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1. INTRODUCTION

A major problem in the provision of short period weather forecasts, that is forecasts for up to 24hr ahead, is the estimation of local rainfall which may be modified by local topographic effects. An important operational requirement is the forecasting of 6hr accumulations of rain in hilly areas. Methods of forecasting based on the extrapolation of the movement of precipitation echos derived from radar observations are only useful for forecasting for a short period ahead and even these forecasts will require modification if local topographic effects are large. The resolution of the present numerical weather forecasting models is too coarse to allow the effects of small scale topographic features to be incorporated into the rainfall forecasts produced by these models but it is probable that sub-grid scale topography does not affect the larger scale flow. We can therefore regard the effects of small scale topography as a process which need not be incorporated into the numerical models but which must be considered when interpreting the output of the models. A further problem with interpreting the output of these numerical models to provide short period forecasts is that, because of the methods used to derive the initial data for the models, the properties of the forecast parameters are not constant throughout the forecast period. These changes are very evident in the first few hours of the forecast. This problem will not be too important because of the time taken after the nominal data time to derive the numerical forecast. This period is of the order of 3hr and it seems probable that for a local forecast valid at, for example, 1800GMT to be of any use it would be necessary to interpret an 18hr forecast based on 0000GMT data rather than a 6hr forecast based on 1200GMT data. The forecast based on the later data time would be issued too late to be of practical use although it is possible that later observations could be incorporated into the method of interpretation. Although it is desirable to improve the early behaviour of the numerical forecast models this will not in practice limit the use of the output of large scale models as input into a local rainfall forecasting model.

Collier (1975) has formulated a model of orographic rainfall which has been used, with some success, to estimate rainfall accumulations in North Wales using observations as input data. As Collier (1976) points out the model can over-estimate precipitation in situations where the orographic clouds are of limited extent as under these conditions the droplets do not have sufficient time to grow to precipitation size. In these cases the model of Bader and Roach (1976) is more applicable as this calculates orographic precipitation enhancement assuming that rain from upper levels falls through, and removes water from, low level orographic cloud. An additional deficiency of Collier's model is that it assumes after the slightest ascent, all of the depth of the model will be saturated with precipitation from all levels where there is rising motion. In contrast to this simple approximation to the various cloud physical processes, the model proposed by Collier provides for a sophisticated method of estimating the vertical velocity in the vicinity of topographic features. The results suggest that, when observations are used as data, this treatment provides an improvement in the accuracy of the accumulations of rainfall over areas of a few tens of square kilometres. However, the accuracy achieved using forecast large scale data would be determined to a great extent by the accuracy of the large scale numerical forecast and it was thought that such sophistication was, at least initially, not necessary and that the cloud physical treatment should be improved.

This note describes a simple model of orographic rainfall intended for use in the interpretation of numerical forecast output. It uses a simple formulation for obtaining the local vertical velocity with the formulation of the cloud physics including the effects of finite cloud extent and the effects of the wash-out from low level cloud. The model is intended for investigations of the feasibility of using simple, essentially one dimensional, models to interpret numerical forecasts and is designed to use data of the type available from numerical forecasts. The results can be compared with those obtained using climatological or statistical techniques or with those obtained from mesoscale primitive equation models.

2. THE RAINFALL ENHANCEMENT MODEL.

Several assumptions are made in developing the model of rainfall enhancement. As has been mentioned it is assumed that the data is representative of the large scale flow and that these values are in some way 'mean' values of the flow in the absence of topography. We consider that the local perturbation of the flow due to topography is only to introduce an additional component to the vertical velocity and to modify the relative humidity at fixed pressure levels. The effects of topography are assumed to decrease to zero above 500mb and it is also assumed that the relative humidity can be converted to a mixing ratio at a particular level assuming that the vertical temperature structure is that for an ICAO standard atmosphere although this approximation is easily removed. It is assumed in the present work that precipitation falls vertically although the effects of incorporating precipitation drift will be discussed later, and that there is no evaporation of precipitation.

We can write the equation governing the local rate of precipitation derived by Collier (1975) in pressure coordinates in the form

$$P = -\omega \, \partial r_s / \partial p \qquad (1)$$

where ω is the vertical velocity and $\partial r_s / \partial p$ represents the derivative of saturation mixing ratio with respect to pressure. Collier assumed that $P \equiv 0$ when $\omega \geqslant 0$, that is when there was local descent. In the present model we assume that at any level there are two contributions to the precipitation. The first, P_1, results from the accretion by precipitation of cloud water at the level. This can be written in the form

$$P_1 = P_1(P', r, r_s) \qquad (2)$$

that is, it is a function of the precipitation P' falling through the level and the cloud water content, assumed to be determined by the relative humidity r/r_s. The precipitation formed at a level is written in the form

$$P_2 = -k_1 k_2 \omega \partial r_s / \partial p \qquad (3)$$

where k_1, takes the value 0 if $\omega \geq 0$ or if $r/r_s < 1$. In this way the possibility of producing precipitation from layers which are not saturated, even after they have been raised by topographic effects, is prevented. The constant k_2 is introduced to take account of the effects of the finite extent of the cloud. For cloud of limited extent it is assumed that $k_2 \to 0$ while for more extensive cloud $k_2 \to 1$. It has been assumed that, although the relative humidity used for testing for saturation is representative of a deep atmospheric layer, no precipitation forms until the local relative humidity r/r_s exceeds 100% after topographic uplift. In the atmosphere condensation will occur at a lower mean relative humidity but this is offset by the retention of water in the cloud, neglected in Collier's model. The present approximation is equivalent to taking both of these factors into account. For example if all of the water in excess of a mean relative humidity of 85% is assumed to condense but 15% of the total water content is retained in the cloud we would expect, in an infinite cloud, $k_2 \approx 1$.

In the model the surface rainfall is calculated by summing the contributions P_1 and P_2 from a series of levels at 100mb intervals from 550mb to 950mb. The contribution from each layer is obtained from a series of approximate formulae and requires knowledge of the large scale horizontal and vertical velocity and of the relative humidity at each level in addition to the local topographic height and the gradient of the topography. For the present model as has been mentioned the precipitation is assumed to fall vertically and there is no evaporation in subsaturated layers. The effects of washout on the water content within a cloud is treated in an approximate manner. The model uses a $3\frac{1}{3}$km horizontal grid which is larger than that suggested by Collier (1976) but is convenient for the present calculations since topographic data was readily available on this grid.

It was assumed in these calculations that the vertical velocity is given by

$$\omega = \omega_L + \omega_T \qquad (4)$$

where ω_L is the large scale vertical velocity and ω_T is a topographically induced component. The latter was assumed to be given by $-k_3 \underset{\sim}{V} \cdot \nabla H$ where the factor k_3 was assumed to vary linearly with pressure from zero at 500mb to 0.114mb m^{-1} at 1000mb. $\underset{\sim}{V}$ is the horizontal velocity at the level and ∇H is the gradient of the topography. It is clear that a more complete model would include the effects of stability on the magnitude and vertical extent of the ascent.

For simplicity an empirical relation was derived for the variation of saturation mixing ratio with pressure. For an approximately ICAO atmosphere we have

$$\frac{\partial r_s}{\partial p} \approx 10^{-2} \left(\frac{p}{300} - 0.7 \right) \qquad (5)$$

where $\partial r_s / \partial p$ is expressed in $g\ kg^{-1}\ mb^{-1}$ and the pressure p is in millibars.

In order to calculate the precipitation it remains to calculate the constants k_1 and k_2. The constant k_1 is set to one if $\omega \leq 0$ and if after ascent the air becomes saturated. Assuming that the ascent due to topography decreases from H for air initially at 1000mb to 0 for air initially at 500mb, and that the temperature profile is similar to the ICAO profile it is possible to derive empirical relations for the increase in the relative humidity due to forced ascent over hills of arbitrary height assuming that no water is lost. There are approximately linear relationships, at least for hills of less than about 800m height. The relative humidity of the air after ascent, h_a, is given in terms of the large scale relative humidity h_o and the topographic height H

$$h_a \approx h_o + f_1(p) H \tag{6}$$

where the values of $f_1(p)$ are given in the table below:

p (mb)	950	850	750	650	550
$f_1(p)$ (m^{-1})	4.4×10^{-4}	4.0×10^{-4}	3.4×10^{-4}	2.45×10^{-4}	0.95×10^{-4}

Using (6) h_a can be calculated and if $h_a < 1.0$, k_1, is set to zero.

Estimation of k_2 is rather more difficult since it involves knowledge of the extent of the cloud. For the present calculations a rough approximation can be made. We assume that the time taken for a drop to grow in the cloud is given by

$$t = -(h_T - h_s)/\omega \tag{7}$$

where h_T is the extent of the topographically induced ascent and h_s is the height at which the air becomes saturated and it assumed that the droplets follow the air motion. The approximation arises from the use of the local value of ω rather than the mean value during the motion through the cloud and will result in an overestimation of the life of a drop in regions close to the maximum topography. We know that $h_T = \rho g H(p-500)/500$ and that the topographic height H' over which the air just becomes saturated is $H' = (1-h_o)/f_1(p)$. If we assume $h_s = \rho g H'(p-500)/500$ we have

$$t = -\frac{\rho g}{\omega}\,\frac{(p-500)}{500}\left[H - \frac{1-h_o}{f_1(p)}\right] \tag{8}$$

But $H f_1(p) = h_a - h_0$ hence

$$t = -\rho g \frac{(p-500)}{500} \left[h_a - 1 \right] \Big/ (\omega f_1(p)) \tag{9}$$

which can be written in the form

$$t = (h_a - 1) f_2(p) / (\omega f_1(p)) \tag{9a}$$

where the values of $f_2(p)$ are given in the table:

p (mb)	950	850	750	650	550
$f_2(p)$ (mb m^{-1})	-10.12×10^{-2}	-7.97×10^{-2}	-5.62×10^{-2}	-3.37×10^{-2}	-1.12×10^{-2}

The value of k_2 is determined by t. If t is less than 5 min, k_2 is set to zero and if t is greater than 20min, k_2 is set to 1. A linear interpolation is made between these values.

Equation 2 gives the rate of removal of water in the form of precipitation in terms of mixing ratio changes. This can readily be converted to the precipitated water from a 100mb deep layer assuming that ω is constant throughout the layer. Incorporating this conversion into the approximate equation 5 for $\partial r_s / \partial p$ we can write

$$ppn = k_1 k_2 f_3(p)\, \omega \tag{10}$$

where $f_3(p)$ is given in the table and ppn is the precipitation rate from the layer due to P_2.

p (mb)	950	850	750	650	550
$f_3(p)$ (mm mb^{-1})	-2.47×10^{-2}	-2.13×10^{-2}	-1.80×10^{-2}	-1.47×10^{-2}	-1.13×10^{-2}

A simple approximation to the enhancement of rainfall falling through a lower level cloud can be obtained assuming that the drops collect all of the cloud droplets lying in their path. In this case it may be shown that for monodispersed raindrops of initial radius a_0 the ratio of the final rainfall to the initial rainfall is

$$\text{enhancement ratio} = \left(1 + \frac{l \rho d}{4 \rho_w a_0}\right)^3$$

where l is the cloud water mixing ratio, ρ and ρ_w the densities of air and water and d is the depth of the cloud layer. Assuming that $a_0 \approx 250\mu m$ and that for 100mb deep layers $\rho d = 10^3 \ kg \ m^{-2}$ we have

$$\text{enhancement ratio} \approx (1 + l)^3 \tag{11}$$

where l is expressed in $g \ kg^{-1}$. For water contents of about $1 \ g \ kg^{-1}$ we see that an eightfold enhancement is possible for rain falling through a 100mb deep cloud layer. There remains the problem of calculating l. For the present calculations we assume that the contribution to the total mixing ratio in excess of 90% of the saturated mixing ratio is condensed. The liquid water mixing ratio is assumed to be limited by precipitation and washout in a layer to 10% of the saturated mixing ratio and that this figure is achieved when the mean relative humidity of the air after forced ascent reaches 110%. We have therefore

$$l = k_4 r_s \tag{12}$$

where $k_4 = 0$ for $h_a < 0.9$, $k_4 = 0.1$ for $h_a \geq 1.1$ with $k_4 = (h_a - 0.9)/2.0$ otherwise. The values of r_s used are those appropriate for an ICAO atmosphere. While this formulation has been used in the present calculations it is clear that it will tend to overestimate the amount of washout in occasions where heavy precipitation falls through lower cloud. In these conditions the liquid water content will be reduced by the washout and will tend to zero for high precipitation rates. A formulation which would be more accurate in these conditions would decrease ϵ as the precipitation rate increases and attempts are being made to improve this aspect of the model.

The local orographically enhanced precipitation can be obtained by calculating the precipitation falling from the top layer of the model, allowing for enhancement in the next lower layer and adding the contribution to the precipitation from the layer. The process is repeated for all of the model layers until the precipitation reaching the surface is obtained. The results of calculations using this very simple model will be described in the next section.

3. RESULTS

The model has been used to calculate orographic rainfall over Wales using a $3\frac{1}{3}$km grid. The area covered by the study is shown in Fig.1 which also shows the topography which was used in the calculations. In order to test the model real data based on that presented by Collier (1975) was used as this enables the predicted and observed rainfall to be compared. It was assumed that the large scale parameters used as input to the model were representative of conditions for the whole area for a 30 min. period. Rainfall accumulations were then derived from the mean rainfall rates predicted by the model for successive 30min periods. The time sections of the input wind components, vertical velocity and relative humidity are shown in Fig.2. In order to obtain results which would be comparable with those of Collier, who used vertical velocities averaged over 6km squares, the grid point values of rainfall accumulation were smoothed using a 1-2-1 numerical filter.

The rainfall accumulations obtained using the model are shown in Fig.3. The results demonstrate the effects of small vertical displacements of the air, which would not be incorporated into synoptic scale models, on local rainfall. Although the large scale relative humidity does not exceed 90% the model predicts rainfall over almost all of Wales. Over most regions the accumulations are small, less than 2mm but much heavier accumulations are predicted over the highest areas. In particular accumulations of 30mm are predicted for the $5\frac{1}{2}$hr period over Snowdonia while accumulations of 20mm are predicted over the Brecon Beacons and in the area to the southwest of Llandegla for which the rainfall observations were presented by Collier (1975). It is difficult to obtain objective verifications of the rainfall except for the area analysed by Collier but it can be argued that there is qualitative agreement between the predictions of the present model and the precipitation distribution which might be expected.

In Fig. 4 a comparison is made between the rainfall accumulations predicted by the present model (a), the rainfall observations (b), and the accumulations predicted by Collier (1975) for the 50x30km area. It can be seen that the variation between the rainfall maxima and minima is much greater in the present model than was observed or was predicted by Collier. It is thought that this results from the neglect, in the present model, of the horizontal drift of precipitation. This has the effect of smoothing the rainfall distribution over distances of several tens of kilometres along the direction of the wind.

As has been mentioned the model used by Collier (1975) did not include the effects of washout but it was assumed that all of the layers were saturated by the slighest ascent so that precipitation fell from all of the layers within which there was rising motion. The finite growth time of the drops was also neglected. The present model was modified by the removal of the washout term and of the factors relating to the relative humidity and droplet growth that is, the constant k_4 in Eq.12 was set to zero and the constants k_1 and k_2 in Eq.10 were set to one. These changes make the present model similar to that used by Collier in the treatment of precipitation, except for the neglect of drift, although it contains the simplified treatment of the vertical velocity. The results obtained with the modified model are shown in Fig.5. It can be seen that the minima in the rainfall accumulations are greatly enhanced by the simplifications and the model predicts accumulations of 5mm over the sea. This results from the much deeper layer from which precipitation is assumed to fall and the general upward motion of the air. These effects more than compensate for the neglect of washout in this version of the model. The rainfall maxima are only slightly increased in the simplified formulation but close inspection of the results shows that they have been moved from the areas of high topography and low gradients to the regions of maximum topographic gradient. This is to be expected since washout depends not on the local vertical velocity but on the cloud water content while the direct precipitation depends on the vertical velocity and not, in this experiment, on the vertical displacement of the air. Inspection of Fig.5 also shows that the rainfall accumulations on the downwind slopes of the mountain are too low suggesting again that the observed precipitation in these areas has drifted from the areas of high rainfall production rate.

A more direct assessment of the effects of washout on precipitation was made by rerunning the original model with the effects of washout only removed. The results of this calculation are presented in Fig.6. The rainfall accumuldation when washout is not included seldom exceeds about 2mm. As would be expected however the regions of the maxima are the same as those calculated using the full model and the regions over which some precipitation is expected are the same using both versions of the model.

4. DISCUSSIONS AND CONCLUSIONS.

The results presented in this note have demonstrated the sensitivity of orographic rainfall to the microphysical processes. In contrast to this, the broad similarity between the forecast results and those of Collier (1975) suggests that a sophisticated dynamical model is not necessary to obtain realistic rainfall accumulations; it could be argued that the present results are better since the orientation of the isopleths in Fig.4(a) is closer to the observed orientation than the results of Collier.

A major deficiency of the present model is the neglect of precipitation drift and an experiment has been carried out to assess the importance of this on the precipitation accumulations. It was assumed that the precipitation fell through each of the lower two 100mb deep layers in 200s and through the upper two layers in 1000s. The layer from 700mb to 800mb was assumed to be traversed in 600s.

Using these figures the horizontal drift of the precipitation was allowed for and with the data described earlier the rainfall accumulations shown in Fig.7 were obtained. It can be seen by comparing these results with those of Fig.3 that the drift has a considerable smoothing effect on the precipitation accumulations and that the maxima are in reasonable agreement with the observations presented by Collier (1975).

It has been demonstrated that data derived from radio-sonde observations can be used with the simple model presented in this note to predict orographically enhanced rainfall accumulations. The data was chosen to be consistent in type and vertical resolution to the data which is obtained from the synoptic forecasts using the Meteorological Office 10-level model (Benwell et al., 1971) and it is suggested that the present model could be used in conjunction with the forecast data to obtain short period rainfall forecasts. Work is at present being carried out on a scheme for interpolating the synoptic scale model data which is derived at 6hr intervals to the 30min intervals used in the present model. Consideration is also being given to a spatial interpolation over the grid of the present model rather than the use of the same large scale parameters at all grid points as in the calculations described here

Various improvements are being made to the model. In view of the sensitivity of the results to washout attempts are being made to improve the formulations of this process and the effects of evaporation are being investigated. The limitation of the cloud water content by the washout process is a further subject for investigation. The vertical motion has been assumed to be independent of stability and this could be improved. The removal of water as precipitation will clearly have some effect on the downwind precipitation and this could be important when calculations are carried out over an extended area. This was neglected by Collier (1975) using observations as data but when using large scale forecast data it may be sufficient to modify the large scale humidity field to take account of the local removal of water.

The orographic influence on rainfall from large scale systems has been considered in this note but such systems often contain meso-scale areas of intense precipitation (see for example Browning and Harrold, 1969). The use of conventional data or of synoptic scale forecast data with the present model cannot enable the rainfall from such systems to be calculated. It may be possible to estimate the movement and development of these systems from radar observations and if this is possible they could be treated as perturbation to the larger scale flow which could be incorporated into the scheme for interpolating the synoptic scale data onto the present model grid. It may therefore be possible to estimate rainfall from such systems with a model of the present type whereas the use of a primitive equation meso-scale model could involve difficulties with the initialisation procedures when small scale features are present as well as being expensive in terms of computational resources.

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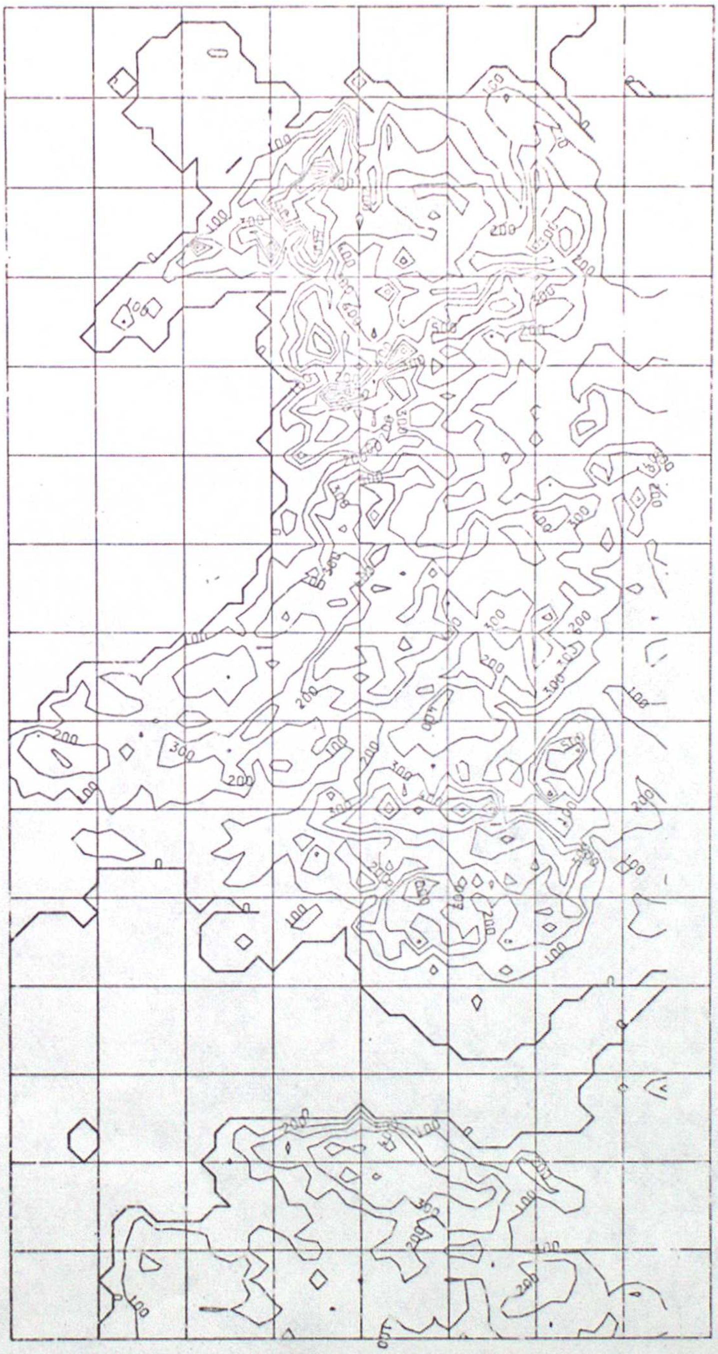
[15].

LIST OF FIGURES

Fig 1

CON INT=100 METRES
TOPOGRAPHY



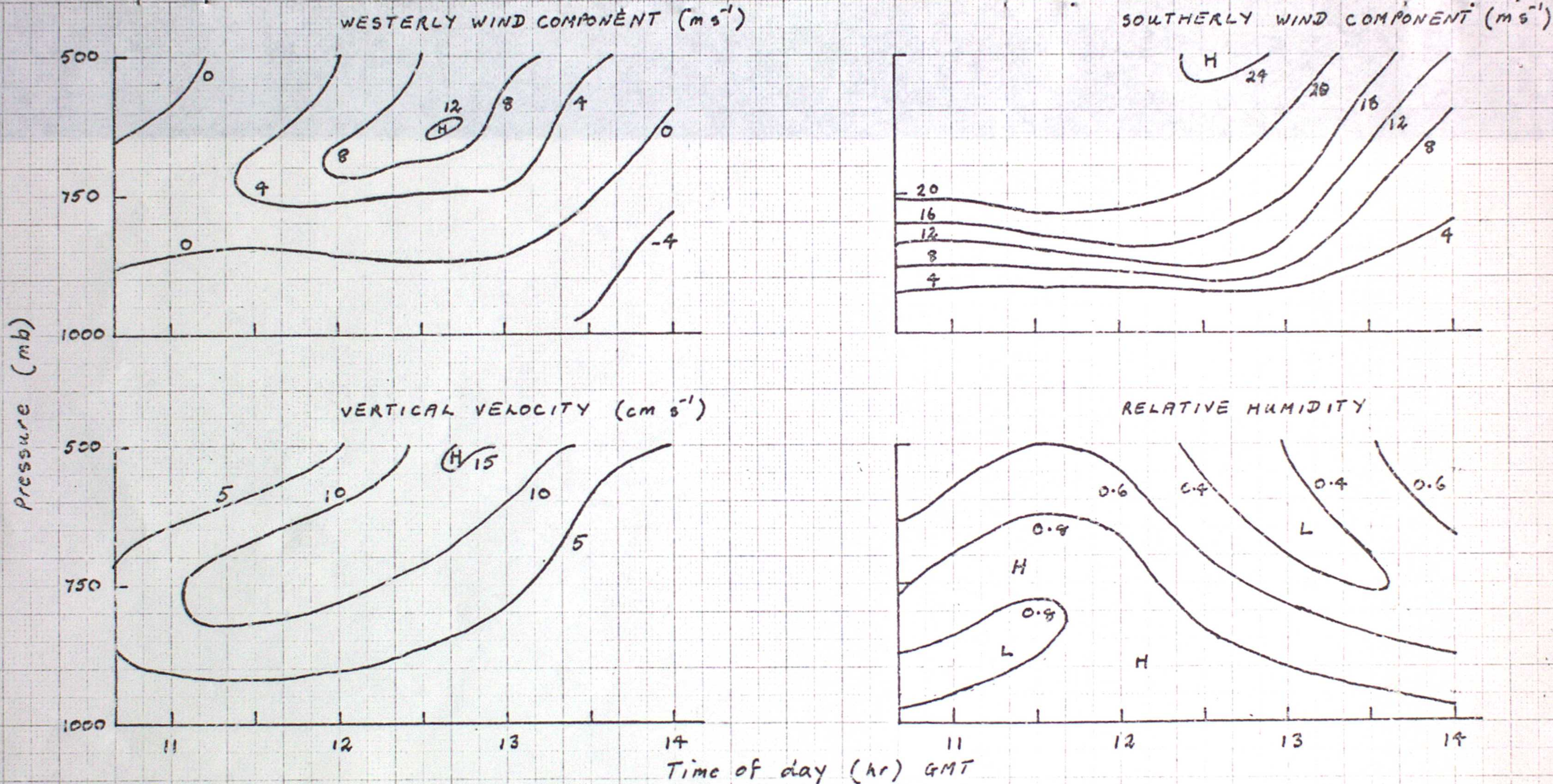

WESTERLY WIND COMPONENT (m s⁻¹)
SOUTHERLY WIND COMPONENT (m s⁻¹)
VERTICAL VELOCITY (cm s⁻¹)
RELATIVE HUMIDITY
Pressure (mb)
500
750
1000
11
12
13
14
Time of day (hr) GMT

Fig 2

Fig 3

5 LEVEL&WASHOUT&AVERAGED
TOTAL RAIN IN 3.5HRS ISOHYET 2.0MM

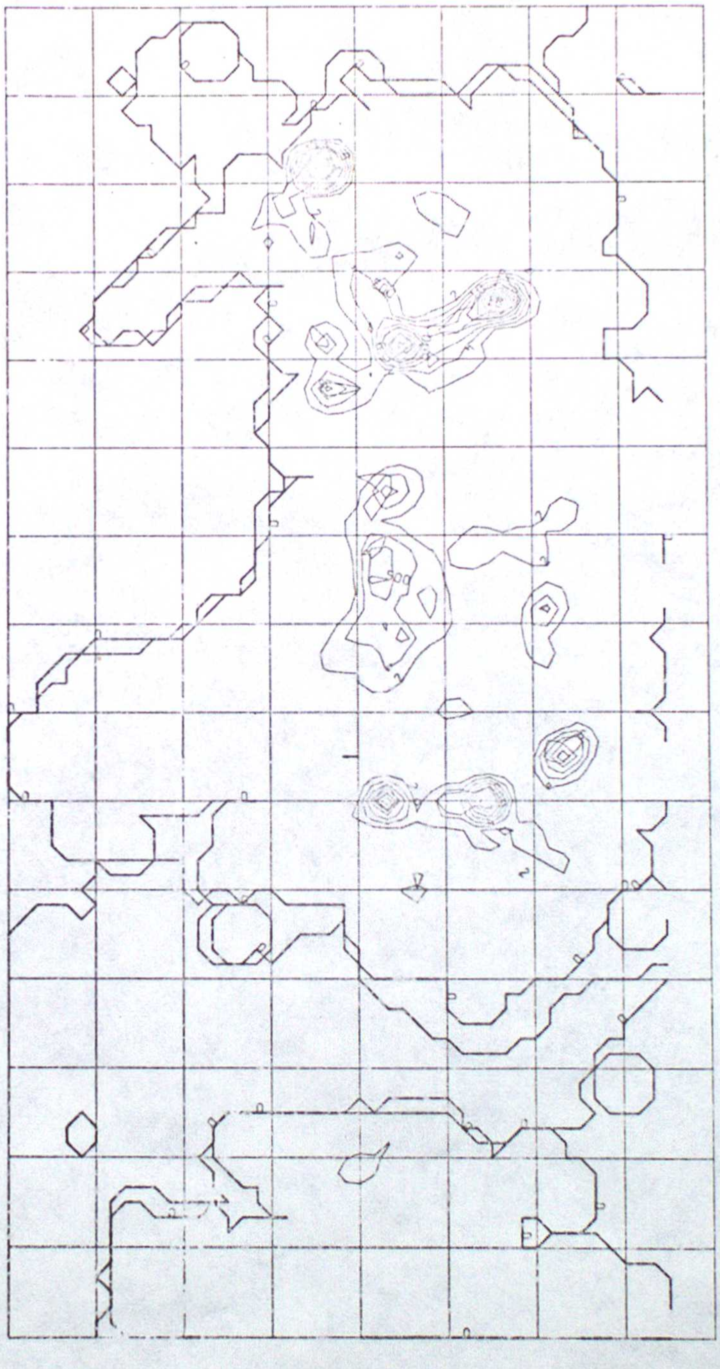
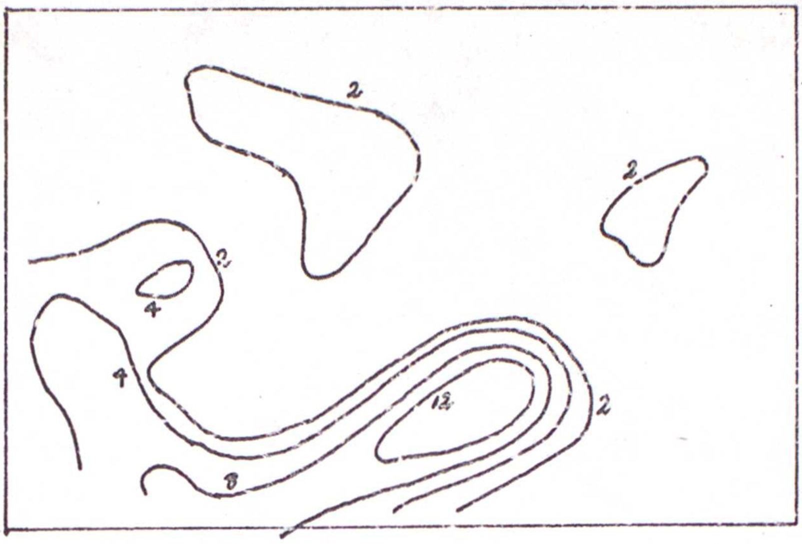
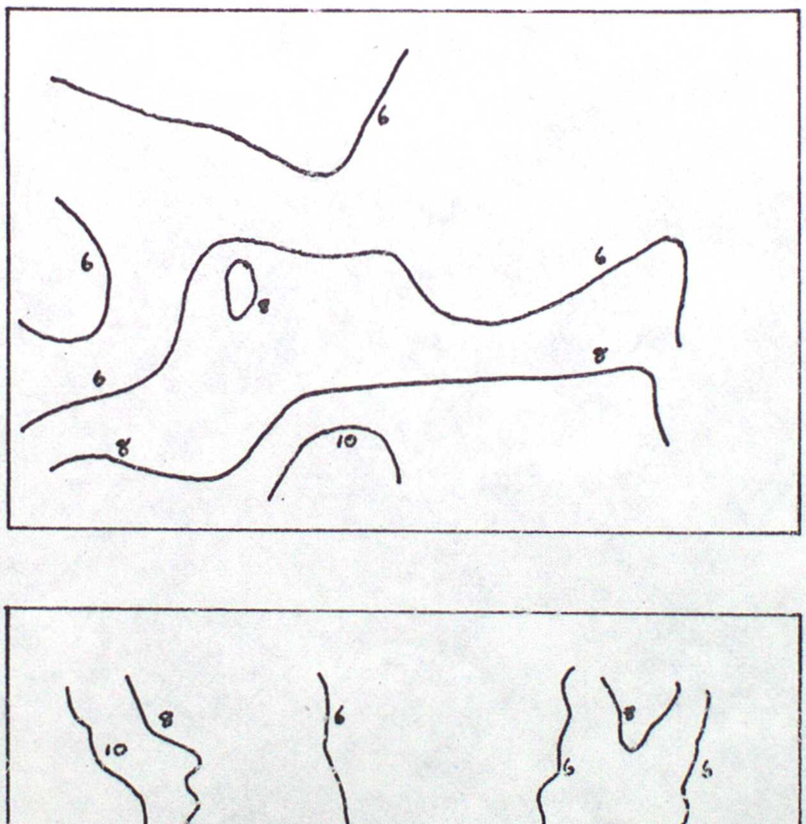


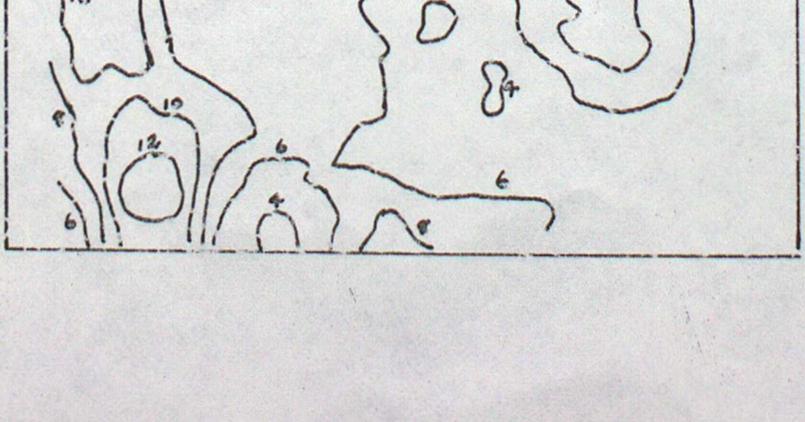
Fig 4



(a)



(b)



(c)

Fig 5

NO WASHOUT&TIME OR TEST FOR RH.GT.1
TOTAL RAIN IN 3.5HRS ISOHYET 2.0MM

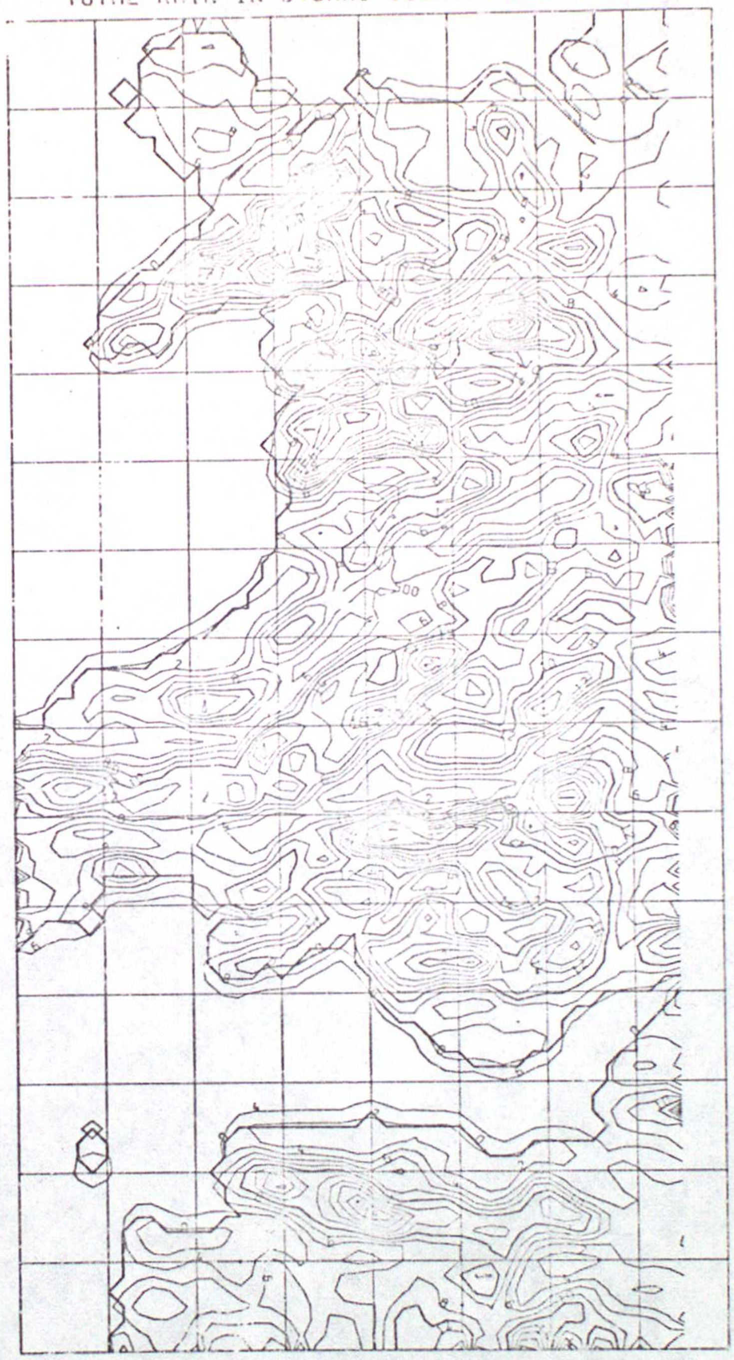


Fig 6

NO WASHOUT

TOTAL RAIN IN 3.5HRS ISOHYET 2.0MM

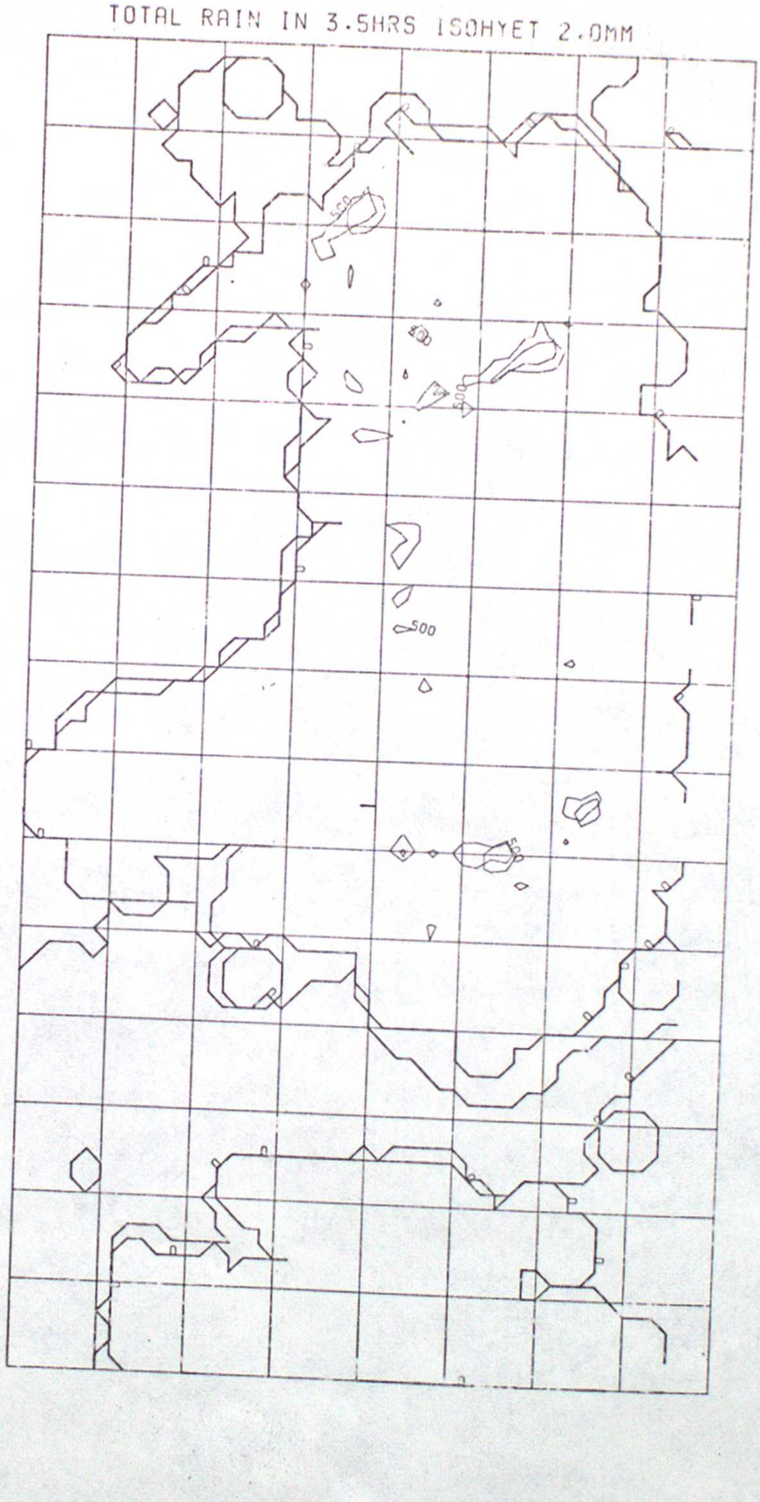


Fig 7

TOTAL RAIN IN 3.5HRS ISOHYET 2.0MM