DESCRIPTION OF BARE ESSENTIALS OF SURFACE TRANSFER FOR THE BUREAU OF METEOROLOGY RESEARCH CENTRE AGCM

May, 1991

A.J. Pitman, Z.-L. Yang, J.G. Cogley and A. Henderson-Sellers

Address for Correspondence

Dr A.J. Pitman
School of Earth Sciences
Macquarie University
North Ryde, NSW, 2109
Australia

Table of Contents

List of Figures

Abstract

This technical note describes in full detail the structure, development and rational behind a new land surface model designed for incorporation into Atmospheric General Circulation Models (AGCMs). The model, Bare Essentials for Surface Transfer (BEST), represents the lowest boundary of the atmosphere. It supports a large number of vegetation and soil types. The surface energy balance is modelled through a series of heat and moisture reservoirs, in particular an explicit canopy layer, three soil layers and a snow mass.

The method for accounting for surface heterogeneity, albedo, roughness length and stability, the canopy, soil and snow are all discussed. Soil characteristics and thermal properties are described as are methods for accounting for snow, snow pack metamorphosis, soil water balance, soil drainage and runoff.

In the appendix derivations of differentials required in the calculation of canopy temperature and canopy fluxes are given, along with the derivation of the exfiltration rate. Some sample simulations from BEST linked to the AGCM are also shown in Appendix 4.

Finally, notes on the implementation of BEST into the BMRC AGCM are provided.

1. Introduction

This document is a detailed description of the philosophy and physics of a new parameterization of the land surface. "**B**are **E**ssentials of **S**urface **T**ransfer", (BEST), is a time-dependent model, in one vertical dimension, of the base of the atmosphere over land. It interacts with a "host" model of the atmosphere, requiring the input of incident shortwave (solar) and longwave (atmospheric) irradiance, precipitation, wind vector components, surface pressure, air temperature and specific humidity. The principal outputs of BEST include the albedo, temperature, wetness and roughness length of the surface. These outputs may be expressed as fluxes of radiation, sensible and latent heat, heat conducted into the sub-surface, momentum and runoff. The host model is assumed to be an Atmospheric General Circulation Model (AGCM), although we use a much simpler forcing model for purposes of testing and validation and it is easy to imagine BEST coupled to host models of intermediate complexity between these extremes.

This description is of the most recent version of BEST (April 1991) which has been linked to the Bureau of Meteorology Research Centre (BMRC) AGCM (Hart *et al.*, 1988). The most complete description of the underlying theory and philosophy of BEST is provided by Cogley *et al.* (1990) which this document does not attempt to duplicate. Rather, this description is designed as a guide for anyone wishing to use BEST within the BMRC AGCM. Some of this description is duplicated in Cogley *et al.* (1990) but we have attempted to simplify this description as far as possible.

The need to consider and model accurately all the different kinds of surface type at the spatial scales typical of AGCMs, (e.g. from wave number 15 upwards means that BEST must accommodate both large-scale and smaller subgrid-scale spatial variability. AGCMs typically have time steps of 15-30 minutes and are used to perform simulations for periods of months to decades, so that BEST must respond well to forcing at the diurnal frequency, at higher frequencies, such as those involved in rainstorms, the passage of fronts and the rising and setting of the Sun and at lower frequencies, particularly the annual period.

The term *bare essentials* incorporated in the name of our model was coined in the belief that a detailed model of land surface climatology will probably never fit into an atmospheric AGCM; such a model has never been written, and if one ever were written it would probably overwhelm its atmospheric host model. On the other hand, present-day AGCMs do not simulate the lower boundary of the atmosphere with enough accuracy for the predictive demands being made upon them. The ideal, therefore, would be a land surface scheme of minimal complexity, where "minimal" means "containing all those features, and only those features, needed to compute fluxes and other land-surface properties to sufficient accuracy for the prediction of climate by AGCMs". While we still hold to these beliefs, we cannot claim that BEST provides only the bare essentials of surface-atmosphere transfer. There is no way of knowing which of the included features in BEST are superfluous or which important features we have left out. However, in designing BEST, the concept of *bare essentials* has been kept as a primary consideration hence we feel that the model is close to some optimum level of complexity.

This description is organised as follows. The next section surveys BEST giving its history, its conceptual structure and its computational structure. The third section explains sub-grid scale processes, and the fourth section describes the organisation of BEST's horizontal or geographical dimension. In Section 5, we detail the treatment of albedo by BEST, and turn in Section 6 to the surface roughness length and drag coefficients. Section 7 deals with the vegetation canopy which overlies the soil surface. Section 8 describes the surface-atmosphere exchanges of energy due to turbulence, and supplements the discussion of the canopy energy balances (Section 7). Section 9 explains the soil and snow pack model, discussing both heat and moisture accounting and transfer and Section 10 discussed the implementation of BEST into the BMRC AGCM. Some concluding remarks constitute the final section.

2. Model Architecture

In this section we explain the ancestry of BEST and its relationship to its host model. We also describe its structure informally as a way of introducing our notation.

2.1 History

Land surface models were originally developed along concepts of monthly water balance (e.g. Thornthwaite, 1948; Budyko, 1956). Manabe (1969) took this methodology and developed the "bucket model" which represented a valuable methodology so long as AGCMs simulated the climate without representing the diurnal cycle in the solar forcing (Dickinson, 1984). Once the inadequacies of the bucket model were realised other methods for representing the land surface were developed. This development took two main routes. First there were, and still are, attempts to improve on the basic bucket design (Hansen *et al.*, 1983; Boer *et al.*, 1984). More significantly, there have been attempts to model the surface physically (e.g. Deardorff, 1977, 1978; Dickinson, 1983, 1984; Dickinson *et al.*, 1986; Sellers *et al.*, 1986; Abramopoulos *et al.*, 1988).

The earliest version of BEST was a modified version of BATS (version 1b), the Biosphere Atmosphere Transfer Scheme of Dickinson *et al.* (1986). BEST has developed in such a way that little of the original design, or indeed the original detail remains of BATS. However, the purpose of BEST is similar to that of BATS, namely to compute scalar properties of, and fluxes at, the land surface with more accuracy than is typical of the land-surface schemes currently implemented in AGCMs. Although the basic framework used by BEST to do this is similar to BATS and most other physical models of the land surface, the methodology used by BEST to compute aspects of the surface energy and water balance are generally quite different.

2.2 Conceptual Structure

We picture the land surface as a complex of irregular layers separated by interfaces which may or may not cover continuously a patch of terrain. The horizontal extent of the patch is indefinite, but it is implicitly "large" or regional in scale. BEST is designed for nominally horizontal resolutions of > 1^o x 1^o although it may be suitable for limited area modelling at spatial scales of > 50km (cf. Giorgi and Bates, 1989).

The basic structure of BEST, and the arrangement of the layers is shown in Figure 1. The layers are identified by names and by mnemonic letters which appear as subscripts when we wish to refer algebraically to a particular layer or interface. The lowest model layer, m, is the lowest layer at which the host model computes dynamic and thermal properties of the atmosphere; it is the source of most of BEST's information about the behaviour of the fluid overlying the land surface. This layer should be as close to the surface as possible or a planetary boundary layer model used to extrapolate the atmospheric quantities required by BEST from the lowest model layer to the surface. The surface air layer, *s*, represents air in intimate contact with the land surface. Its nominal height above the surface is zero. For reasons which will be explained later, it is convenient to have a second reservoir of surface air, labelled *a*, which is in intimate contact with the convoluted surface of the vegetation canopy. We visualise the *s* layer as lying just above a smooth envelope covering the soil and the top of the canopy, if one is present. It should be noted that while layers *m* and *s* should be different, AGCMs often assume that quantities predicted for the lowest model layer are suitable approximations for the surface air level. This simplification is likely to become increasingly important as horizontal resolution increases and surface schemes improve, although increasing numbers of vertical layers will reduce the significance of this simplification.

The land surface itself is a composite of layers which may be partially or wholly exposed to the *s* layer. The BMRC AGCM "sees" an undifferentiated lower boundary (i.e. the AGCM only sees a single temperature, wetness, sensible heat and latent heat flux, albedo and roughness length). The lower boundary is therefore represented as the terrestrial surface or *t* layer. Depending on the context, properties of the *t* layer are computed in different ways, but they are always functions the properties of the upper ground layer *u*, usually the vegetation canopy *c*, and often the properties of a snow layer *n*. The *t* layer, or terrestrial surface is the basic surface for which quantities are predicted for the AGCM. Throughout this document many other layers are referred to (*s, n, u, l, c*). Quantities predicted for these layers form a variably sized component of the terrestrial surface and the AGCM only *ever* `knows' about this single land surface layer.

The canopy is the most dynamic single element in BEST and is assumed to cover a fraction of each grid element. This fraction varies with, among other things, temperature and location on the Earth's surface, and may range from 0 to, at least in principle, 1. The canopy itself is divided into green surfaces (i.e. leaves), which can transpire, and brown surfaces (i.e. branches), which do not transpire, the extent of the two kinds being described by leaf and stem indices respectively. The canopy as a whole may be covered by a variable depth and extent of water which we call dew and is labelled *d*, although the water may come from precipitation as well condensation. Whatever its source, the dew suppresses transpiration where it covers green canopy surfaces, and is important as a fast-response reservoir, allowing a more realistic description of the temporal variations in the land surface water balance.

The snow pack, *n*, covers a variable fraction of the canopy, and a different variable fraction of each ground layer. In practice the snow pack properties which are carried by BEST as global fields are the mass and the density of the snow. These are used to compute a snow depth averaged over the grid element, and the fractional extents of the snow pack which are parameterized in terms of snow depth.

The ground is divided into three soil layers by parallel interfaces, the layers being labelled *u* for upper, *l* for lower and *b* for bottom or base. When there is no snow, the three interfaces are horizontal, but in the presence of a snow pack we imagine them to parallel the snow surface. Thus, depending on the depth of the snow and the fractional horizontal extent of the snow pack, each of the ground layers may consist partly or entirely of snow. To be precise, this description is true only of the treatment of sub-surface heat transfer. For the modelling of water transfer we retain fixed boundaries between snow pack, topsoil and subsoil, and although there is good reason for this discrepancy it requires the interactions between the sub-models for heat and moisture be handled carefully.

2.3 Computational Structure

The organisation of BEST in terms of the order of calculations needed to complete a single time step will largely depend on the sequence in which the host model completes its own calculations. Figure 2 gives an example of the order of calculations currently used in the BMRC AGCM with BEST. The model begins by initializing BEST's grid fields in SURFIN. Then geographical data pertinent to the current grid element are assembled. in BEPREP provided the surface type (land, sea, ice) is land (defined by subroutine LNDSEV). These geographical data are derived from Wilson and Henderson-Sellers (1985) data set and are either constant or slowly-varying functions of temperature. The temperature of the previous model time step is considered an acceptable approximation when the current temperature is needed. Once all the geographical data are assembled, the albedo of the component parts of the land surface is calculated in ALBEDO, followed by soil fluxes and drag coefficients (BEFLUX), canopy quantities (CANOPY), surface hydrology (BEPLUM) and surface thermodynamics (BEHEAT). Table 1 shows all these subroutines and a more complete list of which routine calculates which variable.

Table 1. A list of the subroutines from the BMRC GCM which link with BEST and which subroutines of BEST calculate which basic quantities

Subroutine name	**Called from**	**Purpose (which variables calculated)**
SURFIN	PHYSD	Initialization of surface grid fields (N, ρ_n, T_u, T_l, T_b, T_t, $W_{L,u}$, $W_{L,l}$, $W_{F,u}$, $W_{F,l}$, D)
SURFDRV	SURFIN	Mainly deals with ocean and sea ice grid points
LNDSEV	SURFDRV	Determine whether surface is ocean, ice or land
BESTDRV	SURFDRV	Drives BEST and contains most of the BEST – AGCM linking. Also calculates H_{sm}, E_{sm}, T_t
BEPREP	BESTDRV	Determine the parameter data used from Table 1
BEALBEDO	BESTDRV	$\alpha_{SW,c}$, $\alpha_{LW,c}$, $\alpha_{SW,u}$, $\alpha_{LW,u}$, $\alpha_{SW,n}$, $\alpha_{LW,n}$
BEFLUX	BESTDRV	z_0, C_{Dh}, H_{us}, E_{us}, C_{Dm}, H_{su}, β_u
CANOPY	BEFLUX	r_s, T_c, E_{cs}, H_{cs}, E_{ua}, H_{ua}, R_i, β_c, E_{ca}, E_{da}, f_{wet}, f_{dry}, E_{trca}, E_{uc}, E_{lc}
BEPLUM	BESTDRV	C_{vu}, H_{ul}, β_t, N, R_{su}, R_{ul}, R_{lb}, $W_{L,u}$, $W_{L,l}$, $W_{F,u}$, $W_{F,l}$, D, c_{vu}, c_{vl}, c_{vn}, K_h, K_{Tu}, K_{Tl}, K_{Tn}
BEHEAT	BEPLUM	T_u, T_l, T_b

2.4 *Notation*

Describing a model like BEST requires a large amount of simple but bulky algebra. In an effort to reduce this bulk we have adopted the following conventions.

Differentiation with respect to the vertical coordinate z is denoted by the∇ operator and differentiation with respect to time is denoted by an overdot; that is, $\nabla(\ ) = d\,/\,dz$ and $(\dot{\ }) = d/dt$. In describing the heat balance of the canopy we use a prime and a double prime to denote differentiation with respect to canopy temperature and the drag coefficient respectively.

All fluxes carry two lower case subscripts; the direction of the flux is from the layer denoted by the first subscript to the layer denoted by the second subscript (e.g Z_{us} is the flux of Z from the upper soil layer (u) to the air above the surface (s)). In general, lower case subscripts refer to layers (e.g. u for upper, l for lower, t for terrestrial n for snow, d for intercepted water and b for base) while upper case subscripts distinguish such attributes as phase of water substances, and spectral region (e.g. L for liquid, F for frozen, SW for visible and LW for longwave).

In the soil, volume fractions of soil water and soil ice relative to the soil volume are denoted by the letter X. Fractions relative to the volume fraction of voids (i.e., the porosity) are denoted by the letter W. A list of symbols appearing in this document together with their FORTRAN names in the code can be found in Appendix 1.

3. Sub–grid scale processes

Sub–grid scale processes are processes which operate, and are considered important at spatial scale of less than the AGCM's grid element. In the last few years the realisation that these processes are crucially important has lead to something of a revolution in the parameterization of some land surface processes (see, for example, Mahrt, 1987).

In this section we draw attention to the importance of sub–grid scale processes and indicate how BEST accounts for their influence. The actual parameterizations incorporated into the model are generally discussed in their proper place, but this section identifies them, and indicated where further explanation can be found.

There are basically two approaches to the correction of model results to account for sub–grid scale processes, and BEST incorporates both. The first approach is to regard any AGCM grid element as consisting of several sub–cells (each sub–cell is assumed to be homogeneous). Calculations are performed for each sub–cell and the results are then weighted according to the fractional extent of the individual sub–cells to yield composite estimates of the quantity for the whole grid element. Unfortunately, there is no guarantee that the number of sub–cells required to solve the problem will be small and the number is large then little is gained. The second approach is to correct the transfer coefficients by parameterizing the sub–grid scale variability. Unfortunately it is impossible to determine how realistic the parameterization is and hence the model simulation is effectively tuneable. BEST also accounts for the sub–grid scale with respect to time. The ripening of the snow pack by melt water percolation, leaf drip and stem flow are examples which are discussed later.

BEST presently assumes that each grid element is a composite of two sub–cells, vegetated and bare soil which correspond to the presence or absence of the canopy component of the model. The parameter A_v is defined in Equation (4.1) and depends on the composite land surface type, the soil temperature and the amount of snow. BEST has been extended to account for small amounts of open water within a grid element (i.e. small lakes, Pitman, 1991) but sensitivity tests within an AGCM remain to be preformed.

Snow cover can also occur at sub–grid scales. Its extend depends upon the roughness length of the vegetation and the amount of snow. Equation (4.5) discussed the fractional cover of the canopy while Equation (4.7) discusses the coverage of the ground.

The infiltration of precipitation or water from snow melt infiltrates into the ground at a rate depending on the wetness of the soil and the soil texture. Section 9.5 discusses a simple sub-grid scale parameterization of the spatial heterogeneity in the infiltration rate.

Finally, it has recently been shown that the sub-grid scale variability in precipitation (in particular the differentiation between large scale and convective rainfall) can have profound implications on the simulation of the surface climatology (Pitman *et al.*, 1990). The version of BEST incorporated into the BMRC AGCM does not incorporate the parameterization discussed by Pitman *et al.* (1990) because the sensitivity shown by their simulations were so severe. However, any information the AGCM can provide in the direction of precipitation type, intensity or spatial extent can be incorporated into BEST and sensitivity tests are planned to investigate this area.

There are other important processes which operate at sub-grid scale processes including soil moisture, variations in the roughness length, albedo etc. Each of these needs to be examined, but BEST is forced to assume that most processes are spatially homogeneous since there is insufficient information to do otherwise.

4. Heterogeneous grid-cells

4.1 Surface types

Traditionally the treatment of the Earth's surface in AGCMs has been extremely simple with a homogeneous surface type (either ocean, sea ice, snow or land) has been assigned for each grid element. Some AGCMs have allowed limited heterogeneity to exist, for instance, over sea ice by simulating sea ice leads and over soils by allowing snow to exist. Over the land surface bare soil has been assumed to cover the whole grid element uniformly, which is obviously not the case in the real world.

The ability to account for the effects of surface heterogeneity is one of the principal driving forces behind the development of realistic land surface models. Recently, new land surface models for use in AGCMs have been introduced which incorporate limited heterogeneity at the land surface (e.g., Dickinson *et al.*, 1986; Sellers *et al.*, 1986; Abramopoulos *et al.*, 1988). For example, bare soil, snow and vegetation can all exist simultaneously in a single grid element. In the version of BEST described here, the fraction of the grid element covered by the three surface types have the following relationship,

$$\begin{aligned} A_u + A_v &= 1 \\ A_n + A_u &= 1 \end{aligned} \tag{4.1}$$

where A_u is the fraction of the bare soil, A_n is the fraction of snow and A_v is the fraction of vegetation. All fractions are relative to the grid element and all fractions are ≥ 0. Equation (4.1) incorporates the concept of self-similarity discussed by Cogley *et al.* (1990).

In BEST, the parameterization of A_n is based on the simulated snow depth over the grid element and A_v is specified separately from the look-up table, discussed in detail later. Therefore, A_u is calculated as a residual from Equation (4.1).

The calculation or prescription of A_v, A_n and A_u is clearly important since these terms are the basis for all other modelled quantities. For instance, if soil evaporation and canopy evaporation could be predicted perfectly, the land surface model would only simulate the correct *overall* evaporation flux if the relative fractions in Equation (4.1) were correct. The critical fraction is A_v since $A_n = 0$ when there is no snow, and A_v has to be prescribed from data (e.g. Wilson and Henderson-Sellers, 1985). Although land surface models are qualitatively insensitive to small variations in A_v (Mahfouf and Jacquemin, 1989; Sellers and Dorman, 1987) BEST must be provided with a realistic estimate. In the following two sections the methods for calculating A_v and A_n are described.

4.2 Vegetation fraction

In order to accommodate vegetation into BEST its seasonal fractional extent must be specified. In BEST, the fractional vegetation cover is dependent upon the ecotype derived from the land use data set of Wilson and Henderson−Sellers (1985). The fractional vegetation cover (A_{v0}) varies seasonally and is parameterized following Dickinson *et al.* (1986) as, in the absence of snow

$$A_{v0} = A_{vmax} - S_c[1 - f(T_{soil})] \tag{4.2}$$

where the depth averaged soil temperature T_{soil} is

$$T_{soil} = 0.5\ (T_l + T_b) \tag{4.3}$$

where T_l is the lower soil temperature and T_b is the bottom soil temperature. Note that in Equation (4.3), T_l and T_b are used rather than T_u because by omitting T_u a diurnal cycle in A_{v0} is prevented. The seasonal factor is defined following Dickinson *et al.* (1986) as

$$f(T) = \begin{cases} 0 & T \leq 273\ K \\ 1 - 0.0016\ (298 - T)^2 & 273\ K < T < 298\ K \\ 1 & T \geq 298\ K \end{cases} \tag{4.4}$$

A_{vmax} is the maximum fractional vegetation cover (if $T_{soil} \geq 298$ K), S_c is the difference between the maximum fractional vegetation cover and a cover at a temperature of 269K (A_{vmax} and S_c are tabulated in Table 2).

Table 2. Canopy/Land-cover types and their parameters used in BEST

Code	Ecotype	A_{vmax}	S_c	L_{AImax}	S_l	S_{AI}	d_u	d_l	f_{rootu}	z_{0c}	$\alpha_{SW,c}$	$\alpha_{LW,c}$	f_c
00	Open water	0.00	0.00	0.0	0.0	0.0	0.0	0.0	0.0	0.001	0.09	0.05	0.0
01	Inland water	0.00	0.00	0.0	0.0	0.0	0.0	0.0	0.0	0.001	0.09	0.05	0.0
02	Bog or marsh	0.90	0.00	4.0	3.5	1.0	0.1	0.9	0.80	0.03	0.08	0.16	0.01
03	Ice	0.00	0.00	0.0	0.0	0.0	0.2	1.8	0.0	0.002	0.85	0.65	0.0
04	Paddy rice	0.90	0.80	3.0	2.5	0.5	0.1	0.9	0.90	0.06	0.08	0.16	0.01
05	Mangrove (tree swamp)	0.90	0.00	4.0	1.0	1.0	0.1	0.9	0.9	1.00	0.10	0.15	0.01
10	Dense needleleaf evergreen forest	0.90	0.00	6.0	1.0	2.0	0.1	1.4	0.67	1.00	0.09	0.19	0.03
11	Open needleleaf evergreen woodland	0.90	0.00	6.0	1.0	2.0	0.1	1.4	0.67	1.00	0.09	0.19	0.03
12	Dense mixed needleleaf & broadleaf, evergreen and deciduous forest	0.90	0.00	6.0	3.0	2.0	0.1	1.4	0.67	1.00	0.09	0.18	0.03
13	Open mixed needleleaf & broadleaf, evergreen & deciduous forest	0.70	0.00	6.0	3.0	2.0	0.1	1.4	0.67	1.00	0.09	0.18	0.03
14	Open evergreen broadleaf woodland	0.70	0.00	6.0	1.0	2.0	0.1	1.4	0.75	1.00	0.09	0.19	0.03
15	Evergreen broadleaf cropland	0.90	0.00	4.0	1.0	1.5	0.1	1.4	0.80	0.10	0.12	0.22	0.02
16	Evergreen broadleaf shrub	0.90	0.00	4.0	1.0	1.5	0.1	1.4	0.80	0.10	0.12	0.22	0.02
17	Open deciduous needleleaf woodland	0.90	0.00	6.0	1.0	2.0	0.1	1.4	0.80	1.00	0.12	0.22	0.03
18	Dense deciduous needleleaf forest	0.90	0.00	6.0	5.5	2.0	0.1	1.4	0.67	1.00	0.09	0.17	0.03
19	Dense evergreen broadleaf forest	0.90	0.00	6.0	1.0	2.0	0.1	1.4	0.75	1.00	0.09	0.19	0.03
20	Dense deciduous broadleaf forest	0.90	0.00	6.0	5.5	2.0	0.1	1.4	0.67	1.00	0.09	0.17	0.03
21	Open deciduous broadleaf woodland	0.70	0.00	6.0	5.5	2.0	0.1	1.4	0.67	1.00	0.09	0.17	0.03
22	Deciduous tree crops (temperate)	0.70	0.00	4.0	3.5	2.0	0.1	1.4	0.67	1.00	0.09	0.17	0.03
23	Open tropical woodland	0.90	0.10	6.0	5.5	2.0	0.1	1.4	0.67	1.00	0.09	0.17	0.03
24	Woodland + shrub	0.90	0.10	5.0	4.5	2.0	0.1	1.4	0.67	0.50	0.09	0.17	0.03
25	Dense drought deciduous forest	0.90	0.00	6.0	5.5	2.0	0.1	1.4	0.60	1.00	0.09	0.17	0.02
26	Open drought deciduous woodland	0.70	0.00	6.0	5.5	2.0	0.1	1.4	0.60	1.00	0.09	0.17	0.02
27	Deciduous shrub	0.90	0.00	4.0	3.5	1.5	0.1	1.4	0.80	0.10	0.10	0.22	0.02
28	Thorn shrub	0.75	0.00	3.0	2.5	1.5	0.1	0.9	0.67	0.10	0.10	0.22	0.02
30	Temperate meadow and permanent pasture	0.75	0.05	3.0	2.5	1.0	0.1	0.9	0.90	0.10	0.13	0.25	0.01
31	Temperate rough grazing	0.75	0.05	4.0	3.5	1.5	0.1	0.9	0.90	0.10	0.14	0.26	0.01
32	Tropical grassland and shrub	0.75	0.05	4.0	3.5	1.5	0.1	0.9	0.90	0.10	0.14	0.26	0.01
33	Tropical pasture	0.75	0.05	4.0	3.5	1.5	0.1	0.9	0.90	0.10	0.14	0.26	0.01
34	Rough grazing and shrub	0.75	0.05	4.0	3.5	1.5	0.1	0.9	0.90	0.10	0.14	0.26	0.01

Code	Ecotype	A_{vmax}	S_c	L_{AImax}	S_l	S_{AI}	d_u	d_l	f_{rootu}	z_{0c}	$\alpha_{SW,c}$	$\alpha_{LW,c}$	f_c
35	Pasture and trees	0.75	0.05	4.0	3.5	1.5	0.1	0.9	0.90	0.10	0.14	0.26	0.01
36	Semi arid rough grazing	0.75	0.05	4.0	3.5	1.5	0.1	0.9	0.90	0.10	0.14	0.26	0.01
37	Tropical savanna (grassland and tree)	0.75	0.05	4.0	3.5	1.5	0.1	0.9	0.90	0.10	0.14	0.26	0.01
39	Pasture + shrub	0.75	0.05	4.0	3.5	1.5	0.1	0.9	0.90	0.10	0.14	0.26	0.01
40	Arable cropland	0.90	0.40	4.0	3.5	0.5	0.1	0.9	0.90	0.06	0.14	0.26	0.01
41	Dry farm arable	0.90	0.40	4.0	3.5	0.5	0.1	0.9	0.90	0.06	0.13	0.25	0.01
42	Nursery and market gardening	0.90	0.40	4.0	3.5	0.5	0.1	0.9	0.90	0.06	0.13	0.25	0.01
43	Cane sugar	0.90	0.40	4.0	3.5	0.5	0.1	0.9	0.90	0.06	0.12	0.22	0.01
44	Maize	0.90	0.40	4.0	3.5	0.5	0.1	0.9	0.90	0.06	0.13	0.25	0.01
45	Cotton	0.90	0.40	4.0	3.5	0.5	0.1	0.9	0.90	0.06	0.13	0.25	0.01
46	Coffee	0.90	0.40	4.0	3.5	0.5	0.1	0.9	0.90	0.06	0.13	0.25	0.01
47	Vineyard	0.90	0.40	4.0	3.5	0.5	0.1	0.9	0.90	0.06	0.13	0.25	0.01
48	Irrigated cropland	0.90	0.40	4.0	3.5	1.0	0.1	0.9	0.90	0.06	0.16	0.34	0.01
49	Tea	0.90	0.40	4.0	3.5	1.0	0.1	0.9	0.90	0.06	0.13	0.25	0.01
50	Equatorial rain forest	0.90	0.00	6.0	5.5	2.0	0.1	1.4	0.67	1.00	0.09	0.17	0.03
51	Equatorial tree crop	0.90	0.40	4.0	3.5	1.0	0.1	1.4	0.80	0.10	0.13	0.25	0.01
52	Tropical broadleaf forest (slightly seasonal)	0.70	0.10	6.0	5.5	2.0	0.1	1.4	0.67	1.00	0.09	0.17	0.03
61	Tundra	0.75	0.00	4.0	3.5	0.5	0.1	0.9	0.90	0.06	0.13	0.25	0.01
62	Dwarf shrub (tundra transition and high altitude wasteland)	0.75	0.00	4.0	3.5	0.5	0.1	0.9	0.90	0.06	0.13	0.25	0.01
70	Sand desert and barren land	0.00	0.00	0.0	0.0	0.0	0.2	0.8	0.50	0.01	0.30	0.40	0.00
71	Scrub desert and semi–desert	0.20	0.10	1.0	0.5	1.0	0.1	0.9	0.67	0.05	0.2	0.2	0.01
73	Semi–desert + scattered trees	0.20	0.10	1.0	0.5	1.0	0.1	0.9	0.67	0.05	0.2	0.2	0.01
80	Urban	0.20	0.00	4.0	3.5	1.5	0.1	0.9	0.80	1.50	0.18	0.18	0.01

Table 2b Other parameters prescribed in BEST as values independent of canopy/land-cover types

Parameter	Value	SI units
D_{max0}	0.2	kg m^{-2}
E_{tr0}	1.8×10^{-4}	kg m^{-2} s^{-1}
C_f	0.01	m s$^{-1/2}$
S_f	0.04	m
r_{smin}	200	s m^{-1}
r_{smax}	5000	s m^{-1}
U_{cmin}	0.02	m s^{-1}
z_{0n}	0.00093	m
z_{0u}	0.01	m
κ	0.40	–

When snow cover exists, the fraction of the canopy covered by snow is,

$$A_{nc} = \frac{d_n}{d_n + 5z_{0c}} \qquad 0 \le A_{nc} \le 1 \tag{4.5}$$

where z_{0c} is the canopy roughness length (see Table 2) and d_n is the snow depth defined above. Therefore, the actual fractional vegetation cover (A_v) excluding the snow−covered portion is,

$$A_v = (1 - A_{nc})\, A_{v0} \tag{4.6}$$

In this parameterization of snow cover, snow can not completely cover the vegetation. In addition, the depth at which A_{nc} approaches 1 varies according to the roughness length of the canopy. Hence forests require significantly more snow to begin to mask the vegetation as compared to bare soil.

4.3 Snow fraction

The subgrid scale variability in snow cover is represented by the snow fraction A_n in BEST which is parameterized as

$$A_n = \left(\frac{d_n}{2d_{neff}}\right)^{1/2} \qquad (0 \le A_n \le 1) \tag{4.7}$$

where

$$d_n = \frac{N}{\rho_n} \tag{4.8}$$

is snow depth, N is the snow mass on the ground, ρ_n is the snow density. The calculations of ρ_n and N are discussed in a later section. The effective snow depth d_{neff} is

$$d_{neff} = 0.038 + 0.000144\rho_n \tag{4.9}$$

This expression is in a simplified form which parameterizes the damping depth and thermal conductivity of snow and is derived in more detail by Cogley *et al.* (1990). When calculating the snow mass N, it is assumed that the snow falls uniformly over the grid element. It will be shown later that the snow fraction concept is useful in calculating the grid element averaged albedo, roughness length, thermal properties and turbulent fluxes.

Equation (4.2) has important implications for BEST. Mahfouf and Jacquemin (1989) showed that a BEST like surface parameterization was quite sensitive to significant variations in A_v through the simulation of interception. The form of Equation (4.2) follows Dickinson *et al.* (1986) by imposing a seasonal cycle on A_v in association with the deep soil temperature. This appears to work well in general, but depends on the realistic specification of S_c and A_{vmax} for each grid element. The function *f(T)* is also a problem since it requires thresholds to be specified to define the range of the function. Clearly, different types of vegetation reach maximum density or optimise growth at different temperatures. The function *f(T)* is currently general and applies to all vegetation types although this apparent gross simplification is tempered by the ecotype specific definition of S_c. We intend to replace the function *f(T)* by a more appropriate quantity in the future, but Equation (4.4) is satisfactory for the present.

4.4 Data sets needed by BEST

When BEST is accessed by the AGCM, the first priority is to match the land surface type (ecotype) as determined by the AGCM's background geography to specific data required by BEST. This provision of data is the weakest element in land surface models and the lack of sufficient data and the number of "tunable parameters" has often been used as excuses for retaining simpler land surface schemes. However, enough data is available to support all but the most advanced models, and even in the case of Sellers *et al.*'s (1986) model (SiB) it has been shown by Sato *et al.* (1989a,b) that the model performs extremely well. In any case, data are becoming available in increasing amounts through large scale field experiments (FIFE, HAPEX, BOREAS *etc.*) and hence developing models to use this data is essential.

BEST can, in principal, accommodate any number of ecotypes or soil types. BEST requires soil and canopy information in order to determine the values of those variables provided by Tables 2 (canopy) and 3 (soil) for a specific soil texture and/or ecotype.

Table 3. Soil texture classes and parameters

Soil type	K_{H0} (W m^{-1} K^{-1})	B	c_{vm} (J m^{-3} K^{-1})	ψ_0 (m)
Clay	0.001	10.0	2.38×10^6	-0.2
Sandy clay	0.0015	9.2	2.38×10^6	-0.2
Clay loam	0.0021	8.4	2.38×10^6	-0.2
Silt clay loam	0.0032	7.6	2.38×10^6	-0.2
Sandy clay loam	0.0045	6.8	2.38×10^6	-0.2
Loam	0.006	6.0	2.38×10^6	-0.2
Loamy silt	0.0090	5.5	2.38×10^6	-0.2
Silty loam	0.0150	5.0	2.38×10^6	-0.2
Sandy loam	0.0400	4.5	2.38×10^6	-0.2
Sand	0.100	4.0	2.38×10^6	-0.2

The ecotype and soils information is provided by the BMRC AGCM from the global data of Wilson and Henderson‑Sellers (1985).

The canopy or vegetation index derived from the Wilson and Henderson‑Sellers (1985) data set can not be averaged since the index is only a classification. However, quantities in the vegetation look‑up tables (e.g. Table 2) which correspond to the vegetation index or vegetation type are physical quantities which can be aggregated. The usual method for aggregating 1° x 1° data to the AGCM resolution is to calculate which ecotype is most common within an area representing an AGCM grid square, and then to assume that this single ecotype is representative of the entire grid square. This ecotype is then linked to a look‑up table of parameters analogous to Table 2. An alternative and more realistic methodology is used by BEST and the BMRC AGCM. The 1° x 1° vegetation index is associated with a look‑up table and a geometric or arithmetic averaging of parameter values to combine all 1° x 1° sub‑elements within a single grid element is performed.

The effect of the different procedure can be quite significant and sensitivity tests within the AGCM are planned. In particular, note that the full range of 53 ecotypes described by Wilson and Henderson‑Sellers (1985) can be represented as well as small amounts of, say forest within a predominantly grassland grid square. Further, the secondary vegetation type information in the Wilson and Henderson‑Sellers (1985) data set can also be incorporated. Although the averaging procedure can not be shown to produce a "more correct" answer it does reduce the loss of information and produce an intuitively more reasonable estimate for each of the parameters shown in Table 2.

The methodology for calculating values for soil parameters in BEST has been to average arithmetically the texture index (coarse, medium, fine) provided at 1° x 1° resolution by Wilson and Henderson-Sellers (1985). By assigning fine = 1, medium = 2 and coarse = 3, an arithmetic average can then be mapped onto an extended look-up table to provide the necessary information (Table 3). This is satisfactory except where a soil quantity covers a wide range of values, *e.g.* hydraulic conductivity, K_h, which can range over several orders of magnitude in which case a geometric mean is performed.

Wilson and Henderson-Sellers (1985) suggest that their soil data should be aggregated by averaging the soil texture index. It is proposed here that the soil parameters associated with a given texture index should be aggregated instead since this will give a slightly more accurate answer. Although these errors incurred by averaging the index are comparatively small, they are easily avoided by averaging the soil parameters themselves.

The principal disadvantage of averaging parameters from the original data for the AGCM is that there is a loss of the classification information. In BATS, for instance, it is possible to show the spatial distribution of actual vegetation types. In the scheme proposed here this is not possible since canopy type is impossible to determine from the aggregated data. This does not have implications with respect to future developments of interactive biosphere models, but it does have implications with respect to experiments for which land surface type must be determined. This problem would seem minor compared to the advantages of utilising the full characteristics of the 1° x 1° data sets.

The colour index which is provided by Wilson and Henderson-Sellers (1985) is used to calculate the soil albedo (see Section 5). The texture index is used to calculate a variety of soil related parameters such as soil porosity, field capacity and the volume fraction at which permanent wilting occurs (see below). The drainage index is presently not used, although a role is anticipated in future soil drainage and runoff parameterizations. The provision, accuracy and reliability of these data are a major problem (see Wilson and Henderson-Sellers, 1985).

It should be noted that the data presented in Table 2 are derived largely by intelligent guessing, guided by various micrometeorological texts, e.g., Geiger (1965), Monteith (1973, 1976), and other literature, in particular, Federer (1979) for leafless hardwood forests. The soil properties in Table 3 are obtained from a variety of sources, e.g., Campbell (1974) and Clapp and Hornberger (1978). We have tended to incorporate similar data to that used by Dickinson *et al.* (1986) unless more realistic or recent measurements are available.

The data shown in Tables 2 and 3 differs in some respects from the data provided by Dickinson *et al.* (1986) shown in their tables 2 and 3. Most of the significant differences between these tables are in terms of complexity or format. For instance, Dickinson *et al.* (1986) provide a different value for their inverse square root of leaf dimension (in $m^{-1/2}$) for each land cover type although it is a constant for all types other than type 1 (crop); in BEST a constant value (i.e., S_f = 0.04 m) for all land-cover types is assumed. In Dickinson *et al.* (1986), the porosity (X_v) and the moisture content relative to saturation at which transpiration ceases (W_{wilt}) are specified in the table together with other soil properties as functions of texture class. In BEST, these variables are calculated using the soil texture index (t_{ex}), which is obtained from Wilson and Henderson-Sellers (1985). Despite this difference of format, their values are essentially the same. For example, in BEST, the soil porosity, X_v, is given by (see Figure 3)

$$X_v = 0.6 - 0.03\, t_{ex} \tag{4.10}$$

where t_{ex} is the index for soil texture, which ranges from 0 (fine texture, e.g. clay-type soil) to 9 (coarse texture, e.g. sand-type soil). Note that for glacier ice, X_v = 0.07. The volume fraction occupied by soil moisture at field capacity, X_{FC}, is given by (see Figure 3)

$$X_{FC} = (0.95 - 0.086\, t_{ex})\, X_v \tag{4.11}$$

This expression is a simple fit to data used by BEST. The volume fraction at which permanent wilting occurs, X_{wilt}, is given by (see Figure 3)

$$X_{wilt} = X_v - 0.3 \tag{4.12}$$

Note that for glacier ice, X_{wilt} = 0.01. Then, W_{wilt} (required in Equations (7.40) and (7.41) is given by

$$W_{wilt} = \frac{X_{wilt}}{X_v} \tag{4.13}$$

Equations (4.10)-(4.12) simply provide a means to infer values for X_v, X_{FC} and X_{wilt} with the global soils data set provided by Wilson and Henderson-Sellers (1985). An assumption is made that X_v, X_{FC} and X_{wilt} are all related to soil texture, and that a given soil type will (globally) have the same value for X_v, X_{FC} and X_{wilt}. This is a simplification which was unavoidable due to the lack of more reliable data sources.

This section has described the framework of BEST and has shown how it links to the BMRC AGCM for supply of atmospheric and initialisation data. The remaining sections expand on the fluxes, temperatures and moistures are calculated for fractions A_u, A_n and A_v.

5.0 Soil, canopy and snow albedo parameterization

Henderson-Sellers and Wilson (1983) showed based on data from Kondratyev (1969), O'Brien and Munis (1975) and Wiscombe and Warren (1980) that the albedo of many surfaces is a strong function of wavelength. For example, the albedo of plants increases suddenly from about 0.10 in the wavelength region < 0.7 μm to around 0.5 at near-infrared wavelengths (see Figure 4). In contrast, the spectral albedo of snow follows an opposite path. Very high albedos (0.7 - 0.8) in the visible wavelengths drop to lower values (0.3 - 0.4) in the longwave region of 0.7 - 1.0 μm (see Figure 4). As a simplification, therefore, the surface albedo in BEST is calculated in terms of two wavelength bands (visible and near-infrared) split at 0.7 μm. Assuming the incident solar flux is split equally in the two bands, the combined albedo for the total wavelength region of the solar spectrum, α, is

$$\alpha = \frac{1}{2} (\alpha_{SW} + \alpha_{LW}) \tag{5.1}$$

where α_{SW} is the albedo for the visible wavelength region (≤ 0.7 μm), and α_{LW} is the albedo for the infrared wavelength region (> 0.7 μm). A list of the albedos for α_{SW} and α_{LW} for different ecotypes is given in Table 2. The albedos for bare soil (α_u) snow (α_n) and vegetation (α_c) can be written as

$$\alpha_u = 0.5(\alpha_{SW,u} + \alpha_{LW,u}) \tag{5.2}$$

$$\alpha_n = 0.5\ (\alpha_{SW,n} + \alpha_{LW,n}) \tag{5.3}$$

$$\alpha_c = 0.5\ (\alpha_{SW,c} + \alpha_{LW,c}) \tag{5.4}$$

The details of the specification of the notations can be seen in Appendix 1. The albedo for bare soils is given by,

$$\alpha_{SW,u} = 0.10 + 0.1c_{lr} + 0.07(1 - W_{L,u}) \tag{5.5}$$

$$\alpha_{LW,u} = 2\alpha_{SW,u} \tag{5.6}$$

where c_{lr} is the soil colour from Wilson and Henderson-Sellers (1985) which ranges from 0 (the darkest soils) to 1 (the lightest coloured soils). $W_{L,u}$ is the soil wetness factor in the upper soil layer, defined as the ratio of the liquid soil water content to the porosity, which is detailed in the later sections. The soil wetness is included in Equation (5.5) following reference to observations by, for example, Idso *et al.* (1975) who showed that changes in soil wetness can have a significant affect on the visible soil albedo.

The assumption that soil longwave albedo is twice the shortwave albedo is justified by reference to spectral observations such as those reported by Kondratyev (1969). Equations (5.5) and (5.6) together yield broadband albedos consistent with those reported by, for example, Idso *et al.* (1975).

In BEST glacier ice is represented explicitly since the overall climatic influence of Antarctica, Greenland and other smaller ice caps are extremely important. The albedo of glacier ice can be defined following Cogley (1987) as

$$\alpha_{SW,u} = 0.60 + 0.06\ (1 - W_{L,u}) \tag{5.7}$$

$$\alpha_{LW,u} = \frac{1}{3}\alpha_{SW,u} \tag{5.8}$$

These relations express patterns seen in observations by Bolsenga (1977) and in theoretical spectra modelled by Warren (1982). Notice that ice is significantly darker than fresh snow, as can be observed on any valley glacier during the summer. Note that when modelling glacier ice, an upper limit of $W_{FC} = 0.07$ for $W_{L,u}$ to limit the role of liquid water in reducing the albedo of glacier ice.

The albedo of snow is extremely important in AGCMs since the spatial discontinuity due to the presence or absence of snow greatly affects the overall surface albedo. BEST parameterizes the reduction in the snow albedo as the age of the snow increases and as it slowly turns into ice or accumulates surface pollutants (particularly soot). In BEST the snow albedo is given by

$$\alpha_{sw,n} = 0.85 - 0.20T_m^3 \tag{5.9}$$

$$\alpha_{LW,n} = 0.65 - 0.16T_m^3 \tag{5.10}$$

where $T_m^{\ 3}$ is included under the assumption that the thermal metamorphism of snow (i.e., growth of grains) increases from 0 at 263.16 K to a maximum rate at 273.16 K. In accordance with the observation (Figure 5), T_m takes the form of

$$T_m = \begin{cases} 0.001, & T_u \le 263.16K, \\ 0.1\ (T_u - 263.16), & 263.16K < T_u < 273.16K, \\ 1.0, & T_u \ge 273.16K. \end{cases} \qquad (5.11)$$

The form of T_m and the parameters in Equations (5.9) and (5.10) are chosen to give snow albedos in the shortwave and longwave which decrease smoothly from 0.85 and 0.65 when the snow is significantly below 263K to 0.65 and 0.49 when it is at its melting point. Note that the temperature which appears on the right-hand side of Equation (5.11) is T_u not T_s. The equivalent broadband albedos for snow are 0.75 and 0.57 for 263K and 273K respectively.

Allowing the albedo of snow to depend on temperature requires some explanation. In older AGCMs the albedo of snow was usually a constant, while more recently (e.g. Hansen *et al.*, 1983) was modelled as a function of the "age" of the snow surface, that is, of the time elapsed since the last snowfall, and of the solar zenith angle. Sometimes the depth of the snow pack was allowed to influence its albedo on the ground in that the underlying surface may be "visible" through shallow snow (i.e. fractional cover of snow was assumed). Each of these dependencies is known to be significant, and ideally all of them would be modelled explicitly, as would the observed albedo dependence on grain size and liquid water content. Equations (5.9) and (5.10) capture much of the actual variability of snow albedo. They are supported by observations (e.g. Figure 5), while the physical justification for them is that the rate of metamorphism of the snow pack (including the enlargement of grains and the accumulation of liquid water) increases with temperature. (The physical justification for modelling the aging of the snow is that the amount of metamorphism increases with time.)

A different, practical justification for our model of snow albedo is that we have found it important to model the maturation and ripening of the snow pack in springtime as accurately as possible. This means not only keeping track of its rate of thermal metamorphism as above, but also tracking its rate of dynamic metamorphism, or in other words modelling its density carefully. When we faced the decision of choosing to store either snow density or snow age between AGCM time steps we opted for snow density. Further work will tell whether snow age, and thus an additional quantity of computer memory, is also needed. The model for snow density is described in section 9. It should be noted that although Equations (5.9) to (5.11) appear rather elaborate, there is little justification for including complex canopy or soil models in AGCMs unless the seasonal variation in snow distribution can be represented realistically.

The albedos of the canopy, $\alpha_{SW,c}$ and $\alpha_{LW,c}$ are taken via the aggregation procedure (Section 4.4) directly from Table 2. Hence, the albedo for the whole grid element (i.e. the terrestrial surface, α_t [see layer `t' in Figure 1]) may be obtained by taking the weighted average of the albedos of bare soil, snow and canopy,

$$\alpha_t = (1 - A_v^2) + A_v^2 \alpha_c \tag{5.12}$$

where $A_u = 1 - A_n - A_v$ from Equation (4.1). This albedo parameterization incorporates the concept of self similarity into BEST which explains the need for the A_v^2 term in Equation (5.12).

It should be emphasised that this parameterization of albedo is intentionally relatively simple. For instance, there is no explicit parameterization of the effects of the sun's zenith angle on the albedo (in contrast to BATS, Dickinson *et al.*, 1986) because of the amount of computer time required to handle this properly. BEST also fails to parameterize radiative transfer in the canopy realistically. These two areas may be improved in the next version of BEST, in particular by incorporating a more detailed layer system for the canopy (cf. Sellers *et al.*, 1986). However, although the formulation described here for BEST is simple, it does differentiate between soil, snow and vegetation albedo realistically and is therefore probably adequate for climate modelling.

6.0 Surface roughness length and drag coefficients

6.1 Introduction

Simulations with AGCMs have shown that the climate is sensitive to the parameterization of the surface roughness and the drag coefficient (e.g. Sud *et al.*, 1985). However, as Carson (1987a,b) pointed out, there is no rigorous method for dealing with heterogeneous areas within a grid element to derive a single appropriate estimate of roughness. However, Mason (1988), Andre and Blondin (1986) and Taylor (1987) have discussed and made progress on this issue. Garratt (1977) reviewed the formulation and typical values for drag coefficients which was updated by Carson (1987a) with particular reference to the UK Meteorological Office AGCM. We draw on this and the large quantity of other literature for the parameterization of these quantities in BEST.

6.2 Surface roughness length

A surface roughness length must be defined for each element of the terrestrial surface to describe the efficiency at which surface−atmosphere transfers of energy and momentum take place. We define an effective roughness length which takes account of subgrid scale variability following Taylor (1987), by assuming that the local frictional velocity, u^*, is the same above a patchwork of areas of different roughness lengths. Thus the effective roughness length of the grid element may be written as

$$\ln z_{0t} = (1 - A_v) \ln z_{0u} + A_v \ln z_{0c} \tag{6.1}$$

where z_{0u} = 0.01 m is the bare soil roughness length (Oke, 1978; Stull, 1988), z_{0n} = 0.001 m is the snow roughness length (Oke, 1978, Stull, 1988) and z_{0c} is the canopy roughness length which is the function of the specified ecotype (see Table 2) and the method of aggregation (Section 4.4).

6.3 Drag coefficients

The parameterization of the drag coefficient in numerical models was reviewed by Garratt (1977). In this section we first discuss the parameterization of the neutral drag coefficient in BEST and then relate this quantity to stability through the Richardson number. The formulation described here are taken from McFarlane and Laprise (1985) who developed this parameterization for the Canadian Climate Centre AGCM for which BEST was originally developed. The parameterization described here may not necessarily be ideal for all other climate models although it is probably generally applicable and seems suitable for the BMRC AGCM.

6.3.1 Neutral drag coefficient

If the height of the lowest model level, z_m, in the AGCM is sufficiently close to the ground, i.e., within the constant fluxes layer, the neutral drag coefficient (C_{DN}) for the grid element at this level can be derived from surface layer theory,

$$C_{DN} = \left(\frac{\kappa}{\ln z_m - \ln z_{0t}} \right)^2 \tag{6.2}$$

where κ is the von Karman constant ($\kappa = 0.40$) and z_{0t} was defined in Equation (6.1). The lowest model layer height used in th BMRC AGCM (75m) appears to be low enough for Equation (6.2) to be used. However, a lower layer (1–2m) and several layers within the lowest 100m would probably improve the modelling of surface–atmosphere interactions.

6.3.2 Drag coefficients and Richardson number

The drag coefficient is a function of the neutral drag coefficient and the stability parameter (e.g. the Richardson number), which has been studied extensively in the literature based on the Monin-Obukhov similarity theory (e.g., Clarke, 1970; Businger *et al.*, 1971; Dyer, 1974). In BEST, the formulations of drag coefficients are taken from Boer *et al.* (1984) and McFarlane and Laprise (1985); these authors based their parameterizations on work reviewed by Dyer (1974).

In BEST, the drag coefficient for momentum, C_{Dm}, is different from that for heat, C_{Dh} since it is clear that in unstable conditions C_{Dm} can differ from C_{Dh} by a factor of 3 (Carson, 1987a,b). However, $C_{Dh} \equiv C_{Dv}$ (the drag coefficient for water vapour) since it is not clear how to differentiate between them in AGCMs. In general, the assumption that $C_{Dv} \equiv C_{Dh}$ appears to be reasonable (cf. Carson, 1987a, figure 5).

The drag coefficients for momentum, heat and vapour are defined as follows,

$$C_{Dm} = \begin{cases} C_{DN}\left[1 - \dfrac{8\,R_i}{1 + 56.768\,C_{DN}\,(-R_i/\zeta)^{1/2}}\right] & R_i < 0 \\ C_{DN}\,\dfrac{(1 - 4\,\epsilon R_i)^2}{1 + 8(1 - \epsilon)R_i} & R_i \geq 0 \end{cases} \tag{6.3}$$

$$C_{Dh} = C_{Dv} = \begin{cases} C_{DN}\left[1 - \dfrac{12\,R_i}{1 + 41.801\,C_{DN}\,(-R_i/\zeta)^{1/2}}\right] & R_i < 0 \\ C_{DN}\left[\dfrac{1 - 4\,\epsilon R_i}{1 + (6 - 4\epsilon)R_i}\right]^2 & R_i \geq 0 \end{cases} \tag{6.4}$$

where

$$\zeta = \exp\left(\frac{-\kappa}{\sqrt{C_{DN}}}\right) \tag{6.5}$$

is a normalised mixing length, equivalent to z_m / z_{0t}, and $\epsilon = 0.01$ for land and 1.0 for oceans. The stability of the atmosphere near the surface is described by a bulk Richardson number defined as

$$R_i = -\frac{2g\,z_m}{T_s U_m^2}\left[\,(1 - A_v)T_u + A_v T_c - T_s\right], \qquad R_i \leq 0.25\ /\ (1.0004\epsilon) \tag{6.6}$$

where T_s is the temperature of the air above the surface (assumed to be the lowest model level in the AGCM), T_c is the canopy temperature, U_m is the wind speed at the lowest model level, and g is the acceleration due to gravity. In BEST, U_m is defined as

$$U_m = \max\left(U_{mmin}, \sqrt{U^2 + V^2}\right) \tag{6.7}$$

where $U_{mmin} = 2$ m s^{-1}, U, V are the wind components at the lowest model layer.

It should be noted that the "2" in the right-hand side of Equation (6.6) is added only for the host model in which the temperature is located at half levels while the velocity is located at full levels (Boer *et al.*, 1984 discuss this approximation). If in the host model, all the prognostic variables are distributed at the same levels (e.g., Bourke *et al.*, 1977), the "2" should be omitted.

Figure 6 shows the curve of C_{Dh} against R_i as represented by BEST. It can be seen that C_{Dh} is equal to C_{DN} in the neutral condition ($R_i \equiv 0$). C_{Dh} is greater than C_{DN} in the unstable condition ($R_i < 0$), while C_{Dh} is less than C_{DN} in the stable condition ($R_i > 0$). Note that C_{Dh} is not allowed to drop to zero in order to ensure that there are still weak turbulent heat transfers even under extremely stable conditions. The curves in Figure 6 compare favourably with those suggested by Carson (1987a).

6.4 Summary

The parameterization of the roughness length, stability and drag coefficients is important since these quantities can strongly affect the simulation of the surface–atmosphere interaction. In BEST these quantities are parameterized according to McFarlane and Laprise (1985) and appear to be predicted reasonably (Figure 6). However, the parameterization of these quantities seems to be quite variable between AGCMs and since they form part of the iteration to solve the canopy energy balance, care is required to link BEST into AGCMs which do not use the McFarlane and Laprise (1985) approach. The assumption that $C_{Dh} = C_{Dv}$ (Equation 6.4) is reasonable in that any error which is derived from it must be minor. However, as has been shown, differentiating between C_{Dh} and C_{Dm} is necessary.

7.0 Canopy model

Most of the Earth's land surface is, at least partly, covered by vegetation. Verstraete and Dickinson (1986) indicate that there are four basic reasons why vegetation is important and why it needs to be incorporated into AGCMs. These are: radiative interactions (canopies have low albedos in general); water balance (canopies intercept a large but rather variable fraction of incoming precipitation while transpiration moves large quantities of water from within the soil to the atmosphere); energy balance (the leaves offer a large area in contact with the atmosphere enhancing energy exchange) and roughness (plants are aerodynamically rough and are responsible for much of the variability in the surface roughness).

Although there have been numerous attempts to incorporate the characteristics of vegetation into AGCMs without explicitly modelling the canopy (e.g. Hansen *et al.*, 1983; Warrilow *et al.*, 1986) it seems unlikely that the full effects of the biosphere in the atmosphere could be represented without explicitly incorporating a physical model of the canopy. Further, recent attempts to develop "interactive biosphere models" (e.g. Henderson-Sellers, 1990), although preliminary, indicate that a successful model could be based on the type of scheme discussed here. There seems little long term potential in attempts to represent vegetation in AGCMs by extensions to the bucket type model (e.g. Hansen *et al.*, 1983) when more physically realistic models already exist.

It is therefore apparent that vegetation is an important element in land surface-atmosphere interactions and that there are good reasons to incorporate a parameterization of its effects in AGCMs. The number of reasons is increasing as the demands made on AGCMs increase in terms of what they are required to predict at regional scales. This section describes the parameterization of vegetation incorporated by BEST.

7.1 Leaf and stem area indexes

In BEST, the leaf area index L_{AI} is the projected area of transpiring surface per unit area of vegetated ground. It is a function of ecotype and season. L_{AI} is parameterized as

$$L_{AI} = L_{AImax} - S_l\,[1 - f(T_{soil})] \tag{7.1}$$

where L_{AImax} is the maximum leaf area index, S_l is the seasonal range of the leaf area index (L_{AImax} and S_l are both defined in Table 2). The function $f\,(T_{soil})$ was defined in Equation (4.4).

The stem area index S_{AI} is defined here as the projected area of non−transpiring surface (assumed to include dead vegetation) per unit area of vegetated ground. S_{AI} is a constant for each ecotype (see Table 2). The combined leaf and stem area index is defined below which is required in order to calculate the amount of dew formulation and interception of precipitation on the entire surface of vegetation

$$L_{SAI} = L_{AI} + S_{AI} \tag{7.2}$$

Both L_{AI} and S_{AI} are important for calculating the interception and moisture holding store of the canopy. The S_{AI} basically performs only this role, while the L_{AI} dominates the calculation of transpiration, radiation absorption and canopy temperature. The importance of the L_{AI} tends to outweigh that of the S_{AI} for these reasons, while numerically, Table 2 shows that the S_{AI} is generally small. However, during periods when L_{AI} is reduced due to seasonal factors (i.e. grassland and deciduous forest in winter), the S_{AI} can become important in the canopy water balance.

7.2 Atmospheric resistance and stomatal resistance

In BEST, the parameterization of the atmospheric resistance to surface fluxes, r_a, follows Dickinson (1984) and is based on work by Deardorff (1978), Brutsaert (1982) and Gates (1980). Dickinson (1984) adopted the following formulations for canopy wind speed (U_c) and r_a (both in m s^{-1})

$$U_c = u^* = C_{Dh}^{\ 1/2} U_m \tag{7.3}$$

and

$$r_a^{\,-1} = C_f \left(\frac{U_c}{S_f} \right)^{1/2} \tag{7.4}$$

where C_f is a reference leaf conductance which applies when the canopy wind interacts with a typical leaf for 1 second (set equal to 0.01 $m^{-1/2}$) and S_f = 0.04 m is the characteristic dimension of the leaves in the direction of wind flow (Gates, 1980). A minimum canopy wind speed (U_{cmin}) is specified arbitrarily within BEST as 0.02 m s^{-1} to ensure that there are fluxes from the canopy and to avoid numerical problems.

The stomatal resistance (r_s) is defined as the total mechanical resistance to the transfer of moisture from within to outside a leaf. It is one of the hardest quantities to predict in land surface models since all the models and observational data for it are based on small spatial scales (e.g. experimental plots or laboratory experiments). Although tables of "typical" stomatal resistances are available from the literature (Monteith, 1973; Dickinson *et al.*, 1986) may be generally inappropriate for AGCM grid element scales. However, an estimate for the stomatal resistance is essential for BEST like models although simply differentiating between basic ecotypes (e.g. crop, grassland, forest, shrub etc.) may be sufficient.

In BEST a simple approach is taken in that stomatal resistance is a function of canopy temperature, amount of photosynthetically active radiation absorbed by the canopy, supply of water from the soil via the roots and seasonal factors. Although the dependence of stomatal resistance on vapour pressure deficit is likely to be important (see Farquhar, 1978) it is not clear how to incorporate its effects into BEST. We also ignore ambient CO_2 levels for the same reason which can vary dramatically over actively transpiring vegetation.

The form of Equation (7.5) for r_s follows Deardorff (1978) and Dickinson *et al.* (1986) and is discussed in more detail there. Although its form is based on observed data (e.g. Hinckley *et al.*, 1978; Watts, 1977; Denmead and Millar, 1976) it must be re-iterated that these studies were performed at small scales. Jarvis and McNaughton (1986) indicate that processes which operate at these small scales are unlikely to be the same as those operating at coarser scales and hence the parameterization of stomatal resistance in BEST may be inappropriate. However, we defend Equation (7.5) in two ways. First, it is simple and does not require a large number of ill-defined parameters. Secondly, it is physically and observationally based (if at different spatial scales) and appears to perform well. It is also the only expression available which can be applied to every grid element. Although BEST ignores vapour pressure deficit feedback and ambient CO_2 concentrations, Equation (7.5) is probably good enough, although we intend to extend it to incorporate these two factors.

The stomatal resistance, in BEST, is defined following Deardorff (1978) and Dickinson *et al.* (1986) as

$$r_s = r_{smin} \Big/ \left[0.5 \left(\frac{f_t + r_{smin}/r_{smax}}{1 + f_t} + \frac{f_b + r_{smin}/r_{smax}}{1 + f_b} \right) (f(T_c) + r_{smin}/r_{smax}) \right]$$

$$= r_{smin} f_R f_S, \quad r_{smin} \le r_s \le r_{smax}$$

where r_{smin} (equal to 200 s m^{-1}) and r_{smax} (equal to 5000 s m^{-1}) are the minimum and maximum stomatal resistances, respectively (see Table 2). Note that the contribution of soil water potential to r_s is omitted in Equation (7.5) and will be considered at the end of Subsection 7.5. f_S is a seasonal factor defined as

$$f_S^{-1} = f(T_c) + \left(\frac{r_{smin}}{r_{smax}} \right) \tag{7.6}$$

and f_R is a radiation factor and takes the form

$$f_R^{-1} = 0.5 \left(\frac{f_t + r_{smin}/r_{smax}}{1 + f_t} + \frac{f_b + r_{smin}/r_{smax}}{1 + f_b} \right) \tag{7.7}$$

Note that Equation (7.5) is obtained by substituting (Equation (7.6) and (7.7) into the lower expression of Equation (7.5).

Since different leaves in the canopy receive different amounts of radiation some assumption for radiative transfer within the canopy must be incorporated. BEST incorporates the crude simplification that the canopy is divided into two equal parts following Dickinson *et al.* (1986). The upper parts (0.5 L_{SAI}) receive 75% of the absorbed photosynthetically active radiation K_p^*, while the lower parts (0.5 L_{SAI}) receive 25%. This means that BEST deals with intra-canopy radiative transfer rather simply. It is debatable whether at AGCM grid scales a more complex formulation is feasible, although the parameterization described by Sellers *et al.* (1986) may prove more satisfactory. The fraction of K_p^* absorbed in the upper parts of the canopy, f_t, is

$$\begin{aligned} f_t &= \frac{0.75\ K_p^*\ f_c}{0.5\ L_{SAI}} \\ &= 1.5\ f_c \left(\frac{K_p^*}{L_{SAI}} \right) \end{aligned} \tag{7.8}$$

where f_c is the reciprocal of the visible solar flux for which f_R is double its minimum value, as specified in Table 2. If the incident solar irradiance above the canopy is K_{st}, K_p^* may be written as

$$K_p^* = 0.5\ A_v\ (1 - \alpha_{SW,c})\ K_{st} \tag{7.9}$$

Then, f_b, the fraction of visible radiation absorbed in the lower parts of the canopy, is

$$\begin{aligned} f_b &= \frac{0.25\ K_p^*\ f_c}{0.5\ L_{SAI}} \\ &= 0.5\ f_c \left(\frac{K_p^*}{L_{SAI}} \right) = \frac{1}{3} f_t \end{aligned} \tag{7.10}$$

This parameterization of intra−canopy radiation interception is simple. We aim to improve on the parameterization following Dickinson *et al.* (1986) (by increasing the number of layers from 2 to 4) or Sellers *et al.* (1986) who incorporate a more complex formulation.

7.3 Wet and dry−green fractions of the canopy

Following Deardorff (1978), the wet portion of the canopy surfaces, f_{wet}, is parameterized in BEST as

$$f_{wet} = \left(\frac{D}{D_{\max}} \right)^{2/3} \tag{7.11}$$

where D is the mass of liquid water retained on the foliage per unit ground area and D_{max} (in kg m^{-2}) is the maximum value of D beyond which drip to the ground occurs. Although D_{max} should be a function of wind speed, geometry, leaf angle, plant species, seasonal factors *etc.* this represents too complex a system, hence we use

$$D_{\max} = 0.2\ A_v\ \min(3,\ L_{SAI}) \tag{7.12}$$

This simple parameterization is analogous to the formation of surface runoff in the "bucket" hydrology scheme (Manabe, 1969).

In order to calculate the canopy transpiration rate, the fraction of the canopy surface which is dry and green needs to be defined. Following Deardorff (1978)

$$f_{dry} = (1 - f_{wet})\frac{L_{AI}}{L_{SAI}} \frac{r_a}{r_a + r_s} \tag{7.13}$$

Equation (7.13) shows that the dry and green fraction f_{dry} , and hence, the transpiration, decreases (increases) as the stomatal resistance r_s increases (decreases).

7.4 *Canopy wetness factor*

The wetness factor of the canopy, β_c, is required in order to calculate the suppression of transpiration and the enhancement of the re-evaporation of intercepted precipitation. It is defined following Deardorff (1978) as the ratio of the actual evapotranspiration, E_{ca}, to the potential evapotranspiration E_{caP}, i.e.,

$$\begin{aligned} \beta_c &= \frac{E_{ca}}{E_{caP}} \\ &= 1 - \delta(1 - f_{wet}) + \delta f_{dry} \end{aligned} \tag{7.14}$$

where δ is a step function which is 0 if condensation is occurring onto the canopy surfaces (i.e., if $E_{ca} < 0$) and is 1 otherwise. For ease of discussion, we will only consider $\delta = 1$ in the subsequent sections, except where otherwise stated. Hence, Equation (7.14) becomes

$$\beta_c = f_{wet} + f_{dry} \tag{7.15}$$

7.5 *Canopy fluxes*

The fluxes at the top of the canopy (H_{cs}) are composited from fluxes from the canopy (H_{ca}) and from the underlying ground (H_{ua}) (Deardorff, 1978; Dickinson, 1984). For the sensible heat fluxes,

$$H_{cs} \equiv H_{ca} + H_{ua} \tag{7.16}$$

These three fluxes are defined by standard expressions

$$H_{cs} = \rho_s \, c_{pa} \, c_s \, (T_a - T_s) \tag{7.17}$$

$$H_{ca} = \rho_s \, c_{pa} \, c_c \, (T_c - T_a) \tag{7.18}$$

and

$$H_{ua} = \rho_s \, c_{pa} \, c_u \, (T_u - T_a) \tag{7.19}$$

where T_a is the temperature of the air within the canopy, T_s is the temperature above the canopy (taken as the lowest model level temperature), T_c is the canopy temperature and T_u is the upper soil layer temperature. The three conductances denoted by c are given by

$$c_s = A_v \, C_{Dh} \, U_m \tag{7.20}$$

$$c_c = A_v \, \min(3, \, L_{SAI}) \, r_a^{-1} \tag{7.21}$$

and

$$c_u = C_{Dh}[(1 - A_v)U_m + A_v U_c] \tag{7.22}$$

Note that in the right-hand side of Equation (7.20), it is important to include the vegetation fraction A_v to ensure that equation (7.16) still holds in the case of sparse vegetation fraction (i.e., $A_v \rightarrow 0$). Solving Equations (7.16)-(7.19) for the temperature of the air within the canopy, T_a, we have

$$T_a = \frac{c_s T_s + c_c T_c + c_u T_u}{c_s + c_c + c_u} \tag{7.23}$$

hence

$$T_a = r_h (c_s T_s + c_c T_c + c_u T_u) \tag{7.24}$$

where r_h, the resistance of the canopy, its ambient air and the underlying soil to sensible heat transfer, i.e.

$$r_h = \frac{1}{(c_s + c_c + c_u)} \tag{7.25}$$

Substituting Equation (7.24) into (7.18), the sensible heat flux between the canopy and the air within the canopy, H_{ca} can be rewritten as

$$H_{ca} = \rho_s\, c_{pa}\, c_c\, r_h\, [(c_s + c_u)T_c - (c_s T_s + c_u T_u)] \tag{7.26}$$

Similarly for evaporation, we have

$$E_{cs} \equiv E_{ca} + E_{ua} \tag{7.27}$$

with

$$E_{cs} = \rho_s\, c_s\, (q_a - q_s) \tag{7.28}$$

$$E_{ca} = \rho_s\, \beta_c\, c_c\, (q_c^* - q_a) \tag{7.29}$$

and

$$E_{ua} = \rho_s\, \beta_u\, c_u\, (q_u^* - q_a) \tag{7.30}$$

where q_a is the specific humidity of air within the canopy, q_s is the specific humidity of air above the canopy, q_c^* is the saturated specific humidity of the canopy, and q_u^* is the saturated specific humidity of the ground beneath the canopy.

Solving Equations (7.27)-(7.30) for the specific humidity of the air within the canopy, q_a,

$$\begin{aligned} q_a &= \frac{c_s q_s + \beta_c c_c q_c^* + \beta_u c_u q_u^*}{c_s + \beta_c c_c + \beta_u c_u} \\ &= r_e (c_s q_s + \beta_c c_c q_c^* + \beta_u c_u q_u^*) \end{aligned} \tag{7.31}$$

where the resistance of canopy, its ambient air and the underlying ground to the water vapour transfer is given by

$$r_e = \frac{1}{(c_s + \beta_c c_c + \beta_u c_u)} \tag{7.32}$$

Substituting (7.31) into (7.29), the evaporation rate between the canopy and the air within the canopy, E_{ca}, can then be written as

$$E_{ca} = \rho_s \beta_c c_c r_e [(c_s + \beta_u c_u) q_c^* - (c_s q_s + \beta_u c_u q_u^*)] \tag{7.33}$$

From Equations (7.14) and (7.15), E_{ca} may be written as

$$\begin{aligned} E_{ca} &= \beta_c E_{caP} \\ &= f_{wet} E_{caP} + f_{dry} E_{caP} \\ &= E_{da} + E_{trca} \end{aligned} \tag{7.34}$$

where E_{da} is the evaporation of water intercepted on the surface of the canopy given by

$$E_{da} = f_{wet} E_{caP} \tag{7.35}$$

and E_{trca} is the transpiration through the dry and green surfaces of the canopy, defined as

$$E_{trca} = f_{dry} E_{caP} \tag{7.36}$$

The potential evapotranspiration can be defined as

$$E_{caP} = \rho_s c_c r_e [(c_s + \beta_u c_u) q_c^* - (c_s q_s + \beta_u c_u q_u^*)] \tag{7.37}$$

It should be noted that the evapotranspiration from the vegetation follows the rule of supply and demand (Dickinson, 1984). If $E_{caP} \leq 0$, condensation occurs onto the canopy surface, and transpiration is suppressed and assumed to be zero, thus $f_{wet} = 1$, and $f_{dry} = 0$.

If $E_{caP} > 0$, both evaporation and transpiration occurs from the canopy surface, according to Equation (7.34). The evaporation is limited by the supply of water intercepted on the canopy surface. Thus, Equation (7.11) is true only if $f_{wet} E_{caP} < D / \Delta t$ while if $f_{wet} E_{caP} \geq D / \Delta t$ then $f_{wet} = D / (\Delta t E_{caP})$. The transpiration is limited by the supply of water by the roots. Equation (7.13) is true only when $f_{dry} E_{caP} < E_{trmax}$. If $f_{dry} E_{caP} \geq E_{trmax}$, $f_{dry} = E_{trmax} / E_{caP}$.

E_{trmax} is the maximum possible transpiration, under the present soil moisture distribution, and is

$$E_{trmax} = E_{tr0} A_v (r_{rootu} + r_{rootl}) f(T_{root}) \tag{7.38}$$

where E_{tr0} is the maximum rate of transpiration with saturated soil, total vegetation cover and optimum temperature. Following Dickinson *et al.* (1986) E_{tr0} = 1.8 × 10^{-4} kg m^{-2} s^{-1}. The parameter f is a seasonal factor and was defined in Equation (4.4). The term T_{root}, required in Equation (7.38) defines the temperature of the root zone required for the function defined in Equation (4.4). It is given by

$$T_{root} = 0.5[f_{rootu}T_u + T_l + (1 - f_{rootu})T_b] \tag{7.39}$$

where T_u , T_l , T_b are soil temperatures of upper–, lower–, and bottom–layers respectively. f_{rootu} is a fraction of roots in the upper soil layer (given in Table 2). The quantities r_{rootu} and r_{rootl} are soil–root–xylem dimensionless conductances for the upper and lower soil layers, respectively. r_{rootu} is defined following Dickinson *et al.* (1986) as

$$r_{rootu} = \begin{cases} f_{rootu}\left[1 - \left(\dfrac{W_{wilt}}{W_{L,u}}\right)^B\right] & W_{wilt} < W_{L,u} \\ 0 & W_{wilt} \ge W_{L,u} \end{cases} \tag{7.40}$$

and r_{rootl} is defined as

$$r_{rootl} = \begin{cases} (1 - f_{rootu})\left[1 - \left(\dfrac{W_{wilt}}{W_{L,l}}\right)^B\right] & W_{wilt} < W_{L,l} \\ 0 & W_{wilt} \ge W_{L,l} \end{cases} \tag{7.41}$$

where B is a dimensionless soil diffusivity parameter given in Table 3, and $W_{L,u}$, $W_{L,l}$ are the ratios of the volumetric soil water content to the porosity for the upper and lower soil layers respectively described in detail later.

In order to calculate heat and moisture fluxes the net radiation of the canopy must be calculated. The net radiation absorbed by the canopy, R_c^*, is determined using

$$R_c^* = K_c^* - I_c^* \tag{7.42}$$

where K_c^* is the net absorbed solar radiation by the canopy, and I_c^* is the net emitted longwave radiation by the canopy. K_c^* can be represented as

$$K_c^* = A_v(1 - \alpha_c)K_{st} \tag{7.43}$$

where K_{st} is the incident solar radiation at the terrestrial surface, i.e., just above the top of the canopy, which is predicted by the host AGCM. The canopy albedo, α_c, was given by Equation (5.4). The net emitted longwave radiation I_c^* can be written following Dickinson *et al.* (1986) as

$$\begin{aligned} I_c^* &= A_v(2\sigma T_c^{\,4} - \sigma T_u^{\,4} - I_{st}) \\ &= A_v[2\sigma\,(T_c^{\,4} - T_u^{\,4}) + (\sigma T_u^{\,4} - I_{st})] \\ &= A_v[2\sigma(T_c^{\,4} - T_u^{\,4}) + I_u^*] \end{aligned} \tag{7.44}$$

where I_u^* is the net emitted longwave radiation at the exposed ground surface and is defined as

$$I_u^* = \sigma T_u^{\,4} - I_{st} \tag{7.45}$$

where I_{st} is the incident longwave radiation on the terrestrial surface, predicted by the host AGCM. $\sigma = 5.67 \times 10^{-8}$ W m^{-2} K^{-4} is the Stefan−Boltzmann constant. For simplicity, $(T_u^4 - T_c^4)$ is approximated by $4T_u^3\,(T_u - T_c)$, thus, the net emitted longwave radiation of the canopy, I_c^*, can be defined as

$$I_c^* = A_v[8\sigma T_u^{\,3}(T_c - T_u) + I_u^*] \tag{7.46}$$

Substituting (7.43) and (7.46) into (7.42) leads to

$$R_c^* = A_v(R_{cZ}^* - 8\sigma T_u^{\,3}\,T_c) \tag{7.47}$$

with

$$R_{cZ}^* = (1 - \alpha_c)\,K_{st} - I_u^* + 8\sigma T_u^{\,4} \tag{7.48}$$

which is that part of the canopy net radiation which does not depend on T_c.

7.6 Canopy heat balance and canopy temperature

Given G_{sc} denotes the heat flux into the canopy, the energy balance of the canopy can be written as,

$$\begin{aligned} G_{sc} &= R_c^* - H_{ca} - L_v E_{ca} \\ &= A_v(R_{cZ}^* - 8\sigma T_u^{\,3} T_c) - H_{ca} - L_v(E_{da} + E_{trca}) \end{aligned} \tag{7.49}$$

where L_v is latent heat of evaporation (J kg^{-1}). Note that in deriving Equation (7.49), Equations (7.34) and (7.47) have been used.

If we assume the canopy heat storage is zero, i.e., $G_{sc} = 0$, Equation (7.49) can be solved iteratively for T_c by using the Newton−Raphson method, i.e.,

$$T_c^{n+1} = T_c^n - \frac{G_{sc}(T_c^n)}{\dot{G}_{sc}(T_c^n)} \tag{7.50}$$

where n denotes the n^{th} iteration step. The first guess $T_c^{\,0}$ is taken as T_c of the previous time step which means that the AGCM must save T_c as a global field. $\dot{G}_{sc}(T_c)$ is the derivative of G_{sc} with respect to T_c and is as follows,

$$\dot{G}_{sc}(T_c) = -8A_v\sigma T_u^{\,3} - \frac{dH_{ca}}{dT_c} - L_v\left(\frac{dE_{da}}{dT_c} + \frac{dE_{trca}}{dT_c}\right) \tag{7.51}$$

The formulations for the derivatives dH_{ca} / dT_c, dE_{da} / dT_c and dE_{trca} / dT_c can be seen in the Appendix 2. As shown in the Appendix, dC_{Dh} / dR_i for the stable ($R_i > 0$) is identical to that for the unstable ($R_i < 0$), i.e., $dC_{Dh} / dR_i = -12\ C_{DN}$, when $R_i \to 0$. However, dC_{Dh} / dR_i for the neutral case ($R_i \equiv 0$) is equal to zero. Thus, dC_{Dh} / dR_i is not continuous at the point $R_i = 0$. Experiments have shown that when the lower atmosphere is in transition from a stable state to an unstable state, or vice versa (generally, at dawn or dusk) the Newton−Raphson method described above is very inefficient at finding the desired root. Hence, to ensure the iteration for T_c will converge and retain the efficiency of the iteration, a bi−section method is introduced to replace the Newton−Raphson method when the above mentioned transition from one state to another occurs. In the code, the new estimate of T_c is selected randomly within the range defined by the two most recent estimates of T_c.

It is possible that this computationally expensive iterative procedure could be replaced in BEST by using a heat capacity for the canopy (cf. Sellers *et al.*, 1986). However, it is not clear how to calculate an appropriate heat capacity, while the method described above seems to work consistently well.

7.7 Interception, dew and canopy drip

The parameterization of interception and the subsequent redistribution of water to drip or re−evaporation is one of the principal reasons for incorporation canopies into AGCMs. Intercepted water which fails to drip to the surface re−evaporates rapidly due to the large roughness of canopies, high ventilation and large surface area in contact with the atmosphere. Most precipitation intercepted by the canopy re−evaporates within a few hours. In contrast, precipitation which reaches and infiltrates the soil tends to remain in the soil for much longer periods and is therefore effectively lost to the atmosphere.

Canopy parameterizations, including BEST attempt to represent this fundamental characteristic of vegetation as realistically as possible. It is necessary to calculate the fraction of precipitation intercepted and the fraction which infiltrates into the soil as accurately as possible. There are a variety of models for interception (e.g. Rutter *et al.*, 1972; Rutter, 1975; Massman, 1980) which seem to work reasonably well.

The water storage on the surfaces of the canopy is determined from the balance of intercepted precipitation and the evaporation of the retained water. In BEST the approach is similar to that of Deardorff (1978) and Dickinson *et al.* (1986). The retained water is assumed to be distributed uniformly on the total vegetation surface (i.e., $A_v\ L_{SAI}$) and there is no transpiration through the wetted canopy surface. Hence, the water stored per unit area of ground is,

$$\frac{dD}{dt} = A_v P_{st} - E_{da} - P_{cu} \tag{7.52}$$

and

$$P_{cu} = \begin{cases} 0 & D \le D_{\max} \\ D - D_{\max} & D > D_{\max} \end{cases} \tag{7.53}$$

where D_{max} was defined before and P_{st} is the precipitation rate at the top of the canopy. If $D > D_{max}$, D is set equal to D_{max} and the excess water falls to the surface as canopy drip (P_{cu}), which can be either water (if $T_s \geq 273.16$ K) or snow (if $T_s < 273.16$ K).

This model for interception has been examined elsewhere (Mahfouf and Jacquemin, 1989) and seems to perform well. Other methods (e.g. Sellers *et al.*, 1986) could perform more realistically although it is not clear whether more sophisticated models are appropriate, at this stage, for AGCMs.

Although the canopy model described here appears to be computationally expensive it only requires around 20% of the total time required by BEST. In return BEST provides an improved estimate of surface−atmosphere fluxes and surface temperatures at both the diurnal and seasonal timescale. The incorporation of a canopy also improves the level of realism at which the surface is represented. Overall, the increase in computer time needed to represent the canopy and its effects are minor while it offers the potential for far more realistic impact and climate change experiments.

7.8 Summary

Parameterizing the canopy in AGCMs is particularly difficult. The formulation described here attempts to provide a robust and thoroughly tested methodology which is computationally efficient and undemanding. It is clearly necessary to prescribe a number of parameters and the change in these parameters with respect to time of day or season with relatively sparse data. However, the model described here has been examined carefully and appears to work well and reliably. Ignoring the thermal and hydrological significance of canopies in AGCMs leads to an unrealistic parameterization of land surface–atmosphere interactions. Dickinson *et al.* (1986), Sellers *et al.* (1986) and Cogley *et al.* (1990) have provided canopy parameterizations which can form a component part of AGCMs to improve the physical representation of the land surface.

8.0 Surface fluxes and heat balances

The parameterization of the surface energy balance and the fluxes between the surface and the atmosphere was the basic reason behind incorporating the land surface into AGCMs. The original bucket model combined with a single soil model (Manabe, 1969) provided an adequate model for AGCMs which did not incorporate a diurnal cycle, however, these simple models are inappropriate in modern AGCMs. In this section we describe how BEST accounts for the soil and canopy energy balance. The parameterization described here is more complex than in many land surface models, but irrespective of whether a given AGCM incorporates a canopy, the type of soil model described here is still necessary.

8.1 Ground fluxes

Fluxes from the ground for the grid element as a whole consist of fluxes from the non-vegetated ground and the ground beneath the canopy. Thus,

$$H_{us} = H_{gs} + H_{ua} \tag{8.1}$$

$$E_{us} = E_{gs} + E_{ua} \tag{8.2}$$

where H_{us} is the sensible heat flux from the soil surface to the air above the canopy, H_{gs} is the sensible heat flux from the non-vegetated ground to the air above the soil, H_{ua} is the sensible heat flux from the ground beneath the canopy to the air within the canopy. E_{us} is the latent heat flux from the soil surface to the air above the soil, E_{gs} is the latent heat flux from the non-vegetated ground to the air above the soil and E_{ua} is the latent heat flux from the ground beneath the canopy to the air within the canopy. H_{gs} and E_{gs} can be written as

$$H_{gs} = (1 - A_v)\, \rho_s\, c_{pa}\, U_m\, C_{Dh}\, (T_u - T_s) \tag{8.3}$$

$$E_{gs} = (1 - A_v)\, \rho_s\, U_m\, C_{Dh}\, \beta_u\, (q_u^* - q_s) \tag{8.4}$$

H_{ua} and E_{ua} were defined in Equations (7.19) and (7.30), respectively.

The calculation of β_u is a major problem in solving Equation (8.4). Historically β_u has been defined as a ratio of available soil moisture to some maximum possible soil moisture (e.g "field capacity"). This has worked reasonably but is not a particularly realistic approach since it can not simulate the drying of a soil crust which inhibits evaporation. As noted by Dickinson *et al.* (1986), around noon the top few mm of soil often dry out which inhibits evaporation, irrespective of how wet the soil is beneath. Dickinson *et al.* (1986) incorporate this effect into

BATS by calculating a potential soil evaporation flux according to the mechanical rate at which soil can diffuse towards the surface. BEST takes a similar approach.

The ground wetness, β_u, is parameterized as

$$\beta_u = \begin{cases} 1 & \text{if } A_n = 1 \text{ or } q_u^* \leq q_s \\ A_n + (1 - A_n) \min(1, \beta_g) & \text{otherwise} \end{cases} \tag{8.5}$$

where the fractional extent of snow surfaces (A_n) are assumed to be saturated and β_g is the wetness of non-snow covered soil surfaces given by

$$\beta_g = \frac{W_{F,u} L_v}{L_s} + \frac{E_{usmax}}{E_{usP}} \tag{8.6}$$

where L_v is the latent heat of vaporisation, L_s is the latent heat of sublimation, $W_{F,u}$ is the frozen soil moisture at the upper soil layer, E_{usmax} is the maximum exfiltration rate due to the mechanical processes of soil moisture uptake within the soil (see Eagleson, 1970) discussed in Appendix 3. The term containing $W_{F,u}$ in Equation (8.6) represents the rate at which frozen soil moisture can sublimate, while the remaining term defines the ratio of potential exfiltration to potential evaporation.

E_{usP} is the potential evaporation rate at the soil surface, defined as

$$E_{usP} = \rho_s c_u (q_u^* - q_s) \tag{8.7}$$

and E_{usmax}, the exfiltration rate is given by

$$E_{usmax} = K_{HD} \Theta_u^{0.5B + 2} - K_{H0} \Theta_u^{2B + 3} \tag{8.8}$$

where B is a soil diffusivity parameter (Clapp and Hornberger, 1978) and K_{H0} is the hydraulic conductivity at saturation. Both B and K_{H0} are functions of soil type (see Table 3). K_{HD} is a rate of diffusion from a surface "pond" into dry soil and is shown to be represented as (see Appendix 3 for the derivation)

$$K_{HD} = \left(\frac{-4K_{H0} B \psi_0 \rho_w X_v (1 - W_{F,u})}{\pi \Delta t} \right)^{1/2} \tag{8.9}$$

where ρ_w is the density of water, X_v and t_{ex} were defined before, Ψ_0 is the soil water suction at saturation (−0.2m), defined as $(p - p^*)/\rho_w g$, where p is pressure and p^* is the saturated water vapour pressure.

Θ_u in Equation (8.8) is a wetness factor of upper layer soil and is defined as

$$\Theta_u = \frac{W_{L,u} - 0.01}{1 - W_{F,u}} \tag{8.10}$$

Note that in BEST, $W_{L,u}$ and $W_{L,l}$ are not allowed to drop below 0.01, both because totally dry soils are unphysical at the spatial scales of AGCMs and since totally dry soils lead to numerical difficulties.

8.2 Terrestrial surface fluxes

In the previous sections, it was shown that each land surface grid element in BEST consists of three surface types (snow, vegetation and bare soil). However, the host AGCM only requires the fluxes from the grid element as a whole. The average grid element sensible and latent heat fluxes are given by

$$H_{sm} = H_{gs} + H_{cs} \tag{8.11}$$

$$E_{sm} = E_{gs} + E_{cs} \tag{8.12}$$

where H_{gs} and E_{gs} were defined in Equations (8.3) and (8.4) and H_{cs} and E_{cs} were defined in Equations (7.16) and (7.27), respectively. Comparing Equations (8.11), (8.12) with (8.1) and (8.2), it can be seen that the former have considered the additional fluxes from the canopy, i.e., H_{ca} and E_{ca} respectively. The general philosopy of the accumulation of turbulent energy fluxes from the soil and canopy system is shown scematically (for the sensible heat flux) in Figure 7.

The left-hand side of Equations (8.11) and (8.12) may be written as

$$H_{sm} = \rho_s \, c_{pa} \, C_{Dh} \, U_m \, (T_t - T_s) \tag{8.13}$$

$$E_{sm} = \rho_s \, C_{Dh} \, U_m \, \beta_t \, (q_t^* - q_s) \tag{8.14}$$

where T_t, β_t, q_t^* are the temperature, wetness and saturated mixing ratio of the terrestrial system respectively. Substituting (7.17), (8.3) and (8.11) into (8.13), and through some rearrangements, we obtain

$$T_t = (1 - A_v)T_u + A_v T_a \tag{8.15}$$

where T_a is the air temperature within the canopy given by Equation (7.23). Now let us derive β_t and q_t^*. First E_{cs} from Equation (7.28) is rewritten as

$$\begin{aligned} E_{cs} &= \rho_s\, c_s\, (q_a - q_s) \\ &= \rho_s\, c_s\, \beta_a\, (q_a^* - q_s) \end{aligned} \tag{8.16}$$

Using Equation (7.31) and (7.32), $(q_a - q_s)$ may be written as

$$(q_a - q_s) = r_e\, (\beta_c c_c q_c^* + \beta_u c_u q_u^*) - (q_s - r_e c_s q_s) \tag{8.17}$$

$$= r_e\, (\beta_c c_c q_c^* + \beta_u c_u q_u^*) - r_e(\beta_c c_c + \beta_u c_u) q_s \tag{8.18}$$

$$= r_e(\beta_c c_c + \beta_u c_u) \left(\frac{\beta_c c_c q_c^* + \beta_u c_u q_u^*}{\beta_c c_c + \beta_u c_u} - q_s \right) \tag{8.19}$$

Substituting (8.19) into (7.28) and equating it to (8.16), we obtain

$$\beta_a = r_e\, (\beta_c c_c + \beta_u c_u) \tag{8.20}$$

and

$$q_a^* = \frac{\beta_c c_c q_c^* + \beta_u c_u q_u^*}{\beta_c c_c + \beta_u c_u} \tag{8.21}$$

$$= \frac{r_e\, (\beta_c c_c q_c^* + \beta_u c_u q_u^*)}{\beta_a} \tag{8.22}$$

Using Equations (8.14), (8.4) and (8.16) into (8.12) and with rearrangements, β_t becomes

$$\beta_t = A_v \beta_a + (1 - A_v) \beta_u \tag{8.23}$$

and

$$q_t^* = \frac{A_v \beta_a q_a^* + (1 - A_v) \beta_u q_u^*}{\beta_t} \tag{8.24}$$

The momentum fluxes may be defined as

$$U_{sm} = -\rho_s \, C_{Dm} \, U_m \, u \tag{8.25}$$

and

$$V_{sm} = -\rho_s \, C_{Dm} \, U_m \, v \tag{8.26}$$

where u and v are the wind components at the lowest model level.

8.3 Ground heat balances

The net absorbed radiation R_u^* at the ground surface is

$$R_u^* = A_v \, I_{cu} + (1 - A_v) \, (K_u^* - I_u^*) \tag{8.27}$$

where I_u^* is the net emitted longwave radiation from the exposed ground surface (Equation (7.45)) and I_{cu} is the net longwave radiation from the canopy to the overlying ground which, following Dickinson *et al.* (1986) can be approximated as

$$I_{cu} = \sigma(T_c^4 - T_u^4) \tag{8.28}$$

$$\approx 4 \, \sigma \, T_u^3 \, (T_c - T_u) \tag{8.29}$$

K_u^* is the absorbed solar radiation at the ground and is

$$K_u^* = (1 - \alpha_u) K_{st} \tag{8.30}$$

Hence, given that G_{su} denotes the heat flux into the top soil, then the energy balance of the ground is

$$G_{su} = R_u^* - H_{us} - L_v E_{us} \tag{8.31}$$

8.4 Terrestrial surface radiative fluxes

The net absorbed radiation R_t^* by the terrestrial surface is given

$$R_t^* = K_t^* - I_t^* \tag{8.32}$$

with

$$K_t^* = (1 - \alpha_t) K_{st} \tag{8.33}$$

and

$$I_t^* = \sigma T_t^{\,4} - I_{st} \tag{8.34}$$

The parameterization of the fluxes described in this section are basically the standard formulations. However, the definition of β_u by defining an exfiltration rate is different from most previous methods designed for AGCMs which were statistically or non-physically based (e.g. Manabe, 1969). The parameterization described here has the advantage of being a physical approach based on a solution of the diffusive equation for a semi-infinite medium forced by steady flow at the soil surface (see Eagleson, 1970).

9. Soil and Snowpack

9.1 Rational

The movement of heat and water in solid, liquid and vapour phases within the soil are coupled. A full analysis of this coupled system has never been fully described, but Philip and de Vries (1957) and de Vries (1958) developed a rather complete theory in which reasonable assumptions are made about pore geometry although their theory did not allow for soil ice. For climatological purposes it is reasonable to regard soil ice as immobile, but its interaction with the heat, liquid and vapour within the soil must be treated correctly. An example of a theory for moisture migration in a frozen soil was described by Kay *et al.* (1981). It can also be shown that in many situations the coupling between heat and moisture can be neglected hence with one important exception, moisture transfer can be modelled as depending only on moisture concentration gradients and heat transfer as depending only on temperature gradients. The exception is the transfer of latent heat by advection in the form of vapour, which increases the apparent thermal conductivity of soil air and can enhance heat transfer considerably. These principles are developed further by, for example, de Vries (1963), Eagleson (1970) and Brutsaert (1982).

We adopt a simplified and extended version of the Philip–de Vries theory for heat and water transfer within the soil. The main simplification, apart from uncoupling the heat and moisture transfers, is that we set vapour transfer to zero except at the ground surface, while the main extension is that a simple scheme is introduced to handle soil ice. Snow and glacier ice are also treated explicitly. There are heat and water sources at the surface, internal sinks in the form of latent heat of melting of soil ice and withdrawal of water by the roots of plants. Simple boundary conditions exist at the base of the soil: zero heat flux at a temperature equal to the mean annual surface air temperature and a vadose zone at "field capacity" through which excess water can drain to a water table at indefinite depth. Heat diffuses along the soil temperature gradient, and water follows gradients of gravitational and pressure potential.

The details in calculating soil temperature and soil moisture can be found in the subsequent sections. However, the generalised conservation equations which describe the transfers of soil heat and moisture are by

$$C_v \dot{T} = \nabla G + L_f \Gamma + c_{pw} R \nabla T - \dot{T}_{fict} \tag{9.1}$$

$$\rho_w \dot{X}_w = \nabla (R - (1 - f_i) E) - \Gamma - \Pi \tag{9.2}$$

$$\rho_i \dot{X}_i = -\nabla (f_i E) + \Gamma \tag{9.3}$$

where the overdot denotes differentiation with respect to time and the ∇ differentiation with respect to depth z, which we take to be positive upwards towards an origin at the ground (soil or snow) surface. In Equation (9.1), $\bar{T}_{fict}$ represents a correction for "fictitious" advection which appears to occur because the depths at which T is evaluated are slowly variable with time. Other symbols are ρ_i which is the density of ice (900 kg m^3), ρ_w is the density of water (1000 kg m^{-3}), L_f is the latent heat of fusion (3.33 x 10^5 J kg^{-1}), T is the temperature (K), X_w is the volume fraction of water (m^3 of water m^{-3} of soil), X_i is the volume fraction of ice (m^3 of ice m^{-3} of soil), $f_i = X_i / (X_w + X_i)$ and c_v is a volumetric heat capacity for the soil (J m^{-3} K^{-1}). The internal sinks in Equations (9.1−9.3) are the potential ice production rate

$$\Gamma = \frac{c_v}{L_f} \frac{(T_f - T)}{\Delta t} \tag{9.4}$$

where the freezing temperature $T_f = 273.16$ K, and the rate at which water is removed from the soil by transpiration is given by

$$\Pi = \frac{A_v E_c}{\Delta z} \tag{9.5}$$

where E_c is the transpiration rate which is non−negative. Suitable steps are taken to ensure that actual ice production is zero when no water is available for freezing or when there is no ice is available for melting. Both Γ and Π are volumetric rates expressed in kg m^{-3} s^{-1}. The three flux densities in Equations (9.1)−(9.3) are for heat, in W m^{-2}

$$G = K_T \nabla T \tag{9.6}$$

for water, in kg m^{-2} s^{-1}

$$\begin{aligned} R &= \rho_w (D_H \nabla X_w + K_H) \\ &= \rho_w K_H (1 + \nabla \psi) \end{aligned} \tag{9.7}$$

and for water vapour, in kg m^{-2} s^{-1}

$$E = \begin{cases} 0, & z < 0 \\ E_{us}, & z = 0 \end{cases} \tag{9.8}$$

where E_{us} is the surface evaporation rate defined in Equation (8.2).

In Equation (9.6), K_T is the thermal conductivity calculated with allowance for latent heat transfer, which is discussed in detail in Equations (9.10)−(9.27). In Equation (9.7), D_H is the hydraulic diffusivity (m^2 s^{-1})

$$D_H = K_H \frac{d\psi}{dX_w} \tag{9.9}$$

where K_H is the hydraulic conductivity (m s^{-1}) and ψ is the moisture potential (m). Details of calculations of K_H, ψ and D_H can be found in Appendix 3.

9.2 Thermal properties

9.2.1 Soil thermal conductivity

The thermal conductivity of soil depends on the conductivities of its constituents and their geometry (de Vries, 1963; Cogley, 1987). According to de Vries (1963), the bulk thermal conductivity of the soil is a weighted sum of the component conductivities, each weight being the product of a volume fraction and a "shape factor". Mathematically this may be represented as

$$K_T = \frac{\sum_j (h_j X_j K_{T,j})}{\sum_j (h_j X_j)} \tag{9.10}$$

where j is one of the soil components, h_j is the shape factor, X_j is the volume fraction, $K_{T,j}$ is the thermal conductivity and K_T is the bulk thermal conductivity of the soil.

In BEST, only four soil constituents (air, ice, water and mineral) are explicitly considered to evaluate the bulk soil thermal conductivity. This may be represented by using their volume fractions

$$X_v + X_m = 1 \tag{9.11}$$

$$X_a + X_i + X_w = X_v \tag{9.12}$$

where X_a, X_i, X_w and X_m are the volumetric fractions of air, ice, liquid water and soil minerals respectively. X_v is the volumetric fraction of voids (i.e. the soil porosity). Equation (9.12) may also be expressed as

$$W_a + W_f + W_l = 1 \tag{9.13}$$

where

$$W_a = \frac{X_a}{X_v}, \qquad W_f = \frac{X_i}{X_v}, \qquad W_l = \frac{X_w}{X_i} \tag{9.14}$$

It should be noted that in BEST $W_a = 0.01$ or $X_a = 0.01\ X_v$.

Of the four soil components, the soil mineral has the highest thermal conductivity and dry air the lowest. In order of decreasing conductivity the other components are ice and water (Cogley, 1987). According to Cogley (1987), the common mineral components consist of quartz, clays and humus and although these three mineral components are not considered explicitly in BEST, the composite shape factor and thermal conductivity for the mineral fraction are obtained as weighted sums of the three components. In addition it is necessary to consider water vapour, as its movement through soil pores increases the thermal conductivity because it represents a transfer of latent heat (note that we do not model vapour transfers within the soil column explicitly.

Following de Vries (1963), the soil particles are assumed to be spherical, for the shape factor of component j this leads to

$$h_j = \frac{3}{(K_{T,j}\ /\ K_{T,medium}) + 2} \tag{9.15}$$

where j is one of w, a, i and m. $K_{T,medium}$ is the thermal conductivity of the medium which is water in all but the driest soils, where air becomes the medium. The shape factor of the medium is unity. For example, the geometry factor for air (h_a) is,

$$h_a = \frac{3}{1.75K_{Ta} + 2.0} \tag{9.16}$$

where we have used the water as the medium and the thermal conductivity is 0.57 W m^{-1} K^{-1}. K_{Ta} is the thermal conductivity of air parameterized as

$$K_{Ta} = K_{T,air} + K_{T,vap} \tag{9.17}$$

where $K_{T,air}$ = 0.026 W m^{-1} K^{-1} is the dry air thermal conductivity, $K_{T,vap}$ is the thermal conductivity of the water vapour. Following de Vries and Philip (1986) we assume that the vapour transfer is diffusive, but we linearize their expression for the diffusivity of water in saturated air and make some other simplifying assumptions. The result is

$$K_{T,vap} = \frac{q^*(a_1 + a_2 T)}{T^3} \tag{9.18}$$

where $a_1 = -88.4 \times 10^6$ and $a_2 = 0.698 \times 10^6$. q^* is the saturation specific humidity at soil temperature T obtained using Equations (A27) and (A28) in Appendix 2, with $p = 10^5$ Pa.

As shown in Equation (9.15), the shape factor is dependent upon which medium is assumed. Physically we should expect a much lower bulk conductivity when the medium is air than when it is water. Observations suggest that soil water appears to be continuous at volume fractions above field capacity and to lose its continuity gradually as the moisture content drops from field capacity towards zero. Therefore we take water as the continuous medium at and above field capacity, and air as the continuous medium at complete dryness. Between field capacity and dryness we interpolate linearly using the weight or wetness factor

$$\beta = \frac{W_l + W_f}{W_{FC} + (0.95 - W_{FC})W_f}, \quad 0 \le \beta \le 1 \tag{9.19}$$

where X_{FC} was given in Section 4 and W_{FC} is defined as

$$W_{FC} = \frac{X_{FC}}{X_v} \tag{9.20}$$

The water-medium estimates of thermal conductivity is obtained by substituting the shape factor (Equation (9.15)) of each component for the water medium into Equation (9.10). The result is

$$K_{T,wet} = \frac{(0.57W_l + 1.14W_f + h_a K_{Ta}W_a)X_v + h_{km}X_m}{(W_l + 0.51W_f + h_aW_a)X_v + h_mX_m} \tag{9.21}$$

where h_m is a composite shape factor for the mineral fraction which is derived by Cogley (1987) assuming the mineral fraction (X_m in Equation (9.11)) consists of quartz, clays and humus. h_{km} is a product of h_m and K_{Tmnrl}, a composite conductivity for the mineral fraction which is given by Cogley (1987)

$$h_{km} = 0.98 + 0.64X_m \tag{9.22}$$

$$h_m = 0.65 - 0.44X_m \tag{9.23}$$

Similarly the air-medium estimates of thermal conductivity is

$$K_{T,dry} = \frac{(0.008W_l + 0.076W_f + K_{TA}W_a)X_v + h_{km}X_m}{(0.014W_l + 0.034W_f + W_a)X_v + h_mX_m} \tag{9.24}$$

with

$$h_{km} = 0.073 + 0.005X_m \tag{9.25}$$

$$h_m = 0.060 - 0.030X_m \tag{9.26}$$

Then, the bulk thermal conductivity of the soil ($K_{T,sphere}$) is

$$K_{T,sphere} = \beta K_{T,wet} + (1 - \beta)\, K_{T,dry} \tag{9.27}$$

The weighted estimate $K_{T,sphere}$ is significantly too low when compared to observed thermal conductivities. This is because the soil particles were assumed spherical to derive Equation (9.15). This defect may be corrected by multiplying the estimate by a "constant factor". Good agreement can be obtained with observations if this contact factor is made dependent on the liquid water content. The bulk thermal conductivity K_T for the given layer is therefore

$$K_T = (b_1 + b_2W_l)\, K_{T,sphere} \tag{9.28}$$

where $b_1 = 1.25$, and $b_2 = 0.25$. In summary, our estimate of the bulk thermal conductivity of the soil depends on a knowledge of the porosity and the temperature, as well as of the contents of the soil pores. The porosity is used to infer mineral composition and the field capacity, the field capacity to weight the water-medium and air-medium estimates, and the temperature to calculate the thermal conductivity due to water vapour transfer. Therefore, Equation (9.28) may be expressed as

$$K_T = K_T \, (W_l , W_f , T) \tag{9.29}$$

Although this parameterization of the soil thermal characteristics may appear complex, it is not computationally expensive and is based on well understood theory. It seemed worthwhile to calculate K_T accurately due to its role in the soil temperature calculation.

9.2.2 Soil heat capacity

While the thermal conductivity is a "pseudovector" in that it has relevance only where a temperature gradient exists down which energy must be transferred, the heat capacity of a medium such as a soil is a true scalar quantity. Therefore, the calculation of the soil heat capacity is much simpler than for thermal conductivity. A simple weighted average of the heat capacities of its components is all that is required. We take information for common soil components in Equations (9.11) and (9.12) and write the volumetric heat capacities (in J m^{-3} K^{-1}) of the topsoil and subsoil as

$$c_{v0} = (W_{L,u} \, c_{vw} + W_{F,u} \, c_{vi}) \, X_v + (1 - X_v) \, c_{vm} \tag{9.30}$$

$$c_{v1} = (W_{L,l} \, c_{vw} + W_{F,l} \, c_{vi}) \, X_v + (1 - X_v) \, c_{vm} \tag{9.31}$$

where the subscripts 0 are used for the topsoil and 1 for the subsoil. If the terrain type is glacier, c_{vm} should be replaced by c_{vi} (the values of c_{vw}, c_{vi} and c_{vm} are defined in Table 4.

Table 4. Some constants used in BEST

Symbol	Value	SI Unit			
c_{pa}	1005.0	$J\ kg^{-1}\ K^{-1}$	ρ_i	900.0	$kg\ m^{-3}$
c_{pi}	2090.0	$J\ kg^{-1}\ K^{-1}$	ρ_{nmax}	450.0	$kg\ m^{-3}$
c_{pw}	4187.0	$J\ kg^{-1}\ K^{-1}$	ρ_{nnew}	100.0	$kg\ m^{-3}$
c_{vi}	1.885×10^6	$J\ m^{-3}\ K^{-1}$	ρ_w	1000.0	$kg\ m^{-3}$
c_{vm}	2.380×10^6	$J\ m^{-3}\ K^{-1}$	σ	5.67×10^{-8}	$W\ m^{-2}\ K^{-4}$
c_{vw}	4.180×10^6	$J\ m^{-3}\ K^{-1}$	p	1.0×10^5	Pa
g	9.80616	$m\ s^{-2}$	τ	86400.0	s
L_f	0.333×10^6	$J\ kg^{-1}$	ω	7.2722×10^{-5}	$rad\ s^{-1}$
L_s	2.830×10^6	$J\ kg^{-1}$	π	3.14159	
L_v	2.500×10^6	$J\ kg^{-1}$	2π	6.28318	
R_g	2.8704×10^{-2}	$J\ kg^{-1}\ K^{-1}$	$2^{1/2}$	1.414214	
T_f	273.16	K			

9.2.3 Thermal conductivity of snow and ice

The thermal conductivity of snow depends on its constituents, ice and saturated air. There are theoretical reasons (Schwerdtfeger, 1963) and experimental justification (Mellor, 1977) for allowing it to vary with the square of snow density, and we use the expression

$$K_{Tn} = 2.805\text{x}10^{-6}\ \rho_n^2 \tag{9.32}$$

where ρ_n is snow density evaluated in Equation (9.104). For glacier ice we use a constant thermal conductivity

$$K_{T0} = K_{T1} = 2.0\ \mathrm{W\,m^{-1}\,K^{-1}} \tag{9.33}$$

consistent with measurements summarised by Mellor (1977) and Paterson (1981) but allowing for the fact that real glacier ice has a density somewhat less than that of pure ice.

9.2.4 Snow heat capacity

The volumetric heat capacity of the snow pack (c_{vn}) is given by

$$\begin{aligned} c_{vn} &= c_{pi}\ \rho_n \\ &= X_i\ c_{vi} \end{aligned} \tag{9.34}$$

where c_{pi} and c_{vi} are the specific and volumetric heat capacities of ice (Table 4), respectively. We neglect the contribution of liquid water, if any, to the thermal properties of the snowpack.

9.2.5 Thermal properties of soil/snow layer

The thermal conductivity of upper soil/snow layer (K_{Tu}) may be defined as

$$K_{Tu} = A_n K_{Tn} + (1 - A_n) K_{T0} \tag{9.35}$$

where K_{T0} is obtained from Equation (9.27) for given $W_{L,u}$, $W_{F,u}$ and T_u, and K_{Tn} is the thermal conductivity of snow pack.

The volumetric heat capacity of upper soil/snow layer (c_{vu}) may be written as

$$c_{vu} = A_n c_{vn} + (1 - A_n)\, c_{v0} \tag{9.36}$$

where c_{v0} is the volumetric heat capacity of the upper soil layer.

K_{Tl} is the thermal conductivity of lower soil/snow layer, c_{vl} is the volumetric heat capacity of lower soil/snow layer. To evaluate K_{Tl} and c_{vl} the subgrid scale variation of the depth of the snowpack must be calculated.

The surface of the snowpack can be imagined to make an angle with the horizontal whose tangent is approximately $4d_{neff}$ (when $A_{nu} = 1$), and by Pythagoras we find that the fraction of the l surface lying within the snowpack (A_{ln}) is approximately

$$A_{ln} = \text{clip}\left[\, 0,\ \frac{d_{nmax} - d_0}{4\, d_{neff}},\ 1\, \right] \tag{9.37}$$

and the fraction of the l surface lying above the subsoil (A_{l0}) is approximately

$$A_{l0} = \text{clip}\left[\, 0,\ \frac{d_{nmax} - d_u - d_0}{4\, d_{neff}},\ 1\, \right] \tag{9.38}$$

where

$$d_{nmax} = \sqrt{(8\ d_n\ d_{neff})} \tag{9.39}$$

while the fraction within the subsoil is just $1 - A_{l0}$. The function 'clip' is defined in Equation (9.47). The thermal conductivity of the l surface (K_{tl}) may be defined as

$$K_{Tl} = A_{ln}\, K_{Tn} + (A_{l0} - A_{ln})\, K_{T0} + (1 - A_{l0})\, K_{T1} \tag{9.40}$$

where K_{T1} is obtained from Equation (9.29) for given $W_{L,l}$, $W_{F,l}$ and T_l if $d_0 > d_u$, and $K_{T1} = K_{T0}$ otherwise. Note that for glacier ice, $K_{T1} = 2$ W m^{-1} K^{-1}. The volumetric heat capacity of the l surface (c_{vl}) may be written as

$$c_{vl} = A_{ln}\, c_{vn} + (A_{l0} - A_{ln})\, c_{v0} + (1 - A_{l0})\, c_{v1} \tag{9.41}$$

9.3 Soil and snow layer temperature

The calculation of soil temperature is obviously important: even the earliest climate models solved a soil heat balance to predict a temperature, but modern climate models can now incorporate far more advanced and realistic schemes based, in general, on the "force−restore" model. This model was suggested by Bhumralkar (1975) and Blackadar (1976) and is named due to a tendency of the temperature being given by a difference of two terms, one which describes the diurnal forcing and the other which tends to "restore" it to a deep temperature. Bhumralkar (1975) and Blackadar's (1976) model was taken by Deardorff (1977, 1978) and incorporated into a more general framework appropriate to climate models. From this, many basically similar models have been derived to minimise either numerical error (Jacobsen and Heise, 1982; Warillow *et al.*, 1986) or computer time (Dickinson, 1988; Noilhan and Planton, 1989). BEST rests on the same conceptual ground as the above models, but as is described below the model is derived in a slightly different way.

9.3.1 Introduction

In BEST there are some features regarding the treatment of heat transfer within the soil/snow layers which are new to the modelling of the land surface. These feature are plausible and give reasonable simulations.

1. The scheme incorporated to model soil and snow temperatures is an extended version of the well-known Force-Restore method (Bhumralker, 1975; Blackadar, 1976). However, three adjacent layers are used instead of the more usual two permitting a more realistic representation of the diurnal, seasonal and inter−annual temperature waves.

   The forcing terms used to predict the soil temperature (i.e., the conductive heating rate) depend upon the depth of the soil layer under consideration. The upper soil layer temperature (T_u) is subject to the diurnally conductive forcing, the lower soil layer temperature (T_l) is forced by about a third of that of the conductive forcing for the ground temperature and the bottom level temperature (T_b) is influenced by the seasonal forcing.

In addition to the forcing terms, there are some modifications in the restore terms in the equations for T_u and T_l such that the deep temperature is replaced by the temperature at the next lower level. There are a series of parameters which are used to control the frequency and are tunable (Cogley *et al.* (1990) discuss the derivation of this soil temperature model in full detail).

All other heating sources/sinks are taken into account to modify or correct the temperatures in such a way that they are assumed to be additive to the conductive heating rate. These include the latent heat of fusion of snow and soil ice, sensible heat due to advection at the ground surface due to infiltration rate and snow melt rate, and other heat fluxes which are introduced because of the moving boundary (see below).

2. Using different coordinate systems for calculating heat transfer and water transfer results in some discrepancies in coupling the heat, ice and water stores. For example, the soil moisture calculations involve two layers which have fixed boundaries, while the three thermal layers have a mobile coordinate system which is located at the surface of the snowpack which changes with snowfall and snow melt. Also, the soil depths can change due to the time-dependent properties of the soil medium quantities K_T and c_v.

   The soil and snow pack are considered as a whole to have a temperature at one surface. Thus the extent of snow at each temperature level plays an important role in determining the soil and snow packs thermal properties, the coupling of heat transfer between water and ice stores and the heat stores. These combine and eventually affect the temperature calculations.

3. The feedback between the soil ice or frozen soil water and the temperature is included explicitly in BEST because frozen soil moisture content is a prognostic variable. Phase changes in the soil moisture affects the rate of temperature change in realistic ways.

These three concepts, combined with the description below indicate that the treatment of soil temperature in BEST is reasonably advanced.

9.3.2 Ground surface temperature

The rate of temperature change at the ground surface is given by

$$\frac{dT_u}{dt} = Q_{CNu} - Q_{ADu} + Q_{RIu} + Q_{RSu} \tag{9.42}$$

where Q_{CNu} is the heating rate due to conduction, Q_{ADu} is the sensible heating rate by rainwater advected through the ground surface, Q_{RIu} is the heating rate due to refreezing at level l of snow meltwater produced at level u immediately above (this is part of the sub−grid scale processes of ripening and Q_{RSu} is the heating rate due to melting/freezing of soil ice/water at level u. Q_{CNu} is parameterized as

$$Q_{CNu} = \frac{\omega}{2}\left[\frac{4d_0}{\pi K_{Tu}}G_{su} - (T_u - T_l)\right] \tag{9.43}$$

where G_{su} is the conductive heat flux at the surface and is known from solution of the surface energy balance (see Equation (8.31)), $\omega = 2\pi/\tau$, and $\tau = 86400$ s is the length of day, and

$$d_0 = \frac{\pi}{4}\left(\frac{\tau K_{Tu}}{\pi c_{vu}}\right)^{1/2} \tag{9.44}$$

is the damping depth at which the lower soil/snow pack temperature (T_l) is computed, K_{Tu} is the thermal conductivity of upper soil/snow layer and c_{vu} is the volumetric heat capacity of upper soil/snow layer.

The heating rate due to advection at the ground surface is given by

$$Q_{ADu} = 2c_{pw}\, R_{su}\, \frac{T_u - T_s}{c_{vu}\, d_0} \tag{9.45}$$

where c_{pw} is the specific heat capacity of water and R_{su} is the rainfall infiltration rate and is defined by

$$R_{su} = \text{clip}\left[0,\ a_{dry}\, [K_{H0} - K_{HD}(\Theta_u - 1)],\ \ P_{su}\right] \tag{9.46}$$

where

$$\text{clip}\ [a,\ b,\ c] = \max\ [a,\ \min\ [b,\ c]] \tag{9.47}$$

$$a_{dry} = 1 - W_{L,u} - W_{F,u} \tag{9.48}$$

In springtime, before the snowpack begins to yield meltwater runoff, it must first warm up to the melting point and then fill its void spaces to field capacity with meltwater. This penultimate stage in the life of the snowpack is given the name "ripening". We ignore the ability of snowpack to hold liquid water, but we include in BEST the phenomena which eliminate the cold content of the snowpack. It should be noted that no net meltwater is generated in the early stages of ripening. Energy supplied at the snow surface, mainly by conduction, eventually eliminates the cold content of the uppermost snow. Further supplies of energy generate meltwater, but this meltwater, when it percolates to deeper, colder levels in the snowpack, refreezes and surrenders its latent heat of fusion. This process is
parameterized as

$$Q_{RIu} = f_{n1} \frac{\min\ (0\,,T_f - T_u)}{\Delta t} \tag{9.49}$$

where f_{n1} is the fraction of snow in the upper snow/soil layer, and is written as

$$f_{n1} = \begin{cases} 1 & d_n \geq d_0 \\ d_n \,/\, d_0 & d_n < d_0 \\ 0 & d_n = 0 \end{cases} \tag{9.50}$$

Equation (9.43) never heats level u, rather it simply cools "hot" snow to the freezing point.

A process similar to snowpack ripening goes on in the soil when it is thawing or freezing, although its efficiency will be less because in general the hydraulic conductivity of the soil is less than that of the snowpack. The "ripening" of the soil is represented by the tendency

$$Q_{RSu} = (1 - f_{nl})\, s_1 \frac{T_f - T_u}{\Delta t} \tag{9.51}$$

where s_1 is the efficiency factor which measures the accessibility of the "anomalous" phase (ice above the freezing point, and liquid water below) defined as

$$s_1 = \begin{cases} \dfrac{W_{L,u}}{W_{L,u} + W_{F,u}} & T_u \le T_f \text{ , and } W_{L,u} > 0.01 \\ \dfrac{W_{F,u}}{W_{L,u} + W_{F,u}} & T_u > T_f\text{, and } W_{F,u} > 0 \end{cases} \tag{9.52}$$

Note that if $s_1 \geq 0.2$, s_1 is set to 0.2.

9.3.3 Lower soil/snow pack temperature

The lower soil temperature is parameterized as

$$\frac{dT_l}{dt} = Q_{CNl} + (Q_{RIl} - Q_{RIu}) - (Q_{ADu} + Q_{ADRIl}) + Q_{GMu} + Q_{RSl} - Q_{FICTl}$$

where Q_{GMu} is the soil heating rate due to net production of ice in the layer between level u and l. A negative rate implies net melting. Q_{FICTl} is the apparent heating rate because of the moving boundary. The rest of the terms in the right-hand side of the equation have similar meanings as in previous section. The conductive heating at depth (d_0) is

$$Q_{CNl} = \frac{\omega}{2}\left[\frac{4\, d_0}{\pi K_{Tu}}\, G_{ul} - (T_l - T_b)\right] \tag{9.54}$$

where G_{ul} is the dumped heating rate due to conduction at the level l and is given by

$$\begin{aligned} G_{ul} &= \frac{\exp(-\pi/4)}{\sqrt{2}} G_{su} \\ &\approx 0.32\ G_{su} \end{aligned} \tag{9.55}$$

Q_{RIu} was defined before. Similarly Q_{RIl} is given by

$$Q_{RIl} = f_{n2}\frac{\min(0, T_f - T_l)}{\Delta t} \tag{9.56}$$

where f_{n2} is given in Appendix 4. The latent heat source at level u heats only the base of the u layer. This heating rate is

$$Q_{GMu} = \frac{\Gamma_{01}L_f}{c_{v0}} + \frac{\Gamma_{11}L_f}{c_{v1}} - \frac{\Gamma_{n1}L_f}{c_{vn}} \tag{9.57}$$

where Γ_{01} is the topsoil rate of ice production and is

$$\Gamma_{01} = \frac{f_{u1}\ s_1\ c_{v0}\ (T_f - T_1)}{L_f\ \Delta t} \tag{9.58}$$

with the constrains

$$\Gamma_{01} = -\rho_w\ X_v\ f_{u1}\ W_{F,u}\ /\ \Delta t \qquad \text{if } T_1 > T_f \tag{9.59}$$

$$\Gamma_{01} = -\rho_w\ X_v\ f_{u1}\ (W_{L,u} - 0.01)\ /\ \Delta t \qquad \text{if } T_1 \le T_f \tag{9.60}$$

and Γ_{11} is for subsoil rate of ice production

$$\Gamma_{11} = \frac{f_{l1}\ s_1\ c_{v1}\ (T_f - T_1)}{L_f\ \Delta t} \tag{9.61}$$

with the constraints

$$\Gamma_{11} = -\rho_w\ X_v\ f_{l1}\ W_{F,l}\ /\ \Delta t \qquad \text{if } T_1 > T_f \tag{9.62}$$

$$\Gamma_{11} = -\rho_w\ X_v\ f_{l1}\ (W_{L,l} - 0.01)\ /\ \Delta t \qquad \text{if } T_1 \le T_f \tag{9.63}$$

and Γ_{n1} is the ice production rate per unit of volume for the snowpack defined as

$$\Gamma_{n1} = -\frac{f_{n1}\ c_{vn}\ (T_f - T_1)}{L_f\ \Delta t} \tag{9.64}$$

with

$$0 \leq \Gamma_{n1} \leq \rho_n / \Delta t \tag{9.65}$$

and

$$T_1 = \frac{1}{2}(T_u + T_l) \tag{9.66}$$

The values of f_{u1} and f_{l1} can be found in Appendix 4. Equations (9.59) and (9.62) mean that no more soil water can freeze than is available to be frozen (if $T_1 \leq T_f$), and Equations (9.60) and (9.63) mean no more soil ice can melt than ice available to be melted (if $T_1 > T_f$). Equation (9.65) means no more snow may be melted than exists.

The formulations of Q_{ADu} and Q_{ADRll} need more explanations here. They are introduced in Equation (9.53) for calculating T_l because the origin of the coordinate system is at the surface of the snowpack, which rises and sinks as snow falls and melts. The consequence of this is that the depth at which T_l is evaluated varies slowly with time.

The mobility of the origin is readily accommodated by imagining that snowfall and snow melt give rise to an apparent advective heating rate. Newly-fallen snow has a density of ρ_{new} and arrives at the rate P_{sn}. It therefore translates the origin of coordinates upwards with an apparent velocity ($A_{nu}\, P_{sn} / \rho_{new}$) and appears to heat the level l at a rate of the form

$$Q_{ADu} = \frac{P_{sn}}{\rho_{nnew}} A_n \left(\frac{T_u - T_l}{d_0} \right) \tag{9.67}$$

On the other hand snow melt at the rate ($\Gamma_{n1}\, d_0$) reduces the mass and the depth of the snowpack, so the origin of coordinates also has an apparent velocity (- $A_n\, \Gamma_{n1}\, d_0 / \rho_n$) and heats the level l at a rate of the form

$$Q_{ADRll} = -\frac{\Gamma_{n1}\, d_0}{\rho_n} A_n \left(\frac{T_l - T_b}{d_1} \right) \tag{9.68}$$

where

$$d_1 = \frac{\pi}{4} \sqrt{\frac{365\ \tau\ K_{Tl}}{\pi\ c_{vl}}} \tag{9.69}$$

is the damping depth which originates from the depth d_0 and the bottom soil/snow layer temperature is evaluated at this new level.

The formulation of the ripening of the soil at level l is similar to that at level u, and is given as follows: Q_{RSl} is the ripening of the soil at level l and is defined in the similar way to Q_{RSu},

$$Q_{RSl} = (1 - f_{n2})\, s_2\, \frac{T_f - T_l}{\Delta t} \tag{9.70}$$

where the efficiency factor s_2 is

$$s_2 = \begin{cases} \dfrac{W_{L,l}}{W_{L,l} + W_{F,l}} & T_l \leq T_f \text{ , and } W_{L,l} > 0.01 \\[2ex] \dfrac{W_{F,l}}{W_{L,l} + W_{F,l}} & T_l > T_f\text{, and } W_{L,l} > 0 \end{cases} \tag{9.71}$$

Note that if $s_2 \geq 0.2$, s_2 is set to 0.2.

The apparent heating rate at level l due to changes in d_0 is

$$Q_{FICTl} = \begin{cases} \dfrac{(T_l - T_b)}{d_1} \dfrac{\mathrm{d}\,d_0}{\mathrm{d}\,t} & \text{if } \dfrac{\mathrm{d}\,d_0}{\mathrm{d}\,t} > 0 \\[2ex] \dfrac{(T_u - T_l)}{d_0} \dfrac{\mathrm{d}\,d_0}{\mathrm{d}\,t} & \text{if } \dfrac{\mathrm{d}\,d_0}{\mathrm{d}\,t} \leq 0 \end{cases} \tag{9.72}$$

where the tendency of d_0 is a moderately complicated function of tendencies in the thermal conductivity of soil and snow, the water and ice contents of the soil, and the extent (and therefore the mass and density) of the snow pack. This information can be assembled rather laboriously from the appropriate defining equations.

9.3.4 Bottom soil/snow pack temperatures

The change rate of soil/snow pack temperature at level b is

$$\frac{dT_b}{dt} = Q_{CNb} + (Q_{RIb} - Q_{RIl}) - (Q_{ADl} + Q_{ADRIb}) + Q_{GMl} + Q_{RSb} - Q_{FICTb}$$

where all the terms in the right hand side of the equation have the similar meanings as in previous section. The conductive heating rate at depth ($d_0 + d_1$) is

$$Q_{CNb} = \frac{30\ \omega}{365} \left[\frac{4\ d_1}{\pi\ K_{Tl}}\ G_{lb} - (T_b - \bar{T}) \right] \tag{9.74}$$

where

$$G_{lb} = \frac{\sqrt{2}\ \pi\ K_{Tl}\ \exp(-\pi/4)\ (T_l - \bar{T})}{4\ d_1} \tag{9.75}$$

The latent heat source at level l heats only the base of the l layer. This heating rate is

$$Q_{GMl} = \frac{\Gamma_{02} L_f}{c_{v0}} + \frac{\Gamma_{12} L_f}{c_{vl}} - \frac{\Gamma_{n2} L_f}{c_{vn}} \tag{9.76}$$

where Γ_{02} is the topsoil rate of ice production and is

$$\Gamma_{02} = \frac{f_{u2}\ s_2\ c_{v0}\ (T_f - T_2)}{L_f\ \Delta t} \tag{9.77}$$

with the constrains

$$\Gamma_{02} \leq \rho_w\ X_v\ f_{u2}\ (W_{L,u} - 0.01)\ /\ \Delta t, \quad \text{if} \quad T_2 \leq T_f \tag{9.78}$$

$$\Gamma_{02} \geq -\rho_w\ X_v\ f_{u2}\ W_{F,u}\ /\ \Delta t, \quad \text{if} \quad T_2 > T_f \tag{9.79}$$

and Γ_{12} is for subsoil rate of ice production

$$\Gamma_{12} = \frac{f_{l2}\ s_2\ c_{vl}\ (T_f - T_2)}{L_f\ \Delta t} \tag{9.80}$$

with the constrains

$$\Gamma_{12} \leq \rho_w\ X_v\ f_{u2}\ (W_{L,l} - 0.01)\ /\ \Delta t, \quad \text{if} \quad T_2 \leq T_f \tag{9.81}$$

$$\Gamma_{12} \geq -\rho_w\ X_v\ f_{l2}\ W_{F,l}\ /\ \Delta t, \quad \text{if} \quad T_2 > T_f \tag{9.82}$$

and Γ_{n2} is the ice production rate per unit of volume for the snow pack and is

$$\Gamma_{n2} = -\frac{f_{n2}\, c_{vn}\, (T_f - T_2)}{L_f\, \Delta t} \tag{9.83}$$

with

$$0 \le \Gamma_{n2} \le \rho_n / \Delta t \tag{9.84}$$

and

$$T_2 = \frac{1}{2}\, (T_l + T_b) \tag{9.85}$$

The values of f_{u2}, f_{l2} can be found in Appendix 4. The apparent heating at level b due to the mobility of the origin is

$$Q_{ADl} = \frac{P_{sn}}{\rho_{nnew}}\, A_n\, \frac{T_l - T_b}{d_1} \tag{9.86}$$

Similar to Q_{RIu} and Q_{RIl}, the snow ripening rate at level b is

$$Q_{RIb} = f_{n3} \frac{\min(0\,, T_f - T_b)}{\Delta t} \tag{9.87}$$

The ripening of soil at level b is

$$Q_{RSb} = (1 - f_{n3})\, s_3 \frac{T_f - T_b}{\Delta t} \tag{9.88}$$

where the efficiency factor is

$$s_3 = \begin{cases} \dfrac{W_{L,l}}{W_{L,l} + W_{F,l}} & T_b \le T_f \text{, and } W_{L,l} > 0.01 \\[2ex] \dfrac{W_{F,l}}{W_{L,l} + W_{F,l}} & T_b > T_f\text{, and } W_{F,l} > 0 \end{cases} \tag{9.89}$$

Note that if $s_3 \ge 0.2$, s_3 is set to 0.2. f_{n3} can be found in the Appendix 3.

Similar to Q_{ADRIl}, the apparent heating rate at level b due to snow melt is

$$\Gamma_{12} \geq -\rho_w X_v f_{l2} W_{F,l} / \Delta t \qquad \text{if } T_2 > T_f \tag{9.90}$$

$$\Gamma_{02} \leq -\rho_w X_v f_{l2} (W_{L,l} - 0.01) / \Delta t \qquad \text{if } T_2 \leq T_f \tag{9.91}$$

where d_2 is the depth at which the soil temperature becomes constant and equal to the mean annual surface air temperature, and is specified as

$$d_2 = 112 d_0 + d_n \tag{9.92}$$

The apparent heating rate at level l due to changes in d_0 is

$$Q_{FICTb} = \begin{cases} (365)^{\frac{1}{2}} \dfrac{(T_b - \overline{T})}{d_2 - d_1} \dfrac{\mathrm{d}\, d_0}{\mathrm{d}\, t} & \text{if } \dfrac{\mathrm{d}\, d_0}{\mathrm{d}\, t} > 0 \\ (365)^{\frac{1}{2}} \dfrac{(T_l - T_b)}{d_0} \dfrac{\mathrm{d}\, d_0}{\mathrm{d}\, t} & \text{if } \dfrac{\mathrm{d}\, d_0}{\mathrm{d}\, t} \leq 0 \end{cases} \tag{9.93}$$

and $S_2 \leq 0.2$.

9.4 Snow model

Snow is an important quantity to simulate realistically in AGCMs due to its effect of the surface energy balance, albedo and the soil moisture budget. At high latitudes, snow and snow melt dominate the hydrological regime while the spatial extent of regional snow cover may have considerable influence on climate (Morris, 1985). Unfortunately, models of snow mass and extent incorporated into AGCMs have been, in general, unnecessarily simplistic. Simple snow mass balance models which uniformly cover a grid element above some critical depth are common, while only the effects of snow on albedo are incorporated into some models. While attempts to improve these models have helped (e.g. Hansen *et al.*, 1983; Dickinson *et al.*, 1986), AGCMs have never attempted to incorporate a model of snow dynamics which is comparable to the complexity shown by soil models. BEST attempts to provide a first step towards accounting for the characteristics of snow and its effects on hydrology, energy balance and roughness.

9.4.1 Criteria of snowfall

The precipitation falling at the ground is either snow or rain, depending on whether certain temperature criteria are satisfied. Basically, if the air above the ground falls below the freezing point, precipitation falls as snow. This can be expressed as

$$P_{sn} = \begin{cases} 0 & T_s \geq T_f \\ P_{sg} & T_s < T_f \end{cases} \tag{9.94}$$

and

$$P_{su} = \begin{cases} P_{sg} & T_s \geq T_f \\ 0 & T_s < T_f \end{cases} \tag{9.95}$$

where the precipitation falling at the ground P_{sg} is

$$P_{sg} = P_{cu} + (1 - A_v)\, P_{st} \tag{9.96}$$

BEST can incorporate improved criteria for snow fall should the AGCM be able to provide them.

9.4.2 Snow pack Metamorphism

As observed by, for example, Nishimura and Maeno (1985), the density and hence the depth of a mass of snow changes with time as the cumulative load of overlying snow and air increases. At first the conversion of irregular snowflakes to more rounded grains causes the density to increase rapidly. Later the creep of grains into more compact configurations gives a slower densification rate, and later still sintering and internal deformation of grains become important. Eventually the snow ceases to be an assemblage of ice grains in air, and becomes an assemblage of air bubbles in a continuum of ice. In seasonal snow packs sintering and internal deformation can usually be neglected, but all the other processes deserve consideration. They all occur more rapidly at higher temperatures, and if parts of the snow pack reach the freezing point then regelation – the refreezing of meltwater – may become important as a rapid mechanism of densification.

Kojima (1967) presents a model, calibrated from extensive measurements in Hokkaido, for snow density as a function of depth and time. It requires information on the loading history of the snow pack and relates this history to that of the vertical compressive strain rate through a coefficient of "compactive viscosity", η. Unfortunately the historical dimension, and the fact that it leads to exponential integrals in the solution for density, make this model impractical for large-scale applications but it is useful as a starting point for a simple snow densification model.

Considering only vertical forces, and taking compression to be positive, the Kojima stress – strain-rate relation is

$$\dot{\rho}_n = \rho_n \dot{\epsilon} = \rho_n \sigma_v / \eta \tag{9.97}$$

where ρ_n is snow density (kg m^{-3}), ϵ is strain rate (s^{-1}), σ_v is stress (Pa) and η is the compactive viscosity (η = 8.48 M Pa s at 273 K following Kojima, 1967).

All the variables in Equation (9.97) are functions of depth, time and time since deposition, t_0, yet we require an estimate of the densification rate of the whole snow pack knowing only its current mass N, density ρ_n and temperature T. At times close to t_0 strain rates are very rapid. We choose not to model this transient phase, which means ignoring the chance to model the rapid decrease in the albedo of new snow surfaces. However it does not seem practical to keep track of both these early events and the bulk density of the snow pack with a single time-dependent variable; two AGCM spatial fields would be required if we were to model snow age as well as snow density.

We equate the stress to half of the load ρ_n exerted by the snow pack on its base

$$\sigma_v = \frac{1}{2}\rho_n = \frac{1}{2}gN \tag{9.98}$$

The viscosity depends on both the load and the temperature. Kojima (1967) gives

$$\eta_r = \eta_0 = \exp(\rho_n / \rho_0) \tag{9.99}$$

for the viscosity in the reference temperature range 268–273 K, with ρ_0 about 50 kg m^{-3} and η_0 between about 5 and 14 M Pa s. These numbers describe well an extensive series of measurements in the reference temperature range. The temperature dependence of η is less well known, but Kojima suggests

$$\frac{\Delta(\ln\eta)}{\Delta(1/T)} = k_n \approx 2600 - 4600 \tag{9.100}$$

Taking η_0 = 10 M Pa s, k_n = 4000 m, and a reference temperature T_f = 273.16 K, and substituting (9.99) into (9.100),

$$\ln\eta = \ln\eta_r + k_n\left(\frac{1}{T} - \frac{1}{T_f}\right) \tag{9.101}$$

and

$$\begin{aligned}\eta &= \eta_0 \exp(\rho_n / \rho_0 + k_n / T - k_n / T_f) \\ &= 10^7 \exp(0.02\rho_n + 4000/T - 14.643)\end{aligned} \tag{9.102}$$

Thus the expression for the densification rate of the snow pack due to self-loading is

$$\begin{aligned}\frac{d\rho_n}{dt} &= \rho_n \sigma_v / \eta \\ &= \rho_n g N / 2\eta \\ &= \frac{0.5\,\rho_n g N}{[10^7 \exp(0.02\,\rho_n + 4000/T - 14.643)]}\end{aligned} \tag{9.103}$$

This expression does not include the reduction of bulk density due to the addition of new snow with density assumed to be $\rho_{nnew} = 100$ kg m^{-3}. We assume that this newly-advected density is "mixed" instantaneously with the old density and thus that its effect is additive. The change rate of snow density may be parameterized as

$$\frac{d\rho_n}{dt} = \frac{1}{2}\rho_n\, g\, N\, 1.0\mathrm{x}10^{-7}\exp\left[14.643 - \frac{4000}{\min(T_u, T_f)} - 0.02\rho_n\right] + \frac{P_{sn}(\rho_{nnew} - \rho_n)}{N} \tag{9.104}$$

and the rate of snow mass change with time is

$$\frac{dN}{dt} = P_{sn} - A_n E_{us} - M_{nu} \tag{9.105}$$

with the snow melt rate given by

$$M_{nu} = \Gamma_{n1}\, d_0 + \Gamma_{n2} d_1 + \Gamma_{n3} d_2 \tag{9.106}$$

where Γ_n is the ice production rate in a given soil layer

$$\Gamma_{n3} = - \frac{f_{n3}\ c_{vn}\ (T_f - T_3)}{L_f\ \Delta t} \tag{9.107}$$

with $0 \leq \Gamma_{n3} \leq \rho_n/\Delta t$, and

$$T_3 = \frac{1}{2}(T_b + \bar{T}) \tag{9.108}$$

With the snow melt added to the rainwater infiltration, R_{su} may be rewritten as

$$R_{su} = \text{clip}\ [0,\ a_{dry}[K_{H0} - K_{HD}(\Theta_u - 1)],\ P_{su}] + clip\ [0,\ a_{dry}[K_{H0} - K_{HD}(\Theta_u - 1)] - P_{su},\ M_{nu}] \tag{9.109}$$

Runoff is discussed in more detail in Section 9.5.3.

9.5 Soil moisture calculations

Modelling soil moisture variations in space and time was shown by, for example, Mintz (1984) to be climatically important. Rind (1988) showed that soil moisture anomalies can be propagated through seasons while these are an increasingly large number of AGCM studies which show that soil moisture is important at global, continental and regional scales (e.g, Walker and Rowntree, 1977; Kurbatkin *et al.*, 1979; Rind, 1982; Yeh *et al.*, 1984). It is clear that a land surface scheme must be able to account for diurnal, seasonal and perhaps inter-annual variations in soil moisture. However, as indicated by Wetzel and Chang (1988) regional and even catchment scale variations in soil moisture are extremely important in the surface energy balance. Our methodology for describing the space and time variations in soil moisture are discussed in this section.

Unfortunately, the types of simplifying assumption used for soil temperature (i.e. sinusoidal forcing) are not available for the modelling of soil moisture movements. BEST is therefore based on a conventional finite-difference approximation to predict the soil moisture distribution in the soil. In BEST soil water can move downwards due to gravity, or in either direction due to sorption (the attraction to drier pores of water occupying wetter parts of the soil) depending on the direction of the moisture potential gradient. Water can also be moved out of the soil by evaporation and by root uptake.

The full model of soil hydrology is discussed below, although most of the basic philosophy is omitted to aid brevity and clarity (Cogley *et al.* 1990 discuss this side in more detail).

9.5.1 Upper soil moisture calculations

The rates of liquid, $W_{L,u}$, and frozen, $W_{F,u}$ soil moisture in the upper soil layer can be written as

$$X_v \frac{dW_{L,u}}{dt} = \frac{R_{su} - R_{ul} - (1 - f_i)(1 - A_n)E_{us} - A_v E_{uc} - \Gamma_u}{\rho_w d_u} \tag{9.110}$$

$$X_v \frac{dW_{F,u}}{dt} = \frac{\Gamma_u - f_i\,(1 - A_n)\,E_{us}}{\rho_w\, d_u} \tag{9.111}$$

where

$$f_i = \begin{cases} 1 & W_{F,u} > W_{L,u} \\ 0 & W_{F,u} \le W_{L,u} \end{cases} \tag{9.112}$$

and an ice production rate Γ_u can be defined as

$$\Gamma_u = (\Gamma_{01} + \Gamma_{02} + \Gamma_{03})\; d_0 \tag{9.113}$$

where the terms were calculated in Equations (9.58), (9.77) and similarly for Γ_{03}.

In order to model the movements of water within the soil it is convenient to calculate the hydraulic conductivity (K_H). The capillary rise of soil water, R_{ul}, is written as, according to Darcy's law,

$$R_{ul} = K_{Hu}\left[1 - \left(\frac{\Delta\psi}{\Delta z}\right)_u\right] \tag{9.114}$$

where the hydraulic conductivity in the upper soil layer is expressed following Clapp and Hornberger (1978) as

$$K_{Hu} = K_{H0}\left(\frac{\Theta_u + \Theta_l}{2}\right)^{2B+3} \tag{9.115}$$

where

$$\Theta_u = \frac{W_{L,u} - 0.01}{1 - W_{F,u}} \tag{9.116}$$

$$\Theta_l = \frac{W_{L,l} - 0.01}{1 - W_{F,l}} \tag{9.117}$$

The sign of Equation (9.114) is reversible in that water can flow between the upper and lower soil layers in either direction. The total moisture potential (ψ) represents the combined influence of gravity and water pressure forces. The water pressure force has a variable force in the vertical and under certain circumstances (saturated soil under free drainage with steady infiltration at the surface) water flows downward at a positive velocity K_H. We assume that the material properties of the soil (K_{H0}, B and X_{void}) are a function of texture, but the variation in ψ is independent of soil type. With these simplifying assumptions, the moisture potential gradient can be written as

$$\left(\frac{\Delta\psi}{\Delta z}\right)_u = B\ \psi_0 \left(\frac{\Theta_u + \Theta_l}{2}\right)^{-B-1} \frac{\Theta_u - \Theta_l}{d_u} \tag{9.118}$$

In Equation (9.111), if $W_{F,u} < 1 \text{ x } 10^{-5}$, $W_{F,u}$ is set to 0 in order to prevent the computational expense of keeping account of trivial amounts of frozen soil moisture. If $W_{F,u} > 0.99$ then Equation (9.114) is modified so that the 0.01 fraction of the total soil moisture can remain liquid

$$R_{ul} = K_{Hu}(1 - \left(\frac{\Delta\psi}{\Delta z}\right)_u) + \frac{(W_{F,u} - 0.99)\rho_w\ d_u\ X_v}{\Delta t} \tag{9.119}$$

and then $W_{F,u}$ is set to 0.99. This prevents numerical difficulties which occur if $W_{L,u}$ or $W_{L,l}$ fall to zero. If the available space in the soil for liquid water is exceeded (i.e. if $W_{L,u} > 1 - W_{F,u}$), overflow from top soil, R_{ux}, is simulated

$$R_{ux} = \frac{(W_{L,u} + W_{F,u} - 1)\ \rho_w\ d_u\ X_v}{\Delta t} \tag{9.120}$$

and then $W_{L,u}$ is set to $(1 - W_{F,u})$. Equation (9.120) is purely precautionary since the circumstances for which it applies are extremely rare. Note that for glacier ice,

$$\frac{dW_{F,u}}{dt} = 0 \tag{9.121}$$

9.5.2 Sub-soil moisture calculations

The rate of change of liquid and frozen soil moisture in the lower soil layers may be written as

$$X_v \frac{dW_{L,l}}{dt} = \frac{R_{ul} - R_{lb} - A_v E_{lc} - \Gamma_l}{\rho_w \; d_l} \tag{9.122}$$

$$X_v \frac{dW_{F,l}}{dt} = \frac{\Gamma_l}{\rho_w \; d_l} \tag{9.123}$$

where the ice production rate Γ_l is

$$\Gamma_l = (\Gamma_{11} + \Gamma_{12} + \Gamma_{13}) \; d_1 \tag{9.124}$$

where the terms were defined in Equations (9.61), (9.80) and similarly for Γ_{13}.
and the capillary rise of soil water R_{lb} is written as, (cf. Equation (9.114) according to Darcy's law,

$$R_{lb} = K_{Hl} \; (1 - \left(\frac{\Delta\psi}{\Delta z} \right)_l) \tag{9.125}$$

where

$$K_{Hl} = K_{H0} \; \Theta_l^{2B + 3} \tag{9.126}$$

$$\left(\frac{\Delta\psi}{\Delta z} \right)_u = \begin{cases} B \; \psi_0 \; \Theta_l^{-B-1} \left(\dfrac{\Theta_l - \Theta_b}{d_l} \right) & \Theta_l > \Theta_b \\ 0 & \Theta_l \le \Theta_b \end{cases} \tag{9.127}$$

where

$$\Theta_b = \frac{W_{L,b} - 0.01}{1 - W_{F,b}} \tag{9.128}$$

with

$$W_{L,b} = (1 - f_{il})\, W_{FC}\ , \qquad W_{F,b} = f_{il} W_{FC} \tag{9.129}$$

and

$$f_{il} = \frac{W_{F,l}}{W_{L,l} + W_{F,l}} \tag{9.130}$$

In Equation (9.123), if $W_{F,l} < 1 \times 10^{-5}$, $W_{F,l}$ is set to 0. If $W_{F,l} > 0.99$,

$$R_{lb} = K_{Hl}\,(1 - \left(\frac{\Delta\psi}{\Delta z}\right)_l) + \frac{(W_{F,l} - 0.99)\ \rho_w\ d_l\ X_v}{\Delta t} \tag{9.131}$$

and then $W_{F,l}$ is set to 0.99. If $W_{L,l} > 1 - W_{F,l}$, the overflow from subsoil R_{ux}, is

$$R_{lx} = \frac{(W_{L,l} + W_{F,l} - 1)\rho_w\ d_l\ X_v}{\Delta t} \tag{9.132}$$

and then $W_{L,l}$ is set to $(1 - W_{F,l})$. Note that for glacier ice,

$$\frac{dW_{F,l}}{dt} = 0 \tag{9.133}$$

9.5.3 Runoff

The within soil moisture fluxes R_{ul} and R_{lb} have been discussed above. This section describes the genuine runoff terms. The actual infiltration rate at the surface must be no greater than the total flux of water to the surface (a combination of precipitation, leaf drip and snow melt). Following Eagleson (1970) and Carslaw and Jaeger (1959) a *potential* rate of flow at the surface (positive downwards) is defined as

$$\frac{1}{\rho_w} R^*_{su} = K_{H0} - 2\,(X_v - X_{i,u})\,(\Theta_u - 1)\ \sqrt{D_{H0} / (\pi\,\partial t)} \tag{9.134}$$

where $X_{i,u}$ is the volume fraction of ice in the upper soil layer, Θ_u is the relative concentration of water in the upper soil layer and Equation (9.134) represented the
solution described by Eagleson (1970) integrated over a time step (∂t) and averaged. The saturation hydraulic diffusivity (κ_{H0}) is defined as

$$D_{H0} = -B\ \Psi_0\ K_{H0} / (X_v - X_i) \tag{9.135}$$

Equation (9.134) rests on a solution of the diffusion equation for a semi-infinite medium forced by a steady flow at the soil surface. The solution is reasonable only if infiltrating water travels a short distance in the time span ∂t, compared to the depth of the soil layer (0.1 m). This condition is satisfied for all but the coarsest of the soils incorporated into BEST in their driest states.

Since part of the rational in developing BEST is to parameterize sub-grid scale variability, the actual infiltration rate (R_{su}) must be less than R^*_{su} due to sub-grid scale heterogeneity in the soil wetness. Since the term Θ_u in the calculation of R^*_{su} is an moisture content for the upper soil layer averaged over the entire grid element it is unlikely to be close to saturation. Consequently, R^*_{su} will normally exceed the net flux of water at the surface, hence the soil will soak up all the water. In reality, part of the grid elements surface should be saturated while most of the grid element should be comparatively dry, implying sub-grid scale variation in the infiltration rates. We therefore assume that there is a wet portion of the grid element (a_{wet}) where there is no infiltration, while over the remaining fraction of the grid, a_{dry}, infiltration occurs at the potential rate R^*_{su}.

The terms a_{wet} and a_{dry} are given by

$$\begin{aligned} a_{wet} &= W_{L,u} + W_{F,u} \\ a_{dry} &= 1 - a_{wet} \end{aligned} \tag{9.136}$$

where a_{dry} is the fraction of the grid element where infiltration can occur.

In BEST, the fractions A_{wet} and A_{dry} are only important in the parameterization of infiltration. The grid element is not split geographically into wet and dry sub-elements as this would require prognostic variables and thereby excessive computational expense. Within this context, the equation which describes the actual infiltration rate is

$$R_{su} = \mathrm{clip}\left[0\,,\; A_{dry} R^*_{su}\;,\;\; P_{su} + R_{nu}\right] \tag{9.137}$$

Overland flow R_{sb} is of considerable importance. Although its definition is simple,

$$R_{sb} = P_{su} + M_{nu} - R_{su} \tag{9.138}$$

it represents the only output from BEST which might be validated by routinely observed quantities at temporal and spatial scales appropriate to an AGCM. R_{sb} is lost to the land surface and is not forwarded to another grid element. The total runoff is

$$R_{tb} = R_{sb} + R_{lb} + R_{ux} + R_{lx} \tag{9.139}$$

although as pointed out above, the terms R_{ux} and R_{lx} are likely to always be 0 in an AGCM.

9.5.4 Root withdrawal

The sub-section describes the parameterization of root uptake of water incorporated by BEST. The parameterization of this process is rather important to the overall quality of the simulation of the land surface since when the upper soil layer is dry, the root-xylem-leaf route is the only one available for water to pass from the soil to the atmosphere. It is discussed here as it represents a major mechanism for soil moisture loss. E_{uc} is the withdrawal of water from the top soil layer by the transpiring leaves, and is parameterized as

$$E_{uc} = \frac{r_{rootu}}{\max\left(1\,x\,10^{-12}, r_{rootu} + r_{rootl}\right)} E_{trca} \tag{9.140}$$

where r_{rootu} and r_{rootl} are the soil-root-xylem dimensionless conductances for the upper and lower soil layers respectively. E_{lc} is the withdrawal of water from the lower soil layer by the transpiring leaves and is parameterized as

$$E_{lc} = E_{trca} - E_{uc} \tag{9.141}$$

which is simply the total water taken up by the roots minus that taken up by roots in the upper soil layer.

The calculation of E_{uc} requires a calculation of a root and xylem conductance: in BEST these are combined and depend only on the soil moisture suction potential. The terms r_{rootu} and r_{rootl} are given by Equations (7.40) and (7.41). This parameterization of root conductances follows earlier models described by Deardorff (1978) and Dickinson *et al.* (1986).

10. Implementing BEST into BMRC AGCM

The linking of BEST to the BMRC AGCM requires the removal of the old land surface physics, the embedding of the new code and the setting up of initialisation and diagnostic routines. This section describes these procedures and reference to Figure 2 and Table 1 should prove useful.

10.1 Rational

In order to link BEST to the AGCM, BEST needs the input of a number of atmospheric quantities from the AGCM. These are the incident short wave flux (W m^{-2}),
the downward IR incident upon the terrestrial surface (W m^{-2}), the precipitation rate at the top of the canopy (kg m^{-2} s^{-1}), the northerly and easterly wind components at lowest model level (m s^{-1}), the surface pressure (Pa), the temperature of the air at the lowest model level (K) and the specific humidity of air at the lowest model level (kg kg^{-1}).

Most of these are readily available in BMRC AGCM, but the incident short wave flux and the downward IR incident upon the terrestrial surface (K_{st} and I_{st} respectively) are not directly available. Rather, a combination term (K_{st} (1 - α_t) + I_{st}) is given (where α_t is the surface albedo of the grid square) for their surface layer scheme. Therefore an additional array which contains I_{st} has to be made available from the radiation code. The BMRC AGCM also needs the momentum flux in x- and y− direction from ground and canopy (kg m^{-1} s^{-2}), the evaporative flux from ground and canopy (kg m^{-2} s^{-1}) and the sensible heat flux from ground and canopy (W m^{-2}). These four fluxes are defined in Equations (8.25), (8.26), (8.14) and (8.13) respectively.

10.2 Initialisation of BEST

A series of studies have indicated that the initial values of soil moisture have played an important role in the modelled climate, especially in the "spin−up" period (Walker and Rowntree, 1977; Rind, 1982; Rowntree and Bolton, 1983; Yang, 1989). It is therefore important to initialise the soil moisture as realistically as possible. Unfortunately, these experiments were conducted with relatively simple land surface models which were likely to be less sensitive to the initial conditions that BEST. The initialisation of the top soil layer is unimportant since after a matter of a few days this layer can equilibrate with the atmosphere. The lower soil layer is far more important and should take a season or more to reach an equilibrium with the climate forcing. In terms of soil temperature, the deep soil layer may take years to loose knowledge of its initial condition but has proportionately less influence on the diurnal and seasonal predictions from the AGCM. We intend to re−initialise BEST from the climatology produced by the longest integration performed as soon as possible. In the meantime we use the following initialisation routine to allow that climatology to be produced.

Soil moisture

Following Bourke (1987) and Hart *et al.* (1988), where the soil moisture was routinely initialised with the monthly normal fields of soil moisture generated by Mintz and Serafini (their personal communication), we derive the initial wetness for the upper- and lower-soil layers. This procedure is complicated by the need to initialise both liquid and frozen soil moisture fractions.

In BEST we assume that the mean annual air temperature is a good approximation for the temperature of the soil below the depth that the seasonal temperature wave is felt. If the mean annual air temperature is above the freezing point, the liquid soil wetness in the two soil layers is given by

$$W_{L,u} = \frac{d_{L,u}}{d_{FC,u}} = \frac{d_{L,u}}{W_{FC,u}\, d_u} \tag{10.1}$$

$$W_{L,l} = \frac{d_{L,l}}{d_{FC,l}} = \frac{d_{L,l}}{W_{FC,l}\, d_l} \tag{10.2}$$

and the frozen soil wetness for upper- and lower-soil layers may be assumed to be zero, i.e.,

$$W_{F,u} = W_{F,l} = 0 \tag{10.3}$$

where d_L is the observed soil moisture content (in m) generated by Mintz and Serafini for the upper (u) and lower (l) soil layers, $d_{FC,u} = W_{fc,u}\, d_u$ and $d_{FC,l} = W_{fc,l}\, d_{u,l}$ are the field capacities (in m), and W_{fc} can be specified from the global soil data set provided by Wilson and Henderson-Sellers (1985).

If the mean annual air temperature is below 263K then we assume that all soil moisture is frozen, and that

$$W_{F,u} = 0.75$$
$$W_{F,l} = 0.75 \tag{10.4}$$

and $W_{L,u}$ and $W_{L,l}$ are set to zero. Finally, if the deep soil layer temperature is between 263K and 273K we use

$$W_{L,u} = \frac{1}{2}\left(\frac{d_{L,u}}{d_{FC,u}} = \frac{d_{L,u}}{W_{FC,u}\, d_u}\right)$$
$$W_{L,l} = W_{L,u} = W_{F,u} = W_{F,l} \tag{10.5}$$

which simply splits the available soil moisture equally between liquid and frozen water.

A different approach with regard to the initialisation of soil moisture by Sato *et al.* (1989) when linking their Simple Biosphere (SiB) model to a AGCM may also be employed to compare the results with those from the method proposed above.

Canopy temperature and soil temperatures

Following Hart *et al.* (1988), many of the temperatures initialised within BEST are obtained by interpolating between the lowest model level temperature and the mean annual air temperature. The upper soil layer temperature, T_u and the canopy temperature T_c are therefore initialised as

$$T_u = T_c = 0.5\,(T_s + \bar{T}) \tag{10.6}$$

where $\bar{T}$ is the annual mean air temperature and is supplied by the BMRC AGCM. The remaining temperatures (T_l and T_b) are obtained from

$$T_l = \frac{d_u T_s}{15} + (15 - d_u)\frac{\bar{T}}{15}$$
$$T_b = \frac{(d_u + d_l)\, T_s}{15} + (15 - d_u - d_l)\frac{\bar{T}}{15} \tag{10.7}$$

Snow mass (N), sea ice mass (I), snow density (ρ_n) and dew depth (D)

Sea ice mass (*I*) and snow mass (*N*) are initialised as in the current BMRC AGCM. Snow density (ρn) is initialised according to the depth of the snow mass. If this depth exceeds 500 km m^{-2} we assume the snow is "old" and assign the maximum snow density (450 kg m^{-3}). If the snow depth is less than 200 kg m^{-2} the snow is assumed to be "new" and the minimum snow density is assigned (100 kg m^{-3}). For intermediate cases (when the snow depth is between 200–500 kg m^{-2}) an average snow density (275 kg m^{-3}) is assigned. Although this is not a sophisticated initialisation, the snow density decreases quickly if new snow falls and increases quite rapidly if no further snow falls, hence an incorrect initialisation will not propagate for more than a few weeks. The depth of water intercepted onto the canopy (D) is set to zero but becomes modified to the "correct" value the first time it rains in the model.

10.3 Grid variables and history tape archive

During the time integration of the AGCM, the prognostic variables need to be saved for each time step. In the BMRC AGCM, these variables include the spectral values of the atmospheric dynamic variables such as streamfunction. velocity potential, temperature, humidity and pressure at different levels and the grid point values of surface climatic variables. The common block which carries all the grid variables used in the AGCM for each time step is

```
COMMON /INCGRD/ SVGRID(JF,JG2,JNGRID)
```

where JF is the number of grid points for longitude, JG2 is the number of gaussian latitude points and JNGRID is the number of grid variables. In the BMRC AGCM, there are 10 grid variables which are

Surface air temperature (K),
Accumulated precipitation rate (mm),
Soil moisture content (mm),
Surface albedo,
Snow mass (depth of water equivalent in mm),
Sum of absorbed solar radiation and the incident longwave radiation (W m^{-2}),
Surface emissivity,
Upper layer soil temperature (K),
Lower layer soil temperature (K),
Drag coefficient for momentum fluxes.

When including BEST into the BMRC AGCM 12 other variables must be accounted but as some of them are already available in the current AGCM, only 8 extra variables need to be included in the grid-point value common block in order to be carried between time steps and archived in the history tape file. All 12 variables together with their FORTRAN names are tabulated in Table 5.

Table 5. List of prognostic variables in BEST

Symbol	FORTRAN Name	Quantity	Units

T_u	TEMPU	Upper soil and snow layer temperature	K
T_l	TEMPL	Lower soil and snow layer temperature	K
T_b	TEMPB	Bottom soil and snow layer temperature	K
T_c	TEMPC	Canopy temperature	K
$W_{L,u}$	WLRU	Liquid soil water ratio in the upper soil layer	–
$W_{L,l}$	WLRL	Liquid soil water ratio in the lower soil layer	–
$W_{F,u}$	WFRU	Frozen soil water ratio in the upper soil layer	–
$W_{F,l}$	WFRL	Frozen soil water ratio in the lower soil layer	–
N	SNO	Snow mass	kg m^{-2}
ρ_n	RHON	Density of snow	kg m^{-3}
D	DEW	Water stored on the leaf surface	kg m^{-2}
I	SEAICE	Mass of sea ice	kg m^{-2}

$W_{L,u}$ is assumed to be equivalent to the soil moisture content in the AGCM. Hence, there are 8 extra variables. We decided that all the 12 variables should be stored together following the 6 unchanged variables in the common block /INCGRD/.

11. Summary

This document has described a version of BEST adapted for the BMRC AGCM. It is slightly simplified from the standard version of BEST described by Cogley *et al.* (1990) but is basically the same.

BEST represents a versatile and economical model for representing the land surface in climate models. While we do not claim that BEST is a complete model, it does offer a conceptual advance and, we believe, a considerable improvement in performance over the similar schemes currently incorporated in most models. BEST is an addition to the group of models with the complexity of BATS and SiB although it only requires approximately 60% of the computational time compared to BATS and probably a smaller percentage compared to SiB. BEST requires approximately 5–10% of the total computer time required by the climate model.

The marginal cost of 5–10% seems surprisingly small to represent the surface–atmospheric interactions realistically. Indeed, interactive biosphere models may require much more effort, even approaching the ~40% required to do radiative transfer.

The next stage in the development of BEST is to examine the performance of the model against observational data. This is remarkably difficult to do since so few long term observational data sets at spatial scales comparable to AGCMs exist. However, the large scale field experiments currently being performed or planned do offer hope that reasonable model verification might be achieved in the next few years.

In extensive stand–alone testing BEST has been shown to perform realistically. Extensive sensitivity experiments within an AGCM will help to develop BEST even further.

Acknowledgments

The early development of BEST followed closely from the work by R.E. Dickinson on BATS. Although BEST is now fundamentally different from BATS, we owe a considerable debt to work by R.E. Dickinson over the last decade. We also thank N. McFarlane, M. Lazare and D. Verseghy and from the Canadian Climate Centre. The bulk of the expense in the development of BEST was funded by a Science Subvention from the Atmospheric Environment Service of Environment Canada. Financial and practical support was also provided by the National Engineering Research Council of Canada, Trent University, Liverpool University, Macquarie University and The Australian Research Council.

Appendix 1. Nomenclature

The use of mathematical symbols in this report is mainly based on the standard notations in the micrometeorological text (cf. Oke, 1987). However, as the subscripts are used extensively in this text, their meanings, except those which are self-explained, are described in the following.

A	For the vertical levels, *m* denotes the lowest model level, *s* screen level, *t* terrestrial level, *c* canopy layer, *a* air within canopy, *u* ground level or upper soil layer, *l* lower soil layer, *b* bottom soil layer.
B	For the surface types, *c* or *v* refers to canopy, *i* for sea ice or glacier ice, *n* for snow, *u* for bare soil.
C	For the soil medium, *m* refers to soil minerals, *i* soil ice, *a* air with the soil, *v* soil voids, *w* soil water.
D	For fluxes, the direction is indicated from the first subscript to the second subscript.
E	The capital letters in the subscripts have special meanings. $_{SW}$ and $_{LW}$ refer to the solar wavelength region and near-infrared wavelength region, respectively. $_{L}$ and $_{F}$ refer to the liquid and frozen soil moistures, respectively. $_{FC}$ refers to the field capacity. max and min represent the maximum and minimum values, respectively.

Symbol	FORTRAN name	Quantity (SI Units)
a_c	AC	Partial derivative of c_c wrt C_{Dh} (m s^{-1})
a_{cloud}	ACLOUD	Fractional areal extent of cloud
a_{dry}	ADRY	Fractional extent of soil NOT supporting overland flow during infiltration; equated to X_a / X_v
a_u	AU	Partial derivative of c_u wrt C_{Dh}
α_c	ALBC	Allwave albedo of canopy
α_t	ALBT	Allwave albedo of model region
α_u	ALBU	Allwave albedo of ground
$\alpha_{LW,c}$	ALBLWC	Albedo (canopy) near-IR, direct
$\alpha_{SW,c}$	ALBSWC	Albedo (canopy) visible, direct

$\alpha_{LW,n}$	ALBLWN	Albedo (snow) near−IR, direct
$\alpha_{SW,n}$	ALBSWN	Albedo (snow) visible, direct
$\alpha_{LW,u}$	ALBLWU	Albedo (soil) near−IR, direct
$\alpha_{SW,u}$	ALBSWU	Albedo (soil) visible, direct
L_{AI}	ALEAF	Leaf area index
L_{AImax}	ALFMAX	Maximum leaf area index
a_s	AS	Partial derivative of c_s wrt C_{Dh} (m s^{-1})
A_n	ASNOWU	Snow cover over ground
S_{AI}	ASTEM	Stem area index
β_a	BETAA	Wetness factor for canopied portion of model region
β_c	BETAC	Canopy wetness factor
β_t	BETAT	Terrestrial surface wetness factor
β_u	BETAU	Ground wetness factor
B	BSW	Clapp and Hornberger's "b", a soil diffusivity parameter
c_c	CC	Canopy conductance (m s^{-1})
C_{Dm}	CDM	Surface drag coefficient for momentum
C_{Dh}, C_{Dv}	CDH	Surface drag coefficient for heat and vapour
C_{DN}	CDN	Neutral surface drag coefficient
C_{DON}	CDON	Neutral drag coefficient over the ocean
C_f	CLEAF1	A characteristic value used in Equation (28) (m $s^{-1/2}$)
c_{ieff}	CESICE	Effective heat capacity of sea ice (J m^{-2} K^{-1})
c_{lr}	SOLOUR	Soil colour factor
c_u	CU	Soil conductance (m s^{-1})
r_a^{-1}	CLEAF	Leaf aerodynamic conductance (m s^{-1})
r_s^{-1}	CPORE	Stomatal conductance (m s^{-1})
c_p	CPAIR	Specific heat capacity of dry air (J kg^{-1} K^{-1})
c_{pi}	CPICE	Specific heat capacity of ice (J kg^{-1} K^{-1})
c_{pw}	CPWTR	Specific heat capacity of water (J kg^{-1} K^{-1})
r_{smin}^{-1}	CPMAX	Maximum stomatal (canopy) conductances (m s^{-1})
r_{smax}^{-1}	CPMIN	Minimum stomatal (canopy) conductances (m s^{-1})
C_s	CS	Air conductance (m s^{-1})
C_{vi}	CVICE	Volumetric heat capacity of ice (J m^{-3} K^{-1})
C_{vn}, C_{vu}, C_{vl}	CVN,U,L	Soil or snow volumetric heat capacity (J m^{-3} K^{-1})
C_{vw}	CVWTR	Volumetric heat capacity of water (J m^{-3} K^{-1})
C_{vm}	CVMNRL	Volumetric heat capacity of soil minerals (J m^{-3} K^{-1})
dA_n	DANUDT	Rate of change of A_n (s^{-1})
Δt	DELT	Model time step (s)
d_u, d_l	DEPTHU,L	Upper and lower soil depths (m)
d_n	DEPTHN	Snow depth averaged over entire grid cell (m)
D	DEW	Water stored on the leaf surface (kg m^{-2})

D_{max}	DEWMAX	Capacity per unit area of leaf water store (kg m^{-2})
D_{neff}	DNEFF	Effective thickness of snow for heat transfer (m)
d_{nmax}	DNMAX	Maximum snow depth (m)
d_0	D0	Thickness of upper layer (m)
d_1, d_2	D1,D2	Thicknesses of lower and bottom layers (m)
ϵ	EPS	1.00 for ocean, otherwise 0.01; used to adjust dependence of C_{dh} on R_i
E_{da}	EVAPDA	Canopy evaporation (not transpiration) (kg m^{-2} s^{-1})
E_{ua}	EVAPUA	Evaporation from soil+snow beneath canopy (kg m^{-2} s^{-1})
E_{us}	EVAPUS	Evaporation from soil+snow (kg m^{-2} s^{-1})
E_{uc}	EVAPUC	Transpiration from upper soil via canopy (kg m^{-2} s^{-1})
E_{lc}	EVAPLC	Transpiration from lower soil via canopy (kg m^{-2} s^{-1})
E_{sm}	EVAPSM	Evaporative flux from ground plus canopy (kg m^{-2} s^{-1})
E_{caP}	EVPTDA	Potential evaporation from a wet canopy (kg m^{-2} s^{-1})
E_{usP}	EVPTUS	Potential evaporation from the soil (kg m^{-2} s^{-1})
E_{usmax}	EVUSMX	Maximum exfiltration from saturated soil (kg m^{-2} s^{-1})
f_b,f_t	FB,FT	Fractional intercepted photosynthetically active radiation per unit (leaf & stem) area in the lower, upper canopies
f_{ka},f_{kv}	FKA,V	Thermal conductivities of moist air
f_{kair}	FKAIR	Thermal conductivities of dry and moisted air
f_c	FC	Reciprocal of that photosynthetically−active flux for (m^{2}W^{-1}) which f_R is about double its minimum value
f_{dry}	FDRY	Fraction of canopy that is green and dry
f_i	FICE	Fraction of soil water which is frozen
f_k	FK	Thermal conductivities of soil components (Wm^{-1}K^{-1})
I_u^*	FLNTU	Net longwave flux from the ground surface (W m^{-2})
I_{cu}	FLWVCU	Long wave flux between canopy and ground (W m^{-2})
I_{st}	FLWVST	Downwelling IR incident upon the terrestrial surface (W m^{-2}
K^*_p	FPARC	Net photosynthetically active radiation (W m^{-2})
f_{rootu}	FROOTU	Fraction of roots in upper layer
K_c^*	FSNTC	Shortwave absorbed by canopy (W m^{-2})
K_u^*	FSNTU	Shortwave absorbed by ground (W m^{-2})
K_t^*	FSNTT	Shortwave absorbed by terrestrial surface (W m^{-2})
K_{st}	FSWVST	Incident short wave flux (W m^{-2})
A_v	FVEG	Fractional shading of soil beneath canopy
f_{wet}	FWET	Fraction of canopy covered by water
f_R		Radiation factor in calculating r_s
f_S		Seasonal factor in calculating r_s
R_g	GASK	Gas constant of dry air (J kg^{-1} K^{-1})
g_c	GC	Ground cover index; −1 for land

Γ_u	GAMMAU	Soil ice production rate (kg m^{-3} s^{-1})
Γ_l	GAMMAL	Soil ice production rate (kg m^{-3} s^{-1})
Γ_n	GAMMAN	
Γ_0	GAMMA0	
Γ_1	GAMMA1	Ice production rates for current (kg m^{-3} s^{-1}) heat balance layer
f_n, f_u, f_l	GN,U,L	Fraction of snow, topsoil or subsoil in current heat balance layer
g	GRAV	Acceleration due to gravity (m s^{-2})
G_{lb}	HEATLB	Heat conducted across L/B interface (W m^{-2})
G_{sc}	HEATAC	Heat conducted from air into canopy (W m^{-2})
G_{su}	HEATSU	Heat conducted from air into soil or snow (W m^{-2})
G_{ul}	HEATUL	Heat conducted across U/L interface (W m^{-2})
h_a	HA	Geometry factor h for air fraction of soil
h_m	HMNRL	Geometry factor h for mineral fraction of soil
L_f	HTFU	Latent heat of fusion (J kg^{-1})
L_s	HTSB	Latent heat of sublimation (J kg^{-1})
L_v	HTVP	Latent heat of vapourization (J kg^{-1})
L_{SAI}		Leaf and stem area index (= LAI + SAI)
K_h	HYCON	Hydraulic conductivity (contains a factor of ρ_w (kg m^{-2} s^{-1})
K_{H0}	HYCON1	Hydraulic conductivity at saturation (kg m^{-2} s^{-1})
K_{HD}	HYDIF1	Rate of diffusion from surface "pond" into dry soil (kg m^{-2} s^{-1})
ω	OMEGA	$2 \pi / \tau$ (rad s^{-1})
τ	DAYLEN	Length of day (s)
ψ_0	PSI1	Soil water suction at saturation (m)
P_{cu}	PRCPCU	Leaf drip rate (kg m^{-2} s^{-1})
P_{st}	PRCPST	Precip. rate at the top of the canopy (kg m^{-2} s^{-1})
P_{su}	PRCPSU	Rain fall rate at the ground (kg m^{-2} s^{-1})
P_{sn}	PRCPSN	Snow fall rate at the ground (kg m^{-2} s^{-1})
q_a	QA	Specific humidity of air within the canopy (kg kg^{-1})
q_s	QS	Specific humidity of air at the lowest model level (kg kg^{-1})
q_a^*	QSTARA	Flux-weighted saturated specific humidity (kg kg^{-1}) for canopied portion of model region
q_c^*	QSTARC	Saturated specific humidity of canopy (kg kg^{-1})
q_u^*	QSTARU	Saturated specific humidity of ground (kg kg^{-1})
q_t^*	QSTART	Saturated specific humidity of system (kg kg^{-1})
r_e	RE	System resistance for vapour (s m^{-1})
r_h	RH	System resistance for heat (s m^{-1})
r_{smin}, r_{smax}		Minimum and maximum stomatal resistance (s m^{-1})
ρ_{nmax}	RHNMAX	density of old snow (kg m^{-3})
ρ_{nnew}	RHNMAX	Densities of snow (kg m^{-3})
ρ_i	RHOI	Density of ice (kg m^{-3})

ρ_n	RHON	Density of snow (kg m^{-3})
ρ_s	RHOS	Density of surface air (kg m^{-3})
ρ_w	RHOW	Density of water (kg m^{-3})
RH	RHUM	Relative humidity
R_i	RIB	Surface bulk Richardson number
R_c^*	RNETC	Net radiation of canopy (W m^{-2})
R_u^*	RNETU	Net radiation of soil (W m^{-2})
R_t^*	RNETT	Net radiation of terrestrial system (W m^{-2})
R_{cZ}^*	RNETCZ	Part of canopy net radiation independent of T_c (W m^{-2})
R_{lb}	RNOFLB	Runoff from lower soil (kg m^{-2} s^{-1})
R_{lx}	RNOFLX	Overflow from subsoil (kg m^{-2} s^{-1})
M_{nu}	RNOFNU	Snow melt rate (kg m^{-2} s^{-1})
R_{sb}	RNOFSB	Overland flow (kg m^{-2} s^{-1})
R_{su}	RNOFSU	Input of rain and meltwater to upper soil (kg m^{-2} s^{-1})
R_{tb}	RNOFTB	Total runoff (kg m^{-2} s^{-1})
R_{ub}	RNOFUB	Runoff from upper soil (kg m^{-2} s^{-1})
R_{ul}	RNOFUL	Drainage (capillary rise) from upper to lower soil (kg m^{-2} s^{-1})
R_{ux}	RNOFUX	Overflow from topsoil (kg m^{-2} s^{-1})
z_{0c}	ROUGHC	Canopy roughness length (m)
I	SEAICE	Mass of sea ice (kg m^{-2})
S_c	SEASC	Seasonal range of A_v
S_l	SEASL	Seasonal range of L_{AI}
H_{ca}	SENSCA	Sensible heat flux, canopy to air (W m^{-2})
H_{ua}	SENSUA	Sensible heat flux from soil+snow beneath canopy (W m^{-2})
H_{us}	SENSUS	Sensible heat flux, soil+snow to screen (W m^{-2})
H_{sm}	SENSSM	Sensible heat flux, screen to lowest model level (W m^{-2})
S_f	SIZLEF	Leaf dimension, used to calculate aero-dynamic (m) resistance of canopy
N	SNO	Snow mass (kg m^{-2})
P	SPRESS	Surface pressure (Pa)
t_{ex}	ITEX	Texture index of soil
σ	STEFK	Stefan-Boltzmann constant (W m^{-2} K^{-4})
$\bar{T}$	TBAR	Mean annual air (and therefore deep soil) temperature (K)
T_a	TEMPA	Flux-weighted average for canopy of T_c, T_u, and T_s (K) (Equation (43))
T_c	TEMPC	Temperature of canopy (K)
T_s	TEMPS	Temperature of the air at the lowest model level (K)
T_t	TEMPT	Temperature of the system (flux-weighted average of T_a and T_u) (K)
T_u	TEMPU	Soil temperatures for U layer (K)

T_l	TEMPL	Soil temperatures for L layer (K)
T_b	TEMPB	Soil temperatures for B layer (K)
T_f	TFREZ	Freezing point of water (K)
K_{Tu}	THCONU	Thermal conductivity of soil interfaces (W m^{-1} K^{-1})
K_{Tl}	THCONL	Thermal conductivity of soil interfaces (W m^{-1} K^{-1})
K_{Tb}	THCONB	Thermal conductivity of soil interfaces (W m^{-1} K^{-1})
K_{Tn}	THCONN	Thermal conductivity of snow (W m^{-1} K^{-1})
K_{T0}	THCON0	Thermal conductivity of topsoil (W m^{-1} K^{-1})
K_{T1}	THCON1	Thermal conductivity of subsoil (W m^{-1} K^{-1})
K_{TD}	THDIFU	Thermal diffusivity ($m^{-2}s^{-1}$)
E_{trca}	TRANCA	Transpiration rate (kg m^{-2} s^{-1})
E_{trmax}	TRCAMX	Maximum possible transpiration,given the (kg $m^{-2}s^{-1}$) soil moisture distribution
E_{tr0}	TRNSMX	Maximum rate of transpiration with saturated soil, (kg $m^{-2}s^{-1}$) total vegetation cover and optimum temperature
u , v	U,V	Wind velocities at lowest model level (m s^{-1})
U_{ms}	UMOMMS	Momentum fluxes (kg m^{-1} s^{-2})
V_{ms}	VMOMMS	Momentum fluxes (kg m^{-1} s^{-2})
U_c	UC	Canopy wind speed (m s^{-1})
U_m	UM	Wind speed at lowest model level (m s^{-1})
U_{cmin}	UC	Intercanopy wind speed (m s^{-1})
U_{mmin}	UMMIN	Minimum ("ventilation") windspeeds (m s^{-1})
κ	VKC	von Karman constant
Θ	VL	$X_w / (X_v - X_i) = W_l / (1- W_f)$
Θ_u, Θ_l, Θ_b	VLU,L,B	THETA for U, L and B layers
W_a	WA	$1- W_l - W_f =$ air−filled fraction of voids
W_{FC}	WFCAP	X_{FC} / X_v
r_{rootu} , r_{rootl}	WILTL,U	Upper− and lower−soil dimensionless root conductances
$W_{F,u}$, $W_{F,l}$	WFRU,L	Ratio of soil ice to saturation (= X_i / X_v)
$W_{L,u}$, $W_{L,l}$	WLRU,L	Ratio of soil water to saturation (= X_w / X_v)
W_{wilt}	WWILT	X_{wilt} / X_v
X_{FC}	XFCAP	Fractional moisture at field capacity
X_m	XMNRL	$1 - X_v$
X_v	XVOID	Soil porosity
X_{wilt}	XWILT	Volumetric soil moisture at which permanent wilting occurs
X_a	XAIR	Volumetric soil air content (= m^3 of soil air / m^3 of soil)
X_i	XICE	Volumetric soil ice content (= m^3 of soil ice / m^3 of soil)
X_w	XWTR	Volumetric soil water content (= m^3 of soil water / m^3 of soil)
z_m	ZM	Height of lowest model level for wind (m)
ζ	ZETA	Normalized mixing height, equal to z_m / z_{0t}

z_0	ZOUGHT	Grid−cell roughness length (m)
z_{0n}	ZOUGHN	Snow surface roughness length (m)
z_{0u}	ZOUGHU	Bare soil surface roughness length (m)

Appendix 2. Derivations of $\frac{dH_{ca}}{dT_c}$, $\frac{dE_{da}}{dT_c}$ and $\frac{dE_{trca}}{dT_c}$

A2.1. Derivation of $\frac{dH_{ca}}{dT_c}$

From Equation (7.26), H_{ca} may be written as

$$H_{ca} = [\rho_s c_{pa} r_{hc} (c_s + c_u) T_c] - [\rho_s c_{pa} r_{hc} c_s T_s] - [\rho_s c_{pa} r_{hc} c_u T_u]$$

$$= A + B + C \tag{A1}$$

where

$$r_{hc} = r_h c_c \tag{A2}$$

According to the chain rule, dH_{ca} / dT_c may be written as

$$\frac{dH_{ca}}{dT_c} = \left(\frac{dH_{ca}}{dT_c} \right) + \frac{dH_{ca}}{dC_{Dh}} \frac{dC_{Dh}}{dT_c} \tag{A3}$$

with

$$\frac{dH_{ca}}{dC_{Dh}} = \left(\frac{dH_{ca}}{dC_{Dh}} \right)_A + \left(\frac{dH_{ca}}{dC_{Dh}} \right)_B + \left(\frac{dH_{ca}}{dC_{Dh}} \right)_C \tag{A4}$$

Then $\frac{dH_{ca}}{dT_c}$, $\left(\frac{dH_{ca}}{dC_{Dh}} \right)_A$, $\left(\frac{dH_{ca}}{dC_{Dh}} \right)_B$ and $\left(\frac{dH_{ca}}{dC_{Dh}} \right)_C$ can be written as

$$\left(\frac{dH_{ca}}{dT_c} \right) = \rho_s c_{pa} r_{hc} (c_s + c_u) \tag{A5}$$

$$\left(\frac{dH_{ca}}{dC_{Dh}} \right)_A = \rho_s c_{pa} [r_{hc} (a_s + a_u) + DHC (c_s + c_u)] T_c \tag{A6}$$

$$\left(\frac{dH_{ca}}{dC_{Dh}}\right)_B = -\rho_s\, c_{pa}\,[r_{hc}\, a_s + DHC\, c_s\,]\, T_s \tag{A7}$$

$$\left(\frac{dH_{ca}}{dC_{Dh}}\right)_C = -\rho_s\, c_{pa}\,[r_{hc}\, a_u + DHC\, c_u\,]\, T_u \tag{A8}$$

where a_s, a_c, a_u and DHC are defined as,

$$a_s = \frac{dc_s}{dC_{Dh}} = \frac{c_s}{C_{Dh}} \tag{A9}$$

$$a_c = \frac{dc_c}{dC_{Dh}} = \frac{c_c}{4\, C_{Dh}} \tag{A10}$$

$$a_u = \frac{dc_u}{dC_{Dh}} = \frac{c_c}{C_{Dh}} + \frac{1}{2} A_v\, U_c \tag{A11}$$

$$DHC = \frac{dr_{hc}}{dC_{Dh}} = r_h[a_c - r_{hc}\, (a_s + a_c + a_u)] \tag{A12}$$

where Equations (7.20)−(7.22), (7.25) and (A2) have been used.

The remaining term in (A3) is dC_{Dh} / dT_c, which can be written as

$$\frac{dC_{Dh}}{dT_c} = \frac{dC_{Dh}}{dR_i} \frac{dR_i}{dT_c} \qquad \textbf{(A13)}$$

According to the definition of R_i in Equation (6.9), dR_i / dT_c is

$$\frac{dR_i}{dT_c} = -A_v R_{i0} \qquad \textbf{(A14)}$$

with

$$R_{i0} = \frac{2gz_m}{T_s U_m^2} \qquad \textbf{(A15)}$$

dC_{Dh} / dR_i can be obtained from Equation (6.7).

When $R_i < 0$,

$$\frac{dC_{Dh}}{dR_i} = -12\, C_{DN} \frac{1 + 0.5R_{i1}}{(1 + R_{i1})^2} \qquad \textbf{(A16)}$$

where

$$R_{i1} = 41.801\, C_{DN} (-R_i / \zeta)^{1/2} \qquad \textbf{(A17)}$$

when $R_i > 0$,

$$\frac{dC_{Dh}}{dR_i} = -2C_{Dh}\left[\frac{4\epsilon}{1 - 4\epsilon R_i} + \frac{6 - 4\epsilon}{1 + (6 - 4\epsilon)R_i}\right] \qquad \textbf{(A18)}$$

when $R_i = 0$,

$$\frac{dC_{Dh}}{dR_i} = 0 \qquad \textbf{(A19)}$$

A2.2. Derivations of dE_{da} / dT_c and dE_{trca} / dT_c

From Equation (7.35), dE_{da} / dT_c is

$$\frac{dE_{da}}{dT_c} = f_{wet} \frac{dE_{caP}}{dT_c} \tag{A20}$$

from Equation (7.36), dE_{trca} / dT_c is

$$\frac{dE_{trca}}{dT_c} = \frac{df_{dry}}{dT_c} E_{caP} + f_{dry} \frac{dE_{caP}}{dT_c} \tag{A21}$$

where

$$\frac{df_{dry}}{dT_c} = \frac{df_{dry}}{dC_{Dh}} \frac{dC_{Dh}}{dT_c} \tag{A22a}$$

with

$$\frac{df_{dry}}{dC_{Dh}} = -\frac{f_{dry}}{4\,C_{Dh}}\frac{r_s}{r_a + r_s} \qquad \textbf{(A22b)}$$

Now the remaining task is to find dE_{caP} / dT_c. According to Equation (7.37), E_{caP} may be written as

$$\begin{aligned} E_{caP} &= [\rho_s r_e c_c (c_s + \beta_u c_u) q_c^*] - [\rho_s r_e c_c c_s q_s] - [\rho_s r_e c_c \beta_u c_u q_u^*] \\ &= A + B + C \end{aligned} \qquad \textbf{(A23)}$$

According to the chain rule, dE_{caP} / dT_c may be written as

$$\frac{E_{caP}}{dT_c} = \left[\frac{dE_{caP}}{dT_c}\right] + \frac{dE_{caP}}{dC_{Dh}}\frac{dC_{Dh}}{dT_c} \qquad \textbf{(A24)}$$

with

$$\frac{dE_{caP}}{dC_{Dh}} = \left(\frac{dE_{caP}}{dC_{Dh}}\right)_A + \left(\frac{dE_{caP}}{dC_{Dh}}\right)_B + \left(\frac{dE_{caP}}{dC_{Dh}}\right)_C \qquad \textbf{(A25)}$$

Then $\frac{dE_{caP}}{dT_c}$, $\left(\frac{dE_{caP}}{dC_{Dh}}\right)_A$, $\left(\frac{dE_{caP}}{dC_{Dh}}\right)_B$ and $\left(\frac{dE_{caP}}{dC_{Dh}}\right)_C$ can be written as

$$\begin{aligned} \left[\frac{dE_{caP}}{dT_c}\right] &= \frac{dE_{caP}}{dq_c^*}\frac{dq_c^*}{dT_c} \\ &= \rho_s r_e c_c (c_s + \beta_u c_u)\frac{dq_c^*}{dT_c} \end{aligned} \qquad \textbf{(A26)}$$

The expression q_c^*, as for all saturated specific humidities, is calculated using

$$q_* = \frac{0.622\, p_*}{p - 0.378\, p_*}$$

with p_* calculated from Tetens' formula

$$p_* = 611 \exp\left[\frac{A\,(T - T_f)}{T - B}\right] \tag{A28}$$

where

$$A = \begin{cases} 21.874 & T \le T_f \\ 17.269 & T > T_f \end{cases}$$

and

$$B = \begin{cases} 7.66 & T \le T_f \\ 35.86 & T > T_f \end{cases}$$

where $T_f = 273.16$. Hence,

$$\frac{dp_*}{dT} = \frac{A(T_f - B)}{(T - B)^2}\, p_*$$

$$\begin{aligned} \frac{dq_c^*}{dT_c} &= \frac{0.622\, p}{(p - 0.378\, p_*)^2}\, \frac{dp_*}{dT_c} \\ &= \frac{p\, q_c^*}{p - 0.378\, p_*}\, \frac{A\,(T_f - B)}{(T_c - B)^2} \end{aligned} \tag{A29}$$

$$\left[\frac{dE_{caP}}{dC_{Dh}}\right]_A = \rho_s\,[a_c\, r_e\, (c_s + \beta_u\, c_u) + c_c\,(DES + DEU)]\; q_c^* \tag{A30}$$

$$\left[\frac{dE_{caP}}{dC_{Dh}}\right]_B = -\rho_s\,[a_c\, r_e\, c_s + c_c\, DES]\; q_s \tag{A31}$$

$$\left[\frac{dE_{caP}}{dC_{Dh}}\right]_C = -\rho_s \, [a_c \; r_e \; \beta_u \, c_u \; + \; c_c \; DEU \;] \; q_u^* \tag{A32}$$

with

$$DES = \frac{d(r_e \, c_s)}{dC_{Dh}} = r_e \, a_s + c_s \, DRE \tag{A33}$$

$$DEU = \frac{d(r_e \, \beta_u \, c_u)}{dC_{Dh}} = (r_e \, a_u + c_u \, DRE) \; \beta_u \tag{A34}$$

where

$$\begin{aligned} DRE &= \frac{dr_e}{dC_{Dh}} = -r_e^2 \left[a_s + a_c \, \beta_c + c_c \, \frac{d\beta_c}{dC_{Dh}} + \beta_u \, c_u \right] \\ &= -r_e^2 \, [a_s + (\beta_c + DFD4CD) \, a_c + a_u \, \beta_u] \end{aligned} \tag{A35}$$

Note in (A35),

$$\frac{d\beta_c}{dC_{Dh}} = \frac{df_{dry}}{dC_{Dh}} = -\frac{f_{dry}}{4\,C_{Dh}}\,\frac{r_s}{r_a + r_s}$$

and

$$DFD4CD = 4\,C_{Dh}\,\frac{d\beta_c}{dC_{Dh}} = -f_{dry}\,(1 - c_{dry})$$

with

$$c_{dry} = \frac{r_a}{r_a + r_s}$$

Appendix 3. Derivation of exfiltration rate

According to Eagleson (1970), the exfiltration rate E_{usmax} is

$$E_{usmax} = (X_{w1} - X_{w0})\ (D_H / \pi t)^{1/2} \qquad \textbf{(A36)}$$

where X_{w1} is the initial volumetric water content within the soil layer, X_{w0} is the initial volumetric water content at the surface. D_H is the hydraulic diffusivity and is

$$\begin{aligned} D_H &= K_H \frac{d\psi}{dX_w} \\ &= -B\, K_H\, \psi / X_w \\ &= -B\, K_{H0}\, \psi_0\, \Theta^{B+3} / X_w \end{aligned} \qquad \textbf{(A37)}$$

where we have used

$$\psi = \psi_0\, \Theta^{-B} \qquad \textbf{(A38)}$$

and

$$\frac{d\psi}{dX_w} = -B\, \psi / X_w$$

and

$$K_H = K_{H0}\, \Theta^{2B+3} \qquad \textbf{(A39)}$$

Using (A37), (A38) and (A39) into (A36), we obtain

$$\begin{aligned} E_{usmax} &= \left(\frac{(X_{w1} - X_{w0})^2}{X_w} \ \frac{-B\, K_{H0}\, \psi_0}{\pi\, \Delta t} \right)^{1/2} \Theta^{0.5B + 1.5} \\ &= K_{HD} \Theta^{0.5B + 1.5} \end{aligned} \qquad \textbf{(A40)}$$

where K_{HD} is

$$K_{HD} = \left(\frac{(X_{w1} - X_{w0})^2}{X_w} \ \frac{-B\, K_{H0}\psi_0}{\pi \Delta t} \right)^{1/2} \qquad \textbf{(A41)}$$

Appendix 4 Preliminary results from BEST and the BMRC AGCM

A4.1 Introduction

In determining the success of incorporating BEST into the BMRC AGCM two principal sets of results need to be examined. The established technique of analysing global maps, upper atmospheric statistics and the large scale flows is crucial to the general assessment of whether incorporating BEST improves the climate simulation. However, BEST is a land surface model and therefore the simulation of the characteristics of a variety of land surface types is also important. Linking BEST to the AGCM will modify the surface–atmosphere interaction quite substantially. The canopy increases the heterogeneity of the surface and increases the speed at which the temperatures and fluxes between the surface and the atmosphere can change. The result is a surface which is dynamic and can show large changes between individual time steps. The first objective is therefore to simulate the surface realistically. The second objective is to ensure that the lowest model layer of the AGCM can cope with the new land surface scheme, and finally, to ensure that the global and continental scale simulations are improved.

In this section, a small set of results is presented, which show how well BEST is performing. The global results will be presented when a long enough simulation has been performed to generate meaningful statistics. The results described here are for a series of grid squares for January. Northern and Southern hemisphere grid points are shown in order to show both summer and winter simulations. In each case only three days are shown, although the full month of January has been examined to ensure that the following results are typical. These results are only intended to show that BEST is performing realistically in terms of how the model deals with precipitation and radiation, and that it predicts energy fluxes and temperatures consistent with the quantities provided by the AGCM.

Simulations for five ecotypes are shown in the remainder of this appendix. The location of these ecotypes are listed in Table A4.1 together with the latitude and longitude of the grid squares and their approximate geographical location.

Table A4.1 List of the location and broad ecotype description of the selected grid squares

Ecotype name	Latitude and longitude	Approximate location
Tropical forest	2°S 62°W 2°N 56°W	Central Amazonia Northern Amazonia
Coniferous forest	59°N 56°E 59°N 51°E	near Perm, USSR near Kirov, USSR
Temperate grassland	49°N 39°E 49°N 45°E	near Char'kov, USSR near Volgograd, USSR
Australian desert	24°S 129°E 27°S 129°E	Gibson desert, W.A Great Victoria desert, S.A.
Tundra	65°N 101°W 65°N 96°W	N.W. of Baker Lake, Canada N.W. of Baker Lake, Canada

A4.2 Results

Figures A4.1–A4.10 show a series of simulations for the ecotypes described above. In all cases the x–axis is for three diurnal cycles representing 180 AGCM time steps. It should be noted that although these simulations are broadly based on the named ecotype the characteristics of the surface parameters at the locations listed above are not usual for the given ecotype. For instance, Table A4.2 shows the "typical" roughness length and albedo for the grid square classification in addition to the albedo and roughness length used by the AGCM in the simulations described here.

The roughness lengths are typically much lower in the simulations described here because the aggregation procedure (Section 2) means that there is no assumption that each grid element is homogeneously covered by a particular ecotype. Hence, in the tropical forest simulations, a roughness length of 0.5m is used rather than a more typical value of 1m since the grid element is partly covered by grassland and partly by tropical forest. Similarly, the albedo is increased from the very low tropical forest values due to the presence of significant amounts of more reflective grassland.

Table A4.2 "Typical" albedo and roughness values (from Table 2) and albedo and roughness values from "actual" grid points which show the effect of the aggregation procedure.

Ecotype name	Latitude and longitude	"Typical albedo" (%)	"Actual" albedo (%)	"Typical" roughness length (m)	"Actual" roughness length (m)
Tropical forest	2°S 62°W	14	17	1	0.5
	2°N 56°W	14	17	1	0.5
Coniferous forest	59°N 56°E	14	52–53	1	0.055*
	59°N 51°E	14	50–52	1	0.055*
Temperate grassland	49°N 39°E	20	42–48	0.1	0.015*
	49°N 45°E	20	38–48	0.1	0.02*
Tundra	65°N 101°W	19	71	0.06	0.004*
	65°N 96°W	19	71	0.06	0.004*
Australian desert	24°S 129°E	35	34	0.01	0.015
	27°S 129°E	35	34	0.01	0.015

* Indicates that the grid square is snow affected and that the albedo and roughness lengths are therefore modified.

A4.3 Tropical forest

The two locations shown in Figures A4.1 and A4.2 represent quite different climatic regimes during January. The first simulation (Figure A4.1a) shows an environment where the available energy is dominated by the sensible heat flux with relatively small latent heat fluxes. The temperatures (Figure 4.1c) are dominated by a large diurnal range in the canopy temperature which reaches 312K. However, the fact that the terrestrial temperature is only about 1K warmer than the upper soil layer temperature indicates that there is little vegetation over this grid square which is surprising considering the geographical location of the grid element. The terrestrial temperature only exceeds the air temperature by 5K which seems realistic. Figure A4.1c shows the soil hydrology. The soil is rather dry and β is very small during the hours of daylight which explains the low latent heat fluxes. Figure A4.1d shows that two small precipitation occur during this three day period, but since both fall during the night they do not lead to significant pulses in the latent heat flux.

Figure A4.2 shows a very different hydrological regime. Figure A4.2a shows a land surface dominated by large latent heat fluxes and small sensible heat fluxes (basically the reverse of the previous example). This leads to a reduction in the maximum canopy temperatures reached (304K) (Figure A4.2b). Figure A4.2c shows the explanation for this pattern. The soil is very much wetter (60%) and hence β is much higher supporting the bigger latent heat flux. Finally Figure A4.2.d shows the moisture fluxes. The evaporation dominates this figure, with only a couple of small showers. Gravitational drainage is significant throughout the three day period since the soil is above its field capacity.

The pattern of β in Figure A4.2c is rather interesting and deserves some further explanation. The decrease in β after sunrise is due to the evaporation of water off the canopy and in the first few mm of the soil, leading to a general drying of the surface. As the sun sets moisture recharge from within the soil leads to an increase in β. The second pulse in the wetness near sunset is probably due to moisture condensation onto the canopy. This rapidly evaporates leading to a small increase in evaporation shown by Figure A4.2a. Finally the surface dries until recharge and further condensation re−saturate the surface.

A4.4 Coniferous forest

Figure A4.3 shows a simulation for a coniferous forest located at 59ºN 56ºE. Figure A4.3a shows the energy fluxes and, in particular, that there is very little solar radiation absorbed by the surface. The panel shows three diurnal cycles with maximum solar radiation reaching only about 50 W m^{-2}. The sensible heat flux is small but positive (warming the surface) and the latent heat flux is small and negative (drying the surface). Figure A4.3b shows the temperatures for this grid element. For the first few hours the soil and terrestrial temperatures are much colder than the air temperature but they subsequently warm and closely match the air temperature. The exception is the lower soil layer temperature which is rather warmer than the other temperatures. Figure A4.3c shows the moisture regime. Deep snow covers the grid element and the soil contains quite a lot of moisture although all soil moisture is frozen. β shows little variability and is typically around 80%. Since, for all snow surfaces β = 100% the fact that β < 100 indicated that the coniferous forest is not entirely snow covered and reduces the overall wetness of the grid element which is quite realistic. Finally, Figure A4.3d shows the moisture fluxes. Two precipitation events occur (both are snow fall leading to slight increases in the snow depth) but evaporation is rather suppressed and runoff is zero as expected.

The second coniferous forest location (59ºN 51ºE) shows a more confused but basically similar pattern. The energy fluxes (Figure A4.4a) show the same pattern as in the previous example although the sensible heat is negative for the first and third day. Figure A4.4b shows that the temperatures are quite similar in magnitude to the other coniferous forest grid square. Snow again covers much of the grid square (Figure A4.4c) and increases through the three day period. The soil is completely frozen and β is higher than the previous example at 90%. Finally Figure A4.4d shows the moisture fluxes. Evaporation is negligible until after the precipitation event when small fluxes occur for about a day.

A4.5 Temperature grassland

The selected grassland grid elements are 10º south of the coniferous forest locations which is clearly shown by Figure A4.5a which is for a grassland located at 49ºN 39ºE. The absorbed solar radiation is much higher on days 2 and 3 compared to the coniferous forest which leads to a more interesting pattern of turbulent fluxes. The latent heat flux is consistently negative and around 50–100 W m^{-2}. The sensible heat flux is more variable. It is only about 20 W m^{-2} on the first day, but increases to about 60 W m^{-2} on day 2 and 90 W m^{-2} on the third day where it exceeds the latent heat flux. This flux pattern is supported by the temperatures (Figure A4.5b) which shows that the temperature difference between the terrestrial temperature and the air temperature gradually increases over these three days. In addition β slowly declines from around 85% to 60% (Figure A4.5c) over this period until the final day. The soil is quite wet and most of the soil moisture is not frozen. A small amount of snow exists and snow depth increases slightly due to precipitation (Figure A4.5d). This precipitation also leads to a small pulse in evaporation which reduces β. On the final day both turbulent energy fluxes fall to zero just after noon which permits β to increase back to 100%.

Figure A4.6 shows a second grassland grid element located at 49°N 45°E. The energy fluxes (Figure A4.6a) and temperatures (A4.6b) are quite similar to the previous example although the sensible heat flux is slightly larger. The soil is quite wet (Figure A4.6c) which leads to significant runoff (Figure A4.6d). The runoff rate slowly dries the soil but is also partly due to very slow snow melt. The pattern of β and evaporation are closely correlated with β showing a sudden increase between days two and three due to the sudden reduction in evaporation. Two snowfalls help to maintain overall high β values.

A4.6 Tundra

The grid squares chosen to represent tundra are very far north and are therefore subject to large amounts of snow and very little solar radiation. The first tundra grid element (65°N 101°W) shown in Figure A4.7 receives less than 10 W m^{-2} of absorbed solar radiation (Figure A4.7a) and negligible turbulent energy fluxes. The heat flux into the soil (G) is quite large (25–50 W m^{-2}). The temperatures are quite low, the air temperature (Figure A4.7b) is relatively constant (257K) but the terrestrial temperature falls to 238K on the third day, mostly due to the fall in the upper soil layer temperature to 240K, but small amounts of canopy still exposed fall to 230K which forces the terrestrial temperature to cool to below the upper soil layer temperature. Figure A4.7c shows the moisture regime for this grid square. Deep snow almost completely covers the grid square and thus β is almost 100% at all times. All available soil moisture is frozen. Figure A4.7d shows the moisture fluxes and illustrates that precipitation occurs for most of the three day period leading to a steady increase in the snow depth. Evaporation is zero for all but a short period on the first day while runoff is always zero.

The second tundra grid square is located at 65°N 96°W. The patterns shown are almost identical to the previous tundra grid square with very small amounts of solar radiation and turbulent energy fluxes, cold temperatures and large amounts of snow. These simulations are not illustrated due to their similarity with Figure A4.7.

A4.7 Australian desert

Figure A4.9 shows the simulation for a desert located at 24ºS 129ºE. Figure A4.9a shows the energy fluxes which, over the full diurnal cycle, are dominated by the sensible heat flux. The latent heat is small and consistent at about 25 W m^{-2}. The sensible heat flux is only about 150 W m^{-2} on the first day due to the relatively low amounts of solar radiation, but on days 2 and 3 the sensible heat flux reaches 300 W m^{-2} and 350 W m^{-2}. The soil heat flux shows short intense pulses on each day which reach a maximum of 400 W m^{-2} on day 3 for a few time steps. These pulses occur before the sensible heat flux becomes established and are coincident with the sudden and artificial rise in the solar radiation due to 3 hour time steps used to calculate radiation in the AGCM. BEST needs a few time steps to adjust to the sudden 500 W m^{-2} increase in solar radiation.

Figure A4.8b shows that canopy temperatures in the desert reach extremely high values (>330K). However, there is virtually no vegetation present hence the terrestrial temperature only reaches 310K. The predicted air temperatures are quite low for a desert since they only reach 300K on the second day. Figure A4.8c shows that there is virtually no moisture available. The soil is virtually dry and β is only significantly above zero at night and rapidly falls as the sun rises. The very low holding capacity of the desert soil leads to small amounts of gravitational drainage (Figure A4.8d) even though the soil is very dry. The second desert grid element selected (27ºS 129ºE) shows a very similar pattern and is therefore not included here.

A4.8 Summary and discussion

BEST has been shown to perform coherently within the BMRC AGCM for a number of selected grid elements. Although the period of simulation presented was only three days, these three days come at the end of a 240 day simulation and are therefore presumably independent of the initial conditions. The true test of BEST will be in its long term performance and its ability to predict realistic seasonal and regional variability. This section has not attempted to demonstrate that this is the case, rather it has concentrated on the simulation by BEST of the diurnal cycle.

The examination of the sensitivity of BEST to different climate forcing in a "stand-alone" mode has been performed over the last few years. The exercise discussed above demonstrates to the authors that BEST performs within the AGCM in similar ways to its performance in these off-line tests. These results are non-conclusive in terms of whether BEST produces the right answer in all circumstances, but they do show that the model is performing reasonably well in a relatively difficult test.

Appendix 5. $f_{n2}, f_{n3}, f_{u1}, f_{u2}, f_{l1}$ and f_{l2}

The geometrical relationship between the layers bounded by the temperature surfaces and the water store bounded by fixed surfaces are expressed in terms of three coefficients f_n, f_0, and f_{l1} which are respectively the fractions of the snow pack, the top soil and the subsoil lying within the current thermal layer.
The general functions of f_n, f_u and f_l are

$$f_n = \text{clip}\ [0.0, \frac{\min\ (z_t, d_n) - \max\ (z_b, 0.0)}{d}, 1.0] \tag{A42}$$

$$f_u = \text{clip}\ [0.0, \frac{\min\ (z_t, 0.0) - \max\ (z_b, -d_u)}{d}, 1.0] \tag{A43}$$

$$f_l = \text{clip}\ [0.0, \frac{\min\ (z_t, -d_u) - \max\ (z_b, -d_u - d_l)}{d}, 1.0] \tag{A44}$$

In the case of one,

$$z_t = d_n,\ z_b = d_n - d_0,\ \ d = d_0 \tag{A45a}$$

and substituting them into (A42)−(A44), f_{n1}, f_{u1}, and f_{l1} can be obtained. The nice form of f_{n1} was given in Equation (9.44) as an example.

In the case of two,

$$z_t = d_n - d_0,\ \ z_b = d_n - d_0 - d_1,\ \ d = d_1 \tag{A45b}$$

and substituting them into (A42)−(A44), f_{n2}, f_{u2}, and f_{l2} can be obtained.

In the case of three,

$$z_t = d_n - d_0 - d_1,\ \ z_b = d_n - d_0 - d_1 - d_2,\ \ d = d_2 \tag{A45c}$$

and substituting them into (A42)−(A44), f_{n3} can be obtained.

References

Abramopoulos, F., Rosenzweig, C. and Choudhury, B., 1988, Improved ground hydrology calculations for global climate models (GCMs): Soil water movement and evapotranspiration, *J. Climate*, **1**, 921-941.

Andre, J.C. and Blondin, C., 1986, On the effective roughness length for use in numerical three-dimensional models, *Bound. Layer Meteor.*, **35**, 231-245.

Bhumralkar, C. M., 1975, Numerical experiments on the computation of ground surface temperature in an atmospheric general circulation model, *J. Appl. Meteor.*, **14**, 1246-1258.

Blackadar, A.K., 1976, Modelling the nocturnal boundary layer. In *Proceedings of the Third Symposium on Atmospheric Turbulence, Diffusion and Air Quality*, pp.46-49. American Meteorological Society, Boston, Ma.

Boer, G.J., McFarlane, N.A., Laprise, R., Henderson, J.D. and Blanchet, J-P., 1984, The Canadian Climate Centre spectral atmospheric general circulation model, *Atmosphere -Ocean*, **22**, 397-429.

Bolsenga, S.J., 1977, Preliminary observations on the daily variation of ice albedo, *J. Glaciol.*, **18**, 517–521.

Bourke, W., McAvaney, B., Puri, K. and Thurling, R., 1977, Global modelling of atmospheric flow by spectral methods, *Methods in Comp. Phys.* **17**, General Circulation Models of the Atmosphere, ed. J. Chang, Academic Press, New York, 267–324.

Brutsaert, W., 1982, *Evaporation into the Atmosphere. Theory, History and Application*, D. Reidel Publ. Comp., 299pp.

Budyko, M.I., 1956, *The Heat Balance of the Earth's Surface*, Gidrometeoizdat, Leningrad, 255pp. (In Russian).

Businger, J.A., Wyngaard, J.C., Izumi, Y. and Bradley, E.F., 1971, Flux–profile relationships in the atmospheric surface layer, *J. Atmos. Sci.*, **28**, 181–189.

Cambell, G.S., 1974, A simple method for determining unsaturated hydraulic conductivity from moisture retention data, *Soil Sci*, **117**, 311–314.

Carslaw, H.S. and Jaeger, J.C., 1959, *Conduction of Heat in Solids*, Clarendon Press, 150pp.

Carson, D.J., 1987a, An introduction to the parameterization of land-surface processes Part 2. Soil heat conduction and surface hydrology, *Meteorological Magazine*, **116**, 263-279.

Carson, D.J., 1987b, An introduction to the parameterization of land-surface processes Part 1 Radiation and turbulence, *Meteorological Magazine*, **116**, 229-242.

Clapp, R.B. and Hornberger,G.M., 1978, Empirical equations for some soil hydraulic properties, *Water Resources Research*, **14**, 601-604.

Clarke, R.H., 1970, Recommended methods for the treatment of the boundary layer in numerical models of the atmosphere, *Aust. Meteor. Mag.*, **18**, 51–73.

Cogley, J.G., 1987, The thermal properties of solid subsurfaces in General Circulation Models, *Trent Tech. Note*, 87-1, 15pp.

Cogley, J.G., Pitman, A.J. and Henderson-Sellers, A., 1990, A land surface for large scale climate models, Trent University technical note, 90–1, Trent University, Peterborough, Ontario, K9J 7B8, Canada.

De Vries, D.A., 1958, Simultaneous transfer of heat and moisture in porous media, *Trans. Amer. Geophys. Union*, **39**, 909–916.

De Vries, D.A., 1963, Thermal properties of soils, in *Physics of Plant Environment*, (ed.) W.R. Van Wijk, North Holland, Publ. Co., 595pp.

De Vries, D.A. and Philip, J.R., 1986, Soil heat flux, thermal conductivity and the null–alignment method, *Soil Sci. Soc. Amer. J.*, **50**, 12–18.

Denmead, O.T. and Millar, B.D., 1976, Water transport in plants (wheat) in the field, *Agron. J.*, **68**, 297-303.

Deardorff, J.W. 1977, A parameterization of ground-surface moisture content for use in atmosphere prediction models, *J. Appl. Meteor.*, **16**, 1182-1185.

Deardorff, J.W. 1978, Efficient prediction of ground surface temperature and moisture with inclusion of a layer of vegetation, *J. Geophys. Res.*, **83**, 1889-1903.

Dickinson, R.E., 1983, Land-surface processes and climate, surface albedos and energy balance, *Advances in Geophysics*, **25**, 303-253.

Dickinson, R.E., 1984, Modelling evapotranspiration for three dimensional global climate models, in *Climate Processes and Climate Sensitivity*, (eds. Hansen, J.E. and Takahasi, T.), Geophys. Monograph, **29**, vol. 5, Am.Geophys.Union, 58-72.

Dickinson, R.E., 1988, The force-restore model for surface temperatures and its generalizations, *J. Climate*, **1**, 1086 -1097.

Dickinson, R.E., Henderson-Sellers,A., Kennedy, P.J. and Wilson, M.F., 1986, Biosphere Atmosphere Transfer Scheme (BATS) for the NCAR Community Climate Model, *NCAR Technical Note, NCAR, TN275+STR*, 69pp.

Dyer, A.J., 1974, A review of flux-profile relationships, *Bound. Layer Meteor.*, **7**, 363-372.

Eagleson, P.S., 1970, *Dynamic Hydrology*. McGraw–Hill Book Company, New York/St. Louis/San Francisco/London/Sydney/Toronto/Mexico/Panama, 462pp.

Farquhar, G.D., 1978, Feedforward responses of stomata to humidity, *Aust. J. Plant Physiol.*, **5**, 787-800.

Federer, C.A., 1979, A soil-plant-atmosphere model for transpiration and availability of soil water, *Water Resources Research*, **15**, 555-562.

Garratt, J.R., 1977, Review of drag coefficients over oceans and continents, *Mon. Wea. Rev.*, **105**, 915-929.

Gates, D.M., 1980, *Biophysical Ecology*, Springer-Verlag, New York, NY, 611pp.

Geiger, R., 1965, *The Climate near the Ground*, Harvard University Press, Cambridge, Massachusetts, 611pp.

Giorgi, F. and Bates, G.T., 1989, On the climatological skill of a regional model over complex terrain, *Mon. Wea. Rev.*, **117**, 2325-2347.

Hart, T.L., Bourke, W., McAvaney, B.J., Forgan, B.W. and McGregor, J.L., 1988, Atmospheric general circulation simulations with the BMRC spectral model: The impact of revised physical parameterizations, *BMRC Research Report*, **12**, 77pp.

Hart, T.L., Bourke, W., McAvaney, B.J., Forgan, B.W. and McGregor, J.L., 1990, Atmospheric general circulation simulations with the BMRC spectral model: The impact of revised physical parameterizations, *J. Climate*, **3**, 436–459.

Hansen, J.E., Russell, G., Rind, D., Stone, P.H., Lacis, A.A., Lebedeff, S., Ruedy, R. and Travis, L., 1983, Efficient Three Dimensional Global Models for Climate Studies: Models I and II, *Mon. Wea. Rev.* **111**, 609-662.

Henderson-Sellers, A., 1990, "Coming of age of land surface climatology, 1990. *Global Planet. Change*, **82**, 291–319.

Henderson–Sellers, A. and Wilson, M.F., 1983, Surface albedo data for climatic modeling, *Rev. Geophys. Space Phys.*, **21**., 1743–1778.

Hinckley, T.M., Lassoie, J.P. and Running, S.W., 1978, Temporal and spatial variations in the water status for forest trees, *Forest Science Monograph*, **20**, Society of American Foresters, 72pp.

Idso, S.B., Jackson, R.D., Reginato, R.J., Kimball, B.A. and Nakayama, F.S., 1975, The dependence of bare soil albedo on soil water content, *J. Appl. Meteor.*, **14**, 109-113.

Jacobsen, I. and Heise, E., 1982, A new economic method for the computation of the surface temperature in numerical models, *Beitr. Phys. Atmosph.*, **55,** 128-141.

Jarvis, P.G. and McNaughton, K.G., 1986, Stomatal control of transpiration: scaling up from leaf to region, *Adv. in Ecol. Res*, **15**, 1-49.

Kay, B.D., Fukuda, M., Izuta, H. and Sheppard, M.I., 1981, The importance of water migration in the measurement of the conductivity of unsaturated frozen soils, *Cold Reg. Sci. Technol.*, **5**, 95–106.

Kojima, K., 1967, Densification of seasonal snow cover, Inst. Low Temp. Sci., Sapporo, 929–952.

Kondratyev, K. Ya., 1969, *Radiation in the Atmosphere*. Academic Press, New York, 912pp.

Kurbatkin, G.P., Manabe, S. and Hahn, D.G., 1979, The moisture content of the contents and the intensity of summer monsoon circulation, *Meteorologiya i Gidrologiya*, **11**, 5-11, translated as *Soviet Meteorology and Hydrology*, **11**, 1-6, 1979.

McFarlane, N. and Laprise, R., 1985, Parameterization of sub-grid scale processes in the AES/CCC Spectral G.C.M., *CCC Report No.* **85-12**, CCRN 17, 70pp. (Unpublished manuscript).

Mahfouf, J-F. and Jacquemin, B., 1989, A study of rainfall interception using a land surface parameterization for mesoscale meteorological models, *J. Appl. Met.*, **28**, 1282-1302.

Mahrt, L., 1987, Grid–averaged surface fluxes, *Mon. Wea. Rev.*, **115**, 1550–1560.

Manabe, S., 1969, Climate and the ocean circulation: 1, The atmospheric circulation and the hydrology of the earths' surface, *Mon. Wea. Rev.*, **97**, 739-305.

Mason, P.J., 1988, The formation of areally-averaged roughness lengths, *Quart. J. Roy. Met. Soc.*, **114**, 399-420.

Massman, W.J., 1980, Water storage on forest foliage, a general model, *Water Resour. Res.*, **16**, 210–216.

Mellor, M., 1977, Engineering properties of snow, *J. Glaciol.*, **19**, 15–66.

Mintz, Y., 1984, The sensitivity of numerically simulated climates to land-surface boundary conditions, *The Global Climate*, Houghton, J.T., (ed.), Cambridge University Press, 79-105.

Monteith, J.L.(ed.), 1976, *Vegetation and the Atmosphere, Volume 2 Case Studies*, Academic Press, London New York San Francisco, 439pp.

Monteith, J.L., 1973, *Principals of Environmental Physics*, Edward Arnold, London, 241pp.

Morris, E.M., 1985, Snow and ice, in *Hydrological Forecasting*, (eds.), Anderson, M.G. and Burt, T.P., J.Wiley and Sons Ltd., 153-182.

Nishimura, H. and Maeno, N., 1985, Studies on structures and physical properties of snow on Mizuho Plateau, *Antarctica. Ann. Glaciol.*, **6**, 105–107.

Noilhan, J. and Planton, S., 1989, A simple parameterization of land surface processes for meteorological models, *Mon. Wea. Rev.*, **117**, 536–549.

O'Brien, H.W. and Munis, R.M., 1975, Red and near infra–red spectral reflectance of snow, *CRREL Res. Rep. 3.32*, U.S. Army Cold Reg. Res. and Eng. Lab., Hanover, N.H., 1975.

Oke, T.R., 1978, *Boundary Layer Climates*, J.Wiley and Sons, 372pp.

Paterson, W.S.B., 1981, *The Physics of Glaciers* (2nd ed.), Pergamon, Oxford, 380pp.

Philip, J.R. and de Vries, D.A., 1957, Moisture movement in porous materials under temperature gradients, *Trans. Am. Geophys. Union*, **38**, 222–232.

Pitman, 1991, A simple parameterization of sub–grid scale open water for climate models, *Climate Dynamics*, Accepted for publication.

Pitman, A.J., Henderson–Sellers, A. and Yang, Z.–L., 1990, Sensitivity of regional climates to localized precipitation in global models, *Nature*, **346**, 734–737.

Rind, D., 1982, The influence of ground moisture conditions in North America on summer climate as modelled in the GISS GCM, *Mon. Wea. Rev.*, **110**, 1487-1494.

Rind, D., 1988, The doubled CO_2 climate and the sensitivity of the modeled hydrological cycle, *J. Geophys. Res.*, **93**, 5385-5412.

Rowntree, P.R. and Bolton, J.A., 1983, Simulation of the atmospheric response to soil moisture anomalies over Europe, *Quart.J.Roy.Meteor.Soc.*, **109**, 501-526.

Rutter, A.J., 1975, The hydrological cycle in vegetation, in *Vegetation and the Atmosphere*, **1**, Monteith, J.L., (ed.), 111- 154, Academic Press, 278pp.

Rutter, A.J., Kershaw, K.A., Robins, P.C. and Morton, A.J., 1972, A predictive model of rainfall interception in forest, *Agric. Met.*, **9**, 857-861.

Sato, N., Sellers, P.J., Randall, D.A., Schneider, E.K., Shukla, J., Kinter, J.L., Hou, Y.-T. and Albertazzi, E., 1989a, Implementing the Simple Biosphere Model (SiB) in a General Circulation Models: Methodologies and Results, *NASA Contractor Report 185509*, August, 1989, 76pp.

Sato, N., Sellers, P.J., Randall, D.A., Schneider, E.K., Shukla, J., Kinter III, J.L., Hou, Y.–T. and Albertazzi, E., 1989b, Effects of implementing the Simple Biosphere Model in a general circulation model, *J. Atmos. Sci.*, **46**, 2757–2782.

Schwerdtfeger, P., 1963, Theoretical derivation of the thermal conductivity and diffusivity of snow, *Insl. Assoc. Sci. Hydrol. Publ.*, **61**, 75–81.

Sellers, P.J. and Dorman, J.L., 1987, Testing the Simple Biosphere Model (SiB) using point micrometeorological and biophysical data, *J. Clim. Appl. Meteor.*, **26**, 622-651.

Sellers, P.J., Mintz, Y., Sud, Y.C. and Dalcher, A., 1986, A Simple Biosphere model (SiB) for use within general circulation models, *J. Atmos. Sci.*, **43**, 505-531.

Stull, R.B., 1988, *An introduction to Boundary Layer Meteorology*, Kluwer Academic Publ., 666pp.

Sud, Y.C., Shukla, J. and Mintz, Y., 1985, Influence of land- surface roughness on circulation and rainfall: A sensitivity study with a GCM, Third Conf. Climate Variations, Los Angeles, *Amer.Meteor.Soc.*, 93-94.

Taylor, P.A., 1987, Comments and further analysis on effective roughness lengths for use in numerical three-dimensional models, *Bound. Layer Meteor.*, **39**, 403-418.

Thornthwaite, C.W., 1948, An approach toward a rational classification of climate, *Geogr. Rev.*, **38**, 55-94.

Verstraete, M.M. and Dickinson, R.E., 1986, Modeling surface processes in atmospheric general circulation models, *Annales Geophysicae*, **4**, 357-364.

Walker, J. and Rowntree, P.R., 1977, The effect of soil moisture on circulation and rainfalln a tropical model, *Quart.J.Roy. Meteor.Soc.*, **103**, 29-46.

Warren, S.G., 1982, Optical properties of snow, *Rev. Geophys. Space Phys.*, **20**, 67-89.

Warrilow, D.A., Sangster, A.B. and Slingo, A., 1986, Modelling of land surface processes and their influence on European climate, Dynamic Climatology Tech.Note No.38, Meteorological Office, MET O 20. (Unpublished), Bracknell, Berks, 94pp.

Watts, W.R., 1977, Field studies of stomatal conductance, in *Environmental Effects in Crop Physiology*, (eds.) J.J. Landsberg and C.V. Cutting, Academic Press, N.Y., 388pp.

Wetzel, P.J. and Chang, J.T., 1988, Evapotranspiration from nonuniform surfaces, a first approach for short term numerical weather prediction, *Mon. Wea. Rev.*, **116**, 600-621.

Wilson, M.F. and Henderson-Sellers, A., 1985, A global archive of land cover and soil data sets for use in general circulation climate models, *J. Climatol.*, **5**, 119-143.

Wiscombe, W.J. and Warren, S.G., 1980, A model for the spectral albedo of snow I, Pure snow, *J. Atmos. Sci.*, **37**., 2712-2733.

Yang, Z-L., 1989, The sensitivity of a general circulation model to soil hydrology parameterization, Master of Science thesis, Melbourne University.

Yeh, T.-C., Weatherald, R.T. and Manabe,S., 1984, The effect of soil moisture on the short-term climate and hydrology change - a numerical experiment, *Mon. Wea. Rev.*, **112**, 474-490.