USING TREE RINGS, LAKE SEDIMENTS, AND ANCIENT SHORELINES TO RECONSTRUCT PAST CLIMATE AND LAKE LEVELS FOR DAY COUNTY, SOUTH DAKOTA

TASK 3 FINAL REPORT

Prepared for the Day County Risk Assessment Study
Federal Emergency Management Agency Region VIII
Denver, Colorado

September 2000

in cooperation with

SOUTH DAKOTA STATE UNIVERSITY

SDSU

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EXECUTIVE SUMMARY

1. The Federal Emergency Management Agency declared several counties in northeastern South Dakota a disaster area on June 1, 1998 (FEMA-1218-DR). Roads, farmland, and towns became flooded when heavy rains began to fill the many closed-basin lakes. Waubay Lake in Day County rose more than 5 meters and more than doubled in area between 1993 and 1999. Other closed-basin lakes in the northern Great Plains flooded during this time, such as Devils Lake in north-central North Dakota.
2. An inter-disciplinary and inter-agency study was initiated shortly after the disaster declaration to understand the causes of the flooding, the chances for recurrence, and possible mitigation alternatives. The team of scientists and engineers divided the project into five tasks. Final reports have been issued for tasks 1, 2, 4 and 5. This final report contains the findings and conclusions of task 3, which used several investigative techniques to reconstruct the past climate and lake levels of Day County.
3. The specific purpose of task 3 was to determine how often in the past the closed-basin lakes of Day County have risen to levels comparable to those of today. Scientific approaches included using tree rings to indicate past climate patterns, examination of outflow channels and radioisotope dating of ancient shoreline formations to identify previous high-water periods, and coring of Spring Lake, one of the Waubay Lakes chain, to examine the sedimentary record of seed shrimp (ostracode) shells to indicate past water levels.
4. A range of climatic and ecological conditions have presided over northeastern South Dakota since continental glaciers began their retreat 18,000 years ago. A cold, dry climate supporting spruce forest and dilute lakes prevailed 12,000 years ago. The climate then warmed and the ecosystem changed to tall grass prairie and windswept alkaline lakes. About 2,000 years ago the modern climate arrived supporting prairie with patches of deciduous forest and highly variable precipitation. Two samples from the Bitter Lake basin that were identified as lakebed and till deposits contained pine *(Pinus)*, spruce *(Picea)*, and other pollen types suggestive of middle-Wisconsin conditions; the pollen of neither sample is "typical" of Holocene nor late glacial pollen assemblages in the region. The presence of pine in the Bitter Lake basin samples indicates an age of perhaps 15,000 to 18,000 years.
5. The lakes of Day County are very changeable in depth and in area. Some were dry in the 1930s. In 1998, total surface area of the Waubay Lakes chain was 3.5 times the lake area in 1939 and 1.3 times the surface area in 1868-1877 when the lakes were first surveyed. Waubay Lake itself was 7.6 times larger in 1998 than in 1939.
6. A tree-ring chronology was constructed from 16 bur oak cores collected in 1999 on islands in Waubay Lake and from 4 bur oak cores collected previously from building timbers at Fort Sisseton State Park. The chronology based on these cross-dated cores spans 325 years (1674-1998). The chronology identifies a wet period at about year 1725, possibly strong enough to have produced water levels comparable to those of the 1990s. Thus, two wet periods with flood potential appear in the chronology (one known in the 1990s and one indicated in the 1720s), with an average frequency of 1 in 163 years. The same frequency occurred in the chronology for extreme droughts, one in the 1930s (known) and another in the 1750s (indicated).
7. Nearby regions have similar climatic patterns. Tree-ring widths from Day County corresponded to lake level patterns for Devils Lake, North Dakota, a regional closed-basin lake with a long data record. Also, a strong correlation exists between the Day County tree-ring chronology and another chronology at Lake Herman (Madison, S.D.).
8. The Day County chronology suggests steadily increasing favorability of weather for tree growth since the 1930s. This 60-year-long pattern is unprecedented in the 325 year chronology and suggests steady increases in subsoil moisture and in shallow aquifers, contributing to the extreme flooding when the high rainfall and cool temperatures of the 1990s arrived.
9. Two sediment cores [4 m (13 ft) and 1.5 m (5 ft) in length] were collected from Spring Lake in June, 1999. The cores were analyzed for dry-density, water content, organic content, carbonate content, sediment mineralogy, pollen, and ostracodes. Both Lead-210 and C-14 methods were used to date the core. Elemental analysis was conducted on ostracode shells. Five ostracode taxa were found in the Spring Lake cores. Ratios of Mg/Ca and Sr/Ca were determined from elemental analyses conducted on *Candona rawsoni.* High Mg/Ca ratios indicated drought periods, while low ratios indicated wet periods.
10. The ostracode analysis indicated that Spring Lake sustained freshwater episodes (highstands) centered on 1250, 1400, and 1660 AD. High salinity (low water) periods were centered on 1330, 1480, 1600, and 1740 AD. After 1800 AD, salinity events decreased in period. Generally, there was good

correspondence between the post-1650 AD Spring Lake salinity record and the local tree-ring record of moisture. The long-term highstand frequency of 163 years derived from the tree ring chronology over a 325-year period was similar to the approximately 140-year frequency derived from the lake sediment record for a 1,000-year period.

11. Ostracode shell chemistry did not track the known rising water levels in Spring Lake in the 1990s. Rather, the data indicated increasing salinity, suggesting declining lake levels. This result may have been caused when the more saline water of Swan and Hillebrands basins coalesced with the fresher water of Spring Lake. The specific conductance of Spring Lake more than doubled between 1995 and 1999.
12. The lake sediment and tree-ring records provide strong evidence that the Waubay Lake complex has experienced a cycle of repeated and sustained highstands punctuated by periods of significant drought. Prior to 1800 AD, highstands persisted from several decades to a century. The low to moderate lake levels with high variability, which characterized the period since white settlement (but before the recent flooding), are not typical of the longer data record. Additionally, the steadily increasing growth of trees in northeastern South Dakota may indicate the advent of a wetter climate. The EPA reports that precipitation along the northern tier of states has increased by 10-15 percent in this century. As a result, extrapolating from weather records for the period of record (the twentieth century) may underestimate the future occurrence and duration of highstand conditions of the Waubay lakes chain.
13. Aerial photographs and field observations reveal numerous beachlines formed in response to past fluctuations in lake level. No shorelines higher than an elevation of about 549.6 m (1803 ft) have been preserved in the lower lake basins. Two channelways identified as possible flow paths from Bitter Lake to the Big Sioux River were examined for evidence of flow features and fluvial landforms. In both cases, the channelways were dominated by glacial till and sandy outwash presumably deposited at the time of glacial retreat. No lacustrine sediment suggestive of levels of Bitter Lake between 549.6 and 552 m (1803 and 1811 feet) was found. These channelways exhibit no evidence of modification by flowing water. Thus, field examination of the two channelways indicates that significant spillage from Bitter Lake to the Big Sioux River has not occurred since deposition of glacial till and outwash.
14. Most or all major lakes of the Waubay chain except Bitter Lake periodically have had water levels high enough to cause overflow. Several of the upper lakes, such as Enemy Swim Lake and Pickerel Lake, have outlets with well-defined stream-channel shapes. The same is not as apparent for spill paths between several of the lower lakes. Only after Blue Dog, Rush, Spring, Hillebrand, Minnewasta, and Waubay Lakes have merged into a composite lake at the 548.3 m (1799 ft) level, can water spill into Bitter Lake. This condition occurs following the filling of Rush Lake and spillage from it into Waubay Lake. After water reaches the level of 548.3 m (1799 ft) in the Waubay lake basin, spillage can occur from the Rush Lake portion of the composite lake into the Bitter Lake basin. Thus, under natural (pre-development) conditions, the upper lakes act as a single input rather than as separate controls on the water level, area, and volume of Bitter Lake.
15. The water level of Rush Lake, the nearest lake to Bitter Lake, recently reached a partially-controlled (post-development) height great enough to spill limited flows into Bitter Lake, but this condition has probably been infrequent during Holocene time because observations show that the channelway lacks well-defined features. A substantial amount of water must flow from Rush Lake (and the rest of the single-lake complex) into the voluminous Bitter Lake basin before spillage of Bitter Lake can occur. This extensive surface area provides the means by which a significant amount of water is lost by evaporation and perhaps infiltration to the ground-water system. Evidence for high evaporative losses from the Bitter Lake basin is its high concentration of dissolved solids, typically 4 to 60 times greater than higher-elevation lakes. Waubay Lake generally has concentrations of dissolved solids 2 to 15 times greater than higher lakes of the chain. It follows that spillage from Rush Lake (the next highest lake) into Waubay Lake has been unusual during Holocene time. Elevated concentrations of dissolved solids in water of Waubay Lake, combined with the water-quality data of Bitter Lake, suggest that mixing of water during periods of high lake levels and the formation of a single composite lake have occurred infrequently or rarely.
16. If Bitter Lake fills to a level of 548.3 m (1799 ft), further inflow from higher lakes cannot occur because at that level the large, single-lake basin includes that of Bitter Lake. When this lake level is reached, the entire complex of lakes is probably controlled mostly by the balance between direct precipitation falling on the lake and evaporation (including transpiration from associated wetlands) from the lake surface. If so, this control is likely one of strongly stabilizing the hydrology (especial-

ly movement of water), area, and volume of the composite lake. For such a lake to surpass 548.3 m (1799 ft), a large excess of direct precipitation relative to evaporation would be essential.

17. Combined geomorphic observations, geologic and climatic information, water-quality data, and pollen and radiocarbon analyses suggest that (1) Bitter Lake has not spilled to the Big Sioux River since glacial retreat, and (2) the highest lake stand, resulting in preserved shorelines around several lakes, was at an elevation of about 549.6 m (1803 ft). Based on a single radiocarbon date and results of other studies, it is speculated that this ancient shoreline may have been formed between 2600 and 4400 years BP.
18. The extreme nature of the flooding in the 1990s may have been caused or at least worsened by human actions, namely the damming effects of roads and increased runoff to the Waubay Lake complex from farmed land and drained wetlands. The pre-settlement lake complex often reached highstand conditions but rarely produced spillage between the terminal (lowest) lakes in the chain.

INTRODUCTION

The Federal Emergency Management Agency declared several counties in northeastern South Dakota a disaster area on June 1, 1998 (FEMA-1218-DR) because of severe flooding of roads, farms, and towns from rising lake levels. Day County was the hardest hit because of its numerous and large closed-basin lakes, whose natural outlets often are much higher than road surfaces and human settlements. When prolonged wet periods occur, as in the early to mid-1990s, some of these lakes can rise by many meters over several years. Waubay Lake in Day County rose by 5.7 m and more than doubled in area from 1993 to 1999 (Niehus *et al.* 1999a). Other closed-basin lakes in the northern Great Plains have flooded during this wet weather extreme, most notably Devils Lake in North Dakota which rose over 5 m between 1990 and 1997 causing considerable economic damage (Wiche *et al.* 1997, Winter and Rosenberry 1998).

Rising lake levels were caused by a combination of above average precipitation and below average air temperatures (i.e., reduced evaporation). For example, annual precipitation at the Webster, South Dakota weather station averaged 69.6 cm between 1991 and 1998, 16.0 cm above the 1960-1990 normal (NGPWRRC 1999). Cool and more humid air produced low potential evaporation during the 1991-98 period and contributed to 225.8 cm of land-surface hydrologic loading (difference between evaporation and precipitation) (NGPWRRC 1999). Normally, evaporation exceeds precipitation in this region.

Current levels of Day County lakes are unprecedented in recorded history, although systematic measurements began only in 1960 for the largest closed-basin lakes (Waubay Lake, Bitter Lake) (Niehus *et al.* 1999a). Scattered, older records indicate that the lowest levels occurred in the 1930s. The longer continuous records at Devils Lake, North Dakota, indicate moderately high lake levels around 1900 (but lower than those of the 1990s); anecdotal records suggest high water levels in the 1860s (Winter and Rosenberry 1998). Government reports from Ft. Sisseton, South Dakota, near the Waubay Lake chain, also indicate high water conditions in the 1860s (Bennett 1878), but no actual measurements were made. These scattered records suggest a high lake-stand recurrence frequency of approximately 100-150 years.

A comprehensive, inter-disciplinary study was initiated shortly after the disaster declaration to determine the historical context of the rising lake levels, estimate the time expected to return to more normal levels, and to explore mitigation alternatives. FEMA needed this information to assess and possibly to mitigate further damage to public and private property in the affected areas (Niehus *et al.* 1999a). An interagency investigative team divided the project into five tasks:

- Task 1: History of inundation (1992-1998)
- Task 2: Predictive modeling based on climate scenarios
- Task 3: Long-term estimates of historical lake-level chronologies
- Task 4: Water balance and lake-level frequency analysis
- Task 5: Hydrologic modeling and evaluation of mitigation alternatives

Final reports have been issued for tasks 1 and 2 (Northern Great Plains Water Resources Research Center 1999), task 4 (Niehus *et al.* 1999a,b), and task 5 (U. S. Army Corps of Engineers 1999).

This report contains the findings of the third task, which used paleoecological and geomorphological methods to reconstruct the past climate and lake levels of Day County. The specific purpose was to determine how often in the past the Waubay Lake complex rose to levels comparable to those of today. Paleoecological methods have enabled scientists to put current weather extremes, such as the flooding in Day County, into historical perspective. Clearly, such events are rare enough that their long-term frequency cannot be estimated well from the relatively short, post-settlement period. Determining the long-term frequency of such events is key to predicting the occurrence of the next highstand or flood, once the current high lake levels recede.

Three methods were used to reconstruct past climatic conditions and lake levels. These were: (1) collection of cores from living trees, linking of these to cores collected from military fort timbers, and construction of a tree-ring width chronology to extend weather conditions for Day County back several hundred years, (2) identification and aging of ancient outflow channels and shoreline features formed during previous high-water periods, and (3) coring of Spring Lake (one of the Waubay Chain lakes; Fig. 1) to extend the climate records back a thousand years using the chemical composition of biogenic carbonates accumulated in the lake sediments. Integration of these diverse approaches provided the best, cross-checked evidence for reconstructing climate for Day County and northeastern South Dakota.

CLIMATE AND LAKE LEVEL HISTORY

Post-Glacial

Much is known about the climate history of northeastern South Dakota, primarily because Pickerel Lake in Day County has been the subject of numerous paleoecological investigations (e.g., Watts and Bright 1968, Haworth 1972, Smith 1993, Schwalb and Dean 1998). Its relatively deep water (13 m) has yielded an uninterrupted record of environmental history from its sediments. Pickerel Lake appears not to have gone dry since deglaciation. Other regional lakes have been the focus of more recent studies, such as Moon (e.g., Laird *et al.* 1996a, 1998, Fritz *et al.* 1994, Winter and Rosenberry 1998) and Devils Lake in North Dakota (e.g., Fritz *et al.* 1994, Haskell *et al.* 1996) and Elk Lake in Minnesota (e.g., Bradbury and Dean 1993).

Paleoecological studies have shown northeastern South Dakota to have been a region of environmental extremes, encompassing glaciation, droughts, and floods. Approximately 18,000 years B.P. (before present), northeastern South Dakota was under full-glacial conditions. After deglaciation, pollen assemblages from Pickerel Lake (Day County) indicate that a cool and dry climate supported extensive spruce forest about 12,000 B.P. (Watts and Bright 1968). As the climate warmed further, a deciduous woodland developed (9,000 B.P.), followed by mixed-grass prairie (7,000 B.P.), and finally by tall-grass prairie intermixed with areas of deciduous woodland (2,000 B.P. to present).

While the pollen data indicate the types of terrestrial vegetation present in the region since glaciation, diatoms (algae) and ostracodes (seed shrimp) from the bottom sediments of Pickerel Lake indicate its specific water quality (and water volume) history (Haworth 1972, Schwalb and Dean 1998). Shortly after glaciation (12,000 B.P.), Pickerel Lake was dilute due to a cold climate and acidic soils associated with spruce forest. The next period (10,000 B.P.) was considerably drier, producing alkaline water and probably the highest salinity in the history of the lake. A prairie lake with greater influence from groundwater became established about 9,000 B.P. and lasted until about 2,000 B.P. The climate of the region has been relatively stable (i.e., no consistent directional change) in the last two millennia; however, there have been noteworthy periods, including the Little Ice Age (1300-1850 A.D.) characterized by cooler and wetter weather, the Medieval Warm Period (1000-1300 A.D.) with warmer and drier weather and lower lake levels (Laird *et al.* 1998), and approximate 400-yr cycles of aridity during which windy conditions created high inputs of dust to Pickerel Lake (Schwalb and Dean 1998).

Woodhouse and Overpeck (1998) have shown from multiple data sources, including an extensive tree-ring network, that drought dominates the climate history of the Great Plains during the past 2,000 years. Some droughts prior to 1600 A.D. were more severe (decades in duration) and of greater geographical extent than those of the 1930s and 1950s. Tree-ring evidence suggests that droughts since 1600 A.D. have tended to be a decade or less in duration.

Paleolimnological studies of other regional lakes have been conducted, yielding a number of insights into the past climate of the northern Great Plains. A high-resolution reconstruction of salinity fluctuation in Devils Lake, North Dakota, based on fossil diatoms, ostracode-shell chemistry, and bulk carbonate geochemistry, suggests an arid climate and drought conditions at least as extreme as those of the Dust Bowl from the 16th through mid-19th centuries (Fritz *et al.* 1994). Low salinities, and by inference, wetter periods occurred between 1700 and 1750, the late 1800s, and the last three to four decades. These intervals correspond in a general fashion with less arid periods interpreted from diatom stratigraphy (Bradbury 1988) and charcoal records of lowered fire frequency (Clark 1990) in the Itasca region of western Minnesota. The Devils Lake data also imply that present-day fresh conditions are unprecedented during the preceding 500-years of record, although the introduction of European agriculture in the 1880s may have modified the hydrological response of the lake to climate change.

The most detailed account of late Holocene climate for this region published thus far is that from Moon Lake (Barnes Co.) in southeastern North Dakota (Laird *et al.* 1996a, 1998). Here diatom-based salinity reconstructions (with sub-decadal resolution) for the last two millennia reveal repeated oscillations between high and low salinity that indicate shifts between dry and moist periods. This record shows persistently high salinity prior to AD 1200, suggesting frequent and severe droughts of greater magnitude and duration than those of the 20th century. A major decline in mean salinity occurs after AD 1200 with two major freshwater events centered around AD 1350 and 1850. The overall decline in salinity is roughly coincident with the onset of the Little Ice Age (AD 1300-1850) and implies enhanced effective moisture within the last 750 years relative to the previous 1500 years of record. This interpretation is consistent with pollen records that show Little Ice Age expansion of mesic tree species in forested regions of eastern Minnesota (Grimm 1983), but is opposite in direction from that revealed in an oxygen-isotopic record from Lake Mina in west-central Minnesota (near the town of Alexandria). Here, a high-resolution O^{18} signal from endogenic carbonates increases after AD 1300, and the higher values imply enhanced evaporative effects (Stevens 1997).

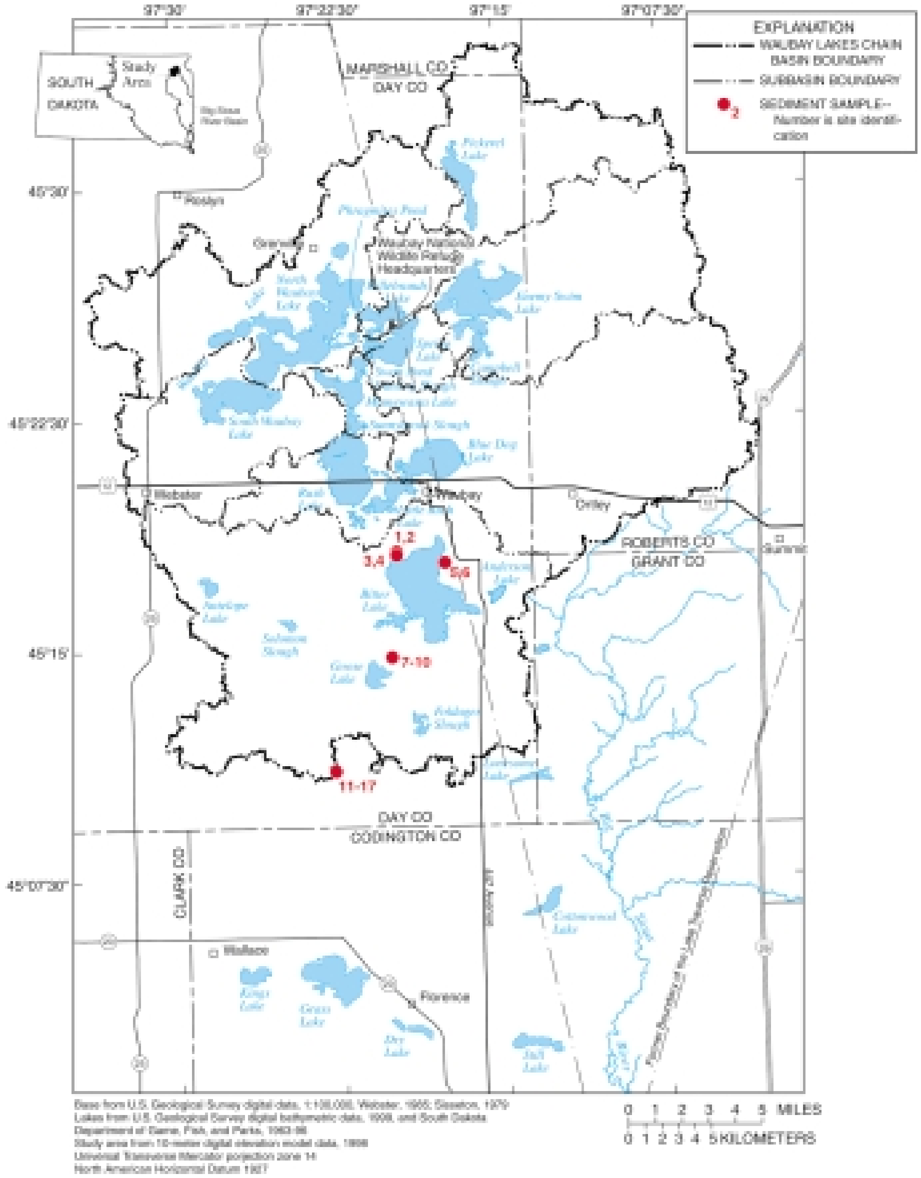

Figure 1. Waubay Lakes study area and locations of sediment samples taken for radiocarbon dating and/or pollen analysis (after Niehus *et al.* 1999a).

More recently, the Moon Lake core has been compared with two new high-resolution sediment records, one from Coldwater Lake (MacIntosh Co.) also in southeastern North Dakota, and the other from Rice Lake (Ward Co.) in north-central North Dakota (Fritz *et al.* 2000). Both records span the last 2000 years at roughly decadal resolution and utilize ostracode shell chemistry to infer lakewater Mg/Ca ratios (which in turn track salinity). At Coldwater Lake, diatoms were analyzed as well as ostracodes, and the salinity reconstructions from the two methods are coherent, both in terms of direction and magnitude of change. The reconstructions from all three lakes indicate that regional climate of the last two millennia was hydrologically complex, with large oscillations between low-salinity wet phases and high-salinity dry phases. These hydrological shifts are broadly synchronous among sites, despite fine-scale differences in both timing and magnitude that are due, in part, to errors inherent in core chronology. All sites show intervals of very fresh conditions (similar to or lower than present day), suggesting high precipitation, sometime between AD 1330 and 1430 and in the mid-1800s. Wet conditions are also evident between 1650 and 1700 in two of the records, Coldwater and Rice, but there is a gap in the Moon Lake core during this interval. The data also indicate persistent and severe drought around AD 1500, 1600, and 1800.

In summary, published studies of the post-glacial climate of northeastern South Dakota indicate considerable change in regional vegetation, lake size/depth, and water quality. Spruce forests and dilute lakes that formed shortly after deglaciation were succeeded by prairie vegetation surrounding windswept alkaline lakes. The modern condition of relatively high water, grassland vegetation, and scattered deciduous forests in protected areas has existed for the past 2,000 years. The climate in northeastern South Dakota during the modern period is typical of the Great Plains, with considerable inter-annual variation in temperature and precipitation with periods of drought (Borchert 1950).

Modern Period

Water levels and surface area of many Day County lakes have changed dramatically over the past one-and-a-half centuries. Although the data record is fragmentary, there is little doubt that current levels are maximum for the period, while levels in the 1930s are the minimum. Lakes were full and large at the start of the period during early European settlement based on the meander survey conducted by the General Land Office survey of 1868-1877. Lakes at that time were similar in size to those indicated by LANDSAT images in 1994 (Table 1), just after the lakes had begun to rise sharply from heavy rains in 1993 (Fig. 2). Total lake area was reported to be 10,343 ha in 1868-1877 and 8,549 ha in 1994 (Table 1).

In contrast, several lakes were dry or nearly dry in 1939 when the first aerial photographs were taken. Total water surface area of the lakes after the 1930s drought was about one-third of that in 1868-77 (Table 1). Lake levels probably were lower several years earlier at the height of the drought. Bitter Lake was dry in 1934 (Rothrock 1935).

Systematic measurement of lake levels for the closed-basin lakes with high outlet levels began only recently. The data for Waubay Lake indicate relatively stable levels from the early 1960s until 1993 (Fig. 2). Between 1993 and 1998, Waubay Lake rose by about 6 meters. Total lake area in 1998 was 13,433 ha, 3.5 times the area in 1939, and 1.3 times the area in 1868-77 (Table 1). The specific lakes which have been the most dynamic are Waubay and Bitter. For example, the water surface area of Waubay (South and North lakes combined) in 1998 was 7.6 times its area in 1939. Bitter Lake was 3.2 times greater in area in 1998 than in 1939.

FIELD AND LABORATORY METHODS

Tree-Ring Studies

Prior to this study, northeastern South Dakota presented a large gap in the continent-wide tree-ring and climate data base (Cook *et al.* 1996). The natural scarcity of trees in the central U. S. grasslands and heavy timber exploitation during initial European settlement have left few living trees much older than meteorological records (approximately 100 years). To make up for the dearth of old trees from which to reconstruct paleoclimates, the Laboratory of Tree-Ring Research (University of Arizona-Tucson) searched for old timbers in Fort Sisseton in northeastern S.D. (Sieg and Meko 1995). Two oak timbers were found in the Commanding Officer's Residence which probably date back to the original construction about 1864. Two cores were extracted from each timber in 1992. These cores were included in the Laboratory's data base, but no living trees in northeastern S.D. were located or cored at the time to link with the Fort trees to provide a continuous record of ring widths and climate analogues.

Our goal was to find living trees which would overlap in date with the Fort cores. Trees at least 150 years old were necessary to construct a useful

Table 1. Lake area based on General Land Office meander survey, 1939 aerial photography, and 1994 & 1998 LANDSAT imagery.

Unit Name	*1868-1877 Meander Survey (hectares)*	*1939 Aerial Photography (hectares)*	*1994 LANDSAT (hectares)*	*1998 LANDSAT (hectares)*
Pickerel	437	398	370	406
North Waubay	2,926	622	2,694	3,759
Enemy Swim	1,166	652	844	1,179
Hillebrands	271	103	183	379
Spring	477	257	379	523
Blue Dog	624	572	587	625
Swan Pond	64	0	43	82
South Waubay	1,292	123	1,271	1,934
Minnewasta	222	16	234	260
Rush	1,145	43	63	1,311
Bitter	1,717	923	1,381	2,976
TOTAL	*10,341*	*3,709*	*8,549*	*13,434*

chronology. Beginning in fall, 1998, large trees in wooded areas near Waubay Lake were found and cored. Forty-five trees were cored at five sites (Sica Hollow, Fort Sisseton State Park, Pickerel Lake, Oak Island, and Waubay Lake) (Table 2). The largest number of samples (32) was collected at Waubay Lake itself. Trees were cored at breast height and the cores were returned to South Dakota State University for preparation and initial analysis.

The cores were mounted in wooden channels, filed flat, and sanded with increasingly finer sandpaper until the rings were clearly distinguishable (Stokes and Smiley 1996). Ring widths on cores were measured to the nearest 0.01 mm using a standard encoder/translator mechanism connected to a computer at the U. S. Geological Survey's Midcontinent Ecological Science Center in Fort Collins, Colorado.

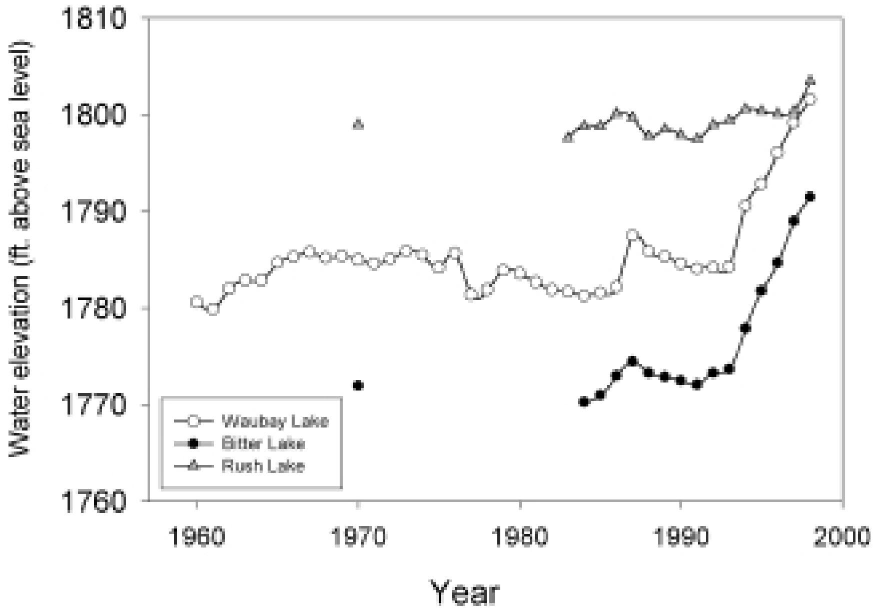

Figure 2. Lake level data record for Waubay, Rush, and Bitter Lakes.

Ring series on individual cores were standardized using quotients between annual ring widths and overall means of ring widths. Also, the natural effect of slower growth with age was removed using detrending curves. Ring widths were detrended using best-fit straight lines constrained to horizontal or negative slopes or negative exponential curves. The outermost rings of the Fort cores could not be dated precisely, but could be linked to the modern series by matching overall growth patterns using skeleton plots (which display ring width patterns among cores) and correlation coefficents between ring measurements. The modern and Fort cores were the same tree species (bur oak; *Quercus macrocarpa*) growing near each other in the same physiographic and climatic region, the *Coteau des Prairies.*

Sediment Cores from Spring Lake

General

The geochemistry of fossil ostracodes in a sediment core from Spring Lake served as a proxy for paleosalinity and lake-level change. Ostracodes accumulating in the lake sediments retain evidence of water salinity at the time they were alive. Water salinity is an indicator of water depth and volume and can be estimated from the Mg/Ca and Sr/Ca ratios and stable isotopes (O^{18} and C^{13}) of ostracode shells.

Bur oak on island of Waubay Lake cored by W. Carter Johnson, February 1999.

Table 2. Summary of core samples collected from trees in northeastern South Dakota.

Core Sample #	*Date Cored*	*Location*	*Species*	*Tree Diameter (cm)*	*# Annual Rings*
1	30 Nov 98	Sica Hollow, Roberts Co.	Green Ash	74.5	80
2	30 Nov 98	Sica Hollow, Roberts Co.	Bur Oak	51	74
3	30 Nov 98	Fort Sisseton State Park, Marshall Co.	Bur Oak	77	113
4	11 Dec 98	Pickerel Lake, Day Co.	Bur Oak	71.8	96
5	11 Dec 98	Pickerel Lake, Day Co.	Bur Oak	73	109
6	11 Dec 98	Pickerel Lake, Day Co.	Bur Oak	53.3	101
7	11 Dec 98	Pickerel Lake, Day Co.	Bur Oak	55	84
8	11 Dec 98	Pickerel Lake, Day Co.	Bur Oak	51.5	80
9	17 Dec 98	Oak Island, Roberts Co.	Bur Oak	55	71
10	17 Dec 98	Oak Island, Roberts Co.	Green Ash	59	59
11	17 Dec 98	Oak Island, Roberts Co.	Bur Oak	51	78
12	17 Dec 98	Oak Island, Roberts Co.	Bur Oak	56	54
13	17 Dec 98	Oak Island, Roberts Co.	Bur Oak	53.5	72
14	14 Jan 99	Waubay Lake-Island A, Day Co.	Bur Oak	54.5	99
15	14 Jan 99	Waubay Lake-Island A, Day Co.	Bur Oak	56.5	93
16	14 Jan 99	Waubay Lake-Island A, Day Co.	Bur Oak	63	94
17	14 Jan 99	Waubay Lake-Island A, Day Co.	American Elm	83.5	78
18	14 Jan 99	Waubay Lake-Island A, Day Co.	Green Ash	61.5	90
19	14 Jan 99	Waubay Lake-Island B, Day Co.	Bur Oak	58.8	145
20	15 Jan 99	Waubay Lake-Island B, Day Co.	Bur Oak	60	130
21	15 Jan 99	Waubay Lake-Island B, Day Co.	Bur Oak	59	132
22	15 Jan 99	Waubay Lake-Island B, Day Co.	Bur Oak	65.5	120
23	15Jan 99	Waubay Lake-Island C, Day Co.	Bur Oak	72	81
24	15 Jan 99	Waubay Lake-Island C, Day Co.	Bur Oak	74.5	136
25	15 Jan 99	Waubay Lake-Island C, Day Co.	Bur Oak	66	93
26	15 Jan 99	Waubay Lake-South Shore, Day Co.	Bur Oak	70.5	127
27	17 Feb 99	Waubay Lake-Staunton Island, Day Co.	Bur Oak	57.5	105
28	17 Feb 99	Waubay Lake-Staunton Island, Day Co.	Bur Oak	55	97
29	17 Feb 99	Waubay Lake-Staunton Island, Day Co.	Bur Oak	57	113
30	17 Feb 99	Waubay Lake-Staunton Island, Day Co.	Bur Oak	58.5	106
31	17 Feb 99	Waubay Lake-West Woods Island, Day Co.	Bur Oak	59.5	161
32	17 Feb 99	Waubay Lake-West Woods Island, Day Co.	Bur Oak	61	106
33	17 Feb 99	Waubay Lake-West Woods Island, Day Co.	Bur Oak	61	109
34	17 Feb 99	Waubay Lake-West Woods Island, Day Co.	Bur Oak	61	133
35	17 Feb 99	WaubayLake-West Woods Island, Day Co.	Bur Oak	60.5	112
36	17 Feb 99	Waubay Lake-West Woods Island, Day Co.	Bur Oak	73.5	136
37	17 Feb 99	Waubay Lake-West Woods Island, Day Co.	Bur Oak	56.5	110
38	17 Feb 99	Waubay Lake-West Woods Island, Day Co.	Bur Oak	64	120
39	18 Feb 99	Waubay Lake-West Woods Island, Day Co.	Bur Oak	64.5	143
40	18 Feb 99	Waubay Lake-West Woods Island, Day Co.	Bur Oak	64.5	95
41	18 Feb 99	Waubay Lake-West Woods Island, Day Co.	Bur Oak	62.5	148
42	18 Feb 99	Waubay Lake-West Woods Island, Day Co.	Bur Oak	64	147
43	18 Feb 99	Waubay Lake-West Woods Island, Day Co.	Bur Oak	60	114
44	18 Feb 99	Waubay Lake-West Woods Island, Day Co.	Bur Oak	68	147
45	18 Feb 99	Waubay Lake-West Woods Island, Day Co.	Bur Oak	78.5	104

Core Collection and Processing

A 4-m long sediment core was collected with a square-rod (modified Livingstone) piston corer from the apparent deep point of Spring Lake (determined by acoustic sounding) on June 14, 1999 (Fig. 3). At the same location a piston corer equipped with a clear polycarbonate core barrel was used to collect a 1.5-m long surface core preserving the sediment-water interface. Livingstone-core sections were collected in 1-m overlapping drives, extruded on site, and wrapped in plastic food wrapping and aluminum foil for transport to the University of Minnesota, Limnological Research Center (LRC). The surface core was extruded vertically and sectioned into 2-cm increments on site and stored in thread-capped polypropylene specimen cups for transport to the St. Croix Watershed Research Station, where subsampling for ^{210}Pb chronology took place. All subsequent description and subsampling of both the 4 m core and the surface core were performed at the LRC Core Lab Facility.

At the time of the Spring Lake coring, a single water sample for analysis of major-ion chemistry was collected from the well-mixed near-surface waters at the coring site. We subsequently assembled from various sources additional existing chemical data collected from Spring Lake, Waubay Lake and surrounding water bodies.

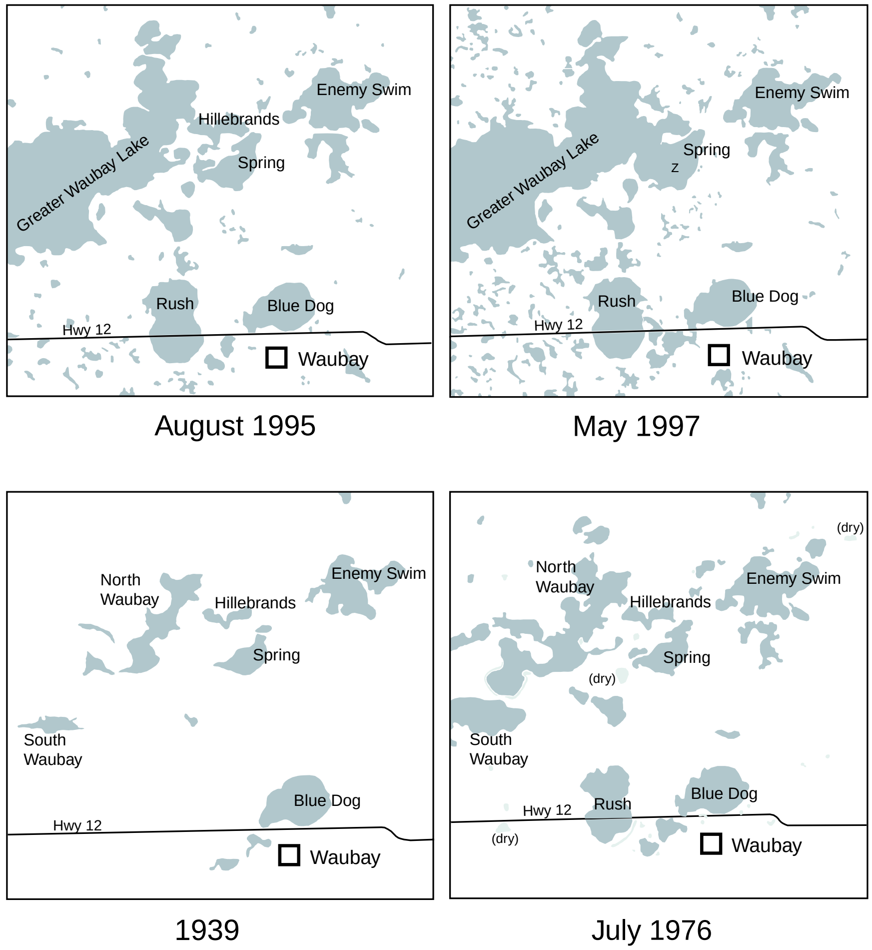

Figure 3. Historical changes in surface-water coverage in the Lake Waubay region interpreted from Landsat imagery and aerial photography compiled by the U.S. Geological Survey (http://gisdata.usgs.gov/nesdflood/). The May 1997 image shows the location of the coring site (+) in the Spring Lake basin.

Loss-on-Ignition

Dry-density (dry mass per volume of fresh sediment), water content, organic content, and carbonate content of the surface core and the first two sections of the 4-m core were determined by standard loss-on-ignition techniques (Dean 1974). Sediment samples of 4-5 g were dried overnight at 100° C and ignited at 550° and 1000° C for 1 hr each. Mass measurements were made of the wet samples and after each heating on an electronic analytical balance. Dry density was calculated from water content and fixed densities for organic, carbonate, and inorganic fractions. The resulting LOI stratigraphies were used to cross-correlate core sections and correct for small errors in depth measurement that occur during core collection.

XRD Mineralogy

Sediment mineralogy was measured by X-ray diffraction (XRD) every 6 cm in the analyzed core. Subsamples of bulk sediment (approximately 0.5 cm^3) were ground in ethanol with a mortar and pestle. Slides prepared from the resulting slurry were scanned over 5° to 65° in 2θ on a Siemens D500 diffractometer with automatic sample changer. The Reference Intensity Ratio (RIR) method was indexed to naturally occurring quartz to generate semi-quantitative estimates of mineral proportions from the XRD spectra. The XRD stratigraphy was normalized to a 3-phase system (aragonite + calcite + quartz), ignoring

(Above) A 1.5 m long surface core collected from Spring Lake by Mark Shapley, June 1999. (Below) Livingstone core collected from Spring Lake, June 1999, by Dan Engstrom and Mark Shapley.

the minor dolomite and albite present in the XRD spectra.

Lead-210 Dating

Twenty-three depth intervals from the surface core were analyzed for ^{210}Pb activity to determine age and sediment accumulation rates for the past 150 years. Lead-210 was measured through its grand-daughter product ^{210}Po, with ^{209}Po added as an internal yield tracer. The polonium isotopes were distilled from 0.6-2.3 g dry sediment at 550° C following pretreatment with concentrated HCl and plated directly onto silver planchets from a 0.5 N HCl solution (Eakins & Morrison 1978). Activity was measured for 1-3 x 10^5 s with ion-implanted or Si-depleted surface barrier detectors and an Ortec alpha spectroscopy system. Unsupported ^{210}Pb was calculated by subtracting supported activity from the total activity measured at each level; supported ^{210}Pb was estimated from the asymptotic activity at depth (the mean of the lowermost two samples in the core). Dates and sedimentation rates were determined according to the c.r.s. (constant rate of supply) model (Appleby & Oldfield 1978) with confidence intervals calculated by first-order error analysis of counting uncertainty (Binford 1990).

AMS 14C Dating

Two core levels were selected for AMS (accelerator mass spectrometry) ^{14}C dating based upon extrapolation of the ^{210}Pb dating curve for the surface core. All available sediment from the selected intervals was sieved through 250-µm and 125-µm screens, and the wet sieve residue was immersed in distilled water for microscopic examination and selection of ^{14}C dateable material. AMS samples were assembled from very fine, delicate charcoal believed to be of airfall origin, primarily from the 250 micron fraction. Any abraded, rounded, or discolored material that suggested possible fluvial transport or subaerial weathering was rejected. These samples were then submitted to the LRC AMS Target Preparation Lab for pretreatment (HCl and NaOH) and reduction to graphite AMS targets. Measurement of ^{14}C was conducted by the NSF-Arizona Accelerator Mass Spectrometry Laboratory. Radiocarbon dates were converted to calendar years AD using the CALIB program of Stuiver and Reimer (1993) and calibration data sets from Stuiver *et al.* (1998).

Pollen Analysis

Pollen analysis was conducted on selected intervals to identify the stratigraphic horizon for the onset of European settlement in the Waubay region, and specifically the arrival of Russian thistle *(Salsola iberica)* in 1885-1890, which is known precisely for the Dakota region from USDA records (Jacobson & Engstrom 1989). Pollen samples from six surface-core intervals were prepared according to procedures described by Faegri and Iversen (1975). Residues were mounted in silicon oil, and pollen identified under magnifications of 400X and 1000X. At least 300 terrestrial pollen grains were counted in each sample.

Ostracodes Abundance and Elemental Analysis

Subsamples of each 2-cm core interval were wet-sieved through 250-µm and 125-µm screens using a manual plant sprayer and warm tap water. After removal of the fine sediment, ostracode shells and other residue were rinsed in distilled water, collected onto cellulose filters and air-dried under cover. The screens were examined microscopically between samples, and any adhering shells were manually removed to prevent cross contamination. The dried sieve residues were examined for ostracode shells and ranked semi-quantitatively for abundances of the five most common ostracode species; identifications were confirmed by Rick Forester (USGS, Denver). Forty juvenile shells (A-1 and A-2 instars) of *Candona rawsoni* were selected from each core interval for pretreatment and chemical analysis. Abundance of *C. rawsoni* (and of total ostracode shells) varies substantially in the core, so that in some intervals of low abundance all available well-preserved A-1 and A-2 shells present in our analytical samples were included. In all cases, we excluded broken, visibly etched, and taxonomically uncertain shells from the treated samples.

All shell pretreatments and sample preparations were conducted in the University of Minnesota Stable Isotope Laboratory. Shells were cleaned for 5 to 7 hours in a solution of 50% commercial bleach followed by a sequence of at least 6 immersion rinses in high purity deionized water. Under a dissecting microscope, shells (7 shells per depth increment) destined for elemental analysis were given a final rinse in deionized water followed by reagent grade absolute alcohol and stored in acid-washed vials. The cleaned shells were then dissolved in acid-washed glass vials by reaction with high-purity anhydrous H_3PO_4. Following dilution with a Rh-spiked standard solution, elemental analysis (Ca, Mg, Sr and Ba) of the dissolved shells was performed by inductively-coupled-plasma mass-spectrometry (ICP-MS) in the Department of Geology and Geophysics Geochemistry Lab (University of Minnesota). Standard QA/QC procedures included standard and blank analyses every fourth sample and 5 replicate measurements per sample element.

Shoreline Dating and Channelway Morphology

General

Other measures, such as dating paleo-beach and spill-site sediment, assisted in dating and estimating the frequency of previous high water periods. Methods involved identifying and characterizing paleo-beachlines and other geomorphic features from aerial photographs, topographic maps, field observations, and soil and geologic surveys, and sampling for datable carbon (mostly decayed plant carbon) and pollen in beach sand, lakebed deposits, outwash, and till. Other features of significance used in dating past events included channelways entering and leaving lake basins, lake-water chemistry, and identification and comparison of spill-site elevations for several lakebed basins.

Waite Osterkamp near Bitter Lake, September 1999.

Aerial photographs and topographic maps

Aerial photography and 7.5-minute topographic maps were examined to identify likely flow paths out of the Bitter Lake basin if past spillage had occurred, to provide preliminary topographic and geomorphic indications of whether discharge occurred along these paths, and to identify potential sites for collection of sediment samples for radiocarbon and pollen analyses. Photography and field checks upon which the topographic maps were based had been conducted 1970 through 1973. Aerial photography of July, August, and September, 1939 (scale 1:20,000), obtained from the National Archives, was used to determine a likelihood of whether streamflow had occurred along possible spillage routes and to yield initial sites for sediment sampling. Photographs of summer, 1939, the earliest available, were selected because lake levels were near historic lows, agriculture and other land uses caused less interference for identification of ancient shorelines than it did in subsequent photography, and sparse vegetation and low lake levels allowed sampling sites to be readily identified. Many potential sites exposed in 1939, however, have been under water in recent years.

The identification of possible spillage routes from the Bitter Lake basin during the last 10,000 years is based on assumptions that the topographic maps are accurate, that elevations along the possible flow paths have not changed significantly by human activity, and that isostatic rebound following deglaciation was largely complete by the beginning of Holocene time. Of numerous flow paths considered, the two identified as most likely for spillage to the Big Sioux River have modern divides at 45° 11' 1.9", 97° 22' 55.2" (T 120 N, R 55 W, sec. 24, NW ¼ of SE ¼ of SW ¼, Florence NW Quadrangle), and 45° 14' 39.7", 97° 13' 53.7" (T 121 N, R 53 W, sec. 24, SW ¼ of NE ¼ of SE ¼, Lonesome Lake Quadrangle). Elevations at the sites (identified as sites 1 and 2; Fig. 1), respectively, are 552.02 m and 552.33 m (Kim Kempton, NRCS, personal commun., 1999).

Collection of sediment samples

Twenty-eight sediment samples were collected from beach and lakebed deposits and at potential spillage site 1 in and around the Bitter Lake basin. Eighteen of the samples were from cores recovered as part of the Day County investigations. Seventeen samples (Fig. 1) were selected for radiocarbon dating, pollen analyses, or both. The radiocarbon dating was conducted by Geochron Laboratories, Cambridge, MA, and the samples for pollen analysis were submitted to the Bilby Research Center, Northern Arizona University (NAU), Flagstaff, AZ. The pollen analyses given here are in some cases abstracted directly from a written report of March, 2000, describing results by Dr. R. Scott Anderson, NAU. Many of the samples were submitted for analysis, although it was anticipated that reported radiocarbon dates could be much too young (due to death and decay of modern vegetation) or because sandy beach and outwash deposits might contain little or no pollen. Analyses for these samples were requested to help ensure that interpretations of the depositional environments from which the samples were collected were correct.

Field studies

Field studies, supported by photo and map interpretations, were designed to appraise geomorphic and hydrologic conditions of the Waubay Lakes area of Day County. Owing to unusually high lake levels at times of the field investigations, observations were not extensive, and many sites at which visits were desired were either inundated or otherwise inaccessible. Most attention during field investigations (other than for the collection of samples for radiocarbon or pollen analyses) was given to spill paths between lakes, the positions and persistence of shorelines, and the geomorphic characteristics along those paths identified as potential natural spillways from the Bitter Lake basin to the Big Sioux River (sites 1 and 2; Fig. 1).

RESULTS

Tree-Ring Chronology

General

Tree cores ranged in age from 54-161 years. None of the green ash or elm cores collected was older than 90 years (Table 2). The oldest trees found were bur oak, growing on the islands of Waubay Lake. The oldest non-Waubay Lake core was 113 years collected from Fort Sisseton State Park. The largest number of old oak trees was cored on West Woods Island in Waubay Lake, including the oldest (161 years; first ring in 1839) core in the sample. Eight cores in the total sample had their first growth ring in or prior to 1864, the year Fort Sisseton was built.

Seventeen of the oldest cores were used to construct the chronology (core numbers 14, 15, 16, 19, 20, 21, 22, 24, 29, 31, 32, 34, 36, 39, 41, 42, 44; Table 2). Each core was standardized and detrended before averaging ring widths for each year. The cores from Fort Sisseton were processed in the same manner. The Fort and living-tree series appear to overlap by 13 years. The overlap occurs with one living tree (#31; first ring in 1839). With this overlap, the Fort Sisseton series dates to 1851. The complete chronology (Fig. 4) covers years 1674-1998, a total of 325 years.

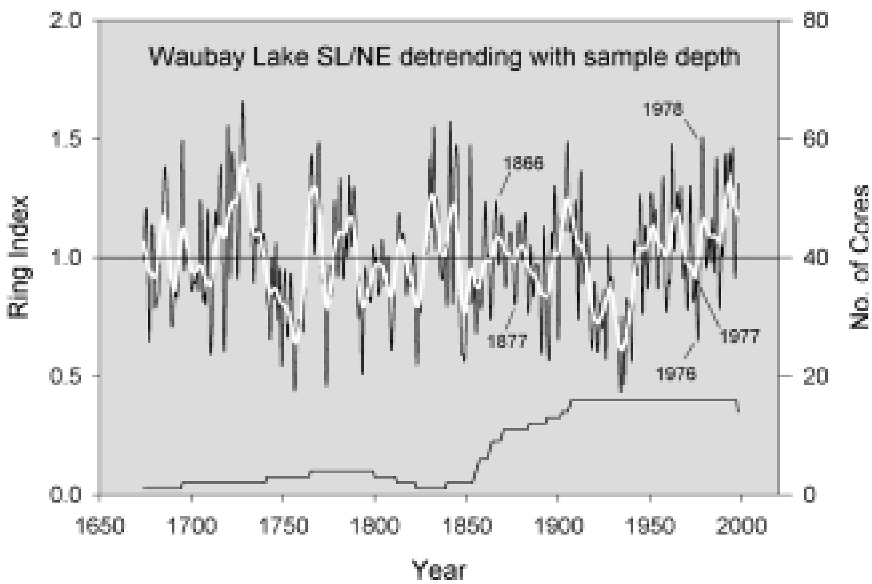

Figure 4. Waubay Lake chronology w. 10-yr spline curve and number of cores for each year (sample depth).

Ring width and precipitation records

The ring-width patterns of the chronology correlate well with precipitation records and closely match known wet and dry periods. The correlation coefficient between ring-index values in the chronology and annual precipitation at Webster is positive (more precipitation, more growth) and statistically significant ($r=0.4911$; $p<0.0001$). The severe drought of the 1930s is clearly evident in the chronology as a trough of the lowest growth rates during the 20th century (Fig. 4). The known low rainfall years in the 1970s appear as a less extreme trough in growth. Single years in the chronology also match up with known precipitation extremes. For example, a very low ring index value for 1976 coincided with the driest year this century (19.6 cm of precipitation; Webster weather station) (Fig. 4). The next year (1977) was very wet (76.2 cm) and initiated a two-year response of increasing ring width. In 1978, ring index was one of the largest for this century (Fig. 4). Finally, the series of years with consistently wide growth rings in the mid-1990s corresponds to some of the wettest years in the climate record. Overall, the ring-width patterns consistently reflect the known precipitation patterns in this century, supporting use of the chronology to identify wet and dry periods prior to climate records. The portion of the chronology using only Fort cores, however, needs to be interpreted with caution because of the small sample size and short overlap with the modern cores.

An Army officer's correspondence and scattered weather records from Fort Sisseton in the 1870s provide an additional check on the reliability of the chronology. Capt. Clarence E. Bennett, corresponding with the Adjutant General of the Army on November 23, 1878, noted "I am informed that when the Post was located [1864] it was surrounded by a chain of lakes and the road to the Post was through water, over a bar, connecting two of the lakes—now that road is high & dry far above high water mark. These lakes are many of them now dry, and grass grows luxuriantly, where once, for years & years were lakes … Since the establishment of this Post, these lakes have been steadily drying up … These Lakes have fallen perpendicularly 12 feet or more in the last twelve years." Capt. Bennett reports rainfall (exclusive of snow) from 1874-1878 to have been 40.64, 37.64, 38.81, 45.92 (below average), and 60.81 cm (average), respectively. These records indicate a series of moderately dry, not exceptionally dry, years up to 1878. The chronology (Fig. 4) indicates increasing landscape wetness in the few years leading up to the establishment of the Fort followed by a series of years of increasing landscape dryness, much the same as suggested in Capt. Bennett's report. The tree ring data indicate neither extreme wetness prior to establishment of the Fort as occurred in the 1990s nor extreme drought in the 1870s as severe as in the 1930s. The Waubay Lakes chain surveyed by the government in 1868-1877 had relatively high water levels (comparable to those in 1994; Table 3). However, the steadily declining water levels mentioned by Capt. Bennett for small lakes near Fort Sisseton from the late 1860s through the late 1870s match declining growth rates of bur oak trees in the chronology.

Wet/Dry Periods

The smoothed chronology indicates the occurrence of 5 wet periods over the 325 year period, i.e., average frequency of one extended wet period every 65 years (Fig. 5). The wet period peaks are nearly evenly spaced. Each of the peaks appears to have been wet enough to have produced lake levels well above average. Peak 4 (Fig. 5) is the least in breadth and height, followed with increasing strength by Peak 3 (peak split by several years of only moderate wetness), and Peak 2 (interrupted more strongly mid-way by a short drought). Peaks 1 and 5, the largest peaks, share certain characteristics. Both curves have wide breadth (40-50 years), sprinkled by occasional dry years but

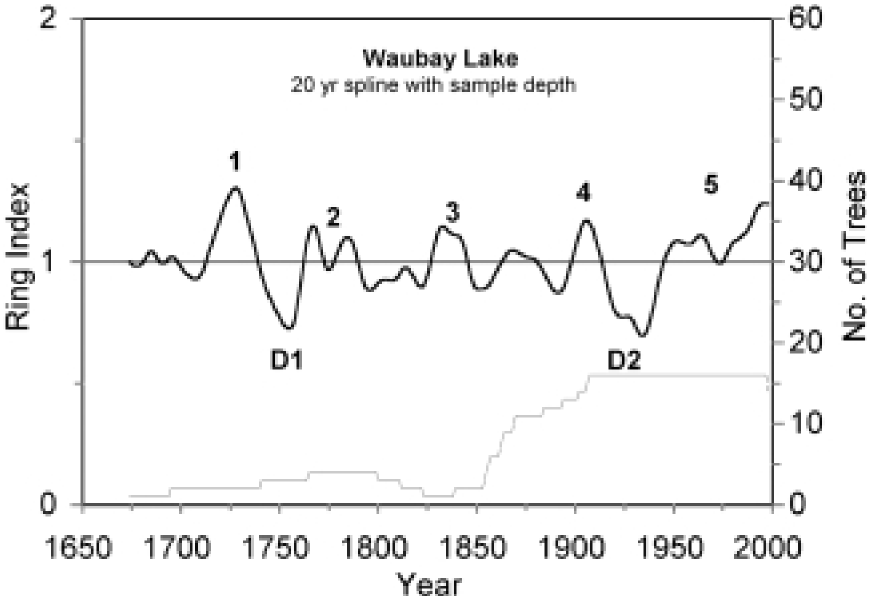

Figure 5. Waubay chronology 20-year spline fit and number of cores for each year (sample depth). Numbers 1-5 refer to wet periods. D1 and D2 refer to drought periods.

not disrupted by intervals of dry years. Maximum growth rates are higher in Peak 1, but Peak 5 has greater breadth.

Only Peak 1 resembles the duration and intensity of Peak 5 (known to have produced extreme lake flooding). Thus, twice in the 325 year chronology have conditions possibly been wet enough and long enough to have produced extremely high lake levels. This translates into minimum average frequency of once in 163 years. Two strong drought periods are evident in the chronology. The years with lowest growth rates in the chronology correspond to the 1930s drought period. However, a similar period of low growth rates occurred between 1740-1760 (Figs. 4 & 5). The drought indicated in the mid-1700s may have been longer (more consecutive years of below average moisture) but perhaps less intense (fewer extreme values) than the 1930s drought. Thus, the average frequency of severe droughts on the scale of the 1930s drought is one event in 163 years, the same calculated for extreme wet periods.

Comparison between chronology and regional lake levels

An interesting and informative comparison is made between the ring-width chronology and the lake-level history of Devils Lake, North Dakota. Devils Lake is a large prairie lake on the *Coteau du Missouri* approximately 200 miles northwest of Waubay Lake. Like Waubay Lake, it has risen dramatically in the 1990s flooding farmland, roads, farms, and towns. Continuous water-level records for Devils Lake go back to 1901, much farther than for Waubay Lake (1960).

The water-level data from Devils Lake form a single, flattened v-curve spanning the 20th century (Fig. 6). The water level was high at the start of the record, but fell steadily to a record low about 1940, after which the lake level began to rise. The rise has been progressive except for two short-term declines in the 1960s and late 1980s. The steepest rise in lake levels occurred in the 1990s.

The low-frequency trajectory of the tree ring chronology tracks that of the water level of Devils Lake (Fig. 6). Growth was highest early and late in this century, with a deep trough of record slow growth bottoming out in the late 1930s. The consistently highest tree growth this century was in the 1990s, coinciding with the period of rapid rise in the level of Devils Lake. The post-1930s pattern of tree growth is unprecedented in the 325 years of the chronology. For 60 years, environmental conditions were increasingly favorable for tree growth. Certainly there were scattered dry years and a short dry period in the 1970s; however, the consistent upward trend in growth is unmistakable (Figs. 6 & 7). The trees indicate an increasingly favorable soil moisture balance, most probably in subsoils and shallow aquifers, during this extended period.

The current flooding in Day County has focused attention on the high rainfall years of the 1990s. Certainly, these record wet years are a large factor because it was the recent wet weather that pushed lake levels beyond the historically known limit. However, the tree growth data point out something not so evident from the weather data alone, namely, that the climate had been providing an increasingly more favorable moisture regime during the 60 years following the 1930s drought. Although only a surrogate, this result suggests that antecedent conditions of steadily rising water levels, higher subsoil moisture, and slow aquifer recharge may have increased the risk of flood damage following an extremely wet period. Examination of long-term records from shallow wells may support this conclusion.

The increasing moisture favorability indicated by tree growth since the 1930s drought may indicate the advent of a wetter climate. The EPA reports on its Global Warming Website (www.epa.gov/oppeoee1/globalwarming/index.html) that precipitation has increased by about 1 percent over the world's continents in the last century. In

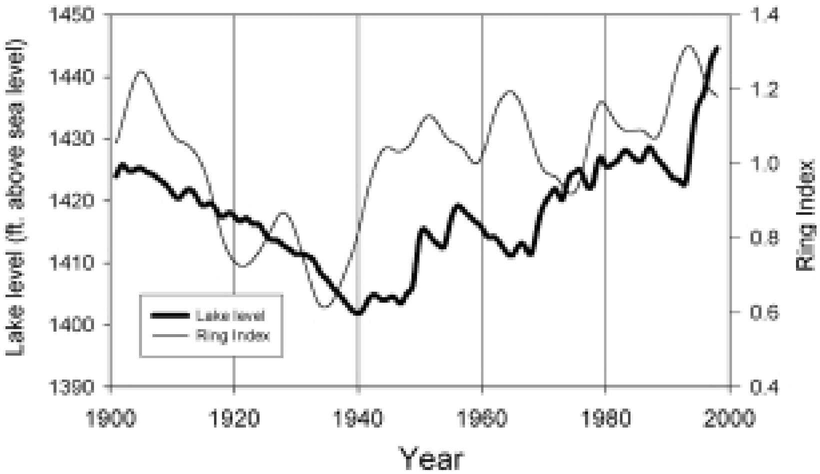

Figure 6. Historic Devils Lake water levels compared to Waubay Lake tree-ring chronology (10-yr spline, 1901-1998).

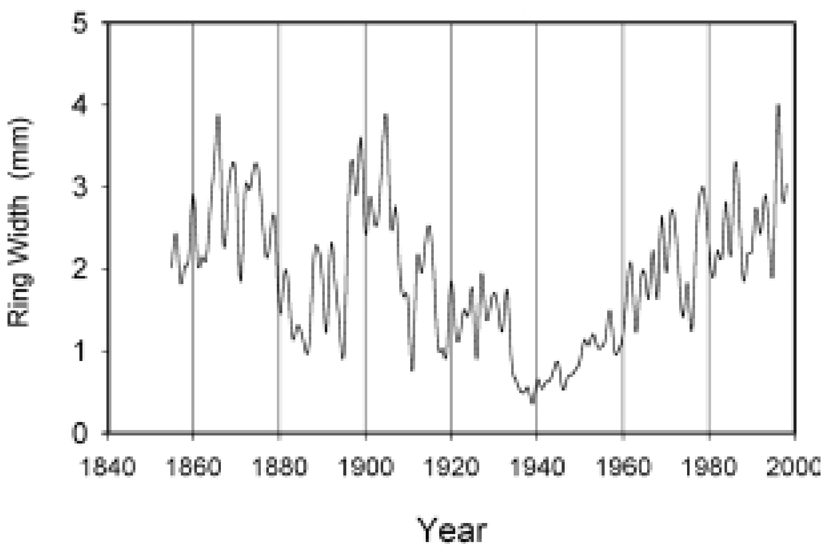

Figure 7. Unadjusted ring widths for core no. 44 (West Woods Island, Waubay Lake, 1855-1998) showing increasing growth rate from 1940-1998.

North America, precipitation has increased by an average of about 5 percent in the last century. Along the northern tier of states, rainfall has increased 10-15 percent. Much of the increase in rainfall has been occurring between September and November and is more concentrated in heavy downpours. If we are indeed in a wetter climate (rather than just in a wet phase within the normal climate), the climate record may seriously underestimate the persistence of current high water levels and future flood frequencies of the Waubay Lakes chain.

Regional patterns

The striking similarity between the South Dakota tree ring chronology and North Dakota lake levels in the 20th century suggests considerable regional uniformity in climate. Diatom inferred salinity indices from Devils Lake also reflect high water in 1725 and low water in the early 1750s (Fritz *et al.* 1994), matching conditions for these years in the Waubay chronology. The similarity continues south of Waubay Lake, based on correlation coefficients (r=0.55) calculated between tree-ring chronologies from Waubay Lake and Lake Herman (Madison, S.D.) (Sieg and Meko 1995); the latter is approximately 160 km south of the former. Finally, the Waubay chronology correlates well (r=0.58) with annual July Palmer drought severity indices (PDSI) derived from tree rings in other locales and extrapolated to northeastern S.D. by Cooke *et al.* (1996).

Lake Sediments

Core chronology

LOI results show the Spring Lake core to be composed of 10-30% organic matter, 25-35% carbonate minerals, and 40-55% inorganic matter (primarily detrital silicates). The organic content increases irregularly up-core, beginning well below the onset of European settlement (64 cm), rising sharply from 20 to 30% in the uppermost 16 cm (Fig. 8). The cause of the change in organic content is uncertain, but the small-scale variations are reproduced exactly in the overlaps between core sections, allowing for detailed adjustment of field-measured core depth. The LOI stratigraphy indicates a 2-cm displacement between drive-1 and drive-2 of the Livingstone core sections and a progressive compression of 2-6 cm toward the base of the surface core that resulted from piston leakage caused by sediment friction along the core-tube. Depth-adjusted core stratigraphies show almost perfect alignment of LOI profiles (Fig. 8). Ostracode analyses were performed on the upper 200 cm of core: 0-138 cm (adjusted depths) of the surface core and 6-70 cm (136-200 cm adjusted depths) of Livingstone drive-2.

The profile of total ^{210}Pb activity shows a monotonic decline from 12.2 pCi/g at the surface to a near-constant 0.83 pCi/g below 78 cm core depth (Fig. 9). Applying the constant rate of supply (c.r.s.) model to these data yields a basal date of 1849 ± 22 years at 78 cm and a sedimentation rate that varies between 0.06 and 0.12 g cm^{2} yr^{1}. Dating uncertainty increases sub-

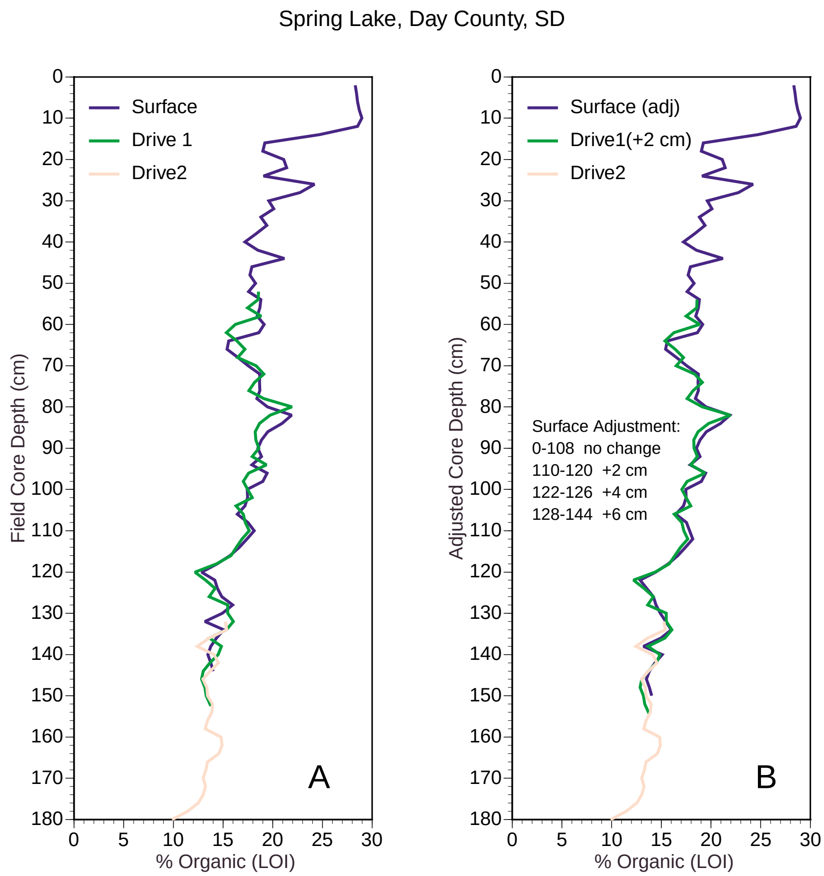

Figure 8 Loss-on-ignition profiles of % organic matter for the surface core and the upper two sections (drive-1 and drive-2) of the 4 m Livingstone core. Field depths (A) were adjusted for coring errors as indicated (B) to match stratigraphic detail at the overlap between core sections (see text).

stantially for strata older than about 1900, as total ^{210}Pb activity approaches supported values (0.83 pCi/g). Nineteenth-century sedimentation rates in particular show large error terms, making extrapolation to older strata fairly unreliable.

Despite such statistical uncertainties, results from pollen analysis provide strong independent support for the ^{210}Pb chronology. Abrupt changes in pollen percentages beginning at 64 cm core depth—in particular the decline in oak *(Quercus)* and other riparian forest types (elm, ash, birch) and the corresponding rise in grasses (Poaceae) and weedy taxa (*Ambrosia,* chenopods)—indicate the late-1800s spread of European agriculture into the Waubay region (Fig. 10). The most precise pollen marker is provided by Russian thistle *(Salsola),* an aggressive Eurasian weed that spread explosively across the northern Great Plains and arrived in Day county around 1885 (Jacobson and Engstrom 1989). Its appearance in the Spring Lake core at 64 cm corresponds to a virtually indistinguishable 210Pb date of 1886 ± 7.5 years (Fig. 9).

The chronology of pre-1850 sediments is provided by linear interpolation between the two AMS ^{14}C dates and the basal ^{210}Pb date at 78 cm (1849). The ^{14}C date at 148-150 cm (adjusted depth) is 710 ± 40 yr BP (AA37099), while that at 226-228 cm is 1250 ± 40 yr BP (AA37098). Calibration of these ^{14}C dates yields calendrical dates of 1287 AD (1235-1390 AD, ± 2 sigma) and 775 AD (685-877 AD, ± 2 sigma), respectively. Plots of calibrated date vs. depth or vs. cumulative dry mass (which corrects for changes in water content) indicate near-constant sediment accumulation rates between dated intervals: 0.14 cm yr^{1} or 0.044 g cm^{-2} yr^{1} (Fig. 11). These rates are substantially slower than the average accumulation rate for the ^{210}Pb dated section (0.51 cm yr^{1} or 0.11 g cm^{-2} yr^{1}). This apparent acceleration of recent sedimentation reflects both the upward increase in sediment porosity (a problem for linear accumulation rates only) and an acceleration of sediment flux to the lake caused by post-European land-use changes. We ultimately chose to use the dry-mass accumulation rate (0.044 g cm^{-2} yr^{1}) for interpolating between 14C dates, although the chronology provided by the depth/age model is quite similar.

Ostracode abundance

Figure 12 shows the semi-quantitative abundance of the 5 dominant ostracode taxa identified in the Spring Lake core. *Candona rawsoni,* a widespread species of the northern Great Plains having broad ecological tolerance, is the most common fossil taxa in most core intervals examined. This species is ubiquitous throughout the core, becoming relatively rare only in intervals where total ostracode abundance is very low (possibly due to poor shell preservation). Between 1000 AD and 1500 AD, *Candona ohioensis* comprises an important component of the fossil ostracode fauna. *Candona ohioensis* experiences three distinct peaks in abundance centered approximately on the years 1150, 1250, and 1400 AD. At the peak of these cycles, *C. ohioensis* shell abundance approaches and occasionally exceeds that of *C. rawsoni.* From 1500 AD to the present, *C. ohioensis* appears only rarely and in low abundance. *Limnocythere itasca* occurs at low to moderate numbers throughout most of the record.

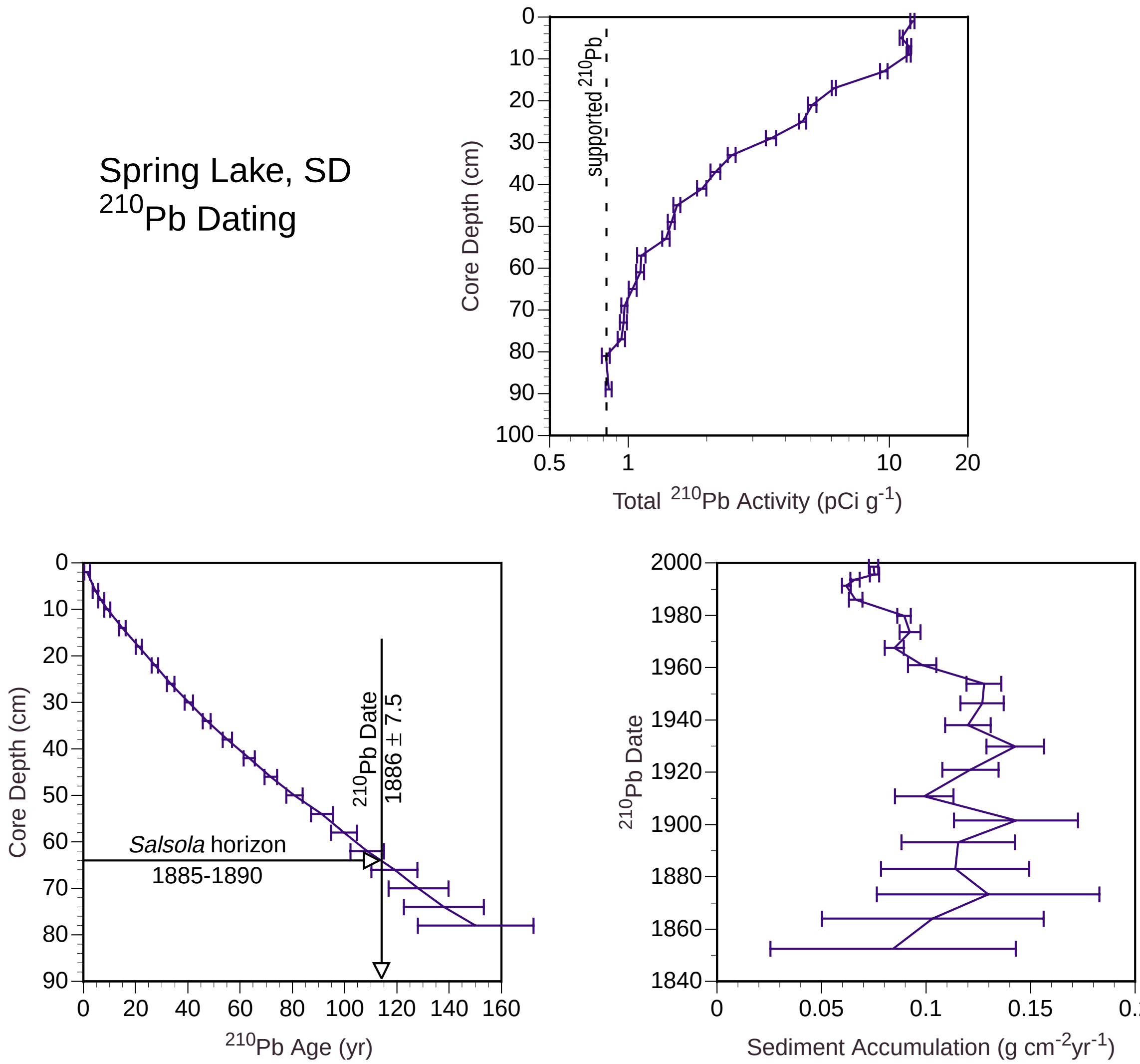

Figure 9. Lead-210 activity profile for the Spring Lake surface core and the resulting age/depth profile and sediment accumulation rates determined by the constant rate of supply (c.r.s.) mode; error bars represent ± 1 s.d. The depth at which Russian thistle *(Salsola iberica)* first appears in the pollen record corresponds to a ^{210}Pb date of 1886; Russian thistle spread throughout Day County between 1885 and 1890 (see text.)

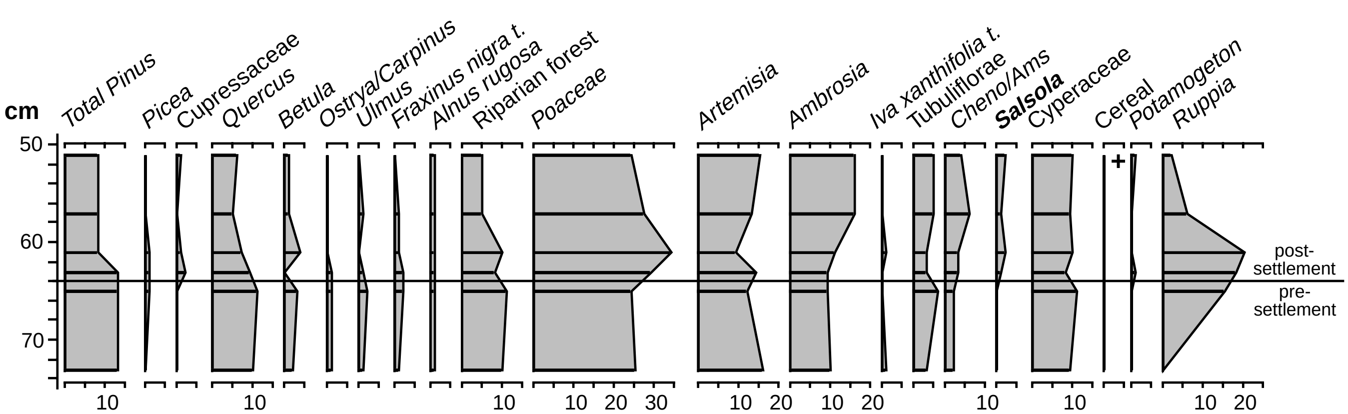

Figure 10. Pollen percentage diagram from the Spring Lake surface core showing changes in major pollen types that correspond to the arrival of European agriculture in the Waubay region.

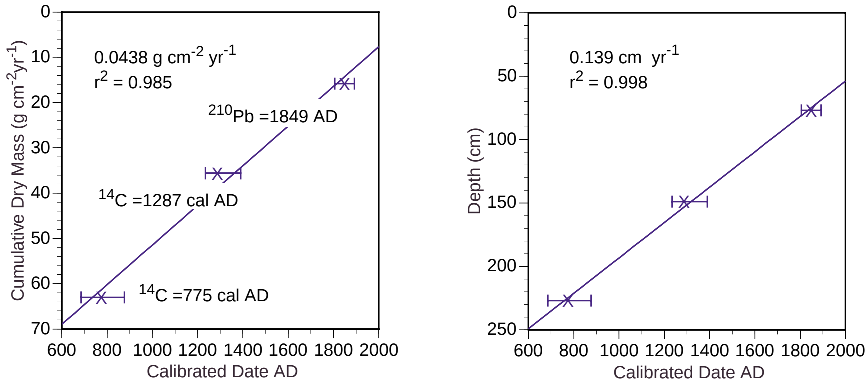

Figure 11. Linear regression models for dating interpolation between 14C-dated core intervals; radiocarbon dates have been calibrated to a calendrical time scale. The plot for cumulative dry mass corrects for changes in sediment porosity (water conent) and was used as the chronological model for stratigraphic interpretation.

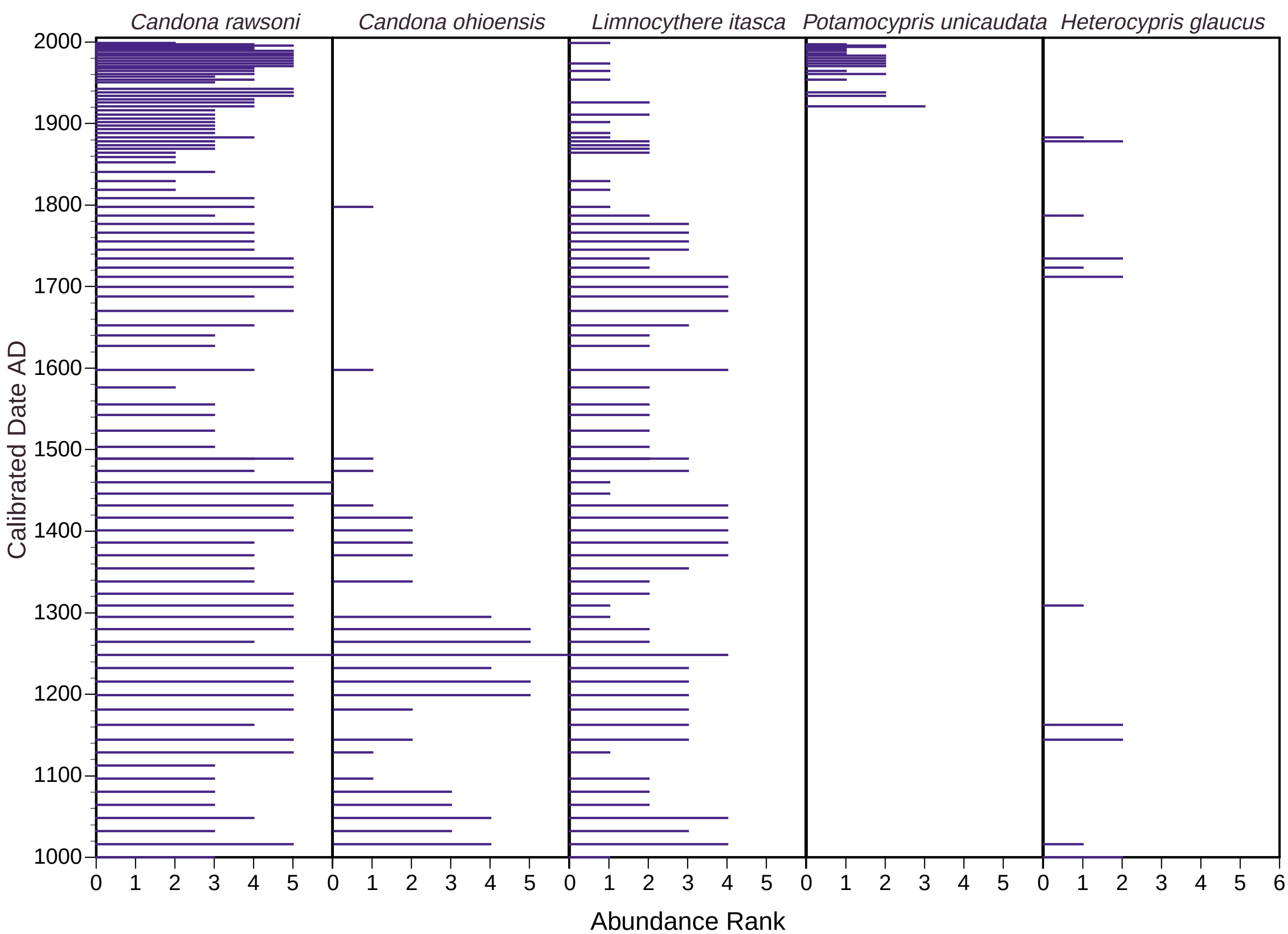

Figure 12. Abundance profiles for the five major ostracode species found in the Spring Lake core.

Between 1000 AD and 1500 AD, its abundance peaks coincident with that of *C. ohioensis.* After 1500 AD, when *C. ohioensis* essentially disappears from the fossil record, *L. itasca* displays another broad peak centered on ca. 1700 AD, coincident with a peak in *C. rawsoni* numbers and in total ostracode shells. From 1800 AD to the present, *L. itasca* occurs frequently but in low numbers. *Potamocypris unicaudata* is apparently absent from the record from 1000 AD to approximately 1900, after which it occurs at low abundances. *Heterocypris glaucus* occurs sporadically and at low numbers prior to the late 19th century.

Elemental shell chemistry

The molar ratios of Mg and Sr to Ca in the calcite shells of *C. rawsoni* are shown stratigraphically in Figure 13. From 1000 to 1500 AD, ostracode Mg/Ca expresses a sequence of well-defined fluctuations with a period of approximately 150 to 200 years. The three low-Mg/Ca intervals, centered on 1050, 1250 and 1400 AD correspond temporally with periods of increased abundance of *L. itasca* and *C. ohioensis* shells in the Spring Lake core (Fig. 14). Between c. 1500 and 1800 AD, the periodicity of Mg/Ca fluctuation appears to decrease to approximately 100 years and from c. 1800 to the present, the ostracode shell record shows even shorter period variation. Historically recorded droughts of the 1890s and 1930s are expressed as maxima in ostracode-shell Mg/Ca during those periods, while historical wet periods during the early 20th century and in the 1960s are reflected as minima in the Mg/Ca curve. Contrary to a straightforward model of lake dilution, however, the youngest intervals of the core (interpreted by ^{210}Pb as dating from the current highstand) do not show declining Mg/Ca values.

Ratios of Sr/Ca, plotted on an inverted scale (Fig. 13), display a negatively covariant relationship to

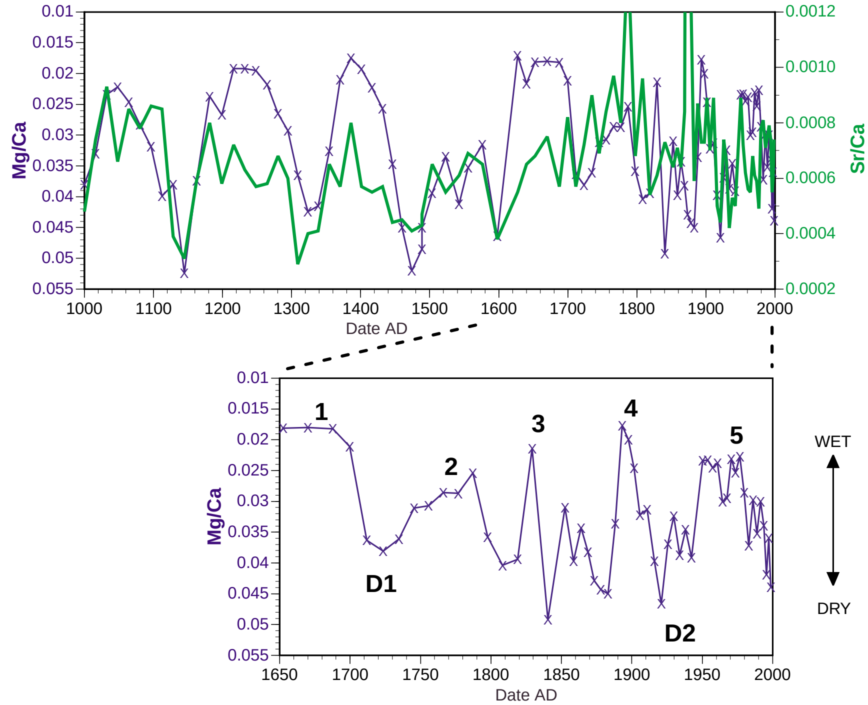

Figure 13. Stratigraphic changes in the trace-metal composition of calcite shells from *Candona rawsoni* in the Spring Lake core. The scale for ostracode Mg/Ca is inverted so that freshening events (wet intervals) are represented by positive excursions in both Mg/Ca and Sr/Ca. Major wet periods (1-5) and drought intervals (D1, D2) identified from local tree-ring data are compared with Mg/Ca events on the expanded time scale.

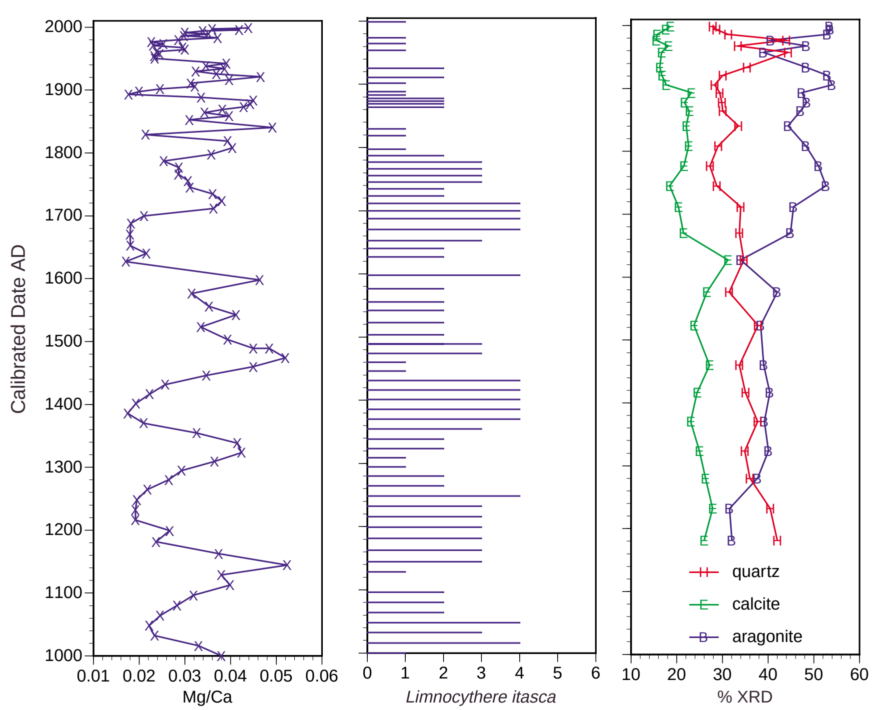

Figure 14. Mineralogical variations in the Spring Lake short core as determined by semi-quantitative XRD compared with profiles for ostracode Mg/Ca and the abundance of *Limnocythere itasca*, a generally fresh-water species.

Mg/Ca over most of the evaluated core. All of the 100-150 year Mg/Ca events from 1000 AD to 1800 AD show corresponding inverse Sr/Ca trends. From 1800 AD to the present, the inverse relationship is occasionally interrupted by episodes of coherent increase in Mg/Ca and Sr/Ca, notably during the wet period of the late 19th century.

Mineralogy

Figure 14 shows a semi-quantitative XRD plot of sediment mineralogy together with the record of ostracode Mg/Ca and *L. itasca* abundance. Aragonite (presumably endogenic) displays an overall increase in abundance over time (1180 AD to the present), while quartz (presumed to be detrital) displays a corresponding decline. Throughout the record the two minerals show an antiphase relationship indicative of variable detrital dilution of the endogenic carbonate. This pattern is particularly evident in the 20th century quartz maximum which likely reflects the combined effects of Dust Bowl climate and agricultural practices. The calcite component displays a more complex relation to the other two phases, in keeping with multiple origins of sedimentary calcite (endogenic, biogenic, and detrital). The low-resolution calcite curve from the Spring Lake core is not readily related to the other mineral phases analyzed nor to the indicators of climatic conditions provided by ostracode shells, although there may be some tendency for calcite to increase at the expense of aragonite during periods of low ostracode-shell Mg/Ca. The predominance of aragonite throughout the record indicates maintenance of relatively high lake-water Mg^{2+}/Ca^{2+}, even during periods of lake highstand and low salinities (Müller *et al.* 1972)

Water chemistry

The single water sample collected from Spring Lake in June, 1999 had a salinity of 1.65 ‰ and was dominated by Mg^{2+}, Na^{+}, and SO_4^{2-} ions, similar to other regional lakes. Molar ratios for Mg^{2+}/Ca^{2+} and Sr^{2+}/Ca^{2+} were 4.52 and 1.9 x 10^{-3}, respectively (Table 3). We found only limited additional data from prior years for Spring Lake and the surrounding water bodies. Field measurements of specific conductance on the Waubay Lakes system by the U. S. Geological Survey (USGS) in 1995 show Spring Lake to have been dilute relative to Lake Waubay early in the present lake highstand, prior to the coalescence of Waubay and Spring Lakes (Table 4). In 1995 the Spring Lake specific conductance of 1040 µS cm^{-1} was lower than that for Waubay (2490 µS cm^{-1}) and lower than the 2050 µS cm^{-1} we measured in 1999, following lake coalescence. Compiled specific conductance data for Lake Waubay dating back to 1971 show specific conductance values ranging between approximately 4000 and 11,000 µS cm^{-1}. Anecdotal information from the 1930s drought indicates that Spring Lake remained flooded and relatively fresh, suggesting the importance of groundwater through-flow to this lake during arid periods (Rothrock 1935).

Salinity and hydrologic reconstruction

The Spring Lake ostracode record provides an internally coherent, high-resolution reconstruction of hydrochemical changes in the Lake Waubay system. Our primary climate proxy, the Mg/Ca content of ostracode shells, is a well-established method for reconstructing salinity changes in closed-basin lakes based on both empirical and experimental evidence (Chivas *et al.* 1986, Engstrom & Nelson 1991, Holmes 1996, De Deckker *et al.* 1999). In semi-arid regions such as the northern Great Plains, changes in moisture balance directly influence lake salinities through the evaporative concentration or dilution of dissolved salts. Because Mg^{2+} behaves conservatively in moderately saline waters and Ca^{2+} is held constant by carbonate precipitation, lakewater Mg^{2+}/Ca^{2+} closely tracks salinity, and by extension hydrologic and climatic change. Ostracodes, in turn, incorporate Mg into their calcite shells in direct proportion to the

Table 3. Major ion Chemistry of Spring Lake, June 1999

Ion	*ppm*	*meq/L*
Na^{+}	100	4.36
K^{+}	46	1.19
Mg^{2+}	194	15.94
Ca^{2+}	71	3.52
Sr^{2+}	0.288	0.0066
Ba^{2+}	0.075	0.0011
Cl^{-}	35	0.98
SO_4^{2-}	776	16.16
HCO_3^{-}	424	6.96
Cations		25.01
Anions		24.09
Salinity	1.65 ‰	
Conductivity	2050 µS cm^{-1}	

Table 4. Specific Conductance of Waubay-region Lakes, October, 1995

Hillebrands/Waubay	2490 µS cm^{-1}
Swan Pond	2440
Spring Lake	1040
Rush Lake	461

Mg^{2+}/Ca^{2+} ratio of the water from which the shells are precipitated. Although Mg uptake can be described in thermodynamic terms—as influenced only by temperature and Mg^{2+}/Ca^{2+} of the host water—vital effects arising from the biotic nature of shell calcification complicate the picture and preclude a precise relationship between shell chemistry and lake salinity (Xia *et al.* 1997a, Wansard *et al.* 1998). Nonetheless, recent studies have demonstrated that ostracode Mg/Ca ratios closely track other robust salinity proxies such as diatom-based salinity inference models (Fritz *et al.* 2000).

In the Spring Lake core both ostracode Sr/Ca ratios and ostracode abundance data are coherent with the Mg/Ca record and its paleo-hydrologic interpretation. Ostracode Sr/Ca is negatively covariant with ostracode Mg/Ca in Spring lake because of preferential uptake of Sr by aragonite precipitation, which lowers lakewater Sr^{2+}/Ca^{2+} as salinity increases. An antipathetic relationship between ostracode Mg/Ca and Sr/Ca ratios has been noted in other sediment records from Dakota lakes in which aragonite is the dominate endogenic carbonate (Fritz *et al.* 1994, Haskell *et al.* 1996, Xia *et al.* 1997b). High abundances of *Limnocythere itasca* (and in some instances, *Candona ohioensis*), during low Mg/Ca events centered on 1050, 1250, 1400, and 1660 AD also support the interpretation of Mg/Ca minima as freshening lake highstands. Both taxa prefer less saline conditions than does *C. rawsoni* (Smith 1993).

Regional comparisons

The reliability of ostracode-shell chemistry in evaluating the frequency and severity of Spring Lake highstands hinges on (i) the coherence of the reconstruction with other regional records, and (ii) correlation with instrumental data and shorter-term proxy records from the immediate vicinity of Waubay Lake. Regional context is provided by the paleo-hydrologic synthesis of Fritz *et al.* (2000), wherein late Holocene salinity histories are compared among three North Dakota sites (Rice, Moon, and Coldwater lakes; Fig. 15). Each of these records constitutes a high-resolution reconstruction of lake chemistry during the past two millennia, obtained from analysis of fossil diatom assemblages and the chemical composition of ostracode shells. These records show considerable regional coherence in direction and duration of inferred climatic transients, with distinctions related to position along regional climatic gradients and hydraulic idiosyncrasies of the lakes.

The Spring Lake Mg/Ca profile is remarkably similar to these three records, especially those from Rice and Coldwater lakes. However, temporal correlation among individual lake records carries uncertainty; the standard error of AMS ^{14}C dates reported by Fritz *et al.* is 50 years, and the two AMS dates controlling our Spring Lake chronology carry standard errors of 40 years. Similarities in duration and frequency of geochemically important hydrologic events between Spring Lake and these other northern Great Plains records are nevertheless clear, and we can propose correspondence between particular Spring Lake events and those elsewhere in the regional record with some confidence. Sustained freshwater episodes (highstands) centered on 1250, 1400, and 1660 AD are particularly well expressed regionally, as are the sustained high-salinity events centered on 1330, 1480, 1600, and 1740 AD.

After c. 1800 AD, salinity events observed in the Spring Lake record decrease in period, a trend not obvious in the records examined by Fritz *et al.* (2000). A possible explanation for this change lies in the up-core increase in resolution (inherent in our shell sampling approach) coupled with the probable complexity of interactions between lakes of the Waubay Lake system during periods of water-balance transience (discussed below). Nevertheless, major salinity events of the post-1800 record (e.g., the low Mg/Ca event centered on c. 1890 AD and the Dust Bowl high Mg/Ca event) can be regionally correlated

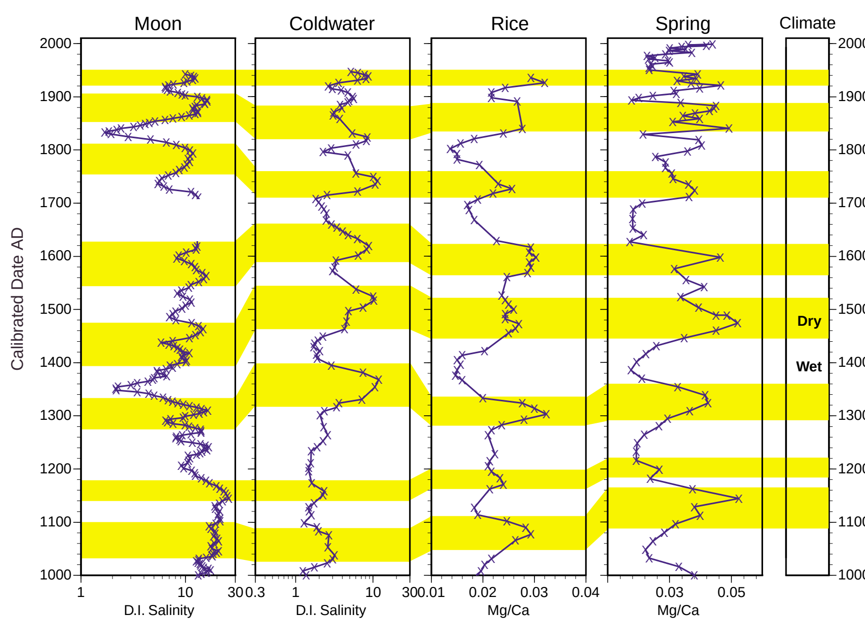

Figure 15. A comparison of the ostracode-Mg/Ca profile from Spring Lake with other high resolution paleo-hydrologic records from regional lakes, compiled by Fritz *et al.* (2000).

with geochemical evidence from the other paleohydrologic records shown.

Equally important, we can show good correspondence between the post-1650 AD Spring Lake salinity record and the local tree ring record of moisture. Figure 13 shows our proposed correlation to the tree ring-based moisture supply curve. From the mid-19th century to the present, extremes of moisture supply deduced from the tree ring record correspond to hydrologic events deduced from ostracode shell chemistry within the standard error of our ^{210}Pb chronological model. Between 1650 AD and 1850 AD, major events can be correlated with confidence, although chronological discrepancies increase somewhat, reflecting the uncertainty inherent in our AMS dates.

Basin coalescence and hydrological modification

The response of *C. rawsoni* shell chemistry to the highstand of the 1990s represents an apparent departure from the earlier record analyzed. Despite the historically high lake stage and volume achieved in 1999, the short core record of Mg/Ca maintains a generally upward trend (first established in the 1980s) through the most recent samples analyzed. These shells date from as recently as 1998, when the volume of Spring Lake had increased by a factor of 2.5 from that at the beginning of the 1990s stage increase (Niehus *et al.* 1999a).

We hypothesize that the initial rising-stage control of salinity and Mg^{2+}/Ca^{2+} in Spring Lake rests with a complex set of interactions between the waters of Spring, Hillebrands, Waubay, and Rush lakes and Swan Pond, which coalesced into a single interconnected water body in the late 1990s. Table 4 shows that as these lakes approached coalescence in 1995, they carried a substantial range of dissolved solids, reflected in the variable specific conductance shown. Thus when Spring Lake merged with the Swan and Hillebrands basins, more saline water probably bearing higher Mg^{2+}/Ca^{2+} began mixing with Spring Lake (the specific conductance of Spring Lake doubled between 1995 and 1999). Moreover, the introduction of water of higher alkalinity from the Hillebrands/Waubay system may have driven precipitation of calcium carbonate from Spring Lake waters, further enhancing Mg^{2+}/Ca^{2+} during lake mixing.

Over the same period, spill from the merged lake system to down-gradient Bitter Lake, which results in export of solutes and freshening of the Waubay system, was delayed due to the elevation control imposed by artificial barriers (Niehus *et al.* 1999a). In addition, the sequencing of water exchange between the adjoining lakes, each with it's own hydraulically determined salinization trend, has probably been altered by other barriers and conduits (Niehus *et al.* 1999a). Thus the short-term response of Spring Lake's chemistry may be contrary to a simple salinity-dilution model, and may also differ from the response exhibited prior to hydrologic modification of the Waubay Lake system. Calculated annual inflows for the period 1993 to 1999 are roughly a tenth of the recent lake system volume, suggesting that several years of spill and solute export would be required to equilibrate lake chemistry to a lake highstand. As spill continues, chemically less evolved inflow will result in decreasing lakewater Mg^{2+}/Ca^{2+} which will be reflected in decreasing Mg/Ca in ostracode calcite. It is also possible that similar short-term complexity in salinity response to earlier transients (e.g. the post-Dust Bowl rising lake stage of the 1940s) may be concealed by our sample resolution. Chronological uncertainty inherent in the record of earlier geochemical events precludes evaluating the possibility of such short-term effects.

Highstand frequency and duration

Overall correspondence with regional Late Holocene geochemical records and agreement with local tree-ring moisture history indicate that the Spring Lake Mg/Ca profile is a highly reliable record of the hydrologic status of the Lake Waubay system. The hydrologic complexities described above and the geochemical complications of ostracode shell magnesium uptake (Xia *et al.* 1997a) discourage quantitative interpretation of aqueous chemistry from these data. We interpret the geochemical record of the past millennium as providing strong evidence for a cycle of repeated and sustained highstands punctuated by periods of notable drought. Prior to 1800 AD, geochemically dilute conditions tended to persist for several decades or more, suggesting a pattern of wet cycles considerably longer than the present highstand to date. The consistency of Mg/Ca maxima achieved during the dilute events centered on 1050, 1250, 1400 and 1660 AD suggests that some hydrologic boundary condition, possibly the spill elevation of the Waubay lake system, was approached and maintained during these periods. The recurrence frequency of these pluvial peaks averages ~140 years prior to 1800 AD; hydrologic transients of the past two centuries appear to have been more frequent, with peak recurrences on the order of 60 years. At the same time, the duration of chemically inferred climatic cycles appears to have been considerably longer prior to 1800 AD. Thus the 1000 year lake-geochemical record suggests a peak recurrence of wet periods averaging 140 years, with four dilute periods interpreted to represent highstands of comparable volume and considerably longer duration than the present condition.

The multi-decadal to century-long nature of the dilute episodes prior to 1800 AD indicates that the annual probability of highstand conditions was higher during this earlier period than the probability reflected in the historical record. Approximately 300 years of the ostracode record appears to reflect dilute water-chemistry conditions, suggesting an annual probability of high lake-stand conditions around 0.3. This result contrasts with the stochastic water-balance modeling of Niehus *et al.* (1999a), which suggest lower probabilities and shorter duration of highstand conditions, based on analysis of instrumental lake-stage and meteorological data.

Shoreline and Geomorphological Indicators

Shoreline/channelway dating

Samples 1-9-1-99 and 2-9-1-99 are from an assumed beach deposit, the highest (elevation approximately 549.2 m) and therefore probably the oldest deposit identified in the Bitter Lake basin. Sample 2-9-1-99, from a depth of 0.3 m, yielded a radiocarbon age of 1730 +/- 110 years, which may be substantially less than the actual age of the beach deposit owing to recent additions of decayed plant roots. Sample 1-9-1-99, as expected, was nearly barren of pollen. If significant amounts of pollen had been recognized, disturbance by cultivation practices would have been indicated, further limiting the reliability of the radiocarbon date.

Samples 3-9-1-99 and 4-9-1-99 were collected from an assumed beach deposit, about a meter lower and 50 m nearer Bitter Lake than that of the first two samples. The samples were submitted for analyses to help confirm the assumption that all identifiable beach deposits signify the oldest lake level for that elevation, and that any beach deposit inundated and reworked by a subsequent high lake level could not persist as an individual geomorphic feature. Analytical results showed the samples to be barren of pollen and less than 100 years in age; these results are, of course, inconclusive but do not refute the above assumption.

Analyses of sample 5-9-1-99, from the presumed highest highest beach deposits above the east-northeast shoreline of Bitter Lake, yielded no pollen and a radiocarbon age of less than 100 years. The sample was obtained at a site that appeared to be undisturbed by recent agricultural activities but which was subject to additions of fresh carbon. Sample 6-9-1-99, collected from a probable beach deposit of intermediate position and younger age, about 20 m downslope from the site of sample 5-9-1-99, was likewise nearly barren of pollen and yielded a radiocarbon age of less than 100 years.

Sample 7-9-1-99 from lakebed sediment immediately overlying outwash was submitted for pollen analysis. The sample was extracted from a core recovered at a high but intermediate shoreline, in the southwestern part of the Bitter Lake basin (Kim Kempton, NRCS, personal commun., 1999). Although pollen concentration in the sample was high, Dr. Anderson could identify only 127 pollen grains due to deterioration. The sample was dominated by pine *(Pinus)* (66.9 %), a pollen type tentatively identified as *Shepherdia* (11.2 %), fungi (8.7 %), and Cyperaceae (6.3 %); also present were spruce *(Picea),* Cruciferae, and Onagraceae. Spores were common. Owing to its stratigraphic position above glacial outwash, it is inferred that pollen of this sample is representative of early lacustrine conditions, shortly after glacial retreat and formation of Bitter Lake.

Pollen analyses were conducted on sediment samples 8-9-1-99 and 9-9-1-99, which were extracted from higher parts of the core that yielded sample 7-9-1-99. Sample 8-9-1-99 represents the uppermost sub-aqueous deposits of the core, and sample 9-9-1-99 was interpreted to be from beach deposits of Bitter Lake and thus may be equivalent to samples 3-9-1-99, 4-9-1-99, and 6-9-1-99. No preserved pollen was observed in sample 8-9-1-99. Pollen was rare in sample 9-9-1-99, but grains of pine *(Pinus),* sagebrush *(Artemisia),* ragweed *(Ambrosia),* and *Chenopodium-Ambrosia* types (principally the goosefoot family) were identified by Dr. Anderson.

Sample 10-9-1-99 was collected from a beach deposit about 20 m upslope from the core samples 7-9-1-99, 8-9-1-99, and 9-9-1-99, and was interpreted to represent the highest preserved shoreline of Bitter Lake. The samples, possibly equivalent to samples 1-9-1-99, 2-9-1-99, and 5-9-1-99, gave a radiocarbon age of 380 +/- 100 years but yielded only a single grain of ragweed *(Ambrosia)* pollen and a single spore. As for sample 1-9-1-99, the radiocarbon age may be much younger than the true age of the deposit.

Core samples 2-9-7-99, 4-9-7-99, and 5-9-7-99, respectively, were from depths of 0.6, 1.2, and 1.5 m at potential spill site 1, southwest of Bitter Lake; the first two are interpreted to represent glacial outwash and the third to be glacial till. Pollen concentrations in sample 2-9-7-99 were very low, only five grains being recovered—one of ragweed *(Ambrosia)* and four of the sunflower family (Asteraceae). Sample 4-9-7-99 was small in volume, which, combined with a laboratory accident, resulted in no reported pollen. Only seven pollen grains of pine *(Pinus),* sagebrush *(Artemisia),* and sunflower (Asteraceae) were recovered from sample 5-9-7-99. Carbon in sample 4-9-7-99 yielded an age of 2770 +/- 230 years. This age, which is due to a continuing influx of plant carbon, is assumed to be a minimum, coupled with the inference that the deposit is glacial outwash, suggests that streamflow from the Bitter Lake basin to the Big Sioux River has not occurred at this site since deglaciation.

A second core was recovered from potential spill site 1. Slopewash, colluvium, and possibly eolian sand extended from the surface to a depth of about 2.0 m; from 2.0 to 2.6 m was sand, possibly outwash, which overlay till. Samples 9-9-7-99, 11-9-7-99, 15-9-7-99, and 18-9-7-99, from depths of 0.6, 1.2, 2.4, and 3.4 m, respectively, were analyzed for pollen. Preserved pollen was not recognized in samples 9-9-7-99 and 15-9-7-99, and in sample 11-9-7-99 only a few grains of sunflower (Asteraceae) and dandilion (Liguliflorae) were recovered. Sediment barren of pollen is consistent with an outwash depositional mode. A full pollen count suggesting "some antiquity" (R. S. Anderson, NAU, written commun., 2000) was obtained from sample 18-9-7-99, which was inferred to be of glacial-till origin. Dominating the assemblage was pine *(Pinus)* (33.7 %). Pollen of other tree types was common and included spruce *(Picea)* (6.7 %), aspen *(Populus)* (8.3 %), ash *(Fraxinus)* (1.6 %), and birch *(Betula)* (2.8 %). Other pollen taxa identified were juniper *(Juniperus)* (9.1 %), hazelnut *(Corylus)* (4.4 %), *Shepherdia* (tentative identification), and members of the evening primrose family (Onagraceae). Sample 15-9-7-99, presumed to be of outwash origin, yielded a radiocarbon age of 8090 +/- 40 years that, as a minimum age, may reflect a depositional environment of outwash during deglaciation.

Lake salinity

Historical and recent records, geomorphic considerations, and modeling results (Niehus *et al.* 1999a,b) provide substantial evidence that most or all major lakes of the Waubay chain except Bitter Lake periodically have had water levels high enough to cause overflow. The water level of Rush Lake, the nearest lake to Bitter Lake, recently reached a height great enough to spill limited flows into Bitter Lake, but geomorphic and water-quality data (Petri and Larson 1968, Leap 1988, Steuven and Stewart 1996), as well as climatic interpretations based on lake cores, suggest that this condition has probably been unusual during Holocene time. Because there appears to be no doubt that surface movement of water from one lake to another has occurred repeatedly (but at differing frequencies), attention here largely is restricted to whether overflow has occurred from Bitter Lake into the Big Sioux River drainage system, and if so, at what frequency.

Occasional high levels and overflow from lakes of the Waubay chain above Bitter Lake do not suggest that spillage occurred from Bitter Lake during the same episodes of high lake surfaces. Owing to a much larger basin area for Bitter Lake than the other lakes of Day County, a substantial volume of water can flow from Rush Lake into the Bitter Lake basin before spillage of Bitter Lake is feasible. This extensive surface area provides the means by which a significant amount of water is lost by evaporation and perhaps infiltration to the ground-water system before spillage can occur. Direct evidence of long-term water loss by evaporation from the Bitter Lake basin is provided by water-quality data showing that concentrations of dissolved solids (in mg/l) in water samples collected from Bitter Lake are typically 4 to 60 times greater than of water of the higher lakes. The lowest lake of the Waubay chain (excepting Bitter Lake), to which other lakes contribute water at times of high levels, is Waubay Lake, which generally has concentrations of dissolved solids 2 to 15 times greater than higher lakes of the chain. It is inferred, therefore, that spillage from Rush Lake (the next highest lake) into Waubay Lake has been unusual during Holocene time.

Photographic and map analyses

Examination of photographs and maps describing sites 1 and 2 suggested that discharge has not occurred along either potential flow path in recent centuries or millennia. Evidence includes occurrences of marshes and ponded water of apparent glacial origin, irregular valley sides that are not indicative of fluvial processes, and especially elevational differences along the flow paths too great to be explained as thalweg variations (such as those of pool-riffle sequences).

Numerous sites for possible sediment sampling were identified on aerial photographs. Of those not inundated in recent years, most were along ancient beach deposits and at spillage sites 1 and 2. One in particular, on a point of land extending as a peninsula or island into the middle of Waubay Lake (T 123 N, R 55 W, sec. 34), is significant because of preserved shorelines. The peninsula/island exhibits well-defined shoreline features as high as approximate elevation 549.6 m; shorelines could not be recognized at higher elevations of the island, which has an estimated maximum elevation of 552.9 m. At this shoreline elevation of about 549.6 m and slightly lower, stored water (for pre-development conditions) overtopped divides separating Blue Dog, Waubay, Rush, Little Rush, and Bitter Lakes, thereby forming one large lake with a shoreline of common elevation.

Channelways

Under recent high lake-level conditions, flow paths, especially for several relatively high lakes such as Enemy Swim Lake and Pickerel Lake, have conveyed overflow, have developed gravel beds, and thus have well defined stream-channel shapes. The same is not as apparent for spill paths between several of the lower lakes. Prior to development, Blue Dog Lake was separated from the Rush Lake complex only by marshy lowlands, and similar separations occurred between Waubay, Hillebrands, and Spring Lakes and between Minnewasta Lake and Rush Lake. The channelway from Little Rush Lake to Bitter Lake has the appearance of being occupied by flowing water only recently and otherwise only infrequently; well defined channel features are lacking. In general, field observations suggest that water levels infrequently, if not rarely, are high enough to cause spillage among the lower lakes.

Both from aerial photography of 1939 and field observations, it is obvious that the lower lakes have had a range of water levels and that numerous beachlines have formed in response to the fluctuations in lake level. Inspection of the aerial photography and supporting field evidence, however, indicate that no shorelines higher than an elevation of about 549.6 m have been preserved in the Bitter Lake/Waubay Lake complex of lake basins (the lower lake basins).

The two channelways identified as possible flow paths (sites 1 and 2) from Bitter Lake to the Big Sioux River were examined for evidence of flow features and fluvial landforms. In both cases, the hummocky sideslopes and gently undulating bottomland deposits of the channelways appear to be dominated by glacial till and sandy outwash presumably deposited at the time of glacial retreat; except where depressions remain in the glacial-till landforms, no lacustrine sediment suggestive of levels of Bitter Lake between 549.6 and 552 m was recognized. Irregular sideslopes of the channelways maintain a relict morphology of glacial-till geology and exhibit no evidence of modification by flowing water. Small lakes and marshes occupy the channelways, which are interpreted to be remnant water courses that conveyed meltwater during de-glaciation, but no landforms along the channelways have the appearance or sediment characteristics of either modern or ancient flood plains. Thus, field examination of the two channelways is consistent with an interpretation that significant spillage from Bitter Lake has not occurred at sites 1 or 2 since deposition of glacial till and outwash.

Pollen analyses

Sandy deposits, such as beach sediment and outwash, often have poorly preserved pollen if water passing through the deposit is aerated and causes oxidizing conditions. Sediment with significant amounts of clay and silt, however, inhibits water movement, helps protect the organic content, and may yield well preserved pollen. Lacustrine deposits and glacial till typically have a range of particle sizes and often, therefore, contain well preserved pollen. Most of the samples submitted for pollen analysis for this investigation contained little or no preserved pollen, conforming to interpretations of the depositional origin of the sediment. Samples 7-9-1-99 and 18-9-7-99 were identified as lakebed and till deposits, respectively (C. A. Niehus, U. S. Geological Survey, written commun., 1999) and contained sufficient pollen to allow interpretations. Both samples contained pine *(Pinus),* spruce *(Picea),* and other pollens suggestive of middle-Wisconsin conditions; the pollen of neither sample is "typical of Holocene nor late glacial pollen assemblages in the region" (R. S. Anderson, NAU, written commun., 2000). In support of this conclusion, late-Wisconsin sediment of Pickerel Lake, to the north, is dominated by spruce *(Picea)* pollen but is nearly barren of pine *(Pinus)* pollen (Watts and Bright, 1968); a similar pattern was observed for Moon Lake, in North Dakota, about 200 km (120 miles) north of Bitter Lake (Laird *et al.* 1996a). Pine *(Pinus),* however, appears to have been common in the northern Great Plains during mid-Wisconsin time (R. S. Anderson, written commun., 2000), and its presence in samples 7-9-1-99 and 18-9-7-99, in turn, indicates an age of perhaps 15,000 to 18,000 years. This conclusion is in agreement with glacial geology studies of Flint (1955). Radiocarbon dates of nearly 3000 years and about 8000 years for samples 4-9-7-99 and 15-9-7-99, both interpreted to be outwash, do not confirm this interpretation but are consistent if new plant carbon was added to the sediment after deposition.

Lake spillage

A likely key to understanding whether Bitter Lake has spilled in the past or will in the future is that the natural (pre-development) spillway level between Rush Lake and Bitter Lake is approximately 548.3 m (R.D. Benson, USGS, personal communication, 2000); according to topographic maps of the area, natural spillway levels between most of these lakes range between 547 m and 549 m, the lowest being between Rush and Waubay Lakes. Only after Blue Dog, Rush, Spring, Hillebrand, Minnewasta, and Waubay Lakes have merged into a composite lake at the 549.74-m level can water spill into Bitter Lake. Thus, the upper lakes act as a single input rather than as separate controls on the water level, area, and volume of Bitter Lake.

Of perhaps greater importance is the recognition that if Bitter Lake fills to a level of about 548.3 m, further inflow from higher lakes cannot occur but that the large single-lake basin then includes that of Bitter Lake. When it occurs, this lake level, 548.3 m, of the entire complex of lakes is probably controlled mostly by the balance between direct precipitation falling on the lake and evaporation (including transpiration from associated marshes and shallow-water areas) from the lake surface. If so, this control is likely one of strongly stabilizing the hydrology (especially movement of water), area, and volume of the composite lake. To result in a significantly higher lake level than 548.3 m, a large excess of direct precipitation relative to evapotranspiration would seem essential. Consistent with this viewpoint are field observations showing that the highest recognized beach deposits rimming various lakes (including Bitter Lake and Waubay Lake) are at an elevation of approximately 549 m. Thus, the shoreline-elevation data suggest (1) that water levels of the lake complex never significantly exceeded an elevation of 548.3 m, and (2) that following glacial retreat, the deposition of outwash during glacial melting, the formation of the Bitter Lake basin, and the stabilization of lake levels through a balance between precipitation and evaporative losses, spillage of water from Bitter Lake at a spillway level of about 552 m to the Big Sioux River has not occurred.

The above interpretations are supported by water-quality data. The accumulation of salts in the water of Bitter Lake has been far greater than that of the other lakes that can spill into Bitter Lake, indicating that long-term water losses from Bitter Lake have occurred principally by evaporation. Elevated concentrations of dissolved solids of Waubay Lake, combined with the water-quality data of Bitter Lake, suggest that mixing of water during periods of high lake levels and the formation of a single composite lake have occurred infrequently or rarely since formation of the Coteau des Prairies.

Data are insufficient to ascertain whether the inferred maximum lake level of slightly more than 548.3 m occurred more than once after glacial retreat. The radiocarbon date of 1730 +/- 110 years for beach-sand sample 2-9-1-99 seems strongly indicative that the most recent, or only, high lake stand occurred significantly earlier than 17 centuries ago. Published results of other studies suggest that a period of moist Holocene climate conditions in the Northern Great Plains occurred 2600 to 4400 years ago. Because numerous remnants of the 548.3 m shoreline(s) have not been badly eroded or destroyed by weathering, but remain easily recognized, an age much greater than 4400 years seems improbable. Thus, the period during which the most recent or only high-water stand occurred in the Waubay chain of lakes may have been roughly 3 to 4 millennia ago.

The combined geomorphic observations, geologic and climatic information, water-quality data, and pollen and radiocarbon analyses provide persuasive evidence (1) that Bitter Lake has not spilled to the Big Sioux River since glacial retreat, and (2) that the highest lake stand, resulting in preserved shorelines around several lakes, following glacial melting was at an elevation of slightly more than 548.3 m. Whether this lake level was attained more than once cannot presently be determined, but, based on a single radiocarbon date and results of other studies, it is speculated that preserved beach sand may have been deposited between 2600 and 4400 years BP.

CONCLUSIONS

Wet and dry periods indicated by the tree ring chronology closely match those indicated by the lake core results. Periods of wide growth rings occurred during periods of high lake levels. The long-term highstand frequency of 163 years derived from the tree ring chronology over a 325 year record was similar to the approximate 140 year frequency derived from the lake sedimentrecord for a 1,000 year period. These studies indicate that under long-term mean conditions, a person would have a 50:50 chance of experiencing a highstand peak event during their lifetime, assuming an average lifespan of about 75 years. Data from both tree rings and lake sediments support this estimate.

The prolonged dilute episodes (century length) in Spring Lake prior to 1800 AD indicates a higher annual probability of high-stand conditions than during the period of historical record. Approximately 300 years of the ostracode record appear to reflect dilute water-chemistry conditions, suggesting an annual probability of high lake levels of approximately 0.3. The generally low lake levels that have occurred since white settlement but before the recent flooding, in addition to the post-1930s pattern of steadily increasing water availability and favorableness for tree growth, are not typical of the long-term record. Analyses based on available weather and lake stage records from this period (Niehus *et al.* 1999a,b) may have underestimated the long-term probability and duration of high-stand conditions.

The frequent highstand conditions in the past apparently were reached with little lake spillage. The geomorphic evidence from the channel connecting Rush and Bitter lakes indicates infrequent and only small flow events. The outlet from Bitter Lake, the terminal lake in the Waubay Lake system, appears not to have functioned since the start of the Holocene. Thus, hydrologic controls have kept this lake system often nearly full of water but rarely overflowing. One control mechanism could be the increasing evaporation that occurs as Waubay and Bitter Lakes expand in area during filling. Deluges during the last millenium simply have not been enough to overwhelm the forces of evaporation in these large, shallow basins.

The paleoecological methods used in this study cannot, however, discriminate between highstand conditions that reach flood proportions and those that only approach them. For example, we could not determine if the prolonged high water conditions seen repeatedly in the lake sediment record would have flooded farms, towns, and roads, had they been present at the time. This is a particularly difficult question to answer, since the presence of our human infrastructure itself must affect lake levels. The extreme nature of the flooding in the 1990s may have been caused, or at least worsened, compared to pre-settlement conditions, by past wetland drainage, greater runoff from agricultural lands in the basin, and by the blocking or slowing of flow by hundreds of miles of roadbeds. Although highstand conditions in the basin were not unusual in the past, spillage was, suggesting that the "hand of man" may be contributing to contemporary spillage and flooding.

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