Lake-gauge evidence for regional postglacial tilting in southern Manitoba

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ABSTRACT

The network of lake-level gauges in place for Lake Winnipeg, Lake Winnipegosis, Lake Manitoba, and Lake of the Woods presents an opportunity to study the contemporary pattern of postglacial rebound in a North American continental-interior region. We used a regional trend-surface model based on mean winter-season differences of lake level between pairs of gauges to extract the postglacial rebound signature recorded in the lake-gauge data. The results from a best-fitting planar model indicate that tilting is up to the northeast with a bearing of N(42.7 ± 11.8)°E and a rate of (10.7 ± 2.2) × 10^{-9} rad·yr^{-1}. In spite of the fact that both the ICE-3G and ICE-4G load models are derived from the inversion of relative sea-level histories obtained from marine coastal sites, the pattern of rebound that we computed from these two ice models for this continental-interior region agrees reasonably well with the pattern derived from the lake-gauge data. Rates computed from an ICE-3G load are in close agreement with those computed from the lake gauges, whereas rates derived from ICE-4G are 50% too slow. When the planar pattern of postglacial rebound derived from the lake gauges in southern Manitoba and Lake of the Woods is linked to the lake-gauge–derived pattern of rebound over the Great Lakes region, the two-region pattern suggests influential source regions of rebound that include southern and southwestern Hudson Bay as well as more local sources in western Ontario. It is possible that the spatial arrangement of ice domes predicted in the multidomed paleotopography of ICE-4G could explain the two-region pattern of rebound if more ice were added to Ontario in the model, or if Earth structure assumptions could be modified, or both.

As postglacial tilting continues to shift the region's lakes to the southwest, there is an increased risk of flooding and wave erosion along the south shores over the long term. In addition, postglacial tilting deforms the datums to which the lake-level gauges are referenced. Thus, where and how the levels of these lakes are measured is an important operational issue. These factors present challenges to water-resource managers who are responsible for monitoring the levels and controlling the outflows of these lakes.

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INTRODUCTION

Observations of vertical surface motions induced by glacial loading can be used to model glacial-loading history, constrain upper-mantle viscosity and lithosphere thickness, and predict future land-surface tilting that can deflect the water bodies of large lakes and the courses of rivers. Many geodynamic studies have relied on elevations of raised marine shorelines and tide gauges as a measurement of load-induced vertical motion for the North American interior. Although significant work has been done in the Great Lakes region (Clark and Persoage, 1970; Coordinating Committee, 1977; Hansel et al., 1985; Larsen, 1985, 1987, 1994; Tushingham, 1992; Clark et al., 1994; Lee and Southam, 1994) and for the tilted shorelines of former Lake Agassiz in Manitoba (Upham, 1895; Johnston, 1946; Teller and Thorleifson, 1983), relatively few observational constraints come from the North American continental interior. An opportunity to remedy that situation exists in southern Manitoba.

Tackman et al. (1998) extended the postglacial tilting history of the region from the time of Lake Agassiz (about 9000 ^{14}C yr B.P.) through to this millennium by using the tilted paleoshorelines of Lakes Winnipegosis and Dauphin (Fig. 1). In this paper, we extend that work through to the present by using a regional trend-surface model to extract the postglacial rebound signature from the lake-gauge records of the region's large lakes. Specifically, we present a reconstruction of the historic spatial pattern of postglacial rebound for the continental-interior region of southern Manitoba and western Ontario using the lake gauges from Lake Winnipeg, Lake Winnipegosis, Lake Manitoba, and Lake of the Woods. We also compare the lake-gauge results to the regional pattern of deformation that we computed from ICE-3G and ICE-4G loading histories and extend our rate contours to the lake-gauge–derived pattern over the Great Lakes computed by Clark and Persoage (1970). Finally, we include a discussion of the effects that postglacial rebound has had in the past, and is likely to have in the future, on the levels of the region's large lakes. This discussion should be of interest to water managers, planners, engineers, and other decision makers interested in regional lake-level management issues.

VERTICAL CRUSTAL MOTION RECORDED BY LAKE GAUGES

The dominant signature recorded in each lake-gauge record is the change in hydrologic regime between wet and dry seasons and wetter or dryer years or

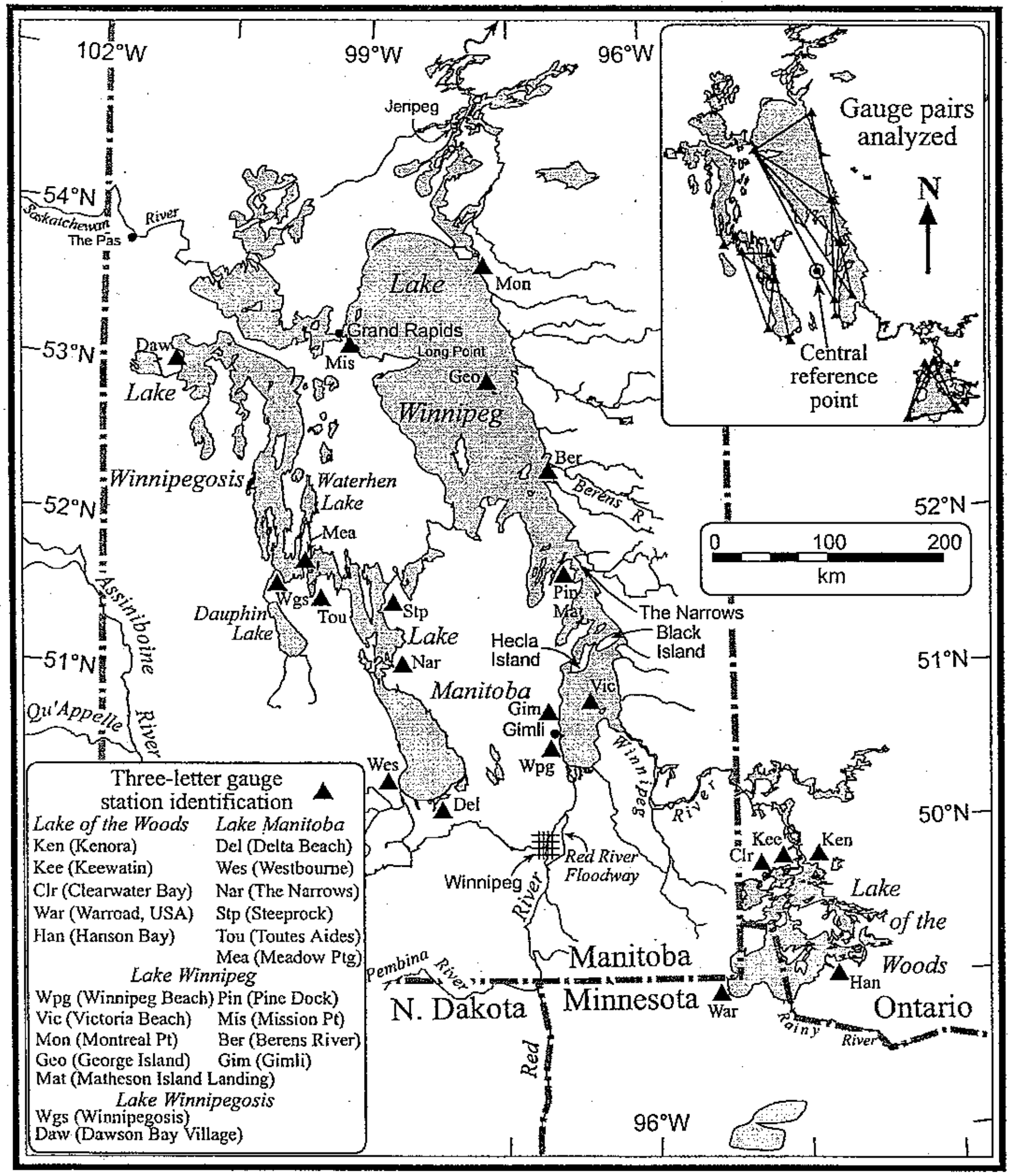


Figure 1. General location map showing lake-gauge stations and other places discussed in text. Note the shorthand three-letter identification for each gauge station. This notation is used throughout the paper. For example, the Gimli minus Pine Dock time series is referred to as GimPin. Similarly, the Winnipegosis minus Dawson Bay time series is referred to as WgsDaw, etc.

decades (Fig. 2). Most of that signature is shared by all gauges on a given lake. Taking the simple difference of two gauge records on the same lake (a procedure termed *differencing*) removes most of that hydrologic-regime signature (also referred to as lake signature in this paper) and reveals a secular trend that, in this region, we attribute to postglacial rebound. This secular trend represents the relative, or differential, vertical motion between two gauges. The secular trend does not represent absolute vertical motion relative to some stable reference. Absolute vertical motion cannot be measured directly from lake gauges or tilted paleoshorelines. However, the differential vertical motion measured by the apparent change in lake level between two gauges is due to, and is a direct measure of, the deformation of the datum to which the lake gauges are referenced. Furthermore, if it is known that the reference datum is a good approximation to the geoid, then the differential vertical motion can also be interpreted as the separation between surface topography and the deforming geoid. In other regions of the world, this differential vertical motion can be due to other processes, such as subduction-zone deformation.

With these ideas as background, consider the expected hydrographic and geodynamic responses for Lake Winnipeg, Lake Manitoba, and Lake of the Woods. All three lakes have positive water budgets and are controlled at their northern outlets by dams. If lake level is held constant relative to a mark on the dam, then postglacial tilting up to the northeast will shift the water bodies of these lakes to the southwest, and the datums to which the gauges are referenced will be tilted down to the southwest. For each lake, gauges that are located south or west of the tilt contour that passes through the outlet will record an apparent increase in lake level relative to the deformed datum. Gauges located north or east of the tilt contour that passes through the outlet will record an apparent decrease in lake level relative to the deformed datum. Lake level at the dam will appear to remain at a constant elevation relative to the deformed datum. Consequently, the difference between any pair of gauge records in which the baseline between the pair is not oriented parallel, or nearly parallel, to the tilt contours should show a secular trend that represents the differential

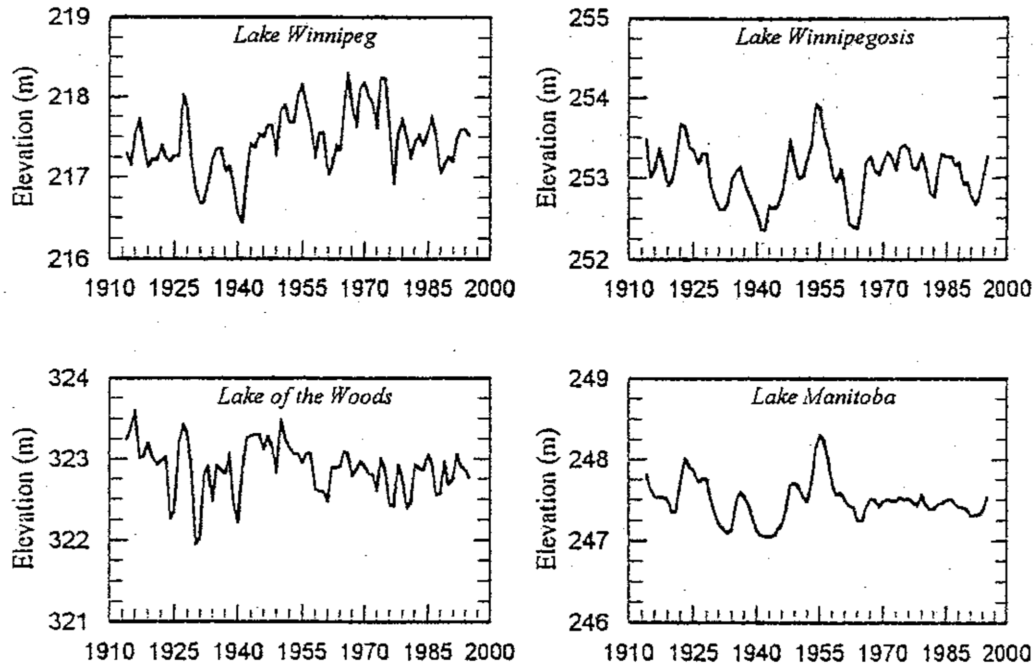


Figure 2. Annual mean lake-level histories.

vertical motion between the pair of gauges, as long as each gauge is referenced to the same datum for its entire length of record.

DATA

The network of lake gauges for Lake Winnipeg, Lake Winnipegosis, Lake Manitoba, and Lake of the Woods (Fig. 1) is in place to monitor the levels and help manage the outflows of these lakes. The Water Survey of Canada is responsible for the data collection and monitoring of the levels of these lakes. The data are distributed on compact disk (CD) under the commercial name HYDAT by EarthInfo, Inc. We used HYDAT version 4.95, which includes daily average lake levels, river discharges, and some river sediment data for all of Canada for various periods of record through 1995. We also acquired hourly lake levels for the three Manitoba lakes for the years 1993 and 1994 directly from the Water Survey of Canada in Winnipeg to evaluate the effect that short-term lake fluctuations might have on the long-term rebound signature. Bench-mark and datum histories were also acquired to verify that each gauge is referenced to the same datum for its entire period of record and to assess bench-mark integrity. The Water Survey of Canada and Manitoba Hydro supplied the bench-mark and datum histories for the three Manitoba lakes. The Lake of the Woods bench-mark and datum histories were acquired through the Lake of the

TABLE 1. BASIC STATISTICS FOR THE GAUGE STATIONS CONSIDERED IN THIS STUDY

Station name	Three-letter code	North Lat	West Long	Length of record	Uncertainty (mm)	Mean (m)	Datum*
Lake Winnipeg (regulated since 1975)							
Winnipeg Beach	Wpg	50°30′22″	96°57′55″	1913–1966	60–100	217.387	GDS
Gimli	Gim	50°37′50″	96°58′58″	1966–1995	3–5	217.616	LWD (1986)
Victoria Beach	Vic	50°42′30″	96°34′10″	1959–1995	3–5	217.599	LWD (1986)
Pine Dock	Pin	51°38′30″	96°47′45″	1958–1995	15–30	217.549	LWD (1986)
Berens River	Ber	52°21′20″	97°00′15″	1914–1995	15–30	217.431	LWD (1986)
Mission Point	Mis	53°11′00″	99°13′30″	1953–1995	15–30	217.630	LWD (1986)
Montreal Point	Mon	53°37′30″	97°50′40″	1969–1995	?	217.602	LWD (1986)
Lake Manitoba (effectively regulated since 1963)							
Delta Beach	Del	50°11′15″	98°19′00″	1914–1969	?	247.488	GDS
Westbourne	Wes	50°15′15″	98°34′50″	1964–1995	15–30	247.452	GDS (1935)
The Narrows	Nar	51°04′50″	98°46′50″	1958–1995	15–30	247.462	GDS (1964)
Steep Rock	Stp	51°26′26″	98°48′13″	1923–1995	3–5	247.514	GDS (1964)
Meadow Portage	Mea	51°37′15″	99°34′30″	1952–1968	?	247.568	GDS
Toutes Aides	Tou	51°31′18″	99°32′23″	1969–1993	60–100	247.380	GDS (1928)
Lake Winnipegosis (unregulated)							
Winnipegosis	Wgs	51°38′45″	99°55′07″	1913–1995	3–5	253.091	GDS (1929)
Dawson Bay	Daw	52°58′30″	100°58′30″	1962–1993	15–30	252.939	GDS (1962)
Lake of the Woods (regulated since 1898 at the Norman Dam)							
Warroad	War	48°54′20″	95°19′00″	1916–1993	3–5	322.885	LoWD
Hanson Bay	Han	49°07′34″	94°17′16″	1962–1995	15–30	322.841	LoWD
Clearwater Bay	Clr	49°43′06″	94°48′20″	1963–1995	3–5	322.804	LoWD
Keewatin	Kee	49°45′50″	94°33′15″	1913–1995	15–30	322.811	LoWD

*LWD (1986) = Lake Winnipeg datum, which is BM78M079, GSD (1960), at Berens River. LoWD = Lake of the Woods datum, which is the same as U.S. Coast & Geodetic Survey datum (1912). GDS (year) = Geodetic Survey of Canada datum and the year the bench mark was tied to the datum.

Woods Secretariat, Ottawa, Ontario. A summary of the lake-gauge stations considered for this research is presented in Table 1.

Potential Noise Sources

At the shortest time scales of interest (hours to days), the water level in these lakes fluctuates in response to atmospheric disturbances such as wind, air-pressure differences, and high-rainfall events. This pattern is especially true during the ice-free months when wind setup, seiching, and cobasin oscillations are common on these lakes (Figs. 3 and 4). The pattern of pronounced oscillation typically persists from about late April, when thawing begins to expose the surfaces of these lakes, to late November, when the lake surfaces become frozen again. To minimize these short-term effects, we used a four-month winter-season average (December through March) when these lake surfaces are considerably calmer under the cover of about 1 m of ice (Fig. 4). This choice of measuring period is contrary to the normal practice in which the ice-free season is considered the quiet time and a four-month summer-season average (June through September) is used. For example, most investigations of Great Lakes gauges (e.g., Clark and Persoage, 1970; Coordinating Committee, 1977; Tait and Bolduc, 1985; Tushingham, 1992) have used a four-month summer-season mean to minimize the effects of gales, ice, and spring runoff (Tait and Bolduc, 1985). The advantages of using the calmer winter-season lake-level data for the four lakes analyzed in this study are (1) the variance of a four-month winter-season average is one-third less than that of a summer-season average; (2) after differencing all possible (i.e., 28) pairs of gauges, six of the summer-season time series of differences still contained some lake signature—as measured by correlations with lake level that were in excess of 0.50—whereas none of the winter-season time-series of differences had correlations in excess of 0.50; and (3) the reduced variance and lack of residual lake signature in the winter-season averages produced better model fits as measured by higher coefficients of determination (r^2) and lower standard errors.

The effects of regulation are also present in the data. This finding is most apparent when comparing the unregulated lake-level history of Lake Winnipegosis with its nearby, regulated, downstream neighbors Lakes Manitoba

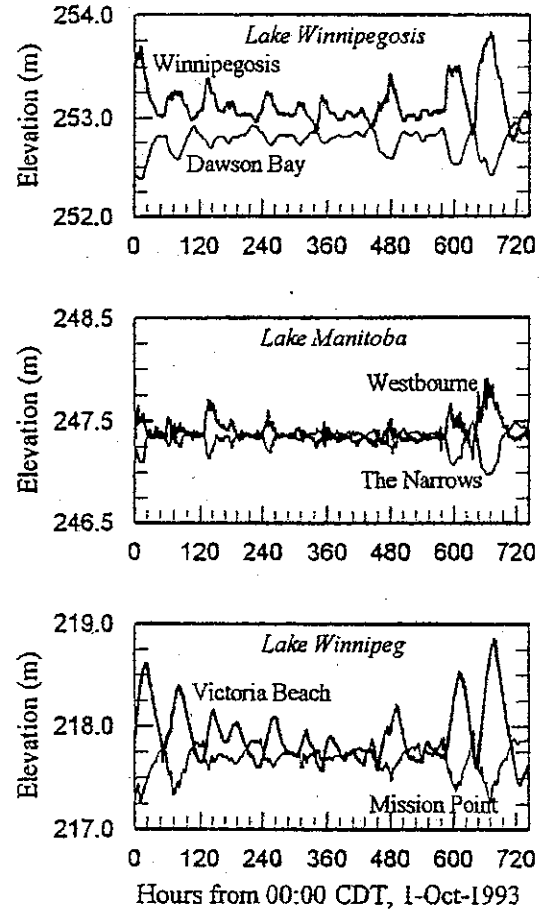


Figure 3. Hourly lake levels for October 1993, showing a pattern of persistent seiching.

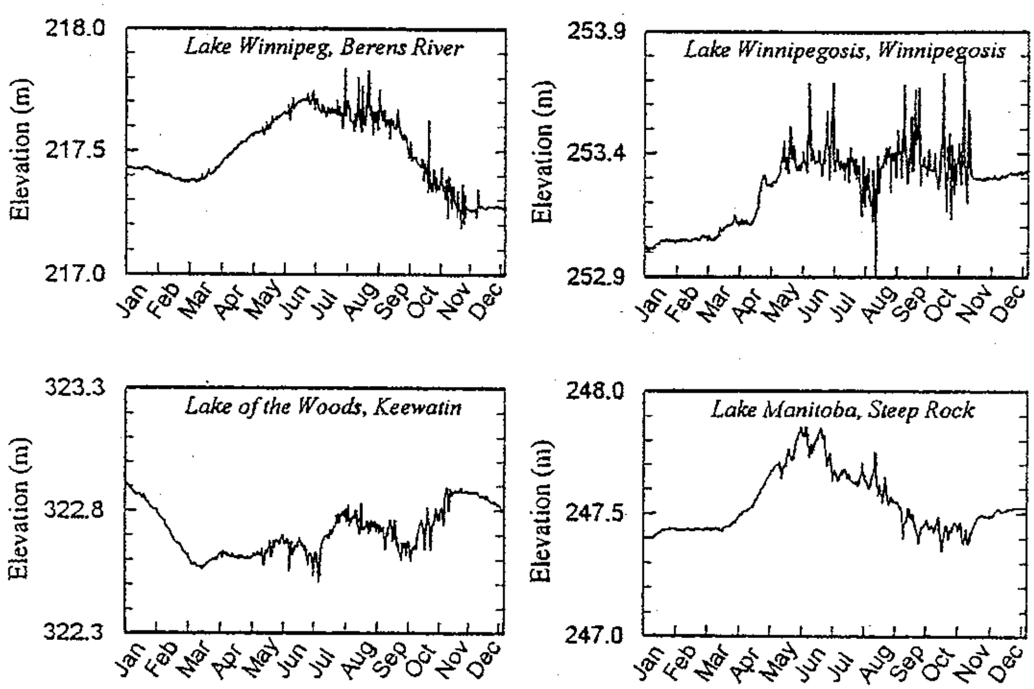


Figure 4. Daily lake levels for 1995. Note the rapid variation in lake level during the ice-free months. The contrast between the quiet winter season and active summer season was the deciding factor for choosing a winter-season mean for analysis over a summer-season mean.

and Winnipeg (Fig. 2). Before 1963, the level of Lake Manitoba was not actively managed, and its record compares well with that of Lake Winnipegosis ($r = 0.91$; $p < 0.05$; $n = 50$). Since 1963, when the small check dam at the Fairford River outlet was improved, Lake Manitoba water level has been actively managed, and, as a result, the Lake Manitoba pattern is no longer similar to that of Lake Winnipegosis. Instead, the lake-surface elevation is stabilized at about 247.5 m (812 ft), whereas Lake Winnipegosis continues to fluctuate freely in response to climate variability. Prior to 1976, Lake Winnipeg was not directly regulated, although after 1965, flow from the Saskatchewan River (about 33% of total river input) into the north basin of Lake Winnipeg was regulated, and flow from the Winnipeg River (about 38% of total river input) into the south basin of Lake Winnipeg has been regulated since 1892 at Lake of the Woods. Lake Winnipeg's pre-1976 record compares reasonably well with that of Lake Winnipegosis ($r = 0.61$; $p < 0.05$; $n = 63$). Since 1976, the level of Lake Winnipeg has been actively managed to maintain an elevation of 216.7–217.9 m (711–715 ft) by controlling outflow at the Jenpeg dam (Fig. 1) in accordance with the Manitoba Hydro water-power license.

Another potential source of noise is dynamic response to river input for those gauges located near the mouths of rivers. A simple correlation analysis of the residuals obtained from the linear regression of winter-season gauge-pair differences with winter-season river input revealed that no correlations exist with an absolute value greater than 0.20. Because river input correlates highly with lake level, it is removed along with the lake signature when taking the difference of the gauge pairs.

Lake-basin constrictions and other geometries that affect circulation might also produce unwanted short-term effects. Such effects may have occurred in the early 1980s when the causeway between the mainland and Hecla Island in the south basin of Lake Winnipeg was constructed. The causeway may have affected circulation through the already restricted area of the Hecla–Black Island channels (Fig. 1), which in turn could have affected water levels in the south basin. An abrupt excursion of about 40–55 mm is present for the early 1980s in the time series of gauge-pair differences between the south and north basins (presented later). Another explanation for this excursion might be human error. The magnitude of the excursion is similar to the usual 43 mm adjustment required to convert most Lake Winnipeg gauges to the Lake Winnipeg datum. However, the Water Survey of Canada is not aware of any erroneous adjustments (Jim Way, 1996, personal commun.), and none could be found in a cross-check of the published book values with the HYDAT database. Other potential error sources include individual gauge-site characteristics such as maintenance history, site reconfigurations, and bench-mark stability, all of which are difficult to correct without detailed knowledge of each gauge's history.

Lake Winnipeg Gauges

Eight gauges are currently monitored on Lake Winnipeg by the Water Survey of Canada (Fig. 1). Three other discontinued gauges (Anama Bay, Sans Souci, and Traverse Bay) were not considered because of their extremely short records. The current gauges are referenced to the Lake Winnipeg datum that was implemented in 1986. The Lake Winnipeg datum master reference point is the 1960 Geodetic Survey of Canada bench mark 78M079 at Berens River.

Of these eight gauges, five were suitable for analysis without any adjustments, and a sixth (Victoria Beach) required an adjustment. The George Island gauge was not used because its 12 yr recording history (through 1995) is too short for this type of analysis. Only gauges with at least 25 yr histories, and preferably more than 40 yr (Coordinating Committee, 1977), were included for analysis. This requirement allows the slow process of postglacial rebound enough time to accumulate a statistically significant signature into the gauge records. The Matheson Island gauge was also rejected because it was relocated from Matheson Island to the mainland in 1962 and has been operated on only a seasonal basis since 1979. This is not a significant loss because the analysis includes the more complete record of the nearby (12 km south) Pine Dock gauge, which has been operated on a continuous basis since 1959. Two entries in the Victoria Beach bench-mark history state that bench mark 703-C was damaged in 1961. In October 1970, the 1960 Geodetic Survey of Canada datum at Victoria Beach was adjusted up by 91.4 mm from the 1962 level. However, the time series of any pair of gauges involving a difference with Victoria Beach exhibited a pronounced offset of about 38 mm for the time interval 1961–1969 relative to the 1970–1995 linear trend line. We attribute this offset to the damaged bench mark. On this basis, we adjusted the differenced time series of VicPin, VicBer, and VicMis (we are using the three-letter gauge notation introduced in Fig. 1, e.g., VicPin means the Victoria Beach time series minus the Pine Dock time series) down by 38 mm for the years 1961 through 1969. The GimVic pair was excluded from the analysis because of its short baseline length (30 km). Finally, the gauge at Winnipeg Beach was discontinued in July 1966 and replaced by a gauge at Gimli in August 1966. Winnipeg Beach data were retained and combined with the Gimli record to create an 83 yr history. Although there is no overlap period from which to positively link the two records, the Water Survey of Canada converted the Gimli gauge to the Lake Winnipeg datum by subtracting 43 mm. We applied the same 43 mm subtraction to the Winnipeg Beach part of the record in order to align it with Gimli. The combined history yields an 82 yr record when compared with the similarly long record at Berens River.

Lake Manitoba Gauges

Six gauges are available on Lake Manitoba (Fig. 1). All gauges are referenced to the Geodetic Survey of Canada Datum that is adjusted locally at irregular intervals. A cross-check of the published lake levels with those of the HYDAT database confirmed that all adjustments indicated in the bench-mark histories were performed and that each gauge is referenced to the same datum for its entire period of record. The Narrows gauge has a volatile 1960–1966 record that might be related to construction of a causeway in the early 1960s to support Manitoba Hydro power-line towers. Later, in 1966, the causeway was extended to support the highway bridge across The Narrows. As a result of these two projects, the width of The Narrows was reduced by about 75% (Jim Way, 1993, personal commun.). We do not consider the 1960–1966 data at The Narrows an accurate measure of lake level and exclude those years from the analysis. The Delta Beach gauge was discontinued in 1969 and replaced by the Westbourne gauge, for which data collection began in 1964. The 5 yr overlap showed that the two gauges could be joined to form an 81 yr history that when differenced with the Steep Rock gauge yielded a 72 yr time series. The Meadow Portage gauge was relocated 16 km south to Toutes Aides in 1969. These two gauges were combined to obtain a 41 yr history. No adjustments were required. The Toutes Aides gauge was discontinued by the Water Survey of Canada in 1994, leaving the entire northwestern area of Lake Manitoba ungauged.

Lake Winnipegosis Gauges

Three gauges are available on Lake Winnipegosis. The discontinued gauge near Meadow Portage has an extremely spotty record. It was maintained on a seasonal basis with readings taken only once or twice per week, mainly during the summer months, and thus is not suitable for analysis. The Winnipegosis and Dawson Bay gauges both have good records and are referenced to the locally adjusted Geodetic Survey of Canada Datum. The crosscheck of published book values with those of the HYDAT database confirmed that all adjustments in the bench-mark histories had been performed and that each gauge is referenced to the same datum for its entire period of record. The Dawson Bay gauge was discontinued by the Water Survey of Canada in 1994, leaving the Winnipegosis gauge as the single point from where the level of Lake Winnipegosis is measured. The differenced time series of the north-northwest–oriented WgsDaw gauge pair represents the highest rate of differential rebound (4.5 mm·yr^{-1}) between any pair of gauges examined in this study (Fig. 5). This rate is consistent with the uptilt direction's being more due north and with greater tilts observed for its paleoshorelines as reported in Tackman et al. (1998). Taken together, both the shoreline and lake-gauge data indicate that tilting in the Lake Winnipegosis basin is directed along a radius closer to due north and greater in magnitude than that of its immediate neighbors. Whether this is simply a local anomaly or is a change in rebound pattern at the regional scale is uncertain. Continuation of long-term recording at Dawson Bay would be most useful in this regard.

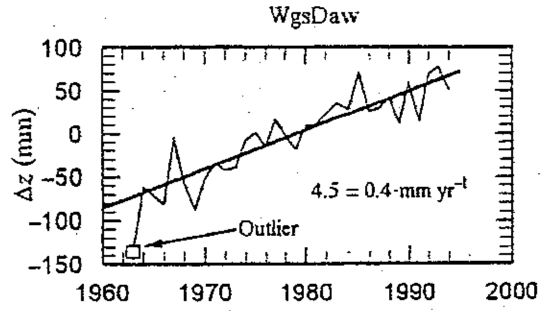


Figure 5. Time series of mean winter-season differences for the Winnipegosis and Dawson Bay pair (WgsDaw) of gauges on Lake Winnipegosis.

Lake of the Woods Gauges

Seven gauge stations are available for Lake of the Woods. The Warroad and Springsteel Point gauges are under control of the U.S. Geological Survey, whereas the other five are maintained and monitored by the Water Survey of Canada. All gauges are referenced to the Lake of the Woods Datum, which is the same as the U.S. Coast and Geodetic Survey Datum of 1912. The Springsteel Point and Cyclone Island gauges were both installed in the early 1980s and are not used because of their short records. The Keewatin and Kenora gauges are separated by a distance of only 4.5 km. Kenora has a record back to 1915 but was discontinued in 1959. Keewatin has a record from 1913 to the present. Both have essentially the same azimuth with the other gauges, and linear regression of the pairwise differenced series between either Keewatin or Kenora and the Warroad gauge (the only other gauge that overlaps in time with Kenora) showed they both have similar relative vertical-motion rates of about 1 mm·yr^{-1}. To avoid redundancy, the longer record of the Keewatin gauge was retained for analysis over the discontinued Kenora gauge. The ClrKee pair is also excluded from the analysis because of a short baseline length (20 km).

In 1978 the relationship between the Warroad gauge and the others changed. In the first few months of the year, the Warroad gauge was tracking with the Clearwater and Hanson Bay gauges under the cover of ice. After the volatile summer season, associated with seiching and wind setup, the Warroad gauge remained at an elevation of 322.893 m for six consecutive days (Nov 21–26), while the other gauges continued to record the normal seasonal decline in lake level. This resulted in an apparent offset of 33 mm that has persisted through to the present. Clearly, the gauge needs to be adjusted. Both Canadian and U.S. officials are aware of the problem but do not know the cause.

At first it is tempting to simply apply a 33 mm correction. However, examination of the December daily averages for the 3 yr prior to 1978 and for the succeeding 3 yr shows that the differences between Warroad and the other gauges vary from year to year. For the years 1975 through 1977, Warroad had December daily average levels that were 28.4, 10.7, and 7.2 mm higher than the nearest gauge at Hanson Bay. These values suggest a correction of less than 33 mm. For December 1978, Warroad was 45.3 mm higher than Hanson Bay, and for the years 1979 through 1981, the Warroad December daily average remained consistently higher than the pre-1978 levels (38.3, 37.4, and 46.0 mm) relative to Hanson Bay. During the six days in question, Hanson Bay fell 28 mm while the two more distant gauges in the north at Clearwater Bay and Keewatin fell 43 and 45 mm, respectively. This pattern suggests that the correction should be closer to 28 mm. Accordingly, we reduced all elevations at the Warroad gauge by 28 mm from November 27, 1978, forward.

COMPUTATIONAL METHODS

Trend-Surface Model

We assume that the lake-surface elevation at each gauge can be separated into three components: a lake signature (L_k) that is common to all gauges on a particular lake, a postglacial rebound signature that is unique to each gauge and varies slowly over space and time, and random noise. The general mathematical form of the model is

$$z(\Delta x_i, \Delta y_i, \Delta t) = A_i + L_k(\Delta t) + F_1(\Delta x_i, \Delta y_i)^* \Delta t + \varepsilon_i(\Delta t), \tag{1}$$

where z is the lake-surface elevation measured at gauge i at time Δt, Δx_i and Δy_i are the relative Cartesian coordinates of gauge i (UTM zone 14 easting and northing are used in this analysis), A is the local fixed elevation bias at gauge i, L_k is the lake signature common (to a first order) to all gauges on lake k, and the subscript i is the random noise at gauge i. F_1 is the assumed spatial form of the regional rebound signature, and for this analysis, both linear and quadratic forms are explored. Although more complex mathematical forms could be used, such as a cubic or spatial harmonic series, the noise level inherent in the lake-gauge data does not justify the assumption that detailed spatial patterns can be extracted. The linear and quadratic forms meet the objectives of capturing the dominant trend and proved to be less influenced by edge effects in the mapped patterns (presented later) than were the more complex mathematical forms.

F_1 depends explicitly on the gauge location relative to a central, but arbitrary, reference point

$$\Delta x_i = x_i - x_0, \quad \Delta y_i = y_i - y_0, \tag{2}$$

where x_i and y_i are the Cartesian coordinates for gauge i, x_0 and y_0 are coordinates for the central reference point that are taken as the spatial mean computed from the gauge station locations (UTM coordinates x_0 = 601 247 m east and y_0 = 5 669 621 m north). Similarly, time is also referenced to a centrally situated, but arbitrary, epoch

$$\Delta t = t - 1950.0, \tag{3}$$

which approximates the midpoint of the data and is conveniently the epoch of the ^{14}C time scale.

The explicit forms of the linear and quadratic components, respectively, of F are

$$F_{1,1}(\Delta x_i,\Delta y_i) = S_x^1 * (\Delta x_i) + S_y^1 * (\Delta y_i) \tag{4}$$
$$F_{1,2}(\Delta x_i,\Delta y_i) = S_{xx}^1 * (\Delta x_i)^2 + S_{xy}^1 * (\Delta x_i) * (\Delta y_i) + S_{yy}^1 * (\Delta y_i)^2,$$

where S_x, S_y, S_{xx}, S_{xy}, and S_{yy} are the rebound-related parameters that must be estimated and interpreted.

A model in which the time dependence includes both linear and quadratic terms was also explored:

$$z(\Delta x_i,\Delta y_i,\Delta t_m) = A_i + L_k(\Delta t_m) + F_1(\Delta x_i,\Delta y_i) * \Delta t_m + F_2(\Delta x_i,\Delta y_i) * (\Delta t_m)^2 + \varepsilon_i(\Delta t_m), \tag{5}$$

where the functions F_1 and F_2 have the same functional form (i.e., either both linear or both quadratic). The quadratic time term allows us to test for variable rates of rebound. We presume that the steady, or time-linear, part of the rebound signature is postglacial rebound, whereas any change in rebound rate is more likely due to other influences; one likely source would be variation in the lake loads themselves.

Before proceeding, it is convenient to introduce a notational scheme to refer concisely to the different variations of the model. A two-letter designation consisting of L for linear and Q for quadratic suffices. The first letter refers to the space order and the second to the time order. For example, an LQ designation refers to a model that is linear in space and quadratic in time. A QL designation refers to a model that is quadratic in space and linear in time, etc. To further simplify notation, the relative site location is represented by a vector **P**

$$\mathbf{P}_i = (\Delta x_i, \Delta y_i). \tag{6}$$

Forming a difference (Δz) between all combinations of gauge pairs on a lake, for each Δt, will remove most of the common lake signature, L_k, and reveal the trend in the changing water level between pairs of gauges that we attribute to postglacial rebound.

$$\Delta z(\mathbf{P}_i,\mathbf{P}_j,\Delta t_m) = z(\mathbf{P}_i,\Delta t_m) - z(\mathbf{P}_j,\Delta t_m) \tag{7}$$
$$= \Delta A + \Delta F_1(\mathbf{P}_i,\mathbf{P}_j) * \Delta t_m + \Delta F_2(\mathbf{P}_i,\mathbf{P}_j) * (\Delta t_m)^2.$$

The mean winter-season differences are actually computed directly from the difference of daily averages for the months of December (of the previous year), January, February, and March. Daily averages for all pairwise combinations on each lake were extracted from the HYDAT database, differenced, and averaged according to

$$\Delta\bar{z}_{t,i,j} = \frac{1}{q}\sum_{a}^{4}\sum_{b}^{days}\left(z_{1ab} - z_{2ab}\right), \tag{8}$$

where q = number of valid-day pairs for the season (number of overlap days), a = month number (December of previous year = 1, January = 2, February = 3, and March = 4), b = number of *days* in month a, the subscript t indicates the year, the subscript i = the first gauge, and the subscript j = the second gauge.

Equation 8 has the advantage of constructing a mean difference for each season—in this case, the winter season—based on the exact time intersection of gauge pairs. If data are absent for one or both days, a daily difference is not formed. Thus, equation 8 addresses the issue of time intersection—allowing the formation of mean winter-season differences when data are missing. This approach has an advantage over the usual practice of forming a difference from winter-season mean values. In that method, monthly means are formed from daily means, and winter-season means are formed from monthly means. Then, the winter-season difference is formed by subtracting winter-season means of pairs of gauges. If the requirement for no missing days is relaxed, as it is for this analysis, differencing winter-season mean lake levels could result in differences between gauge pairs that do not exactly intersect in time. For example, if 10 missing days are allowed and the first 10 days of gauge 1 are missing and the last 10 days of gauge 2 are missing, a difference between winter-season means is actually a difference between the mean of the last 111 days (assuming 121 days in a winter season) of the first gauge and the mean of the first 111 days of the second gauge. If lake elevation is significantly different between the beginning and end of the winter season, unnecessary error can be introduced. Use of equation 8 guarantees time intersection and, in this example, would form a winter-season mean based on the middle 101 common days.

For the present analysis, we have adopted a minimum standard of 10 days of overlapped record for each of at least three of the four months before a winter-season mean difference can be formed. This apparently permissive standard is not possible for the summer-season data, which are noisier. The standard is based on a statistical analysis whereby days were randomly deleted from seasons with no missing days and pairwise single-season mean differences were formed according to equation 8. Fifty runs for each gauge pair for each year on the three Manitoba lakes were analyzed. The threshold for determining the maximum allowable number of missing days is that the mean difference formed with missing days had to be within 20 mm of the mean difference formed with no missing days. The 20 mm constraint represents a reasonable uncertainty estimate of the overall gauge measurement accuracy according to Table 1. It was found that 90% of the mean winter-season differences formed at this minimum standard were within the prescribed 20 mm of the mean winter-season difference formed with no missing days.

After forming the time series of mean winter-season differences between all gauge pairs, the linear-regression line for each series was computed, and the residuals relative to the regression line were examined for influential outliers and any remaining correlation with lake level or river input. This is an important step in the analysis because we interpret the slope of the linear-regression line through the time series of these mean winter-season differences as the differential rate of crustal motion between the gauge pairs. As previously mentioned, no significant correlation was found between any of the residual time series and river input; however, the residual time series from six gauge pairs still correlated with lake level. On Lake Manitoba, the residual time series from the gauge pairs DelStp, WesNar, and WesTou had correlations of –0.44, –0.37, and –0.40, respectively, and on Lake of the Woods, the pairs WarClr, HanKee, and ClrKee had correlations of 0.45, 0.48, and 0.48, respectively. To remove the remaining lake signature, another linear-regression line was computed for each of these six gauge pairs by using the residuals from the previous regression as the dependent variable and lake level as the independent variable. Then, the predicted residual, representing the remaining correlation with lake level, was subtracted from the mean winter-season difference. This correction reduced the slope of the linear-regression line through the time series of mean winter-season differences for DelStp from 1.037 to 1.002 mm·yr^{-1} but did not affect the slopes of the other five gauge pairs. Removal of the remaining lake signature for these gauge pairs improved r^2 values by 5%–10% and reduced standard errors of the slope by 10%–15%. After the correction, none of the residual time series had an absolute correlation value greater than 0.20 with either lake level or local river input.

Next, influential outliers were eliminated, which were defined as those residuals with an absolute standardized value greater than 2.0 and an absolute DFBETAS value greater than $2/\sqrt{n}$, according to a rule given by Neter et al. (1990). DFBETAS is a measure of the influence that a single data point has on each of the regression coefficients (Neter et al., 1990). Thus, outliers are those residuals that are at a significant vertical distance from the regression line and at a location (usually near the ends of the time axis) where they disproportionately influence the slope. Out of ~1000 data points, 33 influential outliers were eliminated.

Next, the rebound-related parameters—S_x, S_y, S_{xx}, S_{xy}, and S_{yy}—can be solved for. In the time-linear models, the slope of the regression line (the differential rate of motion between gauge pairs), fit to each time series of pairwise differences computed according to equation 8, is equated to the time derivative of the constant and the time-linear term [$\Delta A + \Delta F_1(\mathbf{P}_i,\mathbf{P}_j) * \Delta t_m$] from the right side of equation 7:

$$\frac{d\Delta z}{d\Delta t} = \text{slope of linear-regression line} = \Delta F_1\left(\mathbf{P}_i, \mathbf{P}_j\right). \qquad (9a)$$

In the time-quadratic models, the time derivative of a quadratic trend fit ($b_0 + b_1\Delta t + b_2\Delta t^2$) to each series of differences computed from equation 8 is equated to the time derivative of the entire right side of equation 7:

$$\frac{d\Delta z}{d\Delta t} = b_1 + 2 * b_2 \Delta t = \Delta F_1\left(\mathbf{P}_i, \mathbf{P}_j\right) + 2 * \Delta F_2\left(\mathbf{P}_i, \mathbf{P}_j\right) * \Delta t. \qquad (9b)$$

ΔA is conveniently eliminated when computing the derivatives of equation 7 because it is a constant. Then, a least-squares solution is computed for the rebound-related parameters by forming a vector of rates ($d\Delta z/d\Delta t$) from the left side of equation 9 and a corresponding matrix of spatial differences from the right side of equation 9.

Having solved for the rebound-related parameters in ΔF_1 and ΔF_2, the spatial pattern of differential rebound can be recovered by differentiating equation 1 or 5 with respect to time. The time derivative of equation 1 is simply

$$-\frac{dz\left(\mathbf{P}_i\right)}{dt} = -F_1\left(\mathbf{P}_i\right), \qquad (10)$$

and that for equation 5 is

$$-\frac{dz\left(\mathbf{P}_i, \Delta t_m\right)}{dt} = -F_1\left(\mathbf{P}_i\right) - 2 * F_2\left(\mathbf{P}_i, \Delta t_m\right) * \left(\Delta t_m\right), \qquad (11)$$

where the negative sign reflects the fact that an apparent rise in lake level is actually a decrease in the elevation of the land surface to which the gauge is rigidly attached.

Computation of Rebound Rates From ICE-3G and ICE-4G Model Loads

ICE-3G (Tushingham and Peltier, 1991) and ICE-4G (Peltier, 1994) are global models that describe the thickness and extent of ice sheets from the last glacial maximum (LGM) at ca. 18 000 ^{14}C yr B.P. to the present. The ice models are used as input to a surface-loading calculation to determine crustal displacement, change to the Earth's gravitational potential, and sea-level change, assuming an Earth model with a viscous, or more commonly Maxwell viscoelastic, mantle. Constraints on Earth rheology (i.e., mantle viscosity and lithosphere thickness) are determined primarily through comparison with observed relative sea-level changes.

The surface-loading calculations performed here feature a spherically symmetric Earth model with fluid outer core and compressible Maxwell viscoelastic mantle and elastic lithosphere. The surface response, composed of a spectrum of decay times and associated amplitudes, or residues, is found by using the normal mode method (Peltier, 1985). The Maxwell viscoelastic Earth model has a lower-mantle (below 670 km depth) viscosity of 2×10^{21} Pa·s, upper-mantle viscosity of 10^{21} Pa·s, and a 120-km-thick lithosphere, consistent with ICE-3G (Tushingham and Peltier, 1991). Density and elastic parameters vary according to the 1066B seismic Earth model of Gilbert and Dziewonski (1975). Surface displacements and changes to the gravitational potential, from which lake tilt rates can be determined, were calculated by using the methods described by James and Ivins (1998).

The ICE-3G tilt rates were directly calculated by using the ice history provided by Tushingham and Peltier (1992). The calculations assume initial isostatic equilibrium at 18 000 ^{14}C yr B.P. and incorporate gravitationally self-consistent ocean loading without water in-filling for Hudson Bay. Although the in-filling of Hudson Bay with ocean water in postglacial times is important for determining uplift rates in the center of Hudson Bay, it has less effect on land.

In contrast, ice thicknesses are not available for the ICE-4G model, and no surface-loading calculation was attempted. Instead, uplift rates for ICE-4G were obtained from digital files of topography computed by Peltier (1993). These files give the surface elevation (topography) relative to sea level, at 1000 yr intervals since 21 000 calendar yr B.P., resulting from using the ICE-4G chronology as the input for a Maxwell viscoelastic surface-loading calculation. Values are given on a 1° × 1° global grid. At each grid point, the change in topography was determined between the present and 1000 calendar yr B.P. and then from 1000 to 2000 calendar yr B.P. From these values, present-day vertical crustal motion was determined.

Rates derived from the ICE-4G load are considerably less than those based on ICE-3G. Over Lake Winnipeg, the ICE-4G–derived rates are, on average, 43% less; over Lake Manitoba, they are 61% less; over Lake Winnipegosis, they are 53% less; and over Lake of the Woods, they are 32% less. For the most part, this effect reflects the general trend in current thought toward a thinner multidomed ice sheet, but also represents a measure of the uncertainty associated with the task of trying to reconstruct continental-scale ice sheets. The results presented in the next section can be used as geodynamic constraints for constructing and refining future Laurentide ice models.

RESULTS AND DISCUSSION

The results of the least-squares solutions are summarized in Table 2. The LL and QQ fits (the least and most complex of the regional trend-surface models, respectively) are plotted in Figure 6 along with the simple linear-

TABLE 2. LEAST-SQUARES SOLUTIONS FOR REBOUND-RELATED PARAMETER VALUES FOR FOUR REGIONAL TREND-SURFACE MODELS

Parameter	LL	QL	LQ		QQ	
	(yr^{-1})	(yr^{-1})	(yr^{-1})	(yr^{-2})	(yr^{-1})	(yr^{-2})
S_x	*(7.3 ± 2.3) × 10^{-9}	(7.2 ± 4.1) × 10^{-9}	*(5.1 ± 0.7) × 10^{-9}	*(6.4 ± 3.0) × 10^{-11}	*(3.3 ± 0.9) × 10^{-9}	(5.0 ± 3.0) × 10^{-11}
S_y	*(7.9 ± 2.1) × 10^{-9}	*(6.9 ± 3.0) × 10^{-9}	*(4.2 ± 0.4) × 10^{-9}	(–1.3 ± 1.1) × 10^{-11}	*(3.4 ± 0.5) × 10^{-9}	(–1.5 ± 1.0) × 10^{-11}
S_{xx}		(–4.3 ± 25) × 10^{-15}			(0.05 ± 4.3) × 10^{-15}	(5.8 ± 6.2) × 10^{-17}
S_{xy}		(5.1 ± 28) × 10^{-15}			(5.2 ± 4.7) × 10^{-15}	*(45 ± 7.3) × 10^{-17}
S_{yy}		(6.0 ± 9.3) × 10^{-15}			(0.8 ± 2.0) × 10^{-15}	*(17 ± 3.2) × 10^{-17}
r^2	0.25*	0.25	0.25		0.25	

Note: Values are given relative to the motion of the land surface. Values marked with an asterisk are statistically significant ($p < 0.05$).

regression trend line for each pair of time series of mean winter-season differences. Each of the four regional trend-surface models accounts for 25% of the aggregate variance present in the time series of mean winter-season differences as measured by r^2. This percentage is somewhat less than that for the simple linear-regression fits, which account for 39% of the aggregate variance, and the quadratic fits, which account for 42%. If one can accept that the linear and quadratic fits to the time series of mean winter-season differences represent the postglacial rebound signature between each gauge pair, then the regional models capture most of that signature. The results indicate that adding parameters over a simple planar fit (the LL model) does not statistically improve the fits. Therefore, nearly all the postglacial rebound signature that can be identified by using these regional trend-surface

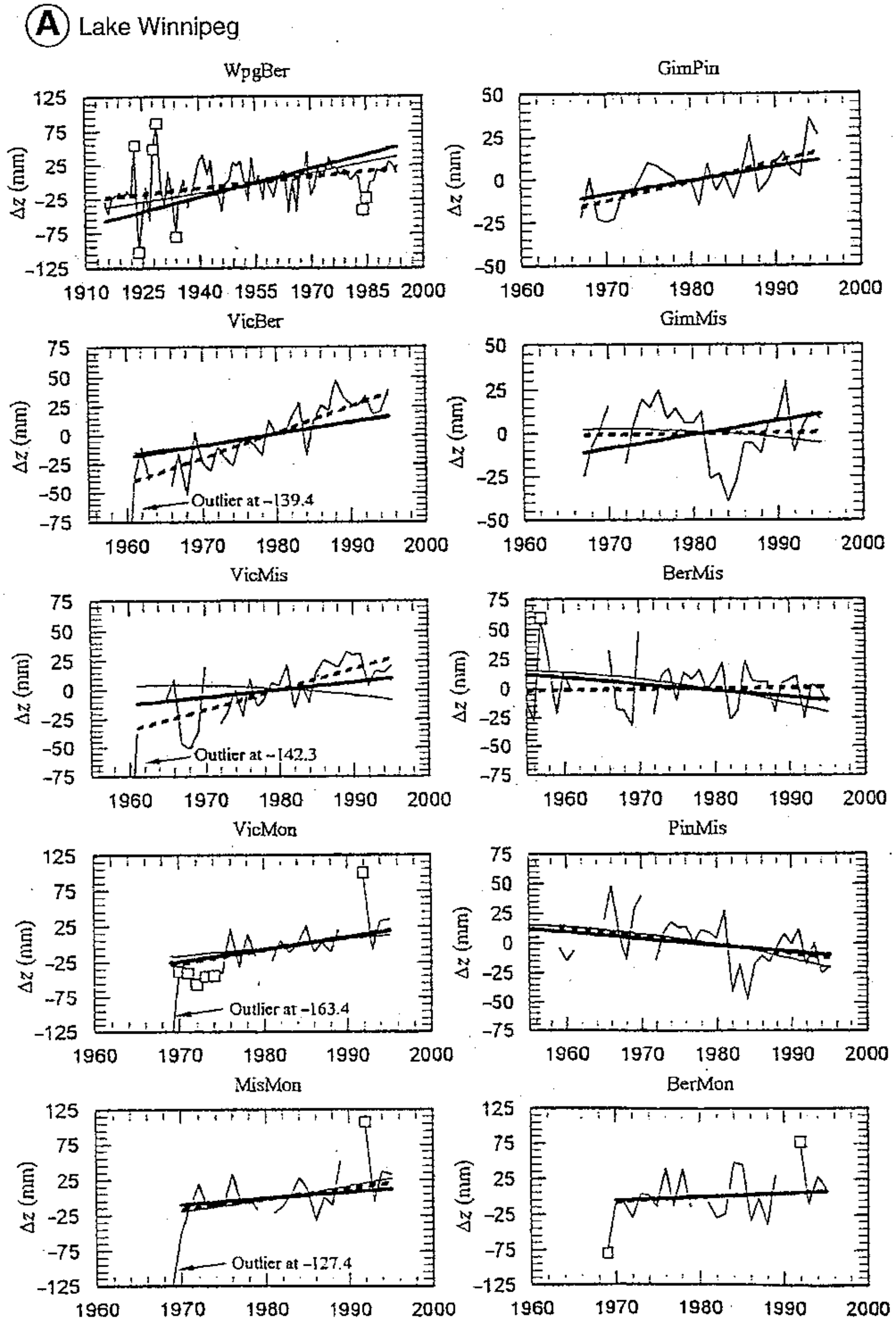
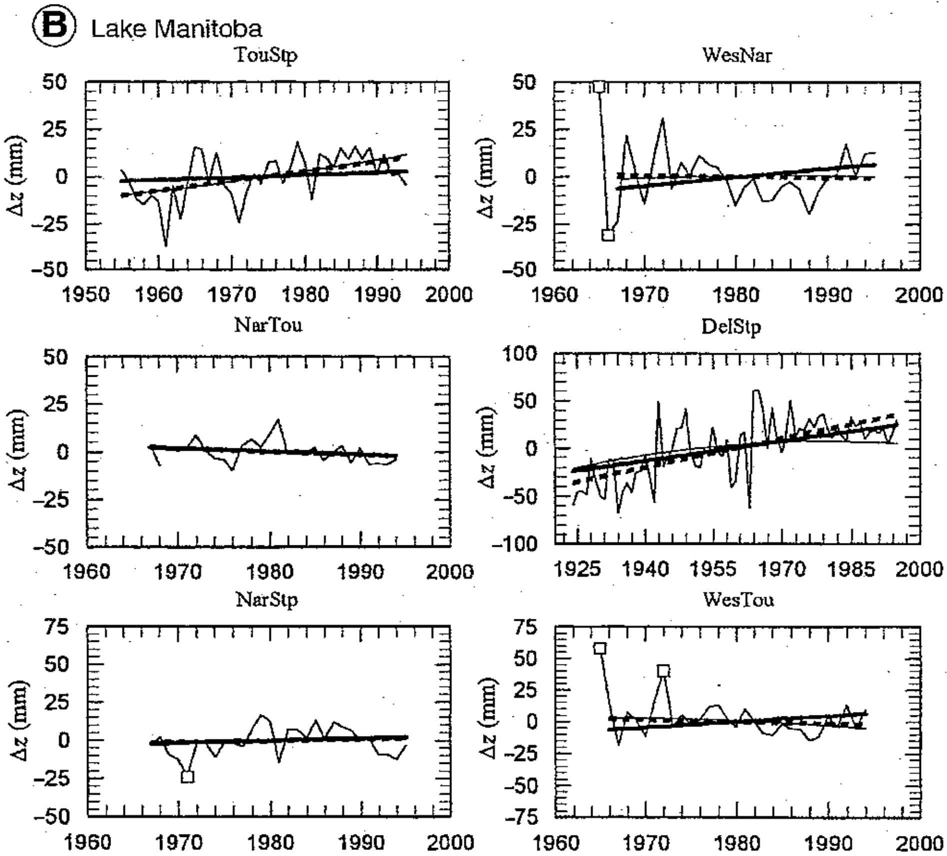
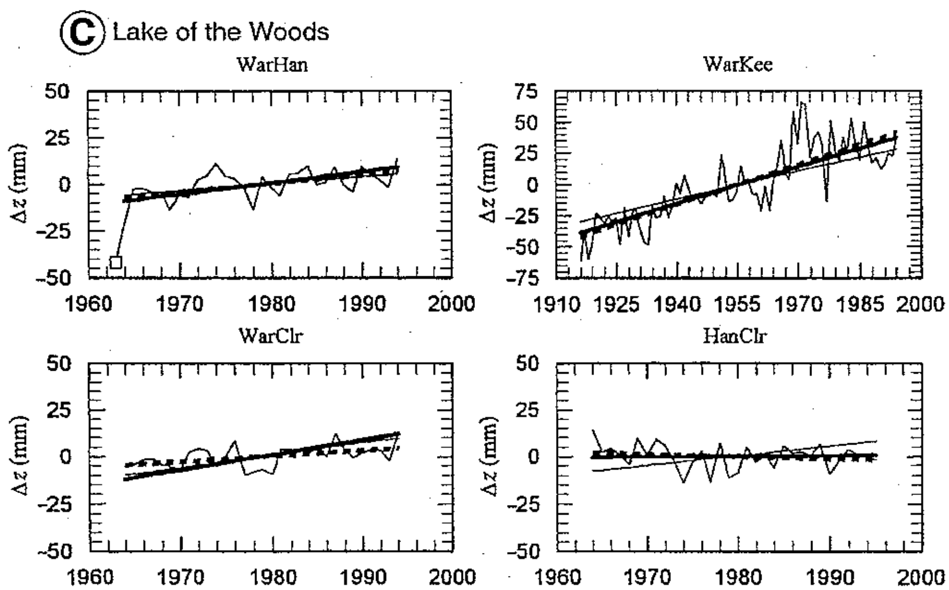


Figure 6. Time series for mean winter-season differences between gauge pairs (fluctuating line) and model fits for (A) Lake Winnipeg, (B) Lake Manitoba, and (C) Lake of the Woods. The heavy dashed line is the simple linear-regression line, the heavy solid line is the LL model fit, and the light solid line is the QQ model fit. The open squares are influential outliers not used to compute any of the trend-surface models or pairwise linear-regression lines.





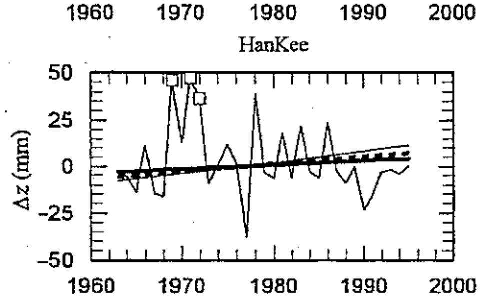


Figure 6. ***(Continued).***

models is captured in the two time- and space-linear parameters of the LL model, S_x and S_y. According to the LL model, postglacial tilting is up to the northeast with a bearing of N(42.7 ± 11.8)°E at a rate of (10.7 ± 2.2) × 10^{-9} rad·yr^{-1}. Relative to the lake-surface elevations from equations 1 and 5, the regional tilt signature extracted from the data by any of the regional trend-surface models accounts for less than 0.5% of the total lake-level fluctuation as measured by r^2. This small percentage is understandable when one considers that the dominant signature in the lake-surface elevation histories is the climate-induced lake excursion, L_k (Fig. 2), which can vary from year to year by as much as 2 m. In contrast, the regional tilt signature is measured in millimeters to centimeters over decades.

The WgsDaw and VicPin pairs were excluded from the final analysis because they are influential outliers according to the criteria previously described. The differential rates of rebound indicated from the slopes of a simple linear regression for these two pairs (4.5 and 3.0 mm·yr^{-1}, respectively) are anomalously high and inconsistent with the remaining network of gauges. Inclusion of these two pairs resulted in regional trend-surface model r^2 values between 4% and 7%, whereas the exclusion of these two pairs increased r^2 values to 25%. If the high rates for these two pairs are real, then a more detailed model is required to capture the rapid changes in the spatial pattern. Examination of the other three time series with the Victoria Beach gauge in Figure 6 shows that the regional models underpredict both the VicBer and VicMis pairs, but yield a good fit to the VicMon pair. These observations suggest that there might be a local subsidence problem at the Victoria Beach station, which artificially increases relative rates of rebound between it and other gauges to the north. A similar situation could exist at either (or both) the Winnipegosis or Dawson Bay gauges. With only two gauges on Lake Winnipegosis, it is impossible to determine which one might be in error, if at all. Overall, visual inspection of the plots in Figure 6 indicates that the regional models fit the time series of mean winter-season differences reasonably well, including sufficiently good fits to the longer time series of WpgBer, DelStp, and WarKee.

Another important result is that because the time-quadratic models (LQ and QQ) did not significantly improve the fits over the time-linear LL model, there is little to no time variation present in the postglacial rebound signature. Thus, it does not appear that there is any change in the rate of rebound due to other sources such as variation in the lake loads themselves. Rather, the postglacial rebound signature appears to be quite steady and is best expressed as a simple linear trend (Fig. 6).

Comparison with Rates Derived from ICE-3G and ICE-4G Models

In Figure 7, the planar pattern of postglacial rebound according to the LL model is overlain onto the patterns of rebound that we computed from the ICE-3G and ICE-4G load models. All three patterns indicate that tilt is up to the northeast, which is consistent with other previously published patterns of postglacial rebound for this region (Teller and Thorleifson, 1983; Peltier, 1986; Andrews and Peltier, 1989; Sjoberg et al., 1990). The rebound patterns derived from the two ice models and the lake-gauge data (according to the LL model) suggest that the tilt pattern over the region is rather smooth and unidirectional. This result is in contrast to what was reported in Tackman et al. (1998), where post–Lake Agassiz paleoshorelines indicated a deformation pattern of tilting up to the east-northeast over Dauphin Lake and a change to a faster rate of tilting up to the east-northeast over Lake Winnipegosis. If the determination of the more rapid rate of tilting up to the north-northeast over Lake Winnipegosis, indicated by the WgsDaw gauge pair, could be augmented and verified with additional gauges, or by other means, then the regional pattern measured by lake gauges in southern Manitoba should show more variation than is suggested by the simple planar pattern of the LL regional model.

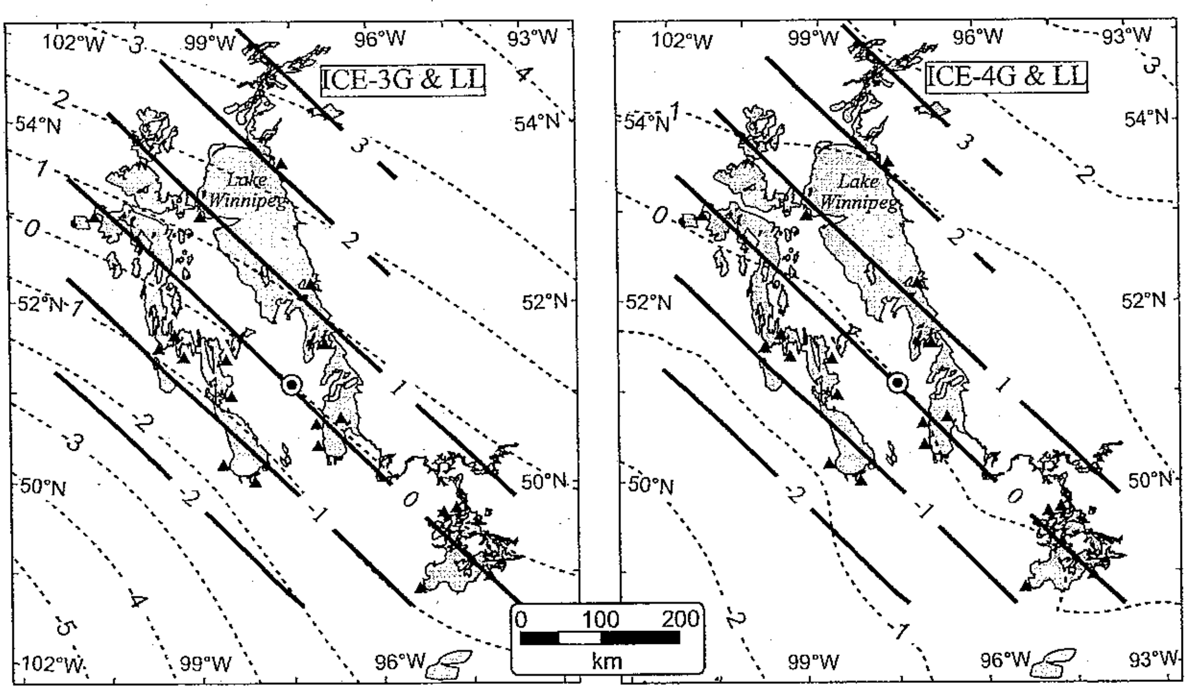


Figure 7. Spatial pattern of tilting computed from the LL regional trend-surface model (solid lines) overlaid on the pattern computed from the ICE-3G and ICE-4G loading models (dashed lines). Rate-contour spacing is 1 mm·yr^{-1}. Triangles are the gauge stations used to compute the regional model pattern (see Fig. 1 for gauge names). All patterns are referenced to the central reference point (lat 51.1710°N, long 97.5518°W) indicated by the bull's-eye symbol.

The rate-contour spacing in Figure 7 reveals that the rate of tilting observed from the lake gauges is in close agreement with those computed from ICE-3G, but is almost twice as fast as the pattern computed from ICE-4G. Neither the ICE-3G nor ICE-4G patterns, or regional trend-surface models, capture the tilting that is apparently occurring over Lake Winnipegosis; this tilting is faster, and is oriented more to the north. Table 3 is a summary of differential rates of rebound between gauge pairs computed from a simple linear-regression line, our regional trend-surface models, the ICE-3G and ICE-4G loading models, and those given in Carrera et al. (1991). The latter are included as an additional comparison to the linear-regression rates computed in this study. When uncertainties are taken into account, there is general agreement between this study and that of Carrera et al. The differences that do exist are likely due to different data-handling methods and time intervals used in the two studies. First, there are corrections made in this study that were not made in Carrera et al. (the Victoria Beach bench-mark adjustment; the 28 mm adjustment to post–November 1978 values for Warroad; and the exclusion of pre-1966 data for The Narrows). Second, in this study, some gauges were combined that were not in Carrera et al. (Winnipeg Beach with Gimli, Westbourne with Delta Beach, Meadow Portage with Toutes Aides). Third, the study of Carrera et al. included data through 1990, whereas this study extends through 1995. Fourth, Carrera et al. used a summer-season mean, whereas we used a winter-season mean. One notable exception to the general pattern of agreement between this study and that of Carrera et al. is the significantly higher rate for WgsDaw determined by Carrera et al. relative to our results. We have no definite explanation for this difference, but suggest as a possibility that Carrera et al. may have inadvertently used the data prior to October 1963 for the Dawson Bay gauge that the Water Survey of Canada indicates is unreliable and should not be used.

In an attempt to summarize the agreement, or disagreement, between rates computed from the lake gauges and those computed from ICE-3G and ICE-4G loads, we simply regressed the rates given in Table 3 for the simple linear regression and the LL model onto those we computed from the ice models (Fig. 8). To understand the rationale for this test, one should consider the case where total agreement exists. Both the slope of the regression line and correlation coefficient would be one, and the standard error of the regression-line slope would be small. In other words, the ice model would exactly predict, or agree with, the rates computed from the lake gauges. When the regression-line slope is greater than one, rates observed from the lake gauges exceed ice-model predictions on an aggregate basis (the ice-model underpredicts), whereas when the slope from the regression line is less than one, rates observed from the lake gauges are less than the ice-model predictions on an aggregate basis (the ice model overpredicts).

In the top two plots of Figure 8, the rates derived directly from the slopes of simple linear-regression lines through the mean winter-season differences agree better with those computed from ICE-3G than with those from ICE-4G. In the comparison with ICE-3G, the regression-line slope is closer to one, has a smaller uncertainty, and a higher correlation than the plot with ICE-4G. In the lower two plots, the agreement between the LL model and ICE-3G is again superior to that with ICE-4G. The slope is near one, the uncertainty is small, and the correlation high. These plots quantitatively confirm the conclusion, previously made from visual inspection of Figure 7, that the pattern computed from an ICE-3G load is in closer agreement with the pattern observed from the lake gauges than is the pattern computed from ICE-4G.

In spite of the fact that these two ice models are derived mainly from the inversion of relative-sea-level histories obtained from marine coastal sites, they do a reasonably good job in the continental-interior region of southern Manitoba, although ICE-4G appears to be considerably too slow in this continental-interior region. The fact that the observed pattern of rebound derived from the lake gauges agrees better with a rebound pattern computed from a single-domed ICE-3G load, compared to a pattern computed from the multidomed ICE-4G load, should not be interpreted to mean that the Laurentide ice sheet was thick and single-domed. Rather, it may mean that the ice domes predicted by ICE-4G in Ontario and/or western Quebec should be increased, or certain Earth structure assumptions modified, or both. In the next section, we offer more specific suggestions for how these models might be modified based on a region-wide pattern of uplift that includes the Great Lakes.

TABLE 3. DIFFERENTIAL RATES OF REBOUND BETWEEN PAIRS OF GAUGES

Gauge pair	Linear	LL	QL	LQ	QQ	ICE-3G	ICE-4G	Carrera
Lake Winnipeg								
GimPin	−1.143 ± 0.253	−0.803	−0.804	−0.613	−0.749	−0.928	−0.569	
GimMis	−0.055 ± 0.403	−0.928	−0.852	0.338	0.273	−1.525	−0.730	
VicBer	−2.232 ± 0.250	−1.023	−1.084	−0.461	−0.881	−1.194	−0.379	−1.59 ± 0.42
VicMis	−1.753 ± 0.319	−0.642	−0.604	0.538	0.356	−1.258	−0.467	
VicMon	−1.873 ± 0.769	−1.700	−1.905	−0.278	−1.141	−2.047	−0.860	
PinMis	0.735 ± 0.334	0.038	0.015	0.700	0.659	−0.597	−0.161	
BerMis	−0.064 ± 0.280	0.544	0.543	0.786	0.897	−0.064	−0.088	0.29 ± 0.33
BerMon	−0.592 ± 0.791	−0.514	−0.757	−0.059	−0.481	−0.853	−0.481	
MisMon	−1.528 ± 0.646	−0.895	−1.237	−1.156	−1.866	−0.789	−0.393	−0.46 ± 0.95
WpgBer	−0.567 ± 0.105	−1.406	−1.409	−0.927	−0.976	−1.551	−0.706	−0.50 ± 0.43
Lake Winnipegosis								
WgsDaw	−4.500 ± 0.425	−0.519	−0.380	−0.046	0.512	−1.234	−0.588	−6.21 ± 0.50
Lake Manitoba								
WesNar	0.045 ± 0.280	−0.459	−0.360	−0.322	−0.049	−0.690	−0.213	0.11 ± 0.56
WesTou	0.158 ± 0.215	−0.466	−0.244	−0.021	0.291	−0.780	−0.352	
NarStp	−0.094 ± 0.197	−0.142	−0.185	−0.265	−0.222	−0.366	−0.120	1.64 ± 0.24
NarTou	0.179 ± 0.143	0.157	0.179	0.162	0.117	−0.090	−0.139	
TouStp	−0.510 ± 0.150	−0.135	−0.302	−0.535	−0.545	−0.276	0.019	−0.97 ± 0.56
DelStp	−1.003 ± 0.135	−0.685	−0.512	−0.470	−0.367	−0.948	−0.261	−0.84 ± 0.34
Lake of the Woods								
WarHan	−0.440 ± 0.133	−0.600	−0.396	−0.893	−0.374	−0.476	−0.317	−1.88 ± 0.21
WarClr	−0.296 ± 0.103	−0.803	−0.586	−0.745	−0.642	−0.745	−0.265	
HanKee	−0.392 ± 0.404	−0.218	−0.247	−0.191	−0.585	−0.390	−0.095	
HanClr	0.124 ± 0.128	−0.040	−0.127	0.005	−0.512	−0.269	0.052	
WarKee	−1.077 ± 0.083	−0.981	−0.706	−0.830	−0.747	−1.163	−1.165	−0.93 ± 0.11

Notes: Data were determined according to the slope of a simple linear regression line (for column 1), computed from the four regional trend surface models (for columns 2–5), computed from ICE-3G and ICE-4G ice load models (for columns 6 and 7), and from simple linear regression taken from Carrera et al. (1991) (for column 8). All rates are in mm yr^{-1} and represent the motion of the land surface between gauge pairs. Negative (positive) rates indicate the land surface at the second gauge is rising (subsiding) relative to the first.

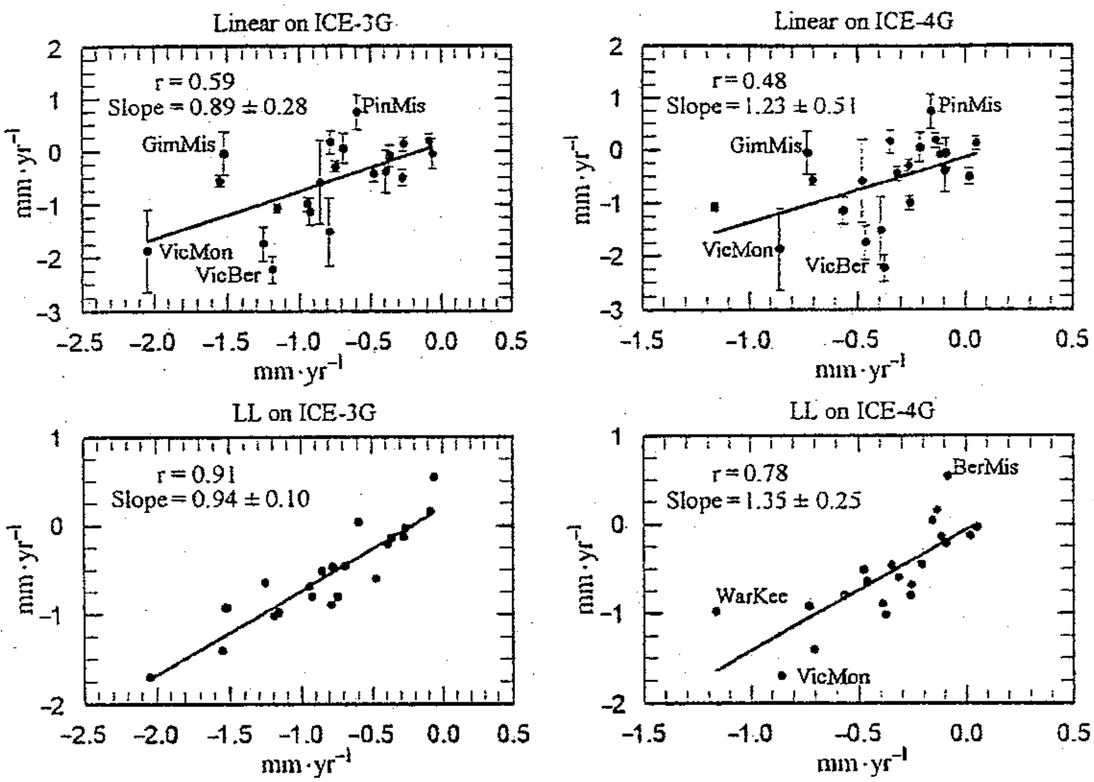


Figure 8. Diagnostic scatter plots for assessing the agreement (or disagreement) between rates predicted by various means. Top two plots: Rates predicted from simple linear regression through winter-season mean differences vs. rates computed from ICE-3G and ICE-4G load histories. Bottom two plots: Rates predicted from the LL model vs. rates computed from ICE-3G and ICE-4G load histories. All rates (both axes) are relative to the motion of the land surface. All the correlations and slopes are statistically significant ($p < 0.05$).

Extension of Pattern to the Great Lakes

In an effort to explore still larger-scale rebound patterns, we extended the LL pattern from the southern Manitoba–Lake of the Woods region to the pattern computed over the Great Lakes from lake gauges by Clark and Persoage (1970) (Fig. 9). We digitized the Great Lakes pattern from their Figure 2 and converted our rates to the same units (feet per century; 1 ft/century = 0.3048 m per 100 yr) used by Clark and Persoage. The rate contours were extended from both patterns by continuing the trend and then linking the extensions approximately halfway between the two regions. Although the extension and linking procedure must be viewed as somewhat subjective, we have strived to be faithful to the individual patterns and believe the linked pattern is a reasonable representation of the region-wide pattern.

On the basis of the rate-contour spacing in Figure 9, the present tilt rate is about the same for both regions, with the exception of the southern half of the lower peninsula of Michigan and much of Lake Erie, where it is less than half the rate seen in the rest of the pattern. Over the Great Lakes, tilt is up to the northeast in apparent response to major ice loading that, according to the ICE-4G model, existed over James Bay and western Quebec. The southern Manitoba region is apparently responding to James Bay ice and an ice dome that, according to the paleotopography of ICE-4G, existed over Churchill. In addition, the overall pattern also suggests a more local source region of isostatic deflection in western Ontario, possibly related to the lobate structure of the former Laurentide ice sheet. In particular, we are inclined to invoke an influence from the Des Moines lobe, but the Rainy and Superior lobes, among others, would also be involved (if we use the paleogeography and terminology of Dredge and Cowan, 1989, their Fig. 3.22). Another ice-sheet feature that might be related to the rebound pattern is a relatively small ice dome (~2600 m in height) present in ICE-4G, which was centered just north of Lake Nipigon. Overall, the spatial arrangement of several ice domes predicted in the paleotopography of the ICE-4G model is consistent with the region-wide contemporary tilt pattern of Figure 9. However, rates derived from the ICE-4G load model underestimate the observed pattern in Manitoba, as previously discussed. Any attempt to modify ICE-4G should retain a similar spatial arrangement of ice domes, but increase ice thickness for (1) the ice lobes that flowed through western Ontario and southern Manitoba or (2) the small ice dome north of Lake Nipigon—or both—in order to increase uplift rates to more closely match those observed in southern Manitoba and, if necessary, also for the Great Lakes. Earth structure assumptions may also need to be modified. In particular, no lateral variations in Earth structure were included in the derivation of ICE-3G and ICE-4G, though such variations certainly exist and can significantly influence patterns of vertical motion. We choose these continental-interior ice-sheet features for modification because we have assumed that the ICE-3G and ICE-4G models are in close agreement with Hudson Bay tide gauges and relative-sea-level histories and that any further modifications of modeled Hudson Bay ice (i.e., the Churchill or Cape Henrietta Maria [lat 55°N, long 83°W] ice domes) would bring these ice models out of agreement with the marine data. Furthermore, we have also assumed that the modeling of these interior ice-sheet features was done without the benefit of continental-interior data such as the lake gauges of southern Manitoba and Lake of the Woods.

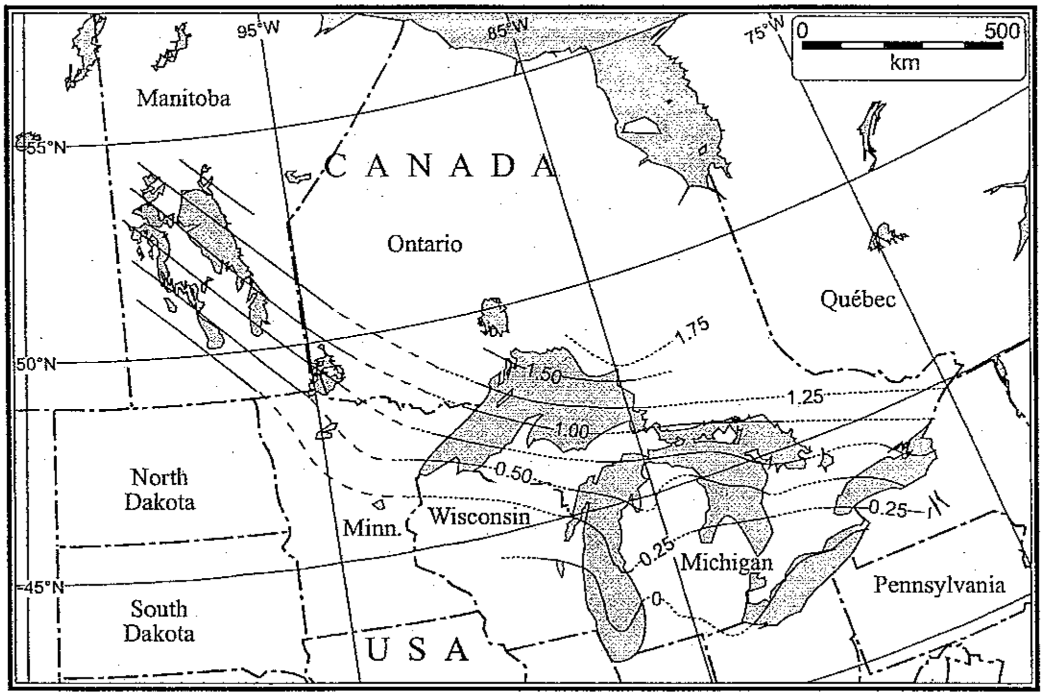


Figure 9. The postglacial rebound pattern over the four lakes in this study computed from the LL regional trend-surface model extended to the lake-gauge–derived pattern over the Great Lakes computed by Clark and Persoage (1970). The short-dash rate contours over the Great Lakes were extended by Clark and Persoage. The connecting long-dash rate contours between the two regions were extended in this study from the two patterns. All rates are expressed in feet per century (1 ft/century = 0.3048 m per 100 yr) to conform to the units used by Clark and Persoage.

Suggesting that additional ice be added to models for western Ontario is not without precedent. Clark et al. (1996) developed a numerical reconstruction of the Laurentide ice sheet for the LGM that is based on mechanical effects associated with deformable subglacial sediments. Their reconstruction included three ice domes with elevations exceeding 3500 m above sea level—one centered ~400 km east of James Bay (about lat 53°N, long 73°W), a second centered halfway between James Bay and Lake Winnipeg in western Ontario (about lat 53°N, long 89°W), and a third in the Keewatin region centered at about lat 62°N, long 105°W. The western Ontario dome featured a lobe that extended southwest through the region between western Lake Superior and Lake of the Woods. However, in a subsequent comment and reply forum (Mooers, 1997; Clark et al., 1997), a second reconstruction was computed by using a different ice margin in Minnesota that reduced the size of the western Ontario dome and consequently also removed the prominent southwest lobe. In either reconstruction, ice height in western Ontario is greater than that of ICE-4G. It would be interesting to compute rates of rebound from the Clark et al. (1996) reconstruction and compare them with those presented in this study and with those from the Great Lakes. However, the Clark et al. (1996) reconstruction is for the LGM only. The entire history of deglaciation is required to compute postglacial rebound patterns. Further support for an enhanced Laurentide ice sheet over that predicted by ICE-4G comes from Pollard and Thompson (1997), who used the GENESIS (version 2.0.a) global climate model with sea-surface temperatures prescribed from CLIMAP (1981), a mixed-layer ocean model, and the CLIMAP (1981) and ICE-4G ice-sheet reconstructions at the LGM. They found that ICE-4G would have had a significantly negative mass balance at the LGM. In summary, the results of three studies—this one, based on the observed pattern of rebound in the continental interior, the modeling of Clark et al. (1996), based on mechanical effects associated with deformable subglacial sediments, and the modeling by Pollard and Thompson (1997) using mass-balance criteria—all suggest that more ice may have been present in the Laurentide ice sheet than what ICE-4G predicts. In particular, both this study and that of Clark et al. (1996) suggest more ice in western Ontario.

Regional Lake-Level Management Implications

Since reliable recording began on these lakes about 1915, postglacial tilting has slowly shifted these water bodies to the south and west, leaving behind a clear signature of the water-body translation in the lake-gauge records (Fig. 6). As the change in the land-to-water-level relationship became significant, different gauges on a lake recorded different levels for the lake, and uncertainty associated with the "mean lake level" computed from the network of gauges increased. The best example of this existed on Lake Winnipegosis while the Dawson Bay gauge was still operational.

In the winter of 1963, when reliable recording began at the Dawson gauge, the recorded lake level was 110 mm lower than that measured at the Winnipegosis gauge. In December 1994, the year the Dawson Bay gauge was discontinued, postglacial tilting—and possibly other factors such as local subsidence at the Winnipegosis gauge site—had substantially increased the difference so that the lake level recorded at Dawson Bay was 270 mm lower than that recorded at Winnipegosis. This result gave the impression that the datum at Dawson Bay was too low, which prompted the Water Survey of Canada to attribute the difference to survey error in either the Dawson Bay or Winnipegosis datum, or both. Because a difference existed in 1963 when the Dawson Bay gauge bench marks were initially tied into Geodetic Survey of Canada bench mark 1384C, there is some justification for believing that a datum error existed. However, the reason for the difference may not be due to survey error alone but is also likely due, at least in part, to postglacial rebound. The last time the datum was adjusted at Winnipegosis was in 1947. Applying a differential rebound rate of 4.5 $mm \cdot yr^{-1}$ (Fig. 5) to the 1947–1963 interval indicates that the bench mark at Winnipegosis would have subsided about 80 mm relative to Dawson Bay. Assuming no survey error in the Winnipegosis datum, the relative subsidence due to postglacial tilting explains most of the initial 110 mm difference observed between the two sites in 1963. Thus, the datum problem may not be a datum problem at all, but rather, a postglacial rebound signature. Because the Dawson Bay gauge was discontinued by the Water

Survey of Canada in 1994, further verification of the exceptionally high rate of rebound and apparently rapid change in tilt direction over Lake Winnipegosis will have to be determined by other means.

Measuring Mean Lake Level. Computing a mean lake level is central to the monitoring and regulation of these lakes and is the main reason for maintaining the network of gauges. Because postglacial rebound deforms the datums to which the gauges are referenced and shifts the water bodies to the southwest, where and how mean lake level is computed determines how meaningful the measure is.

Considerable effort has been devoted to determining methods for computing a meaningful mean lake level. The methods developed to date address short-term fluctuations, such as those presented in Figure 3, but none of the methods explicitly corrects, or monitors, for the effects of postglacial rebound over the long term. For example, the Ad Hoc Committee (1982) on the Lake Winnipeg datum studied various methods for computing a mean lake level. They recommended a measure called the "wind effect eliminated level" to estimate the still-water level of Lake Winnipeg. This measure serves as the official level of Lake Winnipeg for regulation purposes. The method effectively reduces the short-term fluctuations seen in Figure 3 but does not address the long-term effects of postglacial rebound.

The committee also established the Lake Winnipeg datum, which was implemented in January 1986. All active gauges were adjusted by the method of water-level transfer to the level of the gauge at Berens River, which is the location of the agreed-upon master bench mark for the Lake Winnipeg datum (BM78M079, Geodetic Survey of Canada [1960]). The realignment temporarily corrected the gauges for the accumulated effects of postglacial rebound. However, examination of Figure 6 shows that the accumulated effects of postglacial tilting became apparent again within about 10 yr. Therefore, part of the solution for achieving a meaningful measure of mean lake level would be to realign the gauges to their respective datums more frequently than they have been in the past. How often realignment should occur depends on the rate of vertical motion, which is variable over the region. A realignment interval of 10 yr would be a good starting point and should keep the gauges in reasonable agreement. However, we suggest realignment every 5 yr because of our assumptions that (1) the older gauges will eventually be replaced by newer more accurate gauges capable of measuring lake level within 3–5 mm (i.e., Table 1) and (2) a rate of differential vertical motion of 1 $mm \cdot yr^{-1}$ is a reasonable estimate for the region and will become apparent in the measurements of the newer gauges within 5–10 yr after a realignment. We caution that the adjustment for every realignment must be applied to the entire length of record for each gauge in order to preserve the secular trend due to postglacial rebound.

The datums for all these lakes are several decades old (Table 1) and have been deformed significantly over the years by postglacial rebound. Thus, regional water managers should consider referencing the datums to the North American Vertical Datum of 1988 (NAVD 88) or the International Great Lakes Datum of 1985 (IGLD 1985). In either case, both datums are currently less deformed by postglacial rebound than are the datums to which the lakes are currently referenced. Furthermore, adopting either datum would better link the southern Manitoba lakes and Lake of the Woods to the Great Lakes. Finally, the current practice of realignment by water-level transfer should be verified with other leveling techniques that rely on current technologies, such as standard leveling techniques using geodetic total stations and the Global Positioning System (GPS).

Possible Increased South-Shore Flooding. Transgression of the region's lakes along their south and west shores poses a potential risk for increased flooding and wave erosion over the long term. This danger is especially true for Lake Winnipeg and Lake of the Woods. If the water budgets for these two lakes remain positive and the lake levels are held constant at their dams over the long term, possibly to maintain outlet capacity, then postglacial tilting to the northeast will cause a slow, but steady, transgression along the south shores of these lakes. The level of Lake of the Woods at Warroad, Minnesota, has increased about 80 mm relative to the dam at Keewatin over the past 80 yr (Fig. 6). Smaller increases can be estimated for most other locations on Lake of the Woods and Lake Winnipeg. Thus, up to the present, the relative change in lake level since the construction of these dams apparently does not constitute a significant south-shore flooding hazard. Nevertheless, long-range planning for the shorelines of the region's lakes, as well as what level these lakes should be maintained at, should include the effects of ongoing postglacial rebound.

Postglacial Rebound and the Red River Floodway. Another potential water-management issue is whether the rate of postglacial tilting is rapid enough to effectively decrease the gradient of the Red River floodway (Fig. 1) within the foreseeable future. The Red River floodway is a 47.3-km-long diversion channel that was constructed in the 1960s to protect the city of Winnipeg from flooding by the Red River. The floodway is designed to carry 1700 $m^3 \cdot s^{-1}$. During flood stage, some of the Red River flow is diverted into the floodway at a point just south of the city of Winnipeg. The floodway runs north along the east side of the city and then rejoins the natural channel of the Red River north of Winnipeg at Lockport. The floodway has effectively reduced flood damage in Winnipeg for 18 spring high-water events from 1969 to 1997, including the much publicized 1997 Red River flood. Although the gradient varies for different reaches of the channel, the average gradient for the entire length of the floodway is about 9×10^{-5}. The tilt rates determined from the lake-gauge data are on the order of 10^{-9} $rad \cdot yr^{-1}$. At this rate, it would take about 90 000 yr for postglacial tilting to completely reduce the floodway channel bottom to a level surface.

SUMMARY

In this paper, regional trend-surface models were used to extract the postglacial rebound signature from 19 lake gauges on Lake Winnipeg, Lake Winnipegosis, Lake Manitoba, and Lake of the Woods. A winter-season mean difference between pairs of gauge records, when these lake surfaces are relatively calm under the cover of ice, gave better fits to the models and yielded smaller standard errors than the same analyses using a summer-season mean difference, when these lake surfaces are active because of meteorological disturbances. The contemporary postglacial rebound pattern over this region appears to be steady with no significant time-varying component. As a result, the regional pattern of rebound is best explained by the simple planar representation of the LL model. The pattern consists of a northeasterly uptilt direction with a bearing of N(42.7 ± 11.8)°E and a rate of $(10.7 \pm 2.2) \times 10^{-9}$ $rad \cdot yr^{-1}$. The tilt direction is in basic agreement with those computed from the ICE-3G and ICE-4G load histories for the region. Rates of rebound derived from an ICE-3G load are in close agreement with those computed from the lake gauges, whereas rates computed from an ICE-4G load are about 50% too slow. Although the observed rebound pattern based on the lake gauges agrees better with the pattern derived from an ICE-3G load than the pattern derived from an ICE-4G load, we think that the modeled ice domes present in ICE-4G in Ontario and/or western Quebec should be increased or certain Earth structure assumptions should be modified—or both—to bring the ICE-4G rebound pattern into better agreement with observations in this continental-interior region. We select the ICE-4G model for modification over its predecessor ICE-3G because ICE-4G reflects more recent knowledge of ice-sheet dynamics, climate change, and internal Earth structure compared to the ICE-3G reconstruction.

The main reason the network of gauges is in place is to provide ongoing and historical lake-level elevations to monitor the levels and help control the outflows of these lakes. As postglacial tilting shifts these water bodies to the south and west and deforms the datums to which the gauges are referenced,

different gauges on a lake will record different water-surface elevations. Thus, steps must taken to correct the datums of these lakes in order to obtain accurate and equivalent lake levels at all gauge stations. The gauges should be realigned to their respective datums at least every 10 yr and more frequently if budgets and resources allow. In addition, the lake-gauge data for these lakes should be reanalyzed at least every 10 yr to track the progress and effects of postglacial rebound. Finally, the value of long-term record keeping cannot be overstated. Long-term water-level histories not only will benefit those interested in monitoring and managing the water levels of these lakes, but are also an important data source for scientists interested in climate change and, in our case, crustal motion due to postglacial rebound. We hope that in the near future, the value of long-term record keeping will prevail over the short-term need to reduce budgets and that the gauge stations at Dawson Bay and Toutes Aides will be reinstated as well as additional gauges added to all the lakes, especially Lake Winnipegosis.

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