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"An Introduction to the Parametrization of
Land-Surface Processes"

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# AN INTRODUCTION TO THE PARAMETRIZATION OF LAND-SURFACE PROCESSES

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ABSTRACT

This paper introduces the sub grid-scale, land-surface processes which, it is generally acknowledged, need to be included by parametrization in three-dimensional, numerical models for studying climate and climate change and for numerical weather prediction.

The discussion is restricted, in the main, to the relatively simple case of non-vegetated, land surfaces. The general boundary conditions for momentum transfer and the balance equations for energy and mass (moisture) transfer at a bare-soil surface are identified. The physical character and the parametrization of the varied flux-terms at the surface are considered systematically under the headings: Surface Radiative Properties and Fluxes; Surface Turbulent Exchanges; Soil Heat Conduction and the Land-surface Temperature; and Surface Hydrology and the Soil Water Budget.

Some of the particular problems associated with snow-covered, non-vegetated, land surfaces are described very briefly.

## 1. Introduction

The atmospheric boundary layer is the lowest layer of the atmosphere characterized by significant vertical flux divergences of momentum, heat and moisture which result directly or indirectly from interactions between the atmosphere and the underlying surface. The turbulent nature of boundary-layer flows is a vital factor in the efficient exchange of momentum, heat and moisture between the Earth's surface below and the 'free' atmosphere above. In general, up until fairly recently, designers and users of global atmospheric general circulation models (AGCMs) and operational numerical weather prediction models (NWPMs) have not been concerned with the details of boundary-layer and surface properties and processes in their own right but mainly for the influence they exert on weather systems and circulation characteristics on the much larger, synoptic or even global scales. However, the recent upsurge in the simultaneous developments of three-dimensional AGCMs for the study of climate and climate change and of increasingly sophisticated and more highly resolved operational NWPMs has resulted in more effort now being directed towards delineating details in boundary-layer structure and in the characteristics of surface climatologies. Studies with AGCMs have indicated considerable sensitivity of their simulations to changes in surface properties such as albedo, soil moisture and surface roughness. Also, some NWPMs now in operational service are expected to forecast the near-surface meteorological variables, and even changes in surface properties. The importance then of 'land-surface processes' and the need to understand and represent them better in AGCMs and NWPMs are now well established. The respective rôles of these processes in the wider climatological context have been discussed elsewhere.

Following the Joint Scientific Committee Scientific Steering Group on Land-Surface Processes of the World Climate Research Programme (WCP, 1985), I shall adopt the pragmatical definition of land-surface processes as those phenomena which control the fluxes of heat, moisture and momentum between the surface and the atmosphere over the continents. These processes influence both the circulation of the atmosphere, often remotely, and the climate of the surface.

Many important dynamical and physical processes are governed by spatial (and temporal) scales very much smaller than the typical limits of resolution of either a numerical model or an observing system. Such sub grid-scale processes cannot be dealt with explicitly in the models; however, their statistical effects at the resolved scales must be included and are determined in terms of the explicitly resolved variables. This technique is called parametrization and usually introduces empirical terms (parameters) into a model's prescription of the processes. For a fuller discussion of parametrization in numerical models see, for example, Smagorinsky (1982).

My aim here is to introduce the range of sub grid-scale land-surface processes which it is generally recognised need to be represented by parametrizations in climate and numerical weather prediction models. Discussion is restricted in the main to non-vegetated land surfaces and focusses in particular on the surface energy and mass (moisture) fluxes. A more general and fairly comprehensive review of the then current practices in AGCMs was provided by Carson (1982), with an update for Meteorological Office models only in Carson (1986a). As implied above, the parametrization of land-surface processes is a very active field of research and model development and methods labelled 'current' may quickly become susperseded. New approaches ('schemes') are being developed and tested continuously. A single paper cannot do justice to the range and complexity of tried schemes and unresolved problems even in the apparently restricted topic of land-surface processes. The special characteristics and problems of vegetated land surfaces, ice-covered surfaces and the ocean surface are being dealt with by other lecturers here at the Summer School. It should be assumed then throughout Sections 2-6 that discussions refer only to non-vegetated, snow-free, land surfaces, unless explicitly stated otherwise. Some of the particular problems associated with snow-covered, non-vegetated, land surfaces will be described briefly in Section 7.

It should also be stressed that there are many factors in a typical AGCM or NWPM which will have a direct or indirect bearing on the character and performance of the land-surface processes but which are not themselves governed directly by, nor specified explicitly in terms of, surface properties. Obvious examples amongst the other physical parametrizations include: components of the radiation scheme; the cloud scheme; the representation of rainfall and snowfall; the delineation of the atmospheric boundary layer and the parametrization of turbulent mixing within it away from the surface; deep convection; etc. A numerical model's general structure with respect to, for example: horizontal domain; spatial and temporal resolutions; distribution and number of surface types; specification of orography; etc will also determine to some extent the quality of its simulations or predictions of the surface and near-surface climatologies. Such considerations of the general problem of representing the effects of land-surface processes in AGCMs and NWPMs are beyond the scope of this introductory paper.

## 2. The Boundary Conditions for Momentum, Energy and Mass Transfer at a Bare-soil Surface

A natural and instructive way to delineate and introduce the various land-surface processes of interest is through the boundary conditions for momentum and the balance equations for the energy and mass (moisture) that apply at the surface. Most of the current generation of AGCMs and NWPMs involve such boundary conditions but with varying degrees of complexity and sophistication in their use and in the parametrizations chosen to represent individual components of the system. For the moment, let us consider in turn, the boundary constraints and relations between the momentum, energy and mass fluxes as depicted schematically in Figure 1 as a simplistic, air-soil interfacial problem.

Notation: The subscript o is used to denote surface values of variables and parameters, but only where necessary. In general, terms referred to only at the surface will not be given a subscript; soil fluxes and prognostic surface variables will be subscripted.

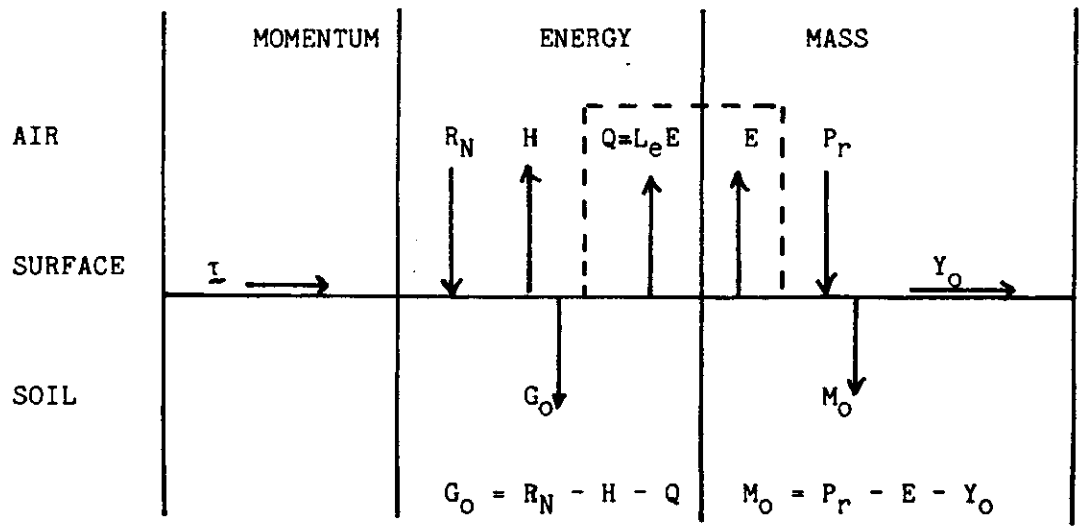


Figure 1. Schematic representation of the fluxes of momentum, energy and mass at a bare-soil surface.

### 2.1 Surface momentum flux ($\underline{\tau}$)

In an aerodynamic sense the atmospheric boundary layer is simply the lowest layer of the atmosphere under the direct influence of the underlying surface from which momentum is extracted and transferred downward to overcome surface friction. Thus the aerodynamically rough land surface provides a sink for atmospheric momentum, the removal of which at the surface is represented by the viscous drag, or horizontal shearing stress, $\underline{\tau}$, which, by convention, is a vectorial measure of the downward flux of horizontal momentum.

The surface boundary conditions for momentum transfer are:

a. NO-SLIP CONDITION: ie, the mean horizontal wind vector is zero at the surface.

b. $\underset{\sim}{\tau}$ (at the surface) is parallel to the limiting wind direction as the surface is approached.

$\underset{\sim}{\tau}$, the horizontal shearing stress, has SI units of $Nm^{-2}$.

## 2.2 The surface energy flux balance

The energy flux balance at a bare-soil surface may be expressed as

$$G_o = R_N - H - Q \qquad (1)$$

where $R_N$ is the net radiative flux at the surface (defined positive towards the surface);

$H$ is the turbulent sensible heat flux (defined positive when directed upward from the surface into the atmosphere);

$Q=L_eE$ represents the latent heat flux due to surface evaporation (defined positive when directed upward from the surface), where E is the turbulent water vapour flux (see Eqn (2)) and $L_e$ is the latent heat of evaporation; and

$G_o$ represents a flux of heat into the soil at the surface, and which, conventionally, is defined to be positive when directed into the soil.

The flux terms in Eqn (1) have SI units of $Wm^{-2}$.

## 2.3 The mass flux balance at the surface

For our purposes the mass flux balance at a bare-soil surface will be taken to be simply the moisture flux balance expressed as

$$M_o = P_r - E - Y_o \qquad (2)$$

where $P_r$ is the intensity of surface rainfall;

$E$ is the surface evaporation rate (turbulent flux of water vapour);

$Y_o$ denotes intensity of surface runoff; and

$M_o$ represents the net mass flux of water into the soil layer.

As defined, the flux terms in Eqn (2) strictly have SI units of kg $m^{-2}$ $s^{-1}$; however, it is fairly common practice to refer to the rates involved in terms of a representative depth (of water) per unit time.

Note:

1. The evaporative flux, E, appears explicitly in both Eqns (1) and (2) and thus provides a direct and important coupling between the surface heat and moisture budgets.

2. A knowledge of heat conduction and water transport in the soil is needed to parametrize the terms $G_o$ and $M_o$, respectively. In AGCMs and NWPMs this leads usually to the reformulation of Eqn (1) as a prognostic equation for the 'surface temperature', $T_o$, and of Eqn (2) as a prognostic equation for the mass of water stored in a specified depth of surface soil layer, ie the 'soil moisture content'. Further details of these soil processes and their representation in Eqns (1) and (2) are described more fully in Sections 5 and 6.

The boundary conditions and surface balance equations of Sections 2.1-2.3 involve a wide range of sub grid-scale physical and dynamical processes in both the atmosphere and the soil. It is convenient to consider the nature and parametrization of the various individual components under the following Section headings:

Section 3 Surface Radiative Properties and Fluxes (vid. $R_N$);

Section 4 Surface Turbulent Exchanges (vid. $\underset{\sim}{\tau}$, H, Q and E);

Section 5 Soil Heat Conduction and the Land-Surface Temperature (vid. $G_o$); and

Section 6 Surface Hydrology and the Soil Water Budget (vid. $P_r$, $Y_o$ and $M_o$).

## 3. Surface Radiative Properties and Fluxes

Since solar radiation provides most of the energy needed to maintain the general circulation of the atmosphere and since the major input of this energy to the Earth-atmosphere system occurs at the surface, it seems natural to start a discussion of land-surface processes by considering the surface radiative properties and fluxes. The term $R_N$ in Eqn (1) acknowledges the importance of, and the need to determine, the net imbalance of radiative fluxes to and from the land surface expressed simply here as the sum of the net short-wave radiative flux, $R_{SN}$, and the net long-wave radiative flux, $R_{LN}$, ie

$$R_N = R_{SN} + R_{LN} \qquad (3)$$

The components of $R_{SN}$ and $R_{LN}$ are shown schematically in Figure 2. Note the convention that the net radiative fluxes are positive when directed towards the surface.

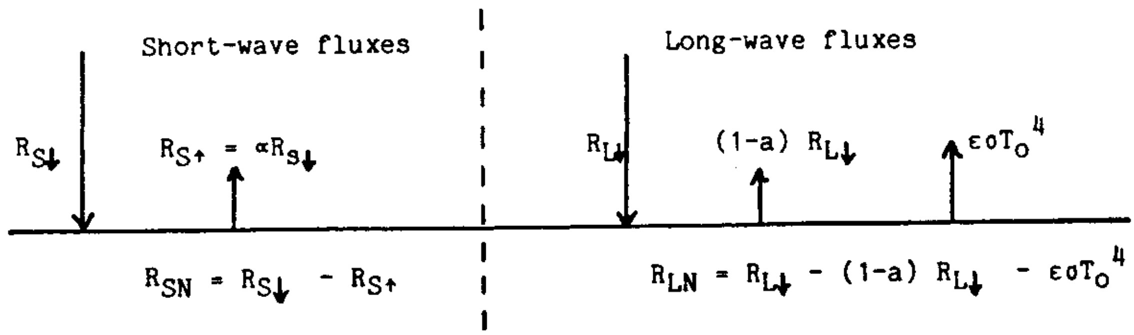


Figure 2. Schematic representation of the short- and long-wave radiative flux balances at a bare-soil surface.

### 3.1 Surface short-wave radiation balance

$$R_{SN} = (1-\alpha)\ R_{S\downarrow} \qquad (4)$$

where $R_{S\downarrow}$ is the downward short-wave radiative flux, including both the direct solar flux and diffuse radiation from the sky, and

$\alpha$ is the surface short-wave reflectivity or albedo.

### 3.2 Surface long-wave radiation balance

$$R_{LN} = a\ R_{L\downarrow} - \varepsilon\sigma T_o^4 \qquad (5)$$

where $R_{L\downarrow}$ is the downward long-wave radiative flux,

$a$ is the surface absorptivity to long-wave radiation,

$\varepsilon\sigma T_o^4$ is the long-wave radiative flux emitted at the surface,

$T_o$ is the surface temperature,

$\varepsilon$ is the long-wave emissivity of the surface, and

$\sigma$ is the Stefan-Boltzmann constant.

It is common practice to simplify Eqn (5) further by combining the definition of $\varepsilon$ with Kirchoff's law to give $a=\varepsilon$. Eqn (5) then reduces to

$$R_{LN} = \varepsilon(R_{L\downarrow} - \sigma T_o^4) \qquad (6)$$

and Eqn (3) becomes

$$R_N = (1-\alpha)\ R_{S\downarrow} + \varepsilon\ (R_{L\downarrow} - \sigma T_o^4) \qquad (7)$$

The parametrization of the radiative fluxes $R_{S\downarrow}$ and $R_{L\downarrow}$ is beyond the scope of this discussion. They are not normally classed as land-surface processes and may be regarded here as externally given forcing factors. It should be stressed though that a correct evaluation of $R_{S\downarrow}$ and $R_{L\downarrow}$ is a crucial element in establishing sensible energy and moisture balances at the surface. The prediction of $T_0$ is dealt with in Section 5. The remainder of this section concentrates on the surface radiative parameters, $\varepsilon$ and $\alpha$.

## 3.3 Surface long-wave emissivity ($\varepsilon$)

$\varepsilon$ is known to have a wavelength dependence and to vary according to the character of the surface as discussed for example by Buettner and Kern (1965), Kondratyev (1972), Paltridge and Platt (1976) and Kondratyev et al (1982). Values quoted for the vertical emissivity range from 0.997 for wet snow to 0.71 for quartz. Kondratyev et al (1982) comment that, on average, the relative emissivities of natural underlying surfaces lie within the range 0.90-0.99 and they cite several authors who have inferred that 0.95 may be assumed as the mean relative emissivity of the Earth's surface. They do caution however that the problem of measuring the emissivity of natural surfaces is far from solved and that existing techniques will need to be improved to make such measurements on a large scale.

Although there are exceptions, the most common practice in AGCMs and NWPMs is still to assume explicitly or implicitly that all surfaces act like perfect black bodies for long-wave radiation with $\varepsilon=1$. To a large extent this simply reflects the preoccupation of numerical modellers with other apparently more important and immediate problems with their physical parametrizations. I am sure that the increasing complexity and sophistication of land-surface descriptions in models will also generate more critical and discriminatory approaches to the specification of $\varepsilon$. This is most likely to be the case, for example, with the further development of models which attempt to include the explicit effects of vegetation in the climate system (see, for example, the models of Deardorff (1978) and Sellers et al (1986)).

## 3.4 Surface short-wave albedo ($\alpha$)

$\alpha$ depends on the solar zenith angle, the spectral distribution of the solar radiation incident on the surface and whether that radiation is direct or diffuse, as well as on the character of the surface as determined by the vegetation (its type, density and state), the soil type, the soil moisture and whether the surface is snow- or ice-covered. Although generally a long way removed from representing the full complexity of its functional dependence on all such quantities, nevertheless $\alpha$ in AGCMs and NWPMs is usually accorded some variation with the broad character of the surface. In AGCMs it has a specified geographical dependence (see, for example, Carson (1982)) and in many models it is still the only land-surface or soil parameter which is given such a geographical variation (see, for example, Carson (1986a)).

A good illustration of the current status of the global specification of $\alpha$ suitable for use in large-scale atmospheric models is the recent work of Wilson and Henderson-Sellers (1985) on which is based the distribution of grid-box, snow-free, land-surface albedos used in the Meteorological Office operational weather forecasting and climate models (Carson, 1986a). Wilson and Henderson-Sellers (1985) have compiled detailed, global, 1° x 1°, latitude-longitude data sets of land cover and soils, respectively. These data can be manipulated to provide the corresponding characteristics for each model grid-box.

Table 1 gives their proposed albedo values, with a seasonal variation, for each of 23 selected land types; Table 2 gives typical bare-soil albedos as a simple function of soil colour and state of surface wetness.

| Land Type Component | Annual | Summer | Winter |
|---|---|---|---|
| 1 Water | 0.07 | 0.07 | 0.07 |
| 2 Ice | 0.75 | 0.60 | 0.80 |
| 3 Inland lake | 0.06 | 0.06 | 0.06 |
| 4 Evergreen needleleaf tree | 0.14 | 0.14 | 0.15 |
| 5 Evergreen broadleaf tree | 0.14 | 0.14 | 0.14 |
| 6 Deciduous needleleaf tree | 0.13 | 0.14 | 0.12 |
| 7 Deciduous broadleaf tree | 0.13 | 0.14 | 0.12 |
| 8 Tropical broadleaf tree | 0.13 | 0.13 | 0.13 |
| 9 Drought deciduous tree | 0.13 | 0.13 | 0.12 |
| 10 Evergreen broadleaf shrub | 0.17 | 0.17 | 0.17 |
| 11 Deciduous shrub | 0.16 | 0.17 | 0.15 |
| 12 Thorn shrub | 0.16 | 0.16 | 0.16 |
| 13 Short grass and forbs | 0.19 | 0.20 | 0.18 |
| 14 Tall grass | 0.20 | 0.17 | 0.22 |
| 15 Arable | 0.20 | 0.25 | 0.16 |
| 16 Rice | 0.12 | 0.12 | 0.12 |
| 17 Sugar | 0.17 | 0.17 | 0.17 |
| 18 Maize | 0.19 | 0.22 | 0.16 |
| 19 Cotton | 0.19 | 0.22 | 0.17 |
| 20 Irrigated crop | 0.25 | 0.25 | 0.25 |
| 21 Urban | 0.18 | 0.18 | 0.18 |
| 22 Tundra | 0.15 | 0.17 | 0.12 |
| 23 Swamp | 0.12 | 0.12 | 0.12 |

Table 1. Short-wave albedos proposed by Wilson and Henderson-Sellers (1985) for 23 different types of surface cover.

| Colour class | Light | Medium | Dark |
|---|---|---|---|
| Moisture state | wet dry | wet dry | wet dry |
| Albedos | 0.18 0.35 | 0.10 0.20 | 0.07 0.15 |
| Average | (0.26) | (0.15) | (0.11) |

Table 2. Short-wave albedos proposed for bare soils by Wilson and Henderson-Sellers (1985).

Wilson and Henderson-Sellers propose that each grid-box effective $\alpha$ can be calculated from the algorithm -

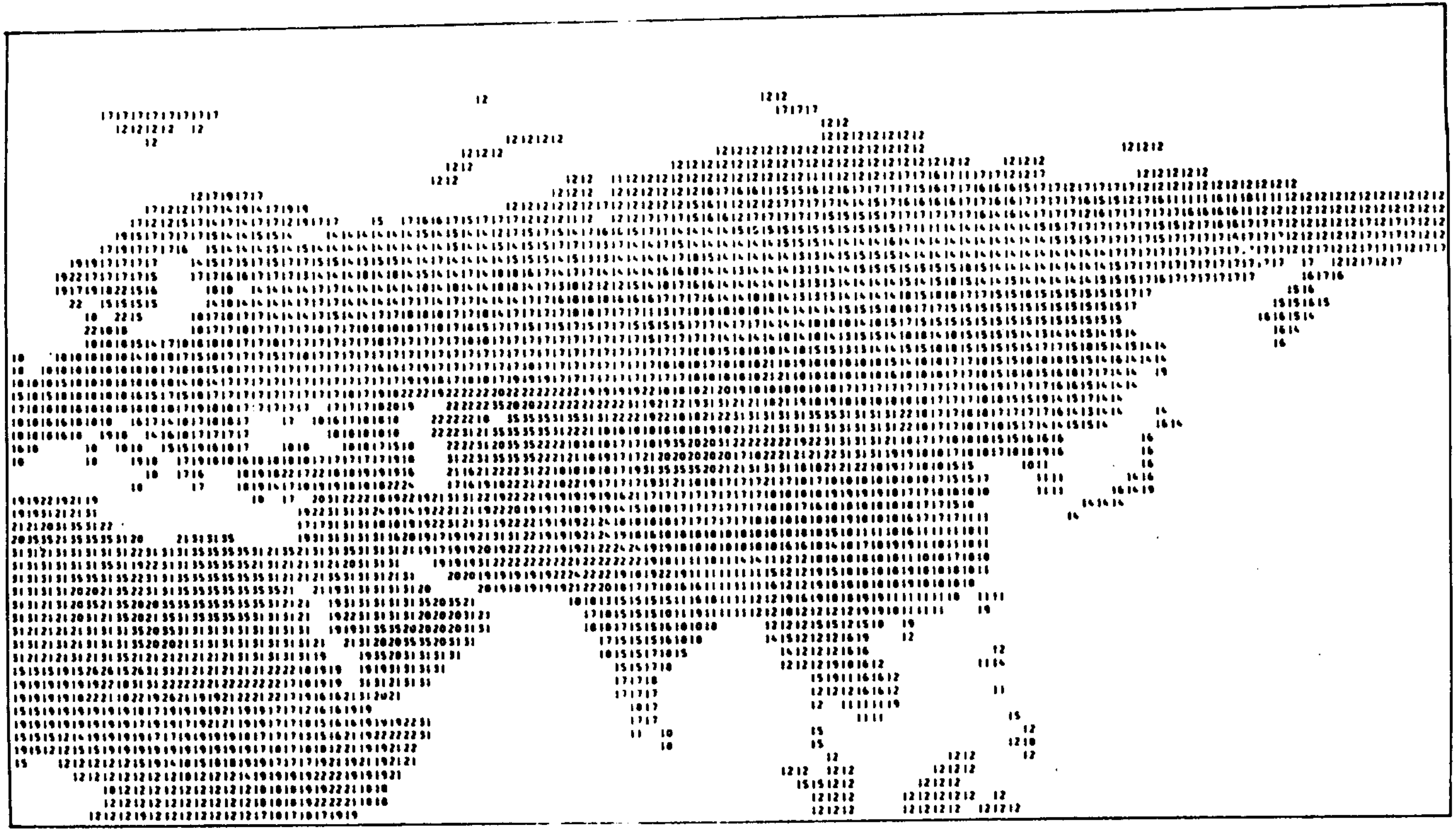

Figure 3. Snow-free land-surface albedos (%) used in the Meteorological Office's 15-level, global operational NWPM.

$$\alpha = \sum_{i=1}^{23} (f_{vi} \alpha_{vi}) + f_s \alpha_s \quad (8)$$

where $\alpha_{vi}$ are the albedos of the 23 different land-cover types in Table 1 and $f_{vi}$ are the corresponding fractions of grid-box covered; $\alpha_s$ is the albedo for the dominant soil type in the grid-box and $f_s$ is the fraction of exposed bare soil. Figure 3 illustrates a section of the particular distribution of snow-free, land-surface albedos used currently in the Meteorological Office's 15-level, global, operational weather prediction model which has a regular, 1.5° x 1.875°, latitude-longitude horizontal grid, ie the typical mid-latitude grid-length is about 150 km.

## 4. Surface Turbulent Exchanges

### 4.1 Definition of the surface turbulent fluxes

The atmospheric boundary layer (planetary boundary layer; mixing layer) is the lowest layer of the atmosphere under the direct influence of the underlying surface. The flow in the atmospheric boundary layer is turbulent except possibly in very stable conditions, for example, such as those that prevail often at night in the presence of strong surface-based temperature inversions. The velocity, temperature, humidity and other properties in a turbulent flow can be considered as random functions in space and time and it is usually necessary to resort to a statistical approach to the calculation of many boundary-layer properties. In particular this introduces the concepts of mean values, fluctuations and variances into the description of the turbulent properties of the flow. For example, if $\xi$ is some conservative quantity which fluctuates because of the turbulent motion, then it is usually written as

$$\xi = \bar{\xi} + \xi' \quad (9)$$

where $\bar{\xi}$ is some suitably defined mean value of $\xi$ and $\xi'$ is called the turbulent or eddy fluctuation (see schematic illustration in Figure 4).

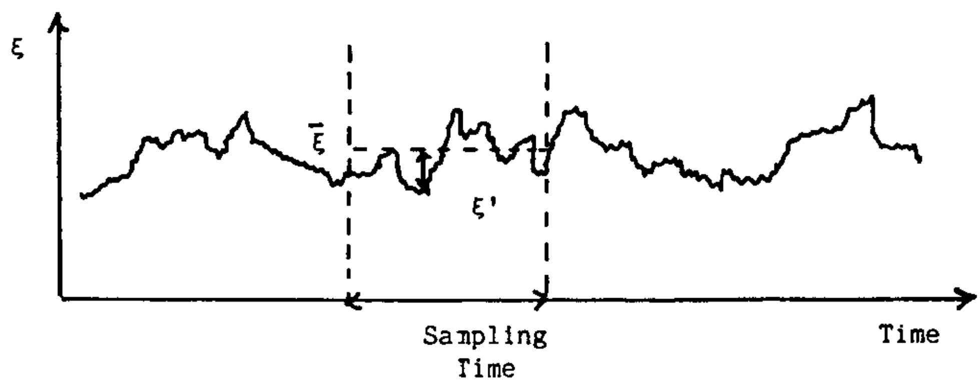


Figure 4. Schematic representation of the mean value, $\bar{\xi}$, and the eddy fluctuation, $\xi'$, determined for a particular sampling time from a time-trace of the fluctuating quantity $\xi$.

In the notation of Eqn (9), the term $\overline{w'\xi'}$, represents the eddy covariance of $\xi$ and the vertical velocity component of the flow, w, and denotes the mean vertical turbulent flux of $\xi$ at a given height in the atmospheric boundary layer. Let

$$F_\xi = (\overline{w'\xi'})_o \qquad (10)$$

denote the surface value of the mean vertical turbulent flux of $\xi$, then, in the context of our discussion of land-surface processes, the surface turbulent fluxes of particular interest are:

a. <u>Momentum flux ($\underset{\sim}{\tau}$)</u>

$$\underset{\sim}{\tau} = \rho(-(\overline{w'u'})_o, -(\overline{w'v'})_o) \qquad (11)$$

where u, v are the components of the horizontal wind vector, $\underset{\sim}{V}$, and $\rho$ is a representative mean air density near the surface (the bar notation to denote a mean value will be dispensed with except where essential to the interpretation of the terms involved). The conventional interpretation and vectorial character of the surface shearing stress, $\underset{\sim}{\tau}$, were discussed in Section 2.1. The direction of $\underset{\sim}{\tau}$ is determined by the limiting wind direction as the surface is approached. An important parameter, the <u>surface friction velocity</u> $u_*$, is defined in terms of the magnitude of $\underset{\sim}{\tau}$ such that

$$|\underset{\sim}{\tau}| = \rho\, u_*^2 \qquad (12)$$

b. <u>Sensible heat flux (H)</u>

$$H = \rho c_p(\overline{w'\theta'})_o = -\rho\, c_p\, u_*\theta_* \qquad (13)$$

where the potential temperature $\theta$ is used as the temperature which is conserved in the large-scale mixing and $c_p$ is the specific heat of air at constant pressure. $\theta_*$ (like $u_*$ in Eqn (12)) is introduced as a scaling parameter defined in terms of H and $u_*$, and is negative for a positive H (ie upward from the surface). The role of H in the surface energy balance is seen in Eqn (1).

c. <u>Water vapour flux (E)</u>

$$E = \rho(\overline{w'q'})_o = -\rho u_* q_* \qquad (14)$$

where q is the specific humidity and $q_*$ is the corresponding surface scaling parameter (defined negative for positive evaporation from the surface). E, the surface evaporation rate, is not only an important direct component of the moisture flux balance at the surface (vid. Eqn (2)) but also appears in the latent heat flux term $Q = L_eE$ in the surface energy balance (vid. Eqn (1)).

From Eqns (11)-(14) our general expression Eqn (10) for the surface turbulent flux $F_\xi$ can be extended to

$$F_\xi = (\overline{w'\xi'})_o = -u_*\xi_* \qquad (15)$$

which defines the surface scaling value $\xi_*$ (for example as for $u_*$, $\theta_*$ and $q_*$) in terms of $u_*$ and the mean vertical turbulent flux of $\xi$ at the surface.

### 4.2 The surface-flux layer

Adjacent to the surface we can identify a shallow layer in which the turning of the wind with height may be ignored and the vertical fluxes of momentum, heat and water vapour may be approximated closely by their surface values (ie for many practical purposes the turbulent fluxes in this layer may be assumed to be virtually constant with height). The layer so-defined is often referred to as the constant-flux layer. However, this terminology can mislead the unwary (note, for example, that the turbulent fluxes generally have their largest vertical gradients at the surface) and it is better to use the more appropriate term of surface-flux layer.

### 4.3 Monin-Obukhov similarity theory

The Monin-Obukhov similarity hypothesis for the surface-flux layer is the most widely accepted approach for describing the properties of the surface layer. Brought down to the very simplest terms, similarity methods depend on the possibility of being able to express the unknown variables in non-dimensional form, there being suitable argument for saying there exist a length-scale, a velocity-scale (or time-scale) and a temperature- (and humidity-) scale relevant in doing this. The non-dimensional forms are then postulated to be universal in character and this will hold as long as the scales remain the relevant ones.

The Monin-Obukhov similarity hypothesis for the fully turbulent surface-flux layer (where the Coriolis force is neglected) states that for any transferrable property, the distribution of which is homogeneous in space and stationary in time, the vertical flux-profile relation is determined uniquely by the parameters

$$\frac{g}{T},\ \frac{|\tau|}{\rho},\ \frac{H}{\rho c_p},\ \frac{E}{\rho} \qquad (16)$$

where $g/T$ is the Archimedean buoyancy parameter, $g$ is the acceleration due to gravity and $T$ is a representative air temperature in the surface layer. From Eqns (12)-(14) these are equivalent to the set

$$\frac{g}{T},\ u_*,\ \theta_*,\ q_* \qquad (17)$$

where $\theta_*$ and $q_*$ can be combined to give

$$\psi_* = \theta_* + 0.61T\, q_* \qquad (18)$$

which is very akin to a virtual potential temperature scaling value.

Instead of using the buoyancy parameter $g/T$ it is convenient to use the length-scale, $L$, defined uniquely by $g/T$, $u_*$ and $\psi_*$ by the relation.

$$L = \frac{T u_*^2}{kg\ \psi_*} = -\frac{\rho c_p T\, u_*^3}{kg\ (H + 0.61 c_p\, TE)} \qquad (19)$$

and called the Monin-Obukhov length. k is the von Kármán constant ($\simeq 0.4$) and is conventionally introduced solely as a matter of convenience. L is effectively constant in the surface-flux layer. The turbulent flow is classed as unstable when $L < 0$ (ie when the net surface buoyancy flux is positive); stable when $L > 0$ (ie when the surface buoyancy flux is negative); and neutral when $|L| \to \infty$ (ie when the surface buoyancy flux is zero).

Thus L, $u_*$, $\theta_*$, $q_*$ may be taken as the set of basic parameters which uniquely determine the relationships between the surface-layer vertical gradients of wind, potential temperature and specific humidity to the corresponding surface turbulent fluxes. Dimensional analysis leads to the vertical flux-gradient relationship expressed in the general form

$$\frac{\partial \xi}{\partial z} = \frac{\xi_*}{kz} \phi_\xi(z/L) \qquad (20)$$

where z is height above the surface. $\phi_\xi(z/L)$ is hypothesized to be a universal function of z/L only which may be of different form for each mean transferrable property, $\xi$, and which has to be established empirically from analysis of surface-layer data. The overall observational evidence is that the $\phi_\xi$ decrease with unstable stratification (ie when $L < 0$) and increase with stable stratification ($L > 0$). For specified functions for $\phi_\xi$, Eqn (20) can be integrated to provide flux-profile relationships for the surface layer, viz:

$$\frac{k[\xi(z)-\xi(z_r)]}{\xi_*} = \int_{\zeta_r}^{\zeta} \frac{\phi_\xi(\eta)d\eta}{\eta} \equiv \Phi_\xi(\zeta,\zeta_r) \qquad (21)$$

where $\zeta = z/L$ and $\zeta_r = z_r/L$, where $z_r$ is some reference height at which $\xi$ is known. In practice, Eqn (21) used in conjunction with Eqns (12)-(14) allows us to estimate the surface turbulent fluxes of momentum, heat and moisture from a knowledge of the corresponding surface-layer profiles of wind, potential temperature and humidity.

### 4.4 The Monin-Obukhov similarity functions ($\phi_\xi$)

The general character of the similarity functions is fairly well established over a limited range of stability conditions, centred on neutral, but their specification for extreme stability conditions (both stable and unstable) is much more debatable and uncertain. The particular specifications of $\phi_\xi$ listed below are subjectively selected, albeit typical, examples of the type of formulae commonly adopted as the basis of parametrizations for the surface turbulent fluxes in numerical models. For fuller discussions of the variety of postulated, empirical forms of $\phi_\xi$ see, for example, Chap 6 of McBean et al (1979).

The general behaviour is that $\phi_\xi$ increases with increasing stability; ie decreasing turbulence decreases the mixing and hence increases the normalised gradient of $\xi$. Figure 5 illustrates schematically the changing character of the surface-layer wind profile throughout a clear day and a clear night. For details see, for example, Chap 6 of Panofsky and Dutton (1984).

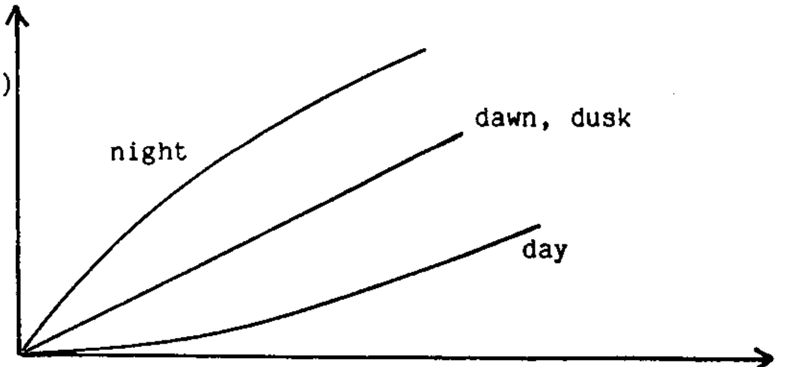


Figure 5. Schematic representation of diurnal variation of surface-layer wind profile.

4.4.1 Unstable and neutral conditions ($z/L \leq 0$)

Dyer and Hicks (1970):

$$\phi_H = \phi_E = \phi_M^2 = (1-16\ z/L)^{-1/2} \quad 0 \geq z/L \geq -1 \qquad (22)$$

where $\phi_M$, $\phi_H$ and $\phi_E$ are the respective $\phi_\xi$ for the turbulent transfers of momentum, sensible heat and water vapour. Note that, strictly, the Dyer and Hicks (1970) formulae are limited to $|z/L| \leq 1$ and so other empirical approaches may need to be invoked for more unstable conditions. For a particular choice of extrapolation beyond the Dyer and Hicks limit towards the free-convection limit see Carson (1982, 1986a).

4.4.2 Stable conditions ($z/L > 0$)

Webb (1970):

$$\phi_H = \phi_E = \phi_M = \begin{cases} 1 + 5\ z/L & 0 < z/L \leq 1 \\ 6 & 1 < z/L < 6 \end{cases} \qquad (23)$$

The problem of extending the functional form of $\phi_\xi$ to highly stable conditions was discussed by Carson and Richards (1978).

## 4.5 The bulk transfer coefficient, $C_\xi$, and the aerodynamic resistance, $r_\xi$

It is standard practice, particularly in AGCMs and NWPMs, to represent the mean vertical surface turbulent flux, $F_\xi$, by

$$F_\xi = -C_\xi V(z_\ell)\ \Delta\xi(z_\ell) \qquad (24)$$

where

$$\Delta\xi(z_\ell) = \xi(z_\ell) - \xi_0 \ . \qquad (25)$$

$z_\ell$ is some specified height above the surface and within the boundary layer (and which may be regarded, without loss of generality, as the notional height of a particular numerical model's first level above the underlying surface); $V(z_\ell)$ is the mean horizontal wind speed (ie

$|\underline{V}(z)|$) at $z_\ell$; $\xi(z_\ell)$ is the value of the property $\xi$ at $z_\ell$ and $\xi_o$ is its surface value. (Note again that the bar notation to denote mean values (see Eqn (9)) has been omitted to simplify the symbolism). $C_\xi$ is the so-called <u>bulk transfer coefficient</u>, defined in a strictly mathematical sense by Eqn (24), and which, in general, is a complicated function of height, atmospheric stability, surface roughness and, for a vegetated surface, of other physical and physiological characteristics of the surface vegetation.

In bulk-aerodynamic form the surface turbulent fluxes of Eqns (11), (13) and (14) are:

a. <u>Momentum flux</u> $\underset{\sim}{\tau} = \rho C_D V(z_\ell)\ \underset{\sim}{V}(z_\ell)$ (26)

where $C_D$ is the traditional '<u>drag coefficient</u>'.

b. <u>Sensible heat flux</u> $H = -\rho c_p C_H V(z_\ell)(\theta(z_\ell)-\theta_o)$ (27)

where $C_H$ is the bulk transfer coefficient for heat transfer.

c. <u>Water vapour flux</u> $E = -\rho C_E V(z_\ell)(q(z_\ell)-q_o)$ (28)

where $C_E$ is the bulk transfer coefficient for water vapour transfer.

To determine the fluxes from Eqns (26)-(28) the $C_\xi$ must be prescribed or expressed in terms of modelled variables and parameters and, in addition to the variables modelled explicitly at $z_\ell$, the surface temperature and humidity need to be known. The prediction of surface temperature, $T_o$ (simply related to $\theta_o$), is discussed in Section 5. The surface specific humidity, $q_o$, is not so easy to predict explicitly and its implied value is inextricably linked to the parametrization of the surface hydrology, which is discussed in Section 6. The Monin-Obukhov theory of Sections 4.3 and 4.4 provides a basis for a fairly sophisticated specification of the $C_\xi$ which is described in the next section.

A related approach to Eqn (24) for the representation of the turbulent fluxes at natural surfaces is the so-called <u>resistance approach</u>. Turbulent transfer in the atmospheric boundary layer is seen as a process analogous to the flow of electric current and, in the spirit of Ohm's Law, $F_\xi$ is written as

$$F_\xi = -\frac{\Delta\xi}{r_\xi} \qquad (29)$$

where, in a similar manner to $C_\xi$ in Eqn (22), Eqn (29) can be regarded as defining $r_\xi$, the <u>aerodynamic resistance</u> to the 'flow' of $F_\xi$.

The resistance approach has particular appeal when dealing with the complicated and multiple routes for sensible heat transfer and evaporation from vegetated surfaces (see, for example, Monteith (1965), Perrier (1982) or Rosenberg et al (1983)). It was, however, felt instructive to mention it here. Also, comparison of Eqns (24) and (29) yields

$$r_\xi = [C_\xi V(z_\ell)]^{-1} \qquad (30)$$

## 4.6 $C_\xi$ from Monin-Obukhov similarity theory

For a discussion of the large variety of specifications of $C_\xi$ then in current use in AGCMs see, for example, Carson (1982). However, discussion here is limited to the approach most acceptable to boundary-layer experts and increasingly more prevalent in the current generation of AGCMs and NWPMs, viz, that based on the Monin-Obukhov similarity theory.

From Eqns (24) and (15) it is seen that

$$C_\xi = \left(\frac{u_*}{V(z_\ell)}\right)\left(\frac{\xi_*}{\Delta\xi(z_\ell)}\right) \tag{31}$$

For Monin-Obukhov theory to be appropriate then $z_\ell$ must be fully within the surface layer so that Eqn (21) can be invoked in the particular form

$$k\,\frac{\Delta\xi(z_\ell)}{\xi_*} = \int_{\zeta_\xi}^{\zeta_\ell} \frac{\phi_\xi(\eta)}{\eta}\,d\eta \equiv \Phi_\xi(\zeta_\ell,\ \zeta_\xi) \tag{32}$$

where $\zeta_\ell = z_\ell/L$ and $\zeta_\xi = z_\xi/L$ is defined such that

$$\xi(z_\xi) \equiv \xi_o \tag{33}$$

The nature of the similarity formulation implies a logarithmic singularity in $\Phi_\xi$ as $z \to 0$. This is avoided by defining the level $z_\xi$ as the virtual height at which the $\xi$-profile, defined by Eqn (21) and extrapolated towards the surface, attains the actual surface value $\xi_o$. For momentum transfer, this level, denoted $z_o$, is defined as the virtual height at which $V=0$ on the postulated wind profile. $z_o$ is called the surface roughness length and over a bare soil surface is a characteristic of the surface and is usually independent of the flow. There are also corresponding characteristic 'surface roughness lengths' for heat and water vapour transfer. The problems of evaluating effective areal roughness lengths and of discriminating between them for the different properties are complex and it remains common practice in large-scale numerical models to use the estimate for $z_o$ for all three profiles. This aspect of the overall problem is discussed below in Section 4.7.

From Eqns (31) and (32), $C_\xi$ can be specified in terms of finite integrals of the Monin-Obukhov similarity functions, thus,

$$C_\xi = k^2\,\Phi_M^{-1}(\zeta_\ell,\zeta_o)\,\Phi_\xi^{-1}(\zeta_\ell,\zeta_\xi) \tag{34}$$

$$\text{where } \Phi_M(\zeta,\zeta_o) = \int_{\zeta_o}^{\zeta} \frac{\phi_M(\eta)}{\eta}\,d\eta = \frac{kV(z)}{u_*} \tag{35}$$

and $\zeta_o = z_o/L$. In general, with $\phi_\xi$ specified as discussed, for example, in Section 4.4, then Eqn (34) gives $C_\xi$ as a function of $\zeta_\ell$, $\zeta_o$ and $\zeta_\xi$.

It is generally more convenient for modelling purposes to express $C_\xi$ directly as a function of the explictly modelled variables $V(z_\ell)$ and $\Delta\xi(z_\ell)$. This can be achieved by using a bulk Richardson number for the surface layer, $Ri_B$, instead of $\zeta_\ell$ as the stability indicator, such that

$$Ri_B = \frac{gz_\ell}{T} \frac{[\Delta\theta(z_\ell)+0.61T\Delta q(z_\ell)]}{V^2(z_\ell)} \quad (36)$$

This can be related implicitly to $\zeta_\ell$ through

$$Ri_B = \frac{\zeta_\ell}{k} \frac{C_D{}^{3/2}}{C_H} \quad (37)$$

For a full description of the method and the assumptions made, see, for example, Carson and Richards (1978).

As an example, Figure 6 depicts the surface-layer bulk transfer coefficients used in the Meteorological Office 11-layer AGCM which are based on Monin-Obukhov similarity theory and in particular for part of the range of $Ri_B$ (corresponding to a very small section of the abscissa in the Figure), on the specifications of $\phi_\xi$ given in Eqns (22) and (23). In that particular model $z_\ell = 100$ m, $z_o$ over land is 0.1 m and $z_o$ over sea is $10^{-4}$m. The bulk transfer coefficients in Figure 6 are used in Eqns (26)-(28) to provide estimates of the surface turbulent fluxes $\underset{\sim}{\tau}$, H and E, respectively.

4.7 Surface roughness length ($z_o$)

$z_o$, like the surface albedo of Section 3, is a land-surface characteristic which has a marked geographical variation. In most of the current generation of AGCMs and NWPMs, $z_o$ has direct and indirect effects on the surface turbulent exchanges of sensible heat and moisture as well as on the surface shearing stress (see comments above in Section 4.6). However, the evaluation of an effective areal surface roughness length for heterogeneous terrain is an important practical issue that poses a variety of as yet unsatisfactorily resolved problems.

The effective areal $z_o$ for natural surfaces is rarely estimated from the wind profile and/or surface shear stress measurements. Instead, it is most likely to be determined indirectly from a knowledge of, for example: terrain relief (elevation, slope, etc); land use; type and distribution of the surface roughness elements. Algorithms, however qualitative, are needed to perform this function sensibly, at least in a fairly local (1 x 1 $km^2$) sense. The pros and cons of alternative approaches to the question of how to average over larger areas has been discussed by Carson (1986b).

Most standard boundary-layer text books provide a table of values of $z_o$ as a function of terrain type described qualitatively in terms of relief and vegetative characteristics (see, for example, Table 6.2 in Panofsky and Dutton (1984)). Such traditional relationships may well be adequate on the very local scale for the smoother, quasi-homogeneous types of terrain but can be expected to be less well founded for areal averages over rough, heterogeneous terrain, typical say of a European semi-rural landscape with small hills, woods, fields, crops, hedges, towns, lakes, etc. Wieringa (1986) has addressed this problem and produced a table giving effective areal $z_o$ in terms of a terrain classification when there are no significant orographic effects (see Table 3).

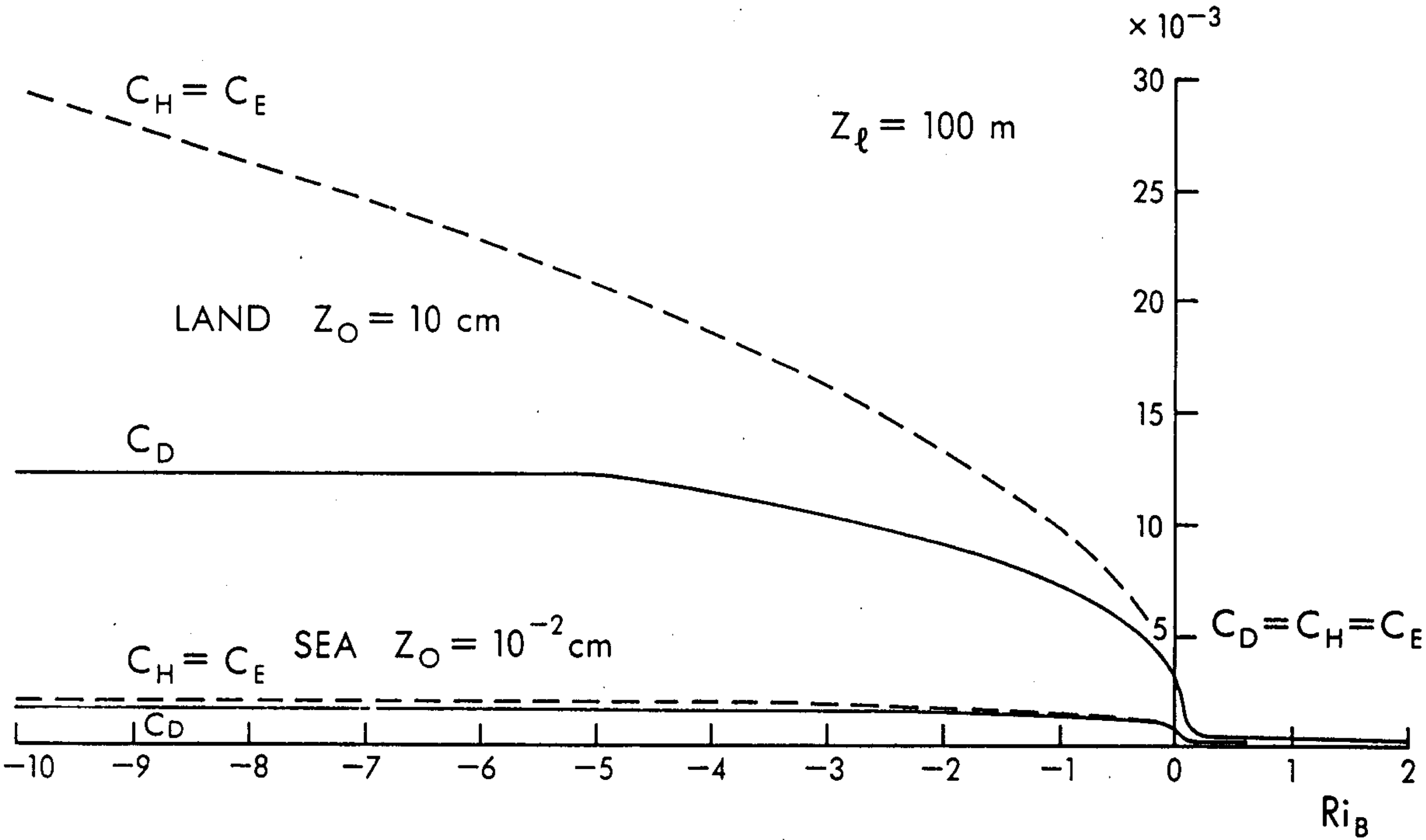


Figure 6. Surface layer bulk transfer coefficients derived from Monin-Obukhov similarity theory and used in the Meteorological Office's 11-layer AGCM.

For fuller discussions of issues concerning the evaluation of effective $z_0$ the reader is referred to the recent papers by Smith and Carson (1977), Mason (1985), Carson (1986b), Wieringa (1986) and André and Blondin (1986).

| Land use category | $z_0$ (m) |
|---|---|
| Sea (minimal fetch 5 km) | 0.0002 |
| Small lake, mud flats | 0.006 |
| Morass | 0.03 |
| Pasture | 0.07 |
| Dunes, heath | 0.10 |
| Agriculture | 0.17 |
| Road, canal (in Dutch landscape tree-lined) | 0.24 |
| Orchards, bushland | 0.35 |
| Forest | 0.75 |
| Residential built-up area ($H \leq 10$ m) | 1.12 |
| City centre (high-rise building) | 1.6 |

Table 3. Effective mesoscale surface roughness length, $z_0$ (m), expressed as a function of land use and proposed by Wieringa (1986). H is the height of the major surface obstacles.

## 5. Soil Heat Conduction and the Land-surface Temperature

In our formulation of the energy flux balance at a bare-soil surface, Eqn (1), $G_0$ the sensible heat flux in the soil is equated to the net imbalance in the energy fluxes between the surface and the atmosphere. If the aim was solely to evaluate $G_0$, then use of the surface energy balance, as depicted in Eqn (1), would be a legitimate method for obtaining such an estimate. Indeed, in principle, the energy balance method can be invoked to estimate any one of the terms in Eqn (1) if all the others are known by some other means.

The more direct, microphysical approach to understanding the soil heat flux term $G_0$ is through the study of heat transfer in the soil itself, a process which is predominantly that of heat conduction. In general, $G_0$ will depend in a complicated way on the soil's thermal properties which in turn depend on, for example, the type of surface, the type of soil and whether it is wet, dry, frozen or snow-covered, and whether it is bare soil or vegetation. In simple, general terms a thin surface layer of the soil stores heat during the day (strictly, from Eqn (1), when $R_N > H + Q$ ie $G_0$ is positive) and acts as a source of heat energy to the surface at night (strictly, when $R_N - H - Q < 0$ ie $G_0$ is negative). On longer, seasonal and annual time scales deeper soil layers act as a reservoir of heat which may be replenished during warm seasons and depleted during the cold seasons.

Good estimates of the detailed behaviour of $G_o$ throughout the day and throughout the year are now recognised as important to include in NWPMs, which attempt to forecast the characteristic diurnal cycle of land-surface temperatures, and also in climate models which need to simulate realistically and interactively the heat-storage properties of the soil over periods ranging from less than a day to at least several years.

Implicit in a knowledge of heat transfer through the soil is a knowledge of the soil temperature profile with depth. In particular, the land-surface temperature, $T_o$, features in each of the terms in Eqn (1) and it is now the common practice in AGCMs and NWPMs to invoke the surface energy balance as a diagnostic relation or prognostic equation for evaluating $T_o$. The variety of techniques commonly used in such models for representing $G_o$ in the surface energy balance has already been reviewed fairly comprehensively by, for example, Bhumralkar (1975), Deardorff (1978) and Carson (1982, 1986a). In order to illustrate the relationships between soil heat flux, soil temperature profile and the thermal properties of the soil, I shall restrict my discussion to those methods which rely on a knowledge of heat conduction in the soil and invoke either simple, one-dimensional, analytical models or attempt to model explicitly the soil heat transfer in a multi-layer soil model.

### 5.1 Heat transfer in a semi-infinite homogeneous soil

Most parametrizations of $G_o$ are now based on considerations of heat conduction and conservation in the soil. The problem is usually simplified by assuming a semi-infinite, spatially homogeneous soil layer with no horizontal heat transfer and no melting or freezing within it. This restricted and idealised one-dimensional problem is governed by:

a. the soil heat conservation equation

$$\frac{\partial T_g}{\partial t} = -\frac{1}{C}\frac{\partial G}{\partial z_g} \qquad (38)$$

where $T_g$ is the soil (ground) temperature, G is the soil heat flux, C is the volumetric heat capacity of the soil (SI units: J $m^{-3}K^{-1}$), $z_g = -z$ is the vertical co-ordinate in the soil layer and t is time; and

b. the flux-gradient relation for heat conduction

$$G = -\lambda \frac{\partial T_g}{\partial z_g} \qquad (39)$$

where $\lambda$ is the thermal conductivity of the soil (SI units: $Wm^{-1}\,K^{-1}$).

Substitution of Eqn (39) into Eqn (38), with the assumption of homogeneity, yields the one-dimensional equation for conduction of heat in the soil, viz.

$$\frac{\partial T_g}{\partial t} = \kappa \frac{\partial^2 T_g}{\partial z_g^2} \qquad (40)$$

where $\kappa$ is the thermal diffusivity of the soil (SI units: $m^2\,s^{-1}$) such that

$$\kappa = \lambda/C = \lambda/\rho_g c_g \qquad (41)$$

where $\rho_g$ is the uniform soil density and $c_g$ is the specific heat capacity (SI units: J $kg^{-1}$ $K^{-1}$).

The definitions and characteristics of the soil thermal properties C, $\lambda$, $\kappa$ and $c_g$ can be found in standard text books such as Geiger (1965), Sellers (1965), Oke (1978) and Rosenberg et al (1983). The values in Table 4 are given in Oke (1978) and illustrate the typical magnitudes of these terms for a few simple soil types (and for snow) and also indicate their sensitivity to how wet or dry the soil is.

| Material | Remarks | $\rho_g$ kg $m^{-3}$ x $10^3$ | $c_g$ J $kg^{-1}$ $K^{-1}$ x $10^3$ | C J $m^{-3}$ $K^{-1}$ x $10^6$ | $\lambda$ $Wm^{-1}$ $K^{-1}$ | $\kappa$ $m^2s^{-1}$ x $10^{-6}$ | $\delta_d$ m | $\delta_a$ m |
|---|---|---|---|---|---|---|---|---|
| Sandy soil (40% pore space) | Dry | 1.60 | 0.80 | 1.28 | 0.30 | 0.24 | 0.08 | 1.55 |
| | Saturated | 2.00 | 1.48 | 2.96 | 2.20 | 0.74 | 0.14 | 2.73 |
| Clay soil (40% pore space) | Dry | 1.60 | 0.89 | 1.42 | 0.25 | 0.18 | 0.07 | 1.34 |
| | Saturated | 2.00 | 1.55 | 3.10 | 1.58 | 0.51 | 0.12 | 2.26 |
| Peat soil (80% pore space) | Dry | 0.30 | 1.92 | 0.58 | 0.06 | 0.10 | 0.05 | 1.00 |
| | Saturated | 1.10 | 3.65 | 4.02 | 0.50 | 0.12 | 0.06 | 1.10 |
| Snow | Fresh | 0.10 | 2.09 | 0.21 | 0.08 | 0.10 | 0.05 | 1.00 |
| | Old | 0.48 | 2.09 | 0.84 | 0.42 | 0.40 | 0.10 | 2.00 |

Table 4. Thermal properties of natural materials (from Oke (1978)). $\rho_g$, $c_g$, C, $\lambda$ and $\kappa$ are defined in Section 5.1. $\delta_d$ and $\delta_a$ are the e-folding depths of the diurnal and annual soil temperature waves and are defined in Section 5.2.

A standard practice is to combine Eqns (38), (39) and the surface energy balance, Eqn (1), to produce a prognostic equation for $T_o$ (usually assumed equivalent to the soil surface temperature $T_{go}$). The simplest approaches of this kind introduce the concept of an effective depth of soil D and an effective surface thermal capacity

$$C_{eff} = CD = \rho_g c_g D \qquad (42)$$

defined such that

$$G_o = C_{eff} \frac{\partial T_{go}}{\partial t} \equiv CD \frac{\partial T_o}{\partial t} \qquad (43)$$

Many AGCMs and NWPMs contain rather arbitrary and empirical selections of $C_{eff}$ (see, for example, Carson (1982)) and with $G_o$ replaced by the RHS of the surface energy balance, Eqn (43) can be solved for $T_o$.

Note however that $C_{eff}$ (and D) can be defined more formally from consideration of the soil heat conservation equation (38). On the assumption that $G \to 0$ as $z_g \to \infty$ then Eqn (38) can be integrated to give

$$G_o = C \int_o^\infty \frac{\partial T_g}{\partial t} \, dz_g \qquad (44)$$

which, when used to replace $G_o$ in Eqn (43), allows D to be defined in a strictly mathematical sense as

$$D = \left(\frac{\partial T_o}{\partial t}\right)^{-1} \int_o^\infty \frac{\partial T_g}{\partial t} \, dz_g \qquad (45)$$

The following section describes a popular analytical approach in which Eqn (45) may be invoked to good advantage.

### 5.2 One-dimensional heat transfer in a semi-infinite, homogeneous soil whose surface is heated in a simple periodic manner

One simple, attractive and commonly adopted method of determing D in Eqn (43) is by appealing to the theory of heat transfer in a semi-infinite homogeneous medium when the surface is heated in a simple periodic manner (as discussed, for example, in Sellers (1965)).

If it is assumed that the surface temperature

$$T_o \equiv T_g(0, t) = \hat{T}_g + a_o \sin \omega t \qquad (46)$$

where $\omega$ is the angular frequency of oscillation, $\hat{T}_g$ is the mean soil temperture (over the period $P = 2\pi/\omega$), assumed to be the same at all depths, and $a_o$ is the amplitude of the surface temperature wave, then the solution of Eqn (40) is

$$\begin{aligned} T_g(z_g,t) &= \hat{T}_g + a(z_g) \sin (\omega t - z_g/\delta) \\ &= \hat{T}_g + a_o \exp(-z_g/\delta) \sin(\omega t - z_g/\delta). \qquad (47) \end{aligned}$$

$$\delta = \left(\frac{\kappa P}{\pi}\right)^{1/2} = \left(\frac{2\lambda}{C\omega}\right)^{1/2} \qquad (48)$$

is the e-folding depth of the temperature wave of period P, ie it is the depth where the amplitude of the oscillation is reduced to $^1/_e$ (ie 0.37) times its surface value. Values of the e-folding depths corresponding to the diurnal and annual periods, respectively, are given for a range of soil types in Table 4.

The effective depth D corresponding to the soil temperature profile Eqn (47) is, from Eqn (45),

$$D = \frac{1}{a_o \cos \omega t} \int_o^\infty a(z_g) \cos(\omega t - z_g/\delta) \, dz_g$$

$$= \frac{1}{\cos \omega t} \int_o^\infty \exp(-z_g/\delta) \cos(\omega t - z_g/\delta) \, dz_g$$

$$= \frac{\delta}{\sqrt{2}} \quad \frac{\sin(\omega t + \pi/4)}{\cos \omega t} \tag{49}$$

Therefore D as defined in Eqn (43) is not only a function of the thermal diffusivity of the soil and the single frequency assumed for the simple periodic forcing at the surface but also varies with time according to Eqn (49). Substituting for D from Eqn (49) in Eqn (43) gives a prognostic equation for $T_o$, viz.

$$\frac{\partial T_o}{\partial t} = \frac{\sqrt{2}\, G_o \cos \omega t}{C\delta \sin(\omega t + \pi/4)} \tag{50}$$

which in turn can be expanded easily to give

$$\frac{\partial T_o}{\partial t} = \frac{2G_o}{C\delta} - \frac{2\pi}{P} (T_o - \hat{T}_g) \tag{51}$$

This, I believe, is a relatively neat way of deriving Eqn (51) which was proposed by Bhumralkar (1975) and has come to be referred to as the 'force-restore method', a term introduced by Deardorff (1978).

The period of the diurnal temperature oscillation is normally used in Eqn (51) as that appropriate for determining the thermal capacity of the effective surface layer. Additional information about $\hat{T}_g$ is required to solve Eqn (51). $\hat{T}_g$ may be fixed or diagnosed over short periods of a few days but would need to be determined prognostically over the much longer periods of integration involved, for example, in climate modelling. Deardorff (1978) has suggested a second prognostic equation for $\hat{T}_g$ analogous to Eqn (51) but with the appropriate effective depth determined by the e-folding depth of the annual temperature wave. Although there is some useful mileage in extending this simple, analytically-based method further (see, for example, Deardorff (1978) and Carson (1982)), such parametrizations soon become analogous to the more elaborate schemes which explicitly model the temperature profile through several soil layers.

## 5.3 Multi-layer soil models

The somewhat idealised analytical assumptions underlying the force-restore method and other simpler parametrizations can be avoided in principle by explicit modelling of the soil temperature profile and soil heat conduction with a multi-layer soil model of specified depth and with appropriate vertical resolution and boundary conditions. One approach, for example, would be to invoke Eqn (39) to evaluate $G_o$ explicitly from the modelled soil temperature profile such that

$$G_o = \left[-\lambda \frac{\partial T_g}{\partial z_g}\right]_{z_g=0} \tag{52}$$

With this representation of $G_o$, Eqn (1) could then be solved diagnostically for $T_o$.

$G_o = R_N - H - Q$

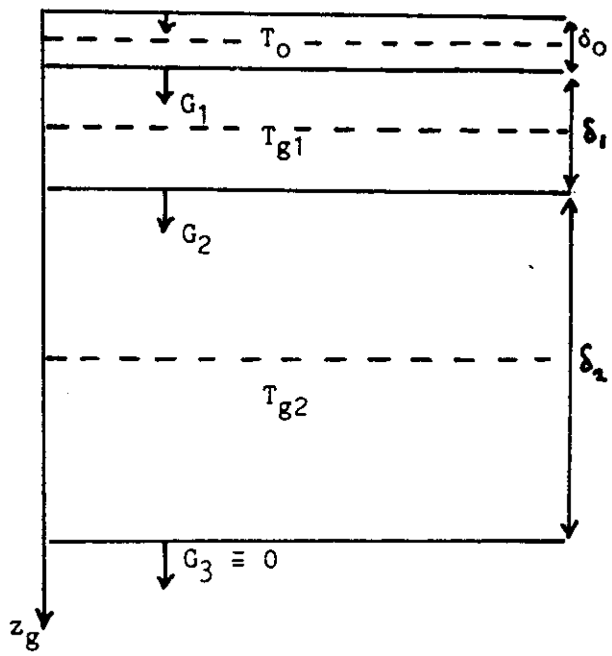


Figure 7. Schematic representation of a 3-layer, soil-temperature, finite-difference model. $T_o$, $T_{g1}$ and $T_{g2}$ are the representative temperatures in the soil layers of depth $\delta_o$, $\delta_1$ and $\delta_2$, respectively. $G_o$, $G_1$, $G_2$ and $G_3$ are the corresponding soil heat fluxes at the respective layer boundaries.

An alternative approach is represented schematically for a 3-layer soil-temperature model in Figure 7. Here the surface temperature $T_o$ is represented by the mean temperature of a very thin surface soil layer of depth $\delta_o$. The rate of change of $T_o$ with time is given by the soil heat flux divergence in the surface layer according to a simple finite difference form of Eqn (38), i.e.

$$\frac{\partial T_o}{\partial t} = \frac{G_o - G_1}{C\delta_o} \quad . \tag{53}$$

$G_o$ is as usual the net imbalance of the terms on the RHS of Eqn (1) and $G_1$, the soil heat flux into the next layer down, is determined from the explicitly modelled soil temperature profile from the heat conduction Eqn (39) written simply as

$$G_1 = \frac{2\lambda(T_o - T_{g1})}{\delta_o + \delta_1} \quad . \tag{54}$$

Therefore, from Eqns (53) and (54),

$$\frac{\partial T_o}{\partial t} = \frac{G_o}{C\delta_o} + \frac{2\,\kappa\,(T_{g1} - T_o)}{\delta_o(\delta_o + \delta_1)} \quad , \tag{55}$$

and the same general technique is used to provide the corresponding predictive equations for the temperatures of the other soil layers.

In the 3-layer soil model of Figure 7, all three soil temperatures are treated as prognostic variables during integration of the model, with the boundary condition that the soil heat flux is zero at the lower soil boundary (ie $G_3 \equiv 0$ in Figure 7). An alternative, popular lower boundary condition is to hold the bottom-layer soil temperature constant at its initialised value. This latter boundary condition is used, for example, in the 4-layer soil model in the current Meteorological Office fine-mesh operational forecasting model (Carson, 1986a) and also in the 3-layer soil model used at ECMWF (Blondin, 1986).

The selection of 'representative' soil thermal characteristics C and $\lambda$ (and hence $\kappa$) and suitable soil-layer depths, $\delta_0 \ldots \delta_i \ldots \delta_{n-1}$ where n is the number of explicitly resolved layers in the soil; remains a difficult, empirical and highly subjective business. On the basis of a comprehensive study of the amplitude and phase responses of multi-layer soil schemes to periodic surface temperature forcing, Warrilow et al (1986) have recommended a 4-layer soil-temperature scheme of the type depicted in Figure 7 for use in the Meteorological Office AGCM. Their paper gives full description of how the appropriate soil-model parameters were selected. For further discussion of values used in specific models see, for example, Blondin (1986) and Carson (1982, 1986a). Table 5 gives values of the main parameters likely to be incorporated into the most recent control version, the so-called 'Fourth Annual-Cycle Version', of the Meteorological Office 11-layer AGCM used for climate modelling research (Warrilow, private communication).

| SOIL THERMAL PROPERTIES | Value |
|---|---|
| Volumetric heat capacity, C ($J\ m^{-3}\ K^{-1}$) | $2.34 \times 10^6$ |
| Thermal conductivity, $\lambda$ ($W\ m^{-1} K^{-1}$) | 0.56 |
| Thermal diffusivity $\kappa = \lambda/C$ ($m^2 s^{-1}$) | $2.39 \times 10^{-7}$ |
| Thermal inertia $\gamma = (\lambda C)^{1/2}$ ($J\ m^{-2} K^{-1} s^{-1/2}$) | 1145 |
| Soil-layer depths:<br>$\delta_o$ (m)<br>$\delta_1$ [$r_1 = \delta_1/\delta_o$]<br>$\delta_2$ [$r_2 = \delta_2/\delta_o$]<br>$\delta_3$ [$r_3 = \delta_3/\delta_o$]<br>$\sum_{i=o}^{3} \delta_i$ | <br>0.037<br>0.143 [3.91]<br>0.516 [14.05]<br>1.639 [44.65]<br>2.335 |
| Depth of the surface layer of soil determined by<br>$\delta_o = (\frac{2\lambda}{C\omega_o})^{1/2}$ where<br>$\omega_o$ ($s^{-1}$)<br>$P_o = 2\pi/\omega_o$ (day) | <br><br>$3.5509 \times 10^{-4}$<br>0.2048 (ie 4.8 hr) |
| Thermal capacity of surface layer of soil: $C\delta_o$ ($J\ m^{-2} K^{-1}$) | $8.59 \times 10^4$ |

Table 5. Physical properties selected by Warrilow (private communication) for use in the 4-layer soil-temperature model to be used in the Fourth Annual-Cycle version of the Meteorological Office 11-layer AGCM. The general approach is described fully in Warrilow et al (1986).

5.4 The land-surface temperature, $T_o$

Throughout this paper, following the general practice in AGCMs and NWPMs, it has been assumed that the land-surface temperature is a well-defined and unique property of any natural land surface and that the same '$T_o$' is appropriate as: the radiative surface temperature of Eqn (7); the surface temperature as used in the extrapolated atmospheric boundary layer profiles and surface-flux formulae of Section 4; and the surface soil temperature related to the soil heat flux as introduced above in Section 5. The 'surface temperatures' implied by these different physical processes at the surface must be closely related but they are not necessarily all the same. The ambiguity and difficulty in defining surface temperature become even greater when the surface has a vegetative canopy. Suffice it to state here that at present the problem is very poorly understood and that more observational and theoretical studies are needed before any significant differences between the '$T_0$' can be clearly delineated and incorporated sensibly in AGCMs and NWPMs.

## 6. Surface Hydrology and the Soil Water Budget

Most of the current generation of AGCMs and NWPMs now include some form of 'interactive' surface hydrology, usually of a very rudimentary nature. Such parametrizations are termed 'interactive' in the sense that the soil has some recognised hydrological property that is allowed to vary in response to the model's continuously evolving atmospheric state and surface boundary conditions and which in turn exerts both direct and indirect influences on the surface fluxes themselves. The most common practice is to define a variable 'soil moisture content' for some notional depth of surface soil layer which is constrained at all times to satisfy the surface moisture flux balance as expressed in Eqn (2).

In direct analogy to the need to study heat conduction in the soil to provide a sound physical basis for evaluating the 'surface temperature', so also is there a corresponding need to understand more about the dynamics which govern the movement of water in the soil in order to model changes in the profile of 'soil moisture content'. Since the concepts of 'surface temperature' and 'soil moisture content' have been introduced here independently and in different sections, it is perhaps worth emphasizing again the strong, interactive coupling between the thermal and hydrological properties and processes in the soil. Not only does E (or Q) appear explicitly in both Eqns (1) and (2) but most of the other surface fluxes (including the momentum flux $\underline{\tau}$) depend to varying degrees on both the 'surface temperature' and the 'soil moisture content'. Indeed, in a model with both interactive surface hydrology and interactive land-surface temperature, the value of the 'soil moisture content' has an important bearing on the evaluation of $T_0$, and vice-versa.

As in the case of the surface radiative fluxes $R_{S\downarrow}$ and $R_{L\downarrow}$ in the context of the surface energy balance (discussed in Section 3), the surface rainfall rate $P_r$ is regarded here as an externally determined component of the surface moisture balance, Eqn (2). Accurate evaluation of $P_r$ is of course of crucial importance in establishing a realistic surface moisture balance and also, through the coupling discussed above, a realistic surface energy balance. The other processes involved in the hydrology of a bare soil, including evaporation, surface runoff, and transport and storage of water in the soil are generally very complex and not so well understood nor as simple to parametrize sensibly as the individual terms in the surface energy balance. The very small-scale spatial inhomogeneities within a typical soil layer appear to be more important in the determination of soil moisture movement than for the heat flow and this presents formidable difficulties when trying to formulate a parametrization based soundly on underlying physical and dynamical hydrological principles. This is particularly so when one-dimensional hydrological models are applied to catchment-sized or typical AGCM/NWPM grid-box areas. Hence the importance of the HAPEX-MOBILHY project (André et al, 1986) aimed at studying the hydrological budget and evaporation flux at the scale of an AGCM grid square, ie $10^4$ $km^2$. A two-and-a-half-month special observing period should provide detailed measurements of the relevant atmospheric fluxes and intensive remote sensing of surface properties. The main objective of the programme is to provide a data base against which parametrizations of the land-surface water budget can be developed and tested.

A proper discussion of the surface and sub-surface hydrology of natural soils is beyond the scope of this paper. For this the reader is referred to the recent fuller expositions by, for example, Brutsaert (1982a, b), Dooge (1982), Eagleson (1982) and Dickinson (1984) in which the problems of areal representation of hydrological processes are specifically discussed. The remainder of this section is restricted to an introduction to the most simple form of the basic equations which govern the movement of water in the soil and brief descriptions of some specific formulations for soil-water transport, evaporation and surface runoff. These examples although chosen quite subjectively should nevertheless give an indication of the general tenor and level of many of the current attempts to parametrize grid-scale hydrological processes.

### 6.1 Water transport in a homogeneous soil

There are various inter-related measures of soil moisture content, two of which are:

a. $\chi$, the soil moisture concentration, defined as the mass of water per unit volume of soil. (SI units: kg $m^{-3}$), and

b. $\chi_v$, the volumetric soil moisture concentration, defined as the volume of water per unit volume of soil and therefore dimensionless.

Therefore

$$\chi = \rho_\omega \chi_v \qquad (56)$$

where $\rho_\omega$ is the density of water. These are very appropriate measures in parametrizations based on simulating changes in the water mass of a specified layer of soil.

In general, several different forces are acting to bind the water to the soil and a less direct but nevertheless very useful measure of soil moisture content in the context of water movement is the soil moisture potential $\Psi$ (also termed soil moisture tension, soil moisture suction, etc) which may be thought of as the energy needed to extract water from the soil matrix. It is common practice to express $\Psi$ as a length, in a fashion analogous to the concept of a pressure head in hydraulics, such that at a level $z_g$ in the soil

$$\Psi = \psi - z_g \qquad (57)$$

where $z_g$ represents the gravitational component of the moisture potential and $\psi$, the so-called matric potential, is the contribution to $\Psi$ due mainly to capillarity and adsorption.

In an analogous fashion to the treatment of soil heat conduction in Section 5.1, consider the grossly simplified hydrology of a spatially homogeneous soil layer with no horizontal water movement and no melting or freezing within it. This restricted and idealised one-dimensional problem is governed by:

a. the equation of continuity

$$\frac{\partial \chi}{\partial t} = \rho_\omega \frac{\partial \chi_v}{\partial t} = - \frac{\partial M}{\partial z_g} \tag{58}$$

where M is the vertical mass flux of water and $\chi$, $\chi_v$ and M are functions of $z_g$ and t; and

b. the flux-gradient relation (Darcy's Law)

$$M = -\rho_\omega K(\psi)\frac{\partial \Psi}{\partial z_g} \tag{59}$$

$$= -\rho_\omega K(\psi) \left(\frac{\partial \psi}{\partial z_g} - 1\right)$$

where K, the hydraulic conductivity of the soil (SI units: $ms^{-1}$), is a function of $\psi$, which in turn is a function of $z_g$ and t. Combining Eqns (58) and (59) yields the Richards' equation for the vertical movement of water in an unsaturated soil, viz.

$$\frac{\partial \chi_v}{\partial t} = \frac{\partial}{\partial z_g} \left[K(\psi) \left(\frac{\partial \psi}{\partial z_g} - 1\right)\right] \tag{60}$$

Solving even the idealised Eqn (60) for reasonable boundary conditions is by no means a trivial matter. Proposals do exist which express $\psi$ and $K(\psi)$ as functions of $\chi_v$, although it should be stressed that these are highly empirical and difficult to justify in all but the most idealised circumstances. In such cases Eqn (60) takes the form of a diffusion equation for soil water

$$\frac{\partial \chi_v}{\partial t} = \frac{\partial}{\partial z_g} \left(\kappa_\omega(\chi_v)\frac{\partial \chi_v}{\partial z_g}\right) - \frac{\partial K}{\partial z_g}(\chi_v) \tag{61}$$

where $\kappa_\omega$ is a moisture diffusivity of the soil (SI units: $m^2s^{-1}$) defined by

$$\kappa_\omega = K(\chi_v)\frac{\partial \psi}{\partial \chi_v} \tag{62}$$

Prognostic equations for soil moisture content based on the idealised hydrology of this section are beginning to appear in AGCMs and NWPMs; see, for example, the particular examples discussed in Dickinson (1984) and Warrilow et al (1986). One particular multi-layer soil hydrology scheme which has attracted considerable support from numerical modellers is the force-restore treatment of Deardorff (1978) in which he postulates equations for soil moisture transport of a form directly analogous to the corresponding force-restore equations for soil temperatures (vid. Eqn (51)). An effective 3-layer version of Deardorff's approach is used, for example, in ECMWF models (Blondin, 1986). However, the most common current approach to modelling soil moisture content is probably still that based on a single surface soil layer and a more detailed discussion of only that example will suffice here.

## 6.2 Single-layer soil hydrology models

A common, rudimentary approach to the parametrization of the hydrological processes at a bare-soil surface is to monitor the change of soil moisture content in a single, shallow surface layer of soil of notional depth $\delta_\omega$, as depicted schematically in Figure 8.

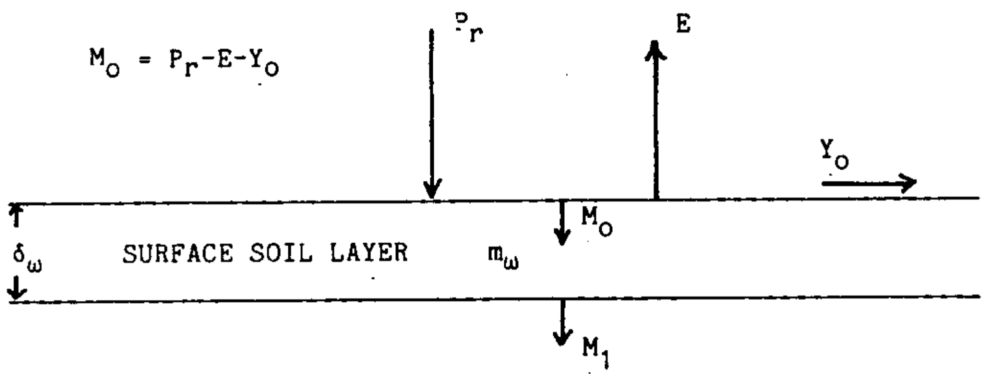


Figure 8. Schematic representation of the moisture balance of a surface layer of soil.

Let $m_\omega$ denote the mass of liquid water per unit lateral area in the soil layer of depth $\delta_\omega$, ie

$$m_\omega = \int_0^{\delta_\omega} \chi \, dz_g = \hat{\chi} \, \delta_\omega = \rho_\omega \hat{\chi}_v \delta_\omega = \rho_\omega d_\omega \qquad (63)$$

where Eqn (63) also defines a layer-mean soil moisture concentration, $\hat{\chi}$, a corresponding layer-mean volumetric soil moisture concentration, $\hat{\chi}_v$, and $d_\omega$, a representative depth of water in the layer. Integration of Eqn (58) over the layer depth gives, from Eqn (63), the surface layer water mass balance equation in the form

$$\begin{aligned} \frac{\partial m_\omega}{\partial t} &= M_O - M_1 \\ &= P_r - E - Y_O - M_1 \end{aligned} \qquad (64)$$

when $M_O$ is substituted from Eqn (2).

$M_1$ is the vertical mass flux of water at the base of the surface layer. Apart from the surface runoff term $Y_O$, all other horizontal fluxes of soil water have been neglected. With $P_r$ regarded in the present context as determined externally, then it remains here to illustrate with the aid of specific examples some of the problems of formulating parametrizations for E, $Y_O$ and $M_1$.

### 6.3 Evaporation at a bare-soil surface (E)

In principle, the surface evaporation rate E can be obtained as the residual flux from either the surface energy balance, Eqn (1), or the surface moisture balance, Eqn (2), and there are many empirical formulae for estimating E based on such approaches. A very useful introduction to the large variety of methods available can be found, for example, in Rosenberg et al (1983) and for more detailed discussions see, for example, Eagleson (1982) and Brutsaert (1982a, b). However, in this introduction to interactive soil temperature and soil moisture content parametrizations in AGCMs and NWPMs, I have selected the soil-flux terms $G_o$ and $M_o$ as the residual components in the surface balance equations (1) and (2), (see, for example, Eqns (51), (55) and (64)) and assumed implicitly that E can be evaluated in some independent manner.

Indeed, the method of estimating E has already been implied in principle in Section 4 where E as one of the main surface turbulent fluxes was ultimately parametrized in the bulk aerodynamic form of Eqn (28) as

$$E = -\rho C_E V(z_\ell)(q(z_\ell)-q_o) \tag{65}$$

with the recommendation that the bulk transfer coefficient $C_E$ be evaluated from Monin-Obukhov similarity theory. It was however also noted that the surface value of the specific humidity $q_o$, required explicitly in Eqn (65) (and also, for example, in determining the bulk Richardson number defined by Eqn (36), and hence $C_E$) is not easy to determine. To overcome this problem it is standard practice to imply a value of $q_o$ through relations with $q_{sat}$ $(T_o)$, the saturation specific humidity at the surface which is readily determined as a function of surface temperature (and pressure) via the Clausius-Clapeyron relationship

$$\frac{dq_{sat}}{dT} \simeq 0.622 \frac{L_i\, q_{sat}}{R\, T^2} \tag{66}$$

where T is temperature, R is the specific gas constant for dry air and $L_i$ is the appropriate latent heat (ie $L_e$ when the surface is not frozen).

Two common methods are:

a. to specify a surface relative humidity, $r_o$, such that

$$q_o = r_o\, q_{sat}(T_o); \tag{67}$$

b. to evaluate a potential evaporation rate

$$E_p = -\rho C_E\, V(z_\ell)(q(z_\ell)-q_{sat}(T_o)) \tag{68}$$

and to specify an empirical 'moisture availability function', β, (usually ranging from 0 for an arid surface to 1 for a saturated surface) such that the actual evaporation rate is given by

$$E = \beta\, E_p \tag{69}$$

The second method is by far the more commonly adopted. For a discussion and comparison of the two approaches see, for example, Nappo (1975) and for examples of their use in specific AGCMs see Carson (1982). It is worth noting in passing that an alternative relation in the spirit of Eqn (69) is used for computational convenience in some models and that is

$$\Delta q(z_\ell) = q(z_\ell) - q_o = \beta(q(z_\ell) - q_{sat}(T_o)) \qquad (70)$$

which implies that

$$q_o = \beta\, q_{sat}(T_o) + (1-\beta)q(z_\ell) \qquad (71)$$

The reasons for preferring Eqn (70) to Eqn (69) are discussed in Carson (1982) (and more fully in Carson and Roberts (1977)).

The most common method now employed is to express $\beta$ as a simple linear function of the variable soil moisture content in the surface soil layer such that

$$\beta = \begin{cases} \hat{\chi}_v/\hat{\chi}_{v,c} & 0 \le \hat{\chi}_v < \hat{\chi}_{v,c} \\ 1 & \hat{\chi}_v \ge \hat{\chi}_{v,c} \end{cases} \qquad (72)$$

where $\hat{\chi}_{v,c}$ is a critical value of the mean volumetric soil moisture concentration below which $\beta < 1$, and is usually expressed as some fraction of a maximum allowable value of $\hat{\chi}_v$ ie some nominal 'field capacity' $\hat{\chi}_{v,f}$. Eqn (72) for $\beta$ can of course be simply reformulated in terms of any of the other standard measures of soil moisture content (see Eqn (63)) the most common of which is probably $d_\omega$.

A slight modification of Eqn (72), due to Warrilow et al (1986), is currently used in the Meteorological Office 11-layer AGCM, viz.

$$\beta = \begin{cases} 0 & 0 \le \hat{\chi}_v < \hat{\chi}_{v,\omega} \\ \dfrac{\hat{\chi}_v - \hat{\chi}_{v,\omega}}{\hat{\chi}_{v,c} - \hat{\chi}_{v,\omega}} & \hat{\chi}_{v,\omega} \le \hat{\chi}_v < \hat{\chi}_{v,c} \\ 1 & \hat{\chi}_v \ge \hat{\chi}_{v,c} \end{cases} \qquad (73)$$

where $\hat{\chi}_{v,\omega}$ is called the 'wilting point'.

The critical value $\hat{\chi}_{v,c}$ used in Eqn (73) is not well defined but for simplicity, and in line with previous practice (see, for example, Carson (1982)) it is given by

$$\hat{\chi}_{v,c} = \hat{\chi}_{v,\omega} + \frac{1}{3}(\hat{\chi}_{v,f} - \hat{\chi}_{v,\omega}) \qquad (74)$$

where $\hat{\chi}_{v,f}$ is a nominal 'field capacity' used only to define $\hat{\chi}_{v,c}$. The particular values of $\hat{\chi}_{v,\omega}$ and $\hat{\chi}_{v,f}$ being used globally in the Meteorological Office AGCM, which assumes for hydrological purposes only a single surface layer of soil of nominal depth $\delta_\omega = 1$ m, are listed in Table 6. With these values Eqn (73) reads

$$\beta = \begin{cases} 0 & 0 \leq \hat{X}_v < 0.08 \\ 20\,\hat{X}_v - 1.6 & 0.08 \leq \hat{X}_v < 0.13 \\ 1 & \hat{X}_v \geq 0.13. \end{cases} \quad (75)$$

For a surface soil layer 1 m deep Manabe (1969), who first introduced interactive surface hydrology with Eqns (69) and (72) into an AGCM, originally selected 15 cm as his field capacity (ie for $d_{\omega,f}$ in terms of $d_\omega$ used in Eqn (63)) and took $d_{\omega,c}/d_{\omega,f}$ (= $\hat{X}_{v,c}/\hat{X}_{v,f}$) = $^3/_4$. Carson's (1982) review of AGCMs indicates field capacities, $d_{\omega,f}$, in the range 10-30 cm and $^1/_3$ - $^3/_4$ for the ratio $d_{\omega,c}/d_{\omega,f}$.

It should be borne in mind that the more complex and real practical issue to be addressed is that of determining the actual evapotranspiration from partially vegetated surfaces and not simply the evaporation from a bare-soil surface.

| | |
|---|---|
| Depth of the surface layer of soil: $\delta_\omega$(m) | 1 |
| Characteristics of the mean volumetric soil moisture concentration: $\hat{X}_v$ | |
| Wilting point: $\hat{X}_{v,\omega}$ | 0.080 |
| 'Critical' point: $\hat{X}_{v,c}$ | 0.130 |
| Nominal field capacity: $\hat{X}_{v,f}$ | 0.230 |
| Saturation value: $\hat{X}_{v,s}$ | 0.445 |
| Saturated hydraulic conductivity: $K_s$ ($mmh^{-1}$) | 13.0 |
| Surface infiltration rate: F (SI units: kg $m^{-2}s^{-1}$) | 13.0 $mmh^{-1}$ equivalent (= $F/\rho_\omega$) |
| Exponent in Eqn (82): c | 6.6 |

Table 6. Soil hydrological characteristics used in the Warrilow et al (1986) hydrological scheme in the Meteorological Office 11-layer AGCM.

## 6.4 Surface runoff ($Y_o$)

Surface runoff is yet another of the complex surface hydrological processes which is treated very simplistically in current AGCMs and NWPMs. In models employing the single-layer water mass balance Eqn (64) the simplest approach is the so-called 'bucket model' for runoff (usually implicitly combining both $Y_o$ and $M_1$ in Eqn (64) into a single 'total runoff' term). In this case, rainfall (modified by the evaporation loss) is allowed to increase the soil moisture content until the field capacity $d_{\omega,f}$ (or $\hat{\chi}_{v,f}$) is reached. Any further attempt to increase $d_\omega$ (or $\hat{\chi}_v$) beyond the field capacity is implicitly assumed to be runoff water (including percolation to deeper layers) which plays no further part in the model's hydrological cycle. This identifies the original rôle played in these simple hydrological parametrizations by the field-capacity term, in addition to its use to define $d_{\omega,c}$ (or $\hat{\chi}_{v,c}$), as in Eqns (72)-(74). For a selection of the crude and highly empirical formulations used in specific AGCMs see, for example, Carson (1982).

A novel, but still relatively simple, parametrization has been developed by Warrilow et al (1986) for use in the Meteorological Office AGCM. It is based, with considerable simplification, on a scheme proposed by Milly and Eagleson (1982). An attempt has been made to allow for the spatial variability of rainfall since use of grid-box averages would give marked underestimation of the surface runoff. The rain is assumed to fall over a proportion $\mu$ of the grid-box where at present $\mu$ is chosen arbitrarily as 1 for the model's so-called 'large-scale dynamic rain' and as 0.3 for its 'convective rain'. These are thought to be conservatively high values. Eagleson and Qinliang (1985) have explored the likely coverage of a rainfall area for different catchment sizes and suggest that for an AGCM grid-square more appropriate values for $\mu$ are 0.6 and 0.05, respectively. The local rainfall rate, $P_{r\ell}$, throughout a grid-area is treated statistically as represented by the probability density function

$$f(P_{r\ell}) = \frac{\mu}{\bar{P}_r} \exp\left(\frac{-\mu P_{r\ell}}{\bar{P}_r}\right) \qquad (76)$$

where $P_r$ is the model's grid-point rainfall rate which is taken to represent the average grid-box rainfall.

The local surface runoff, $Y_{o\ell}$, is defined by

$$Y_{o\ell} = \begin{cases} P_{r\ell} - F & P_{r\ell} > F \\ 0 & P_{r\ell} \le F \end{cases} \qquad (77)$$

where F is a surface infiltration rate, deemed constant for a given soil, and at present given a fixed global value (equivalent to 13 mm $h^{-1}$). Integration of $Y_{o\ell}$ over all values of $P_{r\ell}$ yields an expression for the total surface runoff rate for a grid-area, viz,

$$Y_o = P_r \exp(-\mu F/P_r) \qquad (78)$$

### 6.5 The vertical mass flux of water at the base of the surface layer ($M_1$)

As indicated in the previous section, the simplest single-layer approaches typically assume explicitly that $M_1$ in Eqn (64) is negligible or implicitly that it combines with $Y_o$ to give a 'total runoff'. In the scheme of Warrilow et al (1986), adapted from Milly and Eagleson (1982), $M_1$, referred to as the gravitational drainage from the base of the surface layer, is acknowledged as a separate hydrological component of Eqn (64) that has to be parametrized.

Reference to Eqn (59) shows that

$$M_1 = -\rho_\omega \left[K(\psi)\left(\frac{\partial\psi}{\partial z_g} - 1\right)\right]_{z_g=\delta_\omega} \tag{79}$$

Warrilow et al (1986) have argued, somewhat speculatively, that for horizontal averaging over a typical AGCM grid-area, the term $[K(\psi)\partial\psi/\partial z_g]_{zg=\delta\omega}$ is small and that $M_1$ in Eqn (79) can be represented simply by

$$M_1 = \rho_\omega [K(\chi_v)]_{zg=\delta\omega} \tag{80}$$

with the further assumption that $\chi_v$ is effectively spatially homogeneous in the surface soil layer so that

$$M_1 = \rho_\omega K(\hat{\chi}_v) \tag{81}$$

Their particular prescription of the hydraulic conductivity as a function of $\hat{\chi}_v$, attributed to Eagleson (1978), is

$$K(\hat{\chi}_v) = K_s\left(\frac{\hat{\chi}_v - \hat{\chi}_{v,\omega}}{\hat{\chi}_{v,s} - \hat{\chi}_{v,\omega}}\right)^c \tag{82}$$

where $\hat{\chi}_{v,s}$ is termed the saturation value of $\hat{\chi}_v$, $K_s$ the saturation conductivity (ie $K(\hat{\chi}_{v,s})$) and c is an empirically derived constant. For particular values of these quantities adopted globally by Warrilow et al (1986) see Table 6.

With the terms E, $Y_o$ and $M_1$ evaluated according to a particular model's approach selected from the wide range of methods implied and discussed in Sections 6.3-6.5, and with $P_r$ determined by some other parametrization in the model, then Eqn (64) can be solved either in simple explicit fashion or by more subtle implicit methods to determine the change in $m_\omega$ (and hence in $\hat{\chi}_v$, $\hat{\chi}$, $d_\omega$, etc). This concludes the introduction to parametrization of land-surface hydrological processes in AGCMs and NWPMs.

## 7. Snow-covered Surfaces

A particular class of non-vegetated land surfaces which have their own very special characteristics and exercise significant influence on the climate system over a wide range of time-scales is that comprised of snow- (and ice-) covered surfaces. As in the case of land-surface hydrology, it is generally true that little attention has yet been given to the representation in AGCMs and NWPMs of the special physical processes associated with such surfaces. However, I am confident that this particular area of the wider problem will receive increasing attention in the near future.

According to Kuhn (1982), in the course of the year about 50% of the Earth's land surface is covered by snow or ice. He also comments that, although the polar ice sheets contain about 99% of the Earth's fresh-water ice by mass, nevertheless the seasonal snow cover with its large areal extent and its high spatial and temporal variability may have an equal or even greater impact on the atmospheric circulation. Undoubtedly then, a key issue will be how to deal sensibly with partial and rapidly changing snow cover, particularly in complex terrain, over the area of a typical grid-box in a large-scale numerical model. The proper treatment of the processes associated with snow-covered surfaces is a major topic in its own right. The brief comments here are no more than a postscript to the main discussion of bare-soil surfaces in Sections 2.6. For fuller expositions of the varied and complex characteristics and the effects of snow and its associated physical processes see, for example, Martinelli (1979), Male (1980), Gray and Male (1981) and IGS (1985). For discussions of snow covered surfaces aimed specifically at the AGCM parametrization problem see, in particular, Kuhn (1982) and Kotliakov and Krenke (1982).

7.1 Special conditions at snow- and ice-covered surfaces

Kuhn (1982) has listed the special conditions for snow and ice layers as:

a. the surface temperature cannot exceed the melting temperature of ice;

b. evaporation and sublimation take place at the potential rate;

c. the short-wave albedo is generally high;

d. the medium is permeable to air and water and transparent to visible radiation;

e. the snow pack is a good thermal insulator;

f. the layer has a high storage capacity for heat and water;

g. the roughness of the surface is extremely low (but see comments below at Section 7.2c); and

h. generally, the atmospheric surface layer over snow or ice is stably stratified.

Note that conditions a-h impinge on every aspect of the parametrization problem already discussed in Sections 2-6. The remainder of this section retraces our previous route and indicates briefly where modifications to the parametrizations are typically introduced into AGCMs and NWPMs in recognition of snow (or ice) covering the surface. In general the thermal and hydrological properties of the snow pack are represented very simply and crudely in such models.

7.2 The physical properties of snow- and ice-covered surfaces

a. Short-wave albedo ($\alpha$). It is firmly established that the physical coupling between snow and ice cover, albedo and the surface temperature is one of the most important feedback mechanisms to include in an AGCM. As indicated on the list above (7.1c), an important characteristic of snow- and ice-covered surfaces is their high reflectivity compared with other natural surfaces such that even a thin covering of fresh snow can alter significantly the albedo of a landscape. The local albedo of a snow-covered surface is very variable and a complicated function of many factors including the age of the snow pack ($\alpha$ decrease markedly as the snow becomes compacted and soiled), the wavelength and angle of the incident radiation and even diurnal cycles in the state of the snow surface, particularly when conditions are right for surface melting. The albedo may lie anywhere in the range from 0.95 for freshly fallen snow to about 0.35 for old, slushy snow (see, for example, Kondratyev et al (1982)).

At present the coupling between snow and ice and the surface albedo is generally prescribed very simply. Three types of snow- or ice-covered surfaces are generally acknowledged, viz:

(1) surfaces with instantaneously variable depth of snow either predicted or implied;

(2) permanent or seasonally prescribed snow- and ice-covered land surfaces; and

(3) permanent or seasonally prescribed areas of sea-ice.

The third category is not the concern of this paper. For models that 'carry' a snow depth a common approach still used is that of Holloway and Manabe (1971) who, following Kung et al (1964), introduced the following simple dependence of albedo on snow depth into an AGCM:

$$\alpha = \begin{cases} \alpha_\ell + (\alpha_s - \alpha_\ell)\, d_{s\omega}^{1/2} & d_{s\omega} < 1 \text{ cm} \\ \alpha_s & d_{s\omega} \geq 1 \text{ cm} \end{cases} \qquad (83)$$

where $\alpha_\ell$ is the snow-free land surface albedo (see Section 3.4); $\alpha_s$ is the albedo of a deep-snow surface (assumed in this case to be 0.60); and $d_{s\omega}$ is the water equivalent depth of snow (here expressed in cm). No allowance is usually made for the varying density of a snow pack and $d_{s\omega}$ is assumed typically to be about $^1/_{10}$ of the actual snow depth (see further comments in Section 7.3). Therefore the assumption is that when the grid-point snow depth is greater than about 10 cm, then $\alpha$ is independent of snow depth and equal to 0.60. Eqn (83) is designed supposedly to take account of the fact that as the mean snow depth increases, not only does the snow cover surface irregularities more completely but also the area of the grid-box which is snow-free is likely to decrease.

In some models which predict and monitor snowfall, a single albedo value is used for any non-zero depth of snow (see Carson (1982, 1986a)). The first snowfall on a previously snow-free surface results in an immediate increase in surface albedo which will tend, at least initially, to accelerate the positive feedback of a further lowering of the surface temperature with an enhanced probability of further snow accumulation.

Typical model values for land- and sea-ice are in the range 0.5-0.8 (see Carson (1982, 1986a)).

b. Long-wave emissivity ($\varepsilon$). Kuhn (1982) states that this can be assumed to be unity for all practical purposes.

c. Surface roughness length ($z_0$). The effective $z_0$ for extensive, uniformly covered snow and ice fields and the 'local' value of $z_0$ for snow-covered, simple heterogeneous terrain may indeed be very small ($O(10^{-3}$m) or less). However, in general, the effective areal $z_0$ of natural, heterogeneous and complex terrain with varied relief and vegetation is very difficult to determine (see Section 4.7) and may be affected greatly or insignificantly by different degrees of snow cover. There is little scope for useful discussion of this problem in a global, large-scale modelling context except to note that, in principle, snow and ice cover can alter $z_0$.

d. Thermal properties of snow. As noted above, a snow pack is generally a good thermal insulator for the soil below but to capture this effect in a climate model implies a delineation and explicit modelling of the heat conduction (and the hydrology) in and between the two media. In general the thermal and hydrological properties of snow and ice layers are treated very simply, if at all, in AGCMs and NWPMs (see below). The thermal properties of a snow pack will, like its density and albedo, depend in a complicated fashion on many factors. Values thought to be appropriate for snow are given in Table 4 for comparison with the range of soil values also included there.

## 7.3 Surface energy and mass flux balances at a snow-covered surface

The surface energy flux balance (Eqn (1)) is modified for complete snow cover such that

$$G_0 = R_N - H - Q_s - Q_f \qquad (84)$$

where $Q_f = L_f S$ represents the latent heat flux required to affect phase changes associated with melting or freezing at the surface, where S is the rate of snowmelt (or ice melt) and $L_f$ is the latent heat of fusion;

$Q_s = L_s E_s$ represents the latent heat flux due to surface sublimation by turbulent transfer, where $E_s$ is the rate of sublimation and $L_s$ is the latent heat of sublimation ($L_s = L_e + L_f$);

and $G_0$ is now strictly the flux of heat into the snow layer at its upper surface.

A simple budget equation, corresponding to that used for soil moisture content in a single soil layer (Eqn (64)), is also used for snow on the 'surface', viz.

$$\frac{\partial m_s}{\partial t} = P_s - E_s - S \qquad (85)$$

where $P_s$, the only undefined term on the RHS, is the intensity of snowfall at the surface and $m_s$ is the mass of snow lying per unit area of the surface. $m_s$ is therefore treated like $m_\omega$ as a surface prognostic variable and is often represented as a snow depth, $d_s$, or more commonly as an equivalent depth of water, $d_{s\omega}$, (cf Eqn (63)) such that

$$m_s = \rho_s d_s = \rho_\omega d_{s\omega} \tag{86}$$

where $\rho_s$ is the density of snow. Although it is recognised that the density of a snow pack varies, this again is a complicated issue in its own right and it is quite common practice in large-scale numerical models to assume simply that $\rho_s \simeq 0.1\ \rho_\omega$.

Eqn (85) is usually complemented by the surface-layer balance equation for the soil moisture content (Eqn (64)) modified to include the snowmelt term, ie

$$\frac{\partial m_\omega}{\partial t} = P_r - E + S - Y_o - M_1 \tag{87}$$

Each model has its own system of checks and algorithms for deciding which of the terms in Eqns (85) and (87) are in force simultaneously. One popular approach is as follows. When snow is lying $T_o$ is not allowed to rise above 273 K and the snow depth accumulates without limit or decreases according to the net value of $(P_s - E_s)$. If, however, snow is lying and the solution of the heat balance Eqn (84), excluding the terms $Q_f$, produces an interim surface temperature value $T_o' > 273$ K then sufficient snow (if available) is allowed to melt to maintain $T_o = 273$ K. The heat required to melt the snow and reduce $T_o$ to 273 K can be evaluated by specifying an effective surface thermal capacity of the snow pack (cf Eqn (43)) such that

$$Q_f = L_f S = C_{eff,s} \left(\frac{T_o' - 273}{\Delta t}\right) \tag{88}$$

where $\Delta t$ is the appropriate model time step. The change in the water equivalent snow depth, $\Delta d_{s\omega}$, resulting from the melting is determined from Eqn (85) and (88) such that

$$\Delta m_s = \rho_\omega \Delta d_{s\omega} = -S\ \Delta t$$

$$= -\frac{C_{eff,s}}{L_f} (T_o' - 273) \tag{89}$$

It is usually assumed that the snow pack has no moisture holding capacity; all melted snow is added directly to the soil moisture content (through Eqn (87)) following the corresponding reduction ($\Delta d_{s\omega}$) in the snow depth. In all cases it is only when the snow disappears through melting or sublimation that evaporation of moisture is allowed to resume at the surface.

## 8. Concluding Remarks

It should be evident from Sections 2-7 that, in many respects, the representations of land-surface processes in AGCMs and NWPMs are still rather crude and simple. The demands for improvements will come from both climate modelling studies and numerical weather forecasting. Indeed, the steadily increasing number of studies with AGCM's has already amply demonstrated the sensitivity of such models to surface properties and processes (see, for example, recent reviews by Mintz (1984), Rowntree (1983, 1984) and Rowntree et al (1985)). Parametrizations thought adequate at present will undoubtedly be seen to be deficient in models which couple interactively further components of the climate system. This is already apparent with respect to air-sea interactions in coupled ocean-atmosphere models. Although the major developments in the longer term are more likely to come from climate modelling studies, nevertheless valuable feedback is being obtained from the continuous close scrutiny of the various models' performances in the acutely critical arena of operational weather forecasting - especially of local, near-surface variables such as wind and temperatures.

A schematic resumé of the processes, variables and parameters introduced in this discussion of the specification of parametrizations for simple, non-vegetated land surfaces is given in Figure 9.

| | Radiative | Thermal | Hydrological | Dynamical |
|---|---|---|---|---|
| 'External forcing' | $R_{S\downarrow}$, $R_{L\downarrow}$ | | $P_r$, $P_s$ | |
| Atmospheric variables | | ← $\theta(T)$, | $q$, | $\underset{\sim}{V}$ → |
| Surface variables | ← $T_o$ | → | $q_o$, $d_s$ | |
| Surface parameters | $\alpha$, $\varepsilon$ | ← | $z_o$<br>$\beta$ | → |
| Surface fluxes | $R_N$ | $H$<br>$Q = L_e E$<br>$Q_f = L_f S$<br>$Q_s = L_s E_s$<br>$G_o$ | $E$, $E_s$<br>$S$<br>$M_o$, $Y_o$ | $\underset{\sim}{\tau}$ |
| Sub-surface fluxes | | $G$ | $M$ | |
| Sub-surface parameters | | $\lambda$, $C$ | $\delta$<br>$K$<br>Significant values of $\chi_v$ | |
| Sub-surface variables | | $T_g$ | $\chi_v$ | |

Figure 9. Schematic resumé of the processes, variables and parameters involved in the specification of parametrizations at simple, non-vegetated land surfaces.

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