Sensitivity of a Simulated Water Cycle to a Runoff Process with Atmospheric Feedback

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Abstract

The significance of runoff processes in the atmosphere-land water cycle is investigated using an atmospheric general circulation model with simplified boundary conditions. Two runoff schemes representing two distinct runoff mechanisms, the subsurface drainage runoff and the surface saturation runoff, are incorporated separately in the model, and the simulated results are compared.

The sensitivity of the water cycle to the runoff scheme is significant in the regions of large runoff, namely the tropics, the rainy-season subtropics, and high latitudes with large snowmelt. The surface runoff scheme produces less runoff than the drainage runoff scheme when the ground is sufficiently wet and the precipitation (or snowmelt) is relatively small, and, conversely, more runoff when the ground is dry and the precipitation is large. This difference in runoff causes a systematic difference in ground wetness during the wet seasons. After that, this difference in ground wetness gradually decays due to the modification of evaporation and runoff. When the atmospheric feedback processes are neglected, the time scale of decay is estimated to be less than one month. With the atmospheric feedback, however, this time scale is significantly elongated and the difference in ground wetness persists for a few months. This modification is considered to be principally due to a strong feedback between evaporation and precipitation.

1. Introduction

The ratio of the energy used for evaporation to the radiative energy received by the land surface is strongly dependent on the wetness of the land surface. This wetness is dynamically determined by the interaction between the atmosphere and the land surface. Manabe (1969) treated this effect as part of the climate system by using a conceptual model of surface hydrology (the so-called “bucket model”) in combination with a climate model and succeeded in reproducing some basic features of the climate over the land surface. Since then, a considerable number of studies has been made on the interaction between land surface hydrology and the atmosphere by using atmospheric general circulation models (AGCMs). Shukla and Mintz (1982) compared two simulations with completely wet and dry land surface conditions and showed that evaporation from the land surface may have significant effects on precipitation, air temperature, and atmospheric circulation. The response of the atmosphere to an initially prescribed wetness of the land surface has been examined in many studies (e.g., Walker and Rowntree, 1977; Rowntree and Bolton, 1983; Yeh et al., 1984). These works suggest that a wet surface generally induces an increase in precipitation. Numerical experiments also suggest that the land surface hydrology plays an important role in controlling the natural variability of the climate near the surface (Delworth and Manabe, 1988, 1989; Koster and Suarez, 1995), and particularly, that the soil moisture as well as the snowcover may significantly control the inter-annual variation of the Indian monsoon (Barnett 1989; Yasunari et al., 1991). Milly and Dunne (1993) compared the simulated climate with different values of water-holding capacity in the bucket model and demonstrated the importance of a proper treatment of land surfaces in the simulation of intra- and inter-seasonal variations.
Many land surface schemes have been developed aiming at more realistic representations of land surface hydrology in AGCMs (e.g., Sellers et al., 1986; Dickinson et al., 1986; Abramopoulos et al., 1988). The main purpose of these studies is to represent the effects of vegetation canopy on evaporation, including transpiration through stomata, root uptake of soil water, and canopy interception of precipitation. Although there are two land surface processes that control the ground wetness, evaporation and runoff, previous studies using AGCMs usually stress the evaporation process only. (Note that we use the term “evaporation” to include the transpiration through vegetation throughout this paper.) This is probably because the runoff processes affect the atmosphere only indirectly through the change in ground wetness, in contrast to evaporation, which affects the atmosphere directly. On the other hand, the runoff process has been extensively studied in the field of hydrology. It is recognized that there are many kinds of runoff mechanisms and that the spatial heterogeneity is essential in these mechanisms (see Famiglietti and Wood, 1991, for a review). In recent years, the scope of the research in hydrology has expanded to global and continental scales, and many authors apply hydrological ideas to climate modeling (e.g., Entekhabi and Eagleson, 1989; Famiglietti and Wood, 1991; Liang et al., 1994).

However, at present, the studies from climatological and hydrological perspectives do not seem to be in harmony, as summarized by Dickinson (1992): “Climatologists have often viewed runoff as simply the residual after evapotranspiration requirements have been satisfied; conversely, hydrologists have often perceived runoff as a direct response to precipitation, with evapotranspiration a residual”. The incorporations of comprehensive schemes of hydrology do not necessarily produce better simulations than those obtained using the conventional bucket model (e.g., Stamm et al., 1994). One considerable difficulty arises from the application of basin-scale hydrological results to a much larger scale (the AGCM grid scale). Another difficulty is an incomplete understanding of the mutual interaction between runoff and other components of the climate system. A question now arises: to what extent and between runoff and other components of the climate system? The interaction between the runoff process and the climate is discussed after isolating several feedback processes. The two runoff schemes used in this study are described in Section 2, while other parts of the model and the experimental design are described in Section 3. The experimental results are shown in Section 4, while interpretations of the results are described in Section 5. Section 6 is devoted to the summary and concluding remarks.

2. Runoff mechanisms and their treatments

The occurrence of runoff is based on various mechanisms (Famiglietti and Wood, 1991). One of the important mechanisms is saturation excess runoff (Dunne and Black, 1970), which is generated by precipitation over streams or water-saturated areas around the streams. Another important mechanism is subsurface runoff, which is generated by the subsurface movement of soil water through the pores of the soil or as ground water flow.

Soil water can be classified as active or inactive soil water based on its interaction with the atmosphere. Since soil water below the wilting point of plants or that existing under the rooting depth of plants does not effectively contribute to evaporation, this soil water is considered inactive. The amount of active soil water is referred to as “ground wetness”. The two runoff mechanisms mentioned above are simply modeled as a function of the thus-defined ground wetness.

The area of saturated soil around the streams is expected to increase as the mean wetness of a surrounding region increases. If the saturated area is assumed proportional to ground wetness, the runoff due to the saturation excess can be expressed as

\[ R_s = P_e \cdot \left( \frac{W}{W_s} \right), \tag{1} \]

where \( P_e = P + S_m \) is the sum of the rates of rainfall (liquid precipitation) \( P \) and snow melt \( S_m \) reaching the soil surface and is referred to as “effective precipitation”. \( W \) and \( W_s \) are ground wetness and its saturation value, respectively. In the experiments described in Section 3, \( W_s \) is set to 300 mm, assuming a rooting depth of 1 m and a value of 30% for the saturation volumetric water content of soil above the wilting point. The runoff estimated by this mechanism will be referred to as “surface runoff”, whose schematic representation is shown in Fig. 1a. The scheme is similar to that proposed by Hansen et al. (1983).

When soil water moves below the rooting depth of plants, the water is no longer available for evaporation. The amount of active soil water (ground wetness) decreases by this process and, thus, the drained water can be regarded as runoff. The rate of this drainage depends on the speed of water movement in the soil. According to Clapp and Hornberger (1978), hydraulic conductivity (i.e., the speed of the gravitational flow of soil water) is given by

\[ K = K_s \cdot (\theta/\theta_s)^{2b+3}, \tag{2} \]

where \( K_s \) and \( \theta_s \) are hydraulic conductivity and saturated volumetric water content, respectively. \( b \) is a constant parameter, which depends on soil type and texture. This relationship can be used to describe the rate of water movement in the soil. The rate of drainage can be expressed as

\[ Q = K \cdot \nabla z, \tag{3} \]

where \( Q \) is the rate of water movement, \( \nabla z \) is the gradient of pressure head, and \( z \) is the elevation of the water surface. The amount of water drained can be calculated by integrating this equation over the area of the drainage path.
where \( \theta \) and \( \theta_s \) are the volumetric water content and its saturation value, respectively, \( K_s \) is the hydraulic conductivity at saturation, and \( b \) is a parameter dependent on soil texture. The exponent \( 2b + 3 \) varies with soil type but is generally estimated to be around ten. Based on this formula, the rate of drainage runoff is given by

\[
R_D = R_{Ds} \cdot \left( \frac{W}{W_s} \right)^n,
\]

Fig. 1. Schematic representations for the two runoff schemes: (a) surface runoff and (b) drainage runoff.

Fig. 2. Runoff rate estimated by the two runoff schemes as a function of effective precipitation and ground wetness: (a) surface runoff and (b) drainage runoff. The contour interval is 20 mm/month. The difference between the runoff rates for the two schemes is shown by a thick dotted line; the contour interval is 50 mm/month. To the upper-left of the thick solid line, drainage runoff is larger than surface runoff, while to the lower-right, surface runoff is larger.
site result, smaller surface runoff, occurs when the ground wetness is large but the effective precipitation is relatively small.

3. Model and experiments

3.1 Model description

The model used in the present study is the CCSR/NIES AGCM developed by the Center for Climate System Research, University of Tokyo and the National Institute for Environmental Studies. The model uses a spectral transformation method in the horizontal and a grid differentiation on sigma coordinates in the vertical. The resolution chosen is T21 horizontal truncation (approximately 5.6° latitude by 5.6° longitude) and 20 vertical levels. The physical parameterizations include a two-stream k-distribution scheme for radiative transfer based on Nakajima and Tanaka (1986), a simplified Arakawa-Schubert cumulus scheme (Arakawa and Schubert, 1974), prognostics of cloud water based on Treut and Li (1991), a surface flux scheme based on Louis (1979), the Mellor-Yamada level 2 turbulence scheme (Mellor and Yamada, 1982), and a three-layer heat diffusion scheme for the land surface energy budget.

The ground wetness is calculated by a simple water balance model with a single-layer reservoir:

$$\frac{\partial W}{\partial t} = P_e - E - R,$$

(4)

where $W$ is the ground wetness, $P_e$ the effective precipitation rate (rainfall and/or snowmelt reaching the soil surface), $E$ the evaporation rate, and $R$ is the runoff rate. This scheme is the same as the bucket model (Manabe, 1969) except for the treatment of runoff.

The evaporation rate $E$ is given by

$$E = f(W) \cdot E_p,$$

where $E_p$ is the potential evaporation rate and $f(W)$ is the water stress factor. $E_p$ is determined by a bulk formula:

$$E_p = \rho (q^* (T_w) - q_a)/(r_a + r_c),$$

(6)

where $\rho$ is the density of air, $T_w$ the hypothetical surface skin temperature solved with the wet-surface assumption (Milly, 1992), $q^* (T_w)$ the saturated specific humidity for $T_w$, $q_a$ the specific humidity at the lowest model level, $r_a$ the aerodynamic resistance, and $r_c$ is the canopy resistance. For the water stress factor $f(W)$, the same treatment as in the bucket model (Manabe, 1969) is used:

$$f(W) = \begin{cases} 1 & \text{if } W > W_c, \\ W/W_c & \text{if } W \leq W_c, \end{cases}$$

(7)

where $W_c$ is a critical value of ground wetness.

When snowcover is present, the precipitation or snowmelt reaching the soil surface is reduced due to runoff over the snowcover. This runoff over snowcover is excluded from the effective precipitation, $P_e$.

3.2 Experimental design

To examine the response of the atmosphere-land water cycle to the change in runoff treatment, three experiments summarized in Table 1 have been performed. The first two experiments, DRN and SFC, are AGCM runs incorporating the drainage and surface runoff schemes, respectively. Though these two types of runoff exist together in nature, each scheme is examined separately in order to clearly illustrate the sensitivity of the climate to the runoff. The experiment DRN may represent an extreme case with highly permeable soil and a flat surface, while SFC may represent another extreme with poorly permeable soil and a sloping surface. The difference between the results of SFC and DRN [SFC − DRN] can be regarded as a response of the climate to a change in the dominant mechanism of runoff from drainage runoff to surface runoff.

To clarify the roles of atmospheric and land surface branches of the water cycle in the response, the climatic variables are divided into two parts: "surface hydrological variables", which consist of ground wetness, evaporation, and runoff, and "atmospheric conditions". which include the remaining climatic variables (Note that ground temperature is not treated as a "surface hydrological variable" but included in "atmospheric conditions" for convenience.) It is the surface hydrological variables that are directly altered by the runoff processes. A change in evaporation then affects the atmospheric conditions, which in turn indirectly change the surface hydrological variables even further (i.e., atmospheric feedback).

In order to isolate the internal feedback mechanisms of surface hydrology, the third experiment, S(D), is conducted. By substituting (5) into (4), the water balance equation becomes

$$\frac{\partial W}{\partial t} = P_e - f(W) \cdot E_p - R(W, P_e),$$

(8)

which shows that atmospheric conditions for surface hydrology are contained in two variables: effective precipitation, $P_e$, and potential evaporation, $E_p$. The experiment S(D) is an off-line calculation of surface hydrology using the surface runoff scheme with inputs of prescribed time sequences of $P_e$ and $E_p$ derived from the results of DRN which used the drainage runoff scheme. The difference between S(D) and DRN [S(D) − DRN] represents the direct response of the surface hydrology to the change in runoff treatment without considering the changes in $P_e$ and $E_p$. The difference between SFC and S(D) [SFC − S(D)] represents, on the other hand, the indirect response due to the changes in $P_e$ and $E_p$. 
A similar method was adopted by Milly and Dunne (1993).

The boundary conditions for the experiments are highly idealized for the sake of simplicity. The land-ocean distribution is set to be meridionally uniform, as seen in Fig. 3: two flat continents and two oceans of the same longitudinal width, 90°, are placed alternately. As can also be seen in Fig. 3, idealized seasonal variations of sea surface temperature and sea ice distributions are prescribed. They are constructed from the zonally-averaged observed sea surface temperature. Land albedo for snow-free regions, land roughness length, and canopy resistance are set globally uniform and constant with respect to time. The parameters in (3) are tuned so that the two experiments SFC and DRN give almost the same rates of precipitation, evaporation and runoff in the global and annual average. Model parameters are summarized in Table 2.

The model is first integrated for more than ten model years from an isothermal atmosphere at rest with a uniformly wet ground. For each experiment, another four-year integration is performed for spin-up and a ten-year integration is performed for the analysis. The time step for the integration is 40 minutes for SFC and DRN, and 1 day for S(D). Owing to the hemispheric symmetry and the longitudinal periodicity of the boundary conditions, four samples of the seasonal variation for one hemispheric continent are obtained from each single year integration; forty samples for each experiment are obtained from a ten-year integration. Statistical significance is established by a t-test (Chervin and Schneider, 1976).

4. Experimental results

4.1 Global fields

The statistics of the simulated water cycles are summarized in Table 3. The annual and global mean water cycles for the experiments DRN and SFC are fairly similar because of the tuning of parameters in (3). The simulated zonally-averaged precipitation over land and corresponding observed climatology (Legates and Willmott, 1990) are shown in Fig. 4. Only the results of experiment DRN are shown, because the results of SFC are qualitatively similar. In spite of the highly idealized boundary conditions, the model reproduces the observed field fairly well except for a little-too-heavy simulated precipitation in the extra tropics.

The latitude-time sections of the zonal-mean ground wetness and runoff are shown in Fig. 5. Both are large in the tropics throughout the year and in the subtropics during summer and autumn, where heavy precipitation occurs. In high latitudes,
Fig. 4. Latitude-time cross sections of zonally averaged precipitation on land for a comparison between the AGCM experiment and an observation: (a) simulated in the experiment DRN and (b) the observed climatology by Legates and Willmott (1990). 30-day running mean is applied to (a). The contour interval is 20 mm/month. Values greater than 50 mm/month are lightly stippled and greater than 100 mm/month are densely stippled. Only the Northern hemisphere and the tropics in the Southern hemisphere are shown.

Fig. 5. Latitude-time cross sections of zonally averaged surface-hydrological variables on land simulated in the experiment DRN: (a) ground wetness and (b) runoff rate. 30-day running mean is applied to the data. The contour intervals are 15 mm for (a) and 20 mm/month for (b). In (a), values greater than 50 mm are lightly stippled and greater than 75 mm are densely stippled. In (b), values greater than 25 mm/month are lightly stippled and greater than 50 mm/month are densely stippled. Regions for detailed analysis is designated by A, B, and C.
ground wetness is large during the snow-covered season, while runoff is especially large in the snowmelt season. It is in these regions where the changes in runoff treatment possibly affects the water cycle to a considerable extent. Three regions are selected for detailed analysis, as indicated in Figs. 3, 4 and 5. The longitudinal ranges of the regions A and B are chosen so that the seasonal variations of simulated precipitation for them are similar to those observed in certain reference regions of a real continent, i.e., the western part of tropical Africa for A, and the subtropical Sahel region of Africa for B.

4.2 Tropics

In the tropical region A (Fig. 6), precipitation (solid line in Fig. 6a) is maximum in April and October with the passage of the ITCZ. Simultaneously, evaporation (dashed line), runoff (dotted line) and ground wetness (Fig. 6b) become maximum in these rainy seasons. On the other hand, potential evaporation (dotted-dashed line) is relatively small in the rainy seasons.

The differences in surface hydrological variables among the three experiments are plotted in Fig. 7. The solid lines show the difference between SFC and DRN, which represent the total responses to the change from drainage (DRN) to surface runoff (SFC). The dashed lines, the difference between S(D) and DRN, represent the direct responses to the scheme change without considering atmospheric feedback processes. Finally, the dotted lines, the difference between SFC and S(D), represent the indirect responses due to atmospheric feedback. The thick portions of the lines indicate statistically significant differences at a 95% confidence level.

The significant features of the total response (solid line) in the runoff field (Fig. 7a) are negative anomalies during the later periods of the rainy seasons (April–May, October–November) and positive anomalies during the rest of the year. This means that the surface runoff scheme results in a smaller amount of runoff than that resulting from the drainage scheme during the latter half of the rainy seasons, while the opposite is true during other
For ground wetness (Fig. 7b) and evaporation (Fig. 7c), the results for the surface runoff scheme are larger than those for the drainage scheme just after the rainy seasons, and, on the other hand, smaller at the beginning of the rainy seasons. By dividing into direct (dashed lines) and indirect (dotted) responses, it is found that the total response is essentially determined solely by the direct response. A notable exception is the evaporation in May and November, in which direct and indirect effects cancel each other.

The direct response of runoff can be understood from Fig. 2 (see the discussion in Section 2). During April–May and October–November, the soil is initially very wet due to large precipitation; however, precipitation is gradually decreasing. In Fig. 2, that condition is represented by a shift of the state to the upper-left part of the figures, where the surface runoff scheme estimates less runoff than the drainage scheme. This smaller runoff results in a larger ground wetness and then a larger evaporation with some time lag (about a month).

Though the atmospheric feedbacks are for the most part statistically insignificant, positive anomalies in potential evaporation and negative anomalies in precipitation tend to cause negative indirect anomalies in ground wetness during February–May and August–November. The negative indirect anomalies in evaporation in May and November, which almost cancel the positive direct anomalies, are caused by these negative indirect anomalies in ground wetness. In this tropical region, the changes in atmospheric conditions are generally not statistically significant, partially because of large internal variability. Thus, the indirect responses of surface hydrology play relatively minor roles, except for evaporation in May and November.

4.3 Subtropics

In the subtropical region B, there are distinct rainy and dry seasons as shown in Fig. 8. Precipitation, evaporation, runoff, and ground wetness all have their maxima during the rainy season (April–October), while they are almost zero in the dry season (December–February). Potential evaporation is relatively small in the rainy season and is exceeded by precipitation.

As shown in Fig. 9a (solid line), the runoff for the surface runoff scheme is larger than that for the drainage scheme in the beginning of the rainy

Fig. 7. Differences in hydrological variables between the experiments for the tropical region A: (a) runoff, (b) ground wetness, (c) evaporation, and (d) precipitation (P) and potential evaporation (Ep). In (a), (b), and (c), solid, dashed, and dotted lines are total response (SFC – DRN), direct response (S(D) – DRN), and indirect response (SFC – S(D)), respectively. In (d), SFC – DRN for each variable is shown. See text for the explanation for each response. A 30-day running mean is applied to the data. Periods where the difference is statistically significant at a 95% confidence level (Chervin and Schneider, 1976) are drawn with a thick line.

Fig. 8. Same as Fig. 6, but for the subtropical region B.
season (April–May) and is smaller during the rest of the rainy season (July–October). Ground wetness (Fig. 9b) and evaporation (Fig. 9c) for the surface runoff scheme are smaller than those for the drainage scheme during most of the rainy season (April–September), while they are larger at the end of the rainy season (October–November). During the middle of the rainy season (June–September), the responses are strongly affected by the atmospheric feedback; a considerable portion of the total responses in runoff and ground wetness can be explained by the indirect responses (dotted lines). A small negative anomaly (surface < drainage) in the direct response (dashed line) in ground wetness is amplified by the indirect effect, resulting in the significant negative anomaly in the total response (solid line). For the runoff, the positive anomaly in the total response during April–May becomes negative in June while the direct response remains positive until August. At the end of the rainy season and during the dry season, however, the indirect responses are generally not significant.

The behavior of the direct response during the last three months of the rainy season (September–November) can be explained in an analogous way to the behavior in the tropics (see Subsection 4b). During the rest of the rainy season, however, the opposite change occurs. Namely, the ground wetness is small but the precipitation increases rather rapidly (a shift to lower-right in Fig. 2) at the beginning of the rainy season, causing a larger runoff for the surface runoff scheme than for the drainage scheme.

It is also notable that the direct positive anomaly in ground wetness is diminished during November by a positive anomaly in evaporation. A positive anomaly in potential evaporation and a negative anomaly in precipitation (Fig. 9d) are seen in the middle part of the rainy season and represent the responses of atmospheric conditions. This precipitation anomaly is the principal cause of the significant indirect response in ground wetness. On the other hand, the indirect change in evaporation through the positive anomaly in potential evaporation is small due to a cancelation by the negative anomaly in ground wetness. A significant negative anomaly in potential evaporation exists during October–December. However, this does not result in a negative indirect anomaly in evaporation in November and December, because evaporation in this dry season is strongly controlled by remaining ground wetness rather than potential evaporation.

4.4 High latitudes

In the high-latitude region C, the atmosphere-land water cycle in the liquid phase is active only during spring to autumn, as shown in Fig. 10. The effective precipitation (rainfall + snowmelt – runoff over snowcover) has a maximum due to the
The runoff and ground wetness have their maxima simultaneously because of heavy snowmelt in this season. The potential evaporation becomes maximum in midsummer, strongly controlled by the radiative energy received by the land surface. The peak of evaporation comes in late June, between the peaks of ground wetness and potential evaporation. The effective precipitation exceeds the potential evaporation twice a year, during the snowmelt season (around May) and in the beginning of the snow-cover season (around October).

For the total response between the experiments, the surface runoff scheme estimates a smaller amount of runoff than the drainage scheme (solid line in Fig. 11a) during the snowmelt season (May–June). This is followed by significant positive anomalies (surface > drainage) in ground wetness (Fig. 11b) and evaporation (Fig. 11c). A positive anomaly in runoff is also seen during July and August. The decomposition of the response into direct and indirect responses suggests that a strong atmospheric feedback is present during June–September. Because of the positive indirect anomaly in ground wetness (dotted line in Fig. 11b) in late June to September, the total anomaly (solid line) persists longer than the direct anomaly (dashed line). The total anomaly in evaporation also persists longer but is less in magnitude than the direct anomaly owing to the compensative effect of the indirect response.

The direct responses in the snowmelt season in high latitudes is also explained by Fig. 2. The ground wetness continues to be large from the snow-cover season, while the effective precipitation due to snowmelt begins to increase but is not sufficiently large. This results in a smaller runoff for the surface runoff than for the drainage runoff, as indicated by the upper-left part of Fig. 2. In October, however, the negative anomaly in runoff is not significant in spite of the large ground wetness and decreasing precipitation, because the drainage runoff is suppressed by the freezing ground. Significant changes in the atmospheric conditions during May–August are seen in Fig. 11d. The negative anomaly in potential evaporation is especially large in magnitude and is responsible for the compensating effect of evaporation in June. The positive anomaly in precipitation, as well as this reduced evaporation, is the cause of the indirect effect of ground wetness which elongates the persistence of the anomaly. The changes in snowmelt and runoff over snowcover (not shown) are not significant.

5. Discussion

The results of the experiments indicate that the surface runoff scheme estimates a smaller runoff than the drainage scheme as a direct response when ground wetness is sufficiently large and effective precipitation is relatively small. An anomaly in ground wetness is induced by this runoff anomaly, but it tends to decay through the enhancement of evaporation and runoff. This process can be, however, strongly modified by atmospheric feedback through changes in precipitation and potential evaporation.

5.1 Processes of anomaly decay

An anomaly in ground wetness induced by the different runoff schemes tends to decay when the dependence of runoff on the schemes ceases to be significant. This can be quantitatively understood using the water balance equation (8). Let \( \delta W \) be the anomaly in ground wetness, i.e., the difference between the ground wetness for the surface and drainage runoff schemes. Assuming atmospheric feedback is negligible and the dependence of runoff on the schemes is insignificant, the time development of \( \delta W \) is described by

\[
\frac{d\delta W}{dt} = -\left( \frac{\partial S}{\partial W} \right) \delta W - \left( \frac{\partial E}{\partial W} \right) \delta W
\]

where \( \frac{\partial S}{\partial W} \) and \( \frac{\partial E}{\partial W} \) are the sensitivities of surface runoff and evaporation to ground wetness, respectively.

Fig. 11. Same as Fig. 7, but for the high-latitude region C. In (d), \( S_m \) denotes snowmelt. In (b), apparent time scales for the decay of anomaly in ground wetness is designated by \( \tau_1 \) and \( \tau_2 \) for the direct and total responses, respectively.
\[
\frac{d\delta W}{dt} = -\left(\frac{df}{dW} E_p + \frac{\partial R}{\partial W}\right) \delta W + \frac{1}{\tau_1} \delta W,
\]
where \(\tau_1\) is the decay time scale (e-folding time) of \(\delta W\).

The direct response (dashed line) of ground wetness (Fig. 11b) in July is a good example of a decaying anomaly. By substituting representative values of \(E_p \approx 100 \text{ mm/month}\), \(df/dW \approx 0.01 \text{ mm}^{-1}\), and \(\partial R/\partial W \approx 1\) month into (9), we obtain \(\tau_1 \approx 0.5\) month. This roughly coincides with the apparent decay time of the direct anomaly in Fig. 11b, demonstrating the accuracy of (9). From Fig. 11b, however, the decay of the total response (solid line) has a much longer time scale owing to atmospheric feedback processes. Considering atmospheric feedback, the time development of \(\delta W\) is described by

\[
\frac{d\delta W}{dt} = -\left[\frac{df}{dW} E_p + \frac{\partial R}{\partial W}\right] \delta W + \left[1 - \frac{\partial R}{\partial E_p}\right] \delta P_e - f\delta E_p,
\]
where \(\delta P_e\) and \(\delta E_p\) are the anomalies in effective precipitation and potential evaporation, respectively. If \(\delta P_e\) and \(\delta E_p\) are assumed to be essentially determined by the anomaly in evaporation \(\delta E\),

\[
\delta P_e = (\partial P/\partial E)\delta E + \delta S_m,
\]

\[
\delta E_p = (\partial E_p/\partial E)\delta E,
\]

where from (5),

\[
\delta E = (df/dW) E_p \delta W + f\delta E_p.
\]

Neglecting \(\delta S_m\) and substituting these expressions into (10), we obtain

\[
\frac{\partial \delta W}{\partial t} = -\left[\frac{df}{dW} E_p \left(1 - f\frac{\partial E_p}{\partial E}\right)^{-1}\right] \times \left[1 - \left(1 - \frac{\partial R}{\partial E_p}\right)\frac{\partial P}{\partial E} + \frac{\partial R}{\partial W}\right] \delta W
\]
\[
\equiv -\frac{1}{\tau_2} \delta W.
\]

The decay time scale of \(\delta W\) including atmospheric feedback is therefore given by \(\tau_2\).

Figure 12 indicates a good correlation between \(\delta P\) and \(\delta E\), and between \(\delta E_p\) and \(\delta E\) calculated from the experiments (SFC–DRN). For the high latitudes, the feedback factors are estimated to be \(\partial P/\partial E \approx 1\) and \(\partial E_p/\partial E \approx -1.5\) by linear regression. Substituting these factors and other representative values, \(\partial R/\partial E_p \approx 0.1\) and \(f \approx 0.5\), into (11), we obtain \(\tau_2 \approx 1\) month as an estimation of the decay time of the total anomaly. This is also consistent.
The decay time scales, $T_1$ and $T_2$, are plotted in Fig. 13 as a function of effective precipitation rate, $P_e$, and potential evaporation rate, $E_p$. For the feedback factors in $T_2$, $\Delta P/\Delta E = 1$ and $3E_p/3E = -1.5$ estimated above are used. The behavior of $T_1$ as a function of $P_e$ and $E_p$ is different for the two schemes in the lower-right and upper-left regions in Fig. 13. In the lower-right regions, the ground is so wet ($W_e > W_c$) that evaporation occurs at its potential rate (energy-limited evaporation). This implies that evaporation is independent of ground wetness and that $T_1$ is thus strongly controlled by the treatment of runoff. In the upper-left regions, on the other hand, the dependence of evaporation on ground wetness is significant (water-limited evaporation), and runoff plays a relatively minor role in determining $T_1$. The decay time with feedback, $T_2$, is much longer and much more sensitive to the runoff schemes than $T_1$ when the evaporation is water-limited (upper-left). This is mainly due to the precipitation feedback. The assumption of $\partial P/\partial E$ being unity implies that the anomalous recycling precipitation exactly cancels the anomalous evaporation. The anomaly in ground wetness is then determined solely by the anomalous runoff.

This consideration may also be applied to the decay of a randomly induced anomaly, i.e., the natural temporal variability. It follows that runoff may play an important role in the natural temporal variability of ground wetness when positive evaporation-precipitation feedback is significant.

### 5.2 Precipitation feedback

In the previous discussion, atmospheric conditions are regarded as a black box, whose input from surface hydrology is evaporation and the outputs are effective precipitation and potential evaporation. However, because the feedbacks through atmospheric conditions play important roles in amplifying or elongating an anomaly in surface hydrology caused by a runoff anomaly, we should further examine the processes that produce these feedbacks. The results from the subtropics (Fig. 14) and high latitudes (Fig. 15), where atmospheric feedback is significant, are examined in the following discussion.

As seen in Fig. 12, a positive anomaly in precipitation is accompanied by a positive anomaly in evaporation, and vice-versa. To examine a breakdown of the anomaly in precipitation the water budget of an atmospheric column is considered:
Fig. 14. Differences in variables between SFC and DRN (SFC – DRN) for the subtropical region B. In (a), \( P_C, P_L, -\nabla \cdot Q, \) and \( E \) denote convective precipitation, large-scale precipitation, atmospheric vapor convergence, and evaporation, respectively. In (b), \( R_S, R_L, \) and \( R_{net} \) denote shortwave, longwave and net radiation balance at the surface (positive downward). In (c), \( T_a \) and \( R_H \) denote temperature and relative humidity at the lowest model level of atmosphere. Daily mean relative humidity is estimated from the daily means of air temperature and specific humidity. A 30-day running mean is applied to the data. Periods where the difference is statistically significant at a 95% confidence level (Chervin and Schneider, 1976) are drawn with a thick line.

Fig. 15. Same as Fig. 14, but for the high-latitude region C.
\[ \frac{\partial W_a}{\partial t} + \nabla \cdot \mathbf{Q} = E - P, \]

where \( W_a \) is the precipitable water in the atmospheric column, \( \mathbf{Q} \) is the vertically integrated vapor flux, \( P \) is the precipitation reaching the surface, and \( E \) is the evaporation flux at the surface. The tendency of precipitable water is ignored here because \( W_a \) and its dependence on experimental conditions are generally small compared with the other terms. Therefore, the change in precipitation, \( P \), can be understood in terms of the changes in evaporation, \( E \), and atmospheric vapor convergence, \(-\nabla \cdot \mathbf{Q}\). On the other hand, \( P \) is modeled by two distinct mechanisms; convective precipitation, \( P_C \), and large scale precipitation, \( P_L \). The terms, \( E, -\nabla \cdot \mathbf{Q}, P_C \), and \( P_L \) are shown in Fig. 14a and Fig. 15a for the subtropical region B and the high-latitude region C, respectively.

In both the subtropics and high latitudes, the change in \(-\nabla \cdot \mathbf{Q}\) (dotted lines in Figs. 14a and 15a) is not significant except in the subtropics in November. This change appears, however, well correlated to the change in \( P_L \) (dashed lines), while the change in \( E \) (dotted-dashed lines) is well correlated to that in \( P_C \) (solid lines).

In the subtropical rainy season (May–September), the negative anomaly in \( W \) (Fig. 9) due to the change in runoff suppresses \( E \). Then this negative anomaly in \( E \) causes a negative anomaly in \( P_C \) (Fig. 14a), resulting in an amplification of the negative anomaly in \( W \). A negative anomaly in \( P_L \) accompanying a negative anomaly in \(-\nabla \cdot \mathbf{Q}\) also amplifies the negative anomaly in \( W \) in July. In the high-latitude summer (Fig. 11 and 15a), conversely, the positive anomaly in \( W \) enhances \( P_C \) through the positive anomaly in \( E \), resulting in an amplification of the positive anomaly in \( W \). In this region, the changes in \( P_L \) and \(-\nabla \cdot \mathbf{Q}\) are small and most of the significant change in \( P \) is due to the change in \( P_C \).

In the subtropical dry season, in contrast to the other seasons described above, a positive anomaly in \( E \) in vapor evaporation does not affect precipitation, but enhances atmospheric vapor divergence. This is attributed to the large-scale subsidence motion in the lower atmosphere. Thus, a positive anomaly in \( W \) is not amplified by the precipitation feedback in this season.

### 5.3 Potential evaporation feedback

As seen in Fig. 12, a negative anomaly in evaporation is accompanied by a positive anomaly in potential evaporation, and vice-versa. To estimate the change in potential evaporation, we use the following formula proposed by Penman (1948), which can be obtained by substituting the bulk formulae of evaporation and sensible heat into the surface heat balance equation:

\[ lE_p = \frac{\Delta}{\Delta + \gamma} (R_n - G) + \frac{\gamma}{\Delta + \gamma} \lambda(1 - RH) q(T_a)/(r_a + r_c), \]

where \( \Delta \equiv dq/dT \), \( \gamma \equiv (c_p/\lambda)(1 + r_c/r_a) \), \( c_p \) is the specific heat of air at constant pressure, \( \lambda \) the latent heat of evaporation, \( R_n \) the net radiation at the surface (positive downward), \( G \) the heat conduction into the ground, and \( T_a \) and \( RH \) are the air temperature and the relative humidity at the lowest model level, respectively. The first and the second terms in the right-hand side of (12) are due to the energy available for turbulence and the saturation deficit of near-surface air, respectively.

In the high-latitude summer (Fig. 15b), where a positive anomaly in evaporation is caused by the change in runoff, the downward anomaly in longwave radiation, \( R_L \), and the upward anomaly in shortwave radiation, \( R_S \), are significant. However, they virtually cancel each other in the net anomaly. Analogous results but in the opposite sign, upward \( R_L \) and downward \( R_S \) anomalies, are seen in the subtropics during the middle of the rainy season (Fig. 14b), where the anomaly in evaporation is negative. In the beginning of the subtropical dry season (Fig. 14b), however, only \( R_L \) has a significant downward anomaly. The change in shortwave radiation can be principally attributed to the change in cloud cover. On the other hand, the change in longwave radiation can be mainly attributed to either the change in cloud cover or the change in surface temperature. In the present experiment, by estimating the upward flux using the Stefan-Boltzmann law, it is found that the change in surface temperature is the dominant cause of the change in longwave radiation. The positive anomaly in evaporation in the beginning of the subtropical dry season causes surface cooling but no cloud formation owing to the large-scale subsidence. Thus it causes the downward anomaly in \( R_L \) and no change in \( R_S \).

The change in net surface radiation (the first term in (12)) is insignificant except in the subtropical dry season. Thus, the change in potential evaporation is principally due to the change in the saturation deficit (the second term). This is attributed to the anomalies in surface-air temperature, \( T_a \), and relative humidity, \( RH \) (Figs. 14c and 15c). In the subtropical rainy season, the negative anomaly in \( E \) causes a positive anomaly in \( T_a \) and a negative anomaly in \( RH \), which result in the positive anomaly in \( E_p \) through a positive anomaly in near-surface saturation deficit. In the subtropical dry season and the high-latitude summer, conversely, the positive anomalies in \( E \) result in the negative anomaly in \( E_p \) through a negative anomaly in \( T_a \) and a positive anomaly in \( RH \).
The effects of $E_p$ tend to compensate the anomalies in $E$ and shift the peaks of the anomalies in $E$ and $W$ to approximately half a month later, as clearly seen in the subtropical dry season (Fig. 9c) and the high-latitude summer (Fig. 11c).

6. Summary and concluding remarks

Numerical experiments using an AGCM with simplified boundary conditions were performed to examine the sensitivity of the atmosphere-land water cycle to the treatment of runoff processes. Two distinct runoff schemes, surface runoff and drainage runoff, are incorporated separately in the model, and the simulated results are compared.

Though the two schemes estimate almost the same runoff rate over an annual average, the estimated seasonal patterns have significant systematic differences in the tropics, the subtropics during the rainy season, and high latitudes. Because of the modeled characteristics of the two schemes, drainage runoff estimates larger runoff than surface runoff when ground wetness is sufficiently large and effective precipitation is relatively small, and vice versa. The larger runoff causes a negative anomaly in ground wetness, which is then followed by a negative anomaly in evaporation and runoff. The negative anomalies in evaporation and runoff tend to diminish the ground wetness anomaly over a time scale of less than one month. A positive anomaly caused by a smaller runoff is also diminished in a similar way.

A positive anomaly in evaporation generally causes a positive anomaly in precipitation and a negative anomaly in potential evaporation. In the high-latitude summer, convective precipitation is dominant in the precipitation anomaly, while large-scale precipitation accompanying atmospheric vapor convergence is also important in the subtropical rainy season. As for the anomaly in potential evaporation, the effect of the net surface radiation is insignificant and the near-surface saturation deficit is the principal cause. The schematic diagram for a summary of the feedbacks is shown in Fig. 16.

The anomaly in ground wetness is generally amplified and elongated through the atmospheric feedback. When the atmospheric feedback processes are neglected, the time scale of anomaly decay is typically less than one month. With the atmospheric feedback, however, the anomaly can persist considerably longer, for a few months. Particularly, the precipitation feedback (recycling), which cancels the anomaly in evaporation, is mainly responsible for this modification. In the subtropical dry season, the precipitation feedback is suppressed by the large-scale subsidence resulting in insignificant atmospheric feedback. According to an estimation using water balance equation, the decay time scale is particularly sensitive to the runoff schemes when the ground is sufficiently wet and evaporation occurs at its potential rate (energy-limited) or precipitation recycling is intense.

The magnitude of the difference in ground wetness, evaporation, runoff, or precipitation due to the different runoff treatments is as much as ten or twenty percent. Thus, to develop an accurate climate model, we should carefully consider the treatment of runoff. It is particularly important in the transition from the dry to rainy seasons and in the snowmelt season in high latitudes, where the difference in responses of runoff schemes and following feedback effects can be large.

Though the results are primarily concerned with model sensitivity, they also provide insights into the role of runoff in nature. When precipitation or snowmelt sufficiently exceeds potential evaporation so that evaporation is energy-limited, ground wetness may be strongly controlled by runoff processes. Furthermore, the time scale of natural variability in ground wetness may also be strongly affected by runoff under the condition of energy-limited evaporation or precipitation recycling. Delworth and Manabe (1988) suggested that the time scale of variability is largely determined by potential evaporation when evaporation is below the potential rate (water-limited). In the present study, however, it is suggested that the runoff process is responsible for the variability when intensive precipitation recycling is present even if evaporation is water-limited.

It is shown that the sensitivity to the runoff process is significantly altered by the atmospheric feedback processes, especially, by the feedback between evaporation and precipitation. However, this evaporation-precipitation feedback is presently not well understood and the feedback may be significantly dependent on the parameterization of precipitation processes in the model. In the analysis of the AGCM experiment of Polcher (1995), convection increases with increasing sensible heat flux. In the present experiment, however, convective precipitation generally increases with increasing evaporation (implying decreasing sensible heat). Further investigation into the nature of the evaporation-precipitation feedback is needed for a proper understanding of the land-atmosphere interaction.

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References


大気のフィードバックを考慮した陸面水循環の流出過程に対する感度実験

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大気陸面水循環における流出過程の重要性を調べるために、境界条件を理想化した大気大循環モデルを用いて数値実験を行った。地中の排水流出および地表の飽和流出を表現する 2 つの流出スキームを別個にモデルに組み込み、シミュレートされた水循環を比較した。

流出スキームに対する水循環の感度は、一般に流出量の大きい熱帯、雨季の亜熱帯、高緯度で大きかった。地表が十分湿潤で降水または融雪が比較的小さいとき、地表流出のスキームは排出流出のスキームに比べて流出を小さく、逆の条件では大きく見積る。この流出量の差により地表湿度の系統的な差がもたらされる。しかし、流出量のスキームによる違いが顕著でなくなると、地表湿度の差は蒸発と流出のフィードバックにより減衰する。大気のフィードバックが無いとすると、この減衰の時間スケールは 1 ヶ月より短いと見積もられる。しかし、大気のフィードバックの効果によって地表湿度の差は数ヶ月程度持続する。この過程で蒸発-降水のフィードバックが特に重要な役割を果たしていると考えられる。