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Diatom-inferred lake level from near-shore cores in a drainage lake from the Experimental Lakes Area, northwestern Ontario, Canada

Kathleen R. Laird · Brian F. Cumming

Abstract We inferred late Holocene lake-level changes from a suite of near-shore gravity cores collected in Lake 239 (Rawson Lake), a headwater lake in the Experimental Lakes Area, northwestern Ontario. Results were reproduced across all cores. A gravity core from the deep central basin was very similar to the near-shore cores with respect to trends in the percent abundance of the dominant diatom taxon, *Cylcotella stelligera*. The central basin, however, does not provide a sensitive site for reconstruction of lake-level changes because of the insensitivity of the diatom model at very high percentages of *C. stelligera* and other planktonic taxa. Quantitative estimates of lake level are based on a diatom-inferred depth model that was developed from surficial sediments collected along several depth transects in Lake 239. The lake-level reconstructions during the past ~3,000 years indicate that lake depth varied on average by ±2 m from present-day conditions, with maximum rises of ~3–4 m and maximum declines of ~3.5–5 m. The diatom-inferred depth record indicates several periods of persistent low levels during the nineteenth century, from ~900 to 1100 AD, and for extended periods prior to ~1,500 years ago.

Periods of inferred high lake levels occurred from ~500 to 900 AD and ~1100 to 1650 AD. Our findings suggest that near-shore sediments from small drainage lakes in humid climates can be used to assess long-term fluctuations in lake level and water availability.

Keywords Experimental Lakes Area · Northwest Ontario · Lake level · Drought · Diatoms · Organic matter · Transfer function

K. R. Laird (✉) · B. F. Cumming
Paleoecological Environmental Assessment and Research Laboratory (PEARL), Department of Biology, Queen's University, Biosciences Complex, 116 Barrie St., Kingston, ON, Canada K7L 3N6
e-mail: lairdk@queensu.ca

Introduction

Sediment records from topographically closed-basin lakes in arid and semi-arid regions have been used to infer past climatic conditions (e.g. Fritz 1996; Fritz et al. 1999). The utilization of sediment records from drainage lakes (i.e. lakes that have a surface outflow) in more humid regions to infer climatic conditions is more complex. The extent to which these lakes respond to climate can be difficult to estimate from the sediment record (Smol and Cumming 2000). Given that drainage lakes are the most common lake type, refining paleolimnological methods to decipher lake level from cores taken in drainage lakes will increase our knowledge of long-term water availability in regions where little information exists (e.g. Wolin and Duthie 1999).

A common method for tracking changes in lake level or shoreline location in drainage lakes is that of Digerfeldt (1986). The technique tracks sediment units across a transect of cores from the shore towards the center of the lake. This method relies on tracking changes in sediment composition and type, and changes in the distribution of littoral vegetation inferred by analysis of aquatic pollen and macrofossils (Dearing 1997). This approach assumes that the strongest evidence of changes in lake level is preserved in the near-shore sediments (Battarbee 2000). In contrast, many paleolimnological studies focus on central, deep cores, thus there have been attempts to reconstruct lake level in open, freshwater lakes from these central cores using ratios between open-water and littoral species of cladocerans, ostracods and diatoms (Battarbee 2000). Because other factors can influence such ratios, such as nutrients, pH, wind or light, reconstructions of lake depth from central cores is often not possible. However, if accumulation of near-shore sediments is adequate, then analyses of these deposits can provide a means of inferring lake level (Battarbee 2000).

Analysis of aquatic macrofossils is one means of utilizing the Digerfeldt method to reconstruct lake levels. The amount of sediment needed for macrofossil analysis, however, can be substantial (e.g. Digerfeldt et al. 1992). There has been good agreement between lake levels inferred from near-shore aquatic macrofossils, and paleoclimatic reconstructions based on pollen analogues, (e.g. Digerfeldt et al. 1992). Problems can arise, however, such as a lack of modern pollen analogues from which to infer past conditions. In such cases, the pollen proxy cannot be utilized. We used a modified Digerfeldt approach and analyzed diatom assemblages, the proportion of scaled chrysophytes to diatom valves, and percent organic matter (OM) in gravity and piston cores taken near the present-day littoral/planktonic diatom ecotone (Moos et al. 2005) in ELA Lake 239 (Rawson Lake) to infer changes in past lake levels (Laird and Cumming 2008). This combination of variables, along with the modern surface-sample analyses of OM and diatoms (Laird and Cumming 2008), proved to be a robust means of inferring past lake levels from near-shore sediments. Many characteristics of Lake 239 make it a good candidate for using the Digerfeldt method, i.e. relatively small size (<50 ha), and a small catchment

area to lake area ratio (<5:1) (Dearing 1997). In addition, direct groundwater flow to Lake 239 is negligible (Schindler et al. 1996). Thus, groundwater interactions likely do not complicate the inferences of climate from the lake-level reconstructions (e.g. Digerfeldt et al. 1992).

A number of drainage lakes in the Experimental Lakes Area (ELA), northwestern Ontario, showed marked climate-induced changes during an extended drier and warmer period in the 1970s and 1980s (Schindler et al. 1996; Schindler 1997; Findlay et al. 2001). Impacts on lake ecosystems included more transparent waters due to reduced inputs of dissolved organic carbon from the catchments, warmer temperatures, deeper thermoclines, increased water renewal time, and small declines in lake level (Schindler et al. 1996; Moos et al. 2005). The responsiveness of ELA Lake 239 to these recent droughts suggests that a longer-term record of climate conditions might be archived in the sediment.

The ELA lies within the Winnipeg River Drainage Basin (WRDB), which provides a major component of the hydroelectric power for Manitoba Hydro (St. George 2007). Little is known about the past climate of this region and long-term planning for future extremes is based on only 80–100 years of instrumental records. A few Holocene pollen records from the basin exist (McAndrews 1982; Kronberg et al. 1998), in addition to a few records from sites farther east (Björck 1985). Supplementary information comes from studies of deglaciation and Lake Agassiz (e.g. Bajc et al. 2000). Recently, a tree-ring network was developed to provide a 200–300-year perspective on drought conditions throughout the WRDB (St. George 2007).

In this study we present analyses from a suite of near-shore cores collected between 7 and 13 m water depth, and compare them to results from a central, deep core collected at ~30 m. Results were reproduced in the transect of near-shore cores, but illustrate that the central, deep core is not appropriate for lake-level inference. Analyses from the transect of cores enabled a more robust estimate of the magnitude of high stands than analyses on cores from a single depth, i.e. the littoral/planktonic diatom ecotone (Laird and Cumming 2008). Results from this study suggest that if appropriate lakes and near-shore locations are chosen, records from small drainage lakes can be produced, which could expand our

knowledge of past lake levels and water availability in many regions.

Study site

Lake 239 (Rawson Lake) is a headwater lake located within the Experimental Lakes Area (ELA) (49°40′ N, 93°44′ W), northwestern Ontario (Fig. 1a), and has been one of the primary reference lakes at ELA since its inception in 1968 (Johnson and Vallentyne 1971). It is a relatively small lake (surface area ∼ 56 ha) with a catchment of ∼ 240 ha and a maximum depth of ∼ 32 m (Fig. 1b). Limnological monitoring data from Lake 239 is extensive and has formed the basis of numerous studies of climatic effects on lake ecosystems (e.g. Schindler et al. 1996; Findlay et al. 2001). The drainage system of Lake 239 flows into Lake of the Woods, which flows into Lake Winnipeg via the Winnipeg River and eventually into the Hudson Bay drainage system (Schindler et al. 1996). The region is dominated by boreal tree species such as jackpine (*Pinus banksiana*) and black spruce (*Picea mariana*). Precambrian shield underlies the area with bedrock dominated by pink granodiorite (granite). Soils are thin and mostly moss-covered (Schindler et al. 1996).

Methods

Sampling

Replicate gravity cores were collected in ∼ 30 m of water near the deep, central basin in July 2003 with a modified K-B corer (Glew 1989). Replicate gravity cores were collected from the northeast basin in July 2004 (Fig. 1b). Cores were taken at ∼ 2 m depth intervals between 7 and 13 m depth, using a modified K-B corer (Glew 1989). The location of the near-shore cores was based on high-resolution seismic reflection data (e.g. Lewis et al. 2001), enabling us to choose an area that had the greatest near-shore sediment deposition along a relatively gentle slope. The gravity cores were sectioned in the field at 0.25-cm intervals for the entire length of the cores. From each water depth, a core was chosen for detailed analysis based on the length of core and the integrity of the core as assessed by ^{210}Pb analyses. The core chosen for analysis at 13 m (core 13C, Laird and Cumming 2008) is our master core for comparisons and chronological correlations across cores. This core was chosen for detailed analysis based on preliminary diatom and chrysophyte scale analyses of the top sediments of the near-shore cores and because sediment accumulation was highest at this location (Laird and Cumming 2008).
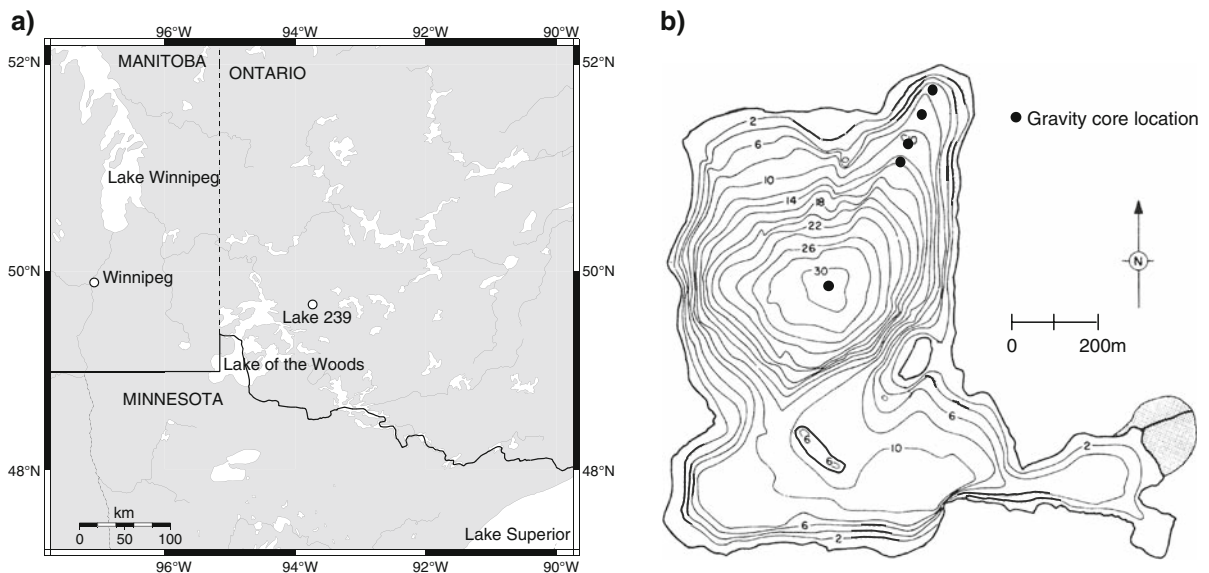

Fig. 1 Location of Lake 239 (**a**) and bathymetric map indicating the location of the gravity cores (**b**)

Chronology

Gravity core chronologies are based on ^{210}Pb for the top ~ 10 cm and stratigraphic correlation with the master gravity core at 13 m. The three radiocarbon dates of the master core were determined by accelerator mass spectrometry (AMS) of isolated pollen (Laird and Cumming 2008). The ^{210}Pb analysis is based on counts from a low-background gamma counter (Schelske et al. 1994) and chronology is calculated with the constant rate of supply model (Oldfield and Appleby 1985). The ^{14}C AMS dates were calibrated to years AD and calendar years BP using OxCal version 3.1. (Stuiver et al. 1998), with 1950 as the reference point. The resulting age model indicates that sedimentation rates have changed little over this time period (Laird and Cumming 2008). A radiocarbon date on macrofossil material ($5,035 \pm 35$ ^{14}C year BP) was consistent with the pollen-dated sample ($4,825 \pm 35$ ^{14}C year BP) from the same interval in a near-shore piston core collected in July 2004 from ~ 13 m depth (Laird and Cumming 2008). A pollen sample ($8,210 \pm 40$ ^{14}C year BP) and macrofossil sample ($8,170 \pm 70$ ^{14}C year BP) from the same interval of a piston core collected from the deep central basin, yielded similar dates.

Diatom analysis

All 186 of the 0.25-cm intervals of the master gravity core from ~ 13 m were prepared for diatom analysis. For the cores retrieved at ~ 7, 9 and 11 m, every 0.25-cm interval in the top 10–12 cm was analyzed. Below 12 cm, samples were taken every 0.5 cm from the core taken at 11 m (98 samples total), and every 1 cm from the cores taken at 7 and 9 m (54 and 74 samples total, respectively). Samples were collected from every 0.5-cm interval throughout the entire core collected at a depth of ~ 30 m (72 samples total). Approximately 0.2–0.3 g of wet sediment was sub-sampled from each core sample. Each sub-sample was weighed and placed in a 20-ml glass vial for acid digestion. The procedure then followed Moos et al. (2005). In addition, an aliquot of a known concentration of microsphere solution was added to each ~ 5-ml diatom slurry following digestion. The microspheres were counted along with the diatoms and used to estimate the concentration of diatoms (Battarbee and Keen 1982). Diatoms were identified and counted along transects

on prepared slides using a Leica DMRB microscope fitted with a $100\times$ fluotar objective (NA = 1.3), and using differential interference contrast optics at $1,000\times$ magnification. Approximately 300 diatom frustules were enumerated per slide. Dominant diatom taxa ($> \sim 5\%$), based on previous counts (Laird and Cumming 2008), were identified to the species level or lower, and rare taxa were counted in appropriate groups (e.g. genera, species with similar life forms and habitats). The main taxonomic references were Krammer and Lange-Bertalot (1986, 1988, 1991a, b), Cumming et al. (1995) and Camburn and Charles (2000). Chrysophyte scales were enumerated, but not identified to the species level.

Major stratigraphic zones were identified using a constrained cluster analysis (CONISS) with a squared-chord distance to estimate dissimilarity with the program TILIA v. 2.02 (Grimm 1987). A Principal Components Analysis (PCA) of the diatom community, using square-root transformed species data, was run using C2 (Juggins 2003) to examine the main direction of variation in taxa composition. A square-root transformation was used to downweight the importance of the dominant taxa.

Organic matter analysis

Organic matter (OM) content in the gravity core samples was estimated by Loss-on-Ignition (LOI) (Dean 1974). All gravity cores were analyzed for OM every ~ 0.5 cm for the length of each core. This resulted in 47 samples for the core from 7 m water depth, 75 samples for the cores from 9 and 11 m, 93 samples for the master core taken in 13 m, and 71 samples for the deep, central core collected in ~ 30 m.

Diatom-inferred depth model

A quantitative depth model was developed using diatom species data from 67 surface sediment samples collected from five water-depth transects (Laird and Cumming 2008), using the computer program C2 (Juggins 2003). The model is based on a modern analogue technique (MAT) on square-root transformed species data. The square-root transformation places more weight on the subdominant taxa, resulting in a reconstruction that does not simply reflect the dominant taxa. A Bray-Curtis dissimilarity index was used to compare down-core diatom assemblages with

surface sample assemblages. The model uses the average depth of the 10 most similar surface samples. The MAT approach was used because the relationship of the distribution of diatoms to depth was not linear over the entire depth gradient (Laird and Cumming 2008). At depths $> \sim 12$ m the relationship plateaus because of the dominance of the planktonic taxon *Cyclotella stelligera* Cleve and Grunow (in Van Heurck). Results from two models, based on either dominant taxa or full species counts (Laird and Cumming 2008) were highly correlated ($r = 0.95$, $P < 0.05$). Thus, results presented here are based on the dominant taxa and groups of rare taxa. This approach provides a good summary of full species counts, with the advantage of much-reduced enumeration time.

Results

Fluctuations in *Cyclotella stelligera*

The dominant taxon in all gravity cores was the planktonic species *C. stelligera* (Fig. 2). There were,

however, distinct changes in the percent abundance of this taxon, along with the other sub-dominant taxa found in the diatom assemblages of the 13-m master core (Fig. 3). The other near-shore cores (not shown) have similar trends in taxonomic composition, although the percentage of *C. stelligera* decreases towards shore and the percentage of tychoplanktonic taxa and benthic taxa compose a higher percentage of the assemblages. In the central core, the subdominant taxa are primarily other planktonic taxa, with littoral taxa being rare. Fluctuations in the percentages of *C. stelligera* are highly consistent across all cores, including the deep, central core (Fig. 2).

A stratigraphically-constrained cluster analysis from 13 m depth indicates three zones of distinct diatom assemblages (Fig. 3). Similar results were found for the other near-shore cores (not shown), whereas the central, deep basin core had a very low total sum of squares and so could not be clearly divided into distinct zones (not shown). The largest change over the last ~ 200–300 years (Zone A, Figs. 2 and 3) occurs in sub-Zone A2, with a large decrease in *C. stelligera*. This decrease is clearly evident in all of the gravity cores (Fig. 2). The onset
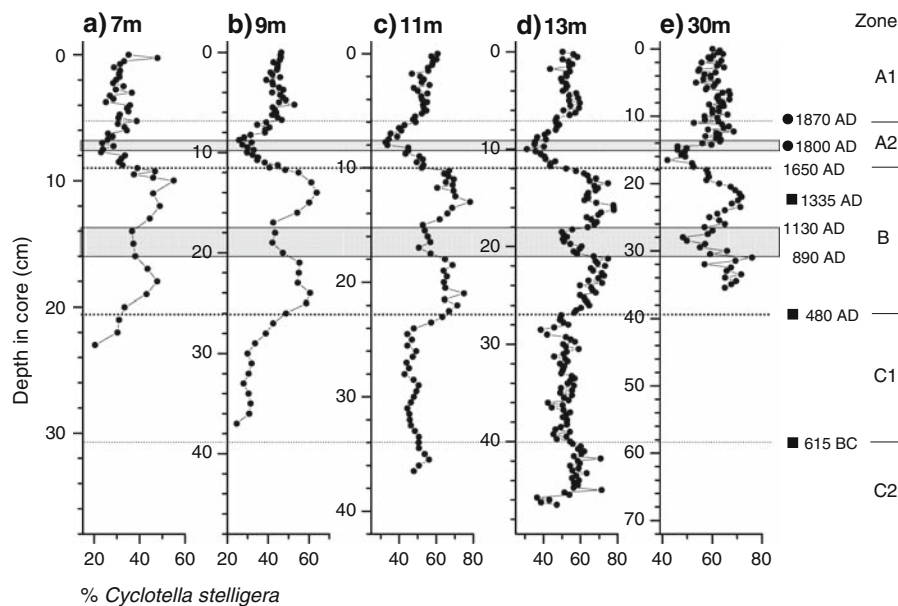

Fig. 2 Percent abundance of *Cyclotella stelligera* in the five gravity cores taken from depths ranging between 7 and 30 m. The early nineteenth century and the Medieval Climatic Anomaly drought intervals are indicated by the gray bars. Approximate ages (years AD) are provided for key intervals: zone boundaries, drought intervals indicated by the gray bars and levels that were carbon dated. Solid circles indicate approximate ^{210}Pb chronology, solid squares indicate intervals that were ^{14}C-dated and other dates are linearly interpolated key dates. Zones are based on a constrained cluster analysis of the diatom assemblage data

Fig. 3 Dominant diatom taxa (>5%) found in the master gravity core taken at 13 m. ^{210}Pb and calibrated AMS ^{14}C dates (cal year BP ± 2 sigma) and year AD are indicated. Diatom zones are based on a constrained cluster analysis. Modified from Laird and Cumming ([2008](#))
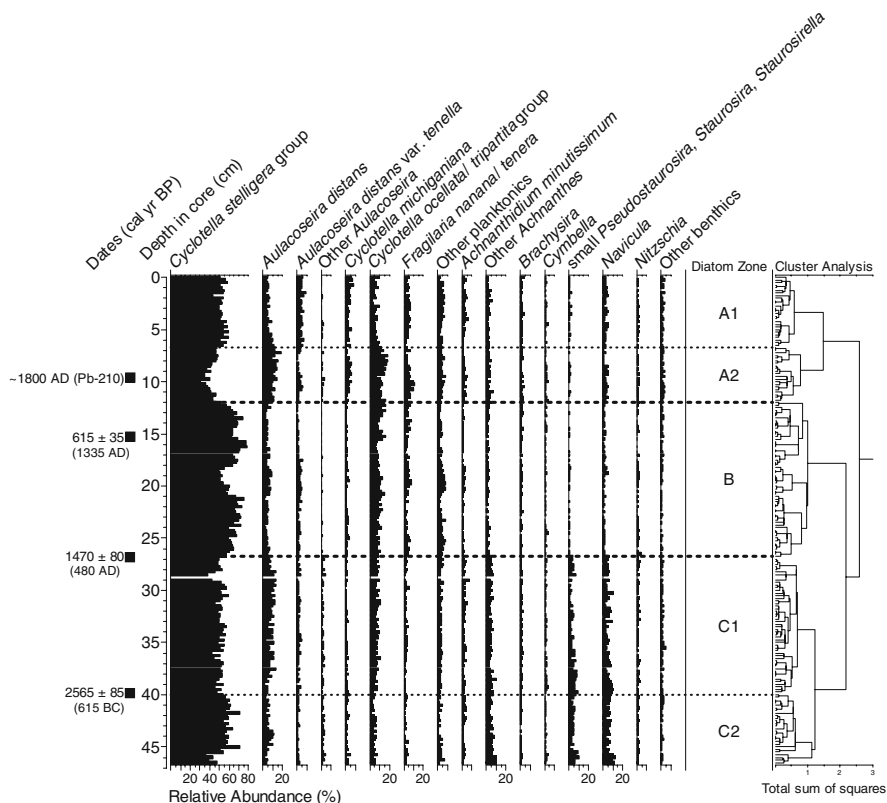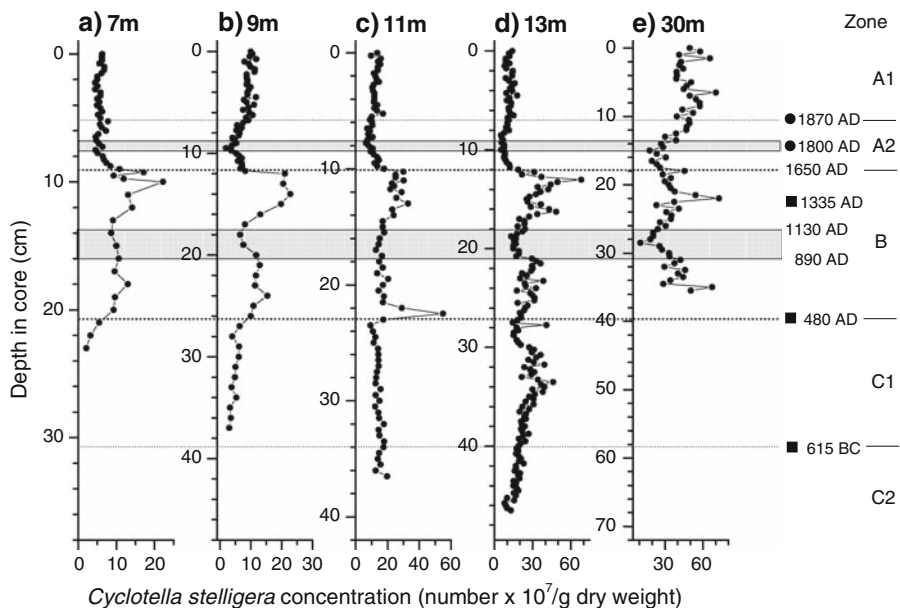

Fig. 4 Absolute abundance of *Cyclotella stelligera* in the five gravity cores taken from depths ranging between 7 and 30 m. Concentration is number of valves × 10^7 per gram dry weight. Gray bars, the indicated dates and zones are as in Fig. 2

of Zone A also indicates a clear shift in the concentration of *C. stelligera*, with the near-shore cores indicating a decrease in concentration and the central core an increase in concentration at sub-Zone A1 (Fig. 4). The total concentration of diatoms is largely represented by the concentration of *C. stelligera*, with correlations between total diatom concentration and the concentration of *C. stelligera*

ranging from 0.87 to 0.95. As a consequence, only the concentration of *C. stelligera* is shown.

Based on initial analyses of ^{210}Pb, the middle of Zone A2 was estimated to be around the 1890s (Laird and Cumming [2008]). This chronological estimate was based on using the average ^{214}Bi as an estimate of background activity. Because the late 1800s approach the limit of ^{210}Pb dating, and carbon dating indicated that sediments below this were older than that estimated by ^{210}Pb, the chronology of this period was examined more thoroughly. Background activity was estimated in two ways: (1) using the average ^{214}Bi (Schelske et al. [1994]), and (2) using the Binford ([1990]) method of estimating supported ^{210}Pb, i.e. by using the mean, down-core asymptotic value for total ^{210}Pb. The age model was rerun with different estimates of supported ^{210}Pb. Depending on which background (supported ^{210}Pb) was used, age estimates for this period varied from the early 1800s to the 1890s. Dating sensitivity in this part of the core is a consequence of low unsupported ^{210}Pb activities near the unsupported/supported ^{210}Pb boundary. Estimates of the early 1800s for the timing of this period (middle of Zone A2, 9.5 cm in the master core from 13 m) are consistent with sedimentation rates estimated from carbon dating this core; however, we can only say confidently that this period corresponds to sometime during the nineteenth century. Interpolation between the bottom ^{210}Pb date at 9.5 cm and the first carbon date at 15.25 cm (615 cal year BP, 1335 AD) yields an estimate for the bottom of Zone A2 of ∼1650 AD.

In Zone B, there is a distinct increase in the percentage of *C. stelligera* across all cores, although the increase is small in the central core (Fig. 2). In general, the percent abundance of *C. stelligera* is highest in this zone, however a distinct decline of *C. stelligera* in the middle of this zone, from ∼890 to 1130 AD, is evident in all cores. All of the near-shore cores indicate an increase in the concentration of *C. stelligera* in Zone B (Fig. 4). However, from ∼890 to 1130 AD, lower diatom concentrations are evident, being most distinct in the cores from 9 and 13 m (Fig. 4). The central core generally indicates lower concentrations of *C. stelligera* in Zone B in comparison to sub-Zone A1, with the lowest concentration between ∼890 and 1130 AD (Fig. 4). In Zone C, prior to ∼1,500 years ago, the diatom flora was quite different (Fig. 3), with higher percent

abundances of benthic taxa (e.g. *Achnanthes*, small *Pseudostaurosira*, *Staurosira*, *Staurosirella*, and *Navicula*), and lower percentages of *C. stelligera* (Fig. 2). The concentration of *C. stelligera* in the near-shore cores is lower throughout this period, with the exception of the core from 13 m, which displays high values in sub-Zone C1 and then low ones in sub-Zone C2 (Fig. 4).

Diatom-inferred depth

The trends in diatom-inferred depth are consistent across all near-shore cores (Fig. 5). Inferences from the central core, however, indicate that this site is not sensitive to fluctuations in lake level. This is due to the insensitivity of the depth model when applied to samples that are dominated only by *C. stelligera* and other planktonic taxa at the greater lake depths (Laird and Cumming [2008]). In addition, the inferred depth of the deep central core is consistently underestimated, never reaching close to the present-day depth of 30 m. Lake depth inferences from all near-shore cores are close to the actual depths at the coring site, with the exception of the inference from the core at 11 m. This, however, may be due to the loss of the very uppermost sediments of that core. PCA axis-1 scores, using square-root transformed species data, were used to summarize the main direction in variation of the taxa. The PCA scores of the near-shore cores are highly correlated with diatom-inferred depth, ranging from $r = 0.84$–0.89, whereas the central deep core has a much lower correlation ($r = 0.55$). The strength of these correlations suggests that diatom-inferred depth is a good reflection of the assemblage changes in the near-shore cores, but not in the central core. This again suggests our depth model is most appropriate for cores collected near and below the ecotonal boundary between the benthic and planktonic assemblages.

In Zone A2, diatom-inferred water depth from the core at 13 m declines, relative to present, an average of ∼0.8 m, with a maximum decline of ∼3.0 m during the early 1800s (Fig. 5). Lake-level decline is corroborated in the other near-shore cores, although the core from 11 m indicates the smallest average declines during this period, and the core from 7 m the smallest maximum declines.

In all near-shore cores, the estimate of lake level suggests that the period represented by Zone B

Fig. 5 Diatom-inferred depth for each of the five gravity cores taken from depths ranging between 7 and 30 m. The vertical lines indicate the present-day depth at the respective coring sites. Gray bars, the indicated dates and zones are as in Fig. 2
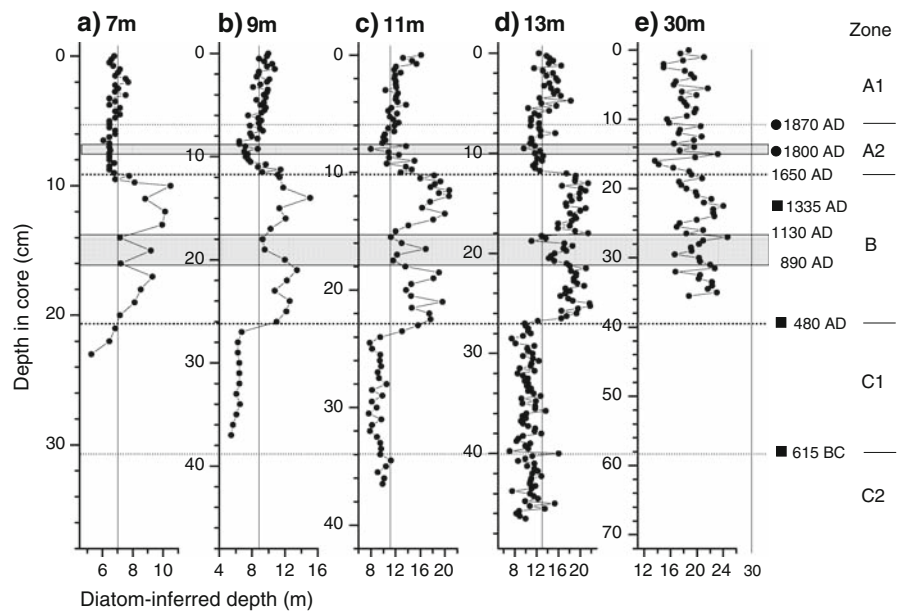
corresponds to the period of highest lake level during the late Holocene (Fig. 5). In our initial study of just the core from 13 m, it was difficult to estimate the magnitude of the rise because our model is most sensitive for estimating lake levels shallower than ~10–12 m (Laird and Cumming 2008). With the series of gravity cores we can better estimate the magnitude of this lake-level rise. The deeper cores at 13 and 11 m suggest the average rise was ~5.0 m, whereas the shallower cores at 9 and 7 m yield inferences for an average rise of ~2 m, with short-term, maximum rises of 3–4 m. The shallower cores give a more realistic estimate because even with the 2-m rise, the inferred depth is around ~10–12 m at the coring sites and is thus within the range of sensitivity of our depth model. The period from ~890 to 1130 AD (~21–18 cm depth in the 13-m depth, master core) indicates distinct declines of ~2.5–3.0 m from the generally higher lake levels of this period.

In Zone C, the near-shore cores display declines in the percentage of *C. stelligera* (Fig. 2) and increases in littoral taxa similar to that in the core from 13 m of water (Fig. 3). Inferred depth was at its lowest of the last ~3,000 years (Fig. 5). All cores except the core from 7 m depth suggest that the average decline during this period was ~2 m. The core from 7 m, however, only encompasses the very top of Zone C and thus does not fully represent this period.

Estimates of maximum declines range from ~3.5 to 5 m (Fig. 5).

Organic matter

The amount of organic matter (OM) in sediments can vary tremendously between lakes; however, within a lake the profundal sediments (deep sediments) typically vary by only 2–5% (Shuman 2003). In contrast, at the lakeward boundary of the littoral zone large changes in OM occur (Shuman 2003). The position of this boundary changes with changes in lake level. When lake level declines, the boundary of the littoral zone moves toward the center of the lake (Dearing 1997). In Lake 239, the OM content of profundal surface sediments varies from ~22% to 27%, and sharply declines towards shore starting at ~10 m (Laird and Cumming 2008). In the near-shore gravity cores, fluctuations in percent OM are small in Zones A and B (Fig. 6), whereas sharp declines occur in Zone C. The OM content of ~16–20% in Zone C of the 13-m depth master core is indicative of OM content found in the surface samples of Lake 239 from ~7 to 9 m depth (Laird and Cumming 2008). Evidence from diatoms and OM together suggests these findings are primarily due to lake lowering and an inward shift of the littoral zone.

The diatom-inferred lower lake levels in Zone A2 are not clearly apparent in the OM records, except for

Fig. 6 Percent organic matter in the five gravity cores taken from depths ranging from 7 to 30 m. Gray bars, the indicated dates and zones are as in Fig. 2
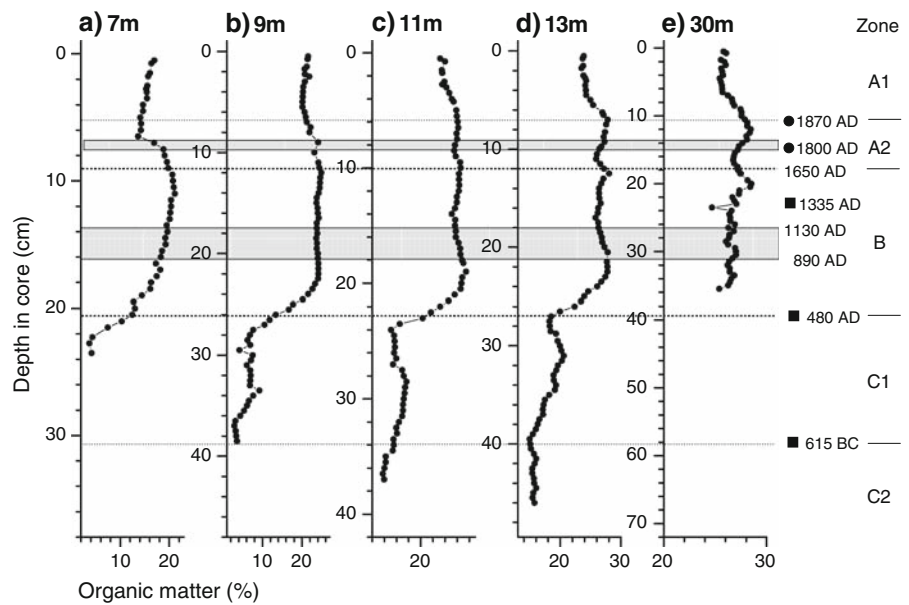
the shallowest core at 7 m (Fig. 6). The diatom-inferred decline of lake level between ~890 and 1130 AD is also undetectable in the OM records.

Discussion

Near-shore core analysis for reconstruction of lake levels

Many paleolimnological studies have used changes in planktonic and non-planktonic (benthic and epiphytic) diatom species to infer past lake-level fluctuations qualitatively (e.g. Owen et al. 1982; Gasse et al. 1989; Wolin and Duthie 1999). Quantitative assessment of lake-level fluctuations inferred from the diatom assemblages in sediments from freshwater lakes has been less common. The two main approaches for developing inference models have been: (1) analysis of diatoms collected from the central basins of lakes that vary in maximum depth, and (2) analysis of diatoms from the shallow littoral zone to the deeper pelagic zone within a lake. Predictive, diatom-based depth models based on sampling the central basins of a suite of lakes are not always applicable to the lake under investigation because of complex interactions between diatoms and depth, distribution of aquatic macrophytes, water clarity, and other variables (e.g. Barker et al. 1994; Brugam et al. 1998; Moser et al. 2000). As a

consequence, a number of studies have used the relationship between the diatom species assemblages and water depth within the study lake as a basis for reconstructing changes in past lake levels (e.g. Barker et al. 1994; Yang and Duthie 1995; Brugam et al. 1998; Nguetsop et al. 2004; Punning and Puusepp 2007). Several of these studies were based on assigning a life-form category (i.e. epiphytic, benthic, tychoplanktonic, planktonic) to the diatom species, but this can often be difficult (Barker et al. 1994; Yang and Duthie 1995). Models based on weighted averaging (WA) of depth versus diatom species along a depth gradient within the study lake, have been used with success in a number of studies (Yang and Duthie 1995; Duthie et al. 1996; Brugam et al. 1998). All of these studies found strong relationships between the diatom-inferred depth and the observed depth, with the coefficients of determination (r^2) ranging from 0.71 to 0.92. These studies, along with our recent analysis of samples along a depth gradient in ELA Lake 239 (Moos et al. 2005), and the development of a quantitative depth model based on the modern analogue technique (MAT) (Laird and Cumming 2008), clearly indicate that the distribution of diatom species within a lake basin can be related to changes in water depth and used to infer past fluctuations in lake level.

Although *C. stelligera* often comprises 40–50% (upwards of 60%) of the near-shore core assemblages, our depth model is extremely robust in

regions of the lake that are near or below the ecotonal boundary between the littoral and planktonic zones (where *C. stelligera* can reach abundances of 60–80%). The diatom taxa have distinct distributional patterns with depth (Laird and Cumming 2008). For example, planktonic taxa, such as *C. stelligera* begin to decrease in percent abundance around 10 m, the tychoplanktonic *Aulacoseira distans* (Ehrenberg) Simonsen peaks in percent abundance between 4 and 7 m, and benthic taxa have distinct distributions at depths shallower than ∼5–6 m. The strength of the relationship between the main direction of variation and inferred depth of the near-shore cores, in comparison to the weaker relationship with the central core, provides evidence that assemblages near the ecotone are highly related to depth. In the deeper environs, additional factors, such as nutrient status, species interactions, water clarity, etc. can have a larger influence on species composition.

The difficulty with using near-shore deposits is that a good sediment archive of past conditions may be lacking (Battarbee 2000). For example, Wolin (1996) attempted to analyze a series of near-shore cores; however, a number of the coring sites had little soft sediment accumulation due to the abundance of sand. Thus the determination of sediment accumulation and the slope of the near-shore environments is a key component in lake-level studies that use marginal sediments. The selection of near-shore cores for this study was based on high-resolution seismic reflection data (e.g. Lewis et al. 2001), enabling us to choose an area that had the greatest near-shore sediment deposition along a relatively gentle slope. Seismic profiling of the lake bottom was key in our being able to infer lake-level changes from the near-shore sediments of Lake 239. Without the seismic reflection data, we would have selected the original transect of the surface samples (Moos et al. 2005), which was chosen for its gentle slope. However, the seismic data indicated that the degree of sediment accumulation in this northwestern basin was small, whereas the northeastern basin had much greater sediment accumulation.

Chronology

The average resolution of each 0.25-cm interval of the 13-m master core for the ^{210}Pb-dated portion (top ∼10 cm) is ∼5 years. It is ∼20 years for each interval within the carbon-dated profile (∼10–46 cm). This resolution enables an excellent general account of lake-level conditions throughout the past ∼3,000 years, whereas higher-frequency fluctuations are not distinguished in these analyses. The period between the lower limit of the ^{210}Pb (∼10 cm) and the first carbon date at 15.25 cm is a particularly difficult part of the record for which to provide a good chronology. Different estimates of supported ^{210}Pb in the dating model can cause variations in dates at low unsupported ^{210}Pb activities. As a consequence, the estimated dates from ∼1650 to 1790 AD (∼10–12 cm) must be considered with caution. For example, extrapolated ^{210}Pb estimates at 12 cm were ∼1760 AD, whereas interpolation between the bottom ^{210}Pb date and the first carbon date estimated this interval to be ∼1650 AD. Below this interval, the carbon-dated chronology suggests that sedimentation rates of the last ∼4,000–5,000 years (Laird and Cumming 2008) have been stable and thus likely provide a good estimate of time within the constraints of ^{14}C dating errors and calibration to calendar years BP (e.g. Guilderson et al. 2005).

Coherency with other records and low stands of the past ∼300 years

The Quetico-Superior area has been described as a regional-scale ecotone between three major vegetation communities, with the boreal forest to the north (which includes the ELA region), the northern temperate forest to the south, and the prairie to the west (Kronberg et al. 1998). One of the primary influences on this ecotonal region is the confluence of three major airstreams; warm and dry Pacific air, cold and dry Arctic air, and moist tropical air from the Gulf of Mexico. In addition, within the boreal forest there is an east-west gradient from moist to dry; this region of the boreal forest is considered to be a transitional region between the humid east and arid west. Because of the proximity of the ELA to this ecotonal boundary, droughts during the twentieth century in the ELA region were often synchronous with arid periods of the west (e.g. 1930s, 1980s). Widespread and persistent North American droughts of the mid- to late-nineteenth century were not, however, always evident in the ELA region. Examination of gridded station data for precipitation

anomalies during these nineteenth century droughts in Herweijer et al. (2006) suggests that the ELA region was clearly encompassed only within the major drought of the 1870s, but near the edge of the major droughts of the mid-1850s to mid-1860s, and the 1890s drought. However, much of this data is based on extrapolation from a few climate stations, particularly in the ELA region, and thus must be viewed with caution. Nonetheless, the main point is that the ELA region often lies on the edge of steep climatic gradients and thus is in and out of phase with other regions of North America. As a consequence of its location, the ELA is a key region for examining the long-term susceptibility to drought and water availability in a region which today is often viewed to be more typical of the humid east.

Historical lake-level data have been recorded at Lake 239 from 1969 to present. During the warmer and drier 1970s and 1980s (Schindler et al. 1996), measured lake-level declined by 0.2–0.3 m at Lake 239 (Moos et al. 2005). Our records do not distinguish this period, most likely because the measured changes in lake level were small. The largest decline in the past ~200–300 years occurs in the nineteenth century (middle of Zone A2) with maximum declines of ~2–3 m and average lake-level declines of approximately 0.8 m (Fig. 5). The bottom portion of Zone A2 is estimated to be ~1650–1790 AD. All cores, including the deep central core recorded a sharp decline in the percentage of *C. stelligera* during sub-Zone A2. However, because of the high percentage of *C. stelligera* and other planktonic taxa in the central core, the site is not sensitive for inferring lake-level changes. The concentration of *C. stelligera* also declines during this period. In the near-shore cores, the trend of low *C. stelligera* concentration remains until present, whereas concentration increases in the most recent sediments of Zone A1 of the central core (Fig. 4).

Several periods of drought in western North America during the nineteenth century have been recorded in historical, proxy, and instrumental records. Explorer accounts in the early 1800s (Zebulon Pike) and 1819–1820 (Stephen Long) describe the North American Great Plains being like a desert (Cook et al. 2007). Records of crop failure and tree-ring data document drought in the Canadian prairies from ~1815 to 1819 (Case and MacDonald 2003). Tree-ring records and historical accounts of eolian

activity document droughts in the 1800s and 1820s in many regions of central North America (Muhs and Holliday 1995; Woodhouse and Overpeck 1998). Periods of drought in many regions of western North America during the mid- and late-1800s have also been documented. The explorer, Captain John Palliser, considered the central Canadian prairies as 'useless land' during his exploration from 1857 to 1860 (McDonald and Case 2000). Historical data from eolian activity, newspaper accounts, early meteorological records and drought reconstructions from tree rings indicate that drought was widespread throughout much of central and western U.S. in the 1860s (Woodhouse and Overpeck 1998). Herweijer et al. (2006) document several widespread and persistent droughts in the U.S. during the latter half of the nineteenth century, with periods of drought from 1856–1865, 1870–1877 and 1890–1896. Tree-ring records within the Winnipeg River Drainage Basin suggest intermittent lower-growth years during much of the 1880s and 1890s, but these were not as extreme as in the 1860s and 1910s (St. George 2007). These data do not correspond completely with the mapping of nineteenth-century droughts in Herweijer et al. (2006); however, this is likely due to the extrapolation of precipitation data from a few climate stations. Tree-ring records farther to the west indicated that the 1890s was a period of drier conditions in many parts of the North American prairies (e.g. Sauchyn et al. 2003; Herweijer et al. 2007).

Given the uncertainty of the Zone A2 chronology, it is difficult to pin-point exactly when the nineteenth-century, maximum low lake levels occurred. However, based on the carbon dates and rate of sedimentation, it is likely closer to the early 1800s than the 1890s as first thought (Laird and Cumming 2008). In general, the nineteenth century (encompassed within sub-Zone A2) had lower lake levels than the twentieth century. The bottom of this zone (~11–12 cm in the 13-m master core) is estimated to be ~1650–1720 AD, which may correspond to an arid period documented in other proxy records. Severe drought from ~1660 to 1710 AD was documented in the pollen record from varved Lake Mina, Minnesota, adjacent to the present-day prairie/deciduous forest boundary (St. Jacques et al. 2008). Further to the northeast, peaks in eolian activity were documented in Elk Lake, MN from ~1670 to 1750 AD (Dean 1997) and inferences from calcite $\delta^{18}O$ at

Steel Lake, MN indicated an arid period from ∼1650 to 1700 AD (Tian et al. 2006). In southern Manitoba in the Red River Valley, tree-ring data indicate an extremely dry period from ∼1670 to 1775 AD (St. George and Nielsen 2002). These periods of drought overlap with the second colder phase of the so-called Little Ice Age (LIA). Pollen data from Lake Mina indicate a very cold phase from ∼1625 to 1775 AD (St. Jacques et al. 2008). The expansion of the Big Woods in MN at ∼1650 AD has been related to the LIA cooling (Brugam and Swain 2000). Perhaps in the ELA region, the apparent arid second phase of the LIA is the result of long periods of dominance of cold, dry Arctic air.

Late Holocene lake-level history
∼1050 BC–1650 AD (∼3,300–300 cal year BP)

The lake-level record of Lake 239 indicates generally high lake levels between ∼480 and 1650 AD (∼1,470–300 cal year BP, Zone B). During this period, both the percentage and concentration of C. stelligera was generally higher in the near-shore cores. A late sixteenth century megadrought indicated in many tree-ring records across North America (Woodhouse and Overpeck 1998; Woodhouse 2004) is not apparent in the Lake 239 lake-level record. However, it is also not apparent in adjacent regions in Minnesota (Dean 1997; St. Jacques et al. 2008), nor in the Red River Valley of Manitoba where inferred precipitation was above normal from ∼1580 to 1620 AD (St. George and Nielsen 2002). This combined data suggests that this region is beyond the northeastern limit of the sixteenth century megadrought. There is, however, a period of much lower lake levels earlier in Zone B.

There was a distinct decline in lake level of ∼2.5 to 3.0 m from ∼890 to 1130 AD (∼1,060–820 cal - year BP). This interval corresponds to an epic drought recorded in many regions of North America from ∼800 to 1400 AD. It is often referred to as the Medieval Climatic Anomaly (MCA) or the Medieval Warm Period, and encompasses "The Great Drought" of the thirteenth century (Woodhouse and Overpeck 1998; Woodhouse 2004; Herweijer et al. 2007). The estimates of timing of this epic drought depend on the geographical location and the chronological constraints of the proxy records. Tree-ring records from western North America indicate

prolonged, severe aridity from ∼900 to 1300 AD (Cook et al. 2004). In Minnesota, several proxy records indicate droughts within the MCA interval. A century-scale drought recorded during the 1300s in a 900-year pollen record from Mina Lake was the most severe of the record period with sand deposits suggesting large declines in lake level (St. Jacques et al. 2008). Eolian deposition was high at Elk Lake from ∼1280 to 1410 AD (Dean 1997) and an arid period from ∼1100 to 1400 AD is inferred from calcite δ^{18}O at Steel Lake (Tian et al. 2006). As in the Steel Lake record and other records where wetter periods occur within generally more arid intervals, the lake-level decline from ∼890 to 1130 AD in Lake 239 is interrupted by wetter conditions from ∼1000 to 1050 AD (19.0–19.75 cm in the master 13-m core). This short rise is most apparent in the more highly-resolved 13-m core. Another smaller decline in lake level is centered at ∼680 AD (24 cm in the master 13-m core). This period is out of the range of tree-ring records, and few other high-resolution records that are adjacent to our study region cover this period. Records of inferred peatland water-table depth do not indicate the period around 680 AD as a dry interval, however widespread drought events are recorded during ∼950–1250 AD (the MCA period) and water-table levels were generally low from ∼1,900 to ∼3,500 cal year BP (Booth and Jackson 2003, Booth et al. 2006). This latter period overlaps with the low lake levels in Lake 239 in Zone C from ∼3,300 to ∼1,470 cal year BP.

Inferred depth was at its lowest of the last ∼3,000 years during Zone C. This is clearly apparent in both the diatom and OM records of all the near-shore cores. The OM records reflect the large changes in lake lowering which resulted in an inward shift of the littoral zone. However, smaller declines inferred earlier in the record (i.e. Zone A2 and from ∼890 to 1130 AD) are not apparent in the OM records. Thus the OM records appear not to be sensitive enough, in comparison to the diatoms, to pick up smaller, short-lived declines in lake level. Average declines during Zone C were ∼2 m, with maximum declines of up to ∼3.5–5.0 m. A decline of ∼2 m would cause a ∼14% decrease in surface area and an ∼18% decrease in the volume of Lake 239. Declines of 4 m would cause decreases in surface area of ∼23% and a reduction in volume of ∼33% (Laird and Cumming 2008). However, in comparison to the

much lower lake levels of the mid-Holocene, conditions of the past ∼3,000 years were more similar to present-day conditions (Laird and Cumming 2008).

The timing of the onset of 'modern' lake-level fluctuations in Lake 239 corresponds to the timing of broad-scale syntheses of lake-level conditions during the Holocene in North America (Ritchie and Harrison 1993; Thompson et al. 1993). These studies suggest that the transition to generally cooler, moister conditions occurred over a protracted period which began ∼5,000–6,000 cal year BP, but was not established until ∼2,000–3,000 cal year BP. This is similar to what the Lake 239 record indicates, with a transitional period from the extreme aridity of the mid-Holocene beginning ∼5,500 cal year BP, with more modern conditions commencing ∼3,300 cal year BP (Laird and Cumming 2008). Sites further west and east suggest that modern lake conditions were not established until ∼2,000 cal year BP (Michels et al. 2007). The variation in timing of the onset of modern lake conditions likely reflects the variability in geological and hydrological setting, as well as the geographic variance in timing of climatic change at specific locations.

Conclusions

The detailed study of the near-shore sediments from Lake 239 indicates that lake levels during the past ∼3,000 years have varied on average by ±2 m from present-day conditions, with maximum rises of ∼3–4 m and maximum declines of ∼3.5–5 m. The series of near-shore cores indicates high reproducibility in the percent and total abundance of the dominant taxon *Cyclotella stelligera*, diatom-inferred depth, and percent organic matter. This transect of near-shore cores enables a better constraint on the magnitude of lake-level fluctuations, particularly rises in lake level, than analyses only at one depth. High-resolution analysis of the core nearest to the present-day ecotone between benthic and planktonic taxa, at 13 m, provided an excellent view of declines and trends in lake-level increases during the late Holocene (Laird and Cumming 2008). Although the deep central basin core had very similar trends in the percentage of *C. stelligera* to the near-shore cores, it does not provide a sensitive site for analysis of lake-level changes because of the insensitivity of the diatom model at very high percentages of *C. stelligera* and other planktonic taxa. The robustness of our findings suggest that near-shore sediments from small drainage lakes in humid climates, with adequate sediment accumulation in the near-shore environs, can provide an excellent means of assessing long-term fluctuations in lake level and water availability.

Our lake-level record indicates several periods of persistent low levels during the nineteenth century, from ∼900 to 1100 AD, and for extended periods prior to ∼1,500 years ago. Water availability under future human-induced warming is of concern to industries dependent upon water resources, such as agriculture and hydro-generation of electricity, and to society in general for future drinking water and other consumptive uses. Many regions of Canada are projected to have increased precipitation under increasing CO_2 conditions; however, there is great uncertainty in both the temporal and spatial distribution of future precipitation based on General Circulation Models (Schindler 1997; IPCC 2007). Even if there is an increase in precipitation, there will likely be less water availability due to higher air temperatures and periods of extended evaporation and transpiration (Schindler 1997). For example, at the ELA during the warm 1970s and 1980s, average evaporation increased by ∼35 mm per 1°C increase in annual temperature and was nearly double this rate (∼68 mm per 1°C increase) during the summer months (Schindler 1997). Consequently, without a substantial increase in precipitation, it is projected from the empirical data from ELA that lake levels, stream flows and groundwater levels will all decline under a warmer climate (Schindler et al. 1996; Schindler 1997). Conditions during the arid mid-Holocene have been surmised to be a good analog for future climatic conditions under greenhouse gas forcing. However, even the more modest declines in water availability during the late Holocene may make it difficult for society to cope, particularly if water management continues to be based on the short-term instrumental records, which clearly do not encompass the full range of natural climate variability.

The Canadian prairies are already experiencing reductions in surface-water availability due to climatic warming and human withdrawals (Schindler and Donahue 2006) and many regions in the western U.S. have experienced water supply deficits in reservoir storage with the recent multi-year drought

(e.g. Cook et al. 2007). These severe multi-year drought conditions pale in comparison to the many widespread megadroughts that persisted for decades and sometimes centuries in many parts of North America over the last millennium (e.g. Woodhouse 2004). It has been observed with the recent drought in western North America that agricultural and hydrological systems do not have the resilience to survive such conditions as inferred during past megadroughts, at least under our current use of water resources.

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