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Quantification of surface water and groundwater flows to open- and closed-basin lakes in a headwaters watershed using a descriptive oxygen stable isotope model

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[1] Accurate quantification of hydrologic fluxes in lakes is important to resource management and for placing hydrologic solute flux in an appropriate biogeochemical context. Water stable isotopes can be used to describe water movements, but they are typically only effective in lakes with long water residence times. We developed a descriptive time series model of lake surface water oxygen-18 stable isotope signature (δL) that was equally useful in open- and closed-basin lakes with very different hydrologic residence times. The model was applied to six lakes, including two closed-basin lakes and four lakes arranged in a chain connected by a river, located in a headwaters watershed. Groundwater discharge was calculated by manual optimization, and other hydrologic flows were constrained by measured values including precipitation, evaporation, and streamflow at several stream gages. Modeled and observed δL were highly correlated in all lakes (r = 0.84–0.98), suggesting that the model adequately described δL in these lakes. Average modeled stream discharge at two points along the river, 16,000 and 11,800 m^3 d^{-1}, compares favorably with synoptic measurement of stream discharge at these sites, 17,600 and 13,700 m^3 d^{-1}, respectively. Water yields in this watershed were much higher, 0.23–0.45 m, than water yields calculated from gaged streamflow in regional rivers, approximately 0.10 m, suggesting that regional groundwater discharge supports water flux through these headwaters lakes. Sensitivity and robustness analyses also emphasized the importance of considering hydrologic residence time when designing a sampling protocol for stable isotope use in lake hydrology studies.

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1. Introduction

[2] Accurate descriptions of water movements through lakes are important to effective lake management as well as appropriately weighing the role of lakes with respect to carbon and nutrient cycling. Maintaining proper water flows is commonly a lake management goal [*Gibson and Edwards*, 2002] and influent water can deliver nutrients or other pollutants which can affect the lake ecosystem [*Tomassoni*, 2000]. Lakes also play an important role in large-scale biogeochemical cycles [*Cole et al.*, 2007], but calculating the retention or transformation of elements, such as carbon, in lakes depends upon accurate estimates of water fluxes [*Stets et al.*, 2009].

[3] Groundwater exchange can be a large component of water flux in lakes, but it is difficult to measure accurately and is often calculated as the residual of the hydrologic budget or sometimes ignored [*Winter*, 1981; *Hunt et al.*, 1996]. Direct measurements of seepage can be made using seepage meters or minipiezometers, but these measurements are labor intensive and only provide information about single points spatially and temporally [*Lee*, 1977; *Rosenberry et al.*, 2008]. Seepage fluxes also are commonly calculated by analytical or numerical mathematical models using Darcy's law, where hydraulic gradients and hydraulic conductivity need to be known or estimated [*Merritt and Konikow*, 2000]. The three-dimensional geologic framework through which groundwater must move is virtually impossible to map accurately so simplifying assumptions must be made concerning the geologic structure of an area in using such models.

[4] Chemical and isotopic tracers greatly aid the effort to quantify groundwater fluxes in lakes and stream systems [*Gibson and Edwards*, 1996; *Walker and Krabbenhoft*, 1998]. Stable isotope studies have been used to expand limited field data to larger spatial and temporal scales [*Gibson and Edwards*, 2002; *Sacks*, 2002], but quantification of hydrologic terms is most effective when performed in conjunction with other field-based data so that only one component of the hydrologic budget is estimated and/or multiple sources are used to corroborate results [*LaBaugh et al.*, 1997]. Both steady state [*LaBaugh et al.*, 1997; *Walker and Krabbenhoft*, 1998] and non steady state [*Gibson and Edwards*, 1996; *Gurrieri and Furniss*, 2004] isotopic

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models have been used to quantify portions of lake hydrologic budgets. Steady state models provide the benefits of simplicity and precision [*Sacks*, 2002], but may not apply to lakes with large water fluxes or extreme seasonal variations.

[5] In this paper, we present complete hydrologic budgets for six lakes located in the Shingobee River headwaters study area (Minnesota, United States). The budgets were calculated by combining field observations with a non steady state descriptive model of lake surface water δ^{18}O. Water oxygen stable isotopes have been shown to be effective in this watershed [*LaBaugh et al.*, 1997], but have only been used in one lake, Williams Lake, partly because several other lakes in this watershed violate assumptions of steady state. The model presented in this study aims to expand our understanding of the hydrologic fluxes in this watershed by using an approach which is more broadly applicable. We used manual optimization to determine the value of groundwater flux that produced the best fit between modeled and observed lake surface δ^{18}O signature (δL). We assessed the accuracy of our isotope model by comparing the results to field-based observations of streamflow and water yield. We also analyzed model sensitivity and robustness to assess model performance in open-basin versus closed-basin lakes. The data were collected from the beginning of 2004 to spring of 2005 when an extensive effort was made to sample a time series of all components of the hydrologic system in order to obtain detailed knowledge of the hydrologic dynamics of the lakes.

2. Site Description

[6] The Shingobee watershed is located in north central Minnesota and is part of the larger upper Mississippi River watershed, with hydrologic flows generally from south to north (Figure 1). The boundary of the Shingobee River headwaters upstream of the outlet of Shingobee Lake (SLO) shown in Figure 1a was drawn on the basis of land surface topography and therefore depicts the surface water watershed. It is likely that the groundwater watershed extends beyond this in places, as indicated by *Winter et al.* [2003].

[7] Approximately 120 m of sand and silt overlay thick deposits of carbonate-rich glacial till [*Winter and Rosenberry*, 1997]. Advective groundwater transport occurs throughout the watershed and enters surface water bodies as either diffuse seepage in areas with higher hydrologic conductivities (8.0×10^{-5} m s^{-1}) or focused spring water discharge where hydraulic conductivities are lower ($1–2 \times 10^{-5}$ m s^{-1}) [*Filby et al.*, 2002]. Crystal and Williams Lakes are closed-basin lakes located in the upper part of the watershed (Figure 1b). Hydrologic exchange in these lakes occurs entirely through diffuse groundwater seepage, precipitation, and evaporation. The other lakes considered in this study (Mary, Island, Steel, and Shingobee) are connected by the Shingobee River. These lakes are located in sediments with lower hydraulic conductivity than the closed-basin lakes and groundwater flux tends to be focused into visible springs around lake edges. The Shingobee River Headwaters originate as a spring downgradient of Williams Lake (Figure 1) and enter Mary Lake after traversing a small pond. Water and inorganic carbon stable isotope data suggest that the water in the Shingobee River Headwaters is composed primarily of groundwater rather than Williams Lake outflow (P. F. Schuster et al., unpublished data, 1995). The Shingobee River gains hydrologic inputs throughout the watershed from groundwater and surface runoff and exits the watershed below Shingobee Lake with an average discharge of 0.3 to 0.4 m^3 s^{-1} [*Rosenberry et al.*, 2003].

[8] The Shingobee watershed has been the focus of intense hydrologic and biogeochemical studies for more than 30 years [*LaBaugh et al.*, 1995; *Winter*, 1997] and groundwater flows, surface water flows, and meteorological conditions are monitored mostly around Shingobee and Williams Lakes. Groundwater wells are located upgradient from Williams, Mary, and Shingobee Lakes and downgradient from Crystal and Williams Lakes. Figure 1 depicts the locations of the four permanent stream gages (defined here as a location where river stage is recorded continuously, discrete current-meter discharge measurements are made, and a stage-discharge relation is developed and used to obtain daily discharge values) at the outlet of Little Shingobee Lake (Shingobee River Tributary (SRT)), on the Shingobee River just upstream from this tributary (Shingobee River Inlet (SRI)), at the outlet of Shingobee Lake (SLO), and where Shingobee River exits the watershed 2 km below Shingobee Lake (SRO). Gaged land-based meteorological stations and floating instrumented rafts are located at Shingobee and Williams Lakes (Figure 1a). We used field data collected in the watershed to constrain our hydrologic model and for purposes of comparison.

3. Methods

[9] A total of 598 water stable isotope samples were collected and analyzed throughout the watershed during 2004–2005 and include lake surface waters, stream segments, groundwater, springs, and precipitation. A similar, although far less comprehensive, water stable isotope sampling effort began in this watershed in 2001. Data from calendar year 2002 were also used as a way of testing overall model performance. Approximately 400 individual measurements which were most relevant to constructing the model presented in this study are available in the auxiliary material.[1] Unfiltered water was collected into cleaned and rinsed 60 mL clear glass bottles fitted with black polycone caps (Scientific Specialties Service, Inc.). The bottles were filled completely and carefully to avoid the introduction of air bubbles.

[10] Water table wells were constructed by augering a hole to about half a meter below the water table and inserting a well screen and casing assembly into the hole so the screen fully filled with groundwater near the water table. All of the wells used in this study were constructed prior to 2004. For each sample, a hand pump fitted with silicone tubing was used to remove 3 to 4 volumes of well water and fill sample bottles. The wells used in this study were each sampled 5 to 10 times in 2004 and springs entering Shingobee Lake, Mary Lake, Spring Lake, and Shingobee River downstream from Steel Lake were sampled once or twice. Lake surface water samples were collected biweekly at 0.2 m depth near the center of each lake during the ice-free season. In winter, water samples were collected monthly by drilling a hole in the ice with a manual ice auger and sampling water at 0.2 m below the ice using a hand crank pump fitted with silicone tubing. Stream segments were sampled by hand biweekly during the ice-free season and approximately monthly in the winter.

[1] Auxiliary material data sets are available at ftp://ftp.agu.org/apend/wr/2009wr007793. Other auxiliary material files are in the HTML.

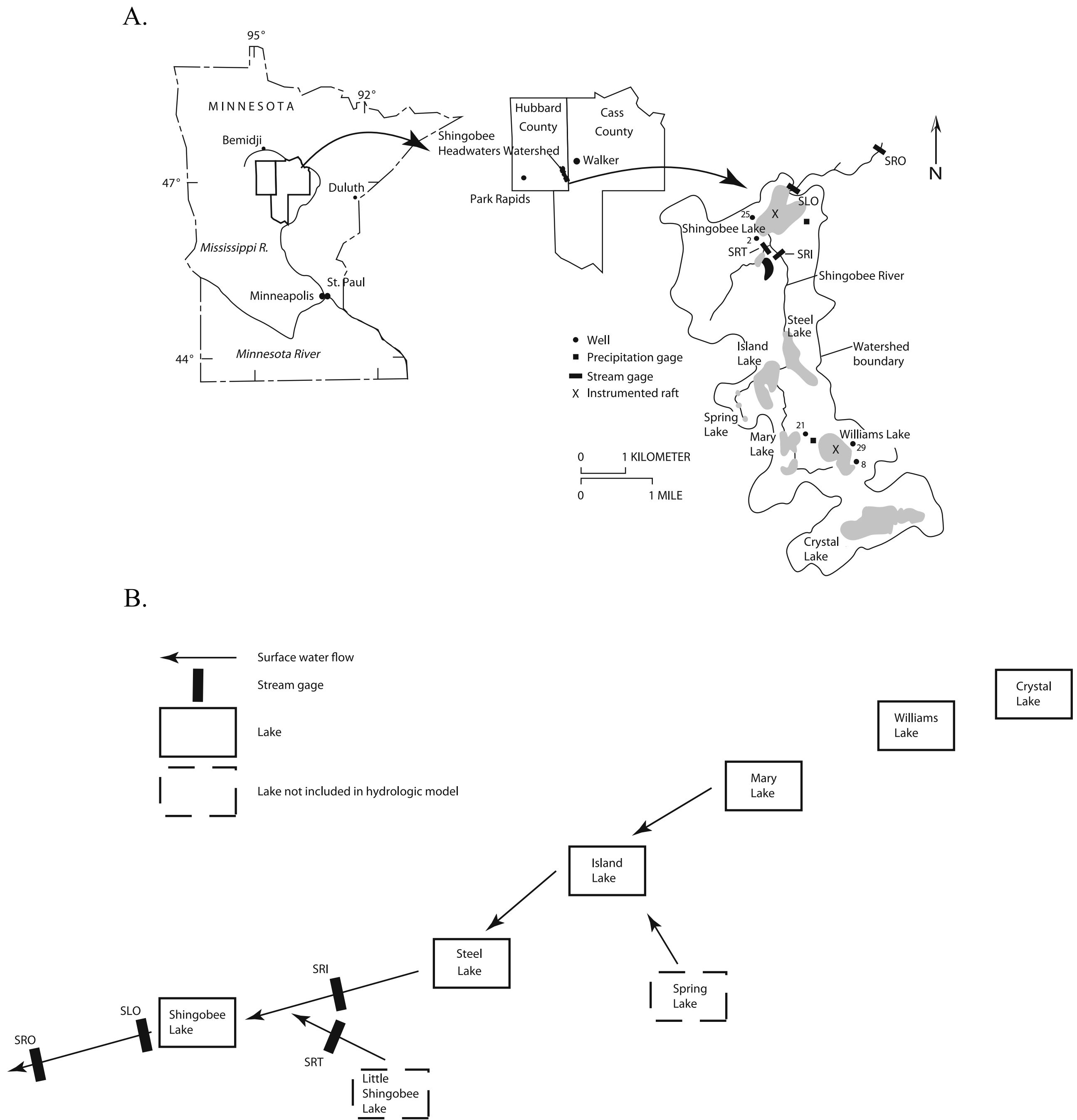

Figure 1. (a) The Shingobee headwaters watershed. Groundwater wells used in this study are denoted as numbered circles. (b) Diagram showing the conceptual framework used for developing the hydrologic model of the Shingobee headwaters watershed.

These included Mary Lake outlet, Island Lake inlet, Island Lake outlet, Steel Lake outlet, Shingobee Lake inlet, and the four permanent stream gages (SRT, SRI, SLO, and SRO) (Figure 1a). Additional samples were collected daily during snowmelt and following major precipitation events (>12.5 mm) at Crystal Lake, Mary Lake, Island Lake outlet, SRI, and SLO. Sampling in calendar year 2002 was much less dense with samples collected only at Shingobee and Williams Lakes and at regular 2 week intervals at the Williams precipitation gage.

[11] The δ^{18}O analyses were performed at the International Atomic Energy Agency Isotope Hydrology Laboratory in Vienna, Austria by a Finnigan DeltaPlus dual inlet mass spectrometer with associated self-constructed 48 port water-gas equilibration unit. Three milliliters of water were used for the equilibration process, where the sample was first equilibrated with hydrogen gas for 1 h using a platinum catalyst and subsequently with carbon dioxide gas for 5 h. This procedure yielded results for both δ^2H and δ^{18}O, but only the δ^{18}O results are considered in this manuscript. The standard uncertainty of a sample measurement for δ^{18}O was 0.07‰ versus Vienna Standard Mean Ocean Water (VSMOW). Samples were calibrated using two internal laboratory water standards with isotopic compositions well

constrained by several calibrations, carried out under repeatability conditions directly in reference to VSMOW and Standard Light Antarctic Precipitation (SLAP).

3.1. Precipitation

[12] The volume of precipitation reaching a lake (*Pr*) was assumed to be equal to the sum of precipitation falling directly on the lake surface and precipitation entering the lake as overland or interflow. Precipitation was measured at land-based meteorological stations located at Williams and Shingobee Lakes using continuously recording weighing-bucket gages and standard bulk nonrecording gages. The volume of precipitation falling directly onto lake surfaces was calculated as the product of lake surface area (m^2) and precipitation depth (m). Overland and interflow were calculated to be 5% of total precipitation based on streamflow recession studies at SRI following major precipitation events. The volume of water reaching lakes in this way was calculated as the product of watershed area (m^2, lake surface area exclusive) and 5% of precipitation depth (m). Samples of precipitation for oxygen stable isotopes were taken from the bulk collector at Williams Lake following each precipitation event. Evaporation of precipitation samples prior to collection could affect isotopic signatures. No steps were taken to prevent this evaporation, but the slope and intercept of the local meteoric water line were 8.1 ± 0.2 and 11.2 ± 2.3 ($R^2 = 0.99$, $P < 0.0001$, $F_{20,\,1} = 1521$), suggesting that evaporation of these samples was minimal [*Kendall and Caldwell*, 1998].

3.2. Stream Gauging

[13] Stream discharge was determined from measured stage and a stage-discharge relation developed for each permanent stream gage. Manual measurements of discharge using a Price pygmy current meter, and concurrent observations of stage, were made over a wide range of discharges to develop a stage-discharge relation for each gauging station. Additionally, occasional manual measurements of stream discharge were made on the Shingobee River between the lakes within the headwaters study area. Lake stage was measured daily at Williams and Shingobee Lakes.

[14] Data from four U.S Geological Survey stream gages on the Crow Wing River (stations 05247500 at Pillager, Minnesota, and 05244000 at Nimrod, Minnesota) and Upper Mississippi River (stations 05200510 at Bemidji, Minnesota, and 05211000 at Grand Rapids, Minnesota) were used to calculate regional water yield, defined as average annual stream discharge divided by watershed area. All of the data used in the regional water yield calculation are available at the U.S. Geological Survey National Water Information System (NWIS) (http://waterdata.usgs.gov/nwis). Details of streamflow data collection at these four sites are available in work by *Mitton et al.* [2006].

3.3. Evaporation

[15] Evaporation was determined at Shingobee and Williams Lake by the Priestley-Taylor method,

$$E = \frac{\beta\left(\frac{s}{s+\gamma}\right)(Q_n - Q_x)}{\Lambda} \tag{1}$$

where

- E evaporation from the lake surface (m d^{-1});
- β an empirically derived constant, usually 0.26, dimensionless;
- s the slope of the saturated vapor pressure-temperature curve at mean air temperature, determined using an equation presented by *Lowe* [1977];
- γ the psychrometric constant, and was determined using an equation presented by *Fritschen and Gay* [1979];
- Q_n net radiation (W m^{-2});
- Q_x the change in heat stored in the lake between thermal surveys (W m^{-2});
- Λ the latent heat of vaporization (W m^{-3}).

Evaporation rates were averaged over the days bounded by thermal surveys. Thermal surveys are temperature-depth profiles made at about 10 locations in the lake. Monthly evaporation was calculated as time-weighted daily values summed over the month, then multiplied by the area of the lake. Evaporation rate at Mary, Island, and Steel Lakes was assumed to be identical to that of Shingobee Lake, whereas evaporation rate at Crystal Lake was assumed to be identical to Williams Lake. No samples of water vapor were collected for isotope analyses.

3.4. Isotopic Composition of Evaporation

[16] Evaporation preferentially removes isotopically lighter water and causes the remaining lake water to become enriched in $H_2^{18}O$. Direct observation of the isotopic composition of the evaporating water, δ_E, is not technically feasible, but can be estimated from the linear resistance model formulated by *Craig and Gordon* [1965]:

$$\delta E = \frac{(\alpha^* \delta L - h\delta A - \varepsilon)}{[1 - h + 10^{-3}\kappa(1-h)]} \tag{2}$$

where δA is the isotopic composition of the local atmospheric moisture, h is the relative humidity at the lake surface temperature, α^* is the equilibrium isotope fractionation factor at the temperature of the air-water interface, $\varepsilon = 1000(1 - \alpha^*) + [\kappa(1 - h)]$ is the total fractionation factor. κ is an empirical constant relating the kinetic fractionation factor and relative humidity and was 14.3 [*Gilath and Gonfiantini*, 1983]. All δ and ε values are in per mil (‰). Calculation of α^* was performed by relating it to the equilibrium water liquid-vapor transition ($\alpha_{l\text{–}v}$) such that $\alpha^* = 1/\alpha_{l\text{–}v}$ [*Kendall and Caldwell*, 1998]. We assumed that $\alpha_{l\text{–}v}$ was linearly related to temperature as $\alpha_{l\text{–}v} = 1.0117 - [9.5 \text{ X } T_a/100000]$ [mathematical relation derived from *Krabbenhoft*, 1990]. We calculated δA as $\delta A = \alpha^*\delta Pr - \varepsilon^*$. The Tetens equation (described by *Horita et al.* [2008]) was used to calculate saturated atmospheric vapor pressure, $e_A{}^*$, and saturated vapor pressure at lake surface temperature, $e_0{}^*$. The term h was then calculated from relative humidity measured at 2 m above the lake surface (RH) as $h = (e_A{}^*\text{RH})/(100{}^*e_0)$. RH and h are sometimes considered to be equivalent [*Hostetler and Benson*, 1994], but *Horita et al.* [2008, p. 29] describe h as "the RH of ambient air normalized to the saturation vapour pressure at the temperature of the surface, not at air temperature." We chose to calculate h according to this verbal description because the process of evaporative modification depends most directly upon conditions in the boundary layer above the lake surface, which could be presumed to have a temperature equivalent to the lake surface, but a water vapor pressure similar to ambient air 2 m above the lake surface.

[17] This formulation of δE assumes that the upward evaporative fluxes do not significantly perturb the atmospheric boundary layer [*Horita et al.*, 2008], an assumption that is violated in large lakes with high evaporation rates such as the Laurentian Great Lakes [*Gat*, 2000] and Pyramid Lake [*Hostetler and Benson*, 1994]. However, working at a small lake in northern Wisconsin, *Krabbenhoft* [1990] found that δA was not substantially affected by lake-derived water vapor, suggesting that for small lakes such as the ones included in this study, equation (2) gives a suitable approximation of δE.

[18] The δE values were evaluated daily in the model. Interpolated values of δPr were used to calculate δA between precipitation sampling throughout calendar year 2004. Values of δE were sometimes volatile, particularly in early spring and late autumn. Therefore, we constrained δE values used in equation (3) to the 5th through the 95th percentile of all calculated δE values for each lake.

4. Model Description

[19] We developed a stable isotope-based hydrologic model for these six lakes using a time series of δ^{18}O water isotope samples collected throughout the watershed during 2004–2005. The model estimates surface water and groundwater fluxes and is constrained by gaged river flow at SRI (Q_{SRI}), measured Pr in the watershed, and E calculated using the Priestley-Taylor method.

[20] The δ^{18}O composition of lake surface water is determined by δ^{18}O of source water and isotopic modification through evaporation or freezing [*Gat*, 1996]. Quantification of evaporative modification of lake water allows source water inputs to be quantified as well. For example, a steady state isotope mass balance model can be used to estimate groundwater input (Gi) to lakes. Typically Gi is evaluated by solving a steady state mass balance equation for Gi and then calculating monthly or annual average values for the other terms in the mass balance equation [*Krabbenhoft et al.*, 1994]. This approach works well in lakes with long water residence time and moderate seasonal variability, but is less useful for lakes with large water fluxes or extreme seasonal variability, in which the assumption of steady state may be a poor one.

[21] For this study, we developed a descriptive model of lake surface δ^{18}O (δL) and inferred Gi by selecting the value which achieved the closest match between observed and modeled δL. The model was evaluated at daily time steps and includes terms for all processes known to affect δL:

$$\delta L_t = \delta L_{t-1} + \left[\begin{array}{c} (\delta Gi \times Gi) + (\delta Si \times Si) + (\delta Pr \times Pr) - (\delta L_{t-1} \times Go) - (\delta L_{t-1} \times So) \\ -(\delta E \times E) + (\delta Hypo \times \Delta Hypo) - (\delta Ice \times \Delta Ice) \end{array} \right] \times \frac{\Delta t}{L_{t-1}} \quad (3)$$

where δL_t is the lake surface δ^{18}O at time (t), δL_{t-1} is the lake surface δ^{18}O at time ($t-1$), Gi is groundwater inflow, Si is surface water inflow, Pr is precipitation (unsaturated zone flow inclusive), Go is groundwater outflow, So is surface water outflow, E is evaporation, $\Delta Hypo$ is the volumetric rate of hypolimnetic water mixed into the epilimnion, ΔIce is volumetric rate of ice formation (positive) or ablation (negative), Δt is the time step in days, and L_t is the volume of the surface mixed layer at time t. All fluxes are given in m^3 d^{-1} and L_t is given in m^3. The term δX denotes δ^{18}O of component X in ‰. This model encompasses all terms we believe to be influencing ^{18}O signature in the lakes. Some terms, such as E, Pr, $\Delta Hypo$, and ΔIce, are set to zero for part of the year. Other terms are only applicable to certain lakes in the watershed. For example, Si and So are both set to zero in the closed-basin lakes, Williams and Crystal.

5. Quantification of Surface Water and Groundwater Fluxes

5.1. Surface Water

[22] All quantified surface water fluxes in this model occur at lake outlets or specific reaches of the Shingobee River. We relied on Q_{SRI} as a constraint on the overall surface water fluxes throughout the watershed because this stream gage has been in continuous operation for many years and has a well-developed stage-discharge relation.

[23] Surface water flux into Shingobee Lake (Q_{SLI}) was quantified as the sum of Q_{SRI} and streamflow at SRT (Q_{SRT}). Surface water outflow from Shingobee Lake (Q_{SLO}) was estimated at SLO. The stage-discharge relation at SLO was less well developed and was significantly affected by the presence of beaver dams in 2004. We compared modeled discharge from Shingobee Lake, Q_{SLO}, to field observations of outflow from Shingobee Lake. We modeled Q_{SLO} using mass balance

$$Q_{SLO} = Q_{SLI} + Gi + Pr - E \quad (4)$$

[24] Upstream from SRI, the Shingobee River gains water through groundwater input, overland flow, and precipitation-associated unsaturated zone flow (Pr, see section 3.1). It was necessary to calculate the magnitude of these water sources to Shingobee River in order to adequately quantify streamflow in the watershed. We began by developing estimates of Pr and groundwater inflow (Gi) between SRI and the outlet of Steel Lake. In contrast to the lakes considered in this study, we assumed that direct precipitation onto this section of the stream surface was negligible and therefore Pr was entirely due to precipitation-associated unsaturated zone flow. This allowed us to calculate daily streamflow values at the outlet of Steel Lake (Q_{STLO}) using the relation

$$Q_{STLO} = Q_{SRI} - Gi - Pr \quad (5)$$

Q_{SRI} was obtained from stream gauging measurements, *Pr* was assumed to be 5% of the precipitation volume on the watershed directly adjacent to this stretch of the Shingobee River, and *Gi* was modeled using stable isotope data. We were able to calculate Q_{STLO} independently of *Gi* for 15 dates in 2004 for which isotopic data were available for both SRI and the outlet of Steel Lake, δSRI and $\delta STLO$, respectively using the equation

$$Q_{STLO} = \frac{Q_{SRI}(\delta SRI - \delta Gi) + Pr(\delta Gi - \delta Pr)}{(\delta STLO - \delta Gi)} \quad (6)$$

Groundwater isotopic composition, δGi (−10.16‰), was measured directly from groundwater springs located along this reach of the Shingobee River. The isotopic composition of unsaturated zone flow (δPr) was assumed to be identical to precipitation collected in the watershed during each precipitation event exceeding 12.5 mm. This procedure allowed us to model Q_{STLO} for 15 dates in 2004, which we then used to model *Gi* in this stretch of the river by rearranging equation (5) to solve for *Gi* for each of the 15 dates. *Gi* varies by <10% on an annual basis in this watershed [*LaBaugh et al.*, 1997] so we assumed that *Gi* was a constant equal to the average value. We then estimated Q_{STLO} on a daily basis in 2004 using equation (5). For the purpose of modeling hydrologic flows in Steel Lake, we used 7 day moving averages of calculated Q_{STLO} because of the uncertainties associated with this type of measurement.

[25] Surface water flow into Steel Lake (Q_{STLI}) was then calculated from the hydrologic budget of Steel Lake as

$$Q_{STLI} = Q_{STLO} + E - Gi - Pr \quad (7)$$

Surface water outflow from Island Lake (Q_{ILO}) was assumed to be identical to Q_{STLI} because the two lakes are separated by a short (<200 m) reach of the Shingobee River and *Gi* is believed to be minimal in this section of the river.

[26] Mary Lake is a headwater lake which receives a large input of focused spring water discharge and diffuse groundwater seepage. Therefore, surface water input to Mary Lake was assumed to be negligible and surface water outflow (Q_{MLO}) was calculated from other quantified components of the hydrologic budget

$$Q_{MLO} = Gi + Pr - E \quad (8)$$

[27] Island Lake surface water inputs arise from multiple sources including the Shingobee River via Mary Lake, the outlet of Spring Lake, and a series of upgradient wetlands. Therefore, we did not quantify Island Lake surface water inputs directly. Instead, we compare Q_{ILO} to other components of the Island Lake hydrologic budget to infer a reasonable partitioning of surface water inputs between the various sources to Island Lake.

5.2. Groundwater

[28] Groundwater exchanges with surface water throughout the Shingobee headwaters watershed and is a substantial part of the hydrologic budget of all six lakes considered in this study. Groundwater flow into both Shingobee Lake and Williams Lake has been considered in other studies [*LaBaugh et al.*, 1997; *Rosenberry et al.*, 2000; *Rosenberry et al.*, 1997]. However, all *Gi* and *Go* values presented in this study are derived from the application of equation (3) to each lake.

[29] Groundwater isotopic composition, δGi, was determined from groundwater wells and springs sampled throughout the watershed. Overall, δGi ranged from −9.1‰ in wells 29 and 8 to −12.9‰ in well 21 (Figure 1a). Stable isotope data from wells and springs used in this study are presented in the auxiliary material. For Crystal Lake, δGi was set equal to −12‰ based on a study of shallow (20–60 cm deep) groundwater adjacent to the upgradient portion of that lake (P. F. Schuster, unpublished). For Williams Lake, δGi was assumed to be −9.1‰ based on samples taken from wells 29 and 8 ($n = 25$, auxiliary material, Figure S1). In the lower portion of the watershed, in which the open-basin lakes reside, δGi in spring and groundwater well samples was much more uniform, −11.1 to −11.7‰. For Shingobee Lake, δGi was assumed to be equal to the mean value of all samples taken from wells 2 and 25, −11.5‰ ($n = 44$, auxiliary material, Figure S1). This value was also similar to average stable isotopic composition of springs sampled on Shingobee Lake. Steel Lake was also assumed to have a δGi of −11.5‰ because of its proximity to Shingobee Lake and because of the relative consistency of δGi in the lower part of the watershed. Springs sampled on the southwestern edge of Island, Mary, and Spring Lakes had stable isotopic composition between −11.7 and −11.8‰, so δGi was assumed to be −11.7 for both Mary and Island Lakes. In addition to the numerous springs on the southwestern shore of Mary Lake, groundwater inputs to Mary Lake come from the shoreline segment adjacent to well 21 (−12.9‰) and the spring which constitutes the Shingobee River Headwaters (−10.8‰). However, previous modeling in this watershed has shown steep water table gradients west and southwest of Mary Lake [*Filby et al.*, 2002], suggesting that groundwater inputs could be dominated by fluxes similar to the springs on that shore. Also, even if we assume that groundwater fluxes to Mary Lake from all three sources are similar in magnitude; average δGi would be −11.7‰.

[30] We assumed that *Gi* and δGi were constant throughout the year because groundwater residence time exceeds 1 year in most parts of the watershed [*Reddy et al.*, 2006] and *Gi* is believed to be relatively constant intra-annually [*Filby et al.*, 2002]. The value of *Gi* that produced the best fit between the observed and modeled δL values was selected. Similar methods were used to calculate *Go* in Crystal and Williams Lakes except that at time *t*, δGo_t was assumed to be equal to δL_t. A fuller explanation of the statistical methods used in this study appears in section 6. Variations of equation (3) were used to reflect the hydrologic setting of each lake included in the study (Table 1). In cases where more than one component of the water budget was unknown, we used algebraic substitution based on the steady state water budget to develop equations in which the groundwater flux term of interest (*Gi* or *Go*) was the only unknown. This procedure is described in greater detail by *Walker and Krabbenhoft* [1998] and *Sacks* [2002]. The principal assumptions were that the closed-basin lakes had no surface water flows; Mary Lake received water only as *Gi* and *Pr* and exported water only as *So* and *E*; all other open-basin lakes received water through *Si*, *Gi*, and *Pr* but exported water only as *So* and *E* (Figure 1 and Table 1).

Table 1. Model Equations Used to Estimate *Gi* for All Lakes and *Go* for Williams and Crystal Lakes in This Study

Lake	Model
Williams and Crystal	$\delta L_t = \delta L_{t-1} + \begin{bmatrix} Gi(\delta Gi - \delta L_{t-1}) + Pr(\delta Pr - \delta L_{t-1}) \\ +E(\delta L_{t-1} - \delta E) + \Delta Hypo(\delta Hypo - \delta L_{t-1}) - \Delta Ice(\delta Ice - \delta L_{t-1}) \end{bmatrix} \frac{\Delta t}{L_{t-1}}$ [a]
	$\delta L_t = \delta L_{t-1} + \begin{bmatrix} Go(\delta Gi - \delta L_{t-1}) + Pr(\delta Pr - \delta Gi) \\ +E(\delta Gi - \delta E) + \Delta Hypo(\delta Hypo - \delta Gi) - \Delta Ice(\delta Ice - \delta Gi) \end{bmatrix} \frac{\Delta t}{L_{t-1}}$ [b]
Shingobee	$\delta L_t = \delta L_{t-1} + \begin{bmatrix} Q_{SLI}(\delta SLI - \delta L_{t-1}) + Gi(\delta Gi - \delta L_{t-1}) + Pr(\delta Pr - \delta L_{t-1}) \\ +E(\delta L_{t-1} - \delta E) + \Delta Hypo(\delta Hypo - \delta L_{t-1}) - \Delta Ice(\delta Ice - \delta L_{t-1}) \end{bmatrix} \frac{\Delta t}{L_{t-1}}$ [c]
Steel	$\delta L_t = \delta L_{t-1} + \begin{bmatrix} Pr(\delta Pr - \delta ILO) + Gi(\delta Gi - \delta ILO) \\ +E(\delta ILO - \delta E) + Q_{STLO}(\delta ILO - \delta L_{t-1}) + \Delta Hypo(\delta Hypo - \delta L_{t-1}) - \Delta Ice(\delta Ice - \delta L_{t-1}) \end{bmatrix} \frac{\Delta t}{L_{t-1}}$ [d]
Island	$\delta L_t = \delta L_{t-1} + \begin{bmatrix} Pr(\delta Pr - \delta MLO) + Gi(\delta Gi - \delta MLO) \\ +E(\delta MLO - \delta E) + Q_{ILO}(\delta MLO - \delta L_{t-1}) + \Delta Hypo(\delta Hypo - \delta L_{t-1}) - \Delta Ice(\delta Ice - \delta L_{t-1}) \end{bmatrix} \frac{\Delta t}{L_{t-1}}$ [e]
Mary	$\delta L_t = \delta L_{t-1} + \begin{bmatrix} Gi(\delta Gi - \delta L_{t-1}) + Pr(\delta Pr - \delta L_{t-1}) \\ +E(\delta L_{t-1} - \delta E) + \Delta Hypo(\delta Hypo - \delta L_{t-1}) - \Delta Ice(\delta Ice - \delta L_{t-1}) \end{bmatrix} \frac{\Delta t}{L_{t-1}}$ [f]

[a]Used to calculate *Gi* in both lakes; independent of *Go*; assumed $\delta Go = \delta L_{t-1}$.
[b]Used to calculate *Go* in both lakes; independent of *Gi*.
[c]Independent of Q_{SLO}; assumed $\delta SLO_{t-1} = \delta L_{t-1}$.
[d]Independent of surface inflow; assumed $\delta Si_{t-1} = \delta ILO_{t-1}$; assumed $\delta STLO_{t-1} = \delta L_{t-1}$.
[e]Independent of surface inflow; assumed $\delta Si_{t-1} = \delta MLO_{t-1}$; assumed $\delta ILO_{t-1} = \delta L_{t-1}$.
[f]Assumed *Si* was negligible.

5.3. Lake Volumes and Mixing Depths

[31] Lake volumes were obtained from planimetered bathymetric maps for Crystal, Williams, Island, Steel, and Shingobee Lakes. A detailed bathymetric map does not exist for Mary Lake so we estimated lake volume by applying a general lake area depth model developed for Minnesota lakes [*Hondzo and Stefan*, 1993]:

$$\frac{Area_z}{Area_0} = 1.14\exp\left(-2.1\frac{z}{z_{\max}}\right) - 0.15 \tag{9}$$

where $Area_z$ is lake cross-sectional area (km^2) at depth z (m), $Area_0$ is lake surface area (km^2), and $z_{\max}$ is maximum lake depth. Lake volume at each depth was then calculated as a truncated cone.

[32] The lakes in this watershed are dimictic and experience thermal stratification in both the summer and winter [*Striegl and Michmerhuizen*, 1998]. During stratification the processes modifying δL in these lakes, such as evaporation and hydrologic exchange are restricted to the surface mixed layer. Therefore, we used the volume of the surface mixed layer for L_t in equation (3). Oxygen-temperature-depth profiles were obtained from Shingobee and Williams Lakes throughout 2004 (n = 16) and were used to calculate L_t. No depth profiles were collected from the other lakes during 2004, but previous work in this watershed showed that stratification patterns in Williams were similar to Crystal, whereas the stratification patterns in Shingobee were similar to Mary, Island, and Steel Lakes. Therefore, we used the depth profiles from Shingobee to calculate L_t in Mary, Island, and Steel Lakes and the depth profiles from Williams to calculate L_t in Crystal Lake.

5.4. Hypolimnetic Mixing

[33] As mentioned in section 5.3, the processes modifying δL act principally on the lake surface and so hypolimnetic $\delta^{18}O$ signatures remain largely unaltered during periods of stratification. This assertion is confirmed by $\delta^{18}O$ samples collected from 1 m and 8 m deep in Shingobee and Williams from 2001 to 2005. Erosion of the thermocline in late summer mixes hypolimnetic water with a different $\delta^{18}O$ signature into the lake surface thereby influencing lake surface $\delta^{18}O$. Therefore, we included a term in equation (3) to account for this process. For each lake, we assumed that $\delta Hypo$ was equal to δL at the time the lake stratified, mid-May 2004. So that δL continued to change throughout the stratified season, but $\delta Hypo$ did not. As the thermocline eroded in early autumn, we assumed that the hypolimnetic water mixed conservatively into the epilimnion and influenced δL. $\Delta Hypo$ was calculated from the observed thermocline depth in temperature profiles obtained between 2 August and 21 October 2004.

5.5. Ice Formation and Isotopic Signature

[34] Lakes in north-temperate regions can produce large volumes of ice during the winter and the isotopic fractionation associated with ice freezing can affect δL. Freezing selectively removes $H_2^{18}O$ from lake surface waters, causing ice to be enriched in ^{18}O relative to lake water. We incorporated an isotopic fractionation of 3.5‰ for the formation of ice in these lakes [*Kendall and Caldwell*, 1998] into equation (3). ΔIce was calculated from several winter 2003–2004 observations of ice thickness, which revealed a maximum ice thickness of 655 mm on 17 March 2004. We assumed that ice grew at a linear rate of 5.8 mm d^{-1} from 25 November 2003 until 17 March 2004 and then decayed at a rate of 21.3 mm d^{-1} from 17 March 2004, when average daily temperatures began exceeding 0°C, until 17 April 2004, when the lakes became completely free of ice. Melting ice returns water enriched in ^{18}O back to the lake surface so the effect of ice on δL is transient but important for accurately describing temporal patterns in δL in these lakes. Sublimation of ice from the lake surface was assumed to be negligible.

6. Statistics and Model Conditions

[35] The value for *Gi* which provided the best fit between observed lake surface isotopic composition, δL_{OBS}, and modeled δL was found by an unweighted manual optimization. First we ran the model with 200 values of *Gi* which

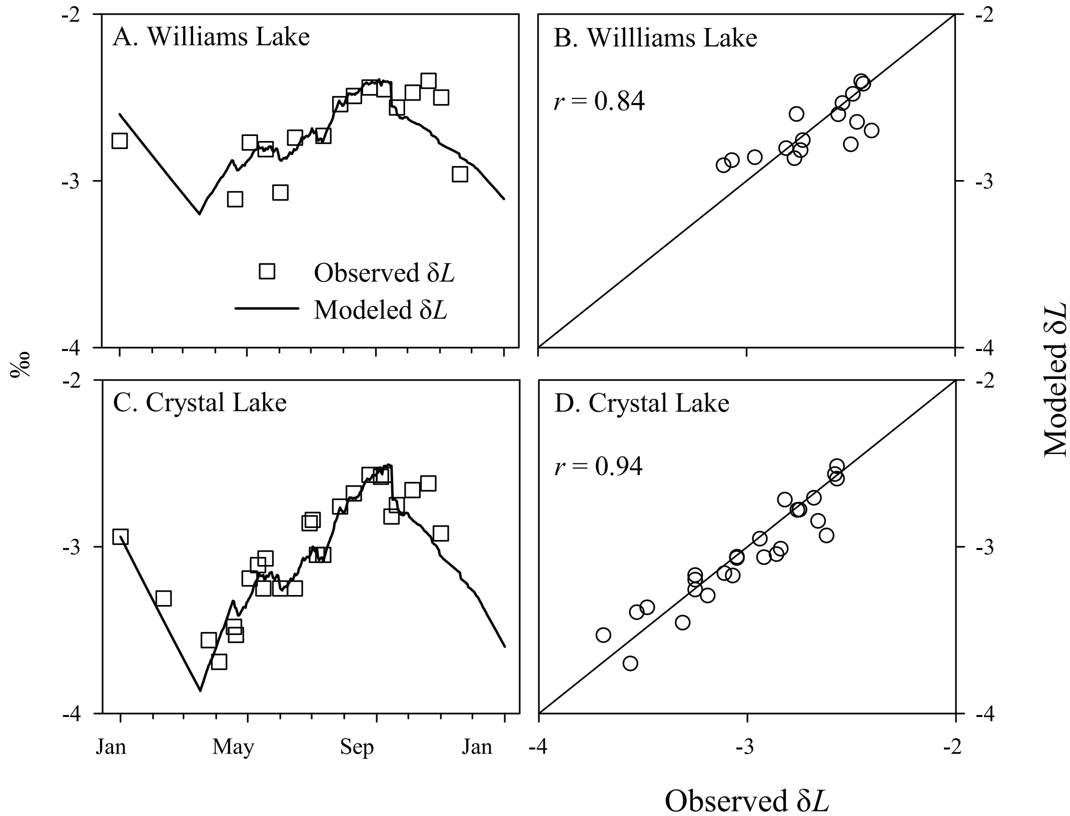

Figure 2. Comparison of modeled and observed lake surface δ^{18}O water isotopic signature in the closed-basin lakes (a, b) Williams and (c, d) Crystal. Open squares in Figures 2a and 2c show observed δ^{18}O, and the solid line shows daily modeled δ^{18}O. Observed δ^{18}O values appear on the *x* axis in Figures 2b and 2d, and modeled δ^{18}O values from corresponding dates appear on the *y* axis. The simple correlation coefficient (*r*) between these values is also shown.

we believed bracketed the *Gi* value for each lake. We calculated the variance for the fit between the observed and modeled data set at each *Gi* value as $\sigma_{\mathrm{Gi}} = \Sigma\,[\delta L_{\mathrm{t}} - \delta L_{\mathrm{OBS(t)}}]^2/n$, where n is the number of δL_{OBS} for each lake. The calculated variance reaches a minimum at the *Gi* value which provides the best fit between the modeled and observed data sets. We then fit a Gamma distribution to the 200 values of σ_{GI} and present the central 66 percentile of this distribution as a range of likely values of *Gi*. We selected this range because it should give a distribution similar to that of the standard deviation. An identical routine was used to estimate *Go* in Crystal and Williams Lakes. The Gamma distributions and quantile calculations were performed using JMP IN 5.1.2 (SAS Institute, Inc.).

[36] The model was run in Berkeley Madonna 8.3.1.1 modeling software (R.I. Macey and G.F. Oster, UC Berkeley, California, United States) with δL_0 set to $\delta L_{OBS(0)}$ and evaluated at a daily time step thereafter (Δt = 1). *Pr* and *E* were set to zero from 1 January 2004 to 1 April 2004 and 28 November 2004 to 31 December 2004, the periods of complete ice cover.

[37] A sensitivity analysis was conducted for Shingobee and Williams Lakes, the two lakes with the most complete data sets. The magnitude of the sensitivity analysis was meant to convey a sense of the uncertainty associated with this model and its basic assumptions. For example, the uncertainty in *E* is likely to be 15% while the uncertainty in *So* (Shingobee Lake only) is likely to be 5%. Less is known about the uncertainty associated with δGi, so we used ±1‰ for demonstrative purposes.

[38] We also tested model performance by repeating model calculations for calendar year 2002 in Shingobee and Williams Lakes using identical assumptions, equations, δGi, and calculated *Gi* for calendar year 2004. Other model inputs, including initial δL values, lake mixing depths, and all relevant meteorological data were collected at Shingobee and Williams Lakes during calendar year 2002. We describe the results of this exercise below and detailed information is given as auxiliary material.

7. Results and Discussion

7.1. Groundwater Model Results

[39] The best fit *Gi* values and model output are summarized in Table 2 and Figures 2 and 3. Calculated *Gi* varied from 800 to 5,000 $m^3\ d^{-1}$ in Williams and Mary Lakes, respectively. Modeled and observed δL were highly correlated in all lakes with the simple correlation coefficient (*r*) ranging from 0.84 to 0.98 in Williams and Island Lakes, respectively, (Figures 2 and 3 and Table 2). Modeled *Go* was 790 and 1,300 $m^3\ d^{-1}$ with correlation coefficients of 0.81 and 0.90 for Williams and Crystal Lakes, respectively (data not shown).

[40] The results of previous studies investigating groundwater flux in Shingobee and Williams Lakes compare favorable with the present study. *Gi* in Williams Lake

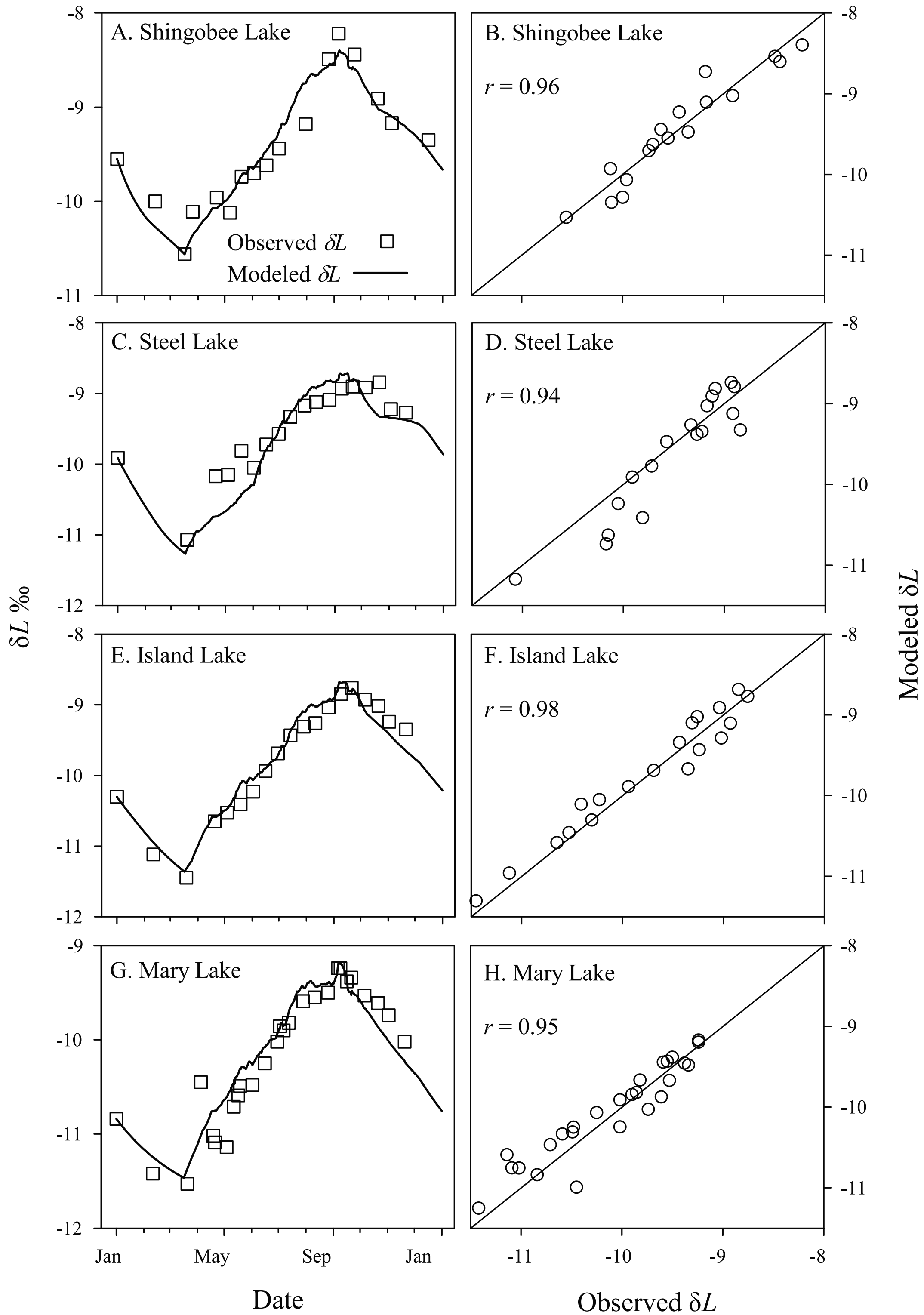

Figure 3. Comparison of modeled and observed lake surface $\delta^{18}O$ water isotopic signature in open-basin lakes (a, b) Shingobee, (c, d) Steel, (e, f) Island, and (g, h) Mary. Open squares in Figures 3a, 3c, 3e, and 3g show observed $\delta^{18}O$, and the solid line shows daily modeled $\delta^{18}O$. Observed $\delta^{18}O$ values appear on the *x* axis in Figures 3b, 3d, 3f, and 3h, and modeled $\delta^{18}O$ values from corresponding dates appear on the *y* axis. The simple correlation coefficient (r) between these values is also shown.

Table 2. Summary of Modeled Groundwater Inflow, Gi, in All Lakes as Determined by Manual Optimization[a]

Lake	Gi^a (m^3 d^{-1})	Minimum $\sigma_{GI}{}^b$	r^c
Crystal Lake	1,400 (1,200–1,700)	0.012	0.94
Williams Lake	800 (600–1,100)	0.018	0.84
Mary Lake	5,000 (4,500–5,700)	0.053	0.95
Island Lake	4,700 (4,300–5,200)	0.032	0.98
Steel Lake	2,900 (1,500–4,400)	0.082	0.94
Shingobee Lake	4,800 (4,400–6,900)	0.033	0.96

[a] Gi confidence intervals appear in parentheses and were calculated as the lowest 33rd percentile of calculated σ_{GI}, the variance between modeled and observed δL values, fit to a Gamma distribution (see section 3).

[b] Minimum σ_{GI} represents the σ_{GI} associated with the model run which provided the best fit between observed and modeled δL.

[c] The correlation coefficient, r, between observed and modeled δL for each lake.

has been found to be approximately 1,000 to 1,400 m^3 d^{-1} [*LaBaugh et al.*, 1997; *McConnaughey et al.*, 1994; *Stets et al.*, 2009], which is similar to our finding of 600 to 1,100 m^3 d^{-1} (Table 2). Gi has been studied much less in Shingobee Lake, but has been previously reported as 4,900 to 7,400 m^3 d^{-1} [*Rosenberry et al.*, 2000; *Striegl and Michmerhuizen*, 1998], whereas our model result was 4,400 to 6,900 m^3 d^{-1} (Table 2). However, groundwater fluxes have not been estimated previously for the rest of the lakes in this study (Crystal, Mary, Island, and Steel). A primary benefit of this study was that it allowed us to expand our hydrologic understanding beyond Shingobee and Williams Lakes to the rest of the watershed.

[41] The model developed for this study was flexible enough to apply to both open- and closed-basin lakes, but was constrained by the assumption that Gi was constant throughout the year. Other descriptive models have been developed with the goal of predicting long-term changes in δL as a result of climate change [*Hostetler and Benson*, 1994; *Shapley et al.*, 2008], and as a way of calculating E in arctic ponds [*Gibson et al.*, 1998]. More commonly, researchers solve isotope mass balance models for one component of the hydrologic budget time-averaged over a period of weeks or longer [*Gurrieri and Furniss*, 2004; *Sacks*, 2002]. This approach works well for short periods of time or in environments that can be assumed to reach steady state over the course of the time averaging. It is necessary to perform calculations in this way partly due to the uncertainty in the other measured components of the hydrologic budget [*LaBaugh et al.*, 1997]. In the approach used in the present study, all of the uncertainties are considered and then a best overall Gi is calculated. One obvious drawback is that our approach only provides a single value to Gi. This assumption is acceptable in the Shingobee River headwaters watershed [*Filby et al.*, 2002; *Reddy et al.*, 2006], but may not be applicable to areas in which groundwater discharge is more variable over short time periods.

[42] The model also performed well using data from calendar year 2002 with correlation coefficients between modeled and observed δL of 0.97 and 0.91 for Shingobee and Williams Lakes, respectively (see auxiliary material). Because this model was constructed with consideration given to all of the factors we believe influence δL in lakes, the ability of the model to perform using both 2002 and 2004 input data is not surprising. However, despite the high correlation coefficient, the absolute value of modeled δL did not correspond as closely to observed δL in Shingobee Lake, mostly giving higher (less negative) values late in the year. Problems modeling δL in Shingobee Lake may have arisen from interannual variation in Gi [*LaBaugh et al.*, 1995].

7.2. Surface Water Flows

[43] Developing hydrologic budgets for Steel and Island Lakes depended upon quantifying Q_{STLO} as outlined in section 5.1. Gi to Shingobee River between the outlet of Steel Lake and SRI was estimated to be 3,900 ± 500 m^3 d^{-1}, mean and standard error (Figure 4). Our result was reasonably close to the finding of *Jackman et al.* [1997] that Gi was approximately 3,100 m^3 d^{-1} in this stretch of the Shingobee River based on a tracer addition experiment. Although the model shows some seasonal variation in Gi along this portion of the Shingobee River, there was no corroborating evidence that the water table was substantially higher early in the year. The model is sensitive to errors in $\delta^{18}O$ signature, particularly when the difference between δGi and $\delta STLO$ is small, as is the case in winter and early spring. The short-term (i.e., monthly) fluctuations in Gi (Figure 4) may also have resulted from precipitation-generated displacement of groundwater which would re-

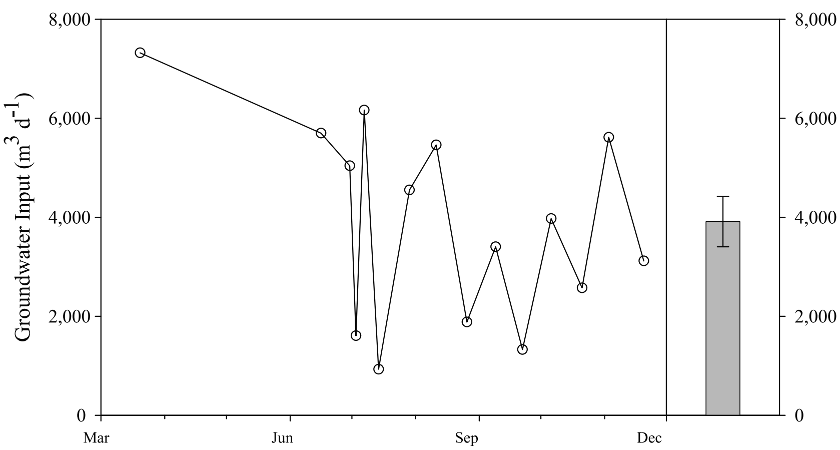

Figure 4. Calculated Gi between Steel Lake Outlet and the Shingobee River Inlet Flume, expressed as m^3 d^{-1}. Time series appears on the left, and the annual average and standard error appear on the right.

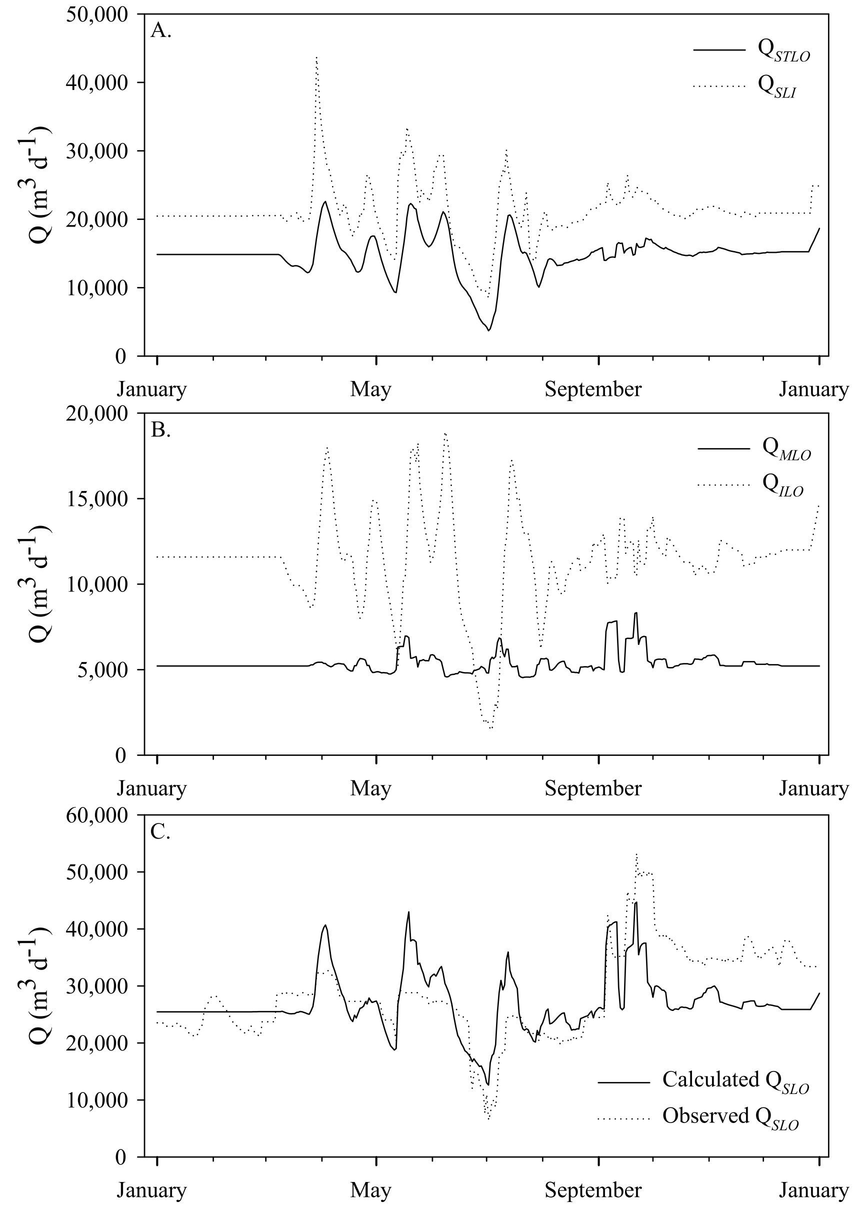

Figure 5. Calculated and observed stream discharge at several points in the Shingobee headwaters watershed, expressed as $m^3 d^{-1}$. (a) Modeled discharge at Steel Lake Outlet (Q_{STLO}) and measured stream input to Shingobee Lake (Q_{SLI}). (b) Modeled discharge at Mary Lake Outlet (Q_{MLO}) and modeled Island Lake Outlet discharge (Q_{ILO}). (c) Modeled and observed discharge out of Shingobee Lake (Q_{SLO}). Seven day moving averages are presented for all modeled discharge values.

sult in short-term storm-associated increases in *Gi* that had the isotopic signature of groundwater rather than precipitation water. Although this phenomenon could be short lived and deliver only small amounts of additional groundwater to the stream, our sampling schedule, which emphasized sampling during storm events, could have inflated the importance of this mechanism. Therefore, we believe that the annual average and standard error provide the best overall estimate of *Gi* for this portion of the Shingobee River.

[44] Q_{STLO} ranged from 3,700 to 22,600 $m^3 d^{-1}$ in 2004 with a median daily flow of 15,200 $m^3 d^{-1}$ (Figure 5a and Table 3). Q_{SLI} was obtained from stream discharge measurements and the median daily flow was 20,800 $m^3 d^{-1}$ (Figure 5a and Table 3).

Table 3. Modeled and Observed Surface Water Discharge at Various Points Along Shingobee River[a]

Section	Median Daily Flow	Minimum Daily Flow	Maximum Daily Flow
Modeled Q_{MLO}	5,100	4,400	7,500
Modeled Q_{ILO}	11,900	1,800	19,300
Modeled Q_{STLO}	15,200	3,700	22,600
Observed Q_{SLI}	20,800	8,500	43,700
Observed Q_{SLO}	27,300	6,400	53,300
Modeled Q_{SLO}	26,600	12,500	39,300

[a]The data presented include Mary Lake Outlet (Q_{MLO}), Island Lake Outlet (Q_{ILO}), Steel Lake Outlet (Q_{STLO}), Shingobee Lake Inlet (Q_{SLI}), and Shingobee Lake Outlet (Q_{SLO}). Observed values are obtained from stream gages present at the inflow and outflow of Shingobee Lake. Discharges are presented in $m^3 d^{-1}$.

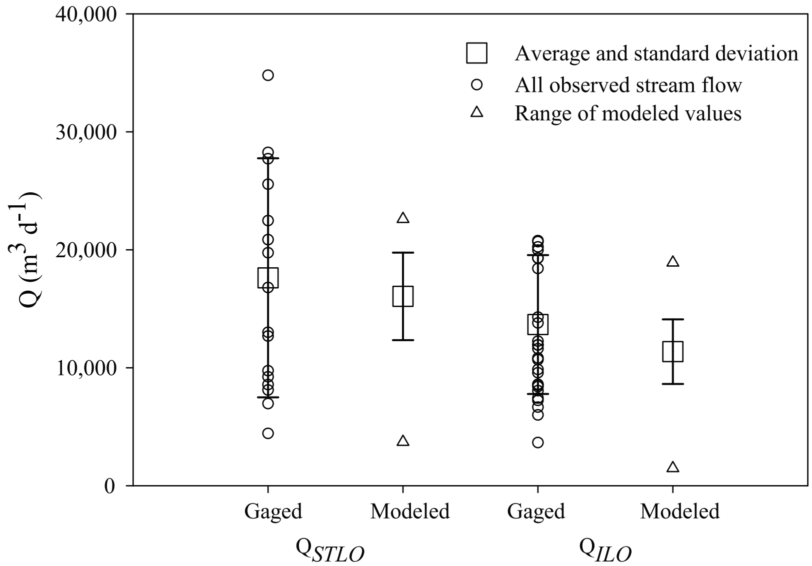

Figure 6. Comparison of modeled and gaged streamflow at Steel Lake Outlet (Q_{STLO}) and Island Lake Outlet (Q_{ILO}). Gaged data are from 1997, whereas the modeled data are from 2004. Boxes and whiskers show the average and standard deviation. For modeled values the circles depict the maximum and minimum 7 day moving average discharge. For measured values circles depict all individual measurements.

[45] Q_{ILO} was calculated by accounting for the hydrologic flows in Steel Lake and had a median discharge of 11,900 m^3 d^{-1} in 2004 with a range from 1,800 to 19,300 m^3 d^{-1} (Figure 5b and Table 3). Q_{MLO}, calculated from the hydrologic budget of Mary Lake, ranged from 4,400 to 7,500 m^3 d^{-1} (Figure 5b and Table 3). The small range of values in Q_{MLO} reflects the fact that hydrologic flow through Mary Lake is dominated by spring and groundwater discharges which are relatively constant throughout the year.

[46] Q_{SLO} was measured at a permanent stream gage at the outflow of Shingobee Lake and we were also able to model Q_{SLO} based on our hydrologic budget for Shingobee Lake. Median daily measured Q_{SLO} was 27,300 m^3 d^{-1} compared with our modeled value of 25,900 × 10^6 m^3 d^{-1} (Figure 5c

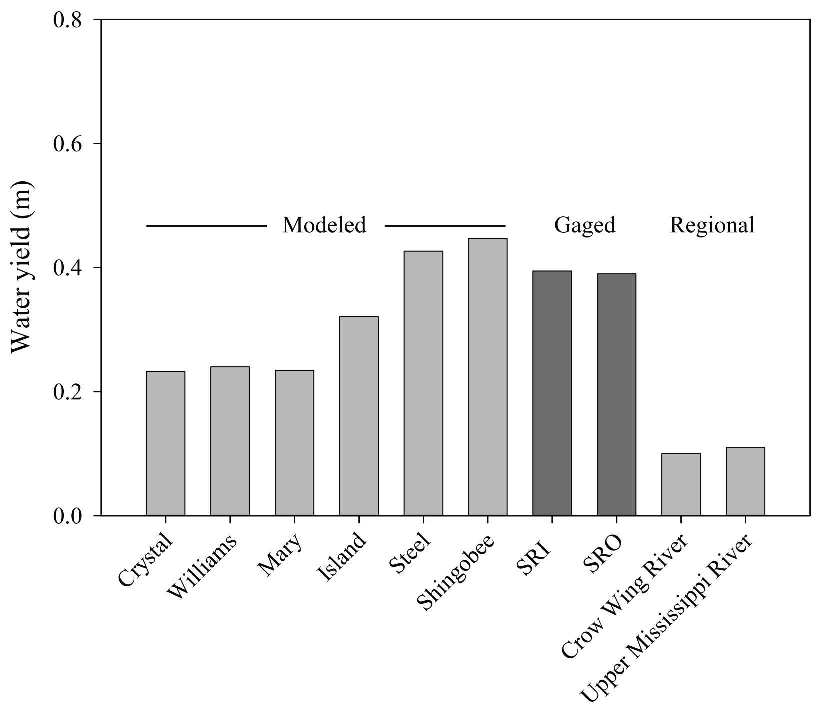

Figure 7. Water yield from model output in the Shingobee headwaters watershed, stream gages in the Shingobee headwaters watershed (SRI and SRO), and regional data from the U.S. Geological Survey National Water Information System (http://waterdata.usgs.gov/nwis).

Table 4. Complete Water Budgets for All Lakes Included in This Study[a]

Lake	V	*Area*	*Si*	*Gi*	*Pr*	*So*	*Go*	E	ΣI	ΣO	$\Sigma I - \Sigma O$	τ
Crystal	2.48	0.77	NA	0.51	0.47	NA	0.47	0.43	0.99	0.90	0.09	2.50
Williams	2.04	0.40	NA	0.30	0.25	NA	0.29	0.22	0.54	0.51	0.03	3.75
Mary	0.53	0.14	NA	1.86	0.11	1.90	NA	0.08	1.97	1.98	−0.01	0.27
Island	1.25	0.32	1.90[b]	1.72	0.19	4.31	NA	0.14	3.81	4.45	−0.63	0.33
Steel	1.90	0.25	4.31	1.07	0.30	5.43	NA	0.14	5.69	5.56	0.12	0.33
Shingobee	4.02	0.65	7.77	1.75	0.44	9.61	NA	0.34	9.96	9.95	0.00	0.40

[a]All water fluxes are in 10^6 m^3 yr^{-1}; lake volume, V, is in 10^6 m^3; lake area, *Area*, is in km^2; and residence time, τ, is in years. ΣI and ΣO are total water fluxes in and out of each lake, respectively. NA means not applicable.
[b]From Mary Lake Outlet only.

and Table 3). A comparison of the seasonal variability in Q_{SLO} shows that the calculated and measured values were similar from January to late September, and diverged from October to December (Figure 5c). This result is consistent with field observations that beaver dams present in autumn of 2004 downstream from SLO maintained elevated lake stage in Shingobee Lake and led to artificially high estimates of Q_{SLO} based on the stage-discharge relation.

[47] Streamflow was measured at the outlet of Steel Lake and Island Lake 16 and 26 times, respectively, since 1997 providing an opportunity to check the accuracy of our model-based streamflow calculations. Modeled Q_{STLO} was 16,000 ± 3,700 m^3 d^{-1} (average and standard deviation, n = 366, Figure 6) in 2004, which compares favorably with measured streamflow values of 17,600 ± 10,100 m^3 d^{-1} (average and standard deviation, n = 16, Figure 6). The modeled and measured values at Q_{ILO} were also similar, 11,800 ± 2,800 (n = 366) and 13,700 ± 5,900 m^3 d^{-1} (n = 26), respectively (Figure 6). There are two likely reasons that the standard deviation of the measured values was greater than the modeled values. First, we present modeled river flow values as 7 day moving averages, which tends to dampen extreme flows. Second, flow measurements were often conducted with the goal of adequately describing the range of streamflows in this watershed and so emphasized measurement of extreme values.

[48] Water yield, calculated by dividing water discharge (*So* or *Go*) by the drainage areas based on surface topography, varied from 0.23 to 0.45 m and generally increased from upstream to downstream in the watershed (Figure 7). Yields calculated from model output compared favorably with field-based observations of water yield at two points in the watershed, but were much higher than water yield measured regionally in the Crow Wing and Upper Mississippi River watersheds, approximately 0.10 m (Figure 7). The coherence between measured and modeled water yields in the Shingobee River watershed suggests that these data are correct and that there are real discrepancies between local and regional water yields based on surface water drainage basins.

[49] Elevated water yield from the Shingobee River watershed is most likely the result of regional groundwater flow into this area. The Shingobee River headwaters watershed is situated in an area with high permeability and low to moderate topographic relief and as such may be susceptible to large water inputs from the regional groundwater system [*Winter et al.*, 2003]. Regional groundwater flux was also implicated in controlling some of the chemical differences between Shingobee and Williams Lakes [*Dean et al.*, 2003]. For the present study, water yield was calculated in the Shingobee River headwaters watershed using the watershed delineated from ostensible surface topography. If groundwater enters the Shingobee River headwaters watershed from a significantly larger area than the one used to calculate yield, then the yield will be artificially high. The discrepancy between local and regional water yield suggests that groundwater discharging in the Shingobee River headwaters watershed may originate from an area several times larger than the surface water watershed (Figures 1 and 7).

7.3. Complete Water Budgets

[50] We calculated complete water budgets for the six lakes included in this study by compiling *Pr* and E with model-based estimates of groundwater and surface water flux (Table 4). Water residence time, calculated as total water input divided by lake volume, ranged from 0.26 to 3.75 years in Mary and Williams Lakes, respectively (Table 4). With the exception of Island Lake, the water inputs and outflow were balanced to within 10% (Table 4).

[51] Island Lake receives water from groundwater discharge, the outlet of Mary Lake, and the outlet of Spring Lake (Figure 1). Q_{MLO} and *Gi* were quantified, and we expected that the hydrologic imbalance, −0.63 × 10^6 m^3 yr^{-1} would be approximately equivalent to the surface inflow from Spring Lake. Measured streamflow into Island Lake from Spring

Table 5. Sensitivity Analysis for Shingobee and Williams Lakes[a]

Parameter	Scenario	*Gi*
	Shingobee Lake	
Best estimate	-	1.75 (1.62–2.52)
E	+15%	2.50 (2.20–2.80)
	−15%	1.04 (0.78–1.31)
δGi	= −10.5‰	3.36 (3.03–3.67)
	= −12.5‰	1.14 (0.88–1.43)
Si	+5%	1.71 (1.54–1.87)
	−5%	1.81 (1.64–1.97)
L	no stratification	2.27 (1.96–2.58)
	summer mixed layer 2 m	2.32 (1.94–2.70)
	Williams Lake	
Best estimate	-	0.30 (0.21–0.40)
E	+15%	0.38 (0.23–0.54)
	−15%	0.22 (0.04–0.41)
δGi	= −8.1‰	0.36 (0.20–0.53)
	= −10.1‰	0.26 (0.09–0.44)
L	no stratification	0.28 (0.12–0.45)
	summer mixed layer 2 m	0.31 (0.16–0.48)

[a]The parameters evaporation (E), groundwater stable isotope composition (δGi), surface water inflow (*Si*), and lake epilimnetic volume (L) were varied, and the model was rerun to obtain a new value of groundwater inflow (*Gi*) and a corresponding level of certainty (displayed in parentheses). We also present our best *Gi* estimate for the purpose of comparison.

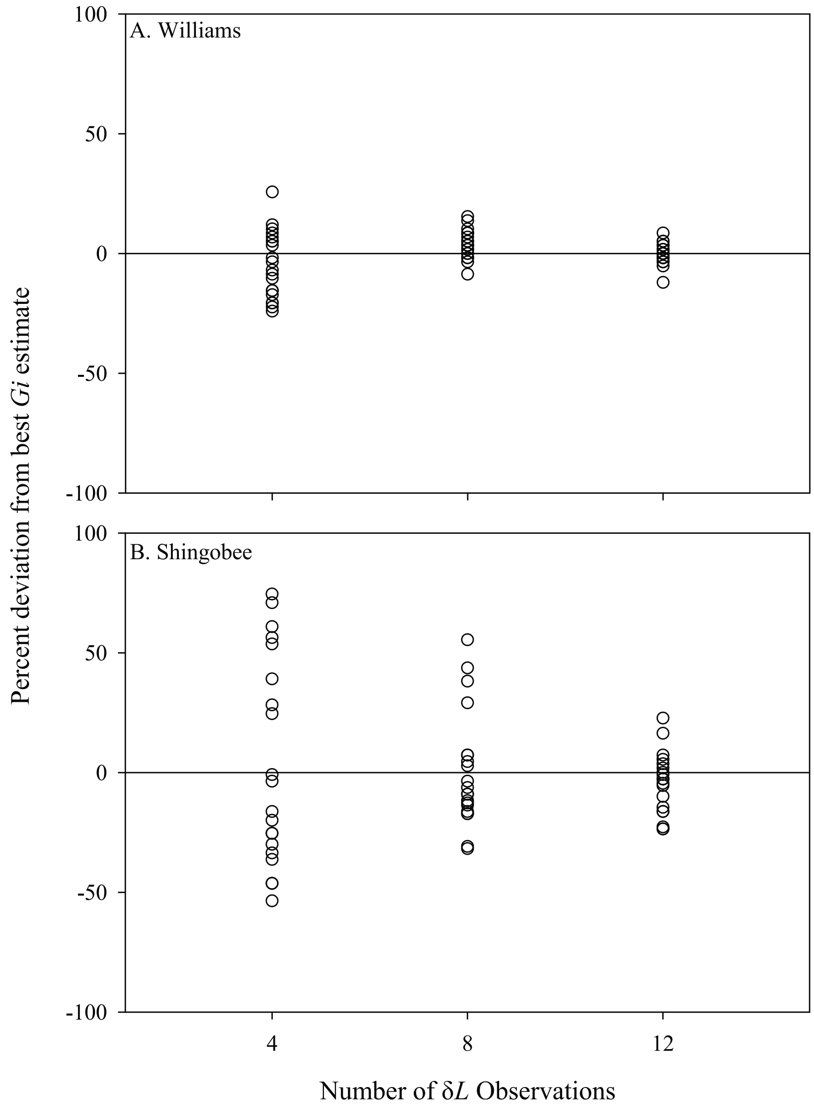

Figure 8. Demonstration of how estimated Gi responds to changes in the number of δL observations used. The number of randomly selected δL_{OBS} appears on the x axis, and the percent deviation from the best Gi estimate, obtained using all available data, appears on the y axis. The results of each of 20 runs are depicted by open circles. Results are shown for (a) Williams Lake and (b) Shingobee Lake.

Lake was $0.71 \pm 0.37 \times 10^6$ m^3 yr^{-1} (average and standard deviation, $n = 4$), which is close to the hydrologic imbalance in Island Lake and suggests that the model provided a reasonable estimate of the total water input to Island Lake.

7.4. Sensitivity Analysis

[52] Modeled Gi in Williams Lake is less sensitive than in Shingobee Lake (Table 5). For example, a 15% increase or decrease in E leads to a 26% change in modeled Gi in Williams Lake versus a 40–43% change in modeled Gi in Shingobee Lake (Table 5). A one per mil change in δGi resulted in a 35–92% change in modeled Gi in Shingobee Lake versus a 13–20% change in modeled Gi in Williams Lake (Table 5).

[53] The differing sensitivity of the model in Shingobee and Williams Lake is a result of the difference in the magnitude of hydrologic flux in the lakes. The hydrologic residence time of Williams Lake is 3.75 years (Table 4), meaning that δL is controlled primarily by the long-term (>1 year) balance between groundwater input and evaporative modification rather than small annual differences in hydrologic flux. By comparison, hydrologic inputs replace the entire volume of Shingobee Lake approximately 2.5 times every year (Table 4), so short-term alteration of any of the components has a large effect on δL and, consequently, modeled Gi.

8. Model Robustness and Applicability to Other Lake Stable Isotope Studies

[54] This study relied upon numbers of $\delta^{18}O$ water isotope samples beyond feasibility for most lake studies. But how many samples are needed to use this model effectively to obtain estimates of Gi? We investigated this question as it relates to the number of lake surface water samples by ran-

domly selecting a specified number of δL_{OBS} data points in Shingobee and Williams Lakes to see how the estimate of *Gi* responded to changes in the number of δL_{OBS} samples used. We randomly selected δL_{OBS} data for 4, 8, and 12 dates for Shingobee and Williams Lakes and observed the distribution of *Gi* estimates over 20 simulations.

[55] In Williams Lake, a closed-basin lake with a hydrologic residence time of 3.75 years, using as few as 4 samples gave a reasonable estimate of *Gi*, with all of the simulations lying within 25% of our best estimate (Figure 8a). Using 8 samples improved the precision to approximately ±10% of the best estimate (Figure 8a). Little additional benefit was gained by using 12 samples (Figure 8a). In contrast, 12 randomly selected δL_{OBS} values were required in Shingobee Lake to ensure that estimated *Gi* was within 25% of the best *Gi* estimate (Figure 8b). Using fewer δL_{OBS} values resulted in a very wide distribution of *Gi* estimates (Figure 8b). The need to sample Shingobee Lake more often in order to use this model effectively is a result of the large hydrologic fluxes in this lake and the large variation in δ^{18}O values occurring throughout the year (see Figures 3a and 3b). However, we were able to obtain a *Gi* estimate within 4% of the best estimate by carefully selecting 7 sample dates. The dates were selected to bracket periods of relatively constant change in δL. They were: 1 January (early winter), 17 March (late winter), 20 May (early summer stratification), 1 July (midsummer), 26 August (late summer), 24 September (early autumn), and 5 November (late autumn). Using these seven dates produced a *Gi* estimate of 4,618 m^3 d^{-1}, which is very close to the best estimate of 4,799 m^3 d^{-1}.

9. Conclusions

[56] Annual variation in δL_{OBS} was predictable in these lakes and allowed quantification of *Gi* (Table 2 and Figures 2 and 3). We were also able to calculate surface water flows from the hydrologic budgets of these lakes (Table 3 and Figure 5). The model depended upon accurate measurements of *E* and δGi (Table 5), but the coherence between modeled water flows and field-based observations (Figures 6 and 7) suggested that the model provided a reasonable estimate of the hydrologic budgets for these lakes. Model results for Williams Lake, a closed-basin lake with long hydrologic residence time, was less sensitive and required fewer δL_{OBS} values than Shingobee Lake, an open-basin lake with large annual variation in hydrologic flux and lake surface δ^{18}O. So while model results in closed-basin lakes were more robust and less sensitive to changes in *E* and δGi, the model performed well in open-basin lakes providing that an adequate number of lake surface water samples were collected.

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