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Sensitivity of the surface hydrology to the complexity of the land-surface parameterization scheme employed

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RESUMEN

Se investiga la sensibilidad de la hidrología de la superficie al esquema de parametrización incorporado en tres diferentes modelos de Circulación General de la Atmósfera (MCGA). Se encuentra que las simulaciones de la hidrología de la superficie hechas por los modelos NCAR, CCM0 y GFDL (que incorporan parametrización hidrológica tipo "bucket") difieren marcadamente del modelo CCM1B (que incorpora el esquema de transferencia Biosfera Atmósfera, BATS). Los MCGA que incorporan las parametrizaciones más simples de la superficie de la tierra simulan en forma irreal en números de fenómenos físicos. Muchos de estos grandes problemas de la climatología de la superficie de la tierra son eliminados en el modelo CCM1B que incorpora BATS. Cuando se comparan los escenarios correspondientes al contenido de CO_2 de la época presente y la duplicación de éste, se demuestra que las diferencias entre los resultados de los modelos incorporando diferentes esquemas en la superficie son mayores que las diferencias enre $1 \times$ y $2 \times CO_2$. Estos resultados implican que los campos hidrológicos de los MCGA, y sus predicciones de cambio climático en la *superficie de la tierra*, son probablemente poco confiables. Investigaciones del impacto de cambios climáticos asociados con cambios en la climatología de la superficie de la tierra deben emplear una de las parametrizaciones avanzadas de la superficie de la tierra acoplada con el modelo de circulación general, y desarrollar y validar aquellos componentes del modelo atmosférico que son importantes para derivar procesos en la superficie de la tierra, tales como forzamiento radiacional en la superficie, contribuciones regionales de la precipitación, y el ciclo diurno dentro de la capa límite planetaria.

ABSTRACT

The sensitivity of the surface hydrology to the parameterization scheme incorporated in three different AGCMs is investigated. It is found that the simulations of the surface hydrology by the NCAR CCM0 and GFDL models (which incorporate the bucket type hydrological parameterization) differ markedly from the CCM1B model (which incorporates the Biosphere Atmosphere Transfer Scheme, BATS). The AGCMs which incorporate the simpler parameterizations of the land surface simulate a number of physical phenomena unrealistically. Many of these gross problems in the land surface climatology are eliminated in the CCM1B model which incorporates BATS. When present-day and doubled CO_2 scenarios are compared, it is shown that the differences between model results incorporating different surface schemes are greater than the differences between $1 \times$ and $2 \times CO_2$. The implications of these results are that the surface hydrological fields from AGCMs, and their predictions of climatic change *at the land surface*, are likely to be unreliable. Investigations of the impact of climatic change consequent upon changes in the land surface climatology should employ one of the advanced parameterizations of the land surface coupled into the general circulation model, and develop and validate those components of the atmospheric model, such as the surface radiative forcing, regional precipitation patterns and the diurnal cycle within the planetary boundary layer, that are most important for deriving land surface processes.

1. Introduction

To date the continental surfaces have been somewhat neglected by climate modellers: the parameterization of the land surface is highly simplistic in most atmospheric general circulation climate models (AGCMs) and evaluation of the present-day continental surface climate usually consists of perfunctory comparisons of temperature and precipitation. Despite the highly simplistic nature of the land surface parameterization employed, AGCMs exhibit considerable sensitivity to changes in surface specification. An increase in surface albedo can reduce simulated precipitation (Charney *et al.*, 1977), while extreme changes in surface evaporation (e.g., zero to potential) can also have large effects (e.g., Shukla and Mintz, 1982). More recently the climate has been shown to be sensitive to the parameterization of the surface roughness length (Sud *et al.*, 1985; Sud and Smith, 1985). Furthermore, it has been recognized that predictions of climatic changes are highly sensitive to apparently minor differences in the implementation and initialization of land surface schemes (e.g. Hunt, 1985; Manabe and Wetherald, 1987; Mitchell and Warrilow, 1987; Rind, 1988).

Over the last three years the parameterization of the land surface in AGCMs has improved considerably. To date, three examples of this new generation of land-surface schemes, which include an explicit parameterization of the canopy, have been coupled with AGCMs. These are the Simple Biosphere model (SiB) of Sellers *et al.* (1986), the Bare Essentials of Surface Transfer model (BEST) of Pitman (1988) and the Biosphere Atmosphere Transfer Scheme (BATS) of Dickinson *et al.* (1986). It is the latter scheme, BATS, coupled into a host AGCM, version one of the NCAR Community Climate Model (CCM1) which has been used for this experiment.

BATS has been extensively tested in a stand-alone mode (e.g., Wilson *et al.*, 1987a) and has been used to examine the climate of tundra regions (Wilson *et al.*, 1987b) and the impact of Amazonian tropical deforestation (Dickinson and Henderson-Sellers, 1988), by coupling it into the previous version of the CCM (CCM0B). In Section 2, three different model integrations are described: firstly, the CCM0 (without BATS) coupled to a simple mixed layer ocean; secondly a three year integration of NCAR CCM1 + BATS (hereafter CCM1B) and finally a simulation of surface hydrology from the GFDL AGCM. These have given rise to monthly means of a large number of surface climate parameters.

In Section 3, the sensitivity of the surface hydrology is investigated using results from these 3 multi-year simulations. Finally, in Section 4, the effects of doubling atmospheric CO_2 on the surface hydrology are investigated for a number of regions. The primary aim was to investigate whether the changes induced at the land surface by doubling CO_2 were large enough to mask intermodel differences.

2. Simulations of land-surface hydrology

2.1 CCM0 plus a simple mixed layer ocean model

In this version of the CCM, the land surface hydrology is modelled by a simple bucket (or Budyko) parameterization in which the bucket depth (or the 'field capacity') is a uniform 15 cm over the globe (e.g. Manabe, 1969; Meehl and Washington, 1988). The land surface albedos are specified as 0.13 for all non-desert and snow-free areas, 0.25 for deserts and 0.80 for snow. Cloud formation is determined interactively using the Ramanathan *et al.* (1983) radiation model. The planetary boundary layer is represented by eddy diffusion with essentially a constant coefficient. The diurnal cycle is not resolved. This version of the CCM has been coupled to a simple mixed layer ocean model that computes seasonal heat storage based upon the assumption of an isothermal ocean layer of fixed depth equivalent to 50 m. The ocean model, described in Washington and Meehl (1984), does not include horizontal or vertical heat transport or changes in salinity.

Two experiments were performed using this coupled atmosphere plus mixed layer ocean model: a control climate simulation (also termed $1 \times CO_2$) and a doubled CO_2 simulation ($2 \times CO_2$). In each case, the experiment comprised three parts: a two phase 'spin-up' followed by simulation of 11 full annual solar cycles. The results used here, described in full in Washington and Meehl (1984), are three-year averages from the final three years of the 11 year simulation.

The simulations from the GFDL model were generated using an essentially identical parameterization of land surface hydrology as was used in the CCM0 model but with only a superficially similar atmospheric model (Manabe and Wetherald, 1987).

2.2 CCM1 plus the Biosphere-Atmosphere Transfer Scheme (BATS): CCM1B

A diurnal cycle version of the CCM1, now incorporating a planetary boundary layer with stability-dependent diffusion inferred from Monin-Obukhov similarity theory and an updated radiative transfer package (Williamson *et al.*, 1987), has been coupled to the Biosphere Atmosphere Transfer Scheme of Dickinson *et al.* (1986) in which prognostic equations are solved for the temperatures and water contents of a surface and deep soil layer and a canopy (e.g., Wilson *et al.*, 1987b). Foliage and canopy air temperatures are calculated diagnostically using an energy balance equation which includes stomatal resistance, transpiration and direct evaporation of intercepted precipitation from the leaves.

The BATS package uses a vegetation/land cover classification derived from Matthews (1984) and Wilson and Henderson-Sellers (1985), containing 17 land use classes plus an ocean class. Each of the 4.5° by 7.5° grid elements is associated with one of these classes. The depth of the upper and total soil layers, the roughness length, the percentage vegetation cover, leaf area index and hence the surface radiative, momentum exchange and hydrological properties are all dependent upon the land surface class prescribed within the BATS submodel.

2.3 Soil moisture evaluations

The amount and seasonal distribution of soil moisture has been identified as an important parameter in determining the land surface climate. Great care must be taken in comparing different model estimates of soil moisture (*cf.* Mintz and Serafini, 1984). The CCM0 and GFDL surface schemes are essentially a bucket-type hydrology. This parameterization follows Manabe (1969) by defining an equation for the rate of change of soil moisture, W, with time, t, as

$$\frac{\partial W}{\partial t} = P - E + S_m \tag{1}$$

where P is precipitation, E is evaporation, S_m is snow-melt. If the soil moisture exceeds 15 cm then runoff occurs. The evaporation rate is a function of the soil moisture such that if the soil moisture is greater than a critical value (75% of "field capacity" i.e., 11.25 cm) potential evaporative demand is satisfied.

It is possible to normalize the bucket soil moisture to produce a soil wetness fraction by dividing the soil moisture, W, by the maximum soil wetness or "field capacity" which in these schemes is 15 cm at all locations, i.e.

$$W_{norm} = \frac{W}{W_{fc}} \tag{2}$$

where W_{norm} is the normalized soil moisture and W_{fc} is the "field capacity".

The total soil water computed by BATS, on the other hand, includes a moisture term which is unavailable to the vegetation. This is the soil moisture which remains when the plants are deemed to have wilted and hence ceased transpiration. For comparison, we determined normalized BATS total soil moisture values less this "unavailable"soil water (i.e., the total soil moisture available for transpiration or soil evaporation) which is

$$W_{norm} = \frac{(W - W_{un})}{W_{max}} \tag{3}$$

where W_{un} is the "unavailable" soil water and W_{max} is the maximum possible soil moisture (in cm) defined as the soil depth divided by the soil porosity. The two normalized soil moistures given in Equations (2) and (3) are similar except that Equation (3) would usually be expected to be smaller since values of soil water W near W_{max} are reduced rapidly by model subsurface drainage.

The surface hydrological fields derived from the three models for selected regions have been investigated.

3. Regional analyses

Although it has generally been established that AGCM simulations represent the Earth's macroclimatology adequately, the analysis of AGCM results at a finer spatial scale reveals some inadequacies. For example, Gutowski *et al.* (1988) established that regional scale surface energy flux differences as large as 50 W m^{-2} exist between three different AGCM simulations; this was more than twice the difference found amongst the same AGCMS for globally averaged fluxes. Still more importantly, their analysis showed that these models even differed over the sign of the regional surface flux changes when CO_2 is doubled. They attributed these differences to differences in the model hydrologies.

For the comparison to be conducted here, three regions have been selected for analysis. These are two regions selected by Meehl and Washington (1988) and Manabe and Wetherald (1987) [i.e. northern Canada and central North America] plus an additional area in eastern Australia (Figure 1). Note that the GFDL simulations used fixed zonally averaged cloudiness compared to the predicted cloudiness in the CCM0 and CCM1B experiments.

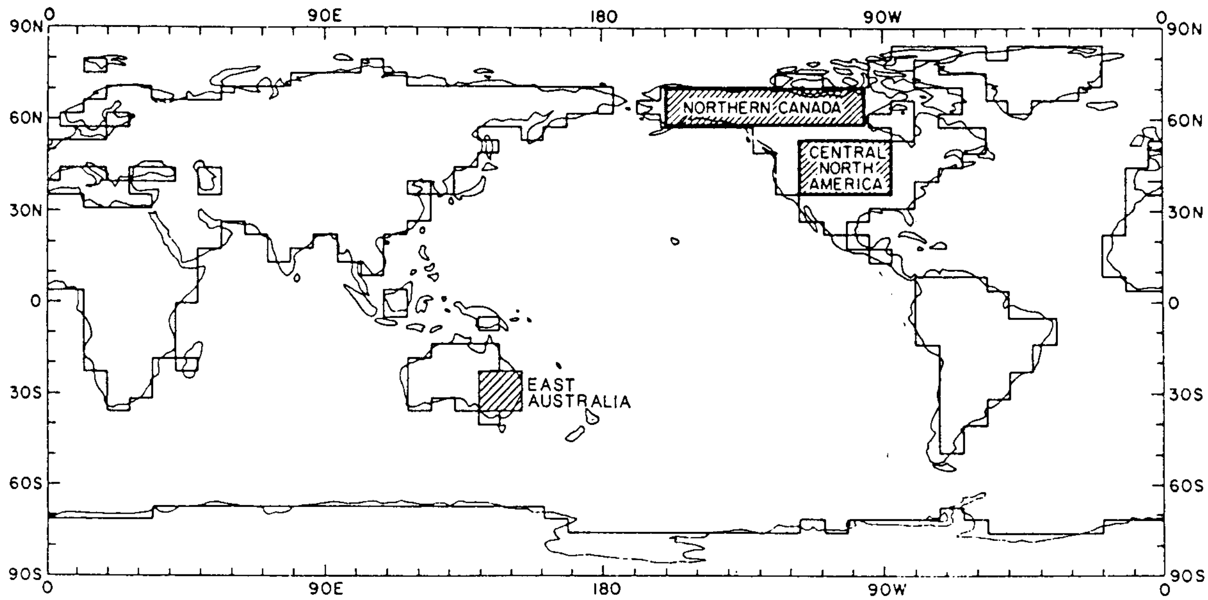


Fig. 1. The location of the three regions examined: northern Canada, central North America and east Australia.

3.1 Northern Canada

The annual cycles of soil moisture derived for northern Canada for each model are shown in Figure 2. The critical difference between results from the CCM0 and GFDL models is the

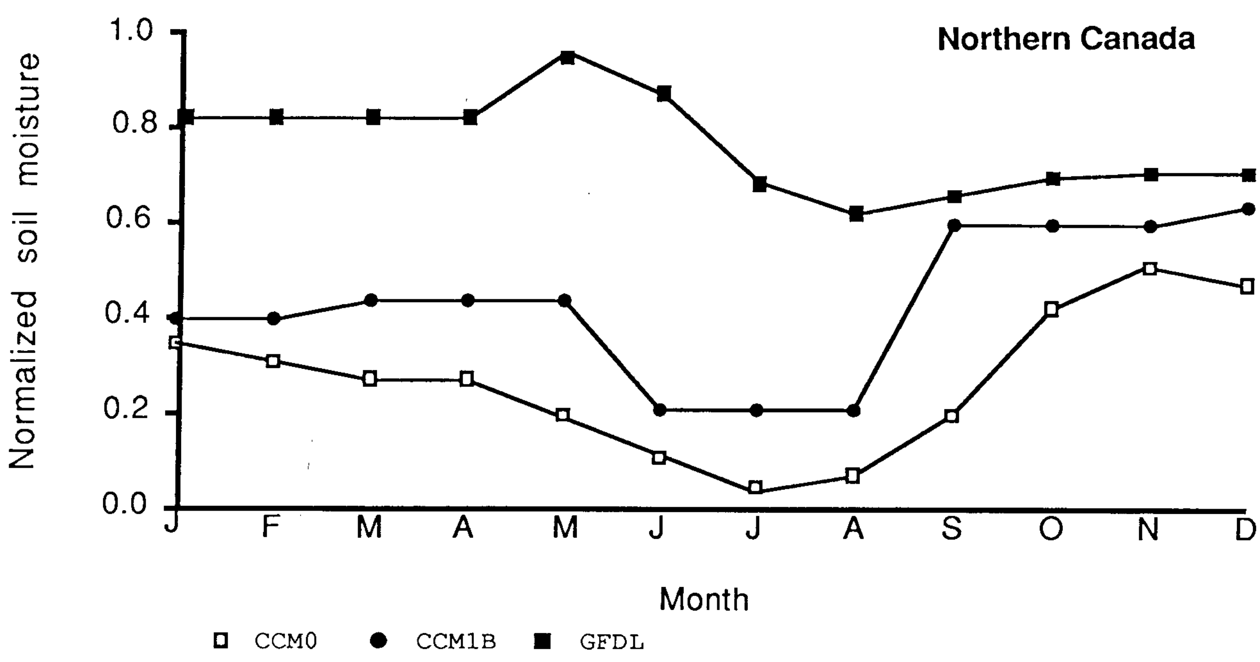


Fig. 2. Annual soil moisture distribution for northern Canada for 1 × CO_2 from the CCM0 and GFDL models (after Meehl and Washington, 1988), and from the CCM1B model. The soil moisture is expressed as a fraction of field capacity, which is normalized in the case of the CCM1B.

distribution of soil moisture in the spring. In the CCM0 model, soil moisture levels slowly decline through the year, due to evaporation, until July, when higher precipitation begins to replenish the soil moisture store. In contrast, the GFDL model simulates constant soil moisture levels until April, when due to snow-melt (Figure 3a) the soil moisture store reaches about 0.95 (i.e. 95% of field capacity). Evaporation then reduces the soil moisture level until August, when the moisture flux into the soil becomes positive again. The CCM1B model simulates a similar amount of moisture in the soil compared to the CCM0 model; hence the GFDL model is substantially wetter than the other two models.

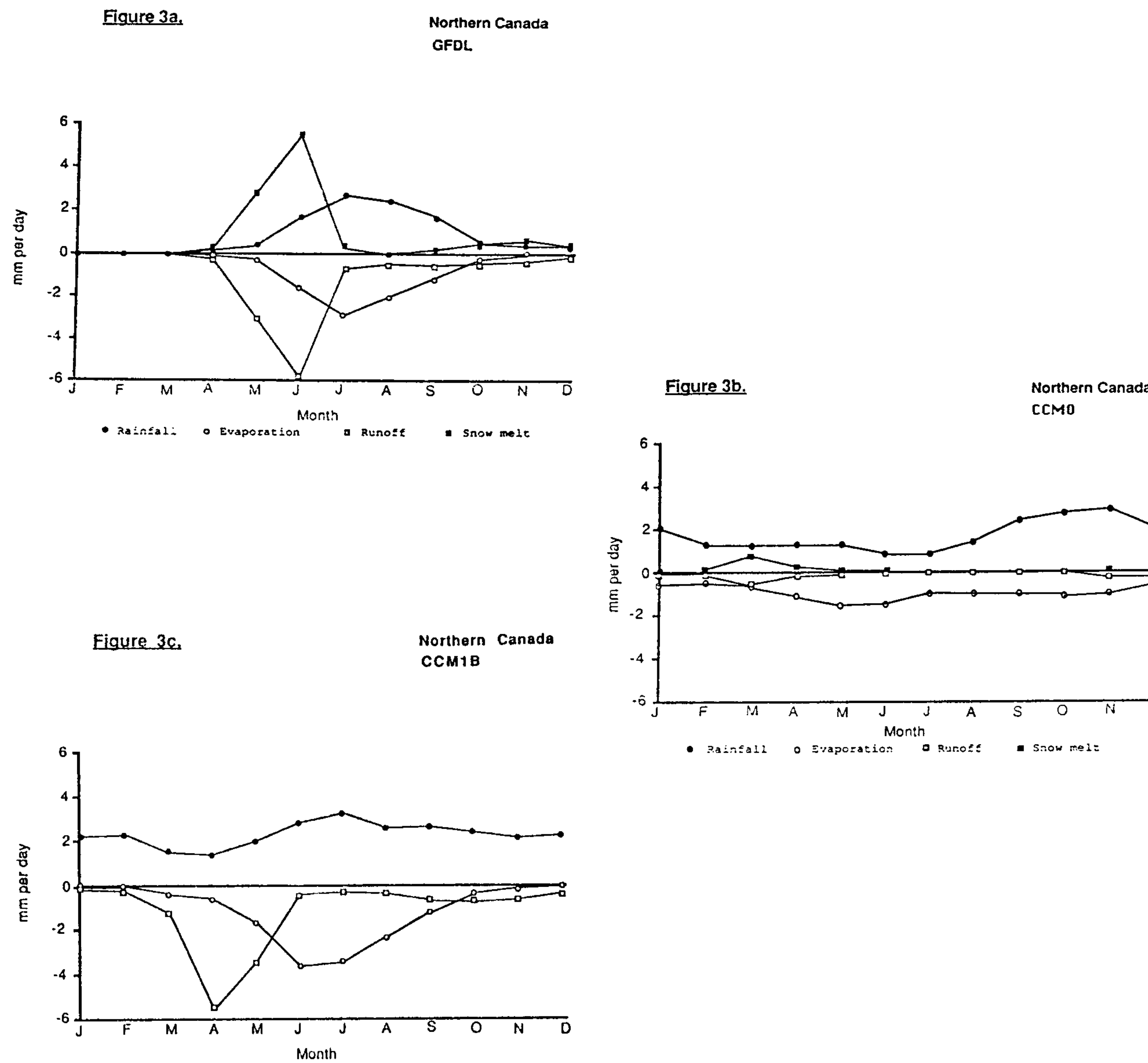


Fig. 3. Annual variation of surface hydrological fluxes for northern Canada for a) the GFDL model, b) the CCM0 model (both after Meehl and Washington, 1988), and c) for the CCM1B model. The fluxes are monthly averages, expressed in mm d^{-1}.

A most surprising feature of the CCM0 soil moisture distribution is the lack of a spring runoff peak. A comparison of Figures 3a, b, c shows that, whereas in the GFDL and CCM1B model spring runoff peaks at about 5.5 mm d^{-1}, in the CCM0 model, runoff never exceeds 0.5 mm d^{-1}. This is largely the result of the low snow-melt rate in the CCM0 model (approximately 1 mm d^{-1} in comparison to greater than 5 mm d^{-1} in the GFDL model). This in turn is the result of too little snow accumulation in the CCM0 model, continuous runoff through the winter and comparatively high mid-winter evaporation rates. These processes combine to reduce the snow pack, thereby preventing the simulation of a satisfactory spring snow-melt peak. Indeed, the simulation of runoff by the CCM0 model is generally poor. Runoff is not simulated between June and September which is incorrect (UNESCO, 1971). In contrast both the CCM1B and GFDL models simulate runoff throughout the year.

There are other important differences between GFDL and CCM1B models in the simulation of runoff. Although both models represent the gross hydrology reasonably well, the phase of the spring runoff peak is quite different. In the GFDL model, snow-melt induced runoff begins during March, increasing in April to peak in June. By July, the influence of snow-melt has ceased to be important. In contrast, in the CCM1B model, runoff increases from background (sub-surface) runoff in February, reaching a peak in April, and falling back to pre-snow-melt levels by June. There is therefore a clear *one month* difference between these two models when the seasonality of the spring snow-melt is analyzed. As discussed by Mitchell and Warrilow (1987) and Rind (1988), this difference could be crucial to the hydrology (i.e. soil moisture distribution) of the following year. It seems likely that the hydrological model based on Manabe (1969) responds more slowly to seasonal events, compared to the improved parameterization of soil hydrology incorporated in the BATS submodel. However, the lack of observational data at an appropriate scale prevents a quantitative assessment of the relative merits of the two models.

The simulation of runoff in the CCM0 model has a profound effect on the soil moisture amounts throughout the year. Since the spring snow-melt fails to replenish the soil moisture (Figure 2) (as in the other two models), the soil moisture falls to extremely low levels in mid summer (0.1 in comparison to 0.4 in the CCM1B and 0.66 in the GFDL model). The resulting effect is an absence of significant seasonality in the monthly evaporation rate (Figure 3b). The GFDL and CCM1B models both show the summer evaporation rates to be 3 to 4 times the winter rate. It is possible that the lack of seasonality in soil moisture and evaporation influences precipitation, which in both the GFDL and CCM1B model shows some seasonality, but such seasonality is largely absent in the CCM0 model.

3.2 Central North America

Figure 4 shows the seasonal distribution of soil moisture simulated by each model for central North America. In the CCM0 model the soil is relatively dry throughout the year, becoming desiccated between June and October. The GFDL model shows a much greater seasonality, from near saturated soils in winter to relatively dry soils during the summer. The CCM1B model simulation of soil moisture is similar to the CCM0 model. The CCM1B and GFDL models simulate a small runoff peak in spring (associated with snow-melt, see Figures 5a and 5c) which is not simulated particularly well by the CCM0 model (Figure 5b). The CCM0 model simulates a runoff peak in May (although this is an order of magnitude smaller than that simulated by the other two models).

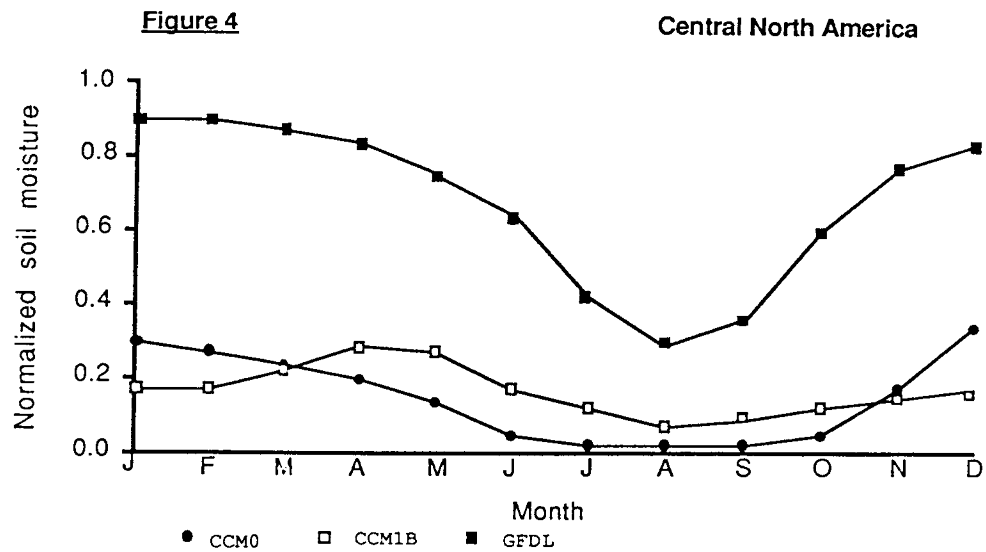


Fig. 4. Annual soil moisture distribution for central North America for 1 × CO_2 from the CCM0 and GFDL models (after Meehl and Washington, 1988), and from the CCM1B model. The soil moisture is expressed as a fraction of field capacity, which is normalized in the case of the CCM1B.

In both the GFDL and CCM1B models, runoff occurs throughout the year. In the case of the CCM1B this is gravitational drainage. The CCM0 model does not simulate runoff between July and October, but this is not necessarily unrealistic, although the extreme soil desiccation simulated by this model (Figure 4) is probably unrealistic.

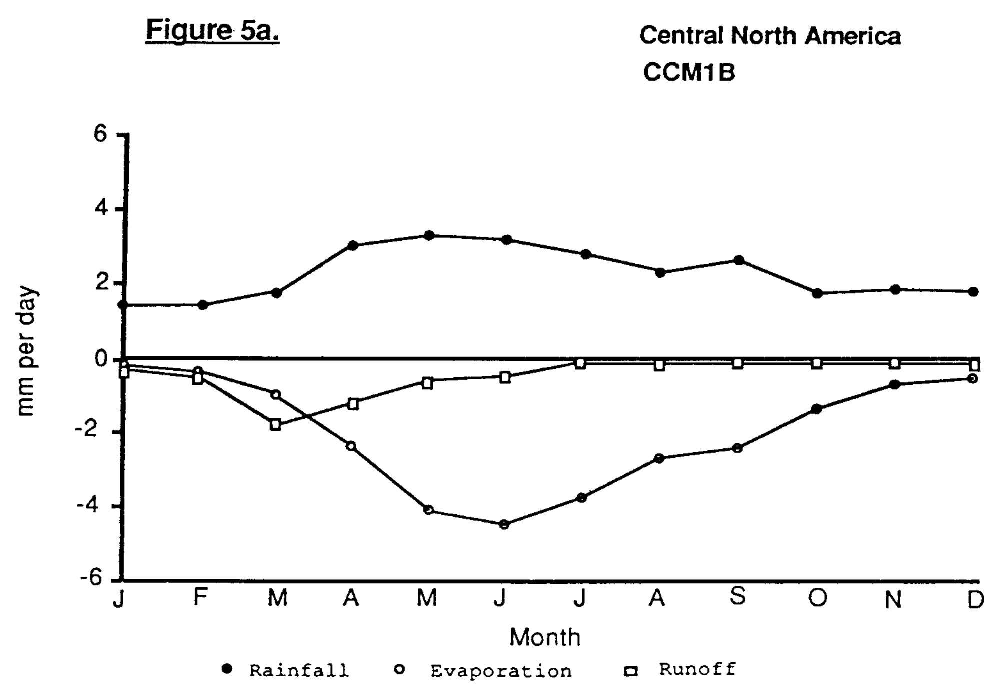


Fig. 5. Annual variation of surface hydrological fluxes for central North America for a) the CCM1B model, b) the CCM0 model and c) the GFDL model (results from the CCM0 and GFDL models are after Meehl and Washington, 1988). The fluxes are monthly averages, expressed in mm d^{-1}.

Figure 5b. Central North America CCM0

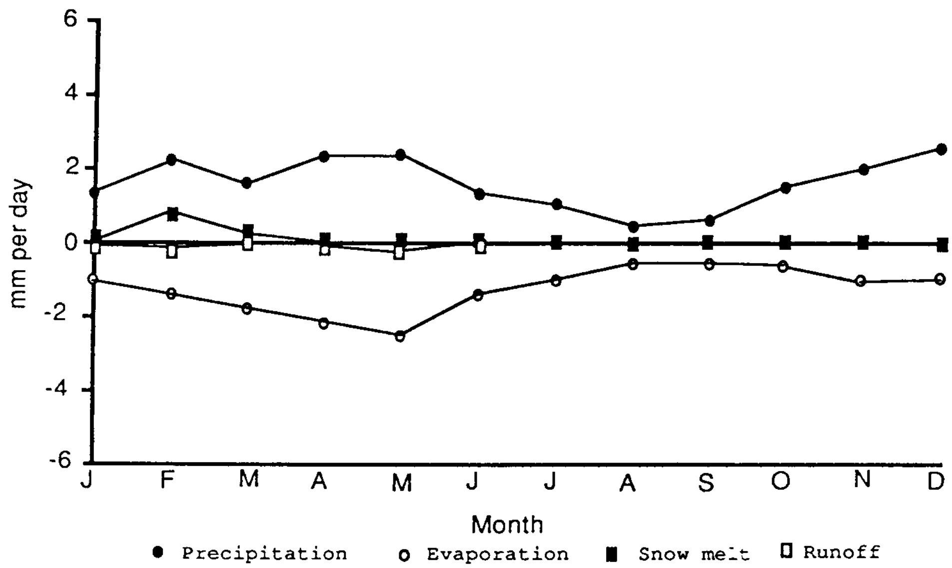
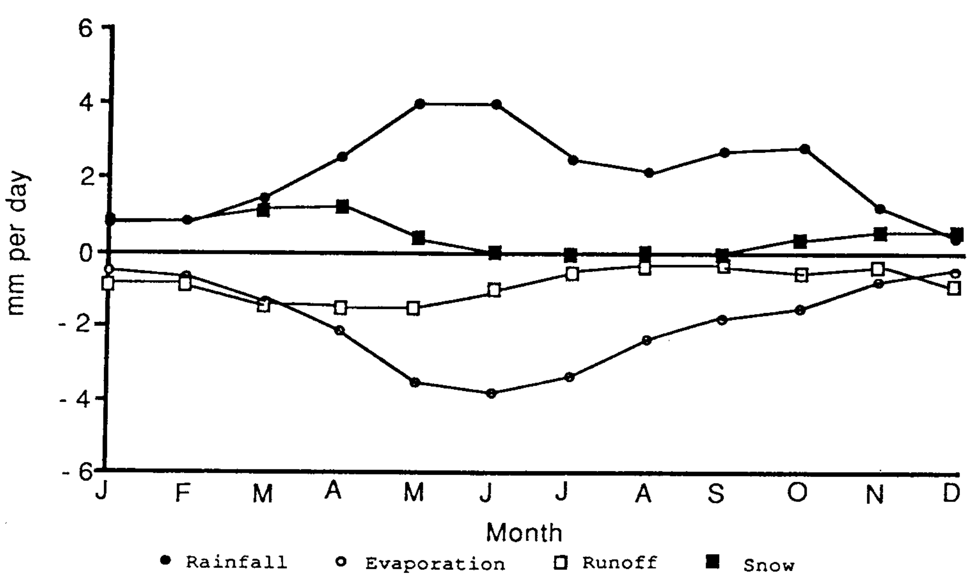


Figure 5c. Central North America GFDL



The very dry soils simulated by the CCM0 model are partly due to relatively low rainfall rates, which do not exceed 2.5 mm d^{-1} (Figure 5b). According to Baumgartner and Reichel (1975) annually averaged precipitation rates over central North America are approximately 1.5 mm d^{-1}. (Note that this figure masks extreme spatial inhomogeneity in the precipitation field). Therefore, for this region, the CCM0 model reproduces the observational data reasonably well. The model rainfall evaporates quickly leaving the soil dry and therefore preventing runoff, since gravitational drainage is not parameterized by the CCM0 model. In contrast, the GFDL and CCM1B models simulate summer precipitation rates in excess of 4 mm d^{-1} (Figure 5a, 5c) which appear to be excessively high. Most of the precipitation evaporates, but enough moisture is retained in the CCM1B model to prevent the extreme mid summer desiccation simulated by CCM0 model (see Figure 4). The advanced hydrological parameterization incorporated into the BATS submodel used in the CCM1 integration avoids waterlogging from the excessive precipitation by simulating gravitational drainage. In the GFDL model, the simpler hydrological parameterization responds to the excessive precipitation by filling the soil with water. No moisture can drain out until field capacity is exceeded.

3.3 Eastern Australia

Although we do not have access to results for the surface hydrology of eastern Australia from the GFDL AGCM, it is interesting to analyze the results from the CCM1B model in order to see whether it can reproduce the broad surface hydrology and climatology of this heterogeneous region.

Figure 6a shows that the annual variation of soil moisture is very different when the CCM0 and CCM1B models are compared. The CCM0 model shows normalized soil moisture as a fraction of field capacity varying between 0.2 and 0.55 while the soil moisture levels in the CCM1B model vary from almost zero to only about 0.05.

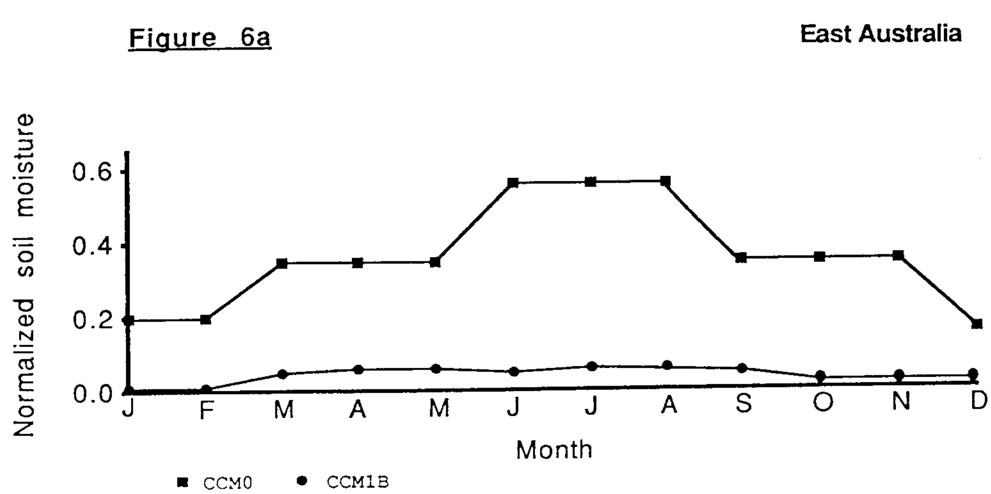


Fig. 6a. Annual soil moisture distribution of east Australia for 1 × CO_2 from the CCM0 model (after Meehl and Washington, 1988), and from the CCM1B model. The soil moisture is expressed as a fraction of field capacity, which is normalized in the case of the CCM1B.

Due to the extreme dryness of the CCM1B model, runoff (Figure 6b) is low (less than 0.2 mm d^{-1}). It is apparent from Figure 6b that most of the precipitation evaporates quickly, while a sizable fraction of any net flux of moisture to the soil runs off. During the first few months of the year (autumn) there is still a small positive flux of water to the soil, which leads to a small increase in soil moisture levels (Figure 6a). For most of the months after May, evaporation exceeds precipitation, drying the soil.

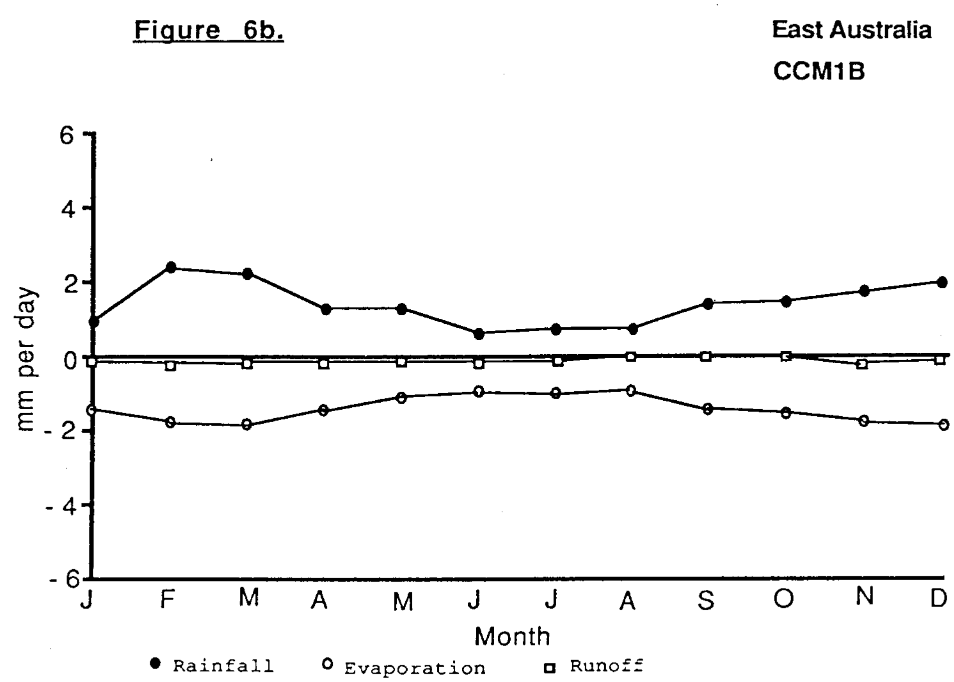


Fig. 6b. **Annual variation of surface hydrological fluxes for east Australia for the CCM1B model. The fluxes are monthly averages, expressed in mm d^{-1}.**

The annual total runoff simulated by the CCM1B model is shown in Figure 7 for each model grid square. This seems reasonable in comparison with observed data from the Australian Water Resources Council (1978). The CCM1B model differentiates between the high coastal and low continental interior runoff. For grid squares 1, 3, 4 and 8 the CCM1B and observations agree well. However, the model does not seem to simulate adequately the high runoff rates observed along the south-eastern seaboard: in box 6 the model value of 80 mm y^{-1} overestimates the observation of 35 mm y^{-1} while in box 9 the prediction of only 15 mm y^{-1} is less than the 55 mm y^{-1} observed. Most of the other grid boxes are simulated adequately, although runoff in box 5 is rather too high, and runoff in box 7 is probably rather low.

The CCM1B runoff compares with that observed in this region about as well as might be expected considering the coarse model resolution. The size of the grid squares in the CCM0 and CCM1B (4.5° by 7.5°) is far too large to account for the spatial variability of the relief, precipitation and runoff. However, as the grid resolution of AGCMs becomes finer (e.g., the resolution of the UK Meteorological Office 11-layer model is 2.5° by 3.75°, Mitchell *et al.*, 1987) it might be anticipated that the simulation of the surface hydrology would improve considerably. With a finer grid resolution, the spatial variation in precipitation as a result of coastal orographic uplift might be more realistically simulated.

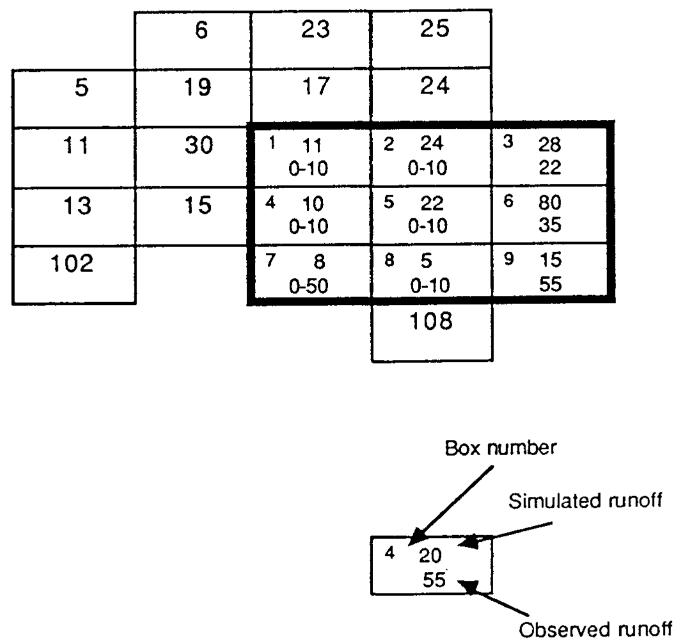


Fig. 7. Annual total runoff of east Australia simulated by the CCM1B model and observations from Australian Water Resources Council (1978). Total runoff includes surface and subsurface runoff. The flux is expressed in mm.

4. Effects of changing CO_2 concentrations on the simulation of surface hydrology

Forecasting climatic changes such as those resulting from increasing levels of atmospheric CO_2 and other trace gases in one of the practical reasons for developing AGCMs. However, investigation of CO_2 induced climate change is dependent upon the quality of simulation in areas which traditionally have been ignored by modellers, especially the climatology of the land surface.

In this section the effects of doubling the atmospheric CO_2 concentration on the surface hydrology are investigated using two models: the CCM0 (incorporating the bucket type hydrological parameterization) and the CCM1B (incorporating BATS). The same three regions will be discussed as in Section 3. Results from the CCM0 ($1 \times CO_2$) and the CCM0 ($2 \times CO_2$) are compared with results from the CCM1B $1 \times CO_2$ experiment.

4.1 Northern Canada

Figure 8a, b shows the seasonal distribution of soil moisture for the three different sets of model results. The primary effect of doubling CO_2 on the soil moisture levels simulated by the CCM0 model is to increase the wetness of the soil, but not to change the seasonal distribution of soil moisture. Surprisingly the soil moisture level in the CCM1B ($1 \times CO_2$) model are more similar to the CCM0 ($2 \times CO_2$) results, than to the CCM0 ($1 \times CO_2$) results.

Figure 8b shows the surface hydrological fluxes simulated by the CCM0 ($2 \times CO_2$) experiment which can be compared to Figure 3b (CCM0 $1 \times CO_2$). The principal hydrological response to doubling CO_2 is to increase precipitation by about 1 mm d^{-1} throughout the year. Consequently snow-melt is greater in spring, evaporation is increased due to the much wetter soils, but runoff

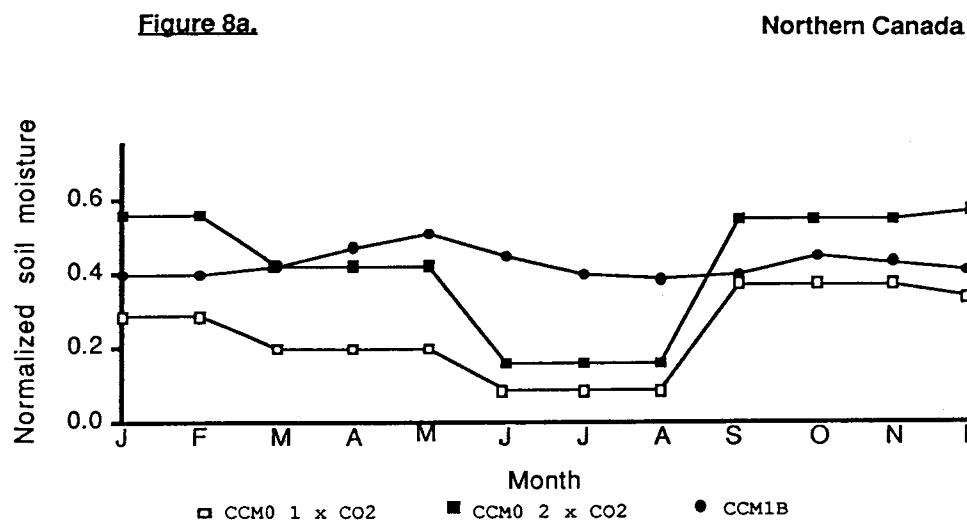


Fig. 8a. Annual soil moisture distribution for northern Canada for 1 × CO_2 and 2 × CO_2 from the CCM0 model (after Meehl and Washington, 1988), and for 1 × CO_2 for the CCM1B model. The soil moisture is expressed as a fraction of field capacity, which is normalised in the case of the CCM1B.

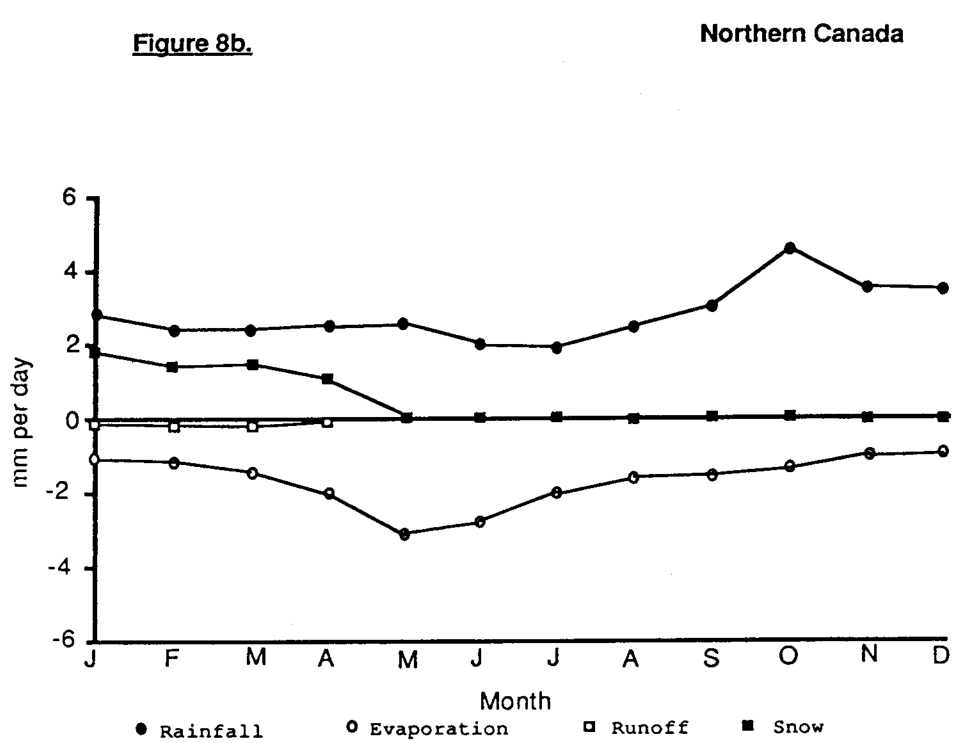


Fig. 8b. Annual variation of surface hydrological fluxes for northern Canada for the CCM0 model for 2 × CO_2. The fluxes are monthly averages, expressed in mm d^{-1} (after Meehl and Washington, 1988).

changes little (Figure 8b). In the bucket type model, incorporated in the CCM0 model, the higher soil moisture levels simulated, due to the doubling of CO_2, do not lead to increased runoff (compare Figures 3b and 8b). This is physically incorrect, since higher soil moisture levels should lead to increased runoff as a result of greater gravitational drainage and a larger proportional area of saturated soil. In the bucket model, runoff only occurs if the area average field capacity is exceeded. Thus it is evident that the simulation of the soil hydrology (in this case runoff) does not respond to a new hydrological state because the level of physical realism achieved by the bucket model is insufficient. It is also clear from a comparison of Figures 3b and 8b that overall, the differences between simulations by CCM1B and CCM0 for $1 \times CO_2$ are greater than the differences between $1 \times$ and $2 \times CO_2$.

The implication of this result is crucially important for attempts to predict the land surface response to doubling CO_2. It is clear that the differences between model simulations incorporating different land surface schemes are greater (in some regions) than the changes induced by doubling CO_2, i.e. the predicted near surface response to doubling CO_2 is probably at least partially dependent on the land surface parameterization incorporated into the AGCM.

4.2 Central North America

Figure 9a shows the soil moisture distribution for the central North America region for the CCM0 ($1 \times CO_2$) and ($2 \times CO_2$) model results, in comparison to the CCM1B ($1 \times CO_2$) experiment results.

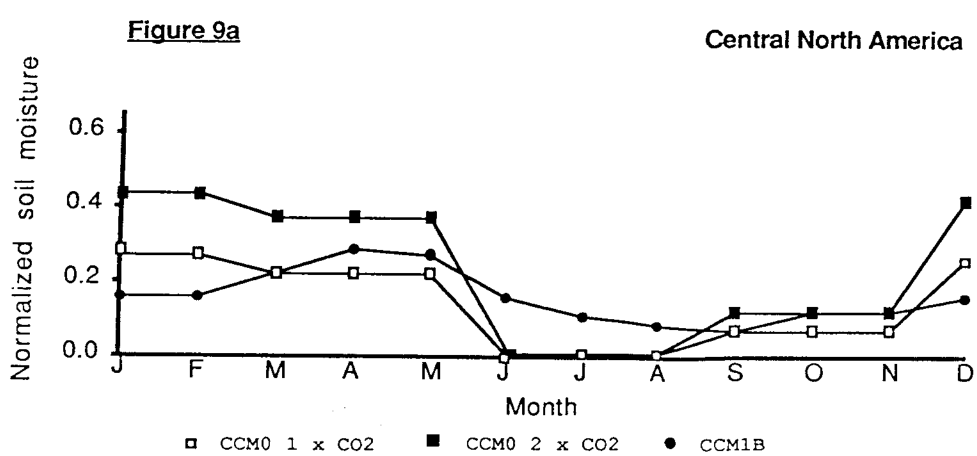


Fig. 9a. Annual soil moisture distribution for central North America for $1 \times CO_2$ and $2 \times CO_2$ from the CCM0 model (after Meehl and Washington, 1988), and for $1 \times CO_2$ for the CCM1B model. The soil moisture is expressed as a fraction of field capacity, which is normalized in the case of the CCM1B.

Doubling CO_2 leads to a much wetter soil in winter (from 0.28 to 0.42 of field capacity). However, in both cases with the CCM0 model, the soil becomes virtually dry in summer. Thus doubling CO_2 increases the seasonal range of soil moisture by about 14%. The soil moisture results from the CCM1B model compare reasonably well with the CCM0 ($1 \times CO_2$) results.

Comparison of Figure 5a with Figures 5b and 9b shows that the surface hydrological fluxes

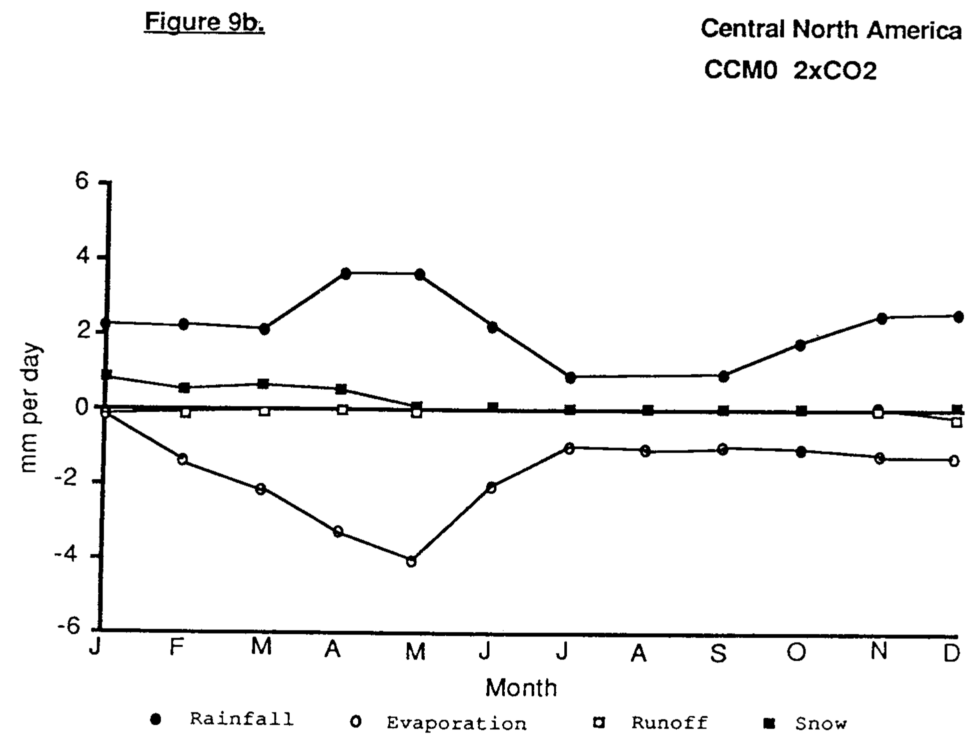


Fig. 9b. Annual variation of surface hydrological fluxes for central North America for the CCM0 model for 2 × CO_2. The fluxes are monthly averages, expressed in mm d^{-1} (after Meehl and Washington, 1988).

simulated by the CCM1B model are, in terms of magnitude and "seasonal shape" far more similar to the CCM0 (2 × CO_2) than to the CCM0 (1 × CO_2) results. For instance, the seasonal precipitation in the CCM1B model ranges between about 1.5 to 3.25 mm d^{-1}. This compares with 0.5 to 2.5 mm d^{-1} in the CCM0 (1 × CO_2) and 1.0 to 3.5 mm d^{-1} in the CCM0 (2 × CO_2) model. There is also a much closer similarity between the evaporation simulated by the CCM0 (2 × CO_2) and the CCM1B models than between the CCM1B and CCM0 (1 × CO_2) models, although this similarity is probably primarily the result of the closer agreement between the predictions of precipitation.

4.3 Eastern Australia

It was shown in Section 3 that the simulation of the hydrology of eastern Australia in the CCM0 (1 × CO_2) was poor because the soil was too wet. In contrast, the CCM1B model seemed to represent the hydrology of this region reasonably well (Figure 6a).

Figure 10 shows the effect on soil moisture of doubling CO_2 in the CCM0 model, in comparison to the CCM1B (1 × CO_2) simulation. It is clear from Figure 6a and 10 that doubling CO_2 leads to a marked ***drying*** of the soil in the CCM0 model. For all months except June, July and August, the soil moisture falls to between 50% and 65% of the CCM0 (1 × CO_2) results. The direction of this change is opposite to the change in the two North American regions presumably because the changes in large-scale atmospheric regime have caused a decrease in precipitation in this region (Washington and Meehl, 1984).

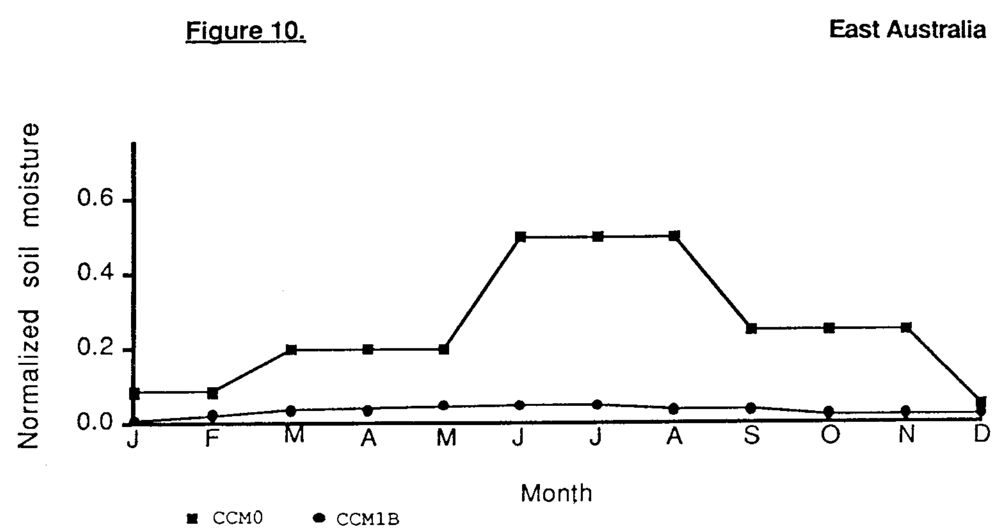


Fig. 10. Annual soil moisture distribution for east Australia for 2 × CO_2 from the CCM0 model (after Meehl and Washington, 1988), and for 1 × CO_2 for the CCM1B model. The soil moisture is expressed as a fraction of field capacity, which is normalized in the case of the CCM1B.

Most significantly, both CCM0 1 × CO_2 and 2 × CO_2 results show a marked seasonality which is not reproduced in the CCM1B results. This is another example where the differences in the surface hydrology between 1 × CO_2 and 2 × CO_2 scenarios are smaller than the differences between results generated using different land surface hydrological parameterizations.

5. Discussion

It has been shown for three regions that the CCM1B model produces different simulations of the regional scale hydrology compared to the CCM0 and GFDL models. These three simulation differ in various important ways: (i) GFDL and CCM0 incorporate a simple mixed layer ocean while CCM1B uses specified sea surface temperature; (ii) the models use different boundary and radiation schemes and GFDL specifies cloudiness and (iii) GFDL and CCM0 use bucket model surface hydrologies whereas CCM1B includes a more complete representation of the land surface. The latter surface scheme incorporates most of the features identified individually as important for the land surface hydrology e.g., a canopy evapotranspiration parameterization, increased surface runoff when the ground surface is frozen and the dependence of surface and subsurface runoff on soil texture (Mitchell and Warrilow, 1987).

The intercomparison of surface hydrology described in Section 3 shows that considerable care has to be taken if AGCMs are to be used for continental surface impact studies. It is clear that the conceptual differences between the simple bucket model and the more advanced surface schemes result in differences in the surface climate simulations. In particular, the seasonal variation in soil moisture distribution simulated by the CCM1B model is generally quite different to that of the other two models (e.g., Figures 2, 4 and 6). Rind (1988) has shown that the seasonal distribution of soil moisture is an important variable in AGCM simulations causing regional-scale feedbacks into the atmospheric component of the hydrological cycle.

With existing observational data, it cannot be proved that the CCM1B regional scale predictions are quantitatively improved by the inclusion of BATS, but the CCM1B simulations do seem to avoid many of the errors exhibited by the GFDL and CCM0 models (e.g., mid-summer desiccation (CCM0) and very wet soils year-round (GFDL), Figure 4; little seasonal variation in evaporation, Figure 3b; no spring runoff peak, Figure 3b; very wet Australian soils, Figure 6a).

This study therefore supports the inclusion of more advanced parameterizations of the land surface in AGCMs, since they appear to lead to more physically plausible simulations of the continental surface climate (although this does not mean *per se* that they produce better quantitative results).

The results described in Section 4 included comparisons between $2 \times CO_2$ with the CCM1B and $1 \times CO_2$ and $2 \times CO_2$ simulations with the CCM0. Generally, doubling CO_2 led to a wetter soil and a more active hydrological cycle. In contrast, for Australia, doubling CO_2 led to a decrease in soil moisture and, presumably, a less active hydrological cycle. This is in agreement with other regionally specific studies (e.g., Zhao and Kellogg, 1988a, b) showing that the direction (and magnitude) of the change due to doubling atmospheric CO_2 is not regionally dependent.

There is a second, but much more tentative, conclusion to be drawn from these CO_2-dependent comparisons. As the differences between the surface hydrology simulated by the different models are greater than the differences in the results from one AGCM when CO_2 was doubled, it is clear that AGCMs, or at least those which still incorporate simplistic parameterization of the land surface, may be generating both models dependent and unreliable surface hydrology fields (runoff, soil moisture, evaporation). This assertion is, of course, only true if the other differences between the AGCMs, especially (i) use of a mixed layer ocean versus specified ocean temperature and (ii) differences in the radiation and cloud schemes, are less important to the continental surface climate than the type (level of complexity) of the land surface submodel employed.

Overall, it has been demonstrated that it is possible to improve the parameterization of the land surface and the new generation of schemes perform better than the bucket hydrology. However, other components of atmospheric models are now seen to be crucial in computing the land surface climate and these remain neglected and largely unvalidated. These features include (i) the surface radiative forcing, especially as modified by clouds, (ii) regional precipitation patterns, especially the details of temporal distribution and spatial statistics and (iii) the realism of the exchanges within the planetary boundary layer (PBL) and between the PBL and the free atmosphere.

It is now essential that the answers to crucial questions concerning the greenhouse effect, as well as other climate change issues, are addressed using advanced land surface parameterizations. The more complete schemes, such as BATS, SiB and BEST already produce improved physical simulations of the land surface hydrology (e.g., Dickinson and Henderson-Sellers, 1988; Pitman, 1988; Sato *et al.*, 1989). They therefore offer much greater potential for climate impact studies. The maximum benefit from these models can only be obtained once other crucial features of global climate models are commensurately improved. Sensitivity studies and validation exercises are urgently required.

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