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The forecasting of orographically enhanced rainfall accumulations using 10-level model data

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1. Introduction

A major problem in the provision of short period weather forecasts is the estimation of local rainfall which may be modified by local topographic effects. An important operational requirement is the forecasting of rainfall accumulations in hilly areas (Holgate 1973). Methods of forecasting based on the extrapolation of the movement of precipitation echoes derived from radar observations (Hill, Whyte and Browning, 1977) are only useful for forecasting up to 6 hr ahead and even these forecasts will require modification if the local orographic effects are large. For forecasts up to 36 hr ahead a numerical model is required. Unfortunately the resolution of present numerical models is one to two orders of magnitude greater than the scale of the orographic features, and the effect of these features is not incorporated into the resulting rainfall forecasts. One solution is to use the output from a numerical model as input to a separate, numerically simple model which has a grid size small enough to resolve the topography adequately and which contains the essential physics of the orographic rainfall process.

Collier (1975) has formulated a model of orographic rainfall which provides for an elaborate method of estimating vertical velocity in the vicinity of orographic features. This model has been used, with some success to estimate rainfall accumulations in North Wales using radiosonde observations as input data. As Collier (1977) indicates, the model results are only valid in special circumstances when there is efficient conversion of orographic cloud to liquid rain water. The cloud microphysical processes which are not parameterised in Collier's model play an important role in determining the extent of the orographic influence on the rainfall. The model about to be described is similar in concept to Collier's model but a simpler formulation for obtaining local vertical velocity is used together with a scheme for treating the cloud physics processes. It is based on the original scheme devised by Jonas (1976). The method described by Bader and Roach (1977) for calculating the washout of droplets in a low level orographically produced cloud by raindrops from a higher cloud has been adapted for use in this model.

2. Model Features

(a) General Description

The model has been designed to use the output from the fine mesh version of the Meteorological Office 10-level model (Benwell et al 1971) as input. Although the 10-level model does include topography, it is in a very smooth form. The local perturbations of the flow due to small sub-grid scale variations in topography are assumed not to affect the large scale flow, they merely introduce an additional component to the vertical velocity and modify the relative humidity at the fixed pressure levels.

Precipitation is formed at a given level by adiabatic ascent. It is assumed that the rate of rainfall so formed may be written as

$$P_{,} = -k_1 k_2 W \left(\frac{\partial r_s}{\partial P}\right)_{SALR} \rho \Delta z \qquad gm/sec.m^2 \qquad (1)$$

This is similar to the expression derived by Collier (1975) $W = {}^{DP}/Dt$, is the vertical velocity in P-coordinates, $\left(\frac{\partial r_s}{\partial P}\right)_{SALR}$ is the derivative of saturated humidity mixing ratio with respect to pressure for a saturated adiabatic ascent, ρ is the air density and Δz the thickness of the layer or, in the case of the lowest level, the height of the level above the surface. The parameters k_1 and k_2 depend on the way in which the relative humidity is modified by ascent, this will be described in section 2(c).

The precipitation so formed falls to the next layer and is further enhanced by the accretion of cloud water. This washout process increases the rainfall rate by P_2 to give a total rainfall rate $P=P_1+P_2$, we assume

$$P_2^{(k)} = P_2\,(P^{(k-1)}, Q^{(k)}) \qquad (2)$$

that is $P_2^{(k)}$ is a function of the precipitation rate $P^{(k-1)}$ from the layer (k-1) above, and of the cloud liquid water content $Q^{(k)}$ in the layer k. Details of this process will be considered later, see section 2(d).

The surface rainfall rate is calculated by summing the contributions of P_1 and P_2 from five 100 mb levels, from the top of the model at 500 mb to the surface.

The large scale velocity components (u,v,w), the layer thickness h' and the humidity mixing ratio r, available from the 10 level model are interpolated to the orographic model grid points. The resolution of the orographic model is $3\frac{1}{3}$ km and the topographic height is defined at each grid point.

(b) Formulae for W and $(\partial T_s/\partial p)$.

A very simple parameterisation of the effect of topography has been adopted. The vertical velocity W is given by

$$W = W_L + W_T \quad \text{mb/s} \qquad (3)$$

W_L is the large scale vertical velocity.

W_T is a topographically induced component which is assumed to be proportional to the scalar product of horizontal velocity $\underline{V}$ and local topographic gradient ∇H

$$W_T = -k_3\,(\underline{V}.\nabla H)\,\rho g \quad \text{mb/s} \qquad (4)$$

The effect of topography is assumed to reduce linearly to zero at 500mb. Accordingly the factor k_3 varies linearly with the height of the level above the surface, from zero at 500mb to one at the surface. An elaborate treatment of the dynamics of airflow over hills has not been attempted because the accuracy of the forecast input parameters is the dominating factor in determining the accuracy of the model results and the improvements gained by such sophistication are likely to be small in comparison. The effect of stability on the magnitude and vertical of the ascent is also neglected.

The rate of change of saturated mixing ratio with respect to pressure along a saturated adiabat may be readily derived from thermodynamic arguments. The energy conservation equation may be written approximately as

$$L\, dr_s = C_p dT + RT/p\, dp \qquad (5)$$

where L is the latent heat of vaporisation and C_p the specific heat at constant pressure.

In terms of the saturated vapour pressure e_s, the humidity mixing ratio is $r_s \approx \varepsilon e_s/p$, where ε=.622, in differential form this becomes

$$dr_s/r_s = \partial e_s/e_s - \partial p/p \qquad (6)$$

e_s may be eliminated from equation (6) using the Clausius-Clapeyron relation to give

$$\partial r_s/r_s = \varepsilon \frac{L\partial T}{RT^2} - \frac{\partial p}{p} \qquad (7)$$

manipulation of equations (5) and (7) gives

$$\frac{\partial r_s}{\partial p} = r_s \frac{RT}{p}(\varepsilon L - C_p T)/(L^2 \varepsilon r_s + RC_p T^2) \qquad (8)$$

(c) Modification of the relative humidity by forced adiabatic ascent

The topographically induced displacement is parameterised in the same way as the topographically induced component of vertical velocity. At a given level, the displacement is given by $H_T = k_3 H$ where k_3 is defined in (2b) and H is the topographic height.

We can use approximately linear relations for a dry adiabatic ascent relating the final relative humidity X_f to the initial relative humidity X_i and the topographic height.

$$X_f = X_i\,(1 + \alpha(P,T)H_T) \qquad (9)$$

α is a function of temperature and pressure only.

K_1 is assumed to depend on the vertical velocity and the relative humidity equations (3) and (9). Thus k_1=0 if $W \geqslant 0$, that is where there is local descent or if $X_f<1$ when the air is unsaturated even after ascent.

Otherwise k_1 is set equal to one and precipitation is allowed to form.

The parameter k_2 is introduced in an attempt to take into account the depth of the cloud. k_2 = 1 if the air is already saturated before topographic uplift. Otherwise k_2 is determined by the length of time the growing droplets spend in the orographic cloud. The vertical displacement H_s for which saturation occurs is given by substituting $X_f = 1$ in equation (9).

$$Hs = (1-X_i)/ \quad X_i \tag{10}$$

The droplets are assumed to follow the air motion with a vertical velocity $- W/_{\rho g}$ m/s for the remainder of the forced displacement $(H_T - H_s)$.

Hence the time taken for droplets to grow in the cloud is given by

$$t = -\rho g/_{\omega} \left(H_T - \frac{(1-X_i)}{\alpha X_i} \right) \tag{11}$$

A typical time scale for droplet growth is 20 minutes, thus we assume

$t > 1200$ secs,	$k_2 = 1$
$300 \leq t \leq 1200$ secs	$k_2 = (t-300)/900$
$t < 300$ secs	$k_2 = 0$

If the air becomes saturated during ascent the remainder of the ascent is taken along a saturated adiabat. The final value of the saturated mixing ratio r_s is found using equation (8). Using this value of r_s and the initial value of the mixing ratio r (considered as the sum of cloud rain and vapour mixing ratios) an apparent value of $X_f = \left(\frac{r}{r_s}\right)$ is found which is greater than unity. This is used to derive an estimate of the cloud liquid water content q, assuming 10% of the condensed water is retained in the cloud and allowing cloud to form at 90% relative humidity to allow for partial saturation of the grid point

$$q = 0.1\ (X_f - 0.9)\ r_s \times 10^{-3} \rho \quad kg/m^3 \tag{12}$$

(d) <u>Washout</u>

Now that an estimate of cloud liquid water has been made the rainfall P_2 due to accretion of cloud liquid water may be calculated. For simplicity a single cloud drop radius of 10μ is assumed. Now for a raindrop of radius a, with a fall speed V_a and a collection efficiency E_a, the rate of accretion W^1 of cloud liquid water is simply $W^1 = E_a V_a \pi a^2 q$ since $\pi a^2 V_a$ is the volume of air swept per second. If the raindrop distribution is $N_a \Delta_a$ then the total washout summed over all drops is given by

$$W = q \sum_a N_a E_a V_a \, \pi a^2 \, \Delta a \tag{13}$$

This expression is used by Bader and Roach in their model. $N_a \Delta_a$ is defined by the 'Best' dropsize distribution which is a function only of precipitation rate (see Mason 1971). The terminal velocities and efficiency factors are also taken from Mason. The summation in equation (13) is then only a function of precipitation rate. The increase in precipitation rate is $P_2 = W \Delta z$ where Δz is the layer thickness.

Thus the total precipitation derived within a layer is

$$P = P_1 + P_2 = -k_1 k_2 W \left(\frac{\partial T_s}{\partial p}\right) \rho \Delta z + W \Delta z \qquad gm/sec.\,m^2$$

This is subject to the constraint that the maximum allowable rainfall rate is $P = - W \left(\frac{\partial T_s}{\partial p}\right) \rho \, \Delta z$

(e) Other features in the model

Three other features included are evaporation, precipitation drift and spatial averaging.

If rain falls through a layer of unsaturated air, some or all of it will evaporate. The evaporation scheme uses empirically derived relationships due to Best (1952), a version of this scheme is also used in the 10-level model. A raindrop of radius a_1 reduces to a radius a_2 when falling from height Z_1 to height Z_2. The empirical relation between a_1 and a_2 is

$$a_1^2 - a_2^2 = E(z_1 z_2)(1 - X)^{1.13} \tag{14}$$

where X is the relative humidity and E is an empirical parameter derived by Best which depends on the height of the levels. The initial precipitation rate is resolved into a dropsize spectrum as in 2(d), a new dropsize spectrum can be computed using equation (13) and from this the final precipitation rate derived.

The horizontal drift of precipitation as it falls can have a considerable effect on the distribution of surface rainfall. Although the thermodynamics of the ice phase is not included in the model, the precipitation is considered as snow for the purposes of calculating drift if the temperature of the layer is less than 273K. For example, we readily calculate that the time taken for snow to fall from one level to the next is about 1000 secs assuming a fall speed 1m/s. In a strong wind of 25m/s the snow will have drifted 25kms (ie about 8 grid lengths). If the precipitation is in the form of rain with a fall speed of approximately 5 m/s, the drift is correspondingly reduced but is still appreciable.

Finally to avoid unrealistic gradients which might arise since the rainfall rate at each gridpoint is calculated independently of its neighbours, a 1-2-1 smoothing function is applied in both the x direction and the y direction at each point.

3. Results

The aim of this study has been to provide a potential forecasting model. To assess the performance of the model it was tested over an extended period using input data derived from a 10-level model rectangle forecast. Had the input data been perfectly accurate the results from the model would no doubt have been better, however an accurate analysis of the large, scale variables is difficult to obtain on an hourly basis. The method of assessment used here shows how the model would perform in an operational forecasting situation.

The model was used to provide forecasts of orographic rainfall over Wales and Central England. The area covered by the study is shown in Figure 1. The orientation of this area is the same as that of the fine mesh 10 level model whose grid points are marked by crosses. Figure 1 also shows the topography which was used in the calculations and the 23, 50 km X 50 km areas used for verification. The trial was for a period of fourteen consecutive days starting on the 3 October 1976.

Throughout the period a westerly or south westerly situation persisted with fronts and depressions crossing all areas, notably on the 14th when a deep depression reached S W England and moved slowly Northeast.

The input data for the orographic model was extracted hourly from the 10 level model fine mesh forecast which was based on a midnight analysis. The data from T+9 to T+32 hrs was used, thus the forecast period of the orographic model coincides with a rainfall day. The large scale parameters used as input to the model were assumed to be representative of conditions for a sixty minute period and were updated each hour. 24-hour accumulations were obtained by summing the hourly rainfall rates predicted by the model.

The only suitable observational data with which to compare the model results was the daily rainfall totals from the national raingauge network. Programs developed by the hydrometeorology section of the Meteorological Office produce objective estimates of the areal rainfall totals (Shearman 1975). These estimates may be in error over data sparse areas but in the absence of any other information they will be considered as 'truth'. A further problem with the verification is that the observed rainfall is the sum of convective type and dynamic type. The orographic model only caters for dynamically induced ascent (both large scale and topographically induced) therefore when comparing area totals a contribution to the total due to convective processes was included.

This convective contribution was derived from the 10-level model deep convection scheme (Hayes 1977) which added about 20% to the rainfall totals for low lying areas. This figure is in broad agreement with reports from synoptic stations although is probably an overestimate for inland stations.

Since the accuracy of the results from the orographic model depends directly on the quality of the forecast input data derived from the rectangle, verification of the results would be incomplete without a comparison with the results from the present operational scheme.

Figure 2 shows the difference between model and raingauge estimates of the 2 week totals for each of the 23, 2500 km^2 areas as a function of gauge estimate. Figure 2a refers to the 10 level model and Figure 2b refers to the orographic model. The areas are numbered as in Figure 1 for reference, low land areas are marked with a cross and areas where a significant percentage of the land is above 200 m are marked with a dot. The closer the points are to the x-axis the better the forecasts. Several conclusions can be drawn from Fig 2a. It is evident that although the 10 level model forecasts rainfall fairly accurately for most of the low lying areas, as the orographic influence becomes more pronounced it increasingly it underforecasts the rainfall. The improvement given by using the orographic model is greatest for the hilly areas. Considering the eight areas where the rainfall exceeded 100 mms in two weeks, the 10-level model underforecast the rain in these areas by between 45-60%. On the other hand for the orographic model (Fig 2b) the errors were less than 18% for all but one area and five areas had errors below 10%.

Figure 3 shows the daily rainfall for both models together with raingauge estimates for area 20 in South Wales. Figure 3a compares 10 level model results with raingauges and Figure 3b compares orographic model results with raingauges. Figure 4 shows a similar plot for area 5 in North Wales.

These results are typical of the results from the hilly areas. For both areas the orographic model produces a better forecast on almost everyday. The percentage of observed rainfall forecast by the 10-level model for area 5 was 51% for the two week period compared with 94% for the orographic model forecasts. For area 20 the same figures were 40% and 89%. For all 23 areas together the rectangle forecast 62% of the observed rainfall and the orographic model forecast 106%.

Figures 5 and 6 show the rainfall fields as depicted by the raingauge network and as produced by the orographic model respectively for the rainfall day commencing at 9Z on the 14 October 1976, the wettest day during the trial period. Bearing in mind that the basic input to the model was forecast data and that data was on a scale of 100 kms so that many of the mesoscale features producing intense precipitation (such as those discussed by Browning et al 1974) were not defined, the rainfall field forecast by the orographic model fits encouragingly well with that derived from raingauges. The distribution and intensity of rainfall in Snowdonia, Exmoor and Pembroke is particularly good. Note the drift of precipitation over the Conway Valley to the east of Snowdon and also the accurate forecast in Pembroke of an enhancement by a factor of three on fairly low hills indicating the importance of the gradient of the hills as well as their height. Although the rain shadow effect giving drier areas to the immediate lee of the hills is fairly realistic there is a general tendency to overforecast the rainfall in the area to the east of the Welsh Mountains and it seems that the removal of additional water by precipitation upwind may be a contributory factor. An improved scheme might be one which modifies the large scale humidity field in the 10 level model to take into account the local removal of water by the topographic model. This would lead to a feedback mechanism between the two models. On the, 14th October the rain area was forecast by the 10 level model to be rather too far eastward and this is reflected in the poor rainfall forecast for the west Midlands (20 mm forecast by both orographic model and 10 level model compared with an observed 5 mm). Since this area is low lying the orographic contribution is minimal but a scheme for removing water as just described might have produced better results.

4. Conclusion

It has been shown that the orographic influence on rainfall from large scale systems is well reproduced by this relatively simple model used in conjunction with the 36 hr forecast produced by the fine mesh version of the 10 level model. Several improvements might be envisaged in the future. The dynamics has been treated very simply and an improved scheme on the lines developed by Collier might be of use. Inclusion of a scheme to handle orographically triggered convection and improved modelling of the rain shadow effect would also be beneficial.

References

Bader, M.J. and Roach, W. T.	1977	Orographic rainfall in warm sectors of depressions. Quart. J. R.Met.Soc. 103 pp.269-280.
Benwell, C. R.R., Gadd, A.J., Keers, J. F., Timpson, M.S. and White R W.	1971	The Bushby-Timpson 10 level model on a fine mesh. Met Office Sci Paper No 32 HMSO, London
Best, A.C.	1952	The evaporation of raindrops. Quart.J.R. Met.Soc. 78 pp 200-225.
Browning, K.A., Hill, F. F. and Pardow C.W.	1974	The structure and mechanism of precipitation and the effect of orography in a wintertime warm sector. Ibid 100 pp. 309-330.
Collier, C.G.	1975	A representation of the effects of topography on surface rainfall within moving baroclinic disturbances. Ibid. 101 pp. 407-422.
	1977	The effect of model grid length and orographic rainfall 'efficiency' on computed surface rainfall. Ibid. 103 pp247-253.
Hayes, F. R.	1977	A new parameterisation of deep convection for use in the 10-level model Ibid. 103 pp 359-367.
Hill, F. F., Whyle, K.W. and Browning K.A.	1977	The contribution of a weather radar network to forecasting frontal precipitation - a case study. Met. Mag. 106 pp 69-89

Holgate, H.T.D.	1973	Rainfall forecasting for river authorities. Ibid. 102 pp 33-38
Jonas, P.J.	1976	A method of estimating local topograghically enhanced rainfall accumulations using 10 level model forecast data. Met O 11 Tech Note 66, Met Office Bracknell.
Mason, B.J.	1971	The Physics of clouds. Clarendon Press.
Shearman, R.J.	1975	Computer quality control of daily and monthly rainfall data. Met. Mag. 104 pp 102-108

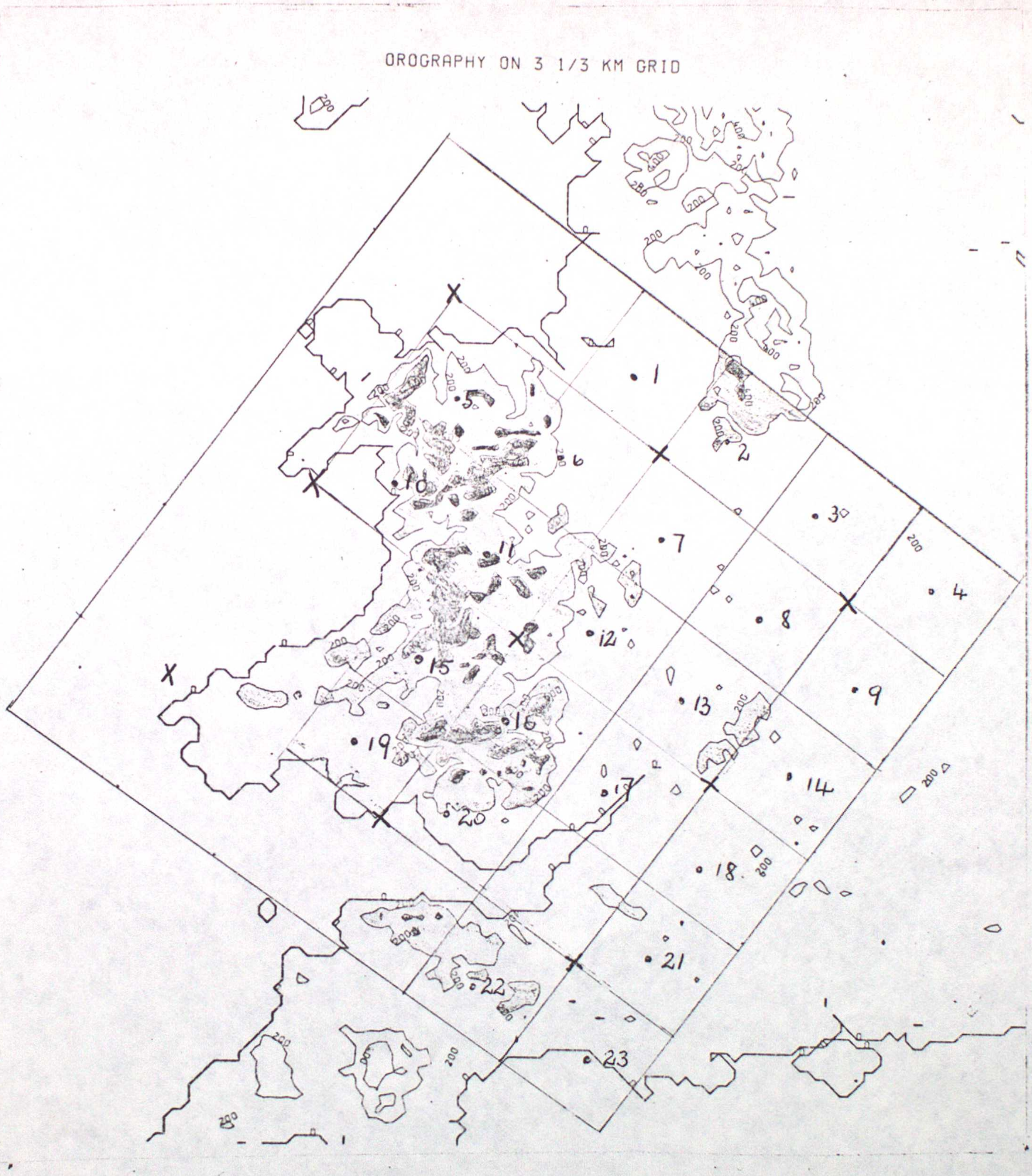


FIG. 1 — light shaded area - land between 200-400m above m.s.l.

dark shaded area - land above 400m above m.s.l.

X rectangle grid point

• 50 x 50km areas used for verification

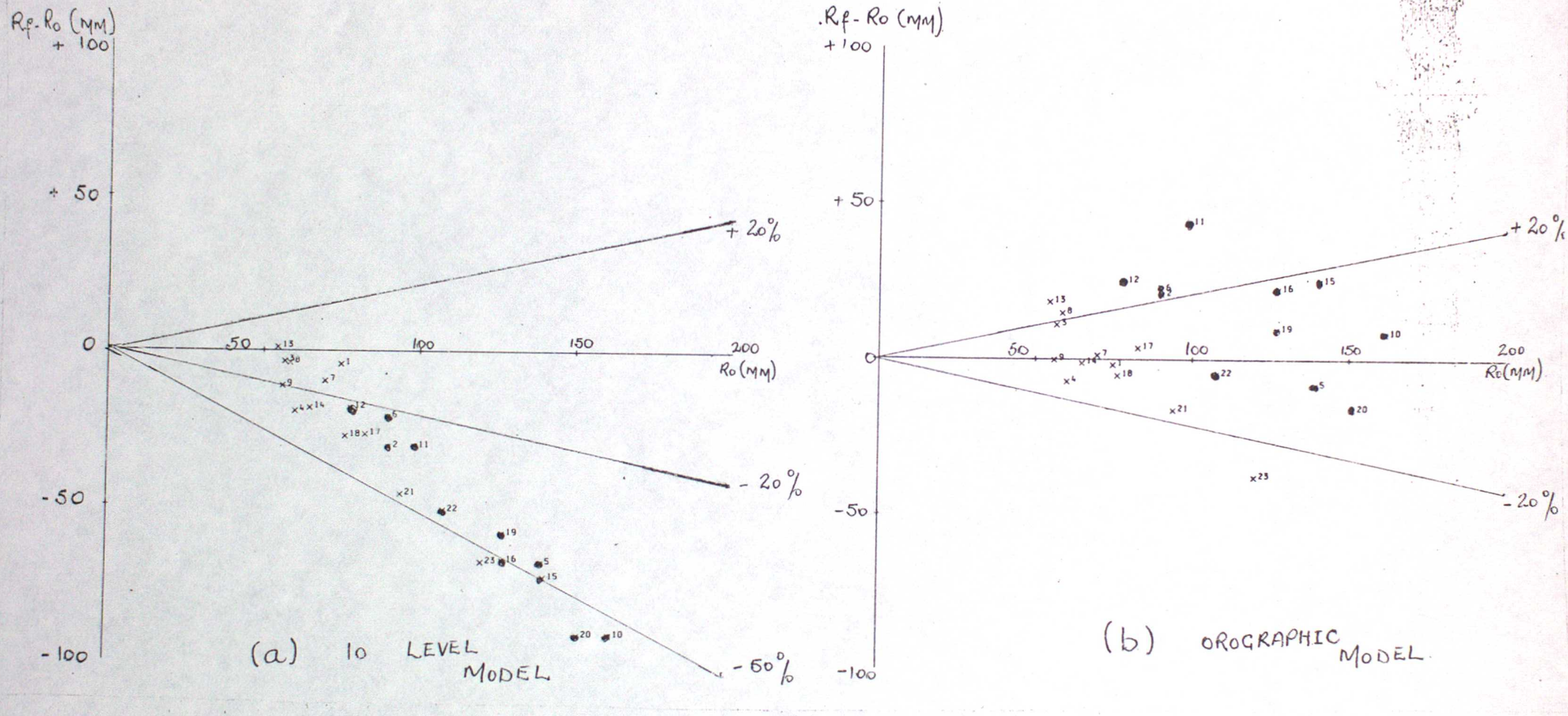
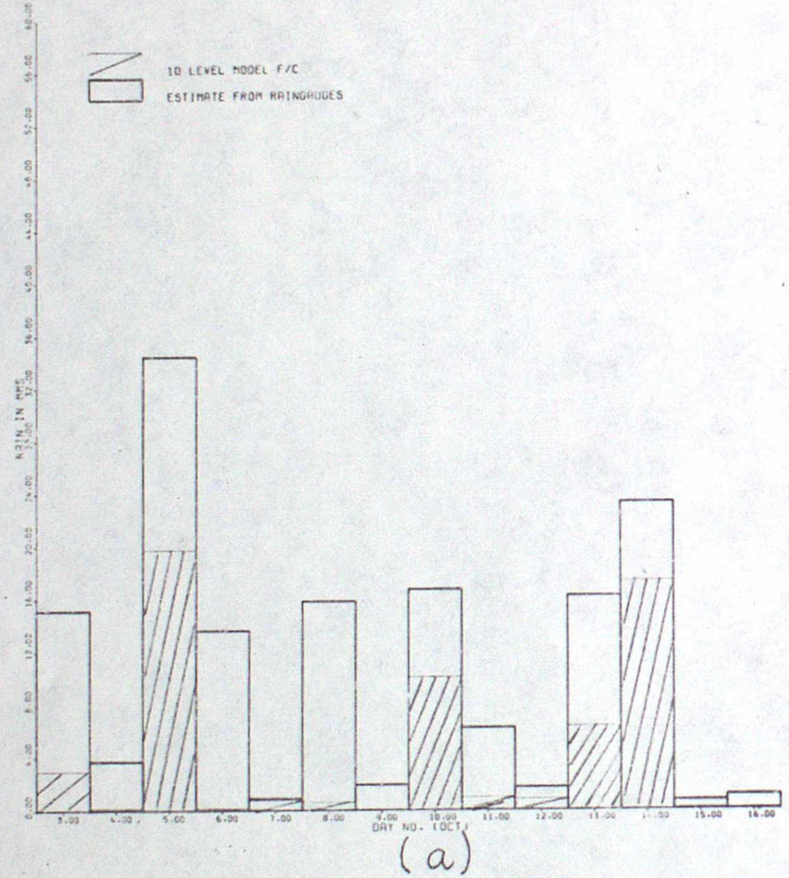


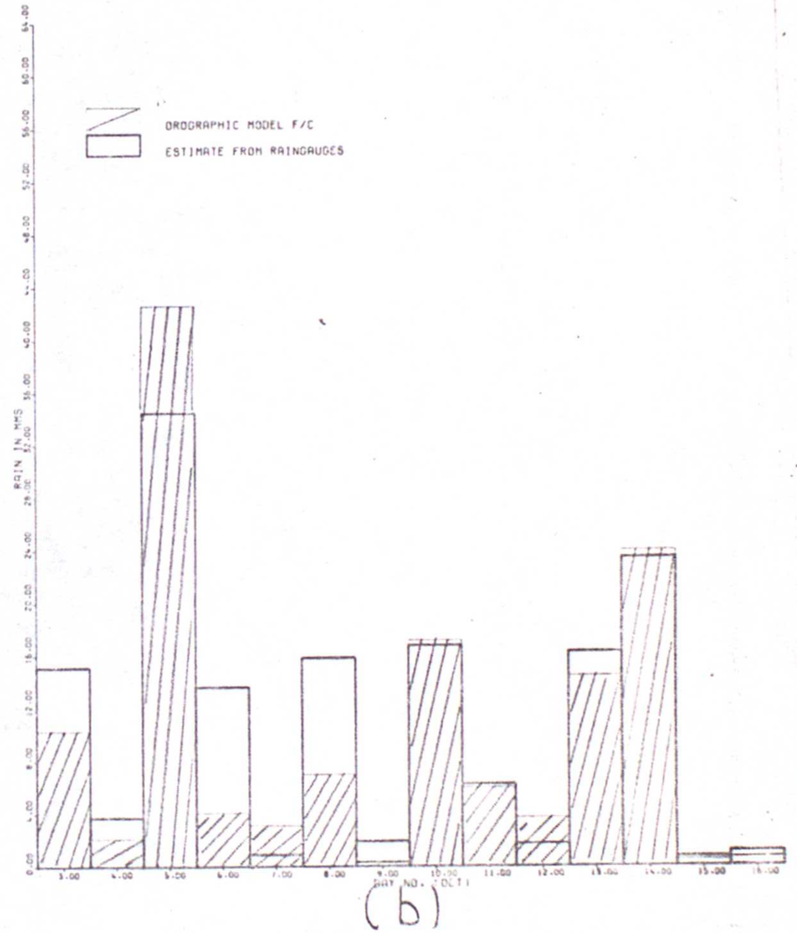
FIG.2 For two week period, plot of (forecast rainfall - observed rainfall) against (observed rainfall) for the 23 areas shown in Fig.1.

Fig.2a compares results from 10level model with observed rainfall. Fig (2b) compares results from orographic model with observed rainfall.

- • hilly areas
- x lowland areas.

The straight lines show the bounds for forecast errors of 20% and 50% respectively.





Area 20 (South Wales)

FIG 3a. daily rainfall f/c by 10 level model

FIG 3b. daily rainfall f/c by orographic model

(observed rainfall superimposed on both Figures)

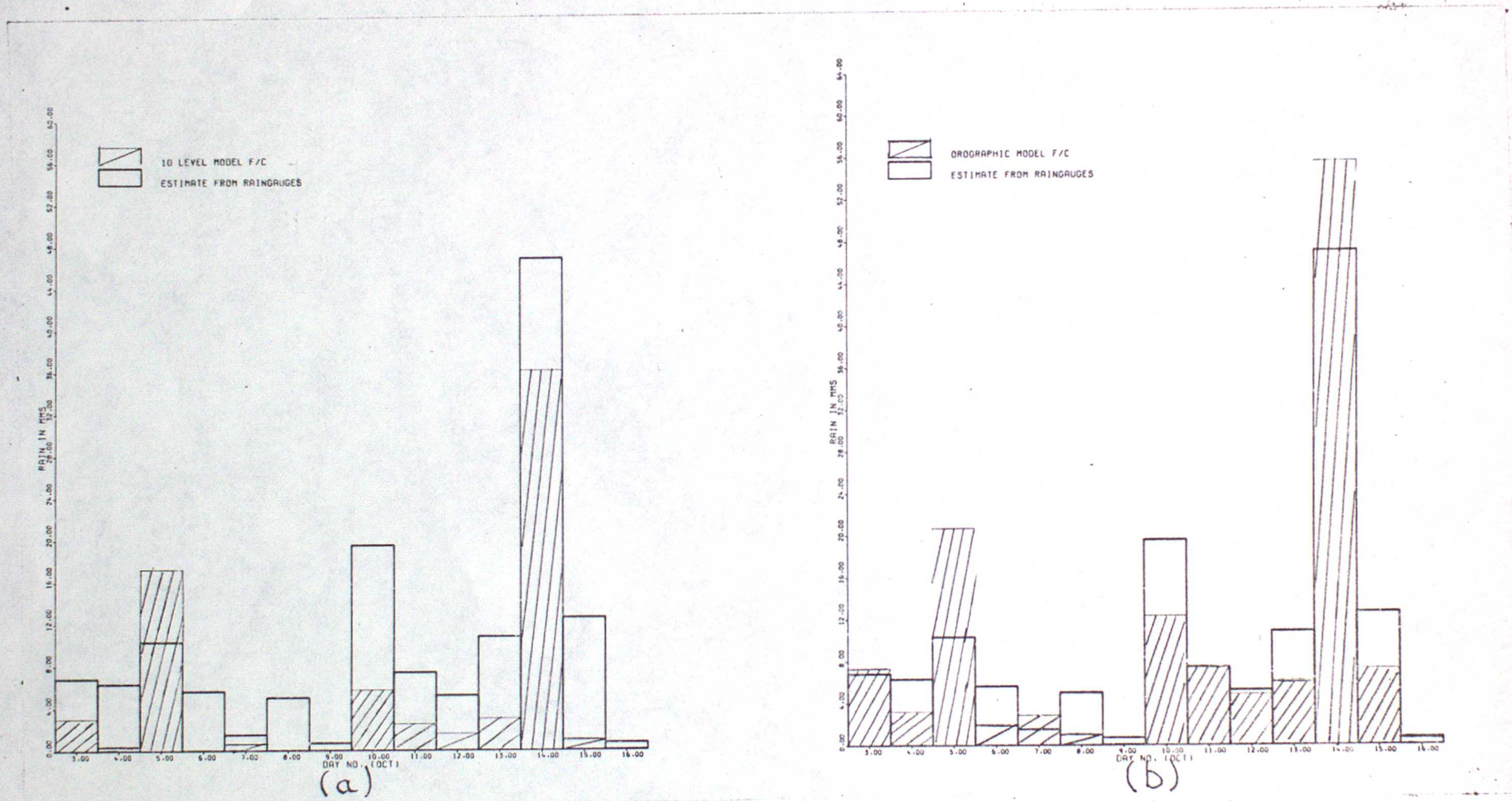


FIG. 4.   Area 5 (North Wales)

FIG 4a. daily rainfall f/c by 10 level model

FIG 4b. daily rainfall f/c by orographic model

(observed rainfall superimposed on both Figures)

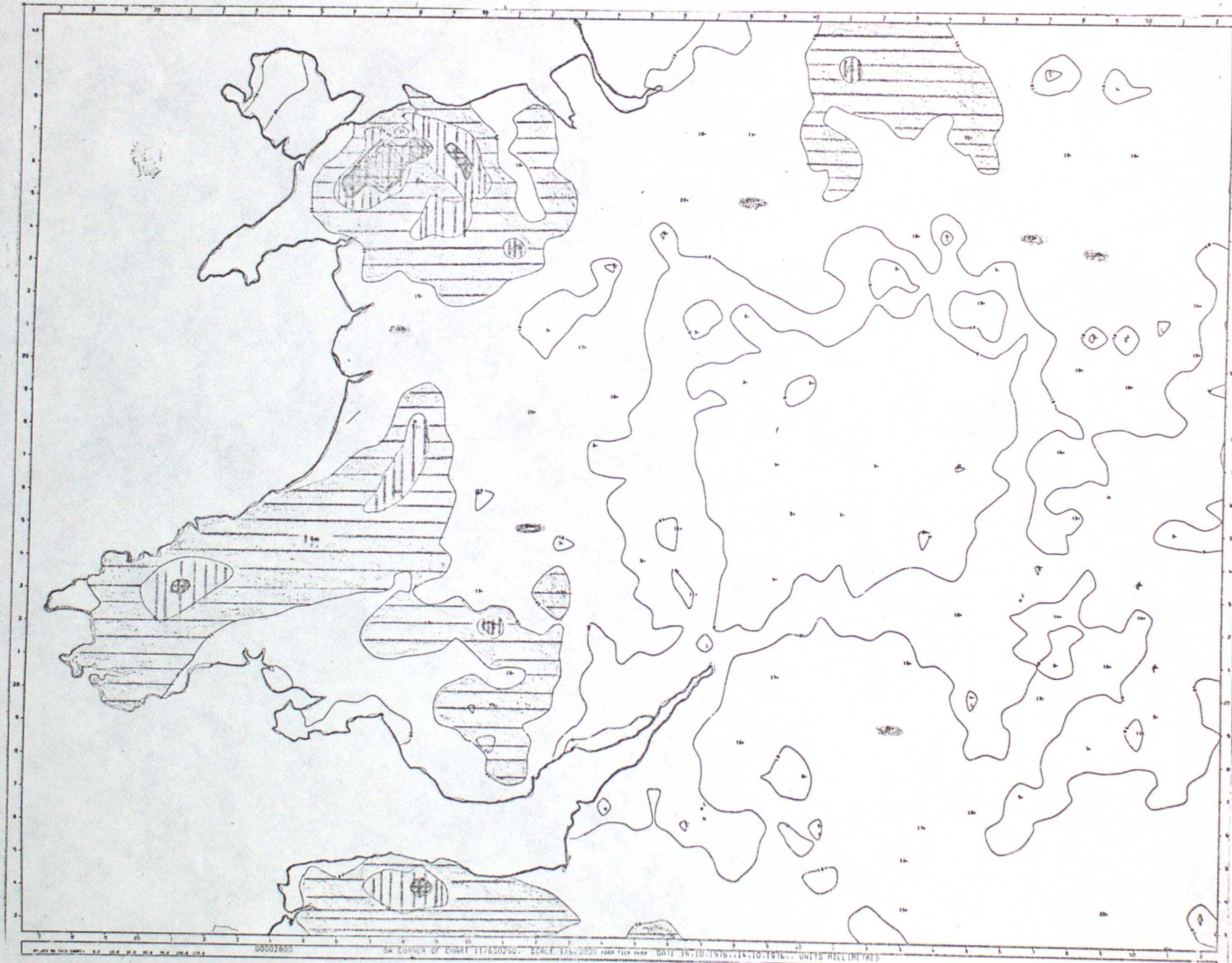


FIG.5

25-50mm

50-75mm

75mm <

Rainfall estimated from rainguage for 14/10/76

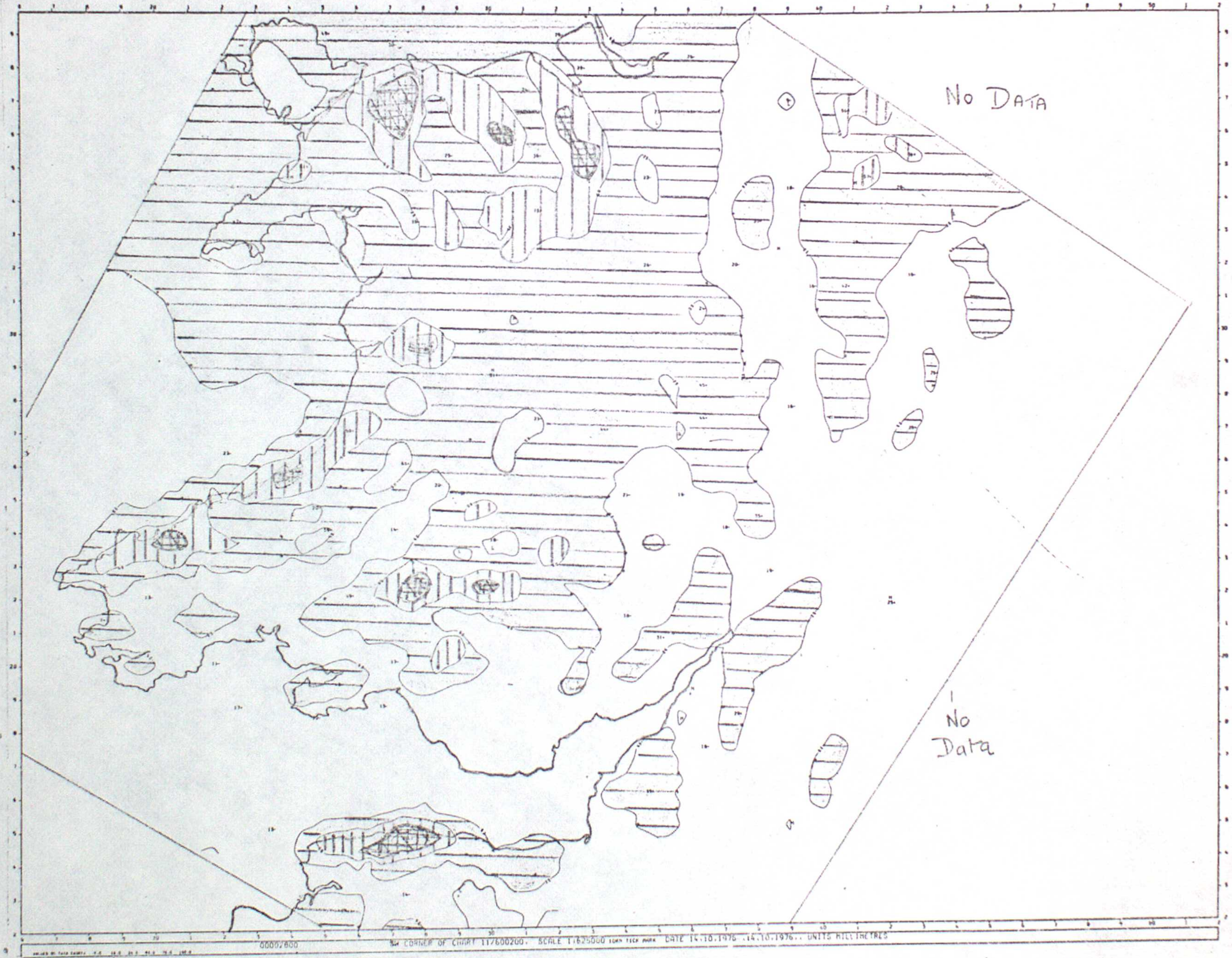


FIG. 6 25-50mm 75mm < 50-75mm

Rainfall forecast by orographic model on 14/10/76