Groundwater-supported evapotranspiration within glaciated watersheds under conditions of climate change

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Abstract
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Keywords
Groundwater flow, Evapotranspiration, Richard's equation, Climatic change, Crow Wing watershed

Disciplines
Atmospheric Sciences | Climate | Environmental Indicators and Impact Assessment | Fresh Water Studies | Hydrology

Comments

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Groundwater-supported evapotranspiration within glaciated watersheds under conditions of climate change

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Abstract

This paper analyzes the effects of geology and geomorphology on surface-water/groundwater interactions, evapotranspiration, and recharge under conditions of long-term climatic change. Our analysis uses hydrologic data from the glaciated Crow Wing watershed in central Minnesota, USA, combined with a hydrologic model of transient coupled unsaturated/saturated flow (HYDRAT2D). Analysis of historical water-table (1970–1993) and lake-level (1924–2002) records indicates that larger amplitude and longer period fluctuations occur within the upland portions of watersheds due to the response of the aquifer system to relatively short-term climatic fluctuations. Under drought conditions, lake and water-table levels fell by as much as 2–4 m in the uplands but by 1 m in the lowlands. The same pattern can be seen on millennial time scales. Analysis of Holocene lake-core records indicates that Moody Lake, located near the outlet of the Crow Wing watershed, fell by up to 4 m between about 4400 and 7000 yr BP. During the same time, water levels in Lake Mina, located near the upland watershed divide, fell by about 15 m. Reconstructed Holocene climate as represented by HYDRAT2D gives somewhat larger drops (6 and 24 m for Moody Lake and Lake Mina, respectively). The discrepancy is probably due to the effect of three-dimensional flow. A sensitivity analysis was carried out to study how aquifer hydraulic conductivity and land-surface topography can influence water-table fluctuations, wetlands formation, and evapotranspiration. The models were run by recycling a wet year (1985, 87 cm annual precipitation) over a 10-year period followed by 20 years of drier and warmer climate (1976, 38 cm precipitation). Model results indicated that groundwater-supported evapotranspiration accounted for as much as 12% (10 cm) of evapotranspiration. The aquifers of highest hydraulic conductivity had the least amount of groundwater-supported evapotranspiration owing to a deep water table. Recharge was even more sensitive to aquifer hydraulic conductivity,

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1. Introduction

Atmospheric scientists have demonstrated that spatial variations in soil moisture influence atmospheric processes such as cumulus convective rainfall (e.g. Pielke, 2001). While global circulation models are yet to include the effects of groundwater hydrodynamics on soil-moisture conditions, simplified representations of groundwater and surface-water flow, such as the statistical–dynamical method of TOPMODEL (e.g. Wigmosta and Lettenmaier, 1999; Beven, 1997), have been incorporated into surface–vegetation–atmosphere transfer schemes (e.g. Famiglietti and Wood, 1994a,b; Stieglitz et al., 1997) and mesoscale atmospheric models (York et al., 2002; Gutowski et al., 2002). Shallow water tables can sustain high soil-moisture levels, enhancing evapotranspiration (referred to herein as groundwater-supported evapotranspiration; Fig. 1). To date, however, none of the above studies has rigorously represented the dynamics of evapotranspiration and unsaturated/saturated flow under conditions of long-term climatic change.

Regions of high soil moisture, such as wetlands, are characterized by areas where the water table is close to the land surface. Within most glaciated watersheds, these sites generally occupy areas of low topography. In addition to topography, stratigraphic variations across a watershed can control the height of the water table. The purpose of this paper is to assess quantitatively how climatic change and geology influence the height of the water table, wetlands, lake levels, and groundwater-supported evapotranspiration within glaciated watersheds.

![Fig. 1. Schematic diagram illustrating groundwater-supported evapotranspiration (areas where plant roots can access the water table directly and by wicking) during wet and dry climatic periods. During wet conditions, the water table encroaches on the land surface in topographic depressions within the uplands. The flow system has components of both local and regional flow. During dry climatic periods, the water table falls below local topographic depressions in the uplands, significantly reducing the amount of groundwater-supported evapotranspiration. The characteristic length of local and regional flow systems (L) is also indicated.](image-url)
Toward this end, we present a new quasi-two-dimensional numerical model of unsaturated/saturated flow that fully couples the surface–subsurface hydrologic cycle (HYDRAT2D). The model is validated in part by Bowen-ratio measurements of evapotranspiration from a field site on the edge of the Crow Wing watershed in north-central Minnesota. Next, we present generic numerical experiments using climatic and hydrologic conditions of the Crow Wing watershed to guide selection of model parameters. Hydrologic and paleolimnologic data from this watershed help to demonstrate that aquifer water levels and lakes in different regions of the watershed respond to climatic forcing differently on both historical and geological time scales.

Finally, we present a field application of the model to reconstruct evapotranspiration and watertable conditions for the mid-Holocene. In the remainder of this paper, we describe the present-day and Holocene hydrology of the Crow Wing watershed, the transport equations and evapotranspiration scheme used in our model HYDRAT2D, and a sensitivity study using the model in which climatic, topographic, and hydrogeologic parameters are varied. A comparison between present and Holocene lake levels and evapotranspiration for a cross-sectional model of the Crow Wing watershed is given. The implications of our study to atmosphere/land-surface interactions are then discussed.

Fig. 2. Location of lakes and wetlands (gray) across the Crow Wing watershed. County names are listed in italics. The location of selected wells (circles) and lakes (black polygons) is also presented. Numbers near wells indicate well ID. For example, Well 1 in Cass County is thereafter referred to as Cass-1.
2. Study area

The Crow Wing watershed in central Minnesota is situated within a climatically sensitive region of the north-central USA. The 120×150 km study area is part of the upper Mississippi River watershed (Fig. 2). The watershed is characterized by end moraines along its western and northern boundaries, and till plains and outwash sands in the remainder. Total thickness of glacial deposits ranges from 30 to 180 m (Winter, 2001). The glacial deposits consist largely of till, a heterogeneous mixture of clay through boulders. Glacial outwash sand and gravel deposits of limited extent are also present within the till, and sand and gravel outwash deposits cover the surface of the watershed over about half of its area (Lindholm, 1970). The hydraulic properties of the outwash sands have been measured by over 20 aquifer tests. Hydraulic conductivity values measured by single and multi-well aquifer tests performed at the northern edge of the watershed within the outwash sands ranged from 9.15×10⁻⁷ to 1.72×10⁻³ m s⁻¹ (Filby et al., 2002).

Typical of young glaciated terrains, the area is characterized by hummocky topography, including many depressions that contain lakes (about 1% by area) and wetlands (about 8.5% by area, Fig. 2) with no surface-water outlets. Total relief across the watershed is about 120 m; local relief in most of the area varies by less than 20 m. The inferred water table for the Crow Wing watershed is shown in Fig. 3 and cross-sections in Fig. 4. The regional water table presented in Figs. 3 and 4 is based on a calibrated free-surface groundwater flow model (MWT3D) described by Corbet and Knupp (1996). As part of the model calibration exercise, 3037 water-table elevation measurements from wells drilled across the Crow Wing watershed and reported by the US Geological Survey (http://nwis.waterdata.usgs.gov/mn/nwis/gwlevels) were used. Residual distribution (measured minus modeled heads) across the watershed showed no apparent bias. The average residual was −1.3 m.

The study area is located in a climatic transition zone, where air masses from the Arctic, Pacific, and Gulf of Mexico interact (Bradbury and Dean, 1993). Average monthly temperatures range between −13 and +18 °C. About 75% of the annual precipitation, which averages about 70 cm, is returned to the atmosphere by evapotranspiration (Baker et al., 1979). Average streamflow for the Crow Wing River at its confluence with the Mississippi River is about 28 m³ s⁻¹ (Lindholm et al., 1972). While most precipitation occurs primarily in summer (Fig. 5A), groundwater recharge occurs in the spring as a result of snow melt prior to the growing season.

![Fig. 3. (A) Geology of the Crow Wing watershed, MN. Then gray lines are surface watershed divides. (B) Contour map of calculated water-table elevations of the Crow Wing watershed. Positions of cross-sections A–A’ and B–B’ presented in Fig. 4 are also shown.](image-url)
Climatic change has occurred at different time scales. During the 1930s drought conditions in the Midwestern United States, lake levels and water tables declined in western Minnesota by up to 5.1 m, with all but the deepest lakes going dry (Donovan et al., 2002). In the subsequent decade, lake levels recovered their normal conditions (Donovan et al., 2002) as wetter climatic conditions returned. Analysis of modern climate record for Minnesota suggests that climatic fluctuations persist for at most 3–4 years before changing (Fig. 5B).

Historical lake and water-table records across the watershed indicate that the magnitude of water-table fluctuations in wells (Fig. 6A) and lakes (Fig. 6B) are sensitive to the position within the watershed. The largest annual water-table and lake-level fluctuations (2–3 m) occur in the upland portions of the watershed. Fluctuations within the low-lying regions of the watershed exhibit higher frequency and smaller amplitude (less than 1 m). High-frequency fluctuations in low-lying regions are also due to the rapid infiltration through thin unsaturated zones.

Pollen-based paleoclimate reconstructions indicate that the middle Holocene was dry and warm in Minnesota over a much longer time scale (over 3000 years). For example, Locke (1995) used pollen-transfer functions described by Bartlein and Whitlock (1993) to reconstruct Holocene mean January and July temperatures and mean annual precipitation for the Shingobee River watershed with pollen data from Williams Lake sediment cores. The Shingobee watershed lies on the northern border of the Crow Wing watershed. The results shown in Fig. 7 indicate that within the interval from 9500 14C yr BP to present the precipitation was at its lowest point 7700 14C yr BP, about 25 cm less than today, and that mean July temperatures were 4°C warmer than today and January temperatures were about 3.5°C higher (Fig. 7). Buried layers of littoral sand in Williams Lake (Locke, 1995; Filby et al., 2002) indicate lake levels 2.5–4.5 m lower during the warm dry period (4000 and 7700 14C yr BP) dominated by prairie vegetation.

Changes in Minnesota lake levels by as much as 15 m in response to mid-Holocene climatic change were also inferred by paleolimnological data and Holocene paleohydrologic numerical models of the
Elk Lake Chain (Donovan et al., 2002; Smith et al., 2002). In the present study, Lake Mina in the upper part of the Crow Wing Watershed and Moody Lake in the lower part were investigated by transects of sediment cores from shallow to deeper water to explain the presence of sand layers buried in the fine organic sediments. The sand layers are interpreted as shore-sand deposits at times of low lake levels, subsequently buried when lake levels rose again. The pollen stratigraphy indicates that the sand layers were formed during the mid-Holocene dry period. Eolian deposition was rejected as an alternative explanation because of the difficulty of transporting sand across water in summer, or across ice in winter when frozen ground or a snow cover must have prevailed in potential source areas. Deposition during strong runoff events was rejected because the lakes have no entering streams in the area of transects, and the sand layers do not have textural features common to turbidity flows. These results suggest that the level of Lake Mina at the head of the watershed was lowered by as much as 15 m in the mid-Holocene (Fig. 8) and at Moody Lake near the basal part of the watershed by about 4 m (Fig. 9).

Fig. 6. Long-term (1925–2002) water-table fluctuations in wells (A) and lakes (B) across the Crow Wing watershed. As with water-table elevations, lake levels in lower portions of the watershed display smaller amplitude and higher-frequency fluctuations relative to wells located in higher positions within the watershed. The response time of lake levels to wetter climate following the dry conditions of 1923–1938 appears to be slower than later climatic events. The location of these lakes and wells is shown in Fig. 2.
3. Model

To quantitatively study the effects of climatic changes on water-table position, recharge, and evapotranspiration, we constructed a numerical model (HYDRAT2D) that couples the saturated and unsaturated zones and includes evapotranspiration. The model follows the work of Yakirevich et al. (1998): a series of vertical one-dimensional columns representing flow in the unsaturated zone are coupled to a one-dimensional horizontal saturated flow model (see Fig. 10). Unlike the study of Yakirevich et al. (1998), however, our model includes the effects of soil evaporation and plant transpiration on soil-moisture redistribution by adding a sink term in Richards equation (see Appendix A for details). The model assumes vertical flow in the vadose zone and the Dupuit assumption of horizontal flow in the saturated zone. Water-table elevation, evapotranspiration, and recharge are calculated as a function of time and position along the landscape (see Appendix A).

To validate in part the model’s calculation of evaporation rate, we compare model results with Bowen-ratio measurements of actual evaporation rate from a field site near Williams Lake, which lies on the north-central edge of the Crow Wing Watershed. Table 1 shows values of parameters used in the computation. Measured evaporation for 1999 (corrected for missing days) was 0.80 m while the simulated value is 0.88 m. This 10% difference is due in part to the differences between actual radiation near Williams Lake (incomplete measurements) and radiation data used in the model (1990 hourly solar-radiation measurements from St Cloud, MN, located about 145 km southeast from the study site). Calculated recharge was 0.37 m a$^{-1}$ or 32% of surface application. Evapotranspiration accounted for 68% of total precipitation, in accord with measurements and with the record from Minnesota (Baker et al., 1979). Our model also reproduces the increase in groundwater recharge during spring as a result of snow melt. Water mass-balance errors over the cross-section did not exceed 1%.

4. Sensitivity study

In this section, HYDRAT2D is applied in a sensitivity study to understand the effects of long-term climatic change on lake-level fluctuations, wetlands preservation, runoff, and groundwater-supported evapotranspiration. An idealized landscape, 5 km long, delimited on the left by a lake and on the right by the watershed divide, was used to represent a
hummocky topography. A hill and a valley separate the lake from the divide (see Fig. 11). The height of the lake is kept constant through time while at the divide the no-flux boundary condition is imposed. The total land-surface elevation change is 35 m (average slope 0.7%). Aquifer hydraulic conductivity was varied over three orders of magnitude and the specific yield was set to 0.1. The topographic variations across the model domain are representative of many glaciated regions in the Midwestern United States.

Climatic forcing for the simulation involved selecting 2 years of climatic data, one wet, one dry. We chose the climatic record of St Cloud, MN, USA, located 80 km southeast of the Crow Wing watershed, for its extensive record from 1948 to 2003. Climatic input to the vadose-zone model includes smoothed hourly precipitation, temperature, radiation, and relative humidity. Smoothing hourly precipitation was necessary to eliminate numerical oscillations associated with hourly time-steps. Soil and vegetation properties were identical to those used in the validation models (see Table 1).

For the sensitivity study, the model was spatially discretized laterally into 31 equally spaced nodes, with 31 vertical columns at each node. In each vertical column, node spacing was 2 cm resulting in the deepest columns having about 3000 vertical nodes. This discretization also produced mass-balance errors.
of less than 1%. The model was run for 10 years under ‘wet’ climatic conditions followed by 20 years of ‘dry’ climatic conditions. We used the 1985 (0.87 m of precipitation) and 1976 (0.38 m of precipitation) climatic records to represent ‘wet’ and ‘dry’, conditions, respectively. Note that 1976 was drier than during the mid-Holocene (0.5 m of precipitation). The initial water-table elevation used in the model increased linearly from the lake to 20 m below the water divide (see Fig. 11).

Fig. 11 shows the elevation of the water table and the amount of annual evapotranspiration as a function of spatial position along the landscape for three values of aquifer hydraulic conductivity. Fig. 11A shows the
Table 1
Parameters for soil and vegetation used in the model validation and sensitivity study

<table>
<thead>
<tr>
<th>Soil</th>
<th>Plant</th>
</tr>
</thead>
<tbody>
<tr>
<td>$K_h = 4.47 \times 10^{-6}$ m s$^{-1}$</td>
<td>Growing season: May 1 to October 1</td>
</tr>
<tr>
<td>$\theta_{sat} = 0.3$</td>
<td>LAI = 3</td>
</tr>
<tr>
<td>$\theta_r = 0.0633$</td>
<td>Albedo = 0.2</td>
</tr>
<tr>
<td>$\alpha = 3$</td>
<td>Vegetation height = 2 m</td>
</tr>
<tr>
<td>$n = 2.5$</td>
<td>Root depth = 0.5 m</td>
</tr>
<tr>
<td></td>
<td>Limiting point = $-3$ m</td>
</tr>
<tr>
<td></td>
<td>Wilting point = $-140$ m</td>
</tr>
</tbody>
</table>

Unsaturated soil properties were specified according to van Genuchten’s hydraulic function (van Genuchten, 1980). $K_h$ is the aquifer hydraulic conductivity, $\theta_{sat}$ is the saturated water content, $\theta_r$ is the residual water content, and $n$ and $\alpha$ are the empirical constants used in van Genuchten’s model. Leaf-area index (LAI) and albedo values are consistent with a grassland-type vegetation (Dingman, 2002). Values for the limiting and wilting points are taken from Feddes et al. (1978).

Fig. 11. Water-table elevation and evapotranspiration (ET) for three different aquifer hydraulic conductivities. (A) End of wet period and (B) end of dry period.
result at the end of the wet period; Fig. 11B after 20 years of ensuing dry conditions. At the end of the wet season, evapotranspiration for all three cases of hydraulic conductivity is around 0.8 m a\(^{-1}\). Spatial variation in evapotranspiration is small (less than 5%) with highest evapotranspiration occurring for the case of smallest hydraulic conductivity where the water table is closest to the land surface. Because of high precipitation, soil moisture near the surface is high and plant evapotranspiration is not groundwater-supported. Evapotranspiration is independent of the thickness of the vadose zone. After 20 years of dry conditions, a relatively thick (greater than 10 m) unsaturated zone forms under high grounds for even the lowest conductivity system. Evapotranspiration drops and shows greater spatial variability. Because of lower precipitation, plants now rely more heavily on groundwater. Thus, evapotranspiration is highest where the vadose zone is thinnest beneath the valley and near the lake, and lowest beneath the water divide and the hill. This difference in evapotranspiration between thick and thin vadose zones is about 17% for the lowest and intermediate hydraulic conductivities. For a high hydraulic conductivity aquifer, the shape of the water table is controlled by lateral water movement: low water-table elevation causes plant evapotranspiration to be relatively independent of spatial position.

Fig. 12. Temporal changes in water-table elevation for (A) water divide, (B) valley, and (C) hill, for three different aquifer hydraulic conductivities.
Temporal variations in the water table beneath the watershed divide, the valley, and the hill are shown in Fig. 12A–C, respectively, for three different aquifer hydraulic conductivities presented in Fig. 11. Steady state is reached quickly for the high hydraulic conductivity. From the initially sloped water table, only 3 years are needed to equilibrate during wet conditions (years 0–9). About 20 years are necessary when switching from wet to dry conditions (years 10–30). For the two lower hydraulic conductivities, steady-state conditions take longer to attain and are not always reached during the simulations (e.g. below high grounds). During the wet climate, the water table in the case of small and intermediate aquifer hydraulic conductivity rises continuously. Slow lateral water movement, in comparison to high infiltration rate, causes the increase in water-table elevation. In the valley, the water table eventually intersects the land surface and cannot rise higher. Only during the summer, plant evapotranspiration draws the water table down by about 1 m.

When the dry period begins, the water table drops for all cases. Interestingly, the water table below the valley drops faster for the case of lowest conductivity than for the intermediate conductivity (Fig. 12B). For intermediate conductivity, lateral water transfer brings water from beneath the high grounds toward the valley and keeps the water table high there. For the lowest conductivity, lateral groundwater flow is too slow in comparison to surface evaporation, and thus water level initially drops more rapidly there. These results highlight the trade-off in low-lying areas between strong lateral flow sustaining high water levels and weak lateral flow favoring evapotranspiration and contributing to a lowering of the water table.

Seasonal fluctuations in water levels are largest when the water table is close to the land surface. The dampening of seasonal oscillations in the water table as it drops is well illustrated in the record of low and intermediate conductivities beneath the valley (Fig. 12B). As the vadose zone thickens, the precipitation signal is dissipated, resulting in water-table fluctuations that are dampened.

Temporal variations in average annual evapotranspiration and recharge for the three different aquifer hydraulic conductivities are presented in Fig. 13. For all cases, evapotranspiration is nearly constant to about 0.81 m a\(^{-1}\) (Fig. 13A) during the wet period (years 0–9). During the dry period (years 10–30), the average annual evapotranspiration drops for all cases:
the largest decline occurs within the most conductive aquifer (about 0.16 m a\(^{-1}\) or about 20%). The change in evapotranspiration from wet to dry climate is due to groundwater-supported evapotranspiration induced by shallow water tables. Interestingly, the lower hydraulic conductivity aquifer, while transferring water more slowly to the valley has the highest levels of evapotranspiration during the dry climate. This is because the water table remains shallower under low-conductivity conditions (see Fig. 11). Higher conductivity materials have deep water tables and thus less groundwater support.

Recharge (Fig. 13B) during both wet and dry conditions varies dramatically with changes in hydraulic conductivity due to the influence of watertable shape (Appendix A, term \(q_v\) in Eqs. (A5) and (A7)). Concave water table beneath the valley during the wet season causes water movement toward the valley, which translates into upward water movement or negative recharge. When the water table reaches the land surface, this becomes runoff. The phenomena is amplified in the case of low hydraulic conductivity at the beginning of dry conditions: low precipitation and slow lateral water movement keep the shape of the water table concave and cause large negative recharge. As the water table declines, this effect diminishes. Negative recharge when the water table is pinned at the land surface (runoff) discharges to the stream network (not represented).

5. Model of crow wing watershed during the mid-holocene

A 60-km cross-sectional model (Fig. 14) was developed for the Crow Wing watershed along section B–B’ (Fig. 3) using HYDRAT2D to assess whether pollen-based paleoclimate reconstructions for the mid-Holocene (Fig. 7) could predict the paleowatertable configuration inferred from sediment-based lake records (Figs. 8 and 9). The model was first calibrated using present-day water-table elevations across the watershed (B–B’, Fig. 4) and historical 1951–1980 climatic data for the watershed (Olcott, 1992). Best agreement between computed and observed water levels was obtained using an aquifer hydraulic conductivity of \(3.8 \times 10^{-4} \text{ m s}^{-1}\) and a specific yield of 0.1 consistent with measured values for glacial outwash sand within the watershed. Using pollen-based estimates of mid-Holocene, climate produced a water-level drop of 24 m the water divide (section B–B’, 0 km, Fig. 4) and 6 m in the lowland (55 km) if precipitation was 70% of the modern average annual precipitation and temperature 4°C warmer than today, consistent with the paleoclimatic reconstruction for the mid-Holocene. Calculations indicate that evapotranspiration was 28% lower during the mid-Holocene (0.55 m a\(^{-1}\)) than it is today (0.77 m a\(^{-1}\)). Variations in evapotranspiration were larger during the mid-Holocene because of...
greater variations in the thickness of the vadose zone. Recharge, however, did not vary significantly across the model domain as almost all of precipitation is returned to the atmosphere via evapotranspiration. The simulated adjustment from present to mid-Holocene climate took about 250 years.

The calculated drop in water levels for the mid-Holocene for cross-section B–B' in Fig. 14 is consistent with but not identical to the observed values. The larger simulated drop in water levels may be due to the inability of our model to account for three-dimensional effects.

6. Discussion and conclusions

Our findings that lakes and wetlands respond differently across watersheds in response to climatic change have implications for atmospheric sciences, paleolimnology, and watershed management. Analysis of historical lake-level and groundwater-level data indicates that the response of surface-water bodies to climatic change is controlled in part by groundwater hydrodynamics and position within the watershed. Lakes and wetlands located at higher elevations within the watershed have a larger response of water-level fluctuations to a change in climate than lakes located in the lowlands (Almendinger, 1989; 1993). For example, Lake Mina located high in the Crow Wing watershed dropped 15 m in response to the mid-Holocene warm period. In contrast, Moody lake, located near the confluence of the Mississippi and Crow Wing rivers, experienced a drop of 4 m in lake level. This hypothesis is also supported by an analysis of hydrologic records across the Crow Wing watershed. We found that modern water-table levels measured between 1970 and 1993 and lake levels recorded between 1924 and 2002 indicated that larger-amplitude fluctuations occur within the upland portions of watersheds.

Our results indicate that different lakes within a watershed will provide a different response to climate. Results of our sensitivity study indicate that groundwater-supported evapotranspiration varied with topography and aquifer hydraulic conductivity. Higher amounts of groundwater-supported evapotranspiration (about 10 cm a⁻¹ or 12% of total evaporation) occurred in lowlands and when the hydraulic conductivity of the aquifer was relatively low. Larger-scale flow systems (15 km instead of 5 km) also produced more groundwater-supported evapotranspiration, since this resulted in the water table being held at higher elevations. The highest levels of groundwater-supported evapotranspiration occur for intermediate hydraulic-conductivity conditions, which optimize lateral transport of water to lowlands.

These results suggest that a small yet important feedback exists between groundwater and atmospheric processes on decadal to longer time scales. There is a clear need to develop regional and global atmospheric circulation models that couple these processes (e.g. Gutowski et al., 2002).

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Appendix A. Model description

A.1. Governing equations

In each vertical column (see Fig. 10), we solve the one-dimensional Richards equation (Freeze, 1969)

\[ C(\psi, z) \frac{\partial \psi}{\partial t} = \frac{\partial}{\partial z} \left[ K(\psi, z) \left( \frac{\partial \psi}{\partial z} + 1 \right) \right] - S_{\text{root}}(z), \]

\[ b \leq z \leq H, \tag{A1} \]

where \( \psi \) is the pressure head, \( z \) is the vertical direction pointing upward, \( t \) is the time, \( C(\psi, z) \) and \( K(\psi, z) \) are the specific moisture capacity and the hydraulic conductivity, respectively, \( S_{\text{root}}(z) \) is a sink term representing water uptake by roots, \( H \) is the land-surface elevation, and \( b \) is the bottom elevation of the vertical column below the water table. Depth-varying soil properties can be included by making \( C \) and \( K \) dependent on \( z \). Eq. (A1) applies to both unsaturated
and saturated conditions \((C(\psi, z)=0\) and \(K(\psi, z)=K_{\text{sat}}\)).

At the land surface \((z=H)\), the application rate (or flux), \(\mathcal{A}\), is
\[
\mathcal{A} = R + M - E_{\text{soil}},
\]
where \(R\) is the rain, \(M\) is the snow melt, and \(E_{\text{soil}}\) is the actual soil evaporation (defined later). The soil infiltration capacity, \(C\), is
\[
C = \left| K(\psi, H) \left( \frac{\partial \psi}{\partial z} + 1 \right) \right|.
\]

At a time step \(j+1\), if \(\psi^j < 0\) or if \(\psi^j = 0\) and \(\mathcal{A} < C\), we apply the flux boundary condition (assuming no ponding)
\[
\frac{\partial \psi^{j+1}}{\partial z} = -\left( 1 + \frac{\mathcal{A}}{K} \right).
\]
If \(\psi^j = 0\) and \(\mathcal{A} \geq C\), we fix the pressure head at zero, i.e. \(\psi^{j+1} = 0\).

Thus, during a rain event, a flux condition is imposed as long as the pressure head at the surface is negative. When the pressure head reaches the value of zero, the boundary condition will remain a flux condition if soil infiltration capacity exceeds application rate. If not, the pressure head is fixed at zero. At the end of a rain event, when water has drained, soil infiltration capacity becomes higher than the application rate, and the boundary condition switches back to a flux boundary condition.

As in Yakirevich et al. (1998), we obtain the vertical flux, \(q_v(x)\), at the bottom of the unsaturated zone by vertically integrating the two-dimensional saturated flow equation over the depth of the aquifer. Assuming no sources nor sinks and no flow through the bottom of the aquifer, this integration yields
\[
q_v(x) = K_h \frac{\partial}{\partial x} \left[ (h - \eta) \frac{\partial h}{\partial x} \right],
\]
where \(h\) is the elevation of the water table, \(\eta\) is the elevation of the bottom of the aquifer, and \(K_h\) is the hydraulic conductivity of the aquifer. Thus, the flux boundary condition applied at the bottom of the unsaturated zone is
\[
\frac{\partial \psi}{\partial z} = -\left( 1 + \frac{q_v}{K_{\text{sat}}} \right),
\]
where \(K_{\text{sat}}\) is the saturated vertical hydraulic conductivity. This boundary condition should in principle be applied at the phreatic surface \(z=h\). However, as done in Yakirevich et al. (1998), we apply it at the bottom of the unsaturated model, \(z=b\) (bottom of vertical column), which is below the phreatic surface. This procedure allows us to solve Richards equation in a fixed domain and avoid cumbersome numerical procedure to solve a moving boundary-value problem.

Given boundary conditions at the land surface and at the bottom of the vertical columns, the solution of Richards equation yields the pressure-head profile, \(\psi(z)\) at each column from which the elevation of the phreatic surface, \(h\), can be calculated as the elevation, where \(\psi(z)=0\). The recharge to the phreatic surface obtained by comparing Boussinesq equation (e.g. see Bear, 1972) and Eq. (A5) is
\[
R = S_y \frac{\partial h}{\partial t} - q_v,
\]
where the first term depends on the specific yield, \(S_y\), and is due to the vertical motion of the phreatic surface while the second term comes from the actual vertical flux of water through the phreatic surface. This definition of recharge agrees with that of Freeze (1969) which defines recharge as the sum of ‘water made available at the water table surface’ (\(q_v\) in Eq. (A7)) and ‘flow away from the water table within the saturated zone’ (term \(S_y \partial h / \partial t\) in Eq. (A7)). Appendix B outlines the numerical algorithm used to solve this system of equation.

### A.1.1. Evapotranspiration

Evapotranspiration is calculated from the sink-term method of Feddes et al. (1978). A volumetric sink term in the unsaturated-flow equation represents water uptake by roots \((S_{\text{root}}(z)\) in Eq. (A1)). This term depends on the potential evapotranspiration computed using the model of Priestley and Taylor (1972). Actual evaporation, which depends on leaf-area index (LAI), the vapor pressure at the soil surface and in the air above the soil, and the saturation vapor pressure at the soil surface, was calculated with the method outlined in Feddes et al. (1978).
A.1.2. Snow-cover calculation

Our goal is to represent long-term ground-water recharge and water-level fluctuations in the saturated zone. Hence, we use a simple snow-cover model: if air temperature falls below the freezing point, precipitation falls as snow and the surface flux is set to zero; if the air temperature is above freezing, we use a degree-day method to calculate snow melt \( m \)

\[
m = m_T,
\]

where \( T_a \) is the temperature above freezing and \( m = 3.6 \text{ mm d}^{-1} \text{ °C}^{-1} \) (Dingman, 2002) is the melt coefficient. The snow melt is added to the total amount of precipitation to compute the application rate at the soil surface. The application rate \( A \) is set to zero if \( T_a \) is negative (i.e. frozen soil). This approach is simpler than heat budgets at the snow surface and produces results that are sufficiently accurate for our purpose.

Appendix B. Numerical algorithm

Because of the non-linearity of Richards equation and because the model couples the solutions of Richards equation at several vertical columns through the flux applied at the bottom of the columns, an iterative scheme is needed to reach convergence of the pressure-head field at each column consistent with the prescribed boundary fluxes. Our algorithm does not differ significantly from that of Yakirevich et al. (1998), but we describe it in this section for the sake of completeness.

At a new time step \( j + 1 \) we take, as an initial guess for the pressure head at each column, the solution of the pressure head from the previous time step \( j \), i.e. \( \psi_{0,j}^{j+1} = \psi^{j} \). We then follow an iterative procedure indicated by an index \( k \):

1. From the pressure head profile at the previous iteration, \( \psi_{0,j}^{j+1,k} \), we compute the phreatic surface elevation \( h_{j}^{j+1,k+1} \) as the vertical position, where \( \psi_{0,j}^{j+1,k+1} = 0 \) for each column.
2. We compute the vertical flux \( q_{b,j}^{j+1,k+1} \) from Eq. (A5) using a centered finite difference approximation. We also compute the application rate and use these two fluxes as boundary conditions for Richards equation, again at each column.
3. We then solve Richards equation and obtain a new pressure profile \( \psi_{j}^{j+1,k+1} \).
4. Steps 1 – 3 are iterated until the pressure head profile at each column has converged.

Richards equation is solved in its mixed form by means of the modified Picard iteration method of Celia et al. (1990). A fully implicit centered finite-difference approximation is used to discretize the equations. In cases where convergence is difficult (as when a dry soil receives precipitation), we switch to an explicit scheme for computing the pressure head of the top surface node. In this explicit scheme, the volumetric water content of the surface node \( \theta_N \) at time step \( j + 1 \) is set to

\[
\theta_N^{j+1} = \theta_N^j - \Delta t \left( \frac{\partial q}{\partial z} \bigg|_{N} + S_{\text{root}} \right),
\]

where \( \Delta t \) is the time-step increment. The newly calculated water content at \( j + 1 \) is not allowed to exceed 0.1 times the saturated water content of the soil.

References


