Mid-Holocene Hydrologic Model of the Shingobee Watershed, Minnesota

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Abstract
A hydrologic model of the Shingobee Watershed in north-central Minnesota was developed to reconstruct mid-Holocene paleo-lake levels for Williams Lake, a surface-water body located in the southern portion of the watershed. Hydrologic parameters for the model were first estimated in a calibration exercise using a 9-yr historical record (1990–1998) of climatic and hydrologic stresses. The model reproduced observed temporal and spatial trends in surface/groundwater levels across the watershed. Mid-Holocene aquifer and lake levels were then reconstructed using two paleoclimatic data sets: CCM1 atmospheric general circulation model output and pollen-transfer functions using sediment core data from Williams Lake.

Calculated paleo-lake levels based on pollen-derived paleoclimatic reconstructions indicated a 3.5-m drop in simulated lake levels and were in good agreement with the position of mid-Holocene beach sands observed in a Williams Lake sediment core transect. However, calculated paleolake levels based on CCM1 climate forcing produced only a 0.05-m drop in lake levels. We found that decreases in winter precipitation rather than temperature increases had the largest effect on simulated mid-Holocene lake levels. The study illustrates how watershed models can be used to critically evaluate paleoclimatic reconstructions by integrating geologic, climatic, limnologic, and hydrogeologic data sets.

Disciplines
Atmospheric Sciences | Climate | Fresh Water Studies | Hydrology

Comments

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Key Words: Groundwater water; paleohydrology; pollen transfer functions; paleolimnology; climate change.

INTRODUCTION

Lake-core records and atmospheric global circulation models (GCMs) have been extensively used to reconstruct past climatic conditions across continents (e.g., COHMAP Members, 1988). Climatic reconstruction methods based on lake-sediment cores have utilized temporal changes in the assemblages of aquatic plant macrofossils, ostracodes, diatoms, pollen, and chironomids, as well as sediment geochemistry and sediment facies information to infer changes in lake levels, lake salinity, precipitation, and temperature (e.g., Bernabo and Webb, 1977; Fritz et al., 2000). These methods provide an independent means of
HYDROLOGIC MODEL, SHINGOBEE WATERSHED, MINNESOTA

FIG. 1. Location of Elk Lake and the Shingobee watershed (approximately 46°57′30″N, 94°42′W) in Minnesota. Location of lakes, streams, and wells within the Shingobee watershed are shown. Average (1990–1998) lake levels and water-table contours (where known) are presented in meters. "USGS"-United States Geological Survey; "U of MN"-University of Minnesota.

estimated select climatic variables. Taken alone, however, each method provides limited insight into overall climatic conditions. For example, reconstructions of past climate based on sediment facies data provide reconstructions of paleo-lake levels but fail to provide direct information regarding precipitation or temperature. Transfer functions developed using faunal or pollen assemblages, on the other hand, provide estimates of summer/winter paleo-temperatures and paleo-precipitation but do not provide insights on changes in lake levels, extreme climatic events, or other important climatic variables such as humidity. GCMs provide more information on a wide range of climatic variables than can be gained from lake-core records. However, GCMs utilize land-surface parameterization schemes that do not incorporate small-scale (1–10 km) features of the hydrologic cycle such as streams, lakes, and aquifers, owing to their relatively coarse-spatial scale (typically 400 × 400 km). Thus, it has been difficult to make direct comparisons between GCM paleo-climatic reconstructions with those obtained from lake-sediment corse.

Physically based, distributed-parameter hydrologic models developed at the watershed scale can overcome some of the limitations of the above-mentioned approaches by representing the paleo-hydrologic cycle at relatively fine spatial and temporal resolution. They allow for the quantitative evaluation of whether various paleoclimatic reconstructions are self-consistent. In this study we present a paleo-hydrologic model for the Shingobee watershed (Fig. 1) in north-central Minnesota that incorporates streams, lakes, and subsurface aquifers at a horizontal spatial resolution down to 10 × 10 m. We use this model to critically evaluate paleoclimatic reconstructions for the mid-Holocene dry period (6000 14C yr B.P.) in Minnesota by comparing the position of the beach sands, pollen-transfer functions, and GCM model output.

HOLOCENE CLIMATIC RECONSTRUCTIONS OF MINNESOTA

The climatic history of Minnesota has been reconstructed from lake core records from different regions of the state. These indicate that the middle Holocene was dry and warm. One of these records comes from varved sediment cores extracted from Elk Lake (e.g., Dean et al., 1984), which is near our study area (Fig. 1). Bartlein and Whitlock (1993), using pollen records extracted from these cores, found that the middle Holocene in Minnesota was characterized by annual precipitation of 100 mm less and temperatures of about 2.0°C warmer than present.

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 watershed using pollen data from Williams Lake sediment cores in the Shingobee watershed (Fig. 1). The results indicate that within the interval of 9500 14C yr B.P. to present the precipitation was at its lowest point 7700 14C yr B.P., about 250 mm less than today. In addition, mean July temperatures were 4°C warmer than today and January temperatures were about 3.5°C higher. Locke (1995) used sediment-facies data to infer a drop in Williams Lake level between 2.5 and 4.5 m during the warm dry period (4000 and 7700 14C yr B.P.) dominated by prairie vegetation in the mid-Holocene (Fig. 2).

The middle Holocene has also been noted as a time of decreased regional precipitation (e.g., Wright et al., 1963) and increased summer insolation as calculated by paleo-GCM experiments (COHMAP Members, 1988). In this study, we utilized output from the paleo-GCM model CCM1 data (Kutzbach et al., 1998) from a single grid cell (ca. 400 × 400 km) which overlies the Shingobee Watershed in Minnesota.

STUDY AREA

The Shingobee watershed in north-central Minnesota (Fig. 1) is part of the larger upper Mississippi River watershed. Surface
water bodies in the watershed include 11 lakes and 10 stream segments. It is the site of the U.S. Geological Survey (USGS) Interdisciplinary Research Initiative (IRI), which was established to advance the understanding of the hydrologic process through lakes (Averett and Winter, 1997). As a result of this effort, Winter (1997), Rosenberry et al. (2000), and others have studied the watershed’s hydrology, biology, chemistry, and geology.

The Laurentide Ice Sheet repeatedly overran Minnesota during the Pleistocene, leaving highly heterogeneous sediment and glacial landforms. The Shingobee watershed is characterized by hummocky topography and lies near the convergence of two major glacial landforms in the region, the Itasca and St. Croix moraines, which were formed during the St. Croix phase of the Wisconsin glaciation (Wright, 1972).

Lakes have accumulated deposits throughout the Holocene (Wright et al., 1993). Within the watershed, Crystal Lake and Williams Lake are the only major water bodies without input or output streams. As a result of their position, farthest from the outlet of the watershed, these lakes should be most sensitive to changes in climate (Almendinger, 1989).

Recharge during the summer is negligible, because evapotranspiration exceeds available moisture from precipitation (Baker et al., 1979). Average daily and annual precipitation and temperature as well as annual runoff are presented in Figure 3. Within the Shingobee watershed, the average annual surface water runoff is about 24% of precipitation. The surface water and groundwater fluctuations are in phase within the watershed, which implies a good connection between these two reservoirs.

**MATHEMATICAL MODEL**

The Shingobee watershed has been the topic of several prior modeling studies (Karls, 1982; Locke, 1995; Gerla, 1999). For this study, a one-dimensional evapotranspiration/unsaturated flow/heat transfer model for the vadose zone was constructed to calculate temporal variations in recharge to the aquifer. Soil-moisture redistribution in the model was approximated with an implicit finite-difference scheme that relates pressure head to unsaturated hydraulic conductivity and specific capacity (e.g., Selker et al., 1999). Evapotranspiration is represented using a modified Penman-Montieth equation (Dingman, 1994). The calculated recharge from this one-dimensional vadose-zone model was used as input to a three-dimensional surface water/groundwater model (MODFLOW; McDonald and Harbaugh, 1988) to calculate lake and aquifer levels.
MODFLOW input data sets were generated using the ArgusONE™ preprocessor with a modified Graphical User Interface developed by Richard Winston (unpublished, 1999) that incorporates the MODFLOW lake (Cheng and Anderson, 1993) and stream (Prudic, 1989) packages. The model was discretized to a 34 × 71 cell grid (each cell is approximately 10 × 10 m) with 8 vertical layers, each approximately 14 m in depth.

The lateral and basal model boundaries are all no-flow boundaries except for the northern boundary. The northern boundary is specified as a constant-flux boundary as a result of the steep gradient of the water table into the watershed at this boundary. For this reason, a conservative flux of 6.12 × 10⁻⁵ m³ s⁻¹ was determined from Darcy’s Law and applied to the Well Package in MODFLOW at this boundary. The outlet of the watershed (Shingobee River out of Shingobee Lake) was specified at 405.7 m, which is the only fixed head in the model.

MODEL CALIBRATION TO THE SHINGOBEE WATERSHED

The one-dimensional vadose-zone model was calibrated using hourly time steps for period between 1990 and 1998, chosen for the availability of climatic data. Climatic input to the vadose-zone model includes precipitation, maximum daily temperature, radiation, temperature, dew-point temperature, relative humidity, wind speed, and snow depth. These data were available from the USGS meteorological stations within the watershed. During periods when hydrologic data were missing, they were estimated using observed data from the nearby Alexandria Airport. Leaf-area index (LAI) was assigned using published values (Dingman, 1994; maximum LAI of 4 for northern hardwood forest, 3 for grasslands). Albedo levels were varied based on land cover for new snow, old snow, grass, and trees (e.g., Dingman, 1994). The sediment type used for the recharge calculation in the unsaturated zone was homogenous sandy silt, which is characteristic of the sediment in the Shingobee watershed area. The relationships for soil moisture and conductivity as a function of pressure head were found experimentally for the soil and incorporated into Richards’ equation (Richards, 1931) by linear interpolation.

The depth of the unsaturated zone varies significantly within the Shingobee watershed. Adjusting the depth of the unsaturated zone used in the model influenced the lag time between snow melt/precipitation and recharge at the water table. A water-table depth of 15 m was found to produce the best agreement between simulated and observed lake and aquifer water levels. Yearly average recharge in the area has been estimated to range from 75 to 350 mm yr⁻¹ based on groundwater-model calibration to lake and water-table levels by a number of previous investigations (Karls, 1982; Locke, 1995; Gerla, 1999).

The model was calibrated based on the magnitude and timing of recharge and its associated effect on simulated lake and aquifer levels. Simulated values were compared to observed water levels in monitoring wells and lakes. While this approach can yield a nonunique solution (varying either the hydraulic conductivity or the recharge rate can produce a good fit to aquifer and lake water-level data), numerous transient aquifer tests provide some independent constraints on the magnitude of hydraulic conductivity for the study area. Single and multiwell aquifer tests in and around the University of Minnesota well field, located in the southern portion of the watershed (Fig. 1), resulted in a measured range of hydraulic conductivities from 9.15 × 10⁻⁰⁷ to 1.72 × 10⁻³ m s⁻¹.

Calculated recharge for the Shingobee watershed was found to be 241 mm yr⁻¹ (Fig. 4). This represents about 30% of average annual precipitation, a reasonable value when compared to surface water runoff estimates from Bartlein et al. (1984). The peaks in recharge correspond to spring snowmelt events. The highest peak in recharge is attributed to high observed snow depth in 1997 in the study area (Fig. 4).

As part of the groundwater/surface water model calibration observed lake levels, representative water-table levels from five wells, and Shingobee River stream flow were compared with simulated values. The locations of the five wells throughout the watershed and the Shingobee stream gauge used in the calibration procedure are shown in Figure 1. Shingobee and Williams Lake records were evaluated for the period of 1990–1998.

The parameter varied in the model calibration exercise which had the greatest impact on simulated surface water and groundwater levels was the hydraulic conductivity of the aquifer. Glacial materials in particular are difficult to characterize because of the complex depositional environments and associated permeability heterogeneity of the sediments (Winter, 1975). The model best matched observed data with three main hydraulic conductivity zones, which varied from south to north across the watershed. The zones are consistent with the north-to-south trend of coarse sand, fine sand, and silt sediments. For the purpose of this study, the till at the base of the aquifer system is considered to be the bottom of the aquifer, as the conductivity of the till is at least two orders of magnitude greater than that of sand, essentially removing it from the system (Freeze and Witherspoon, 1967).

Since the lakes appear to be in good connection with the aquifer,
lake-bed sediments were given the same hydraulic conductivities as the aquifer in which the lake is located (Fig. 5).

Observed and calculated water levels for five wells in the watershed were in good agreement (Fig. 6). Simulated fluctuations in water-table levels of about 1 m were largely in phase with observed variations during the model-calibration period. The poor fit between the observed and modeled values of well SW25 is probably due to spatial heterogeneities in hydraulic conductivity that are not represented in the model and the close proximity to the boundary between hydraulic conductivity zones. The simulated and observed spatial water-table patterns agree well (Fig. 7). The areas where lakes occur correspond to the regions of extremely low water-table gradients in Figure 7.

Simulated lake levels for Williams Lake (which has no surface-water inflows or outflows) agree with observed levels (Fig. 8). However, the agreement between simulated and observed Lake levels is not as good for Shingobee Lake; a lake with surface water inflows and outflows. In addition, the calculated average annual Shingobee Lake outlet discharge is $2.34 \times 10^6$ m$^3$ yr$^{-1}$, which is low compared with observed average annual surface water discharge of $3.5 \times 10^6$ m$^3$ yr$^{-1}$.

**FIG. 5.** Cross-section showing the vertical extent of hydraulic conductivity layers used in MODFLOW model of Shingobee watershed. Insert shows lateral extent of hydraulic conductivity zones and location of cross section.

**FIG. 6.** Observed (dashed) and calculated (solid) water levels in wells from 1990 to 1998. The locations of the wells are shown in Figure 1. Note that the vertical axis is not continuous between individual well plots.

**FIG. 7.** Average (1990–1999) calculated (solid) and observed (dashed) water-table elevation. The contour map is based on water levels measured in 23 wells in the southern portion of the study area. The locations of the wells are shown in Figure 1.
Lake levels and outflow at the outlet correlate poorly with the observed data for Shingobee lake for many reasons. To begin, the observed stream-discharge data are measured downstream of the outlet of Shingobee Lake and the Howard Lake tributary joins the Shingobee River before the USGS gauging station (Fig. 1). No observed data exists for this tributary, accounting for a portion of the error. In addition, numerous beaver dams exist along the river and at the outlet of the lake and have a large effect on the stream flow and muting the lake level response to climate (Dallas Hudson, personal communication, 1999). As a result, matching Shingobee Lake levels was not a primary goal of the model calibration exercise. However, the simulated and observed surface-water fluxes had the same order of magnitude.

RECONSTRUCTION OF PALEOHYDROLOGY AND RESULTS

The calibrated hydrologic model was used to reconstruct paleohydrologic conditions in the Shingobee watershed for the mid-Holocene. Two sources of precipitation and temperature data were evaluated for the middle Holocene in this study. Both the CCM1 output and pollen data of Locke (1995) provide estimates for precipitation and temperature for 6000 $^{14}$C yr B.P. However, the vadose-zone model needs additional climatic input data. The vadose-model requires leaf area index, rooting depth, temperature, precipitation, wind speed, snow depth, radiation, and humidity.

Snow depth was found by converting precipitation to snow when the temperature was below 0°C. The leaf area index for the mid-Holocene was estimated to be that for grass, and the rooting depth was estimated to be about 1 m (Locke, 1995). It is not possible to predict humidity and wind speed for the paleoclimatic forcing; hence, we assumed modern values for these variables. This assumption is a potential source of error.

Webb et al. (1987) and Harrison et al. (1998) describe two approaches for reconstructing past climatic forcing: the direct and the anomaly approach. In the direct approach, the actual output of temperature and precipitation from the GCM is used to drive the watershed hydrologic model for the control (modern) and the 6000 $^{14}$C yr B.P. simulations. We rejected the direct approach because the CCM1 modern run resulted in unrealistic declines in simulated aquifer and lake levels. Instead, following the “anomaly” approach of Webb and Bartlein (1987), we superimposed the differences between the CCM1 modern and scenario (i.e., the 6000 $^{14}$C yr B.P. CCM1 climate forcing; Fig. 9) data sets on the observed modern climate record (1990–1998) to obtain the paleoclimatic forcing for 6000 $^{14}$C yr B.P. Recharge rates from the vadose zone model which incorporated this paleoclimatic forcing were used as input to MODFLOW to reconstruct paleo-aquifer and paleo-lake levels. The results of
FIG. 10. Calculated Williams Lake level (solid line) using recharge from regression analysis in the anomaly method, run for 160 yr. The dashed line indicates estimated mid-Holocene lake level of Locke (1995) inferred from elevation of beach sand. About 100 yr are required for the hydrologic system to equilibrate to mid-Holocene climatic conditions.

The “anomaly” approach show little change (≤0.05 m) in simulated Williams Lake levels between the modern and 6000 14C yr B.P.

Locke used a method described by Bartlein et al. (1984) and Bartlein and Webb (1985) to predict precipitation, January temperature, and July temperature for the middle Holocene drought. Changes in pollen abundances indicate that annual precipitation 6000 14C yr B.P. was about 250 mm less than today (about 62% of modern precipitation). In addition, January temperatures were about 3.5°C warmer and July temperatures were about 4.0°C warmer than present. For consistency, the changes in precipitation and temperature were applied in our model using the anomaly approach of Harrison et al. (1998). Modern precipitation was decreased by 62% and temperatures increased 3.5°C in the winter and 4.0°C in the summer. The calculated paleo-recharge from the vadose-zone model, using Locke’s (1995) results as input, had an average value of $4.63 \times 10^{-9}$ m s$^{-1}$.

When the monthly average recharge values produced by the GCM output and pollen-transfer functions were applied to MODFLOW, Williams Lake dropped 3.5 m (Fig. 10). This lake-level drop in Williams Lake is within the range of values predicted by examining paleo-shoreline migration in Williams Lake (2.5–4.5 m; Fig. 2). Inspection of Figure 10 indicates that more than 100 yr are required for the calculated lake levels to equilibrate to the new climatic conditions.

The calculated recharge based on the two paleoclimatic reconstructions were quite different from each other for the Shingobee watershed (Table 1). A sensitivity study using our unsaturated zone model was carried out in which temperature and precipitation were independently varied to determine their relative effects on calculated recharge. These parameters were varied over a range of uncertainty indicated by the different paleoclimatic reconstructions presented in this study. Results suggest that the most important factor in aquifer recharge and paleo-lake levels is the winter precipitation. Changing precipitation with fixed temperature resulted in a much more significant change in recharge of $6.16 \times 10^{-9}$ m s$^{-1}$ (two orders of magnitude). Changes in summer precipitation rates had only a minor effect on calculated recharge because plant transpiration could utilize almost all of the additional water delivered to the root zone. Changes in temperature (holding precipitation fixed) only resulted in a $4 \times 10^{-11}$ m s$^{-1}$ change in recharge.

**DISCUSSION**

The value of the approach presented in this study is that we were able to quantify the impact of various paleoclimate reconstructions on the hydrologic cycle directly by reconstructing paleo-lake levels and then comparing them to an independent record of climate change, lake shoreline stratigraphic records. Watershed-scale mathematical modeling could also be used to critically evaluate paleoclimatic reconstructions based on isotopic and geochemical proxies. For example, isotopic composition of ostracodes from Shingobee and Williams Lake cores reported by Locke and Schwab (1997) show a variation that is linked to the changes in pollen assemblages. Although paleoecologists often use lake chemistry and salinity to reconstruct climate (e.g., Fritz et al., 2000; Laird et al., 1996), the relationships between these proxies and annual temperature and rainfall are not well understood. The isotopic composition of lakes can be reconstructed for the Holocene by adding advection-dispersion isotope transport and lake fractionation equations (Cross et al., 2001) to watershed hydrologic models.

While there is some concern in using GCM and pollen-based paleoclimatic data sets to drive a watershed-scale hydrologic model, our results are consistent with more regional studies. For example, Webb et al. (1998) showed that CCM1 paleoclimatic data for the Eastern USA did not represent the mid-Holocene drought based on their evaluation of regional, pollen-based reconstructions of precipitation and temperature for that time period.

**CONCLUSIONS**

In this study, a high-resolution, physically based, groundwater/surface water model was used to reconstruct paleo-recharge, lake levels, water-table configuration, and stream flow for the Shingobee watershed in north-central Minnesota. Hydrologic parameters were first estimated in a calibration exercise using hydrologic and climatic data between 1990 and 1998. The calibrated model was then used to reconstruct paleohydrologic conditions for 6000 14C yr B.P.
Two methods for reconstructing paleo-climatic conditions and lake levels were assessed to determine if they could predict the observed range of 2.5–4.5 m decline in Williams Lake, within the Shingobee watershed, as reported by Locke (1995) for the middle Holocene based on beach sand facies. The first approach utilized paleo-GCM output (CCM1; Kutzbach et al., 1998). These paleoclimate simulations did not accurately represent the mid-Holocene drought conditions in Minnesota and calculated water levels for Williams Lake were only about 0.05 m lower than present-day lake levels. We also evaluated paleoclimate reconstruction using pollen data from a Williams Lake core. Paleo-climatic conditions using pollen-transfer functions of Bartlein et al. (1984) and Bartlein and Webb (1985) were used by Locke (1995) to predict Holocene winter temperature, summer temperature, and annual precipitation for the Shingobee Watershed. The watershed model calculated a lake-level drop of 3.5 m in Williams Lake using these pollen-derived estimates, which is consistent with the observed decrease for the mid-Holocene. This study demonstrates that watershed-scale mathematical models represent a powerful tool for in paleoclimate studies. Our watershed-scale hydrologic model was able to critically evaluate whether different paleo-climatic reconstructions were internally consistent with one another. This approach also provides a means to evaluate how regional climate estimates will impact local hydrologic conditions at the watershed scale.

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