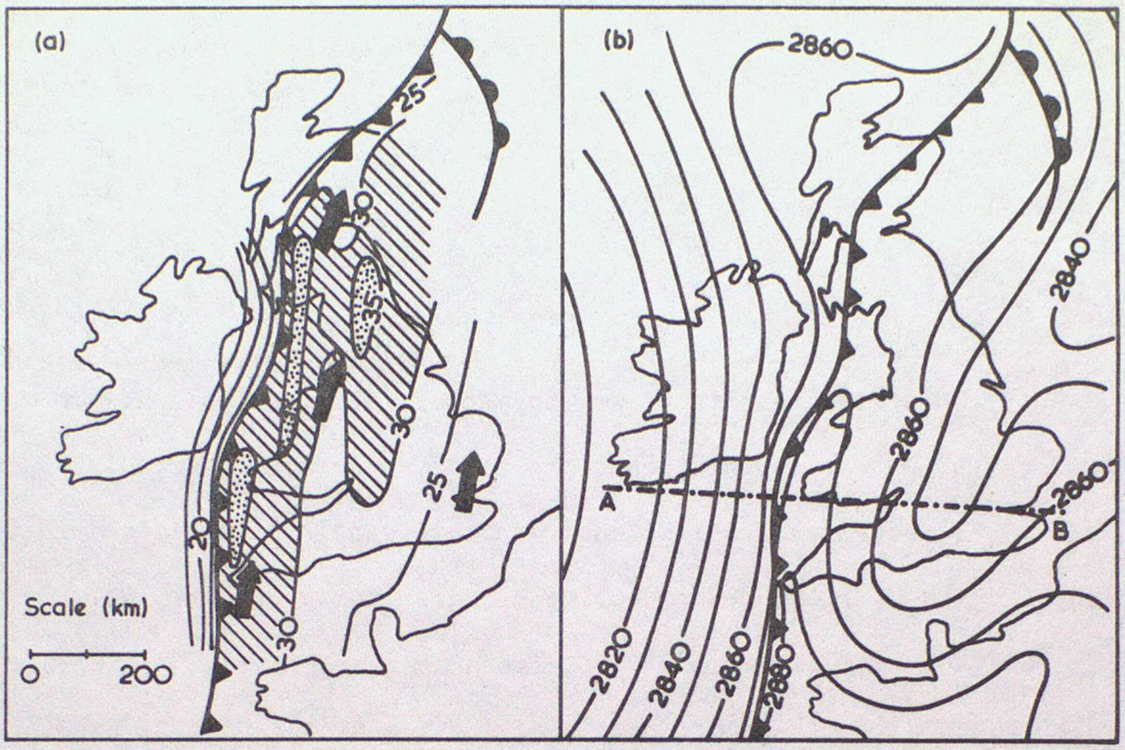
Fig 3 Comparison of the magnitude of the wind speed at 900 m in the warm-sector low-level jets with the magnitude of the surface geostrophic winds.

Figs 4 (a and b)

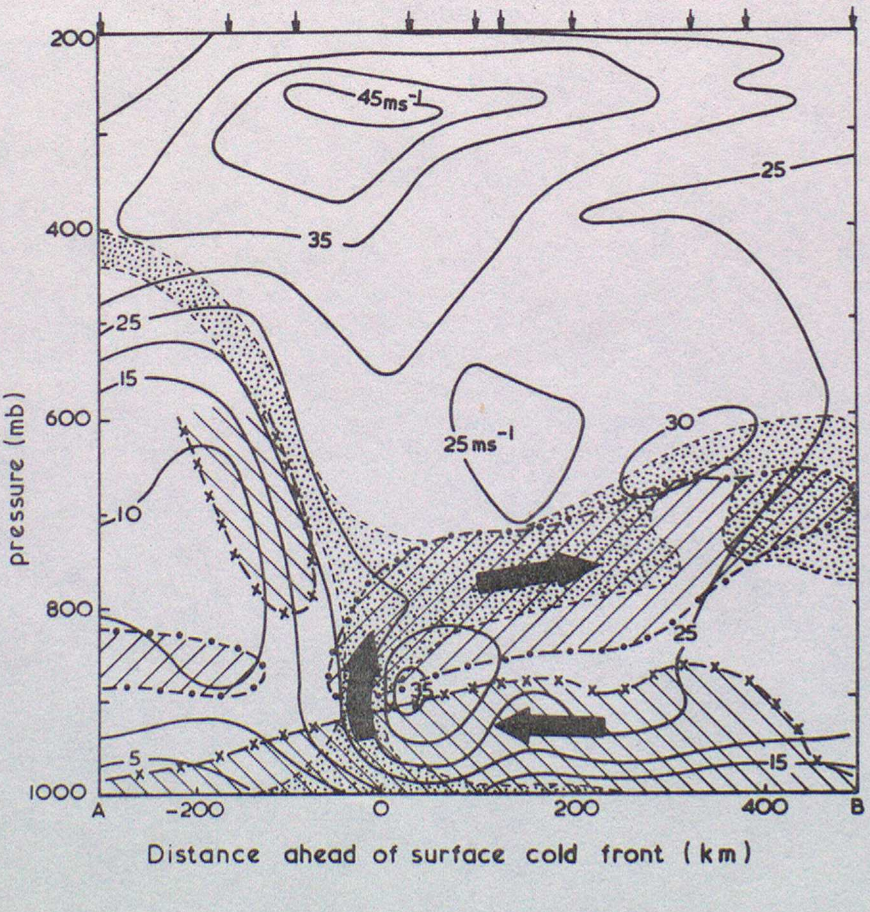
(a) Wind field at 900 m at 12 GMT, 30 January 1974. Isotachs are in $m\ s^{-1}$. Bold arrows indicate the wind direction at 900 m.

(b) 1000-700 mb thickness at 12 GMT, 30 January 1974.

Fig 5 Vertical section normal to the axis of the low-level jet(s) at 12 GMT, 30 January 1974. The solid lines are isotachs of the absolute magnitude of wind speed labelled in $m\ s^{-1}$. The regions below 600 mb in which the transverse velocity relative to the motion of the low-level jet axis exceeded 2 $m\ s^{-1}$ are indicated by hatching and the regions with θ_w between 8 and 9°C are stippled. The bold arrows indicate the sense of the transverse circulation. The positions of soundings are indicated by the small arrows along the top of the figure.



(a)
(b)
25
30
35
30
25
20
30
Scale (km)
0
200
2860
2840
2860
2860
A
B
2820
2840
2860
2880



200
400
600
800
1000
pressure (mb)
45ms⁻¹
35
25
25
15
10
25ms⁻¹
30
25
35
5
15
A
-200
0
200
400
B
Distance ahead of surface cold front (km)

Figs 6, 7 and 8 Distributions in an x-y format of (a) wind speed at 900 m, (b) 1000-700 mb thickness, and distributions in a y-t format of (c) wind speed at 900 m and (d) adjusted rainfall intensity at orographic sites (resolved along $Y_1 Y_2$ in (b)). Figs 6, 7 and 8, respectively, refer to 16 October 1967, 27 February 1967 and 17 December 1965. Isotachs are at 5 m s^{-1} intervals and isohyets are at 3 mm h^{-1} intervals. See Key for explanation of shading. The system velocity is given for each case and this can be used to transform the time co-ordinate in (c) and (d) to an approximate spatial co-ordinate. Arrows on the right side of (d) indicate location of the gauges used in the rainfall analysis; the Treherbert gauge is labelled Tr.

<u>Key to Figs 6 to 12</u>

Wind speed at 900 m:

30 to 35 ms^{-1}

35 to 40 ms^{-1}

40 to 45 ms^{-1}

Adjusted rainfall intensity:

0 to 3 mm h^{-1}

3 to 6 mm h^{-1}

6 to 9 mm h^{-1}

> 9 mm h^{-1}

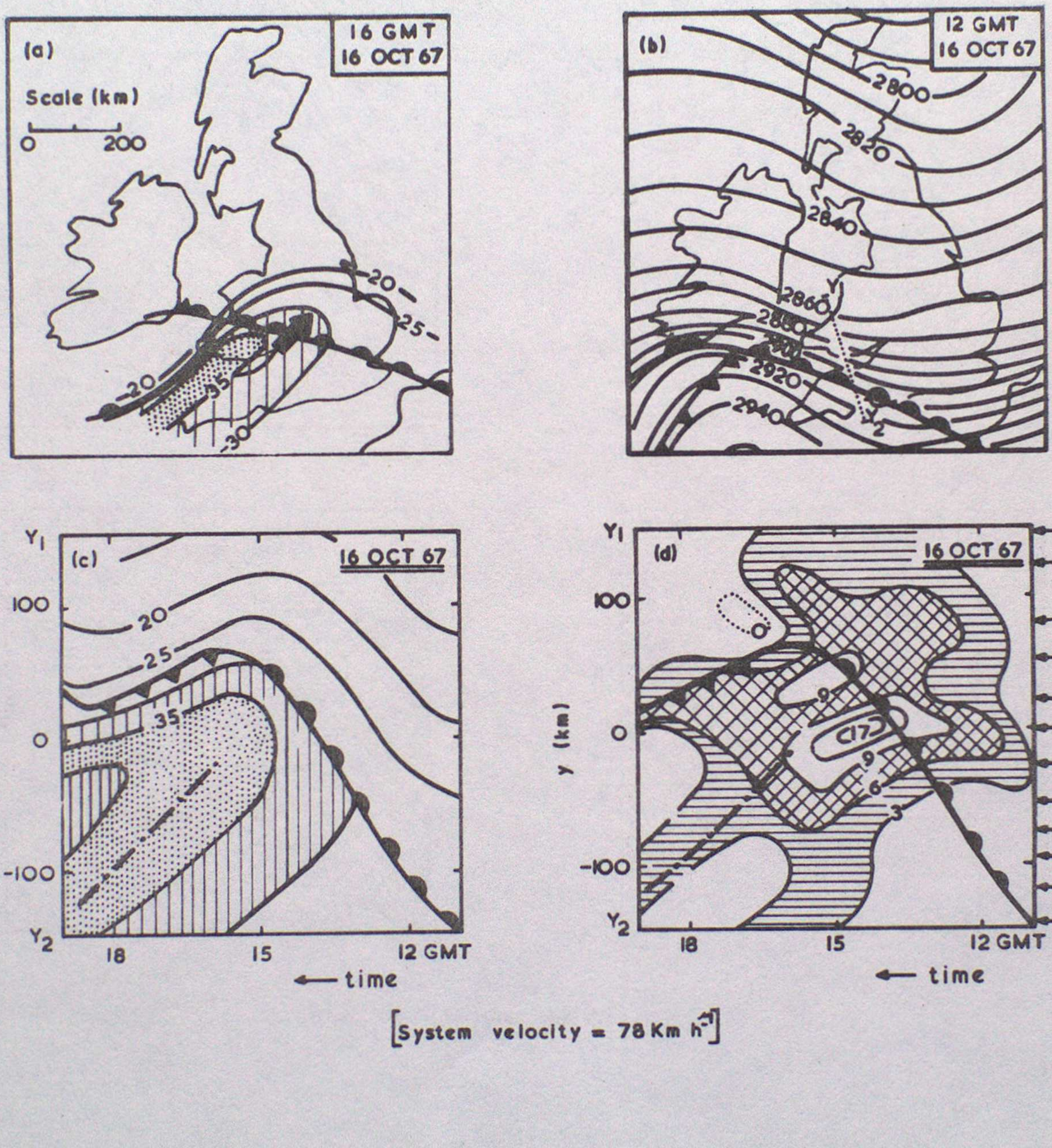

(a)
16 GMT
16 OCT 67
Scale (km)
0
200
20
25
35
30
(b)
12 GMT
16 OCT 67
2800
2820
2840
2860
2880
2900
2920
2940
(c)
16 OCT 67
Y1
100
0
-100
Y2
y [340°] (km)
20
25
35
18
15
12 GMT
← time
(d)
16 OCT 67
y (km)
9
17
6
3
Tr
← time
[System velocity = 78 Km h-1]

Fig 6

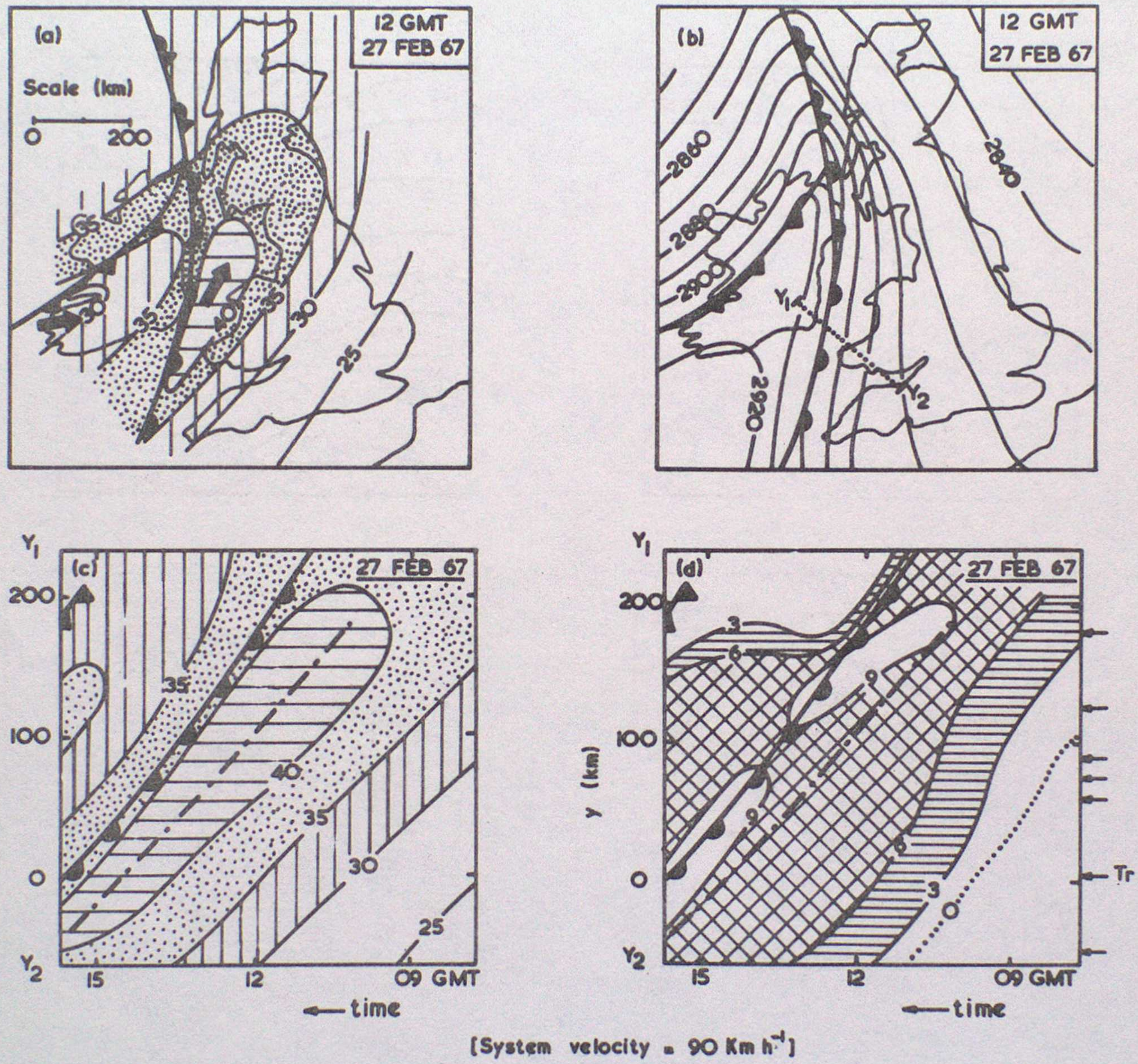

(a)
12 GMT
27 FEB 67
Scale (km)
0
200
35
30
35
40
35
30
25
(b)
12 GMT
27 FEB 67
2860
2880
2900
2920
2840
Y_1
Y_2
(c)
27 FEB 67
200
100
0
y [310°] (km)
35
40
35
30
25
15
12
09 GMT
time
(d)
27 FEB 67
3
6
9
9
6
3
0
Tr
200
100
0
y (km)
15
12
09 GMT
time
[System velocity = 90 Km h⁻¹]

Fig 7

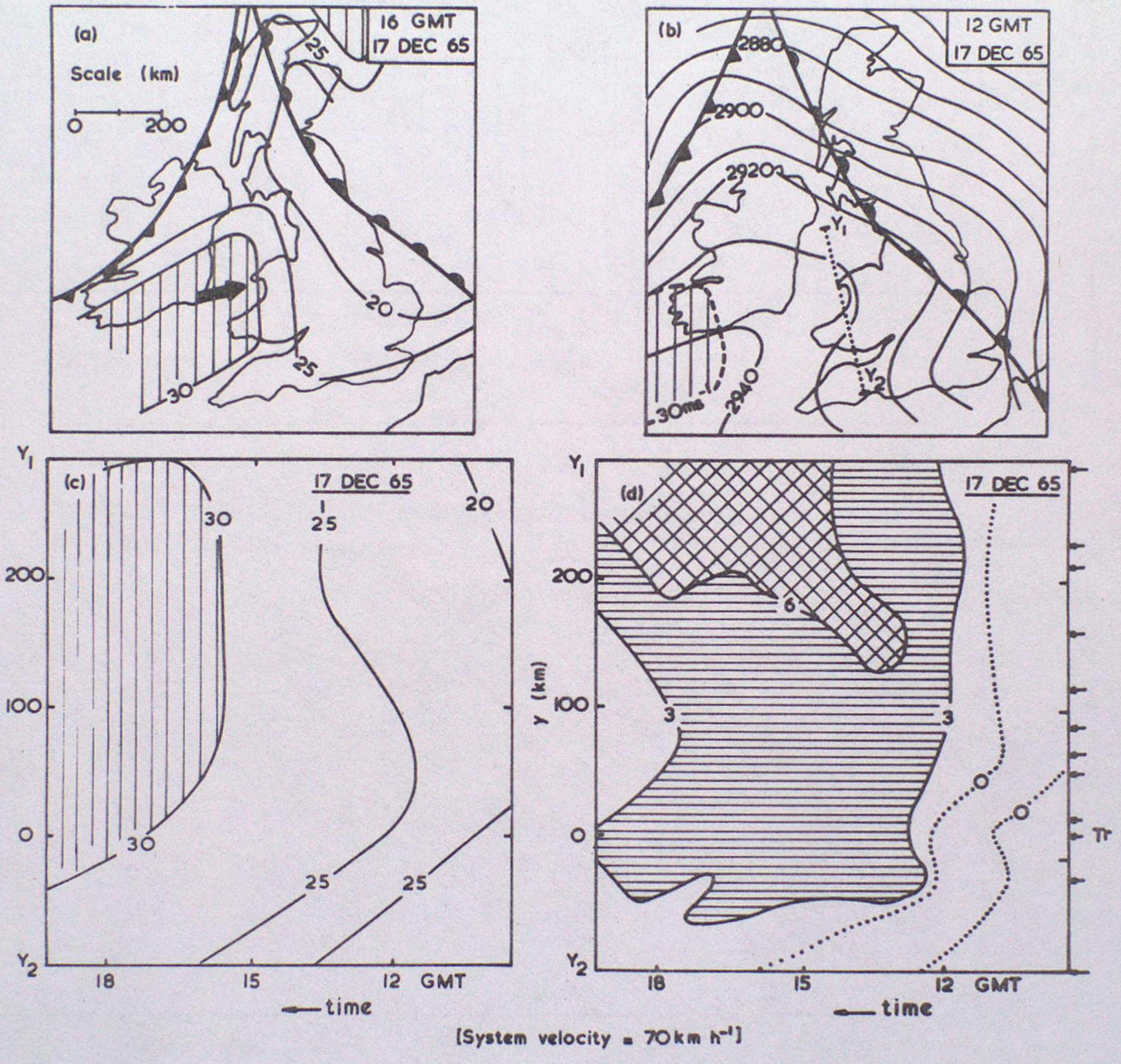

(a)
16 GMT
17 DEC 65
Scale (km)
0
200
25
20
25
30
(b)
12 GMT
17 DEC 65
2880
2900
2920
2940
30ms
Y_1
Y_2
(c)
17 DEC 65
30
25
20
30
25
25
Y_1
Y_2
200
100
0
y [345°] (km)
18
15
12 GMT
time
(d)
17 DEC 65
6
3
3
0
0
Tr
y (km)
18
15
12 GMT
time
[System velocity = 70 km h^{-1}]

Fig 8

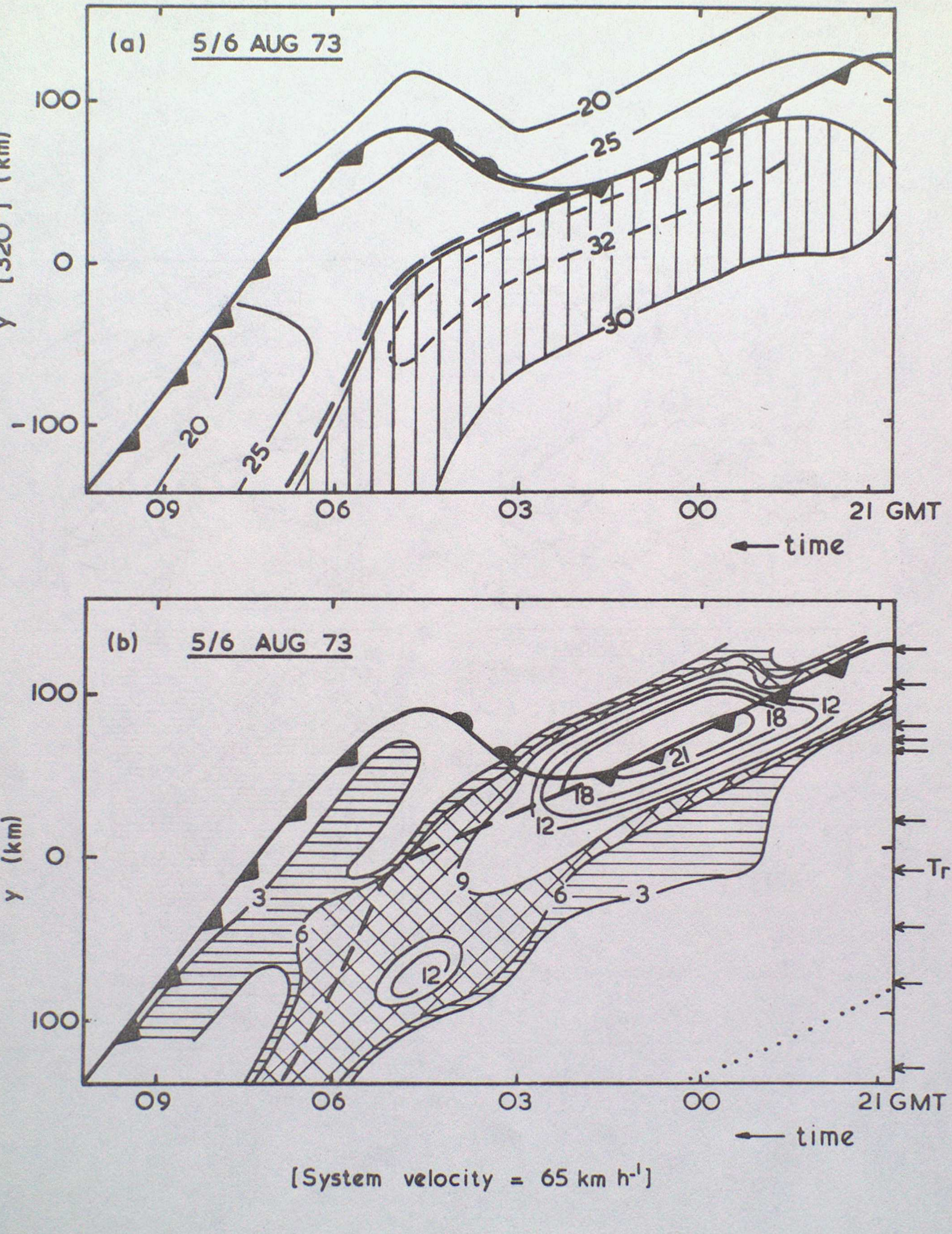


Fig 9 Distribution in a y-t format of (a) wind speed at 900 m and (b) adjusted rainfall intensity at orographic sites for 5/6 August 1973. See Key for explanation of shading. The indicated system velocity can be used to transform the time co-ordinate to an effective spatial co-ordinate. Arrows on the right side of (b) indicate locations of the gauges used in the rainfall analysis; the Treherbert gauge is labelled Tr.

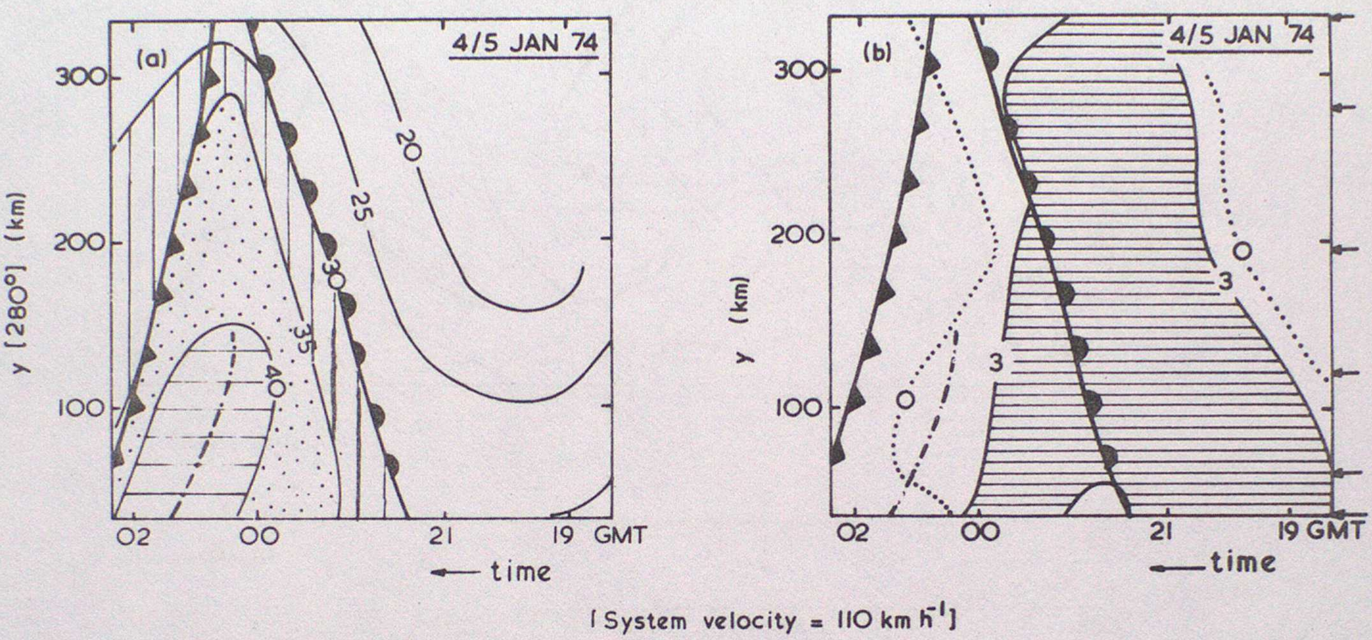


Fig 10 Legend as for Fig 9 but for 4/5 January 1974

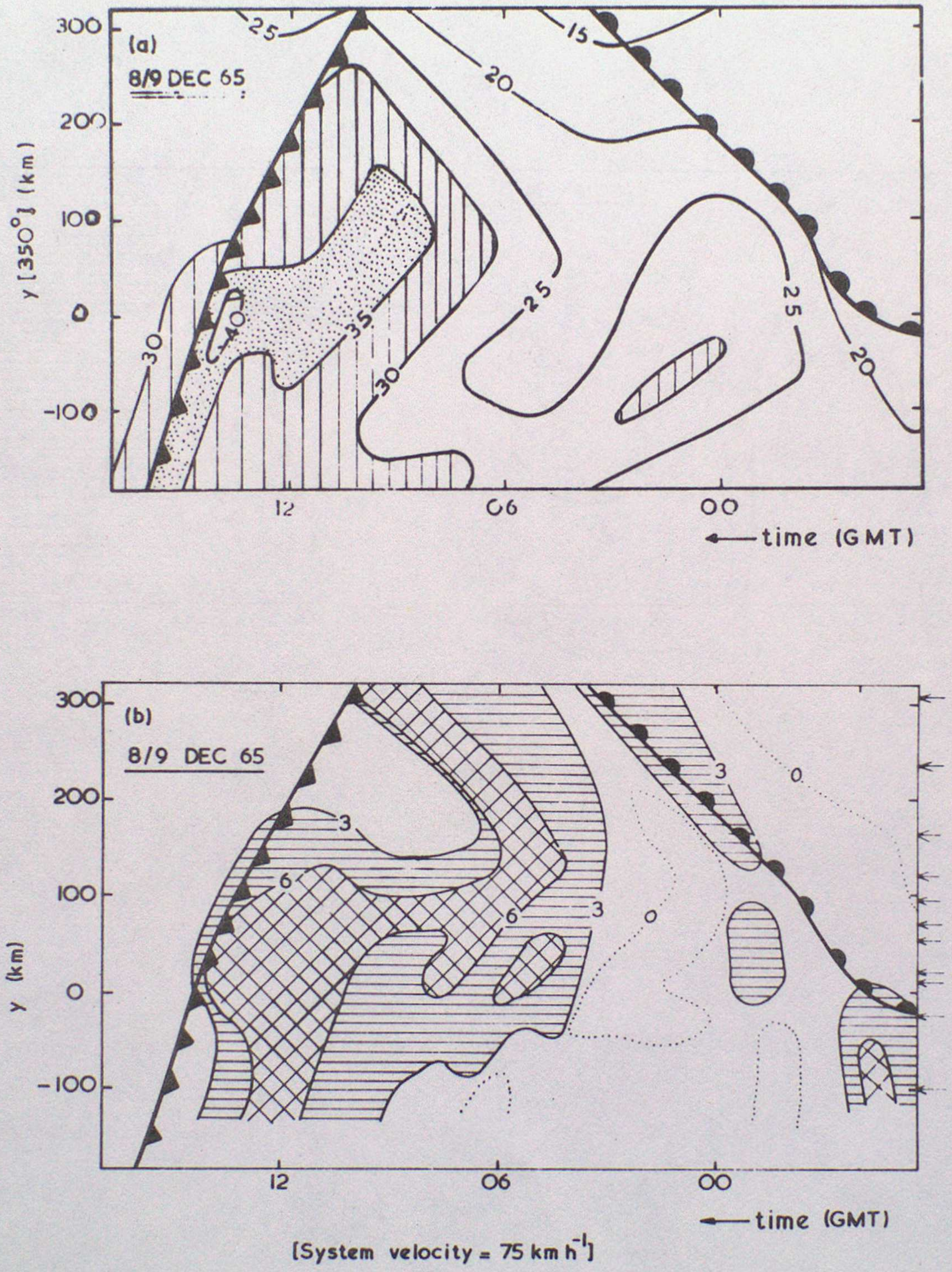


Fig 11 Legend as for Fig 9 but for 8/9 December 1965

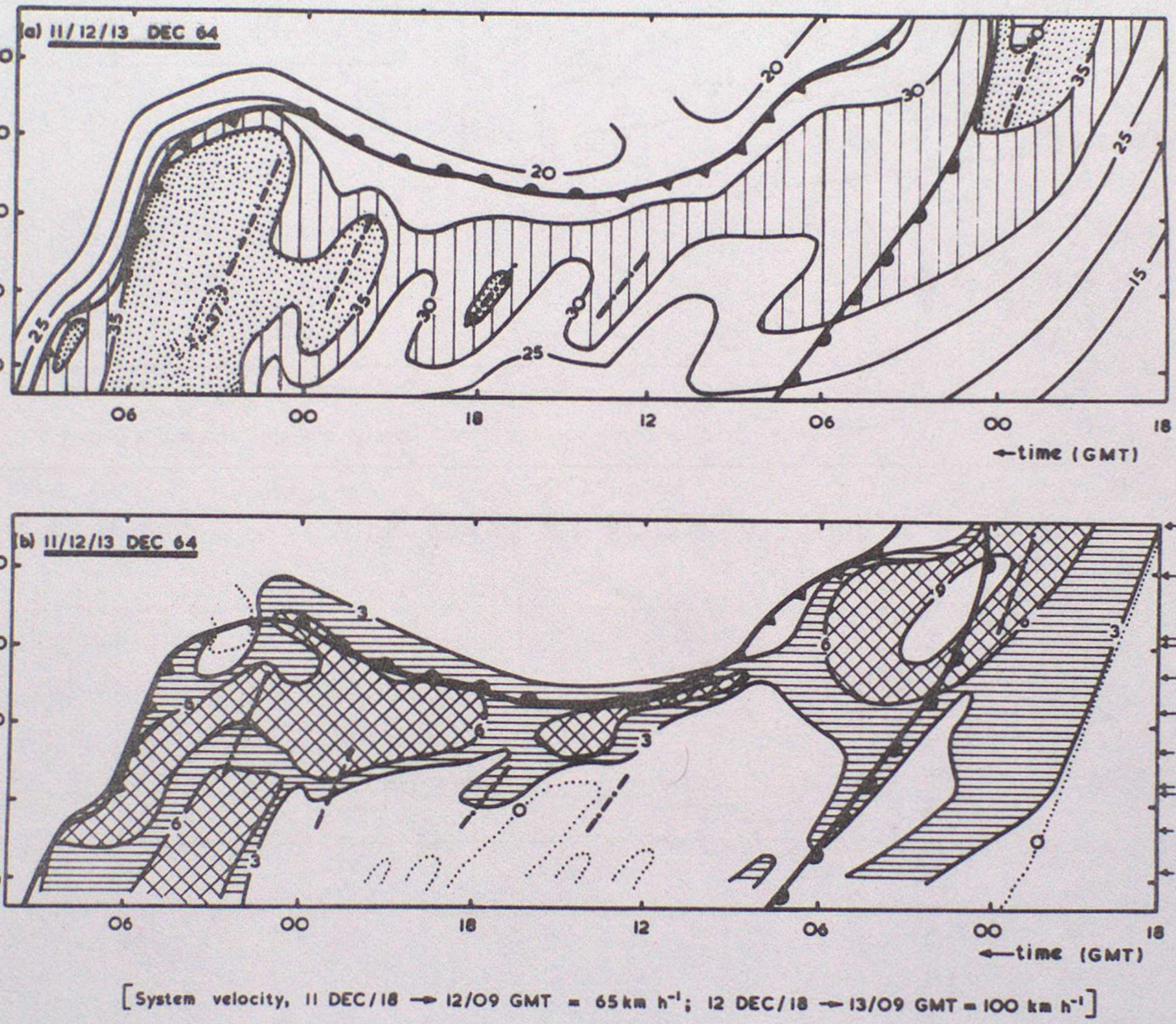
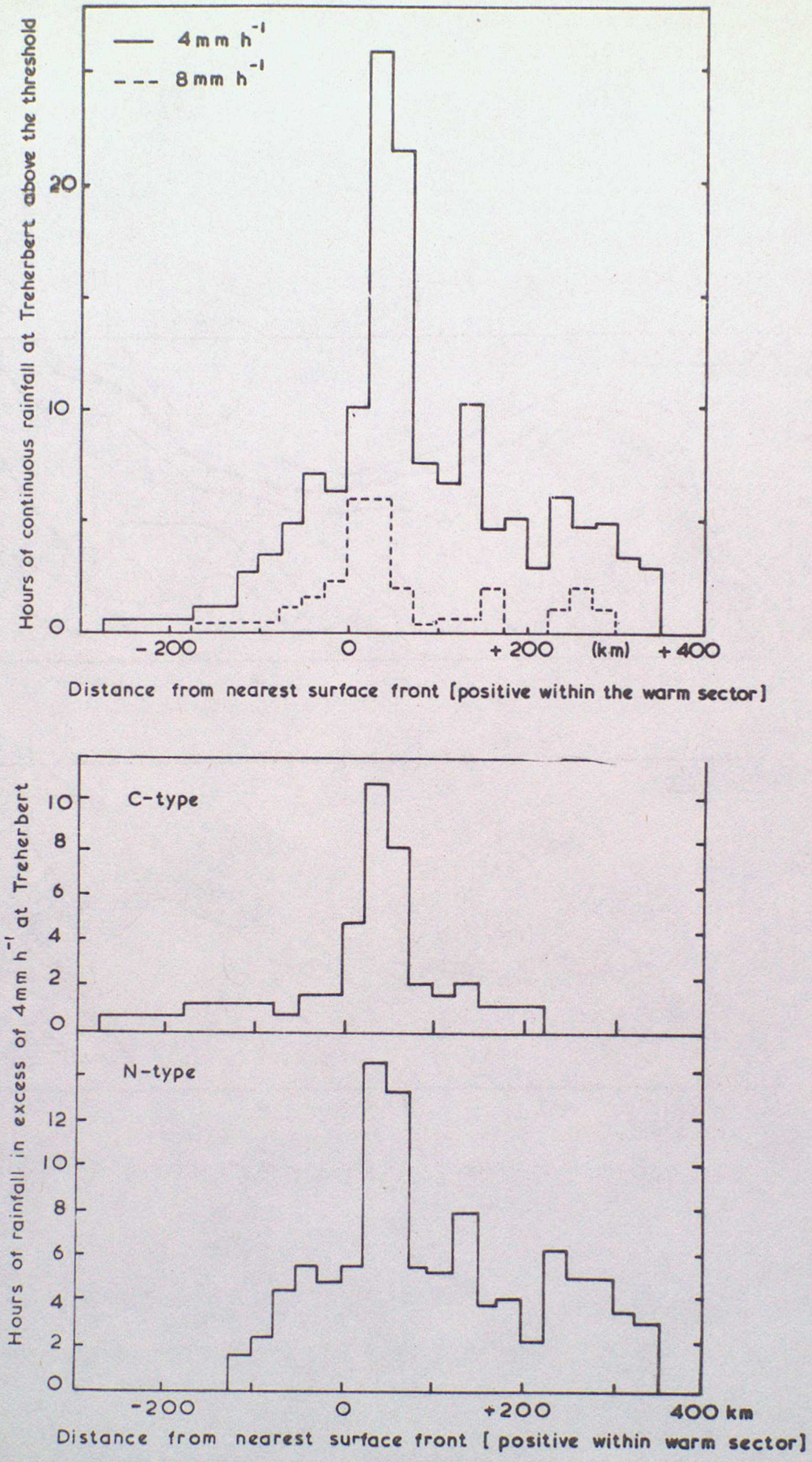


Fig 12 Legend as for Fig 9 but for 11/12/13 December 1964.

(a)



(b)

Fig 13 Distribution of heavy rainfall at Treherbert with respect to distance from the closest surface front for the 18 conveyor belt situations analysed in this study. (a) All situations combined (b) Situations with respect to the two main categories of mesoscale circulation.

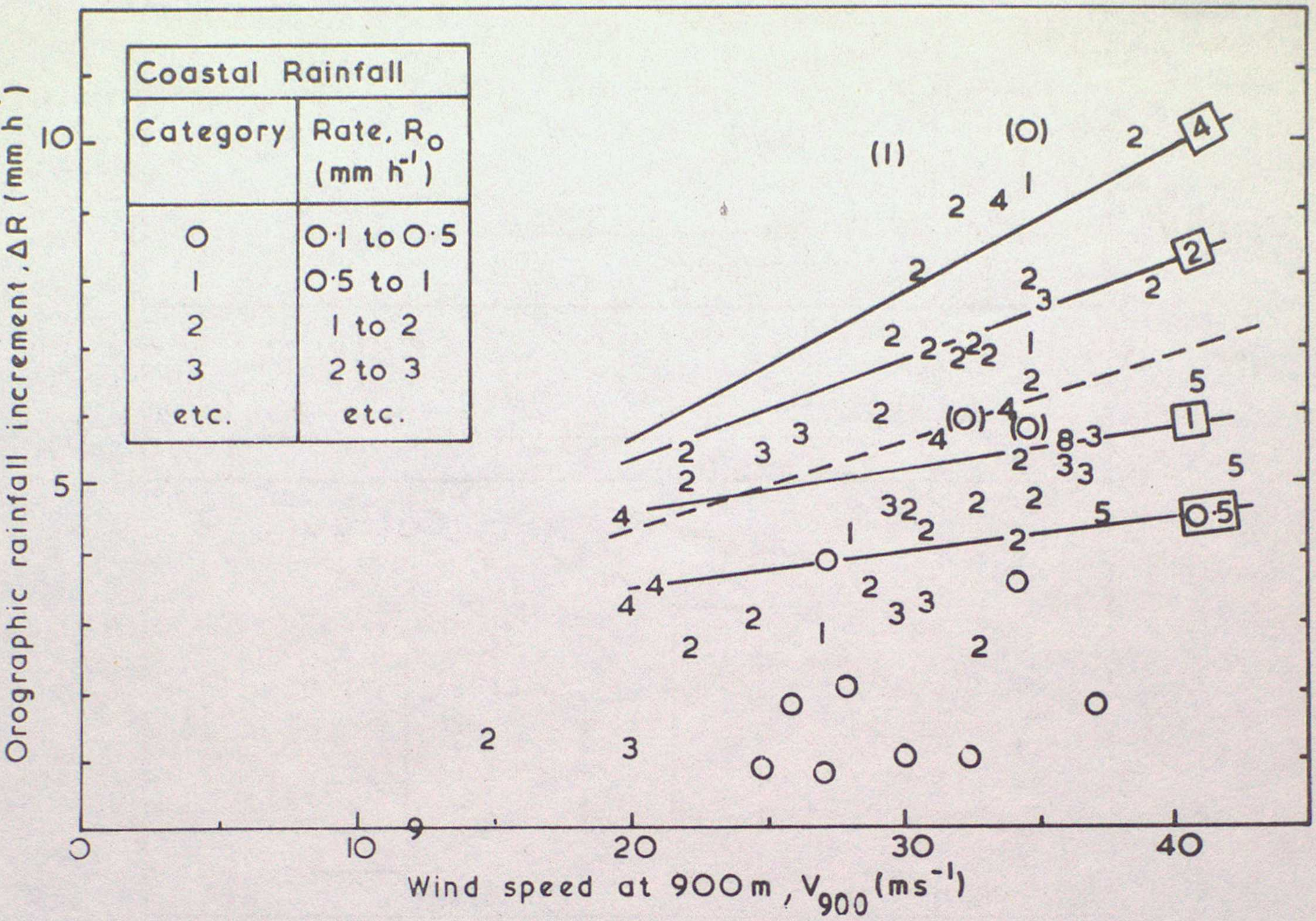


Fig 14 Relationship of the orographic increment (ΔR) at Treherbert (ie rainfall intensity at Treherbert minus the coastal rainfall intensity upwind) to the magnitude of the low-level wind (V_{900}) at Mumbles for different values of the coastal rainfall intensity (R_o). The observational results (point values) are compared with the predictions of Bader and Roach (solid curves). The observational values of ΔR are plotted as single digits (0,1,2,3 etc) according to whether the coastal rainfall rate R_o falls in the ranges 0.1-0.5, 0.5-1, 1-2, 2-3 mm h^{-1} etc. The theoretical curves correspond to different seeding rates at the top of the feeder cloud, as indicated in mm h^{-1} in the square boxes. Assuming negligible low-level growth or evaporation near the coast, the labels in the square boxes can be interpreted as representing the surface coastal rainfall rate R_o. The dashed curve applies to the observational values and represents the linear regression of ΔR on V_{900} for R_o >).5 mm h^{-1} and $V_{900} \geqslant 20$ m s^{-1}.

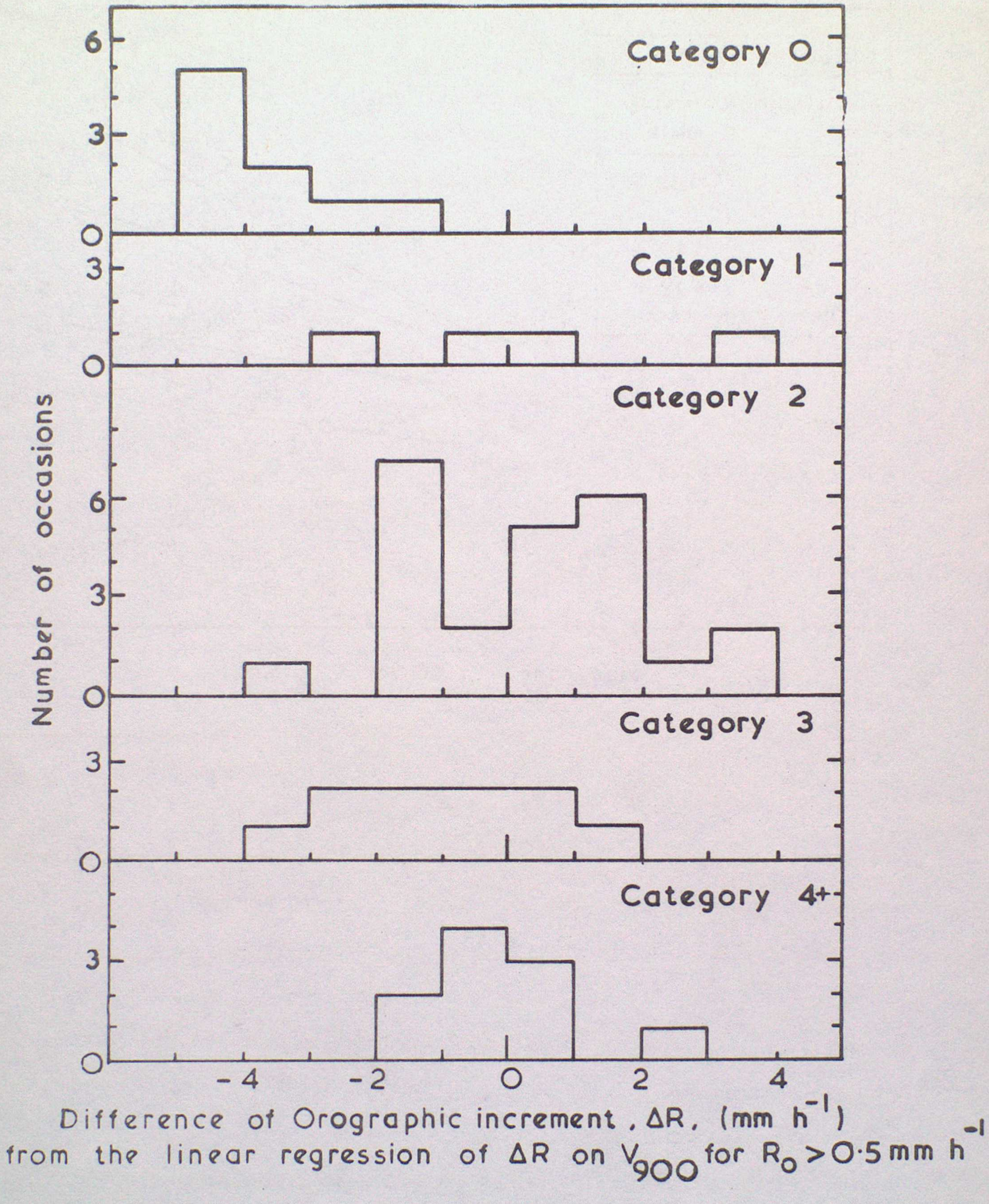


Fig 15 Histograms showing the deviations of the observed orographic increments ΔR at Treherbert from the dashed curve in Fig 14 for 5 different categories of coastal rainfall rate R_o.

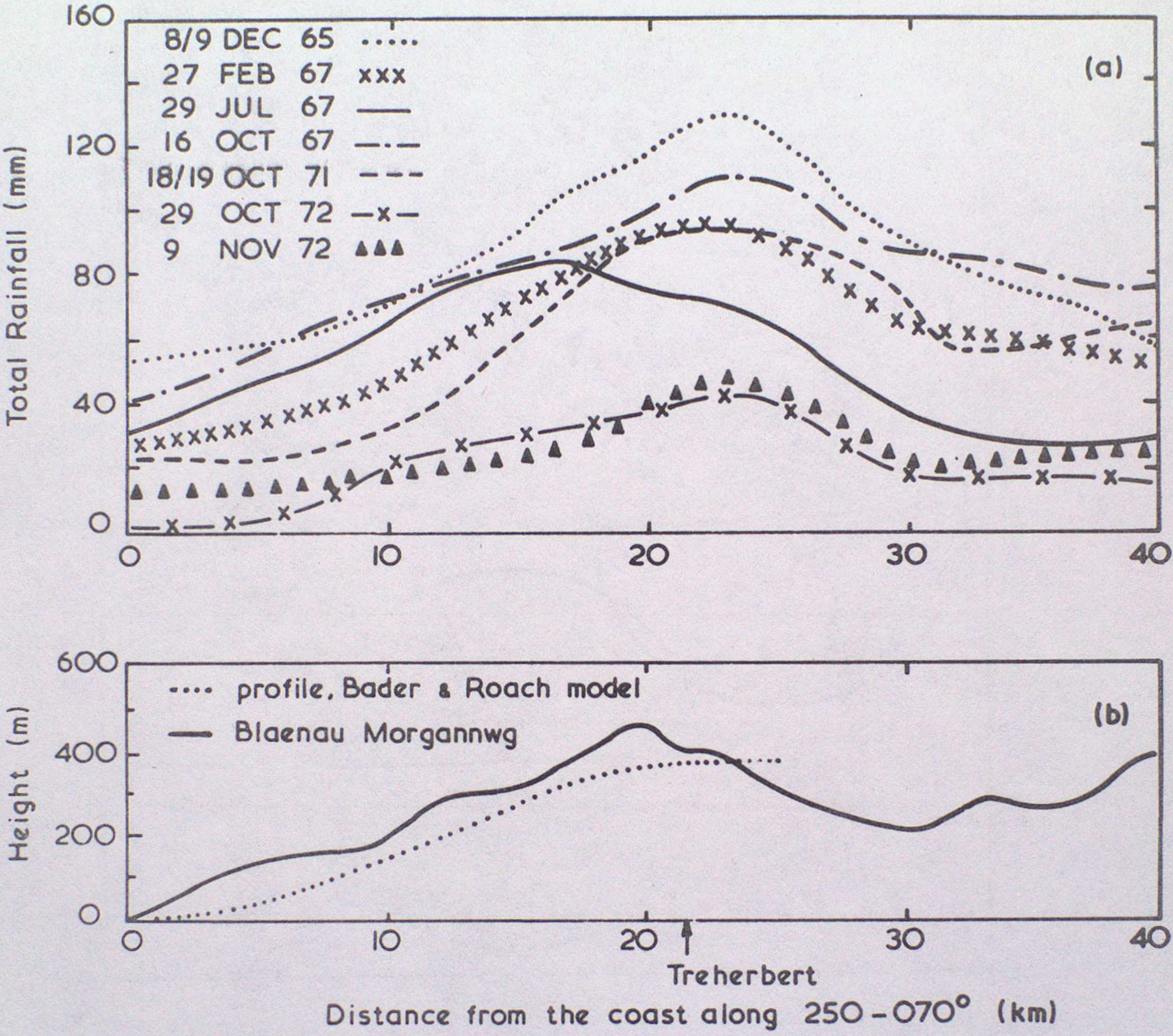


Fig 16 Distribution of orographic rainfall with topography within sections across the Blaenau Morgannwg, as derived from 24-hour raingauge records.

(a) Rainfall profiles for 7 occasions. (b) Hill profile along 250°.

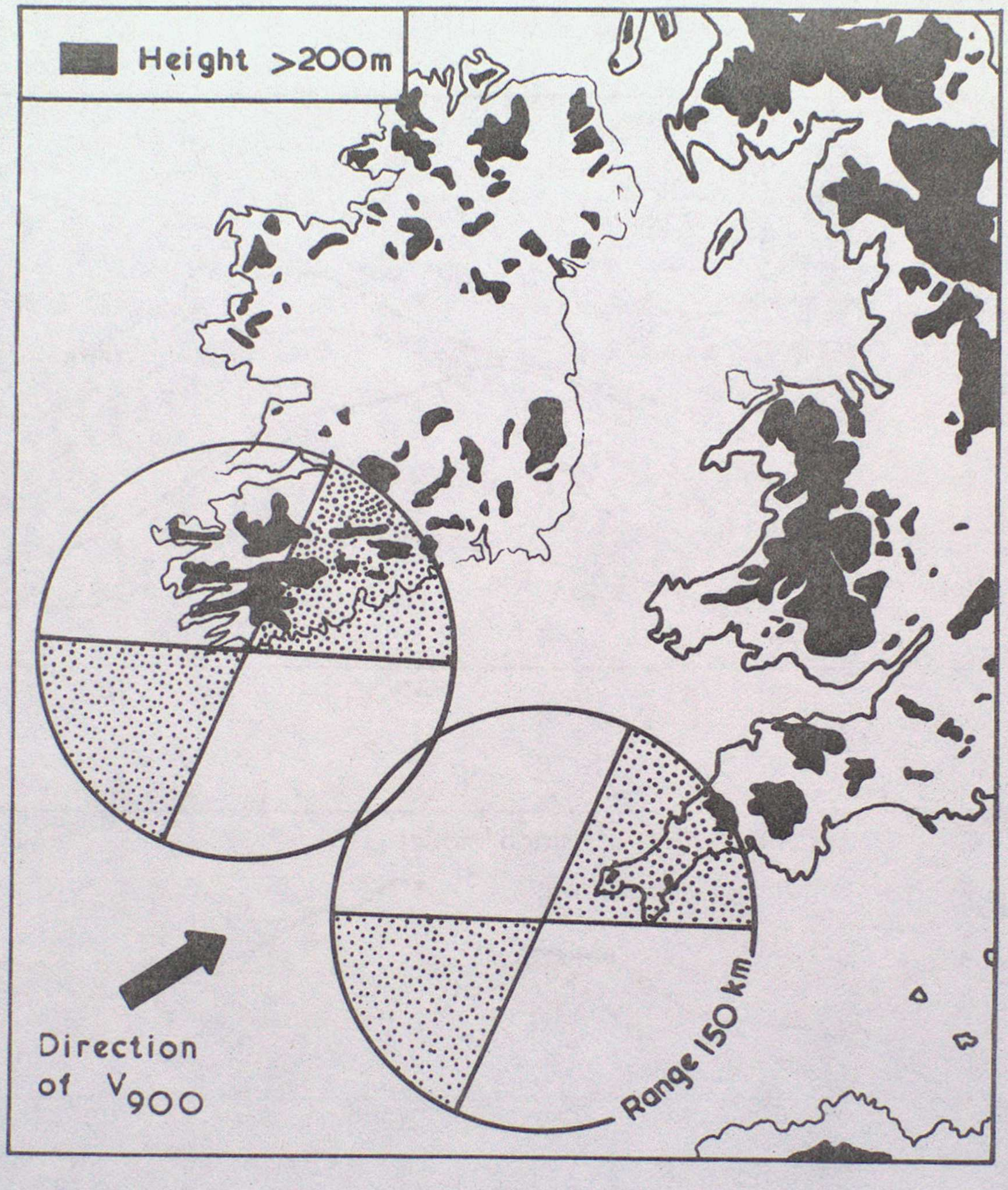


Fig 17 Technically ideal radar sites for use in forecasting heavy orographic rainfall over Wales and southwest England. Stippled shading represents areas where velocity can be measured by a single Doppler radar to an accuracy better than 20% assuming the incident wind direction to be known to $\pm$ 15° from synoptic scale weather information.

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