Structure and mechanism of precipitation and the effect of orography in a wintertime warm sector

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SUMMARY

A case study is presented showing the three-dimensional structure and evolution of precipitation upwind, over, and downwind of the south Wales hills during the passage of a wintertime warm sector that gave rather heavy and prolonged rainfall. The precipitation structure was synthesized from a network of weather radars and autographic rain gauges; it is interpreted within a dynamical framework derived from routine upper air soundings supplemented by serial ascents upwind and downwind of the hills.

The warm sector was characterized by a fast-moving airstream with potential instability (PI) not only at low levels due to its passage over a warm sea but also in the middle troposphere. The middle-level PI occurred extensively and played a significant role in determining the distribution and amount of precipitation. It has not received much attention until now, mainly because the limitations of the humidity element in conventional radiosondes tend to cause its magnitude to be underestimated. The middle-level PI was due to differential thermal advection in an intense and nearly vertical baroclinic zone that extended ahead of the surface cold front at middle levels.

Large-scale ascent was slight in the warm sector but, nevertheless, middle-level PI was realized even over the sea in scattered 'Mesoscale Precipitation Areas' (MPAs) which travelled rapidly at the speed of the winds in the middle troposphere (about 120 km hr⁻¹). Once the airstream experienced orographic uplift, fresh outbreaks of middle-level convection occurred extensively between existing MPAs. The fresh outbreaks were observed first as middle-level precipitation echoes over the sea up to 100 km upwind of the hills, indicating that the orographic ascent aloft began far upwind of the hills. Thus some of the precipitation generated by orographic ascent in the middle levels reached low levels over the hills despite the drift of the precipitation in the strong winds; there it seeded low-level orographic cloud which gave heavy rain over the hills. Some of the convective precipitation generated aloft as a result of the middle-level PI also reached the ground downwind of the hills, thereby displacing to some extent the commonly-observed rain shadow.

The importance of seeding from aloft for releasing heavy orographic rainfall has long been recognized. Sometimes the seeding is achieved by precipitation generated aloft by large scale ascent; however, the present study suggests that in certain circumstances the hills may be able to generate their own seeds when they are not being generated by large scale ascent. In the case study the occurrence of heavy warm-sector rainfall over the hills appears to have been favoured by the existence of middle-level PI which, although not being realized generally by large scale ascent, required only a small amount of local orographic ascent to realize it. Thus forecasting techniques that predict rainfall on the basis of the forecast large-scale vertical ascent may fail to identify some important situations of orographic warm sector rain unless the interaction of PI and orography is realistically taken into account.

I. INTRODUCTION

Prolonged falls of rain sometimes occur in the warm sectors of depressions and in the area just on the warm side of slow moving or waving cold fronts. In the hilly western parts of the British Isles orographic effects can give heavy rainfall in these situations. Douglas and Glasspoole (1947) showed that the heaviest orographic rainfalls are associated with strong south-west to west winds in the warm sector together with a deep layer of moist air. Generally the rain in the warm sector is limited to within about 300 km of a front; farther south in most warm sectors the air is moist to a depth of only two kilometres or so.

During the last quarter century there has been little progress toward a physical understanding of the structure and mechanism of warm sector rain and its dependence on orographic effects. This slow progress is particularly disappointing in view of the practical importance of forecasting warm sector rainfall. Holgate (1973) has reviewed rainfall forecasting for River Authorities in the British Isles, mainly for the hilly parts of north Wales.
and north-west England. He showed that prolonged heavy rainfall rather than thundery activity is the principal cause of flooding; an average rainfall rate of only 3 to 4 mm hr$^{-1}$ over 10 or more hours is often sufficient to produce flooding in broad river valleys (depending on antecedent precipitation, soil moisture deficit, and so on). Holgate also showed that the ten most outstanding occasions of prolonged heavy rain (satisfying the criterion of more than 50 mm at a mean rate of at least 8 mm hr$^{-1}$) that occurred over the period 1954 to 1963 at a hilly site in the English Lake District all were associated with the eastward passage of a warm sector of a depression enclosing moist air of subtropical origin. At present the techniques for forecasting heavy prolonged rainfall in these hilly areas are empirical refinements of the results of Douglas and Glasspoole’s study.

In a discussion at the Royal Meteorological Society as long ago as 1945, static instability was considered to be an important condition for orographic rainfall (Bonacina 1945). Since then the role of instability in orographic rainfall has not received much attention in this country. Douglas and Glasspoole, for example, while noting the existence of shallow unstable layers embedded within much deeper layers of nearly constant wet bulb potential temperature, concluded that potential instability is not a necessary factor for heavy orographic rain. The purpose of this paper is to describe new evidence which shows that, even in winter, potential instability can in some circumstances play a significant role in determining the structure and amount of precipitation in a warm sector.

An exercise was mounted during October–December 1972 involving a number of weather radars and special radiosonde ascents covering areas upwind and downwind of the south Wales hills. In this paper we present a case study of a wet warm sector that crossed Wales on 5 December 1972. Rainfall totals for the entire system were large but not extreme on this occasion, with maximum values of about 50 mm over parts of the Welsh hills and 5 to 10 mm over central England.

2. The data

Locations of the principal observations used in this study are shown in Fig. 1. A common direction of the winds in wet warm sectors is from west-south-west and, as shown

![Figure 1](image-url). Locations of principal observations during Autumn 1972 exercise. See text for details.
by the large arrow in Fig. 1, rain-bearing systems in these circumstances approach the hills of south Wales and south-west England from the Atlantic without prior modification by topography (refer ahead to Fig. 6(a) for map showing high land). The hatched regions in Fig. 1 depict the extent of qualitative precipitation surveillance by the weather radars. High-power research radars at Malvern and nearby Defford (D) gave coverage downwind of the hills of south Wales. Data over north Wales were provided by a Plessey 43S radar at Llandegla (L). Precipitation data also were obtained using a mobile 43S radar at Port Eynon on the Gower Peninsula (G). The Gower site had a horizon everywhere below $\frac{1}{2}$ degree and it provided good coverage over the hills of south Wales and upwind over the sea. All the radars gave PPI data showing the horizontal distribution of precipitation, mainly to about 200 km (hatched areas in Fig. 1), and the Defford and Gower radars gave RHI data showing the vertical precipitation structure out to 100 km (cross-hatched areas in Fig. 1). Although the slow-scanning Defford radar gave RHI data only at 90 deg intervals of azimuth, the Gower radar provided detailed RHI data at 4 deg intervals. All radars completed a scanning sequence at intervals of 15 min or less. Owing to the large beamwidth of the 43S radars, no attempt was made to use them to derive surface rainfall patterns quantitatively beyond 50 km. Instead, quantitative patterns of surface rainfall were derived using a dense network of existing autographic raingauges operated by many different authorities.

Routine radiosonde observations from a large area were used to provide a broad upper air description; the nearest sonde stations are indicated by X in Fig. 1. In addition special radiosondes were released from Camborne (C) and Defford (D). The latter ascents (obtained at 1-hr intervals, with detailed wind soundings using a precision tracking radar) clarified the upper air structure downwind of the hills; the former (obtained at 6-hr intervals) provided an indication of the structure upwind of the main hills. Comparisons of soundings from Camborne with soundings made simultaneously from the Isles of Scilly in earlier studies indicated that the small hills (altitude less than 270 m) in extreme south-west England upwind of Camborne are likely to have caused only small modifications to the soundings at Camborne. One significant effect at Camborne which sometimes occurs in an unstable westerly airstream is a decrease in the magnitude of low-level potential instability.

Apart from the careful siting of the principal observations to isolate the effects of topography, and the selection of a weather system travelling from a suitable direction (west-south-west), the data used in this study are fairly standard. As is usual in studying synoptic scale phenomena, we have been faced with the problem of making the best possible use of an inadequate spatial distribution of upper air data whose accuracy is only marginally adequate (especially in regard to relative humidity). Following the normal meso-analysis procedure, we have simulated a better spatial distribution by using sequential data from key stations (C and D). These data have been interpreted in the context of a description of the system based upon true spatial cross-sections. The selection of appropriate cross-sections is critical for, just as the silhouette of an elephant resembles the actual animal only when viewed from a single aspect, so too an atmospheric entity must be properly sectioned to reveal its salient features.

3. THE CASE STUDY: SYNOPTIC SITUATION

A wave depression which was just south of Newfoundland at 00 GMT on 4 December moved very rapidly east-north-eastward, reaching Scotland by 12 GMT on 5 December. During the 24-hr period up to 00 GMT* on 5 December the surface pressure at the low centre

* Henceforth all times refer to 5 December.

D
Figure 2(a). Conventional surface analysis for 12 GMT on 5 December 1972. Solid lines are sea-level isobars at intervals of 4 mb. Stippled shading denotes the approximate extent of rain.

Figure 2(b). Synoptic situation at 12 GMT on 5 December 1972. Solid lines are sea-level isobars at intervals of 16 mb. Dashed lines are isopleths of 1000–500 mb thickness at intervals of 3 decametres. Heavy arrows represent the axis of the upper tropospheric jet stream; speeds are indicated in m s$^{-1}$. The jet core was at 300 mb over Ireland; further east it broadened, the northern part descending to 400 mb over the North Sea and the southern part ascending to 230 mb over France. The shaded area denotes extensive upper cloud.
decreased by as much as 40 mb but during the next 24-hr period the central pressure remained almost constant while the pressure in the warm sector continued to fall slowly (typically 0-3 mb hr⁻¹). Fig. 2(a) shows the conventional frontal analysis for 12 GMT and a simplified depiction of the extent of rain. Although the depression was characterized by strong baroclinic zones, the fronts were indistinct and the conventional frontal analysis is rather arbitrary. The surface warm front (SWF) in Fig. 2(a) marks the warm side of a zone 300 to 500 km wide where the wet bulb potential temperature \( \theta_w \) rose from 4°C to 8°C behind it there was a more gradual rise of \( \theta_w \) to about 11°C at coastal stations. The surface cold front (SCF) was ill-defined, marking the leading edge of a gradual temperature fall associated with a broad and slow-moving baroclinic zone; in the conventional analysis it has been drawn to correspond to a burst of heavier rain at the end of the period of warm sector rain.

A more revealing analysis of the synoptic situation at 12 GMT incorporating some suggestions of Sawyer (1964) is shown in Fig. 2(b). The stippled shading shows the region of extensive upper cloud derived from satellite data. Part of this cloud area, ahead of the 543 decametre contour of the thickness ridge, was associated with the warm frontal zone; behind there was a long belt of upper cloud situated on the warm side of the slow-moving cold baroclinic zone. The main rain area, centred over central England at 12 GMT, was near the right-hand jet exit rather than the left exit. Although this is not what might have been expected on the basis of empirical forecasting rules (Benwell 1967) it is consistent with the finding of Lowndes (1968) that from September to February in the hills of north Wales most heavy rain days are associated with a right-hand jet-stream exit. An assessment of the large scale vertical velocity using an approximate form of Sutcliffe's (1947) development equation given by Sawyer (1952; p. 238) suggested very weak large scale descent throughout much of the warm sector. Isentropic analyses, on the other hand, suggested very weak ascent (\(< 2 \text{ cm s}^{-1}\)) in some parts of the warm sector. The version of the Meteorological Office 10-level model (Bushby and Timpson 1967) in use during December 1972 gave weak large scale vertical motion at all levels in the warm sector as it crossed England and Wales; vertical velocities were in the range \( +10 \text{ to } -10 \text{ mb hr}^{-1} \), with a slight preponderance of upward motion in the range 0 to \(-5 \text{ mb hr}^{-1}\), and the forecast dynamic rainfall rate throughout the warm sector was in the range 0 to 0.5 mm hr⁻¹. It appears, therefore, that none of these analyses indicated appreciable vertical motion, and the main conclusion to be drawn from them is that large scale vertical motion in the warm sector was very weak.

4. THE CASE STUDY: PRECIPITATION STRUCTURE

(a) Surface rainfall distribution

The depression of 5 December brought rain to all parts of England and Wales. A special feature was the long duration of the rainfall; rain began ahead of the SWF and continued for many hours in the warm sector. In most places the rain was intermittent but it fell almost continuously over the Welsh hills. Data from autographic raingauge stations have been analysed to provide the rainfall distribution in a strip 150 km wide across Wales and England, oriented along 250°. The totals were divided into three categories: (i) rain that fell in an approximately 3-hr period ahead of the conventionally analysed SWF, (ii) rain that fell in the broad warm sector, and (iii) a relatively heavy burst of rain of typically 1-hr duration that occurred close to the conventional SCF. The distribution of rainfall is shown in detail only for the warm sector rain (Fig. 3); the other rainfall categories are compared with the warm sector rainfall by means of a summary table. Fig. 3 shows that heavy falls of warm sector rain accumulated over the Welsh hills while eastern England remained almost dry. The total duration of warm sector rain inclusive of dry intervals in the region
analysed was about 8 hr; this implies mean hourly rainfall rates of 2 to 4 mm hr$^{-1}$ over the higher hills, 0·3 to 0·5 mm hr$^{-1}$ along the Welsh coast and over central England, and a trace over eastern England. The area of heaviest rainfall in Fig. 3 corresponds closely with the smoothed pattern of surface topography (to within 10 km).

Rainfall totals for the warm front, warm sector and cold front are compared in Table 1 for three regions. Regions selected are: (a) the south coast of Wales including Lundy Island, (b) the south Wales hills, and (c) Norfolk, near the east coast of England. Actually the south Wales hills reach an altitude of over 800 m, whereas the available autographic sites are located below 400 m, and so the data in column (b) may underestimate the higher rainfall totals on the hilltops. Pedgley (1970) has shown that the rainfall in deep valleys in an area of generally high ground does not necessarily underestimate grossly the rainfall over the neighbouring high ground. In the present case, however, measurements from non-autographic gauges showed that the highest parts of the hills experienced 40 per cent more rain than the highest autographic gauges over the 24-hr period starting 0900 on 5 December; unfortunately a further period of rain which occurred behind the cold front prevented any more quantitative use of the 24-hr totals.

TABLE 1. AVERAGE RAINFALL (mm) FOR WARM FRONT, WARM SECTOR AND COLD FRONT, INFERRED FROM FIVE REPRESENTATIVE STATIONS IN EACH OF THREE TOPOGRAPHICAL REGIONS. STATION LOCATIONS ARE PLOTTED IN FIG. 3 AS (a) CIRCLES, (b) TRIANGLES AND (c) SQUARES. ALTITUDES OF THE STATIONS ARE IN THE RANGE:
(a) 6–137 m, (b) 262–366 m, and (c) 23–60 m

<table>
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<th>(a) South Wales coast</th>
<th>(b) South Wales hills</th>
<th>(c) Norfolk</th>
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<tr>
<td>Warm front</td>
<td>4·3</td>
<td>12·5</td>
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<tr>
<td>Warm sector</td>
<td>4·4</td>
<td>25·6</td>
<td>Trace</td>
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<td>Cold front</td>
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Table 1 shows that the rainfall over the hills was much heavier than that upwind and downwind for all categories of rainfall. In the warm sector the rainfall increased by a factor of six over the hills compared with that upwind on the Welsh coast. At the warm and cold fronts the corresponding increase was rather less, between a factor of three and four. The subsequent decrease in rainfall downwind toward the east coast of England was greatest in the warm sector and least at the cold front. Because of the large spatial variability and short duration of the cold frontal precipitation, the values in Table 1 for the cold front are liable to large errors due to possibly unrepresentative gauge locations. The warm sector values, on the other hand, are thought to be more meaningful, especially since radar observations showed that the north-south distribution of precipitation approaching south Wales, integrated over the 8-hr period of the warm sector, was almost uniform.

The weather radars showed that the rain in the warm sector was broken up into regions of relatively heavy rain, so-called mesoscale precipitation areas (MPAs), with diameters 10 to 100 km. Throughout most of the warm sector the MPAs showed no tendency to be organized into bands. The MPAs existed over the sea as well as over the land. Individual MPAs often could be tracked across the entire radar network (see Fig. 1) over a distance of about 600 km (about 5-hr duration). Their direction of travel was from 250° and their speed varied from 100 to 160 km hr⁻¹ (28 to 44 m s⁻¹). At one time a speed differential of 60 km hr⁻¹ occurred over a north-south distance of only 100 km just ahead of the SCF,

Figure 4. z-t diagram showing movement of mesoscale precipitation areas (MPAs) along line GDF in Fig. 3. The x-axis represents distance from G along the direction of travel of MPAs (250°); t represents time at locations along the line GDF. Over the sea, areas of moderate rainfall have been inferred from radar. Surface rainfall intensity over land has been derived from a line of 18 autographic gauges, all located within 10 km of the section. The height of each gauge is shown at the foot of the diagram. The conventional surface warm front (SWF) and surface cold front (SCF) are also indicated. Dashed lines (based upon isohyets at more closely spaced intervals than shown in the Figure) represent the axes of areas of heavier rainfall.
the higher speeds occurring farther north. As we show later, the velocity of the MPAs corresponded to the winds at the 600 mb level (Fig. 9).

Fig. 4 is an \( x-t \) diagram derived using data from autographic raingauges and radar to show the variations of surface rainfall rate resulting from the movement of MPAs along a 250° line passing close to Gower (G) and Defford (D). The orientation of short-term rainfall maxima in the \( x-t \) diagram represents the speed of the MPAs. The pattern of rainfall is dominated by the passage of MPAs travelling with the mid-tropospheric winds (except close to and ahead of the SWF upwind of the south Wales hills where a few showers occurred travelling at the speed of the low-level winds). Near the Welsh coast and over central England the MPAs produced occasional bursts of moderate rain. Over the hills, however, the variation in rainfall intensity due to the passage of MPAs was modified by a stationary orographic component. Thus, although individual MPAs retained their identity as they crossed the hills, the maximum rainfall rate increased to 4 to 6 mm hr\(^{-1} \) during their passage.

Not only did the intensity of the rainfall from pre-existing MPAs increase over the hills, but new precipitation areas formed between them causing virtually continuous rain in the warm sector over the hills. The origin of these new precipitation areas is explained in Section 4(b). According to Fig. 4, the rainfall regime was similar throughout almost the entire warm sector. The primary difference between the character of warm frontal rain and the warm sector rain at the surface was in the higher intensity of the warm frontal rain when it first arrived at the Welsh coast and later again as it neared the east coast of England. Over the hills, the intensity of warm frontal rain was no greater than that associated with MPAs in the warm sector.

Although most of the warm sector rain occurred as irregular arrays of MPAs, there were two rather poorly defined bands of MPAs just ahead of the SCF. As shown in Fig. 4, one of these gave moderate or heavy rain over eastern and central England as well as over the hills (between 1730 and 1900 GMT) and this was somewhat arbitrarily identified as the conventional SCF. The other generally weaker band which occurred 2 hr earlier was classified along with the warm sector rain. An MPA within this band gave rise to the single swath of 5 to 10 mm rainfall that can be seen in Fig. 3 extending to Lincolnshire on the east coast. Even discounting this swath, Fig. 3 shows that the area experiencing moderate falls of say 5 mm extended farther downwind than upwind of the hills. That is to say, far from there being a rain shadow in the lee of the hills in the warm sector, there was a slight increase in rainfall there compared with that over the sea. As we show later, this was due to the downwind drift of precipitation generated in the middle troposphere by orographic uplift over the hills; the more familiar effect of the area of heaviest rainfall being tied closely to the hills was due to orographic growth at low levels.

(b) Vertical structure of the precipitation

The high speed of travel of the MPAs noted in the previous Section is consistent with the associated precipitation being generated mainly in the middle troposphere, or alternatively, where there is known to have been a large low-level increment of growth over the hills, it implies that the low-level growth was being seeded or triggered by the precipitation generated in the middle troposphere. This view is supported by RHI radar observations from Gower and Defford which showed that precipitation associated with the MPAs extended to 6 km (about 450 mb) with abundant convective generator cells between 4 and 6 km (Douglas et al. 1957; Wexler and Atlas 1959). Later we show that this convection was due to a distinct layer of potential instability in the middle troposphere. Away from the hills at least, the passage of MPAs produced negligible fluctuation of surface wind, indicating that the associated convection did not extend to the ground.
Precipitation within the MPAs was, according to observations with the high-power Defford radar, embedded in a deep and extensive area of ice crystal 'cloud' which reached up to 6 km (occasionally to 7–8 km) throughout most of the period of precipitation. This cloud was dense at about 6 km, i.e. at the top of the middle-level convection. A similar cloud canopy was also detected above a region of middle-level convection in a wet warm sector studied by Browning and Harrold (1969; see their Fig. 12). The intensity of the radar echo from the ice-crystal cloud within the warm sector diminished steadily with distance downwind of the Welsh hills, suggesting that the ice-crystal cloud gradually evaporated in a region of zero ascent or weak descent covering much of central England.

The Gower radar was not sensitive enough to detect the ice-crystal canopy; however, its three-dimensional scanning programme gave detailed information about the changes in horizontal and vertical extent of precipitation in MPAs as they travelled from over the sea to the hills of south Wales. An example of the development of precipitation as it approached the coast is shown in Fig. 5. Three MPAs - A, B and C - are represented; they were identified as individual entities on the basis of constant-altitude PPI maps reconstructed from

![Diagram](image)

**Figure 5.** Series of sections at approximately 11-min intervals showing vertical distribution of precipitation echo along a 200 km line through Gower (G) oriented along the direction of travel of mesoscale precipitation areas (MPAs). Each vertical section has been derived from a pair of full-gain RH1 photographs obtained with the Gower radar along azimuths 250° and 070°. As shown at the foot of the diagram, the region upwind of G lies over the sea whereas the region downwind of G is over the south Wales hills. Individual MPAs, A, B and C, are identified by hatched, stippled and cross-hatched shading, respectively, as they travelled from left to right in the manner indicated by the arrows. The radar data are not range-normalized; although this accounts for the echo often being detected to a higher altitude close to the radar, it only partly accounts for the rapid increase in the extent of MPAs A and B as they approached the radar. The surface precipitation patterns from MPAs A, B and C are labelled in Fig. 4.
RHI data at 4° azimuth intervals. According to Fig. 4 only C was giving rain at the surface when it crossed the coast. Although A and B did not begin to produce rain at the surface until about 20 km inland, Fig. 5 shows that the precipitation in A and B originated aloft up to 50 km upwind of G, giving first radar echoes between 3 and 4 km. (The simultaneous formation of numerous echoes aloft over a large range interval is best seen in the 1056 GMT section in Fig. 5.) A detailed analysis of first echoes revealed many such echoes forming aloft in the same general region: these echoes grew rapidly and gave precipitation that reached the ground over the hills. It is important to note that these echoes were being initiated preferentially in the region upwind of the hills rather than as a continuation of widespread precipitation initiation which might have been occurring farther out to sea. This is confirmed by the enormous increase in horizontal extent of echo aloft in this region as shown by Fig. 6.

Fig. 6(b) shows a comparison of the horizontal extent of precipitation echo at 3 km in the two areas whose locations are shown in Fig. 6(a). One of the areas (Box S) is over the sea, the other (Box H) is over the south Wales hills. The boxes are directly upwind and downwind from the Gower radar so that individual MPAs can be identified first in one box, then in the other. (The time scale in Fig. 6(b) has been adjusted so that peaks associated with a given MPA passing through the two boxes are superimposed.) The boxes are also

Figure 6(a). Map of south Wales and the Bristol Channel showing locations of Box S (over the sea) and Box H (mainly over the hills), respectively upwind and downwind of the radar at Gower (G). Low land is stippled; areas over 150 m (500 ft) are hatched. Precipitation echo data within Boxes S and H are shown in Fig. 6(b).

Figure 6(b). Percentage of area of Boxes S and H occupied by precipitation echo at altitude 3 km, derived from RHI data from the 43S radar at Gower operating at full sensitivity. The time scales for Boxes S and H have been adjusted to correspond to the time when the echoes would have passed Gower.
at equal ranges from the radar to avoid the effect of the dependence of echo dimensions on range. The apparent discrepancy between the occasionally small amounts of echo in Box H and the known occurrence of continuous rain over the hills is partly a result of the low radar sensitivity and of the radar measurements being made in snow above the melting level; it is also partly due to the occurrence of considerable orographic growth at low levels. However, the most interesting aspect of Fig. 6(b) is that, despite the modest extent of echo in Box H, the amount of echo in Box H is still far greater than that in Box S. The fractional increase amounts to a factor of 4·1 averaged over the warm sector compared with 1·5 for the cold front. This indicates that the orography triggered new precipitation areas extensively in the warm sector. Fig. 6(b) shows that some of the new precipitation areas developed in the vicinity of pre-existing MPAs whereas others were triggered at times when no pre-existing MPAs were detected in Box S. The only significant period with no echo aloft over the hills occurred at about 1630 GMT; Fig. 4 shows that at this time the surface rainfall rate over the higher hills decreased to a very low value (~0·3 mm hr⁻¹).

A feature of this situation that is typical of most orographic situations was the close spatial agreement between the heavy surface rainfall over the hills and the surface topography. This suggests that, although there were important orographic effects at middle levels, the main orographic component of rainfall originated at low levels in what Bergeron (1965) called the ‘feeder’ clouds. Assuming a mean 1 m s⁻¹ fallspeed for precipitation particles above the melting level, any increase in the precipitation intensity induced over the hills at, say, 4 km would have taken 2500 s to descend even to the melting level at 1·5 km, during which time the 35 m s⁻¹ wind would have carried it 90 km downstream of the hills. The inferred downstream displacement of less than 10 km for the area of heaviest rainfall is instead consistent with most of the orographic component originating below 1·5 km and falling to the ground as raindrops with a mean fallspeed of, say, 5 m s⁻¹ in the presence of the observed 25 m s⁻¹ low-level wind. As emphasized by Bergeron and by Sawyer (1956), orographic growth at such low levels cannot take place efficiently unless the area of concentrated orographic uplift is seeded by pre-existing particles at least 100 μm in diameter. In the present warm sector we have seen that abundant particles were available for seeding the low-level growth as a result of middle-level convection; some of this convection was triggered dynamically but most of it was triggered orographically. It is most significant that the orographically triggered middle-level convection developed far enough upwind of the hills to be effective in seeding the low-level orographic cloud directly over the hills.

We show later that many aspects of the precipitation distribution can be accounted for in terms of potential instability. Before considering the effects of potential instability on the precipitation and its interaction with orography, however, we describe next the distribution of potential instability itself and attempt to explain how it was generated.

5. The case study: upper air structure

(a) Three-dimensional structure

Fig. 7 shows the 1000–500 mb thickness analysis and surface fronts at 24 GMT on 5 December 1972 when the warm sector was retreating over Scandinavia. The purpose of this Figure is to show the location of cross-sections in subsequent figures that depict the three-dimensional structure of the airstreams entering and bordering the warm sector.

Figs. 8(a) and (b) are time sections at Camborne and Deford respectively; they show the distribution of wet-bulb potential temperature (θₑ), potential instability (dθₑ/dz < 0) and the location of layers of marked static stability. When converted to spatial representations using the synoptic scale system velocity of 23 m s⁻¹/255°, these sections lie along C₁C₂.
Figure 7. Routine 1000-500 mb thickness analysis for 24 GMT on 5 December 1972, showing locations of cross-sections in Figs. 8 to 10. Surface fronts are also shown.

Figures 8(a) and (b). Time-height sections at Camborne and Delford, respectively, equivalent to spatial sections along (a) C1C2 and (b) D1D2 in Fig. 7. Solid lines are isopleths of $\theta_w$ at intervals of 1 deg C. Hatched shading denotes potential instability. The dashed line in Fig. 8(a) indicates the axis of maximum $\theta_w$. Double lines represent boundaries of stable layers; these cap regions of convective mixing associated with the two layers of potential instability. Arrows at top of diagrams indicate times of soundings. Note different time scales in Figs. 8(a) and (b).
and D_1D_2 in Fig. 7. Both sections are situated mainly in the conventionally-analysed warm sector and extend from just ahead of the SCF into the warm baroclinic zone. C_1C_2, which is twice as long as D_1D_2, extends a considerable distance into the warm baroclinic zone.

Fig. 8 shows that, at first, potential instability (PI) occurred mostly at low levels beneath the warm frontal zone. In the warm sector at both Camborne (Fig. 8(a)) and Defford (Fig. 8(b)) there were two distinct layers of PI – one at low-levels, capped by a stable and relatively dry layer between 800 and 700 mb, the other at middle-levels, capped by a stable layer near 500 mb. Although the paucity of soundings at Camborne has allowed some freedom in constructing Fig. 8(a), there is nevertheless a marked similarity in the patterns of PI in the warm sector in Figs. 8(a) and (b). In view of this and the large number of soundings at Defford, we can be confident of the reality and representativeness of these two layers of PI. The occurrence of two similar layers of PI was also identified in the wet warm sector studied by Browning and Harrold (1969). Although the radiosonde data suggest that the middle-level PI was weak (\(\partial \theta_w / \partial z \approx -0.5 \text{ deg C km}^{-1}\)), limitations of the radiosonde humidity element imply it was significantly greater than this. This view is supported by the study of Browning, Hardman, Harrold and Parjoe (1973) in which a fully-instrumented Canberra aircraft reported cumulus-scale updraughts of up to 5 m s\(^{-1}\) in a rainband where radiosondes indicated middle-level PI comparable with that in the present study. As we show later, it is the middle-level PI that was responsible for the generation of convective precipitation in MPAs, but the low-level PI may also have had an important effect.

The two layers of PI were separated by a relatively dry statically stable layer. For this to have persisted despite prolonged erosion by moist low-level convection over the warm sea suggests the air had been subsiding at low levels during much of its passage across the Atlantic. This was probably true also in the wet warm sector studied by Browning and Harrold (1969) in which Doppler radar measurements revealed descending air motion of about 10 cm s\(^{-1}\) in the lowest 3 km during the passage of MPAs. Existence of this low-level subsidence is a necessary condition for permitting intense low-level PI to build up in the

![Figure 9](image-url)  
Figure 9. Cross-section along A_1 A_2 in Fig. 7. Solid lines are isotachs of wind component, u, along 250° at 10 m s\(^{-1}\) intervals. Dashed lines are isentropes at 10°C intervals. The shaded area is the main baroclinic zone within which the horizontal temperature gradient exceeds 2°C over 200 km. MM represents the approximate location of middle-level convection.
warm sector. The zone with PI at two levels in the warm sector was associated not with the
travelling warm baroclinic zone but, rather, with the trailing cold baroclinic zone oriented
along $240^\circ$ over England. The structure of this zone is best shown by means of cross-sections
at right angles to it. Sections $A_1A_2$ and $B_1B_2$ in Figs. 9 and 10 are true spatial sections;
$B_1B_2$ is based upon routine radiosonde soundings at 24 GMT from Long Kesh, Valentia,
Camborne, Brest and Bordeaux, upwind of the principal orographic influences, and $A_1A_2$
is based upon soundings at 24 GMT from stations downwind of the orographic effects.

Fig. 9 shows the distribution of wind speed and baroclinity. The patterns were much
the same in sections $B_1B_2$ as in $A_1A_2$ and only section $A_1A_2$ is shown. The baroclinic zone
was almost vertical below 350 mb, with an average horizontal temperature gradient of
2 deg C over 200 km in the stippled parts of the figure, exceeding 4 deg C over 200 km in
many places. A tendency for cold baroclinic zones to be vertical has been reported by
Elliott and Hovind (1965) for wet cold fronts in southern California. Schwerdtfeger and
Strommen (1964) have reported a vertical cold frontal zone but the baroclinic zone in the
present study was broad and relatively uniform without major hyperbaroclinic (frontal)
zones in it. Fig. 9 shows that the component of wind parallel to the baroclinic zone at low
levels was strong over a zone 300 km wide ahead of the SCF, exceeding 25 m s$^{-1}$ at the top

![Diagram](image-url)

Figures 10 (a) and (b). Cross-sections along (a) $A_1A_2$ and (b) $B_1B_2$ in Fig. 7. Solid lines are isopleths of
$\theta_a$ at intervals of 1°C. Hatched shading denotes potential instability. Pockets of relatively dry air are also
shown. Arrows at top of diagram indicate locations of soundings.
of the friction layer. In contrast with sharp ana-cold fronts (Browning and Pardoe 1973), there was no very strong horizontal gradient of this wind component at the SCF. The maximum horizontal shear on both sides of the strong low-level flow was only $5 \times 10^{-5} \text{ s}^{-1}$, and so frictional convergence near the SCF was slight. The dashed line MM in Fig. 9, following the surface $\theta = 27^\circ \text{C}$, indicates the approximate location of the convection cells in the MPAs. $\partial^2 u/\partial z^2$ was negative just ahead of the SCF but this does not appear to have prevented orographic uplift from occurring over a deep layer and triggering additional convection in the region of MM.

Figs. 10(a) and (b) show the distribution of $\theta_w$, PI and dry air (RH < 60%) within sections A, A, B, and B, B. The two sections are similar in a number of respects. For example, both intersect a deep tongue of air of relatively high $\theta_w$ which extends down to the surface ahead of the SCF with a pronounced decrease in $\theta_w$ on either side, especially to the north. Both sections also reveal the two layers of PI identified in Figs. 8(a) and (b). The low-level PI is mainly between the surface and 800 mb. The middle-level PI is centred between about 700 and 600 mb in a narrow strip, which was situated on the cold side of the tongue of air at high $\theta_w$ but mostly ahead of the SCF. Relative humidity was high at all levels except in a few pockets aloft particularly above the SCF. Among the several differences between Figs. 10(a) and (b) are the tongue of air of high $\theta_w$ ahead of the SCF, which is warmer in section B than in A, and the magnitude of $\partial \theta_w/\partial y$ in the low and middle troposphere above the SCF, which is larger in B than in A (the $y$-axis is taken parallel to A and B). Also, the low-level PI within 200 km on either side of the SCF decreased between sections B and A owing to its partial release during forced ascent over the hills. Figs. 8(a) and (b) show that the mean $\theta_w$-lapse in the lowest 2 km diminished from $-1.7 \text{ deg C km}^{-1}$ just ahead of the SCF at Camborne to about $-0.8 \text{ deg C km}^{-1}$ at the corresponding position at Dofford.

We have shown the existence of appreciable PI in the warm sector and have established its relationship to the cold baroclinic zone. The next problem is to show how the two layers of PI were generated.

(b) Differential advection and the generation of the potential instability

Fig. 11 is a hodograph of winds obtained with the precision tracking radar at Dofford at about 21 GMT. The flow was mainly from 248° below 500 mb but there were layers characterized by winds veered (V) or backed (B) with respect to the mean direction. Except for winds in the friction layer (B), angular deviations of the wind from 248° did not exceed 5° below 400 mb; they correspond to transverse flows of only $\pm 3 \text{ m s}^{-1}$. Although these differential flows were weak, they were a persistent feature of the part of the warm sector characterized by section D-D. They could be traced in six successive soundings throughout the entire period during which PI was occurring at two levels. Differential flows resembling B and V were also evident in the 18 GMT sounding at C, and so they are believed to have been a characteristic feature of the airstream. The same was not true for the backed flow B which was detectable only at Dofford; this flow was associated with the dense ice crystal canopy above the layer of middle-level PI and may have been due to the outflow from intensified middle-level convection over the Welsh hills.

The axes of the veered flows V and V corresponded to two distinct minima of $\theta_w$ in the vertical throughout the period 13 to 21 GMT at Dofford (Fig. 8(b)). This is hardly surprising since these data were obtained in a region where there was a strong horizontal gradient of $\theta_w$ transverse to the mean wind direction. Thus the differential thermal advection between flows V and B, and between V and B, was an important mechanism leading to the generation of the low-level and middle-level PI, respectively. The low-level PI was
enhanced by the diabatic heat source at the sea surface, especially when the low-level flow left the north-east coast of the USA with a temperature much lower than the sea surface temperature of the Gulf Stream. The middle-level PI, on the other hand, apart from relatively slow-acting radiative effects, was due almost entirely to differential advection. Although the transverse circulations responsible for generation of the PI at the two levels corresponded to a vertical wind shear component, $\partial v/\partial z$, of only about 6 m s$^{-1}$ over 1 km, the fact that it was operating in a region where $\partial \theta_w/\partial y$ was as large as 2 deg C per 100 km implies a large rate of destabilization, given by:

$$\frac{\partial}{\partial t} \left( \frac{\partial \theta_w}{\partial z} \right) = - \left( \frac{\partial v}{\partial z} \right) \left( \frac{\partial \theta_w}{\partial y} \right) = 0.4 \text{ deg C km}^{-1} \text{ hr}^{-1}.$$ 

This value is comparable with values measured by Elliott and Hovind (1964) in Pacific storms approaching southern California; however, the intense differential thermal advection in the Californian storms was maintained over deeper layers than the 100 mb layers in the present study.

6. **Discussion of the Dependence of Warm Sector Rainfall on Potential Instability**

We have presented a case study in which differential advection in an intense and nearly upright baroclinic zone led to generation of layers of PI (potential instability) in a warm sector in the low and middle troposphere. Miller (1955) has pointed out that differential advection leading to PI can give rise to higher spatially averaged rainfall rates when the instability is realized by large scale ascent than would occur in the presence of the same large scale ascent without PI. The gain in precipitation efficiency is achieved at the expense of a drying out of air descending outside the areas of convective updraughts. In the present
study large scale ascent in the warm sector was too weak to be detected but, nevertheless, PI was realized on the mesoscale both in isolated MPAs (Mesoscale Precipitation Areas) over the sea and by forced orographic ascent over hills. The release of this PI and the resulting localized outbreaks of convection led to prolonged rainfall at rates between 2 and 6 mm hr\(^{-1}\) over the Welsh hills and extensive light rain over central England.

A model illustrating the dependence of the rainfall distribution in the warm sector on the release of PI has been synthesized from the results in Sections 4 and 5. The model is shown in Fig. 12 and an explanation of the symbols used in the model is given in the Key.

![Diagram of Orographic Rain in a Warm Sector](image)

**Figure 12.** Model showing dependence of warm sector rainfall on potential instability and orography. See Key for explanation of symbols.

- **Mean streamlines within the strong west-southwest flow crossing the Welsh hills, drawn to be consistent with the observed pattern of precipitation development; although the precise form of the streamlines is arbitrary, notice that the middle-level air begins to ascend far upwind of the hills.**

- **Layer with rather high static stability separating the potentially unstable air at low levels from potentially unstable air at middle levels.**

- **Base of the region of high static stability that extends throughout the upper troposphere.**

- **Small scale convection occurring where the low-level or middle-level potential instability (PI) is realized by general ascent.**

- **Ice crystal (anvil) 'cloud' resulting from the middle-level convection and perhaps also, above 500 mb, from stable ascent over the hills.**

- **Precipitation trajectories relative to the ground, strongly inclined because of the high winds; the change in slope occurs at the melting level at about 840 mb.**

- **Middle-level convection within isolated MPAs due to areas of mesoscale ascent that occur in the warm sector even over the sea.**

- **Abundant middle-level convection triggered by orographic uplift over the hills, occurring as fresh outbreaks within and between existing MPAs.**

- **Decaying middle-level convection mainly associated with MPAs previously in existence far upwind of the hills (i.e. \(M_1\)).**

- **Rapid low-level growth of precipitation falling from aloft, producing a large increment in rainfall rate tied closely to the hills.**

- **Evaporation in the lee of the hills, decreasing the amount of precipitation from middle-levels that reaches the ground over central England; however, because of the enhanced generation of precipitation over the hills (\(M_2\)) widespread rain continues to fall up to 100 km downwind of the hills.**
PI occurred at two levels and we consider first the role of the middle-level PI. The two modes of release of this instability are represented in Fig. 12 by $M_1$ and $M_2$. $M_1$ represents the scattered MPAs with dimensions typically 50 km which existed over the sea; these gave rise to what might be called the dynamical component of rainfall in the warm sector. It is difficult to assess accurately the amount of surface rainfall due to $M_1$ because quantitative measurements are not available far enough away from orographic influences, but in the present study it appears to have been rather small. Earlier studies by Browning, Hardman, Harrold and Parode (1973) suggested that MPAs over the sea are characterized by mesoscale updraughts of about 20 cm s$^{-1}$ over a deep layer; outside these the vertical motion is relatively weak. The trend with height of the relative humidity in the present warm sector showed a tendency for the veered flows to be consistently drier than the (nearly saturated) backed flows with the result that the PI generated by differential advection was prevented from being realized except locally in areas of mesoscale ascent.

The rate of release of middle-level PI within scattered MPAs over the sea presumably remained in approximate balance with the rate of generation of PI by large scale differential advection and diabatic effects. Once the airstream encountered orographic uplift, however, the balance was upset; for a time the rate of release exceeded the rate of generation and fresh outbreaks of convection occurred extensively between existing MPAs. The first precipitation echoes from these fresh outbreaks were detected aloft (at 3 to 4 km) between 50 and 100 km upwind of the windward slopes of the south Wales hills, up to 50 km out to sea. This implies that triggering of extensive middle-level convection by lifting over the hills (modified by frictional effects near the coast) began even farther upwind over the sea ($M_2$ in Fig. 12); the orographically-induced ascent in the middle-level mean airflow pattern has been drawn accordingly in Fig. 12. Ascent aloft beginning far upwind of the hills has been predicted by Eliassen and Rekustad (1971) in a theoretical study of mesoscale mountain waves.

Maxima in the surface rainfall rate associated with the passage of individual MPAs travelled at the speed of the wind at their level of generation in the middle troposphere (100 to 160 km hr$^{-1}$). Some MPAs ($M_1$) could be tracked from the sea, over the Welsh hills and thence across England ($M_3$). Eventually they died out in eastern England, having persisted longer than other MPAs ($M_2$) that had been initiated orographically. The middle-level convective generator cells associated with all of the MPAs were embedded within a massive plume of ice crystal 'cloud' which trailed away from the Welsh hills, and this also gradually dissipated with distance downwind of the hills.

The middle-level PI was bounded above and below by statically stable layers. In the 200 mb layer adjacent to the surface, there was another layer of PI which had been permitted to build up as a result of the earlier passage of subsiding low-level air over a warm sea surface. Like the middle-level PI, this low-level PI was ripe for release on encountering orographic uplift. Whether the low-level convection was restrained by the stable layer centred at 750 mb so as partially to overturn only the lowest layer, or whether it occasionally penetrated into the middle troposphere, would have depended on the amount of forced lifting of the lower stable layer and on the rate of entrainment of dry air into buoyant elements penetrating into this layer. One of the Camborne soundings suggested that convective elements occasionally might have penetrated from low to high levels over the hills; however, the fact that the rainfall rate seldom exceeded 8 mm hr$^{-1}$ even over the hills indicates that this probably did not occur.

The low-level convection over the hills may have had two main effects. One effect, predicted theoretically by Elliott and Shaffer (1962), was probably to increase the vertical extent of the orographic ascent; this is especially likely in the case of a fast-moving moist airstream (Sawyer 1956) and would have had the effect of triggering the middle-level convect-
tion more effectively. Certainly the orographic ascent extended over a great depth in this study; however, further observational studies are needed to show whether the low-level instability plays a decisive role in this respect. Another effect of low-level convection may have been to give rise to an increase in the low-level increment in rainfall rate. On this occasion the low-level contribution to the surface rainfall over the hills, averaged over the entire warm sector, is estimated to have been typically 2 to 3 mm hr\(^{-1}\), but this can be attributed only partly to the actual convection at low levels since even a uniform saturated updraught of 10 cm s\(^{-1}\) over the lowest 2 km would have been sufficient to condense water at the rate of 1.5 mm hr\(^{-1}\). However, the very existence of any low-level PI in the first place implies that the low-level flow had been descending and subsaturated upwind of the hills and this will tend to have decreased the liquid water content in the orographic feeder cloud.

Although the surface rainfall in the warm sector was continuous over the hills, it reached a maximum intensity each time an identifiable area of middle-level convection passed overhead (Fig. 4). The amplitude of the modulation of rainfall rate due to this effect was about 2 mm hr\(^{-1}\) over the hills compared with 0.5 mm hr\(^{-1}\) over low land nearby; this indicates that the rainfall maxima over the hills were not simply due to the addition of the middle-level precipitation to an independent low-level source of precipitation. Rather, the middle level precipitation was effectively seeding the low-level clouds. As pointed out by Sawyer (1956) the low-level cloud, although rich in liquid water content, would have had insufficient time in the absence of such seeding for the cloud droplets to have been converted into precipitation in the period taken for air to cross the hills. The efficient conversion of low-level cloud into rain that fell directly on to the hills was ensured in this case by almost continuous generation of middle-level convection by orographic ascent that began upwind of the hills. Despite the long horizontal distance travelled by the middle-level orographic precipitation in falling to the ground in the strong winds, it was generated far enough upwind to be effective in seeding the low-level clouds over the hills. It is possible that, even in the absence of middle-level convection, orographically induced stable ascent would have generated precipitation aloft; however, it is doubtful in this event whether the particles would have grown fast enough to seed low-level clouds over the hills themselves.

Descent of air in the lee of the hills led to at least partial evaporation of existing precipitation, especially of the slowly falling ice particles at levels above 840 mb (E in Fig. 12). As in the cases described by Sawyer (1952), however, the resulting rain shadow was not as marked as might have been expected assuming the air to have descended in the lee of the hills (almost) to the extent to which it had been lifted. This is because the middle-level precipitation generated over the hills (M\(_2\) in Fig. 12) was carried up to 100 km downstream in the strong winds before reaching the ground and, partly because of its convective nature, the orographically-induced middle-level increment of precipitation exceeded the orographically-induced low-level evaporation in the lee of the hills. Thus not only the rain that fell over the Welsh hills but also the rain over central England can be regarded as being mainly an orographic component. Only the occasional MPAs that could be traced giving relatively high rainfall rates as they travelled from the Welsh coast to eastern England may be regarded as essentially dynamical in origin and the surface rainfall even from these was greatly enhanced by orography over the hills themselves. In some earlier studies, large falls of rain have been observed within well-defined swaths due to persistent passage of areas of heavy rain downwind of the south Wales hills (Browning and Harrold 1969; Harrold 1972). This effect was not observed to any marked degree in the present study and it must be left to future research to investigate whether such cases can be accounted for in terms of the nature of PI triggered by the hills.

The passage of MPAs close to the SCF resembled the passage of other MPAs in the
warm sector in that none was accompanied by significant surface wind shifts which would have indicated organized low-level convection. The main difference between the MPAs at the cold front and the MPAs in the warm sector was to be found, rather, in the greater vigour of the middle-level convection at the cold front away from the influence of the hills. This suggests that dynamical lifting in the middle troposphere may have been significant in the vicinity of the cold front even though it appears to have been negligible throughout most of the warm sector. The existence over the sea of a banded organization for the MPAs near the cold front, as distinct from the scattered distribution of MPAs in the warm sector, provides a way of identifying the existence of a well-organized dynamical lifting mechanism. Such precipitation bands can be expected to give rainfall in areas such as eastern England that may escape the orographic rain altogether.

7. Application to forecasting

The conclusion by Sawyer (1952) that the main effect of hills is to intensify rainfall in rain areas already established by dynamical causes needs to be qualified in the light of this study of warm sector rain. Although it is true that the orographic component in the present study occurred in a warm sector already characterized by scattered MPAs, the orographic effects not only increased the intensity of the rain in existing MPAs but also greatly increased the extent of rain outside these MPAs both over hills and up to 100 km downwind.

The importance of seeding from aloft for realizing heavy orographic rainfall has long been recognized. Sometimes the seeding is achieved by precipitation generated aloft by large scale ascent; sometimes, as in the present case study, the hills are able to generate their own high-level seeds even when they are not being generated widely by large scale ascent. In this study the occurrence of orographic precipitation appears to have been favoured by the existence of extensive middle-level PI which, because of earlier large scale ascent, required only a small amount of local ascent to realize it, but which at the same time was not being realized generally by large scale ascent. This finding is relevant to forecasting because rainfall predictions of most models are weighted heavily towards the predicted fields of large scale ascent. Thus, although it is certainly necessary to forecast the area of large scale ascent required to produce a deep moist layer in the warm sector, it is important to recognize that major orographic falls can occur in warm sectors in which large scale ascent has already virtually ceased by the time they reach the British Isles. A possible further implication is that, if a deep layer has been brought close to saturation by earlier large scale ascent, then continued large scale ascent in the warm sector as it approaches the British Isles might actually diminish the orographic component by realizing PI prematurely over the sea instead of allowing it to build up. Such an interpretation would be consistent with the suggestion of Lowndes (1968) that the heaviest orographic rainfalls occur within warm sectors that are not characterized by large scale ascent.

It is clear from the above that the interaction of PI with topography must be carefully taken into account if numerical prediction models are to succeed in forecasting orographic rainfall. Further detailed case studies will be needed to clarify this interaction. Unfortunately, direct observation of PI for forecasting purposes while a system is upwind over the Atlantic is difficult because of the narrowness of the potentially unstable part of the warm sector (~200 km). Therefore it is necessary instead to identify and predict the factors leading to the pronounced differential advection that creates the middle-level PI.

8. Future studies

The case study in this paper is one of a series of investigations into the sub-synoptic
scale structure of different categories of rain-producing situations. The broad objectives of our research are to clarify the physical mechanisms responsible for the observed spatial distribution of rainfall and to derive models that in the long run may prove useful for local forecasting. If these objectives are to be met, further studies will be needed using improved radiosonde and precipitation surveillance facilities. Two complementary programmes are planned by the Meteorological Office. One involves airborne surveillance of systems over the sea to exclude the effects of topography. The other is an extension of the present study and depends largely on ground-based facilities to observe precipitation systems before, during and after their passage over the hills of Wales and south-west England. The principal facilities needed for the latter programme are outlined below.

(a) *The radiosonde network*

The present study has revealed deficiencies in the normal mesoanalysis procedure of converting sequences of soundings from individual stations to spatial cross-sections when the precipitation system has a small component of motion normal to its axis. Since slow-moving baroclinic zones such as the one in this study tend to give rise to the worst situations of prolonged heavy rain, this is an important category of event requiring further investigation. Consequently it is desirable to increase the number of radiosonde stations during project periods especially upwind of the main hills. It is hoped to achieve this partially by using mobile radiosonde units.

(b) *The weather radar network*

The task of manually compositing the data from a network of radars and of subtracting out spurious echoes from the ground renders the radar analysis time-consuming. It is desirable, therefore, to make maximum use of automatic processing techniques to enable larger numbers of situations to be studied in the future. A development programme is now under way at the RRE (Taylor and Browning 1974) by means of which an on-site computer at each of three radars will ‘clean’ the data and transmit digital information on the areal distribution of surface rainfall intensity in rectangular Cartesian format via telephone lines to Malvern where a composite colour television display will be generated in real-time with facilities for ‘action reply’. The resulting integrated ‘mini-network’ of weather radars will provide not only data for the basic precipitation research studies but also practical experience for the possible later implementation of a more extensive operational network that would provide real-time data on the areal rainfall distribution for a variety of users.

Detailed forecasts of local rainfall can be made for only a very limited period by objectively extrapolating existing rainfall patterns; however, it is expected that knowledge of the sub-synoptic scale structure and mechanisms of precipitation systems gained through the combined use of radiosonde and radar networks will enable us to anticipate the development of rainfall patterns over longer periods. In the long run, the operation of a Weather Radar Network would create a computer-archived body of mesoscale experience that would enable local rainfall conditions to be related empirically to parameters forecastable on the larger scale (100 km) using existing numerical-dynamical techniques such as those of Bushby and Timpson (1967).

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