Radar and raingauge observations of orographic rain over south Wales

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SUMMARY

Eight detailed case studies are summarized to clarify the structure and mechanism of orographically enhanced frontal rain over hills of modest height. The observations were obtained as part of a field project in south Wales in which data from a 3-dimensionally scanning radar were combined with autographic raingauge data. The results show that the generation of orographic rain is consistent with Bergeron's seeder-feeder mechanism, according to which raindrops from upper-level (seeder) clouds wash out small droplets within low-level (feeder) clouds formed over the hills. It is demonstrated that the orographic enhancement is strongly influenced by the low-level wind speed. The largest enhancement of rainfall occurred in association with strong winds, and also high relative humidity, below 2km. The radar showed that over 80% of the enhancement occurred in the lowest 1-5 km above the hills. It also showed that the periods of enhanced rainfall were associated with the passage of pre-existing areas of precipitation. The precise value of the upwind rainfall rate was rather unimportant in influencing the orographic increment provided the rainfall rate upwind exceeded about 0.5 mm h$^{-1}$. These findings are compared with the results of theoretical calculations based upon the washout model of Bader and Roach.

1. INTRODUCTION

A field experiment was carried out in south Wales during the winter of 1976/77 to study orographic rain over an area of hills exposed to winds from the sea. The aim was to investigate the 3-dimensional structure of the precipitation and the dependence of the orographic rain on meteorological conditions upwind. As part of the experiment a quantitative weather radar was operated in a 3-dimensional scanning mode and the measurements were combined with information from autographic raingauges. Special rawinsondes were released to determine the upper air conditions just upwind of the hills.

The approach adopted in this study has been to exploit the complementary nature of radar and autographic raingauge data. Gauges usually provide quite accurate point measurements of the rain reaching the ground but they are available at only a small number of locations. The absolute accuracy of radar is poor by comparison. This is due to the influence of variable drop-size distribution and to the radar observations being made some distance above the ground. Radar, however, does provide information on vertical structure which is of interest in its own right, and in addition it provides extensive areal coverage of the low-level rainfall. Thus it enables realistic fields of surface rainfall to be derived by interpolating between the limited number of raingauge observations. Radar also provides a ready means for identifying mechanical faults and timing errors in individual gauges.

The nature of orographic rain and the conditions responsible for heavy falls in south Wales have been reviewed in broad terms by Browning (1980). It is suggested that the generation of orographic rain is consistent with the seeder-feeder mechanism advanced by Bergeron (1965) in which raindrops from pre-existing (seeder) clouds aloft wash out small cloud droplets within low-level (feeder) clouds over the hills. According to this mechanism, the amount of orographic enhancement is determined by the rate at which the low-level flow ascending over the hills can replenish the liquid water content within the feeder cloud. The largest orographic falls can be expected to occur when pre-existing areas of rain associated with fronts and warm sectors are accompanied by strong, saturated low-level flows. The feasibility of the seeder-feeder mechanism of orographic rain has been supported by several
Figure 1. Location of radar project (upper part) and topography of south Wales (lower part). The numbers denote autographic gauges which are referred to in the text and in some figures.

Theoretical studies (e.g. Storebo 1976, Bader and Roach 1977, Gocho 1978) but until now detailed measurements of the rainfall rates aloft above hills have been lacking. This study describes an attempt to measure the precipitation rate at various altitudes and to relate these measurements to the surface rainfall, with a view to determining quantitatively the effect of different parameters on the orographic rainfall.

2. TOPOGRAPHY AND ANNUAL RAINFALL OF THE GLAMORGAN HILLS

The study was centred on the 20 km-wide Glamorgan Hills which are located in south Wales between the Bristol Channel to the southwest and the Brecon Beacons to the northeast (Fig. 1). Despite their modest dimensions (maximum height only 600 m), the average annual rainfall is approximately 2500 mm on the highest ground compared with only 1000 mm on the nearby coast. A pilot study of rainfall data for 1974 and 1975 indicated that about 75% of the total rainfall on the hills (60% on the coast) was associated with low-level winds from the southwest quadrant during the passage of fronts and troughs. This implies an average enhancement factor in these circumstances of about three, although the enhancement can be much more than this in individual cases.

The average annual rainfall for the period 1941-1970 at several long-established sites (locations in Fig. 2(b)) has been plotted against the station height; see crosses in Fig. 2(a).
Figure 2. (a) Relation between average annual rainfall ($R$) at 23 raingauge sites near the Glamorgan Hills and the height ($H$) of the surrounding ground when smoothed over increasingly large areas. The smallest scatter about a linear relationship was obtained for 4 km squares displaced 1-5 km to the southwest of each site. The linear regression is then $R = 1100 + 3.25H_4$. (b) Contours of average annual rainfall derived by using this regression. The numbered locations denote daily raingauges (as distinct from the autographic gauge sites shown in Fig. 1).

These values show considerable scatter because many of the highland gauges are located in steep valleys and are influenced by adjacent higher ground. In an attempt to overcome this difficulty the average height of the ground over most of Glamorgan was calculated over 1, 2, 4 and 8 km squares, and fresh sets of rainfall/height relationships were derived. Figure 2(a) shows that the relationship improves with increasing areal averaging up to 4 km. Further analysis showed that the smallest scatter occurred using the mean height $H_4$ evaluated over 4 km squares centred 1-5 km to the southwest of the gauges (Fig. 2(a)). In view of the high frequency of SWLy winds over south Wales this result indicates the importance of the downwind drift of precipitation. The linear regression $R = 1100 + 3.25H_4$, where $R$ is in mm and $H_4$ in m, is valid between the coast and the eastern slopes of the Glamorgan Hills, but not beyond the next valley because of the rain shadow effect.

Figure 2(b) shows the average annual rainfall over the Glamorgan Hills based on the regression in Fig. 2(a). It suggests that the maximum is 2660 mm and that gauges 22 and 23, just east of the highest ground, record over 90% of the rain that falls on the highest part of the hills. Of course the percentage could be quite different on individual days depending on the direction and strength of the low-level wind (which will determine the magnitude of the downwind transport of precipitation beyond the region of maximum growth). However, Fig. 2(b) is thought to provide a useful guide to the pattern of enhancement we may expect during the passage of frontal rain systems considered in this paper, which had fairly strong winds from a southwesterly quadrant.
3. RADAR MEASUREMENTS AND ANALYSIS PROCEDURES

In order to observe areas of rain while they crossed the Glamorgan Hills and as they approached from the sea, a radar was located on the Gower peninsula, 45 km WSW of the highest ground (Fig. 1). The radar had a wavelength of 10 cm and a 2° conical beam (see footnote* for further details). The basic radar data were averaged over 2 km squares out to 84 km range and were derived in digitized format. The relationship $Z = AR^{1.6}$ was used to convert the radar reflectivity factor $Z (\text{mm}^6 \text{m}^{-3})$ to the rainfall rate $R (\text{mm h}^{-1})$, the constant $A$ being determined using the coastal gauges labelled 1, 2, 3 and 4 in Fig. 1. Radar measurements averaged over a 4 km square centered just upwind of each gauge were compared

* The radar was a mobile Plessey 43S with the following characteristics: frequency 2860 GHz, antenna diameter 3.7 m, polarization vertical, pulse length 1.5 µs, pulse repetition frequency 275 s⁻¹, peak power 650 kW, minimum detectable signal 110 dBm.
Figure 4. Example of the time-integrated rainfall pattern within a vertical section parallel to the mean direction of motion of individual rainfall areas. This is later referred to as case 2. (a) Contours (mm) based on original radar data collected at 8 elevation angles; this figure shows an intense bright band centred at 1.7 km. (b) Similar section to (a) but showing the estimated rainfall after correcting for the bright band and for low radar reflectivity above the 0°C level. Measurements from six raingauges are also incorporated into the analysis.

with the raingauge totals measured over 10 min. It was found that this procedure gave a fairly uniform set of calibration figures within individual areas of rain. For mesoscale studies, a separate calibration figure was used for each rain area; for analyses covering several mesoscale areas, a mean calibration was applied to the radar data.

The radar made a complete set of azimuth (PPI) scans at a sequence of 8 elevation angles from 0.6° to 12°, once every 9 min. As shown by Fig. 3(a), the bottom of the lowest radar beam over much of south Wales and the Bristol Channel was within 500 m of the ground. Clutter (reflections from the ground and sea) precluded the acquisition of rainfall measurements in the hatched parts of Fig. 3(b). Over the sea the clutter could be avoided by raising the axis of the beam above 1°, but the clutter from some of the ground was detectable to beyond 6°, in which case low-level radar measurements were not possible except by horizontal interpolation. The elevation angles used and the vertical coverage achieved are illustrated in Fig. 4(a), which shows an example of the results that were obtained by extracting data from each scan to produce a vertical section from southwest to northeast across the Bristol Channel and the Glamorgan Hills. The isopleths in this figure show the total ‘rainfall’ integrated over a 10 h period as inferred directly from the mean Z–R relationship applicable at the surface. This figure draws attention to one of the major observational difficulties encountered: namely the layer of high reflectivity (the bright band)
between 1 and 2 km which is due to melting snow rather than to any real maximum in rainfall intensity. In this study we have reduced the effect of the bright band using a vertical bright band profile inferred from high resolution data at ranges close to the radar, taking into account the degradation in radar resolution at longer range (see Appendix 1 for details). Above the bright band the relatively weak radar returns from ice crystals has also been allowed for. A modified cross-section corresponding to Fig. 4(a) is shown in Fig. 4(b).

We have used the combined radar/raingauge data sets in three ways.

(1) The changes in rainfall intensity during the passage of rainfall systems from sea to hills have been studied by comparing the gauge-calibrated radar intensities observed at similar heights within an 80 km² area over the hills and a corresponding area upwind over the sea (section 5).

(2) The vertical structure of the precipitation has been studied using radar aloft but with raingauge data being used to provide surface rates of rainfall over the hills where radar failed to detect the full magnitude of the low-level enhancement (section 6).

(3) The change in surface rainfall intensity from the coast to the hills has been studied by comparing radar-vetted autographic raingauge charts at appropriate pairs of coastal and hill sites (section 7).

4. THE SYNOPTIC SETTING FOR THE CASE STUDIES

Of the many rainfall systems for which data were recorded, eight have been selected for detailed study. These were characterized by winds from the southwest quadrant below 700 mb; large falls of orographic rain in this region rarely occur with other wind directions because the region is then sheltered by other areas of high ground over Wales and western England. The cases are summarized in Table 1 in order of decreasing mean low-level wind speed. The low-level wind is defined here as the wind at 600 m, just above the coastal friction layer, as measured by radiosondes launched from the radar site. The gradient wind as estimated from hourly surface charts was used to interpolate between ascents and to identify occasions when the wind over the Glamorgan Hills may have differed from that over the radar site.

<table>
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<tr>
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<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
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<td>1200</td>
<td>2050</td>
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<td>0540</td>
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<td>4.5</td>
<td>8.5</td>
<td>3.7</td>
<td>9</td>
<td>10.5</td>
<td>2.7</td>
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<td>220/26</td>
<td>230/25</td>
<td>245/22</td>
<td>235/20</td>
<td>220/19</td>
<td>195/16</td>
<td>260/14</td>
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<td>235/28</td>
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<td>243/21</td>
<td>260/13</td>
<td>210/14</td>
<td>225/32</td>
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<td>8-0</td>
<td>9-5</td>
<td>8-2</td>
<td>8-5</td>
<td>5-5</td>
<td>10-0</td>
<td>8-3</td>
</tr>
<tr>
<td>Mean RH in lowest 3000 m (%)</td>
<td>13-2</td>
<td>10-2</td>
<td>7-5</td>
<td>9-5</td>
<td>9-5</td>
<td>8-0</td>
<td>10-5</td>
<td>11-8</td>
</tr>
<tr>
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<td>93</td>
<td>85</td>
<td>90</td>
<td>87</td>
<td>92</td>
<td>87</td>
<td>80</td>
</tr>
</tbody>
</table>

Frontal analyses for the eight cases are shown in Fig. 5, each chart being drawn for a time approximately midway through the period of interest. The extent of surface rain at this time is shown by stippled shading. (This shows merely the synoptic scale extent of the rain;
figure 5. Synoptic analyses for the 8 cases: the precipitation areas (shown stippled) are based on radar data over south Wales and the Bristol Channel and on surface reports elsewhere. The double arrows show the extent of the rain which was included in each case study.
as we shall show, there was considerable mesoscale structure embedded within it.) A double line is used to denote the portion of the rainfall system that crossed the Glamorgan Hills during each period of study. The project evidently covered a wide variety of rainfall systems. Most of the systems were analysed as frontal, the rain occurring both at the fronts and in the associated warm sectors. One of the systems (case 3) was a convective trough. Each case has been limited to a period within which the wind direction was fairly uniform. Cases 1 and 8 span the same system, case 8 applying to the rain just behind the cold front.

Upper air conditions are plotted in Fig. 6. Most of these tephigrams represent an average of several ascents launched from the radar site during the passage of the rainfall system. Because of operational difficulties, however, the tephigrams shown for cases 3 and 7 are based on ascents launched from Aberporth and Camborne (located 90 km WNW and 200 km SW of the Glamorgan Hills, respectively). In 6 of the 8 cases the air in the lowest 1 or 2 km was potentially unstable. A layer of weak potential instability also occurred from time to time in the middle troposphere in many of the cases, although it tended to be masked because of the limitations in the humidity measurements from radiosondes.
Figure 7. Rates of rainfall over and upwind of the hills during Case 1 (solid curves) derived from radar measurements in an 80 km$^2$ area over the Glamorgan Hills and in a similar area south of Swansea Bay located upwind of the Glamorgan Hills. Also shown is the mean rate of rainfall recorded by two gauges on the hills (dashed curve). The location of these gauges is given in Fig. 1.

5. COMPARISON OF LOW-LEVEL RAINFALL RATE OVER THE SEA AND HILLS AS MEASURED BY RADAR

For the investigation of changes in rainfall intensity at low levels between sea and hills, a compact zone comprising twenty of the 2 km radar cells was first selected over the eastern half of the Glamorgan Hills, the cells being chosen to avoid those in which there was severe clutter. Then another zone of identical configuration was selected precisely ‘upwind’ (with respect to the motion of the mesoscale rain areas) of the first zone and about 20 km offshore. The mean areal intensity of the rain was calculated in each zone using gauge-calibrated data from the lowest radar scan available at 9 min intervals. The timing and intensity of small-scale features which would have caused peaks and lulls in the rainfall intensity during the 9 min between each scan cycle were estimated by spatial interpolation thereby enabling a continuous curve of the mean rainfall intensity to be drawn as a function of time for each zone.

The curves produced by the above procedure for two of the cases (1 and 4) are shown in Fig. 7 and 8. The abscissa in these figures represents time over the Glamorgan Hills, the curves for the zone over the sea being displaced to allow for the time taken by rain areas to travel between the two zones. From these figures and those for other cases (not shown), we observe (a) that there was nearly always an increase in the rainfall intensity from the sea to the hills and (b) that the periods of heaviest rain over the hills were usually associated with
the passage of identifiable precipitation areas which had previously been observed upwind by radar.

The dashed curves in Figs. 7 and 8 show that the rainfall intensities measured by rain-gauges over the hills were generally higher than the radar-derived values there, sometimes appreciably so. The radar measurements were calibrated using raingauges at coastal sites, and so any substantial differences between the radar and raingauge curves could be partly due to changes in drop-size distribution from the sea to the hills. It is well known that mountain rainfall tends to consist of a high proportion of small raindrops and drizzle (Pedgley 1970). According to Joss et al. (1970), in the case of drizzle, the rainfall intensity responsible for a given radar reflectivity tends to be greater than that for typical widespread rain by a factor of about 1·4. This effect could have contributed to the particularly large discrepancy between the radar and raingauge measurements during the passage of the precipitation area labelled D in Fig. 8, in which the coastal observer reported a continuous mixture of rain and drizzle.

The other cause of disagreement between the radar and gauges over the hills is the effect of strong rainfall gradients in the vertical. Figure 3(a) shows that the base of the lowest beam was approximately 200 m above the ground over the Glamorgan Hills; allowing for the shape of the beam and the occultation of part of it, the effective centre of the lowest beam was 500 m above the general level of the hills. The evidence to be presented in section 6 shows that there may have been considerable enhancement of the rainfall between this height and the surface in situations of strong winds.

6. VERTICAL DISTRIBUTION OF RAINFALL RATE ACROSS THE HILLS

(a) Time sequences

In all of the cases in this study the precipitation extended far above the 0°C level into the middle troposphere. The bulk of the precipitation at mid-tropospheric levels was associated with fast-moving mesoscale areas having dimensions of between 20 and 100 km. This
Figure 9. Vertical sections at 18-min intervals for Case 1 showing rainfall rate derived from calibrated radar measurements after correcting for the bright band. The sections are parallel to the motion of individual rainfall areas and intersect the hills near gauge No. 6 (location in Fig. 1). The dashed contour corresponds to 0-2 mm h⁻¹; solid contours represent 1, 2, etc., mm h⁻¹. Numbers plotted on the hills show the surface rainfall rate at gauge No. 6. Letters identifying individual mesoscale precipitation areas correspond to those used in Fig. 7.
was true both over the sea and over the hills. Within these mesoscale areas, at heights of 3 km or more, the equivalent rainfall rate occasionally reached 2 to 4 mm h$^{-1}$; outside them at these heights it was usually below 0.2 mm h$^{-1}$. Close to the ground, orographic effects led to rainfall rates of over 8 mm h$^{-1}$, with rates of 2 mm h$^{-1}$ often persisting during periods when there was only slight precipitation aloft. We shall now illustrate these features by means of two examples.

Figure 10. Same as Fig. 9 but for Case 2. Gauge No. 5 is used instead of 6, and letters identifying individual mesoscale precipitation areas correspond to those used in Fig. 11.
Figures 9 and 10 show the vertical distribution of rainfall intensity at a sequence of times for cases 1 and 2, respectively. Almost the entire sequence depicted in Fig. 9 was obtained in a warm sector; Fig. 10 was obtained mainly ahead of a warm front. In each case the sections are across the Glamorgan Hills and parallel to the motion of the rain areas. The bright band has been eliminated and the precipitation rates above it increased as described in Appendix 1. The times shown correspond to periods of maximum mesoscale activity. Individual precipitation maxima are identified by letters and can be traced moving from left to right from one time to the next as they travelled first over the sea and then over the hills. The ten-fold exaggeration in the vertical scale draws attention to the variability of the precipitation rate but also gives a spurious impression of convective activity; the radiosonde ascents suggest that convective cells would have been present at times but because of the format of the radar data these were not easy to identify.

Figure 11. x-t sections showing rainfall rates at (a) altitude 500-900 m and (b) altitude 2700-3300 m above MSL, derived from calibrated radar data, during the passage of a succession of mesoscale precipitation areas across the Bristol Channel and Glamorgan Hills. The x-axis corresponds to a direction of approximately 235°/055°. The t-axis represents time in GMT. Letters correspond to those used in Fig. 10. Notice the pronounced orographic maximum over the hills for much of the time in (a) and its virtual absence in (b).
A comparison of Fig. 9 and 10 shows that a more intense mesoscale structure was present in case 2 than in case 1, with a correspondingly larger vertical gradient of precipitation rate from high levels down to about 2 km. However, in neither case was there any overall trend towards intensification of the upper-level precipitation as rain areas travelled from sea to hills. In spite of the modest amount of medium-level intensification above the hills, substantial enhancement occurred at low levels and continuous and often quite heavy rain fell on the high ground in both cases. The contrast between the major orographic effects at

Figure 12. Mean rates of rainfall (mm h\(^{-1}\)) above the Glamorgan Hills and eastern Brecons for the duration of each case study. The mean wind at 600 m above MSL is shown below the case number. Numbers along the base-line refer to autographic raingauge sites (see Fig. 1) which lie close to each section, the orientation of which depends on the mean direction of motion of the individual rain areas.
very low levels and the almost negligible orographic enhancement occurring at rather higher levels is illustrated even more clearly in Fig. 11, which shows x-t representations for case 2 of the rainfall intensity at 500 to 900 m above MSL (Fig. 11(a)) and at 2700 to 3300 m above MSL (Fig. 11(b)).

(b) Time integrations

Figure 12 shows vertical sections of mean rainfall intensity across the Glamorgan Hills for all 8 cases, obtained from a time integration of data of the kind depicted in Fig. 9 and 10. These sections intersect the high ground near the same point (between gauges 5 and 6 in
Fig. 1) but their orientation varies according to the mean direction of movement of the precipitation areas. The mean rate of rainfall, over the sea as well as over the hills, has been calculated by dividing the total fall (measured using both radar and gauges) by the duration of surface rain in excess of 0.1 mm h\(^{-1}\) as measured by gauges over the hills. This definition was used because of the well-known tendency of recording gauges to fail to record the slight rain and drizzle in frontal systems which frequently occurs at coastal sites when heavier and easily recordable rain is falling on high ground.

It is readily apparent from Fig. 12 that the vertical gradient of mean rainfall over the hills fell sharply from case 1 to case 8 as the mean 600 m wind decreased. A high proportion of the enhancement occurred below 1·5 km above MSL. Only in case 1 did the enhancement begin significantly above this level, this case being noteworthy for having the strongest low-level winds and the highest melting-level.

The essential features of the sections in Fig. 12 are synthesized in Fig. 13. This has been done by plotting for each case a curve of the maximum rainfall intensity as a function of height above the hills (note from Fig. 12 that this tended to occur slightly upwind of the

![Diagram](image)

**Figure 13.** Vertical profiles of the rate of rainfall above the Glamorgan Hills averaged over the duration of each case study. Inset shows the enhancement over the lowest 1·5 km compared with the wind speed at 600 m above MSL.
7. **Comparison of Surface Rainfall Rate Over the Coast and Hills Using Gauges**

In order to investigate further the dependence of orographic enhancement on low-level windspeed $V_L$ and also its dependence on the background rainfall intensity $P_0$, the rainfall rate at a coastal site (gauge 4 in Fig. 1) has been compared with that at a hilltop location (either gauge 5 or 6 depending on serviceability) whenever the mesoscale rain areas moved from between $210^\circ$ and $250^\circ$. This directional tolerance is such that the coastal site would on occasions have been as much as 8 km to one side of a line drawn directly upwind of the hilltop site. The justification for allowing this degree of latitude is that in hilly areas the differences in rainfall intensity caused by the topography often exceed the intrinsic spatial variations within the mesoscale precipitation areas, in which case it is sufficient that the pair of gauge sites should be affected by the same general region of a mesoscale precipitation area without necessarily being precisely in line.

The cases listed in Table 1 have been used for this analysis but some of them have been extended to include periods when only partial radar coverage was acquired. Some of the cases have also been subdivided to reveal in more detail the influence of changes in wind speed during each period. This has yielded a total of 14 periods which are listed in Table 2.

![Image](image.png)

**Figure 14.** Mean enhancement of the surface rainfall intensity from the Glamorgan Coast (autographic gauge site No. 4 in Fig. 1) to the Glamorgan Hills (gauge 5 or 6) plotted as a function of the windspeed at 600 m above MSL.
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<th>3b</th>
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<tr>
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together with the corresponding low-level wind velocity. The duration of the rain events is also given in Table 2 and, as before, this is defined as the total period during which the rainfall rate at the hill site exceeded 0.1 mm h⁻¹. The data in Table 2 have been plotted in Fig. 14 to show the strong association between mean enhancement from the coast to the hills \((P_h - P_0)\) and mean low-level wind speed \((V_L)\). The curve drawn in this figure is \((P_h - P_0) = 6.5 \times 10^{-4} V_L^{2.8}\).

The association of orographic enhancement \((P_h - P_0)\) with background rainfall intensity \(P_0\) is more difficult to demonstrate, partly because the dependence is weaker and partly because of the great variability in background rainfall rate. Fortunately there is very little variability in the vertical over the sea and so it has been possible to take the surface rainfall rate at the coastal site as representative of the background precipitation rate associated with the incoming seeder clouds. However, in order to overcome difficulties due to the rapid temporal variability of the background rainfall intensity associated with the swift passage of mesoscale rain areas, it has been necessary to consider the rainfall over short time intervals. Accordingly we have compared the rainfall at the coastal gauge with that at the hilltop gauge evaluated over matching 30 min periods. Appropriate time adjustments have been made to allow for the distance between the two sites. The results, plotted in Fig. 15,

![Figure 15. Orographic increment between the Glamorgan coast and the hills plotted as a function of coastal rainfall intensity \((P_0)\) and the windspeed at 600m above MSL \((V_L)\). These values are based on half-hourly measurements from a coastal raingauge (gauge 4 in Fig. 1) and a hilltop raingauge (gauges 5 or 6). Smooth curves have been drawn so as to fit as many as possible of the data points. Data points are encircled for those cases (Nos. 1, 2, 4a and 6a) with a mean relative humidity of at least 90% in the lowest 2 km. The four data points marked by stars originate from the first two hours of case 1 and are explained in section 8.](image-url)
confirm the strong dependence of enhancement on wind speed. For speeds below 20 m s$^{-1}$ the enhancement was small ($\leq 3$ mm h$^{-1}$) irrespective of the coastal rainfall rate. All occasions of large enhancement ($\geq 5$ mm h$^{-1}$) occurred with wind speeds of at least 23 m s$^{-1}$; although these were usually associated with coastal rates of more than 1 mm h$^{-1}$, some instances of large enhancement were observed when the coastal rate was only 0.3 mm h$^{-1}$.

Occasions when the coastal gauge was recording no rain despite the occurrence of drizzle or very light rain are not plotted in Fig. 15. A comparison of the radar intensities (centred at 600 m above the ground) with the raingauge trace at coastal site 4 showed that seeding rates of up to 0.15 mm h$^{-1}$ did sometimes occur near the coast when no rain was recorded by the gauge. In such cases, however, the subsequent rainfall rate at hill sites 5 and 6 was always found to be less than 2 mm h$^{-1}$. Occasions of true zero rainfall over the coast were almost always found to be associated with zero measurable rainfall at the hill sites.

To test the representativeness of the present results we have compared Fig. 14 with the unpublished results of two earlier studies in the same area (Nash and Browning 1977,* and Hill 1977*) – see Fig. 16. The first of these studies was confined to very wet periods (daily falls exceeding 85 mm on the Glamorgan Hills). The latter took into account all cases of rainfall of more than 20 mm on the hills and, unlike the other study, included many occasions when the low-level wind speeds were less than 15 m s$^{-1}$. Figure 16 shows that agreement between the three studies is quite good.

* Nash, J. and Browning, K. A. 1977 (Structure of the atmosphere associated with heavy falls of orographic rain in South Wales), and Hill, F. F. 1977 (On forecasting rainfall at Llyn Fawr over periods of a few hours); both reports available from the Meteorological Office Radar Research Laboratory.

![Figure 16](image)

Figure 16. Dependence of the orographic increment ($P_h - P_o$) on windspeed, as shown in Fig. 14, compared with the corresponding relationships obtained in previous unpublished research studies.
8. Comparison of Observational Results with Theory

Bader and Roach (1977) have developed a theoretical model of the washout of droplets in an orographically produced low-level feeder cloud by rain-drops falling from a high-level seeder cloud. Their calculations were based on conditions commonly encountered within warm sectors of depressions, as determined during our earlier studies of orographic rain in south Wales. Bader and Roach assumed that the orographic enhancement was associated with an initially saturated layer with $\theta_w = 10^\circ\text{C}$ located entirely below the freezing level, between the surface and 1.5 km. The results in the present paper confirm the appropriateness of these assumptions. A particularly important result, summarized in Table 3, is that most of the increment in rainfall rate between the south Wales coast and the Glamorgan Hills can indeed be explained by enhancement in the lowest 1.5 km.

<table>
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<tr>
<th>Case number (as defined in Table 1)</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
</tr>
</thead>
<tbody>
<tr>
<td>Enhancement of surface rainfall rate between the coastal station and hill-top gauges (mm h$^{-1}$) (as plotted in Fig. 14)</td>
<td>6.6</td>
<td>3.8</td>
<td>4.1</td>
<td>2.5</td>
<td>1.6</td>
<td>1.2</td>
<td>0.7</td>
<td>1.2</td>
</tr>
<tr>
<td>Enhancement in the vertical within the lowest 1.5 km above the hills (mm h$^{-1}$) (derived from same data as plotted in Fig. 13)</td>
<td>6.2</td>
<td>3.3</td>
<td>3.3</td>
<td>2.6</td>
<td>1.6</td>
<td>2.2</td>
<td>1.0</td>
<td>0.3</td>
</tr>
</tbody>
</table>

The published results of Bader and Roach (see their Fig. 3) were for only a limited range of background rainfall rate $P_0$ and low-level wind speed $V_L$. Accordingly the model has been re-run using the same assumptions as before but for a range of $P_0$ and $V_L$ similar to that encountered in the present study. The results are shown in Fig. 17 in the same format as the observational results in Fig. 15. The results in these two figures are similar insofar as the enhancement increases as a function of both $P_0$ and $V_L$. However, the model shows a much smaller dependence of the enhancement on $V_L$ than was actually observed and a much greater dependence on $P_0$. In particular, the theoretical model gives too little enhancement at high windspeeds ($V_L > 20$ m s$^{-1}$). Some of the more likely causes of these differences are now reviewed briefly. The first three effects could explain the higher overall rates of rainfall observed over the hills, while the last three could contribute to the stronger dependence on the wind.

1. Slope of the airflow over the hills

Bader and Roach took an idealized hill with a mean slope of 1 in 100 and assumed that the air flows parallel to it near the surface. A close look at the Glamorgan Hills shows that the terrain is much steeper than this. If we assume that the air follows the terrain smoothed on a scale of 4 km (as implied by section 2 of this paper), then the mean gradient would have been 1 in 40, reaching 1 in 35 from 0 to 10 km upwind of the crest. By re-running the model, we have estimated that the resulting increase in the rate of condensation might cause a 40% increase in orographic enhancement over a small area.

2. Wind speed over the hills

Bader and Roach assumed that the wind speed at the surface drops to one half of its
undisturbed value at 600 m. Whereas this is believed to be a good assumption over the sea and coast, it may underestimate the wind speed near the surface over high ground. If the surface wind speed at the hill tops were equal to the gradient wind speed, then the resulting increase in the vertical component of the air flow would cause the orographic increment to increase by about 25% compared with the values given by Bader and Roach.

(3) Potential instability

Potential instability occurs independently at two levels in the kind of situation studied in this paper. One of the unstable layers is in the middle troposphere; the other occurs close to the ground. Shallow convective cells associated with the upper layer play a role in sustaining the seeder cloud. Browning et al. (1974) found that the ascent of moist warm-sector air over the Glamorgan Hills led to the triggering of fresh convection at middle levels. However, that case was more than usually unstable. In the eight cases described in the present study there was no evidence of sustained enhancement of precipitation rate within the seeder cloud which could be attributed to this effect.

The lower layer of potential instability is thought to have had more impact on the orographic rainfall than the mid-level instability. The air in the lowest 1 to 2 km was potentially unstable for much of the time in 6 of the 8 cases, and in all but one of these the air was sufficiently moist for the instability to be realized during forced ascent over the Glamorgan Hills. As suggested by Browning et al. (1974) and Smith (1979, p. 178) the resulting low-level convection is likely to have led to a local increase in the mean rate of ascent and hence in the rate of condensation near the hilltop.
(4) **Horizontal drift of precipitation**

For a mean fall speed of $5 \text{ m s}^{-1}$ in the presence of a $25 \text{ m s}^{-1}$ wind, rain drifts $7.5 \text{ km}$ downwind during a descent of $1.5 \text{ km}$. Because of the steep gradient of the hills just upwind of the highest ground, rain which reached the surface just to the lee of the crest would have passed through a region of high liquid-water content. Hence, in strong winds, the rainfall recorded by the two gauges (No. 5 and 6 in Fig. 1) located just downwind from the crest would have been near to the maximum that is likely to have fallen on the highest ground. However, in light winds, the heaviest rain is more likely to have occurred upwind of the crest near the position shown by Bader and Roach (who did not include drift in their model). Hence we suspect that the data in Fig. 15 underestimate the true enhancement for winds lighter than about $20 \text{ m s}^{-1}$.

(5) **Unsaturated air**

The theoretical orographic enhancement in Fig. 17 has been calculated assuming initially saturated low-level air. During the case studies the mean relative humidity in the lowest $1 \text{ km}$ varied from 80 to 93\% (Table 1). Moreover, there was a tendency for the data to be biased slightly towards the drier low-level flow occurring in the cases of weaker winds, causing the apparent dependence of enhancement on wind speed to be larger than it would have been in the absence of humidity variations.

(6) **Difference between the surface coastal rainfall rate and the true seeding rate aloft**

The calculations in Fig. 17 are related to the seeding rate $P_0$ at the top of the feeder cloud, whereas in Fig. 15 the rainfall at a coastal site has been used as a measure of $P_0$. The radar evidence is that this is usually a good approximation when the rainfall is averaged over periods of several hours. However, in cases where the low-level flow was significantly under-saturated, the true seeding rate aloft would have been higher than the surface coastal rate and hence the true (vertical) enhancement less than that deduced from the raingauge measurements. This will tend to oppose the effect of (5), which is probably why the influence of humidity does not show more strongly in Fig. 15. On the other hand, if a shallow dry layer is located above a fairly deep layer of moist low-level air, the seeding rate above the hills may be reduced significantly whilst still retaining a rich feeder cloud. In this event a large orographic enhancement may ensue in association with an anomalously low surface rainfall rate on the coast. This occurred during the first two hours of case 1 and gave rise to the data points marked by stars in Fig. 15.

The above factors suggest that the differences between the observations and the model of Bader and Roach can be accounted for partly in terms of limitations in the observational data. Nevertheless there is sufficient evidence here to suggest that some reformulation of the the physical processes is needed to account for the observed stronger influence of wind speed compared with seeding rate.

9. **Conclusions**

The evidence presented in this paper supports the view that orographic rain is largely a low-level phenomenon and that the washout of cloud droplets is likely to be an important mechanism for generating it. More than 80\% of the overall orographic enhancement was concentrated in the lowest $1.5 \text{ km}$ above the hills when the low-level wind was in excess of $20 \text{ m s}^{-1}$. The presence of a low melting-level in most of the cases observed during this
project caused the radar measurements to be distorted by the bright band. Although this has prevented the gradient of precipitation rate above the hills from being determined precisely, we believe that it does not invalidate the broad conclusions regarding the predominantly low-level origin of the enhancement.

Over periods of several hours the enhancement was found to be quite well correlated with the mean wind speed just above the friction layer. The existence of high relative humidity in the lowest 1.5 km was also important. Together these factors ensure the formation of a rich feeder cloud at low levels. The radar confirmed that periods of enhanced rain over the hills were associated with the passage of pre-existing areas of precipitation (seeder clouds), some of which gave rain and drizzle upwind that was too slight to be recorded by coastal gauges. When these gauges showed that the rainfall rate was less than about 0.3 mm h\(^{-1}\), there was usually only a small enhancement over the hills. For wind speeds greater than 20 m s\(^{-1}\), the enhancement was often large and appeared to be influenced more strongly by the low-level wind speed than by the seeding rate, provided the latter exceeded 0.5 mm h\(^{-1}\).

These observations agree with the predictions of the theoretical model by Bader and Roach (1977) in that the enhancement depends both on the low-level wind speed and on the background rainfall rate; however, the observed dependence on background rainfall rate is less than that predicted by the model whereas the dependence on wind speed is greater. A particularly large discrepancy between observation and theory is in the observation of large orographic enhancement in strong winds with fairly light coastal rain.

A useful forecast of the orographic enhancement of rain can be obtained using:

(i) the duration of rain which will cross the hills in the specified forecast period,
(ii) the associated windspeed and direction above the friction layer, and,
(iii) the average relative humidity in the lowest 1.5 km.

The rainfall duration can be predicted using radar up to about 3 h ahead and this period can be extended by analysis of satellite imagery. The wind can be forecast from routine synoptic analysis. The humidity may be difficult to forecast, however, on account of its high variability on the mesoscale. It is clearly most important to discriminate as quickly as possible between moist and fairly dry low-level air when the winds are expected to be strong.

**APPENDIX I**

**METHOD USED TO REDUCE BRIGHT BAND EFFECTS**

In processing the radar data the measurements were converted to a rainfall rate using \( Z = AR^{1.6} \) where \( Z \) is the equivalent reflectivity factor and \( R \) is the rainfall rate, the constant \( A \) being obtained by calibrating the radar data at low levels against coastal gauges. However, the reflectivity from ice crystals is less than from rain at a comparable precipitation intensity whereas the reflectivity from aggregated snow and especially from large melting flakes is much higher. Hence measurements from a radar tend to be too high where the beam intersects the melting layer (the bright band region) and are too low at high levels. The bright band effect is most severe if the beam intersects the melting layer at short range, for the melting layer may then fill most of the beam. Because the melting layer has a finite depth whereas the radar beam broadens with range, the bright band tends to decrease in intensity with range but it extends over a greater depth.

The following procedure was used to reduce the magnitude of the errors. Particular reference is made here to the time-integrated section in Fig. 4(a).

**Step 1.** The height and vertical profile of the bright band were determined from
Figure A1. Radar measurements of the time-integrated rainfall for case 2 as a function of height, taken at a range of 6 to 12 km from the radar (curve 1). Each of the data-points is a radar measurement in a 2 km square. Slight broadening of the bright band (by 100 to 200 m) will have occurred because of the 2° beam-width. The true rainfall total is assumed to follow the dashed line (curve 2), being a linear interpolation between the ice crystal region above and the rain region below the melting layer.

2 km-square data at elevations of 8°, 10° and 12° to the southeast of the radar site. This enabled the structure to be resolved at short range and as nearly as possible upwind of the hills. A smooth profile was drawn through these points (curve 1 in Fig. A1).

Step 2. An estimate of the true precipitation profile over the sea was then made by assuming that, below the bright band, the true rainfall was equal to the surface-calibrated radar measurements and that, above the bright band, the equivalent rainfall was higher than the radar measurement by a factor of 2.5. This figure was used because the reflectivity of spherical water droplets is higher than from idealized, spherical ice particles by a factor of 4.2; thus the radar data needed to be multiplied by $(4.2)^{1/6} = 2.5$ in the ice-crystal region. (Note that errors in this factor, such as may be caused by the ice crystals having various non-spherical shapes and different size distributions compared with the rain would need to be large to distort the vertical profiles seriously in view of the small precipitation rates aloft.) The rainfall in the bright band region was derived by linear interpolation between the values above and below it (curve 2 in Fig. A1).

Step 3. The ratio $f$ of curve 1 to curve 2 was then calculated (see right-hand side of Fig. A2). This profile of $f$ was assumed to remain constant as the precipitation traversed the hills. There was evidence in some sections, including Fig. 4(a), for a lowering of the bright band over the western slopes of the Glamorgan Hills. Indeed a comparison of radar measurements at the melting level over the sea with measurements at the same height and range over the hills showed that the radar intensities often decreased from the sea to the hills even though the precipitation would generally have been higher over the hills. This apparent anomaly could be explained by assuming that the melting layer had fallen by 100 to 200 m. This fall was taken into account in the subsequent steps.

Step 4. The radar measurement from any one beam at a particular range is a combi-
nation of several factors: the precipitation rate, the microphysical factors affecting the relationship between precipitation rate and radar reflectivity, and the distribution of power across the beam. We assumed that the radar beam had a Gaussian profile (this ignores the existence of side-lobes but they constitute a small proportion of the total two-way power). Hence, by taking into account the height of each beam, the distribution of power in the vertical, the height of the melting layer and the bright band profile, as shown in Fig. A2, an estimate could be made of which radar measurements were likely to be too high and which
would be too low. This enabled a subjectively-derived cross-section of the rainfall to be drawn.

Step 5. As a constraint on the subjective analysis, the following objective check was made over the Glamorgan Hills for each integrated section. Let \( P_i \) be the fraction of the two-way power contained in each 0·1° horizontal slice of a 2° Gaussian beam; let \( f_i \) be the bright band correction factor in each slice, and let \( R_i \) be the rainfall in each slice as indicated by the subjective analysis. If the subjectively estimated rainfall is approximately correct, then \( \Sigma P_i (f_i R_i)^{1·6} \) should equal \( (R')^{1·6} \) where \( R' \) is the uncorrected radar measurement. If the summation did not agree with \( R' \), the subjective analysis was revised until reasonable continuity from one beam to the next was attained.

The results are only approximations to the true rainfall. Neither the bright band intensity nor the dropsize distribution will necessarily remain constant as the precipitation moves from the coast to the hills. The most likely trends are for the bright band to intensify (in which case the true rainfall at the level of the bright band would have been less than this procedure indicates) and for the proportion of small droplets to increase at low levels within the orographic cloud (causing the true rainfall to become increasingly more than was estimated by radar as it fell towards the hill tops). With reference to Fig. A2, these effects would mean that the rainfall estimate from the 1·6° beam would be too high and that from the 0·6° beam would be too low. The greatest decrease of rain with height might thus occur between these two beams instead of between the 0·6° beam and the surface. For the correlation of enhancement with wind speed in this paper, we have quoted the mean enhancement of rainfall intensity calculated over a 1·5 km layer above the hills and this avoids the question of precisely how this enhancement was distributed.

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REFERENCES


<table>
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