Simulations of fully coupled lake-groundwater exchange in a subhumid climate with an integrated hydrologic model

B. D. Smerdon,[1] C. A. Mendoza,[1] and K. J. Devito[2]

Received 1 May 2006; revised 13 August 2006; accepted 31 August 2006; published 23 January 2007.

[1] A fully coupled, integrated surface water/groundwater model was used to study hydrologic controls on lake-groundwater interaction in the subhumid, Boreal Plains of northern Alberta, Canada. Findings from a previous water budget study indicate that lakes on the outwash landscape capture groundwater as a major source of water input and function as evaporation windows. Transient hydrologic responses of a flow-through style lake and outwash groundwater flow system were simulated for a three-dimensional model. Hydraulic heads, surface water depth, and the corresponding exchange fluxes between the surface and the subsurface were all simulated simultaneously and compared to field observations for the summers of 2002 and 2003. Replication of the transient flow regime required an anisotropy ratio of 10:1 for the outwash deposits and inclusion of riparian peatlands, which control lake-groundwater interaction and maintain surface water on permeable northern landscapes. Spatially and temporally variable evapotranspiration governed the water table configuration and lake-groundwater seepage patterns.

Citation: Smerdon, B. D., C. A. Mendoza, and K. J. Devito (2007), Simulations of fully coupled lake-groundwater exchange in a subhumid climate with an integrated hydrologic model, *Water Resour. Res.*, *43*, W01416, doi:10.1029/2006WR005137.

1. Introduction

[2] Effective management of water resources on all landscapes requires an understanding of water cycling processes, which are controlled by climate, geology, and the interaction of surface water and groundwater [*Devito et al.*, 2005a]. Research at the Utikuma Research Study Area (URSA), on the Boreal Plains region of northern Alberta, Canada, reveals complex surface and groundwater interactions [*Ferone and Devito*, 2004; *Smerdon et al.*, 2005], which are largely controlled by regional climate [*York et al.*, 2002; *Winter et al.*, 2005] and a heterogeneous landscape. Assessment of impacts to hydrologic systems on this landscape, from the rapid expansion of oil and gas exploration and production infrastructure, to exploitation from the forestry industry, requires models that can adequately represent integrated surface water/groundwater flow; a landscape dominated by widely distributed lakes, ponds and wetlands; and, a subhumid climate. To simulate water cycling processes on the Boreal Plains, and to predict changes to surface water bodies for scenarios of landscape disturbance or variation in climate, spatially distributed models must also be inclusive of the ubiquitous peatlands, which greatly add to landscape heterogeneity and control surface water/groundwater interaction [*Ferone and Devito*, 2004].

[3] The principles of simulating lake-groundwater interaction have been well documented by *Winter* [1976, 1983], *Townley and Trefry* [2000], and *Smith and Townley* [2002], and through development of the Lake Package for the MODFLOW groundwater model (described by *Hunt et al.* [2003]). In this study, we seek to enhance modeling of lake-groundwater exchange further, by using a numerical model capable of determining the location and depth of surface water bodies (i.e., lakes and ponds) and surface-subsurface exchange fluxes, without prior definition (i.e., fully coupled surface water/groundwater interaction). That is, we do not want to specifically define water bodies as hydraulic boundary conditions, but instead have lake location and stage determined as part of the simulation. This methodology is relatively new in hydrologic modeling, and has not been done for a lake- and wetland-dominated area with abundant peatlands and an irregular surface geometry. This fully coupled approach allowed lake-groundwater interaction to be investigated for a study lake at the URSA without defining the study lake as a boundary condition, thereby minimizing the amount of numerical intervention reflected in the solution of governing flow equations [*Loague and VanderKwaak*, 2004].

[4] The objectives of this study were to simulate lake-groundwater interaction for a subhumid hydrologic system on forested, glacial outwash terrain, and to determine the degree of spatially distributed atmospheric flux data (i.e., evapotranspiration and throughfall) and landscape heterogeneity needed to replicate field observations. These objectives were met through modeling groundwater conditions, lake levels, and lakebed seepage fluxes for an instrumented study lake at the URSA, for 2002 and 2003, and represent an intermediate step toward modeling the effects of land use, and climate change at a larger scale. A 3-year water budget study, including the field observations used for estimating parameter values and spatial distribution, and

[1]Department of Earth and Atmospheric Sciences, University of Alberta, Edmonton, Alberta, Canada.

[2]Department of Biological Sciences, University of Alberta, Edmonton, Alberta, Canada.

0043-1397/07/2006WR005137

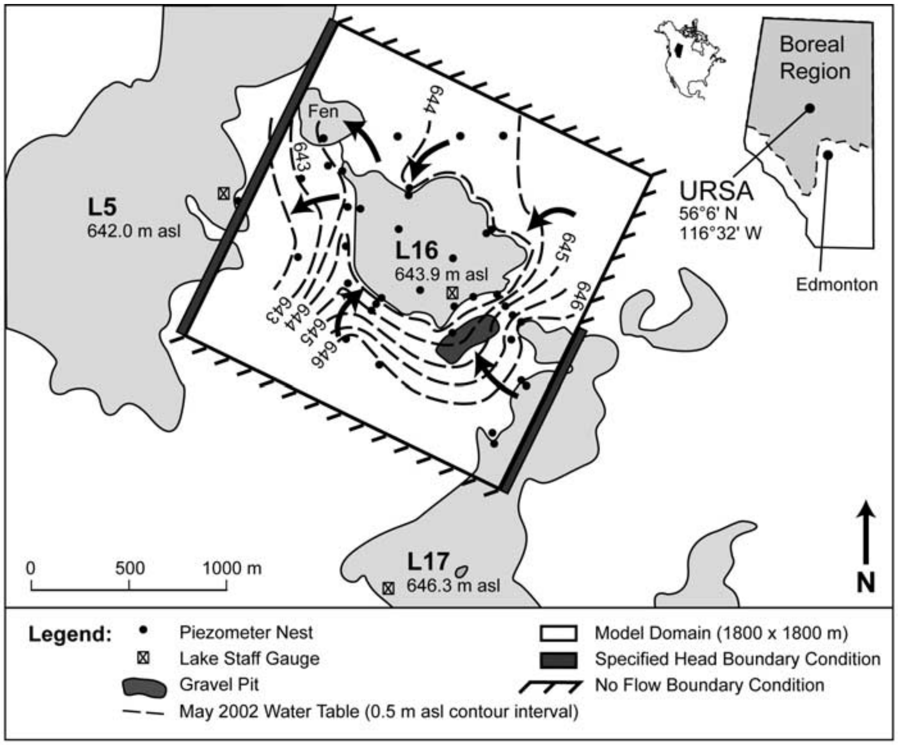

Figure 1. Lake 16 study site with selected field instrumentation, numerical model domain, and boundary conditions. Arrows indicate groundwater flow direction. URSA and Boreal Plains region in Alberta, Canada, is shown on inset.

comparison of modeling results in the present study, has been previously reported by *Smerdon et al.* [2005].

2. Study Area: URSA Lake 16

[5] The Boreal region containing the URSA is characterized by a subhumid climate [*Ecoregions Working Group*, 1989], marine shale bedrock (Upper Cretaceous Smoky Group [*Hamilton et al.*, 1999]) that is covered by 80 to 240 m of heterogeneous glacial sediments [*Pawlowicz and Fenton*, 2002], and low topographic relief. Average annual precipitation (P) and potential evapotranspiration (PET) are 481 mm [*Environment Canada*, 2003] and 518 mm [*Bothe and Abraham*, 1993], respectively; however, annual precipitation varies year to year (318 to 529 mm in 1997 to 2001 [*Devito et al.*, 2005b]). The URSA is approximately 150 km south of the discontinuous permafrost zone [*Woo and Winter*, 1993], and has average monthly temperature that ranges from −14.6°C to 15.6°C [*Environment Canada*, 2003], with the winter season occurring from approximately November to April.

[6] Lake 16 (39 ha) is located within the URSA, 370 km north of Edmonton, Alberta, Canada (56°6′ N, 116°32′ W; Figure 1), on a 200 km^2 coarse-textured glacial outwash plain, which is bounded by fine-textured sediments of a disintegration moraine [*Fenton et al.*, 2003]. In the vicinity of the study area, the outwash landscape is hummocky, with remnants of an east to west trending esker, and a maximum topographic relief of 20 m. Surface drainage is poorly developed, with no evidence of surface inflow to lake 16, and ephemeral outflow through a narrow channel to a fen on the west side of lake 16. The study area contains three lakes that exist in a series of "steps", where lake 16 was 1.8 m higher than lake 5, and 2.4 m lower than lake 17 in early 2002 (Figure 1). Groundwater springs occur along the southeast shore, adjacent to a small gravel pit, which was constructed by removing part of the esker deposit.

[7] Measurements of hydraulic head at 70 piezometers, lake level at each of the 3 lakes, and groundwater seepage estimates from 8 lakeshore segments were made from April 2001 to October 2003 [*Smerdon et al.*, 2005]. Hydraulic conductivity (K) was found to vary over six orders of magnitude (10^{-3} to 10^{-8} m s^{-1}), for 8 soil types (Figures 2 and 3). Annual P was 283 mm and 345 mm (including 44 mm and 83 mm of snow water equivalent), and annual lake evaporation was 336 mm and 455 mm for 2002 and 2003, respectively (Figure 4a). Over the summer of 2002, lake levels declined by 0.26 m, 0.18 m, and 0.34 m for Lakes 5, 16, and 17. Throughout 2003, lake 5 increased by 0.14 m, and Lakes 16 and 17 remained at nearly a constant elevation.

[8] Groundwater moved from southeast to northwest at an average horizontal hydraulic head gradient of 0.002, with the formation of small water table mounds (from snowmelt recharge) downgradient of the lake in spring months of each year [e.g., *Anderson and Munter*, 1981]. Evaporation from the lake was the largest hydrologic flux each year, exposing captured groundwater to the atmosphere via evaporation windows (i.e., lakes), and causing water table hinge lines to be located downgradient from the lake [*Smerdon et al.*, 2005]. This dynamic relationship between precipitation, groundwater interaction, and evaporation controls the degree of lake-groundwater exchange, and is expected to be very

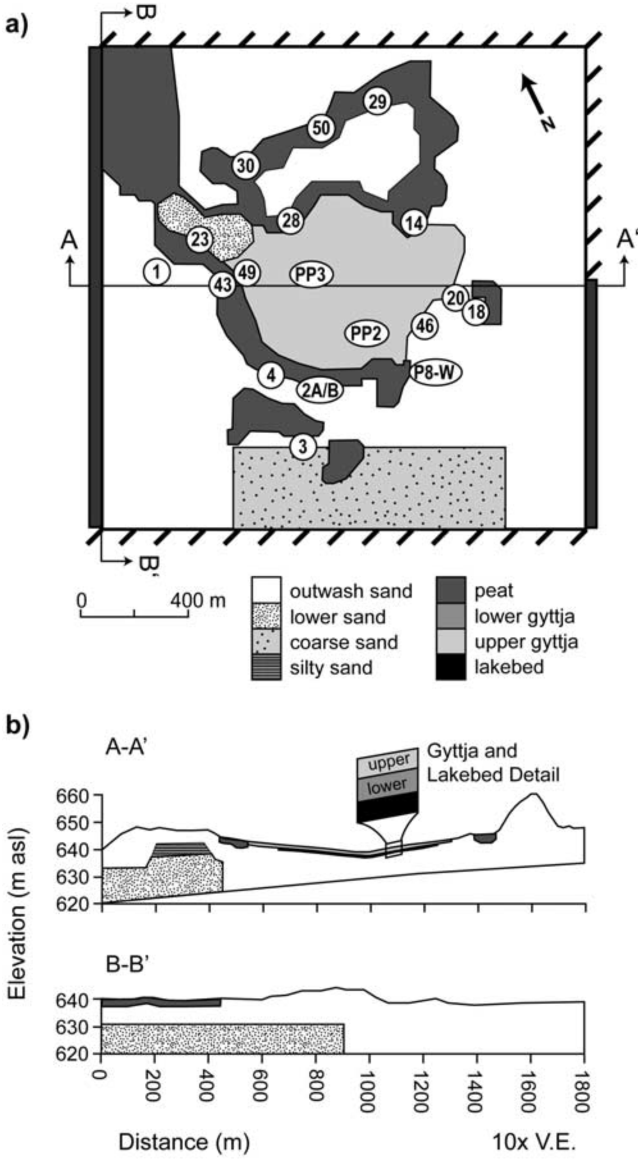

Figure 2. (a) Plan view and (b) cross sections of model domain, showing porous media zones (shaded), observation points (circled), and boundary conditions from Figure 1. Details of upper and lower gyttja and mineral lakebed zones are shown on inset in Figure 2b.

sensitive to regional climate conditions, which may be altered under future land use and climate change scenarios.

3. Numerical Simulations

[9] Lake-groundwater exchange was simulated for a three-dimensional, 1800 m by 1800 m by 30 m (high) section of the outwash landscape, including lake 16 (Figures 1 and 2). Two separate timeframes were considered: March 22 to November 2, 2002, when lake 16 declined by 0.18 m, and April 25 to November 1, 2003, when the level of lake 16 remained nearly stable. These periods coincide with abundant time series field data and the water budget analysis by *Smerdon et al.* [2005], but exclude the winter season, which would have additional cryospheric processes to consider. Results of simulated hy heads, water table elevation, lake 16 level, and seepage fluxes were compared to the field data. For the 2002 timeframe, field measured and/or assumed hydraulic input parameters were initially used, and were then adjusted slightly to better match simulation results with field observations (i.e., calibration); however, for the 2003 timeframe only new boundary conditions (based on field measurements) were specified to assess applicability of the model to represent field conditions without additional calibration (i.e., validation [*Anderson and Woessner*, 1992]).

3.1. Lake 16 Flow Model

[10] The integrated hydrology model (InHM) [*VanderKwaak*, 1999], was used to solve fully coupled equations for variably saturated porous media flow and overland flow, using a first-order flux relationship, as given by

$$-\nabla q \pm q^b \pm q^e_{ps} = \frac{\partial \phi S_w}{\partial t} \tag{1}$$

$$-\nabla q_s \psi_s \pm a_s q^b \pm a_s q^e_{sp} = \frac{\partial \left(S_{ws} h_s + \psi_s^{store} \right)}{\partial t} \tag{2}$$

where q is Darcy flux, q^b is a specified boundary condition (i.e., source or sink within the porous medium or on the land surface), ϕ is porosity, S_w and S_{ws} are water saturation for the porous media and land surface, t is time, q_s is overland flow (implemented as a two-dimensional diffusive wave equation approximated with Manning's equation [see *VanderKwaak*, 1999]), ψ_s is depth of mobile surface water, ψ_s^{store} is depth of immobile surface water stored in centimeter-scale microtopography h_s. The exchange flux between the porous media and land surface is given by q^e, where the subscripts *ps* and *sp* denote flux from the porous media (p) to surface (s), or vice versa, and a water exchange coefficient (a_s). Thus the model for the URSA lake 16 area simulated fully coupled hydraulic head, water saturation, lake depth, and lakebed seepage for a three-dimensional, finite element mesh.

[11] The finite element mesh used in this study was composed of 243,000 subsurface prism elements and 16,200 triangular surface elements (i.e., the top of the prisms) that were uniformly spaced at 20 m horizontally, and variably spaced in the vertical direction. Elements had finer vertical spacing (0.25 m) to a depth of 1 m beneath the lake basin and within the expected range of water table depths. The topography of the surface of the mesh was constructed from a 1 m horizontal resolution digital elevation model of the land surface combined with lake bathymetry measurements [*Smerdon et al.*, 2005], which were smoothed using a Spline algorithm, to 20 m horizontal resolution to reduce the complexity of the surface features. The bottom surface of the mesh sloped from east to west (Figure 2b), corresponding to the base of the glacial outwash as determined from geophysical field investigations [*Domes*, 2004].

[12] Subsurface heterogeneity of the glacial outwash was represented by 8 zones in the model, the parameters of which were based on field investigation. Each zone was specified with a hydraulic conductivity (K) value measured in the field and estimates of specific storage (S_s), ϕ, and ratio of horizontal to vertical hydraulic conductivity (i.e., K_{xy}:K_z; Figure 3a). Within the lake basin, the bathymetry

a)

Zone	K (m/s)	K_{XY}:K_Z	S_S (m^{-1})	ϕ
outwash sand	$1x10^{-5}$	10:1	10^{-5}	0.35
lower sand	$5x10^{-5}$	1:1	10^{-5}	0.35
coarse sand	$1x10^{-4}$	1:1	10^{-6}	0.35
silty sand	$2x10^{-6}$	50:1	10^{-5}	0.35
lakebed	$1x10^{-8}$	1:1	10^{-4}	0.45
upper gyttja	$3x10^{-6}$	1:1	10^{-4}	0.45
lower gyttja	$1x10^{-7}$	1:1	10^{-4}	0.45
peat	$3x10^{-6}$	1:1	10^{-4}	0.90

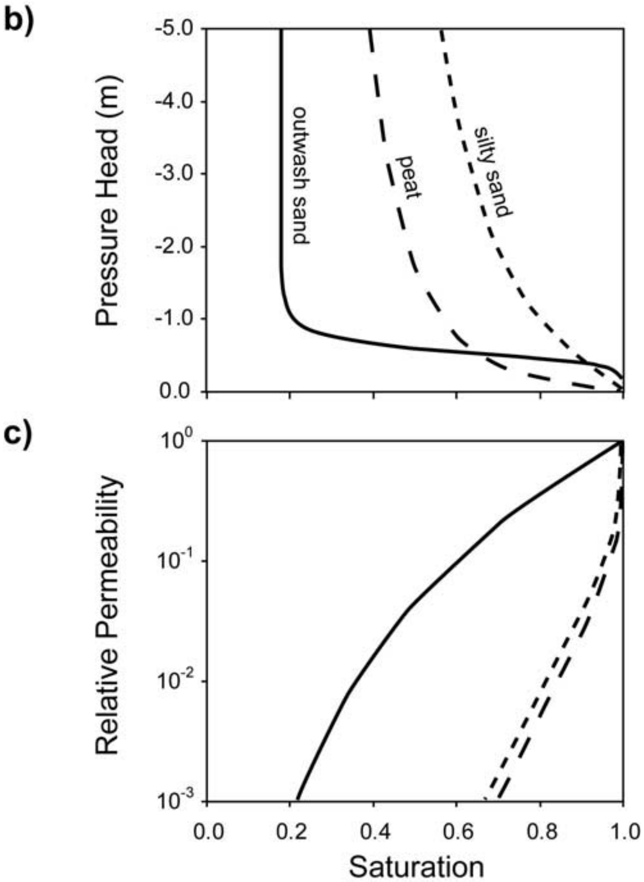

Figure 3. (a) Hydraulic parameters for each porous media zone and (b) pressure head and (c) relative permeability relationships for variably saturated zones.

(i.e., top of model domain) corresponds to the top of the gyttja, thereby allowing fluid flow, and more importantly, lakebed seepage flux, to be simulated through organic and mineral soil layers present in the lake basin [*Squires et al.*, 2006]. Gyttja and lakebed sediment K decreased with depth, such that the gyttja more easily transmits water than lakebed mineral sediments. Tabulated relationships for capillary pressure, water saturation, and relative permeability were required for 3 zones expected to be variably saturated (Figures 3b and 3c), and were based on published data of similar soil texture: outwash sand (Borden sand [*Abdul*, 1985]), silty sand [*Carsel and Parrish*, 1988], and peat [*Silins and Rothwell*, 1998]. Manning's n values used in equation 2 were assumed 0.04 for the narrow stream outlet on the west side of the lake [*Dingman*, 2002], and 0.4 for the remainder of the ground surface [*Woolhiser*, 1975]. Microtopography (h_s) and mobile water depth were each assumed to be 10^{-2} m, and the surface-coupling coefficient (a_s) was set to 10^{-4}.

3.2. Boundary Conditions

[13] Fluid flow boundary conditions include time-varying specified hydraulic heads of Lakes 5 and 17 (Figures 1 and 4c) along the east and west edges of the mesh (measured by *Smerdon et al.* [2005]), and a specified hydraulic head of 640 m within the lower sand zone (Figure 2b) to allow subsurface connection with the larger-scale, regional groundwater flow system existing beneath the west of the study area. The mesh has been oriented in the direction of groundwater flow; therefore the remaining edges parallel to groundwater flow were specified as no-flow boundaries. The bottom of the mesh has also been set as a no-flow boundary because it corresponds with the base of the outwash deposits, which were underlain by clay and have significantly lower K than the outwash sand.

[14] Daily precipitation and evaporation was applied to the top of the model domain as time series, net fluid fluxes (i.e., the sum of positive precipitation fluxes and negative evaporative fluxes). For areas of the ground surface that were saturated or that had surface water, and were not covered by a forest (e.g., fen and lake), specified net fluxes corresponded to the precipitation and open water evaporative fluxes (Figure 4a) measured by *Smerdon et al.* [2005]. Net fluxes specified for the gravel pit area were assumed to consist of precipitation only (i.e., no evaporative flux), because the area is unvegetated. For the remaining forested land surfaces, daily net fluxes (Figure 4b) accounted for interception and evapotranspiration, which were dependent on the time of year. Forest interception and actual evapotranspiration were based on hydrologic research at another site in the Western Boreal Forest [*Price et al.*, 1997]. From March to June 2002, and October 2002 in the first simulated year, and April to June, and October 2003 in the second simulated year, 100% of daily P measured at the URSA was assumed for the forested areas, to account for limited interception and evapotranspiration. For the summer months (i.e., June to August) the net flux was assumed to be 80% of daily P greater than 15 mm, and for September net flux was assumed to be 95% of daily P greater than 5 mm. These net fluxes are lower than observed precipitation throughfall rates at the URSA (85% to 95% of P (K. J. Devito, unpublished data, 2005)), in order to account for canopy interception and actual evapotranspiration, which could not be simulated explicitly for unsaturated subsurface (i.e., porous media) nodes [*Loague et al.*, 2005].

3.3. Initial Conditions

[15] Initial conditions were required for hydraulic head, water saturation, and lake depth for each node in the model domain, and were determined by assigning the initial hydraulic heads of lake 5 and lake 17 to those measured in the spring of either 2002 or 2003, and running the model to steady state. For these steady state simulations, the average sum of atmospheric fluxes (i.e., precipitation and evapotranspiration) was assumed to be zero. The resulting hydraulic heads and lake depths were compared to 22 March 2002 and 25 April 2003 field observations, and

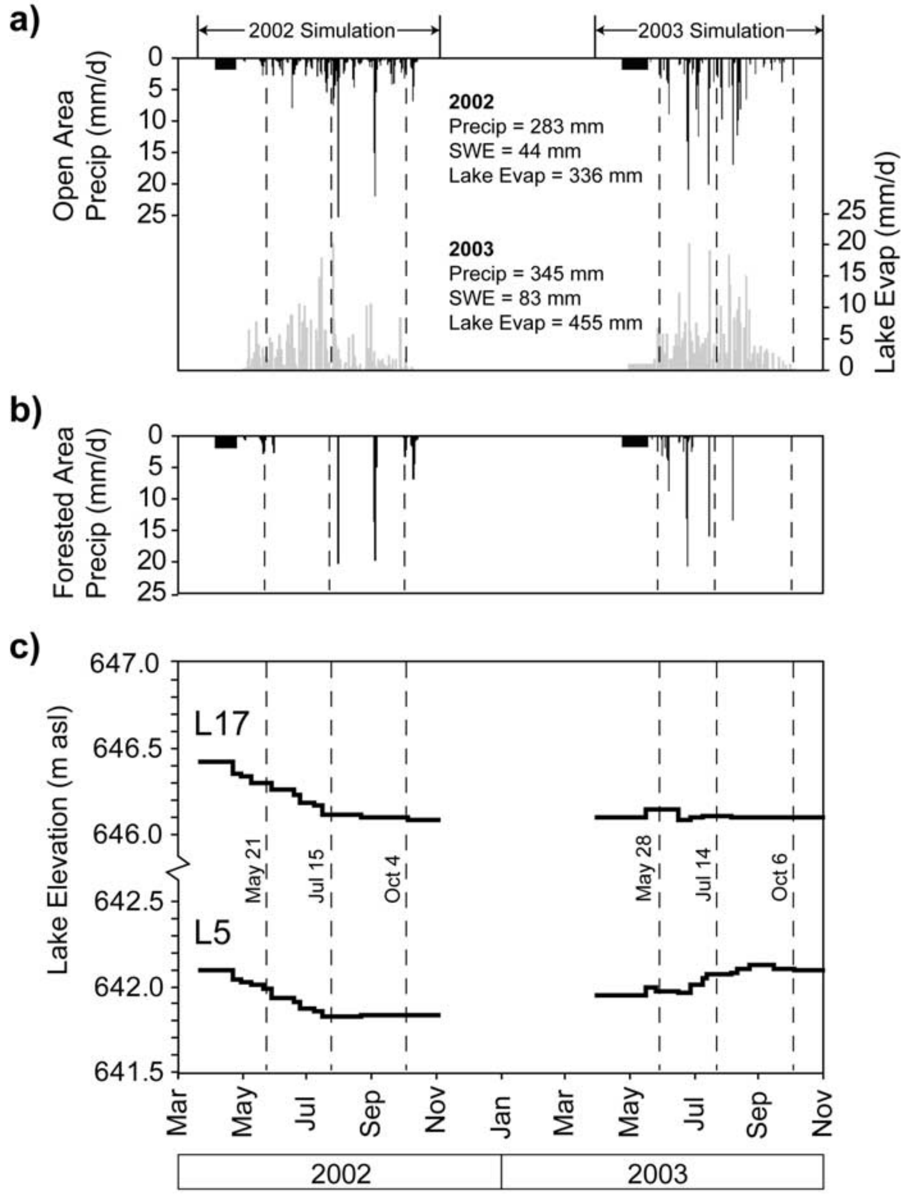

Figure 4. (a) Specified daily precipitation (including snow water equivalent (SWE)) for open areas and specified lake evaporation. (b) Net precipitation for forested areas. (c) Specified hydraulic heads for lakes 5 and 17. Specific output times for each year (May, July, and October) are shown with dashed lines.

subsequently input as initial conditions for the transient simulations.

4. Results

4.1. Steady State Initial Conditions

[16] As the simulation converged from an initially saturated condition, to the steady state solution, the lake level came into equilibrium conditions with the simulated groundwater flow field (Figure 5). At steady state, the water table configuration, and level of lake 16, were similar to field observations for March 2002. For 19 wells/piezometers distributed around the study site and lake 16 (Figure 2a), the root mean squared error (RMSE) was 0.29 m, which is on the order of observed water table and lake level fluctuations. Similar results were obtained for the initial conditions of April 2003 (RMSE = 0.30 m).

[17] The simulated flow system adequately represented the flow-through interaction between the lake and groundwater, as well as the large groundwater capture zone (Figure 5). These initial conditions were sensitive to K and anisotropy of the outwash sand, the dominant zone of porous media. For the field-measured outwash K, an assumed anisotropy of 10:1 provided the most representative results. Variation of the anisotropy ratio to 1:1 or 100:1 (Figure 6a), or variation of K by half an order of magnitude, resulted in an incorrect regional water table gradient and lake level.

4.2. Comparison of Transient Hydraulic Heads

[18] Simulated hydraulic head values for 19 observation locations around the lake, and the computed elevation of lake 16, matched most of the time series data from corresponding field observation points (Figure 7). For the

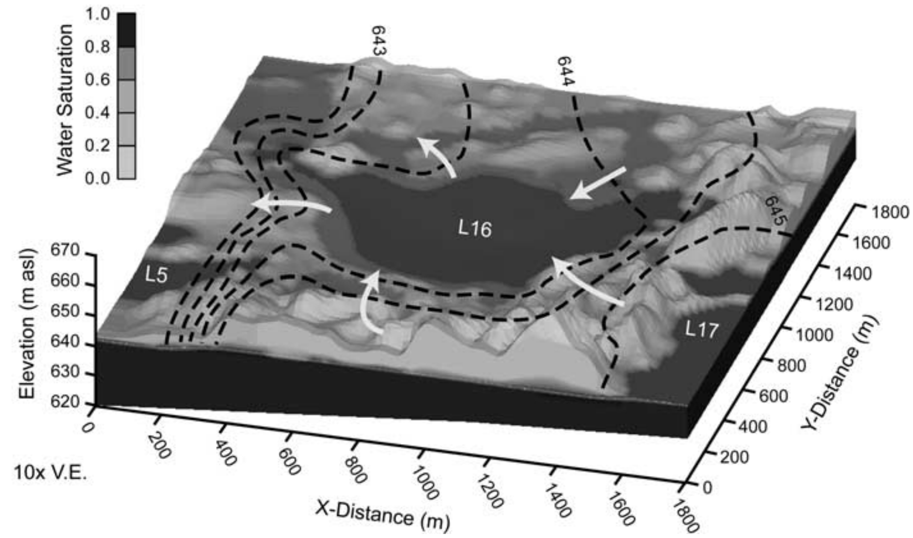

Figure 5. Three-dimensional shaded perspective of simulated water saturation overlain on model topography, simulated water table contours (0.5 m asl dashed lines), and approximate groundwater flow paths interacting with lake 16 (arrows) at model equilibrium for 2002.

riparian peatland zone directly adjacent to lake 16 (2B, 14, 20, 23, 28, 29, 43 on Figure 7), the general trend in hydraulic head response appears to mimic field observations, except during summer months when simulated and observed hydraulic head data were out of phase. Considering that hydraulic heads are represented well for the spring and autumn months, and that the general trend of hydraulic head decrease is similar to field observations, we suspect that an additional transient storage mechanism is operating in the field that is not represented in the simulation. At a location downgradient of lake 16, water table response was not as well represented in the simulation as is observed in the field (1; Figure 7), and improved characterization of the unsaturated hydraulic parameters in this porous media zone is likely warranted.

[19] The RMSE between simulated and observed hydraulic heads were calculated for 3 times in each simulation (May, July, and October). RMSE varied from 0.24 to 0.31 m for the 2002 data, and from 0.21 to 0.42 m for the 2003 data (Figure 8). In each simulation, the RMSE increased by approximately 0.1 m for the July period, suggesting the occurrence of a midsummer phenomenon occurring in the field that might not be replicated suitably in the numerical model.

4.3. Water Table Configuration

[20] Simulated water table contours in the vicinity of lake 16 were plotted for spring (May), summer (July), and autumn (October) months of 2002 and 2003 (Figure 9). As anticipated from the formulation of this boundary value problem, groundwater flow is from southeast to northwest (L17 to L5) across the model domain, with various degrees of interaction between the surface water and groundwater. Because of the dominance of evaporation from the lake, the water table contours wrap more than halfway around lake 16, resulting in hinge lines that point in the downgradient direction, indicative of the lake's subsurface capture zone [*Townley and Davidson*, 1988; *Gosselin and Khisty*, 2001]. The gen hape of the water table compares favorably with the water table maps produced by *Smerdon et al.* [2005] for the same time periods.

[21] In the 2002 simulation, the lakes on the model domain boundary (Lakes 5 and 17) decline due to regional drought conditions (Figure 4c). As expected, the simulated water table configuration adjusts as the level of lake 16 decreased, shown by the position of the 643.5 m asl contour on Figures 9a and 9b. In response to 2 days of heavy precipitation, late in September 2002, the elevation of lake 16 increased by 0.05 m, and the position of the 643.5 m asl contour was relocated downgradient of the lake basin. In the 2003 simulation, the balance between fluxes of groundwater discharge and evaporation maintain the elevation of lake 16 to a more consistent level than observed in 2002 (Figure 7), with nearly a steady water table configuration (Figure 9).

4.4. Lake-Groundwater Seepage

[22] Simulated exchange fluxes, which were calculated between the subsurface and surface elements in the numerical model, represent lakebed seepage for the lake 16 area within the model domain (Figure 9). Groundwater discharge measurements from 3 seepage meters installed along the southeast lakeshore varied from 2.4×10^{-6} to 5×10^{-8} m s^{-1}, with a geometric mean of 6.3×10^{-7} m s^{-1} for the summer months of 2002 and 2003 [*Smerdon et al.*, 2005]. Simulated seepage fluxes at corresponding locations within the flow model were 2.5×10^{-7} m s^{-1} (Figure 10). Although the simulated seepage varied over a narrower range than either of the seepage estimates reported by *Smerdon et al.* [2005], the magnitude of groundwater discharge and spatial pattern of groundwater recharge and discharge across the lakebed, honor field observations.

[23] Shaded contours of lakebed seepage (Figure 9) indicate that groundwater discharge occurs over at least 75% of the lakebed area. Throughout the summer of 2002, and between 2002 and 2003, the fraction of the lakebed area that discharged groundwater to the lake basin increased from 0.75 to 0.83, and is associated with a noticeable

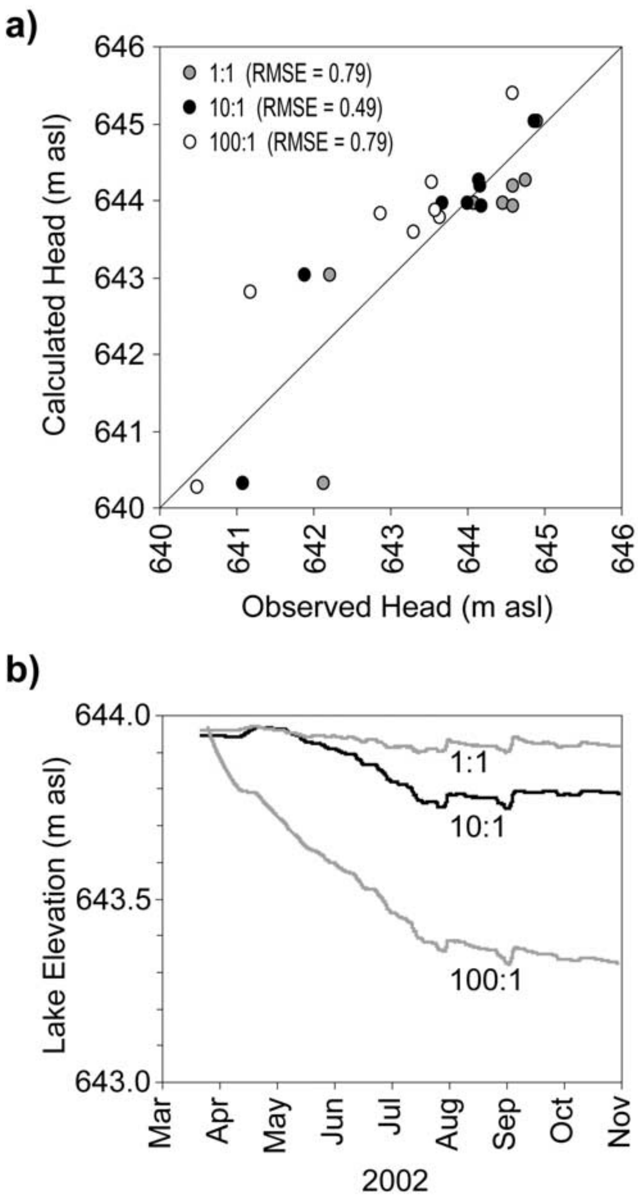

Figure 6. Effect of anisotropy (K_{xy}:K_z) on (a) steady state solution of hydraulic heads and (b) transient response of lake 16 elevation.

movement in the hinge line between discharge and recharge fluxes across the lakebed (Figure 9).

5. Discussion

5.1. Landscape Heterogeneity

[24] Eight zones of heterogeneity were required to represent the hydrologic flow regime of lake 16. The broad-scale flow system was adequately defined by the hydraulic parameters of the most predominant zone of porous media [*Winter and Pfannkuch*, 1984], which in these simulations was the "bulk" outwash K. The outwash sediments at the URSA were predominately well sorted fine sand, with a narrow range in approximate pore throat diameter and average grain size (d_{10} and d_{60} of 0.09 and 0.20 mm). Replicating the field observations of initial conditions and the transient response of lake 16 and hydraulic heads within the surrounding outwash, required an anisotropy ratio of 10:1, which was an order of magnitude lower than values reported by *Winter and Pfannkuch* [1984] for an outwash field site. For the coupled lake-groundwater model developed in this study, a higher anisotropy ratio for the outwash sediments (100:1) resulted in an underestimate of hydraulic heads in 2002, and isotropic conditions resulted in over-prediction of hydraulic heads (Figure 6b). The narrow range in grain size distributions reported for the URSA outwash sediments [*Smerdon et al.*, 2005], with low fractions of silt and clay, corroborate the low anisotropy (10:1) used in this study is representative of the predominant porous medium. Furthermore, *Weeks* [1969] measured aquifer anisotropy at an outwash deposit in Wisconsin, United States, and found that it ranged from 5:1 to 40:1, providing additional corroboration of the 10:1 anisotropy ratio determined from this study.

5.2. Riparian Peatlands

[25] Lake-groundwater interaction is further controlled by the presence of riparian peatlands at the margin of the lake and the stratigraphic sequence of lakebed deposits (upper and lower gyttja overlying fine grained mineral sediments). These heterogeneous features on the outwash landscape were required for the accurate simulation of lake 16's elevation and the surrounding hydraulic heads, especially when transient responses were considered. The most sensitive of these zones of porous medium were the riparian peatlands. Variation of the peatland K by a factor of 5 had a large influence on the transient response of the lake elevation, the water table configuration adjacent to the lake, and nearshore lakebed seepage fluxes. In effect, riparian peatland K governs the hydraulic connection between the lake and adjacent groundwater regime. If the peatland K were increased slightly (i.e., having a lower contrast with the adjacent outwash sediments), the flow system was more efficiently connected laterally. The net effect was enhanced lateral fluid exchange with groundwater at the lake margin, which created more groundwater discharge along the lakeshore than observed in the field. With a slightly lower peatland K, the lake became more isolated from the transience of the outwash groundwater system, and began to appear more hydraulically disconnected. Thus higher K riparian peatlands provide lower buffering ability or hydraulic storage capacity between the groundwater regime and the lake. Limited hydraulic storage capacity will have significant influence on biogeochemical reactions [*Devito et al.*, 2000; *Gibbons*, 2005]. High K riparian peatlands will promote nutrient movement between aquatic (i.e., lake) and groundwater zones, which are different biogeochemical environments.

[26] When field observations of hydraulic head response were compared with transient modeled response (observation points with shaded boxes on Figure 7), it appears that the value of peatland K and S_s could be transient within a seasonal timeframe, as the riparian peat material transitions from a frozen to thawed state in the midsummer. Considering that the riparian peatlands remain nearly water saturated continuously, below freezing air temperatures in the winter months (November to April) will cause these porous materials to freeze in a solid form, with reduced permeability. Field

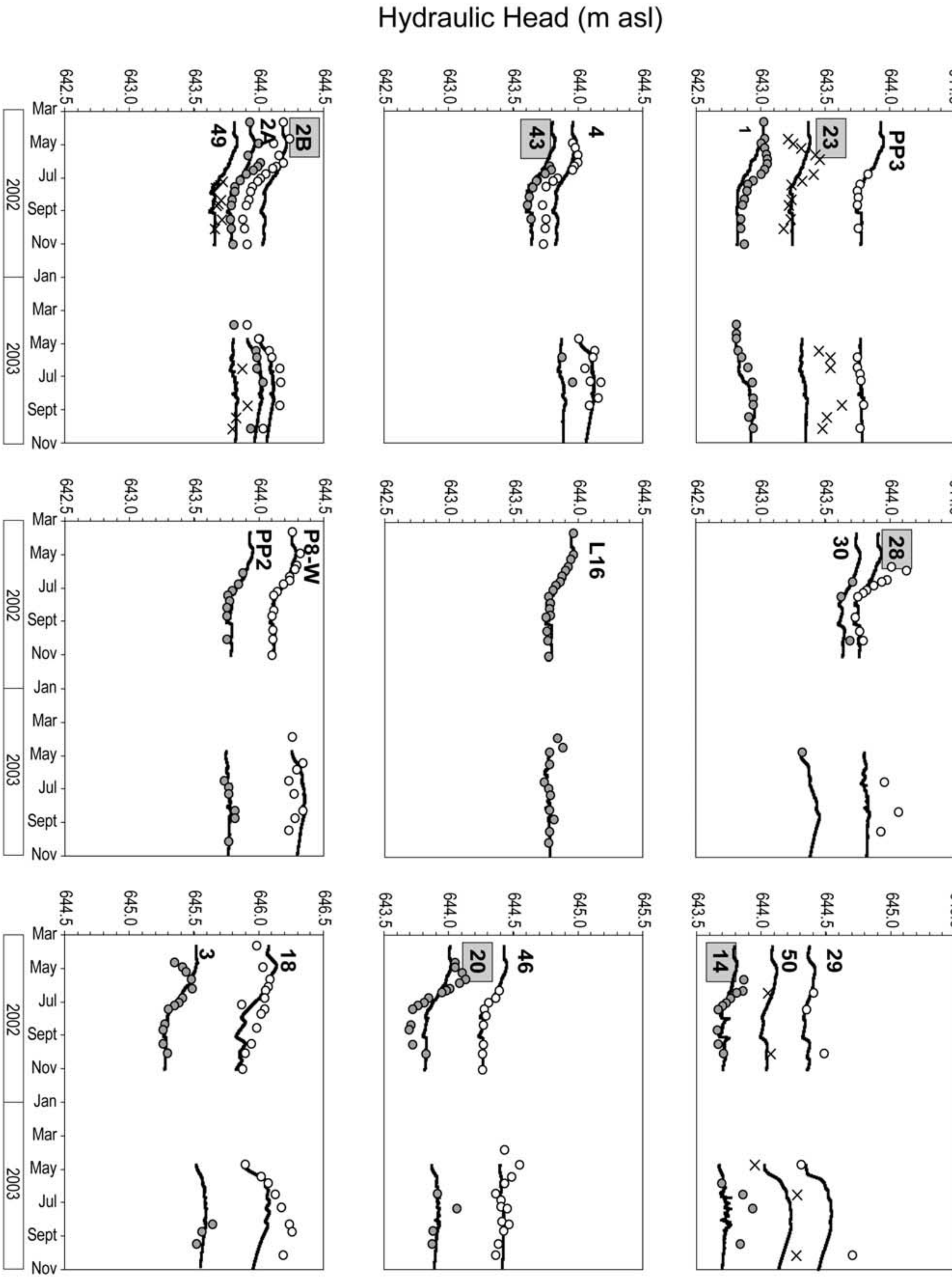

Figure 7. Time series simulation output (solid lines) and field measurements (symbols) of hydraulic head for observations points shown on Figure 2a. Observation points are grouped by spatial location around lake 16 (L16). Shaded boxes denote observations points associated with riparian peatlands.

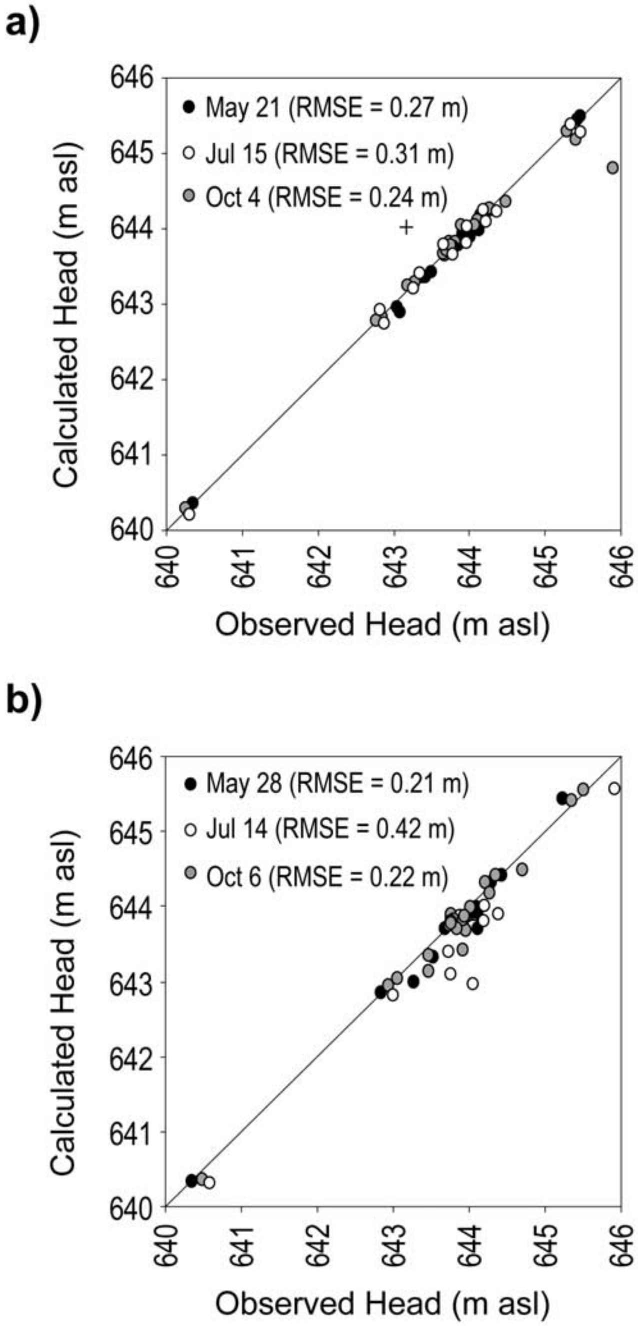

Figure 8. Root-mean-square error (RMSE) for specific output times for (a) 2002 and (b) 2003.

observations and preliminary lab study of the peatland state (B. D. Smerdon, unpublished data, 2004) corroborate the timing of a transition from frozen to thawed states, and further study of the effects of seasonal thermal effects on water transmission properties [e.g., *McCauley et al.*, 2002] is warranted.

[27] The peatland zones were parameterized with relatively high water retention characteristics determined from a different peatland in the same climatic region [*Silins and Rothwell*, 1998]. These water retention characteristics, high water saturation, shallow water table depth (i.e., nearly at surface), and close proximity to the lake, allow the peatland materials to retain wa ring regional water deficit conditions. Although the peatland material has been simulated as a traditional porous medium in this study, *Price* [2003] showed that peatlands will maintain saturation through a complex self-preservation feedback mechanism that utilizes vertical compression during drought, and flotation when conditions are more favorable for sphagnum growth. The presence of peatland material around most of lake 16 created a potential for water preservation on the outwash landscape. When the subhumid climate is considered, these peatland features are responsible for buffering lake-groundwater exchange, and are a critical feature for maintenance of wetlands, ponds and lakes on a permeable outwash landscape.

5.3. Lake-Groundwater Seepage

[28] Findings from other studies of lake-groundwater interaction, including *Winter* [1976, 1983], *Cheng and Anderson* [1994] and *Townley and Trefry* [2000], have shown that lakebed seepage patterns are largely controlled by position of a lake within a groundwater flow system, lakebed slope, and lakebed sediment K. However, the magnitude of seepage is more often controlled by regional hydraulic head gradient [*Smith and Townley*, 2002] and aquifer anisotropy [*Winter and Pfannkuch*, 1984]. In this study, many of these controlling factors (such as the groundwater flow system and lakebed sediment K) were constrained by field data. Spatial distribution of seepage at lake 16 is similar to the seepage patterns predicted by *Townley and Trefry* [2000] for lakes in a flow-through groundwater regime, and the magnitude of seepage flux was limited by the stratigraphy of lakebed deposits (gyttja overlying mineral sediments). Discrepancy between groundwater discharge measured in the field, and seepage flux predicted in the simulations (Figure 10), is within half an order of magnitude, and any subtle difference is attributed to the amount of spatial lakebed heterogeneity not represented in the model. Discrete field measurements capture the range of seepage through the lakebed at specific locations, representing lateral heterogeneity [*Kishel and Gerla*, 2002; *Rosenberry*, 2005] not specified in the model, whereas the numerical model produces an average result.

[29] Previous numerical studies of lake-groundwater seepage have often described hypothetical, circular lakes, in homogeneous aquifers [e.g., *Townley and Trefry*, 2000; *Genereux and Bandopadhyay*, 2001], under steady state flow conditions. Field observations at lake 16 [*Smerdon et al.*, 2005] and the present simulations illustrate that there is a transient interaction between lakes and the outwash groundwater flow system at the URSA, which is a common hydrologic phenomenon for lake-groundwater settings [e.g., *Anderson and Cheng*, 1993; *Gosselin and Khisty*, 2001]. The fully coupled transient solution of hydraulic heads and lakebed seepage fluxes were determined in these simulations simultaneously, which allowed for a more realistic representation of the hydrodynamics of lake-groundwater exchange than previously reported [e.g., *Genereux and Bandopadhyay*, 2001]. The finite element formulation of the InHM is suitable for areas of irregular topography that must be considered in simulation, such as lakebed bathymetry, and for areas of varying seepage face locations [*Romano et al.*, 1999], such as the dynamic lakeshore boundary. At the interface of the lake and the land surface, groundwater, surface water, and lakebed slope control the physics of

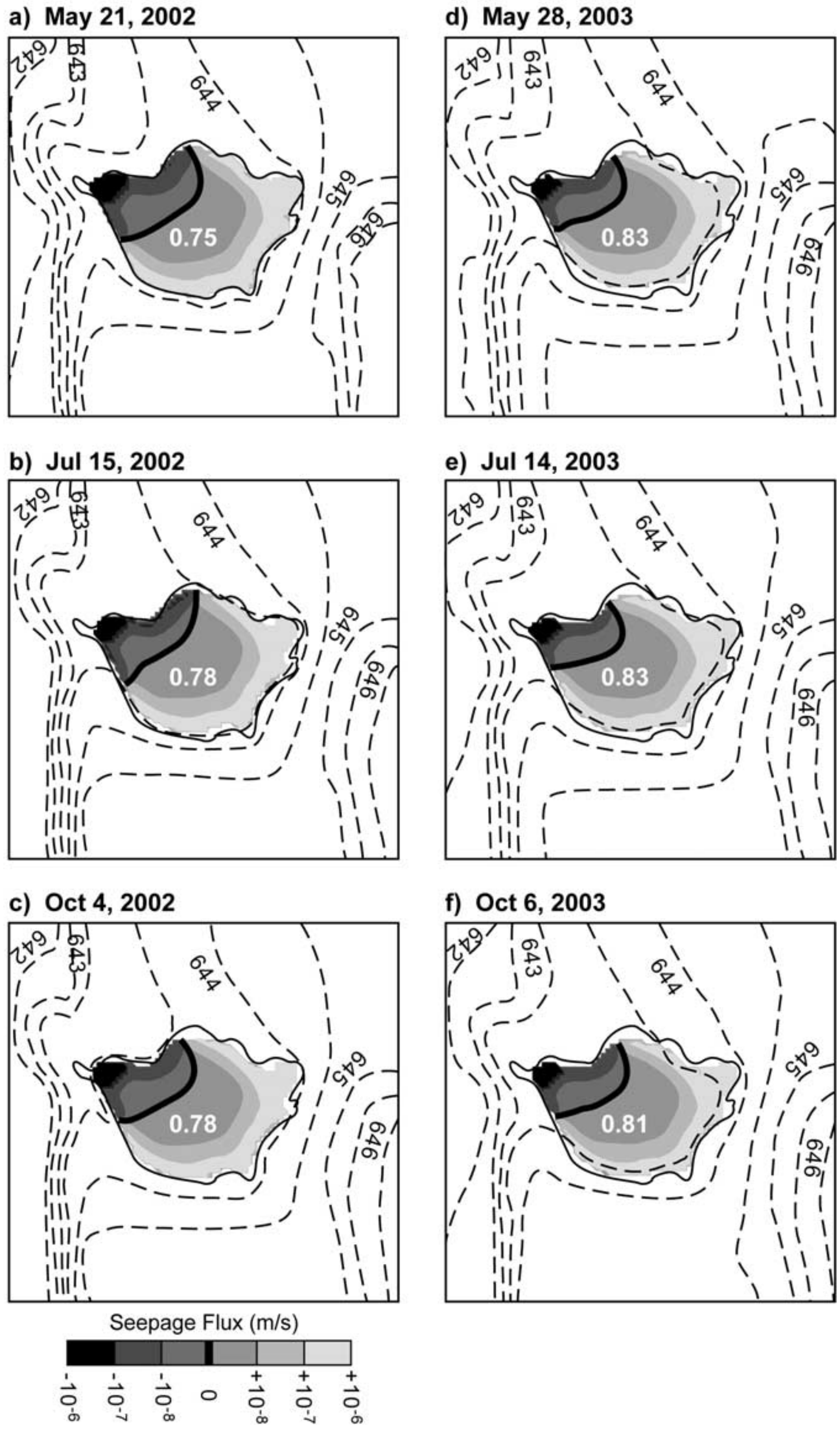

Figure 9. Simulated water table elevation (0.5 m asl contour interval) and shaded lakebed seepage flux. Fraction of lakebed area contributing groundwater to lake 16 is shown by white text. Transition between groundwater discharge areas and groundwater recharge areas is shown as a thick black line.

water flow, and thus lake-groundwater interaction [*Winter*, 1981]. These have all been explicitly considered in this model.

5.4. Climatic Controls on Lake-Groundwater Exchange

[30] Replication of the transient water table configuration, lake level and lakebed see upports the hypothesis that shallow lakes act as evaporation windows on the Boreal Plains landscape. The observed lake level decline of 0.18 m in 2002 was caused by water deficit conditions imposed by the subhumid atmosphere. Although the 2002 drought occurred only over one season, the rate of lake level decline (0.2 m yr^{-1}) was the same as observed during a severe (4 yr) drought at the glacial outwash Williams Lake in Minnesota,

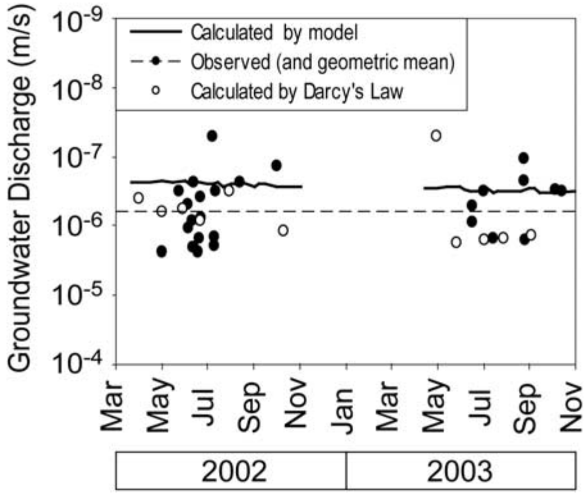

Figure 10. Simulated and observed groundwater discharge fluxes along southeast shore of lake 16. Values observed from individual seepage meter measurements or calculated using Darcy's Law are from *Smerdon et al.* [2005].

United States [*Winter et al.*, 2005]. The similarities in lake level decline between lake 16 and Williams Lake offers the opportunity to elucidate hydrodynamic response of the lake-groundwater system at lake 16 to drought conditions that could develop in many glaciated regions in North America. With a reduction in the volume of lake 16, there was a corresponding increase in lakebed area contributing groundwater discharge (10% increase in seepage area) in the following year, which contributed groundwater inflow to the lake basin. Although simulated groundwater discharge fluxes were nearly constant throughout the simulation timeframes, the reconfiguration of the water table contours and decrease in groundwater outflow from the lake illustrate the role of regional climate on lake-groundwater interaction in this subhumid climatic zone.

[31] The sensitivity analysis by *Genereux and Bandopadhyay* [2001] of lakebed seepage suggests that the seepage patterns observed for lake 16 (i.e., strong exponential decrease with increased distance from shore) are predominantly found in lakes with steeper lakebeds. The shallow slope of lake 16's bed (0.012) appears to have similar patterns as those lakes with moderate (0.013) to steep (0.02) slopes in the work of *Genereux and Bandopadhyay* [2001], suggesting that a large outflow of water from lake 16 exists, and that Lake 16 is being replenished by the groundwater system. Simulation results and the corroborating field data indicate that there is minimal to no surface outflow and diminished groundwater outflow from lake 16 (Figure 9) for 2002 and 2003. Therefore, given the shallow lakebed slope, seepage pattern indicative of a lake that would typically have a steeper slope, and field observations of minimal fluid outflow, the seepage pattern also confirms a dominance of summer evaporation (i.e., outflow via evaporation) and that climate has a significant influence on lake-groundwater interaction.

[32] *Lenters et al.* [2005] suggest that it is critical to understand "why" lake levels change in response to climate variation for effective water resource management. For landscapes composed o -permeability lithology, lake level changes will be an integrated response of atmospheric conditions (precipitation and evaporation), groundwater interaction, and overland flow. In this study, water cycling for a coupled lake-groundwater system was simulated, and results compared favorably with transient observations from an instrumented field site. For coarse-textured landscapes in subhumid climates, lake level changes are controlled by regional climate, and are further maintained by an active groundwater flow regime. The efficiency at which groundwater can maintain lakes, ponds and wetlands on this landscape, is controlled by the presence of landscape features that have high water retention, such as riparian peatlands, and by the duration of active lake evaporation (i.e., length of season). For shallow lakes in Northern Canada (which cover 37% of the landscape), *Rouse et al.* [2005] found that the lake evaporation season was 22 weeks per year, and that shallow lakes have a higher heat capacity and can partition water to the atmosphere more readily than wetlands. In these simulations, evaporation was applied to areas where the water table is at the land surface (similar to *York et al.* [2002]), for 22.8 and 23.3 weeks in 2002 and 2003, respectively. Although the lake level decline occurred in 2002, due largely to a regional drought that also lowered adjacent Lakes 5 and 17, it appears that a lowered lake level and increased water deficit condition (110 mm for 2003, compared to 53 mm for 2002) caused increased groundwater inflow.

5.5. Representing Atmospheric Boundary Conditions

[33] Lake level and hydraulic head response of the outwash groundwater flow systems were sensitive to the timing of the net atmospheric fluxes (described in section 3.2). Although actual evapotranspiration fluxes from the forested uplands were not explicitly simulated in this work, distinct periods of assumed boundary condition data (i.e., net fluxes that accounted for interception and evapotranspiration) had to be specified to accurately match field observations. The timing of these periods correspond to: spring (prior to June), when accumulated snowfall would melt and recharge the subsurface; summer (June to September), when only relatively large precipitation events (80% of daily P greater than 15 mm) were assumed to enter the subsurface; and, fall (from September to the end of each simulation), when reduced forest evapotranspiration and the loss of leaves would allow for more precipitation to reach the subsurface than summer months (95% of daily P greater than 5 mm). Although a good transient water table and lake level response was simulated, based on reasonable assumptions of upland forest water partitioning, future work should explicitly consider throughfall (as net precipitation reaching the ground surface) and actual evapotranspiration fluxes in a subhumid climate, and is presently being studied by the Western Boreal Hydrology Group at the University of Alberta. Improved representation of the atmospheric boundary conditions could be applied to a larger-scale model that extends to natural watershed boundaries, and be used to investigate scenario simulations of landscape disturbances and changes in climate.

6. Conclusions

[34] For lake-dominated hydrologic systems, many different styles of models have been developed to investigate

lake-groundwater interaction. This study investigated simulation of lake-groundwater interaction through use of a fully coupled, finite element, surface water/groundwater model to represent the hydrologic regime of a glacial outwash setting in a subhumid environment. Formulation of the numerical model allowed simultaneous simulation of subsurface hydraulic heads and surface water depths, and the corresponding exchange fluxes between the surface and the subsurface, without assuming any explicit hydraulic exchange fluxes a priori. Replication of three-dimensional, transient fluid flow, for a study area in northern Alberta, Canada, relied on accurately representing the "bulk" porous media with anisotropy that suitably represented structure of the porous media, and two key landscape heterogeneities that control lake-groundwater interaction: riparian peatlands and stratified lakebed deposits. To mimic time series field measurements, a lower anisotropy ratio (10:1) for glacial outwash was required than previously suggested for groundwater flow-through lake environments. The inclusion of riparian peatlands was found to govern lake-groundwater fluid exchange, and the peatlands appear to have seasonally transient hydraulic parameters. We hypothesize seasonal variation of peatland K values (i.e., higher K when thawed and lower K when frozen) as a mechanism for maintaining water in the subhumid, Boreal environment. Field observations and preliminary lab study of the peatland state (B. D. Smerdon, unpublished data, 2004) corroborate the timing of such a transition (i.e., from frozen to thawed states), and further study of the effects of seasonal thermal effects [e.g., *McCauley et al.*, 2002] on lake-groundwater exchange is warranted.

[35] Simulated patterns of lakebed seepage for lake 16 confirm predictions made by others, for hypothetical lakes of more simplified geometry. The magnitude and spatial distribution of seepage fluxes computed in these simulations compare favorably with field observations and, when combined with calibrated hydraulic head data, provide further corroboration that the model for the lake in this study represented the hydrologic regime adequately. The transient flow system is largely driven by high summer lake evaporation, which caused reconfiguration of water table contours around the lake basin and increased lakebed seepage following lake level decline. Simplified atmospheric boundary conditions were assumed for the forested upland areas, indicating that three distinct timeframes (spring, summer and fall) are required in simulations, which correspond to seasonal variation of throughfall and evapotranspiration. Considering the dominance of evaporation from lakes and shallow water table areas in this landscape, further investigation into the relationship between temporal throughfall and evapotranspiration fluid fluxes is needed, to establish the rate and timing of groundwater recharge. In a lake-dominated landscape that appears to be controlled by evaporation in summer months, the timing of groundwater recharge is likely also an important controlling factor for maintenance of lakes and wetlands, which requires further investigation to more fully understand water cycling processes in this region.

[36] This study illustrated that InHM can be applied to lake-dominated hydrologic systems, and used to investigate landscape and atmospheric controls on hydrologic processes. For larger areas, specific hydrologic landscapes [*Winter*, 2001], and longer-term applications (i.e., landscape management and reclamation), the modeling framework presented here would be appropriate, because it is not hampered by excessive numerical intervention (i.e., minimizing a priori assumptions). The physical boundaries of the model domain need not be surface catchments or watersheds [*Winter et al.*, 2003; *Devito et al.*, 2005a]; however, successful application will depend greatly on the ability to define spatially variable, subsurface hydraulic properties. Thus robust models such as this will provide insight for investigating response of hydrologic systems in areas where anthropogenic changes (imposed by landscape disturbance or variation in climate) might be masked or subdued by natural variation in water cycling.

[37] **Acknowledgments.** This research was supported by an NSERC-CRD grant for the HEAD Project, NSERC discovery grants to K. J. Devito and C. A. Mendoza, an NSERC-IPS to B. D. Smerdon (sponsored by Syncrude Canada Ltd.), Circumpolar/Boreal Alberta Research (C/BAR) grants to B. D. Smerdon, and Institute for Wetland and Waterfowl Research (IWWR) grants to B. D. Smerdon and K. J. Devito. We thank I. Creed at the University of Western Ontario for the high-resolution DEM data, J. VanderKwaak for use of the InHM, and R. McLaren at the University of Waterloo for the GridBuilder mesh generator. Comments from S. Tyler, M. Marani, M. Person, D. Rosenberry, and an anonymous reviewer improved this manuscript.

References

Abdul, A. S. (1985), Experimental and numerical studies of the effect of the capillary fringe on streamflow generation, Ph.D. thesis, 210 pp., Univ. of Waterloo, Ontario, Canada.

Anderson, M. P., and X. Cheng (1993), Long- and short-term transience in a groundwater/lake system in Wisconsin, USA, *J. Hydrol.*, *145*, 1–18.

Anderson, M. P., and J. A. Munter (1981), Seasonal reversals of groundwater flow around lakes and the relevance to stagnation points and lake budgets, *Water Resour. Res.*, *17*, 1139–1150.

Anderson, M. P. and W. W. Woessner (1992), *Applied Groundwater Modeling—Simulation of Flow and Advective Transport*, 379 pp., Elsevier, New York.

Bothe, R. A. and C. Abraham (1993), Evaporation and evapotranspiration in Alberta, 1986–1992 addendum, report, Alberta Environ. Protect., Edmonton, Alberta, Canada.

Carsel, R. F., and R. S. Parrish (1988), Developing joint probability distributions of soil water retention characteristics, *Water Resour. Res.*, *24*, 755–769.

Cheng, X., and M. P. Anderson (1994), Simulating the influence of lake position on groundwater fluxes, *Water Resour. Res.*, *30*, 2041–2049.

Devito, K. J., I. F. Creed, R. L. Rothwell, and E. E. Prepas (2000), Landscape controls on phosphorus loading to boreal lakes: Implications for the potential impacts to forest harvesting, *Can. J. Fish Aquat. Sci.*, *57*, 1977–1984.

Devito, K., I. Creed, T. Gan, C. Mendoza, R. Petrone, U. Silins, and B. Smerdon (2005a), A framework for broad scale classification of hydrologic response units on the Boreal Plains: Is topography the last thing to think of?, *Hydrol. Processes*, *19*, 1705–1714.

Devito, K. J., I. F. Creed, and C. J. D. Fraser (2005b), Controls on runoff from a partially harvested aspen-forested headwater catchment, Boreal Plain, Canada, *Hydrol. Processes*, *19*, 3–25.

Dingman, S. L. (2002), *Physical Hydrology*, 2nd ed., 646 pp., Prentice-Hall, Upper Saddle River, N. J.

Domes, F. (2004), 2-D traveltime inversion of near surface refractions and reflections in support of hydrogeological studies, diploma thesis, Karlsruhe Univ., Germany.

Ecoregions Working Group (1989), *Ecoclimatic Regions of Canada*, *Ecol. Land Classification Ser.*, vol. 23, Environ. Can., Ottawa, Ontario.

Environment Canada (2003), Canadian climate normals 1971–2000: Slave Lake, Alberta, report, Ottawa. (Available at http://climate.weatheroffice.ec.gc.ca/climate_normals/index_e.html)

Fenton, M. M., R. C. Paulen, and J. G. Pawlowicz (2003), Surficial geology of the Lubicon Lake area, Alberta (NTS 84B/SW), *Map 266*, Alberta Geol. Surv., Edmonton, Alberta, Canada.

Ferone, J. M., and K. J. Devito (2004), Shallow groundwater-surface water interactions in pond-peatland complexes along a Boreal Plains topographic gradient, *J. Hydrol.*, *292*, 75–95.

Genereux, D., and I. Bandopadhyay (2001), Numerical investigation of lake bed seepage patterns: Effect of porous medium and lake properties, *J. Hydrol.*, *241*, 286–303.

Gibbons, Z. (2005), The influence of groundwater flow on phosphate dynamics in three riparian wetlands surrounding an outwash lake in northern Alberta, M.Sc. thesis, 95 pp., Univ. of Alberta, Edmonton, Alberta, Canada.

Gosselin, D. C., and M. J. Khisty (2001), Simulating the influence of two shallow, flow-through lakes on a groundwater system: Implications for groundwater mounds and hinge lines, *Hydrogeol. J.*, *9*, 476–486.

Hamilton, W. N., M. C. Price, and C. W. Langenberg (1999), Geological map of Alberta, *Map 236*, Alberta Geol. Surv., Edmonton, Alberta, Canada.

Hunt, R. J., H. M. Haitjema, J. T. Krohelski, and D. T. Feinstein (2003), Simulating ground water-lake interactions: Approaches and insights, *Ground Water*, *41*, 227–237.

Kishel, H. F., and P. J. Gerla (2002), Characteristics of preferential flow and groundwater discharge to Shingobee Lake, Minnesota, USA, *Hydrol. Processes*, *16*, 1921–1934.

Lenters, J. D., T. K. Kratz, and C. J. Bowser (2005), Effects of climate variability on lake evaporation: Results from a long-term energy budget study of Sparkling Lake, northern Wisconsin (USA), *J. Hydrol.*, *308*, 168–195.

Loague, K., and J. E. VanderKwaak (2004), Physics-based hydrologic response simulation: Platinum bridge, 1958 Edsel, or useful tool?, *Hydrol. Processes*, *18*, 2949–2956.

Loague, K., C. S. Heppner, R. H. Abrams, A. E. Carr, J. E. VanderKwaak, and B. A. Ebel (2005), Further testing of the Integrated Hydrology Model (InHM): Event-based simulations for a small rangeland catchment located near Chickasha, Oklahoma, *Hydrol. Processes*, *19*, 1373–1398.

McCauley, C. A., D. M. White, M. R. Lilly, and D. M. Nyman (2002), A comparison of hydraulic conductivities, permeabilities and infiltration rates in frozen and unfrozen soils, *Cold Reg. Sci. Technol.*, *34*, 117–125.

Pawlowicz, J. G. and M. M. Fenton (2002), Drift thickness of the Peerless Lake map area (NTS 84B), *Map 253*, Alberta Geol. Surv., Edmonton, Alberta, Canada.

Price, A. G., K. Dunham, T. Carleton, and L. Band (1997), Variability of water fluxes through black spruce (*Picea mariana*) canopy and feather moss (*Pleurozium schreberi*) carpet in the boreal forest of northern Manitoba, *J. Hydrol.*, *196*, 310–323.

Price, J. S. (2003), The role and character of seasonal peat soil deformation on the hydrology of undisturbed and cutover peatlands, *Water Resour. Res.*, *39*(9), 1241, doi:10.1029/2002WR001302.

Romano, C. G., E. O. Frind, and D. L. Rudolph (1999), Significance of unsaturated flow and seepage faces in the simulation of steady-state subsurface flow, *Ground Water*, *37*, 625–632.

Rosenberry, D. O. (2005), Integrating seepage heterogeneity with the use of ganged seepage meters, *Limnol. Oceanogr.*, *3*, 131–142.

Rouse, W. R., C. J. Oswald, J. Binyamin, C. Spence, W. M. Schertzer, P. D. Blanken, N. Bussieres, and C. R. Duguay (2005), The role of northern lakes in a regional energy balance, *J. Hydrometeorol.*, *6*, 291–305.

Silins, U., and R. L. Rothwell (1998), Forest peatland drainage and subsidence affect soil water retention and transport properties in an Alberta peatland, *Soil Sci. Soc. Am.*, *62*, 1048–1056.

Smerdon, B. D., K. J. Devito, and C. A. Mendoza (2005), Interaction of groundwater and shallow lakes on outwash sediments in the sub-humid Boreal Plains of Canada, *J. Hydrol.*, *314*, 246–262.

Smith, A. J., and L. R. Townley (2002), Influence of regional setting on the interaction between shallow lakes and aquifers, *Water Resour. Res.*, *38*(9), 1171, doi:10.1029/2001WR000781.

Squires, M. M., D. Mazzucchi, and K. J. Devito (2006), Carbon burial and infill rates in small Western Boreal lakes: Physical factors affecting carbon storage, *Can. J. Fish Aquat. Sci.*, *63*, 711–720.

Townley, L. R., and M. R. Davidson (1988), Definition of a capture zone for shallow water table lakes, *J. Hydrol.*, *104*, 53–76.

Townley, L. R., and M. G. Trefry (2000), Surface water-groundwater interaction near shallow circular lakes: Flow geometry in three dimensions, *Water Resour. Res.*, *36*, 935–949.

VanderKwaak, J. E. (1999), Numerical simulation of flow and chemical transport in integrated surface-subsurface hydrologic systems, Ph.D. thesis, 217 pp., Univ. of Waterloo, Ontario, Canada.

Weeks, E. P. (1969), Determining the ratio of horizontal to vertical permeability by aquifer-test analysis, *Water Resour. Res.*, *5*, 196–214.

Winter, T. C. (1976), Numerical simulation analysis on the interaction of lakes and ground water, *U.S. Geol. Surv. Prof. Pap.*, 1001.

Winter, T. C. (1981), Effects of water-table configuration on seepage through lakebeds, *Limnol. Oceanogr.*, *26*, 925–934.

Winter, T. C. (1983), The interaction of lakes with variably saturated porous media, *Water Resour. Res.*, *19*, 1203–1218.

Winter, T. C. (2001), The concept of hydrologic landscapes, *J. Am. Water Resour. Assoc.*, *37*, 335–349.

Winter, T. C., and H. O. Pfannkuch (1984), Effect of anisotropy and groundwater system geometry on seepage through lakebeds, 2. Numerical simulation analysis, *J. Hydrol.*, *75*, 239–253.

Winter, T. C., D. O. Rosenberry, and J. W. LaBaugh (2003), Where does the ground water in small watersheds come from?, *Ground Water*, *41*, 989–1000.

Winter, T., D. Rosenberry, E. Kelly, and J. LaBaugh (2005), Comparison of wetlands in different hydrogeological settings under conditions of climate variability, paper presented at Dynamics and Biogeochemistry of River Corridors and Wetlands, Seventh IAHS Scientific Assembly, Foz do Iguacu, Brazil.

Woo, M.-K., and T. C. Winter (1993), The role of permafrost and seasonal frost in the hydrology of northern wetlands in North America, *J. Hydrol.*, *141*, 5–31.

Woolhiser, D. A. (1975), Simulation of unsteady overland flow, in *Unsteady Flow in Open Channels*, edited by K. Mahmood and V. Yevjevich, pp. 485–508, Water Resour. Publ., Highland Ranch, Colo.

York, J. P., M. Person, W. J. Gutowski, and T. C. Winter (2002), Putting aquifers into atmospheric simulation models: An example from the Mill Creek Watershed, northeastern Kansas, *Adv. Water Resour.*, *25*, 221–238.

K. J. Devito, Department of Biological Sciences, University of Alberta, Edmonton, AB, Canada T6G 2E9.

C. A. Mendoza and B. D. Smerdon, Department of Earth and Atmospheric Sciences, University of Alberta, Edmonton, AB, Canada T6G 2E3. (bsmerdon@ualberta.ca)