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Preliminary Considerations for Runoff Modelling in GCMS

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Modelling of runoff in GCMs is a two-stage process: first, a residual water amount resulting from precipitation, evapotranspiration, snowmelt, infiltration and soil moisture movement at each time step must be generated for each grid cell; then a runoff routing model must be applied to this water excess to produce streamflow. Incorporation of the latter into GCMs is as yet in the most initial stages, and will not be reliably achieved for some years to come. However, before this can be accomplished, the water excess that is to be routed must be generated in a physically realistic fashion.

Two land-surface models of widely varying complexity are presently available as options in the Canadian GCM (developed at Atmospheric Environment Service, Canada). The old scheme is based on a bucket approach for soil moisture and on the force-restore method for soil temperature; a more recently developed scheme ("CLASS" – Canadian Land Surface Scheme) incorporates physical modelling of the heat and moisture budgets of three soil layers, the snow cover and the vegetation canopy. Based on comparisons of year-long runs of both models, it is postulated that reliable modelling of the surface water excess cannot be achieved using a "bucket" model. It is also shown that sub-grid scale variability of the precipitation rate must be taken into account, and that the land surface model must allow for surface water retention between timesteps.

Introduction

The treatment of land surface runoff within general circulation models (GCMs) is a topic of increasing interest to researchers involved in the modelling of global climate. The simulation of streamflow over geographical areas provides an output variable which, like surface precipitation, screen temperature and snow depth, can be validated against long-term observations. In support of this goal, the recently-formed Global Runoff Data Centre in Koblenz, Germany, is engaged in compiling a comprehensive database of runoff records for river systems worldwide, and analysis work has already begun on the discharge records assembled to date (*e.g.* Dümenil *et al.* 1993). Moreover, the input of fresh water into the oceans constitutes an important boundary condition for the modelling of oceanic circulation and therefore of sea surface temperatures, which exert a strong influence on climate. Recent studies (*e.g.* Mysak *et al.* 1990) have shown that the historical fluctuations of freshwater discharge into the Arctic, particularly from the Mackenzie River, are highly correlated with variations in sea-ice cover in the Greenland and Labrador Seas.

As a consequence of the increasing recognition of the importance of freshwater runoff in the global climate system, some attempts have begun to be made to incorporate streamflow models into GCMs. Sausen *et al.* (1994) present a linear runoff advection scheme that is based on the slope of the orography modelled by the GCM. Kite (1993) applies a lumped-parameter model to the simulation of runoff in a large river basin in western Canada. Liston *et al.* (1994) describe a mechanistic routing model that simulates flow paths through GCM grid cells based on river networks and streambed geometry. However, all of these models are as yet in the preliminary testing stages, and it is evident that the development of runoff routing models suitable for use in GCMs is still in its infancy.

The present paper takes the problem of runoff modelling in GCMs one step farther back, to a consideration of the actual variables that are to be used as input by the routing schemes under consideration. The main premise of this paper is that before streamflow models can be reasonably implemented in GCMs, there must be some confidence that the total water excess generated over individual grid cells is being simulated with sufficient realism. This grid cell water excess can be defined as the residual of the local precipation, evapotranspiration, snowmelt, infiltration, and soil moisture movement; *i.e.* it is equivalent to the sum of the overland flow and the drainage from the soil column, which are the variables that are used by routing models to generate streamflow. Clearly, if systematic errors are present in the simulation of this total grid cell water excess, reliable estimates of streamflow cannot be generated no matter how sophisticated the routing model.

In the following brief overview, an analysis will be presented of the water excess field generated by two versions of the Canadian GCM, with particular attention to the governing variables of precipitation, surface infiltration and soil moisture.

Description of Land Surface Models

The two versions of the Canadian GCM to be discussed below differ essentially in the land surface schemes that they use. One version incorporates a second-generation land surface model CLASS ("Canadian Land Surface Scheme"), and the other simpler land surface model that was used previously. For the reader's benefit, brief descriptions of the two schemes are given here.

In CLASS (Verseghy 1991; Verseghy *et al.* 1993), the soil column is divided into three layers, whose temperatures and moisture contents are modelled prognostically. Snow cover, if present, is modelled as a separate thermal layer, and heat and moisture transfers between layers are calculated using the thermal diffusion equations and Darcy's law respectively. The surface infiltration rate depends on rainfall rate, soil moisture and soil texture. Water remaining on the surface at the end of each time step is not discarded, but is retained until it either evaporates or infiltrates, to a limiting depth of 0.2 m (after which it is assigned to overland flow). The vegetation canopy, if present, is modelled as thermally separate from the ground; radiation and precipitation cascades through it are explicitly taken into account. Physical characteristics of the canopy are determined as bulk averages of the component vegetation types present on each grid cell. At each time step, grid cells are divided up into four subareas: bare soil, canopy-covered, snow-covered and canopy-and-snow-covered (their respective areas a function of the calculated fractional vegetation and snow covers). Vertical fluxes are calculated separately over each subarea.

The old scheme makes use of the force-restore approach for soil temperature (Deardorff 1978) and the so-called "bucket" model for soil moisture (Manabe 1969). The force-restore method makes the simplifying assumption that the magnitude of the annual thermal forcing term is negligible compared with that of the diurnal term, and calculates the change in surface temperature over a time step as the result of the difference between a "forcing" term, a function of the soil heat flux, and a "restoring" term, dependent on the diurnally-averaged surface temperature. The bucket model treats soil moisture as a single layer, with gains and losses occurring only at the surface *via* infiltration and evaporation; drainage from the bottom of the layer is neglected. The depth of the bucket varies from grid cell to grid cell, and is a function of the primary and secondary vegetation types present. The vegetation canopy is not modelled explicitly, and the effect of snow cover is limited to modification of the surface albedo and the soil bulk heat capacity.

The GCM runs described in this paper are 14- to 17-month simulations, initialized at January 1st with observed data from the 1979 FGGE (First GARP Global Experiment) programme. The runs were performed at a horizontal resolution of T32 (approximately 3.75°×3.75°), with 20 vertical levels in the atmosphere. Sea surface temperatures and sea ice extents were prescribed, and global land cover data were derived from Wilson and Henderson-Sellers (1985). For the run done using CLASS, diagnostic fields were archived over the period from June 1st to the following May

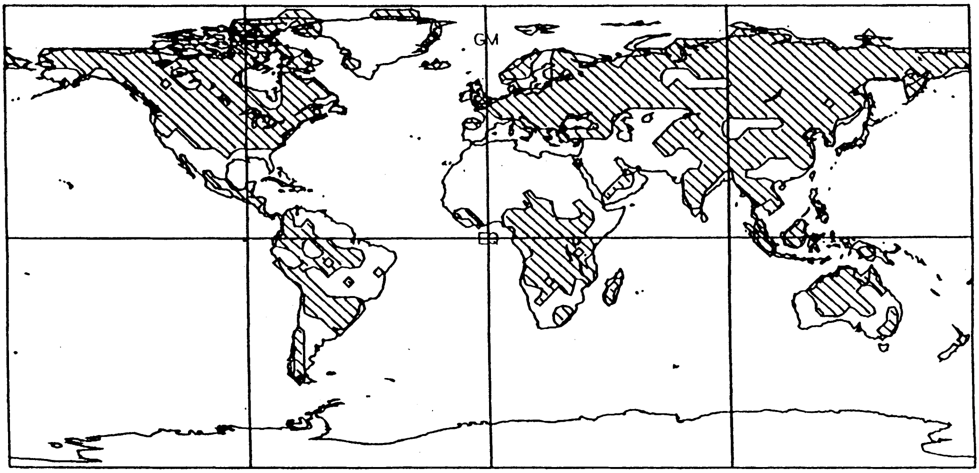
31st (the period conventionally designated by GCM researchers as "summer-autumn-winter-spring"). For the run done using the old scheme, diagnostic fields were only available over the period from March 1st to the following February 28th (the period commonly referred to as "spring-summer-autumn-winter"). However, since a full annual cycle is thus available for both runs, valuable comparisons can be made between the components of the water budget as simulated by the two versions of the GCM.

Modelling of Grid Cell Water Excess

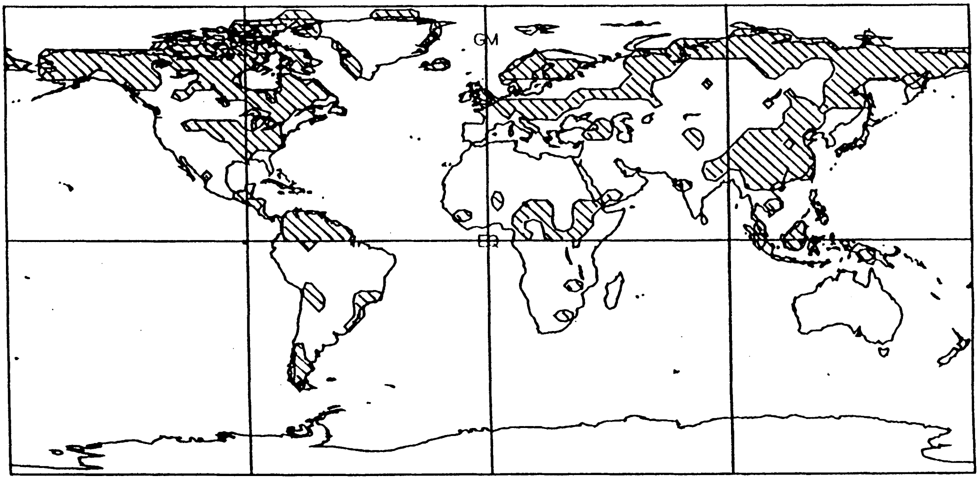
A rough idea of the performance of the GCM with respect to simulating the local water excess over each grid cell may be obtained by referring to the annual, globally-averaged fluxes of precipitation and water excess (which latter must equal the annual global average runoff), and comparing both with measurements or estimates of actual fluxes. According to the precipitation climatology published by Legates and Willmott (1992), the annual precipitation globally averaged over all land surfaces is approximately 2.4 mm d^{-1}. Estimates of global runoff are harder to find, but according to Sellers (1965), the ratio of runoff to precipitation annually averaged over all land areas should be approximately 0.6-0.7. The globally-averaged annual precipitation rate over land areas simulated in the GCM run with the old land surface scheme (GCM-OLD) is somewhat high, at 3.0 mm d^{-1}; that simulated by the GCM coupled with CLASS (GCM-CLASS) comes closer to the observed rate, at 2.6 mm d^{-1}. Yet when the average runoff-to-precipitation rate is calculated for land areas, both runs yield similar values: 0.35 for GCM-CLASS and 0.36 for GCM-OLD, both of which are considerably lower than the expected value.

A clue to the explanation of these low water excess values can be obtained by plotting the grid-cell water excess averaged over the northern hemisphere summer months (June-July-August) for the two model runs. Fig. 1 shows the areas over which the water excess for these three months is greater than zero. It can be seen that even in GCM-CLASS, only about 60% of the land surface area satisfies this criterion; for the old scheme, the proportion drops to about 30%. Over large areas of the globe, therefore, the streamflow generated using the output of either land surface model would be zero.

The first thing to note is that the large discrepancy between GCM-OLD and GCM-CLASS can be attributed to the use of the "bucket" model in the old land surface scheme. As stated above, drainage from the bottom of the soil column is neglected in this model; gains and losses to the soil moisture store only occur *via* surface fluxes. Moreover, the model only generates local water excesses when the "bucket overflows", *i.e.* when rainfall or snowmelt occurs over saturated soil. In Fig. 2, an example is given of the contrast between the output of the two models for a random grid cell centred at 64° 56' N, 11° 15' E, on the western side of Scandina-



a) JJA - AREAS OF WATER EXCESS FOUND USING CLASS



b) JJA - AREAS OF WATER EXCESS FOUND USING OLD SCHEME

Fig. 1. Areas with a modelled water excess for the northern hemisphere summer season (June-July-August), using a) the GCM coupled to CLASS, and b) the GCM coupled to the old land surface scheme.

vian peninsula. For GCM-CLASS, it can be seen that even under dry conditions during the summer, the local water excess over this grid cell never reaches zero; for GCM-OLD, on the other hand, the local water excess is zero for several months. It can be concluded, therefore, that GCM land surface schemes which make use of a simple "bucket" model for soil moisture are unlikely to yield realistic simulations of streamflow.

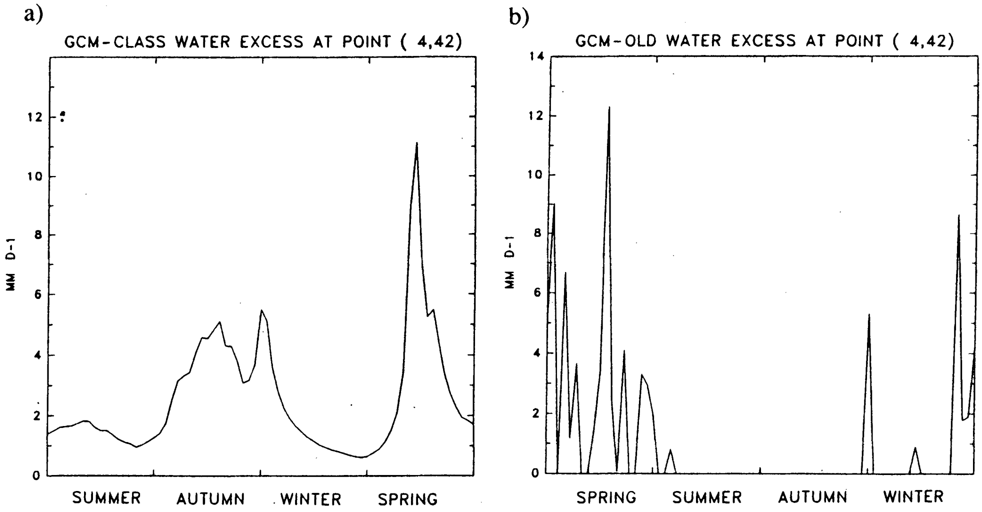


Fig. 2. Annual variation of grid cell water excess (calculated as 5-day averages) for a grid cell centred in western Scandinavia for a) CLASS and b) the old land surface scheme.

Having acknowledged the shortcomings of the old land surface scheme, however, the fact remains that GCM-CLASS also simulates a considerable area of the world as having a water excess of zero during the northern hemisphere summer months. To understand the reason for this, the modelling of precipitation in the GCM must be examined.

Modelling of Precipitation

It has been pointed out that the annual average precipitation rate simulated by GCM-CLASS is quite close to the observed value. However, if annual time series of precipitation are examined over individual grid cells and compared with observed values, a striking discrepancy emerges. Fig. 3 shows two annual time series of daily precipitation totals, the first measured at a site near Prince Albert, Canada, during the 1994 BOREAS campaign (Boreal Ecosystem-Atmosphere Study), and the second modelled for the GCM grid cell most nearly corresponding to this location (53° 48' N, 105° 00' W). While the observed and modelled average annual precipitation rates correspond reasonably closely (1.6 mm d^{-1} for observed and 1.9 mm d^{-1} for modelled), the periodic structure within the time series is quite different. The observed series shows a larger number of high precipitation peaks, but fewer low-magnitude events, while the modelled series is dominated by low-magnitude events. Moreover, in the modelled time series, there are no days at all without rain in contrast to the observations, in which only 28% of the days have precipitation associat-

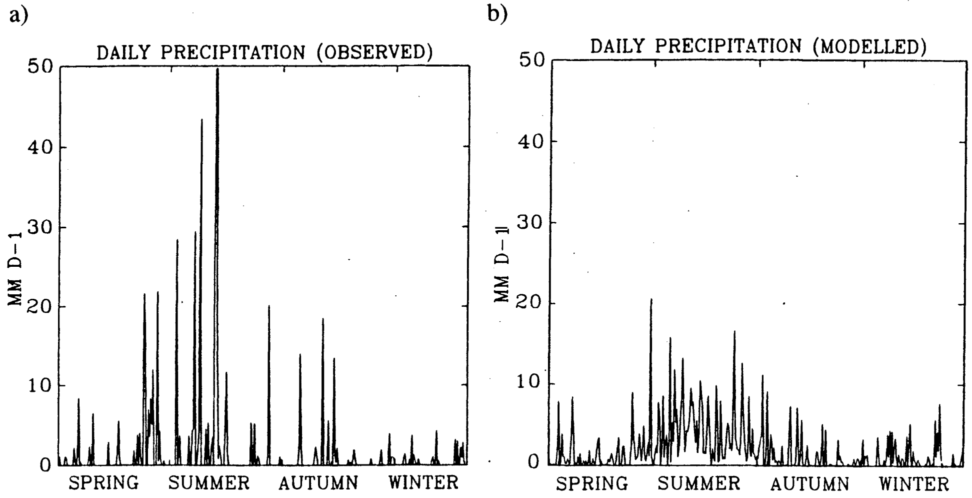


Fig. 3. Annual variation of total daily precipitation, for a) a field site near Prince Albert, Canada and b) the GCM grid cell located nearest that point.

ed with them. This high frequency of low-magnitude events can in fact be shown to be a general characteristic of the precipitation field modelled by the GCM. Fig. 4 contains a plot of the land surface areas which experience precipitation every day during the northern hemisphere summer months. Fully 88% of land areas fall into this category.

The structure of the precipitation field generated by the GCM can essentially be explained as a resolution problem. The runs reported here, as stated above, were

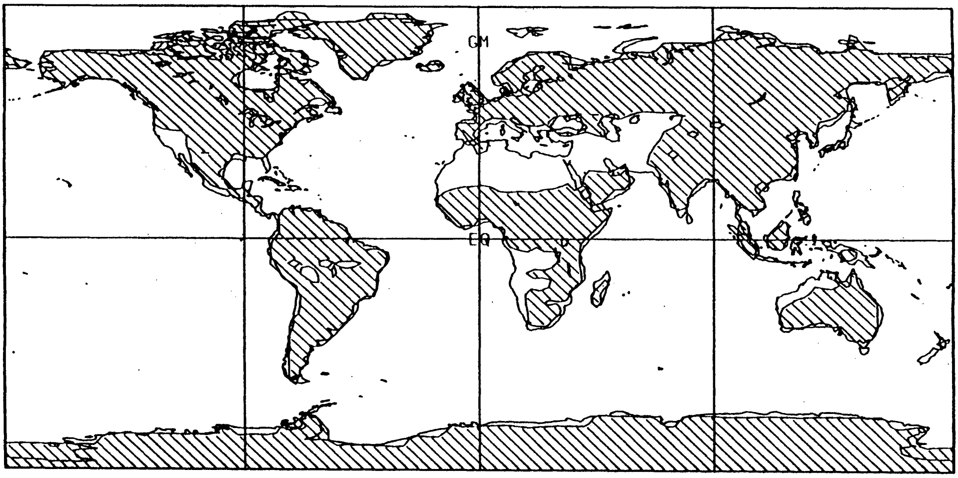


Fig. 4. Areas with precipitation modelled every day during June-July-August.

done at a resolution of 3.75°×3.75°. In most GCMs, including the Canadian one, precipitation generated by the model is implicitly assumed to cover the grid cell uniformly. In reality, of course, much of this precipitation will reflect small-scale or convective activity, and should therefore be modelled as confined to only part of the grid cell. A few GCMs, the UKMO model being the first (Rowntree 1988), have made use of exponential frequency distribution functions to describe the fraction of the grid cell that is covered by convective precipitation. Clearly, from Figs. 3 and 4 the assumption of uniform coverage must lead to serious underestimates of the instantaneous, small-scale precipitation rate.

These low average precipitation rates lead to two results, both of which contribute to explaining the areas of zero water excess shown in Fig. 1. First of all, because little precipitation falls in a given time step and is interspersed with dry periods, the interception capacity of vegetation is less likely to be exceeded, particularly in densely forested areas. The water that falls is therefore likely to be largely intercepted and recycled back to the atmosphere without ever reaching the ground. Second, low rainfall rates also mean that the local infiltration capacity is exceeded only infrequently, which must lead to underestimates of surface overland flow. Several researchers (Wetzel and Chang 1988; Entekhabi and Eagleson 1989; Avissar 1992; Wood *et al.* 1992) have experimented with applying probability density functions to the characterization of sub-grid cell variability of infiltration capacity and/or soil moisture, and incorporating these into large-scale atmospheric models. However, a good deal of development work is still required before these can be implemented operationally in GCMs.

Modelling of Infiltration

Although CLASS does not take into account sub-grid scale variations of surface infiltration capacity, it does incorporate a feature unique among GCM land surface models: the allowance of depression storage between time steps. Most GCMs model infiltration as lasting only as long as the current time step; at the end of the time step (usually of the order of 15-30 minutes), if any water still remains on the surface, it is simply discarded (implicitly, it is assumed to enter the streamflow).

It can be shown that making this simplifying assumption has an adverse effect on the realism of the surface climate simulation. Because the water thus discarded is thereby lost to the soil moisture store, this can in some areas lead to anomalous surface dryness and high screen temperatures. In Verseghy *et al.* (1993), the results of a pair of parallel runs using GCM-CLASS were reported: a control one in which the water ponded on the surface was retained at the end of each time step, and an experiment in which it was discarded to emulate the approach used in most other GCMs. For the northern hemisphere summer months, the control run produced an average screen temperature over land areas (excluding continental ice sheets) of 20.4°C, compared with an observed value, according to Legates and Willmott (1990), of

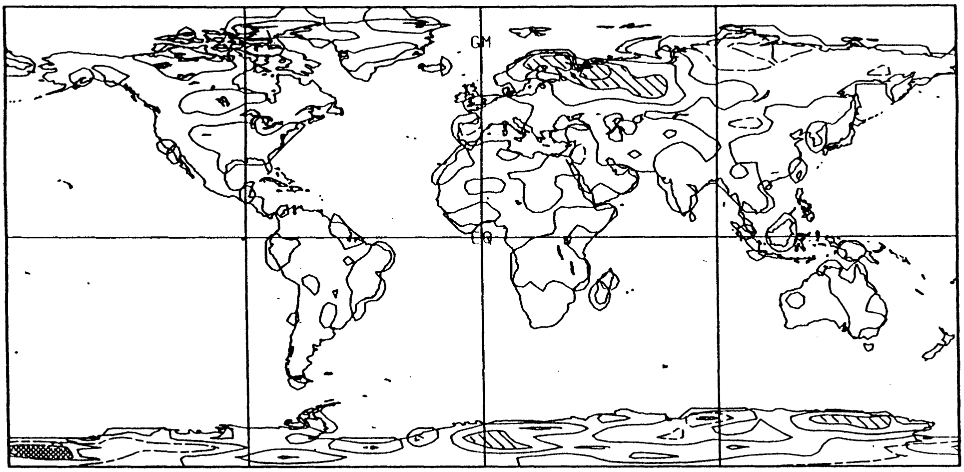


Fig. 5. Temperature difference in the northern hemisphere winter season (December-January-February) caused by the neglect of surface detention of water during the previous year (contour interval 2°C; differences greater than + 4°C are diagonally shaded).

20.2°C; in contrast, the experiment run produced an average screen temperature of 20.8°C. The global distribution of the temperature change is plotted in Fig. 5, with positive differences of greater than 4°C diagonally shaded. The largest differences, reaching 6°C, are found in the large marshy regions of Fennoscandia and northwestern Siberia; clearly, therefore, if a realistic simulation of global wetlands is sought, provision must be made for the long-term retention of water on the surface (in conjunction with this, the capability of modelling sub-grid scale lakes must also be incorporated in the next generation of GCMs, a project which is actively underway in the Canadian modelling community as in others; for example, see Hostetler *et al.* (1993)).

Modelling of Soil Freezing and Thawing

Finally, the soil thermal regime must also be simulated realistically if the surface water excess is to be reliably modelled. Most importantly, the annual march of soil freezing and thawing must be reproduced reasonably well, since the presence of frozen layers in the soil column strongly limits the drainage rate.

It has in fact been demonstrated in another paper (Verseghy 1994) that the use of a "bucket" model for soil moisture in GCMs inherently precludes a realistic simulation of soil freezing and thawing. This is due to the fact that only one soil moisture layer is modelled, leading to the result that in order to preserved physical consistency, the whole soil column must change phase before the surface temperature can pass 0°C. Thus, in the fall when air temperatures are decreasing, the fact that the sur-

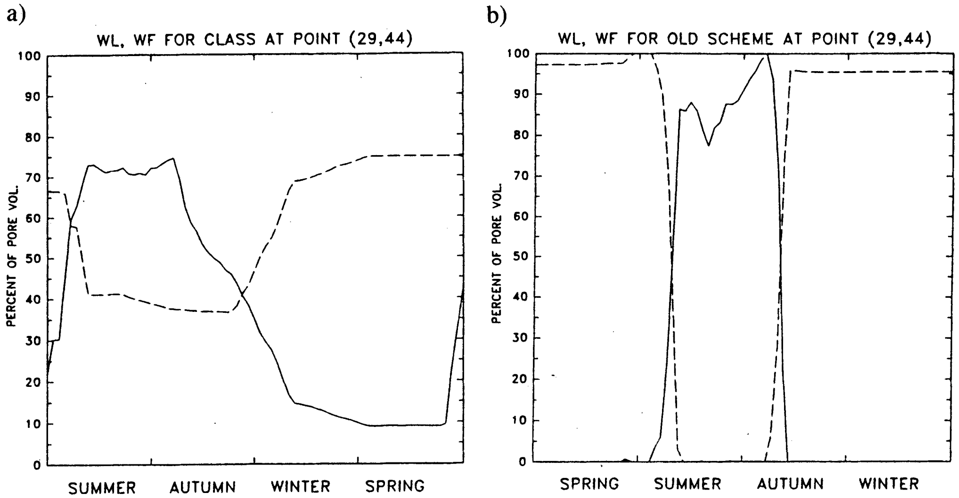


Fig. 6. Annual variation of liquid ("WL") and frozen ("WF") soil moisture averaged over the first metre of the soil column, as modelled for a grid cell centred near Khatanga in northern Siberia, for a) the GCM coupled to CLASS, and b) the GCM coupled to the old land surface scheme.

face temperature is being held at 0°C results in the persistence of unstable conditions in the atmosphere near the surface, resulting in large cooling rates until the ground is frozen. Similarly, in the spring, the fact that the surface temperature is held at 0°C while the overlying air is warming leads to stable atmospheric conditions near the surface and to the suppression of surface fluxes, causing all of the absorbed radiation to be used in thawing the soil.

The net result is that for most areas of the world, the whole modelled soil column (up to two metres deep in places, the depth of the "bucket" being parametrized as a function of the vegetaion rooting depth) freezes in the winter and thaws in the summer; thus, neither shallow frost penetration nor permafrost can be simulated. Fig. 6 shows the annual variation of liquid ("WL", solid line) and frozen ("WF", dashed line) soil moisture averaged over the first metre of the soil column, as modelled by CLASS and by the old land surface scheme, for a grid cell near Khatanga in northern Siberia, centred at 72° 22' N, 105° 00' E in an area dominated by tundra vegetation and underlain by permafrost. It can be seen that while CLASS simulates the normal development of an active layer during the summer, with almost complete refreezing of the soil in the winter (except for a residual amount of water; see *e.g.* Hromadka *et al.* 1981), the old scheme simulates a complete thawing of the soil column in the summer. Moreover, this phase change takes very little time (a little over two weeks). Investigations show that a similar pattern occurs in areas which, according to observations, should experience shallow frost penetration in the winter; in such areas, the old scheme again models the soil column as freezing completely.

Conclusions

In summary, it can be concluded from the foregoing discussion that before runoff routing models can be incorporated into GCMs to provide streamflow simulations for model validation and for input to coupled ocean models, a considerable amount of work must be done to improve the simulation of the surface water excess that feeds into the routing model. As a first conclusion, it is clear that simple soil moisture models of the "bucket" type cannot generate surface water excesses with sufficient reliability for this purpose; the use of a more sophisticated land surface model is definitely required. In addition, in order to provide a reasonable simulation of wetlands, the model should allow for the surface storage of water between time steps. However, perhaps the most important requirement is that subgrid scale variations be taken into account in the model: most importantly of precipitation rate, but soil moisture will also probably be required to enable a rigorous calculation of surface infiltration rate. Only under these conditions will it be possible to regard streamflow simulations made using GCMs as in any way physically meaningful.

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