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THE METEOROLOGICAL OFFICE MESOSCALE MODEL: ITS CURRENT STATUS

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Summary

A numerical forecast model with very fine resolution, is being developed as a short period forecast tool to give detailed guidance on local weather up to a day ahead. The processes represented in the model have been specially developed to take account of the scales represented. Surface synoptic reports are incorporated into the initial data to give mesoscale detail on boundary layer and cloud variables. A weekly trial of the complete system has recently started and is giving encouraging results.

1. Introduction

Numerical models in current operational use give valuable guidance to forecasters on the broad scale atmospheric structure. A gridlength of about 150 km is used for global predictions and half that for the regional model covering the North Atlantic and Europe. However, even this latter model cannot represent the topographic differences between parts of the United Kingdom which are important for short period forecasting. A mesoscale numerical forecast model with very fine resolution is being developed to tackle this problem with the aim of providing guidance to forecasters on the local variations of weather in the period up to a day ahead. This model will be closely tied to the regional model through its boundary conditions so it must be seen as a sophisticated tool for adding detail to the predictions of the coarser models. In particular it will not be able to correct timing errors in systems that are passed through the boundaries. On the other hand, in slow moving situations the topographically induced effects should be well forecast and should be of considerable help to the outstation forecaster. It is widely recognised that model predictions of mesoscale systems that are not forced by topography will be difficult. However the errors will often be in timing or location in the same way that regional scale models predict realistic development of secondary depressions but often at the wrong time or place. It may also be that much of the mesoscale variation in weather from larger scale systems is actually induced by topographic variations, perhaps through the surface temperature or moisture. In these cases the added detail will be of considerable value provided that the regional model has correctly predicted the large scale evolution. In these situations an important task to be performed after the forecast will be to apply gross timing or development corrections which have become apparent through consideration of other observations and forecasts. This will involve the sort of techniques to be discussed in the companion paper by Browning and Golding. In the present paper, the remaining sections will describe the model formulation, the methods currently used for preparing the initial data, and some recent results.

2. The forecast model.

The model is planned to cover the British Isles with a gridlength of 10 km but currently uses a 15 km gridlength (See Fig 1). With this resolution, a reasonably faithful representation of the orography can be given, and the coastline, indicated by the zero contour in Fig 1, has a realistic shape. The mountain ranges are still somewhat lower than reality, eg the Cairngorms reach 750 m rather than the observed 1200 m. Also the valleys which dissect them are not represented and so their local effects on the weather of cities like Sheffield, for instance, cannot be accounted for. A gridlength of under 5 km would be needed to represent such features and is not feasible on a National basis with current computers. Their effects will therefore have to be added to the model guidance by the forecaster.

The basic dynamical equations used by the model have been described in Tapp and White (1976) and Carpenter (1979). In most respects they are the same as those used in the lower resolution operational models. Important differences are that hydrostatic balance is not imposed and that the vertical coordinate is height above land surface rather than a pressure based coordinate. Non-hydrostatic effects are important for small scale thermally driven circulations while the height coordinate is advantageous for prediction of near surface effects. The vertical structure of the model is shown in Fig 2 for the current version with 16 levels. The lowest level is at 10 m and the spacing increases linearly from 100 m to 1500 m at the top. The highest level at 12010 m is in the stratosphere. This arrangement gives 5 levels in the lowest kilometre, and when expressed in terms of the standard atmosphere, an almost constant spacing of 60 mb from there up to the tropopause.

In large scale models, many of the weather producing processes occur at scales much smaller than the model's resolution. They are parametrized in terms of scales that are resolved by assuming that they can be represented by the effects of a statistically homogeneous and stationary ensemble covering a grid square. These models ignore the presence of processes at intermediate scales. It is these intermediate scales that are explicitly forecast by the mesoscale model. Smaller scale processes must still be parametrized and in many cases the same techniques can be applied as in larger scale models. However deep convection occurs on scales close to the model resolution so the statistical assumptions are not tenable in this case. In the following sections, descriptions of these parametrizations are given under the headings of boundary layer, layer cloud, and convective cloud processes.

a) Boundary Layer Processes

The processes involved are illustrated schematically in Fig 3. They may be divided into three groups: radiation, turbulent diffusion in the atmosphere and conduction in the ground. All three are controlled by the characteristics of the ground, eg its wetness, reflectivity, conductivity and porosity, and the vegetation present. At present two characteristics, the albedo and soil conductivity are specified as fixed over all land areas. However two others, the roughness length (z_0) and the surface resistance to evaporation, can be varied. Over the sea, the latter is zero and roughness is related to wind speed through Charnock's formula (Charnock 1955).

$$z_0 = \frac{k u_*^2}{g}$$

with k = 0.035

and u_* calculated from the 10 metre wind, using the previous timestep's drag coefficient. Over land, roughness length is presently fixed at 0.1 m but variations for urban and forest areas will be included soon. The resistance to evaporation is allowed to vary with time over land. The value at night is much greater than in the day to model the effects of darkness on the transpiration of plants, and it is zero when rain is falling or dew is forming. Clearly a desirable improvement will be for this parameter to remain zero after rain has fallen until it has evaporated, percolated into the ground, or run off into rivers.

Most of the heat gain at the surface comes from solar radiation. This is strongly affected by the presence of clouds in the atmosphere and is modelled by applying a transmission function (T) which depends on the integrated density of forecast cloud through a column of the atmosphere. The function has been fitted to data obtained from the radiation scheme of Slingo and Schrecker (1982) and has the form

$$T = \exp\left\{-7.9\, W^{0.5}/(1.84 + \cos^2\gamma)\right\}$$

where W is the total liquid water path in kg m^{-2} and γ is the solar zenith angle. The variation of T with W, for $\cos\gamma = 0.4$, is shown in Fig 4. Clouds also emit long wave radiation and it is the balance between this and the radiation emitted by the ground which determines the surface temperature in overcast conditions. The cloud emission (L) is again dependent on the total liquid water path W and is based on a scheme of Lind and Katsaros (1982) giving

$$L = \sigma\left(1 - \exp(-70W)\right) T_c^4$$

where σ is the Stefan-Boltzmann constant and T_c is the cloud base temperature.

Heat conduction in the ground is crudely modelled by predicting the temperature of a single level in the ground. This varies slowly depending on its difference from the surface temperature.

The final component of heat balance at the surface is the turbulent diffusion through the lowest layers of the atmosphere. In the model, transport between the surface and first level at 10 m is modelled using Monin-Obukhov similarity theory to calculate the mixing coefficient. A full description of the formulation is given in Carpenter (1979). The surface resistance to evaporation, defined above, is important here in determining the relative transports of sensible heat and of moisture. Above the 10 m level, the mixing coefficients are determined from a forecast parameter, the turbulent kinetic energy (TKE), and a diagnosed one, the mixing length. The latter increases above the ground until it reaches an empirically defined fraction of the boundary layer depth. The TKE is generated by shear and buoyancy and can also be transported. In particular, it can be diffused upwards from where it is generated near the ground to the boundary layer top, where the resultant entrainment of air from above is an important factor in the boundary layer evolution. The present formulation of these processes does not account for the reduced stability when saturation occurs. However, a revised formulation including its effects is under test and in particular should improve the prediction of stratus and stratocumulus cloud.

b) Layer cloud processes.

The processes involved in the layer cloud parametrization are depicted in Fig 5 for a region of orographically induced cloud. When moist air is cooled to saturation point in the model, condensate is not immediately rained out as in most large scale models, but is stored as cloud water. When sufficient has accumulated, it will precipitate. Meanwhile, it is advected by the wind and, if warmed, it may re-evaporate. The precipitation process itself is based on a simplified version of a scheme by Sundqvist (1978). It has two components, a local production term and an accretion term.

$$\frac{dP}{dz} = (C_L + C_A P(z))\left(1 - \exp\{-(m/C_M)^2\}\right) m$$

where m is the cloud water mixing ratio, $P(z)$ is the precipitation rate at height z and C_L, C_A and C_M are empirical constants. The exponential term merely ensures that for very low cloud water densities, no rainfall is produced. Above a threshold determined by C_m, the local production depends linearly on m and the accretion term depends on the product of m and P, the precipitation rate from higher cloud. The effect of the accretion term in enhancing the precipitation from "seeder" clouds can be seen in Fig 5. The combined effect of the two terms for clouds of increasing thickness but fixed cloud water mixing ratio is shown in Fig 6. Below cloud base, precipitation is evaporated as it falls to the ground. No specific allowance is made for the physics of solid precipitation in the model. However, the cooling effect due to melting snow is included because of its importance in modifying the low level temperature structure when surface temperatures are near freezing.

c) Convective cloud processes.

In large scale models, cumulonimbus clouds are modelled by parametrizing the mean effect of a large number scattered throughout a general area of instability. This approach is inappropriate for a model with a grid length of the same order as the largest clouds and much smaller than a typical spacing between clouds in an area of instability. It is therefore necessary to model the processes in an individual cloud rather more carefully. The scheme used in the model attempts to do this but is still capable of considerable improvement. It is based on that described by Fritsch and Chappell (1980). Figure 7 shows a schematic of the "typical" cumulonimbus cloud used in the parametrization. An important departure from schemes used in large scale models is that the cloud has a specified lifetime, much larger than the model timestep. Indeed a version currently being tested allows the cloud to move during its life. The details of the cloud's life cycle are not however modelled. Its growth, maturity and dissipation are all averaged out over its lifetime. A major problem for all cumulonimbus parametrizations is to determine the amount of cloud, or more specifically, the mass flux of air through the cloud(s). In the present case this is determined by the maximum deviation of the pseudoadiabat of a parcel lifted from cloud base from the environment temperature sounding. For a given depth of cloud, a standard mass flux is defined taking account of the observation that the aspect ratio of depth to area is of limited variability. If the temperature criterion would give a very tall, thin cloud, the aspect ratio criterion overrides this. Another difficulty in formulating a parametrization is to determine under what conditions a cloud will form. This is sensitive to the formulation of the boundary layer scheme and in the present model is determined by testing the stability to lifting of layers that have already been saturated, normally by upward turbulent transport of moisture.

Other details of the scheme are illustrated in Fig 7. The updraught is modelled as an entraining plume with inflow below cloud base and outflow where the upward momentum created by buoyancy is reduced to zero. The downdraught is forced by precipitation drag and cools by evaporation below cloud base before spreading out in the lowest three layers ie 460 m, of the model. The net mass flux from the updraught and downdraught is balanced by subsidence in the environment. Finally, air from the updraught and downdraught is mixed into the environment to simulate the dissipation process. Rainfall is determined as a proportion of the total moisture condensed in the updraught, the proportion having an empirical dependence on mean shear and humidity. The remaining condensate is mixed into the environment with 60% from the "anvil" and 40% from the lower layers of the cloud. An empirical formula is also used to relate the rain area to the mass flux and mean shear of the cloud so that local rainfall intensity can be diagnosed. Despite this sophistication, the scheme inherits many of the limitations of those in larger scale models. Most important is the assumption that there is no net vertical mass flux in a grid column. This is reasonable for gridlengths of several hundred kilometres but incorrect for 10 km gridlengths. The scheme also lack parametrizations of momentum transport and ice phase effects at present.

3. Initialisation

The representation of the initial state of the atmosphere is of critical importance to the quality of forecast that can be expected from the model. As with large scale models, the constraints of near-geostrophy must be satisfied if a stable forecast evolution is to be obtained. However, a short range forecast model must also be correctly initialised with cloud if the temperature and precipitation are to be realistically forecast. Indeed, the atmosphere "remembers" much of its initial state over a 12 hour period on many occasions and this contributes to the accuracy of subjective forecasts based on modified extrapolation procedures.

In the mesoscale model, the basic specification of initial conditions is obtained by interpolation of a short forecast (usually 6 hours) from the operational regional model. The regional model analysis is not used since that is at present an interpolatio from the global model analysis with a gridlength of 150 km and having a very crude topography specification. The interpolation to the mesoscale model grid is a complex process since the models are based on different map projections, have a different vertical coordinate and different orography as well as the mesoscale model having finer resolution. These initial conditions are then enhanced by the use of surface synoptic observations. At present the techniques used are purely objective but interactive facilities are being developed and it is intended that the human analyst will be able to influence the process at all stages (Browning and Golding 1984). The modifications are made in two stages. First, surface variables and then cloud variables are analysed and incorporated.

The use of surface variables is illustrated in Fig 8. Temperature, humidity and wind observations are first used to correct the interpolated 10 m values of these variables. When a well mixed boundary layer is present in the atmosphere, it can be assumed that information about the surface quickly reaches the boundary layer top. The corrections at 10 m are therefore applied with decreasing weight at higher levels up to a diagnosed boundary layer top. This is defined as the level at which parcels from 10 m, rising with a slightly positive lapse of potential temperature, will cease to be buoyant,provided it is at or above the third model level.

The use of cloud observations is illustrated in Fig 9 and has been described in Higgins and Wardle (1983). Surface observations are used to correct the cloud base, cloud top and precipitation rate, diagnosed from the regional model. The model's precipitation scheme is then used to define the cloud water mixing ratio which, with the analysed cloud depth, will give the analysed rainfall rate. At the 10 m level, fog observations are also used to correct the cloud water values.

Some comparison runs have indicated that the forecast is quite sensitive to the enhancement of initial conditions described above and, in particular, to the cloud data.

4. Examples.

A version of the model containing all of the processes described here was first produced in October 1983. During late 1983 a number of case studies was run which indicated where further work was needed but also showed sufficient skill to justify starting a weekly trial of the model from the start of 1984. Since then, several good forecasts have been obtained although weaknesses remain at present.

An example of a forecast from the model is shown in Fig 10 with a verifying chart in Fig 11. The information displayed there was extracted from charts of several model variables but could, in principle, be obtained automatically. It shows a 6 hour forecast from a data time of 0600 GMT 12/12/1983. An occlusion was moving slowly eastwards with a belt of associated rain and snow and was followed by a cold airstream with showers, especially on coasts. In the observations, Fig 11, the 1° surface isotherm was a good indicator of the boundary between rain and snow. Although somewhat larger in area in Fig 10, the prediction of snow using this indicator would have given excellent guidance to a forecaster. The timing is not quite correct with the frontal rain belt a little slow and too small a gap behind it before the showers start. However, it is encouraging to see the model prediction of a cluster of showers in the Midlands, close to the reported snow. It should be noted here that the model will naturally appear to have a greater density of showers than observed because its resolution is finer than the reporting network. Nevertheless, the predicted showers on the north east coast are clearly erroneous. In the north west, the model has predicted a lot of convective cloud but mainly light rain from stratiform cloud rather than showers. This is because the convection scheme does not account for the effect of the freezing level on shower precipitation and finds insufficient water in the clouds to produce rain. The result is a thick deck of stratocumulus cloud giving drizzle. A simple change to the convection scheme is being tested to correct this behaviour.

During February and March 1984, an extended period of anticyclonic weather affected England with spells of cold northeasterly winds and overcast skies. A number of forecasts were run in this period and they demonstrate the skill of the model in forecasting surface temperature when air mass changes are not occurring. Fig 12 shows the surface temperature curve for Heathrow from the model compared with that observed on 27th February 1984. A thick layer of low stratus persisted throughout the day and although the model cloud was not quite thick enough, the temperatures show very good agreement. In contrast to this case, Fig 13 shows a comparison for the same location on the following Saturday when the cloud was well broken. The agreement is again quite good, the main error being the delay of an hour in the start of the temperature rise. These cases show that the model can correctly represent the effects of the presence or absence of low cloud on the surface temperature.

During the tests, some problems with the model have been identified. These are mainly associated with the boundaries which are of particular importance because of their proximity to the forecast area. Work is in hand to correct these problems. A more difficult challenge is posed by the sensitivity of the model to its initial conditions, especially of cloud. A great deal of work remains to be done to incorporate all of the available information from radar, satellite pictures and radiosonde ascents as well as from the surface reports. As the use of these data improves, the model forecasts can be expected to improve further.

5. Conclusions.

A short range, fine scale forecast model has been developed for forecasting for the British Isles. Many of the physical parametrizations have been specially written to take account of the scales represented by the model. A sophisticated scheme for analysis of surface synoptic reports has been developed for preparing fine scale initial data of the boundary layer and cloud fields. The complete system has been under regular test since the beginning of 1984 and has produced some encouraging results. However, further development and testing are required before it can be used for operational guidance. In particular the format in which the output will be presented to forecasters must be determined. This is a much more complicated task for a model which predicts variables such as cloud, rain and visibility than for one whose main prediction is a pressure pattern. In addition, facilities must be developed for checking the forecast and making any necessary modifications. On the broad scale this may be done centrally but detailed processing for specific requirements will have to be done at the outstation where the guidance is used. The techniques which might be used in these processes form the subject of the companion paper by Browning and Golding.

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Figure Legends

Fig 1. Model domain and orography. The grid points have a 15 Km spacing and the contour interval is 50 m. The bold contour is at zero metres and indicates the model coastline.

Fig 2. Vertical structure of the model. The vertical coordinate is height above ground (η) and there are 16 levels from 10 m to 12010 m. Wind, pressure, temperature, humidity and cloud are carried at the main levels indicated by solid lines. Vertical velocity and turbulent kinetic energy are carried at intermediate levels.

Fig 3. Schematic diagram of processes involved in the surface heat balance of the model.

Fig 4. Transmission of solar radiation by cloud as a function of total liquid water path in kg m^{-2}.

Fig 5. Schematic diagram of processes involved in the layer cloud parametrization. The wind is assumed to be blowing from left to right at all levels.

Fig 6. Rainfall rate in mm hr^{-1} as a function of total cloud thickness when the cloud water mixing ratio is 0.6 g kg^{-1}.

Fig 7. Schematic diagram of the cloud model used in the convection parametrization.

Fig 8. Schematic diagram of the method of incorporating surface observations into the model initialisation.

Fig 9. Schematic diagram of the method of incorporating cloud observations into the model initialisation.

Fig 10. 6 hour model prediction of the weather at 1200 GMT 12/12/1983. Triangles indicate showers and dots are very light layer cloud precipitation. The shaded area in the southeast is the main layer cloud precipitation belt and the dashed lines enclose areas below 1°C where snow is predicted.

Fig 11. Verifying observations at 1200 GMT 12/12/1983. Cloud and weather symbols have their usual meanings and indicate the extent of showery activity. The shaded area in the south east is the main frontal precipitation belt and the dashed lines enclose areas below 1°C where snow was reported.

Fig 12 Verification of 12 hour temperature prediction for Heathrow starting at 0600 GMT 27/2/1984. The dotted line joins predicted values and the full line joins observations.

Fig 13. Verification of 12 hour temperature prediction for Heathrow starting at 0600 GMT 3/3/1984. The dotted line joins predicted values and the full line joins observations.

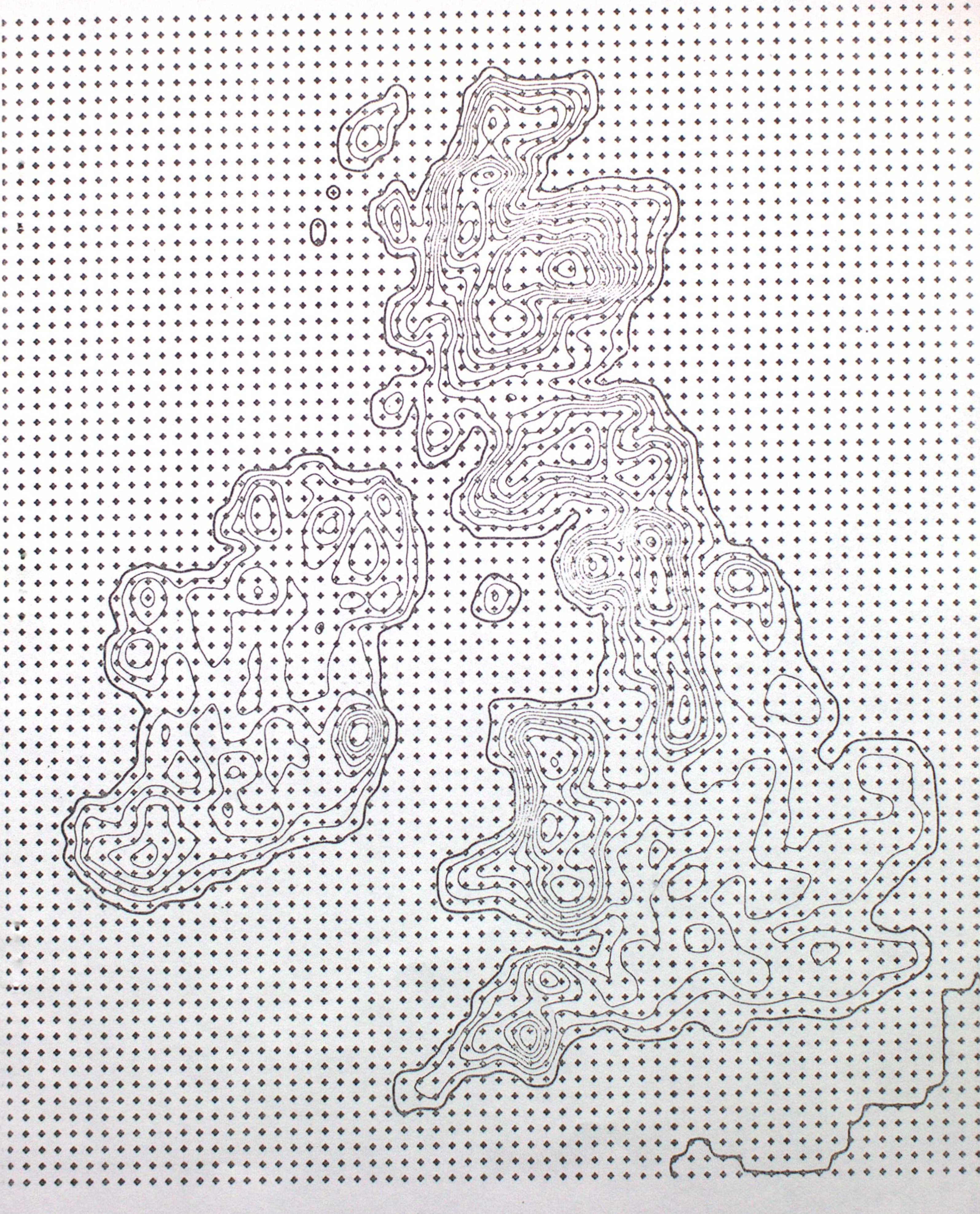


Fig. 1 .

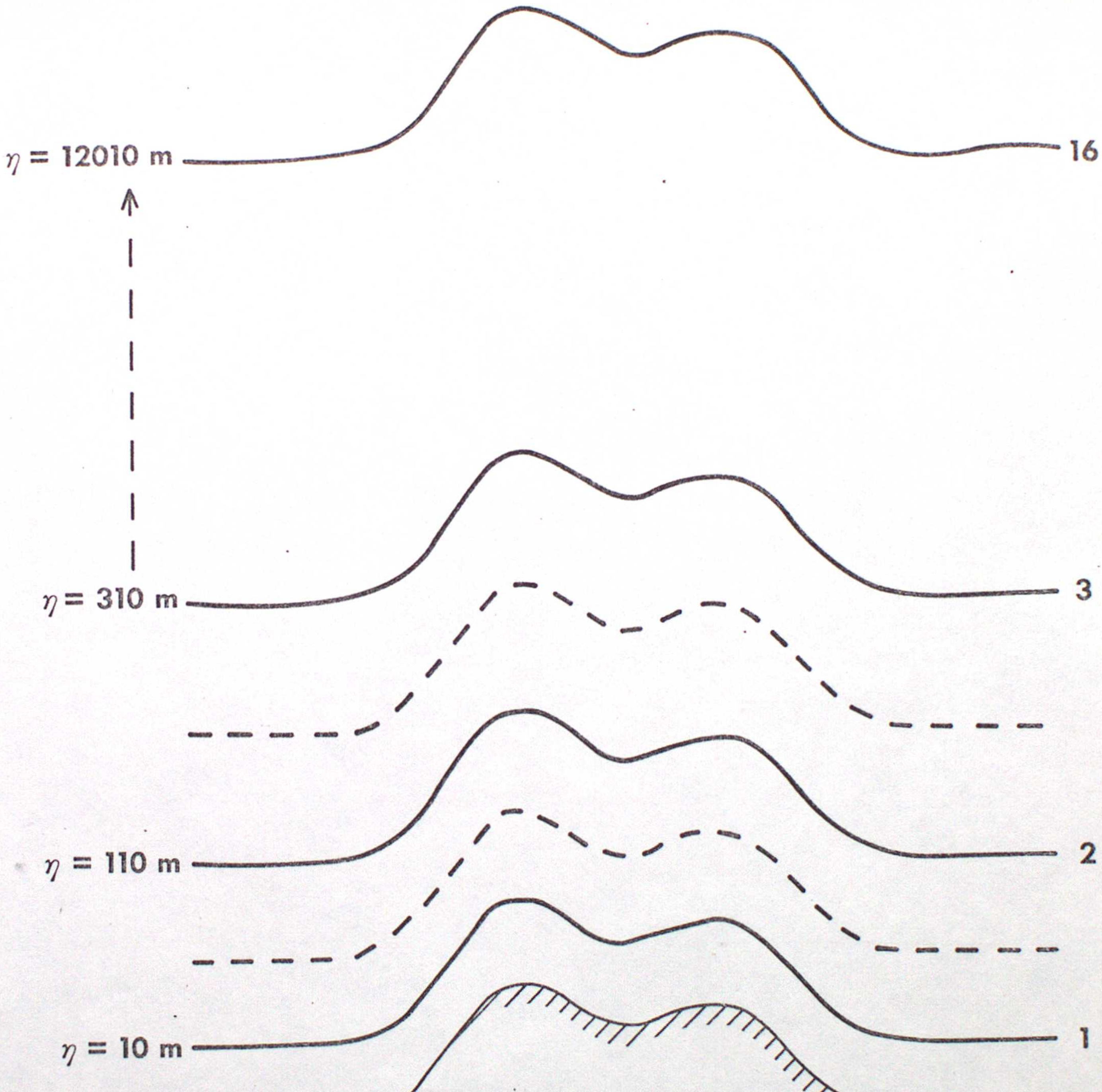


Fig. 2

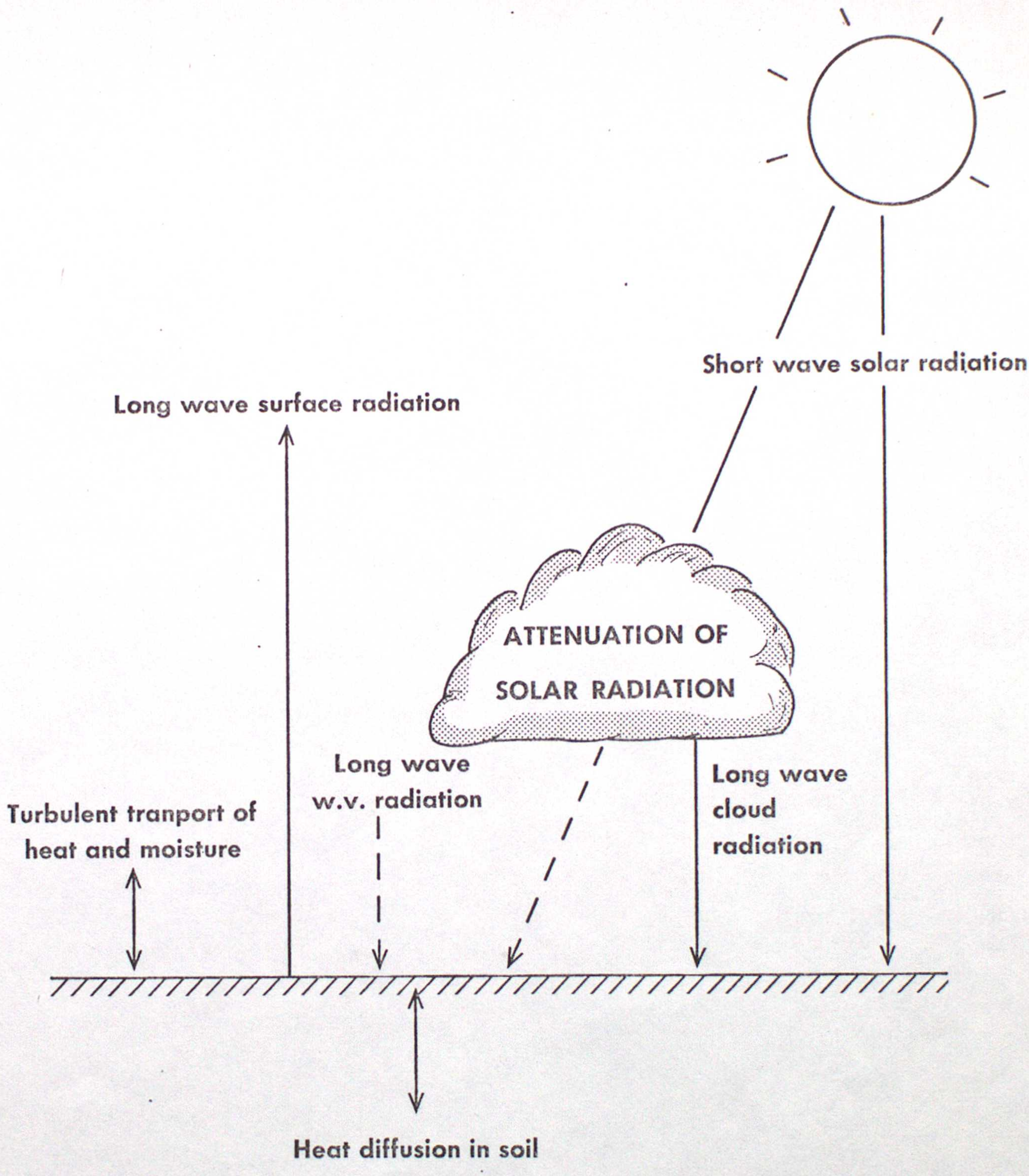

Short wave solar radiation
Long wave surface radiation
ATTENUATION OF
SOLAR RADIATION
Long wave
w.v. radiation
Long wave
cloud
radiation
Turbulent tranport of
heat and moisture
Heat diffusion in soil

Fig. 3

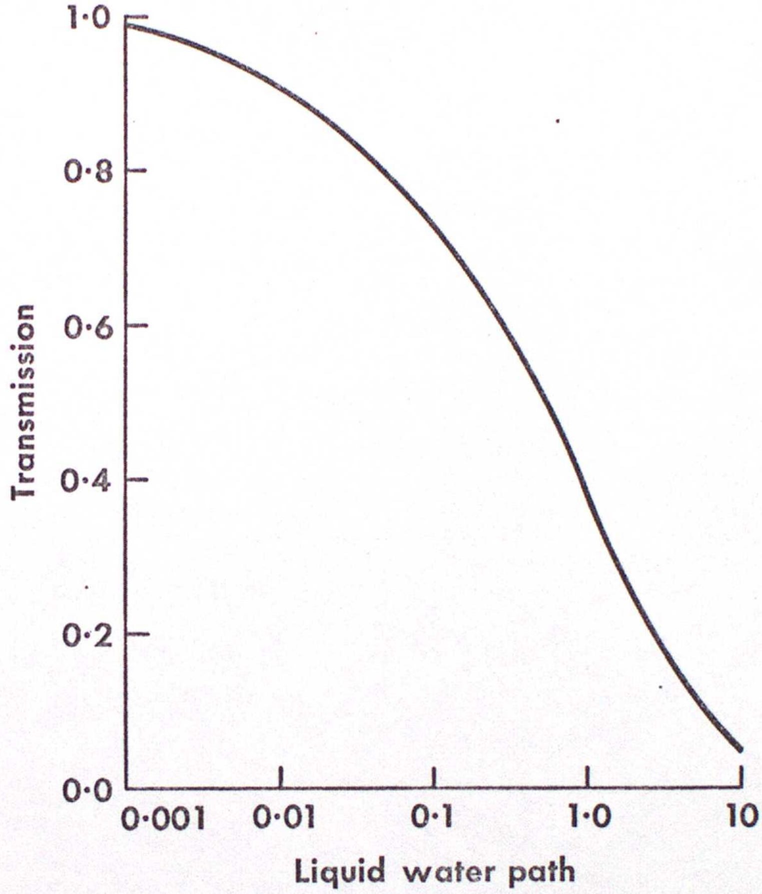
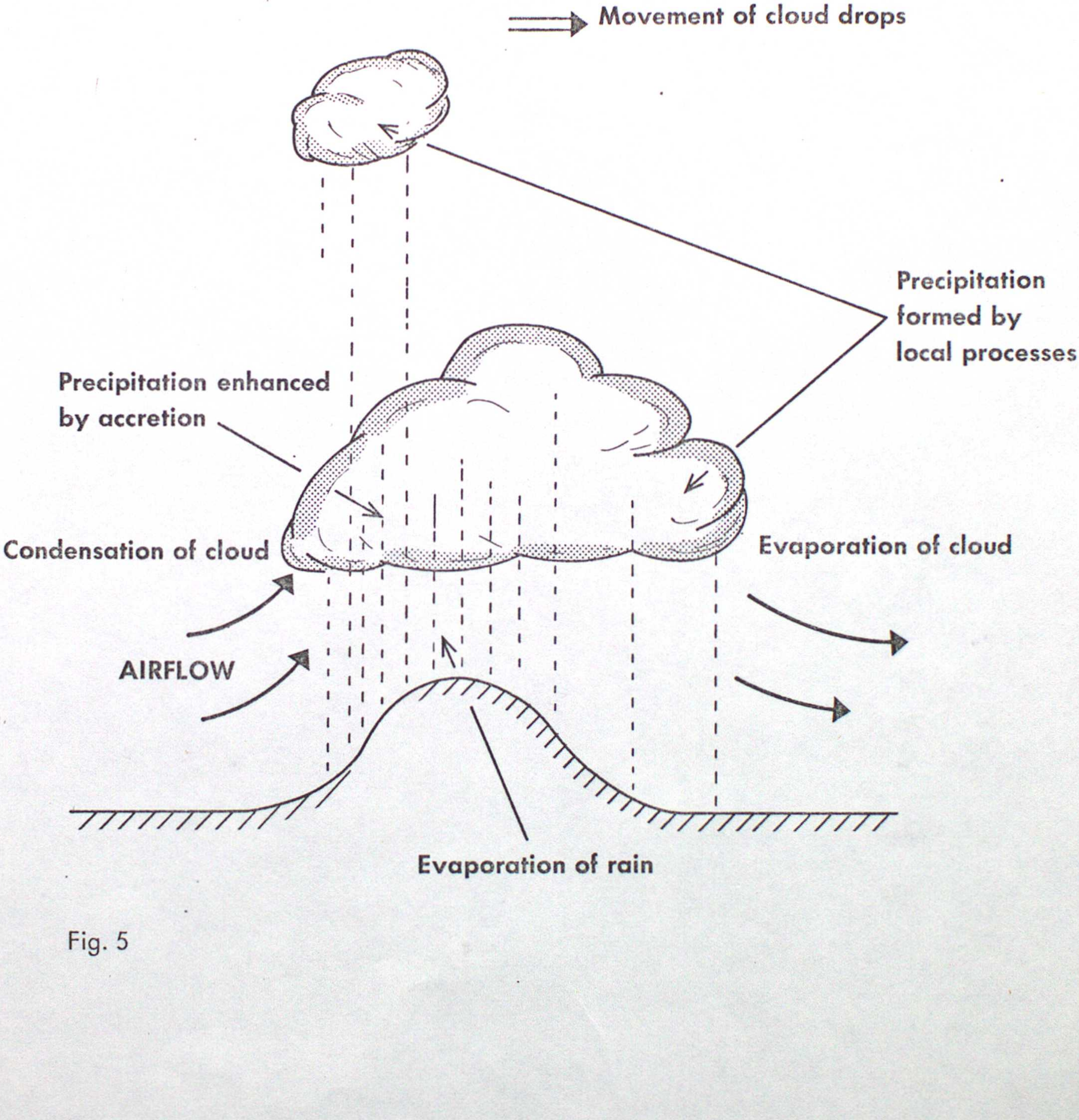

1·0
0·8
0·6
0·4
0·2
0·0
Transmission
0·001
0·01
0·1
1·0
10
Liquid water path

Fig. 4



Movement of cloud drops
Precipitation formed by local processes
Precipitation enhanced by accretion
Condensation of cloud
Evaporation of cloud
AIRFLOW
Evaporation of rain

Fig. 5

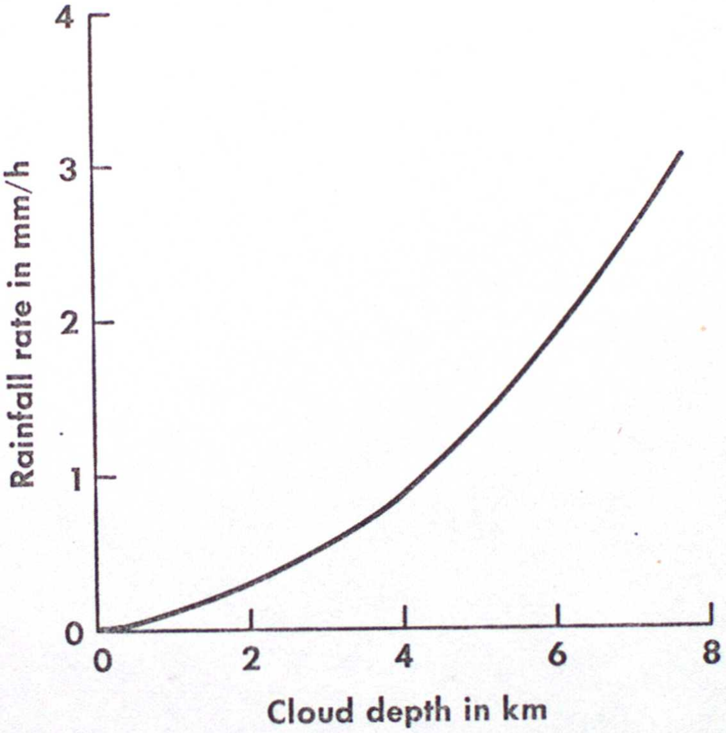


Fig. 6

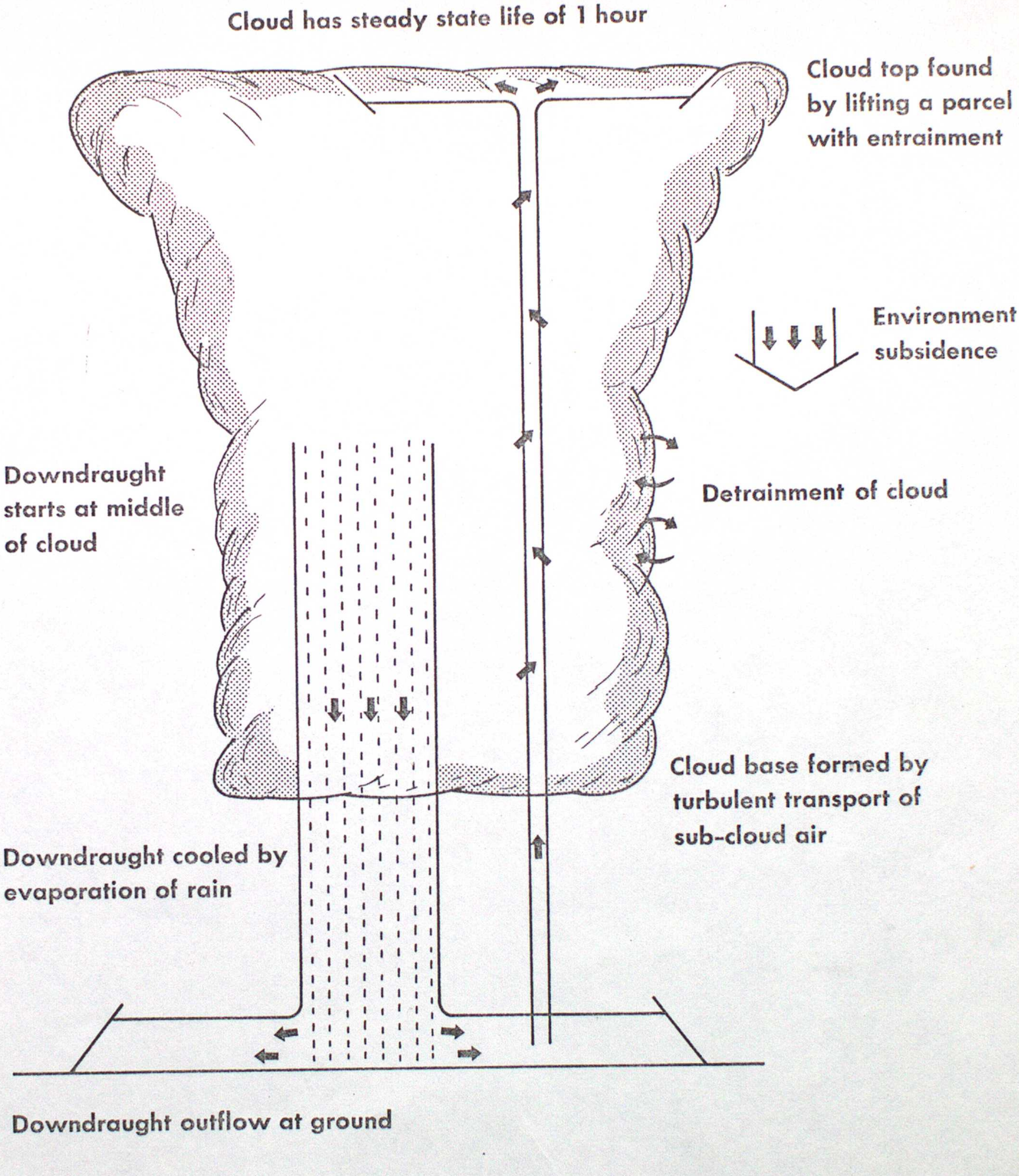

Cloud has steady state life of 1 hour
Cloud top found by lifting a parcel with entrainment
Environment subsidence
Downdraught starts at middle of cloud
Detrainment of cloud
Cloud base formed by turbulent transport of sub-cloud air
Downdraught cooled by evaporation of rain
Downdraught outflow at ground

Fig. 7

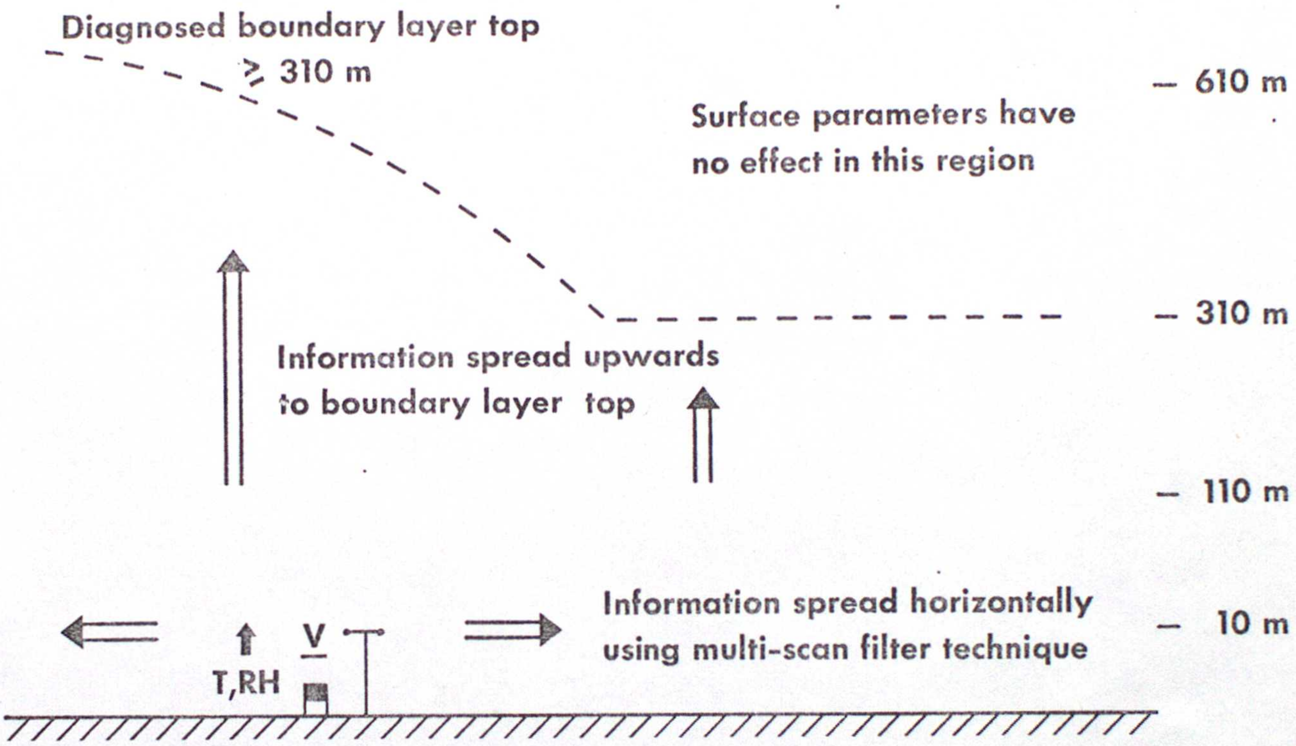

Diagnosed boundary layer top
≳ 310 m
– 610 m
Surface parameters have
no effect in this region
– 310 m
Information spread upwards
to boundary layer top
– 110 m
Information spread horizontally
using multi-scan filter technique
– 10 m
V
T,RH

Fig. 8

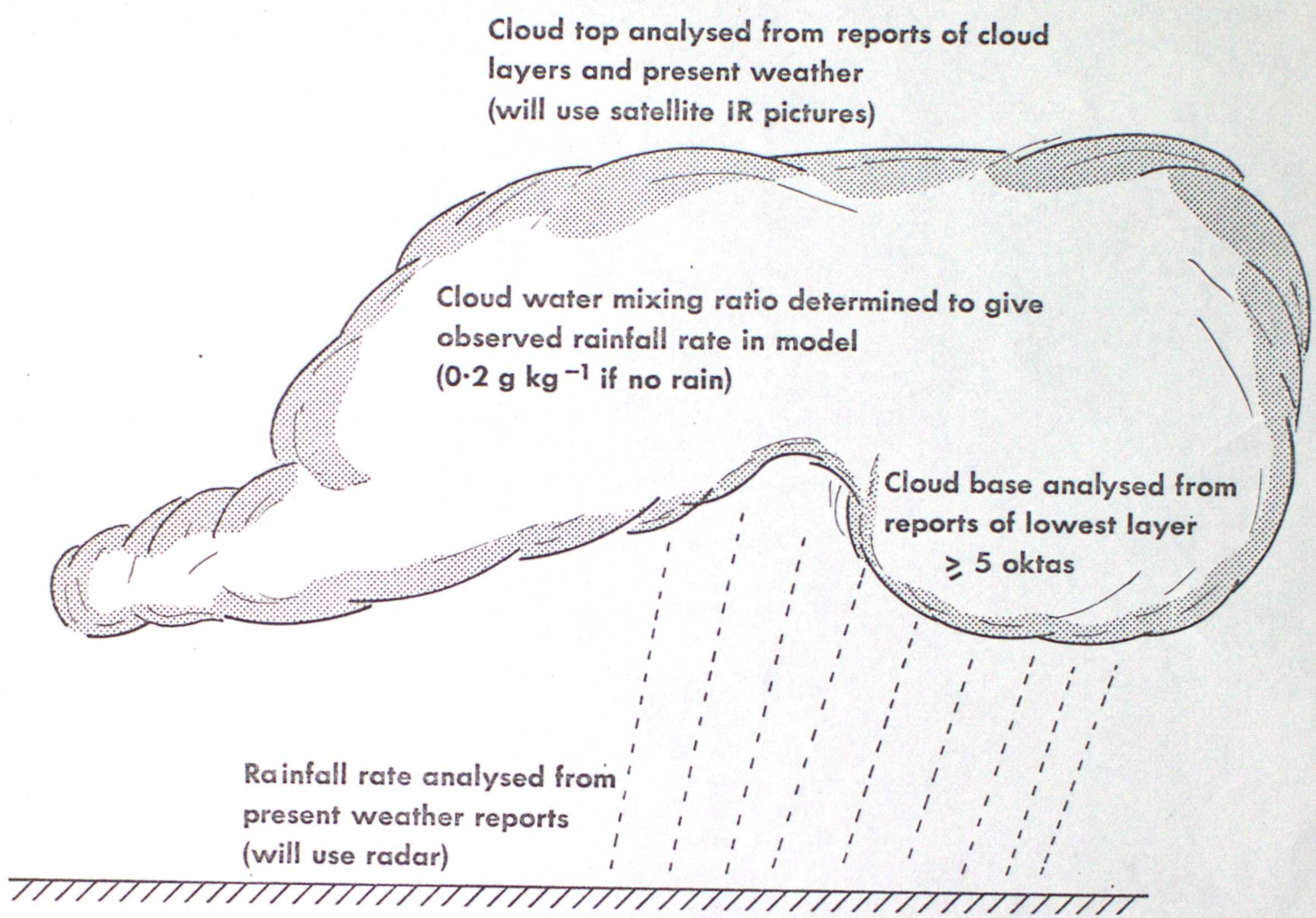

Cloud top analysed from reports of cloud layers and present weather (will use satellite IR pictures)
Cloud water mixing ratio determined to give observed rainfall rate in model (0·2 g kg^{-1} if no rain)
Cloud base analysed from reports of lowest layer ≥ 5 oktas
Rainfall rate analysed from present weather reports (will use radar)

Fig. 9

Fig. 10

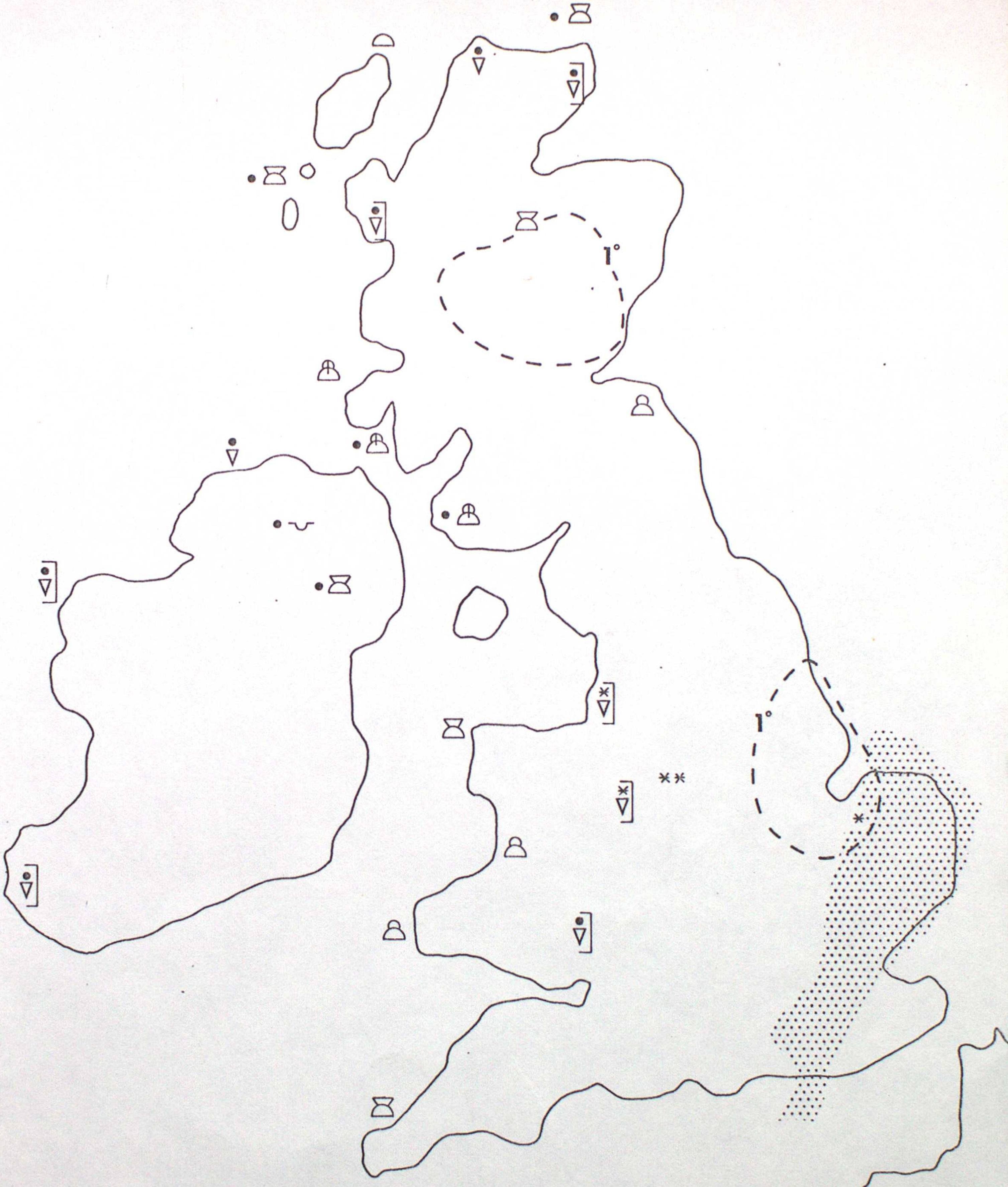


Fig. 11

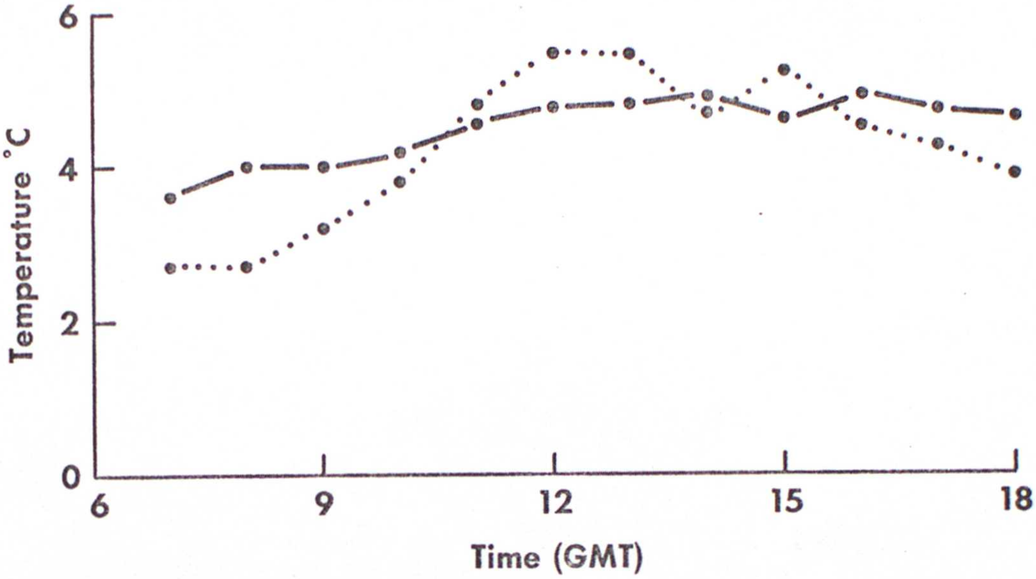

6
4
2
0
Temperature °C
6
9
12
15
18
Time (GMT)

Fig. 12

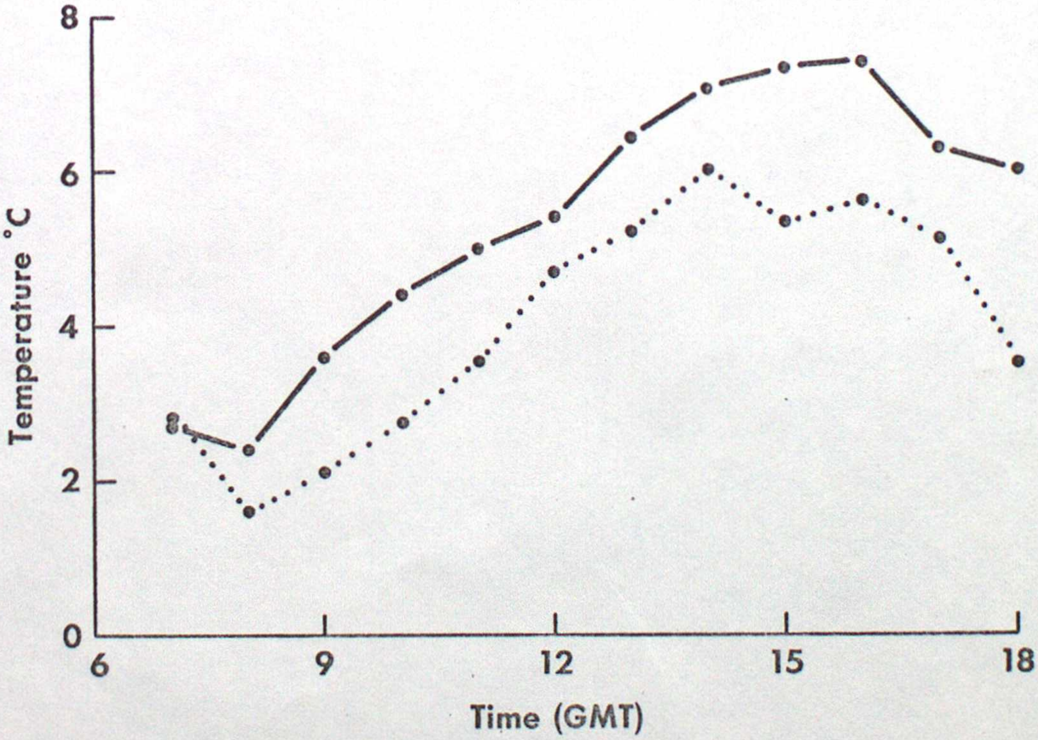

8
6
4
2
0
Temperature °C
6
9
12
15
18
Time (GMT)

Fig. 13