PARAMETERIZATION OF LAND-SURFACE PROCESSES

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1. INTRODUCTION

The parameterization of land-surface processes is a field of growing interest in climate modelling where the forcing by the lower boundary conditions is of prime importance in the determination of the model's steady state. Up till now few studies have addressed this problem in the context of a forecast model, but experimentation undertaken for example at the British Meteorological Office (Rowntree and Bolton, 1983) and at ECMWF gives an idea of how the specification of some of the land surface properties may affect a meteorological forecast within a few days.

The purpose of this paper is to review the aspects of land-surface parameterization which appear to be particularly relevant to medium-range weather forecasting. The physical processes and hypotheses currently considered in numerical global models are discussed, and the present ECMWF formulation is presented as an example. Two key problems are also addressed: the validation of surface schemes and their possible impact on medium-range weather forecasts. Within a forecast model, the representation of surface processes forms the basis for the computation of surface and near surface variables over land, and of surface fluxes, the later being a major element of the boundary layer parameterization. The problem can be summarised as follows: the model computes at each time-step grid-averaged variables at all levels, including the lowest level ℓ which is often supposed to be at the top of the "surface layer" or "constant flux layer" of the atmosphere. Assuming that grid-averaged surface properties are computed by a diagnostic or prognostic method, surface fluxes are obtained from a knowledge of model variables at level ℓ and surface variables by formulations generally based on the surface layer similarity theory. The surface fluxes are assumed to be representative of a model grid area and are used to compute the vertical flux divergence; this gives the contribution to the temporal evolution of the model variables at level ℓ. These fluxes represent the lower boundary input of heat, moisture and momentum, this input being distributed within the atmosphere by the boundary layer and other physical schemes.

The relative importance of the determination of surface fluxes can be appreciated by considering a typical global budget for the various quantities involved. The global energy budget, as displayed in an example of a 10 day forecast (Fig. 1), includes a net radiation term at the top of the atmosphere. This corresponds to a sink of thermal energy which is compensated by heat sources due to parameterized subgrid convection, condensation at resolved scales and the sensible heat flux from the surface. The sensible heat flux (about 15 Wm^{-2}) is only a small fraction of the total which is indeed dominated by the condensation and radiative effects. The heat sources due to condensation (about 80 Wm^{-2} for the grid scale and subgrid scale contributions together) are, on the average, in balance with the latent heat flux from the surface and represent a major component of the global heat budget. If one compares these sources with the mean global energy, 2.6×10^9 J m^{-2} in the present example, this leads to a recycling timescale of the order of 300 days by the surface fluxes, or an average heating of the order of 1°/day for the whole atmosphere. A systematic error in the evolution of those fluxes in the model may then affect the mean temperature by a fraction of a degree per day.

Similar figures for the atmospheric water or kinetic energy budget are more revealing with respect to the relative importance of surface fluxes: in the same forecast, the globally averaged water content was equivalent to a depth of 26 mm, with an average surface flux of 2.6 mm/day, leading to a recycling time scale of 10 days. The mean soil water content over land in the model was 20 mm, which suggests a somewhat slower recycling for the soil water than for the atmospheric water.

A typical recycling time scale for the kinetic energy of the atmosphere would be of the order of 6-8 days with a global kinetic energy of 1.5×10^6 Jm^{-2} and a dissipation rate, mainly due to boundary layer processes, of the order of 3 Wm^{-2}. The above figures vary somewhat with seasons (cf. Oort and Peixoto, 1983) and can be slightly model dependent, but they give an idea of the magnitude of the effects on the surface evolution which can be expected from surface related processes. In this context it is useful to remember that land covers abound one third of the total earth surface and is responsible for about that fraction of the surface heat fluxes, a somewhat larger fraction of the surface energy dissipation due to roughness and orographic effects, and a somewhat smaller fraction of the latent heat fluxes, because of the large vertical exchanges above the main oceans currents, although the exact figures are difficult to estimate.

Another important order or magnitude to keep in mind is the area covered with snow or sea ice which obviously deserves special treatment in atmospheric models. Following Woods (1984) the land area covered with snow varies from 63×10^6 km^2 during the northern hemisphere winter (49×10^6 km^2 in the northern hemisphere and 14×10^6 km^2 in Antarctica) to 16×10^6 km^2 during the northern

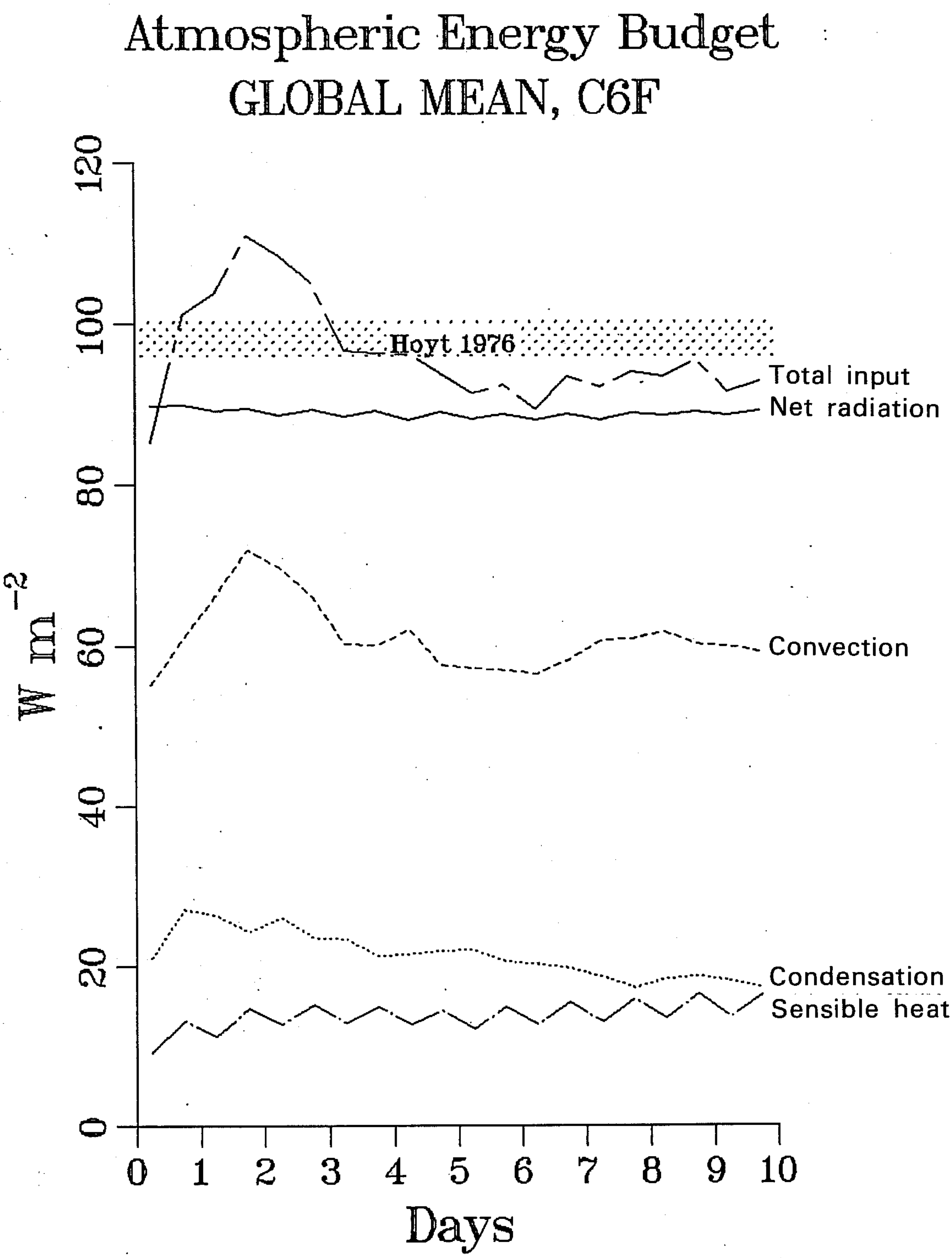

Fig. 1 Global atmospheric energy budget in an example of 10-day forecast (30 May 1985).

hemisphere summer (Antarctica plus Greenland and mountain glaciers) out of the 150×10^6 km^2 of exposed land. The sea ice extent varies for the same seasons from 17×10^6 km^2 to 23×10^6 km^2.

The remainder of this paper will describe the physical basis and usual formulations for present land surface parameterization schemes, leaving aside most of the specific problems related to snow covered land and the important topic of the representation of vegetation, which is treated in the paper by P. Sellers. Some remarks will also be made on the use of satellite data to improve the determination of key parameters and to validate model results, and on the impact of surface schemes within forecast models.

2. MAIN PHYSICAL PROCESSES

2.1 Definition of the processes involved

As mentioned earlier, the physical processes which will be considered here are those which directly affect either the evolution of the surface temperature and moisture or the specification of surface fluxes, and which then have an impact on the meteorological variables. It is also important to stress that those processes, which are often studied at the local scale, need to be averaged over areas of the order of 100x100 km before being considered for parameterization in a large-scale forecast model.

A simple way to summarise the problem is to examine the budget of the various physical quantities which are exchanged at the earth's surface: thermal and radiative energy, water and momentum.

As displayed in Fig. 2, the surface (assumed to be a thin layer of negligible thermal inertia, characterised by a temperature T_s and a specific water content W_s) exchanges radiative energy with the atmosphere and thermal energy with the atmosphere and the underlying ground. If $S\downarrow$ denotes the flux of incoming solar energy (visible and near infra-red), a fraction (α_g) is reflected by the surface, α_g being the albedo. The exchanges of thermal infra-red radiation consist of the net incoming flux from the atmosphere $\varepsilon_g R_L$ and the upward flux $\varepsilon_g \sigma T_s^4$ due to the surface thermal emission, where ε_g is the emissivity of the surface, and σ the Stefan-Boltzmann constant. The overall effect of radiative processes is a net input of energy to the surface which is compensated by the heat diffusion into the ground (F_G) and the turbulent exchanges of sensible and latent heat ($F_H$and LF_W). Here L is the specific latent heat for evaporation. These processes of energy conversion at the surface are essential to the determination and regulation of the earth's climate as about 30% of the total solar incoming radiation at the top of the atmosphere is transformed into sensible heat at the surface, and is returned to the atmosphere by the turbulent fluxes of latent and sensible heat.

As an illustration of a typical diurnal evolution of the above fluxes, Fig. 3 displays the values of the net radiative flux R_N and of the thermal and latent heat fluxes during a clear day over irrigated turf. In this case, where the moisture availability is high, the latent heat flux almost compensates the net radiative input with a slight time lag. The sensible heat flux rises steeply in the morning but decreases and becomes negative in the afternoon due to the strong evaporational cooling. The sensible heat flux in the ground appears to be a small fraction of R_N, and is correlated with it.

SURFACE EXCHANGES

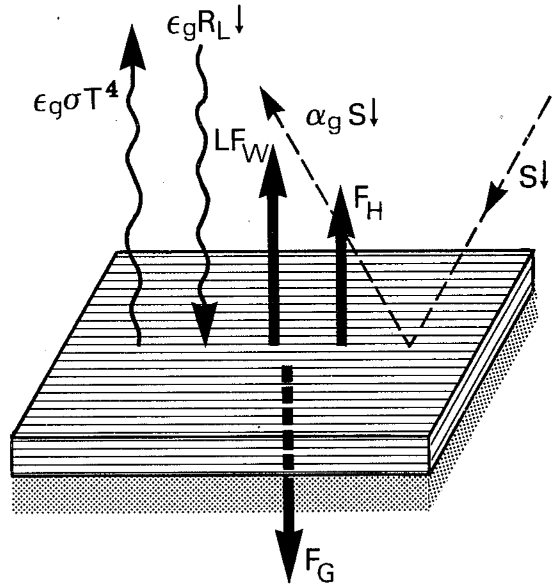

Fig. 2 Schematic illustration of the surface energy budget.

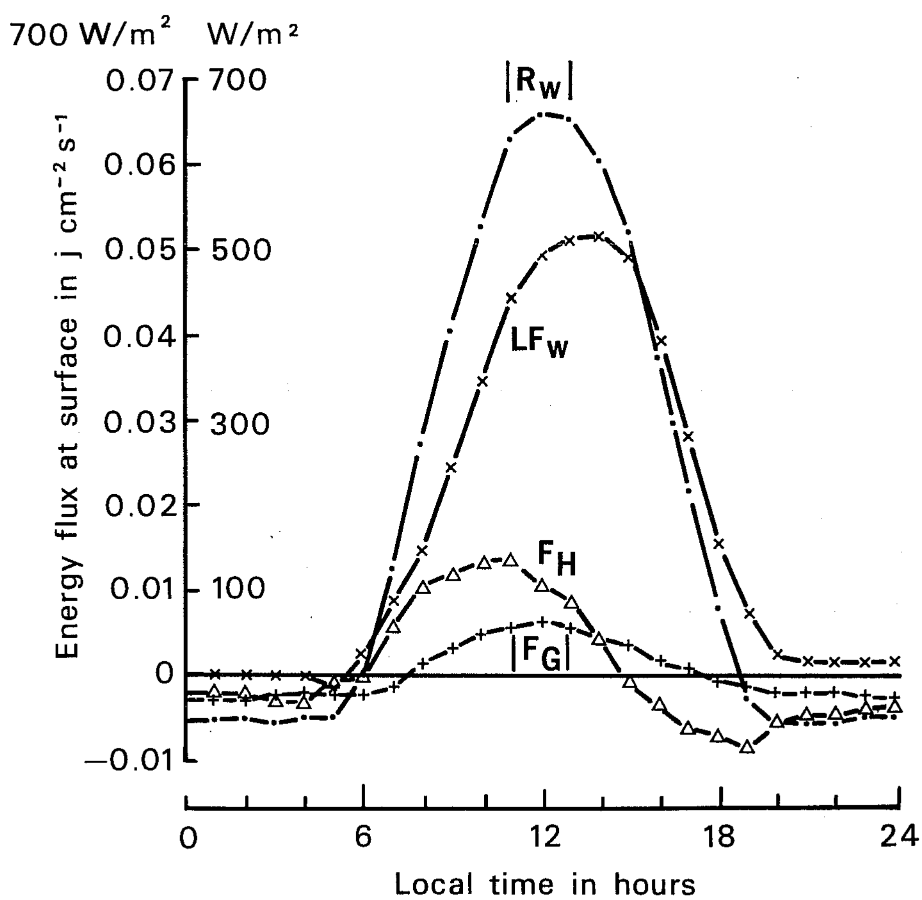

Fig. 3 The diurnal variation of the energy fluxes at the surface as measured by Brooks and Goddard on 3 June 1965 over well irrigated turf at Davis, California.

As for the surface water budget, it is convenient to consider separately the budget for liquid water and snow (Fig. 4). The liquid water input is by liquid precipitation (P_R) and snow melt (M_{Sn}), and this is compensated by percolation into the ground, run-off, and by the turbulent surface moisture flux F_W. The snow mass per unit surface Sn evolves through the effects of solid precipitation P_S, snow melt and the fraction of surface moisture flux F_W which corresponds to sublimation over snow covered areas (L is then the latent heat coefficient for sublimation).

The last budget to be considered is that for momentum, for which the surface acts as a sink because of the turbulent momentum flux denoted later on by u_*^2.

The various processes introduced here will now be reviewed in the context of their representation in large-scale atmospheric models.

2.2 Radiative processes

A good review on this subject is available in Dickinson (1983) and only a few basic ideas will be presented here. The computation of radiative fluxes within the atmosphere is covered elsewhere in this volume and the incoming fluxes $S\downarrow$ and R_L are supposed to be known from other parts of the model.

The incoming solar flux is a combination of a direct beam (that which remains from the solar radiation after partial absorption through the various atmospheric layers) and a diffuse radiative flux. The relative amounts of these within $S\downarrow$ is a function of the diffuse elements encountered by the solar beam (clouds and aerosols), the solar angle and the wavelength considered. It is customary to divide the solar spectrum into two main bands for this purpose, which in the ECMWF model corresponds to a visible window (.24-.78 μm) and a red/near infrared window (.78-4.65 μm).

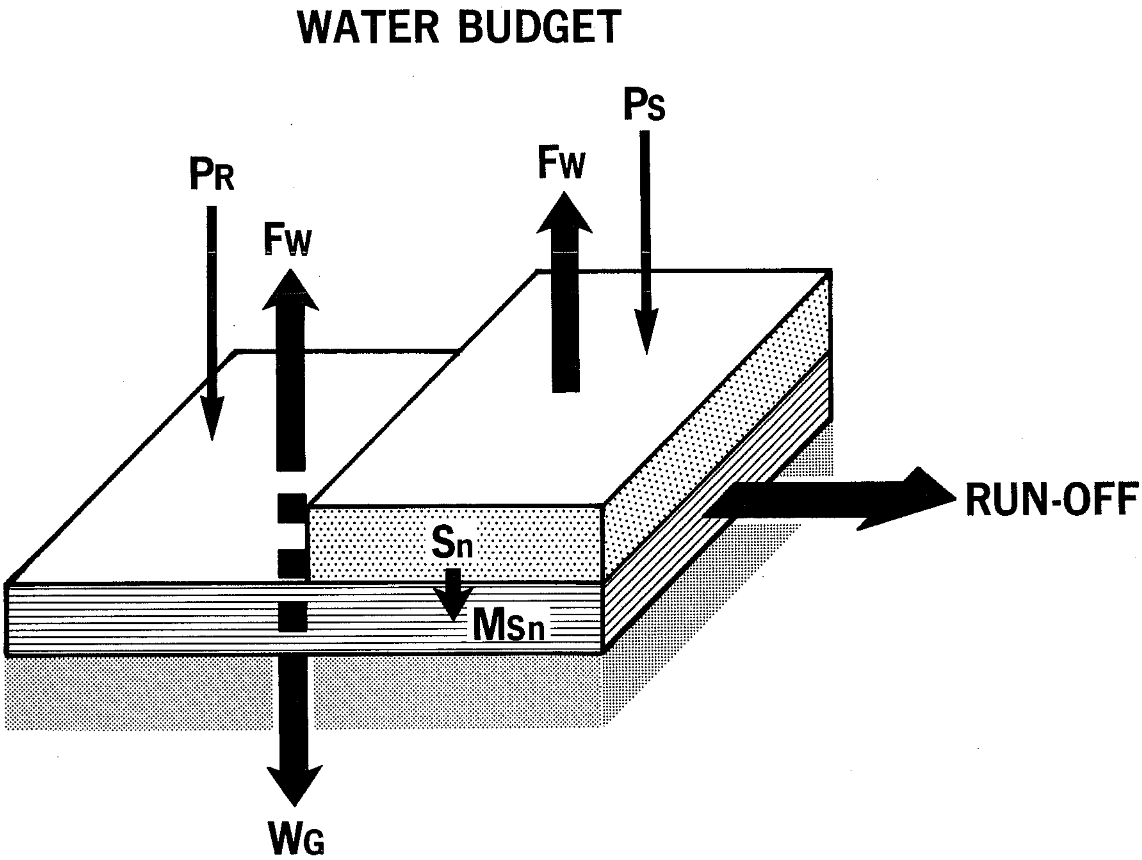

Fig. 4   Schematic illustration of the surface moisture budget.

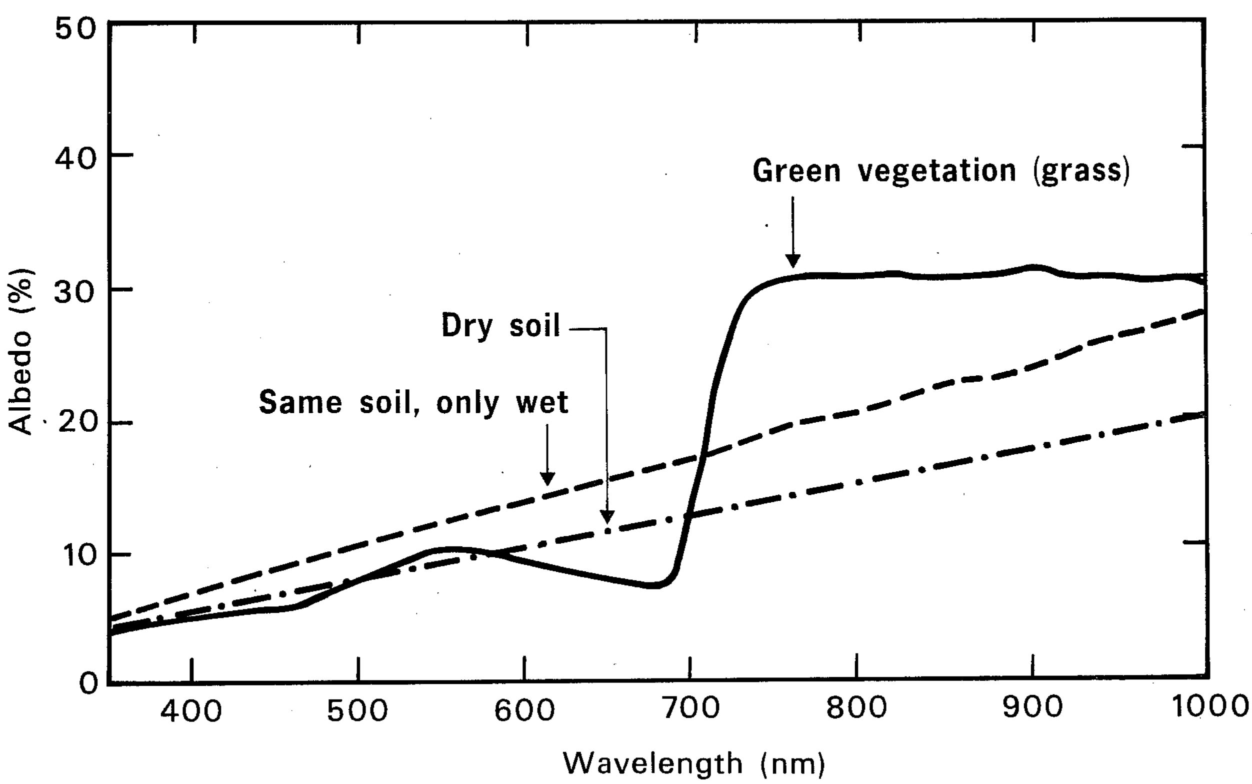

Fig. 5   Spectral reflectances for dry soil, wet soil, and a plot of blue gamma glass (from Tucker and Miller).

The reflective properties of the surface which contribute to the albedo, have to be defined in principle for a given angle of incidence and a given wavelength, and depend upon several of the soil properties, mainly its physical structure (height and shape of roughness elements), chemical composition, moisture content and type of vegetation cover, as illustrated on Fig. 5. The local albedo should in principle be obtained by a spectral reflectance, and again be spatially integrated in order to get a value compatible with a model grid-scale. In practice, as in the ECMWF model, a grid-averaged albedo for diffuse solar light is geographically defined without spectral or moisture dependency. The ratio of direct to diffuse incoming radiation is taken into account as well as the presence of snow. In nature the reflective properties of snow are dependent on snow depth and snow granulation, which varies with the age of snow cover. In the ECMWF model a simple formulation is used which assumes a diffuse albedo of .8 when the grid-averaged snow depth measured in equivalent water depth is larger than 1.5 cm, and uses a linear interpolation with the snow-free value for averaged snow depths up to 1.5 cm.

The second parameter which has to be specified for radiative flux computations is the surface emissivity for thermal radiation ε_g. It should in principle be dependent on the soil type and vegetation cover as well as on the wavelength. Most forecast models, however, use a uniform value of about one (.996 in the ECMWF model). Actual data on emissivities are sparse and are not as reliable as for albedo; a review on this subject is available in Kondratyev et al., 1982 and indicates a natural range of variability between approximately 0.9 and 1.0 with a few percent difference for various infra-red windows. However, according for example to Rowntree et al. (1985), the possible spatial variations of surface emissivity would have a negligible effect on the main features of the atmospheric circulation computed by a G.C.M.

2.3 Turbulent processes at the soil-atmosphere interface

As mentioned earlier, the fluxes which have to be computed at the soil-atmosphere interface are the fluxes of heat, water vapour and momentum. Their formulation with respect to atmospheric variables is discussed elsewhere in these proceedings as part of the boundary layer parameterization problem. What is relevant here is the determination of parameters which enter the surface flux equations and which are dependent on the surface characteristics within a model grid square.

The usual surface flux formulations assume that the first model level corresponds to the top or is within the atmospheric surface layer, where fluxes can be expressed by the Monin-Obukhou similarity laws as a function of surface and first model level variables. A typical formulation, as used in the ECMWF model, is:

$$u_*^2 = a^2 u_\ell^2 \phi_m \qquad (1)$$

$$\frac{F_H}{\rho C_p} = \frac{a^2}{R} u_\ell \Delta\theta \phi_h \qquad (2)$$

$$\frac{F_W}{\rho} = \frac{a^2}{R} u_\ell \Delta q \phi_q \qquad (3)$$

where ρ is the grid-averaged density, C_p the specific heat coefficient at constant pressure, index ℓ denotes values at the first model level, $\Delta\theta$ is a potential temperature difference between the surface and level ℓ, a^2 is a positive constant modified by $\frac{1}{R}$ in the case of heat and water vapour transports, and ϕ_m, ϕ_h and ϕ_q are weighting coefficients function of the thermal stability of the surface layer which take a value of 1.0 for neutral stability (generally $\phi_q=\phi_h$). In the neutral case, this formulation reduces to the usual drag coefficient formulation with:

$$a^2 = \frac{k^2}{\left[\ell n\left(\frac{z_\ell}{z_o}\right)\right]^2} \qquad (4)$$

where k is the von Karman constant and z_o a roughness length. Δq is usually defined as:

$$\Delta q = \beta [q_\ell - q_{sat}(T_s)] \qquad (5)$$

where $q_{sat}(T_s)$ is the saturation specific humidity at the surface and β is an efficiency coefficient which varies between 0 and 1. The main problem discussed below is to specify z_o and β as a function of subgrid surface properties.

(a) Determination of z_o

The first step is to define z_o from micrometeorological considerations, then it has to be generalised for use in a large-scale forecast model. In the lower portion of the surface layer, referred to as the dynamic sublayer, the thermal dependency becomes negligible and simple "law of the wall" similarity formulae can be used such as for the long wind velocity profile:

$$\overline{u} = \frac{u_*}{k} \ln \left(\frac{z}{z_{om}}\right) \qquad (6)$$

where z_{om} is an integration constant which defines a "momentum roughness parameter" a priori dependent on the flow, that is on u_*. This formulation is valid at levels where viscosity is negligible and should be replaced by a viscosity dependent formulation within the viscous sublayer, a thin layer above the surface where the flow is quasi-laminar and dominated by viscosity. The value found for z_{om} is normally a function of the depth of the viscous sublayer, itself related to the height of roughness elements. When the size of the roughness elements is sufficiently large, z_{om} tends towards a value z_o independent of u_*; z_o is characteristic of the surface geometry, that is of the size distribution of roughness elements. A measure of the surface roughness with respect to the flow is the roughness Reynolds number $z_{o+} = \frac{u_* z_o}{v}$. Similarity considerations along with experimental data suggest:

$$z_{om} \simeq .135 \frac{v}{u_*} \quad \text{for } z_{o+} < .13 \text{ (smooth surface)} \tag{7}$$

$$z_{om} \simeq z_o \quad \text{for } z_{o+} > 2.0 \text{ (rough surface)} \tag{8}$$

It is easy to show that most natural surfaces are rough under normal meteorological conditions, which justifies the choice of one specific z_o. When the surface is very rough, and especially when one wishes to define wind profiles at levels commensurate with the size of roughness elements, there is a need to use a displacement height d in the law of the wall, which corresponds to an upward translation of the origin of the order of the size of roughness elements:

$$\overline{u} = \frac{u_*}{k} \ln\left(\frac{z-d}{z_o}\right) \tag{9}$$

This can be important when defining meteorological conditions at 2 or 10 meters over vegetated areas.

As the heat and water vapour transfers in the viscous sublayer are of a different physical nature from the momentum transport, there is a priori no reason for the corresponding constants obtained by the similarity law, z_{oh} and z_{ov}, to be equal to z_{om}. In fact, for rough surfaces, which is the only case of practical interest, and for bluff elements, approximate formulations (from Brutsaert, 1982) are:

$$\frac{z_{ov}}{z_o} = 7.4 \exp\left(-2.85\, z_{o+}^{1/4}\right) \tag{10}$$

$$\frac{z_{oh}}{z_o} = 7.4 \exp\left(-2.46\, z_{o+}^{1/4}\right) \tag{11}$$

Very often natural surfaces are partly permeable or offer greater surface area for heat and water exchanges than for momentum, thus increasing the ratios z_{ov}/z_o and z_{oh}/z_o up to values of the order of 0.3 to 0.5 over forests, a typical value over grass being of the order of 0.1. An extensive account of this subject from the micrometeorological point of view is given in Brutsaert (1982).

The definition of z_o in a large-scale model requires, however, another stage of reasoning in order to extend the local concept to a grid-scale of the order of 100 km. As the flux profile relationships are very non-linear, the portion of the grid which really counts in the averaging of vertical fluxes is that which correspond to large fluxes for a given vertical profile, thus to large z_o. A discussion of this problem is available in André and Blondin (1984) and concludes that an effective z_o for a model grid square should be of the order of the largest local z_o within the area.

Current values used for z_o in large-scale models over flat terrain are of the order of a few centimetres over non-vegetated terrain or over small vegetation, of the order of one metre over forests and a few metres over towns. Those values are artificially increased in some models over orography to take into account the drag effect of small orographic features, leading to the choice of values up to around 10 metres in mountainous areas, but this is without sound theoretical or experimental justification. Very often the same values of z_o are used for heat and water vapour fluxes because there is a lack of precise experimental data at the right scale. When Businger's data from the Kansas experiment are used, which is often the case, slightly larger drag coefficients, (equivalent to larger values of z_o) are applied for heat and moisture; in the case of neutral stability this results has however been questioned and is no longer included in the ECMWF formulation. Now it appears clearly that if z_o is increased for momentum fluxes as an empirical way to take into account subgrid orographic features, there is no reason to use the same procedure for heat and moisture fluxes.

(b) Determination of β

Equation (5) assumes that the surface moisture flux is a fraction β of the flux which would occur over a water surface with the same z_o, referred to as the potential evapo-transpiration E_p:

$$E_p = \rho \frac{a^2}{R} u_\ell [q_\ell - q_{sat}(T_s)] \phi_R \qquad (12)$$

Up till now the usual approach in large-scale models has been to assume that β is a function of the specific moisture content at the surface, W_s, and possibly of the snow cover Sn. The simplest so-called Budyko formulation used in several models is:

$$\beta = \frac{W_s}{W_{cr}} \text{ over snow free areas}$$

$$\beta = 1 \quad \text{over the fraction of the grid assumed to be snow-covered} \qquad (13)$$

W_{cr} is a critical value at which the soil is supposed to behave as a water surface.

An alternate formulation which can be more easily generalised for various soil and vegetation types is based on resistance formulae of the type:

$$E_p = \rho \frac{[q_\ell - q_{sat}(T_s)]}{r_{a\ell}} \qquad (14)$$

where $r_{a\ell}$ is an atmospheric resistance, which is a function of atmospheric properties at level ℓ and of the surface temperature T_s. The evaporation E is then given by:

$$E = E_p \frac{r_{a\ell}}{r_{a\ell}+r_s} \qquad (15)$$

where r_s is a surface resistance, which is a function of variables defined at the surface such as soil moisture, vegetation properties and surface radiative fluxes. An extensive discussion of this approach in the case of a vegetation cover is proposed by P. Sellers in this volume.

In this framework an interesting formulation is the so-called Penman-Monteith approach, which overcomes the need to know the surface temperature T_s. By making use of the heat budget at the surface and using a first order approximation to derive $q_{sat}(T_s)$ from $q_{sat}(T_\ell)$, the latent heat flux is written:

$$LE = \frac{\gamma(R_N-F_G)+\rho C_p[q_{sat}(T_\ell)-q_\ell]/r_{a\ell}}{\gamma+\frac{C_p}{L}(1+r_s/r_{a\ell})} \qquad (16)$$

where $\gamma = (\frac{\partial q_{sat}}{\partial T})_\ell$

(c) Surface and subsurface processes

Hydrological processes

The main processes to consider are run-off, percolation and diffusion within the soil. As a general rule, run-off and percolation are combined with the local loss of water above a given threshold, this water disappearing in the process and not being transported elsewhere as it should be the case in a realistic hydrological model. Very simple formulations for total run-off (horizontal run-off plus percolation) are presently used in all models: run-off takes place either simply to remove any excess of water above a value

$W_s = W_{max}$ as in ECMWF model, or is proportional to $P_R - E$, the net incoming water at the ground, with the proportionality coefficient being a function of the field capacity W/W_{max}. This is a crude way to take into account the inhomogeneity of surface water distribution at the subgrid scale, and it allows run-off before the field capacity is reached over the whole grid square. In all cases, the formulations ensure that $W_s \leqslant W_{max}$. The main practical problem is to determine W_{max} at the grid-scale of the model.

Diffusion into the soil can be either neglected or computed as a diffusion process in the case of a multi-layer soil formulation. This requires the choice of the adequate number of layers, each of them with a maximum field capacity. The field capacities and the diffusion coefficients should in principle be a function of soil properties, but because of lack of adequate data they are generally assumed to be uniform over the globe.

Heat transfers in the soil

The transfer of heat into the soil is assumed to be due to conduction, neglecting possible convective and radiative processes, and a simple conduction equation can be used:

$$F_G = - K_T \frac{\partial T}{\partial z} \tag{17}$$

where K_T is the thermal conductivity of the soil. In models where a multi-layer description of the subsurface is adopted, this leads to a diffusion equation of the type:

$$\rho_G C_G \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} (K_T \frac{\partial T}{\partial z}) \tag{18}$$

where ρ_G is the soil density and C_G a specific heat capacity. If K_T is assumed to be constant with height:

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} \tag{19}$$

where κ is a thermal diffusivity which is the ratio of the conductivity to the volumetric heat capacity. At the local scale both C_G and K_T are dependent on soil type and soil moisture. C_G varies from about 2.0×10^6 for a typical mineral soil to 2.5×10^6 for organic matter and 4.2×10^6 for pure water (units are $J\ m^{-3}K^{-1}$). K_T can vary at least by one order of magnitude between about 0.1 to 0.5 for peat, and 0.5 to 2.5 for sand (values increase with the moisture content, units are $W\ m^{-1}K^{-1}$). In the ECMWF model a value of 2.4×10^6 has been chosen for $\rho_G C_G$ and 1.8 for K_T.

3. EXAMPLES OF FORMULATION

Usually formulations of surface processes within a large-scale model make use of the general ideas discussed above in order to get prognostic equations for three surface variables, T_s, W_s and S_n.

The formulations for T_s have evolved with the complexity of models, a good review is available in Carson (1983). In G.C.M.s without a diurnal cycle, a diagnostic equation for T_s describing the surface energy balance is a fairly good approximation. In this case:

$$F_G = R_N - F_H - LF_W - L_F M_{Sn} = \sum F \qquad (20)$$

where $\sum F$ denotes the sum of all fluxes above the surface and R_N the net radiative flux. The expressions for the various fluxes have been discussed above and make use of T_s. In several models F_G was simply set equal to zero; other possible assumptions were to set F_G proportional to R_N or to F_H ($F_G = \frac{1}{3} F_H$ in the 1971 NCAR model).

An evolution equation for T_s is clearly needed when the diurnal cycle is included. Initially bulk one layer methods were proposed for reasons of economy. For example:

$$\rho_G C_G \ \Delta z \ \frac{\partial T_s}{\partial t} = \sum F \tag{21}$$

was first used in the UCLA model in 1972. An alternative is the so-called force-restore method proposed by Bhumralkar (1978).

$$\frac{\partial T_s}{\partial T} = \frac{\sum F}{\rho_a C_G \Delta z} - \frac{2\pi}{\tau_1} (T_s - \breve{T}) \tag{22}$$

where T_s is forced toward a prescribed $\breve{T}$ temperature with a relaxation time τ_1. These formulations have been compared by Deardorff (1978) who found that the second version is satisfactory for a local computation after a good calibration of $\rho_G C_G \Delta z$ and τ_1.

Some recent models allow for a multi-layer description of the soil which enables a more realistic representation of the physical processes, once the key parameters have been determined. In the ECMWF model (Fig. 6) three layers are considered, the first one sufficiently thin to respond well to the diurnal cycle, the second one evolving with a time-scale of a few days, and the deepest layer being forced by climatic values. This is a compromise which seems adequate if one does not want a detailed description of the subsurface structure but only a realistic evaluation of the surface variables.

The evolution equation for T_s and T_d are then:

$$\frac{\partial T_s}{\partial t} = \frac{\sum F}{\rho_G C_G D_1} + \frac{(T_d - T_s)\kappa}{\frac{1}{2} D_1 (D_1 + D_2)} \tag{23}$$

$$\frac{\partial T}{\partial t} = \frac{-(T_d - T_s) D_T}{\frac{1}{2} D_2 (D_1 + D_2)} + \frac{(T_{c\ell} - T_d)\kappa}{D_2 D_3} \tag{24}$$

where T_d is the temperature of the second layer (deep temperature) and $T_{c\ell}$ the prescribed climatological underground temperature; D_1, D_2 and D_3 are the corresponding depths illustrated on Fig. 6. (D_1 = 7 cm, $D_2 = D_3$ = 42 cm).

Similar equations can be used for the soil water content:

$$\frac{\partial W_s}{\partial T} = (1-C_{sn})\frac{F_W}{\rho_W} + P_R + M_{Sn} + \frac{(W_d-W_s)\lambda}{\frac{1}{2}D_1(D_1+D_2)} \tag{25}$$

$$\frac{\partial W_d}{\partial t} = \frac{(W_d-W_s)\lambda}{\frac{1}{2}D_2(D_1+D_2)} + \frac{(W_{c\ell}-W_d)\lambda}{D_2D_3} \tag{26}$$

where W_s, W_d and $W_{c\ell}$ are specific water contents of the above three layers ($W_{c\ell}$ being specified by climatology), ρ_W the water density, C_{Sn} the fractional snow cover and λ a diffusivity coefficient for water ($\lambda = 1.07 \times 10^{-7}\ m^2\ s^{-1}$ in the ECMWF model). Conditions $W_s < W_{max}$ and $W_d < D_2/D_1\ W_{max}$ are applied in addition to (25) and (26) to represent run-off, as discussed earlier.

A simple evolution equation for snow cover, computed as equivalent water depth, is needed to close the system:

$$\frac{\partial Sn}{\partial t} = C_{Sn}\frac{F_W}{\rho_W} + P_S - M_{Sn} \tag{27}$$

where C_{Sn} is defined by

$$C_{Sn} = \min\left(\frac{Sn}{Sn_{cr}}, 1\right) \tag{28}$$

Sn_{cr} is the critical value above which snow is supposed to cover the entire grid, (Sn_{cr} = .015 m in the ECMWF model) and M_{Sn} is the rate of snow melt which is computed to adjust the surface temperature towards 0°C if T_s is found to be positive under snow cover.

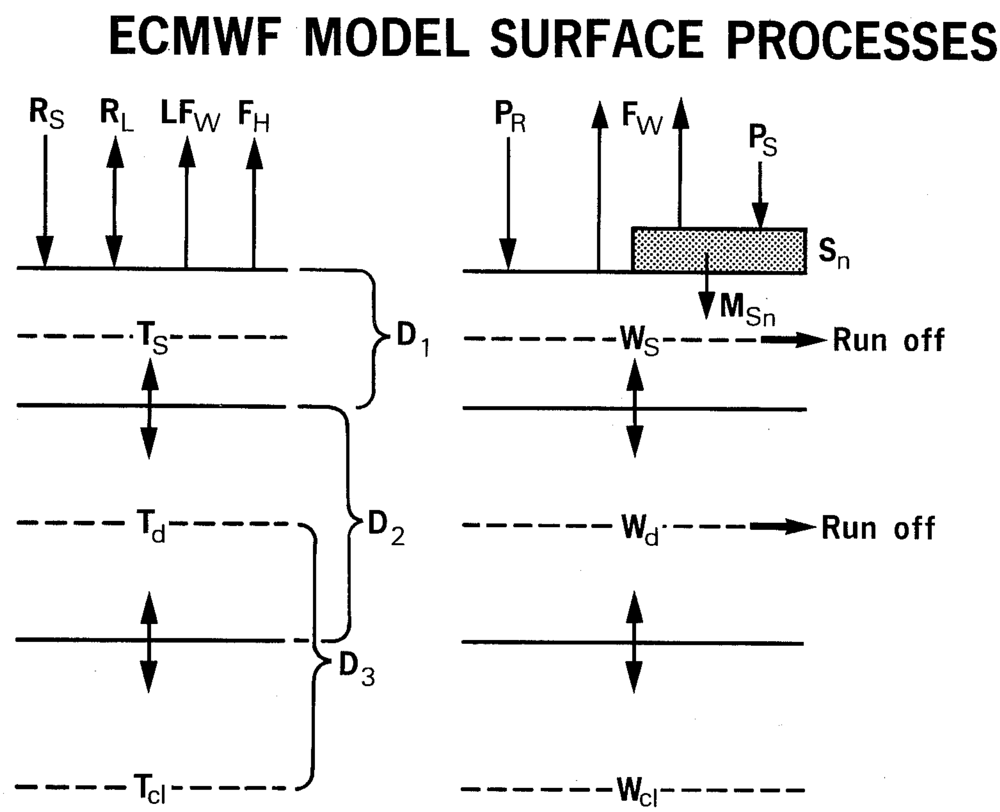

Fig. 6 Schematic representation of the ECMWF surface scheme.

4. VALIDATION OF SURFACE SCHEMES

The problem of validating a specific scheme is directly related to the acquisition of pertinent data, generally on the global scale. A complex scheme is only justified if adequate data are available to justify the hypotheses it involves and the choice of specified parameters, and to verify some of the results. As a complement to observations, detailed numerical models can also be used to validate more synthetic models or parameterization schemes.

One column simulations are the basic tool to first develop a scheme and to check the validity of the basic hypotheses. They can make use of local field experiments which provide a range of detailed information for verification. This type of work has been done for the ECMWF scheme mostly with the O'Neill data set (experiment which took place in the U.S. Great Plains in 1953), where information on the diurnal evolution of surface temperature and moisture, atmospheric temperature, wind and moisture, and estimates of the various surface fluxes are available. More recent experiments emphasize either the evolution of the atmospheric boundary layer structure, such as Wangara (Clarke et al., 1970) which has been used at ECMWF and by many groups for the validation of surface and boundary layer schemes, or provide detailed energy budget estimates at the ground with various vegetation or soil conditions. An observation campaign which was fairly complete with direct flux and boundary layer profile measurements was Voves 1977 described in Goutorbe (1980). However, up to now, no field experiment has been specifically designed to address the problem of surface parameterization at the grid scale of a G.C.M., though this is what is intended in the HAPEX-MOBILHY (Hydrology Atmospheric Pilot Experiment - Modélisation du Bilan Hydrique) experiment which will take place in the South West of France as part of the World Climate Research Programme (the motivation for this approach is outlined in WMO-WCP Vol. 76).

Once a scheme has been validated in a sufficient number of one-column experiments it has to be assessed as an interactive element of the whole model, that is in global forecasts. This can first be done locally, by comparing the diurnal evolution of measurable near surface variables against observations during the first days of the forecast. The 2 metre temperature is presently the only variable which can be used with some confidence for this exercise. This comparison is systematically done at ECMWF on a monthly basis over the European area. A systematic global validation of surface and near-surface variables would be desirable and will probably become possible with the use of adequately processed satellite data.

A third way of validating surface schemes is to study their long term behaviour as one of the elements of the model's climate. Time averaged surface fluxes are the most informative in this respect, but adequate climatic data sets for a quantitative validation are presently not satisfactory though a lot is expected in this domain from satellite data. An international effort as part of the WCRP has been initiated precisely to take advantage of available and future satellite information to provide data sets on land-surface processes. This is the ISLSCP (International Satellite Land-Surface Climatology Project) which coordinates the development of retrieval methods, initiates ground-truth field observations to validate current satellite information, and prepares an operational phase for the aquisition of adequate global data sets. An extensive discussion about what can be expected from this programme for climate models is available in WMO-WCP Vol. 74 (Development of the implementation plan for the ISLSCP) but it may be useful to summarize it briefly here.

Three types of data are needed for developing and validating surface parameterization schemes: external parameters which define the surface properties needed as input for the parameterization schemes, initialisation data for surface fields and validation data.

External parameters in the framework of the above schemes include drag coefficients or z_o, α_G, ε_G, β or R_s, W_{max} and the conductivity and diffusivity properties of the soil. α_G and to a lesser extent ε_G can be derived from satellite radiative flux measurements as a function of geographical position, and possibly including seasonal variations of α_G. They also require snow cover observations which are presently available weekly from NOAA (extent of snow cover but not snow depth). z_o and the other soil physical properties can in principle be derived geographically from global orography, vegetation and soil types information with the help of adequate validation field experiments.

Initialisation data include mostly snow depth and soil moisture which are presently not available but which may be obtainable from microwave measurements. Techniques to derive W_s from infra-red satellite data, using the diurnal T_s amplitude, are presently being tested with N.O.A.A. data at the Jet Propulsion Laboratory and at the Goddard Laboratory for Atmosphere (N.A.S.A.). The initialisation of T_s does not seem to be needed in view of the fast adjustment of this variable within models.

Validation data include basically all surface and near surface variables which are predicted by models. The most useful seem to be T_s or possibly its diurnal amplitude which can already be derived on a monthly averaged basis from NOAA data, and global precipitation for which microwave measurements are probably required. Climatologies of surface variables and fluxes are also required as a general validation tool for climate simulations.

5. IMPACT OF SURFACE SCHEMES WITHIN FORECAST MODELS

Up till now the problem of the impact of surface schemes within a model has mainly been addressed in the framework of climate-type simulations. Sensitivity experiments of various kinds have been carried out in order to determine the impact on the model's steady-state of a modification of specific parameters within surface schemes, or to assess the possible impact in the real world of artificial or natural changes in surface properties on climate characteristics.

Recent sensitivity experiments with climate models have been reviewed by Mintz (1983) and Rowntree et al. (1986). They mostly study the impact of surface albedo modifications and the response of models to changes in the soil moisture availability or in surface moisture flux formulations. A recent experiment by Süd at G.L.A. (N.A.S.A.) deals with the effect of changes in the roughness length z_o over desert areas. On the time scale of climate-type integrations, of the order of several months, the impact of a surface anomaly occurring at the scale of a continent is clearly felt globally both in the water cycle and in the global circulation. Instead of discussing this subject, extensively covered in recent papers, it seems useful in the present context to make a few remarks on the impact of surface formulations within the weather forecast range, that is typically 1-10 days.

One well known forecast experiment concerning the modification of the critical moisture field over part of Western Europe in summer is presented in Rowntree and Bolton (1983). The local impact on precipitations, low level moisture and temperature is felt at the beginning of the forecast and the effect is spread and advected by the circulation over most of Europe within a few days, although the dynamical features of the flow are not strongly altered at this time scale.

Similar results have been obtained at ECMWF for changes in the surface evaporation formulation over temperate regions and some information on this is given here to illustrate the point. This experiment has been undertaken in order to correct excessive convective precipitation which was found to occur in May and June 1985 with the high resolution model which became operational on the 1 May 1985. This model used the surface parameterization described above but with modifications to the other parameterization schemes which allowed a better connection between the surface, boundary layer and convective processes. The excess of precipitation was associated with excessive evaporation, and an underestimation of the maximum diurnal temperature and of the amplitude of the diurnal cycle.

As a temporary solution, a modification of the evaporation efficiency has been introduced, in order to take into account the fact that the resistance of vegetation is larger in temperate regions than in moist tropical areas. Following experimental results by Perrier for moderate wind conditions, evapo-transpiration has been reduced by a factor of 2 in temperate regions (defined by $T_s \leqslant 20°C$), remains unchanged for $T_s \geqslant 30°C$ and the correction factor is linearly interpolated for $20°C \leqslant T_s \leqslant 30°C$.

In the context of this discussions it is interesting to show how this modification affects the various terms of the surface budget and the diurnal cycle locally, and how this is taken into in the model moisture cycle. A local forecast for a grid point near Paris is taken as an example (Fig. 7): the latent heat flux is decreased in magnitude during the day and the sensible heat flux is increased accordingly, thus increasing the magnitude of the diurnal temperature cycle and the ground wetness. A comparison with the

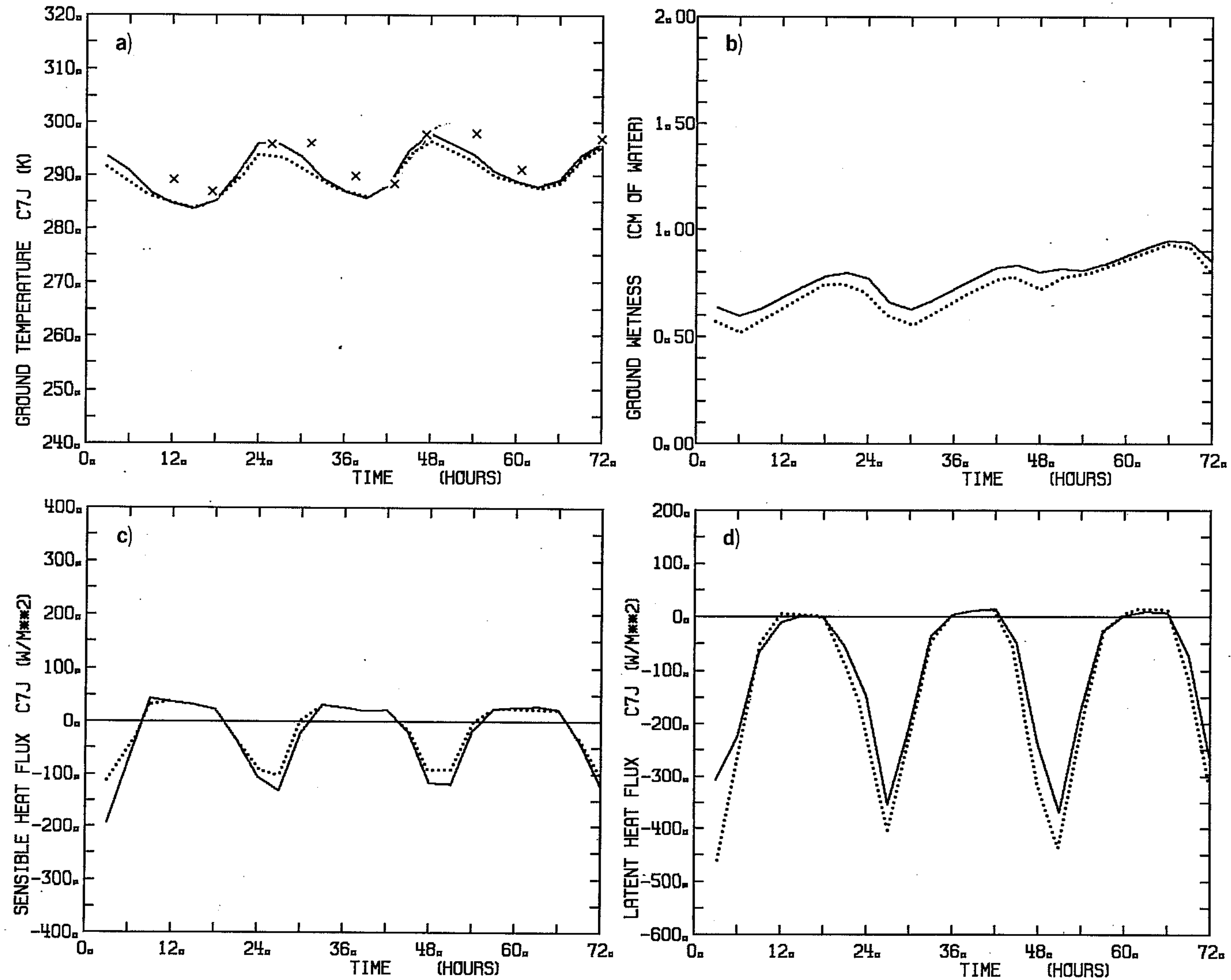

Fig. 7 Comparative local prediction with the basic scheme (dotted lines) and the modified scheme (solid lines) of ground temperature (a), ground wetness (b), sensible heat flux (c) and latent heat flux (d) at a grid point near Paris in the 30 May 1985 forecast. Crosses indicate observed 2 metre temperature.

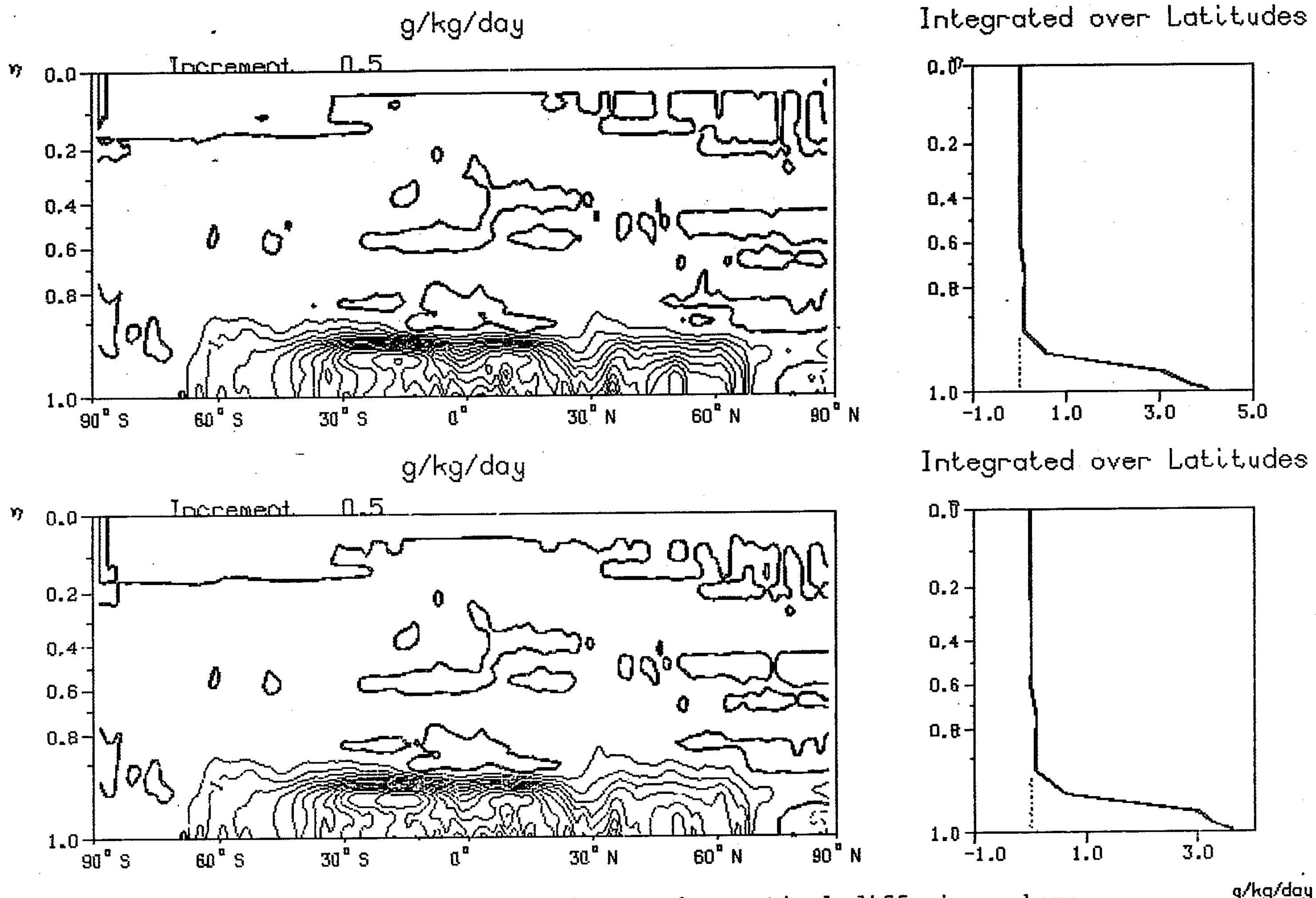

Fig. 8 Mean zonal sources of q due to the vertical diffusion scheme, averaged from day 1 to 5 in the above 10-day forecast.
(a) Basic scheme, (b) Modified scheme.

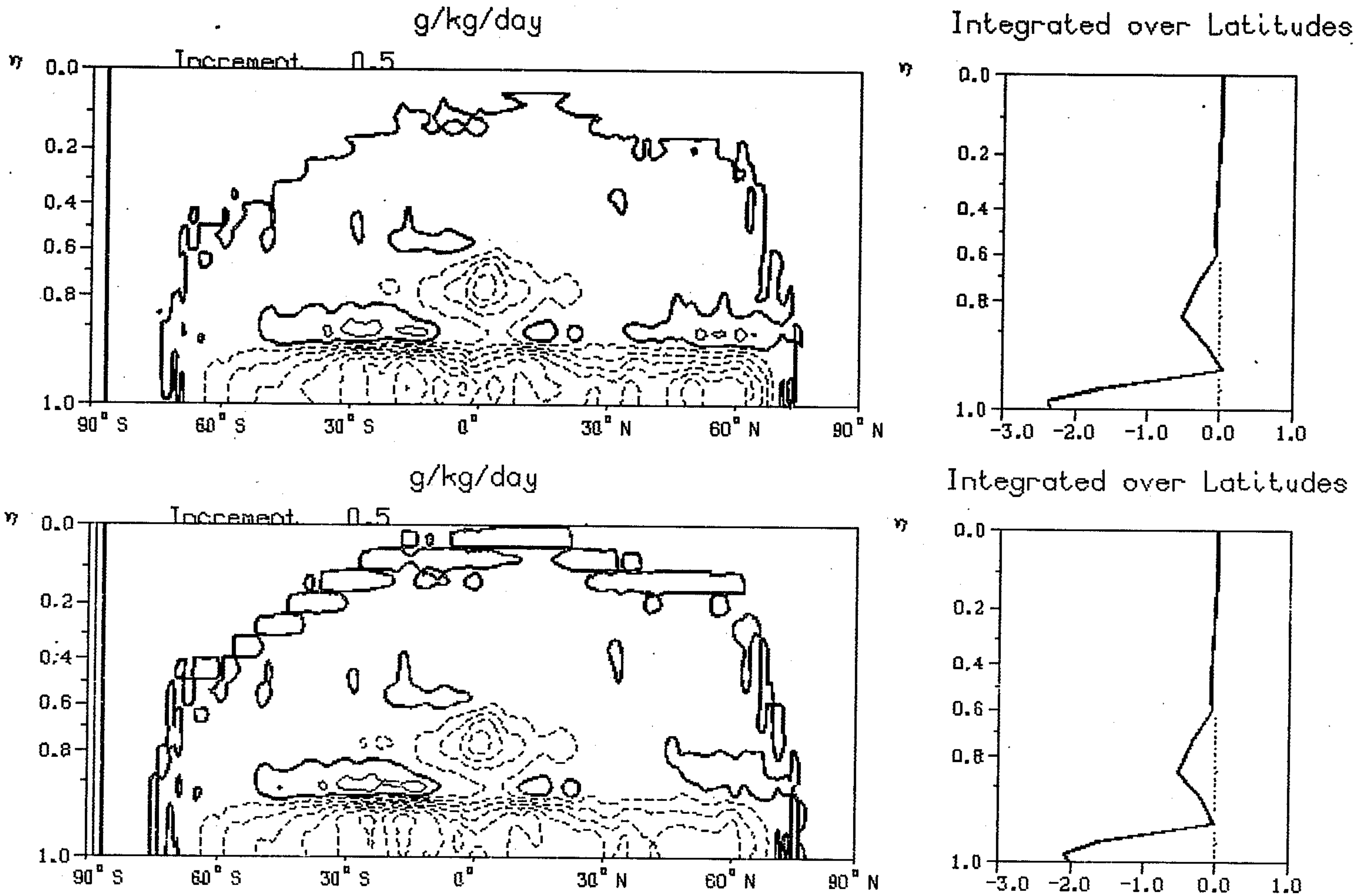

Fig. 9 As Fig. 8, but for the mean zonal sources of q due to the deep convective scheme.

observed 2m temperature at Paris (this is only an indicative comparison due to a lack of detailed surface measurements) suggests that the diurnal cycle may still be underestimated and that the model diurnal cycle displays a slight phase advance.

The above modification of the surface fluxes is directly felt in the atmospheric moisture budget and in the intensity of the various physical processes parameterized in the model. A comparison of the vertical diffusion of moisture before and after modification, zonally averaged and time-averaged over a 5 day forecast (Fig. 8), displays the decrease in the moisture source due to the decrease in surface evaporation, mostly in the 30°N - 60°N belt. This is associated with a decrease in moisture released by the convection parameterization scheme in the same latitude belt (Fig. 9), which is an indication of a decrease in convective precipitations. This is an illustration of a simple interaction between the surface moisture flux and the precipitation processes, affecting the surface and low level thermodynamic fields at a short timescale, as in Rowntree and Bolton's experiment. No significant effect was observed on the circulation itself within a 10-day forecast. Since then further work has been initiated and a vegetation scheme based on a more physical resistance formulation is being developed at ECMWF.

The above results indicate that some aspects of surface parameterization schemes have an impact on weather forecast even in the short-range. The formulation of surface moisture fluxes (or the initialisation of W_s which has similar effects) has a direct impact on surface and near surface variables, and on the water cycle, cloudiness and precipitation, in the latitude belt affected by the modification. The dynamical impact needs to be assessed: it is certainly more important in tropical regions than in temperate regions as shown by climate type experiments. The size of the area affected by a surface flux modification is also important in determining the type of model response, as suggested for example by Mintz (1983). A second surface parameter which may have a significant impact on the flow is z_o which, with the present formulation, affects all surface fluxes. The dynamical effect of an increased z_o is characterised in temperate regions by an increase in the cross-isobaric flow at low level and a filling of depressions, as mentioned for example by Anthes (1978). Experiments are also needed to assess the impact on medium range forecasts of albedo modifications, linked to the snow cover in winter temperate regions or to changes in the soil water content, or alternately to assess the impact on forecasts of errors in the specification of surface albedo.

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