A model study of some aspects of soil hydrology relevant to climatic modelling

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SUMMARY

A considerable international effort is being made to improve all aspects of the physical processes represented in numerical models of climate. This particular paper is concerned with one of those aspects, namely the manner in which soil hydrology is included in such models and how it affects climate. For this purpose a 1-dimensional radiative-convective model has been coupled with various soil hydrology formulations. This simple approach has the combined advantages of permitting a self-consistent energetic evaluation of the impact of the moisture content of the soil, and of providing a very economical means for investigating the model's response to perturbations in selected variables. Three soil hydrology parametrizations suitable for use in climatic models were compared, and their individual characteristics illustrated. A method devised by J. W. Deardorff was judged to be superior and was therefore used in the subsequent experiments. These experiments were designed to clarify the role of soil moisture and its variability on the model's surface temperature and surface heat fluxes. The approach adopted permitted individual responses to be followed in great detail, and highlighted the potential role of the model for assisting in the design and subsequent analysis of appropriate experiments using climatic models. Importantly, the results illustrate the limitations of current formulations for surface hydrology used in climatic models, and the need for better parametrizations to be developed.

1. INTRODUCTION

There is an increasing awareness of the role that surface hydrology, particularly the initial water content of the soil, plays in the evolution of subsequent short-term climatic perturbations. For example Walker and Rowntree (1977) and Rowntree and Bolton (1983) have performed model experiments involving the introduction of wet and dry anomalies in the initial soil moisture content, and have found that they noticeably affected the ensuing synoptic behaviour. Importantly, the anomalies tended to perpetuate themselves. Kurbatkin et al. (1979), Shukla and Mintz (1982) and Yeh et al. (1984), amongst others, have performed similar types of experiments. While such experiments highlight the contribution of soil moisture to the climatic state, they also indicate the need to evaluate carefully the manner in which surface hydrology is incorporated in climatic models in general. A comprehensive documentation of land-surface parametrizations used in current climatic models has been given by Carson (1982), while Calder et al. (1983) have presented an assessment of empirical methods concerned with prediction of soil moisture deficits.

However, with the exception of Hansen et al. (1983), very little experimentation with hydrological formulations in climatic models, and their impacts on such models, appears to have been made. In fact the voluminous output generated in a normal climatic experiment virtually ensures that a detailed evaluation of the sensitivity and response characteristics of an individual hydrological formulation will not be attempted! In order to improve climatic simulations it is clearly desirable to assess the relative merits of surface hydrology parametrizations suitable for climatic models, and to document their responses under a variety of situations. An efficient method of implementing these aims is to restrict the problem to a 1-dimensional method, as this facilitates experimentation and interpretation over a wide range of conditions, albeit in a less realistic overall context. For this purpose a radiative-convective model has been modified to include a number of surface hydrology formulations, as this provides a simple energetically consistent vehicle for the required purposes.

This paper is subsequently concerned with comparing three surface hydrology
formulations due to Holloway and Manabe (1971), Hansen et al. (1983) and Deardorff (1977) respectively. That of Holloway and Manabe has been used almost exclusively to date in climatic models, but, surprisingly, with the assessment of its performance being confined to gross climatic comparisons. The formulation of Hansen et al. has recently been included in the Goddard Institute for Space Studies model, while that of Deardorff, which has response characteristics quite different from those of the other two formulations, has not been used in a climatic model to date. Because of the apparent superior performance of the Deardorff method its response to a variety of situations was studied, particularly those associated with extreme conditions such as drought. The use of a 1-dimensional model provided considerable insight into the behavior of the surface processes in the model in a very economical way. Ultimately it is hoped that this research will result in an improved representation of soil hydrology for use in 3-dimensional model simulations of drought, in continuation of the research reported recently by Voice and Hunt (1985).

2. Model Description

The model was a 1-dimensional radiative–convective model with a fixed relative humidity for use in the radiative calculation, as in Manabe and Wetherald (1967). It differed from their model in having 18 vertical levels extending from the surface to 37.5 km; a self-determined moist adiabatic lapse rate; and surface heat fluxes evaluated from bulk aerodynamic formulae which employed a constant velocity of 7.5 m s\(^{-1}\) at the anemometer level; and a fixed drag coefficient of 0.0015. It was assumed that the sensible plus (released) latent heat fluxes were initially entirely assimilated into the lowest model layer, which was 1.6 km thick. This heat input was then redistributed vertically via the model's convective adjustment mechanism in order to ensure that the temperature profile did not exceed the moist adiabatic lapse rate. It was also assumed that while the latent heat from the evaporating water was released no precipitation occurred locally. In reality this would correspond to the water droplets being advected away and replaced by air of equivalent heat content. No storage or transport of heat into the ground was permitted, and the surface temperature was determined from an instantaneous local heat balance of the various fluxes involved. Details concerning the fixed cloud properties, the radiative scheme, etc. have been given previously in Hunt and Wells (1979) and Hunt (1981). All results are for non-diurnal, bare soil, annually averaged conditions at 35° latitude. A surface albedo of 0.108 was used unless otherwise stated.

The sensible heat flux, \(H\), and potential evaporation rate, \(E_0\), were computed from

\[
H = \rho C_D c_p V (\theta_a - \theta_s) \tag{1}
\]

\[
E_0 = \rho C_D |V| (r_s(T_s) - r_a) \tag{2}
\]

where \(\rho\) is air density, \(C_D\) drag coefficient, \(V\) vector wind velocity, \(\theta_a\) potential temperature and \(r_a\) the humidity mixing ratio, all nominally evaluated at the anemometer level. \(c_p\) is the specific heat of air at constant pressure, \(r_s(T_s)\) is the saturation mixing ratio of water at surface temperature \(T_s\), and \(\theta_s\) is the potential temperature at the surface.

\(\theta_a\) was evaluated from the temperature at the anemometer level, \(T_a\), given by \(T_a = 0.6 T_{10} + 0.4 T_s\), where \(T_{10}\) was the temperature at the lowest model level (0.85 km). The numerical factors in the expression for \(T_a\) are based on the empirical results of Rossby and Montgomery (1935). \(r_a\) was then calculated from the saturation vapour pressure at temperature \(T_a\). An adjustment factor for conditions other than potential evaporation is given below.
For the purposes of this paper the model was extended to include surface hydrology in order to compare some simple representations either in use or suitable for inclusion in general circulation models.

(a) **GFDL method**

Most general circulation models have used the simple 'bucket' formulation first included in the Geophysical Fluid Dynamics Laboratory model by Holloway and Manabe (1971). Surface hydrology was included by assuming a field capacity of 15 cm of available moisture in the soil. Any change in the current soil moisture content, \( W \), is calculated from \( \partial W / \partial t = P - E - R \), where \( P \) is precipitation, \( E \) evaporation and \( R \) run-off. \( E \) was obtained from (2) but with \( r_s \) and \( r_e \) computed using \( e' s = e_s \cdot WETFAC \), where \( e_s \) is the saturation vapour pressure of water vapour at the appropriate temperature and \( WETFAC = 1.0 \) if \( W > 10 \) cm; \( WETFAC = (W/10.0) \) if \( W < 10 \) cm.

This procedure allowed for the reduction in the evaporation rate below the potential value \( E_0 \) as the soil dried.

(b) **GISS method**

The Goddard Institute for Space Studies general circulation model used a two-layer method to represent soil hydrology (Hansen et al. 1983)—the only such model I am aware of in which such a development has been included. In this type of formulation the upper layer is able to respond immediately to evaporation or precipitation while the lower layer acts as a reservoir. The rate of change of water in the upper layer is

\[
\frac{\partial W_1}{\partial t} = \frac{(P - E - R)}{f_1} + \frac{(W_2 - W_1)}{\tau}.
\]

\( W_1 \) and \( W_2 \) are dimensionless water contents of the upper and lower layers respectively, defined as the ratio of available water to field capacity. \( \tau \) is the time-constant for diffusion between layers taken to be one day. For the lower layer

\[
\frac{\partial W_2}{\partial t} = \frac{(f_1/f_2)(W_1 - W_2)}{\tau}
\]

where \( f_1 \) and \( f_2 \) are specified soil depths which vary with land type. For desert and rainforest, respectively, Hansen et al. quote \( f_1 \) values of 0.6 and 6 cm and \( f_2 \) values of 3 and 30 cm. The values used here were those specified for 'other land types', \( f_1 = 2.4 \) and \( f_2 = 12 \) cm. The run-off is defined to be \( R = W_1 P \) with \( W_1 \approx 1 \), while evaporation is \( E = W_2 E_0 \).

Based on information given in Hansen et al. the depth of the upper layer was taken as 10 cm, that of the lower layer 50 cm, with each layer having an assumed water holding capacity of 25%—implying a field capacity of 15 cm. Initial values of \( W_1 \) and \( W_2 \) were taken to be 1.0, indicating saturated soil.

(c) **Deardorff's method**

Deardorff (1977) proposed a two-layer method similar in principle to that developed subsequently at GISS. This approach computes the surface moisture content, \( W_G \), (in the top 1 cm of soil), as well as the 'reservoir' moisture content, \( W_B \), of the first 50 cm of soil. These variables are defined as the ratio of the volume of water in the layer to the volume of soil, and their temporal variations are given by

\[
\frac{\partial W_G}{\partial t} = -\frac{C_1(E - P)/\rho_w d_1 - C_2(W_G - W_B)}{\tau}, \quad 0 \leq W_G \leq W_{MAX}
\]

\[
\frac{\partial W_B}{\partial t} = -(E - P)/\rho_w d_2
\]

\[
E = E_0 \quad \text{WG} > \text{WSAT}
\]

\[
E = (WG/\text{WSAT})E_0 \quad \text{WG} \leq \text{WSAT}
\]
where $W_{\text{MAX}} = 0.40$ and $W_{\text{SAT}} = 0.75 W_{\text{MAX}}$. For recently irrigated soil representative values are $W_B = 0.32$ and $W_G = 0.40$. $W_B$ corresponds to a field capacity of $16 \text{ cm}$ for the selected soil depth of $50 \text{ cm}$. $\rho_w$ is the density of liquid water, $\tau = 1 \text{ day}$, $d_1 = 10 \text{ cm}$ and $d_2 = 50 \text{ cm}$. Deardorff determined $C_1$ and $C_2$ empirically:

\[
\begin{align*}
C_2 &= 0.9 \\
C_1 &= 0.5; \\
C_1 &= 14 - 22.5(W_G/W_{\text{MAX}} - 0.15); \\
C_1 &= 14; \\
WG/W_{\text{MAX}} &\geq 0.75 \\
WG/W_{\text{MAX}} &< 0.75 \\
WG/W_{\text{MAX}} &\leq 0.15.
\end{align*}
\]

Deardorff's method was designed for a diurnally varying model as substantial diurnal variations are observed in $W_G$. However, the present non-diurnal model was found to give better agreement with the limited data available to Deardorff with $d_1 = 12 \text{ cm}$, which was therefore used subsequently.

\[\text{(d) Alternative methods}\]

Numerous ways of representing aspects of soil hydrology exist, for example the empirical approach of Priestley and Taylor (1972) which was essentially confined to estimating evaporation. Calder \textit{et al.} (1983) have given a detailed comparison of a number of methods for computing soil moisture deficits. They found that the elementary prescription of evaporation in conjunction with a more detailed representation of the soil moisture profile gave the best results for their purposes. It should be appreciated that these empirical methods do not compute a self-consistent surface heat balance, as unlike the present model they do not include a radiative flux calculation. Other approaches of increasing complexity exist, for example Mahrt and Pan (1984) have experimented with a two-layer hydrologic model which requires hydraulic conductivity and diffusivity to be defined, while Sievers \textit{et al.} (1983) have developed a complicated scheme based on the hydraulic properties of the soil, involving a number of layers at which both temperature and moisture are calculated. The latter scheme, in particular, is computationally too demanding for use in a general circulation model.

Since this paper is concerned with investigating soil hydrology formulations suitable for current climatic models no attempt will be made to conduct a critical evaluation of all the possible formulations mentioned above. The discussion will be restricted to the first three methods as they represent the range of formulations either in use or likely to be developed further in the immediate future for use in climatic models. Additionally it is not intended to do a generalized comparison over the parameter space of the three surface hydrology models, as in practice many of the parameters have to be specified from observations. Instead, as presented below, after comparing the three models attention will be restricted to just one so that its behaviour and climatic impact can be assessed for a variety of applicable situations. It was considered that this information would be more valuable for the future development of soil hydrology models than a detailed intercomparison of existing models.

3. \text{Comparison of methods}\n
All model runs were started from an atmospheric temperature profile in equilibrium with saturated soil at its field capacity, unless otherwise stated. Individual runs extended over 400 model days, although typically results were presented only for the first 200 days, as this was sufficient to identify the long-term trends. Furthermore it was considered
physically unrealistic to assume that precipitationless conditions would exist over more extensive times.

Figure 1 illustrates the time evolution of four fundamental surface variables of the model using the GFDL method. Since no precipitation was permitted in these comparisons it runs the soil moisture content and the latent heat flux, or equivalently the evaporation rate, decreased monotonically with time. For all practical purposes the soil was dry throughout its depth after about 50 days. The dominant contribution of evaporation to the determination of surface temperature is quite clear in Fig. 1, with a temperature rise of about 8 K occurring as the evaporation rate declined from its potential rate to effectively zero. A compensatory rise occurred in the surface sensible heat flux, but this was insufficient to prevent the surface temperature from increasing as the combined sensible plus latent heat flux also declined to a lower value. A constant value of this combined flux was reached at about 70 days when the surface heat flux was entirely due to sensible heat exchange. A time-invariant surface temperature should then have occurred. The subsequent surface temperature decline after 70 days is attributed to the long response time of this hydraulic formulation, which created an initial extended transient state which was not sustainable once the surface heat flux stabilized and the surface temperature peaked. The subsequent decrease in the surface temperature was caused by a slow growth in the net (upwards) longwave flux at the surface associated with the long-term adjustment of the atmospheric temperature and humidity profiles to the altered, and altering, surface conditions.

The surface temperature response time in Fig. 1 is rather slow, and not particularly realistic, see Geiger (1965), because the hydrologic formulation assumes that all the soil moisture is readily available for evaporation regardless of its depth in the soil. As shown in Fig. 2 the time taken to reach the peak surface temperature with a drying surface decreases as the initial moisture content of the soil is reduced, while the maximum surface temperature attained is reasonably independent of this moisture content. Even though unsaturated surfaces are more representative of the actual world the implied response times in Fig. 2 for such conditions are still long. This suggests that this method should be used with some caution.

The corresponding time series to those in Fig. 1 for the GISS method are given in Fig. 3. Overall the behaviour is surprisingly similar to the GFDL method, apart from an initially more rapid response for the surface temperature and sensible heat flux. The reason for this similarity is attributable to the thickness of the upper layer (10 cm) and the short time-constant for transfer between the layers (1 day) in the GISS model, which maintained a close correspondence in the evaporation rates for the two methods. Although the GISS method offers hardly any advantage over the GFDL method for this particular comparison, despite its added complexity, it does have considerable scope for improvement. In fact, variation of the parameters associated with the GISS formulation could increase the differentiation between the two methods. An advantage of the GISS approach is the ability to represent different soil types and thus climatic regimes, but this particular aspect was not explored here.

Time series for Deardorff's method are illustrated in Fig. 4. The basic responses differ considerably from the other two methods, because of the separation of soil moisture into a very shallow surface layer and a deep reservoir layer. Thus the initial latent heat flux decreased very rapidly after the first five days as the surface layer dried, and the subsequent evaporation rate was determined by transfer from the underlying reservoir. Since virtually all the moisture was contained in the reservoir the soil moisture content decreased much more slowly than in Figs. 1 and 3, typically over 200 days compared to 60 days.
Figure 1. Evolution of the surface temperature, $T_s$, soil moisture content $W$ (left hand scales), latent heat flux, $L$, and sensible heat flux, $S$ (right hand scale) for the GFDL method.

The essential difference between the 2-layer models of GISS and Deardorff is the thickness of their top layers, 10 and 0.5 cm respectively, and the time-constants associated with the replenishment of the moisture in the top layer from the reservoir layer. Deardorff specifically tuned his model's performance to a high resolution data set, whereas a much more generalized approach is used in the GISS model.

The surface temperature in Fig. 4 increased very rapidly initially, by 6 K in eight days, after which it was almost constant, varying by only 0.5 K as the soil continued to dry. Contrast this with Figs. 1 and 3. The peak temperature attained was 2 K lower than in Fig. 1, as might be expected in view of the longer-term role of evaporation in this method. Similarly the sensible heat flux responded on an initially short time scale to the drying of the surface layer, and then increased steadily towards the same asymptotic value as in Figs. 1 and 3. The higher surface temperature and sensible heat flux obtained under drying conditions for all three methods agree with the observations of Ripley (1976).

This hydrologic formulation reproduced the short-term observations quoted by Deardorff (1977) quite well, but longer-term observations corresponding to the artificial precipitationless conditions employed here are not available. In this regard it should be noted that the overall response is determined to a large degree by the transfer coefficient between the two soil layers; this would not necessarily be expected to remain constant under the extreme conditions considered here.

In the remainder of the paper the model response to various perturbations will be presented. Since it would be tiresome to show the responses for each of the three

Figure 2. Peak surface temperature and time taken to reach this temperature for a drying surface for various initial soil moisture contents based on the GFDL method.
formulations, only results for Deardorff’s method will be given. Emphasis is given to this method as it was considered to be the best of the three, primarily because of its ability to reproduce the short-term variations of surface moisture, Deardorff (1977), surface temperature, Geiger (1965), and surface albedo (see below). Future soil hydrology formulations used in climatic models will undoubtedly have two or more layers in the soil, but a very shallow surface layer does appear to be a prerequisite, at least for bare soils. From this viewpoint also it was considered to be counterproductive to expand the effort documenting the rather similar responses of the GFDL and GISS formulations given their limited life expectancies.

4. **Influence of Surface Hydrology on Climate**

(a) **Basic variables**

In Fig. 5 time series for three additional model variables are plotted for the run illustrated in Fig. 4. The Bowen ratio (sensible heat flux divided by latent heat flux)
Figure 5. Evolution of the net upwards longwave flux at the surface, LW, the Bowen ratio, B (left hand scales) and combined latent plus sensible heat fluxes, F (right hand scale) for Deardorff's method.

rapidly changed from negative to positive as the surface layer dried, and within about 10 days exceeded a value of 1.0. Under the conditions of this particular run—no precipitation—the Bowen ratio steadily increased to very high values, indicating the importance of sensible heat flux as a surface cooling mechanism. Values of this ratio greater than about 1.0 would presumably identify a water-deficient stress condition. As demonstrated by Walker and Rowntree (1977) there is a considerably different dynamical impact on the atmosphere depending upon whether the heating is by latent or sensible heat fluxes. The combined (latent plus sensible) heat flux decreased rapidly in the first few days as the surface dried, but eventually attained a nearly constant value as the model adjusted to its long-term situation. The lower combined heat flux, while indicating the superiority of evaporation as a surface cooling mechanism, more importantly demonstrates the change which has to occur in the overall surface heat balance, and therefore the surface temperature, in the transition from wet to dry soil. The net longwave flux at the surface, Fig. 5, also varied rapidly in the first few days as the surface dried and the surface temperature increased. Clearly the model was attempting to counter the reduction in the combined surface heat flux by increasing the outgoing longwave flux. Although the surface temperature declined slightly after 10 days in Fig. 4 the longwave flux increased as the atmospheric temperature fell, see Fig. 6, resulting in less atmospheric water vapour, and thus greater emission to space.

The cooling of the temperature profile in Fig. 6 is somewhat surprising in view of the higher surface temperature at day 200. This arose because, while the falling evaporation rate permitted the surface temperature to increase, it also meant that less latent heat was available for release in the atmosphere.

Atmospheric temperature differences considerably smaller than those in Fig. 6 are potentially of significance in the actual atmosphere because of their ability to perturb the large-scale dynamics. For example, if cooler temperature profiles were produced over a large drought-stricken area the implied latitudinal and longitudinal temperature gradients between this area and adjacent regions would cause variations in the wind distribution according to the thermal wind equation. The latter variations could then further interact to change the overall flow characteristics in the drought-stricken area.

This basic idea was first advanced by Charney (1975) in a somewhat different context. He also sought to identify the subsequent dynamical responses which imply that the impact of a drought can extend well beyond its physical boundaries.
It is apparent from Figs. 5 and 6 that the moist static energy of the atmosphere ($c_p T + Lr + gZ$ in the usual notation) is lower over a drying surface than a moist one. This would be expected to produce less convective activity and, together with the lower absolute humidity, reduced cloud cover. The chance of precipitation would be lessened while more solar radiation would reach the surface leading to an enhancement of the drying conditions. Hence in the absence of airflow from outside the region of concern a perpetuation of drought could be expected.

(b) Surface albedo

A further aspect of drought or desertification is the change in surface albedo, $\alpha_*$, produced by a bare, drying surface, Idso et al. (1975), or due to bio-geophysical factors, Otterman (1974). The impact of the albedo change of a bare, drying surface was investigated in two experiments. It was assumed starting from saturated soil, i.e. $WG = 0.40$, $WB = 0.32$, that

$$\alpha_* = 0.108 \text{ for } WG > 0.30$$
$$\alpha_* \text{ varied linearly between } 0.108 \text{ and } 0.308 \text{ for } 0.30 > WG > 0.10$$
$$\alpha_* = 0.308 \text{ for } WG < 0.10.$$

In the second run $\alpha_*$ was assumed to range from 0.108 to 0.158 for the same WG variations. The $\alpha_*$ values for the first run were based on Idso et al. (1975), although they did not provide corresponding WG values. The second set of $\alpha_*$ values were considered to be more appropriate to vegetated surfaces.

Figure 7 compares the Deardorff control run (Fig. 4) with the two albedo variation experiments for two selected model variables. Clearly, surface albedo changes can have a marked impact in a very short time frame according to the model. In reality the model response is overestimated, particularly for the high albedo case, as other compensatory factors such as atmospheric adiabatic heating might be expected to counter the very large surface cooling obtained. It is for this reason that only the initial responses are shown in Fig. 7, the longer-term trends are apparent and actually resulted in surface temperatures
after 200 days of 272 and 292 K, respectively, for the high and low albedo cases. Nevertheless, the results in Fig. 7 are important as they indicate the magnitude of the compensatory actions which have to come into play. The halving of the sensible heat flux and the cooling of the surface by 10 K over 30 days for the high albedo case are certainly non-trivial changes. After about seven days the surface cooling was a direct response to the higher albedo, and this resulted in the absorbed solar radiation at the surface declining from 172 to 133 W m\(^{-2}\) in this time frame.

While the response time of 6–7 days to achieve the dry surface conditions associated with higher surface albedo may seem inordinately fast it compares extremely well with the observations of Idso et al. (1975). Surprisingly for the first 20 days or so there was relatively little difference in the evaporation rates and soil moisture contents between the three runs. This largely resulted from the mutual compensation between the lower surface temperature and the higher surface moisture availability in the albedo experiments. Eventually the higher surface albedos dominated the model behaviour leading to lower evaporation rates and higher soil moisture contents, by almost a factor of 50% at day 120 for the higher albedo case.

The control run in Fig. 7 indicates that a drier surface is warmer for a low surface albedo, while the high albedo variation experiment shows that a dry surface is cooler than its initial wet surface. The latter finding agrees with the results of Otterman (1974) for the Negev and Sinai where large surface albedo changes were observed, but is the opposite of that found by Ripley (1976). This difference appears to be largely attributable to the rather small albedo change, 0.15 to 0.20, which occurred in the situation considered by Ripley. For example in Fig. 7 the approximate 50% increase in surface albedo, 0.108 to 0.158 in the low albedo experiment was unable, over the 30-day period, to cancel entirely the 5 K temperature rise associated with the drying of the surface. Over a 200-day period the final temperature in the low albedo change experiment equilibrated near 292 K, which was lower than the initial temperature. Nevertheless, for modest surface albedo increases and periods of a few weeks a drying surface should be warmer than a
wet one. These results indicate the care needed in performing soil hydrology and/or surface albedo experiments with climatic models if meaningful responses are to be obtained.

(c) Relative humidity and cloud amount

An associated aspect of drying surfaces is that the relative humidity of the lower troposphere can also decline. Namias (1978) has presented values for normal and drought years for California, U.S.A. showing reductions of greater than 10% in the relative humidity during drought. Since the present model uses a defined relative humidity to compute the atmospheric water vapour content for the radiative calculations the impact of varying relative humidities can be readily investigated. A crude experiment was performed in which, after the initial surface drying phase (the first 6-7 days), the relative humidity in the lowest four model levels was reduced as follows: 45 to 43%; 53 to 45%; 61 to 50% and 77 to 54%. In the subsequent evolution it was found that the major impact was a surface cooling of over 0.6 K in the following two weeks, which increased steadily with time attributable to the increased longwave cooling. A standard climatic model computes its own changes in relative humidity so this effect would be included automatically in the model response. The point emphasized here is that because of the complexity of such models it would probably not be appreciated that the changing relative humidity might be a major cause of, say, surface temperature variations over land, possibly leading to erroneous conclusions as to cause and effect. This highlights the value of identifying the role of individual mechanisms by using the approach outlined here.

Since reducing the relative humidity might also be expected to cause a decrease in cloud amount an experiment was performed in which the low, thick cloud amount in the model was reduced to 90% of its initial value after the first 10 days. As expected a surface warming occurred but this amounted to only about 0.6 K after 10 days and 1.4 K after 50 days. The latter warming was almost cancelled by the 1.2 K cooling produced at the corresponding time by the relative humidity reductions noted above! Thus any climatic model experiments on drought or desertification should preferably be performed with a model in which cloud amount and relative humidity are computed with some accuracy.

![Figure 8. Evolution of the surface temperature, $T_*$, and sensible heat flux, $S$, using Deardorff's method for the control run with WB = 0.32 and a perturbation run with WB = 0.20.](image-url)
(d) Reservoir content

Deardorff’s method also permits studies to be made of the influence of the moisture content of the reservoir layer on the development of surface conditions. Figure 8 compares the evolution of surface temperature and sensible heat flux for the control run and a run with the reservoir moisture parameter \( WB \) reduced from 0.32 to 0.20; \( WG \) was set to 0.40 at the start of each run. The initial temperature profile was equilibrated with this \( WG \) value. As shown in Fig. 8 the surface temperature and sensible heat flux responded almost immediately to the drier reservoir layer with both variables attaining higher values than in the control. This result indicates that superficial wetting of the surface of a basically drying soil has a very transient, but noticeable impact, although it is unable to overcome the underlying moisture deficit of the reservoir layer. In contrast the GFDL formulation would not respond significantly to superficial wetting as the moisture increase has to be assigned to a single layer, where it would usually represent a very small percentage increase to the existing moisture content. Thus the temporal response of the Deardorff and GFDL formulations to small amounts of precipitation on a surface undergoing, say, a seasonal drying trend would be quite different. This again indicates the need for a careful reformulation of surface hydrology used in climatic models.

Although not shown, for simplicity, the evaporation rate for the perturbation run in Fig. 8 after only four days was \( \frac{1}{2} \) of the control value, thus accounting for the surface temperature increase. After 200 days of integration the surface temperatures for these two runs differed by only a fraction of a degree. The model sensitivity is somewhat surprising, but it indicates the importance of moisture transfer from the reservoir into the surface layer in this method. Since such transfer is a function of soil type the current model might be of use in tuning parameters for Deardorff’s method to allow for geographical variability in a climatic model. In addition to being a function of soil type the moisture transfer rate is presumably a function of soil moisture content. However, because of the lack of experimental data for comparison this facet of the model has not been explored.

(e) Precipitation

The impact of precipitation on the behaviour of the model was also investigated, Fig. 9. At day 30 of the control run it was assumed that precipitation of 3 cm d\(^{-1}\) occurred for three days, after which the standard condition of zero precipitation was maintained. The soil saturated within two days, in contrast to the situation discussed in the previous sub-section, and the excess precipitation was assumed to run off. The response to the precipitation was extremely rapid as the surface layer attained its field capacity almost immediately. This caused the evaporation rate to return to its potential value, and thus for the surface temperature to fall 7 K to just below 292 K within one day. This temperature was less than the initial, equilibrated value because of the highly transient nature of the adjustment process. The sensible heat flux became negative while the net longwave flux at the surface returned to very close to its initial value. The subsequent evolution of the evaporation rate, and implicitly the soil moisture content, paralleled that at the start of the control. However, the surface temperature attained as the surface layer dried was less than that of the control and an asymptotic approach to the control value is evident in Fig. 9. This difference was associated with the initial transient ‘overshoot’ of the surface temperature in response to the precipitation. The atmospheric temperatures experienced rather modest changes, with differences of about 0.5 K from the control tropospheric temperatures. Nevertheless, if a sufficiently large land area was affected by precipitation the resulting change in horizontal temperature gradients could influence the atmospheric dynamics.
Although the response in Fig. 9 is rather large it should be noted that bare soil conditions are implied. If the storage of heat in the soil and the variation of surface albedo with surface moisture content, see Fig. 7, were permitted somewhat different results would have been obtained.

A repeat of the precipitation experiment for much drier conditions corresponding to day 130 in Fig. 4 gave very similar results except for a slightly larger temperature drop of 8.5 K. A similar experiment with the GFDL scheme gave an almost identical surface temperature reduction but a much slower temperature rise, as would be expected from Fig. 1.

These large and rapid surface temperature changes caused by precipitation, and the related surface albedo changes for bare soils, constitute a considerable forcing mechanism in a climatic model, similar in principle to the related forcing due to synoptic sea surface temperature variations. Thus the inclusion of a soil hydrology formulation with a thin surface layer might be expected to increase the overall variability in the coupling between the land surface and the lower atmosphere in the model.

5. CONCLUSIONS

It has been shown that the combination of a 1-dimensional radiative-convective model with a surface hydrology formulation provides a cheap, efficient and convenient method for comparing such formulations, and also for developing new ones. The combined approach has the considerable attraction of explicitly computing the radiative terms required for the surface energy balance calculations, thus providing a noticeable advantage over existing empirical methods.

Three hydrologic formulations suitable for use in climatic models were compared. The much-used 1-layer GFDL formulation was found to have undesirably long response times for a drying surface, as did the 2-layer GISS method. The formulation devised by
Deardorff (1977) was preferred because of its very fast (∼6 days) initial response, which agreed well with observed soil moisture and surface temperature variations, at least for bare soils. The principal advantage of this method is associated with its specification of a very shallow surface layer which results in its fast response time. All three methods produced an increase in surface temperature caused by the drying of an initially saturated soil, the associated decrease in evaporation being compensated by an increase in the sensible heat flux.

The behaviour of Deardorff's formulation was explored for a range of situations of climatic interest. For example it was found that rather large (∼7 K) surface temperature fluctuations could be caused by precipitation followed by surface drying, which could constitute a considerable forcing mechanism in a climatic model. All else being the same a general increase in surface temperature was obtained with a drying surface, but this was associated with a cooling of the atmosphere. If such a drying was sufficiently widespread the resulting perturbations in the horizontal temperature gradients could influence the large-scale circulation. Albedo changes attributable to a drying bare surface can more than compensate for the increase in the surface temperature caused by the drying, particularly for large albedo changes. This can lead to a general cooling of the lower troposphere, which could then possibly induce atmospheric changes capable of maintaining a positive feedback, see Otterman (1974) and Charney (1975). Certainly the impact of surface hydrology on the surface albedo, as illustrated in this paper, should be included in climatic models.

These and other variations investigated clearly show that much can be learned from the type of model used here, and that such a model can be used not only in designing appropriate climatic experiments but also in their subsequent analysis. On a more applied note, the potential of the current approach for the evaluation of soil moisture deficits for agricultural purposes needs to be emphasized. With appropriate simplifications of the radiative calculations, and tuning to individual agricultural regions, it should be possible to provide a useful tool.

Finally, it is important to put the Deardorff (1977) scheme into perspective as regards future surface hydrology requirements for climatic modelling. Since it is a bare soil method tuned to one set of observations it would need generalizing for climatic purposes. The constants used in the model would be expected to vary with soil type, so special measurements would be required to derive them. However, the same problem essentially exists with any soil hydrology formulation in which allowance is made for the spatial variability of the soil characteristics. The generality of the model constants under extreme hydrologic conditions, such as severe drought, would also need to be determined. Presumably the formulation could be satisfactorily extended to cope with vegetative cover by using the approach developed by Deardorff (1978), which would make this formulation general enough for use in climatic models.

Certainly Deardorff's method has the major characteristics required for use in a climatic model in that it is simple, does not involve a large number of arbitrary parameters, has minimal storage requirements, is easy to compute, is physically based and is accurate over the timescales for which data are available. In particular the two-timescale response of the method appears to be highly desirable, although based on rather limited data, and it would ensure that greater interaction occurred between the land and the atmosphere than exists in current climatic models.
REFERENCES


