PARAMETERIZATION OF SUB-GRID SCALE PROCESSES.

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1. INTRODUCTION

The role of the parameterization in the forecast model is to take into account the physical forcings which make the equations not purely adiabatic. This includes the exchange of momentum, heat and moisture at the earth's surface, the horizontal and vertical eddy fluxes of the same quantities, the effect of precipitation and radiative exchanges.

I shall not expand on the horizontal diffusion. Although it should represent the effect of the sub-grid scale horizontal fluxes that arise from the discretization of the model, it is in fact used mainly as a mathematical device to control numerical noise in the integration. In the operational model it is done with a linear, fourth order scheme, which is implicit in the east-west direction.

My purpose, in this paper, is not to give a detailed mathematical description of all parts of the parameterization schemes of the ECMWF model. Such a description can be found in the Forecast Model Documentation Manual and in other published papers (Louis, 1979, Geleyn and Hollingsworth, 1979). I shall try, instead, to give the reader an idea of the general principles on which our methods are based, with only enough details to understand the forecast products which are derived from the sub-grid scale parameterization, (namely the near surface temperature and winds, the precipitation and the forecast cloudiness).

2. BOUNDARY LAYER FLUXES

Under this heading I include the surface exchanges as well as the eddy fluxes in the atmosphere. Even though these fluxes are most important in the boundary layer, they are computed throughout the model atmosphere since they can provide a fair amount of dissipation and mixing in region of large vertical shear such as near the jet stream. Results from the surface fluxes computations are also used to determine the two-metre temperature and the ten-metre wind which are among the experimental products of the Centre.

The distribution of levels in the forecast model is such that we can resolve some structure in the boundary layer. As can be seen on figure 1, we have four levels in the lowest 1500 m of the atmosphere, and the lowest level is at about 30 m. Note that the heights shown on the figure are approximate heights above the ground, computed for a standard atmosphere.

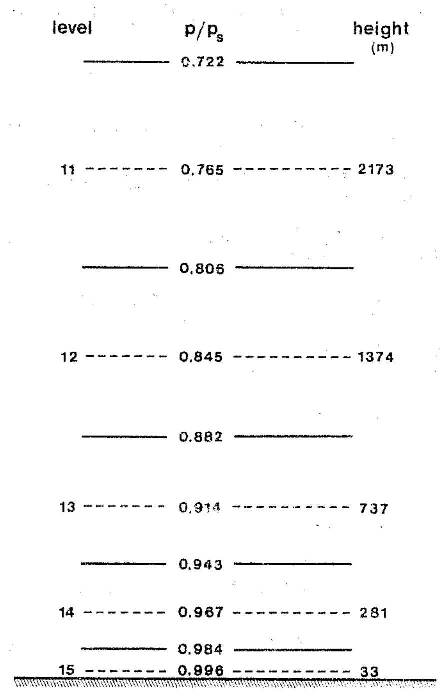


Fig. 1 Vertical distribution of the model levels near the ground

We use the winds and temperature of the lowest level, as well as the ground surface temperature, to compute the fluxes at the surface, according to a drag law. It is well known, however, that surface fluxes are not only dependent on the wind shear at the surface, as in the simplest drag law, but are a strong function of stability. We have used the Monin-Obukhov similarity theory to determine the behaviour of the surface fluxes in terms of the stability.

This theory assumes that, near the ground, the local vertical gradients of potential temperature and wind depend only on the height, the surface fluxes of heat and momentum, and the expansion coefficient of the air. Then, by integrating these flux-gradient relationships between the roughness length z_o and the lowest model level h, it can be shown that the surface fluxes are related to the wind at the lowest model level $\mathbf{V}_h$ and the potential temperature difference $\Delta\theta_h$ in the lowest layer through universal functions of h/z_o and the Richardson number R_i given by

$$Ri = g\ h\ \Delta\theta_h\ /\theta \left|\mathbf{V}_h\right|^2 \tag{1}$$

Hence the expressions for the surface fluxes can be written

$$\begin{aligned} \mathbf{F}_M &= [\ k\ /\ \ln(h/z_o)\ ]^2 \cdot F(h/z_o, Ri) \left|\mathbf{V}_h\right| \mathbf{V}_h \\ F_H &= [\ k\ /\ \ln(h/z_o)\ ]^2 \cdot G(h/z_o, Ri) \left|\mathbf{V}_h\right| \Delta\theta_h \end{aligned} \tag{2}$$

in a way which shows the logarithmic profiles in neutral conditions (i.e. when F=G=1). The constant k is von Karman's constant.

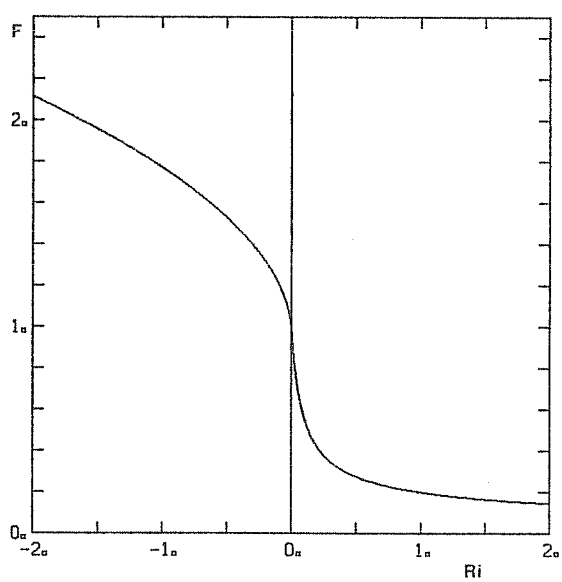


Fig. 2 Variation of the normalized drag coefficient in terms of the Richardson number, for $h/z_o = 50$.

The analytical form of the functions F and G is not uniquely defined. In the course of two years of operational forecasts, we have in fact changed them twice, on the basis of observations and performance of the model. The variation of F with the stability parameter Ri in the current operational model is shown on figure 2 for $h/z_o=50$. This curve represents the ratio between the drag coefficient under varying stability to the one for a neutral atmosphere.

Above the surface layer we have extended the similarity arguments by assuming that, if the model layers are thin enough, the wind and temperature gradients should depend only on the fluxes through the layer, the expansion coefficient, and a mixing length which is a function of height only. The diffusion coefficients are then related to the Richardson number of the layer through formulae very similar to (2).

It should be noted that the effect of moisture on stability must be taken into account since water vapour is lighter than dry air. This is done by using the virtual potential temperature in the definition of the Richardson number.

The two-metre temperature and the ten-metre wind which are disseminated as experimental products of the Centre's forecast should be consistent with the computation of the fluxes described above. In principle (1) and (2) form a system of equations which can be solved for $\mathbf{V}$ and θ at any height, given the surface fluxes $\mathbf{F}_M$ and F_H. However the analytical form of these equations make the solution of this system rather difficult. Hence we do an approximate computation where we first calculate the equivalent roughness length z_o' which would produce the same surface fluxes assuming the logarithmic profile, i.e. z_o' is such that:

$$[\ k\ /\ \ln(h/z_o')\ ]^2 = [\ k\ /\ \ln(h/z_o)\ ]^2 \cdot F(h/z_o,Ri) \qquad (3)$$

Then we use the logarithmic profiles (i.e. Equ.2 with F=G=1) with this equivalent roughness length to compute $\mathbf{V}$ at 2 and 10 m and θ at 2 m. A similar interpolation of the humidity is used to compute the dew-point temperature at 2 m.

It should be noted that, so far, the diurnal variation of the solar radiation has not been included in the operational model: we use, as solar input, the average over 24 hours. This means that one should compare the near-surface variables produced by the model with the averaged observed values and not instantaneous ones. In addition, the absence of diurnal cycle is likely to produce too small a depression of the dew-point temperature.

3. MOIST PROCESSES

We distinguish two kinds of clouds in the forecast model: stratiform and convective.

The treatment of stratiform clouds is quite simple. It is assumed that, whenever the total humidity of a grid point becomes greater than its saturation value, the excess moisture condenses and the corresponding latent heat is released. This excess moisture, however, is not necessarily removed as rain immediately. It only precipitates if either the top level of the cloud is cold enough (below -12 C) or the total liquid water content of the column is large enough (greater than 2 mm). The first criterion takes into account the greater efficiency of ice nuclei at low temperature. The second condition crudely simulates the fact that in a deep cloud the droplet distribution has a wider spectrum than in a thin cloud and the coalescence of droplets into rain drops is more rapid. Finally, as the rain falls, it can re-evaporate in the drier layers below the cloud.

The main weakness of this scheme is the assumption that condensation can occur only when the grid point is entirely saturated, even though it is fairly obvious that stratiform clouds can exist which do not entirely fill a volume nearly 200 km on the side and up to 100 mb thick. In order to overcome this difficulty one would need some additional information such as the mixing ratio of liquid water in the clouds and the variance of humidity in the volume. A drawback of the present scheme is that we only have one prognostic variable for moisture in the model: the total amount of water. Hence we cannot advect separately the water vapour and the liquid water and, at each time step, the liquid water is diagnosed as the amount which is supersaturated. We have preferred to choose 100% relative humidity as the critical value rather than some arbitrary lower value. In a future model we shall have a prognostic variable for the amount of liquid water and we may also be able to relax this 100% relative humidity criterion.

Our convective precipitation scheme follows closely the method proposed by Kuo (1965,1974). A convective cloud exists when there is a net convergence of moisture into a conditionally unstable layer. The amount of cloud air is computed as the ratio of the latent heat contained in the converging water vapor to the excess moist static energy of the cloud. Part of the moisture contained in the cloud is then mixed with the environment, while the rest rains out, releasing its latent heat. This fraction is determined by the relative humidity of the environment. Again, some of the rain can re-evaporate below the cloud.

One should be aware that the division between stratiform and convective precipitation is somewhat arbitrary and model-dependent. As an illustration of this statement we show in figure 3 two forecasts done with the limited area version of the Centre's model. One was done with the operational model. In the other forecast the Kuo convection scheme was replaced by a method proposed by Arakawa and Schubert (1974). It can be seen that, although the

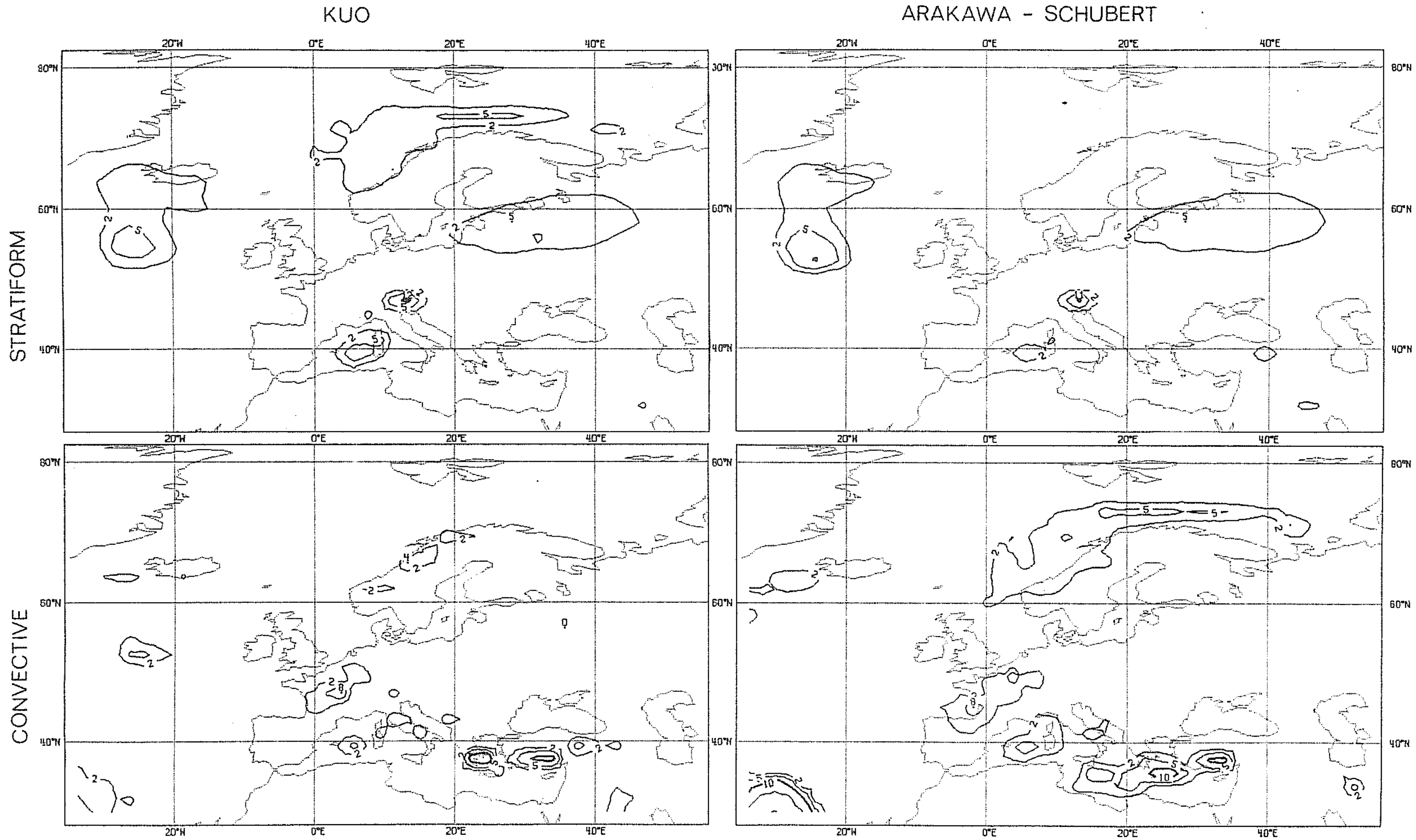


Fig. 3 Comparison of two 24-hour precipitation forecasts

total precipitation is nearly the same in the two forecasts, the partition between stratiform and convective precipitation is very different in the polar region. Another thing which might be worth mentioning is that, at the moment, no distinction is made between snow and rain within the cloud. It is only when the precipitation reaches the ground that it is assumed to be snow if the surface temperature is below 0 C, or rain otherwise. This means that there is a slight energetic inconsistency between the treatment of the clouds and that of the surface processes where we do take into account the heat necessary to melt the snow. This inconsistency will be removed in the future.

4. RADIATION

One can look at the role of radiation in the atmosphere from two different points of view. Globally, radiation provides the primary source of energy for the atmosphere and maintains the thermal gradient between the equator and the poles. Locally, radiation has a large effect on the weather by driving the boundary layer processes and affecting the development of clouds. The importance of the cloud-radiation interactions is obvious for the local weather, but it is no less important for the global heat balance.

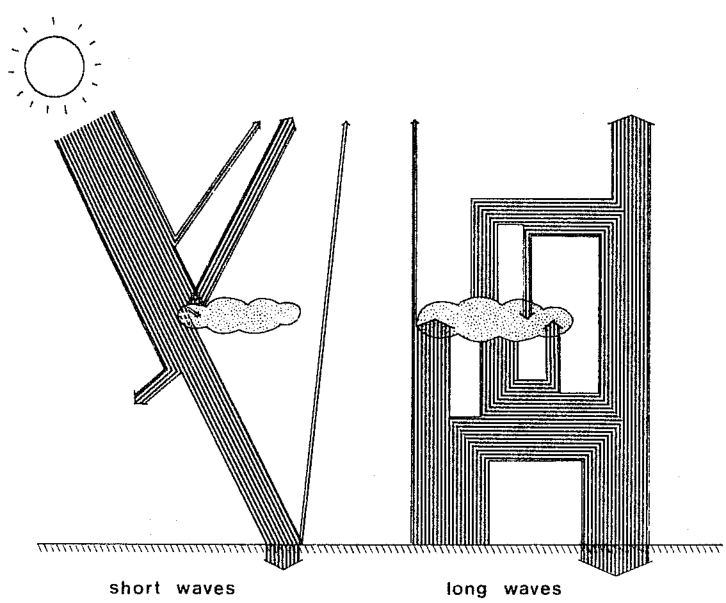


Fig. 4 Qualitative description of the radiative fluxes in the earth/atmosphere system

The diagram of figure 4 illustrates this statement. It describes, in a very schematic way, what happens to the radiative fluxes in the atmosphere. The left side of the figure represents the short wave fluxes, while the right side shows the long wave fluxes. I have not put any numbers on the figure because some of the components are still rather uncertain, but the thickness of the arrows is approximately proportional to the magnitude of the fluxes. The short wave diagram is fairly straightforward, showing that, although little of the sun light is absorbed by the clouds, about one third of the incident solar radiation is reflected by them. The long wave fluxes are a bit more difficult to represent in such a diagram because a lot of absorption and re-emission takes place within the atmosphere, but since the clouds are nearly black bodies in the infra-red, they are clearly important in the transfer of long wave radiation.

In view of the importance of cloud-radiation interactions in both long and short term processes, we have placed a high emphasis on the treatment of the clouds in the radiation scheme. Hence the grey processes are computed first: absorption and scattering by clouds and aerosols, and Rayleigh scattering by the air molecules. In this calculation clouds are allowed to occur in any layer of the model and multiple scattering is taken into account. Then the absorption and emission by the gases (CO_2, H_2O, and O_3) modifies the fluxes resulting from the first part of the computation.

The main problem for the parameterization is to determine the cloudiness of each layer of the model. We cannot, unfortunately, use the information from the parameterization of the moist processes since, with the 100% relative humidity criterion for condensation, we would never have partial cloudiness. Also in its present implementation the Kuo convection scheme does not give us any information on the lateral extent of the cumulus clouds.

We have then used statistics of Pham and Rousseau (1976) on the relationship between relative humidity and the frequency of moist adiabatic lapse rate to evaluate a regression curve between cloudiness and relative humidity. This curve is shown on figure 5 for the 600 mb level. The critical relative humidity below which no clouds are assumed to exist is a inverse function of height. In each layer of the model a fraction of the area corresponding to the curve in figure 5 is assumed to be entirely filled with clouds, and the radiative fluxes are computed separately in the cloudy and clear parts. When clouds exist in two adjacent layers, maximum overlap of the cloudy parts is assumed.

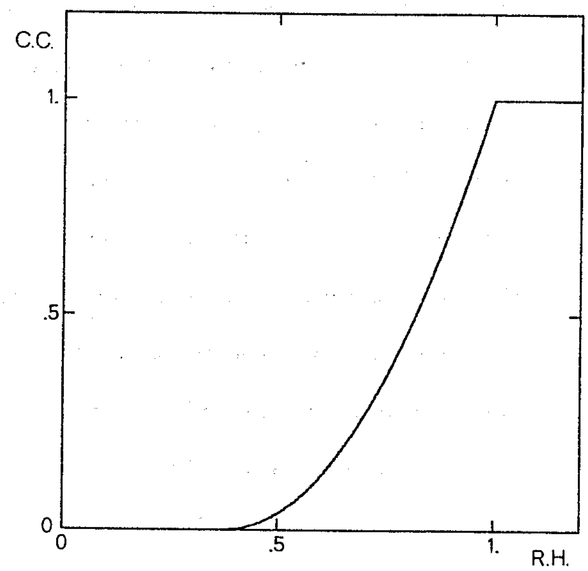


Fig. 5 Assumed relationship between cloud cover and relative humidity at $p/p_s = 0.6$.

It is evident that such a parameterization is very crude, and likely to be poor in the regions which have either tall cumuli penetrating into dry layers, or thin clouds which would not entirely fill a whole layer of the model. The latter problem is particularly bad in the case of thin stratus clouds developing at the top of the boundary layer. Radiation is important in the development of these clouds, with strong cooling at the top and warming within the cloud which destabilises the cloud layer. In the model, however, with a relatively coarse vertical resolution and our assumption that the cloud occupies the whole model layer, the effect of radiation tends to destabilise the whole boundary layer and produce a wrong feed-back involving vertical diffusion and condensation. In order to avoid an exaggerated precipitation in this situation we have been forced to suppress the clouds at the top of an unstable boundary layer, as far as the radiation scheme is concerned.

Once we introduce a prognostic variable for the liquid water content we hope to be able to connect in a more consistent fashion the cloudiness in the radiation scheme and the liquid water determined by the moist processes. At any rate this cloudiness parameter is a quantity which can be directly compared to observations. This is why we routinely produce pseudo-satellite pictures which are simply this radiation cloudiness, integrated over all the levels, and plotted as various shades of grey. An example is given in figure 6, compared with the actual satellite photograph.

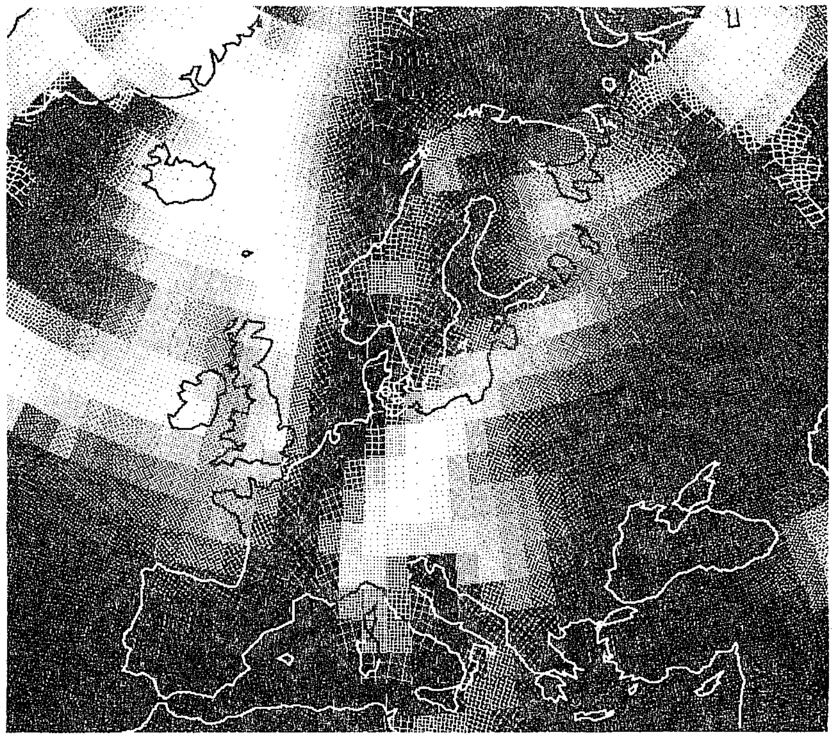
5. GROUND PROCESSES

The model has prognostic equations for the surface temperature and moisture. At the time of writing, the method of computing these variables is about to be modified, and I shall describe the new scheme. It is designed so that we can switch on the diurnal cycle which, as already mentioned, is now suppressed. The scheme is a straightforward simulation of the diffusion equation in the ground, using a finite difference scheme with three layers. The top layer is thin enough to react to the daily cycle. The second layer responds to changes with time scales of about a month. Finally, a climate value is imposed in the bottom layer. The boundary condition at the top is the net flux at the surface: radiative, sensible and latent heat fluxes for the temperature; precipitation and evaporation for the moisture.

Nearly a year ago, a bulk parameterization scheme was introduced, following the ideas of Deardorff (1978). Various problems with our implementation of this scheme have delayed a subsequent introduction of the diurnal cycle. We hope that these difficulties will be avoided with this new scheme so that we can turn on the diurnal cycle soon.

Another element of the parameterization which is important for the radiation is the treatment of the snow, since there is a strong feed-back between the

ECMWF 24 hr forecast for 19-7-81



NOAA-6 satellite photograph [visible]

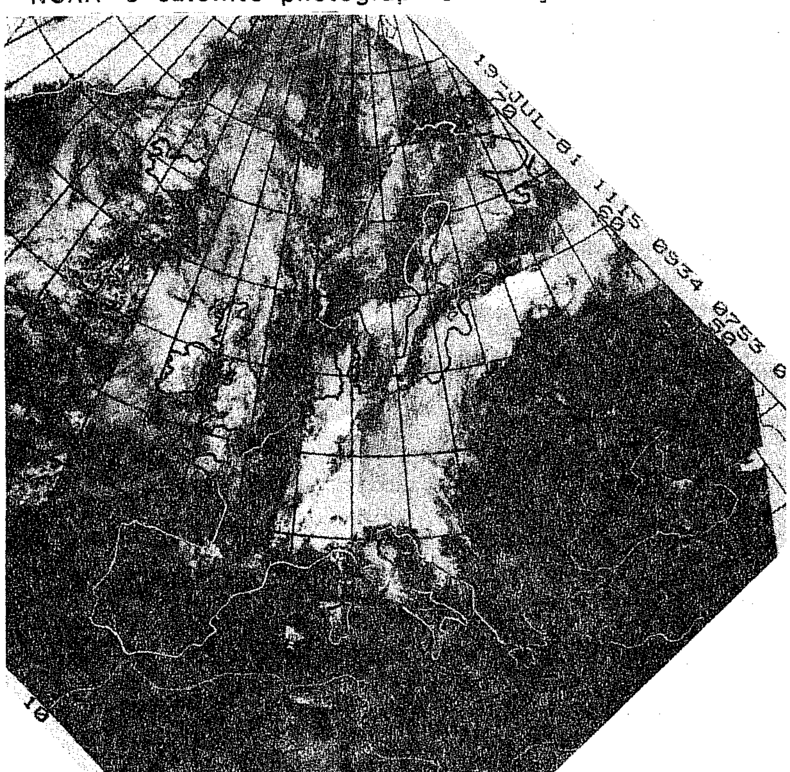


Fig. 6 An example of a 24-hour cloudiness forecast, with the corresponding satellite photograph

albedo and the ground temperature in the presence of snow. Snow is assumed to have an albedo of 0.8. However, in order to avoid shocks which might produce noise in the forecast, we assumed that the albedo changes smoothly between the value of the bare ground and that of the snow, depending of the snow cover which is taken as a monotonic function of snow depth. The albedo value of 0.8 is reached when the snow depth is about 1 m.

There are a number of weaknesses in the treatment of the snow in the operational model. I have already mentioned that precipitation is assumed to be snow only when the ground temperature is below freezing. Another problem, which is due to the fact the we do not carry information about vegetation in the model, is that we do not take into consideration the fact that forest can have a relatively low albedo, even with deep snow, if it is windy. The change of albedo with the age of the snow is not considered either. Finally, and probably most important, is the fact that we do not change the heat capacity and conductivity of the ground when snow is present. This results in frequent overestimation of the snow covered ground temperature and too rapid melting of the snow. We hope to correct this latter problem some time in the near future.

6. CONCLUSION

I hope that this quick overview of the parameterization schemes of our forecast model has given the reader an idea of the methods we use. It is evident that all the schemes are fairly simple, even though they nonetheless represent the current state of the art in large scale model parameterization. One reason for this simplicity is the need for fast computation. In the present operational model the parameterization takes up about 50% of the whole computing time. We would not want this figure to increase too much with more sophisticated methods. Already we are forced to perform the radiation calculation only twice per forecast day because it would be prohibitively expensive to do it more often in its present form.

In addition to the need for fast computation, we also prefer to keep our schemes simple,(with few arbitrary parameters) in order to understand their behaviour more easily; but at the same time we try to include in the schemes all the significant interactions between the various processes, in order to simulate as well as possible the effect of all the feed-back loops. In future development, we shall try first to remove some of the inconsistencies which still exist, such as the different definition of cloudiness in the condensation and radiation, problems with the snow, and the absence of diurnal cycle.

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