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Hydroclimatic controls on water balance and water level variability in Great Slave Lake

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Abstract:

On average, 86% of riverine discharge to Great Slave Lake, Northwest Territories, Canada, was gauged during the period 1964–1998, offering an unprecedented opportunity to study and understand controls on water balance of a large northern lake at the headwaters of the Mackenzie River. A functional daily water balance model, incorporating measurements of riverine inflow, precipitation on the lake surface, evaporation, and riverine outflow was developed, which predicts the amplitude and frequency of annual water level fluctuations, and closes the water balance to within ±6% for 28 of 35 years and ±11% for the remaining 7 years, with an overall systematic error of +2%. Annual water balance estimates for the period 1964–1998 reveal that about 74% of inflow into Great Slave Lake originates from the Peace-Athabasca catchments that enter the lake via the Slave River, whereas 21% is derived from other catchments bordering Great Slave Lake, and 5% from precipitation on the lake surface. An estimated 94% of water losses occur by riverine outflow to the Mackenzie River and 6% by evaporation from the lake surface. The primary driving force behind water level fluctuations in Great Slave Lake, including the post-regulation period following development of the W.A.C. Bennett Dam, is shown to be climate-driven precipitation variability in the Peace-Athabasca basins. A simple precipitation regression model is developed to simulate water level fluctuations in Great Slave Lake over the past 100 years. Copyright © 2006 John Wiley & Sons, Ltd.

KEY WORDS water budget; climate impacts; Great Lakes; evaporation; Arctic; northern hydrology

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INTRODUCTION

The Mackenzie River is distinct among the north-flowing rivers into the Arctic Ocean owing to the presence of several large lakes, Lake Athabasca (7900 km^2), Great Slave Lake (28 568 km^2), and Great Bear Lake (31 328 km^2), which serve to regulate flow, sedimentation, and biogeochemistry along the main stem of the drainage network. These lakes, formed by glacial erosion during the Pleistocene (Hutchinson, 1957), occur along or close to the contact between the western margin of the Precambrian Shield and the adjacent Paleozoic Interior Plains sediment platform to the west. Great Slave Lake (GSL) is among the deepest freshwater lakes in the world (614 m), but is hydrologically dynamic (mean residence time of 16 years; Evans, 2000) because of abundant inflow from a catchment area totalling 949 000 km^2, or 53% of the entire area of the Mackenzie Basin. Roughly three-fourth of the riverine inflow to the lake enters via the Slave R., delivering runoff from the Peace-Athabasca Basins (606 000 km^2) with headwaters in relatively high precipitation areas to the southwest, including the Rocky Mountains (Figure 1a). The remainder of riverine inflows enters directly from numerous Shield catchments situated around the northeastern margin of the lake, and to a lesser extent from wetland-dominated Interior plains watersheds situated around the southeast margin of the lake (Figure 1b). These areas are characterized by gently rolling plateaus and lowlands, with thick glacial drift, sedimentary

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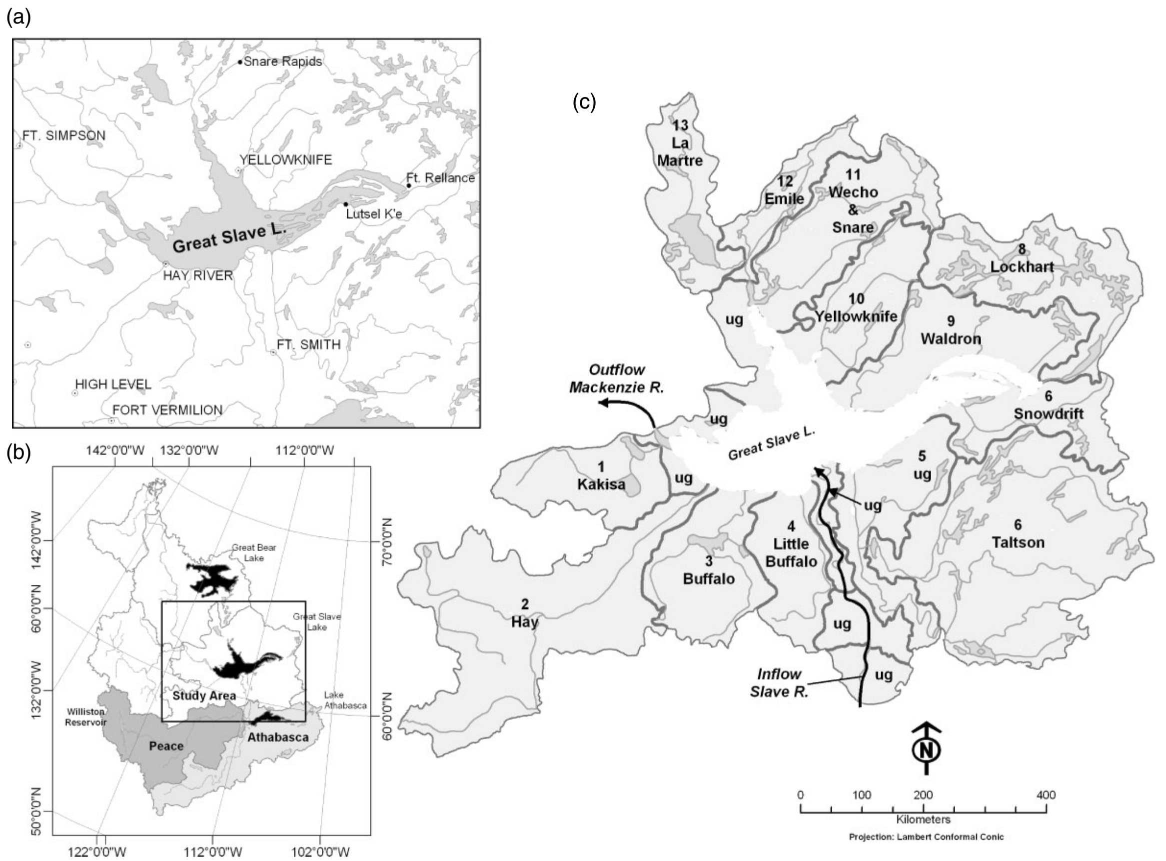

Figure 1. Maps of (a) the Mackenzie Basin showing major sub-basins and large lakes, (b) Great Slave Lake and water level and precipitation monitoring sites, (c) catchments tributary to Great Slave Lake. Note that catchment numbers refer to Table I; ug is ungauged

bedrock, and extensive areas of lakes, muskeg, and swamp, in contrast to a thin, discontinuous cover of glacial drift overlying crystalline bedrock, which creates disorganized drainage and myriad lakes within Precambrian Shield areas (Louie *et al*., 2002).

The most significant flow regulation structure within the GSL drainage is the Bennett Dam, which impounds an estimated $4{\cdot}1 \times 10^{10}$ m^3 of water (Muzik, 1985) in the Williston Reservoir on the Peace River. Despite its considerable distance from Great Slave Lake, filling of the dam during 1968–1971 resulted in diversion of 1·2 times the annual flow from the river (Peters and Prowse, 2001). Regulation of the river also has shifted the pattern of seasonal flows and dampened flow extremes, creating a less variable annual flow regime (Prowse *et al*., 2002a). Such changes may also affect ice conditions, flooding, fluvial geomorphology, and riparian vegetation in the lower Peace River, Peace-Athabasca Delta, and Slave River Delta (Prowse and Conly, 2002b).

While numerous process and modelling studies on hydrology have been undertaken within the Mackenzie Basin, most recently as part of the Canadian Global Energy and Water Cycle Experiment (Stewart *et al*., 1998; Rouse, 2000; Rouse *et al*., 2003a), few studies have focused on the water cycle processes of the large high-latitude lakes. Notable exceptions include Kerr (1997) who summarized the 1973–1992 monthly water balances for Great Slave and Great Bear Lakes, Blanken *et al*. (2000) who provided the first direct measurements of evaporation from a high-latitude northern lake (GSL) using eddy correlation methods, and Menard *et al*. (2002) who modelled GSL ice phenology. Other studies have provided baseline information on related issues such as the effect of ice on the hydraulics of the Mackenzie River at the outlet of GSL (Hicks

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et al., 1995) and preliminary assessment of the potential impacts of climate change on runoff in sub-basins of the Mackenzie (Soulis *et al.*, 1994). One of the major limitations of the study by Kerr (1997) was that the lack of evaporation measurements for high-latitude lakes required that evaporation be incorporated into the residual error, to a large extent preventing rigorous assessment of sources of uncertainty and variability in the water balance. The role of large lakes, a primary unknown in the hydroclimatology of the Mackenzie Basin identified by Lawford (1994), has remained one of the primary objectives of future GEWEX process and modelling studies within the basin (Rouse, 2000). Fortunately, basic monitoring information, such as water level data, are available at four stations on Great Slave Lake from 1938 to the present. The mean water level of Great Slave Lake over the period 1938–99 was $156{\cdot}60 \pm 0{\cdot}22$ (1std) masl, but with considerable variability both annually and interannually. Record high water levels of 157·23 m were observed in August 1997, while record low water levels of 156·03 m were observed in October 1981, giving an historical range of about 1·2 m. High water levels resulted mainly from a management decision to release water from the W.A.C. Bennett Dam because it appeared at the time that the berm was weakening (M.English, pers. comm.) (Note that wind seiche events can produce similar water level changes over short time-periods). Typical annual fluctuations are on the order of 0·4m. It is also significant to note that an average 86% of riverine inflow from the catchment area of GSL has also been gauged by the Water Survey of Canada (WSC) during 1964–98, and similar information is available to estimate outflows. Precipitation monitoring at several stations has also been conducted around the periphery of the lake over the same period.

This paper presents an historical water balance analysis for Great Slave Lake for the period 1964–98, with the aim of examining potential sensitivity of the system to climate and water resources impacts over the past 35 years. One primary motivation for the study is to assess the potential impacts of changing lake levels on the structure and function of the Slave River delta, including depositional energy distribution and location of sedimentation. This work extends the water balance analysis of Kerr (1997) by 15 years, including the period of initial filling of the Williston reservoir, and includes two additional climate-driven drying cycles. In addition, this analysis benefits from knowledge gained through the recent studies on ice phenology and evaporation process by Menard *et al.* (2002) and Blanken *et al.* (2000), respectively, to constrain the evaporation component of the water cycle; which as we show is important for characterizing variability in the terms of water balance and for examining the underlying causes of water balance and water level variability of GSL. A functional daily water balance model of the lake is developed, which is capable of predicting the amplitude and frequency of annual water level fluctuations in Great Slave Lake. Overall, this paper describes the basic characteristics of the water balance of GSL such as throughflow and water level relationships, identifies the principle sources of uncertainty in the water balance, and demonstrates the predominant climatic control of precipitation in the Peace-Athabasca basin on the observed wetting and drying cycles that the lake has undergone over a 35-year period. Knowledge of the basic water balance characteristics of Great Slave Lake also sheds light on the role of large lakes in controlling Mackenzie Basin runoff and highlights the sensitivity of northern Great Lakes to hydroclimatic processes in more southerly areas.

METHODOLOGY AND DATASETS

The basic strategy employed to compute the water balance relies on estimation of total inflow as the sum of riverine inflows and precipitation directly on the lake surface, and estimation of total outflow as the sum of riverine discharge and evaporation from the lake surface, using water level records to characterize storage changes. The potential role of groundwater is also discussed, although direct estimates of subsurface exchange are unavailable. The annual water balance was computed for calendar years (1 January–31 December) on the basis of the relationship

$$I + P - Q - E = \mathrm{d}S/\mathrm{d}t \pm G \pm error \tag{1}$$

where I is the mean annual riverine inflow to the lake (m^3 s^{-1}), P is the mean annual precipitation on the lake surface (m^3 s^{-1}), Q is the mean annual riverine outflow (m^3 s^{-1}), E is the mean annual evaporation

(m^3 s^{-1}), dS/dt is the mean change in lake storage (m^3 s^{-1}), and the residual $\pm G \pm error$ includes error in all measurements including the net groundwater exchange (m^3 s^{-1}). A working model was also developed to test the model's robustness for predicting daily water balance and water level changes. Data used for the computations are described in the subsequent text.

On-lake precipitation

Precipitation data from Environment Canada weather stations in the vicinity of Great Slave Lake were used to estimate the precipitation falling directly on the lake surface. The lake surface area (28 568 km^2) was divided into sub-regions using a Theissen polygon approach (see Dingman, 2002, p. 121) so that precipitation of the whole lake was weighted according to the fraction of sub-basin situated closest to each individual precipitation station. This approach was considered adequate owing to small differences in relief between precipitation stations and the lake, and the small influence of precipitation in the overall water balance. Several combinations were assessed to test the sensitivity of the results by the elimination of one, two, or three stations (Table I) Rouse *et al*. (2003b).

The results shown in Table I indicate that errors associated with the choice of precipitation stations are expected to be less than about 5%, even if up to three recording stations are eliminated. As precipitation accounts for only 5% of total inflows to the lake, errors in the water balance due to the choice of Theissen polygons is evidently negligible (<1%). More significant are the potential errors due to systematic gauge undercatch related to error in snow accumulation measurements, and those due to wetting losses, trace precipitation events, and wind effects (Woo *et al*., 1983; Metcalfe and Goodison, 1993). Metcalfe *et al*. (1999) provide an adjusted archive to account for these effects at 16 stations in the Northwest Territories. These adjusted data suggest a mean annual precipitation of approximately 340 mm at Yellowknife for the period 1968–1996, which is 25% larger than values reported in the standard climate archive. Although the data of Metcalfe *et al*. (1999) likely provide a more accurate view of precipitation accumulation, it is arguable whether such revisions truly reflect available water input unless sublimation losses are taken into account. Sublimation losses during winter can be significant, perhaps 20% of snowfall under optimal blowing snow conditions (Pomeroy *et al*., 1997). The use of unadjusted precipitation in the current analysis, where stations typically receive 50% of annual precipitation as snowfall, is justified as it partially accounts for the combined effect of precipitation undercatch (+25%) and sublimation losses (−10%). As the on-lake precipitation is only a small contribution to the total inflow into GSL, the resulting errors in the water balance are expected to be on the order of 1–1·5%, which is adequate for the purposes of this analysis. No effort was therefore made to account for special factors such as lake-effect, snowfall, and rainfall.

Table I. Mean Annual Precipitation (mm) on Lake evaluated by several Theissen Polygon Combinations. Note that Yellowknife and Hay River were considered critical stations and were used in all combinations. Note that Column 7, the scenario yielding an estimate closest to the calculated mean value, was used in the analysis

Precipitation Station	Annual Precipitation 1964–1998	Theissen Polygon Combination						
		1	2	3	4	5	6	7
Yellowknife A	269·7	0·17	0·19	0·19	0·55	0·19	0·55	0·17
Hay River A	322·0	0·25	0·25	0·23	0·26	0·25	0·28	0·23
Fort Reliance	270·7	0·14	0·14	0·14	0·16	—	0·16	0·14
Fort Providence A	303·8	—	—	0·02	0·02	—	—	0·02
Fort Resolution A	313·7	0·41	0·41	0·41	—	0·55	—	0·41
Snare Rapids	241·7	0·02	—	—	—	—	—	0·02
Predicted precipitation on Lake	—	297·7	298·3	297·9	281·4	304·3	281·8	297·3
% Error (relative to mean = 294·1 mm)	—	1·2	1·4	1·3	−4·3	3·4	4·2	1·1

The following precipitation station weightings were calculated from the Theissen polygon analysis and used to estimate the precipitation over the lake: Yellowknife (17%), Hay River (23%), Fort Reliance (14·2%), Fort Providence (2%), Fort Resolution (41%), Snare Rapids (2%). For running the operational daily water balance model, snow and rain were accounted for separately to allow snow accumulation to occur on lake ice from the time of freeze-up until 1 May. Snow was then melted and added to the lake storage over a period of 15 days, which was intended to roughly simulate the spring melt cycle.

Lake evaporation. Lake evaporation for the entire ice-free periods of 1997 and 1998, the last two years of the water balance period used in this analysis, was taken directly from eddy covariance measurements by Blanken *et al*. (2000). They showed that evaporation of 386 ± 127 mm measured for 1997 was significantly lower than the evaporation of 486 ± 144 mm measured for 1998, (the latter also including 7 days measured during 1 January–7 January 1999) due to a protracted ice-free period that lasted until 12 December 1997 and 8 January 1999, respectively. Blanken *et al*. (2000) also demonstrated that cumulative evaporation was similar in both years between mid-August and mid-November, and that higher totals for 1998–1999 were mainly due to a protracted ice-free season associated with the 1997–1998 El Nino. Rouse *et al*. (2003b) concluded that the date of ice break-up in June exerts the main control on the final freeze-up date and thereby the annual evaporation total because heat storage during this high-sun season drives latent and sensible heat fluxes into the late fall and early winter.

Since evaporation estimates were not available for 1964–1996, an evaporation algorithm was developed to account for variations in annual evaporation that would have occurred owing to variations in the ice-free period, assuming similar monthly evaporation rates in each year, as noted by Blanken *et al*. (2000). For the calculations, ice-free period was characterized from historical records of freeze-up and break-up dates that were measured on Back Bay near Yellowknife (Lenormand *et al*., 2002) and adjusted to match the systematic offset noted between the Back Bay record and whole lake estimates predicted using SSM/I 85 GHz passive microwave imagery (Figure 11 in Menard *et al*., 2002). The algorithm integrates an evaporation function based on monthly values estimated by Blanken *et al*. (2000), combined with knowledge of the shifting ice-free period, to estimate the annual evaporation (Figure 2(a)). The reconstructed estimates of evaporation for the 1964–96 period, which are predictably correlated with the ice-free period (Figure 2(b)), are found to range from 275 to 410 mm yr^{-1} with a mean of 336 mm yr^{-1}. E/P ratios are also within a reasonable range of 0·96–1·86 with a mean of 1·5, and broadly consistent with available annual estimates computed from the Hydrological Atlas of Canada (see Figure 1·10 in Gibson, 1996; see also plates 3 and 17 in denHartog and Ferguson, 1978). Evaporation measured in 1999 by Rouse *et al*. (2003b), a follow-on to the study of Blanken *et al*. (2000), is also shown for comparison.

Riverine inflows and lake levels

Riverine inflows were taken from HYDAT (Environment Canada, 2001) for WSC gauging stations on a daily basis. On average, 86% of the contributing area was gauged during 1964–1998, which is considerably better than for the lower Great Lakes of North America (∼35%). Estimates of flow for an additional 9% of the contributing areas/years were reconstructed with partial records, where available, through comparison with basin hydrographs with similar observed hydrologic responses. Approximately 5% of the contributing area had no representative gauging records, and these areas were assigned values from basins with similar morphology. A summary of riverine inflows (Table II) emphasizes the dominance of the Slave River inflows derived from the Peace-Athabasca basins, both in terms of volumetric inflows and higher mm yr^{-1} runoff from these more southerly basins with headwaters in the Rocky Mountains. Average runoff from Shield basins (i.e. Taltson, Lockhart, Waldron, Yellowknife, Emile, Wecho and Snare) ranges from about 51 to 157 mm yr^{-1} and tends to be higher, in general, than runoff from basins situated on interior plains areas (i.e. Kakisa, Hay, Buffalo, Little Buffalo, La Martre), which ranges from 27 to 83 mm yr^{-1}. Runoff variability among the sub-basins reflects the basin gradient and roughness, fraction of lake cover, evaporation-transpiration losses, and degree

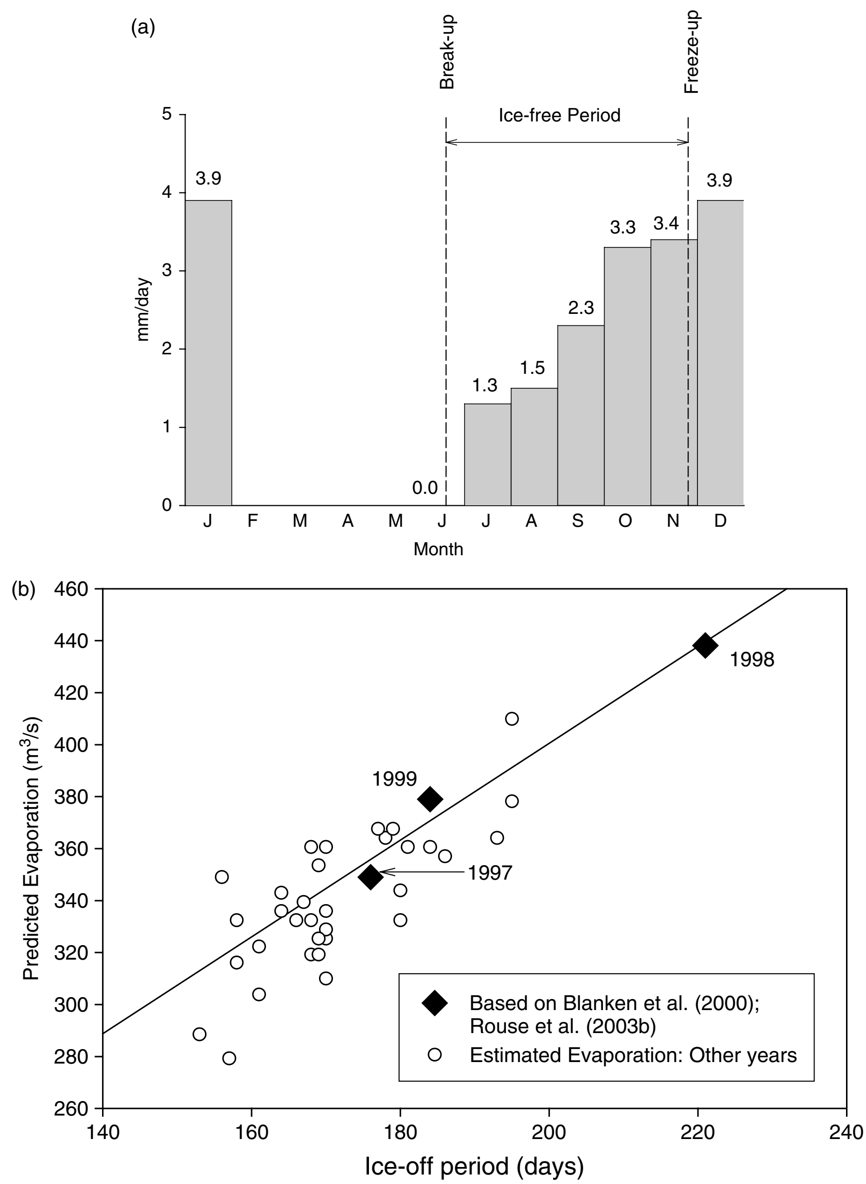

Figure 2. (a) Histogram of monthly evaporation values used to compute annual evaporation based on the length of the ice-free period for Great Slave Lake, (b) Plot of predicted mean annual evaporation rates versus ice-off period used in this analysis compared with values of Blanken *et al*. (2000) and Rouse *et al*. (2003b) for 1997, 1998 (to Dec. 31), and 1999

of connectivity of lake drainage patterns. The seasonal timing of runoff (Figure 3) is characterized in all cases by high-flow during spring freshet in mid-April. Basins with string-of-lakes drainage (e.g. Yellowknife River basin) may exhibit a delay in peak discharge owing to natural regulatory effect of abundant surface storage. Outflow from GSL was estimated from the gauging records for Mackenzie River at Strong Point, subtracting inflows from upstream tributaries, primarily the Trout River.

Average daily water levels for GSL were compiled by averaging WSC records from Yellowknife, Fort Resolution, Hay River, and Lutsel k'e. This provided a smoother record of water level changes than using

Table II. Annual water balance summary, 1964–98

Year	Break-up Date	Freeze-up Date	Ice-free Season days	Inflow $m^3\ s^{-1}$	Outflow $m^3\ s^{-1}$	Evaporation $m^3\ s^{-1}$	Precipitation $m^3\ s^{-1}$	Total Error $m^3\ s^{-1}$	Lake storage change $m^3\ s^{-1}$	Wl change m	Avg Water level m	E/P $m^3\ m^{-3}$	E/I $m^3\ m^{-3}$	Slave R. Inflow Fraction $m^3\ m^{-3}$	Areal Fraction Gauged
1964	6/22	12/8	169	5280	5430	354	193	307	−5	−0·005	156·81	1·81	0·06	0·71	0·84
1965	6/13	11/28	168	4932	5039	319	218	158	−70	−0·077	156·77	1·45	0·06	0·77	0·84
1966	6/23	12/10	170	4804	4821	361	242	178	34	0·037	156·69	1·47	0·07	0·73	0·84
1967	6/25	12/10	168	4641	4768	361	222	158	−93	−0·103	156·69	1·61	0·07	0·71	0·84
1968	6/16	12/11	178	3557	3751	364	231	117	−191	−0·211	156·57	1·56	0·10	0·63	0·84
1969	6/17	11/20	156	3564	3512	349	191	86	−30	−0·033	156·43	1·52	0·08	0·64	0·83
1970	6/8	12/5	180	3356	3532	344	194	312	24	0·026	156·36	1·75	0·09	0·71	0·86
1971	6/18	11/18	153	3675	3598	288	188	48	−10	−0·012	156·51	1·52	0·08	0·76	0·86
1972	6/6	12/9	186	4336	3779	357	222	−230	189	0·209	156·60	1·59	0·09	0·76	0·86
1973	6/11	11/15	157	4876	4584	279	292	−288	70	0·077	156·68	0·94	0·06	0·72	0·86
1974	6/5	12/2	180	5626	4765	332	301	−602	187	0·206	156·79	1·09	0·07	0·70	0·86
1975	6/1	12/11	193	4785	4592	364	217	−177	−144	−0·159	156·75	1·66	0·08	0·57	0·86
1976	6/9	12/10	184	4925	4824	361	208	30	−12	−0·014	156·74	1·71	0·07	0·74	0·89
1977	6/13	11/30	170	4851	4413	325	186	−247	47	0·052	156·71	1·73	0·07	0·75	0·87
1978	6/24	12/5	164	4139	3889	343	210	−252	−141	−0·156	156·64	1·62	0·09	0·77	0·87
1979	6/3	12/15	195	4412	4345	378	201	49	−48	−0·053	156·69	1·86	0·08	0·77	0·88
1980	6/8	11/25	170	3190	3329	310	209	145	−106	−0·117	156·43	1·47	0·09	0·83	0·88
1981	6/22	12/3	164	3354	3378	336	189	131	−40	−0·044	156·43	1·76	0·10	0·73	0·88
1982	6/27	12/2	158	3860	3894	332	205	328	168	0·186	156·49	1·60	0·08	0·68	0·88
1983	6/12	11/28	169	4003	4308	319	231	452	49	0·054	156·56	1·36	0·07	0·69	0·88
1984	6/22	11/27	158	4395	3963	316	270	−198	197	0·217	156·66	1·16	0·07	0·65	0·88
1985	6/20	12/4	167	4623	4775	339	225	153	−119	−0·132	156·73	1·49	0·07	0·62	0·88
1986	6/12	12/10	181	4245	4273	361	212	64	−102	−0·113	156·71	1·68	0·08	0·64	0·88
1987	6/21	11/29	161	4579	4209	322	236	−330	1	0·001	156·62	1·35	0·07	0·73	0·90
1988	6/16	12/3	170	5031	4478	336	318	−322	172	0·190	156·73	1·04	0·07	0·55	0·90
1989	6/14	12/1	170	4501	4521	329	197	77	−71	−0·079	156·77	1·64	0·07	0·70	0·90
1990	6/15	11/23	161	4358	4594	304	280	233	−19	−0·022	156·70	1·07	0·07	0·75	0·89
1991	6/18	12/12	177	4706	4960	368	287	438	85	0·094	156·75	1·27	0·07	0·58	0·85
1992	6/14	11/30	169	4782	5254	325	216	504	−83	−0·092	156·78	1·49	0·06	0·62	0·85
1993	6/12	12/24	195	4064	4331	410	217	264	−202	−0·223	156·64	1·87	0·09	0·70	0·85
1994	6/16	12/12	179	4080	4188	368	224	277	15	0·016	156·61	1·63	0·08	0·80	0·85
1995	6/20	12/3	166	3278	3117	332	178	−105	−76	−0·084	156·40	1·84	0·10	0·78	0·84
1996	6/17	12/2	168	5493	4807	332	181	−115	424	0·468	156·73	1·80	0·06	0·79	0·84
1997	6/19	12/12	176	6121	6003	368	341	31	130	0·144	156·94	0·96	0·05	0·76	0·84
1998	5/31	1/7	221	4691	5265	438	366	129	−507	−0·560	156·77	1·12	0·09	0·70	0·84
mean	6/14	12/4	173	4432	4380	344	231	51	−8	−0·009	156·65	1·50	0·08	0·71	0·86
1 std	7	10	14	690	654	31	47	253	156	0·172	0·13	0·27	0·01	0·07	0·02

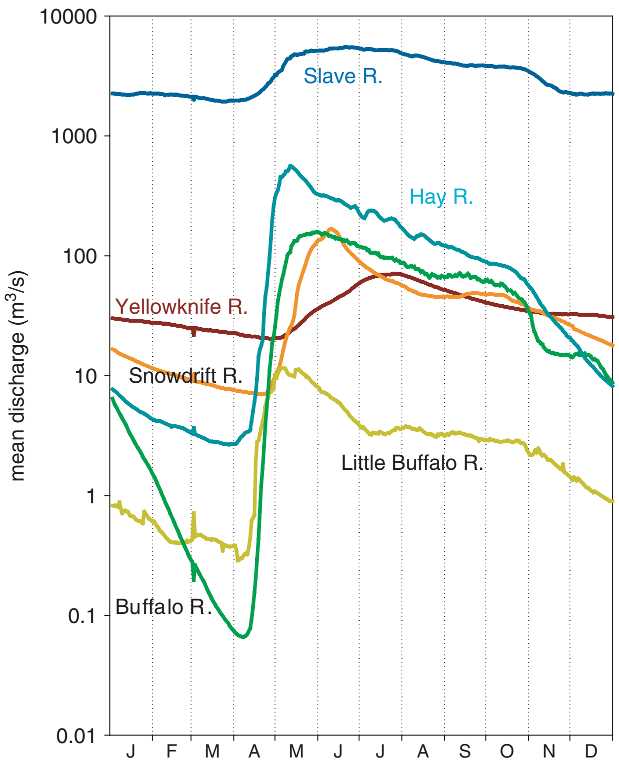

Figure 3. Average hydrographs

a single station alone and served to filter the effect of wind seiche events, which are common on the lake (Sortland, 1994; Gardner *et al.*, 2006). Water level records were used to calibrate the water balance model as described in the following section. The study area, watershed boundaries, and key hydrometric stations are shown in Figure 1.

RESULTS AND DISCUSSION

Model calibration

An operational water balance model based on the datasets described earlier was run on a daily time-step, and averaged to obtain annual and long-term estimates of the water budget components of GSL (Tables II and III). Although seasonal fluctuations in GSL water levels were reasonably replicated using an uncalibrated water balance model, which initially confirmed the dominant control of water balance on lake levels, exaggerated multi-year drift and an overall decline in water levels were predicted by the uncalibrated model (Figure 4(a)). Drift in the lake level predictions is attributable to accumulation of very minor imbalances in the year-to-year inputs and outputs, while long-term decline reflects minor, systematic under-estimation of inputs or over-estimation of outputs to the system. While errors in precipitation and evaporation undoubtedly contribute to the underlying annual and interannual imbalances, the magnitude of the disparities is found to frequently exceed the total estimated evaporation and/or precipitation. Suggested upward adjustments to precipitation for stations in northern Canada (e.g. Metcalfe *et al.*, 1999) were tested and were able to resolve the predicted long-term decline in water levels, but they do not correct for the multi-year (bi-directional) drift (see Figure 4(a)). The dominance of riverine inflow and outflow in the water balance, and the magnitude and bi-directional

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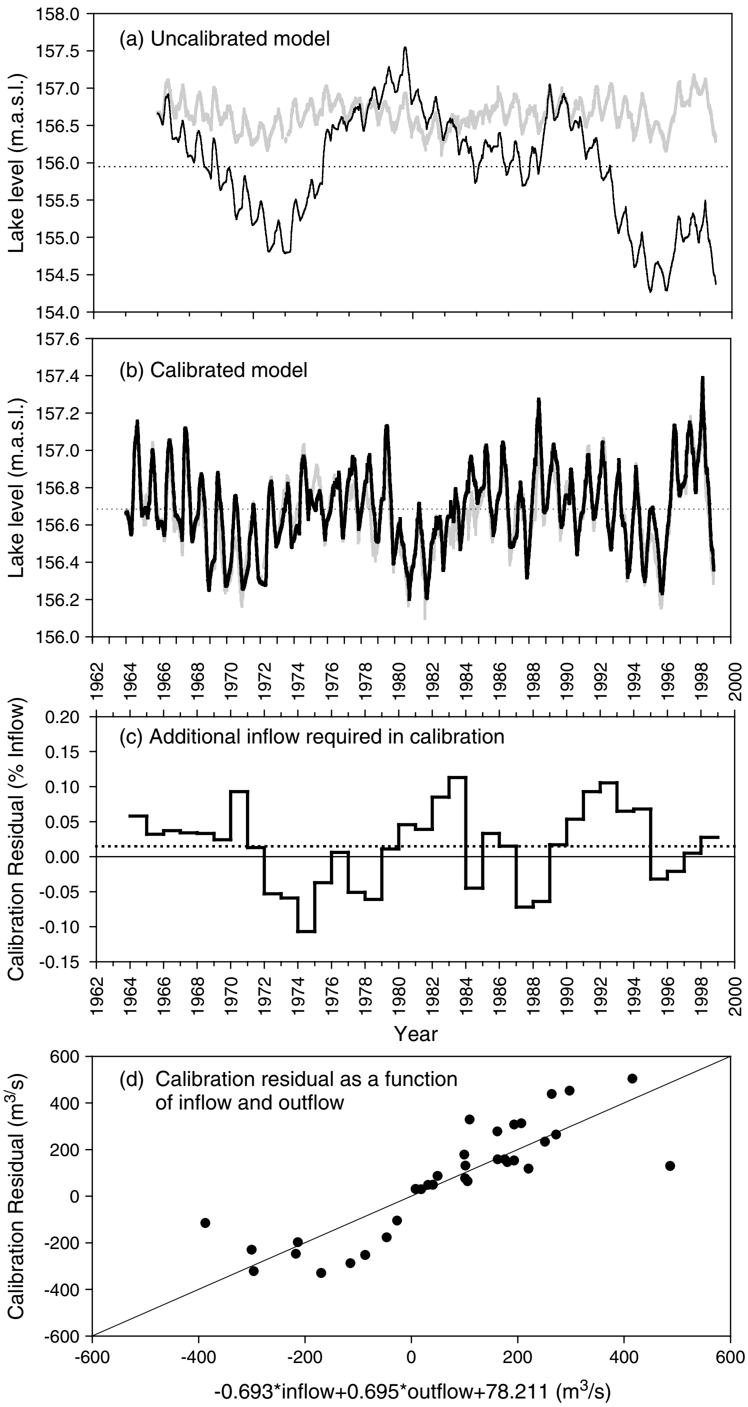

Figure 4. Time-series plots of (a) measured daily GSL lake levels (grey line) and lake levels predicted using the uncalibrated water balance model, 1964–1998; (b) same as (a) but using the calibrated water balance model; (c) additional annual inflow required by the calibration, as a fraction of the riverine inflow (solid stepped line) and mean residual imbalance (dotted line), and (d) calibration residuals predicted from a linear regression of inflow and outflow ($r^2 = 0{\cdot}77$)

nature of the annual imbalances (up to ±11%) suggest that discharge estimates/measurements may be the main source of uncertainty. The observed imbalances are expected, considering that portions of the basin are ungauged, and the uncertainty associated with stream gauging at well-maintained stations typically ranges from ±5% for direct measurements using current meters to ±10% for indirect measurements using rating curves (Tillery *et al*., 2001), with higher potential errors noted under ice conditions (Pelletier, 1990). To maintain a best-fit to the composite water level record for GSL on an interannual basis (Figure 4(b)), it was necessary to calibrate the model for individual years by scaling the inflows by approximately ±10% (Figure 4(c)). Although similar closure could justifiably be achieved by scaling the outflows or using a combination of both approaches, inflow was selected for the calibration as it was considered to be the largest source of uncertainty owing to assumptions inherent in the process of upscaling to representative catchments, and in particular, assignment of proxy runoffs for ungauged basins (see Table IV). As noted previously, ungauged areas accounted for an average of 14% of the drainage area of GSL, varying between 10 and 16% depending on which gauges were operating for individual years. Similar year-to-year scaling of precipitation and/or evaporation was not considered to be a justifiable approach because of their minor roles in the overall water balance (see Table III). Uncertainty in the composite water level record, conversion of water level changes to volumetric equivalents, and groundwater exchange are acknowledged as potential contributory factors, but likely account for a small portion of the overall imbalance. Total inputs (riverine inflow + precipitation) in excess of 2% of the total outputs (riverine outflow + evaporation) are required in the calibrated model to eliminate the long-term decline in the lake level trends for the 1964–1998 period (see Table III). Addition of inflow to achieve the net calibration is considered reasonable as groundwater exchange would likely form a positive addition to the lake, given that it is situated in a regional topographic low. The apparent predictability of calibration residuals on an annual basis from a multi-linear regression of inflow and outflow (Figure 4(c)) substantiates the conclusion that these imbalances arise mainly from uncertainty in the inflow/outflow gauging records from year to year ($r^2 = 0{\cdot}77$). Overall, the calibrated daily model establishes an operational closure of the water budget to within ±10%, which is useful for predicting GSL water levels on a seasonal to interannual basis for the 1964–1998 period. It should be noted that the robustness of the calibrated model in predicting the amplitude of seasonal lake level fluctuations is real, given that a common scaling factor (Figure 4(c)) was applied uniformly to all daily inflows for each year. Of course, the degree of robustness in predicting interannual variations in water balance is largely dependent on the inflow scaling factors. Nevertheless, the calibrated model is shown herein to be a useful tool for investigating the sensitivity of the water balance to hydroclimatic forcings. The calibration exercise also plainly illustrates the inherent sensitivity of GSL water levels to small shifts in inflows and outflows. In the following sections, the mean water balance of the lake, and the seasonal and interannual relationships between water levels and water balance are examined and discussed in more detail using the calibrated water balance model.

Mean water balance

Long-term water balance of GSL, as summarized in Table III, is dominated by throughput of riverine water entering via the Slave River and exiting via the Mackenzie River. Roughly 77% of the total inputs to the lake arise from the Peace-Athabasca drainage basin via the Slave River, while about 18 and 5% of input enters by other rivers and by precipitation on the lake surface, respectively. Among other river basins contributing to GSL, the Taltson (4%), Lockhart (3%), and Hay (2%) account for more than half of these inputs, while numerous smaller basins (Table IV) each contribute <2% to the long-term input. Output of water from GSL is dominated (93%) by outflow to the Mackenzie River, with a small fraction of water lost by evaporation (7%). As noted previously, modelled evaporation exceeds precipitation by approximately 50%, mean evaporation losses (E/I) are close to 8%, and mean throughput (Q/I) is close to 92%.

The mean input of $1{\cdot}47 \times 10^{11}$ m^3 year^{-1} distributed across the lake area ($2{\cdot}856 \times 10^{10}$ m^2) yields an estimate of the equivalent depth of inflow of 5·2 m year^{-1}, about 4 m of which is derived from the Peace-Athabasca basins. Similar calculations for individual years range from about 3·8 to 7·1 m. For comparison,

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Table III. Water balance summary ($m^3 s^{-1}$), 1964–1998. The bracketed values indicate percentage of total input or output for mean or for specified year with maximum or minimum volumetric flux

Component		Mean	Max.		Min.		1 std
			Year		Year		
Inputs							
	Riverine Inflow	4 432 (95%)	**1997**	6 121 (95%)	**1980**	3 190 (94%)	651
	Slave R.	*3432 (77%)*	**1997**	*4922 (76%)*	**1970**	*2594 (73%)*	*553*
	Other Rivers	*999 (21%)*	**1988**	*1560 (29%)*	**1980**	*471 (14%)*	*241*
	Precipitation	231 (5%)	**1998**	366 (6%)	**1995**	178 (5%)	47
Outputs							
	Riverine Outflow	−4 380 (93%)	**1997**	−6003 (94%)	**1995**	−3117 (90%)	654
	Evaporation	−336 (7%)	**1993**	−410 (9%)	**1973**	−279 (6%)	31
Residual		51 (1%)	**1992**	504 (10%)	**1974**	−602 (−10%)	253

Table IV. Summary of watersheds contributing riverine inflow to Great Slave Lake

Ref[a]	Watershed	Gauged area km^2	Upscaled Representative area km^2	Runoff	
				$m^3 s^{-1}$	$mm yr^{-1}$
1	Kakisa	15 600	16 454	43	83
2	Hay	47 900	47 900	114	73
3,3a	Buffalo	18 500	20 504	48	73
4	Little Buffalo	3 330	14 238	12	27
5	Ungauged	16 386	21 422	107[b]	157[b]
6,6a	Taltson	58 700	60 439	195	102
7	Snowdrift	9 110	16 386	81	157
8,8a,8b,8c	Lockhart	26 660	27 327	122	143
9	Waldron	1 830	24 204	78	99
10	Yellowknife	16 300	18 848	31	51
11	Wecho and Snare	18 600	25 482	68	85
12	Emile	4 850	9 895	32	98
13	La Martre	13 900	16 038	33	63
14	Slave	606 000	606 000	3432	179
Ug	*Other Ungauged Basins*		23 812	68[c]	85[c]
	Catchment Area	857 666	948 949	4432	146
	Lake area	28 568	28 568	—	—

[a] See Figure 1.
[b] From Snowdrift record.
[c] From Wecho and Snare record.

reductions in flow during the filling years of the Williston Reservoir equates to an approximate 1·4 m decrease in lake levels. From the estimated mean volume of the lake ($2{\cdot}088 \times 10^{12}$ m^3; Kalff, 2002) and the volumetric inflows estimated herein, the mean residence time (volume/total inputs) is calculated to be 14·2 years for the 1964–1998 period, ranging from 10·4 to 19·7 for individual years. The mean residence time is similar, although slightly lower than values previously reported (16 years; Evans, 2000). It is important to note that the west basin of GSL is expected to flush about 2 times more rapidly than the east arm (Evans, 2000) owing to shallower mean depth and bypass of Slave R. water directly to the outlet of the Mackenzie River (see Figure 1). The drainage basin area to lake volume ratios for GSL are high (454 $m^2 m^{-3}$) as compared to

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the Laurentian Great Lakes (34 m^2 m^{-3}), which explains the higher average rates of throughflow and 50% shorter average residence times for GSL. Note that mean residence times vary between 191 and 2·6 years for the Laurentian Great Lakes, depending mainly on position in the drainage network. Average mean residence times of 30 years reflects an integrated estimate for all lakes.

Overall, the long-term water balance provides a view of the relative magnitude and importance of the riverine inflow and outflow components of the water balance, which are an order of magnitude larger than the estimated evaporation and on-lake precipitation (Table II). It is important to note these features while considering the historical hydroclimatic forcings and regulatory impacts on the lake.

Seasonal and interannual variability

Good reproducibility of the amplitude and timing of seasonal water level fluctuations in most years between 1964 and 1998 (Figure 4) suggests that the calibrated water balance model captures the dominant processes controlling GSL lake levels. The seasonal cycle is largely a product of seasonality in the riverine inflows to GSL, predominantly via the Slave River. The most significant historical changes to the seasonal cycle of lake levels occurred owing to regulation of the Peace River, which has effectively increased mean discharge from the Slave River during winter and reduced flows during the spring freshet and summer periods, apparently affecting seasonal variations in lake levels in a similar way (Figure 5). Precisely defining the effect of regulation on lake levels would require modelling of naturalized (without the effect of impoundment) flows for the post-regulation period; a topic beyond the scope of this investigation.

Typical interannual variations in water balance are described by standard deviations for the individual water balance terms calculated in Table III, which correspond to changes of ±15% in total riverine inflows and outflows, ±20% in precipitation, and ±10% for evaporation. Typical variability in the Slave River inflows

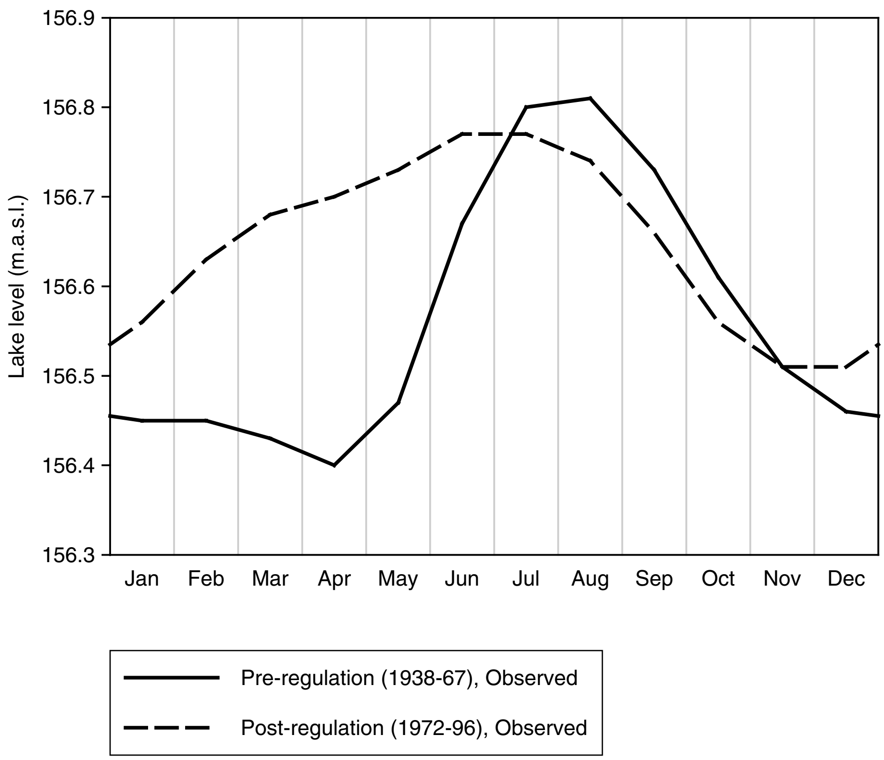

Figure 5. Alteration in mean seasonal water level variations due to regulation, Great Slave Lake

(±16%) is reduced compared to other riverine inflows (±24%), reflecting persistence of discharge from Peace-Athabasca basin, and slightly higher variability in runoff from shield and wetland basins surrounding the lake. The fraction of riverine inflow derived from the Slave River, as shown in Table II, ranges from 57 to 83% in individual years, averaging about 80%. Apart from a temporary decline in Slave River contributions during 1968 and 1969, when significant Peace River discharge was abstracted to fill the Williston Reservoir, variations in this proportion are predominantly influenced by hydroclimatic (precipitation and runoff) variability in the contributing basins, as discussed later. Higher variability for shield basins, where string-of-lakes drainages may serve as sources or sinks of water in individual years, does not produce substantial variations in the annual water budget owing to the dominance and persistence of Slave River contributions. Note that the comparatively low interannual variability in estimated evaporation rates may be an artifact of the ice-free period algorithm used in the calculations. While the evaporation routine is a significant improvement over previous studies, this is an indication of the need for further evaporation studies of the lake.

Evaporation/inflow ratios (E/I) are found to be strongly correlated with mean lake level (Figure 6), reflecting the clear control of lake level on throughflow and hence discharge to the Mackenzie River. Year-to-year variations in E/I (Q/I) are restricted to a fairly narrow range, from approximately 6 to 10% (90 to 94%), which demonstrates that the lake has not experienced drying cycles severe enough to restrict outflows to the Mackenzie River during the 1964–98 period, or is likely during the twentieth century. Future alterations in inflow due to climate or additional regulation could potentially lead to a more critical balance between inflow and outflow in the future, although the hydraulically controlled outlet is at approximately 153·5 masl (Hicks *et al.*, 1995), or 2·7 m below the historical low (Nov. 1980, 156·2 m), and the lake level has apparently remained stable to within 1·2 m during the last 70 years, including the period of filling of the Williston Reservoir and the historical low period in the mid-1940s. Approximately 4% of GSL water is held in hydraulic storage (available for outflow) whereas 96% can be considered dead storage (below the current zero outflow threshold), and yearly turnover is about 7%. Because GSL is a high throughflow system, the continuity of lake outflow in future will depend to a large extent on the changing climate conditions and on the voracity of water abstraction in the contributing basins, particularly in the dominant source regions such as the Peace-Athabasca basin.

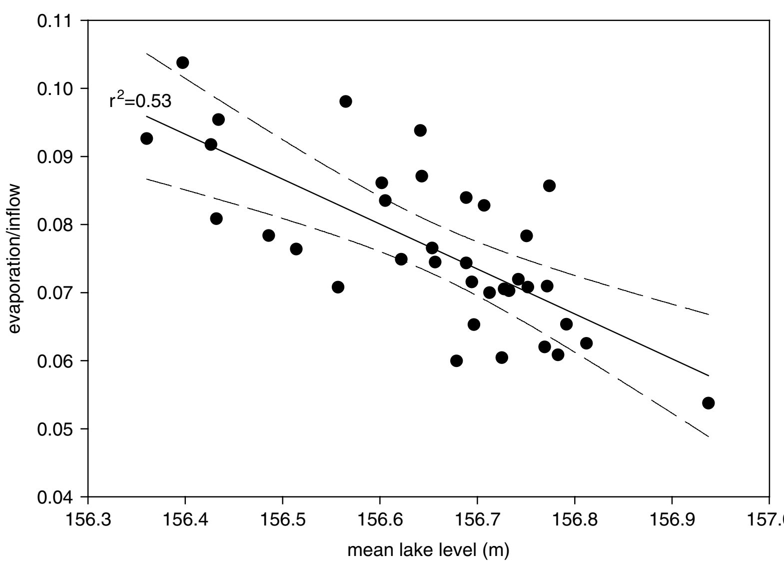

Figure 6. Plot of throughflow index (E/I) and water level relationships in Great Slave Lake. Correlation coefficient and 99% confidence interval for the linear regression is shown

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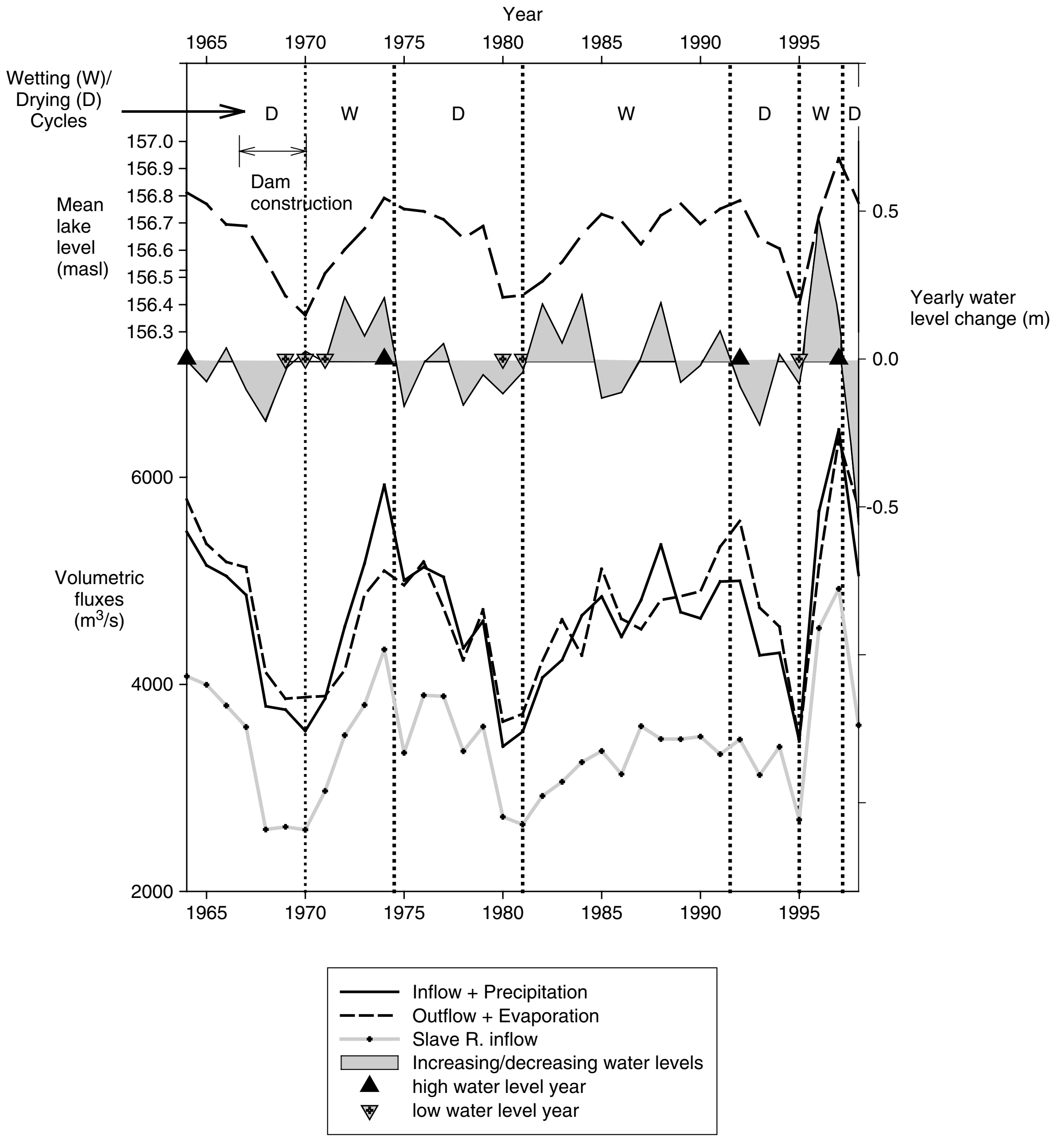

Figure 7. Annual time-series illustrating years of increasing and decreasing GSL lake levels (shaded), and wetting and drying cycles (W/D) identified in the inputs (riverine inflow + precipitation), Slave River inflow, and outputs (riverine outflow + evaporation) from Great Slave Lake, 1964–1998. Years of high and low GSL lake levels, indicated with upward and downward triangles, tend to occur near the end of wetting and drying cycles, respectively

Hydroclimatic controls on interannual wetting/drying cycles

Typical seasonal fluctuation in GSL water levels is close to 0·4 m. Three major interannual wetting (and drying) cycles were also observed during the 1964–1998 period (Figure 7), marked by general increases (declines) in inputs (outputs) and related increases (declines) in GSL lake levels. Periods of wetting and drying, as shown in Figure 7, are typically linked to cyclical variations in riverine inflow from the Slave River, which strongly controls the overall pattern of the total inputs and the resulting pattern of outputs via control of the lake stage. Minimum lake stages (Figure 7) were observed following periods of sustained drying (1980–1981, 1995), and during the filling of the Williston Reservoir at the Bennett Dam (1968–1970). Maximum lake levels were observed following periods of sustained wetting (1964, 1974, 1992, 1997). Interestingly, the most rapid changes in lake level occurred during November 1995–August 1996, when lake levels rose by over 1 m in response to wetter than normal climate conditions and additional water releases from the Williston reservoir. Generally wet conditions and increasing water levels were also observed during much of the 1980s. The pattern that emerges from the time-series analysis is the overriding influence of the Slave River inflow, which is dependent on precipitation and runoff in the Peace-Athabasca Basin, on interannual variability in the GSL water balance and in turn GSL water levels.

A variety of datasets and teleconnection indices were examined to identify potential relationships between the water levels in Great Slave Lake and hydroclimatic conditions in the basin (Figure 8), including

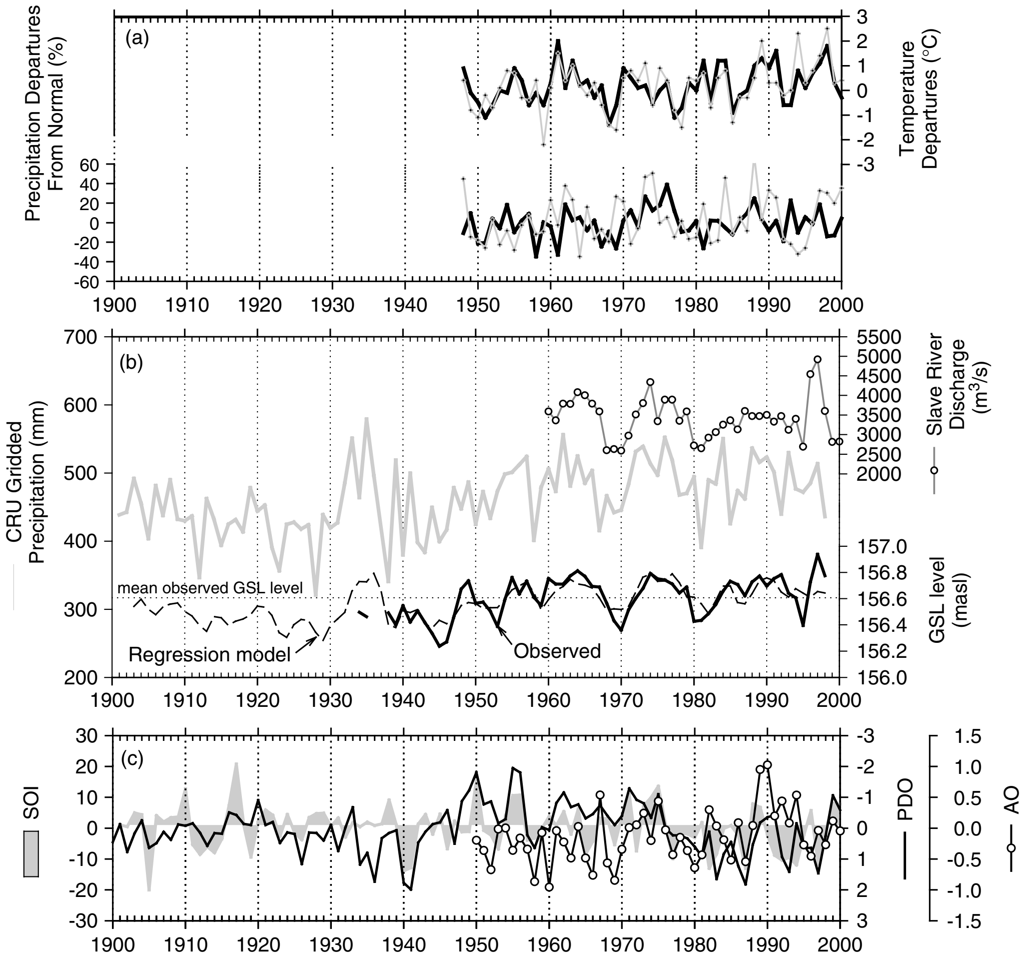

Figure 8. Time-series plots for the twentieth century illustrating (a) temperature and precipitation departures from normal in the Northwestern Forest and Mackenzie District (Environment Canada, 2003), (b) Slave River discharge (m^3 s^{-1}), areally-weighted precipitation for the Peace-Athabasca basin compiled from the CRU gridded dataset (mm), and GSL water levels based on observational measurements and from a simple CRU precipitation regression model, and (c) selected teleconnection indices, namely the South Oscillation Index (SOI), the Pacific Decadal Oscillation (PDO), and the Arctic Oscillation (AO)

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temperature and precipitation departures for the *Northwestern Forest and Mackenzie Districts* (Environment Canada, 2003), Slave river runoff (Environment Canada, 2001), and CRU TS 2.0 gridded ($0{\cdot}5° \times 0{\cdot}5°$) observed precipitation for the Peace-Athabasca basins (Mitchell *et al*., 2004, see also Mitchell and Jones, 2005) (Northwestern Forest and Mackenzie District regions as defined by Environment Canada's Climate Trends and Variations Bulletin, http://www.msc-smc.ec.gc.ca/ccrm/bulletin/). Correlation coefficients between the annual Slave River runoff and GSL water levels (1960–2002) are high ($r^2 = 0{\cdot}76$), although direct correlation between GSL water levels (∇_j) and areally-weighted annual precipitation (P_j) in the Peace-Athabasca basin is modest, largely due to a weaker relationship between basin precipitation and runoff ($r^2 = 0{\cdot}16$).

$$\nabla_j = 155{\cdot}995 + 0{\cdot}00128^* P_j \quad r^2 = 0{\cdot}17$$

Precipitation/runoff relationships are complex within large systems, as in the Peace-Athabasca Basins (606 000 km^2), where the runoff regime is strongly seasonal and includes a range of alpine to lowland sub-basins that individually respond to antecedent moisture conditions, a pronounced snowmelt cycle, summer rainfall events, and significant evapotranspirative losses during summer. Runoff ratios in sub-basins of the Peace-Athabasca basins are known to range from about 0·91 in alpine areas to 0·2 in lowland shield and wetland-dominated areas in the northeastern portions of the basin adjacent to lake Athabasca (Lawford, 1994), and average 0·37 on the basis of the datasets presented herein. While mean monthly runoff and precipitation are strongly correlated for the Peace-Athabasca basin ($r^2 = 0{\cdot}78$), annual runoff/precipitation ratios vary considerably, from 0·29 to 0·5 for individual years during 1960–98, owing to antecedent moisture and other factors mentioned previously. Weaker direct correlations between precipitation and GSL water levels are therefore expected.

A multi-linear regression analysis was also conducted using the gridded CRU TS 2·0 precipitation dataset for the Peace-Athabasca basin, where GSL water level and precipitation relationships (1934–1935, 1938–1998) were interrogated using precipitation in the same year P and in the previous two years as shown below. This analysis produced much higher correlations.

$$\nabla_j = 155{\cdot}246 + (0{\cdot}0011^* P_j) + (0{\cdot}00175^* (P_{j-1}) \quad r^2 = 0{\cdot}48$$

$$\nabla_j = 155{\cdot}068 + (0{\cdot}000883^* P_j) + (0{\cdot}00170^* (P_{j-1}) + (0{\cdot}000653^* (P_{j-2}) \quad r^2 = 0{\cdot}52$$

where ∇_j and P_j are GSL lake level and annual precipitation in year j, P_{j-1} being annual precipitation in the previous year and P_{j-2} being precipitation in the year before the previous year, suggests robust and predictable empirical relationships between precipitation and GSL water levels. Conceptually, this approach accounts for the influence of antecedent moisture conditions and interannual storage of precipitation prior to runoff in the Peace-Athabasca basins. Significant but slightly weaker correlations were found using the precipitation departures datasets (Figure 8(a)) as the regional datasets are not demarcated specifically for the contributing areas of Great Slave Lake. Comparison of several teleconnection indices with GSL water levels and also with precipitation revealed very poor correlations, in part due to the complexity and areal extent of the basin, and presumably due to hydrologic smoothing of the climatic signals over several years. The only significant correlation was found between the Arctic Oscillation Index and annual precipitation over the Peace-Athabasca basin ($r^2 = 0{\cdot}11$), although the correlation was not significantly improved by multi-linear regression using several teleconnection indices or using the previous year approach described above. While there are no simple relationships apparent between El Nino or other teleconnection cycles and water levels in Great Slave Lake, the predictability of water levels from multi-year precipitation in the Peace-Athabasca basin is encouraging, and warrants further analysis to dissect potential seasonal patterns. However, seasonality of lake levels, as shown in Figure 5, mirrors the seasonal shift in river flow that occurred after the Peace River was regulated, a factor that needs to be considered in any future analysis of water balance on a sub-annual time scale.

Forty-eight percent (138 km^3) of the long-term annual discharge from the Mackenzie River to the Arctic Ocean (290 km^3 year^{-1}) consists of outflow from Great Slave Lake, and a total of 57% originates from Great

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Slave and Great Bear Lakes combined. While Great Bear Lake has significantly longer residence times and therefore is expected to have a longer response time to precipitation variability in its contributing basin, Great Slave Lake water levels are found to respond rapidly to changes in Slave River inflow, which appear to be the product of a 2–3 year dampening of upstream precipitation signals in the Peace-Athabasca basins. As a high (93%) throughflow system with a hydraulically controlled outlet, Great Slave Lake has limited hydrologic buffering capacity, despite its mean residence time of 14·2 years. Nevertheless, the lake is expected to play a more dominant role in buffering biogeochemical parameters, which depend more strongly on residence time, and on sediment loading to the downstream reaches of the Mackenzie River and Arctic Ocean.

Concluding remarks

Evaporation remains a poorly known component of the water balance of Great Slave Lake; however, the observed relationship between evaporation and ice-free period has been used in this analysis to constrain estimates and to allow a practical closure of the water balance for annual periods. Groundwater also remains a principal unknown in the water budget, although its magnitude, as with evaporation, is expected to be a small component of the overall water budget. Ongoing studies also include development of a naturalized historical water budget of GSL for analyzing and partitioning the hydroclimatic and regulatory impacts on the water level of GSL (see Gibson *et al*., 2006). This will expectedly improve upon the baseline hydrologic assessment of GSL on interannual and sub-annual time scales, and will provide a basis for evaluating potential impacts on the Slave River Delta. New efforts to monitor the stable isotope composition (^{18}O and ^{2}H) of GSL and the Mackenzie River system are also being undertaken to improve upon estimates of evaporation losses and to constrain subsurface water exchanges. Combined, these efforts will contribute to a more quantitative understanding of the role of large lakes on the Mackenzie River system, a major GEWEX study basin, a primary freshwater source to the Arctic Ocean, and a potential factor driving North Atlantic thermohaline circulation and global climate change.

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