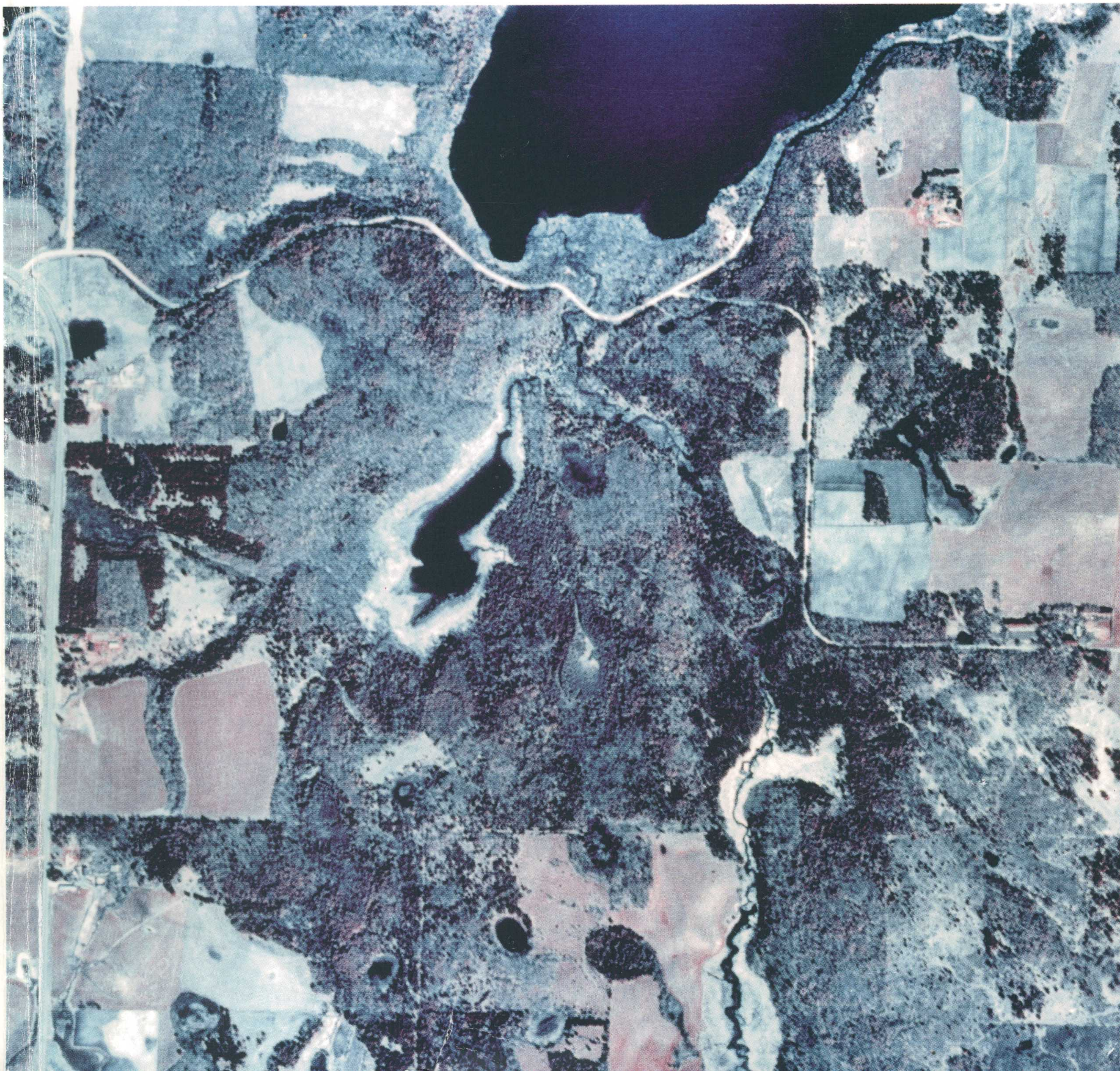
HYDROLOGICAL AND BIOGEOCHEMICAL RESEARCH IN THE SHINGOBEE RIVER HEADWATERS AREA, NORTH-CENTRAL MINNESOTA

U.S. GEOLOGICAL SURVEY
Water-Resources Investigations Report 96–4215



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Edited by Thomas C. Winter

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History and Status of the U.S. Geological Survey Interdisciplinary Research Initiative in the Shingobee River Headwaters Area

By Robert C. Averett and Thomas C. Winter

The availability and quality of water is a fundamental concern to mankind. Water commonly is not available where it is most needed, and its quality commonly is not suitable for use without treatment. Furthermore, land and water developments commonly degrade the quality of water even more. Increased development has resulted in a growing concern about water supply and the general condition of aquatic environments. Historically, most water management has concentrated on surface water or ground water as if they were separate resources. However, development of either water resource eventually affects the other. The growing realization that the hydrologic system is essentially one resource has made it imperative that scientists from many different earth-science disciplines work together to determine the interactions of water with earth materials and organisms throughout the landscape. The goal of working together, however, is often elusive because discipline barriers are difficult to surmount. In some cases, scientists prefer to work with those in their own fields. In other cases, research follows discipline lines because of organizational structure and funding sources. The result is that many scientists from different disciplines do not come in frequent contact with each other. When scientists from different disciplines work at a common site, they commonly continue to work within their discipline, requiring a coordinator to integrate the results. This approach does little to break down discipline barriers. Less commonly is a concerted effort made to focus on a problem, where scientists from different disciplines are strongly encouraged to interact with each other from the outset of the study. Recognizing that most environmental problems require integration of many disciplines in order to understand the linkages between processes and to develop management solutions, a decision was made among U.S. Geological Survey (USGS) scientists to provide an incentive to foster closer interdisciplinary research, in which an attempt to integrate the research would be made from the outset and throughout the study.

In late 1987, a group of USGS scientists formed an Interdisciplinary Research Initiative (IRI). The group prepared a prospectus that had as its major theme that the management of the Nation's water resources requires knowledge of how atmospheric water, surface water, and ground water function as integrated systems. Fundamental attributes for the research initiative included (1) the need for fundamental research on hydrologic and biogeochemical processes related to some of the most pressing environmental problems facing the Nation, (2) selection of ecosystems that had comparative aspects, and (3) a long-term perspective. The fundamental study unit was to be a watershed, or watersheds, that included close interaction between physical, chemical, and biological aspects of atmospheric water, surface water, and ground water. The watershed concept ties directly to USGS expertise in the study of the hydrologic cycle.

A basic premise of the IRI was that participating scientists would become involved on a voluntary basis and that they would provide their own funding. There were to be no constraints placed on the number of scientists participating, their discipline, or on duplication of effort. Indeed, a strength of the effort would be to provide the opportunity to do comparative evaluations of concepts, methods, and models. Annual meetings would be held to keep participants informed on research underway and planned.

It was decided to focus initially on lakes because they integrate many hydrologic processes that are present in their watersheds and they preserve a record of past environmental change in their sediments. The organizing principle was to select one or more watersheds having two lakes with contrasting water residence times, the underlying hypothesis being that many processes in lakes are affected by the water residence time. Lakes having a short residence time generally have stream inputs and outputs and, therefore, are dominated by external fluxes. In contrast, lakes having a long residence time generally are closed and, therefore, are dominated by internal cycling. The contrasting water residence times were hypothesized to result in systemwide differences at all physical, chemical, and biological levels in the lake ecosystems and are, therefore, the impetus for the comparative aspects of the IRI effort. Topics that could be studied stressing interlake comparison include, but would not be limited to, dissolved organic carbon, alkalinity, lake hydrodynamics, composition of the plankton community, population turnover rates, and oxygen dynamics. While the unit of study was to be the watershed, a major focus would be on the lakes.

In the spring of 1988, a nationwide search was conducted to find a study area for the IRI effort. Offices of the USGS and selected universities were sent a list of selection criteria and were asked to propose candidate field sites. Water residence time was listed as the main criterion, but size of the study area and access to the study area were also important constraints to facilitate convenience and affordability. From 61 proposed sites across the United States, the Shingobee River Headwaters Area in north-central Minnesota was selected. This headwaters watershed contains seven lakes, but Williams Lake and Shingobee Lake best fit the selection criteria. Williams Lake has no stream inlet or outlet; it interacts only with the atmosphere and with ground water. Shingobee Lake has a stream inlet and outlet; therefore, the lake interacts with surface water as well as with the atmosphere and with ground water. Water residence time is about 3 years for Williams Lake and about 6 months for Shingobee lake. The location of the study site is shown on plate 1.

Williams Lake had been studied by several members of the USGS for more than a decade, and a continuous data base, as well as instrumentation to measure atmospheric water, surface water, and ground water variables, was in place. Shingobee Lake had not been studied previously. Climate stations and stream gages were established in the vicinity of Shingobee Lake in 1989. Sensors to measure water temperature, air temperature, vapor pressure, and windspeed were placed on a raft in the center of the lake. Sensors to measure long-wave radiation, short-wave radiation, air temperature, vapor pressure, windspeed, and wind direction were placed at a land station near the lake. A stage gage was placed on the lake, and stream gages were placed on the inlet and outlet streams.

Research on Williams and Shingobee Lakes, the Shingobee River, and a fen wetland has been underway since the spring of 1989. Presently, about 20 scientists participate in the IRI effort; most are from the USGS, but others are from the University of Minnesota, Twin Cities; University of Minnesota, Duluth; the University of California, Davis; Michigan Technological University; and the Oak Ridge National Laboratory. To focus the research initially, the participating scientists identified three major study priorities for the IRI effort. These are (1) processes associated with the carbon budgets of the two lakes, (2) physical, chemical, and biological interactions that formed and maintain Little Shingobee Fen, and (3) physical, chemical, and biological interactions that take place between the Shingobee River and its hyporheic zone. A site manager, supported by the USGS, was hired in 1989 to make routine and specialized measurements and sample collections and to manage the general data base.

This document provides a synopsis of the IRI effort, including a description of the area, work underway, and highlights of progress. The papers herein represent only a brief overview of the work underway as of 1995. A goal of this document is to provide information so that others might see the opportunity to complement the present activities and to join the IRI effort.

Glacial Landscape Evolution of the Itasca/St. Croix Moraine Interlobate Area Including the Shingobee River Headwaters Area

By Howard D. Mooers and Arthur R. Norton

INTRODUCTION

The Shingobee River Headwaters Area lies within a physiographic region known as the Itasca/St. Croix moraine interlobate area (Norton, 1983) (fig. 1). The glacial history of this area is particularly complex, and the geomorphology and sedimentology provide a record of glacier dynamics, processes of ice stagnation, and the development of an extensive glacial karst system (Mooers and others, 1990; Wright, 1993).

The Itasca moraine is 160 km long, 20 to 30 km wide (fig. 1) and composed of as much as 250 m of unconsolidated sediment overlying bedrock. This large end moraine marks a lengthy stillstand of the Itasca phase of the Wadena lobe of the Laurentide ice sheet during Late Wisconsin time (Wright and Ruhe, 1965).

The St. Croix moraine of the Rainy lobe extends from its interlobate junction with the Itasca moraine southward to central Minnesota (Wright, 1972; Wright and Ruhe, 1965). Although a prominent geomorphic feature of Minnesota's landscape, the St. Croix moraine is much narrower and has considerably less relief than the Itasca moraine (fig. 1).

The glacial geology of the interlobate junction of the two moraines was the focus of an investigation by Norton (1983). The area is characterized by a myriad of glaciofluvial sediments and landforms including ice-contact sands and gravel, collapsed outwash plains, coalescing outwash fans, outwash terraces, and subglacially carved tunnel valleys (Norton, 1983).

Other than the investigation of Norton (1983), few detailed studies of the glacial geology of this area have been undertaken. The general sequence of glaciation of the region is discussed by Wright (1972), Wright and Ruhe (1965), and Matsch and Schneider (1986). Seaberg (1985) conducted an investigation of the hydrogeological characteristics of the glacial sediments in the area immediately surrounding Williams Lake. Several regional studies were completed for portions of northwestern and central Minnesota. Several lithostratigraphic studies describe and correlate sedimentary units and regional Late Wisconsin glacial events (Anderson, 1976; Harris, 1975; Martin and others, 1989, 1991; Perkins, 1977; Sackreiter, 1975). The area south of the Itasca moraine was the focus of two regional investigations of stratigraphy, sedimentology, and glacial history (Goldstein, 1985; Mooers, 1988).

Recent regional (Dobbs and Mooers, 1990; Mooers and others, 1990) and more detailed local investigations (St. George, 1994) have led to significant revisions of the general glacial history of this area. The discussion below focuses on a general description of the Itasca/St. Croix moraine interlobate area and summarizes the results of recent investigations.

MORPHOLOGY AND SEDIMENTOLOGY OF THE ITASCA / ST. CROIX MORAINE INTERLOBATE AREA

The Itasca moraine is highest along its central axis, which trends west-northwest to east-southeast. From the central crest the moraine slopes downward in both the upglacier (north) and downglacier (south) directions (fig. 1). A short distance to the west of the

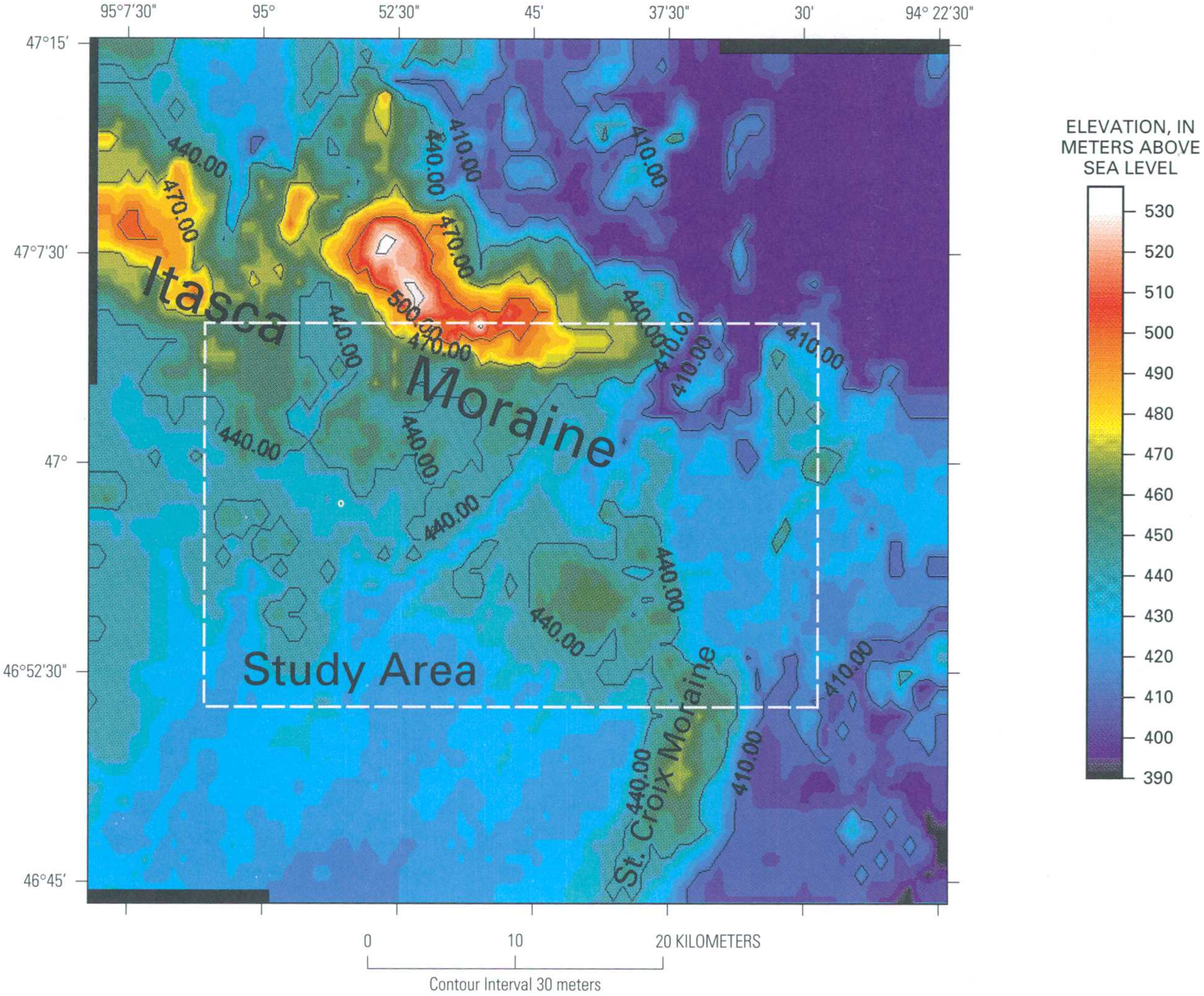

Figure 1. Topography of a portion of the Itasca moraine and vicinity. Itasca/St. Croix moraine interlobate area of figure 2 is indicated by dashed line. Location of study area is shown on figure 3.

study area, the crest of the Itasca moraine reaches an elevation of 605 m, 150 to 200 m higher than terrain a few km to the north or south. The topography is extremely rugged and poorly drained, with hundreds of lakes and thousands of smaller ponds and wetlands.

The crest divides the moraine into two regions of contrasting morphology and sedimentology. The proximal (northern) part of the system is a hummocky terrain composed largely of flow tills, ice-contact sands and gravels, isolated ice-walled lake plains, inverted deltas, and small, isolated outwash plains. This region is very poorly drained, and there is no evidence that a well-developed surface drainage system existed during formation of the moraine.

The distal facies of the moraine also consists of hummocky terrain but is composed of basal and superglacial till, flow till, and glacial thrust systems capped by a thick sequence of collapsed glaciofluvial sands and gravels (Mooers, 1990; Mooers and others, 1990; St. George, 1994). This part of the moraine is also characterized by stagnation features, but the landforms are strikingly different from those of the proximal facies. An integrated drainage network consisting of large, kilometer-wide surface channels and chains of ice-walled lake plains connected by rivers served as drainageways for glacial meltwater. Numerous north-south-trending, subglacially-carved tunnel valleys are now expressed as long valleys often containing chains of lakes (Wright, 1993).

The St. Croix moraine ranges in width from 5 to 7 km and is composed of hummocky stagnation topography similar to the proximal facies of the Itasca moraine; the relief is much less, however. Sedimentary assemblages include interbedded gravel, sand, till, and lacustrine sediments.

SURFICIAL GEOLOGY OF THE ITASCA/ST. CROIX MORAINE INTERLOBATE AREA

A landform-based surficial geologic map of the Itasca/St. Croix moraine interlobate area was compiled from the work of Norton (1983), Mooers (1988), Mooers and others (1990), Wright (1993), and Mooers (unpub. data) (fig. 2). Landforms identified by their distinctive morphology and internal sedimentology and stratigraphy are classified by the parent glacial lobe, the phase of the parent lobe (Wright and Ruhe, 1965), and the landform type.

The topographically higher central axis of the Itasca moraine trends eastward from Mantrap Lake through Lake Alice (fig. 2). The crest then trends toward the southwest and runs along the west side of Shingobee Lake where it turns toward the southeast. Williams Lake lies approximately on this crest. The hummocky topography on either side of the crest comprises the most extensive morainic unit (Wis, fig. 2). Much of this ice-stagnation complex consists of flow till, ice-contact sand and gravel, and lacustrine sediments in complexly interbedded sequences. In addition to this ice-stagnation complex, much of the moraine is composed of laterally extensive collapsed outwash deposits. It is difficult, however, to trace these outwash plains to specific ice margins because of their highly collapsed nature, resulting from the meltout of an underlying relatively continuous blanket of glacial ice.

A striking group of subparallel ridges ($\mathrm{Wis_f}$ on fig. 2) that form a local high occurs in the north-central part of the study area within the ice-stagnation complex. Based on their morphology, these features were identified as glaciotectonic features and referred to as transverse compressional ridges (Sackreiter, 1975). However, the ridges are composed of coarse-grained glacial outwash and appear to be glaciofluvial in origin (S.L. Emery, unpublished data, 1993).

The distal portion of the Itasca moraine complex is composed mainly of collapsed outwash (pitted outwash) ($\mathrm{Wio_{ch}}$, fig. 2). In this area an extensive blanket of sand and gravel, locally more than 30 m thick, was deposited over slowly moving or stagnant ice near the glacier margin. Subsequent meltout resulted in the hummocky nature of this landform. This geomorphic assemblage differs from the collapsed outwash deposits within the stagnation complex (Wis), as the surfaces of the outwash plains can be more easily reconstructed. Within the collapsed outwash, large supraglacial outwash channels can be identified ($\mathrm{Wio_{ch}}$, fig. 2).

The St. Croix moraine consists predominantly of hummocky stagnation topography (Rss), and collapsed outwash composes a relatively small portion of this landform assemblage. Glacial thrust features are present within the St. Croix moraine ($\mathrm{Rss_t}$, fig. 2) (Mooers, 1988, 1990) and are expressed as hill-hole pairs similar to those described by Moran and others (1980). One located a few km east of the study area contains blocks of local bedrock (Mooers, 1990). A series of thrust systems associated with Stony Lake, Pleasant Lake, and Four Point Lake have been named the Pleasant Lake moraine (fig. 2), a minor recessional moraine of the Rainy lobe (Mooers, 1988).

Along the outer margin of the Itasca and St. Croix moraines is a series of large, coalescing outwash fans forming the Park Rapids (Wio) and Oshawa (Rso) outwash plains (Norton, 1983). The particle size of the sediment within these outwash deposits decreases rapidly away from the head of the fans.

Throughout the area, conspicuous chains of lakes outline the numerous tunnel valleys, which represent an englacial or subglacial conduit system that served to drain meltwater from the ice surface.

The Crow Wing outwash terrace occupies a broad valley extending from Leech Lake toward the southwest along the Crow Wing chain of lakes (fig. 2). The development of this outwash plain and its significance are discussed below.

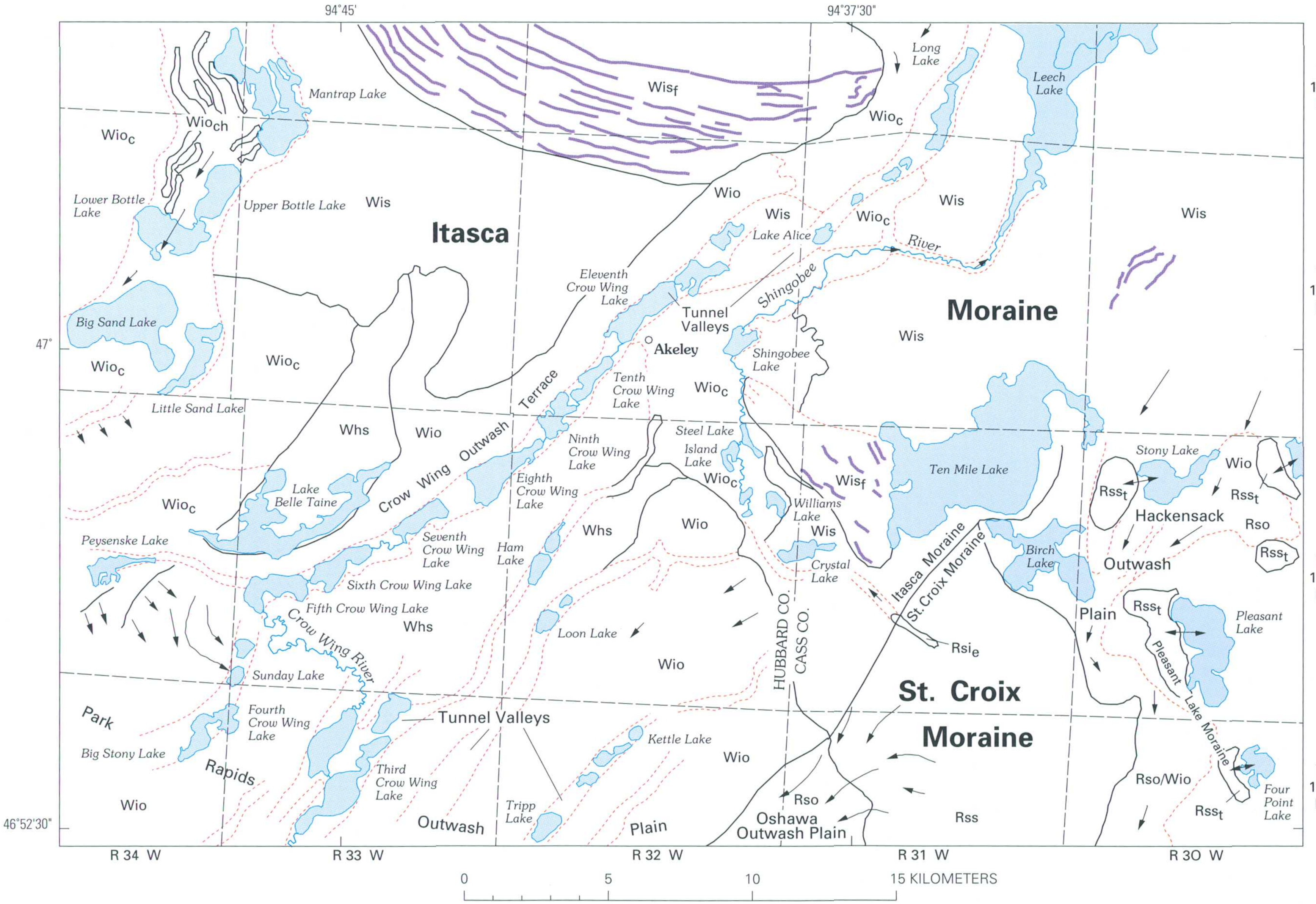

Figure 2. Surficial geologic map of the Itasca/St. Croix moraine interlobate area. Map area corresponds to inset on figure 1. (Compiled from Norton (1983), Mooers (1988), Mooers and others (1990), Wright (1993), and Mooers (unpub. data)).

EXPLANATION

QUATERNARY	Itasca Phase	Wis	Hummocky topography with local relief of 10 to 50 meters deposited in an ice-marginal setting. Hummocks are composed predominantly of subglacial and supraglacial meltout till much of which has been redeposited by mass movement (flow till). Lacustrine sediment deposited in ice-walled lake plains, ice-contact sand and gravel, and glaciofluvial sediment form locally important components of this sedimentary assemblage.
		$\mathrm{Wis_f}$	Glaciofluvial complex characterized by multiple subparallel ridges, Landform assemblage has been extensively modified by collapse during meltout of stagnant ice.
		Wio	Proglacial fluvial sediment of the Park Rapids and Hackensack Outwash Plains, primarily sand and gravel, poorly to moderately well sorted. The surface slopes gently away from the ice margin and an anastamosing pattern of the channels and bars of the glacial braided stream system can be identified on aerial photographs.
		$\mathrm{Wio_c}$	Proglacial fluvial sediment that was deposited over stagnant glacial ice. The surface is characterized by hummocky topography from collapse during the melting of the buried ice. Locally extensive lacustrine sediments composed predominantly of silt.
		$\mathrm{Wio_{ch}}$	Systems of collapsed meltwater deposits associated with the discharge of large volumes of meltwater onto the surface of the ice-cored morainic system.
	Hewitt Phase	Whs	Hummocky topography with local relief of 10 to 20 meters deposited in an ice-marginal setting. Hummocks are composed predominantly of subglacial and supraglacial meltout till and collapsed outwash.

St. Croix Phase	Rss	Hummocky topography with local relief of 10 to 50 meters deposited in an ice-marginal setting. Hummocks are composed predominantly of subglacial and supraglacial meltout till much of which has been redeposited by mass movement (flow till). Lacustrine sediment deposited in ice-walled lake plains, ice-contact sand and gravel, and glaciofluvial sediment form locally important components of this sedimentary assemblage.
	$\mathrm{Rss_t}$	Glacial thrust systems within the hummocky morainic complex. Generally expressed as hill-hole pairs, the thrust systems contain multiple displaced blocks of drift and bedrock.
	Rso	Proglacial fluvial sediment of the Oshawa and Hackensack Outwash Plains, primarily sand and gravel, poorly to moderately well sorted. The surface slopes gently away from the ice margin and an anastamosing pattern of the channels and bars of the glacial braided stream system can be identified on aerial photographs.
	Rsi	Ice-contact sand and gravel. Glaciofluvial systems associated with supraglacial, englacial, or subglacial depositional environments.
	$\mathrm{Rsi_e}$	Eskers

———	Boundaries of geomorphic units, dashed where indistinct or inferred.
- - - - -	Glacial drainage ways, including the traces of tunnel valleys, which often follow the trend of chains of lakes.
━━━	Linear trends within the glacial landforms.
↙	Arrows indicate the direction of glaciofluvial transport.

LATE WISCONSIN GLACIAL HISTORY OF THE ITASCA/ST. CROIX MORAINE INTERLOBATE AREA

The modern glacial history was established largely through the work of Wright (1954; 1957a; 1957b; 1964; Wright and Ruhe, 1965). The first extensive glacial advance of Wisconsin age was that of the Hewitt phase of the Wadena lobe. The ice advanced to the southwest through the prominent topographic low northeast of the Itasca/St. Croix Moraine interlobate area. The ice may have formed much of the Alexandria moraine complex at this time (Wright, 1972), although Goldstein (1985, 1989) identified a belt of stagnation topography along the northeastern margin of the Alexandria moraine in central Minnesota that may mark a recessional position of the lobe during formation of the Wadena drumlin field. Because of the abundance of Paleozoic limestone in the drift, the source of the Wadena lobe was thought to be the Winnipeg lowland to the northwest (Wright, 1957b). The orientation of drumlins in the large Wadena drumlin field (fig. 3) is inconsistent with a northwestern source region, so Wright (1957b) postulated that the advancing Wadena lobe was diverted to the southwest by an ice mass in northeastern Minnesota. Recent workers have presented evidence that the Wadena lobe originated in the Hudson Bay area (Goldstein, 1985, 1989; Martin and others, 1989; Gowan, 1993). The source of the carbonate is still problematic; Goldstein (1985, 1989) suggested that the carbonate was derived by erosion of underlying carbonate-rich drift, whereas Gowan (1993, p. 84) could find no geochemical evidence for mixing. Limestone is the predominant component of the Paleozoic stratigraphy in the Hudson Bay basin, and geochemical analyses of the drift matrix are consistent with a northeastern provenance (Gowan, 1993).

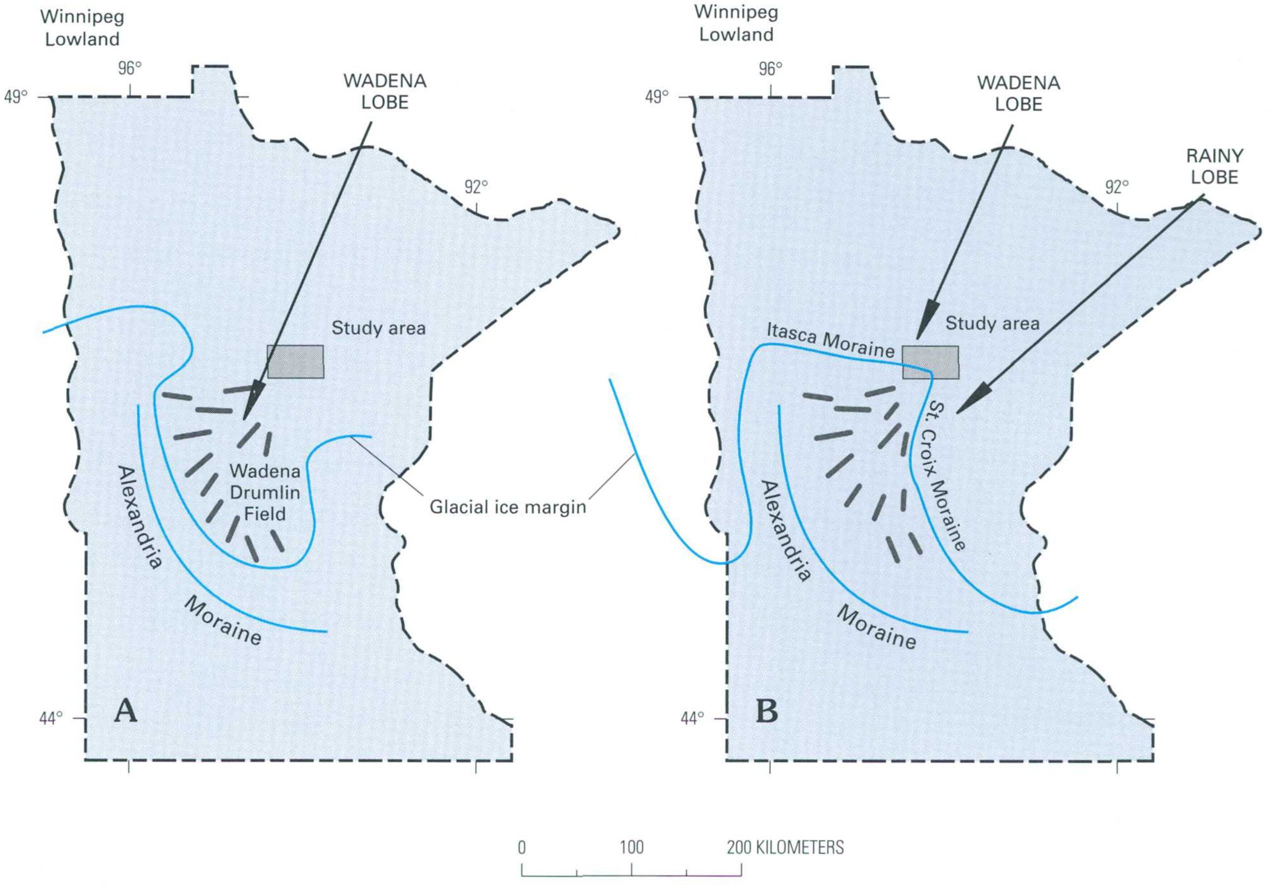

Figure 3. (A) Hewitt phase of the Wadena lobe, (B) Itasca/St. Croix phases of the Wadena and Rainy lobes.

The age of the Hewitt phase is uncertain (Clayton and Moran, 1982; Matsch and Schneider, 1986; Wright, 1993); however, an abundance of stagnant ice related to the Hewitt phase was buried by outwash from the subsequent Itasca and St. Croix phases of the Wadena and Rainy lobes (Goldstein, 1985; Norton, 1983), suggesting that the events may not have been separated by a lengthy period. Landforms described by Goldstein (1985, 1989) indicate that the Wadena lobe may have stagnated and wasted vertically downward rather than undergoing a systematic retreat following the Hewitt phase. In any case the next glacial phase is the advance and stabilization of the Wadena and Rainy lobes at the Itasca and St. Croix moraines (fig. 3), respectively. Wright (1972) suggested that the ice advance during the Itasca/St. Croix phase occurred about 20,500 years before present (BP), whereas Clayton and Moran (1982) suggested a somewhat younger date of about 15,500 years BP.

Following the Hewitt phase, stagnant ice remained in the Itasca/St. Croix moraine interlobate area and was subsequently buried by outwash deposited during the Itasca/St. Croix phase. The tunnel valleys identified in the interlobate area extend beyond the limits of the Itasca and St. Croix moraines; therefore, they were most likely associated with Hewitt phase glaciation (fig. 3). It is probable that the presence of stagnant ice of the Hewitt phase played an important role in controlling the flow direction of the Wadena and Rainy lobes during the Itasca/St. Croix phase and therefore controlled the position of the interlobate junction between the moraines.

The length of the stand of Wadena and Rainy lobes at the Itasca and St. Croix moraines is unknown. However, the Rainy lobe retreated while the Wadena lobe remained at its maximum limit (Norton, 1983; Mooers, 1988). Outwash from the Itasca moraine flowed southward east of the St. Croix moraine (Norton, 1983), and coalescing outwash fans from Wadena and Rainy lobes formed in the vicinity of Pleasant Lake (Mooers, 1988) (fig. 2). The Wadena lobe was apparently still at or near its maximum when the Rainy lobe had retreated to the Outing moraine 50 km to the east of the Itasca/St. Croix moraine interlobate area (Mooers, 1988).

The topographically high axis of the Itasca moraine represents the area of maximum accumulation of debris on the glacier surface and presumably the position of the active margin during maximum glacier activity (Mooers and others, 1990). As the active ice margin retreated northward, meltwater ponded between topographically high portions of the morainic axis and the active ice margin. These temporary glacial lakes commonly can be identified by the presence of fine-grained sediment, particularly silts, although the morphology of the lake basins was destroyed by subsequent meltout of stagnant ice. The Shingobee Lake area is characterized by fine-grained sediment that may represent accumulation in one such ice-marginal lake, and the areas of map units Wio_c and Wis in figure 2 contain numerous other examples of lacustrine sediments deposited by these temporary lakes.

The surface of the Wadena lobe, and to a lesser degree the Rainy lobe, drained by a series of englacial and subglacial meltwater channels (Mooers and others, 1990; Wright, 1993). This system of englacial and subglacial conduits was similar to subsurface channels found in karst terrains. This karst-like system continued to serve as an active subsurface drainage network during the retreat of the ice to more northerly positions within the moraine and probably served to drain water from the area for a substantial period of time after final retreat of glacial ice because of melting of buried ice. The courses of these conduits are readily identifiable as troughs that now contain chains of lakes (Wright, 1993).

During the late stages of the formation of the Itasca moraine, many of these englacial and subglacial conduits discharged onto the surface of the debris-covered stagnant ice of the distal portion of the moraine as fountains, similar to modern drainage systems on the Malaspina Glacier in Alaska (Gustavson and Boothroyd, 1987). Mantrap Lake represents the location of one such discharge area. Tunnel valleys lead into Mantrap Lake from the north (Wright, 1993) and an anastomosing network of surface meltwater channels lead away from the lake to the south (fig. 2).

The interlobate area became a focus for meltwater draining from the ice surface, resulting in the predominance of glaciofluvial sediment and landforms indicating large volumes of water (outwash plains, eskers, and tunnel valleys). The Crow Wing outwash plain drained meltwater from the Wadena lobe and from the melting stagnant ice within the Itasca moraine throughout the episode of moraine formation. This meltwater channel was also occupied after the ice retreated to the north of the Itasca moraine; it is the only large meltwater channel that traverses the moraine from north to south. The Crow Wing chain of lakes formed as blocks of buried ice melted following the abandonment of the meltwater channel.

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Physiographic and Geologic Characteristics of the Shingobee River Headwaters Area

By Thomas C. Winter and Donald O. Rosenberry

INTRODUCTION

Williams Lake was selected in 1977 for long-term study by the U.S. Geological Survey's Hydrology of Lakes Project because of its physiographic and climatic setting (Siegel and Winter, 1980). The selection of this lake was prompted by results of numerical models of hypothetical settings that indicated the interaction of lakes and ground water varied greatly depending on the geologic and climatic settings of the lakes. To evaluate the applicability of the model results to actual lakes, it was decided that field sites needed to be established in different hydrogeologic and climatic settings. Furthermore, the sites needed to be developed using a common conceptual framework and uniform methods to facilitate intersite comparison.

After consideration of a number of sites, Williams Lake was determined to be an appropriate setting because it is underlain by a geologic framework that was likely to have ground-water flow systems of local, intermediate, and regional scale, which is important in assessing the physical and chemical hydrologic processes associated with lakes. The lake and its drainage basin are underlain by thick deposits of glacial drift that were deposited by ice lobes from different sources (Mooers and Norton, this volume), resulting in the drift having contrasting hydraulic and mineralogic characteristics. Furthermore, the lake is located where precipitation is approximately equal to evaporation, in contrast to the sites of similar studies in Nebraska, North Dakota, and Wisconsin.

The purpose of the geologic studies of the Shingobee River Headwaters Area is to determine the three-dimensional geologic framework that affects the surface-water and ground-water processes within the study area.

Six test holes ranging in depth from 116 to 148 m were drilled by the mud-rotary method. Borehole geophysical logs were made by the Minnesota Geological Survey in four of the holes and by the U.S. Geological Survey in two of the holes. The types of logs included electric (spontaneous potential and resistivity), gamma, and density (gamma gamma). Several dozen shallower test holes, as deep as 45 m, were drilled by power auger; most were drilled in the southern part of the study area near Williams Lake. Soils were mapped by hand augering to depths of a meter, or less, throughout the Williams Lake watershed.

PHYSIOGRAPHIC FEATURES

Physiographic features of the Shingobee River Headwaters Area are the result of glacial deposition (Mooers and Norton, this volume). This type of terrain is characterized by hummocky topography, where the drainage network has not yet fully developed, and numerous isolated lakes and wetlands are common. The general study area is located on a small topographic ridge that extends south from the much larger east-west-trending Itasca moraine (see fig. 1). The ridge forms a surface-water divide between the Mississippi River and Crow Wing River drainage basins. Crystal Lake, which has no surface-water inlets or outlets, lies near the drainage divide. Seepage from Crystal Lake moves through the ground-water system to Williams Lake. The Shingobee River begins as a seep in a wetland southwest of Williams Lake. The river then connects a series of lakes (Doe, Mary, Island, and Steel, plate 1) before entering Shingobee Lake. Therefore, the Shingobee River Headwaters Area is defined by both surface-water and ground-water watersheds; it includes a number of contiguous watersheds that extend north from the Crystal Lake watershed to the outlet of Shingobee Lake (plate 1).

Physiographic characteristics of the Shingobee River Headwaters Area that are of direct interest to the present goals of the Interdisciplinary Research Initiative (Averett and Winter, this volume) are those landscapes contiguous to the principal surface-water bodies being studied: Williams Lake, Shingobee Lake, the Shingobee River, and the Little Shingobee Fen. For example, for purposes of watershed modeling and ground-water modeling, the overall relief and landslope are important physiographic characteristics. Topographic relief of the Shingobee River Headwaters Area is about 67 m. Although the highest point in the study area, 472 m (1,550 ft), is a hilltop east of Steel Lake (plate 1), the largest area of high ground, greater than 442 m (1,450 ft) in altitude, is in the southern part of the area near Crystal, Williams, and Doe Lakes. The lowest altitude in the area is the surface of Shingobee Lake, at an altitude of about 405 m (1,329 ft).

The terrane around Williams Lake is relatively gently sloping. The steepest slopes near the lake are on the east and southeast sides of the lake. On the east side of the lake, the land slopes at 0.051 m/m from a hilltop altitude of 463 m (1,520 ft) to Williams Lake. The steepest slope directly adjacent to Williams Lake is a small area on the southeast side, where the slope is 0.097 m/m. In contrast, slopes in the northern and topographically lower part of the study area adjacent to Shingobee Lake are much steeper.

Around the periphery of Shingobee Lake, the land generally slopes about 0.228 m/m from hilltops to the lake. Similarly, the Shingobee River is incised more than 21 m into a steep-walled valley that has side slopes of about 0.15 m/m. Little Shingobee Fen also is bounded by similarly steep slopes on its east, south, and part of its west sides, but it is bounded by a gentle slope to the northwest, where it is contiguous to Little Shingobee Lake. Little Shingobee Fen is only about 305 m west of the Shingobee River, but it is separated from the river by a ridge that is 15 to 18 m high.

The contrast in physiography between the higher (southern) and lower (northern) parts of the Shingobee River Headwaters Area is explained by differences in the geologic substrate across the area. The northern part of the area is underlain by silt and very fine sand, which maintains steeper land slopes than does the sand and gravel substrate in the southern part.

GEOLOGIC DEPOSITS

The glacial stratigraphy in the southern part of the study area in the vicinity of Williams Lake (fig. 4; see plate 1 for location of lines of section) consists of alternating units of sand and gravel and till. The surficial unit is primarily sand, but it also contains beds and lenses of materials ranging in size from silt to gravel. In addition, Seaberg (1985) mapped marl deposits 6 m above the present level of Williams Lake on the east and southwest sides. In general, particle sizes in the surficial sand unit become coarser with depth. This is evident west of Williams Lake, where silt and silty fine sand are common near land surface, but the deposit is largely coarse sand and gravel near its base (fig. 4). The basal coarse sand and gravel is continuous from beneath the northern part of Williams Lake to Mary Lake.

The till underlying the surficial sand is poorly sorted, the matrix consists of silt and clay, and it contains sand-size through boulder-size rocks. The till varies in color from gray to red-brown. Sand units are common within the till, but it is difficult to determine if they are areally extensive. It is conceivable that a relatively thick sand unit in the two deepest test holes at Williams Lake may be continuous. A thick unit of till underlies this sand unit. Bedrock was not penetrated in any of the test holes drilled in the Williams Lake area.

Glacial deposits in the northern part of the study area are not as easy to correlate as they are in the southern part near Williams Lake. Each of the three deep test holes drilled near Shingobee Lake had considerably different stratigraphy. At the deepest test hole drilled, Florence, which was drilled to a depth of 178 m, a thin surficial sand overlies a thin till unit (figs. 5 and 6). These two shallow units are underlain by a thick sand unit, which in turn is underlain by a thick till unit that extends to bedrock. At test hole Mabel (fig. 6), till was penetrated only after drilling through 98 m of sand. At test hole Stenberg (fig. 5), a thin deposit of sand is underlain by 61 m of silt and clay, which is underlain by a thick sand unit. Till was not penetrated in this test hole. Although the identification of bedrock is less certain in test hole Stenberg than it is in test hole Florence, this test hole probably also reached Cretaceous bedrock, but at a shallower depth than at test hole Florence.

The fluvial deposits in the northern part of the study area are considerably more fine grained than the fluvial deposits in the southern part. The steep land slopes in the northern part of the study area are typical of landscapes developed on fine-grained deposits. These deposits become coarser with depth, as they do in the southern part of the study area. However, only in the lowest 46 m of sand at test hole Mabel does the sand become as coarse and gravelly as is found at depth in the southern part. Particle sizes also differ areally in the northern part of the study area. East of Shingobee Lake and Shingobee River, at test holes Florence, T4, and Mabel, (figs. 5 and 6) the uppermost fluvial deposits are primarily fine sand, but they include some silt units. West of Shingobee River and west and north of Shingobee Lake, at test holes Stenberg and T-1, the uppermost fluvial deposits are primarily silt. The finest grained deposit in the entire Shingobee River Headwaters Area is the thick clay and silt deposit at test hole Stenberg. This deposit covers an area of at least several hectares, as indicated by the presence of a clay pit several hundred meters from the test hole on the Stenberg property that was once mined for making bricks.

Glacial deposits in the Steel Lake area and along the Shingobee River between Steel Lake and Shingobee Lake also are fine grained, as indicated by the steep banks along the Shingobee River valley. A geologic feature unique to the Shingobee River valley along the reach from Steel Lake to Shingobee Lake is the presence of a thin till unit at a shallow depth. The upper surface of this till is at a higher altitude than the Shingobee River (fig. 5, test hole Florence). The till underlies a thin surficial sand unit, resulting in many seeps along this reach of the Shingobee River valley at the sand-till contact . (See Triska and others, this volume, and Triska and Duff, this volume, for discussion of physical and chemical processes associated with one of the seeps.)

Until this investigation, characteristics of bedrock were unknown in the Shingobee River Headwaters Area because no wells had penetrated to bedrock in this part of Minnesota. The great thickness of permeable glacial drift has made it unnecessary to drill as deep as bedrock for water supplies. Bedrock at the Florence test site is white kaolin clay (fig. 5), probably a weathering product of Precambrian granite. About 11 m of unconsolidated deposits that are probably Cretaceous in age overlies the white clay. The rock presumed to be bedrock at the Stenberg test site is a blue-green clay containing white and black rock fragments (fig. 5).

PHYSICAL CHARACTERISTICS OF SELECTED SURFACE-WATER BODIES

The surface-water features currently of interest to study participants are Williams Lake, Shingobee Lake, the Shingobee River between Steel Lake and Shingobee Lake, and the Little Shingobee Fen. Comparison of physical characteristics of Williams Lake and Shingobee Lake (table 1 and fig. 7) is of particular interest because of the underlying premise of the IRI effort (see Averett and Winter, this volume). Both lakes have about the same maximum depth, but the area of Shingobee Lake is about 67 percent greater, and the volume of Shingobee Lake is nearly 100 percent greater than Williams Lake. The drainage basin of Shingobee Lake, excluding the lake and the drainage basin area of the inflowing Shingobee River, is only 25 percent larger than the drainage basin of Williams Lake.

The geology of the littoral zones of the two lakes also differs. The littoral zone of Williams Lake generally is sandy, but thin zones of silt and clay are present in limited areas of the east, north, and west sides. A clay zone along the northwest shore of the lake is partly responsible for a steep water-table gradient away from the lake in this vicinity. The littoral zone of Shingobee Lake reflects the fine-grained character of the adjacent hillsides. This type of geologic setting commonly results in discrete zones of spring inflow, as is the case along the periphery of Shingobee Lake. Characteristics of the lacustrine sediments in the two lakes are discussed by Locke and Schwalb (this volume).

The Shingobee River is a relatively small stream, generally only a meter or two wide and less than half a meter deep. The streambed is sand in most places. Discharge of the river is relatively uniform, usually ranging between 0.1 and 0.2 m^3/sec. The river has a number of beaver dams both upstream and downstream from Shingobee Lake. Between the source of the Shingobee River near Williams Lake and Steel Lake, the river consists of short segments that connect a series of lakes and wetlands. Physiography of the drainage basins constituting this upper segment of the river (upstream from Steel Lake) is considerably different than the deep valley in which the river flows between Steel Lake and Shingobee Lake. In this upper segment, the topography has gentler slopes, similar to that described for the Williams Lake area.

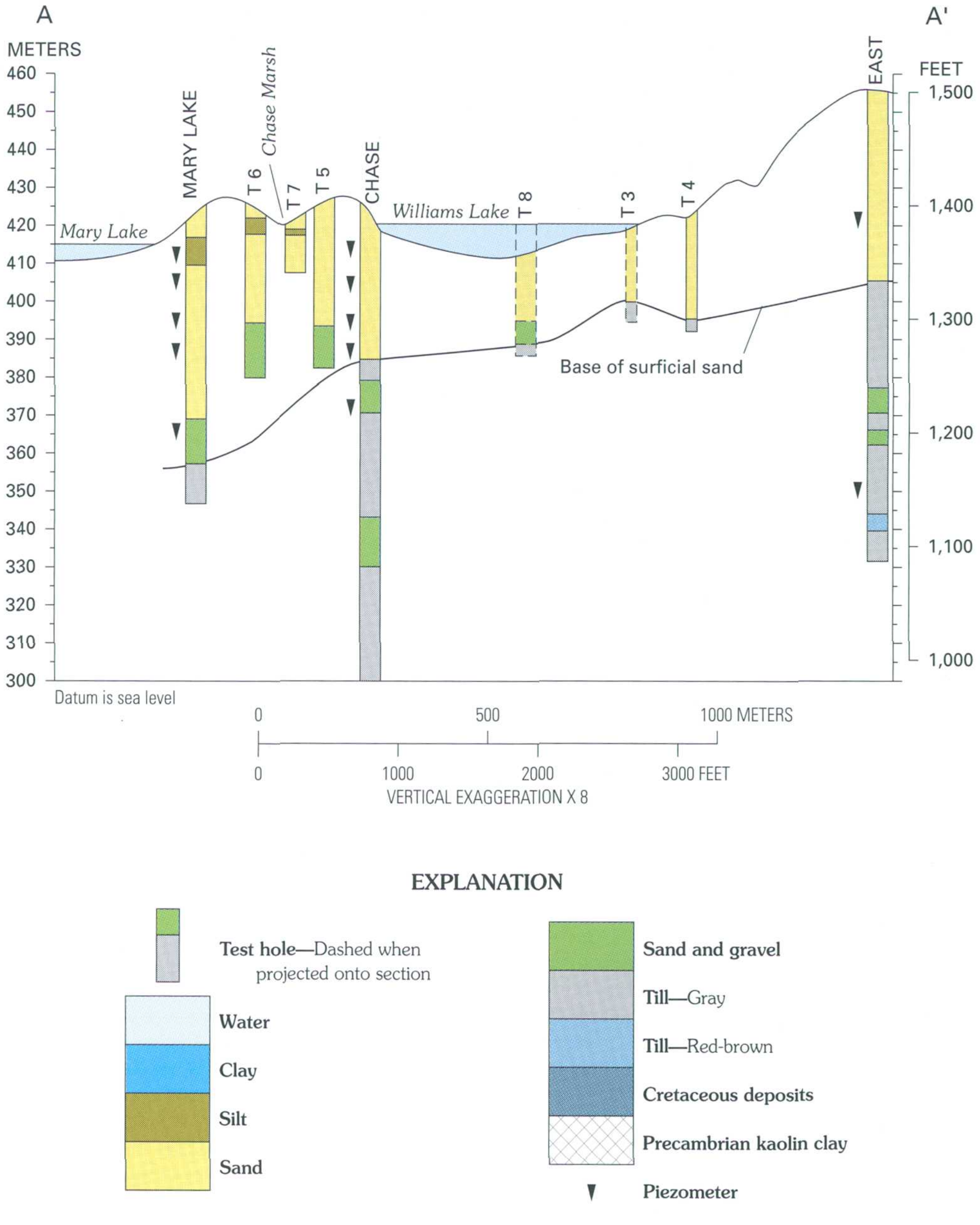

Figure 4. Geologic section A-A' in the Williams Lake area.

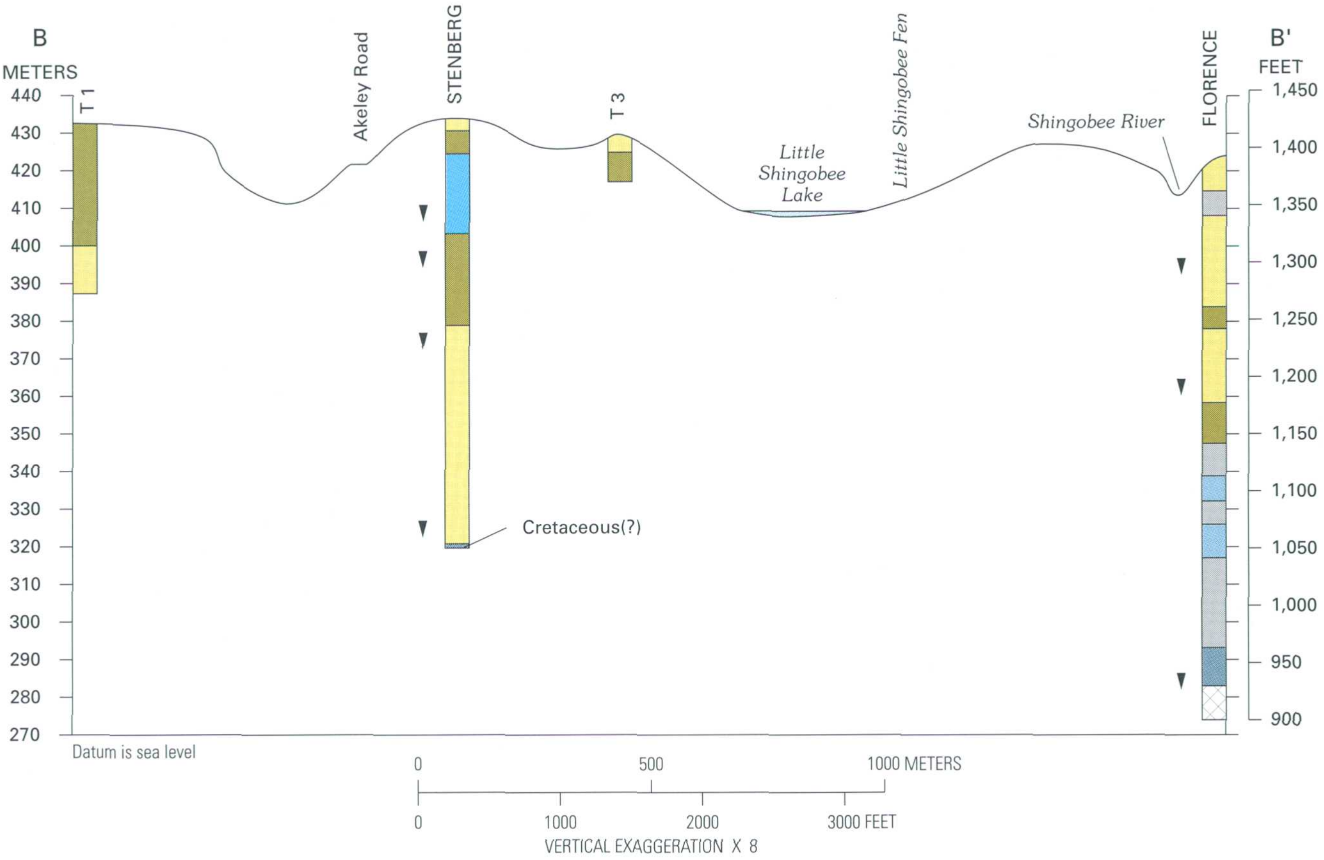

Figure 5. Geologic section B-B' in the Shingobee Lake area

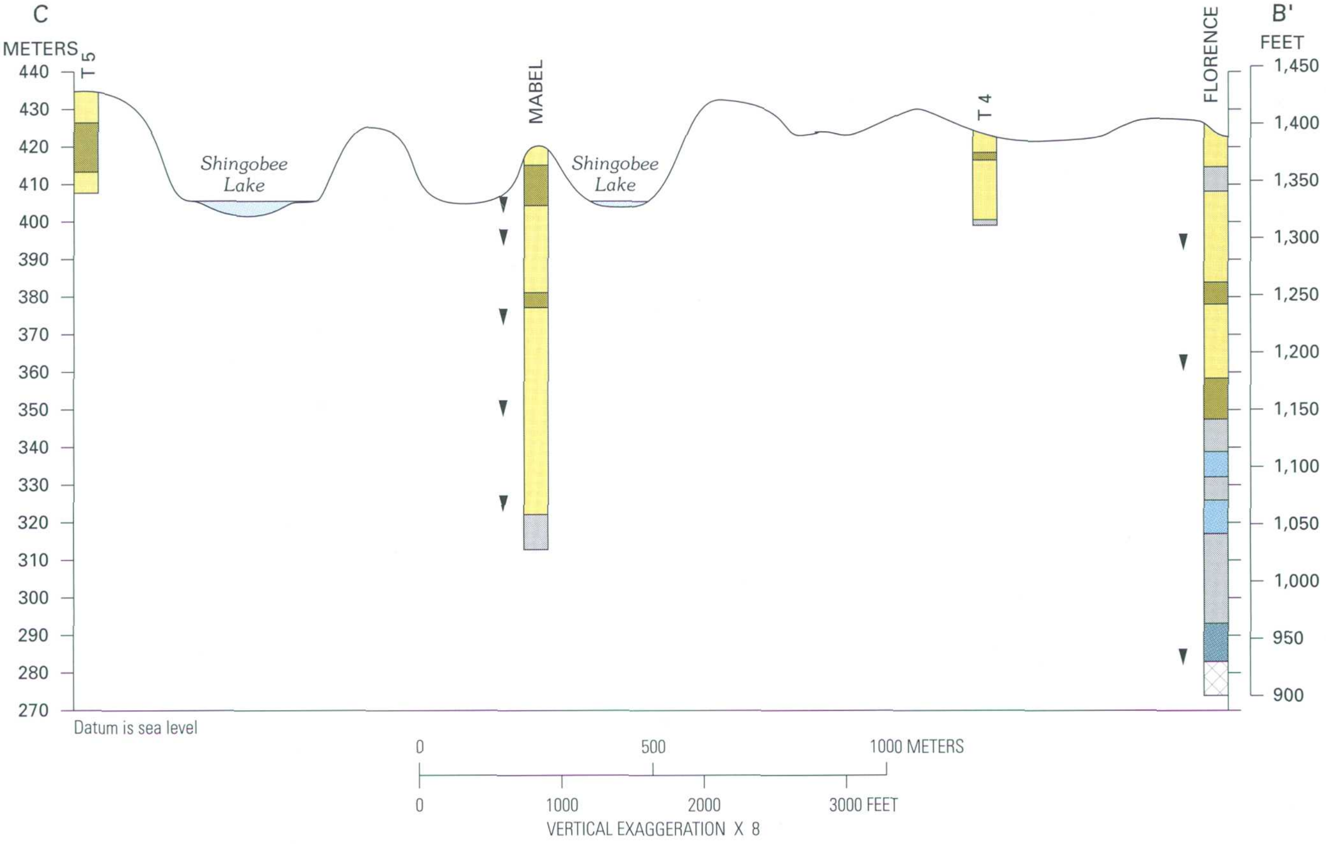

Figure 6. Geologic section *C* -B' *in the Shingobee Lake area.*

Figure 7. Bathymetric map of (A) Williams Lake and (B) Shingobee Lake.

Table 1. Comparison of selected physical and hydrologic characteristics of Shingobee and Williams Lakes

Characteristic	Shingobee Lake	Williams Lake
Lake surface area	65.6 hectares	39.3 hectares
Lake maximum depth	11 meters	10 meters
Lake volume	40.2 x 10^5 cubic meters	20.4 x 10^5 cubic meters
Lake residence time	about 6 months	about 3 years
Watershed area contiguous to lake (excluding lake)	222 hectares	177 hectares
Surface-water inlet	Shingobee River	none
Typical discharge at inlet	0.1 to 0.2 cubic meter per second	none
Surface-water outlet	Shingobee River	none
Typical discharge at outlet	0.11 cubic meter per second	none

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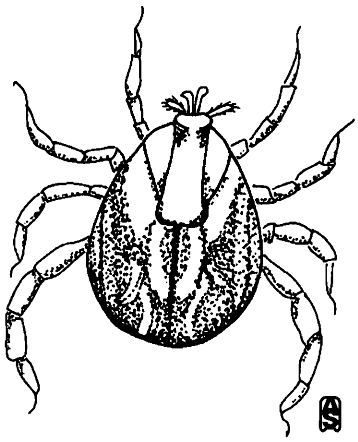

Hydrology of the Shingobee River Headwaters Area

By Donald O. Rosenberry, Thomas C. Winter, Dennis A. Merk, George H. Leavesley and Larry D. Beaver

INTRODUCTION

The determination of accurate water budgets is an important component of understanding lake hydrology. Selection of two lakes in the same watershed, one interacting with a stream and the other not, provides an ideal situation for testing some of the concepts of how lakes are affected by hydrologic fluxes. While all of the hydrologic fluxes that interact with a lake are of interest to the Interdisciplinary Research Initiative (IRI) effort, ground-water fluxes are the most difficult to quantify and are of special interest. Numerical models of ground-water flow near surface water indicate the interaction of ground water and surface water is spatially complex and temporally dynamic. Hydrogeologic setting affects the spatial complexity and temporal dynamics, and the latter also are affected by climate. The Shingobee River Headwaters Area provides a suitable field site to test (1) the effect of a highly variable hydrogeologic setting on the distribution of hydraulic head near lakes, (2) the importance of ground-water fluxes to a lake budget relative to other hydrologic fluxes, and (3) the importance of the geologic framework to design and calibration of ground-water flow models.

Several hypotheses regarding hydrologic fluxes can be tested at the paired-lake study site:

1. A closed lake is more likely to be affected by its interaction with ground water than an open lake.
2. Evaporation is a more significant component of a lake water budget for a closed lake than an open lake.
3. Small-scale reversals of flow at the edges of lakes are caused by focused recharge and evapotranspiration from ground water.

To test these and other hypotheses, it was necessary to obtain accurate water budgets for Williams and Shingobee Lakes. Components of these budgets are compared with respect to the relative significance of each component to the lake water budget. Processes related to ground-water flux to and from lakes as well as ground-water recharge to lake basins also are investigated. The longer term data set for Williams Lake is used to indicate response of lake stage and hydrologic fluxes to variable climatic conditions.

METHODS

Precipitation at Williams Lake has been monitored with a tipping-bucket rain gage during open-water seasons since 1979. Weighted-average values from three nearby National Weather Service stations were used to provide data during periods of missing site data and during winter months. Since 1989, precipitation has been monitored year round with weighing-bucket gages at Williams and Shingobee Lakes (plate 1). When winter data were suspect because of problems associated with accurate measurement of snowfall, National Weather Service data were used.

Evaporation using the energy-budget method was determined at Williams Lake for 1982-86 and at Shingobee Lake from 1989 to the present (1996). A mass-transfer coefficient was determined using energy-budget evaporation data, and evaporation rates using the mass-transfer method were determined for 1980-91 for Williams Lake.

Discharge of the Shingobee River into Shingobee Lake is monitored with a Parshall flume installed 500 m upstream from the lake. Another Parshall flume provides discharge data from a tributary stream that flows from Little Shingobee Lake into the Shingobee River downstream from the main-stem flume but upstream from Shingobee Lake (plate 1).

Because of inaccessibility to the outlet of Shingobee Lake, the stream cannot be gaged until it crosses a county road 2.4 km downstream from the lake. At that location, stream stage is monitored and a stage-discharge relationship is used to provide discharge data.

In the Williams Lake watershed, 80 wells provide data on water-table fluctuations. Piezometer nests provide information on the vertical distribution of hydraulic head at three locations within the Williams Lake watershed. In the vicinity of Shingobee Lake, three piezometer nests and seven water-table wells provide information on hydraulic heads.

Ground-water fluxes to and from Williams and Shingobee Lakes were determined using the Darcy equation: $Q = KIA$, where Q = ground-water flux (L^3/T), K = hydraulic conductivity (L/T), I = hydraulic gradient (dimensionless), and A = cross-sectional area of flux (L^2). Hydraulic gradients were determined by the difference between heads at individual wells and lake stage divided by the distance between the well and the lake. Three wells (locations shown on plate 1) were used to provide data for estimates of ground-water flux to Shingobee Lake. All wells indicate a flux of ground water into Shingobee Lake. Eight wells (locations shown on plate 1) were used to provide data for estimates of ground-water flux to and from Williams Lake; five indicate seepage to Williams Lake, and three indicate seepage from the lake. The cross-sectional area through which ground water flows to or from the lakes was determined based on information from well logs, test holes, and near-shore sediment cores. Hydraulic conductivity of the glacial drift was determined from slug tests of wells. Hydraulic-head data at the wells were collected weekly during spring, biweekly during summer and fall, and monthly during winter.

HIGHLIGHTS OF RESULTS

Hydrologic data presented for Williams Lake are the average values for 1980-91. Errors are assumed to be 5 percent for precipitation and change in storage, 10 percent for streamflow, 15 percent for evaporation, and 50 percent for ground-water fluxes. Values for Shingobee Lake are estimated using data collected in 1990 and 1991. Changes in storage for Shingobee Lake are assumed to be zero to ignore the changes in stage caused by a beaver dam at the outlet. Errors for Shingobee Lake fluxes are likely much larger than for Williams Lake due to fewer data-collection sites and a shorter data-collection period. Ground-water-flux values for both lakes are subject to the greatest errors because of the variability in geologic properties in the lake watersheds, and in the case of Shingobee Lake, because of few wells to characterize both geologic variability and ground-water gradients near the lake.

Hydrologic fluxes to and from Williams and Shingobee Lakes are shown in figure 8. Pie sizes in figure 8 indicate the relative magnitude of the average annual hydrologic fluxes to and from each lake; total fluxes to and from Shingobee Lake were about 8 times the total fluxes to and from Williams Lake. Streamflow dominated the hydrologic fluxes to and from Shingobee Lake, whereas ground water dominated the hydrologic fluxes to and from Williams Lake, supporting hypothesis 1. At Shingobee Lake, streamflow contributed about 72 percent of all fluxes to the lake and about 96 percent of all fluxes from the lake. Ground water seeped into the lake on all sides and is estimated to contribute about 24 percent of fluxes to the lake. Seepage to Shingobee Lake is concentrated in numerous springs along much of the shoreline. Evaporation caused 22 percent of fluxes from Williams Lake and 4 percent of fluxes from Shingobee Lake, supporting hypothesis 2.

Ground water flowed into Williams Lake along the south and east shorelines of the lake, and lake water flowed into the ground-water system along the north and west shorelines. Flow into the lake occurred along 45 percent of the lakeshore, and flow out of the lake occurred along 55 percent of the lakeshore. Ground-water fluxes at Williams Lake are estimated to be 65 percent of all fluxes to the lake, and 75 percent of all fluxes from the lake (fig. 8). Lithologic logs from numerous wells and test holes indicate considerable variability in the type and distribution of geologic materials in the Williams Lake watershed (Winter and Rosenberry, this volume), which makes it difficult to accurately measure ground-water fluxes. In general, the glacial drift near Williams Lake is relatively more conductive than the drift near Shingobee Lake, and this higher hydraulic conductivity allows diffuse fluxes

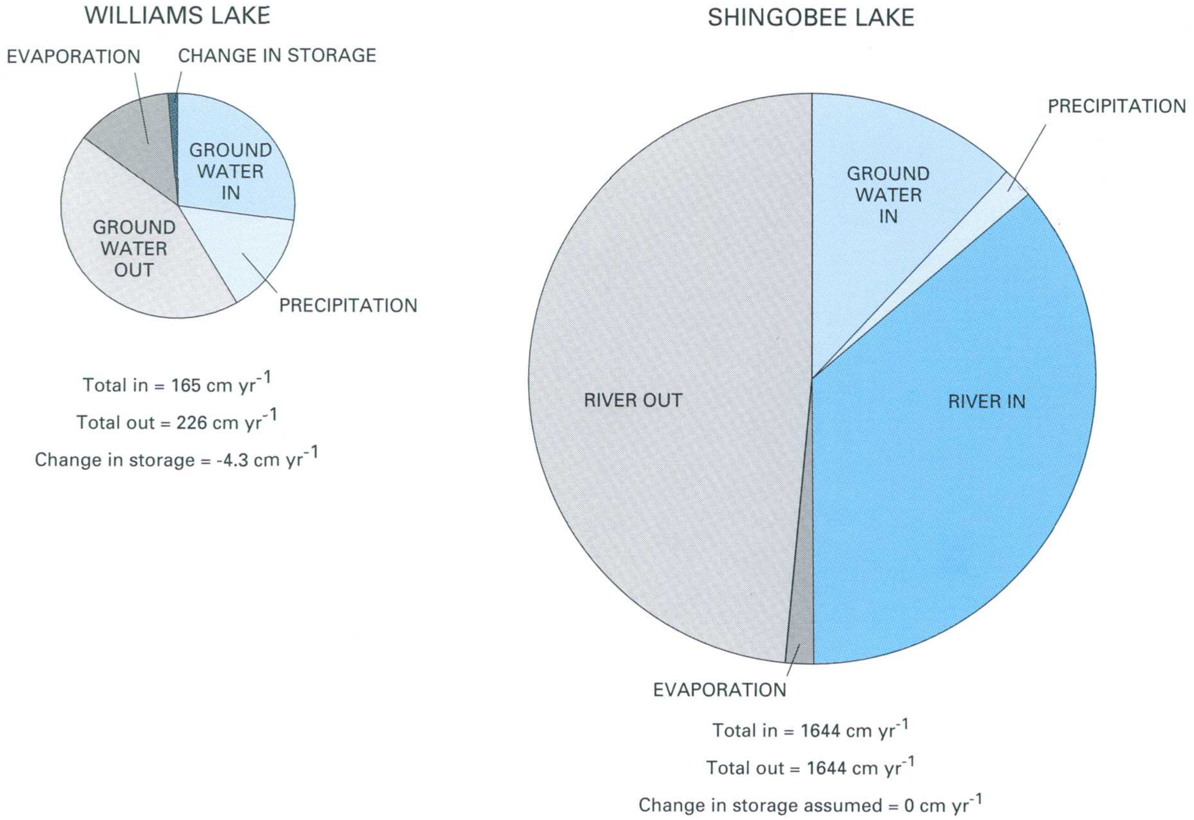

Figure 8. Average annual fluxes for Williams Lake and Shingobee Lake water budgets. [Pie size is proportional to total fluxes to and from each lake. All fluxes are in centimeters of water added to or removed from lake surface.]

of ground water, there are no visible springs in or near Williams Lake.

Occasionally the gradient between the lake and ground water reverses along shoreline reaches of Williams Lake where water normally seeps out of the lake; a temporary ground-water recharge mound directly downgradient from the lake drives the reversals in direction of flow. This water-table configuration has been observed at two sites along the western and northwestern shorelines of the lake (Rosenberry, 1985). This configuration generally is short-lived, lasting from one to two weeks, and is caused by above normal precipitation after a prolonged wet period. These temporary flow reversals likely have little effect on the lake water budget.

If all data were error free, fluxes into a lake would equal fluxes out of a lake plus or minus change in storage. This clearly is not the case for Williams Lake (fig. 8). Fluxes into Williams Lake were only 41 percent of the water budget, and since change in storage was 1 percent, the excess of fluxes out of the lake was 8 percent. Assuming all of the error is in the ground-water estimates, the ground-water-in term would have to be increased by 53 percent, or the ground-water-out term decreased by 33 percent, to balance the total fluxes. Fluxes into Shingobee Lake totaled 50 percent of all fluxes to and from the lake, indicating that the initial estimates are close to balancing. However, this probably is fortuitous considering the uncertainties in the stream discharge measurements associated with the outlet gage. The gage is located 2.4 km downstream from the lake and is affected by hydrologic fluxes along the stream reach between the lake and the gage.

Shingobee Lake stage is controlled by a large beaver dam located downstream from the lake. Therefore, variability of lake stage is not necessarily an indicator of recent hydrologic events. In contrast, Williams Lake stage is relatively sensitive to variability in hydrologic fluxes to and from the lake. Williams Lake stage typically goes through a cycle of rising stage during spring and early summer, falling stage during mid-summer to early fall, rising stage during mid- to late fall, and a stable to slightly rising stage during winter (fig. 9). This annual stage fluctuation pattern is common in closed-basin Minnesota lakes and has been shown to be typical of soil-moisture conditions in Minnesota as well (Baker and others, 1979). During a normal year, lake stage fluctuates from a low of 421.47 m in September to a high of 421.57 m in June, a range of 0.10 m.

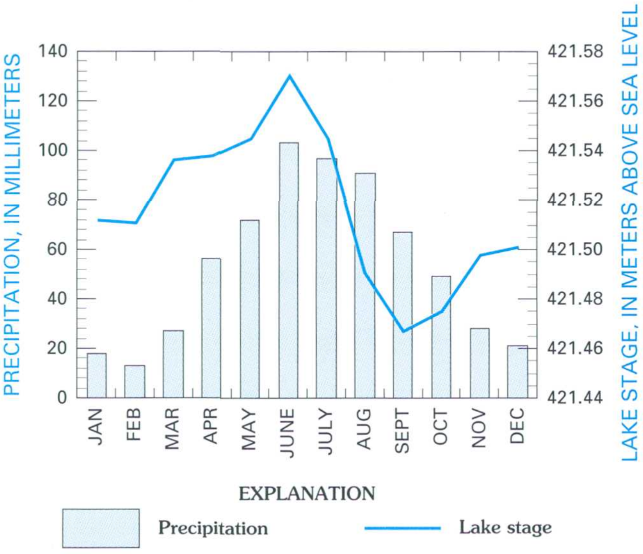

Figure 9. Normal monthly precipitation (1961-90 averages) and normal monthly Williams Lake stage (1980-87 averages).

Average precipitation, as determined using 1961-90 weighted-average data from the three nearby National Weather Service sites, also is shown in figure 9. During an average year Williams Lake receives about 64 cm of precipitation. Nearly one-half of the annual precipitation occurs during June through August, and two-thirds of annual precipitation occurs from May through September. During 1980-87, monthly evaporation commonly exceeded precipitation in July, August, and September, which caused the lake-stage decline during that portion of most years.

The stage of Williams Lake has fluctuated over a range of about 1 m during the 14-year period from 1978 to 1991 (fig. 10). Monthly cumulative departure from normal precipitation (CDNP) during this time also is shown in figure 10. If precipitation were normal for each month of a year, the CDNP curve would plot as a straight horizontal line during that year. The slope of the CDNP line is positive during wetter-than-normal periods and negative during drier-than-normal periods.

Both lake stage and CDNP indicate a dry period during 1979-80, a wet period during 1981-82, a wet period during 1985-86 and a much drier than normal period during 1987-91 (fig. 10). The wet period during 1985-86 caused lake stage to increase by more than 0.4 m, and the dry period from 1987 through 1991 caused lake stage to decline by more than 0.8 m. From 1978 through 1986, the precipitation total was normal, but from 1987 through 1991 the IRI site received 64 fewer centimeters of precipitation than normal. The greatest deficit occurred during 1990, when the site received less than one-half its normal precipitation. From 1986 through 1991, the volume of water in Williams Lake declined by almost 14 percent (LaBaugh and others, 1995) in response to the drier than normal climatic conditions.

The magnitude of climatic and hydrologic variability observed during this study can be compared to a longer period of record using a climate indicator such as the Palmer Drought Severity Index (PDSI). The PDSI is a measure of the severity of agricultural drought (Palmer, 1965), but it also is used to identify and quantify periods of wetness. Monthly PDSI data were available from 1898 to the present. PDSI data for northern and especially northwestern Minnesota indicate that the drought during the late 1980's and early 1990's was the most severe drought since the 1930's. The PDSI data also show a wet period from 1982 to 1986 that was the longest nearly uninterrupted wet period since the mid-1960's. Based on comparison with the nearly 100-year PDSI record, it appears that climatic fluctuations during the period of investigation have been relatively large. Williams Lake stage may have risen to stages higher than observed during the study period, but it is likely that, during the past 100 years, Williams Lake stage has not dropped much lower than that observed during the period of investigation.

An additional process that may act to enhance lake-stage drop during mid- to late summer is near-shore transpiration. During the daytime hours from about June through September, in-lake seepage-measurement transects commonly indicated a gradient from the lake along sections of shoreline where ground-water-level data from wells away from the lake indicated a gradient to the lake. It is likely that these gradients out of the lake represent small, local ground-water flow paths created by transpiration from trees and shrubs growing near the lake. This process has not yet been quantified and is not accounted for in the water budgets for either of the study lakes, but it could amplify lake-stage declines during summer months.

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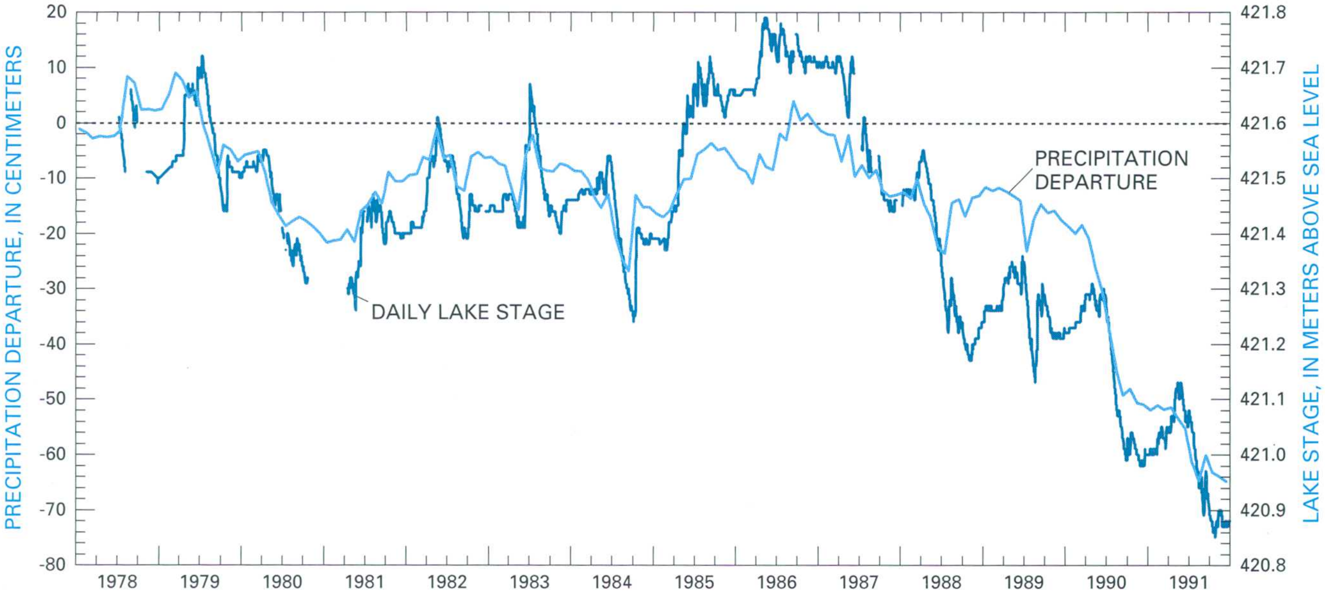

Figure 10. Williams Lake daily stage and monthly cumulative departure from normal precipitation, 1978-91.

Modeling Temperature and Evaporation of Williams Lake

By Steven W. Hostetler

INTRODUCTION

Thermal properties, associated energy and hydrologic balances, and the surface meteorology of Williams Lake have been measured since 1979. From these measurements, accurate estimates of the surface energy budget (SEB, comprising solar radiation, atmospheric radiation, back radiation from the lake surface, the flux of sensible heat, and the flux of latent heat), the storage of heat in the lake sediments, and storage of heat within the lake have been made. Accurate estimates of lake evaporation have also been computed from the energy-budget study (Sturrock and others, 1992). The interannual variation of the measured and computed components of the energy budget and estimates of evaporation provide excellent data with which to describe the thermal characteristics and evaporation of the lake over a number of years. Moreover, they also can be used to investigate temperature-dependent limnologic processes that have been measured within the lake over the same time period.

The measured meteorological and lake temperature data sets also can provide boundary conditions and input data with which to model the thermal response and evaporation of the lake. By applying a model, the possible effects of climate change and the sensitivity of the lake to various climate perturbations can be quantified, thereby allowing the thermal characteristics to be simulated outside the period of measurement. Before such tests can be conducted, the performance of the model needs to be evaluated by comparing simulated values with measured data. Such a model validation is the focus of this chapter. Following discussion of the validation, results from a simple, prescribed climate perturbation are presented.

METHODS

A one-dimensional, eddy-diffusion model is applied to simulate temperature and evaporation of Williams Lake over the ice-free periods of 1983-86 to demonstrate application of a lake model. Although not done here, the model also can simulate lake ice and the behavior of stable isotopes ($\delta^{18}O$ and ^{2}H). Details of the model are presented by Hostetler and Bartlein (1990), Hostetler and Benson (1990; 1994), Hostetler (1991), Hostetler and Giorgi (1993; 1995), Hostetler and others (1993).

The lake model was run on a daily time step using as input the daily average values of solar radiation, long-wave radiation, air temperature, relative humidity, and wind speed that were measured at the lake. Initial conditions for the model were obtained from records of measured depth-temperature profiles.

RESULTS

The time series of measured surface temperature for 1983-86 (fig. 11) display considerable intra- and interannual variability. This variability is also evident in the simulated time series of surface temperature (fig. 11), which compare very well with the measured data. Both time series reflect the interannual variability in the SEB of the lake. Scatter plots of simulated and measured temperatures for each year, together with linear regression of the two data sets indicate a high correlation between the simulated and measured values (fig. 12). In addition to the interannual variability of surface temperature, the model results also indicate substantial interannual variability of internal temperature, the depth and duration of thermal stratification, and the time and temperature of the lake at fall turnover (fig. 13).

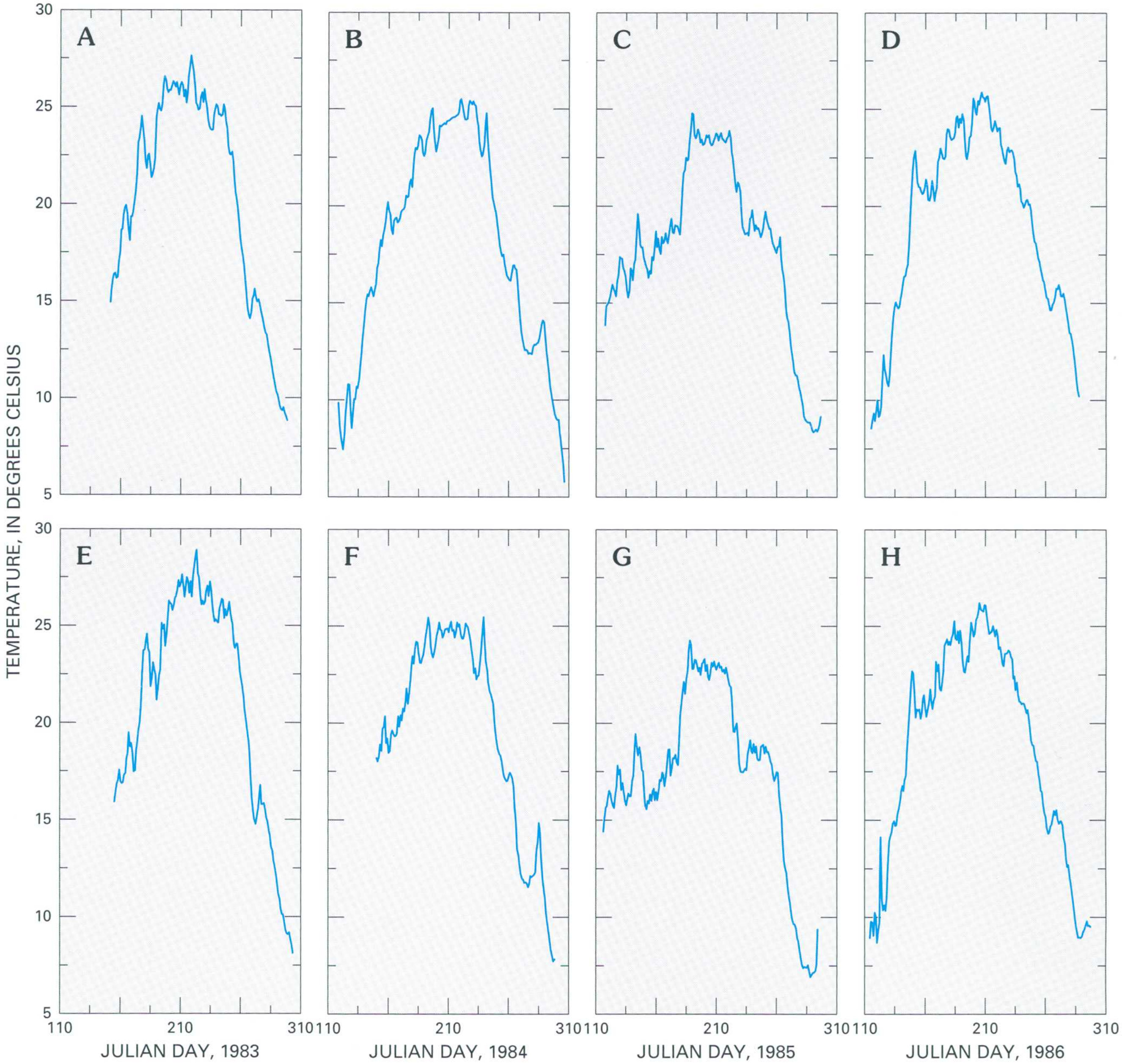

Figure 11. Surface temperature of Williams Lake for the open-water periods of 1983-86. Measured data panels A through D, simulated panels E through H. Measured data modified from Sturrock and others (1992).

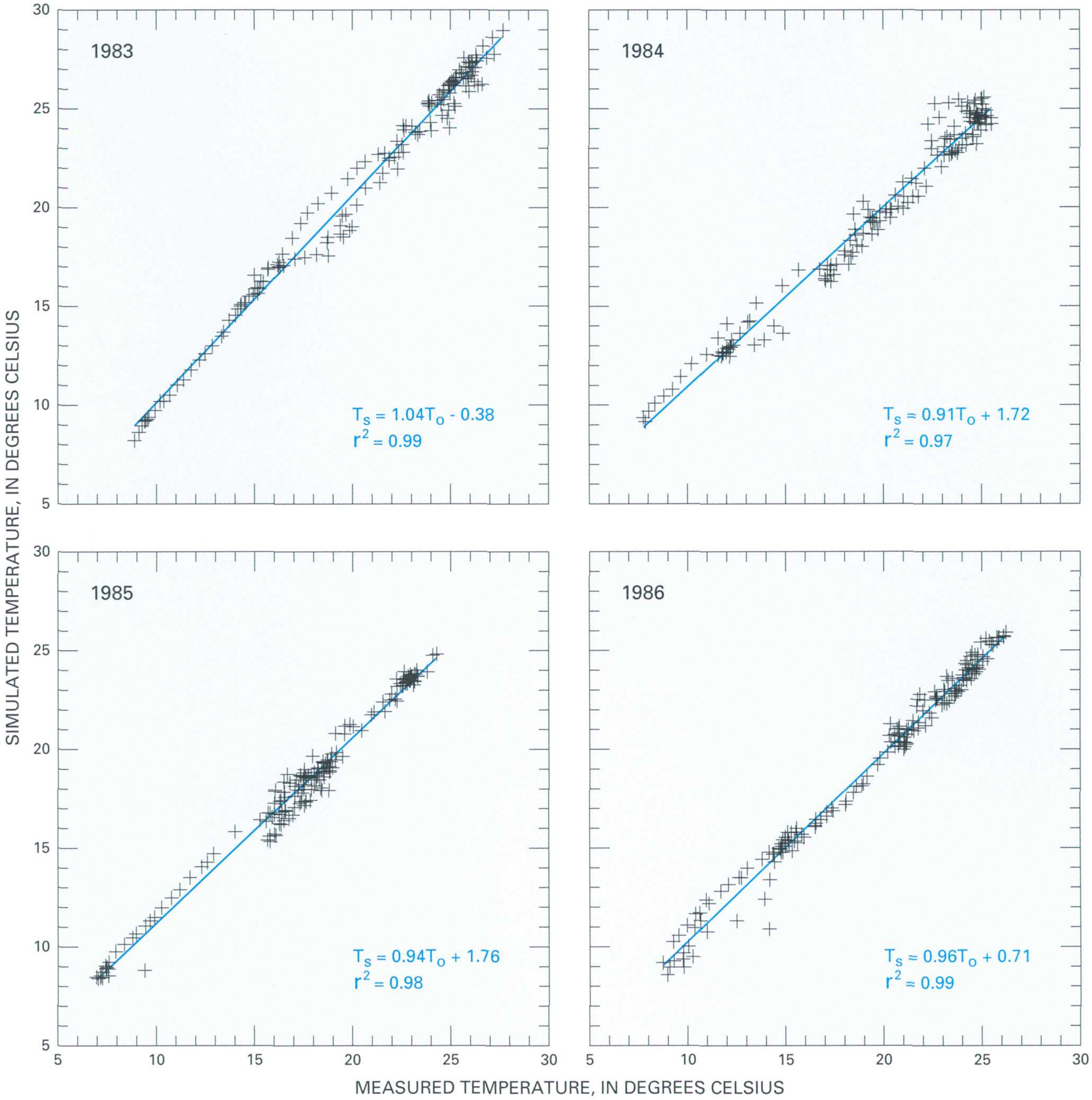

Figure 12. Relation of simulated and measured surface temperatures of Williams Lake for the open-water periods of 1983-86. Lines are linear regression lines fit to data points. T_S is simulated temperature, T_O is measured temperature.

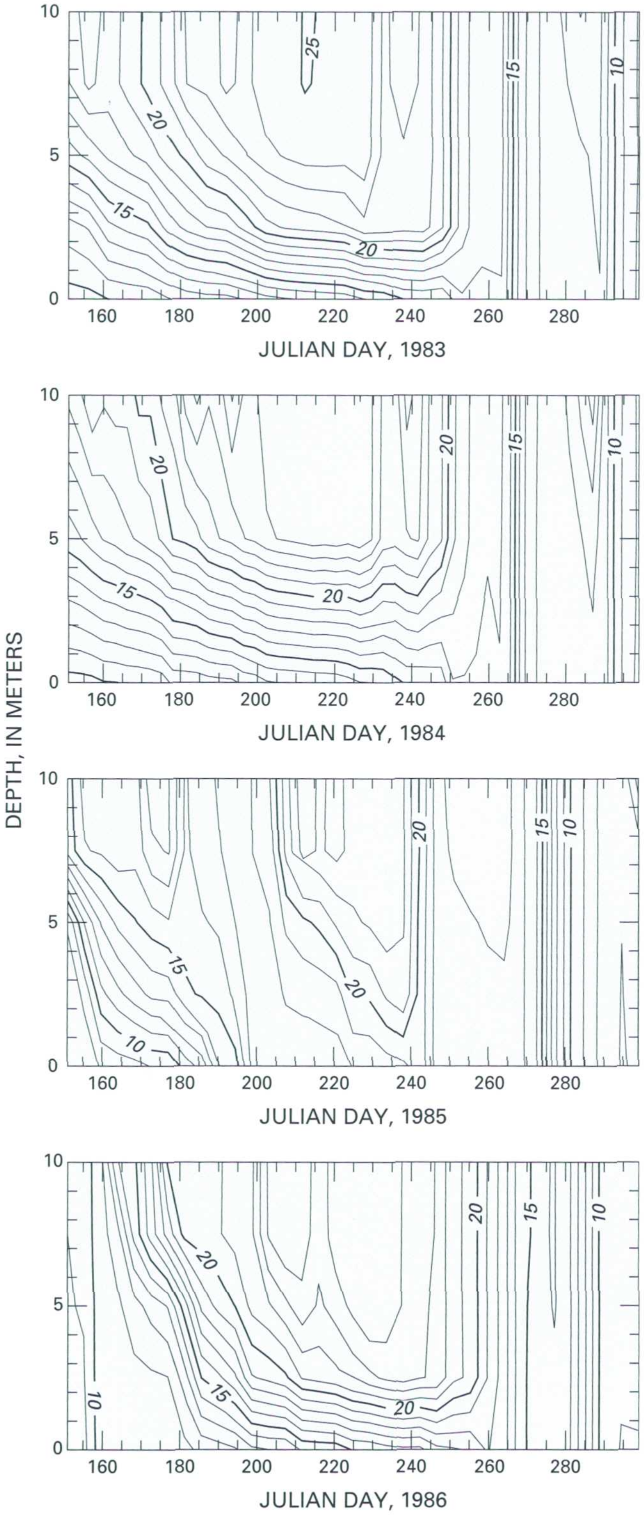

Figure 13. Simulated isotherms of Williams Lake for the open-water periods of 1983-86. (Countour interval is 1°C.)

For each of the ice-free periods, the simulated evaporation totals are in good agreement with the total evaporation estimated in the energy-budget study of the lake (table 2). The results of this model application indicate that the model is capable of simulating well the temperature and evaporation of the lake for the four years considered. The quality of the simulations is a direct reflection of the quality of the measured data that were used as input to the model.

A simple sensitivity experiment demonstrates use of the model and measured data to assess the possible effects of climate change on the temperature and evaporation of the lake. The experiment was conducted using the data set for 1985 which was modified to reflect warmer climatic conditions by increasing average air temperature 3°C and applying an associated 15 percent increase to atmospheric (longwave) radiation. All other variables were held at the measured 1985 values. The surface temperature of the lake increased an average of 3°C over the 1985 average for the open-water period (fig. 14) in response to increased air temperature and atmospheric radiation. Evaporation for the warmer conditions increased 170 mm, or 40 percent, over the 1985 total. Relative to 1985 (fig. 13), simulated thermal stratification (fig. 15A) commences earlier and persists longer, and warmer temperatures penetrate to greater depths. Water temperatures increased by more than 2°C throughout the spring, summer, and fall, with a maximum increase of more than 2.8°C in late August (fig. 15B). Relative to 1985, fall overturn in the warm case is delayed about ten days.

The lake model discussed here allows investigations into the interactions of climate with physical, chemical, and biological processes in lakes. This type of model can be used to gain an understanding of basic processes that are observed at Williams Lake today, to reconstruct the response of the lake to past climates, or to assess what the response of the lake to future climates might be.

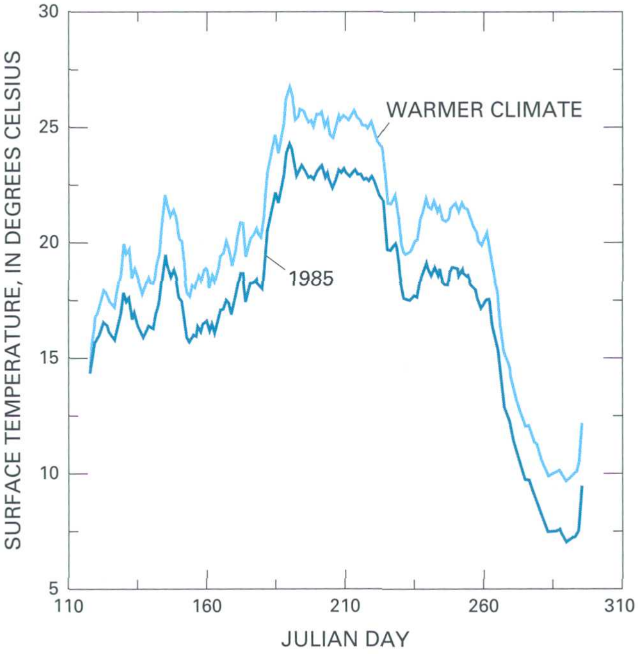

Figure 14. Simulated surface temperatures of Williams Lake for 1985 and for the simple case of a climate that is 3°C warmer and has 15 percent higher atmospheric radiation compared to 1985.

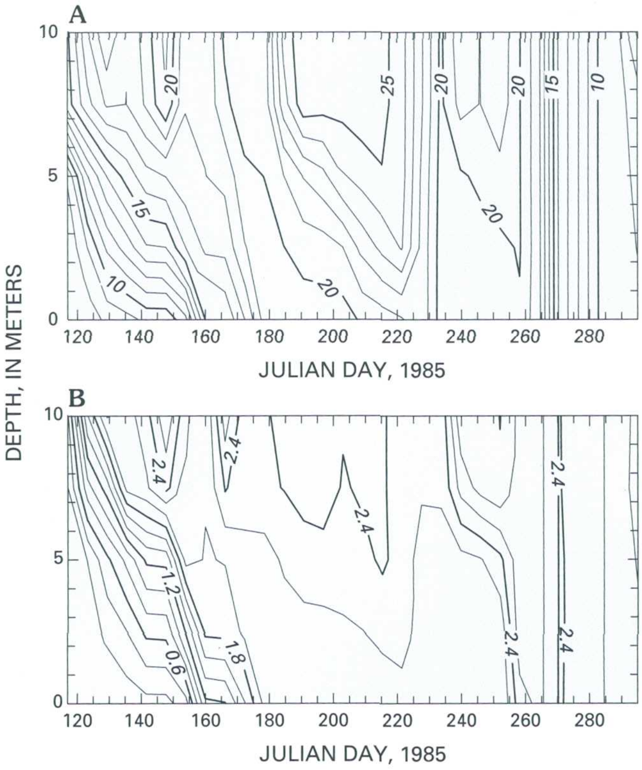

Figure 15. (A) Simulated isotherms of Williams Lake for the warmer climate explained in the text. (Contour interval is 1°C.) (B) Difference (warmer climate case minus 1985) isotherms for the simulated temperature of Williams Lake. (Contour interval is 0.2°C.)

Table 2. Simulated and estimated evaporation totals for the open water periods of 1983-86. Estimated values are from Sturrock and others (1992)

[mm, millimeter]

Year	Simulated evaporation (mm)	Estimated evaporation (mm)	Difference (percent)
1983	390	390	0
1984	370	370	0
1985	420	390	+8
1986	440	450	-2

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Exchanges of Carbon Dioxide and Water Vapor between Williams Lake and the Atmosphere

By Dean E. Anderson, David I. Stannard, Robert G. Striegl, and Ted A. McConnaughey

INTRODUCTION

A thorough study of the exchanges (fluxes) of carbon dioxide (CO_2) and water between a lake and the overlying atmosphere can provide valuable information for use by scientists, engineers, and planners in ecology, hydrology, climatology, and watershed management. Accurate estimation of lake evaporation, using direct methods, is needed to improve the accuracy of commonly used indirect methods that are based on routinely measured parameters. Understanding the mechanisms governing CO_2 flux over water (particularly oceans) can help close the global carbon budget, a problem still unresolved after more than two decades of intense research (Houghton 1993). With regard to lakes, recent work by Cole and others (1994) has attempted to make global estimates of lake fluxes of CO_2 to the atmosphere. Their work was dependent upon indirect methods as no direct measurements were made, further emphasizing the need for the measurements discussed in this chapter.

The theory of how gases like CO_2 and water vapor are transported between lakes (or water bodies in general) and the atmosphere has evolved considerably in recent years. Gases are transported between the atmosphere and a water body through a series of layers. Much of the air above the water is turbulent and gases are quickly transported by eddies to and from the vicinity of the water's surface. Between the turbulent layer and the water's surface exists a thin (a few millimeters thick) non-turbulent, laminar layer in which gases are slowly transported by molecular diffusion. Early theory suggested that immediately below the water-air interface, gases pass through a stagnant film layer of water on the order of micrometers thick (Whitman, 1923). The film's thickness was assumed to remain constant, regardless of wind induced stress. In recent years, the stagnant film model has been shown to be unrealistic and has been replaced by the boundary layer model (Hasse, 1990). This model considers the top most layer of water to be a reactive layer of varying thickness within which molecular diffusion controls transport (Deacon, 1977). An augmentation to the boundary layer theory is surface renewal theory (Hasse, 1990) which allows that established boundary layers may randomly break down as fluid parcels are randomly replaced by parcels from the bulk water column below. This theory embraces a much more efficient process resulting in larger modelled fluxes than those found with either the boundary layer or stagnant film models. Below this topmost layer of water is the bulk water column, dominated by turbulent motions, which quickly transport diffused gases. Recently, substantial improvements to flux estimates have been made on both boundary layer and surface renewal models by Phillips (1994) who defined transport across the interface in terms of more easily measured parameters. Properly understanding the chemical and physical properties of a water body's topmost layer is key to successfully modelling gas transport between the atmosphere and a water body. By far, the greatest resistance to gaseous transport occurs in this layer.

Direct measurements of the fluxes of CO_2 and water vapor between the atmosphere and Williams Lake were made using the eddy correlation method, a micrometeorological technique described by Stannard and others (this volume). Model results estimating water and CO_2 fluxes using routinely made measurements such as wind speed, air and water temperatures, CO_2 concentration, and relative humidity are compared to the measured fluxes. Comparisons are also made with changes in CO_2 concentration of the entire lake volume, which is an indirect method of calculating a flux.The complete data set may also help identify the reasons for the significant differences between fluxes measured by eddy correlation and those estimated from tracer techniques (O'Brian, 1986).

METHODS

A comprehensive discussion of the eddy correlation technique may be found in Baldocchi and others (1988). The following is a very brief introduction to it.

Atmospheric turbulence transports heat and various gases (among other things) between the Earth's surface and the overlying atmosphere through vertical motions. To measure a CO_2 flux, for example, measurements of vertical wind speed and CO_2 concentration are made simultaneously, 10 times per second, over one-half hour. Following this protocol for sampling ensures that most governing atmospheric motions occurring near the Earth's surface have been sufficiently sampled. CO_2 flux is calculated from the covariance of the vertical wind speed and CO_2 concentration measurements. This flux number is then corrected for instrument response and deployment (as outlined in Baldocchi and others, 1988). The corrected number represents the net exchange occurring between the atmosphere and an area of the Earth's surface located at an upwind distance from the sensors over the one-half hour period. The upwind surface area has an elliptical shape. Its position and size vary with wind direction, wind speed, atmospheric thermal stability (farther upwind if the air is stable, such as with warm air over cold water), and the height of the instruments (a larger area is "seen" with higher instruments). Typical upwind distances are about 100 m in this study.

For reliable eddy correlation measurements, vertical wind speed (updrafts and downdrafts) must be accurately measured from a stable platform. In this study, a communications antenna tower was pressed into lake sediments approximately at the center of the lake. Lake depth was approximately 10 m at the site; about 2.5 m of tower extended above the water surface. Below water, guy wires were run from the tower to screw anchors in the lakebed to stabilize the tower. Eddy correlation instruments were affixed to the tower about 1 m above the water and were oriented into the mean wind direction. A low-profile inflatable raft was tethered to the tower and allowed to swing into the downwind direction. The raft held data loggers, amplifiers, power supplies, and related equipment needed to run the eddy correlation array (fig. 16).

Vertical wind speed was measured with a single axis sonic anemometer (Campbell Scientific) except in April 1992 when a tri-axial (three wind components, including vertical) sonic anemometer (model V, Applied Technologies) was used. Temperature was measured with a fine-wire thermocouple on the single axis anemometer and was calculated through speed of sound principles using the tri-axis anemometer. A krypton hygrometer (Campbell Scientific) sensed absolute humidity (water vapor density). Additionally in April 1992 and 1993, open-cell, fast-response infrared gas analyzers (IRGA's) were used to sense absolute humidity. Open-cell configuration IRGA's used in April 1992 (Advanced Systems) and April 1993 (Auble and Meyers, 1992) also sense CO_2 density and have a sensing path or volume, typically the distance between an infrared light source and detector, that is open to the ambient air. In all other measurement periods a closed-cell IRGA (LiCOR model 6252) was used to sense CO_2 density (this instrument does not sense humidity). Air is drawn through an intake tube and through the closed-cell IRGA by a vacuum pump at a flow rate of about 6 L/m. Comparison between the two instrument types (open versus closed-cell) indicate an agreement within a few percent for CO_2 fluxes (Suyker and Verma, 1993). With care, a CO_2 flux of 0.1 to 0.2 $\mu mol/m^2/s$ can be measured.

Mean air temperature, relative humidity, wind speed and direction, net radiation, shortwave radiation (solar), and photosynthetic photon flux density (PPFD) were measured at about 2 m above the water on the tower. Water temperature was measured at several depths at the tower with a thermocouple string (Stannard and others, this volume).

All dissolved gas measurements (gas in water) were made using the following methodology. A 25 ml sample of water was obtained in a 60 ml syringe. To this, 25 ml of air with a known CO_2 concentration was added. The mixture was shaken vigorously for 3 minutes to achieve air-water equilibration of pCO_2 (partial pressure CO_2). The equilibrated air was extracted into a dry syringe and injected into a gas chromatograph (GC) fitted with a methanizer and flame ionization detector. In1994, CO_2 concentration was determined by injecting the equilibrated air of the syringes into a small cell-volume IRGA, instead of a GC.

Figure 16. Meteorology tower, eddy correlation instrumentation, and raft containing support equipment on Williams Lake.

RESULTS

Measured Fluxes and General Conditions

Williams Lake became clear of ice on April 6, 1992. Eddy correlation measurements began in the evening of April 11. Mean conditions are summarized in table 3. As shown in figure 17A, lake-water CO_2 concentrations (CO_{2aq}) decreased rapidly, and air and water temperatures (fig. 17B) rose over the period of measurement. Winds speed (fig. 17C) remained of moderate intensity throughout most of the 7-day measurement period. Fluxes of sensible heat (heat exchanged due to an air-water temperature difference) and latent heat (energy derived from evaporation or condensation) from the lake were quite large on April 11 (fig.18), due to the advection of very cold, dry air. Fluxes in the upward direction are of positive sign. On April 11 the magnitude of CO_2 flux exceeded 2 $\mu mol/m^2/s$, about double of that measured any other day (fig. 18B).

Thaw of lake ice in April 1993 was followed by a very warm, windy period within which measurements began (6 days after thaw). Mean values of sensible and latent heat fluxes (table 3) were comparable to those measured in April 1992 but the range of values, reflected in the respective standard deviations, were smaller. Measured CO_2 flux ranged from -0.06 to +0.06 $\mu mol/m^2/s$ but averaged -0.11. Lake water CO_2 concentration was not measured concurrently with these measurements.

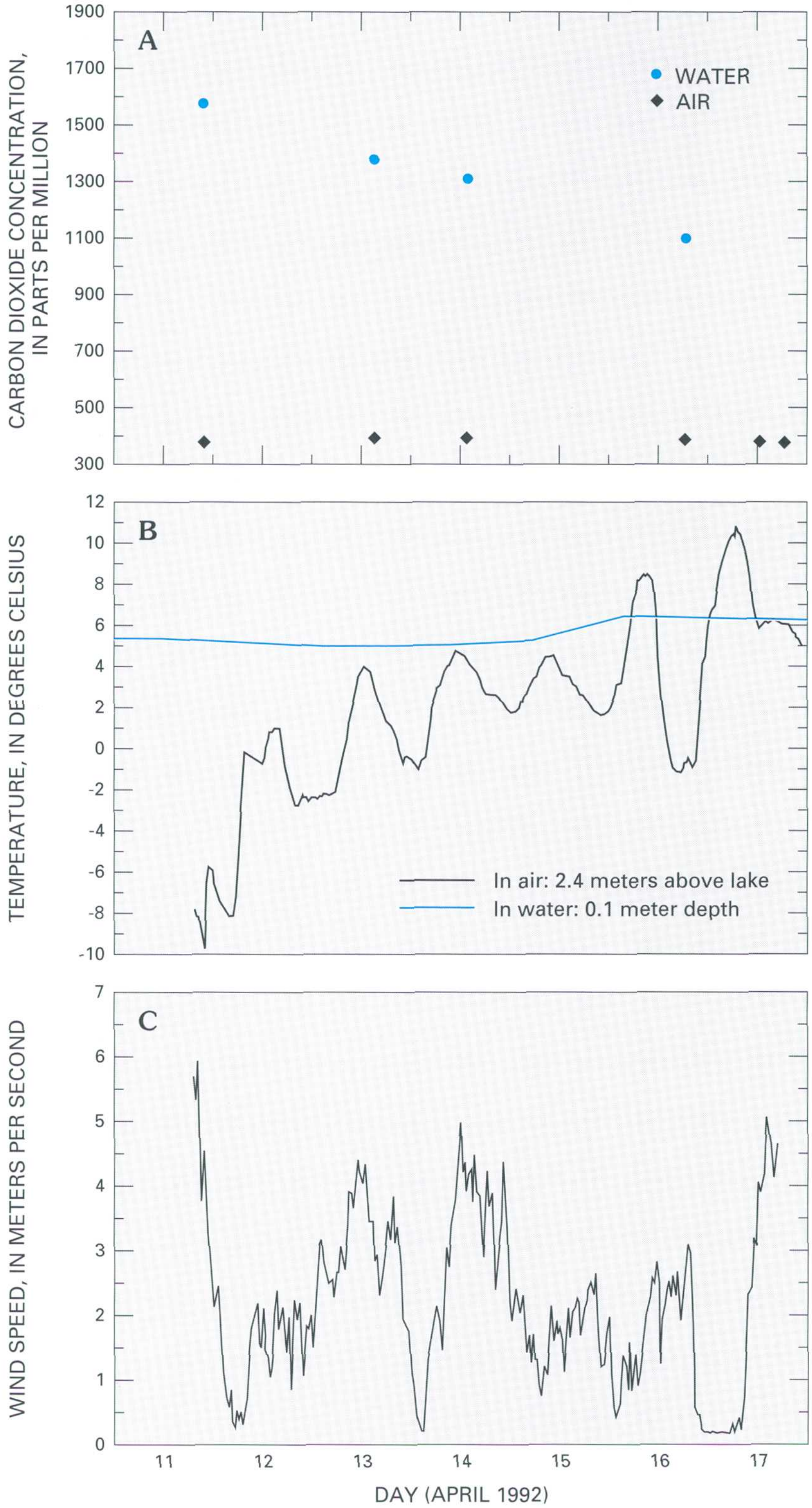

Figure 17. Mean environmental conditions at Williams Lake during April 11-17, 1992.

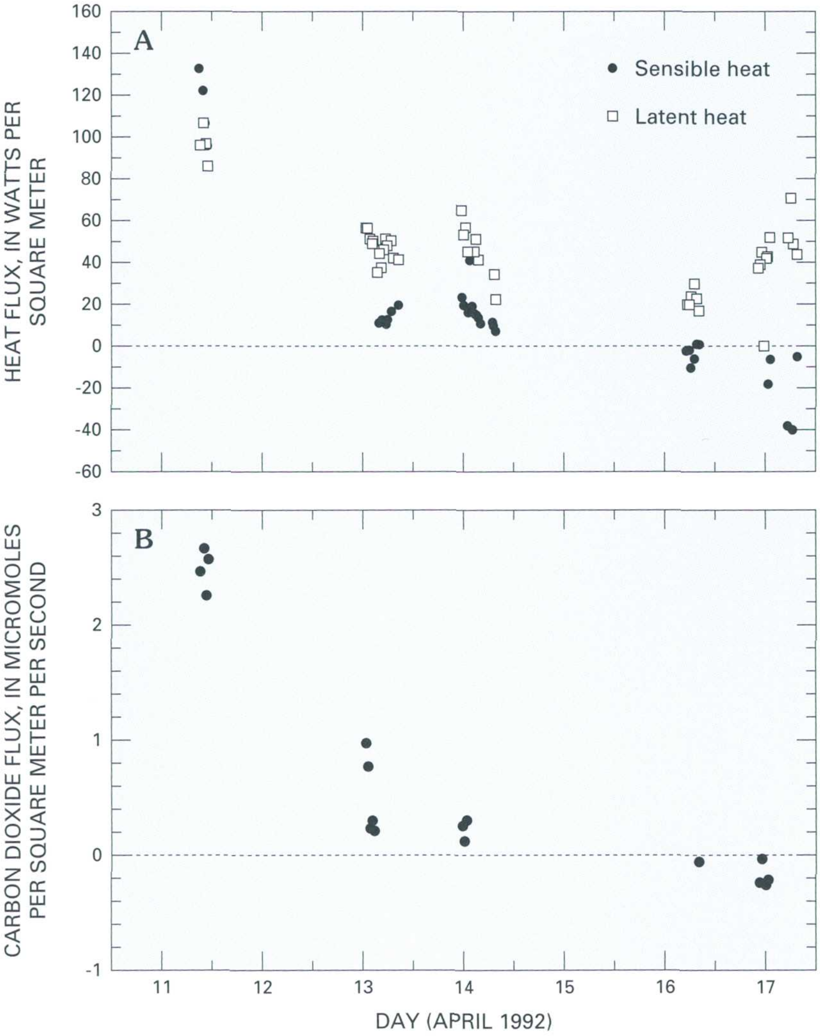

Figure 18. (A) Sensible and latent heat fluxes, and (B) CO_2 flux measured over Williams Lake, April 11-17, 1992.

Mid-summer and fall measurements (July and October 1993) are summarized in table 3. Latent heat fluxes were considerably higher in the summer than in the fall, owing to a greater mean vapor pressure gradient at the time of measurement. Averaged sensible heat fluxes were negligible for both periods. Measured net CO_2 exchange was near zero in summer and was a small efflux in fall. Hesslein and others (1990), working on a series of "productive" lakes in western Ontario, estimated very weak (undetectable) fluxes in summer and a small efflux in fall. A small diurnal trend in daily CO_2 flux was detected on July 29 at Williams Lake (fig. 19A). Flux to the lake may have been in response to presumed photosynthetic activity of the aquatic vegetation. Short-wave radiation usable for photosynthesis (photosynthetic photon flux density, PPFD), plotted in figure 19B, follows a near parabolic trend with time on this clear day. The skewness towards a late afternoon peak in CO_2 flux to the lake may be due to the use of stored lake-water CO_2 in the morning and secondary mechanisms of obtaining CO_2 by plants (McConnaughey and others, 1994).

Conditions following thaw of lake ice on April 16, 1994, were similar to those of 1993 (table 3). On April 23, 1994 (fig. 20A), a net flux of water vapor (latent heat) to the lake (negative values) was caused by a warm, moist air mass that came in contact with

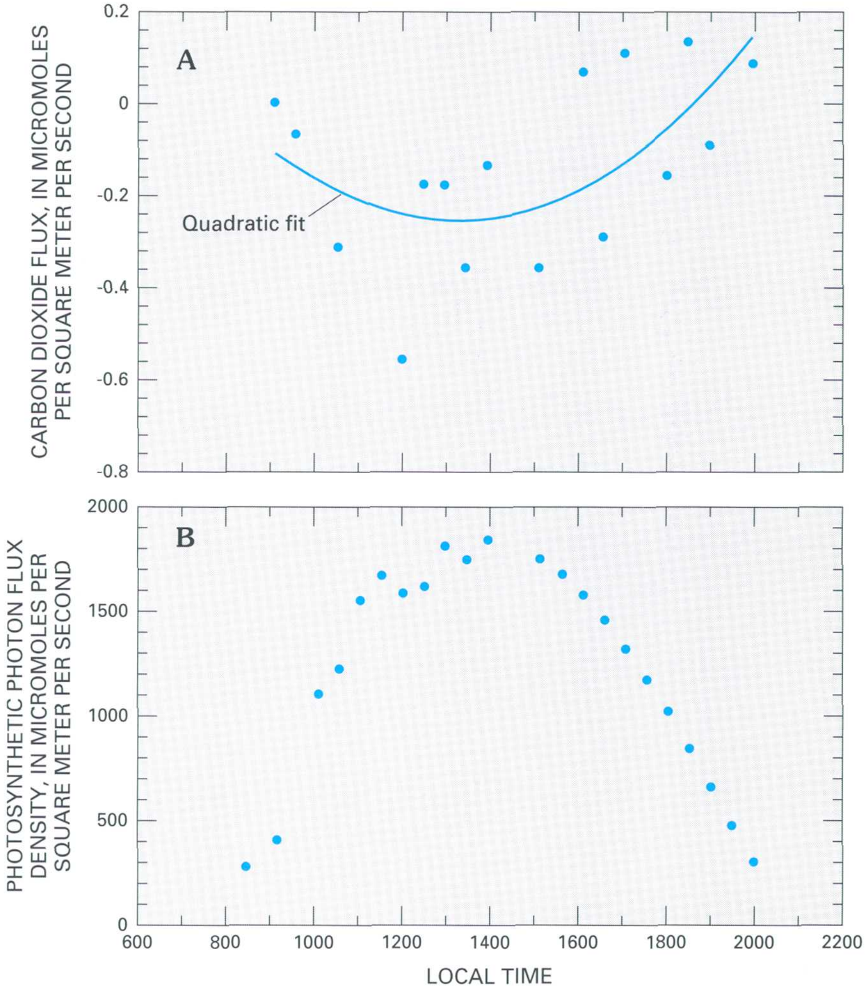

Figure 19. Diurnal trends of (A) CO_2 flux and (B) photosynthetic photon flux density (PPFD) over Williams Lake, July 29, 1993. Curve fitted to CO_2 data is a cubic polynomial used to identify diurnal trend.

the cold lake surface. Rates reached -32W/m^2 (a water depth equivalent of 1.1 mm/d) at around 1800 hours.

A comparison between eddy correlation measurements of CO_2 with those obtained from daily change in lake volume of CO_{2aq} is shown in figure 20B. The latter numbers were calculated from vertical profiles of CO_{2aq} measured once a day, approximately 150 m east of the tower. Most results compare well except those of April 19 which followed a 400 ppmv (parts per million by volume) decrease in CO_{2aq} from April 18 to 19. Beyond measurement errors (CO_2 sensor noise is about 0.1 to 0.2 µmol/m^2/s), one possible explanation for the discrepancy between the flux measurements is that the eddy correlation measurements span only a fraction of the time that lake fluxes occur. Most of the periods between measurements were during the night and early morning when low wind speeds and lower fluxes would be expected. Another, but less plausible, explanation is that horizontal lake CO_2 gradients existed. While CO_{2aq} measurements were consistently made east of the tower, upwind areas were north of the tower on April 18 and 19 but southeast to southwest thereafter.

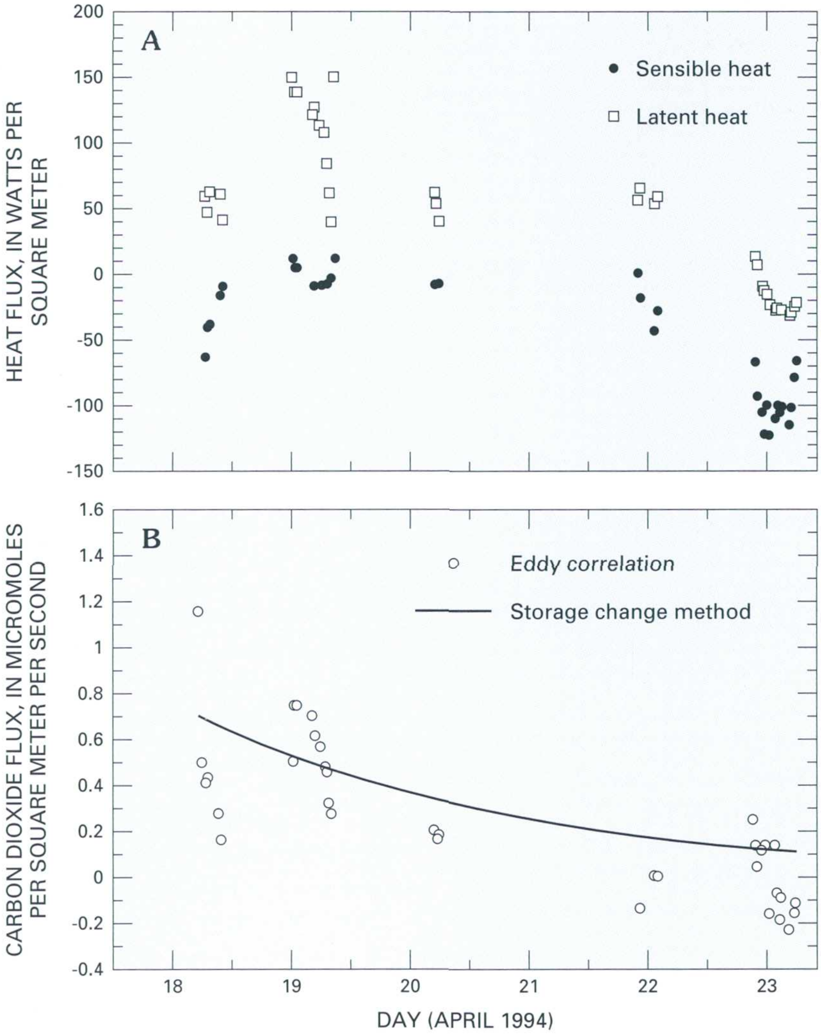

Figure 20. (A) Sensible and latent heat fluxes and (B) CO_2 fluxes measured by eddy correlation and calculated from storage changes in dissolved CO_2 in lake-water volume, Williams Lake, April 18-23, 1994.

Table 3. Summary of environmental conditions, sensible heat, water vapor and CO_2 fluxes at Williams Lake

[Values are mean ± standard deviation; °C, degrees Celsius; m/s, meters per second; W/m^2, Watts per square meter; $\mu mol/m^2/s$, micromoles per square meter per second]

Dates	Air temperature (°C)	Water temperature (°C)	Wind speed (m/s)	Friction velocity, u_* (m/s)	Sensible heat flux (W/m^2)	Latent heat flux (W/m^2)	CO_2 flux ($\mu mol/m^2/s$)
April 11-17, 1992	4.55 ± 5.79	5.48 ± 0.58	2.77 ± 0.58	0.14 ± 0.05	16 ± 37	62 ± 47	0.67 ± 1.05
April 25-26, 1993	8.60 ± 3.92	8.14 ± 0.52	4.07± 0.93	0.13 ± 0.03	7 ± 31	65 ± 20	-0.11 ± 0.61
July 27-29, 1993	19.66 ± 2.41	22.31 ± 0.34	2.71± 0.90	0.08 ± 0.03	7 ± 14	96 ± 29	0.05 ± 0.45
October 13-14, 1993	8.97 ± 1.98	9.35 ± 0.22	2.66 ± 1.01	0.08 ± 0.03	-6 ± 12	4 ± 32	0.18 ± 0.41
April 18-23, 1994	13.76 ± 5.44	7.13 ± 2.4	5.45 ± 1.75	0.28 ± 0.10	-66 ± 73	74 ± 79	0.52 ± 0.55

COMPARISONS WITH MODELS

A number of investigators working with wind tunnels and natural water bodies have attempted to predict gaseous transfer from routine measurements (Peng and Broecker, 1979; Deacon, 1977; Wesely and others, 1982; Wanninhoff and others, 1985; Liss and Merlivat, 1986; Denmead and Freney, 1992). These and other data have been used to test the boundary layer model. The basic model formulation for CO_2 flux (*Fc*) is

$$Fc = V\Delta pCO^2$$

where ΔpCO_2 is the CO_2 concentration gradient between the water's surface (in equilibrium with air at the interface) and that of the CO_{2aq} in the bulk water column. The actual flux is slightly reduced by aerial resistances to transport not included in this equation. The transfer velocity, V, is given by

$$V = 0.082\left(\frac{\rho_a}{\rho_w}\right)^{1/2} Sc^{-b} Sct^{-c} u_*$$

where ρ_a and ρ_w are air and water densities, Sc is the Schmidt number (the ratio of the viscosity of water to the molecular diffusivity of the gas in water), Sct is the turbulent Schmidt number (the ratio of the turbulent transfer coefficient for momentum in the boundary layer to that of the diffusing substance), and u_* is friction velocity (proportional to momentum transfer). Sct is typically set to 1, exponent c to 1/3 (Deacon, 1981), and exponent b, to 2/3. Deacon (1981) points out that setting b to 2/3 can lead to underestimates of V for smooth flow (low wind speeds).

Transfer velocities are modified by a number of phenomena including the presence of surface films (composed of a variety of materials), waves, microcirculations, bubbles (Frankignoulle,1988; Wallace and Wirick, 1992), and chemical enhancements (for CO_2, see Liss, 1983). As described earlier, the surface renewal model's (Liu and others, 1979) transfer velocity varies with random replacement of fluid (water) parcels at the air-water interface. Calculations using the above equations with exponent b set to 2/3 (representing the boundary-layer concept), b set to 1/2 (surface renewal concept), the Wesely and others, (1982) model, eddy correlation measurements and flux calculated from storage changes in lake volume CO_{2aq} are plotted in figure 21. Using storage change measurements as a reference, the surface renewal and Wesely and others models are in best agreement. There is good overall agreement between fluxes obtained by eddy correlation measurements, the storage change method, and the surface renewal model. For as yet undetermined reasons, eddy correlation measurements of flux on April 19 were larger than either of the other methods. The reader should note that there is some autocorrelation between fluxes calculated from lake storage and all model output. All these calculations utilize the same CO_{2aq} measurements.

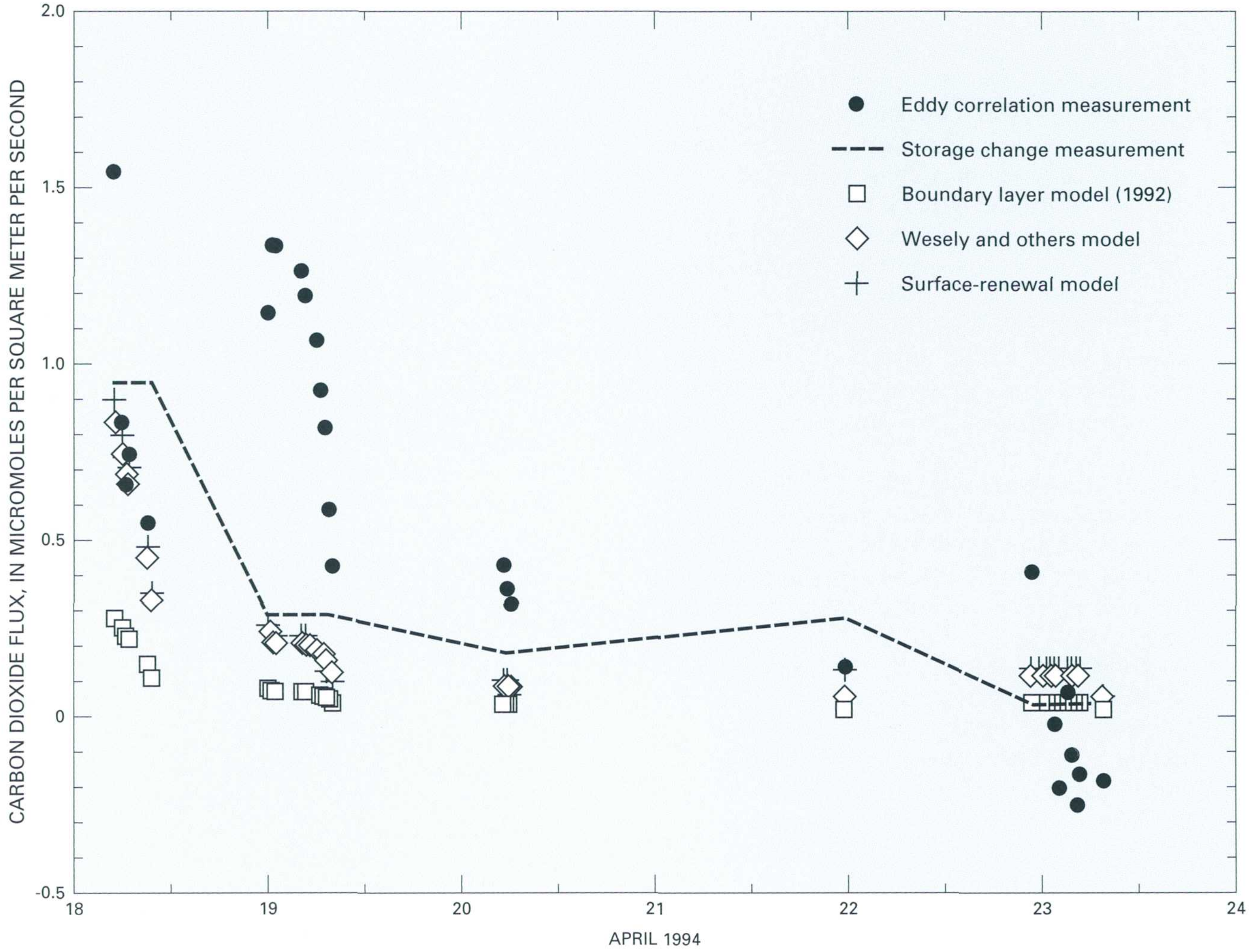

Figure 21. Carbon dioxide flux measurements and model predictions, Williams Lake, April 18-23, 1994.

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Problems in Comparing Short-term Eddy-Correlation and Energy-Budget Measurements of Evaporation from Williams Lake

By David I. Stannard, Dean E. Anderson and Donald O. Rosenberry

INTRODUCTION

Evaporation often is a major component of the hydrologic and energy budgets of a lake (Harbeck and Kennon, 1954; Rodgers and Anderson, 1961; Keijman and Koopmans, 1973; Winter and Woo, 1990; Mahrer and Assouline, 1993). If changes in lake storage are small, the relative importance of evaporation increases as the size of other (surface- and ground-water) outflows decreases. Surface-water outflow from Williams Lake is zero, and evaporation is estimated to be one-third to one-half of ground-water outflow. (Rosenberry and others, this volume). Therefore, evaporation has a significant effect on the thermal structure of the lake through evaporative cooling and on the water chemistry through evaporative concentration of chemical species. Evaporation has been measured routinely at the lake from 1982 to 1986 (Sturrock and others, 1992) using the energy-budget method (Bowen, 1926; Anderson, 1954). These data were used to compute a mass-transfer coefficient, and the mass-transfer method has been used to estimate evaporation at Williams Lake since 1986 (Rosenberry and others, this volume).

During a landmark study at Lake Hefner in Oklahoma (Anderson, 1954), the energy-budget method was first verified for lake use by comparing it to the water- budget method. This study indicated that the energy-budget method "must be used with caution" for a measurement period of less than a week, but that the difference between the energy- and water-budget estimates decreased to ± 5 percent as the measurement period increased beyond a week. The largest source of error for the shorter time periods was considered to be measurement of the change in energy stored in the lake.

Prior to 1980, very few tests of the energy-budget method as applied to lakes were made (Hughes, 1967; Keijman and Koopmans, 1973). However, the literature abounds with reports of studies where the energy-budget method was used as a standard against which other simpler methods were evaluated (for example, Harbeck, 1962; Keijman, 1974; Ficke and others, 1977; Spahr and Ruddy, 1983; Sturrock and others, 1992; Rosenberry and others, 1993). Additionally, the energy-budget method has been applied for periods shorter than a week (for example, Keijman, 1974; Stewart and Rouse, 1976; Strub and Powell, 1987; Lee and Swancar, 1997).

Eddy correlation is a direct method of measuring the turbulent flux of a quantity by rapidly measuring the fluctuations in the concentration of that quantity and in wind velocity components (Swinbank, 1951). Stannard and Rosenberry (1991) made short-term eddy-correlation and energy-budget measurements of evaporation at Island Lake, Nebraska, in an attempt to verify the energy-budget method. Calculated evaporation rates using the eddy-correlation method were 37 percent less than rates using the energy-budget method. It was concluded that errors in the measurement of net radiation may have affected the energy-budget measurements of evaporation, and that evaporative and sensible-heat exchanges between lake and air may have varied significantly across the lake surface.

The energy-budget and eddy-correlation methods were compared for longer time periods at Lake Kinneret, Israel (Assouline and Mahrer, 1993). Again, large differences in calculated evaporation rates were observed. Use of a mesoscale meteorological model suggested that evaporative and sensible-heat exchanges between lake and air varied considerably across the lake surface at this site also (Mahrer and Assouline, 1993).

The hypotheses that led to the present studies are as follows: (1) the energy-budget method is fundamentally sound for short time periods; (2) modern instrumentation can be used to improve the accuracy of the short-term storage measurement; and (3) the eddy-correlation method can be used to confirm energy-budget results, provided that differences in the performance of net radiation sensors can be resolved.

The purpose of these investigations is to make detailed short-term comparisons of eddy-correlation and energy-budget measurements at Williams Lake. Williams Lake is well suited for this comparison because the lake has no surface water interactions. The scope of the work to date has involved deploying sensors from a raft and from a tower located near midlake (plate 1), conducting thermal surveys, and analyzing nearby ground-water-level data. The midlake measurements were recorded every 30 minutes, and thermal surveys were conducted once or twice daily. These measurements were made during site visits, which lasted several days at a time. Site visits were made in April and August 1992, in April, July, and October 1993, and in April 1994. Ground-water levels are measured routinely on weekly to monthly intervals, depending on the season.

METHODS

The energy needed to evaporate water links the water budget and energy budget of a lake and led to the development of the energy-budget method (Bowen, 1926). Processing of short-term Williams Lake energy-budget data for this study follows the method of Sturrock and others (1992). Two of the terms needed in the energy-budget equation are net radiation, Q^*, and change in energy stored in the lake, *S*. These terms typically are the two largest energy-budget components for periods of one-half day or less (Stannard and Rosenberry, 1991).

During a previous study at Island Lake in Nebraska, Q^* was determined using two methods (Stannard and Rosenberry, 1991). The first was the conventional method for lake evaporation studies. It consisted of measuring incoming shortwave and longwave radiation separately, using radiometers on shore, and modeling outgoing shortwave and longwave radiation. The second method consisted of deploying a net radiometer--a sensor that integrates all four of these radiation components into a single measurement--over the lake surface. The second method produced values of Q^* that were on average 67 W/m^2, or 24 percent less than the first method.

Because of the large discrepancy in Q^* values from the Island Lake study, the first method was modified for the Williams Lake study. Four radiometers were deployed over the lake surface from the tower to measure the four components of Q^* separately. Two radiometers faced upward to measure the incoming components, and two faced downward to measure the outgoing components. A net radiometer was again deployed for comparison.

Thermal surveys were conducted once or twice daily to calculate *S*. Thermal surveys consisted of measuring temperature profiles of the water column at 16 stations throughout the lake. This thermal survey schedule limits energy-budget calculation of evaporation to once or twice daily. To allow more frequent determinations of evaporation, a string of 16 thermocouples was deployed from the mid-lake tower to obtain a water-column temperature profile every 30 minutes. To calculate *S*, regressions will be developed between the tower profile temperatures and those from the thermal surveys following the method of Lee and Swancar (1997). However, for this report, *S* was calculated by assuming horizontal thermal homogeneity, based on the earlier research results of Rosenberry and others (1993).

Eddy-correlation measurements were made using a single-axis sonic anemometer and krypton hygrometer, deployed from the tower to measure fluctuations in vertical windspeed and vapor density 10 times per second. Details of the sensors and on-line processing used to compute the direct eddy-correlation measurement of evaporation are given in Stannard and Rosenberry (1991).

PRELIMINARY RESULTS

Although most of the short-term evaporation data from Williams Lake have not yet been processed, some preliminary observations regarding measurement of Q^* and *S* can be made. Measurement of Q^* using four radiometers at Williams Lake appears to be an improvement over the measurement-modeling combination used at Island Lake, based on a comparison against Q^* measured with a net radiometer (fig. 22). At Williams Lake, the net-radiometer measurement was on average 9 percent less than the four-sensor measurement, whereas at Island Lake, the net-radiometer measurement was on average 24 percent less than the measurement-modeling combination. While this constitutes a significant improvement, it is not completely satisfactory. During typical summertime conditions, this difference in measured Q^* would cause about an 8-percent difference in evaporation calculated using the energy-budget method.

The initial attempts to measure the change in energy storage, *S*, on a 30-min basis were unsuccessful. A typical time series of Q^* and *S* is shown for July 1993 in figure 23. The trace of *S* is extremely variable with typical values of plus or minus 1000 W/m^2, a standard deviation of 643 W/m^2, and no apparent diurnal variation related to Q^*. Previous studies (for example, Rodgers and Anderson, 1961; Stannard and Rosenberry, 1991) indicate that Q^* and *S* are relatively well correlated. A multiplexer was used to connect 16 thermocouples to a single data-logger channel to record the temperature profiles used to compute *S*. The thermocouples were scanned once every 30 min to produce a single profile. Sources of variability in the temperature measurements may be (1) electronic, caused by temperature gradients in the multiplexer or by other electronic factors, or (2) real, caused by small water-temperature changes as a result of small- or large-scale turbulence in the lake. A temperature change of 0.1°C throughout the lake during a 30-min period is equivalent to about 1,100 W/m^2, which indicates that the calculated 30-min values of *S* are extremely sensitive to variability from either of the above sources.

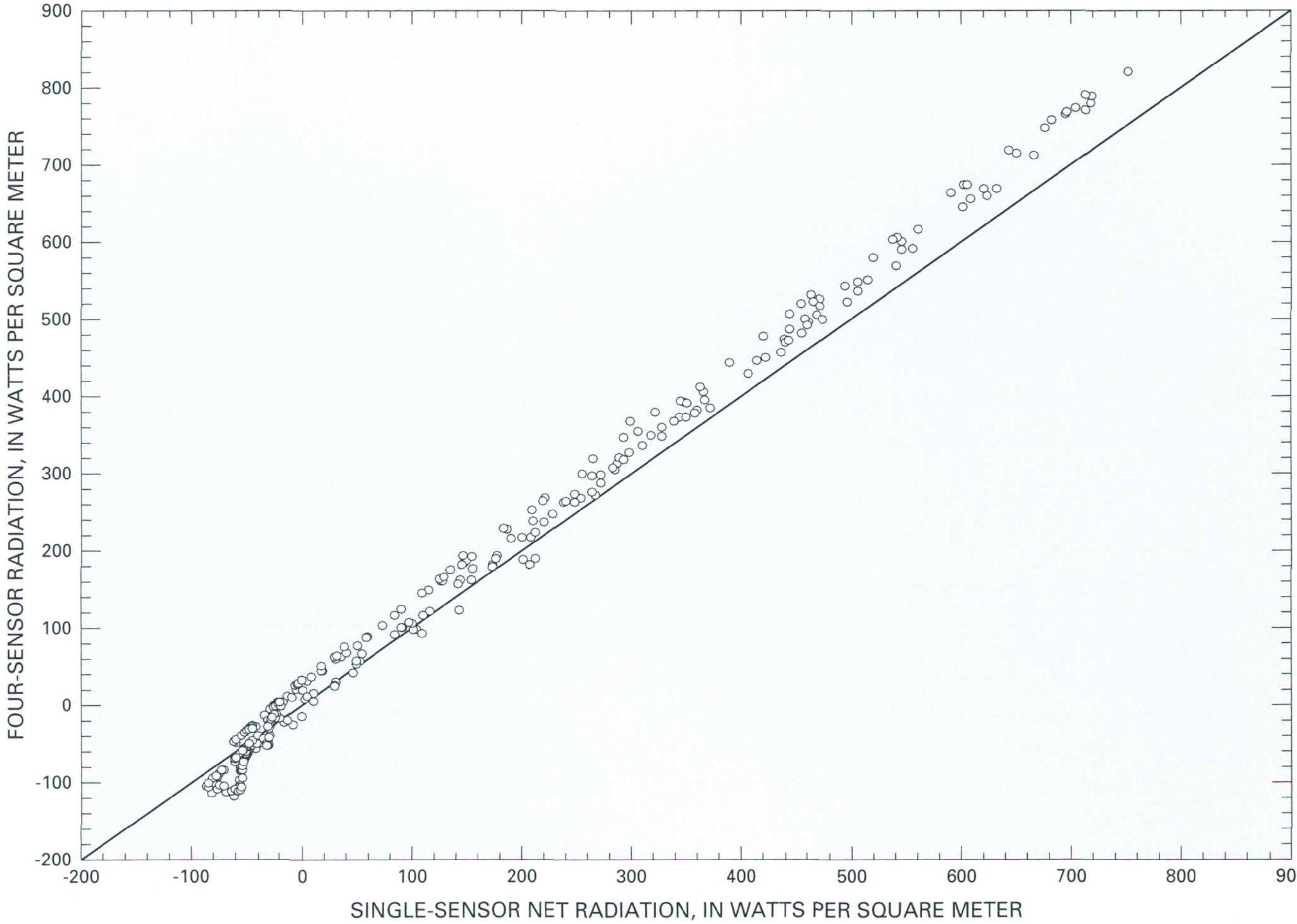

Figure 22. A comparison of 30-minute values of net radiation using a net radiometer and a four-sensor system at Williams Lake, August 1992.

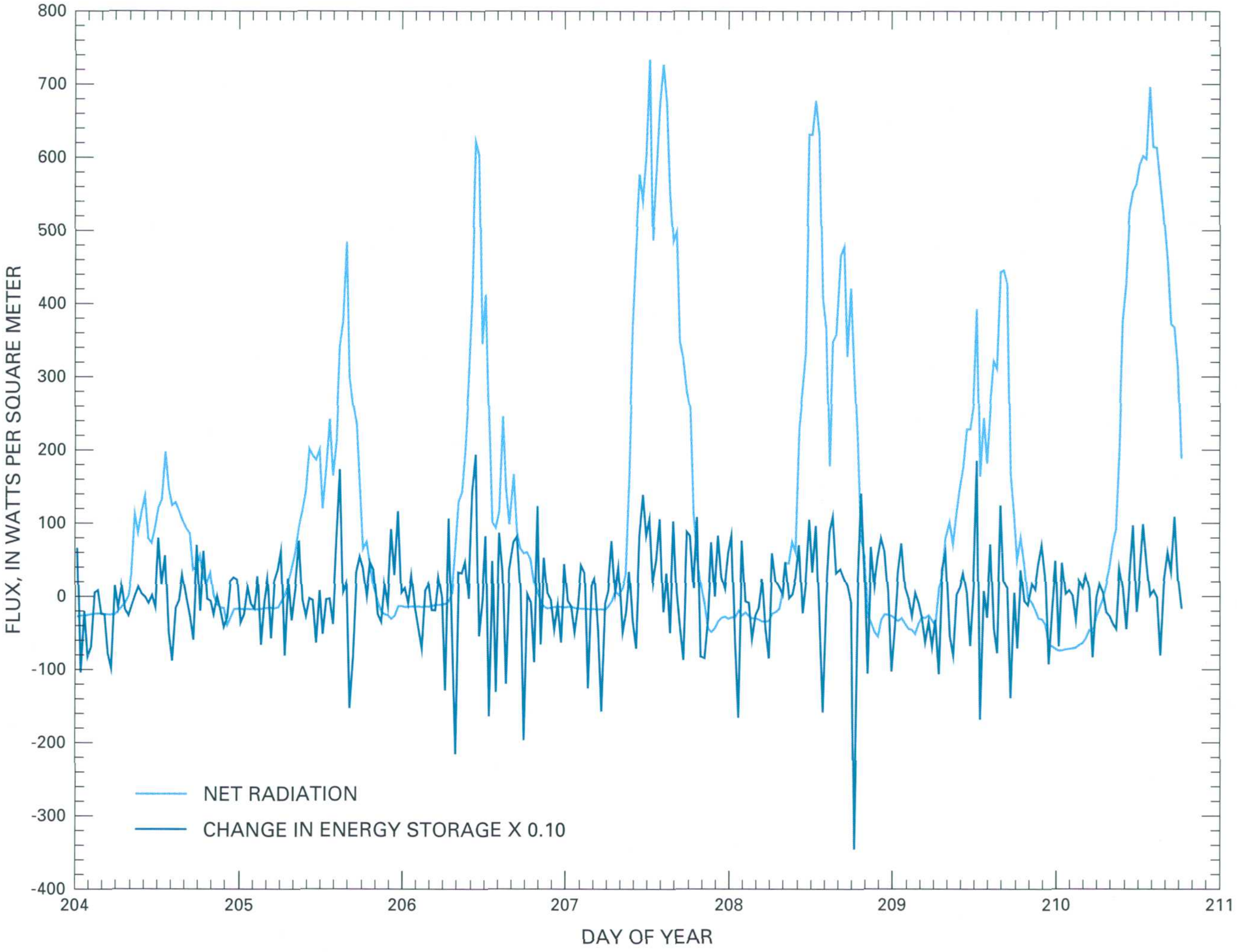

Figure 23. A time series of 30-minute values of net radiation and change in energy storage at Williams Lake, July 1993.

The random variability in *S* can largely be removed by lengthening the measurement interval. Figure 24 shows a time series of 4-hr averages of Q^* and *S*, for the same period as figure 23. The standard deviation of *S* has been reduced to 158 W/m^2. While the correlation between Q^* and *S* is only fair, it is a notable improvement over the 30-min values. These results indicate that midlake temperature profiles used to compute 30-min values of *S* probably need to be measured more frequently than 30 min (for example, 1 min) and averaged, to remove variability in the measurements.

Validation of the short-term use of the energy-budget method by comparison with eddy-correlation measurements has not been accomplished at the time of this report. Processing of short-term energy-budget data is time consuming and will involve investigating several alternatives in the computation of Q^* and *S*.

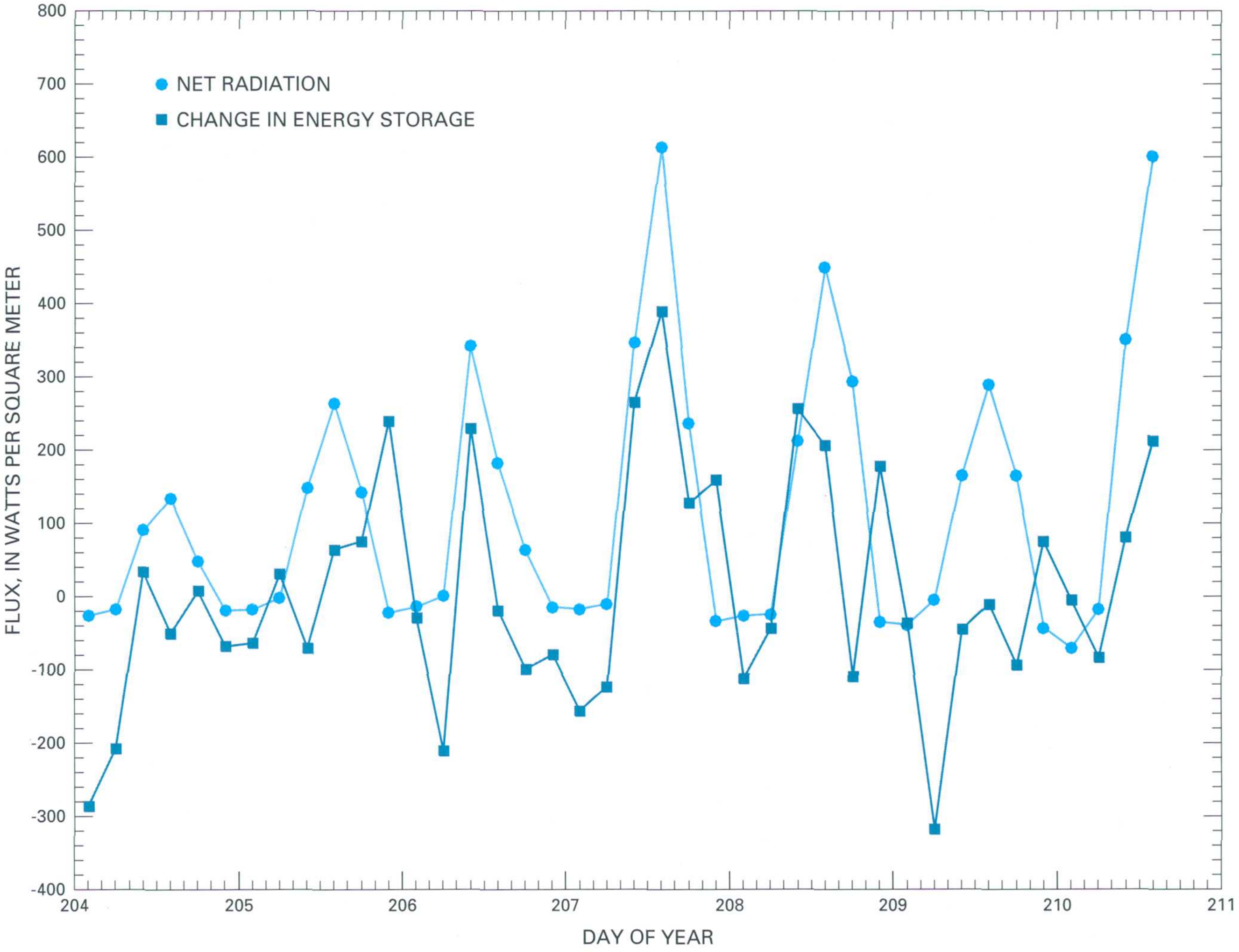

Figure 24. A time series of 4-hour values of net radiation and change in energy storage at Williams Lake, July 1993.

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Atmospheric Input to the Shingobee River Headwaters Area

By Michael M. Reddy, Paul F. Schuster, Larry J. Puckett, and Tilden P. Meyers,

INTRODUCTION

Atmospheric deposition contributes to the chemical and isotopic budgets of forested ecosystems and influences forest decline and lake acidification (Crocker and Forster, 1986; Asbury and others, 1989; Baker and others, 1991; Psenner and Schmidt, 1992). Understanding of nutrient, contaminant, and elemental cycling in northern forested ecosystems requires evaluation and characterization of atmospheric inputs (Shepard and others, 1989).

Atmospheric inputs include wet (rain and snow) and dry (gas and particulate) deposition. Collection-vessel methods, referred to as bulk (open for collection at all times) or wet-only (open for collection only during rain or snow) precipitation methods, measure wet plus dry and wet-deposition fluxes. In contrast, indirect procedures monitor dry-deposition fluxes. Hicks, for example, developed inferential techniques to estimate dry-deposition fluxes, although problems associated with large particle-size material deposition remain to be overcome (Hicks and others, 1991; Meyers and others, 1991). In addition, forest-canopy interception complicates estimation of ecosystem atmospheric input. Canopy interception (throughfall), aided by leaf uptake, modifies inputs of inorganic species.

Several groups have contributed to the understanding of atmospheric deposition. Hicks, for example, pioneered the use of filter packs to infer dry-deposition flux to watersheds (Hicks and others, 1991; Meyers and others, 1991). Puckett (1991) characterized the role of the forest canopy in the chemical budgets of watersheds. Claassen and others (1986) used atmospheric inputs to develop hydrologic budgets.

This report presents a brief review of ongoing field measurements and computational strategies to evaluate wet and dry deposition to the Shingobee River Headwaters Area. These preliminary results allow the evaluation of the wet and dry atmospheric deposition contribution to elemental budgets of the Shingobee River Headwaters Area.

The working hypothesis is that atmospheric inputs to forested ecosystems can be determined with a combination of suitable field measurements and computational techniques. This hypothesis is tested by comparing estimated atmospheric inputs to those obtained by other chemical and hydrological studies at the study site. Nitrogen and sulfur deposition serve as a focus of field-measurement efforts to test the working hypothesis.

Chemical and isotopic composition of wet-deposition inputs are determined by using collection-vessel methods (Claassen and others, 1986; Reddy and Claassen, 1985). Dry-deposition inputs are inferred from the product, $F = V(d) * [C]$ where $V(d)$ is an appropriate deposition velocity and $[C]$ is the atmospheric concentration of the species of interest (Hicks and others, 1991; Meyers and others, 1991). Throughfall measurements evaluate atmospheric input-forest canopy interaction and subsequent bioaccumulation (Puckett, 1991).

This report summarizes measurement protocols for atmospheric input of major chemical species. Research in progress includes precipitation collection at sites in the Shingobee River Headwaters Area followed by analysis of chemical constituents and stable isotopes. Dry deposition and canopy throughfall have been measured on an intermittent basis.

METHODS

The physiography of the Shingobee River Headwaters Area and the physical characteristics of the principal surface-water bodies are described by Winter and Rosenberry (this volume). Local vegetation consists of second-growth aspen and conifer forest interspersed with pastureland and wetlands. The locations of the precipitation-collection vessels and the filterpacks in the Shingobee River Headwaters Area are shown on plate 1. Atmospheric inputs to the Shingobee River Headwaters Area were monitored using an array of sampling devices. A brief description of materials, techniques, and laboratory procedures follows.

Bulk precipitation was collected in polyethylene bags housed inside a steel drum (fig. 25a) (Claassen and others, 1986; Reddy and Claassen, 1985). A restriction incorporated into the bag limited evaporation. Samples were collected biweekly (coinciding with wet-only precipitation collection) during the warmer months and monthly during the colder months (typically December through February). The collection vessel was open to the atmosphere continuously. Volume and pH were recorded at the time of collection. Samples were sent to the laboratory for analysis of chemical constituents.

An Aerochem Metrics wet/dry precipitation sampler (fig. 25b) collected cumulative bi-weekly samples of wet-only precipitation during the warmer months (typically March through November). The collector consists of two polyethylene buckets and a rain sensor. Volume and pH were recorded at the time of collection. Samples were sent to the laboratory for analysis of major ions. The "wet-side" bucket is uncovered only during rainfall and covered during dry periods to prevent evaporation and contamination by dry deposition.

A filterpack system was used to monitor dry deposition. This approach was modified slightly from that used by Barrie and others (1980), Hicks and others (1991), Meyers and others (1991), and the National Dry Deposition Network (NDDN). Air was passed through a series of filters at the rate of 5 L/m using a pump equipped with a special back pressure regulator for exact flow control (fig. 25c). Filters were changed every 2 weeks. Particles larger than 8 µm accumulate on the first filter (Teflon, 8 µm porosity). The second filter (Teflon, 1-µm porosity) collected particulates between 1 and 8 µm in size. The third filter (nylon, 1-µm porosity) removed HNO_3 vapor. Finally, a pair of cellulose filters, pretreated by immersion in a solution of K_2CO_3 and glycerol, removed SO_2.

Throughfall collection used high-density polyethylene (HDPE) funnels inserted in the mouth of 1-liter HDPE bottles (fig. 25d). Collectors attached to stakes stood at a height of about 1 meter above the forest floor. Samples were collected in triplicate under representative coniferous and deciduous canopies and open atmosphere (precipitation). Each set of samples was collected and composited on a single-storm basis. Volume, temperature, and pH were recorded at the time of collection. Samples were sent to the laboratory for analysis of major ions. Collector funnels were left uncovered between storms but were rinsed with deionized water on a regular basis to limit accumulation of material from dry-deposition; time between precipitation events was also noted.

RESULTS

Bulk precipitation has been collected in the Shingobee River Headwaters Area since April 1990. Annual volume-weighted average concentrations of six major ions (chloride, sodium, magnesium, calcium, nitrate, and sulfate) and isotopes ($D/^{18}O$ and tritium) in bulk precipitation during 1991 agreed well between collection site locations (see sites C_L in plate 1). Average ion concentrations during 1992 exhibited somewhat larger differences between collection sites than during 1991. Overall, these results suggest uniformity in precipitation volume and composition within the headwaters area. This result allows comparison of precipitation-input budgets with those developed from other basinwide chemical and hydrological techniques. Ion loadings in the Shingobee River Headwaters Area are similar to loading values at other temperate terrestrial ecosystems.

Wet-only precipitation has been collected at Williams Lake since 1985. There were no significant differences in ion loadings of chloride and sulfur between collection-vessel types (bulk and wet only). Average annual ion loadings calculated from wet-only precipitation for nitrate, sodium, magnesium, and calcium were 17, 26, 27, and 44 percent less, respectively, than average annual ion loadings calculated from bulk precipitation.

Seasonal variation in ion concentrations and loadings in precipitation occurred at both sites. Calcium, nitrate, and sulfate ion concentrations varied by as much as a factor of 10 during the year (fig. 26).

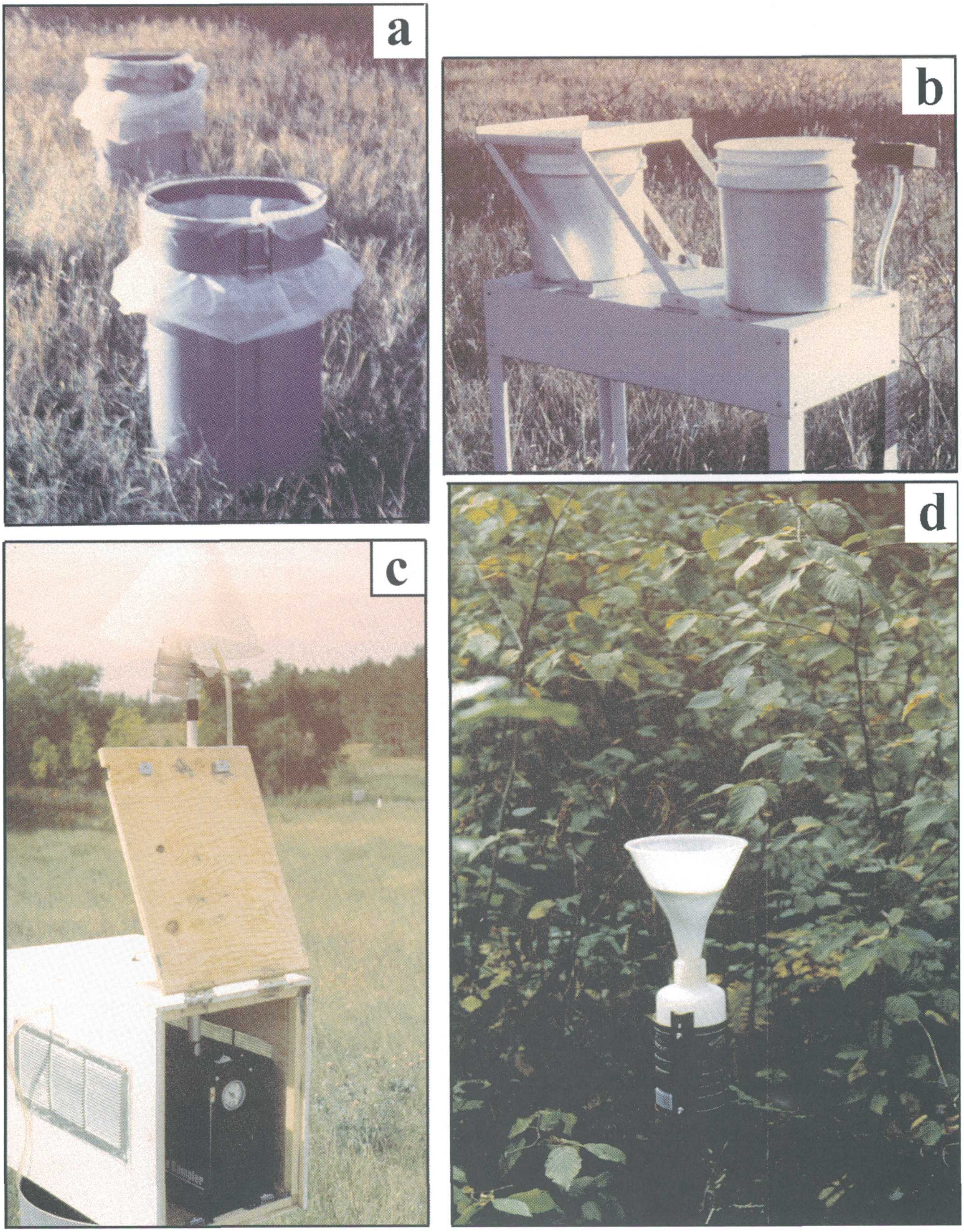


Figure 25. Sampling equipment used to monitor atmospheric inputs to the Shingobee River Headwaters Area: (a) bulk precipitation collectors; (b) wet-only precipitation collector; (c) filterpack system for dry deposition; and (d) throughfall collector.

Concentrations well above the mean value occurred several times throughout the year. Isotope values showed typical seasonal trends and all precipitation isotope data plotted on the meteoric water line.

Dry-deposition loadings of nitric acid, sulfur dioxide, and sulfate aerosol were measured continuously during the late summer and fall of 1991. Nitric acid loadings were the greatest (as much as a maximum of 40 grams/hectare/week). Values peaked at the start of sampling (August 28), decreased regularly thereafter, and increased again in late fall (November 4). Sulfur dioxide loadings, an order of magnitude smaller than those for nitric acid, exhibited a small peak at the start of the sampling period. Sulfate aerosol loadings, consistently the lowest values of the three parameters measured, varied from 0 to 3 grams/hectare/week.

Comparison of wet and dry loading values indicates that gaseous nitrogen (as nitric acid) makes a substantial contribution to the nitrogen atmospheric input budget. Gaseous sulfur dioxide, on the other hand, appears to be quantitatively unimportant. Particulate (> 8 μm) chloride, nitrogen, and sulfur contribute as much as 10 percent of the bulk precipitation concentrations. Particulate calcium and magnesium contribute as much as 20 percent of the bulk precipitation concentrations.

Throughfall precipitation (site P_t, plate 1) was collected during the 1991 and 1992 growing seasons (June-September). Regression analysis of ion deposition against the time between storms and the amount of rainfall during a storm was used to estimate the rate at which those ions accumulate in the canopy between storms and the rate at which they are leached from the canopy during storms (Puckett, 1990). The resulting regression relations were used to estimate ion leaching from and dry deposition to the canopies during the 1991 and 1992 growing seasons.

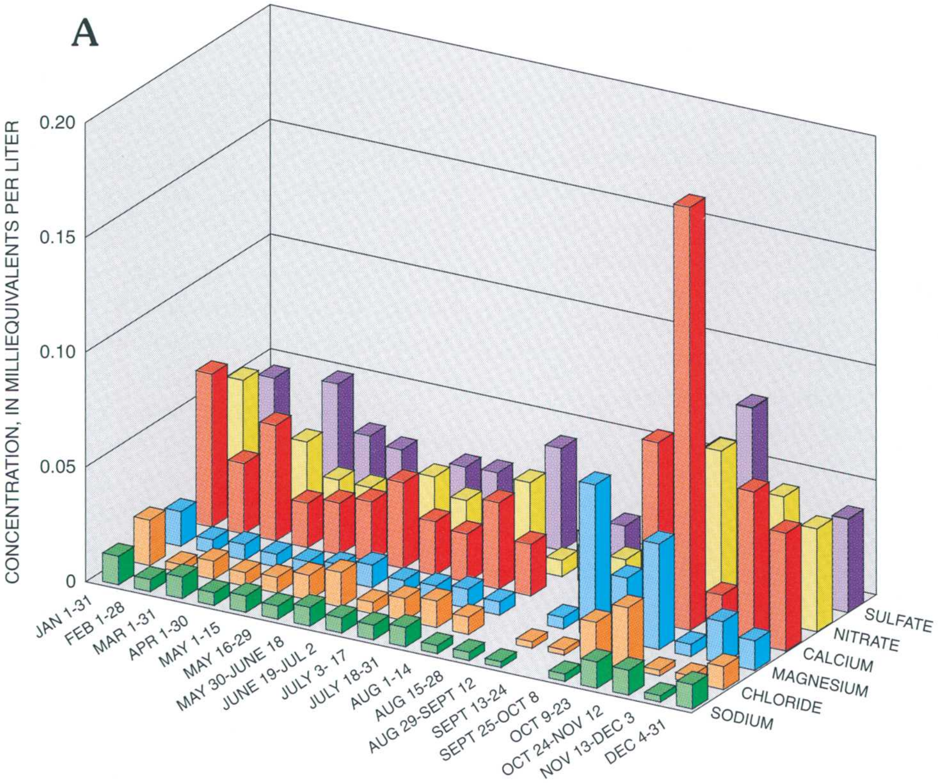
Net fluxes were greatest for calcium ion during both 1991 and 1992 under both canopies. Magnesium and chloride ion net fluxes were about equal under both canopies during the same years with the exception of 1991, when the coniferous canopy chloride ion net flux was essentially the same as that for calcium ion. There was net retention of nitrogen under both canopies during both years, probably due to limited nitrate (as nitrogen) availability in the forest soils. Sodium and sulfate (as sulfur) net fluxes were small, and in 1991 and 1992 they were negative, indicating retention by the canopy during those years.

The throughfall regression analyses against the time between storm events were statistically significant (a = 0.05) for calcium ion and magnesium ion under the deciduous canopy, and sodium ion and chloride ion under both canopies. Canopy leaching rate estimates were statistically significant for calcium ion and magnesium ion, and for nitrate nitrogen under the coniferous canopy only. For both calcium ion and magnesium ion under both canopies, leaching rates were several times those previously observed in the northern Shenandoah Valley of Virginia (Puckett, 1990). Nitrate leaching rates were significant only for the coniferous canopy, and were negative, reflecting net retention by the canopy. The statistically nonsignificant results for sulfate (as sulfur) suggest there is little uptake of sulfur from either the atmosphere or the soil, consequently little is available in either canopy for either washoff or leaching.

Canopy leaching and dry deposition contributions to throughfall were estimated by multiplying the appropriate regression coefficients by precipitation amounts and the time between events during the 1991 and 1992 growing seasons. For both canopies, leaching of calcium ion and magnesium ion and dry deposition of chloride ion accounted for the majority, if not all, of the net flux of those ions during both growing seasons. Dry deposition of calcium ion and magnesium ion was significant under both canopies but, with the exception of magnesium ion under the deciduous canopy in 1991, it represented less than half of the leaching estimate.

The relatively large amount of leaching of both calcium ion and magnesium ion under both canopies is probably related to their availability in the glacially derived soil which contains calcite and dolomite. This soil is also the likely source of the dry deposited calcium ion and magnesium ion in the deciduous canopy. Given the forested nature and general lack of agriculture in much of the surrounding area, suspension of dust from local unpaved roads probably provides most of these two constituents in dry deposition and in precipitation.

In spite of the relatively remote north-central United States location of the Shingobee site, nitrate (as nitrogen) appears to be the constituent delivered in greatest quantities from outside the watershed. During the growing season, this amounted to 0.6-0.8 kg/ha and was 2.1 kg/ha on an annual basis. In comparison, nitrate (as nitrogen) inputs to the northern Shenandoah



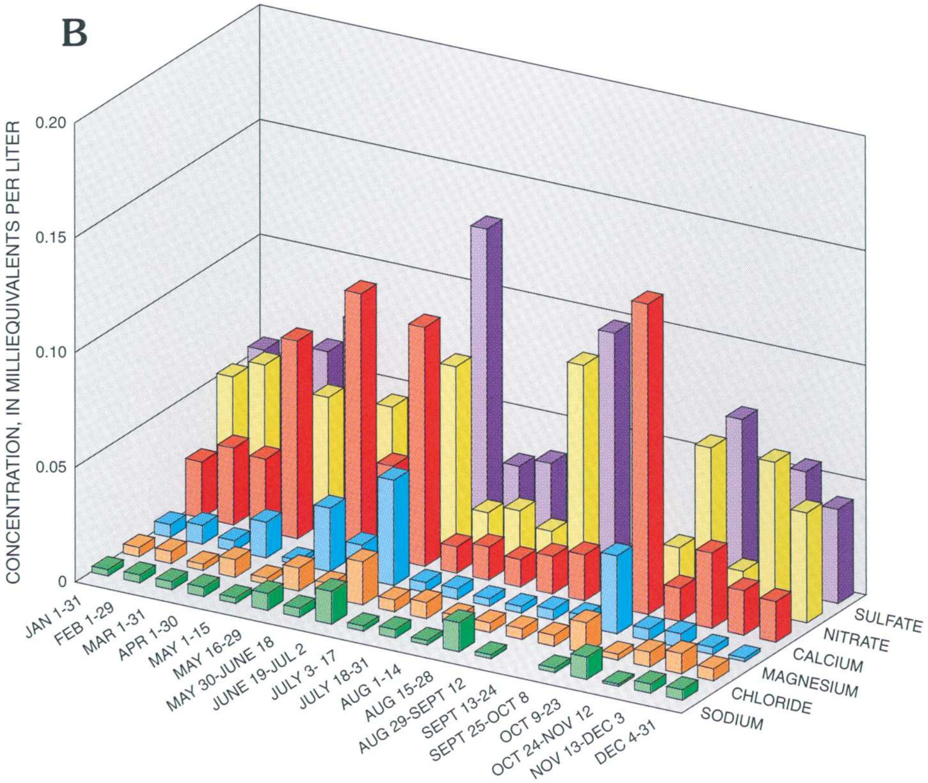


Figure 26. Concentrations of major ions in bulk precipitation in the Shingobee River Headwaters Area during (A) 1991 and (B) 1992.

Valley of Virginia in 1983 were approximately 1.8 and 2.8 kg/ha during the growing season and the entire year, respectively. Based on comparisons of the wet and bulk deposition chemistry it would appear that most (84 percent) of the nitrate (as nitrogen) enters the watershed in wet precipitation, whereas at the Virginia site wet and dry deposition contributions were about equal. It is important to note that the forest canopy at the Shingobee site is much more efficient at retaining nitrate (as nitrogen) inputs than the Virginia site as evidenced by the net retention at the Shingobee site compared to positive fluxes in Virginia. This pattern suggests that these northern Minnesota trees may be more nitrogen limited and therefore dependent on precipitation inputs of nitrogen.

Deposition of sulfate (as sulfur) in precipitation at the Shingobee site was about 0.9-1 kg/ha during the growing season and 3.1 kg/ha on an annual basis. Both deciduous and coniferous throughfall fluxes were about equal to the precipitation inputs. The fact that both the leaching and dry deposition estimates for sulfate (as sulfur) under both canopy types were not significant, suggests that most of the sulfur entering the Shingobee watersheds is in precipitation. This is important in that dry deposition in its various forms in eastern watersheds usually accounts for about one-half to one times the amount of sulfate (as sulfur) as wet deposition. For example, at the Virginia site wet deposition was about 5 kg/ha during the growing season compared to 8.7 and 8.5 kg/ha under the deciduous and coniferous canopies, respectively. Furthermore, wet deposition of sulfate (as sulfur) at the Shingobee site (3.1 kg/ha/yr) was less than half that in Virginia (7.1 kg/ha/yr), and represented about 97 percent of the total (bulk) deposition. Overall, these results suggest that dry deposition of gases, aerosols, or particles is a minor source of sulfur to the Shingobee River Headwaters Area.

CONCLUSION

Atmospheric inputs to the Shingobee River Headwaters Area comprise an important component of the chemical and isotopic budgets for Williams and Shingobee Lakes. Understanding of chemical distributions and fluxes in this forested ecosystem requires knowledge of atmospheric deposition processes throughout the year. These atmospheric inputs, which are one of the sources of chemical constituents in ground-water recharge, establish an end member for chemical and stable isotopic concentrations for the ground water-surface water system in the study area. That is, the source chemical and isotopic composition of recharge water in this area is ultimately controlled by atmospheric inputs and subsequent transformations.

A comparison of samples from 2 collection sites indicates uniformity in precipitation volume and composition within the headwaters area. Loadings calculated from wet-only precipitation were as much as 44 percent less than those calculated from bulk precipitation. Seasonal variation in precipitation ionic composition ranged by as much as a factor of 10. Isotope values showed typical seasonal trends and plotted on the meteoric water line. Nitrogen and to a smaller extent, sulfur, accumulate in the forested ecosystem by gas-phase dry deposition. Particulate (>8 μm) chloride, nitrogen, and sulfur contribute up to 10 percent of the bulk precipitation concentrations. Particulate calcium and magnesium contribute up to 20 percent of the bulk precipitation concentrations. Throughfall precipitation data are consistent with dry and wet deposition data, reflecting the relative contributions of dry deposition and the importance of wet deposition to the Shingobee River Headwaters Area.

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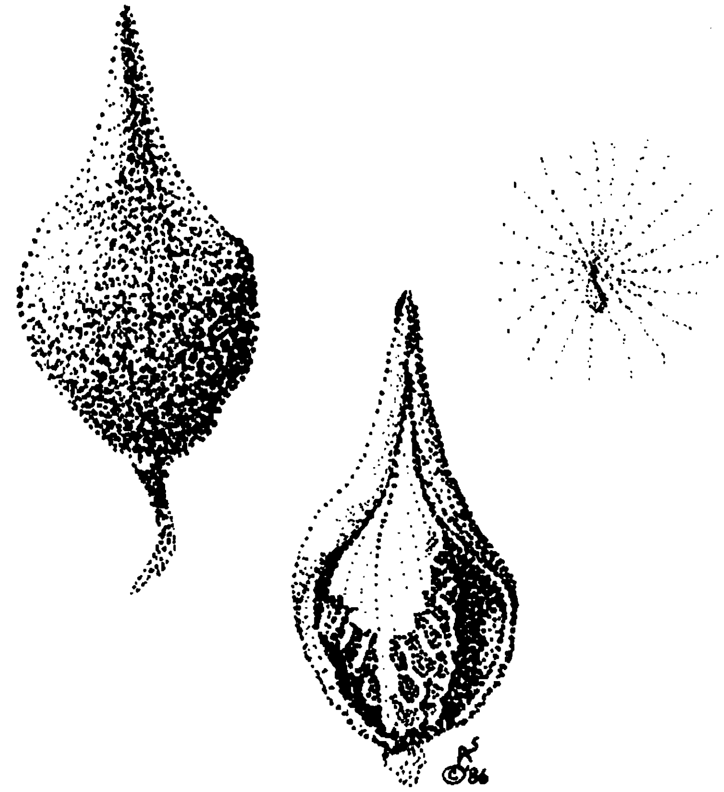
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Chemical Fluxes Between Williams and Shingobee Lakes and Their Watersheds

By James W. LaBaugh

INTRODUCTION

Water transports biologically important chemical elements into and out of lakes and their watersheds (Likens, 1984). In this context, studies of chemical fluxes into and out of lakes rarely have included examination of the role of ground water; the focus commonly has been on surface water flowing to or from lakes. Thus, few comprehensive studies have been made on chemical fluxes in lakes that do not have streams or rivers flowing to or from them. Even in lakes that have channelized surface-water inputs and outputs, little is known about the relative contribution of ground-water fluxes to and from those lakes.

The comprehensive hydrologic research at the Shingobee River Headwaters Area provides the opportunity to address the question: What is the relative contribution of ground water to chemical fluxes to and from Shingobee and Williams Lakes? Other related questions are (1) How does the relative contribution of ground-water chemical fluxes vary as a function of seasonal and inter-annual changes in climate? and (2) What is the effect of external chemical fluxes on processes within the lakes?

The purpose of this research is to quantify hydrologically mediated chemical fluxes to and from Shingobee and Williams Lakes, including interaction with ground water, and the resultant effect on chemical characteristics of the lakes. Natural variability in chemical characteristics over time is used to understand how these lakes respond to changes in hydrologic conditions, in part due to the fact that experimental manipulation at the ecosystem level is not practical or possible at this site. Major ions, silica, nitrogen, and phosphorus are the chemical elements of interest for this research.

METHODS

Samples have been collected on a regular basis from Williams Lake since 1980 and from Shingobee Lake and the Shingobee River since 1989. Samples were collected from Williams Lake, Shingobee Lake and the Shingobee River upstream and downstream from Shingobee Lake approximately every 2 weeks when the lakes were not ice covered, and approximately monthly when the lakes were ice covered. Ground-water samples in the vicinity of Williams Lake were collected at irregular intervals since 1979 from numerous wells; selected wells have been the subject of approximately monthly sample collection when lakes were not ice covered. Wells in the vicinity of Shingobee Lake were added to the sample-collection regime in 1992. Atmospheric deposition also was collected at irregular intervals using a wet-dry collector near Williams Lake (Reddy and others, this volume).

Although Shingobee and Williams Lakes are the focus of detailed study because of the comprehensive hydrologic investigations of those lakes, samples were collected from other lakes and wetlands in the Shingobee River watershed to characterize conditions of water bodies adjacent to those of principal interest. Samples were collected from Mary and Crystal Lakes in 1985, Chase Marsh in 1986, and Little Shingobee Lake in 1991.

All chemical analyses were done at the U.S. Geological Survey National Laboratory. Laboratory analyses included determinations of pH, specific conductance, calcium, magnesium, sodium, potassium, chloride, sulfate, alkalinity, iron, manganese, silica, total phosphorus, soluble reactive phosphorus, total kjeldahl nitrogen, nitrate, nitrite, and ammonia. Methods of analysis are described in Fishman and Friedman (1985).

At the time of sample collection in the lakes, field determinations of water temperature, pH, specific conductance, and dissolved oxygen were made at 1-meter intervals in the water column from the surface to the bottom of the lakes using a submersible multiparameter probe. Field determinations of the same characterstics were made in the Shingobee River using the multiparameter probe at the time of sample collection. Temperature, pH, and specific conductance were measured in the field at the time of collection of ground-water samples for chemical analysis.

RESULTS

A gradient in major ion concentrations in the lakes within the Shingobee River Headwaters Area is related to hydrologic setting. Lakes having smaller concentrations, Crystal Lake and Chase Marsh (table 4), were those in which the water balance is determined by atmospheric deposition, atmospheric exchange, and loss to ground water. Lakes in which the water balance also included ground-water input and/or surface-water input -- Williams, Mary, Shingobee, and Little Shingobee Lakes -- have larger concentrations. Major ion concentrations in Shingobee Lake, through which flows the Shingobee River, were about double those in Williams Lake, which has no surface-water inlet or outlet (table 5).

Lakes without surface-water input had smaller values for total phosphorus than lakes with surface-water inputs. Total phosphorus concentrations in Shingobee Lake were more than twice those in Williams Lake. Hydrologic setting appears to be less of a factor in nitrogen concentrations; total kjeldahl nitrogen concentrations were similar in Shingobee and Williams Lakes, but concentrations of dissolved inorganic nitrogen (nitrate nitrite, ammonium) were larger in Shingobee Lake (table 5).

Preliminary analysis of data from April to September 1991 indicates ground-water input accounted for 53 percent of water input to Williams Lake; the remainder came from rainfall. Ground-water input was 30 percent less than lake discharge to ground water. Measurement error for ground-water fluxes was assumed to be 50 percent. Most of the major ion supply to the lake came from ground-water input, accounting for more than 85 percent of the major ion input. A larger amount of major ions entered the lake than left by ground water; yet, in the short period analyzed, major ion concentrations in the lake declined instead of increasing. Mechanisms that could account for this loss of material include chemical precipitation or biologically mediated precipitation. Additional information on the interaction of Williams Lake and ground water is given by LaBaugh (1988, 1991). In contrast to major ions, almost all of the phosphorus input and most of the nitrogen input to Williams Lake from April to September 1991 came from rainfall.

Analysis of the volume of water moving to and from Shingobee Lake from ground water and the Shingobee River has not been completed to the point where interpretations of water and chemical balances can be made. However, the lake is quite similar in major ion composition to the Shingobee River, which is similar to ground water collected in the vicinity of Williams Lake (table 5). Thus, there may be considerable influence of ground water on the chemical characteristics of the Shingobee River with respect to major ions. Total phosphorus concentrations also were similar in Shingobee Lake and the Shingobee River. Total kjeldahl nitrogen and ammonium nitrogen were larger in the Shingobee Lake than in the Shingobee River.

Chemical flux studies of lakes commonly are of short duration, taking place over the course of a few years. Consequently, few studies span the time over which hydrologic conditions vary naturally in response to variable climate. Recognition that long-term studies are key to understanding complex environmental issues is growing (Likens, 1988), as is emphasis on comparative studies of ecosystems to understand how these systems function (Cole and others, 1991). Thus, the study of Shingobee Lake, dominated by surface water, and Williams Lake, dominated by ground water, provides for the long-term comparisons needed to understand how hydrologic processes affect these aquatic ecosystems. To do this will take the combined, coordinated efforts of many specialists, including hydrologists, chemists, and ecologists.

Table 4. Comparison of chemical characteristics for data obtained on selected dates from lakes and wetlands in the Shingobee River Headwaters Area
[< detection. less than the analytical detection limit]

Lake or wetland	Date of sample collection	pH	Specific conductance, in microsiemens per centimeter	Concentration, in milliequivalents per liter						
				Ca	Mg	Na	K	Cl	SO_4	Alkalinity
Crystal Lake	09 13 85	8.7	140	0.848	0.559	0.060	0.020	0.011	0.079	1.32
Mary Lake	09 12 85	8.3	298	2.04	1.07	0.121	0.033	0.014	0.127	2.90
Williams Lake	09 12 85	8.4	145	0.998	0.567	0.052	0.028	0.011	0.043	1.48
Chase Marsh	06 06 86	6.2	42	0.184	0.090	0.043	0.145	0.042	0.201	0.239
Shingobee Lake	08 15 91	8.0	341	2.09	1.40	0.252	0.043	0.101	0.079	1.40
Little Shingobee Lake	08 08 91	8.0	372	2.40	1.48	0.282	0.017	0.008	0.037	4.12

Lake or wetland	Date of sample collection	Concentration , in millimoles per liters							
		SiO_2	Fe	Mn	Total phosphorous	Ortho phosphorous	Total kjeldahl nitrogen	Nitrate plus nitrite nitrogen	Ammonia nitrogen
Crystal Lake	09 13 85	0.048	0.071	0.018	0.0001	< detection	0.042	< detection	0.0002
Mary Lake	09 12 85	0.183	0.214	0.055	0.0002	0.00006	0.028	< detection	0.0004
Williams Lake	09 12 85	0.015	0.304	0.146	0.0002	0.00006	0.049	< detection	0.0016
Chase Marsh	06 06 86	0.034	14.7	2.91	0.0047	0.0042	0.014	< detection	0.0049
Shingobee Lake	08 15 91	0.183	0.089	1.53	0.0014	0.00009	0.042	0.0022	0.0105
Little Shingobee Lake	08 08 91	0.249	0.179	0.055	0.0004	0.00006	0.042	< detection	0.0004

Table 5. Comparison of chemical characteristics of Shingobee Lake, Shingobee River, Williams Lake and adjacent ground water, 1989-92

[Values are medians for the period of study; < detection, less than the analytical detection limit]

Characteristic	Shingobee River above lake	Shingobee River below lake	Shingobee Lake at 1 meter	Williams Lake at 1 meter	Ground water into Williams Lake
Lab pH	7.9	7.9	8.15	8.1	7.6
Lab specific conductance, microsiemens per centimeter	364	360	366	178	416
Major constituents, milliequivalents per liter					
Calcium	2.47	2.40	2.40	1.25	3.24
Magnesium	1.24	1.40	1.32	0.616	1.23
Sodium	0.180	0.258	0.239	0.060	0.113
Potassium	0.040	0.046	0.040	0.023	0.030
Chloride	0.028	0.035	0.033	0.011	0.047
Sulfate	0.104	0.069	0.082	< detection	0.187
Alkalinity	3.84	3.99	3.90	1.89	4.40
Minor constituents, millimoles per liter					
Silica	0.216	0.232	0.232	0.009	0.266
Minor constituents, micromoles per liter					
Iron	1.19	1.05	0.197	0.161	0.161
Manganese	1.05	1.01	1.03	0.081	0.245
Total phosphorus	0.742	0.775	0.742	0.323	0.291
Ortho phosphorus	0.194	0.178	0.032	< detection	0.097
Total kjeldahl nitrogen	28.5	28.5	42.8	42.8	14.2
Nitrate plus nitrite nitrogen	3.49	2.24	2.28	1.99	7.17
Ammonium nitrogen	2.39	3.07	10.7	2.92	3.60

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Comparison of Dissolved Trace-Metal Concentrations for Selected Lakes in Minnesota (pH ~8-9) and New York (pH ~5-7)

By Alexander van Geen

INTRODUCTION

Adsorption of metal ions such as Cd, Cu, and Zn onto mineral oxide surfaces in laboratory studies depends strongly on pH (Davis and Kent, 1990). While adsorption no doubt is important in natural systems as well, it is not clear which particles in a watershed play a dominant role in regulating dissolved metal concentrations: soil particles, lake and stream particles, or sediment. This needs to be understood to predict the effect of continued acid rain deposition in a poorly buffered region such as the Adirondack State Park, New York (pH 5-7). Williams and Shingobee Lakes (pH 8-9) cover the alkaline end of the pH spectrum of continental waters. Therefore, the composition of these lakes may provide a basis for comparison with acid-rain-impacted systems and help to identify where in the watershed adsorption processes control dissolved metal concentrations.

METHODS

Surface-water samples of Williams and Shingobee Lakes and additional nearby lakes were collected in acid-washed polyethylene bottles from a small moving boat during August 1990. Samples were filtered through acid-washed 0.4-µm Nuclepore polycarbonate filters. Further processing of samples was done in a Class 100 laminar-flow bench. Samples were acidified with 1 mL/250 mL of 12 N Ultrex HCl. Samples from Dart Lake, Moss Lake, and nearby waters in the Adirondack region were collected in a similar fashion in June 1985. Concentrations of Al, Fe, Mn, Cd, Cu, and Zn were determined by direct-injection, graphite-furnace atomic absorption spectroscopy (GFAAS) with calibration by standard additions.

RESULTS

The comparison between the Minnesota and Adirondack lakes is simplified by similarities in their flow regimes. The inflow to Shingobee Lake is the Shingobee River, which flows through several smaller lakes. The residence time of water in Shingobee Lake is on the order of 6 months (Averett and Winter, this volume). In contrast, Williams Lake is a closed-basin lake; the residence time of water is closer to 3 years. In the Adirondack region, Dart Lake is located downstream from Big Moose Lake (fig. 27). The dissolved metal composition of Big Moose Lake and Dart Lake is very similar. While the residence time of water for Dart Lake is only 2 weeks, the residence time for the Big Moose Lake system overall is about 7 months. There is no inlet to Moss Lake, but the residence time of water relative to the outflow estimated from the watershed area is about 11 months. Because suspended particle concentrations in the Adirondack and Minnesota lakes are also similar (~ 2 mg/L), the time of exposure of metals in the water column to potential adsorption by inorganic particles is assumed to be comparable for Shingobee, Big Moose, and Moss Lakes.

For Al, Mn, Cd, and Zn, there is a striking difference between the composition of the Minnesota lakes and the Adirondack lakes, with significantly lower dissolved metal concentrations in the more alkaline system (table 6). For both the Minnesota and the Adirondack regions, several additional nearby lakes were sampled for trace metals. Results very similar to the data in table 6 indicate that the patterns discussed here are representative of a broader area. The range of dissolved Al concentrations in the lakes studied here is also similar to that reported by

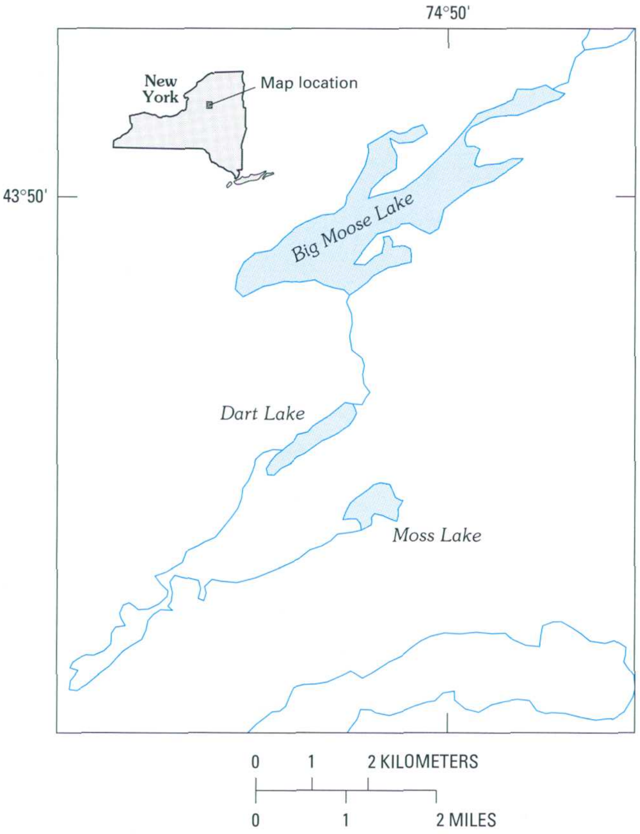

Figure 27. Location of Big Moose, Dart and Moss Lakes in the Adirondack Mountains of New York.

Driscoll and others (1988) in a comparison of Al concentrations in 9 streams in British Columbia (pH, 6.6; Al, 0.7 µM) and Hubbard Brook, New Hampshire (pH, 4.8; Al, 20µM). Dissolved concentrations of Cd and Zn as high as 1 nM and 250 nM, respectively, have been reported for acidified Swedish lakes (Stumm, 1992). Zn concentrations measured in Dart and Moss Lakes (250 and 93 nM, respectively) are comparable to levels measured earlier by White and Driscoll (1987).

Assuming that the bedrock composition is not much different from one region to the other, the high metal to low pH relation is generally consistent with the idea that metal cations tend to adsorb more strongly onto mineral surfaces at elevated pH. Somewhat lower Al, Cd, and Zn concentrations in Moss Lake (pH 6.6) relative to Dart Lake (pH 5.0) indicate that water column pH may play a role in regulating dissolved metal levels. But the much larger regional differences in Al, Zn, and Mn concentrations between Minnesota and the Adirondacks suggest that it is within the watershed soil rather than in the lake water column that the pH sensitivity of metal mobilization has a larger impact.

Mobilization into lake waters does not appear equally sensitive to pH for all metal cations, however. table 6 shows that levels of Fe and Cu in the Adirondack and Minnesota lakes are quite similar. For reasons not clearly understood (possibly organic complexation in the case of Cu), the pH dependence of adsorption of these elements does not appear to be the dominant factor regulating their dissolved concentrations in lake water. Cu concentrations are higher in Williams Lake; this may reflect the longer residence time of water and diagenetic remobilization from sediments, which has been documented for the deep ocean (Klinkhammer, 1980).

Table 6. Selected chemical characteristics of Minnesota lakes and New York lakes

[mg/L, milligrams per liter]

Lake	pH	Total suspended load mg/L	Al	Fe	Mn	Cd	Cu	Zn
			micromoles	micromoles	micromoles	nanomoles	nanomoles	nanomoles
Shingobee	8.4	~2	0.03	0.11	0.019	<0.05	1.6	<0.1
Williams	8.9	~2	0.14	0.06	0.007	<0.05	9.6	<0.1
Moss	6.6	1.9	2.9	0.3	0.3	0.07	2.4	93
Dart	5.0	1.9	9.4	0.2	1.0	0.90	2.6	250

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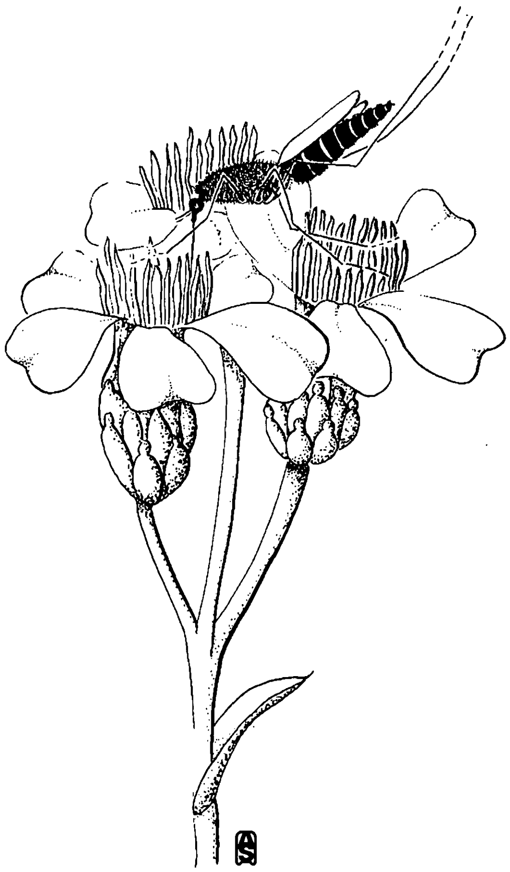

Application of Stable Isotopes in the Shingobee River Headwaters Area

By Carol Kendall, William C. Evans, Michael M. Reddy and Paul F. Schuster

INTRODUCTION

Lakes commonly are in direct hydraulic communication with the surrounding ground-water system. Lakes may serve as recharge or discharge zones for the contiguous ground-water system, or as both if the lake is the surface expression of flow down the hydrologic gradient. Traditional estimates of ground-water-lake exchange are made by careful measurements of hydraulic head around the lake. Because processes such as evaporation, biological activity, chemical reactions, and selective recharge can cause lakes and ground water to be chemically distinctive, chemical tracers can also be used to quantify hydrologic budgets. However, because of reactivity of some dissolved constituents, isotope tracers are often better tools for determining hydrologic budgets.

Isotope techniques take advantage of the fact that lakes, precipitation, and the surrounding ground-water system commonly are isotopically different. Isotopes of oxygen and hydrogen are especially useful tracers of water sources because they are constituents of the water molecule itself. Applications of stable isotopes for the study of lakes are described by Dincer (1968). Later review articles include those of the International Atomic Energy Agency (1979), Gat (1981), Gilath and Gonfiantini (1983), Gonfiantini (1986), and Krabbenhoft and others, (1994). The main applications of isotopes to lake systems are for the determination of water balances, nutrient uptake studies, and paleotemperature reconstructions. In this chapter, a number of useful applications of isotopes in the Shingobee River Headwaters Area are briefly described. Locations of isotope sampling sites are shown in plate 1.

THE LOCAL METEORIC WATER LINE

Isotopic compositions of rain samples and water samples from the Shingobee River, various wells in the area, Williams Lake, and Shingobee Lake, plotted with the Global Meteoric Water Line (GMWL) for reference, are shown in figure 28. The almost identical local meteoric water line (LMWL) defined by 111 samples of rain collected over 10- to 30-day intervals is $\delta D = 7.76\ \delta^{18}O + 6.41$ ($r^2 = 0.97$), which is close to the line for North America: $\delta D = 7.95\ \delta^{18}O + 6.03$ (Gat, 1981). Most of the river, lake, and groundwater samples generally do not plot along the LMWL; instead they define an evaporation line that has a slope of 4.87 and a y-intercept of about -25.04 per mil ($r^2 = 0.99$, n = 156). The water in Williams Lake, a closed-basin lake, has a stronger evaporation signature than water in Shingobee Lake, an open-basin lake. Pairs of samples collected at the inlet and outlet of Shingobee

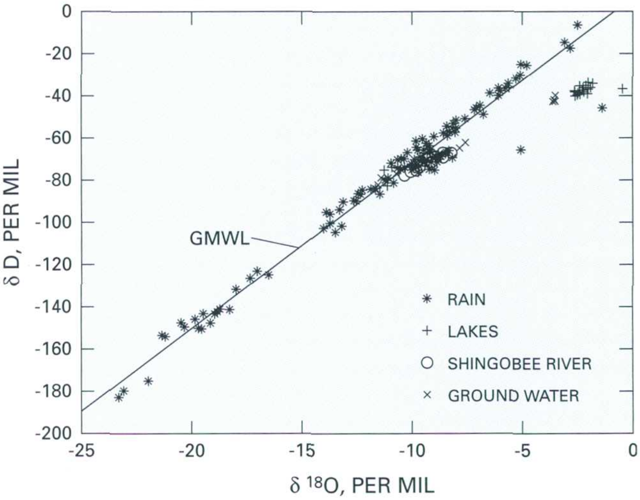

Figure 28. Oxygen and hydrogen isotopic compositions of meteoric-derived waters in the Shingobee River Headwaters Area, 1989-1992.

Lake show that the water at the outlet always has a stronger evaporation signature than the water at the inlet, the average shift is $\delta^{18}O = 0.7$ per mil and $\delta D = 3.5$ per mil in 1992.

Rain samples were collected at both lakes from 1990 to 1993. The collection durations ranged from 2 to 5 weeks and were identical for both collectors. Despite the close proximity of the lakes, the pairs of samples (n = 35) were not identical in isotopic composition. Individual pairs of samples showed differences as large as ±12 per mil in δD and ±3.6 per mil in $\delta^{18}O$, average differences are -3.0 per mil and +0.55 per mil, respectively. However, when these 35 pairs of samples were used to calculate the isotopic compositions of total rain from the two sites, the differences were insignificant: rain from the Shingobee Lake area was only 0.9 per mil enriched in D and 0.11 per mil enriched in ^{18}O relative to the Williams Lake area, a difference well within the 2 σ error bars of the analyses. The equations of the LMWL's calculated for samples from the two collectors were not significantly different. Hence, although rain samples collected from one lake are not suitable as indicators of the composition of rain at the other lake on the scale of 2 to 5 weeks, on a yearly basis, the differences are not significant. These data point out that spatial variability of the isotopic composition of rain on scales of less than 10 km^2 can be significant for short-term studies (Kendall and McDonnell, 1993).

HYDROLOGIC BUDGETS

Our working hypothesis is that a minor amount of ground water flows into Williams Lake and that lake water flows out of Williams Lake, and that this occurs mainly in the littoral zone. We are investigating whether seasonal changes in $\delta^{18}O$, δD, and in conservative chemical species at different depths in the lake and lake sediments can be used as indicators of relative contributions of ground water to lakes. These estimates of ground-water contributions to Williams Lake will be compared to those determined by energy and hydrologic data.

To address these questions, lake water from near the center of Williams and Shingobee Lakes and interstitial pore water from sediments in the littoral zone of Williams Lake were collected for isotopic and chemical analysis. Lake samples taken in April, August, and October of 1989 and February 1990 were analyzed for δD and $\delta^{18}O$ (fig. 29). These preliminary data can be used to calculate the fraction of ground water contributing to well-mixed portions of the lake. Such calculations can be easily applied to (1) the total lake during the winter when the ice cover prevents both evaporative loss and resultant isotope fractionation, and precipitation gain and consequent isotopic mixing; and (2) lake water below the hypolimnion (>7 m) in the summer when there is little mixing of deep water with shallow water above the thermocline. For example, for the period October 1989 to February 1990 and using values of $\delta^{18}O = -11.5$ per mil and $\delta D = -82$ per mil for the ground-water input into Williams Lake obtained by extrapolating the evaporation trend in the lake waters to the LMWL, the calculated amounts of lake water derived from ground water are in the range of 2 to 3 percent. Work is continuing to further refine the estimates of ground-water input to Williams Lake through a geochemical approach.

Pore-water peepers (sediment-water samplers) were installed to collect interstitial pore water from the sediments in the littoral zone of Williams Lake (sites I and X, plate 1) to obtain an independent estimate of ground-water contribution. Peepers (fig. 30A) are constructed by drilling cells into solid blocks of lucite; the cells were covered with semipermeable membranes (Smith and Klug, 1981) that would permit small amounts of ambient water to be collected in each cell. The peepers were installed in the littoral zone at a water depth of about 1m. Two or three cells were exposed to lake water above the sediment-water interface. The peepers equilibrated with in-situ pore waters for about 1 month before being removed (fig. 30B) and sampled. All samples were collected and processed in a nitrogen-filled glove box to limit oxidation of the pore water (fig. 30C).

A two-component mixing model, using the average $\delta^{18}O$ and δD of ground-water and lake-water compositions as end-members, was used to identify the relative contributions to the water in each cell. Figure 31 shows results for samples collected in October 1992 following a prolonged drought at a ground-water inflow site (site I) (plate 1) and a lake-water outflow site (site X). Ground-water inflow areas and lake-water outflow areas were determined using hydrologic data (Siegel and Winter, 1980). At the inflow area (site I) the isotopic composition of pore waters changes sharply in the top 10 cm below the lake-sediment interface. This change in isotopic composition reflects an abrupt change in the

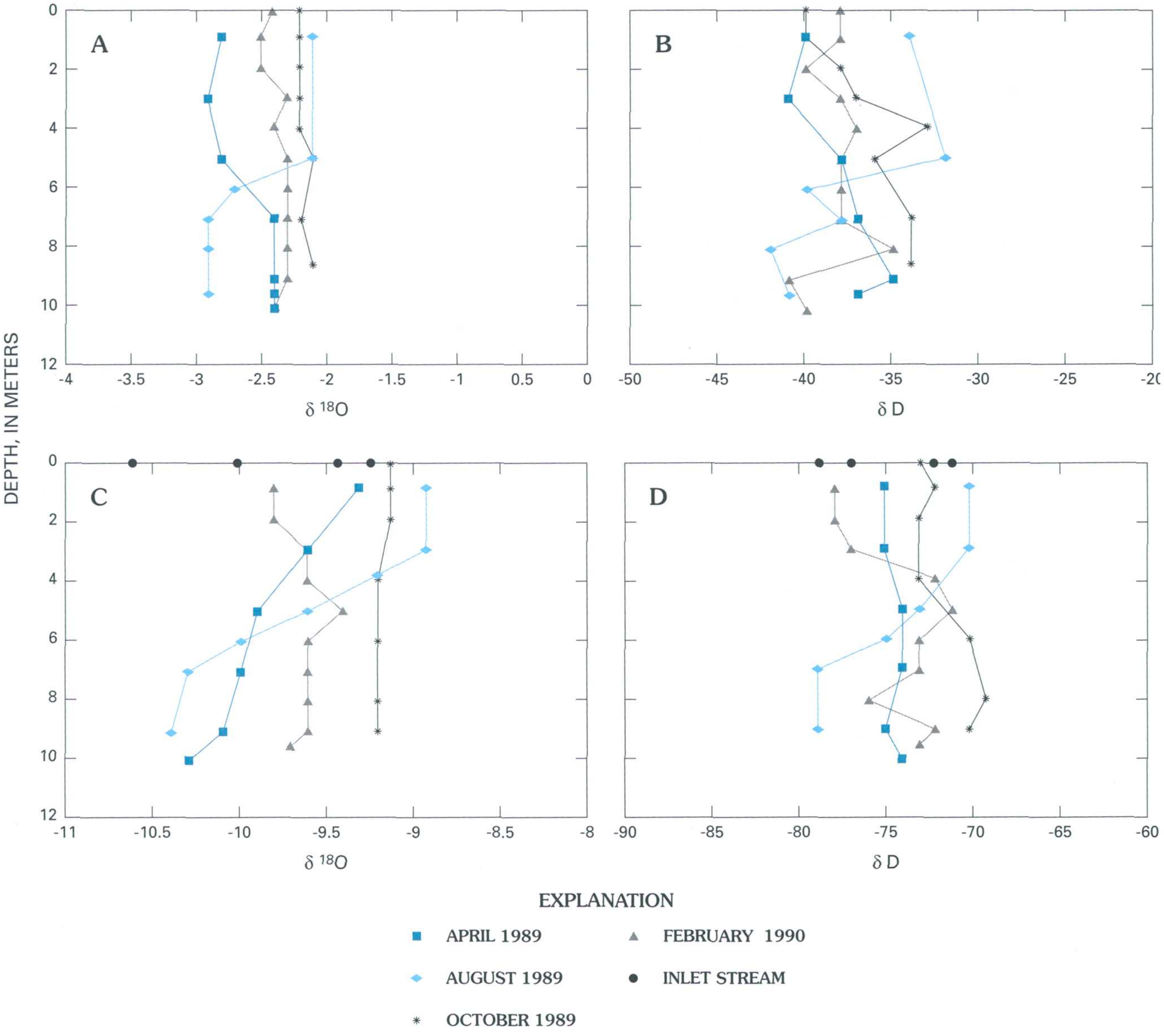

Figure 29. *Changes in $\delta^{18}O$ and δD with depth for lake samples collected at Williams and Shingobee Lakes on four collection dates, (A) Williams Lake $\delta^{18}O$, (B) Williams Lake δD, (C) Shingobee Lake $\delta^{18}O$, and (D) Shingobee Lake δD.*

percentage of lake water in the pore waters, decreasing from about 100 percent lake water at the interface to 20 percent lake water below 10 cm (fig. 31A). In contrast, lake contributions to ground water gradually decrease with depth at the outflow area (site X) for the upper 25 cm of sediments (fig. 31B). The apparent increase in lake-water contribution below 30 cm reflects the assumption that the end-members have seasonally constant composition. As is shown in figure 29, both end members show considerable seasonal variability, which will be taken into consideration in later modeling efforts.

To further constrain the mixing model, bulk-precipitation, lake, and ground-water samples were also analyzed for tritium content (fig.32). Values are reported in tritium units (TU), analytical error bars average ± 0.7 TU. The average tritium content of precipitation in 1992 is 13 TU. This is lower than samples from the lake and all ground-water wells except the piezometer at the CHASE nest (C90 and C120). The samples from these latter two wells (screened at 27 and 37 m) have tritium contents less than 5 TU and consequently are composed largely of water recharged prior to 1960. The other ground waters have tritium contents ranging from 27 to 85 TU, suggesting significant contributions of water recharged after 1960. The high tritium content of the ground-water samples from the outflow area of Williams Lake (wells UM28 and WL18, fig. 32) relative to the current tritium contents of lake and precipitation indicate that water flow in this area is very slow. The very high value from WL18 indicates large contributions from water recharged during atmospheric nuclear-bomb testing in the early 1960's to early 1970's. Later modeling efforts will use tritium data to estimate lake residence times and maximum ground-water flow rates in the outflow zone of the lake.

The tritium content of bulk precipitation samples and lake water collected at 2- and 8-m depths in Shingobee and Williams Lakes (fig. 33) show significant seasonal variability. There is a rough correlation of the variability in precipitation with that of samples from Shingobee Lake. In contrast to the Shingobee Lake data, samples from Williams Lake show less seasonal variability and no apparent correlation with the tritium content in precipitation. The Williams Lake samples collected in the fall show similar patterns, indicating that waters at both depths were affected by the same hydrologic event. The slightly higher and significantly less seasonal variability of tritium content of water at 8 m relative to the content at 2 m supports the hypothesis that lake-outflow rates are slow.

CARBON CYCLING

The $\delta^{13}C$ of dissolved inorganic carbon (DIC) in the Shingobee River Headwaters Area was measured to assess whether the carbon isotopic composition is controlled primarily by hydrologic processes such as mixing or gas exchange, or mainly by biological cycling of carbon. One set of samples collected in September 1990 along a transect down the Shingobee River, across Little Shingobee Lake, and across Shingobee Lake, plus one sample from Williams Lake, exhibits a positive correlation ($r^2 = 0.91$ for $n = 11$) between the $\delta^{18}O$ of water and the $\delta^{13}C$ of DIC (fig. 34). The $\delta^{18}O$ and δD values along the transect plot along an evaporation line that has a slope of 4.9 which is indicative of evaporation along the transect. Longer residence times of water during this dry time of the year result in more evaporation, causing the waters to become more enriched in ^{18}O and D. Increased water residence times also result in more exchange of carbon between the surface-water bodies and the atmosphere ($\delta^{13}C = -7$ per mil), causing an enrichment in ^{13}C of the DIC. If biological processes were the main control of $\delta^{13}C$, productivity variations in different areas might be expected to produce variation in $\delta^{13}C$ unrelated to the degree of evaporation and ^{18}O enrichment. Increased productivity down the transect would cause a progressive enrichment in ^{13}C because of preferential uptake of ^{12}C by organisms; this is a possible but unlikely cause for the regular increase in $\delta^{13}C$ with distance. Hence, it appears that the $\delta^{13}C$ of surface waters during the fall mainly reflects hydrologic processes, not biological recycling. Additional sampling would be needed to determine if biological processes might overwhelm hydrologic processes during other seasons.

The Shingobee River Headwaters Area is well suited for examining the use of isotopic tracers to study watershed processes. For example, do the $\delta^{18}O$, δD, and $\delta^{13}C$ values of lake carbonates, plants, and animal remains reflect annual or seasonal climate variations or do they reflect temporal variations in local environmental conditions that are unrelated to climate? A number of investigators are trying to use the isotopic compositions of ancient materials as indirect or "proxy" records of the ancient climatic conditions, taking advantage of the fact that the isotopic compositions of carbonates, plants, and other materials form in isotopic equilibrium with local waters. A major question for such studies is how to separate isotopic variability caused by climate change from variability produced by a variety of hydrologic mechanisms.

Another question is related to the spatial variation of isotopic tracers. Can the seasonal and spatial variation in the δD and $\delta^{18}O$ of ground water provide more detailed information about the interactions of ground water and surface water at seasonal ponds and ground-water mounds than is currently being generated by hydraulic measurements? A preliminary investigation at a similar site in Delaware (Phillips and others 1993) showed promising results.

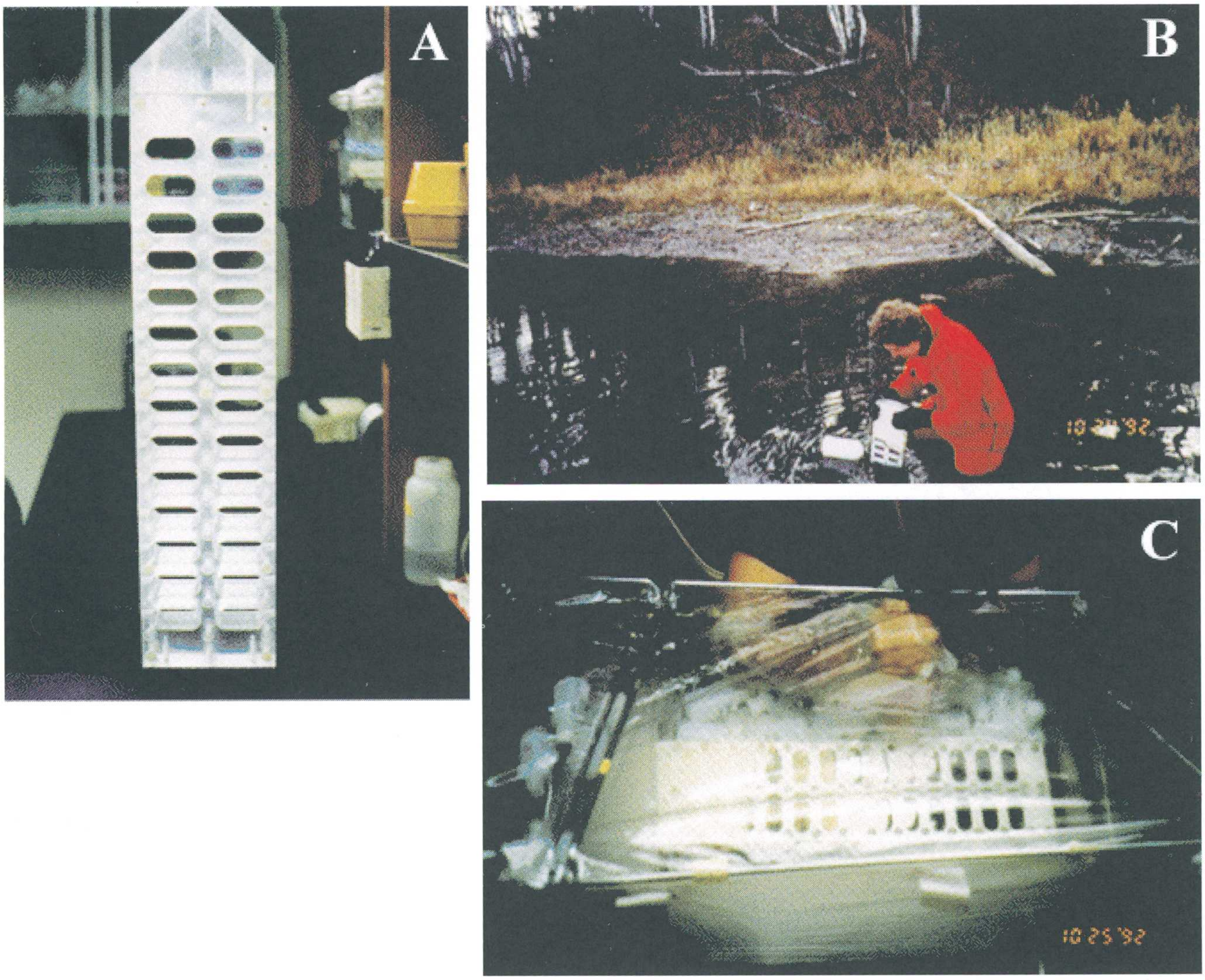

Figure 30. (A) Pore-water peeper used to collect duplicate samples of pore water from multiple depths in the sediments of the littoral zone of Williams Lake, (B) removal of a pore-water peeper from the sediments of the littoral zone of Williams Lake, and (C) collection and processing of pore water from a peeper inside a nitrogen-filled glove box to limit oxidation of the pore water.

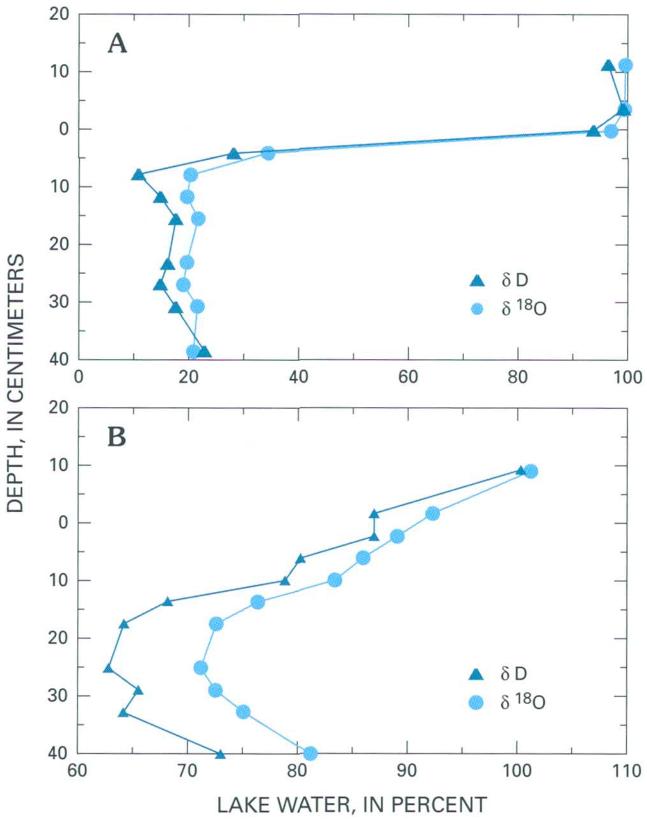

Figure 31. Relative contributions of lake water to littoral-zone pore waters collected at various depths below the lake-sediment interface of Williams Lake at (A) site I (inflow) and (B) site X (outflow).

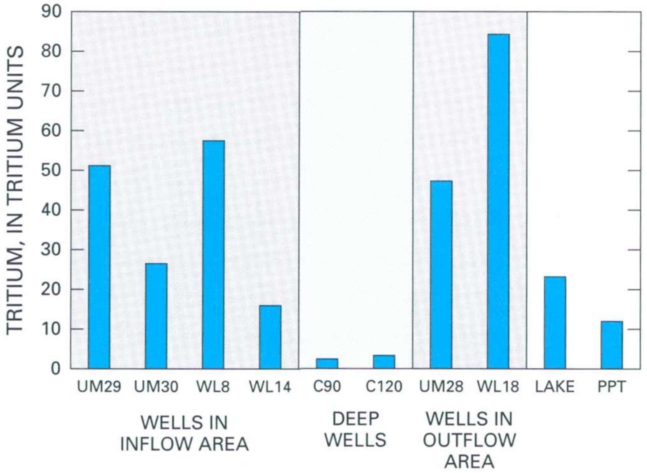

Figure 32. Average tritium concentrations of ground water, Williams Lake water, and precipitation (ppt) in 1992. All wells are completed just below the water table and are less than 5 meters deep except for C90 and C120, whose depths are 27 and 37 meters.

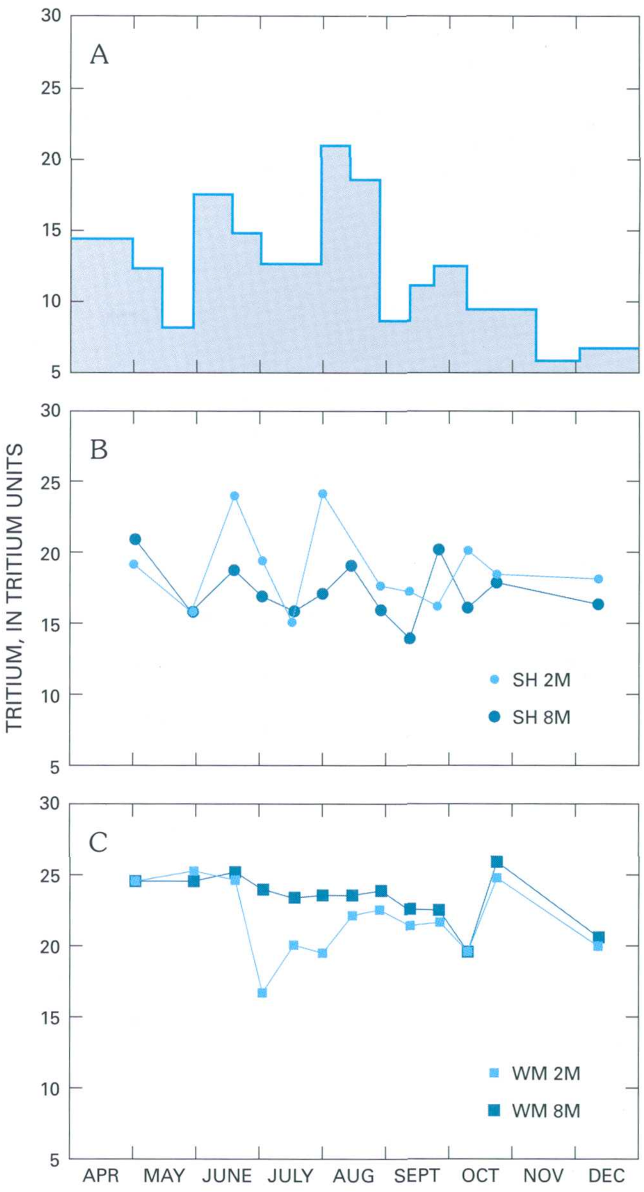

Figure 33. Tritium concentrations in samples collected April to December 1992: (A) bulk precipitation, (B) Shingobee Lake 2-m depth (SH 2M) and 8-m depth (SH 8M), and (C) Williams Lake, 2-m depth (WM 2M) and 8-m depth (WM 8M).

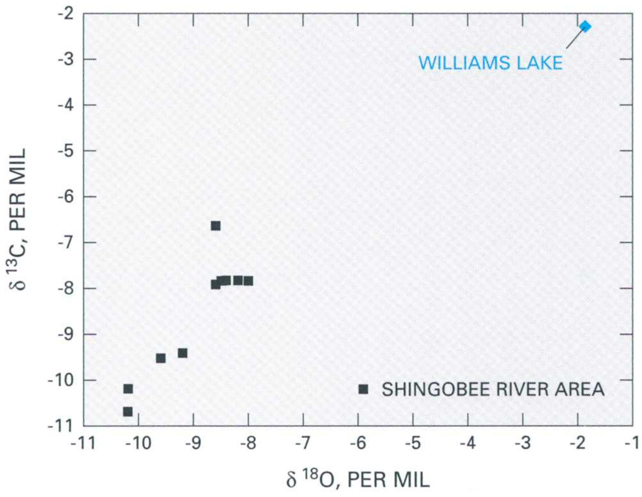

Figure 34. Correlation of $\delta^{18}O$ of water and $\delta^{13}C$ of dissolved inorganic carbon for samples collected in September 1990 from the Shingobee River area and Williams Lake.

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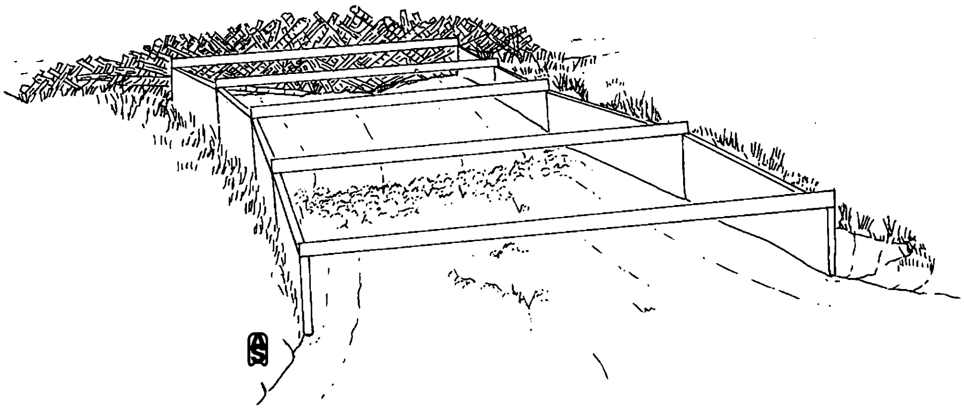

Effects of Hydrologic Factors on Organic Matter in the Williams and Shingobee Lake Systems

By George R. Aiken and Diane M. McKnight

INTRODUCTION

During the past 15 years, environmental scientists have placed increasing importance on understanding the geochemical and ecological roles of organic matter in aquatic ecosystems. The complexity of the biosphere, however, makes study of the nature, chemical reactivity, and ecological roles of organic matter difficult. The lithosphere is not divided into distinct environmental compartments that can be understood independently of neighboring environments. For instance, it is not possible to understand the chemistry of a stream without considering the chemistry of the local soil and ground water. It is also necessary to consider factors such as variations in hydrology, vegetation patterns in the watershed, and airborne inputs to the system. The chemical characteristics of the dissolved organic carbon (DOC) are the net result of all the biogeochemical processes occurring within the system. Changes in hydrology, for example, can influence not only the amount of organic matter within a stream but its chemical characteristics as well. Therefore, understanding the relative importance of different sources and processes pertaining to DOC in an ecosystem is complicated.

The Shingobee River Headwaters Area is a good location to study basic questions concerning the sources, transport, and fate of organic carbon in surface and ground waters. The major differences between Williams Lake and Shingobee Lake for the purposes of studying DOC are the inputs to the lakes and the residence times within the lakes. Williams Lake has no surface-water inlet or outlet, and the estimated residence time for water within the lake is about 3 years. The hydrology of Williams Lake has been extensively studied, and the inorganic chemistry of the water column and sediments have been characterized. Ground-water inflow is the primary factor in the ionic composition of the lake (LaBaugh, this volume). Shingobee Lake, on the other hand, has inflow and outflow streams and a residence time of approximately 6 months. Sources for the DOC and, more specifically, humic substances in Shingobee Lake include not only the ground water and sediments but also material brought into the lake by the inflowing Shingobee River.

In aquatic ecosystems, the sources of DOC can be categorized as (1) allochthonous, entering the system from the terrestrial watershed, and (2) autochthonous, being derived from bacteria, algae, or aquatic macrophytes growing in the water body. In temperate systems, the allochthonous sources are very important. Most of the DOC originates from the degradation and leaching of organic detritus in the watershed and is transported by streams and shallow ground-water flow. The chemical characteristics of the DOC are influenced not only by the source material but also by the biogeochemical processes involved in carbon cycling within the system. These processes include allochthonous flow of organic carbon to the system from the watershed, autochthonous carbon fixation by algae and aquatic plants, transformation and degradation of both autochthonous and allochthonous organic material by heterotrophic microbial activity, transport of particulate organic material to the sediments, remobilization of DOC from the sediments, and photodegradation by incident ultraviolet (UV) light. Each of these processes is likely to affect the DOC chemistry in a different way.

In natural waters, organic acids constitute the most abundant class of DOC. Because DOC accounts for much of the total organic material flux in lakes and streams, these dissolved organic acids are themselves significant in carbon cycling. However, only a small fraction of these compounds are low molecular-weight organic acids that can be identified using current technology. The remaining organic acids can be classified in two operationally defined fractions as fulvic acids and hydrophilic acids. Aquatic fulvic acids are a mixture of yellow organic acids, with number average molecular weights of 500-1,000 daltons. Aquatic fulvic acids are biologically refractory, meaning that they are difficult to degrade, and they usually constitute from 20 to 80 percent of the DOC. Their biogeochemistry has been studied in numerous aquatic environments (Aiken and others, 1985). Hydrophilic acids are more hydrophilic than fulvic acids primarily due to a greater abundance of acidic functional groups (Aiken and others, 1992). They too are heterogeneous mixtures of organic acids. Hydrophilic acids comprise 5-20 percent of the DOC of the relatively few systems in which they have been studied.

RESULTS

Williams Lake has greater DOC concentrations (≈7 mg C/L) than Shingobee Lake (≈4.5 mg C/L), but Williams Lake has a deeper euphotic zone. DOC analyses, UV absorbance measurements of whole water samples, and elemental and ^{13}C-nuclear magnetic resonance spectroscopy (NMR) analyses of aquatic fulvic acid (FA) isolates indicate that the organic matter in Shingobee Lake is dominated by the allochthonous material transported to the lake in the Shingobee River. The organic matter in Williams Lake, on the other hand, is dominated by autochthonous sources. The ^{13}C-NMR method of analysis is an important technique for obtaining structural information about organic molecules. The ^{13}C-NMR data presented in figure 35 show the differences between aquatic FA samples isolated from Williams Lake, a

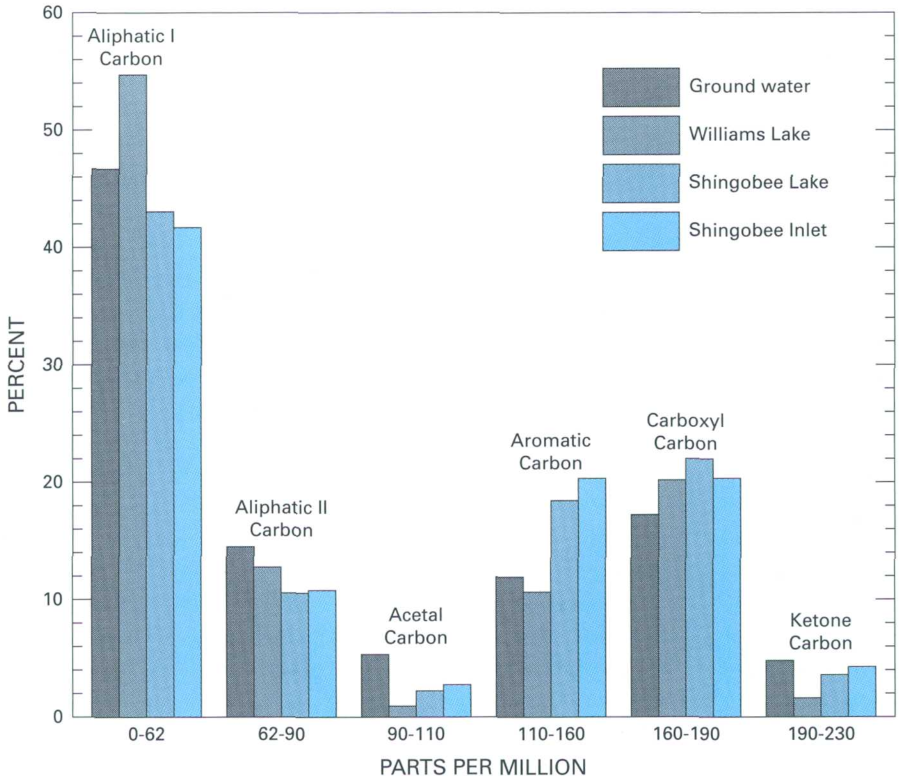

Figure 35. Quantitative ^{13}C-NMR integration data for fulvic acid samples from Williams Lake, ground water entering Williams Lake, Shingobee Lake, and the Shingobee River inlet to Shingobee Lake

nearby shallow ground-water well, Shingobee Lake, and the Shingobee River inlet to the lake. These data indicate that the FA in Williams Lake is more aliphatic (0-62 ppm) and less aromatic (110-160 ppm) than FA samples from Shingobee Lake, the Shingobee River, or ground water entering Williams Lake. The differences in organic matter between the two lakes is, in large part, due to chemical differences in the sources of the DOC. Aquatic FAs derived from different source materials have distinctive characteristics associated with those source materials (McKnight and others, 1991). Aquatic FAs derived from lignin, for example, have relatively large amounts of aromatic carbon and are high in phenolic content.

Another process that may affect the characteristics of the aquatic FAs in the lakes is photochemical oxidation. McKnight and others (1988) showed that the aromatic component of the humic material in Spirit Lake, Washington, resulting from the eruption of Mt. St. Helens in 1980 decreased significantly in the 3 years following the eruption. These losses were thought to be due to photo-oxidative degradation of the humic material in the surface waters of the lake. Other studies have indicated that the aromatic components of marine FA are more likely to be degraded by photo-oxidative degradation than other components (Chen and Bada, 1989). The organic matter in Williams Lake would be subject to a greater amount of photo-oxidation than the material in Shingobee Lake because of the longer residence time in the lake. It is possible that photo-oxidation of the DOC in Williams Lake causes, in part, the loss of aromatic components of the FA.

The transport of allochthonous organic carbon into streams and lakes largely depends on the hydrologic conditions within the watershed. Concentrations of soluble humic substances are substantial in surface soil solutions and vary significantly between different soil and ecosystem strata, vegetation types, and seasons. When the soil is undersaturated with water, precipitation can percolate through the upper soil horizons and the unsaturated zone, eventually recharging ground water. Within the upper soil horizons, DOC concentrations can be fairly high; but as the organic matter is transported through the soil and unsaturated zones, it is subjected to a number of processes, including sorption and biodegradation, that result in decreases in concentration. For example, Cronan and Aiken (1985) noted that, during the summer growing season, DOC concentrations in the interstitial waters of podzol soils in the Adirondack Mountains ranged from 21 to 32 mg C/L in the O/A horizon, from 5 to 7 mg C/L in the B horizon, and from to 2 to 4 mg C/L in ground-water solutions. Under low-flow conditions, ground-water discharge to the stream generally is the major source of water in the stream, and the organic matter in the stream will be representative of that transported by ground water.

The amount and nature of organic matter in the Shingobee River has a strong seasonal variation. Samples collected during the melting of winter snowpack (spring flush) were found to have the greatest DOC concentrations and were the most aromatic of the samples studied. Figure 36 shows the spring flush for the Shingobee River, presenting both DOC concentration data and UV absorbance data collected over a 6-month period. The UV absorbance measurements at λ=254 nanometers (nm) are an estimate of the aromaticity of the DOC in these samples. Each spring the DOC concentrations rise in association with snowmelt. Not only does the amount of organic matter in the river change, but the quality of this material changes as well. While the DOC concentrations increased to approximately twice the background levels, the UV absorbance tripled. This indicates that the organic matter being removed from the soil is more aromatic, and likely of higher molecular weight, than material found in the stream under base-flow conditions. During high-flow conditions, when the soil is saturated with water, the flow of water within the soil column shifts to a more horizontal direction. Shallow subsurface and surface runoff is common. The organic matter transported in these waters has a shorter residence time in the subsurface and is not subjected to the same processes as material transported to the water table. This shift in flow path results in higher DOC concentrations and compositional changes in the organic matter transported to the stream.

Impounding the stream water, whether in a lake or reservoir, averages out the variations in DOC concentration associated with increased stream discharge. Most water bodies are subject to various degrees of physical mixing. Within a lake, for example, physical mixing results in spatially uniform DOC chemistry. This effect can be observed by

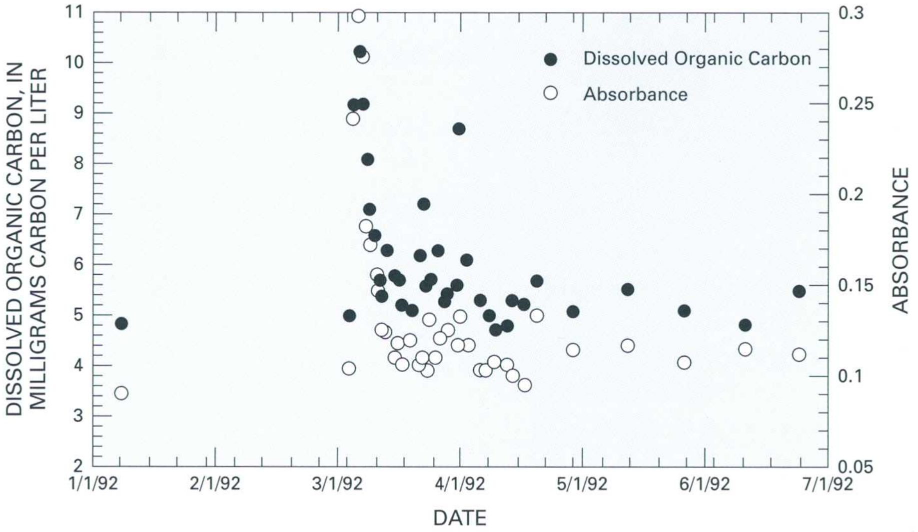

Figure 36. Dissolved organic carbon (DOC) and ultraviolet light absorbance data for the Shingobee River from January 1992 to July 1992

comparing the DOC data for the Shingobee River to the corresponding data for Shingobee Lake (fig. 37). Pulses of organic matter that enter the lake are disseminated in the existing organic matter. The data show that the lake is relatively unresponsive to event-driven changes in the amount of DOC. Rather, the DOC data for the lake resemble the base flow data for the inflow stream.

The ^{13}C-NMR data presented in figure 35 also show the differences between aquatic FA samples isolated from Shingobee Lake, the Shingobee River inlet to the lake, and a shallow ground-water well near Williams Lake. The ground-water FA has greater aliphatic carbon content (0-62 ppm) and less aromatic carbon (110-160 ppm) than does the sample from the Shingobee River. These differences reflect the effects of sorption and biodegradation that occur within the unsaturated zone on the ground-water FA compared to the river sample. The river is in more direct hydrologic contact with the upper soil horizons of the watershed, and it contains organic matter from these zones as well as DOC associated with ground water discharging to the stream. The FA from Shingobee Lake, which receives both allochthonous and autochthonos DOC, is intermediate between these two samples. The effect of this difference in FA chemistry is reflected in the light-absorbance properties of each water body. Compared to the lake sample, the absorbance of UV light (fig. 38) is less for the ground-water sample and greater for the stream sample collected in summer. Also plotted in figure 38 are absorbance data for the Shingobee River collected during the spring flush. This sample had the greatest light absorption of the samples studied, reflecting increased aromatic carbon content.

Variations in the characteristics of the organic matter, especially with respect to changes in aromatic carbon content and molecular weight, are significant factors controlling chemical reactivity. Future research at the site will include studies of the processes that control the fate and transport of organic compounds in ground water and at the sediment-water interface in the lakes and streams. By studying the characteristics and fluxes of DOC in Williams and Shingobee Lakes, it will be possible to more clearly define the processes controlling DOC in both ground and surface waters. Increased understanding of these processes, in turn, will enable scientists and engineers to better understand the ecological and geochemical roles of DOC in aquatic systems and to devise effective strategies to preserve the quality of water resources.

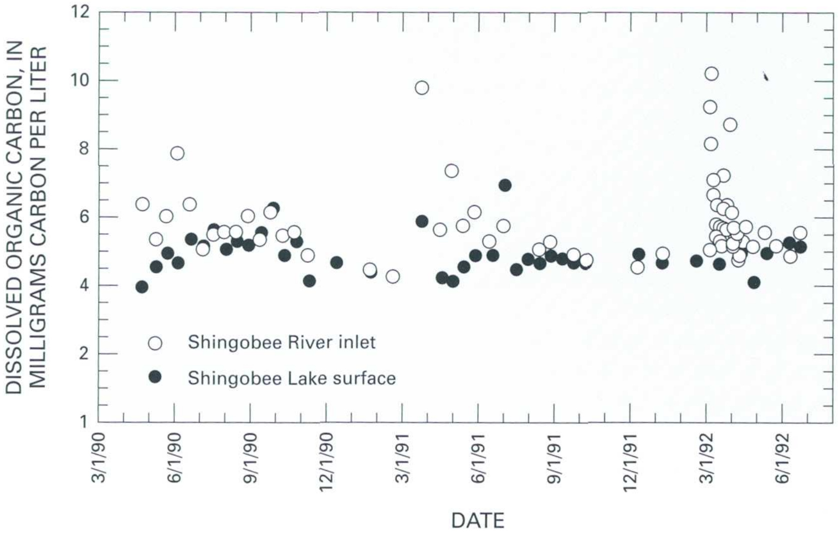

Figure 37. Dissolved organic carbon (DOC) data for the Shingobee River and Shingobee Lake from March 1990 to June 1992

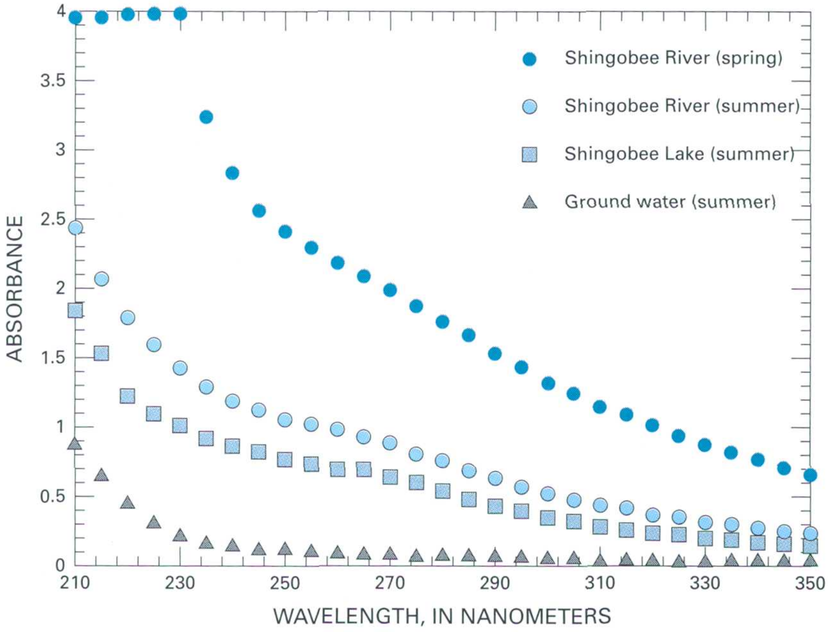

Figure 38. Ultraviolet light absorbance data for water samples from the Shingobee River, Shingobee Lake, and ground water collected near Williams Lake

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Methane Emission from Minnesota and Wisconsin Lakes following Icemelt

By Robert G. Striegl and Catherine M. Michmerhuizen

INTRODUCTION

Organic particles that settle in lakes may be accumulated and stored in lake sediments, disintegrated by weathering and chemical reaction, or fed upon by organisms. The eventual end products of the decomposition of organic matter are carbon dioxide (CO_2) in oxic sediments and CO_2 and methane (CH_4) in anoxic sediments. Lake sediments receiving large amounts of plant and animal debris commonly have anoxic sediments for extended periods of the year, which results in substantial methanogenesis. In such lakes, CH_4 cycling is an important component of the overall lake carbon cycle.

Emission of CH_4 from lakes and wetlands is an important source of atmospheric CH_4, a greenhouse gas that is increasing in concentration in the atmosphere (Raynaud and Chappellaz, 1993; Matthews, 1993). Although lakes are collectively identified as a source of CH_4, the seasonality of CH_4 emission from lakes and the relative importance of morphometric, hydrologic, and trophic factors that potentially affect lake gas emissions are not well known. Because of this, it is difficult to accurately estimate regional CH_4 emissions from areas such as the north temperate lakes region of Minnesota and Wisconsin that have many lakes and wetlands of varying size, setting, and trophic status.

HYPOTHESES TESTED

Because lake CH_4 production is ultimately linked to the amount of organic material that is available for decomposition, we hypothesize that lake CH_4 emission (moles per square meter of lake surface) is proportional to the areally averaged organic carbon loading to the lake-bottom sediments.

Therefore:

1. CH_4 emission should be greater from eutrophic lakes than from mesotrophic and oligotrophic lakes in similar settings,
2. Lakes that have large littoral-zone to total-lake area ratios and extensive macrophyte production should emit more CH_4 to the atmosphere than lakes that do not, and
3. Lakes that receive surface-water organic carbon loads should emit more CH_4 than similar lakes without surface-water input.

The goal was to regionally quantify the potential of lakes in Minnesota and Wisconsin to emit CH_4 to the atmosphere and to relate that potential to commonly observed lake and watershed characteristics. This is a monumental task to accomplish for large numbers of lakes because gas emissions are known to vary greatly by season and are controlled by a variety of factors including temperature, in-lake gas production and consumption rates, episodic emissions caused by winds or by mixing of sediments, and climate. The seasonality of methane emission was tracked at only two lakes, Williams Lake and Shingobee Lake in Minnesota, and potential CH_4 emission was surveyed at an additional 17 lakes and one bog in Minnesota and Wisconsin. The survey was conducted in the spring of 1993.

METHODS

Lake water was pumped from sample depths by peristaltic pump directly into 60-mL polypropylene syringes having three-way nylon stopcocks. Dissolved CH_4 concentrations were determined by equilibrating 25 mL of the lake water with an equal volume of nitrogen and analyzing the head space on a gas chromatograph equipped with a flame ionization detector and a Porapak N column.

Lake-water CH_4 concentrations were determined for the depth of 0.1 m and throughout the water column at the deepest locations in the lakes or lake basins sampled. Total lake storage of dissolved CH_4 (moles of CH_4 per lake) was calculated from the depth relative to concentration profile and from depth relative to lake volume values digitized from lake contour maps.

Time series of CH_4 concentration and storage were tracked in Williams and Shingobee Lakes during 1992 and 1993. Measurements from the spring and winter of 1992 indicated that CH_4 gradually builds up in the lakes under ice cover and is released to the atmosphere following icemelt. The time around icemelt was therefore identified as an ideal time to do a comparative survey of the potential for CH_4 emission from several lakes because all lakes were subject to similar periods of ice cover at nearly identical water temperatures, CH_4 oxidation is slow or negligible in winter, and episodic gas losses from bubbling or wind action are essentially nonexistent when lakes are covered with ice.

Previous experience indicated that the greatest values of springtime CH_4 storage can be measured either just before or just after icemelt, apparently depending on lake mixing conditions during the melt period. Measurements were made at both times and the greater of the two measured storage values were used to calculate potential emission to the atmosphere. Potential emission was calculated as the area-averaged difference between the measured CH_4 storage and CH_4 storage when the lake is at equilibrium with the atmosphere (about 1.75 parts per million by volume).

Nineteen lakes and one bog were selected for the survey (table 7). The lakes ranged over more than four orders of magnitude in size and were located in four separate geographic settings. Settings included carbonaceous glacial sediments in and near the Shingobee River Headwaters Area in Minnesota (SRHA), non carbonate glacial sediments at the North Temperate Lakes Long-Term Ecological Research and Water Energy and Biogeochemical Budgets research site in north-central Wisconsin (LTER), Precambrian Shield and glacial sediments in and near the Boundary Waters Canoe Area of northeastern Minnesota (BWCA), and lakes receiving urban and suburban drainage in the Minneapolis metropolitan area (MMA) (fig. 39).

Figure 39. Location of lakes sampled in Minnesota and Wisconsin.

RESULTS

Methane storage and potential emission have bimodal annual cycles in Williams and Shingobee Lakes (fig. 40). As springtime temperatures rose, CH_4 production increased, and dissolved CH_4 storage increased throughout the summer. Emissions of CH_4 from the lakes to the atmosphere also increased during this period through diffusion, convective mixing, and bubbling. In late summer, CH_4 production decreased as water and sediment temperatures cooled. During early fall, CH_4 emission to the atmosphere exceeded net production and there was a loss of stored CH_4 from the lakes. This loss increased rapidly during and following fall overturn, and the lakes continued to degas until ice formed. Methane emission to the atmosphere ceased when full ice cover was established. This allowed for a buildup of dissolved CH_4 under ice cover that continued until icemelt. Following icemelt and spring overturn, CH_4 was rapidly lost to the atmosphere, followed by a period of more gradual loss until temperatures increased, net CH_4 production exceeded emission, and increases in summertime CH_4 concentration recurred.

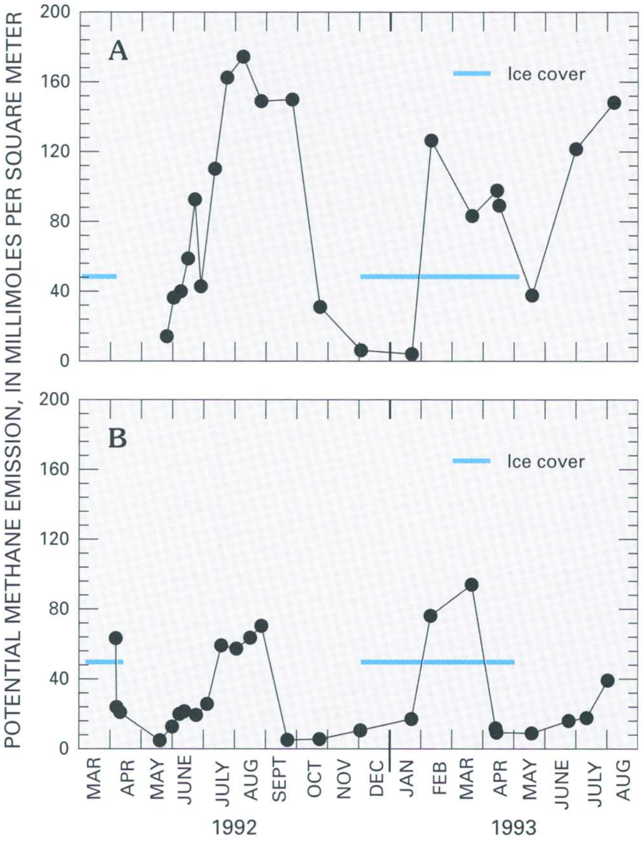

Figure 40. Potential CH_4 emission for (A) Shingobee and (B) Williams Lakes relative to time.

The buildup of CH_4 storage under ice cover was used as a metric to compare potential CH_4 emission from 19 lakes and 1 bog during the spring of 1993. Potential emission ranged from 0.4 mmol/m^2 (millimole CH_4 per square meter) for Trout Lake, the largest lake at the LTER area, to 185.4 mmol/m^2 at Little Shingobee Lake, the smallest lake at the SRHA (table 7). In general, potential emission was inversely proportional to lake area (fig. 41A; $r^2 = 0.33$). This is consistent with the second hypothesis: lakes that have large littoral-zone to total-lake area ratios will have larger CH_4 emissions to the atmosphere than lakes that do not. Among the areas sampled, lakes from the SRHA (Little Shingobee, Williams, Shingobee, Eleventh Crow Wing, and Leech) generally had higher potential emission to the atmosphere and had the most consistent inverse relation between potential emission and total lake area (fig. 41B; $r^2 = 0.987$). Lakes in the SRHA tend to have soft, organic-rich sediments and large macrophyte densities. Inclusion of three additional soft-sediment lakes in the analysis, Hiawatha and Minnetonka (MMA) and Ojibway (BWCA), did not appreciably change the relation (fig. 41C; $r^2 = 0.948$), suggesting that potential emission is strongly linked to the sediment source and that dead macrophytes from littoral zones provide much of the substrate for methanogenesis.

Crystal Bog in the LTER area had appreciably less potential emission than similarly sized lakes having soft sediments (fig. 41A). Reasons for this are not clear but include possible loss of dissolved CH_4 from the bog at some time earlier than the time of sample collection.

Lakes having predominantly sandy or rocky bottoms and small macrophyte densities (Glacier Pond and Tofte, Jasper, and Snowbank Lakes (BWCA), and Crystal, Allequash, Big Muskellunge, and Trout Lakes (LTER) also exhibited a strong inverse relation between potential CH_4 emission and lake area (fig. 41D; $r^2 = 0.636$), but the potentials were smaller than for lakes having soft sediments.

The lines defined in figures 41C and 41D provide a range in potential CH_4 flux following icemelt that could be expected for relatively undisturbed lakes in Minnesota and northern Wisconsin. However, the effects of cultural disturbances on these potentials are uncertain. Nokomis,

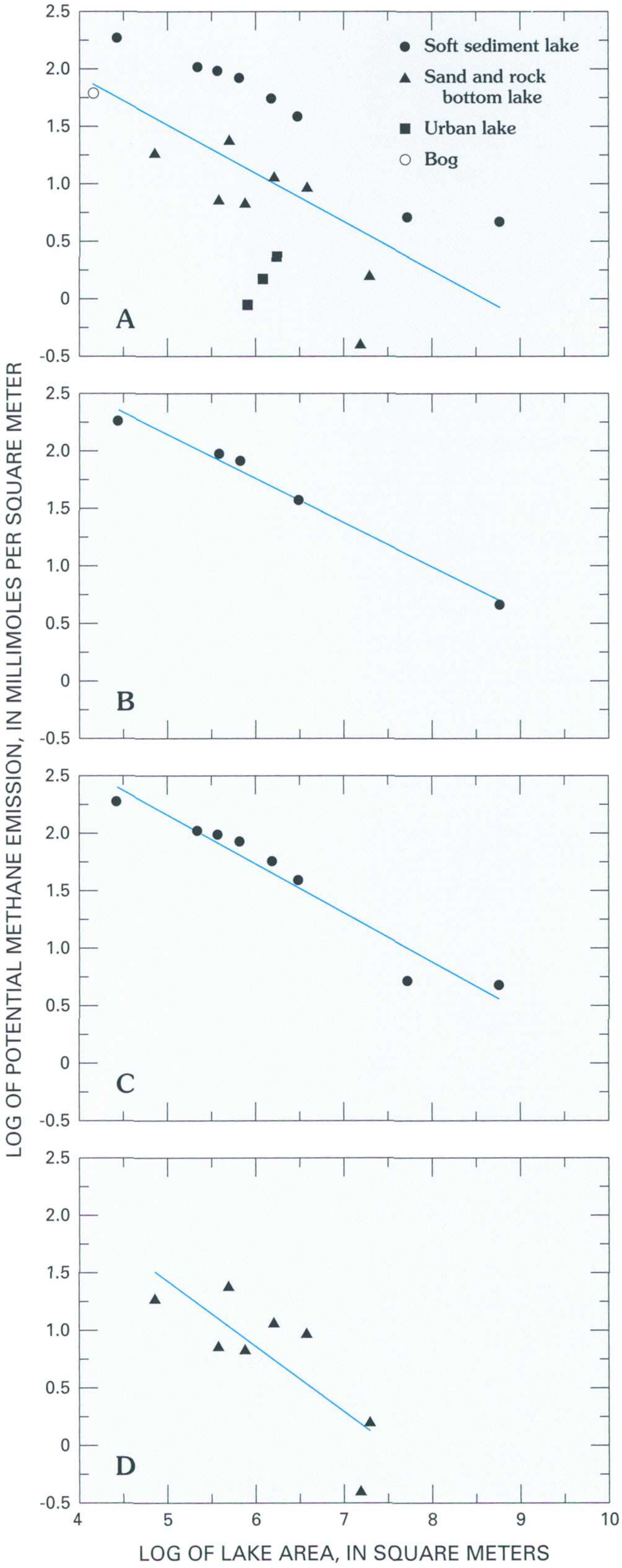

Figure 41. Log of potential CH_4 emission compared to log of lake area for: (A) all lakes, (B) SRHA lakes, (C) eight lakes having soft sediments and high macrophyte densities, and (D) eight lakes having sandy or rocky littoral zones and low macrophyte densities.

Table 7. Lake location, size, and potential methane emission.

[Area, in square meters; methane flux, in millimoles methane per square meter; SRHA, Shingobee River Headwaters Area, Minnesota; LTER, Long-Term Ecological Research / Water Energy and Biogeochemical Budgets research site, Wisconsin; BWCA, Boundary Waters Canoe Area, Minnesota; MMA, Minneapolis Metropolitan Area.]

Lake	Location	Latitude	Longitude	Area (m^2)	Potential methane flux
Little Shingobee	SRHA	46° 59′	94° 41′	27,120	185.4
Williams	SRHA	46° 57′	94° 40′	370,900	94.7
Shingobee	SRHA	47° 00′	94° 41′	655,100	83.2
Eleventh Crow Wing	SRHA	46° 48′	94° 44′	2,997,000	38.3
Leech	SRHA	47°00′	94° 00′	573,400,000	15.6
Crystal Bog	LTER	46° 00′	89° 36′	14,160	61.3
Crystal	LTER	46° 00′	89° 37′	378,900	7.2
Allequash	LTER	46° 02′	89° 37′	1,612,000	11.5
Big Muskellunge	LTER	46° 01′	89° 37′	3,841,000	9.3
Trout	LTER	46° 02′	89° 40′	15,610,000	0.4
Glacier Pond	BWCA	47° 57′	91° 34′	72,300	18.6
Tofte	BWCA	47° 58′	91° 35′	506,100	24.1
Jasper	BWCA	47° 58′	91° 33′	758,400	6.8
Ojibway	BWCA	47° 57′	91° 33′	411,300	54.7
Snowbank	BWCA	48° 00′	91° 25′	20,040,000	1.6
Hiawatha	MMA	44° 55′	93° 14′	216,989	103.6
Nokomis	MMA	44° 54′	93° 14′	805,486	0.88
Harriet	MMA	44° 55′	93° 18′	1,194,762	1.5
Calhoun	MMA	44° 57′	93° 18′	1,720,011	2.3
Minnetonka	MMA	44° 55′	93° 35′	53,010,000	5.3

Harriet, and Calhoun Lakes (MMA) clearly have less potential emission than other lakes of similar size, regardless of bottom type. These lakes have all received urban runoff for decades and have harvesting programs to remove nuisance macrophytes. Removal of macrophytes from lakes must reduce the potential for gas production somewhat, but other factors such as inhibition of methanogenesis by pollutants in urban runoff and historical applications of macrophyte inhibitors such as copper sulfate also need to be considered. Lakes affected by other kinds of cultural disturbances, such as agricultural runoff, were not sampled.

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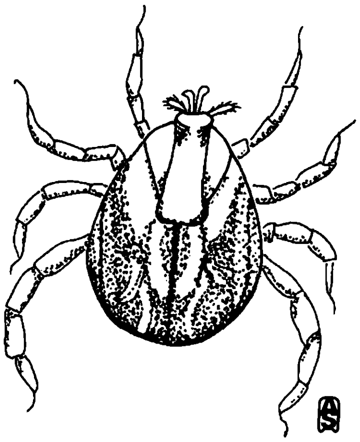

The Aquatic Macrophytes in Williams and Shingobee Lakes--Implications for Carbon Cycling

By Virginia Carter, Nancy B. Rybicki, Robert G. Striegl, and Patricia T. Gammon

INTRODUCTION

One of the significant carbon pools in the Shingobee River Headwaters Area is the biomass of submersed aquatic macrophytes in Williams Lake and Shingobee Lake . During the growing season dissolved carbon dioxide (CO_2) or bicarbonate in the lake water serves as a carbon source for the growth of submersed plants. Floating-leafed and emergent plants in the littoral zone and in the emergent wetland zone surrounding the lakes use atmospheric CO_2 directly. Carbon is transformed into structural and nonstructural carbohydrate in these plants and stored in all parts of the plant: leaves, stems, and rhizomes. This carbon pool is fairly labile; that is, during the growing season there is some turnover in plant tissue, and at the end of the growing season with the onset of colder temperatures, most of the plant material above the sediment-water interface dies. The plant rhizomes, and possibly a small amount of above-sediment plant material, survive over the winter when growth rates are at a minimum. In the lake itself, some of the dead material decomposes in place and reaches the water column in the form of dissolved or particulate organic carbon and dissolved inorganic carbon. The remainder falls to the lake bottom, where it adds to the organic content of the sediments. Sediment organic matter may be stored or further decomposed to dissolved organic carbon, dissolved inorganic carbon, or methane and redistributed to the water column. Water column carbon constituents are exported from the lake by way of ground water, surface water, or exchange with the atmosphere.

Plants in the emergent wetland surrounding the two lakes also die back in winter. Decaying plant material may wash into the lake or decompose in place, adding to the organic soil layer in the wetland. Both Williams and Shingobee Lakes contain a variety of submersed, floating-leafed, and emergent aquatic macrophyte species (Carter and others, 1993) (fig. 42). These plants vary in their ability to utilize dissolved CO_2 and bicarbonate, in the carbon content of their tissues, in the accumulation of other constituents, in their turnover rates, and in their total biomass in each lake.

Figure 42. Floating-leafed zone of Shingobee Lake. This zone contains a variety of submersed, floating-leafed, and emergent aquatic macrophyte species. (Photograph courtesy of Jane Moffett)

HYPOTHESIS, PURPOSE, AND SCOPE

The hypothesis of this study is that aquatic macrophytes are an important part of the lacustrine ecosystem of Williams and Shingobee Lakes and an important component of the carbon cycle in the watersheds of both lakes. The objectives of this research are (1) to document and characterize the aquatic macrophytes of each lake, identifying dominant plants to species where possible, (2) to estimate the in-lake biomass of macrophytes, (3) to determine the concentration of various constituents in plant tissues, and (4) to compare the distribution and species composition of the aquatic macrophytes in the two lakes. The information gathered in the study will be used in development of carbon budgets for the two watersheds and as the basis for analysis of seasonal water-chemistry trends in the two lakes.

METHODS

A survey was conducted to analyze species composition and determine biomass of submersed, floating-leafed, and nonpersistent emergent vegetation in Shingobee and Williams Lakes in August 1991. Five transects in Shingobee Lake and six transects in Williams Lake (plate 1) were placed perpendicular to the shore across the floating-leafed zone to the depth where no appreciable vegetation was found. Samples were collected at depths of 0.5, 1, 2, 3, 4, and 5 m in Shingobee Lake and 0.5, 1, 2, 3, 4, 5, 6, and 7 m in Williams Lake. At each depth three 0.305-m square quadrats were randomly placed on the bottom and all vegetation within the quadrat was collected by divers. Each sample was sorted by species, and submersed plants were soaked in 100 mM acetic acid for 12 to 16 hours to dissolve encrusted calcium carbonate. All samples were air dried for approximately 12 hours (fig. 43) and then dried in a 110°C oven for 12 to 24 hours, after which dry weight in grams was determined. Species were identified according to Fassett (1969) and Fernald (1970). Data obtained from a 0.616-m-interval bathymetric map of Williams Lake and a 1.524-m-interval bathymetric map of Shingobee Lake obtained from the Minnesota Department of Natural Resources were used to calculate the area between metric contours.

Figure 43. Air drying plant samples prior to determination of biomass. (Photograph courtesy of Jane Moffett)

Selected samples of plant material were analyzed by the University of Maryland Cooperative Extension Service. Percent by weight of carbon, hydrogen, nitrogen, phosphorus, potassium, calcium, and magnesium was determined (Carter and others, 1993).

Figure 44. Emergent wetland surrounding Shingobee Lake. This wetland has organic soil and is dominated by grasses and sedges.

RESULTS

Both lakes have a two-part littoral zone containing a floating-leafed zone with submersed, nonpersistent emergent, and floating-leafed species and a submersed zone containing only submersed macrophytes. Each lake is surrounded by a zone of emergent wetland varying in width and species composition (fig. 44). Twenty-five species were identified in the floating-leafed and submersed zones of the two lakes (fig. 45) (Carter and others, 1993). Each lake was dominated by a different suite of species and each lake had six unique species (table 8). Total biomass and total carbon for the two lakes were calculated on a one-time basis without adjustment for below-ground biomass or turnover.

The emergent zone of Shingobee Lake was wider along the south and east shores and narrow along the north and west shores. The soils in this zone were primarily organic. The zone was dominated by

Carex sp. and *Phalaris arundinacea*, with *Typha latifolia* and *Phragmites communis* dominant in some locales. The emergent zone of Williams Lake was narrow and had soils ranging from small cobbles to coarse sand to organic material. The species composition reflected the heterogeneity of the shoreline; different species were dominant in different locales.

Analyses of plant tissues indicated that the mean overall carbon content of the macrophytes in the submersed and floating-leafed zones was 40 percent. Ranges of the nutrient content of plant tissues are listed in table 9; chemical composition varied by species.

Completion of a carbon budget for the Shingobee Lake watersheds will require additional measurements of CO_2 and methane fluxes across wetland and lake surfaces, and estimation of above-ground biomass and net primary productivity in both uplands and wetlands. Measurements or estimates of carbon content of the ground water, the peat underlying the fen and emergent wetlands along the shore of Little Shingobee Lake, and the bottom sediments of both lakes will also be necessary.

Table 8. Summary of vegetation characteristics of the floating-leafed and submersed zones of Shingobee and Williams Lakes

	Shingobee Lake	**Williams Lake**
Total number of species	18	19
Number of submersed species	13	15
Number of floating-leafed species	3	2
Number of nonpersistent emergent species	2	2
Total biomass in metric tons*	77.3	63.5
Carbon in metric tons*	31	25
Dominant species	*Ceratophyllum demersum*	*Megalodonta beckii*
	Chara sp.	*Chara* sp.
	Nymphaea tuberosa *Nuphar variegatum*	*Nymphaea tuberosa* *Nuphar variegatum*
	Scirpus validus	*Potamogeton amplifolius*
Unique species	*Elodea canadensis*	*Heteranthera dubia*
	Lemna trisulca	*Megalodonta beckii*
	Potamogeton pectinatus	*Nitella* sp.
	Sparganium eurycarpum	*Potamogeton gramineus*
	Utricularia vulgaris	*Potamogeton lucens*
	Zizania aquatica	*Sagittaria graminea*

*Does not include below-sediment biomass or turnover estimate.

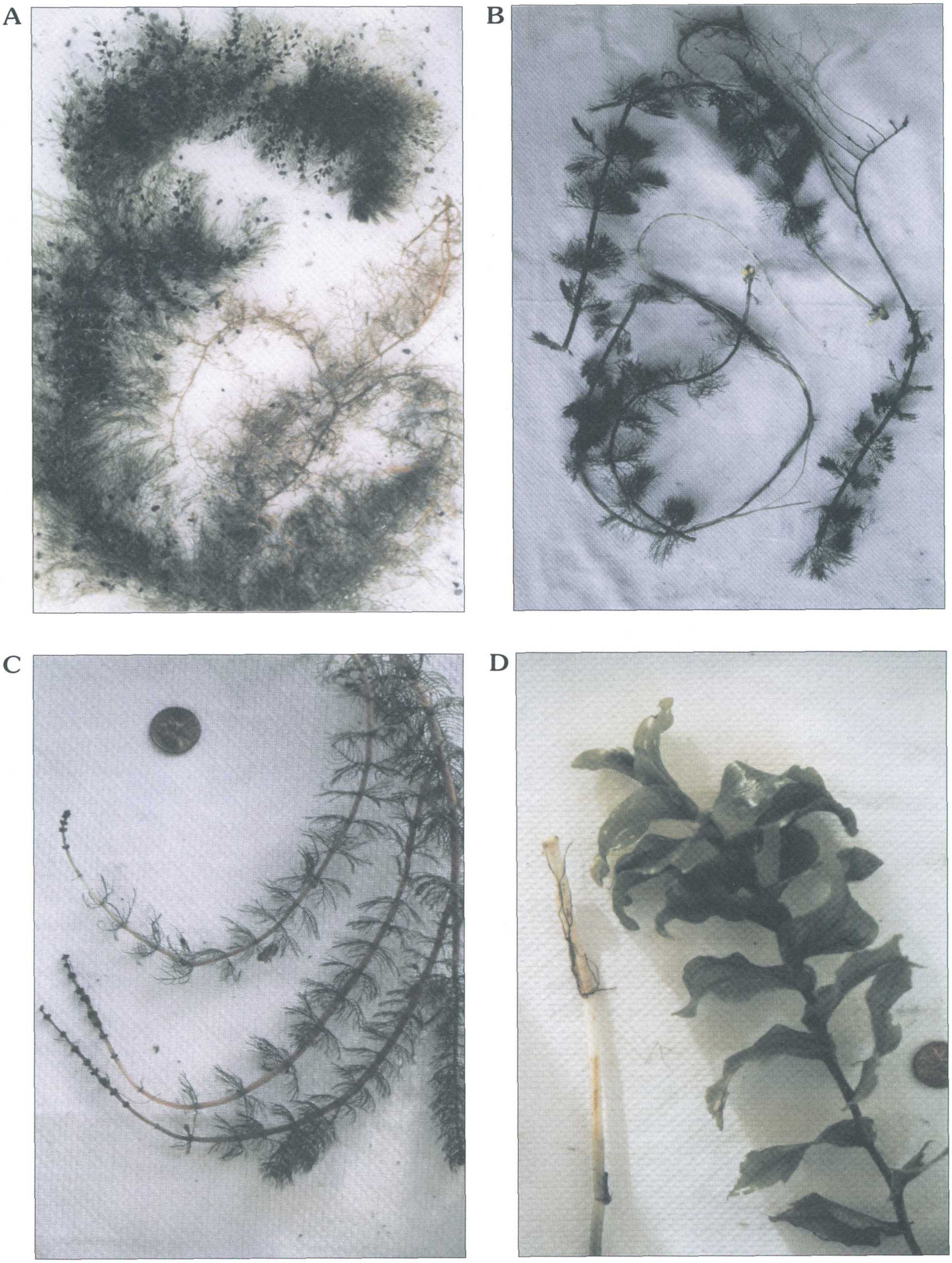

Figure 45. Selected submersed plant species found in Shingobee and Williams Lakes. A. *Utricularia vulgaris*; B. *Megalodonta beckii*; C. *Myriophyllum exalbescens*; D. *Potamogeton richardsonii.* (Photographs courtesy of Jane Moffett)

Table 9. Range of nutrient percentages in plant tissue in Shingobee and Williams Lakes

Nutrients	Shingobee Lake	Williams Lake
Carbon	35.3 - 43.0	32.5 - 46.0
Nitrogen	1.1 - 2.9	1.3 - 2.9
Phosphorus	0.1 - 0.2	0.1
Potassium	0.1 - 3.1	0.1 - 0.9
Calcium	0.2 - 3.3	1.5 - 11.4
Magnesium	0.1 - 0.8	0.1 - 0.4

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The Role of Calcification in Macrophyte Photosynthesis in Williams Lake

By Ted A. McConnaughey, James W. LaBaugh, Paul F. Schuster, Virginia Carter, Robert G. Striegl, Michael M. Reddy, and Donald O. Rosenberry

INTRODUCTION

Many of the submersed aquatic macrophytes in Williams Lake become encrusted with marl (calcium carbonate) during the summer. Alkalinity and calcium concentrations meanwhile decrease by about 15 and 25 percent respectively, after dilution effects and replenishment from ground water are taken into consideration. These observations prompted an examination into the causes of marl precipitation and its roles in plant growth and lake carbon cycling.

A whole-lake carbon budget constitutes the foundation for this analysis. Boundary fluxes include ground-water inputs and outputs and carbon dioxide exchange with the atmosphere. Calcification and calcium carbonate dissolution are estimated from changes in lake storage of calcium (or alkalinity), plus additions from ground-water. Net photosynthesis is estimated from changes in inorganic carbon storage, plus boundary fluxes, minus calcification. The rates and mechanisms of calcification are then addressed through carbonate saturometry for calcification within the water column, and plant incubation experiments for biological calcification. A more comprehensive account of this work has been published (McConnaughey and others, 1994).

Hydrology, geology, plant physiology, ecology, and more than a decade of research all contribute to the analysis. Without the interdisciplinary, cooperative attributes of the Interdisciplinary Research Initiative effort, this analysis would not have been possible.

METHODS

Water samples were obtained from depths of 1 m and 8 m, and temperature, specific conductance, pH, and dissolved oxygen were measured in situ at 1-m depth intervals. Sampling was conducted approximately biweekly during the ice-free part of the year and monthly when the lake was ice covered. Data from 11 years (1980-1990) were composited to produce annual chemical cycles at 1 m and 8 m, and chemistry at other depths was interpolated, based on specific conductance. From these data we calculated water-column chemical inventories.

Two versions of the stagnant film model with and without chemical enhancement of CO_2 diffusion through the surface film were used to calculate carbon dioxide fluxes across the lake surface. Stagnant film thickness was calculated from regressions against windspeed, using wind data collected over Williams Lake.

Ground-water chemical fluxes were estimated from ground-water chemical compositions measured upgradient from the lake, and ground-water flow rates were calculated from hydraulic gradients and hydraulic conductivities. Isotopic methods were also used to estimate ground-water contributions over limited periods (McConnaughey and others, 1994; LaBaugh and others (1995).

Macrophyte biomass was estimated by transect sampling (Carter and others, this volume), and rates of macrophyte calcification and photosynthesis were calculated from changes in pH and alkalinity measured during incubations in shallow water. Authigenic calcification was monitored by repeated weighings of calcium carbonate crystals and blocks suspended within dialysis bags at various points in the lake.

RESULTS

Williams Lake contains a diluted version of upgradient ground-waters. Magnesium and sodium pass through the lake with relatively little chemical reaction; hence, their concentrations in lake waters fall along a mixing line connecting upgradient groundwaters with meteoric waters (fig. 46). Seasonal and interannual chemical variability is evident in both shallow regional ground waters and lake waters. Seasonal variations within the lake can be largely attributed to the springtime melting of ion-depleted snow and ice. All major dissolved chemicals display this springtime dilution (fig. 47). Reactive elements such as carbon, calcium, and potassium exhibit further depletion during summer, which is caused by plant growth and calcification. These nonconservative attributes are more evident when dilution effects are removed by normalizing ion concentrations to a constant magnesium concentration.

Dilution-adjusted calcium and alkalinity concentrations increase from autumn until early spring, then decline rather abruptly during summer (fig. 48B). Calcium and alkalinity display nearly equal rates of wintertime accumulation and summertime depletion, consistent with control by calcium carbonate precipitation and dissolution. When ground-water replenishment is factored in, it appears that the lake loses about 15 percent of its alkalinity and 25 percent of its calcium between late May and August. Calcification rates, calculated from changes in calcium and alkalinity, peak at about 700 micromoles per liter per year from about late June to late July (fig. 48C).

The epilimnion is mildly supersaturated with respect to calcite during summer (fig. 48A); hence, inorganic calcification is possible. However, calcite crystals and blocks suspended in the epilimnion exhibited little or negative changes in weight, aragonite crystals lost weight, and all carbonates suspended in the hypolimnion lost weight. These carbonate saturometry experiments indicated that little authigenic calcification occurs within the water column.

Plant calcification is an obvious alternative, since several submersed macrophytes accumulate visible carbonate encrustations during summer. When incubated in jars containing calcium-supplemented lake water, several species exhibited nearly equal rates of calcification and photosynthesis (fig. 49), as is characteristic of calcify as a precursor to bicarbonate

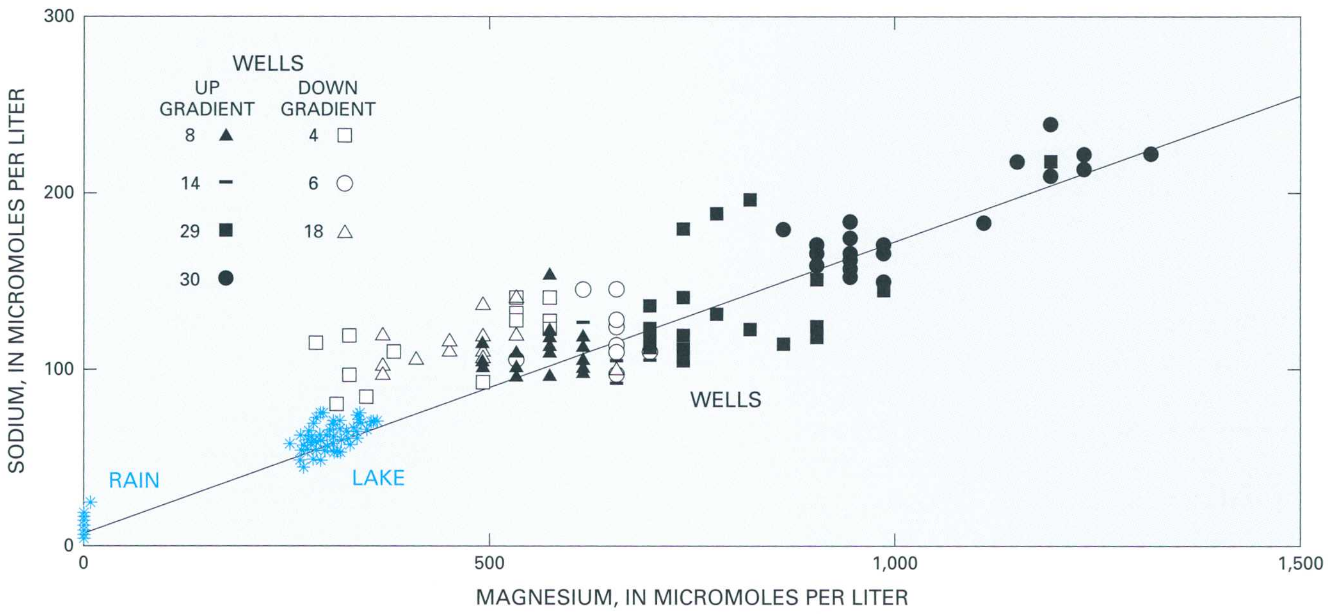

Figure 46. Magnesium and sodium concentrations in Williams Lake, nearby wells, and rain.

assimilation (McConnaughey, 1994). Also shown in figure 49 are contours of pH, carbon dioxide concentrations, and calcite saturation for lake waters subjected to various amounts of calcification and photosynthesis. Large amounts of carbon can be removed from the water, with minimal effect on pH, CO_2 concentrations, or calcite saturation state, under a 1:1 stoichiometry of calcification to photosynthesis.

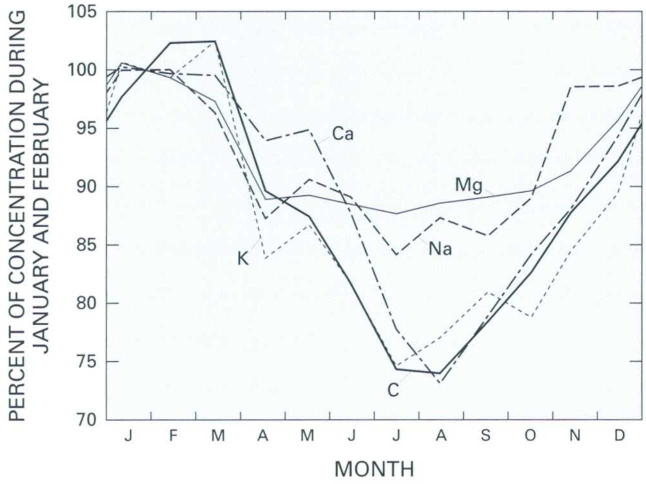

Figure 47. Seasonal cycles of Mg^{2+}, Na^+, K^+, Ca^{2+}, and dissolved inorganic carbon (C) in Williams Lake, expressed as a percentage of average concentrations during January and February.

The physiology linking calcification to photosynthesis in these plants is rather interesting (McConnaughey and Falk, 1991; McConnaughey, 1991). Calcareous plants typically develop spatially alternating alkaline and acidic surfaces, and the alkaline regions become calcified. The giant cells of characean algae commonly display alternating calcified and noncalcified bands, while the submersed leaves of *Potamogeton* become calcified on their upper surfaces but not on their lower surfaces. The calcified surfaces can exhibit pH values as high as 10.5. Carbon dioxide leaks from the plant into these alkaline regions and rapidly precipitates as calcium carbonate. The plant translocates the protons generated during calcification to its acidic surfaces and ejects them into the water, where they convert bicarbonate to carbon dioxide, which the plant absorbs. In mildly alkaline waters, this process nets the plant one carbon dioxide molecule, for photosynthesis, for each carbon dioxide molecule precipitated as calcium carbonate.

Alkaline zones: $Ca^{2+} + CO_2 + H_2O = CaCO_3 + 2H^+$
Acidic zones: $2H^+ + 2HCO_3^- = 2CO_2 + 2H_2O$
Net: $Ca^{2+} + 2HCO_3^- = CaCO_3 + CO_2 + H_2O$

In this sequence, calcification generates the protons needed to convert bicarbonate to carbon dioxide, and functionally precedes photosynthesis.

In the lake carbon budget, net photosynthesis can be calculated from the rate of carbon removal, after allowances have been made for boundary fluxes and calcification. Boundary fluxes include imports and exports through ground-water flow, recycling within the lake-sediment system, and carbon dioxide exchanges with the atmosphere.

Treating magnesium as a non-reactive constituent in the system, the net ground-water influx of carbon to the lake can be estimated as the difference between carbon to magnesium ratios in upgradient ground waters and lake waters, times the magnesium flux through the system. To a first approximation this term is constant throughout the year. Sediment fluxes are estimated as the difference between this number and the total rate of carbon accumulation in the lake. The latter is calculated from the rate of change of water-column carbon inventories during the winter, when biological fluxes are relatively small. (See dashed lines on fig. 48B for the analogous accumulation of calcium and alkalinity in the water column during winter.)

Carbon dioxide exchanges with the atmosphere reverse seasonally (fig. 50). Results from two versions of the stagnant film model are illustrated in figure 50: a standard Fickian version, and a chemically enhanced version, which considers the rate of carbon dioxide reaction within the surface boundary layer. Large losses to the atmosphere during spring and fall bracket a summertime period of carbon uptake by the lake. Carbon dioxide reactions within the surface microlayer significantly enhance summertime carbon dioxide uptake. Whole-lake net photosynthesis (fig. 51) peaks at around 8 moles per square meter per year in early summer.

When plants calcify in order to generate protons for the photosynthetic assimilation of bicarbonate, they typically produce approximately equal molar quantities of calcium carbonate and organic carbon. Summertime photosynthesis can therefore be divided into calcification dependent and independent fractions, which utilize bicarbonate and carbon dioxide, respectively based on rates of calcification. Calcification-dependent photosynthesis accounts for about 40 percent of whole-lake net photosynthesis during the growing season, and more than 50 percent during the midsummer period of maximum plant growth (fig. 51). Calcification-dependent photosynthesis, which uses bicarbonate as its carbon source, occurs mainly during the summertime period of high pH and depressed carbon dioxide partial pressures. It represents an important adaptation to aquatic carbon dioxide depletion.

Ground-water inseepage, augmented considerably by dissolution of sedimentary marls (McConnaughey and others, 1994), supplies the calcium (and alkalinity) needed to support calcification-dependent bicarbonate-based summertime net photosynthesis. From 1980 to 1991 ground water supplied between 654 and 837 thousands of moles per year of calcium and 1,613 and 2,228 thousands of equivalents per year of alkalinity (LaBaugh and others, 1995). This input from ground water represented nearly 100 percent of the annual input of calcium and alkalinity to the lake. Without this seepage input, a significant component of lake productivity would rapidly cease to function.

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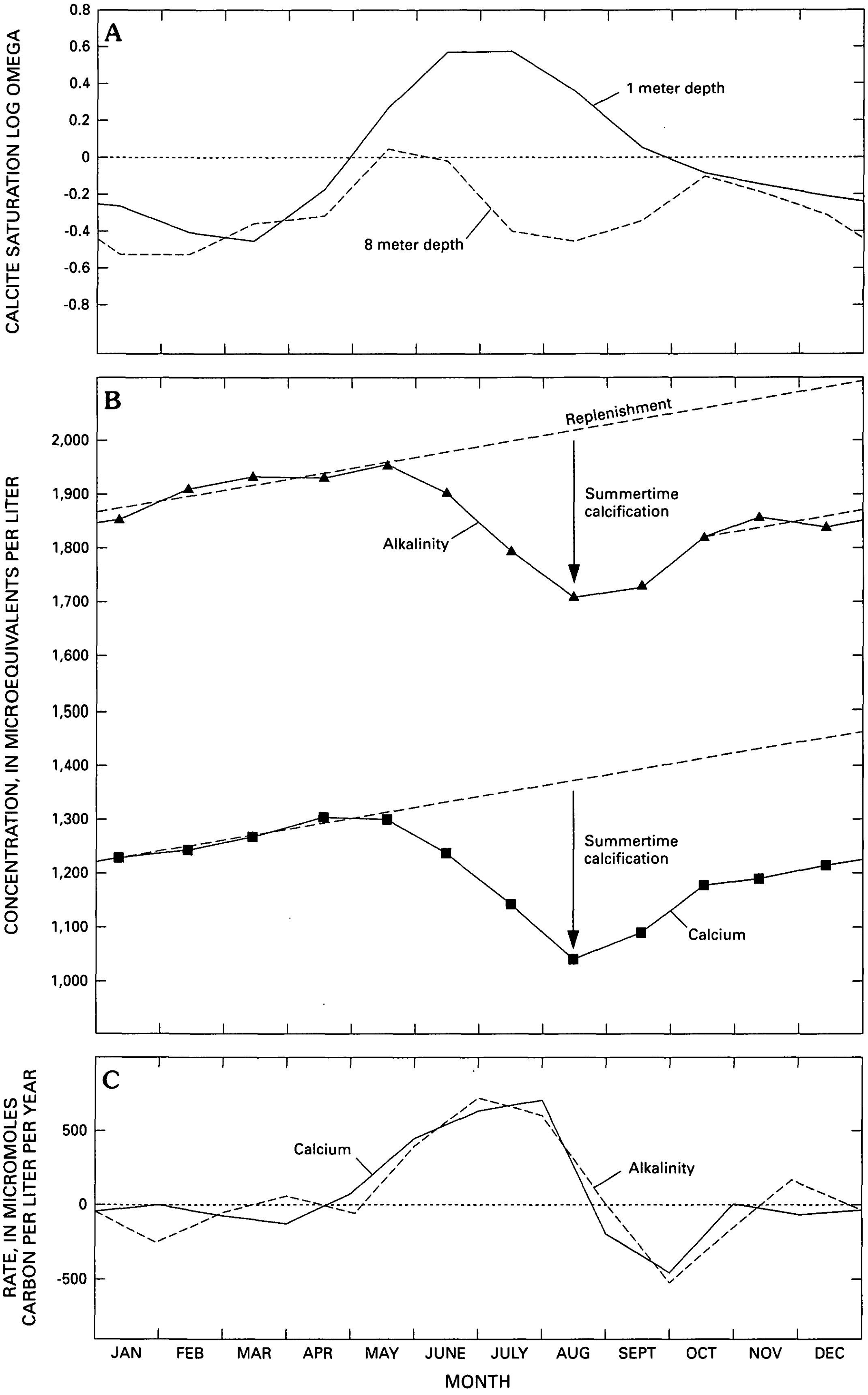

Figure 48. Annual dynamics of calcium carbonate in Williams Lake: (A) Calcite saturation, (B) Alkalinity and calcium concentrations, adjusted for dilution, and (C) Calcification rates calculated from the rates of change of calcium and alkalinity.

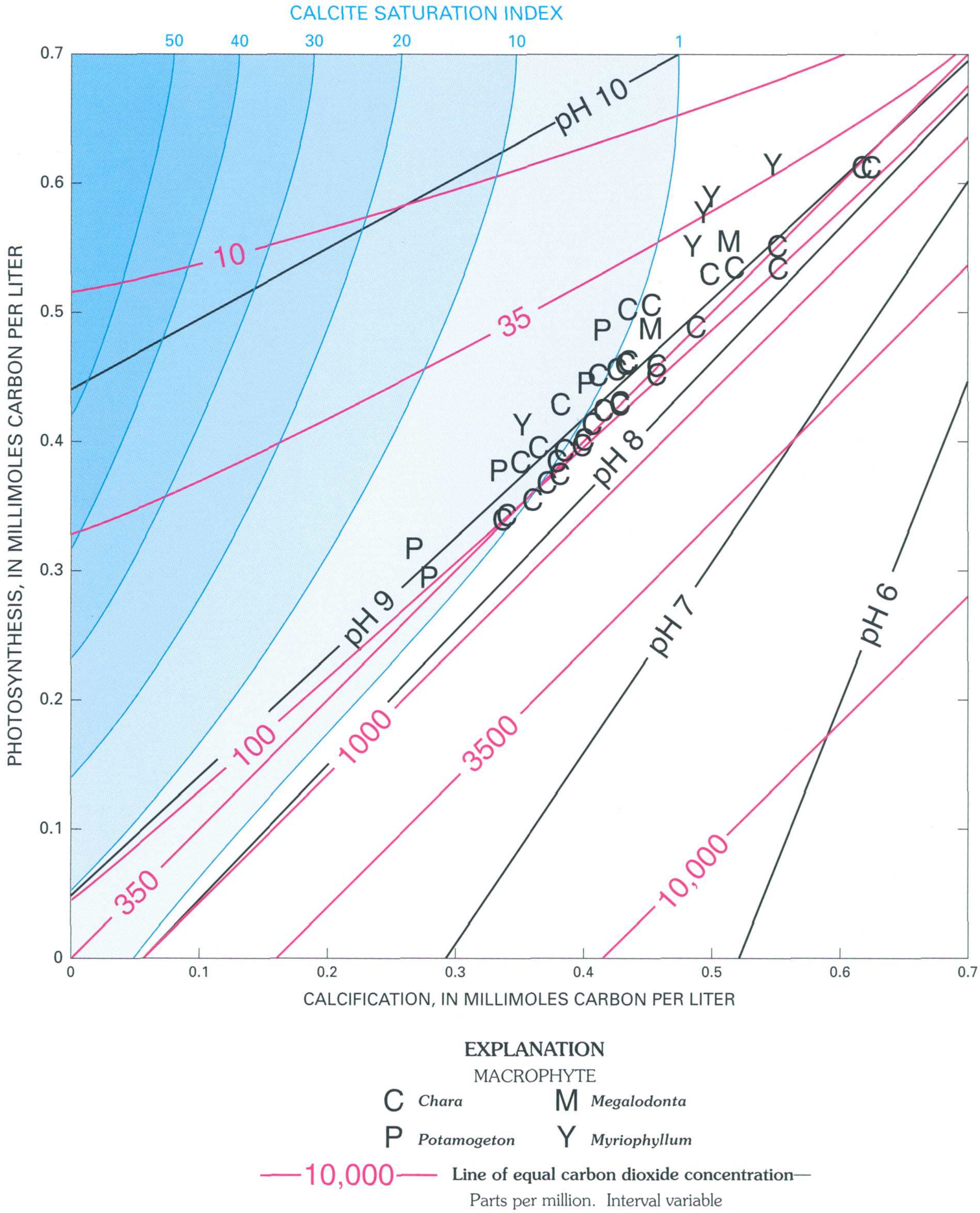

Figure 49. Rates of calcification and photosynthesis by several aquatic macrophytes incubated in calcium-enriched lake water. Also shown are contours of pH, carbon dioxide concentrations, and calcite saturation (shaded) for lake waters subjected to various amounts of calcification and photosynthesis.

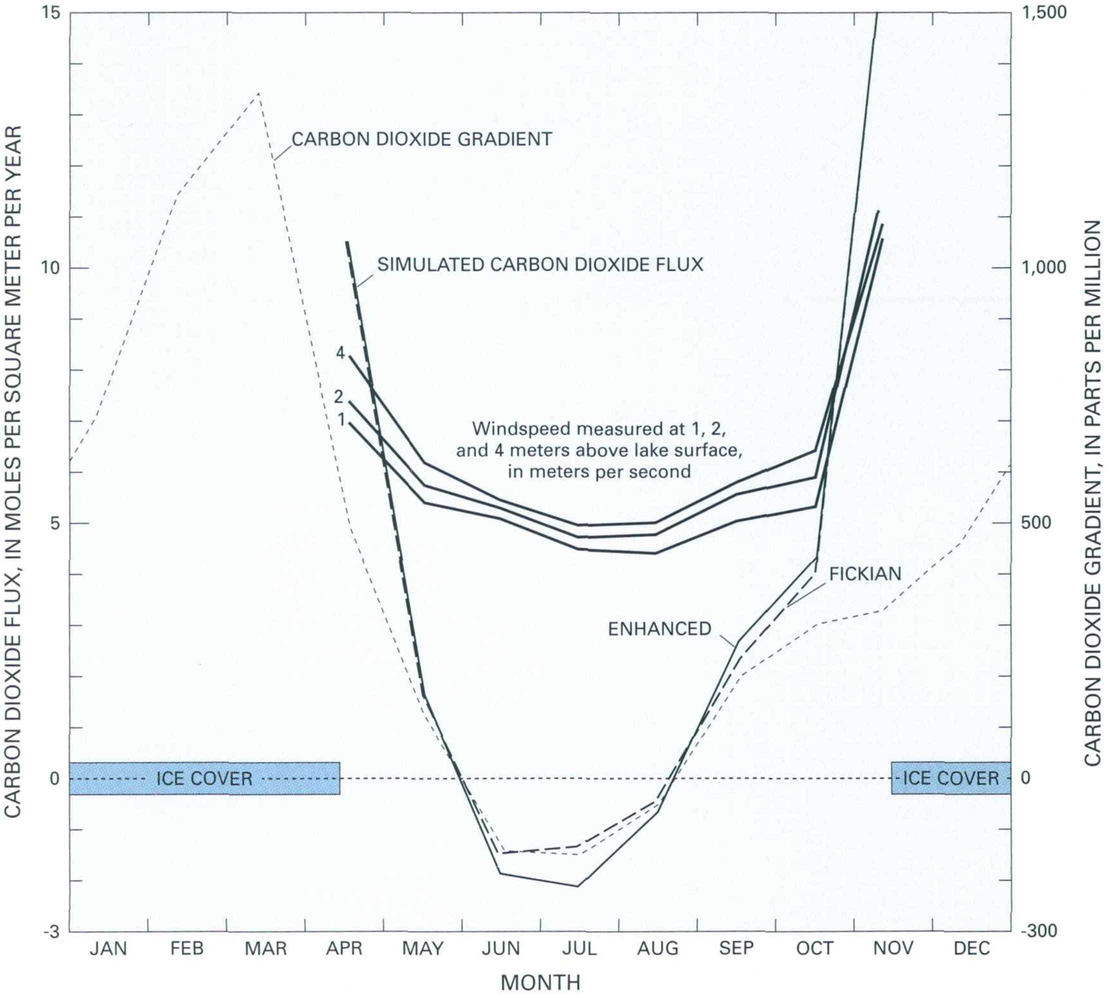

Figure 50. Carbon dioxide exchange between the lake and atmosphere calculated from the gradient in CO_2 partial pressure and windspeed. Results from two versions of the stagnant film model are illustrated: a standard Fickian version, and a chemically enhanced version, which considers the rate of carbon dioxide reaction within the surface boundary layer.

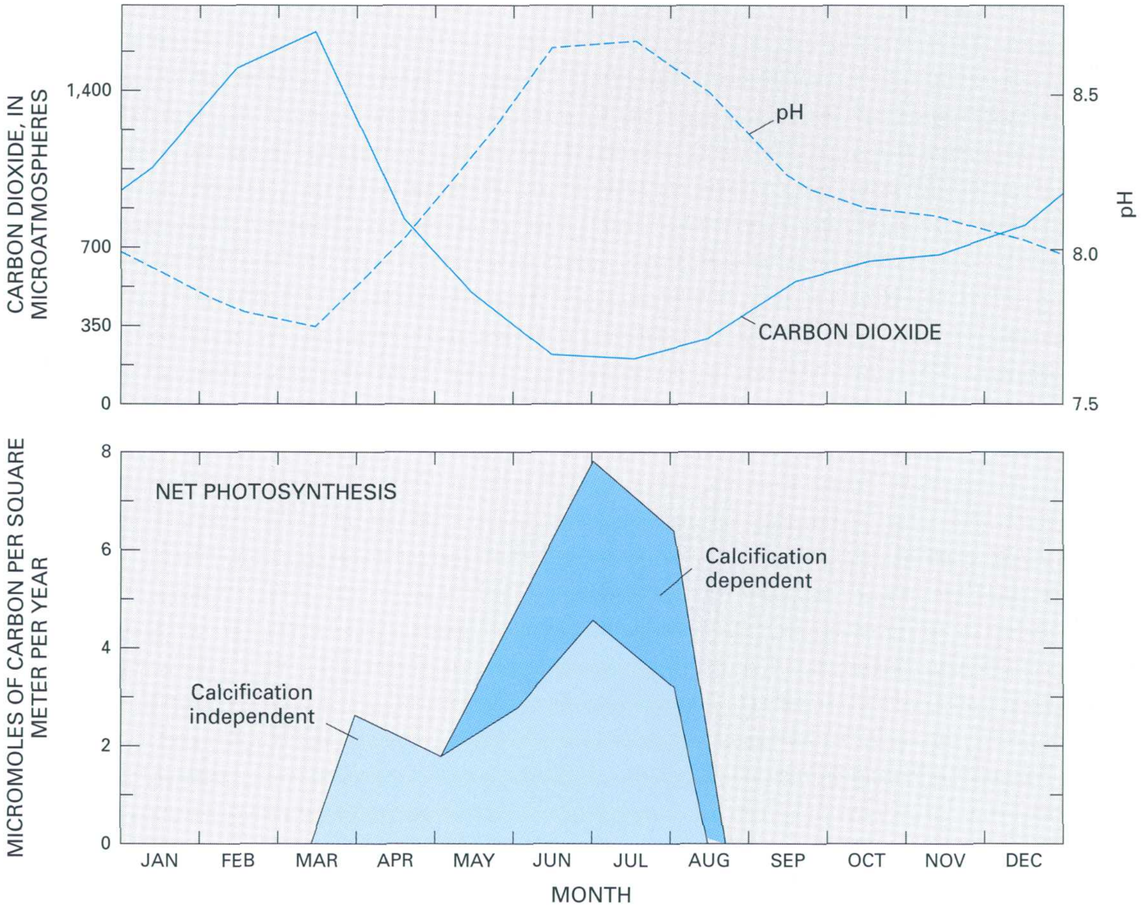

Figure 51. Seasonal cycle of photosynthesis in Williams Lake, calculated using the carbon budget approach and its relation to seasonal variation in CO_2 and pH in the epilimnion.

The Plankton Communities of Shingobee and Williams Lakes

By James W. LaBaugh

INTRODUCTION

Differences in the nutrient content of Shingobee and Williams Lakes (LaBaugh, this volume) reflect differences in nutrient supply and water residence times. The median concentration of total phosphorus in Shingobee Lake is double that of Williams Lake. Differences in nutrient supply control differences between the lakes in the biomass and structure of the plankton communities. Collection of plankton was done at each lake coincidently with collection of water for chemical analysis, and the resulting data provided the opportunity to test the following hypotheses: (1) differences in algal biomass between the lakes are in proportion to differences in nutrient content, (2) differences in zooplankton abundance between the lakes are in proportion to differences in nutrient content, and (3) plankton community structure of the two lakes is similar.

The purpose of these studies is to determine the species composition and temporal variance in algal and microinvertebrate communities found in the pelagic (open-water) area of the lakes in order to evaluate the effect of the different hydrological characteristics of the two lakes on plankton community structure in the lakes. Also, the effect of changes in hydrological fluxes for each lake on the community structure of the plankton is being evaluated. Data presented here are for the period of common record for both lakes, 1989 to 1992.

METHODS

When the lakes were not ice covered, herein referred to as the open-water period, samples were collected at approximately 2-week intervals. When the lakes were ice covered, samples were collected at approximately monthly intervals. Water samples were collected midlake in the vicinity of the deepest part of the lake. Water collected at the surface and at 1-meter intervals through the photic zone was combined for subsequent analysis of chlorophyll-*a* and phytoplankton. The photic zone was defined as twice the depth measured for Secchi disk transparency. For chlorophyll-*a* analysis, a known volume of water was pumped through a Gelman A-E glass fiber filter, and the filter was placed in a glass vial and shipped on ice to the U.S. Geological Survey National Water Quality Laboratory in Arvada, Colorado. Concentration of chlorophyll-*a* was determined at the laboratory by the liquid chromatographic and fluorometric technique described by Britton and Greeson (1988). For phytoplankton analysis, a 1-liter bottle was filled in the field with sample water and preserved with 10 milliliters of Lugol's solution. The contents of this 1-liter bottle were used for species identification, enumeration, and biovolume determination by standard inverted microscopic techniques. The analyses were done by Richard Dufford of Fort Collins, Colorado.

Microinvertebrates found in the open-water area of midlake, also known as zooplankton, were collected at the same time water was collected for chlorophyll-*a* and phytoplankton analyses. An 80-micron mesh Wisconsin plankton net was used to capture zooplankton by hauling the net vertically from 1 meter above lake bottom to the surface. Two hauls of the net were done each time zooplankton were collected, and the contents of each haul were emptied into separate 250-milliliter bottles, narcotized with soda water, and preserved with formalin. Preserved samples were sent to Richard Dufford in Fort Collins, Colorado, for subsequent species identification and enumeration.

RESULTS

Comparison of data for the two lakes indicated that larger values of chlorophyll-*a* were measured in water collected from Shingobee Lake (fig. 52). For the period 1989 to 1992, the median concentrations were 6.2 milligrams per cubic meter in Shingobee Lake and 2.9 milligrams per cubic meter in Williams Lake. Algal biovolume data were available for just two complete years, 1989 and 1990. Median values for algal biovolume in 1989 and 1990 were 5.8 million cubic micrometers per milliliter in Shingobee Lake and 1.9 million cubic micrometers per milliliter in Williams Lake (fig. 53). Based on either measure of algal biomass, median algal biomass in Shingobee Lake was two times larger than in Williams Lake. The difference in the median values of algal biomass between lakes was similar to the twofold difference in median values of total phosphorus (LaBaugh, this volume). Assuming phosphorus limits algal biomass in these lakes, total phosphorus data support the hypothesis that differences in algal biomass between the lakes are in proportion to differences in nutrient concentrations in lake water. There was little difference between Shingobee and Williams Lakes in either total nitrogen or nitrate plus nitrite concentrations. Ammonia nitrogen values usually were larger in Shingobee Lake than in Williams Lake, although concentrations in both lakes declined to near detection limit for most of the spring and summer, when the lakes were thermally stratified.

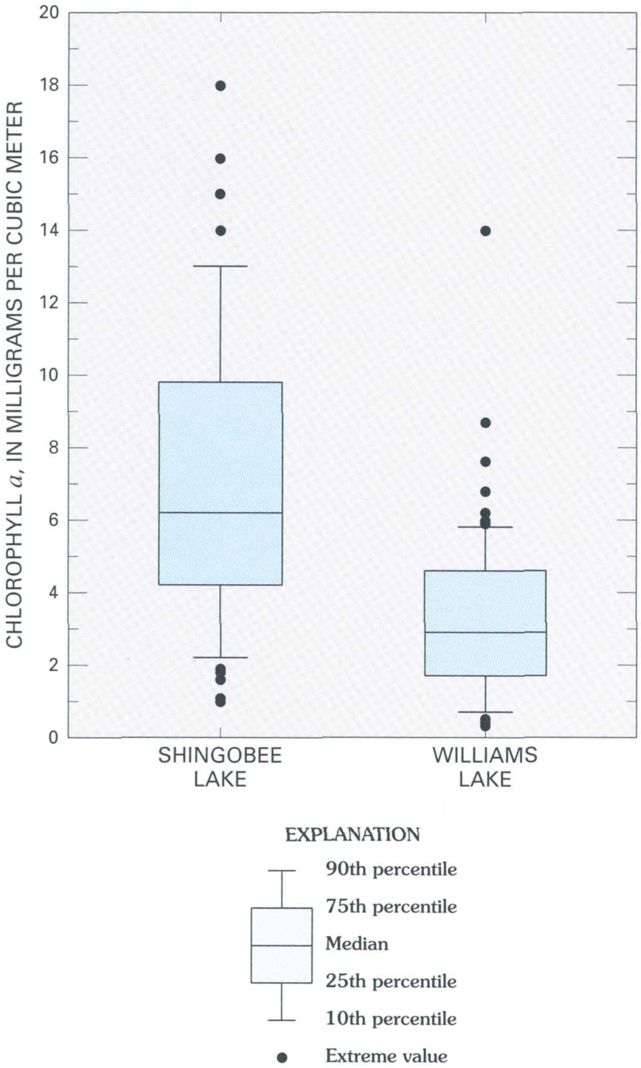

Figure 52. Chlorophyll-*a* data from Shingobee and Williams Lakes, 1989 to 1992.

The algal communities represented by phytoplankton in both lakes were diverse, including nearly 150 species in Shingobee Lake and nearly 200 species in Williams Lake. Ninety species were found in both lakes. Communities in both lakes included representatives from the diatoms, green algae, chrysophytes, cryptophytes, euglenoid and dinoflagellate algae, as well as cyanobacteria, also known as blue-green algae. However, neither the relative contribution to total biovolume nor the seasonal pattern in relative contribution to total biovolume of these taxa were the same for both lakes. The taxon commonly in greatest abundance in Shingobee Lake was cyanobacteria which represented more than 30 percent of the algal biovolume of the entire phytoplankton community in 75 percent of samples collected from the lake (fig. 54). In Williams Lake, dinoflagellate algae commonly were the most abundant taxon, representing more than 20 percent of algal biovolume in the lake in the majority of samples collected from the lake. Green algae and euglenoid algae represented the smallest fraction of algal biovolume in both lakes. Cyanobacteria in Shingobee Lake commonly represented more than 10 percent of the algal biovolume throughout the year, whereas in Williams Lake, cyanobacteria were a large percentage of algal biovolume only in summer (fig. 55). Chrysophytes in Shingobee Lake contributed more to algal biovolume in late spring and early summer than in other seasons. Chrysophytes in Williams Lake commonly contributed more to algal biovolume early to midspring and in the fall than at other times of year, but in some years, such as 1990, they also were a large percentage of algal biovolume in August (fig. 55). The main contribution to algal biovolume by diatoms in Williams Lake occurred in spring and fall, whereas in Shingobee Lake diatoms were more abundant in June. Dinoflagellates were abundant in Williams Lake in winter and in Shingobee Lake in summer and fall. Cryptophyte algae were most abundant in spring and fall in Shingobee Lake and at various times of the year in Williams Lake.

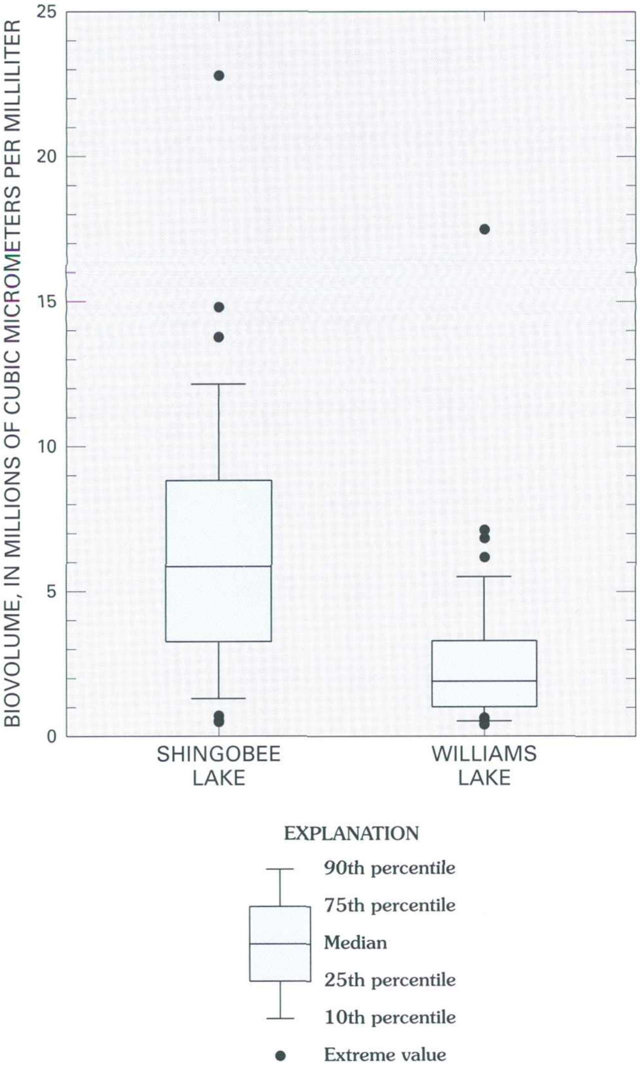

Figure 53. Algal biovolume data from Shingobee and Williams Lakes, 1989 and 1990.

The composition of zooplankton communities in the two lakes was quite similar; most of the crustacean species (table 10) and rotifer species (table 11) were found in both lakes. A variety of crustacean zooplankton were present in each lake. *Bosmina longirostris, Daphnia galeatae mendotae*, and *Chydorus sphaericus* were the most common members of the Cladocera, and *Skistodiaptomus oregonensis, Diacyclops thomasi*, and *Mesocyclops edax* were the most common members of the Copepoda in the lakes. The method of capture provided for semiquantitative information on zooplankton abundance. The definitive test of the hypothesis -- that the difference in zooplankton abundance between the lakes was proportional to differences in nutrient content -- is not possible because zooplankton biomass was not a characteristic measured as part of this study. However, an approximate test of this hypothesis is provided by available data on the number of organisms (fig. 56). The median number of individuals per liter was 271 in Shingobee Lake and 169 in Williams Lake. The median from Shingobee Lake was 1.6 times that from Williams Lake. The larger values from Shingobee Lake may support the hypothesis, given the limitations of the data.

Although the lakes had similar zooplankton communities, there were differences in the composition of the phytoplankton communities that only further study will clarify. The similarity in zooplankton species between the lakes might support the hypothesis that the plankton community structure of the two lakes is similar, but only analysis of the size distribution of both zooplankton and phytoplankton communities can truly test this hypothesis. Further analysis of the food habits and structure of the fish communities as determinants of the size distribution of the phytoplankton and zooplankton in the two lakes may aid in understanding the role of nutrients and consumers in regulating the plankton.

Examples of pertinent questions about the phytoplankton in the two lakes include: (1) Why do cyanobacteria represent so much of the phytoplankton in Shingobee Lake? Little is known about processes controlling nitrogen cycling in these lakes, and such information could be useful in understanding the role nitrogen plays in determining the composition of the phytoplankton community, particularly with respect to the contribution by cyanobacteria. Although the difference in algal biomass between the lakes was proportional to the difference in total phosphorus concentration, experiments need to be done to test the hypothesis that phosphorus limits algal biomass in the lakes. (2) Why do the lakes differ in the timing of increased abundance of diatoms and chrysophytes? It is possible that these differences are related to differences in phosphorus and silica inputs to the lakes. These two groups of algae are used as indicators of past limnological conditions based on their remains left in lake sediments. Differences in hydrological and chemical fluxes to the lakes may provide clues to the mechanisms that bring about changes in the contribution of diatoms and chrysophytes to the phytoplankton community. Knowledge about current conditions affecting the composition of these taxa could improve understanding of how changes in climate translate to changes in hydrological processes that are recorded in the lake sediment by diatoms and chrysophytes.

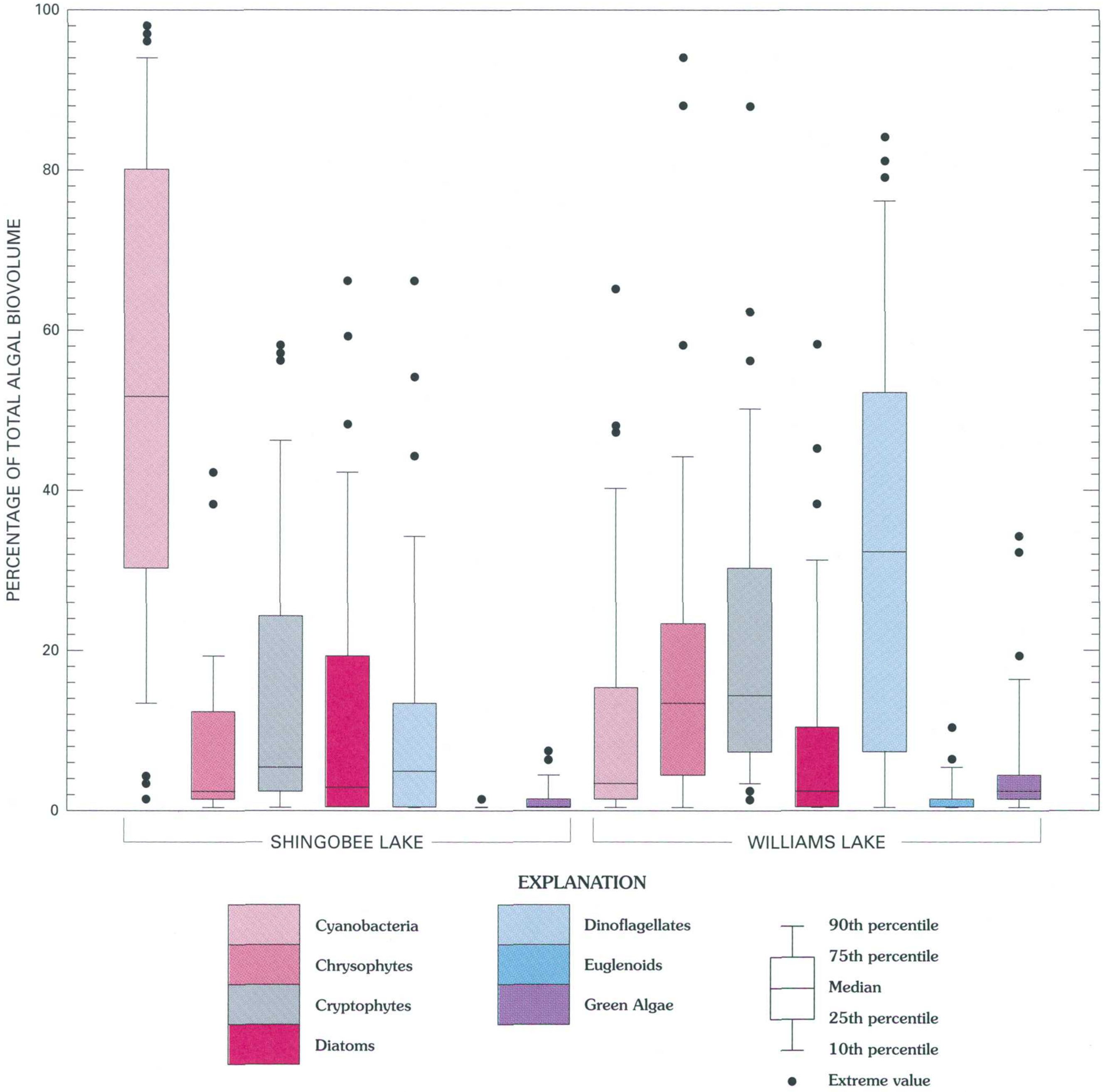

Figure 54. Algal biovolume data for individual taxa, Shingobee and Williams Lakes, 1989 and 1990.

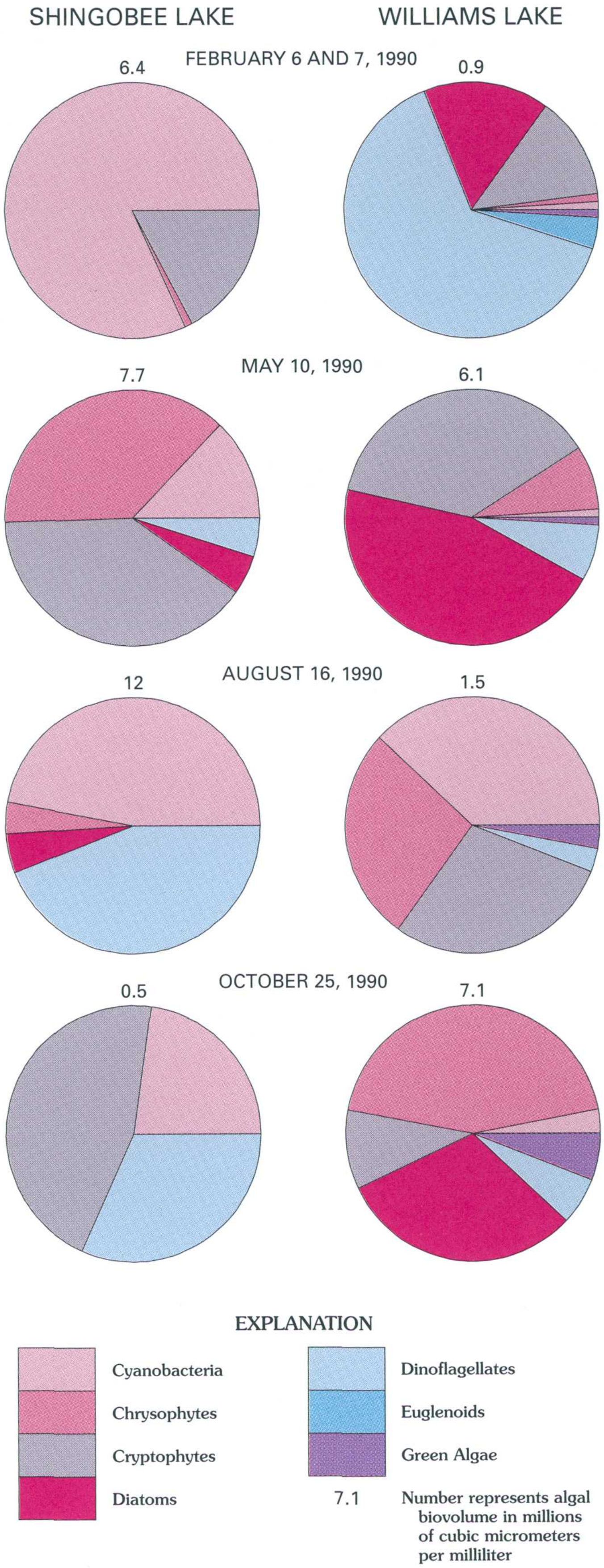

Figure 55. Relative composition of the algal communities in Shingobee and Williams Lakes based on algal biovolume for selected dates representing winter, spring, summer, and fall 1990.

Table 10. Species of crustacean zooplankton found in Shingobee and Williams Lakes, 1989 to 1991.

Common to both lakes	Only in Shingobee Lake	Only in Williams Lake
Cladocera		
Acroperus harpae	*Pleuroxus denticulatis*	*Camptocercus rectirostris*
Alona species	*Sida crystallina*	*Drepanothrix dentata*
Bosmina longirostris		*Eurycercus lamellatus*
Ceriodaphnia lacustris		*Leydigia leydigi*
Chydorus sphaericus		*Ophryoxus gracilis*
Daphnia ambigua		*Pleuroxus* species
Daphnia catawba		
Daphnia galeatae mendotae		
Daphnia parvula		
Daphnia pulex		
Daphnia retrocurva		
Diaphanosoma birgei		
Leptodora kindti		
Simocephalus serrulatus		
Copepoda - Calanoida		
Skistodiaptomus oregonensis		*Epishura lacustris*
Copepoda - Cyclopoida		
Acanthocyclops vernalis		
Diacyclops thomasi		
Ergasilus chatauquaensis		
Eucyclops speratus		
Macrocyclops albidus		
Mesocyclops edax		
Orthocyclops modestus		
Tropocyclops prasinus		

Table 11. Species of Rotifera zooplankton found in Shingobee and Williams Lakes, 1989 to 1991.

Common to both lakes	Only in Shingobee Lake	Only in Williams Lake
Ascomorpha ovalis	*Brachionus calyciflorus*	*Brachionus quadridentatus*
Ascomorpha sultans	*Conochiloides* species	*Dicranophorus* species
Ascomorpha species	*Euchlanis dilatata*	*Euchlanis incisa*
Asplanchna girodi	*Paracolurella* species	*Keratella serrulata*
Asplanchna priodenta	*Trichotria pocillium*	*Notholoca* species
Bdelloid rotifer		*Platyias patulus*
Collotheca species		*Trichocerca lata*
Conochilis natans		*Trichocerca multicrinis*
Conochilus unicornis		
Epiphanes senta		
Filinia longiseta		
Gastropus hyptopus		
Gastropus stylifer		
Kellicottia bostoniensis		
Kellicottia longispina		
Keratella cochlearis var. 1		
Keratella cochlearis var. 2		
Keratella quadrata		
Lecane (M.) species *1*		
Lecane (L.) species 2		
Lepadella species		
Mytilina ventralis		
Notholoca acuminata		
Notommatid rotifer		
Ploesoma hudsoni		
Ploesoma lenticulare		
Polyarthra major		
Polyarthra vulgaris		
Pompholyx sulcata		
Synchaeta species		
Testudinella patina		
Trichocerca cylindrica		
Trichocerca species		
Trichotria tetractis		

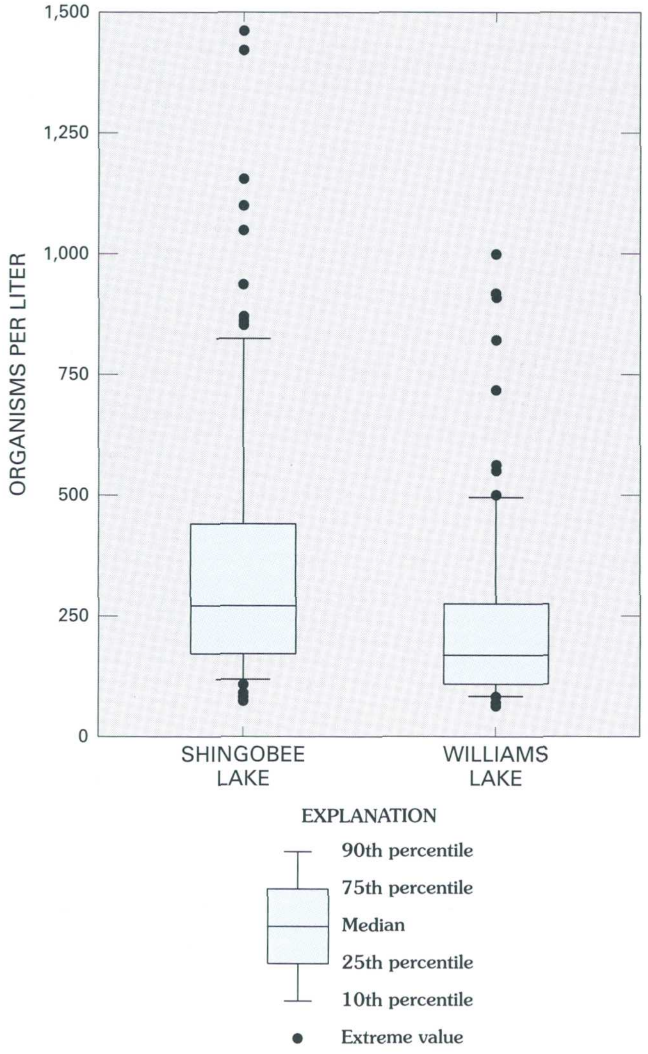

Figure 56. Zooplankton abundance data from Shingobee and Williams Lakes, 1989 to 1991.

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The Photosynthetic Response of Phytoplankton in Shingobee Lake and Williams Lake

By Brian E. Cole, James E. Cloern, and Andrea E. Alpine

INTRODUCTION

An underlying concept of the Interdisciplinary Research Initiative is that the ecological dynamics of lakes reflect biophysical connections between these aquatic systems and atmospheric, terrestrial, and aquatic components of the watershed. From this follows the expectation that many biological processes in lakes, including primary productivity (the incorporation of organic matter by photosynthesis), are influenced by water-residence time. Because organic material is the basis on which the trophic structure of an ecosystem is developed, two defining characteristics of a lake are the seasonal pattern and level of incorporation (for example oligotrophic, eutrophic) of carbon energy into a system through photosynthesis. Differences in physical, chemical, and biological processes within a watershed and between lakes with different residence times are expected to result in unique phytoplankton communities that differ in their ability to incorporate organic carbon into the lake system. Closed-basin lakes such as Williams Lake, with long water-residence time, are expected to exhibit different patterns of carbon uptake than would open-basin lakes such as Shingobee Lake. Different patterns of carbon fixation by phytoplankton are reflected in the photosynthetic response to light and ultimately in the amounts of organic carbon processed.

A cornerstone in studies of phytoplankton ecology is an understanding of the relation between photosynthesis and light intensity. This relation is fundamental to modeling photosynthesis over depth in a water column. A curve relating photosynthesis to light, P versus I curve, can be defined with two parameters, P_m^B, the light-saturated rate of photosynthesis, and α, the light-limited rate of photosynthesis (Platt and Jassby, 1976). This paper describes the comparison of depth-integrated production and the variation in phytoplankton productivity to varying irradiance as measured by the photosynthetic parameters P_m^B and α in lakes of fundamentally different hydrodynamic character. These parameters provide measures of the amount of internally derived carbon in each lake, which enables us to compare the photosynthetic response of the two lakes, and provide the basis for future efforts at modeling carbon flow in the lakes.

STUDY METHODS

Measurements of photosynthetic carbon uptake rate at various water depths (irradiance) were made during three open-water periods in Shingobee and Williams Lakes during 1989. The measurements were made during the spring prestratification period within the first week following ice-out (April), during the summer period of persistent stratification (August), and following destratification due to the fall overturn (October).

Carbon uptake was measured using *in situ* incubations of lake water labeled with a radioisotope tracer, ^{14}C. Replicate experiments were conducted in each lake with a 2- to 3-day interval between incubations. Water samples were collected at eight depths within the photic zone. Aliquots of 300 mL were inoculated with 10 microcuries (μCi) of ^{14}C and then attached to a cable suspended from a float at the water's surface. After 24 hours, the bottles were retrieved and processed as outlined by Cole (1989). Estimates of integral production over the photic depth were calculated by quadrature of the eight measurements of productivity and depth. The photosynthetic parameters P_m^B and α, normalized for the amount of phytoplankton biomass (B), were derived by Gauss-Newton nonlinear curve fitting of the productivity and irradiance data combined from replicate experiments.

Chlorophyll-*a*, a measure of phytoplankton biomass, was measured from water collected at the incubation depths. Aliquots were filtered through 47-mm GF/AE filters and chlorophyll-*a* was determined spectrophotometrically from acetone extracts of the filters (Lorenzen, 1967; Riemann, 1978).

Phytoplankton collected from the epilimnion of each lake were preserved with Lugol's stain, identified, and counted using light microscopy. Biovolumes of the observed taxa were estimated on the basis of cell shape and dimensions of representative specimens. All phytoplankton enumerations and identifications were done by Richard Dufford of Fort Collins, Colorado.

Results of Experiments

The dynamic conditions in the lakes were reflected in variations in incubation conditions between replicate experiments (table 12). For example, differences in surface-light intensity at times varied by a factor of two between experiments. Although the total amount of daily light differed widely, patterns of carbon-assimilation rate with light (P versus I curve) were consistent between replicate experiments. This consistency suggests that the observed P:I responses accurately reflected the algae's response to the incubation conditions. Consequently, data from replicate experiments were combined and a single data set for each sampling period was analyzed

Table 12. Conditions for replicate incubation experiments at Shingobee and Williams Lakes

[mg/m^3, milligrams per cubic meter; m, meter; E/m^2/d, Einsteins per square meter per day; °C, degrees Celsius; Expt, experiment]

		Chlorophyll-*a* Mean (mg/m^3)	Chlorophyll-*a* Range (mg/m^3)	Photic depth (m)	Surface light intensity (E/m^2/d)	Mean temperature (°C)
Shingobee Lake						
April	Exp. #1	8.4	6.7 - 8.9	6.4	25.4	5.2
	Exp. #2	12.4	4.1 - 15.1	4.0	46.8	7.2
August	Expt #1	8.1	7.5 - 12.9	5.6	49.1	20.2
	Expt #2	10.6	7.6 - 19.3	5.8	22.7	20.6
October	Expt #1	8.5	8.0 - 9.3	4.4	23.5	9.1
	Expt #2	8.2	7.1 - 8.6	4.7	19.7	8.9
Williams Lake						
April	Expt #1	5.2	4.3 - 6.8	10.7	20.1	5.1
	Expt #2	6.1	3.8 - 8.3	7.5	47.0	6.2
August	Expt #1	4.0	2.4 - 9.5	9.9	32.8	16.2
	Expt #2	4.7	2.2 - 10.3	10.0	38.3	17.2
October	Expt #1	3.4	3.3 - 3.6	8.6	24.3	9.5
	Expt #2	3.1	3.1 - 3.3	9.2	21.9	9.0

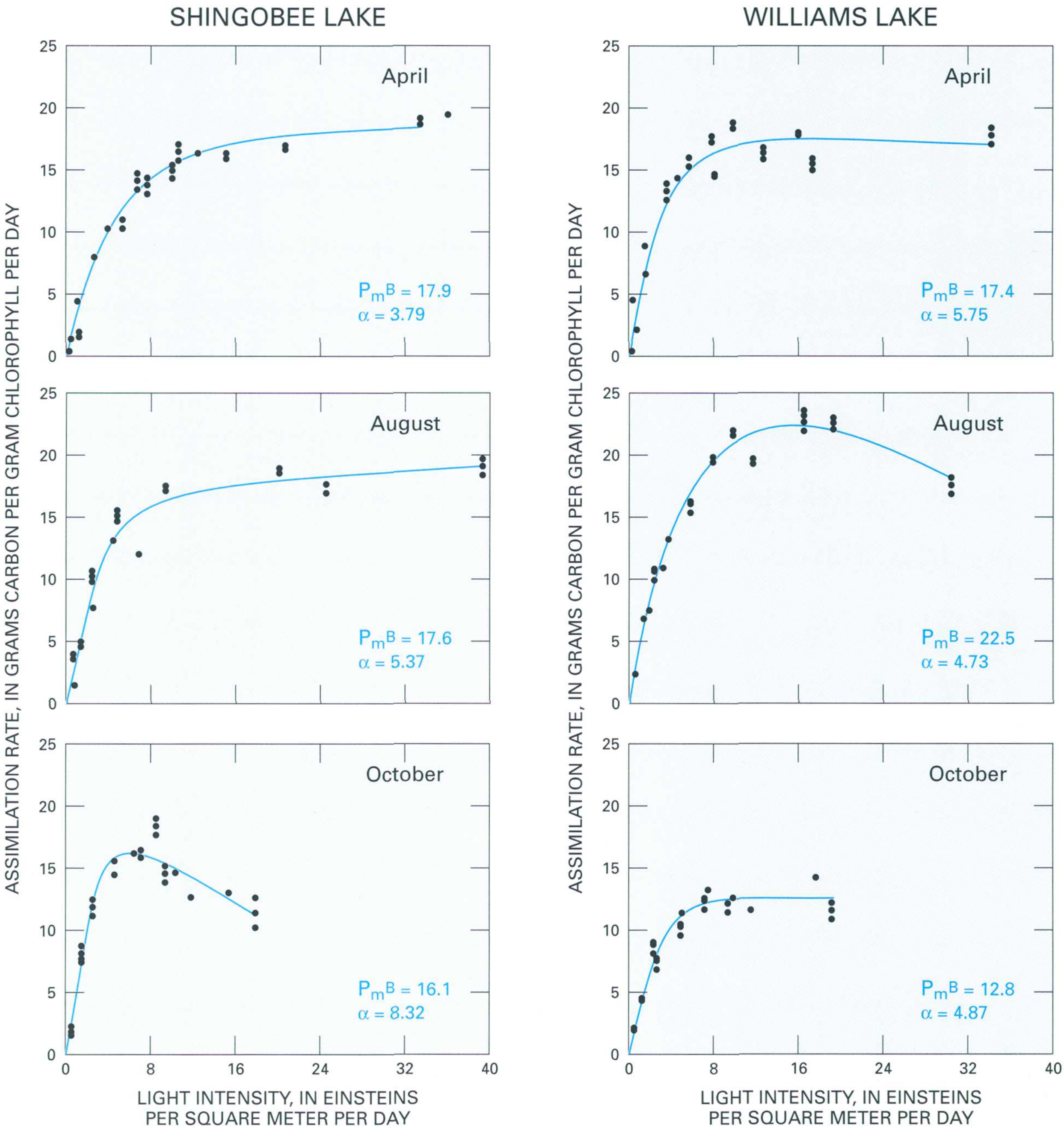

Figure 57. Relation between carbon assimilation rate and light intensity for Shingobee and Williams Lakes

by Gauss-Newton nonlinear curve fitting. The consistency of results observed between replicate experiments was not observed between the two lakes or among sampling dates. In Shingobee Lake, the maximum assimilation rate, about 17 grams carbon per gram chlorophyll per day (gC/gchl/d), was similar during the three periods (fig. 57). However, measures of α varied by a factor of two, from 3.79 grams carbon per gram chlorophyll per Einstein per square meter (gC/gchl/E/m$^{2)}$ in April, to 8.32 gC/gchl/E/m^2 after the fall overturn. These values fall within the range typically reported in the literature (Reynolds, 1984). Uptake at low light levels is a function of the light-absorption characteristics of the pigments contained

Table 13. Percent of total biovolume for each taxon during each sampling period

	Shingobee Lake			Williams Lake		
	April	August	October	April	August	October
Chlorophyceae - green algae	3	5	0	15	3	2
Chrysophyceae - golden algae	1	40	10	1	0	3
Bacillariophyceae - diatoms	6	7	5	2	0	5
Cryptophyceae - cryptophytes	80	11	21	10	2	36
Pyrrophyceae - dinoflagellates	8	18	10	0	84	52
Cyanophyceae - blue-green algae	2	19	54	71	11	3

in the cells (Harris, 1978). Consequently, for a single population, values for α are not expected to vary widely (Reynolds, 1984). The progressive increase in α may reflect changes in the community, from one dominated by cryptophytes in spring, to chrysophytes in summer, to blue-green algae in fall (table 13). However, the reason for the increase in α cannot be determined with any degree of certainty for natural populations with only 1 year of measurements.

In Williams Lake, the maximum photosynthetic rate varied by a factor of two and the initial slope was relatively constant (fig. 57). The light-limited rate of carbon uptake remained at about 5 but P_m^B doubled from a low of 12.8 gC/gchl/d after fall overturn (October) to a high of more than 22 gC/gchl/d during summer stratification (August). Changes in P_m^B can be caused by a number of factors whose relative importance to photosynthetic processes will need to be addressed in future studies. Light history (controlled by mixing processes) of a population, the composition of the phytoplankton community, and the effects of temperature on their physiological condition may be responsible for differences in maximum uptake rates (Reynolds, 1984; Harris, 1978). The high P_m^B level in August occurred when stratification (Cloern and others, 1992) probably retained the phytoplankton in well-lighted surface waters--conditions in which low P_m^B values would typically be expected. High water temperature, a competing factor that tends to increase P_m^B values, was at a maximum in late summer and may have been the cause of the high value.

Changes in the composition of the phytoplankton community in Williams Lake (table 13) also make the causes of variation in carbon assimilation unclear. The population shifted from one dominated by blue-green algae following ice-out to one dominated by dinoflagellates in summer to a mix of dinoflagellates and cryptophytes in fall. Such substantial shifts in community composition were undoubtedly reflected in the phytoplankton's photosynthetic response to light.

Estimates of areal production over the photic depth (table 14) indicate that in both lakes the highest rates of production were during spring, and the production decreased during the following two sampling periods. Although primary production in Shingobee Lake exceeded that in Williams Lake, the variations in areal production between the lakes and among sampling dates were small compared to the range of production typically reported for lakes (Wetzel, 1975). Production levels of 200 to 500 mgC/m^2/d observed in Shingobee and Williams Lakes indicate they are mesotrophic systems, on the basis of Wetzel's criteria for differentiation.

Table 14. Mean estimates of integral carbon production in the photic zone for replicate experiments
[Carbon production values in milligrams carbon per square meter per day]

	April	August	October
Shingobee Lake	524	487	377
Williams Lake	464	377	219

This work provides a basis for additional studies of the relation between primary production and the biophysical components of the lake system. Our observations suggest that the phytoplankton communities in Williams and Shingobee Lakes are different, that the photosynthetic response of the populations reflect this difference, and that depth-integrated production levels do not vary greatly during the open-water period. Additional research will be required to understand how representative the observations made here were of year to-year variations in production and how variability in primary production can be related to biological or physical factors in the lake system. Specific questions include:

How variable is species composition from year to year? Are phytoplankton assemblages related to particular hydrologic conditions?

Do variations in photosynthetic parameters continue to be associated with particular assemblages of phytoplankton or do they vary in response to physical factors?

Is phytoplankton production consistently higher in Shingobee Lake than in Williams Lake? Why?

Are low nutrient concentrations responsible for the moderate production levels in these two lakes?

What is the relation between physical factors (for example wind, temperature, stratification) and primary production?

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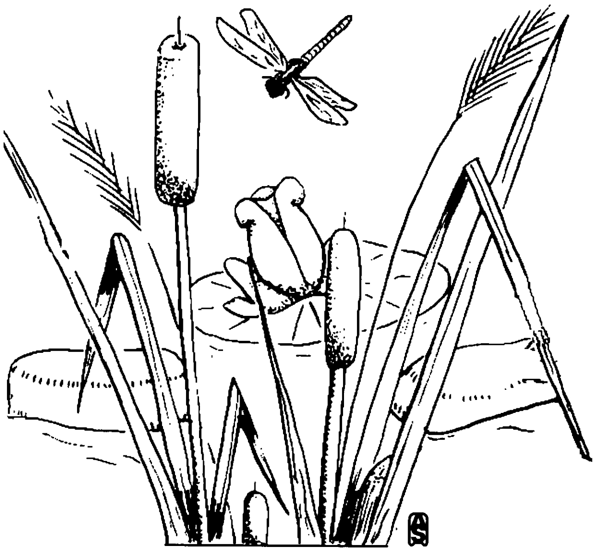

Seasonal Comparisons of Seston Abundance and Sedimentation Rate in a Closed-Basin Lake (Williams) and an Open-Basin Lake (Shingobee)

By James E. Cloern, Andrea E. Alpine, and Brian E. Cole

INTRODUCTION

A contemporary theme of limnology is that lakes strongly reflect their position in the surrounding landscape, and the USGS Interdisciplinary Research Initiative (IRI) is based on the hypothesis that closed-basin lakes such as Williams Lake are fundamentally different ecosystems than open-basin lakes such as Shingobee Lake. Interlake variability can arise from differences in the rate of exchange of materials (including water) between lake basins and their surrounding terrain. Closed-basin lakes have relatively slow exchanges so their temporal dynamics are driven primarily by internal processes and local weather. Open-basin lakes have rapid exchanges, so their dynamics are driven by internal processes plus material inputs from outside the lake basin. If this hypothesis is correct, then we might expect differences in the structure or function of lake ecosystems depending on the degree to which they are influenced by external inputs from their watersheds.

One definition of lake ecosystem structure is given by the abundance and composition of particles suspended in the water column -- the seston. A related function is the sedimentation of seston, the transfer of particles from the water column to the bottom sediments. The abundance, composition, and sedimentation of seston are fundamental features of lake ecosystems that reflect the origins of particles (for example, terrigenous or autochthonous), their chemical nature (for example, organic or mineral), and their fate (export, sedimentation, consumption). These features determine the rates and processes of diagenesis which themselves are fundamental components of lake ecosystem function.

With these definitions as a context, we designed a study to test two hypotheses relevant to the general objectives of the IRI.

Hypothesis 1: Seston abundance and composition are different in the closed-basin Williams Lake compared with the open-basin Shingobee Lake.

Hypothesis 2: If seston abundance and composition are different, then rates of sedimentation also differ between these two lakes.

Our approach was to measure seasonal abundances of different seston constituents within each lake and to measure seasonal rates of sedimentation with sediment traps.

METHODS

Measurements were made four times, representing the seasonal variability in the physical structure of the lakes: April 1989 (immediately following ice-out, during incipient thermal stratification); August 1989 (maximum thermal stratification and hypolimnetic anoxia); October 1989 (fall overturn); and February 1990 (under ice cover). All measurements were made at the deepest location (10 meters) of each lake. Horizontal variability (for example, chlorophyll patchiness) was small during the open-water seasons, but it increased markedly during winter ice cover (Cloern and others, 1992).

Water samples were collected at approximately every meter with a Niskin bottle, and then passed through 200-µm Pecap mesh to screen out the coarse particulates. Aliquots of each sample were analyzed for the following.

1. Total seston -- filtered onto pre-tared 47-mm Nuclepore (0.45-µm pore size) filters; these were air-dried and seston weight was determined gravimetrically (Hager, 1993).

2. Particulate carbon (PC) and particulate nitrogen (PN) -- aliquots were filtered onto precombusted 13-mm GF/AE filters. These were analyzed with a Perkin Elmer 240B Elemental Analyzer (see Cloern and others, 1987).

3. Chlorophyll-*a* (Chl*a*) -- aliquots were collected onto 47-mm GF/AE filters. Chl*a* was determined spectrophotometrically from acetone extracts of the filters (Lorenzen, 1967, Riemann, 1978).

Standing stocks of each constituent were calculated by trapezoidal integration of concentrations measured at depths between 0 and 8 m. Mean annual standing stocks were estimated from seasonally weighted measurements, using the following weightings: April measurements, 2 months; August measurements, 3 months; October measurements, 2 months; February measurements, 5 months. These annual estimates are highly uncertain and valid only if the daily-weekly fluctuations in seston abundance are small relative to the seasonal fluctuations (see LaBaugh, this volume).

Sedimentation rates were measured with paired sediment traps consisting of PVC cylinders (75-cm x 7.5-cm diameter) that were deployed side by side at 8-m depth. Sediment traps were recovered after 5 to 7 days, and their contents were collected into 4-L bottles that were subsampled for replicate determination of seston, PC, PN, and Chl*a* concentrations. Daily sedimentation rates R (mg/m^2/d) were calculated as:

$$R = \frac{M_f - M_i}{A\Delta t}$$

where M_f is final mass (concentration times sediment trap volume) of a constituent, M_i is the initial mass of that constituent at the time of deployment, A is surface area of the sediment trap opening (0.0044 m^2), and Δt is deployment period in days. The daily sedimentation rates reported here are the means of four measurements (duplicate sediment traps with duplicate analyses from each). The coefficient of variation of sedimentation rates (standard deviation divided by mean of four measurements) averaged 14 percent for all constituents.

RESULTS AND CONCLUSIONS

Figure 58 shows seasonal contrasts between Williams and Shingobee Lakes in their standing stocks (mg/m^2) of total seston and three seston constituents (PC, PN, Chl*a*). Estimated mean annual standing stocks are given in table 15. From these annual mean values we calculated the following indicators of seston composition.

1. Phytoplankton fraction of particulate carbon, as [50 x Chl*a*/PC] . This is a lower limit to the percentage of total particulate organic matter comprising phytoplankton biomass, assuming a phytoplankton carbon:Chl*a* ratio of 50 (Cloern and others, 1992) and that PC is mostly organic carbon. This value is an underestimate if particulate inorganic carbon (PIC) comprises a large fraction of PC.

2. Nonphytoplankton fraction of particulate carbon, as [100 percent - phytoplankton fraction]. This fraction is presumably composed of detritus plus microheterotrophs (including bacteria) and an unknown contribution from PIC.

3. C:N ratio [atomic ratio of PC:PN], an indicator of planktonic (low C:N) relative to vascular plant (high C:N) origins of particulate organic matter (Hedges and others, 1988).

4. Abiotic seston, as [total seston - (2 x PC)]. This index of the mineral fraction of seston is based on the assumption that PC is mostly organic carbon, and that organic matter is about one-half carbon by weight (Jordan and others, 1985).

Figure 59 shows seasonal contrasts of sedimentation rates for individual constituents, and table 15 lists the annual sedimentation rates. From these results we conclude the following.

1. Standing stocks of all seston constituents (seasonal measurements and annual means) are higher in Shingobee Lake than in Williams Lake (fig. 58A, table 15).

2. Mean phytoplankton biomass (as chlorophyll-*a*) is about three times higher in Shingobee Lake than in Williams Lake (fig. 58D, table 15).

3. Phytoplankton biomass comprises a larger fraction of the particulate carbon in Shingobee Lake (63 percent) than in Williams Lake (36 percent) (table 15).

4. Ratios of particulate C:N suggest that the seston in Shingobee Lake has a large component of nitrogen-rich organic matter (mean C:N = 7.4), indicative of viable plankton. The seston of Williams Lake

has a larger detrital or mineral component (mean C:N = 11.0) (table 15).

5. The two lakes have different seasonal patterns of changing seston abundance and composition (fig. 58). For example, the standing stock of PC was relatively constant in Shingobee Lake during the open-water seasons, but it was seasonally variable in Williams Lake (fig. 58B).

6. As a first-order estimate, the annual sedimentation rate in Shingobee Lake is about twice that in Williams Lake (table 15). Both lakes have seasonally variable sedimentation, with smallest rates under ice cover (fig. 59).

7. Most of the difference in sedimentation rate between the lakes is attributable to differences in the sedimentation of abiotic particles (table 15).

8. Although phytoplankton standing stock is higher in Shingobee Lake than in Williams Lake, the annual sedimentation rate of chlorophyll-*a* is approximately equal in the two lakes (table 15).

These observations are consistent with the two hypotheses presented above: the abundance, composition, and sedimentation of seston differ measurably between the closed-basin Williams Lake and the open-basin Shingobee Lake. Additional research is required to explain the underlying mechanisms of these interlake differences, and results of our exploratory measurements suggest the following hypotheses:

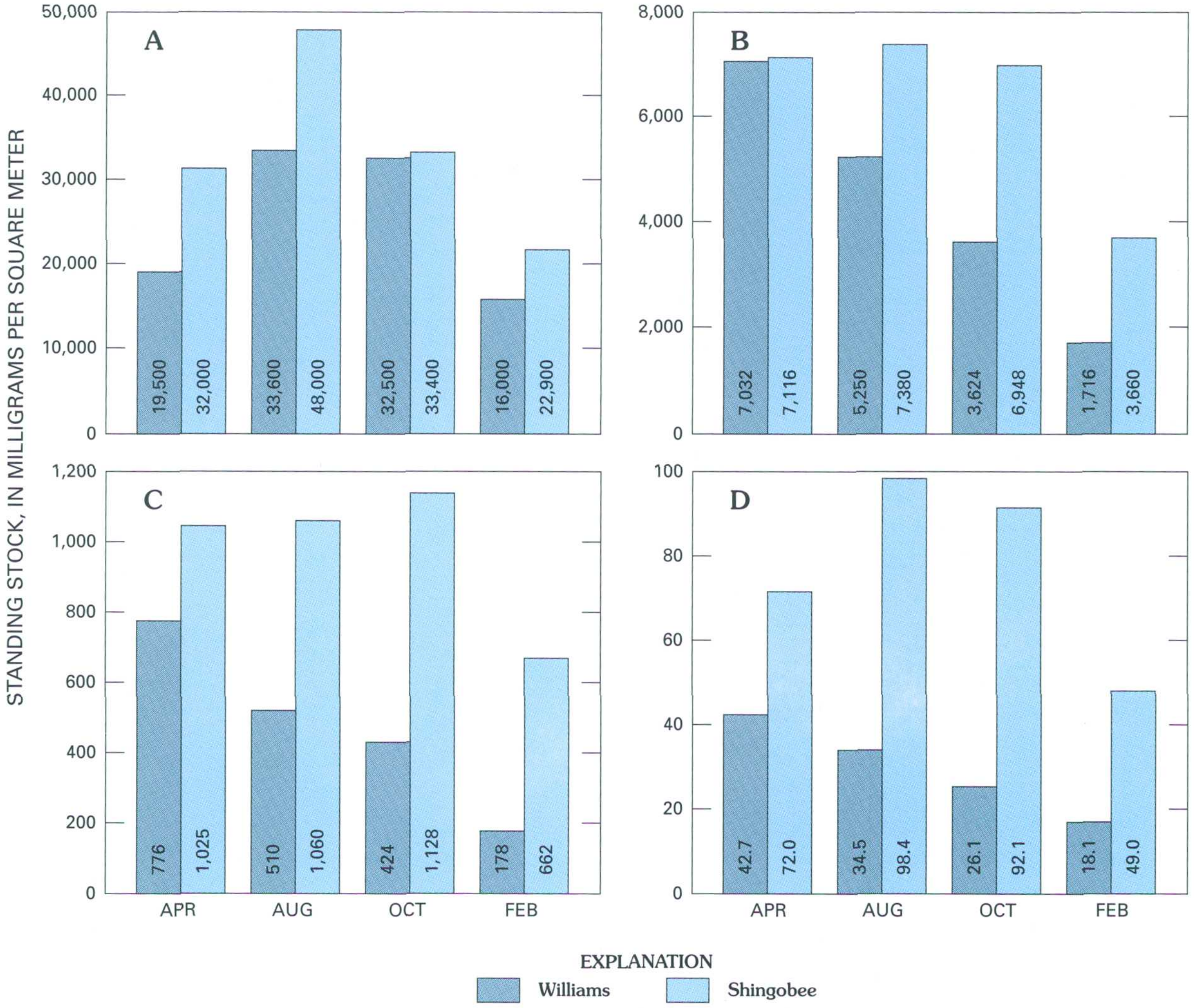

Figure 58. Seasonal contrasts in the standing stocks of (A) total seston, (B) particulate carbon, (C) particulate nitrogen, and (D) chlorophyll-*a* in the upper 8 meters of Williams Lake and Shingobee Lake.

Table 15. Mean seston abundance and composition, and annual sedimentation rates over the deepest parts of Williams and Shingobee Lakes

[Standing stocks are in milligrams per square meter, integrated over the upper 8 meters. Sedimentation rates are in grams per square meter per year.]

	Williams Lake	Shingobee Lake
Mean standing stocks:		
Total seston	23,800	32,148
Particulate carbon, PC	3,795	5,712
Particulate nitrogen, PN	403	901
Chlorophyll-*a*	27	72
Indicators of seston composition:		
Phytoplankton fraction of PC	36 percent	63 percent
Nonphytoplankton fraction of PC	64 percent	27 percent
C:N ratio (atoms)	11.0	7.4
Annual sedimentation rate:		
Total seston	350	757
Abiotic seston	180	531
Particulate carbon, PC	85	113
Particulate nitrogen, PN	9.5	10.6
Chlorophyll-*a*	0.5	0.7

1. Phytoplankton standing stocks are higher in Shingobee Lake because surface-water inflow provides a nutrient source that is absent from the closed-basin Williams Lake (LaBaugh, this volume).

2. Phytoplankton production is a more important component of the carbon budget in Shingobee Lake than in Williams Lake.

3. The large, nonphytoplankton component of particulate carbon in Williams Lake is derived from vascular plants (Carter and others, this volume), either as detritus or biogenic $CaCO_3$ (McConnaughey and others, this volume). Differences in the relative importance of phytoplankton compared with vascular plant production may result partly from differences in lake morphometry (higher ratio of littoral:lake area in Williams Lake; Winter and Rosenberry, this volume).

4. Stream input of terrigenous mineral particles is the major source of seston that gives rise to the enhanced sedimentation rates in Shingobee Lake.

5. Differences in sinking losses of phytoplankton between the two lakes are attributable to differences in their phytoplankton community compositions (Cole and others, this volume).

6. Different seasonal patterns and magnitudes of sedimentation within the two lakes are reflected in their sedimentary records.

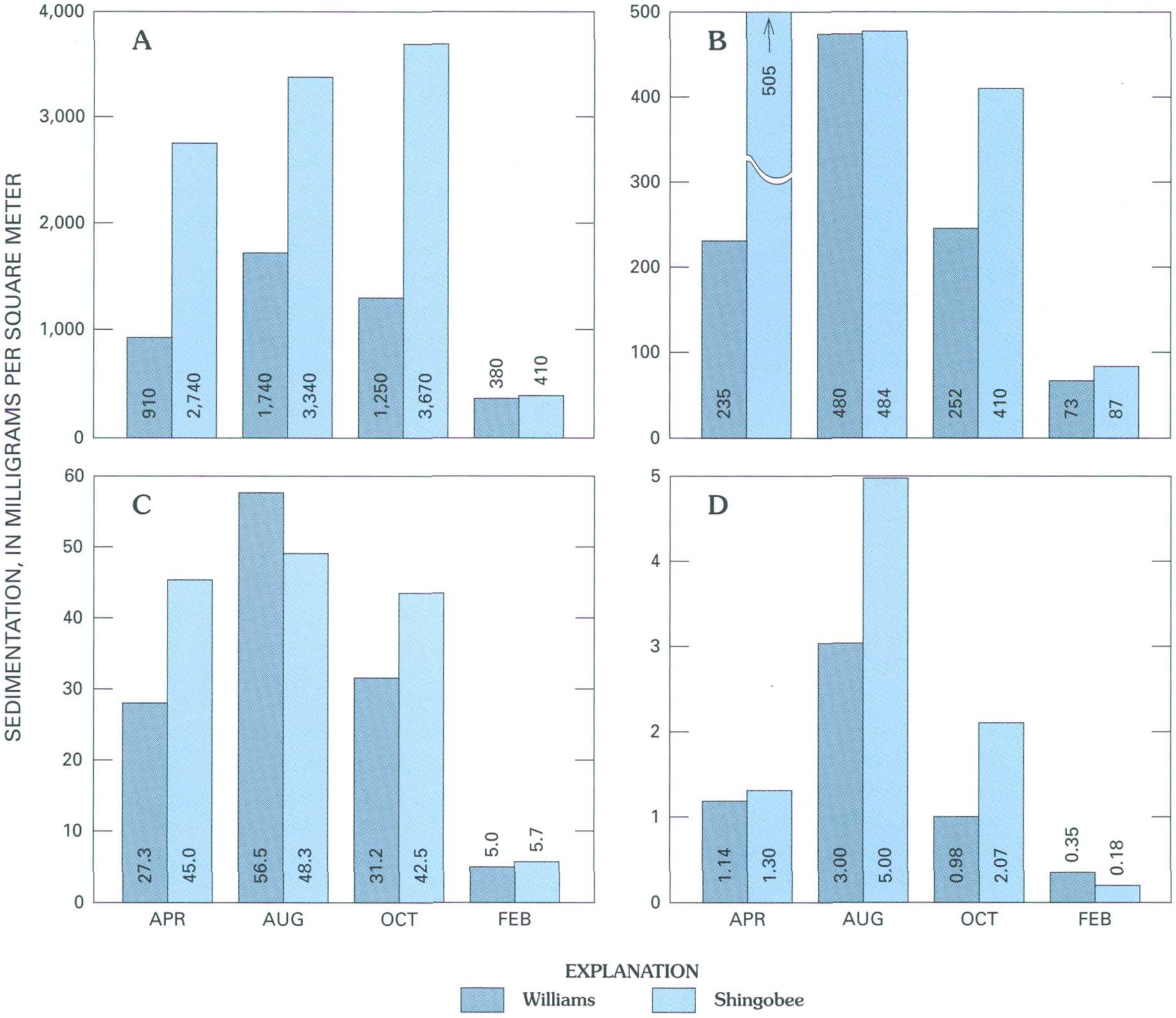

Figure 59. Seasonal contrasts in the sedimentation rates of (A) total seston, (B) particulate carbon, (C) particulate nitrogen, and (D) chlorophyll-*a* from the upper 8 meters of Williams Lake and Shingobee Lake.

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Transects of Organic Carbon, Calcium Carbonate and Diatoms in Surface Sediments of Williams and Shingobee Lakes

By Walter E. Dean and J. Platt Bradbury

INTRODUCTION

Although extensive data on the hydrology, biology, meteorology, chemistry, and limnology of Williams and Shingobee Lakes have been gathered in conjunction with the Interdisciplinary Research Initiative (IRI), little is known about the sediments of these two lakes. Because the main focus of the IRI is on the carbon cycles of these two lakes in particular, and on the Shingobee River Headwaters Area in general, understanding the burial rates of organic carbon (OC) and calcium cabonate ($CaCO_3$) in sediments is an important aspect of the carbon cycle.

Changes in diatom assemblages and abundances of individual diatom species provide measures of past changes in trophic status of lakes. In addition, diatom succession is finely tuned to seasonal fluxes of light, nutrients, and turbulence and is directly or indirectly related to the seasonal climatic, hydrologic, and biological environments both within and outside the lake. By analyzing the distribution of diatoms in the surface sediments of small lakes it is possible to determine areas of the lake bottom where diatom productivity is dominated by benthic communities, as opposed to areas (usually the deeper parts of the lake) where sedimentation of planktonic species predominates. Diatom distribution in surface sediments therefore allows an initial evaluation of the role of photosynthetic activity on the lake bottom in affecting geochemical and hydrochemical budgets in lake systems and the carbon cycle in general.

One of the principal lessons learned from the study of the sediments of Elk Lake, Minnesota, which is 50 km northwest of the IRI site (fig. 60; Bradbury and Dean, 1993), was that the carbon cycle (inorganic and organic) in lacustrine systems is intimately linked with the biogeochemical cycles of other elements, most notably those of calcium (Ca), silica (Si), phosphorus (P), iron (Fe), and manganese (Mn). The sediment components of Elk Lake, a ground-water fed lake much like Williams Lake, are almost entirely produced in the lake. These include calcium carbonate, biogenic silica, iron phosphate, manganese carbonate, and iron and manganese oxyhydroxides. Diatom productivity removes Si (LaBaugh, this volume) and produces OC, some of which is preserved in the lake sediments by burial. Planktonic diatom productivity tracks wind-driven cycling of nutrients, especially P, through time and thereby allows the rate of carbon cycling to be linked to climate change at seasonal and annual scales.

Iron may not be limiting as a micronutrient for algal productivity, but the precipitation of iron phosphate sequesters phosphorus and thereby limits organic productivity. In deep lakes like Elk Lake, much of the sedimentary $CaCO_3$ is precipitated in the water column by algal photosynthesis during the warm summer months. Therefore, one might expect to find a positive correlation between the concentrations of $CaCO_3$ and OC in the sediments. However, the long-term burial rates of these two forms of carbon often are opposite because organic productivity by cyanobacteria and precipitation of $CaCO_3$ during the summer also increases the flux of OC, which increases the carbon dioxide (CO_2) concentration in the hypolimnion and results in dissolution of more $CaCO_3$. In shallower lakes like Williams and Shingobee with extensive littoral vegetation, much of the sedimentary $CaCO_3$ is precipitated on the leaves of aquatic macrophytes (McConnaughey and others, 1994; this volume). The biogeochemical cycles of carbon and other elements integrate many chemical and biological processes operating in complex lacustrine ecosystems and provide the sedimentary record of changes in these processes on human and longer time scales.

METHODS

This paper presents the results of analyses of carbon and diatoms in surface sediments collected along two transects in Williams Lake and one transect in Shingobee Lake (fig. 60). An understanding of variations in the modern sedimentary components in these two lakes is complementary to investigations in progress on the geochemistry, mineralogy, and diatom populations in cores that were collected from each lake for paleohydrological studies based on pollen, ostracodes, and light stable isotopes (Locke and Schwalb, this volume; Schwalb and others, 1995). Surface sediments were collected in July 1994 by using an Ekman box corer; only the top 2 cm of sediment was analyzed. Total carbon and inorganic carbon were determined by coulometric titration following extraction from the sediment by combustion at 950° C and acid volatilization, respectively (Engleman and others, 1985). Percent inorganic carbon was converted to percent $CaCO_3$ by dividing by 0.12, the fraction of carbon in $CaCO_3$. Organic carbon was determined by difference between total carbon and inorganic carbon.

A subsample of water-saturated sediment (about 0.3 cm^3 after centrifugation) was placed in a 50 mL beaker and digested in about 10 mL of concentrated HNO_3 at 100° C for 5 minutes. This digestion removes carbonates and labile organic compounds but leaves diatoms, chrysophyte stomatocysts and scales, opal phytoliths, siliceous clastic particles, and some carbonized plant fragments intact. After digestion and solute removal by centrifugation and decantation, a known aliquot of diluted, suspended residue was settled onto an 18-mm diameter coverslip, dried, and subsequently mounted in a refractive ($n = 1.6$) medium (Battarbee, 1973). Because nearly equal volumes of diatom residue are mounted on each coverslip, the number of diatoms enumerated along a standardized microscope transect gives an approximation of diatom concentration (diatoms per millimeter of transect) in each sample (Bradbury and Dieterich-Rurup, 1993). About 400 diatoms were counted from each sample or until two 18-mm transects were examined at 1,000 x magnification. In all cases, except sample SL4, diatom counts were sufficiently large to reliably calculate percentages. The 59-diatom count of sample SL4 was insufficient for reliable percentage calculation and therefore, the percentages for this sample should not be considered as reliable as the percentages for the other samples.

RESULTS

Organic Carbon and Calcium Carbonate

Results of analyses for carbon are given in table 16 and plotted relative to distance along the three transects in figure 61. The sediments of Williams Lake deeper than 4 m are virtually carbonate free, although small amounts of $CaCO_3$ (up to about 0.7 percent) occur in sediments at the north end of the lake (fig. 61B and C). Shingobee Lake, on the other hand, is carbonate-rich with up to about 73 percent $CaCO_3$ (table 16; fig. 61A). The OC content of Williams Lake sediments (about 17-30 percent) is about three times that of Shingobee Lake sediments (about 4-14 percent), but the sediments of both lakes are within the range typical for lakes in north-central Minnesota (Dean and Gorham, 1976). For the sediments of Shingobee Lake, there is a distinct negative correlation between percent OC and percent $CaCO_3$ (fig. 62).

In the graph of OC relative to $CaCO_3$ (fig. 62), there is nothing particularly unusual about sample SL4, except that it does have a high $CaCO_3$ content (73.2 percent) and a low OC content (3.8 percent). However, when concentrations of OC and $CaCO_3$ are displayed along the Shingobee Lake transect (fig. 61A), sample SL4 , from the deepest part of the lake, clearly stands out as anomalous. Intuitively, one would expect such high-carbonate, low-OC characteristics to be more typical of sediments from a marl bench in the littoral zone where most of the carbonate is biogenic carbonate skeletal debris and carbonate precipitated on leaves and stems of aquatic macrophytes (for example, McConnaughey and others, 1994; this volume). Conversely, we would expect sediments from the deepest part of the lake, with a thicker and more acidic hypolimnion, to have lower carbonate and higher OC content. The abundance of fine-grained (2-5 μm) crystalline calcite in sample SL4 remains unexplained. One possibility is that this part of the lake is recharged by one or more springs consisting of carbonate-rich ground waters having a relatively high CO_2 concentration. Loss of CO_2 when the ground water enters the lake may increase the pH and trigger the precipitation of $CaCO_3$.

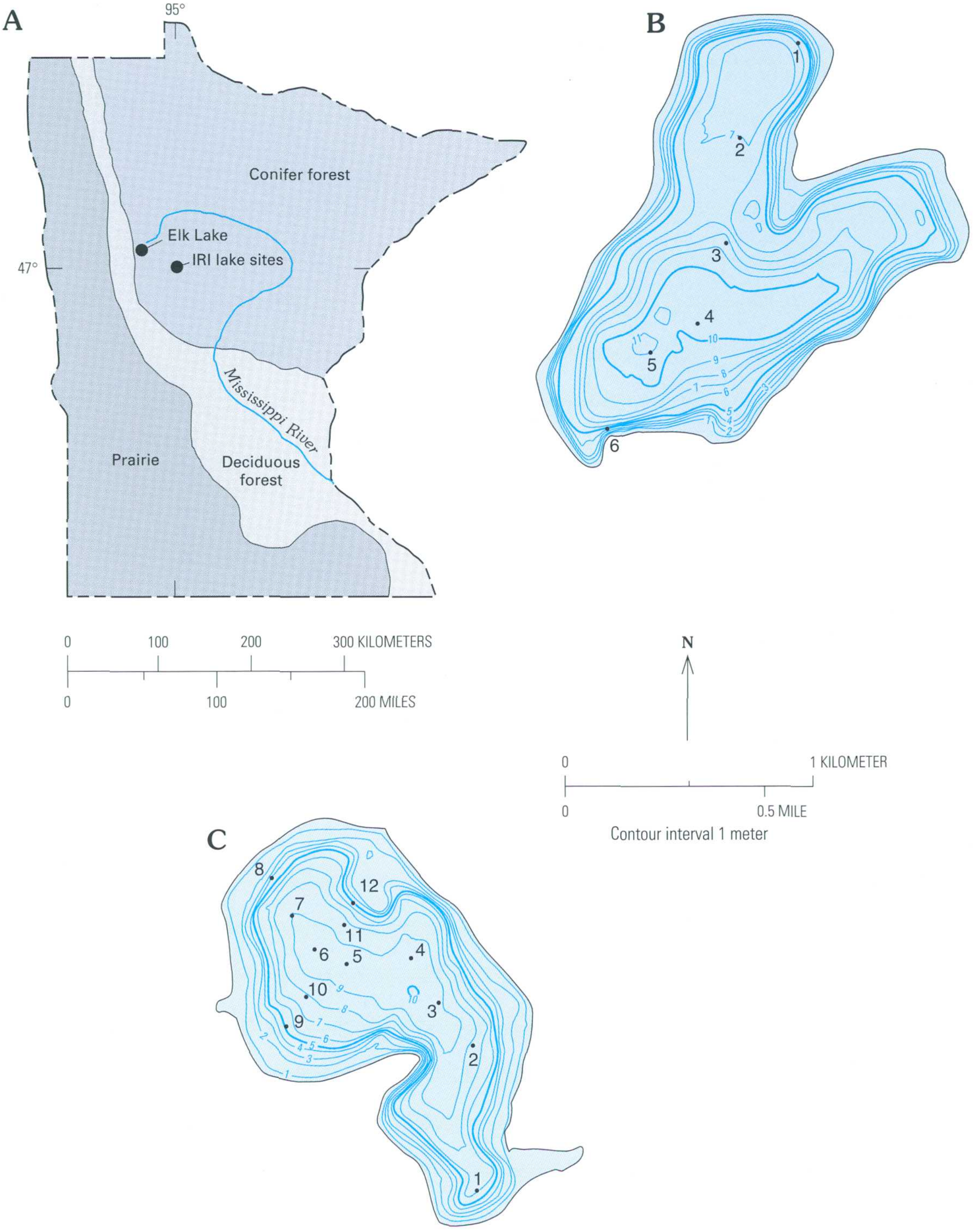

Figure 60. (A) general vegetation zones of Minnesota, and bathymetry and locations of sample sites from (B) Shingobee Lake and (C) Williams Lake.

Table 16. Percentages of dry weight of sediment of total carbon, inorganic carbon, organic carbon, and calcium carbonate in samples of surface sediments collected along transects (fig. 60) in Williams and Shingobee Lakes

Sample number	Depth (meters)	Total carbon	Inorganic carbon	Organic carbon	Calcium carbonate
Shingobee Lake					
SL 1	5.5	14.7	7.14	7.5	59.5
SL 2	6.5	15.6	6.20	9.4	51.7
SL 3	7.0	13.7	5.07	8.6	42.2
SL 4	11.5	12.6	8.79	3.8	73.2
SL 5	10.5	13.3	3.29	10.1	27.4
SL 6	4.5	16.3	2.81	13.5	23.4
Average		14.4	5.50	8.8	46.2
Williams Lake					
WL 1	5.0	23.5	0.01	23.5	0.05
WL 2	7.0	23.9	0.01	23.8	0.11
WL 3	9.0	25.7	0.04	25.7	0.30
WL 4	9.5	30.5	0.02	30.5	0.14
WL 5	10.0	25.0	0.08	24.9	0.69
WL 6	9.5	25.2	0.08	25.2	0.63
WL 7	10.0	21.6	0.04	21.6	0.32
WL 8	5.0	19.9	0.01	19.9	0.12
WL 9	4.5	29.6	0.00	29.6	0.03
WL10	8.0	28.3	0.00	28.3	0.03
WL11	9.0	24.4	0.02	24.4	0.14
WL12	6.0	17.3	0.03	17.3	0.24
Average		24.6	0.03	24.6	0.23

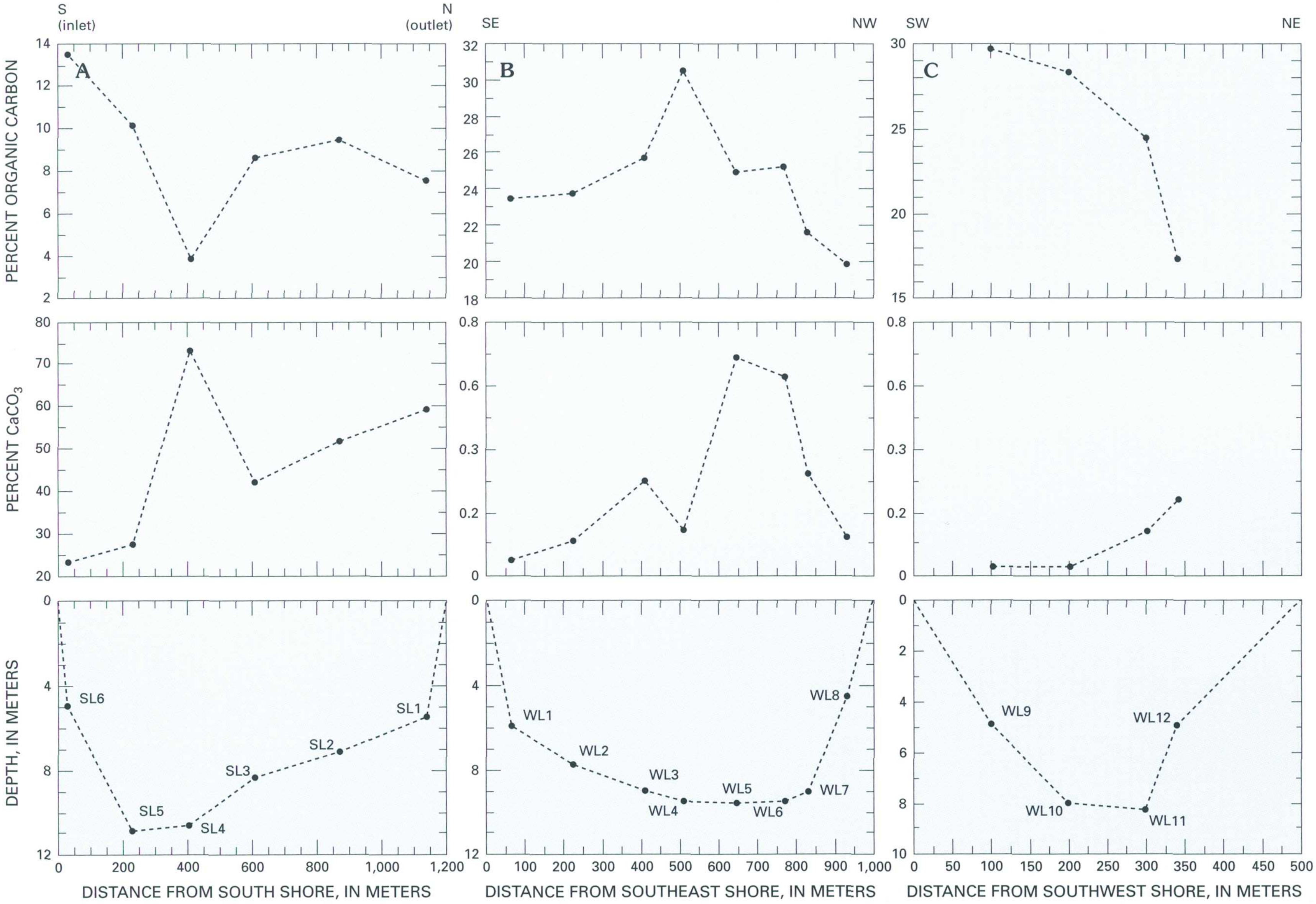

Figure 61. Transects of percentages of dry weight of sediment of organic carbon and $CaCO_3$, and water depths relative to distance from shore going from (A) south to north in Shingobee Lake, (B) southeast to northwest in Williams Lake, and (C) southwest to northeast in Williams Lake. See figure 60 for locations of sampling sites.

Diatoms

Results of diatom analyses are given in table 17 as total diatom concentration (in number of diatoms per millimeter of transect across a microscope slide) and as percentages of ecological groups of diatoms (see discussion below). These same values are plotted relative to distance along the three transects in figure 63. Bottom-dwelling (benthic) diatoms may be classified into three functional groups: (1) attached forms that are fixed to stable substrates such as plant stems and rocks within the photic zone of the lake (for example, *Cymbella*, *Gomphonema*, *Achnanthes*, *Amphora*); (2) motile species that can move on stable substrates or among sediment particles to locate areas of sufficient illumination and nutrients for growth (for example *Navicula*, *Nitzschia*, *Pinnularia*); and (3) benthic taxa that are neither fixed nor motile, but simply rest on the bottom or are loosely attached to underwater substrates (*Staurosira*, *Staurosirella*, *Pseudostaurosira*). The diatoms of the latter group typically form chains or ribbons of attached cells that may help actively growing parts of the colony stay within suitable environments of illumination.

Table 17. Percentages of ecological groups of diatoms in samples of surface sediments collected along transects (fig. 60) in Williams and Shingobee Lakes

Sample number	Depth (meters)	Attached	Benthic	Motile	Planktonic	Diatoms per millimeter
Shingobee Lake						
SL 1	5.5	5.2	4.9	16.0	73.6	9.6
SL 2	6.5	5.0	4.2	4.2	85.6	33.7
SL 3	7.0	8.4	3.6	1.1	86.1	30.9
SL 4	11.5	18.6	33.9	11.8	35.6	1.6
SL 5	10.5	9.2	4.4	4.4	80.5	49.4
SL 6	4.5	31.2	16.6	12.0	34.8	49.2
Williams Lake						
WL 1	5.0	4.2	2.3	6.9	0.5	14.2
WL 2	7.0	4.4	4.6	4.5	0.6	14.1
WL 3	9.0	3.1	22.9	6.4	5.2	38.1
WL 4	9.5	1.7	20.7	4.7	18.4	45.7
WL 5	10	4.0	13.1	9.6	25.3	52.1
WL 6	9.5	2.9	7.3	5.1	11.1	26.4
WL 7	10.0	2.0	8.4	5.7	7.7	24.2
WL 8	5.0	0.5	1.2	1.4	0.1	3.2
WL 9	4.5	20.5	44.9	33.9	0.8	3.7
WL10	8.0	22.6	19.9	45.2	10.4	3.2
WL11	9.0	17.6	12.8	27.9	40.4	13.5
WL12	6.0	24.4	40.6	29.2	5.8	89.5

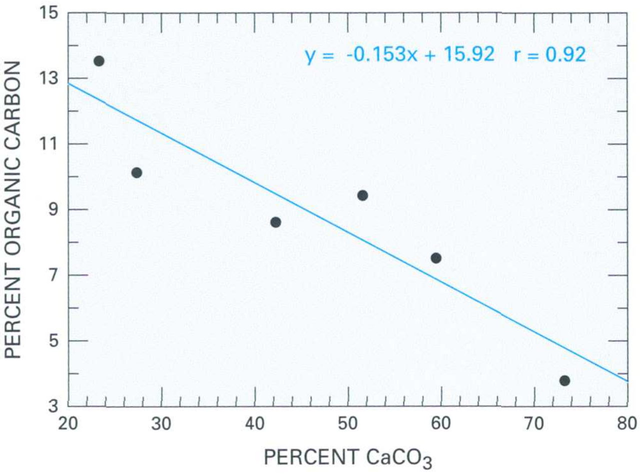

Figure 62. Percent dry weight $CaCO_3$ relative to percent dry weight organic carbon in sediment samples from Shingobee Lake.

The planktonic diatoms in Williams and Shingobee Lakes are dominated by *Fragilaria crotonensis*, *Cyclotella bodanica*, *C. michiganiana*, and *Tabellaria flocculosa*. In addition, several species of *Synedra* are important members of the plankton communities of both lakes, and *Aulacoseira ambigua* is a common planktonic diatom in Williams Lake. These same species characterize the late summer and fall diatom blooms in nearby Elk Lake, particularly if the spring circulation season is short (Bradbury, 1988; Nuhfer and others, 1993). The dominance of the fall bloom in contributing to the diatom assemblages in the sediments of Williams and Shingobee Lakes is further emphasized by the absence or low abundance of spring-blooming diatoms such as species of *Stephanodiscus*, and by the much greater drawdown of silica in the water column of Williams Lake in the fall relative to the spring (LaBaugh and others, 1995).

Percentages of benthic diatoms are usually higher in the shallow-water sediments of both Shingobee and Williams Lakes, whereas planktonic assemblages usually predominate in the deeper regions (table 17). Motile diatoms seem to prefer the shallower water habitats in both lakes, although in general, diatoms are neither very abundant nor well preserved in these environments (table 17), especially on the outflow side of Shingobee Lake. In Williams Lake, the loosely attached benthic diatom taxa are concentrated in relatively deep environments (8- to 10-m depth) in the axial, eight-sample, transect (fig. 63B). Along the four-sample transverse transect of Williams Lake diatoms are only abundant on the northeast side of the lake (fig. 63C). On the southwest end of this transect, diatoms are rare and poorly preserved and the percentage estimates of the ecological groups are probably unreliable and do not accurately reflect benthic diatom habitats.

Because sample SL4 appears to have an anomalously high content of $CaCO_3$ and a low content of OC, it was suggested that a sublacustrine spring might be present at the site where sample SL4 was collected. This sample also has an anomalously low diatom concentration that is nevertheless dominated by planktonic species (fig. 63A) The few diatoms that are present are badly corroded and broken. A sublacustrine spring in Shingobee Lake at the location of sample SL4 may also explain the poor diatom preservation. Sublacustrine springs, with anomalously high rates of flow, may interrupt normal patterns of diatom sedimentation. If such springs exist within the photic zone, benthic diatom growth may be enhanced depending on the nutrient and chemical characteristics of the water.

DISCUSSION

McConnaughey and others (1994 and this volume) showed that there is extensive precipitation of $CaCO_3$ on the leaves of rooted aquatic macrophytes in the littoral zone of Williams Lake, and yet, there is very little $CaCO_3$ in the sediments of Williams Lake (this study; Locke and Schwalb, this volume). McConnaughey and others (1994) further demonstrated that there is considerable recycling of calcium and carbon to the lake from sediments in the littoral zone. LaBaugh (this volume) computed chemical budgets for Williams and Shingobee Lakes and concluded that much of the calcium and most of the silica that enters Williams Lake, mainly from ground water, is retained in the lake by biologically mediated export of calcium and silica to the sediments. Our data, and those of Schwalb and others (1995) and Locke and Schwalb (this volume), show that there is indeed export of silica to the sediments but no calcium export. The calcium budget (and, therefore, the carbon budget) in Williams Lake remains as one of the largest unsolved problems in the IRI study.

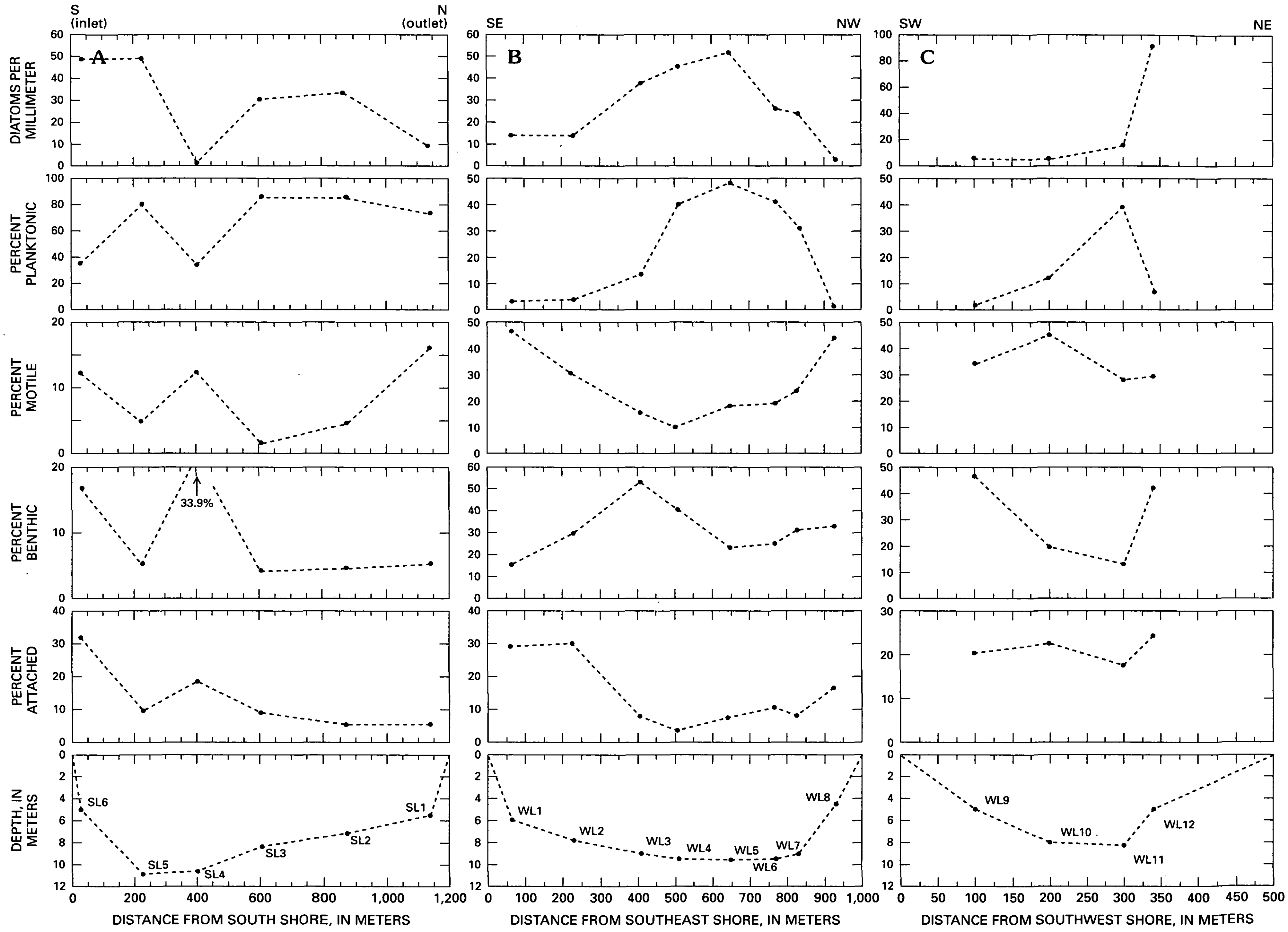

Figure 63. Transects of total diatom concentration in number of diatoms per millimeter of transect across a microscope slide, percent planktonic diatoms, percent motile diatoms, percent benthic diatoms, percent attached diatoms, and water depth relative to distance from shore going from (A) south to north in Shingobee Lake, (B) southeast to northwest in Williams Lake, and (C) southwest to northeast in Williams Lake. See figure 60 for locations of sampling sites.

Chemically, Williams, Shingobee, and other lakes in the Shingobee River Headwaters Area are similar to Elk Lake and other lakes in north-central Minnesota (table 18). These lakes are all classified as Group III lakes (on calcareous substrates with calcium- and bicarbonate-rich waters) under the chemical classification scheme of lakes in north-central United States by Gorham and others (1983). Most of the Group III lakes, based on water chemistry, precipitate $CaCO_3$ and are classified as high-carbonate Group II lakes under the mineral and chemical classification scheme of Minnesota lake sediments by Dean and Gorham (1976) and Dean and others (1993).

The lack of $CaCO_3$ in the sediments of Williams Lake and the distinct inverse relation between percent OC and percent $CaCO_3$ in Shingobee Lake (fig. 62) illustrate how high production of OC can increase dissolution of $CaCO_3$. A high rate of organic productivity in the epilimnion results in a high rate of production of CO_2 by decomposition in the hypolimnion and sediments, which lowers the pH, and increases the rate of dissolution of $CaCO_3$. What is not clear is how much of this dissolution occurs in the hypolimnion and how much occurs in the sediment. This productivity-dissolution hypothesis would suggest that organic productivity is greater in Williams Lake than in Shingobee Lake. However, productivity data for Williams and Shingobee Lakes by LaBaugh (this volume) show that primary productivity in Shingobee Lake is about two times higher than in Williams Lake. Therefore, we might expect carbonate dissolution to be greater in Shingobee Lake than in Williams Lake. The difference between the carbonate budgets in Williams and Shingobee Lakes might simply be due to differences in water chemistry. The total salinity (specific conductance, table 18) and calcium concentrations of Williams Lake are lower than those of Shingobee Lake and most other north-central Minnesota lakes (table 18). Calcite saturation measurements in Williams Lake by McConnaughey and others (1994; this volume) show that at 1 m it is oversaturated with calcite only during the summer and is undersaturated at all times below 8 m. Therefore, the differences in OC and $CaCO_3$ percentages in the sediments of Williams and Shingobee Lakes may simply be due to differences in accumulation rates. In Shingobee Lake, the percentage of OC, which may have a higher benthic accumulation rate than in Williams Lake, is diluted by a very high accumulation rate of $CaCO_3$. In Williams Lake, on the other hand, the percentage of OC, in the absence of dilution by $CaCO_3$, is much higher than in Shingobee Lake, even though the OC accumulation rate may be lower than in Shingobee Lake. Benthic flux studies will be required to resolve these differences.

An inverse relation between percent OC and percent $CaCO_3$ can be seen in surface sediments throughout the north-central Minnesota region (fig. 64), which is similar to that shown for sediments in Shingobee Lake (fig. 62). All of the lakes plotted in figure 64 have water chemistry characteristics that are similar to those of Williams and Shingobee Lakes (table 18), are typical central Minnesota Group III lakes based on the chemical classification scheme of Gorham and others (1982 and 1983), but show a distinct inverse relation between the OC and $CaCO_3$ contents of their profundal sediments. These north-central Minnesota lake sediments seem to form two groups, one with more than 20 percent $CaCO_3$ and less than 13 percent OC, and a second group with less than 20 percent $CaCO_3$ and more than 13 percent OC (fig. 64). The lower percentage of $CaCO_3$ in the profundal sediments of Elk Lake, relative to the other high carbonate, low-OC lakes (fig. 64), may be the result of greater dissolution of $CaCO_3$ in the hypolimnion of this deep, seasonally anoxic lake. Although Elk Lake is preserving seasonally laminated sediments (varves) it is not meromictic. Three lakes within a few kilometers of Elk Lake (Arco, Deming, and Josephine) are meromictic, and their profundal sediments contain 21-25 percent OC but only a few percent $CaCO_3$ (Dean and Gorham, 1976), suggesting that dissolution of $CaCO_3$ would be even more severe with permanent anoxia.

All of the lakes listed in table 18 occur in a northwest-southeast band along the upper Mississippi River drainage from the headwaters (Lake Itasca in Itasca State Park) to the Anoka Sand Plain just north of Minneapolis and St. Paul. This band is roughly parallel to, but east of, the forest/prairie border (fig. 60) and is approximately equivalent to a line between net precipitation (to the northeast) and net evaporation (to the west and south). This listing contains several groups of lakes (for example, Moose and Ball Club; George, Spectacle, Linwood, Ham, and Green) that are in close geographic, geologic, and climatic proximity and have similar water-chemistry characteristics; but some, like Williams, have high OC percentages in their sediments, and others, like Shingobee, are high in sedimentary $CaCO_3$. Because all of these lakes define a distinct trend of decreasing sedimentary $CaCO_3$ with increasing OC suggests that the percentages of these two forms of carbon are causally related. This conclusion is further supported by changes in percentages of $CaCO_3$ and OC in the sediments of Williams and Shingobee Lakes with time.

Analyses of sediment in cores from Williams and Shingobee Lakes by Schwalb and others (1995) and Locke and Schwalb (this volume) have shown that there has been little change in the amounts of sedimentary organic matter and $CaCO_3$ (both determined by loss-on-ignition) over the last 10,000 years in

Table 18. Chemical compositions of water and percentages of organic carbon and calcium carbonate in surface sediments of selected lakes in north-central Minnesota

[μs, microsiemens per centimeter at 25° Celsius;--, no data]

	Specific Conductance	Concentration in milliequivalents per liter							Percent	
Lake	(μS)	Ca	Mg	Na	K	Cl	SO_4	Alkalinity	Organic carbon	$CaCO_3$
Williams[1]	145	0.998	0.567	0.052	0.028	0.011	0.043	1.478	24.6	0.23
Shingobee[1]	341	2.095	1.398	0.252	0.043	0.101	0.079	1.398	8.8	46.2
Crystal[1]	140	0.848	0.559	0.060	0.020	0.011	0.079	1.318	--	--
Mary[1]	298	2.045	1.069	0.121	0.033	0.014	0.127	2.897	--	--
Elk[2]	266	1.646	1.378	0.313	0.045	0.019	0.068	3.250	9.5	19.0
Gladstone[2]	162	1.204	0.534	0.161	0.026	0.010	0.060	1.816	20.9	2.25
George[2]	167	1.016	0.588	0.127	0.021	0.039	0.125	1.588	17.3	3.08
Linwood[2]	171	1.032	0.652	0.115	0.019	0.034	0.122	1.658	15.9	10.5
Spectacle[2]	181	0.956	0.760	0.143	0.039	0.042	0.096	1.747	21.1	3.08
Ham[2]	188	1.048	0.916	0.156	0.039	0.048	0.163	1.924	12.7	37.4
Green[2]	206	1.084	0.984	0.134	0.029	0.028	0.141	2.069	10.6	30.6
Nokay[2]	238	1.828	0.613	0.117	0.027	0.021	0.093	2.464	19.5	15.8
Ball Club[2]	249	1.720	0.944	0.131	0.035	0.017	0.138	2.630	17.1	6.08
Moose[2]	254	1.730	0.880	0.164	0.052	0.023	0.109	2.700	8.4	28.4
Long[2]	274	1.291	1.234	0.157	0.023	0.051	0.073	3.210	6.0	40.2
Itasca[2]	284	1.388	1.820	0.278	0.044	0.030	0.055	3.454	10.8	35.0
Little Pine[2]	311	1.856	1.699	0.192	0.053	0.034	0.124	3.437	8.5	47.0
Sallie[2]	361	1.368	2.377	0.516	0.123	0.167	0.492	3.606	9.5	46.3

[1] Data on water chemistry are from LaBaugh (this volume); OC and $CaCO_3$ values are averages from table 16.

[2] Data on water chemistry are from Gorham and others (1982); data on OC and $CaCO_3$ are from Dean and Gorham (1976).

Shingobee Lake. However, prior to about 4,000 radiocarbon years ago (yr BP) Williams Lake sediments contained abundant $CaCO_3$ and about the same amount of organic matter as Shingobee Lake sediments. Sediments deposited in Williams Lake during the early Holocene (about 10,000 yr BP) had $CaCO_3$ contents of as much as 80 percent, but some of that might be detrital carbonate washed in from calcareous glacial drift. The $CaCO_3$ contents of Williams Lake sediments deposited between about 8,000 yr BP and 4,000 yr BP are fairly constant at about 40 percent. Within a few hundred years, beginning about 4,000 yr BP, the organic content in Williams Lake sediment increased significantly from about 10 percent to about 40 percent, and the $CaCO_3$ content decreased to almost zero. Clearly, 4,000 yr BP marks a major change in the hydrologic regime of Williams Lake. Prior to that time, Williams Lake was more like Shingobee Lake, at least in terms of sediment characteristics. Then, beginning about 4,000 yr BP something happened that caused complete dissolution of $CaCO_3$ in the hypolimnion and sediments of Williams Lake. One possibility is that prior to 4,000 yr BP Williams Lake had a stream outlet. Once the surface outflow ceased, nutrients began to build up, productivity increased, and the increased flux of OC resulted in dissolution of any precipitated $CaCO_3$.

The diatom distributions in the sediments of Williams and Shingobee Lakes reflect the net results of processes of growth, sedimentation, and destruction. Benthic diatoms only are abundant in shallow-water environments at sites where ground- and (or) surface-water inflow occurs. Presumably this results from a larger and more consistent supply of nutrients (including Si) in areas of inflow and implies that diatom growth rates were higher at these localities. The benthic diatom community of *Staurosira* and related genera probably accounts for a major Si (and OC) sink in Williams Lake.

Conversely, diatoms are rare and poorly preserved in shallow lake environments where outflow prevails. This is especially true in Williams Lake where water has a long residence time and ground-water outflow is reduced in nutrients and Si relative to inflow (LaBaugh and others, 1995). Probably, diatoms in such areas of the lakes have low growth rates in general, and dead cells are dissolved by Si-poor water passing through the littoral zone and into the ground water.

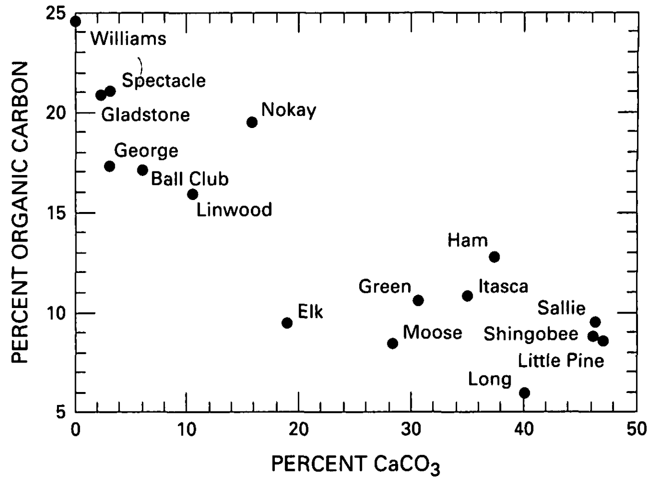

Figure 64. Percent dry weight $CaCO_3$ relative to percent dry weight organic carbon in surface-sediment samples from 16 lakes in north-central Minnesota.

Planktonic diatoms live in the photic zone of the lake when circulation provides turbulence and nutrients to support growth of this community. Biweekly measurements of dissolved silica in Williams Lake (LaBaugh and others, 1995) indicate that during most years the major drawdown of silica occurs in the fall, presumably in response to blooms of diatoms at that time. The importance of fall diatom production, particularly in Williams Lake, is also suggested by the fact that planktonic diatoms dominate the assemblages found in the surface sediments of Williams Lake. The absence of small *Stephanodiscus* species (diatoms that typically bloom during spring circulation) in the Williams Lake transects is consistent with observations of minor spring diatom productivity at Williams Lake (LaBaugh and others, 1981).

The diatom production in both Williams and Shingobee Lakes is ultimately controlled by limiting nutrients, especially phosphorus. The reason spring diatom productivity generally appears to be secondary to the production of chrysophyte algae in Williams Lake (LaBaugh, this volume) may relate to the lower values of phosphorus in the ground water and lake water of Williams Lake compared to Shingobee Lake. In addition, the greater transparency of Williams Lake ensures that phosphorus in or near the bottom sediments can be used by benthic algae (including diatoms) that can grow on the illuminated lake bottom. Higher phosphorus supply rates in

Shingobee Lake (LaBaugh, this volume), on the other hand, stimulate planktonic algal production. One of the effects of greater planktonic algal productivity is a reduction of transparency and consequent diminution of benthic algal growth. The greater relative abundance of planktonic diatoms at intermediate depths in Shingobee Lake compared to Williams Lake (table 17) probably reflects the higher planktonic productivity of Shingobee Lake and reduced growth of benthic diatom communities.

CONCLUSIONS

The present-day sediment characteristics of Williams and Shingobee Lakes are distinctly different. The sediments of Williams Lake contain high concentrations of organic matter and almost no $CaCO_3$. In contrast, the sediments of Shingobee Lake contain much lower concentrations of organic matter but abundant $CaCO_3$, and there is a distinct inverse correlation between these two variables. Water-chemistry budgets for Williams Lake (LaBaugh, this volume) indicate that calcium and silica are both retained in the lake through biologically mediated removal of $CaCO_3$ and biogenic opal to the sediments. However, this study of the sediments shows that silica is retained as remains of planktonic and benthic diatoms, but that the sediments of Williams Lake are almost completely free of $CaCO_3$.

The assemblages of diatoms in the sediments of both lakes reflect the net effects of production, sedimentation, and dissolution. Planktonic diatoms dominate the assemblages found in samples from Shingobee Lake and in the deep-water samples of Williams Lake. Benthic, attached, and motile species are more common overall in intermediate to shallow depths of Williams Lake, especially on the ground-water inflow side of the lake, where they must exert a large influence on benthic cycling of Si, C, P, and other elements.

The carbon budget in Shingobee Lake and, particularly, Williams Lake remains as one of the largest unsolved problems in the IRI research project. More closely coordinated investigations of the water and sediment columns in the lakes and wetlands of the Shingobee River Headwaters Area are needed. Sediment- and water-flux studies using sediment traps and benthic-flux chambers would be particularly useful in these investigations. Growth-rate studies of benthic diatom and other algal communities are necessary to track the removal of Si, C, P, and other elements at intermediate depths in the lakes.

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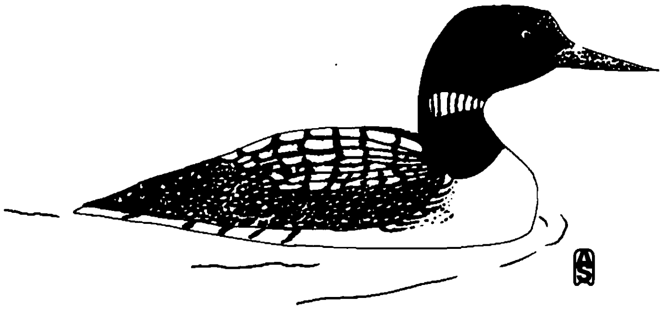

Fish Communities of Williams and Shingobee Lakes

By Robert G. Striegl, Catherine M. Michmerhuizen, and Michael A. McDonald

INTRODUCTION

The abundance and species distribution of biota in lakes are determined by a variety of factors. One concept is that the physics and chemistry of a lake largely constrain its biological identity. The design and setting of lake studies in the Shingobee River Headwaters Area (SRHA) are closely associated with this notion of "bottom up" control, and much of the focus of the IRI effort is on how factors such as hydrology and hydrochemistry affect the cycling of carbon and other elements in a pair of otherwise similar lakes. Another concept is that lake trophic structure and lake productivity are strongly influenced by predators and by predator-prey interaction. Such "top down" control is supported by evidence that fish communities of lakes can have cascading effects on lower trophic levels (Carpenter and others, 1985). For example, a predominance of zooplanktivorous fish in a lake can result in relatively sparse zooplankton densities, correspondingly low grazing pressure on phytoplankton, and high phytoplankton densities. Conversely, a predominance of piscivorous fish may result in sparse populations of zooplanktivorous fish, correspondingly high zooplankton densities, and low phytoplankton densities. Such trophic regulation of algal biomass has been suggested as the mechanism for increased water clarity in a variety of aquatic systems (Scavia and others, 1986; Shapiro and Wright, 1984) and could subsequently affect elemental cycling rates and the thermal and energy budgets of lakes.

Although extremes exist where either bottom up or top down controls dominate, at least in the short term, the trophic dynamics of most lakes are determined by complex interactions of a variety of physical, chemical, and biological factors. For lake comparison studies, such as those in the SRHA, where overall objectives include isolating the influence of hydrology and hydrochemistry on paired lakes that are otherwise similar, it is necessary to ensure that similarity extends not only to climatic, geologic, and morphometric variables but also to other potential controlling variables such as fish-community dynamics.

This study is a first-time comparison of the fish communities of Williams and Shingobee Lakes. Its purposes are (1) to characterize the fish communities of the lakes, (2) to identify similarities and differences between those fish communities in order to aid in hypothesis formulation for future studies at the SRHA site, and (3) to determine if there is any reason to believe that top down interactions are sufficiently different between the lakes to influence the interpretation of phytoplankton and zooplankton data that have already been collected. The primary sources of data for this comparison were 1989 (Shingobee Lake) and 1991 (Williams Lake) fisheries surveys done by the Minnesota Department of Natural Resources (MDNR). The 1991 Williams Lake survey was also compared to a 1982 survey by Taylor and others (1985) to assess possible changes in the Williams Lake fish community over a 9-year period. It is anticipated that for the purpose of long-term monitoring, the MDNR will continue to periodically survey the fisheries of Williams and Shingobee Lakes on similar or identical schedules.

METHODS

Fish were collected by MDNR at Shingobee Lake during August 7-9, 1989, and at Williams Lake during August 12-14, 1991. Four 76-m long nylon experimental gill nets and nine trap nets were set in Shingobee Lake, and three experimental gill nets and five trapnets were set in Williams Lake. We seined the littoral zone of Shingobee Lake using a small haul seine during August 1989. All fish were identified to species, counted, measured, and weighed, and scale samples were taken from selected fish for age determination. The scale samples were used to back-calculate mean length at age according to methods similar to those described by Lager (1975).

Minnesota Department of Natural Resources data (1989, 1991) were used to back-calculate mean length at age and to produce growth curves for five species in each lake. The slopes of these curves showing length relative to age were compared between lakes for four game species common to both lakes. In some cases, fish caught in one lake were older than fish caught in the other lake, therefore, only those fish that had corresponding ages in both lakes were used in the analysis. Differences between growth rates were considered to be significant if probabilities were about 0.05 or less (Weisberg, 1985).

RESULTS

Yellow perch was the most abundant species in Shingobee Lake (table 19), comprising 54 percent of the catch. Bluegill was the most abundant species in Williams Lake (table 19), comprising 82 percent of the catch. Thirteen fish species were caught in Shingobee Lake, and nine species were caught in Williams Lake. This compares to seven fish species caught in Williams Lake in 1982 (Taylor and others, 1985). The two additional species caught at Williams Lake in 1991, walleye and white sucker, represented only 0.6 percent of the total catch, and it is assumed that they were missed in the 1982 survey. Bluegill, pumpkinseed, and largemouth bass were the three most abundant species in Williams Lake in 1982 and 1991. Darters (*percidae*) were observed along sandy shorelines of both lakes, but were not captured and identified to species.

Calculated mean fish lengths relative to age are shown in figure 65. Bluegill growth was clearly faster ($p = 0.0001$) in Shingobee Lake than in Williams Lake. Growth rates for largemouth bass were also significantly faster ($p = 0.0085$) in Shingobee Lake. In addition, mean lengths for year classes 1-4 for northern pike and black crappie were greater in Shingobee Lake than in Williams Lake. Growth measurements were not made for other species common to both lakes.

DISCUSSION

Shingobee Lake has stream inflow and outflow and has more extensive beds of macrophytes than Williams Lake (Rosenberry and others, this volume; Carter and others, this volume). Surface inflow may enhance the lake's productivity by carrying in nutrients and food, and macrophyte beds provide fish habitat and refuge from predators. Both qualities may enhance the ability of a lake to support fish. Unlike Williams Lake, Shingobee Lake supports populations of schooling planktivores (cisco, golden shiner), which are likely to feed heavily on zooplankton in both pelagic and littoral areas of the lake (Scott and Crossman,1973). Shingobee Lake also supports populations of ictalurids (yellow bullhead, black bullhead, brown bullhead), which scavenge nocturnally for benthic invertebrates and various other foodstuffs in or on bottom sediments. Large invertebrates, such as crayfish and unionid clams that appear to be abundant in Williams Lake, are not common in Shingobee Lake. The relationship, if any, between the lack of large invertebrates and feeding pressure by benthivorous fish in Shingobee Lake is not known.

All fish species that were caught in Williams Lake in 1982 (Taylor and others, 1985) were also caught in 1991. However, northern pike were slightly less abundant in 1991 than in 1982. This could be a sampling artifact, but the lake has also become more developed in recent years, and it is likely that it is now more heavily fished. Several northern pike greater than 60 cm in length were caught in Shingobee Lake, but none that large were caught in Williams Lake (fig. 65). This is probably due to fishing pressure. Fishing access to Shingobee Lake is limited by landowners, and the lake appears to be fished less than Williams Lake. Creel census data, however, are not available for either lake.

It is not known if the apparent differences in fish growth rates between Williams and Shingobee Lakes are linked to the greater phytoplankton (Cole and others, this volume; LaBaugh, this volume) and macrophyte (Carter and others, this volume) productivity in Shingobee Lake, to greater production of prey for young fish, or to other factors such as lake hydrology or thermal regime. Such questions need to be approached from bioenergetics and ecosystem modeling perspectives so that transfers of energy and elements among lake trophic levels and between the lakes and their surrounding environments can be quantitatively understood.

Table 19. Fish species captured at Shingobee Lake, 1989, and at Williams Lake, 1991

[Listed in order of decreasing abundance; MDNR, Minnesota Department of Natural Resources]

Species name	Common name	Number caught	Percentage of total
FISH CAPTURED BY MDNR AND LITTORAL SEINE SURVEY, SHINGOBEE LAKE, AUGUST 1989			
Perca flavescens	yellow perch	695	54
Lepomis macrochirus	bluegill	342	26
Lepomis gibbosus	pumpkinseed	68	5.2
Ictalurus natalis	yellow bullhead	63	4.9
Esox lucius	northern pike	28	2.2
Ictalurus melas	black bullhead	23	1.8
Micropterus salmoides	largemouth bass	21	1.6
Ambloplites rupestris	rock bass	15	1.2
Pomoxis nigromaculatus	black crappie	12	0.9
Catastomus commersoni	white sucker	12	0.9
Coregonus artedii	cisco, tullibee	10	0.8
*Notemigonus crysoleucas**	golden shiner	6	0.5
*Ictalurus nebulosus**	brown bullhead	1	0.1
	total catch = 1,296 fish		
FISH CAPTURED BY MDNR, WILLIAMS LAKE, AUGUST 1991			
Lepomis macrochirus	bluegill	609	82
Micropterus salmoides	largemouth bass	59	8.0
Lepomis gibbosus	pumpkinseed	36	4.9
Pomoxis nigromaculatus	black crappie	9	1.2
Perca flavescens	yellow perch	9	1.2
Esox lucius	northern pike	7	0.9
Ambloplites rupestris	rock bass	6	0.8
Stizostidium vitreum	walleye	4	0.5
Catastomus commersoni	white sucker	1	0.1
	total catch = 740 fish		

* Captured only by Minnesota Department of Natural Resources

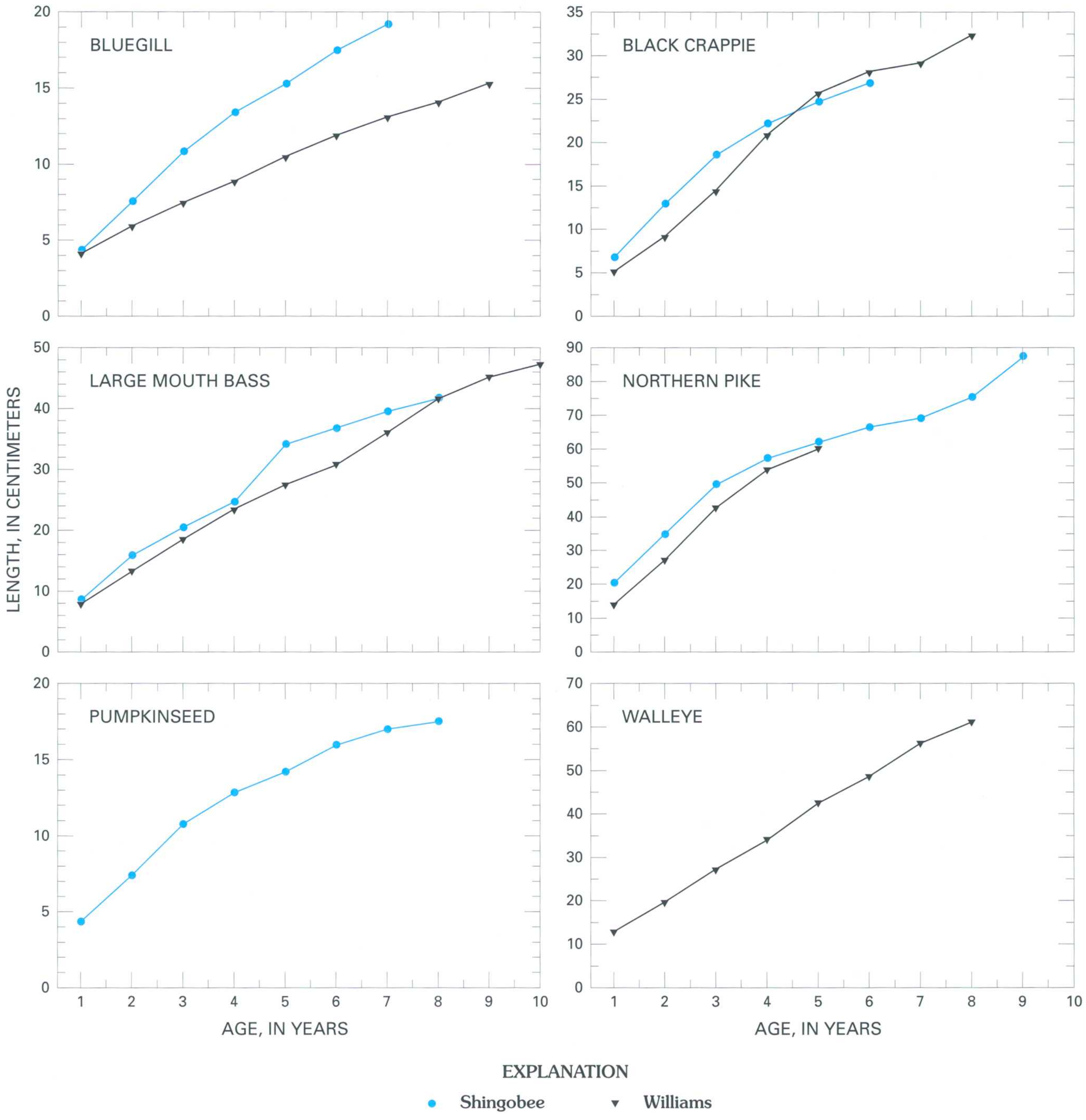

Figure 65. Growth curves for fish from Williams Lake and Shingobee Lake, Hubbard County, Minnesota (data from Minnesota Department of Natural Resources, 1989, 1991).

Table 20. Fish species captured at Williams Lake, 1982

Listed in order of decreasing abundance (from Taylor and others, 1985)

Species name	Common name
Lepomis macrochirus	bluegill
Lepomis gibbosus	pumpkinseed
Micropterus salmoides	largemouth bass
Esox lucius	northern pike
Perca flavescens	yellow perch
Ambloplites rupestris	rock bass
Pomoxis nigromaculatus	black crappie

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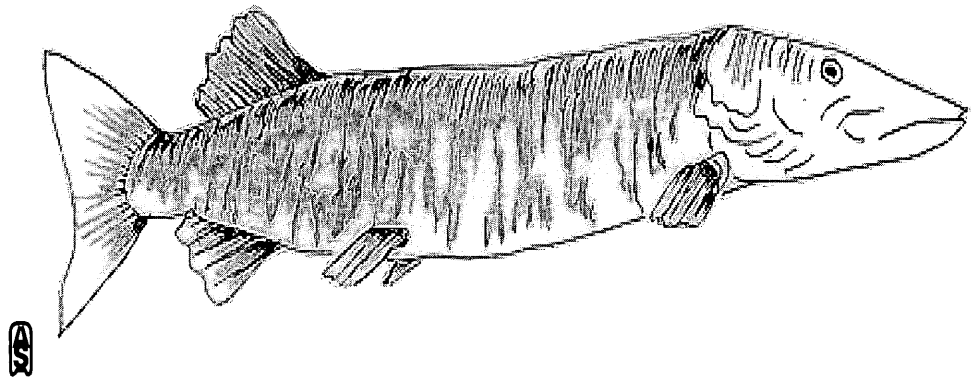

Hydrologic Examination of Ground-Water Discharge into the Upper Shingobee River

By Alan P. Jackman, Frank J. Triska, and John H. Duff

INTRODUCTION

The Shingobee River upstream from Shingobee Lake is a ground-water-dominated river, flowing in a northward direction in a valley that is incised below the level of the adjacent water table (Winter and Rosenberry, this volume). The reach of stream studied here is roughly the lower half of the free-flowing reach between Steel Lake and Shingobee Lake (plate 1). In this reach the wetted channel is generally narrow, ranging in width from 3 to 7 m except in areas ponded by beaver activity. Because there is little runoff from the porous soils, the stream is exceptionally well buffered hydraulically; response, even to large rainfall events, is slow and flow increases are relatively small. The channel bottom is clean, well-sorted sand in most places with a few riffles characterized by poorly sorted sediments ranging from boulders larger than 30 cm to sand. The sandy reaches serve as habitat to several species of submergent aquatic macrophytes, predominantly *Elodea* sp. and *Potamogeton* sp., which occur in hummocks that may cover more that 50 percent of the wetted channel area. Beaver ponds accumulate fine sediments and commonly support a community of emergent aquatic macrophytes, including cattails.

This reach of the Shingobee River is quite different from most montane streams because of the dominant influence of the ground-water discharge into the stream channel and the fine nature of the typical stream sediments. In many montane streams, the coarse bed sediments, steep channel slopes, and limited reaches experiencing ground-water discharge result in a large substream area in which a mixing of streamwater and adjacent ground-water occurs. In streams with large ground-water inputs such as the Shingobee River, the widespread advection of ground-water through the sediments bounding the stream can limit the extent of substream mixing.

Our objectives in this work were to determine the locations and amounts of ground-water input throughout the study reach and to assess the impact of the ground-water input on the extent and characteristics of the hyporheic zone.

METHODS

Tracer-dilution techniques were used to determine flow at various positions along the stream. A chloride injectate solution was prepared by dissolving 490 pounds of sodium chloride and 60 pounds of potassium chloride in approximately 750 L of water. The solution was stored in a plastic swimming pool which was covered to prevent evaporation or dilution by rainfall. This solution was continuously injected into the stream by three battery-operated metering pumps (Fluid Metering Inc.), each set to deliver approximately 50 mL/min. Each pump had a separate discharge nozzle that allowed the injectate to be sprayed rather uniformly across the width of the entire channel. The injection was made at the upstream end of a relatively narrow, steep-gradient reach, which permitted rapid mixing of the injectate. The injection was a broad pulse that lasted 73 hours and resulted in a downstream chloride concentration response consisting of three phases: a rise, a plateau, and a fall.

Streamwater samples were collected upstream and downstream from the mixing reach and at stations designated BMR, 0, 4, 6, 7, and PF (see fig. 66). Station PF is located at the site of the Parshall flume, the downstream end of the study reach (not shown on fig. 66). Samples were collected frequently during the rise and fall. During the plateau phase, samples were collected at least four times daily at all locations. Samples were filtered in the field using 0.45-µm membrane filters. The samples were frozen or refrigerated and returned to the laboratory for analysis. Chloride was determined on a Technicon Autoanalyzer using the method of O'Brien (1962).

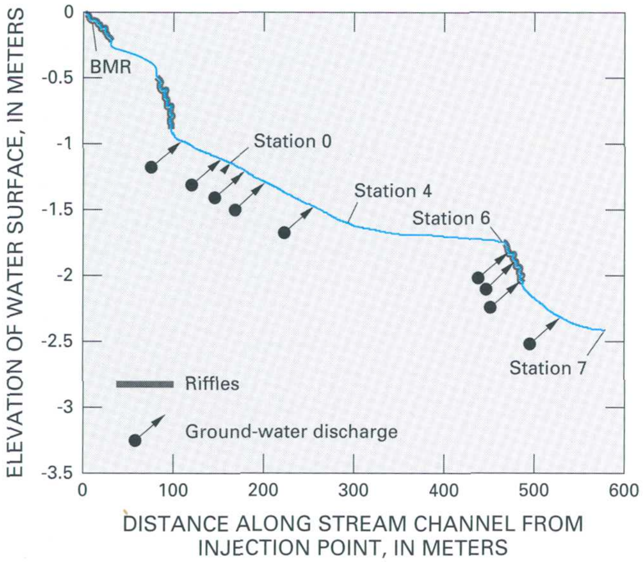

Figure 66. Axial profile of water surface elevation with locations of ground-water discharge (based on temperature measurements) and sampling stations along part of the Shingobee River.

The injection started at 1301 hours on September 9, 1992, and ended at 1400 hours on September 12, 1992. An apparent plateau in chloride concentration was reached at station 7 and all upstream stations in less than 90 minutes and at PF in less than 24 hours. The reach from station 7 to PF, though only about as long as the reach from the injection point to station 7, contains several large beaver ponds that greatly increase the traveltime through this reach. Traveltimes, the time from the start of the injection to the midpoint of the rise of the chloride from background to the plateau, to stations 0, 4, 6, 7, and PF were approximately 15, 26, 43, 52, and 630 minutes.

Ground-water-discharge measurements were made by placing a seepage meter in the streambed sediments. The device consisted of a plastic cylinder with a 19.8-cm inside diameter, open on the bottom end and capped with a flat plate on the top end. The open end of the cylinder had a beveled edge to allow it to be easily pushed into the sediments. The top plate had a hole fitted with a stopper. A short length of 6.34 mm plastic tubing penetrated the stopper and extended less than a centimeter beyond the bottom of the stopper. A condom was fitted over the stopper to collect seepage. The condom was initially evacuated by rolling, then the stopper was placed in the hole in the top of the seepage meter, and a timer was started. When 20-40 mL of seepage had accumulated, the timer was stopped and the seepage volume was determined by transferring the water into a 50-mL graduated cylinder. The seepage flux was calculated as

$$S = \frac{V}{A\Delta t}$$

where S is the seepage flux in L/s/m^2, V is the seepage volume collected in the condom in L, A is the area of the seepage meter (0.0308 m^2), and Δt is the time of collection in s.

RESULTS

Stream discharge at each station was computed using tracer-dilution techniques. Assuming that chloride ions do not react with constituents in solution in the stream, with periphyton and aquatic plants in the stream, or with stream or substream sediments, and that the stream and discharging ground-waters all have the same background chloride concentration, the mass balance for chloride is calculated using

$$Q_s = \frac{Q_{inj}C_{inj}}{(C_s - C_{s,\,bkg})}$$

where Q_s is the stream discharge, Q_{inj} is the injection rate, C_{inj} is the concentration of the injectate, C_s is the concentration in the stream, and $C_{s,bkg}$ is the background chloride concentration of the stream above the injection point and of the ground-water discharging into the stream. In calculating stream discharge, a background concentration of 0.85 mg/L was used, which represents the average value of samples collected just above the injection point on September 12 prior to the end of the injection. Three samples of the injectate were collected during the experiment. The average chloride concentration in the injectate was 188,100 mg/L. The injection rate for each of the pumps was determined by measuring the time required to fill a 50-mL volumetric flask. Pumping rates varied by several percent during the experiment as a result of temperature and battery voltage variations. From 1000 hours on September 10 to 1400 hours on September 12, those variations were within the error limits of the measurement technique and average flows from the pumps were 0.959 x 10^{-4}, 1.066 x 10^{-3}, 1.035 x 10^{-3} L/s for an average total injection rate of 3.060 x 10^{-3} L/s.

The most reliable data for estimating stream discharges are those collected near the end of the plateau. The transient effects associated with storage of solutes in the substream sediments are minimized using these data, and this was a period when the injection rate was particularly stable. Two samples were collected downstream from the mixing reach (station BMR) just prior to injection cutoff. The average chloride concentration was 4.04 mg/L. Using the average background concentration of 0.85 mg/L, a discharge of 180.4 L/s was calculated just downstream from the mixing reach at the end of the injection. This is assumed to be the best estimate of the discharge at the injection point for the last two days of the injection.

On September 11 and 12, samples were collected at all stations with collection times spaced approximately according to traveltimes between the stations. Ratios of concentrations at successive stations were calculated from these data and averaged. Based on these average ratios and on a discharge of 180.4 L/s at station BMR, discharges were calculated for all downstream stations (table 21). Discharge increased by 13.1 percent between the mixing reach and station 7 and by 6.0 percent between stations 7 and PF. Table 22 presents the reach lengths between adjacent stations and the ground-water discharge per unit of reach length in each reach. Clearly, there are great variations in the rate of ground-water discharge on a reach-by-reach basis, from a high of 0.0628 to a low of 0.0057 L/s/m. There is no obvious pattern to the rates of ground-water discharge as one moves downstream. The greatest ground-water discharges are in the most upstream reach and in the reach between stations 6 and 7.

Table 21. Stream discharges at various stations along the Shingobee River

[L/s, liters per second]

Station	Stream discharge (L/s)
BMR*	180.4
0	189.2
4	196.0
6	197.0
7	204.1
PF	216.3

*samples collected just downstream from the mixing reach

A number of discrete measurements of ground-water discharge along the channel bed are presented in table 23. Several readings were taken at the 515 m point (all distances are measured from the injection point) to determine the reproducibility of the seepage-flux measurement. The scatter in the readings, about 4 percent, probably represents the accuracy of the volume and time measurements, suggesting that the flux did not change with time following the emplacement of the seepage meter.

Seepage fluxes were positive at all locations tested but varied over a twentyfold range from a low of 0.0012 to a high of 0.0263 L/s/m^2. Assuming an average channel width of 4.5 m, that is, 4.5 m^2 of bed area per meter of channel length, these correspond to ground-water discharge rates from 0.0054 to 0.1184 L/s/m. The average of the flux measurements in table 23 is 0.0096 L/s/m^2, corresponding to an average ground-water discharge rate of 0.0432 L/s/m for a 4.5-m-wide channel. The measured average ground-water discharge rate based on tracer dilution calculations of increase in stream discharge for the entire reach from the injection point to station 7, the reach in which all seepage measurements were made, is 0.0408 L/s/m. The agreement between the two methods is surprisingly good in light of the small number of seepage flux measurements.

Table 22. Lengths of reaches and average ground-water discharge for each reach

[m, meter; L/s/m, liter per second per meter]

Reach	Length of reach (m)	Ground-water discharge (L/s/m)
BMR to 0	165	0.0533
0 to 4	129	0.0527
4 to 6	174	0.0057
6 to 7	113	0.0628
7 to PF	600	0.0203
entire reach BMR to Pf	1181	0.0304

DISCUSSION

These results clearly indicate that discharge of ground-water into an approximately 1 km reach of the Shingobee River above Shingobee Lake is both very high and extremely variable, ranging from 0.0057 to 0.0628 L/s/m. There are no surface tributaries to the river in the study reach although there are several springs on the banks of the river which flow into the river. Only two of these springs have flows greater than 1 L/s. One is located on the left bank at 138 m, about 27 m upstream from station 0, in a reach that has the third highest ground-water discharge rate. The other is located on the left bank at 408.5 m, in the middle of the reach that was found to have the least ground-water discharge.

To assess the hydraulic potential for ground-water discharge, tubes were driven into the streambed in many locations and the head of water in the tube was measured . Tubes driven 30 cm into the bed often developed static heads of 5-10 cm above the surface of the stream. In some tubes, the static head was above the top of the tube, causing flowing "wells". These observations further support the observation that this reach of the Shingobee River is dominated by ground-water discharge.

Small-scale ground-water discharge and recharge are known to be related to stream geomorphology (Harvey and Bencala, 1993). Hydrodynamic driving forces for local ground-water discharge are typically highest within and immediately downstream from reaches with large changes in elevation, for example, riffles. While the study reach has never been completely surveyed, elevations of the water surface at most of the sampling stations have been determined and the locations of all major riffles are known. Figure 66 presents a profile of the reach based on the available information. Note that the reach from the injection point to station 0 has the largest change in elevation and contains two significant riffles. Also, the reach from station 6 to station 7 has a large change in elevation and one large riffle. These two reaches have the largest amount of local ground-water discharge. The reach from station 4 to station 6 has relatively little elevation change and no riffles. This is also the reach in which there was little local ground-water discharge.

In addition to high hydrodynamic driving forces, high permeability may also be present in reaches that have high local ground-water discharge. High permeability can result from high-porosity geologic units crossing the stream channel, porous alluvial deposits bounding the channel, and so forth. There are several locations along the stream where ground water is entering with sufficient velocity to produce observable "boiling" of the bed sediments. In addition, the temperature of the sediments at 20 cm below the stream channel was surveyed from station 0 to station 4 at a time when ground-water temperatures were about 6-7°C and stream temperatures were about 15°C. Significant ground-water discharge was probably occurring where temperatures measured 20 cm below the sediment-water interface were less than 8°C. The arrows on figure 66 show those locations.

A number of processes can affect the accuracy of the methods used here to measure ground-water discharge. If chloride-enriched water recharges ground water, there will be no effect on the concentration of chloride in the remaining streamwater. Therefore, the technique used here will not detect recharge and will overestimate the net ground-water discharge in cases where true ground-water recharge is present. In some cases, water lost as apparent recharge may return to the channel after being diverted through the interstitial spaces in the sediments beneath and adjacent to the stream. The process of filling these returning flow paths is known as transient storage. Until the transient storage is complete, the returning flow will be low in chloride and will be detected as ground-water discharge. For this reason, it is important to continue an injection long enough to assure that all such subchannel flow paths have been filled with high chloride water and a true plateau

Table 23. Seepage meter measurements of the ground-water discharge flux at various sites

[L/s/m^2, liter per second per square meter]

Measurement site	Seepage meter flux (L/s/m^2)
515 m near center of channel - 1st measurement	0.0127
515 m near center of channel - 2nd measurement	0.0124
515 m near center of channel - 3rd measurement	0.0122
516m near center of channel	0.0106
518 m near center of channel	0.0012
station 4 near center of channel	0.0058
275 m near right bank	0.0027
275 m near center of channel	0.0263
275 m near left bank	0.0023

has been reached. In the present study, only subchannel flow paths with a residence time longer than 3 days will have failed to fill with high chloride water. We believe it is unlikely that such flow paths could have contributed significantly to the measured ground-water discharge.

The very large ground-water discharge rate in the reach just downstream from the mixing reach is questionable. The measurement in that reach is subject to two errors that do not affect other reaches and that could cause the measured value to be inaccurate. The first possible error is incomplete mixing of the injectate in the mixing reach. The samples at station BMR were all collected in the same location near the center of the channel and just below the water surface. As a group, they show relatively little fluctuation, which is encouraging because incomplete lateral or vertical mixing could result in a large standard deviation for samples collected at a single point in the channel. However, the possibility remains that the concentration at the sampling point was either below or above the true average concentration at that cross section. If the concentration at BMR was below (or above) the average, the streamflow would be overestimated (or underestimated) and the ground-water discharge for the reach from BMR to station 0 would be underestimated (or overestimated). The second possible error would result from a subchannel flow path that passes low chloride water from above the injection point to a return point below the station BMR point. All flow in such a subchannel path would be measured as ground-water discharge in the upstream reach, resulting in an overestimate of the ground-water discharge rate in that reach. It is impossible to determine whether either of these possible errors has affected the results reported here.

CONCLUSIONS

Using concentration data from the plateau phase of a long tracer injection, the amount of ground-water discharge into the Shingobee River upstream from Shingobee Lake was measured. These measurements are in good agreement with estimates obtained using seepage meter measurements. Using the tracer concentration measurements at various stations along the reach, together with temperature measurements in the sediments and seepage meter measurements, areas where ground-water discharge is highest were identified. These results show that this reach of the Shingobee River is receiving very large amounts of ground-water discharge, which is concentrated in certain locations along the reach.

ACKNOWLEDGMENT

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The Influence of Streambed Sediments on the Solute Chemistry of Ground-water Discharge in the Upper Shingobee River

By John H. Duff, Frank J. Triska, Alan P. Jackman, James W. LaBaugh

INTRODUCTION

Water from the Shingobee River, which is predominantly ground-water derived (Jackman and others, this volume), is one of the major hydrologic inputs to Shingobee Lake (Rosenberry and Winter, this volume; LaBaugh, this volume). Consequently, ground water discharged to the Shingobee River is one source of solutes to Shingobee Lake, although uptake and modification of streamwater constituents during downstream transport can affect solute composition. Dissolved solutes in ground water also can be modified as they are advected through streambed sediments. In areas like north-central Minnesota, which has large seasonal temperature fluctuations, kinetic constraints imposed by low temperatures may inhibit solute transformations in stream water and streambed sediments. The solute composition of streamwater ultimately entering the lake depends on the expression of instream processes as influenced by the hydrologic, biogeochemical, and climatic setting of the Shingobee River.

One of the objectives of the IRI and, also, a contemporary theme in stream ecology is determining how surface-water/ground-water systems function. Data presented in this report address three premises relevant to the general objectives of the IRI: (1) that ground-water discharge is one source of solutes to the Shingobee River, (2) that biogenic processes in streambed sediments affect dissolved solute composition of ground-water discharge, and (3) that seasonal streamwater temperature fluctuations correspond to ammonium availability in the Shingobee River.

METHODS

Six piezometer nests were installed in the Shingobee Rive along a 100-m reach of streambed approximately 0.5 km above the inlet to Shingobee Lake (plate 1). Each nest consisted of 10 stainless-steel piezometers (6.34 mm inside diameter) ranging in depth from 10 to 100 cm. The reach was free flowing, contained a predominantly sand-and-gravel bed, and connected beaver impoundments about 600 m apart. Piezometric head, temperature, specific conductance, chemically reduced solute, and profiles of biogenically derived gas were measured at each nest in October 1992. Results at each depth were averaged from the six nests. Temperature and ammonium were measured bimonthly or monthly over a 3-year period at the base of the study reach beginning in April 1989.

RESULTS AND DISCUSSION

Vertical hydraulic gradients indicated upward flow of ground water; hydraulic gradients ranged from 0.3 to 0.6 (fig. 67A). The upward gradients were consistent with results presented by Jackman and others (this volume) for chloride-dilution experiments and seepage-meter measurements, which indicated that stream discharge increased by 18 L/sec (10 percent) due to ground-water inflow over a 7,600-m reach of the Shingobee River containing the piezometer nests.

Temperature profiles also indicated upward advection of water in the streambed (fig. 67B). Temperatures increased from about 7°C, at 100 cm below the streambed to about 14°C in stream water. In four of six temperature profiles, ground-water temperatures were uniform to within 10 cm of the streambed indicating that little mixing with surface water occurred as ground water moved up. Above 10 cm, a steep thermal gradient was present, probably resulting from convection of streamwater into surficial streambed sediment.

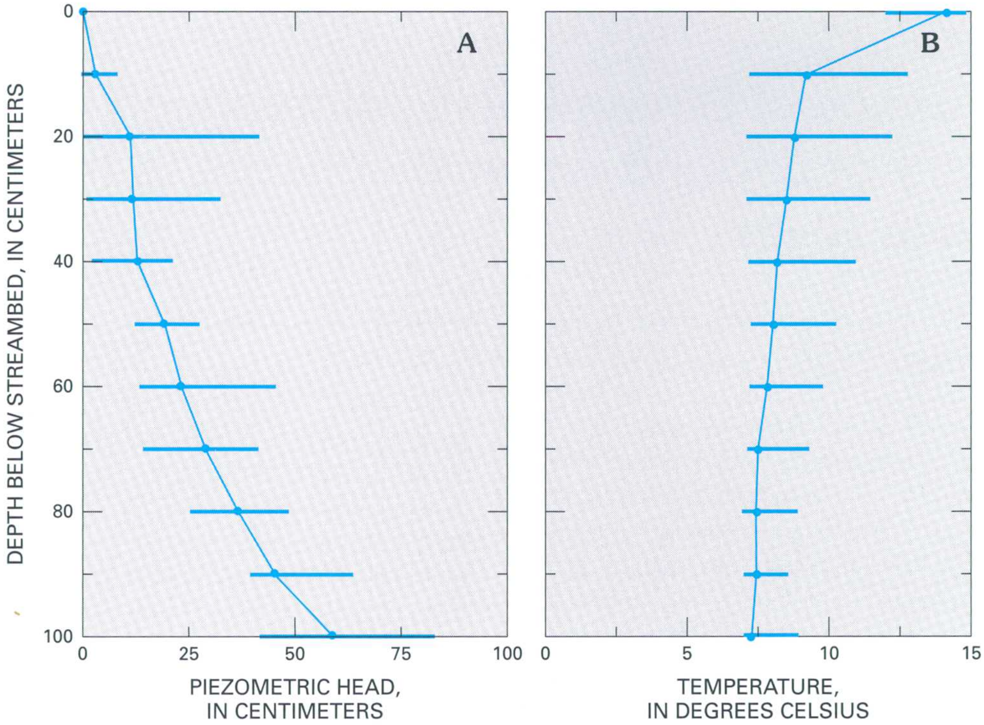

Figure 67. (A) Piezometric head and (B) temperature profiles in the Shingobee River streambed, October 1992. Average (dot) and standard deviation (bar) for six piezometers at each depth are shown.

Concentration profiles of ammonium, methane, and specific conductance (fig. 68A-C) were consistent with the interpretation of vertical hydraulic gradients. An upward advective and dispersive flux of reduced ions occurred from ground water to the stream. In contrast, a diffusive flux of dissolved oxygen occurred into the streambed (fig. 68D). Nitrate was less than 1 µg N/L in these piezometers (data not shown).

Concentration profiles of reduced ions suggest that surficial sediments are biologically and chemically active regions of the streambed where reduced ions in ground water may be oxidized as they are advected upward through the streambed. For example, advection of ammonium through the upper 10 cm of streambed may deplete greater than 95 percent of ammonium originating in ground water. Oxidation of ammonium and other reduced ions may be facilitated by diffusion of oxygen into surficial streambed sediments. These profiles demonstrate that passage of ground water through the streambed may significantly attenuate the effect of ground-water solutes on streamwater chemistry.

Because of the steep ammonium-concentration gradients and knowledge about the role of inorganic nitrogen in biosynthesis, the influence of streambed sediments on inorganic nitrogen chemistry of ground-water discharge was investigated in detailed vertical sampling. Nitrate and ammonium profiles were measured at 2.5-cm intervals in the top 10 cm of sandy sediments with and without *Elodea*, a rooted aquatic macrophyte that grows in the stream channel. Nitrate and ammonium concentrations in this zone were significantly elevated compared with concentrations in interstitial water below and streamwater above. Detailed profiles of nitrate and ammonium indicated that the top 10-cm of streambed were active in ammonium formation, nitrification, and inorganic nitrogen uptake (fig. 69). Furthermore, the presence of *Elodea* may have influenced interstitial ammonium concentrations.

Ammonium concentrations in the Shingobee River fluctuated seasonally, with low levels (about 10-50 µg N/L) persisting May through October (fig. 70). Between November and April, in contrast, ammonium ranged between about 175 and 375 µg N/L, which is

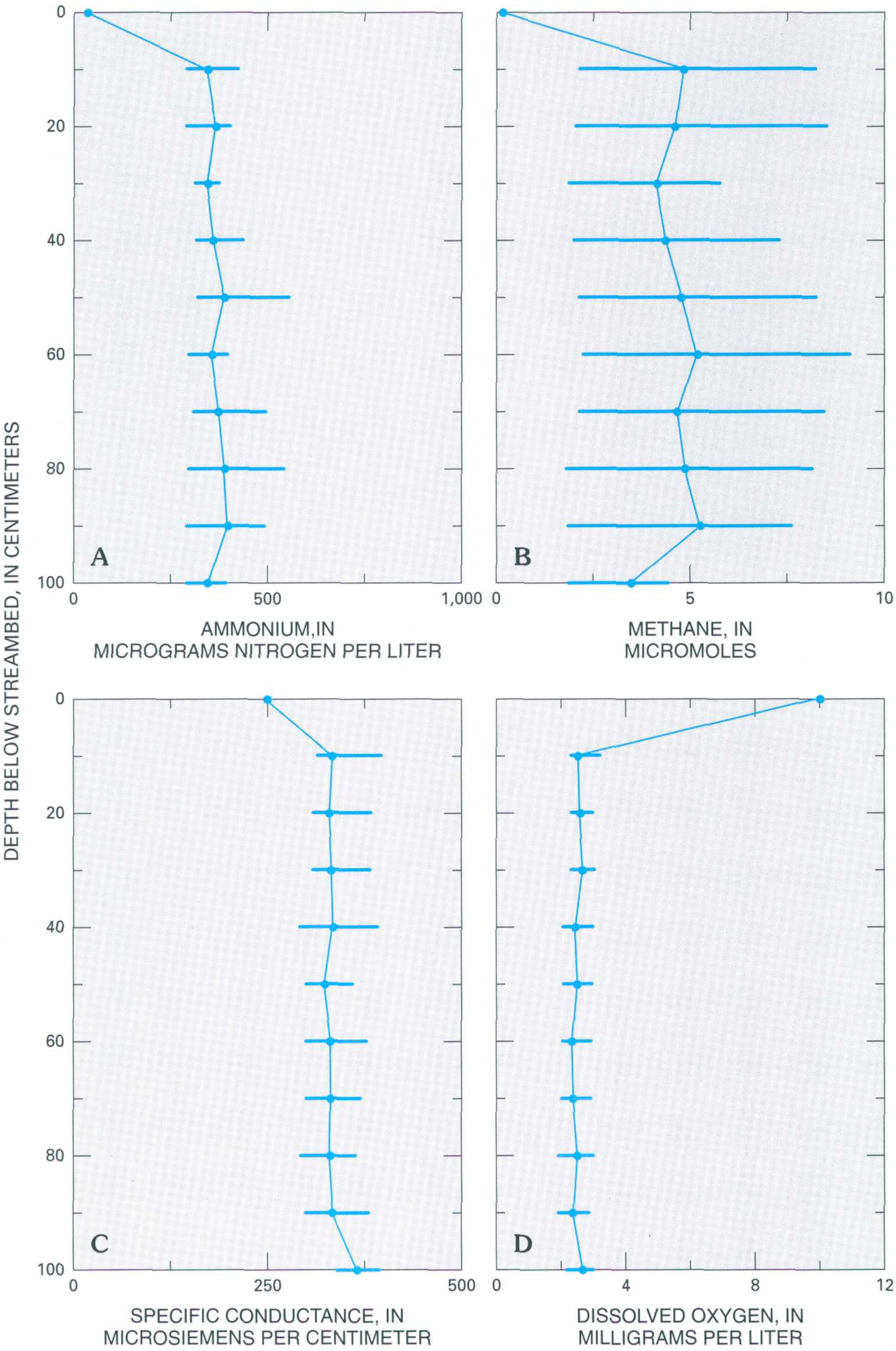

Figure 68. (A) Ammonium, (B) methane, (C) specific conductance), and (D) dissolved oxygen concentration profiles in the Shingobee River streambed, October 1992. Average (dot) and standard deviation (bar) for six piezometers at each depth are shown.

approximately the concentration in ground water. The close agreement between ground water and stream-water ammonium concentrations in winter suggests that biotic processes in the channel and near-surface sediments are inhibited at low temperatures (and possibly by increased light attenuation) and that during winter, ammonium dissolved in ground water may pass directly to the stream without modification. Therefore, between November and April, ground water discharged to the stream may be a significant source of ammonium to the lake. Conversely, the steep ammonium gradient observed in the top 5-cm of the streambed in summer suggests that biotic processes have an effect on the inorganic nitrogen retention in streambed sediments and that the effect may be underestimated if ground-water concentrations are measured without measuring interstitial water concentrations. Between May and October, passage of ground water through the streambed may significantly attenuate the effects of ground-water-derived ammonium on stream and lake water chemistry.

Detailed inorganic nitrogen profiles suggest that the upper streambed sediments are active regions of decomposition and raise several questions for further research on inorganic nitrogen cycling: (1) What is the relative importance of inorganic nitrogen loss through uptake by rooted aquatic macrophytes compared with microbial denitrification tightly coupled to nitrification? (2) What is the influence of rooted macrophytes on microbial processes in the streambed? (3) How do hydrologic and biotic components affect dissolved inorganic nitrogen flux across the sediment-water interface?

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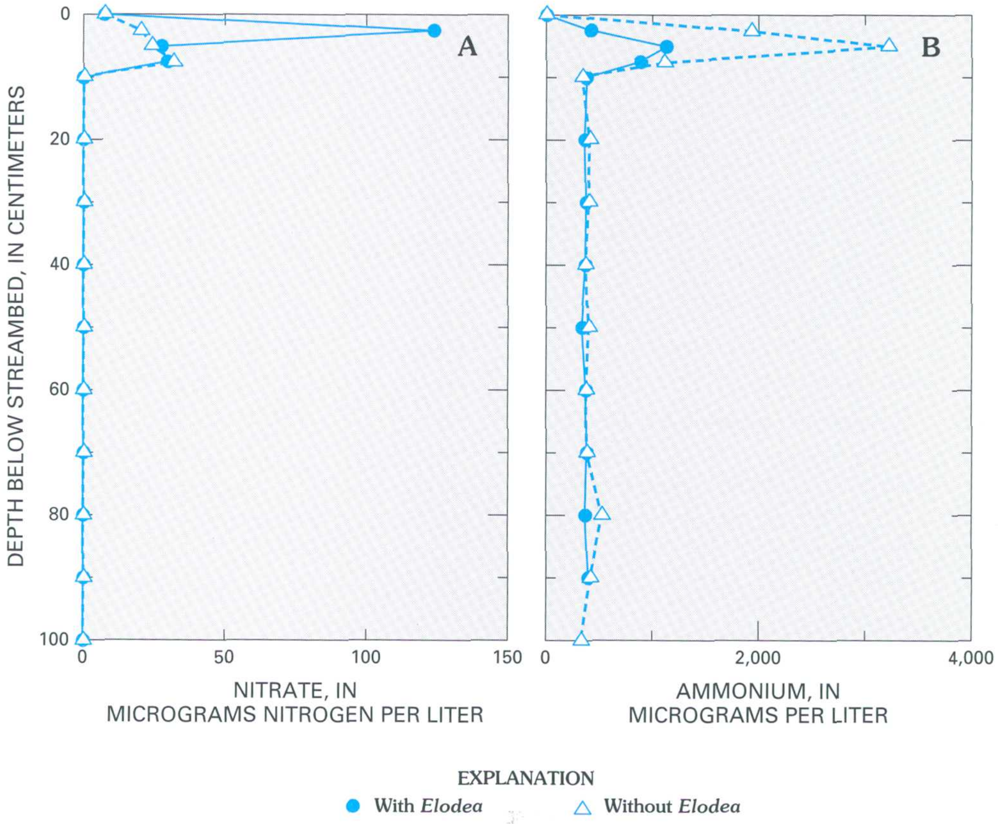

Figure 69. (A) Nitrate and (B) ammonium concentration profiles in the Shingobee River streambed, October 1992.

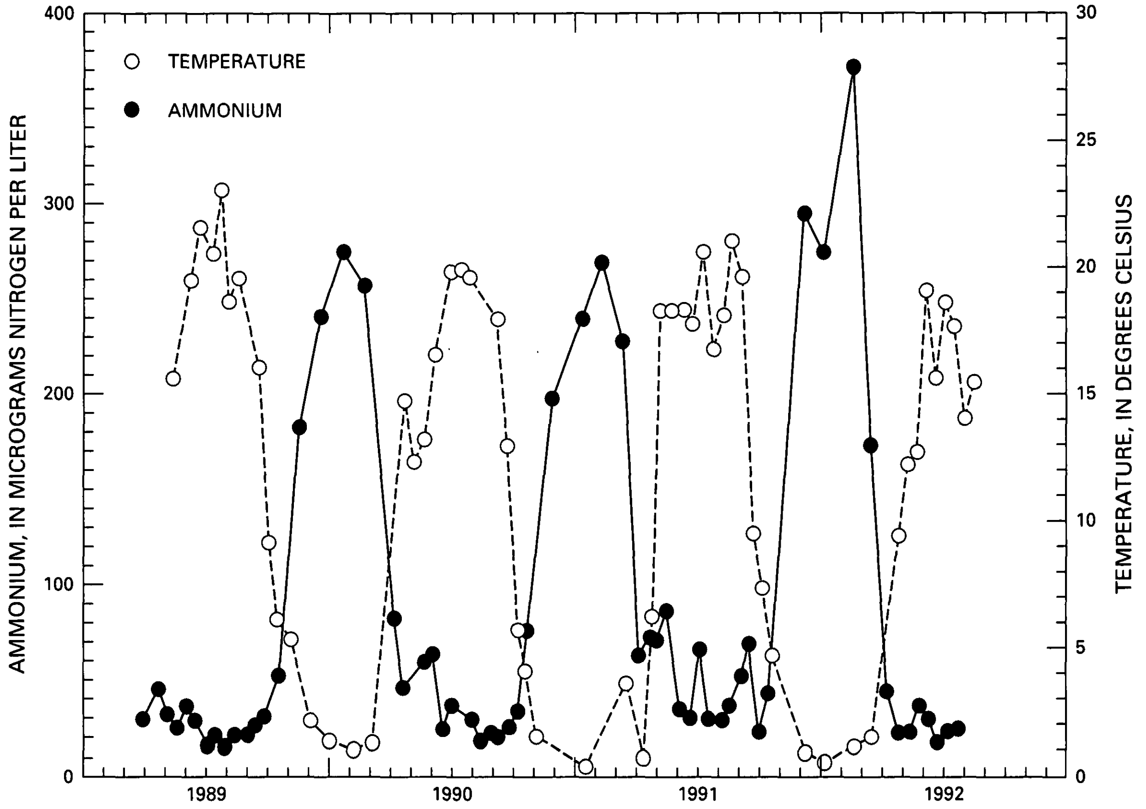

Figure 70. Temperature and ammonium concentrations in the upper Shingobee River, October 1992.

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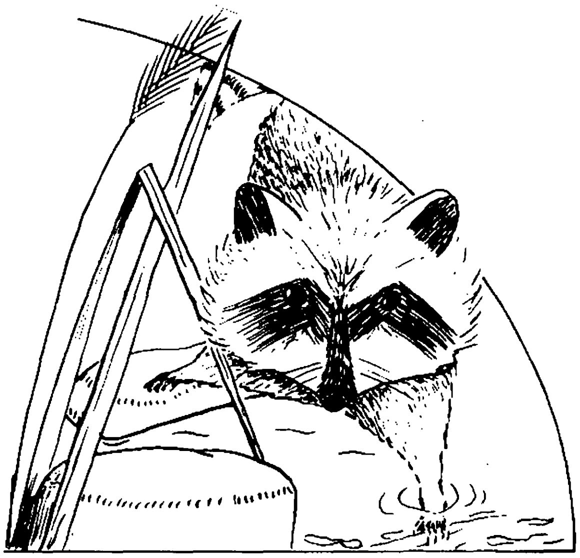

Physical and Chemical Properties of Waters at the Interface Between a Bankside Ground-Water Seep and the Channel of the Shingobee River: A Preliminary Analysis Under Base-Flow Conditions

By Frank J. Triska, John H. Duff, Ronald J. Avanzino, and Alan P. Jackman

IINTRODUCTION

Watersheds are classically viewed as tightly sealed catchments whose hydrologic output is related to the direct interception of atmospheric inputs and internal catchment processes. This conceptual model has been successfully used to estimate water and solute fluxes, especially in steep catchments of the northeastern (for example, Hubbard Brook, New Hampshire), southeastern (for example, Coweeta Hydrologic Station, North Carolina) and northwestern, (for example, H.J. Andrews Experimental Forest, Oregon) United States. In the northern tier of States, however, watersheds commonly lie on deep glacial deposits underlain by extensive regional aquifers. Thus, watershed outputs may not be solely limited to atmospheric inputs and processes within the topographic boundaries of catchments that comprise the headwaters of a river system. For example, along the Shingobee River between Steel Lake and Shingobee Lake (plate 1), a thin till layer underlying surficial sand has a higher surface altitude than the river (Winter and Rosenberry, this volume). The result is emergence of numerous springs and seeps along the banks and beneath the channel, which can significantly increase discharge within short river reaches (Jackman and others, this volume).

A major goal of the Interdisciplinary Research Initiative (IRI) is to assess the impact of regional ground-water flows on surface-water bodies including lakes, streams and wetlands. Understanding biogeochemical linkages between terrestrial and fluvial ecosystems first involves characterizing the origin, magnitude, and physical and chemical properties of various water sources as they cross the land/water interface and become surface water. The objective of our initial study was to characterize the physical and chemical properties of the ground-water/streamwater interface at one of the numerous seeps along the banks of the Shingobee River.

SITE, MATERIALS, AND METHODS

The study seep is a ground-water discharge that emerges approximately 4 m from the wetted channel of the Shingobee River. The seep is located at meter 260 in a 600 m reach described by Jackman, and others (this volume). Water was sampled from 13 piezometer nests and from the seep and stream surface (fig. 71). Piezometers were constructed of 0.25-inch stainless-steel tubing screened with 300 µm nylon mesh. At each sampling site piezometers were inserted to depths of approximately 30 cm, 60 cm, and 90 cm beneath the seep or channel surface. Many piezometers indicated an upward flux of ground water as evidenced by water flowing freely over the top.

Three to five volumes of water were peristaltically pumped prior to filling 50-mL biological oxygen demand (BOD) bottles, which were immediately sealed and returned to a field laboratory for determination of pH, dissolved oxygen, and specific conductance. Temperature was measured in the field. Separate water samples were pumped through 0.45-µm polysulfone membrane filters (Gelman GA6) into new, clean, 60 mL polyethylene bottles. Samples were frozen and returned to the laboratory for analysis of nitrate plus nitrite (cadmium reduction, Technicon Industrial Method # 158-71W, December 1972) and ammonium (automated phenol-hypochlorite

Technicon Industrial Method # 154-71W, February 1973) using a Technicon Autoanalyzer II. For DOC analyses, a separate sample was passed through 0.45 µm silver filters into acid-washed, oven-fired glass bottles then measured by persulfate wet oxidation (Menzel and Vaccaro, 1964). Absorbance at 254 nm was also determined on DOC samples. An extinction coefficient "e" was obtained by dividing absorbance by DOC concentration to provide an estimate of aromaticity. Finally, separate water samples were collected in 50-mL glass-stoppered bottles and analyzed for ferrous iron using a ferrozene reagent method (Stookey, 1970).

RESULTS AND DISCUSSION

Hendricks and White (1988), and White (1990) found that temperature gradients could be used to determine subsurface flow paths in sand-bottom channels such as the Shingobee River. The method is especially effective at summer base flow when a steep thermal gradient may exist between ground water and surface water. At the Shingobee River, surface-water temperature was 21.1°C compared to 8.8 °C at the seep surface (August 1990, fig. 72A). In subseep zones close to the site of emergence, water temperature was approximately 10°C (Sites A and B). Subsurface waters in zones dominated by seep input (Sites A-E) did not exceed 11°C except at Site A (30-cm depth). Interface sites between the seep and stream (Sites G, H and I) were approximately 11° C and varied over a narrow range. Subchannel water temperature (16.2-18.8° C at 30-cm depth) was higher than equivalent subseep locations (10.0-13.5° C). High subchannel water temperatures relative to ground-water temperature may indicate some mixing of surface and subchannel water and warming by insolation and heat convection to depths greater than 1.0 m since temperatures between 13 to 15° C were maintained even to 90-cm depths in late summer.

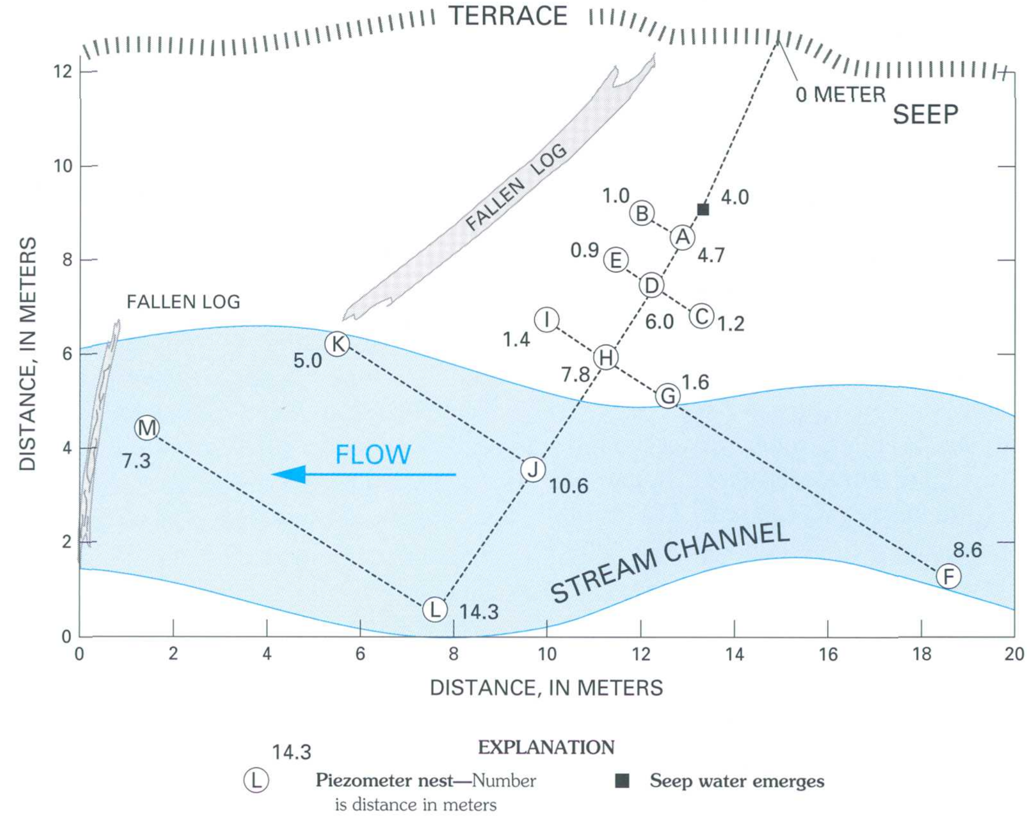

Figure 71. Plan view showing the relative position of a seep input and the location of piezometer nests relative to the seep, a bankside terrace (approx. 1.0 meter high) and the channel. Lettered locations indicate piezometer nests consisting of three samplers inserted to depths of 30, 60, and 90 centimeters into the sediments.

The pH at all sites varied over a narrow range (6.9-7.5) (fig. 72B) and was the only physical and chemical property that was not different between ground water and surface water. The pH was slightly lower at subseep than at subchannel locations, but no significant trend was observed.

Specific conductance was generally lower and more uniform beneath the seep than in subchannel or interface locations (fig. 72C). Specific conductance was lower in surface water than in subsurface locations at both the seep and stream. Subseep values ranged between 337 and 388 µS/cm. The range was greater between channel and subchannel locations (290-445 µS/cm). As with temperature, specific conductance at interface sites was generally intermediate between subseep (Site A) and subchannel (Site M) locations.

With two exceptions (Site C, 60 cm and Site G, 90 cm) highest DOC values were observed beneath seep locations A and B and in channel surface water (fig. 72D; George Aiken, written commun., August 1990). Concentration of DOC in most other subsurface waters (21 of 28 samples) ranged between 4.0 and 4.9 mg/L. Subchannel DOC values at Site L (a depositional area for particulate organic matter) were higher than at Site M which was located in well-washed sands toward the center of the channel. DOC in channel surface water (6.8 mg/L) was higher than in subchannel zones (3.9-5.9 mg/L). The absorbance at 254 nm was similar for all locations except for stream surface water and the 30-cm piezometer at Site L (fig. 73). This may indicate a different composition for DOC in surface water than in ground-water possibly, reflecting the presence of macrophyte biomass both within and adjacent to the channel.

Dissolved oxygen concentration was higher in channel surface water (6.8 mg/L) than at any other location (fig. 74A). Dissolved oxygen generally decreased with depth, and was less than 2.0 mg/L at most locations. The lowest concentration was at Site L at 30-cm depth (0.8 mg/L), the site of high organic matter deposition.

Ammonium concentration was low in channel surface water (17 µg N/L) compared to the seep and all subsurface waters. Subseep locations were generally lower than subchannel locations (fig. 74B). Highest ammonium concentration was at Site L (30-cm depth), the site of observed deposition of particulate organic matter. The opposite was true for nitrate. Although nitrate concentration was low in surface water (4.3 µg N/L), it was much higher than at subsurface locations where nitrate plus nitrite was virtually undetectable (typically less than 1.0 µg N/L, fig. 74C). Concentration data between interstitial water and channel surface water indicate a potentially large sink for dissolved inorganic nitrogen in the top 30 cm of channel sediments, which is related to decreasing dissolved oxygen concentration.

Ferrous iron concentration was much higher in subsurface than in surface waters, indicating the presence of reducing conditions in the subseep and subchannel zones. Iron oxyhydroxide coatings were observed on stream cobbles at many channel locations and on the surface sediments of bankside seeps. Coatings were also commonly observed on the roots and subsurface stems of aquatic macrophytes, indicating a balance between macrophyte photosynthesis (which delivers oxygen to subchannel sediments) and upwelling of deeper ground water either low or lacking in dissolved oxygen. The dissolved ferrous iron, like dissolved oxygen and dissolved inorganic nitrogen, indicate that the upper sediment zone (top 30 cm) may constitute a dynamic aerobic/anaerobic interface maintained by mixing of surface and subsurface waters and biotic processes associated with channel surface sediments. An intensive interdisciplinary approach linking hydrology, biology, and chemistry in the channel surface sediments is a high priority for future research.

SUMMARY AND CONCLUSIONS

Examination of physical and chemical properties of a seep/stream interface indicated differences between seep and stream water and concentration gradients with depth for many measured constituents.

1. Seep water temperature (8.8° C) was significantly lower than stream water (21.1° C) and, temperature of ground-water beneath the stream was higher than at equivalent depths beneath the seep. This may indicate some mixing of stream and substream waters or warming of the top meter of sediment by heat convection.

2. Specific conductance was higher in seep water than in channel surface water. Subsurface values of specific conductance were higher than surface water in both environments. Subchannel values of specific conductance were generally higher than subseep. The pH was similar in all locations.

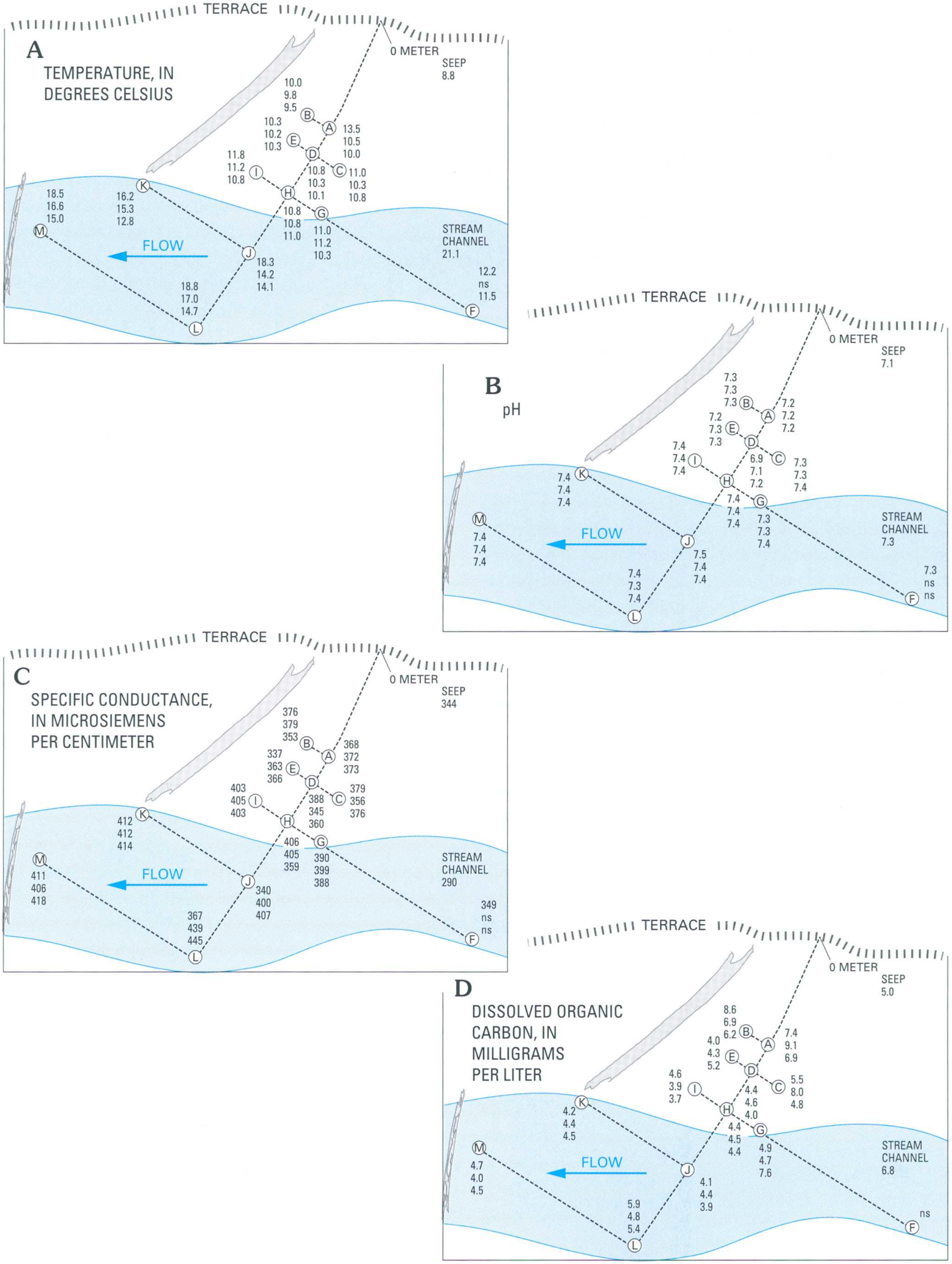

Figure 72. (A) Temperature, (B) pH, (C) specific conductance, and (D) dissolved organic carbon concentration in seep, channel, and subsurface water. Measurements were taken at 30 centimeters(top), 60 centimeters(middle), and 90 centimeters(bottom) beneath the sediment-water interface. Surface-water measurements are indicated beneath the seep and channel designations. ns = no sample.

3. DOC concentration generally was higher in channel surface water than in subchannel, whereas sites nearest the seep had the highest DOC. The DOC signature (absorbance 254) was different for channel water and one subchannel location than for other sites, possibly reflecting DOC generation associated with production or decomposition of macrophyte biomass.

4. Dissolved oxygen concentration was higher in channel water than seep water and generally decreased with depth beneath the sediments.

5. Ammonium concentration was very low in channel surface water and more than an order of magnitude higher in subchannel water. Water at the seep surface was almost an order of magnitude higher than channel surface water, and subseep water was higher than at the surface. Nitrate plus nitrite concentration was very low in surface water of both the seep and channel and was barely detectable at subsurface locations. The steep concentration gradient of nitrate plus nitrite and ammonium between surface and subsurface may indicate an active biogeochemical sink for nitrogen in the top 30 cm of channel sediment.

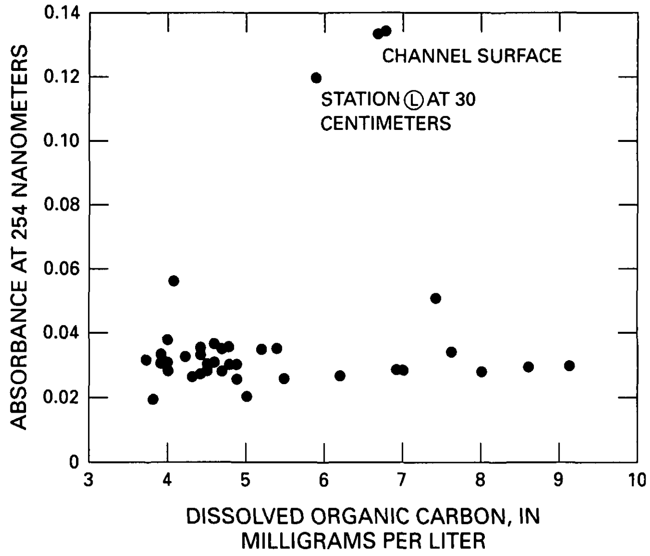

Figure 73. Relation between ultraviolet light absorbance at 254 nanometers and dissolved organic carbon concentration. Data provided by George Aiken (written commun. August, 1990).

ACKNOWLEDGMENT

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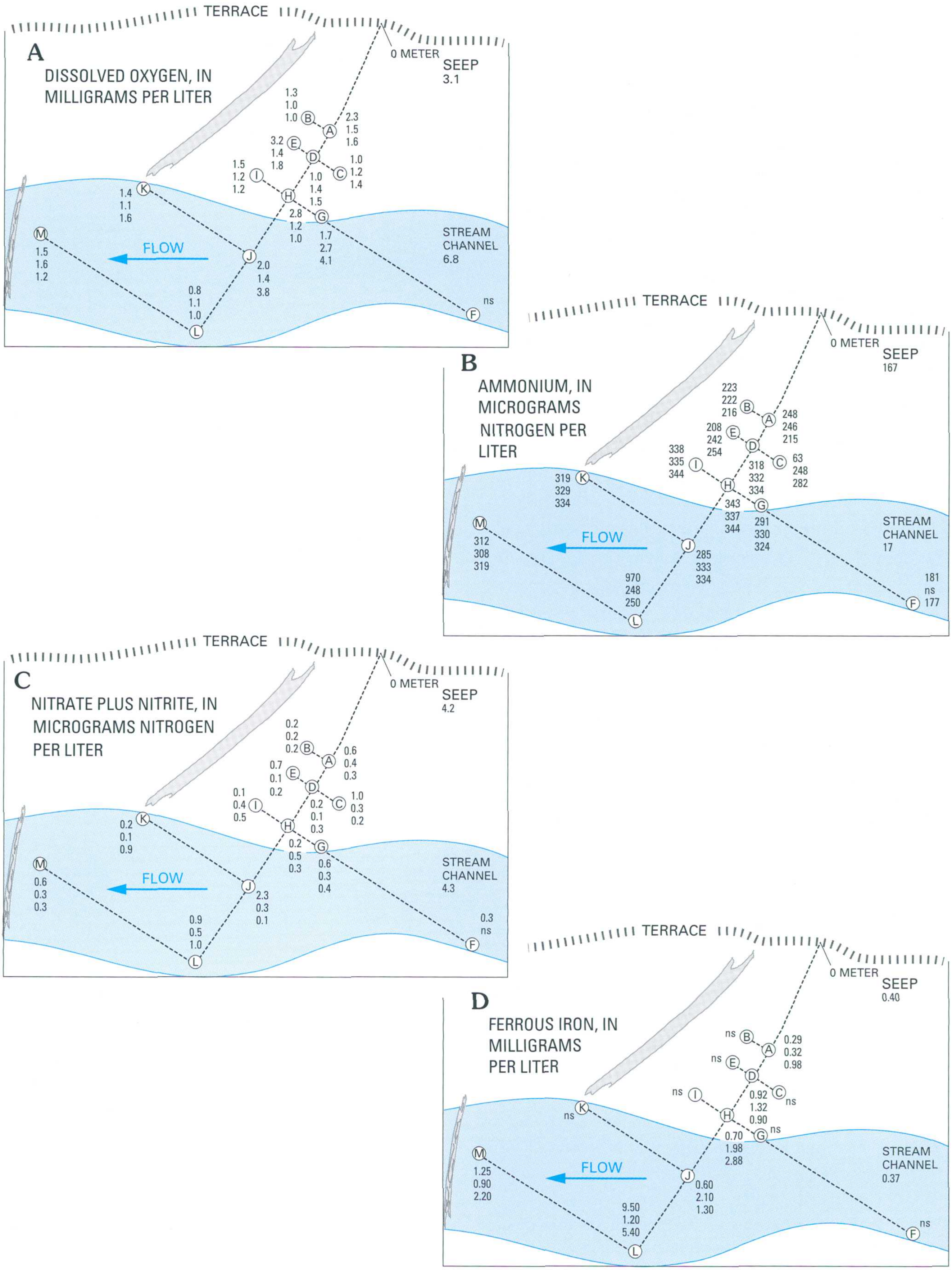

Figure 74. (A) Dissolved oxygen, (B) ammonium, (C) nitrate, and (D) ferrous iron concentrations in seep, channel, and subsurface waters. Measurements were taken at 30 centimeters top), 60 centimeters (middle) and 90 centimeters(bottom) beneath the sediment-water interface. Surface-water measurements are indicated beneath seep and channel designations. ns = no sample.

Sediment-Associated Nitrification and Denitrification Potentials at the Interface Between a Bankside Ground-Water Seep and the Channel of the Shingobee River.

By Frank J. Triska and John H. Duff

INTRODUCTION

A major goal of the Interdisciplinary Research Initative (IRI) is to determine the effect of regional ground water and associated solute transport on surface-water bodies including lakes, streams, and wetlands. Along the banks of fluvial environments, physical and biotic processes can determine the nutrient composition of surface water following ground-water transport through and beneath riparian zones. Recent evidence indicates that bankside processes in both agricultural and forested watersheds control DIN (dissolved inorganic nitrogen) concentrations (Peterjohn and Correll 1984, 1986; Lowrance and others, 1983, 1984, 1985; Schnabel, 1986; Triska and others, 1989a, 1989b; Pinay and Decamps, 1988; Pinay and others, 1990; Haycock and Pinay, 1993), especially through denitrification (Trudell and others, 1986; Duff and Triska, 1990; Pinay and others, 1990).

Hydrologic investigations along the Shingobee River between Steel and Shingobee Lakes (plate 1) indicate significant input of ground water by bankside seeps and ground-water discharge beneath the channel (Jackman and others, this volume; Duff and others, this volume). Ground-water/surface-water mixing results in physical and chemical gradients at the interface between the seep and stream (Triska and others; this volume), as exemplified by steep concentration gradients in ammonium or nitrate within the upper 20 cm of channel sediments (Duff and others, this volume). The observed concentration gradients may result from coupled nitrification and denitrification. Coupling means that the rate of denitrification is controlled by the rate of nitrate production. The cumulative effect can be a large biological sink for DIN in stream sediments, especially during summer. The objective of this preliminary study was to determine nitrification and denitrification potentials associated with saturated-zone sediments in bankside, seep, interface (seep-stream), and channel sediments.

SITE, MATERIALS, AND METHODS

Sediment cores were collected from the seep-stream habitat at 240 m of an 700 m reach of the Shingobee River between Steel and Shingobee Lake described by Jackman and others (this volume). In August 1990, cores were taken by hand using acrylic plastic tubing inserted as much as 30 cm into the sediments then extruded into polyethylene bags in two segments: shallow (0-10 cm) and deep (10-30 cm). Cores of saturated sediments were taken at six sites in four locations: (1) a bankside terrace, (2) the surface of a ground-water seep, (3) an interface between the seep and stream, and (4-6) in channel sediments (fig. 75).

To assay for nitrification potential, 20 mL of fresh sediment was slurried with 150 mL of stream water then amended with ammonium (1.0 mM NH_4Cl), phosphate (1.0 mM NaH_2PO_4) and bicarbonate (1.0 mM $NaHCO_3$) to prevent nutrient limitation. Aliquots of the water phase (30 mL) were removed immediately (0 hours) and after 25 and 50 hours. Samples were filtered through 0.45-µm cellulose triacetate membranes into 60-mL polyethylene bottles then frozen for subsequent nitrate analysis (cadmium reduction, Technicon Industrial Method # 158-71W December 1972). Sample flasks were cotton-stoppered and shaken continuously to maintain aerobic conditions. Flasks of stream water without

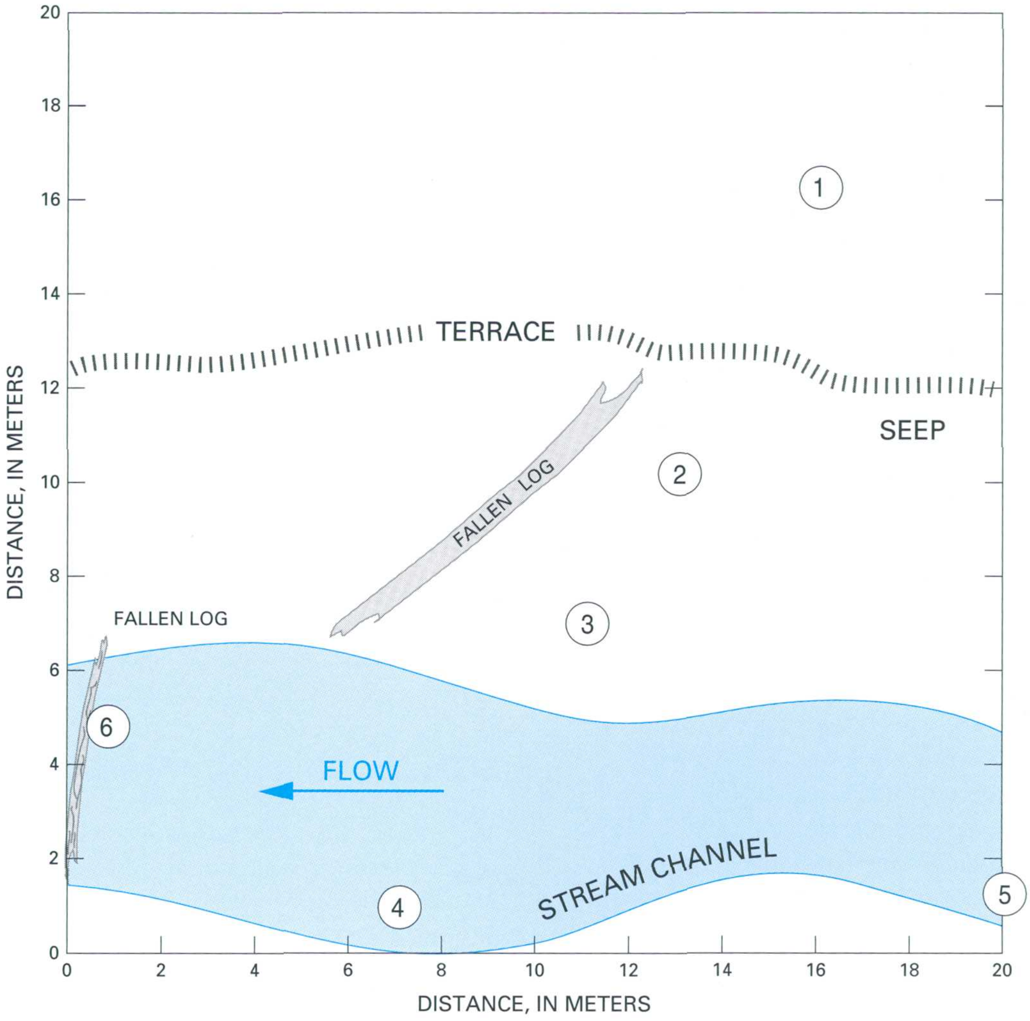

Figure 75. Location of bank (1), seep (2), seep/stream interface (3), and stream channel (4-6) sampling sites.

sediment served as a control for planktonic nitrification. In addition, sediment slurries from two sites were treated with nitrapyrin (10 µL of a 22 percent solution), an inhibitor of ammonium oxidation.

Denitrification was assayed by the acetylene block technique (Yoshinari and others, 1976) using sediments from separate cores from the same sites (fractionated into segments 0-5 cm shallow; 10-14 cm, deep). Four mL of sediment was placed into clean glass scintillation vials to which was added 6 mL of filtered stream water. The vials were stoppered, taped, and flushed with oxygen-free nitrogen for 15 minutes. Then 1.5 mL of $C_2 H_2$ generated from calcium carbide was added to the headspace. Nitrate (1.0 mM $NaNO_3$) and carbon (1.0 mM glucose) were added to all flasks to prevent nutrient limitation, and chloramphenicol (4 mg/mL) was added to inhibit protein synthesis. Flasks were shaken and the headspace sampled immediately and after approximately 8, 18, and 28 hours. Formation of N_2O was measured by electron-capture gas chromatography.

RESULTS AND DISCUSSION

Nitrification

Nitrate increased in all noninhibited sediments over the 50-hour incubation period, but the rate of nitrification was highly variable among the sites (fig.

76). Saturated bank sediments (Site 1) had the least nitrification potential with nitrate increasing from a mean initial concentration of 22 µg N/L to 95 µg N/L after 50 hours. The highest nitrification potential was found in shallow sediments near the site of ground-water emergence (Site 2), where sediments were continually bathed with high concentrations of ammonium (Triska and others, this volume). Nitrate increased from a mean initial concentration of 28 µg N/L to 3,070 µg N/L after 50 hours. The rate of nitrate formation decreased between 25 and 50 hours at Site 2 possibly due to resource limitation, whereas the rate of nitrate formation in sediments from the stream channel (Sites 4-6) and interface (Site 3) accelerated between 25 and 50 hours, possibly due to microbial growth within the flasks. Nitrate concentrations in slurries of stream sediments clustered between 1,200 and 1,600 µg N/L (50 hours) and were distinct from interface, bank, and seep sediments. Flasks treated with nitrapyrin and streamwater controls did not exhibit significant nitrate formation. Calculated mean nitrification rates were 0.14 µg N/cm^3 sediment/hour for stream surface sediments (0-10 cm, Sites 4-6) and 0.09 µg N/cm^3 sediment/hour for subchannel sediments (10-30 cm), based on the 25-hour incubation. Assuming that most nitrification occurred in the top 20 cm of channel sediment, the nitrification potential was 23 mg N/m^2 streambed/hour.

Denitrification

Denitrification, measured as nitrous oxide production, occurred in sediments from all locations (fig. 77). The highest denitrification potential occurred in the samples from the site of ground-water emergence in the bankside seep (Site 2) where nitrification potential was highest; denitrification potentials were higher in surficial (0-5 cm) than in deep sediments (10-14 cm). In the channel, denitrification potential was lower at 0-5 cm depths than in sediments from depths of 10-14 cm. Because nitrate is barely detectable in interstitial water below the streambed (Triska and others, this volume), denitrification may be controlled by NO_3 production.

Concentration of DIN was less than 25 µg N/L in channel surface water and occurred almost exclusively as ammonium. Concentration of ammonium was approximately 325 mg N/L in ground water (Triska and others, Duff and others, this volume) but could exceed 3,000 µg N/L in shallow, organic-rich sediments (Duff and others, this volume). The source of ammonium concentrations <350 mg N/L was probably ammonification associated with the decomposition of particulate organic matter.

Inputs of ground water based on chloride dilution studies (Jackman and others, this volume) were combined with DIN concentration profiles (Triska and others, Duff and others, this volume) and the calculated nitrification-denitrification potentials (this study) to assess the potential for a nitrification-denitrification couple to deplete DIN from ground-water discharged to the Shingobee River. Mean ground-water input to the reach was calculated to be 25 L/m^2 streambed/hour. The concentration of ammonium in ground water was 325 µg. N/L, resulting in a calculated ground-water nitrogen input rate of 8.1 mg N/m^2/hour. In the top 20 cm of channel sediments the nitrification potential was estimated to be 23 mg N/m^2/hour and the denitrification potential was estimated to be 9.8 mg N/m^2/hour. The nitrification potential was sufficient to oxidize any ammonium entering surface sediments from ground water, and the denitrification potential was sufficient to reduce any nitrate generated within upper channel sediments. Thus, coupling of nitrification and denitrification may be an important regulator of surface-water DIN concentration in summer and may cause the low DIN concentration in surface water.

Future research will include seasonal assays of nitrification and denitrification potential to determine the role of temperature on the coupling of these processes. Assays will be conducted using natural communities under in situ conditions of temperature, oxygen, and nutrient concentration. Interstitial water chemistry of subchannel sediments will also be examined seasonally to determine intraannual variability of ammonium, nitrate, dissolved organic carbon, and dissolved oxygen, solutes which regulate the rate of nitrification and denitrification.

ACKNOWLEDGMENT

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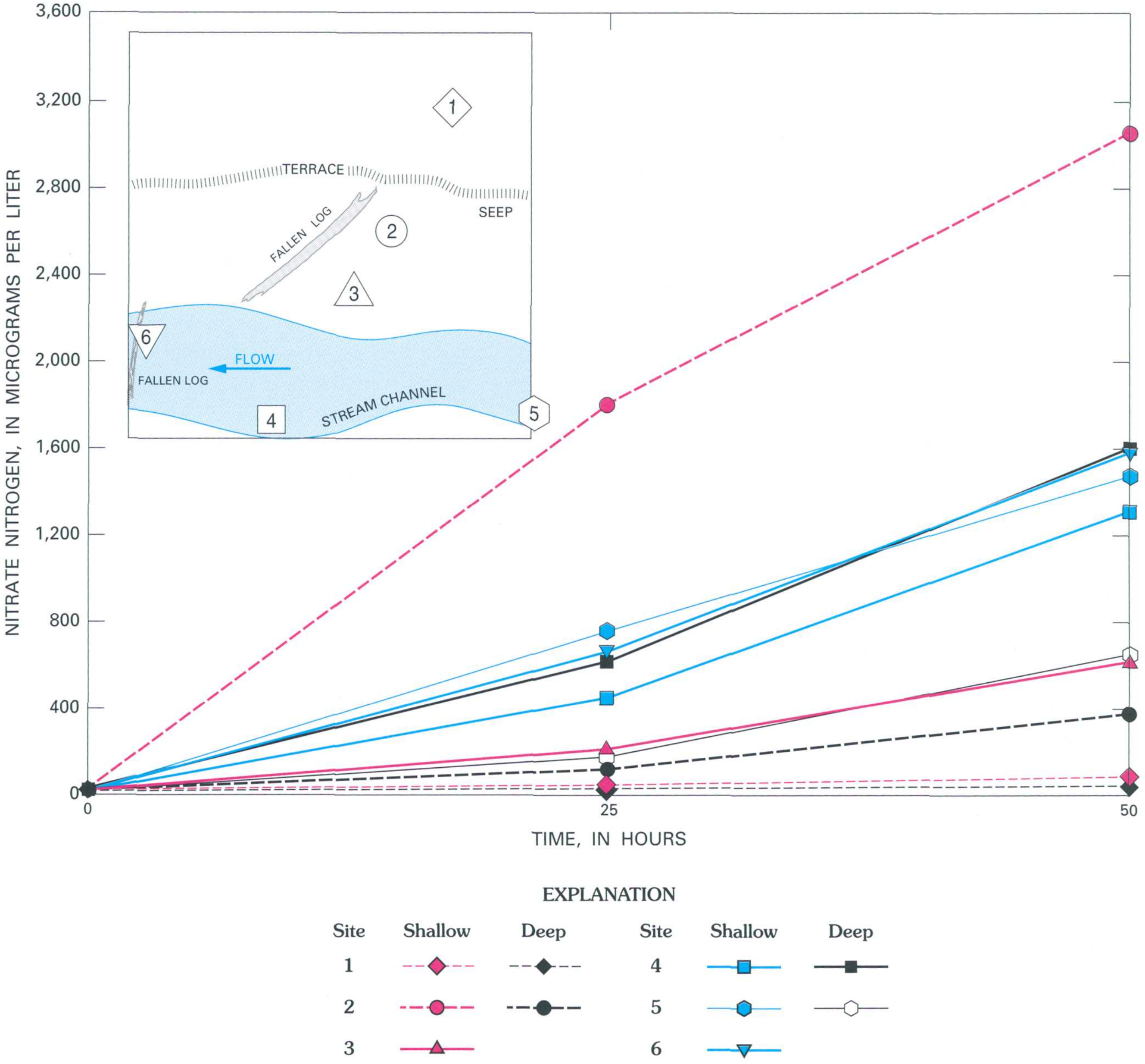

Figure 76. Nitrate-nitrogen production in sediment slurries from the bank (1), seep (2), seep/stream interface (3), and stream channel (4-6) sampling sites. [Shallow = 0-10 cm; Deep = 10-30 cm.]

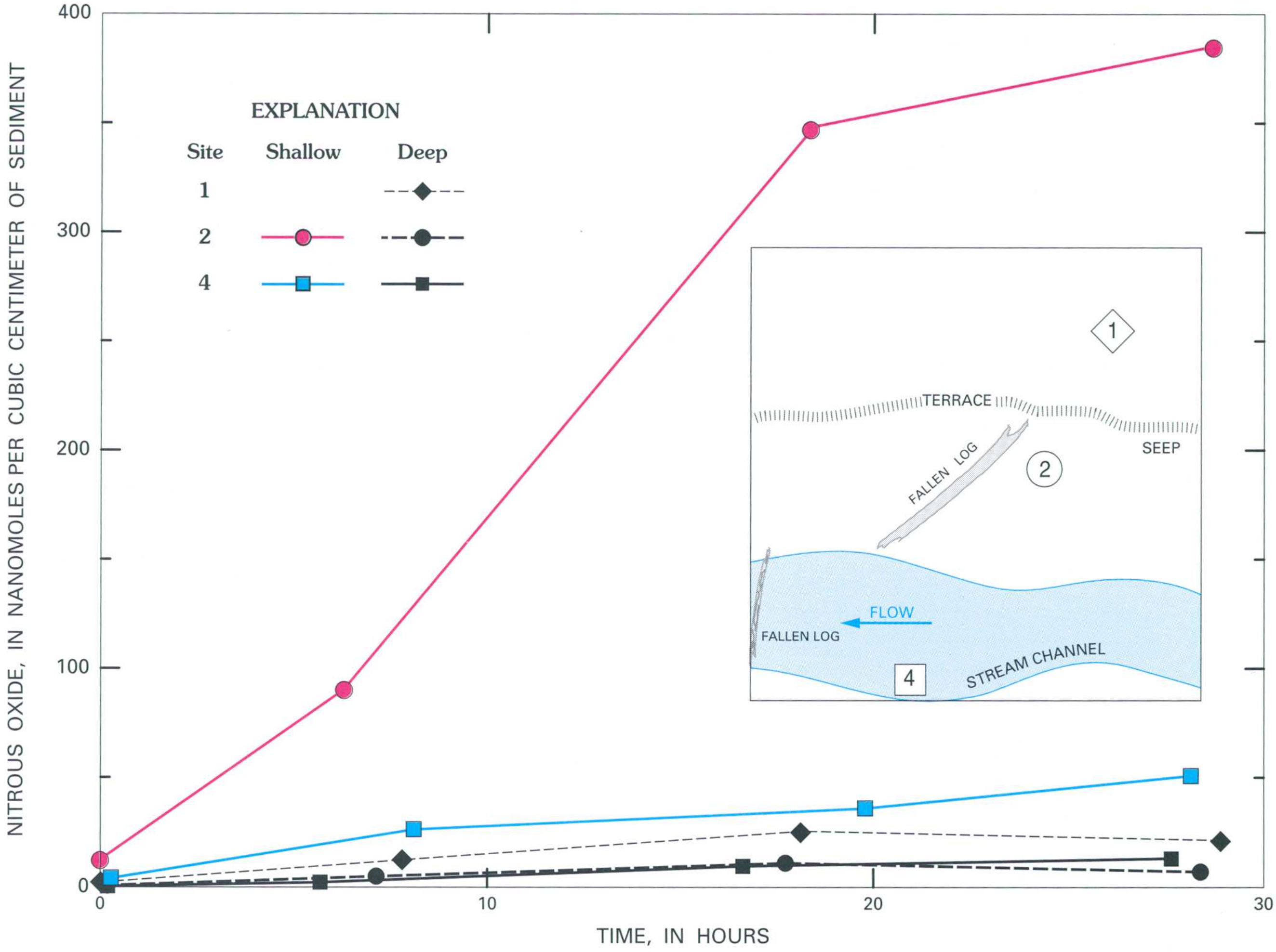

Figure 77. Nitrous oxide production of sediment slurries from the bank (1), seep (2), and stream channel (4) sampling sites. [Shallow = 0-5 cm; Deep = 10-14 cm.]

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Vegetation of Little Shingobee Fen

By Virginia Carter, Patricia T. Gammon, and Nancy B. Rybicki

INTRODUCTION

Little Shingobee Fen surrounds Little Shingobee Lake, which drains northward into Shingobee Lake (pl. 1). It is a typical northern fen with organic soils underlain by marl and lake sediments. A fen is a minerotrophic wetland deriving its water from relatively mineral-rich ground and (or) surface water and typified by a pH greater than 6 and dissolved calcium concentrations greater than 350 µeq/L (Gorham and others, 1985).

The basic hypothesis underlying this study is that the distribution and composition of fen vegetation is controlled by hydrologic and geochemical processes within the wetland. The purpose of this research is to document and characterize the present distribution of vegetation and to correlate these data with wetland hydrology, geochemistry, and stratigraphy. Results will also be used, in conjunction with information on other vegetation in the watershed, to investigate carbon cycling in the Shingobee Lake watershed.

METHODS

Fen vegetation was sampled using quadrat and general survey methods. Trees within the forested wetland were sampled in seven 10-m^2 quadrats. These, in turn, were subdivided into four 5-m^2 quadrats, two of which were selected for the shrub survey. Five 1-m^2 quadrats for herbaceous plants were located randomly within the 10-m^2 quadrats. Two 5-m^2 quadrats were established in the emergent wetland surrounding the lake, and several 1-m^2 quadrats were placed in the open fen within the forested section. All species noted while working in the fen for any purpose were identified in order to have an inclusive species list. Vegetation sampling was conducted in June 1992 and August 1991, 1992, and 1993. Plants were identified to species where possible using Voss (1972, 1985), Newcomb (1977), Gleason (1968), and Fernald (1970).

RESULTS

A vegetation map of the fen shows the major vegetated zones (fig. 78; table 24). The forested part of the fen is dominated primarily by tamarack and black spruce. The water table lies at the land surface in the tamarack-spruce zone; but on the narrow, north-south ridge where the soil is drier, the dominant species are balsam fir and paper birch. Shrubs comprise a minor part of the understory in the tamarack-spruce zone but are more prevalent in the drier fir-birch zone. The ground cover in the tamarack-spruce zone is composed of sphagnum and other mosses, small sedges, and a sparse covering of other herbaceous species; there is relatively little sphagnum in the fir-birch zone, and the ground cover is primarily a mixture of herbaceous species. A large alder-dominated zone with an understory of assorted mosses and herbaceous species lies at the south end of the fen. Both the grass-sedge zone ringing the lake and the open fen are dominated by grasses and sedges. Cattails, wild rice, soft-stemmed bulrush (*Scirpus validus*) and water lily (*Nymphaea tuberosa*) form a floating-leafed zone on the edge of the lake.

The fen contains insectivorous species such as pitcher plants (fig. 79A) and sundews (*Drosera rotundifolia*). Several species of orchids are fairly abundant, including the small yellow lady slipper (*Cypripedium calceolus* var *parviflorum*) (fig. 79F), the pink lady slipper (*Cypripedium acule*), the showy lady slipper (Minnesota's State flower)(fig. 79E), and the small round-leafed orchis (*Orchis rotundifolia*) (fig. 79B). Other wetland species with showy flowers include buckbean (fig. 79C) and wild calla (*Calla palustris*) (fig. 79D) and species with small but unusual flowers such as the naked miterwort are also found in the fen (fig. 79G).

Water levels in Little Shingobee Lake are controlled by beaver dams on the outlet stream. In years of greater than normal rainfall or beaver activity, water levels are high in the lake and surrounding grass-sedge zone. Trees and shrubs that become established in this zone during periods of low water are killed by these high water levels, thus maintaining the herbaceous assemblages in this zone. Zonation of vegetation in the forested fen probably is controlled by ground-water levels. Future data on the hydrology and stratigraphy of the fen will aid in interpretation of these vegetation patterns. Conversely, the vegetation patterns can be used to guide the placement of wells and stratigraphic cores.

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Table 24. Summary of dominant species in the five vegetated zones in the Little Shingoibee Fen

[Herbs and shrubs include all species with percent cover equal to or greater than 20 percent; taxonomy follows Fernald (1970)]

		Zone				
Species	**Common name**	**Fir-birch**	**Spruce-tamarack**	**Alder**	**Open fen**	**Grass-sedge**
Trees and shrubs						
Picea mariana	black spruce	X	X			
Populus tremuloides	aspen	X				
Larix laricina	tamarack		X			
Abies balsamea	balsam fir	X				
Betula papyrifera	paper birch	X				
Fraxinus nigra	ash	X				
Alnus rugosa	alder	X		X		
Ledum groenlandicum	Labrador tea		X			
Acer spicatum	mountain maple	X				
Lonicera canadensis	fly honeysuckle	X				
Salix sp.	willow			X		
Sedges and grasses						
Carex sp.	sedge		X	X	X	
Carex leptalea	sedge		X			
Carex scoparia	sedge			X		
Carex prairea	sedge	X				
Phalaris arundinacea	reed canary grass					X
Calamogrostis canadensis	blue-joint grass					X
Zizania aquatica	wild rice					X
Carex stricta	tussock sedge					X
Carex bebbii	Bebb's sedge					X
Carex lacustris	sedge					X
Calamagrostis neglecta	reed-bentgrass				X	
Carex comosa	bearded sedge				X	
Eleocharis sp.	spike rush				X	
Herbs and mosses						
Sarracenia purpurea	pitcher plant		X			
Menyanthes trifoliata	buckbean		X			
Cornus canadensis	bunchberry	X				
Coptis groenlandica	gold thread	X				
Mitella nuda	naked miterwort	X				
Clintonia borealis	bluebead lily	X				
Aralia nudicaulis	wild sarsaparilla	X				
Maianthemum canadensis	wild lily of the valley	X				
Viola papiloneacea	common violet	X				
Thalictrum sp.	meadow rue	X				
Aster macrophyllus	large-leaved aster	X				
Typha angustifolia	narrow-leafed cattail					X
Typha latifolia	common cattail					X
Circaea alpina	dwarf enchanters' nightshade				X	
Cypripedium reginae	showy lady slipper			X		
Sphagnum and other mosses	mosses	X	X	X	X	

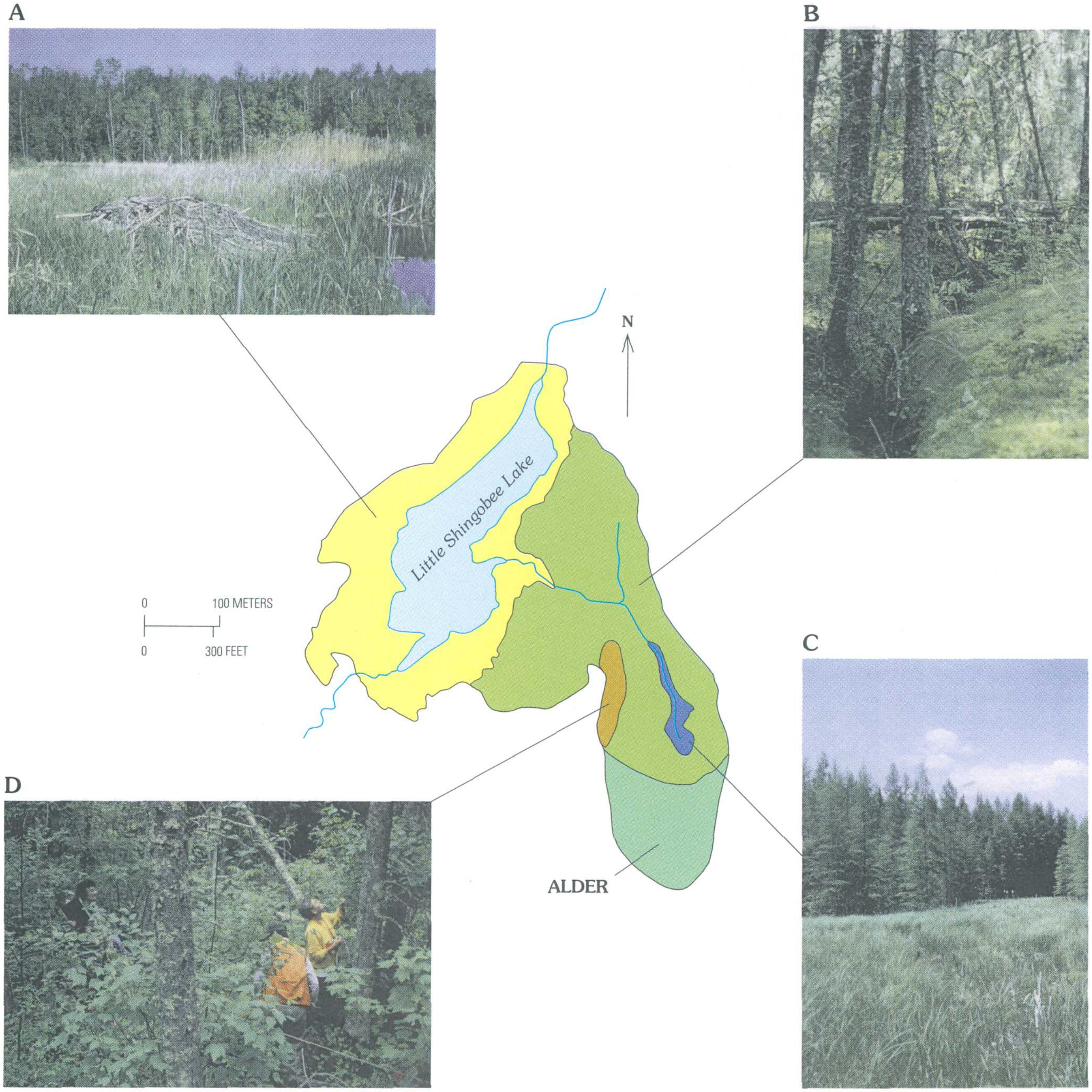


Figure 78. Major vegetated zones of Little Shingobee Fen. A is the grass-sedge zone. Tussock sedge and blue-joint grass comprise most of the foreground vegetation with reeds (*Phragmites australis)* and cattails in the background. B is the tamarack-spruce zone. This small stream drains westward into Little Shingobee Lake. The understory here is primarily sphagnum and Labrador tea. C is the open-fen zone surrounded by tamarack; sedges are dominant in the foreground. D shows the understory in the fir-birch zone which is dominated by the shrubs, mountain maple and fly honeysuckle.

Figure 79. Selected herbaceous species: A. pitcher plant; B. small round-leafed orchis; C. buckbean; D. wild calla; E. showy lady slipper; F. small yellow lady slipper; G. naked miterwort.

Ground-Water Surface-Water Interactions and Biogeochemistry of Little Shingobee Fen

By Larry J. Puckett, Virginia Carter, Donald O. Rosenberry, and Dennis A. Merk

INTRODUC TION

Fens are wetlands having organic soils and mineral-rich ground and (or) surface water. The pH of fens characteristically is greater than 6.0, and they have larger dissolved nutrient and mineral concentrations and greater vegetative diversity than bogs, which are commonly dependent on precipitation inputs for much of their nutrients (Gorham and others, 1985). Fens have distinct plant communities, the development of which are influenced by the hydrology and dissolved mineral and nutrient content of the hydrologic resource. It is important to understand both the chemistry and hydrology of the ground-water, surface-water, and precipitation inputs to understand the biogeochemistry of fens.

STUDY AREA

Little Shingobee Fen (fig. 80) appears to receive a large amount of its water from ground water. Its water has dissolved calcium concentrations exceeding 3,500 µeq/L, which is approximately 10 times the 350 µeq/L value given by Gorham and others (1985) as being typical of fens. Soils in the area are developed on and underlain by outwash and glacial drift associated with the Itasca moraine (Winter and Rosenberry, this volume). Mineralogy of the glacial deposits is poorly characterized but is known to consist of a mixture of quartz, potassium feldspar, plagioclase, and carbonates (Seaberg, 1985). The Itasca moraine has been reported to contain as much as 20 percent carbonate (Norton, 1982).

In contrast to other areas in the Little Shingobee Lake watershed, the fen is saturated to near the surface throughout the growing season and contains the only perennial stream (fig. 81A) feeding the lake, The fen, located southwest of Little Shingobee Lake, consists of a 4.82-ha grass and sedge zone (fig. 81B) that surrounds the lake and a 6.12-ha tamarack and black spruce forest with a ground cover of sphagnum and other mosses. The southeastern corner of the fen contains an 0.26-ha open area with a floating mat of grasses and sedges and a narrow stream channel (fig. 81C) that grades into tamarack and black spruce and, further south, into a 1.58-ha alder swamp. West of the open fen is a 0.32-ha fir and birch stand. To the east, a ridge covered by a mixed-hardwood upland forest separates the fen from the Shingobee River. the fen begins at a break in slope at the base of the ridge where numerous seeps are located.

The fen is underlain by layers of peat, peat and marl, or marl with scattered thin layers of sand and silt; the total thickness is greater than 10 m in the open area of the fen. In the study area, the thickness of the peat and marl layers is about 9 m near the boundary between the tamarack-spruce zone and the grass-sedge zone to a point about 120 m upslope, after which it decreases in thickness until it pinches out at the upland border (Carter and others, this volume). The material directly underlying the peat is sand and silt. The upland soil appears similar in composition to the fine-to-medium sand that overlies a gray till on the east side of the shingobee River (Winter and Rosenberry, this volume). If the stratigraphy is similar on both sides of the Shingobee River, then the sand-till contact could explain the presence of the seeps at the fen-upland border.

Much of the flow for the perennial stream appears to originate in the seep zone and the open area of the fen. Throughout the tamarack-spruce zone are small pools 1-2 m across (fig. 81E) where ground water discharges directly to the surface. These pools are characterized by precipitates of iron oxides that result when the reduced iron in ground water is oxidized by iron-oxidizing bacteria (Robbins and others, 1992).

Field observations suggest that large amounts of ground water are discharged in the fen, and preliminary water budgets indicate that precipitation recharge in the surrounding uplands is inadequate to account for the amount of water leaving the fen. Based on the assumed ground-water source of water entering the fen and the large dissolved concentrations of solutes in fen waters, this study was designed to test two hypotheses: (1) the chemistry of the ground water discharging into the fen is controlled by dissolution-precipitation reactions along the flow path, and (2) the fen is supported primarily by ground water thar originated either as seepage from the Shingobee River or from a regional ground-water flow path. This study contributes to the hydrologic and biogeochemical framework for other studies of vegetation and carbon cycling in the fen and the Shingobee watershed.

METHODS

Nests of piezometers (fig. 81D) were installed at points along a transect approximately 200 m long that began in the grass and sedge zone adjacent to Little Shingobee Lake and extended to the fen-upland forest border (fig. 82). The transect, installed in June 1991, is located about 10 m north of and parallel to the perennial stream. Piezometers constructed of 1.9-cm-diameter PVC pipe were installed in nests of three or four at depths ranging from 0.5 to 10 m. At each site

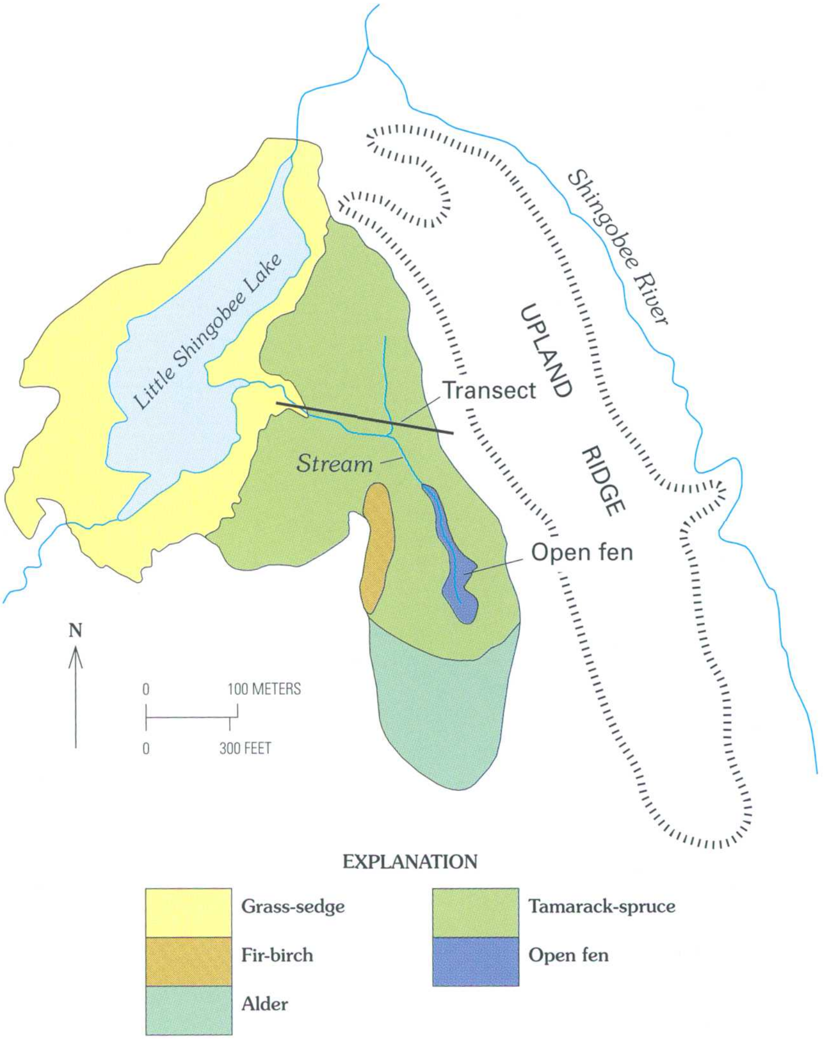


Figure 80. Location and vegetation zones of the Little Shingobee Fen and surroundings.

piezometer 1 was screened throughout the upper 0.5 m, piezometer 2 was screened at 1 m below land surface, piezometer 3 was screened at about 2-3 m below land surface, and when present, piezometer 4 was installed and screened as deeply as possible. Surface-water collection sites were also established on the Shingobee River, on the perennial stream draining the fen, and in Little Shingobee Lake. During June through October 1991, water samples were collected monthly for chemical analyses of major ions and water levels in piezometers were measured weekly (fig. 83).

Water-chemistry speciation was calculated using the WATEQFP program as implemented in the NETPATH program (Plummer and others, 1991). Net geochemical mass-balance reactions were modeled along hypothesized ground-water flow paths using the NETPATH program. Chemical analyses of water from several wells located in the Shingobee area (M. M. Reddy, U.S. Geological Survey, written commun., 1993), the Shingobee River, and precipitation (Reddy and others, this volume) were used as initial input values for the various flow-path models tested.

RESULTS AND DISCUSSION

Median ground-water levels for the piezometers along the transect are shown in figure 83. The piezometer at Site 2 is located on the upland slope about 1 m in elevation above the seep zone, and the water level in it probably indicates the water table. In the piezometers in the fen near the upland edge (Site 3), there was a strong upward ground-water gradient of approximately 0.36 m/m. There was a similar but much smaller upward gradient of 0.07 m/m between the deep and shallow piezometers in the nest at Site 4. At the midslope piezometer nests (Sites 5-6), there were downward gradients of 0.002 and 0.237 m/m respectively. The deepest piezometers in the grass and sedge zone (Sites 7-8) showed upward gradients of about 0.311 and 0.101 m/m, respectively, relative to the shallowest piezometers at the same locations.

The observed upward gradients in the piezometers indicate a strong ground-water-discharge potential under the fen. This upward ground-water gradient probably accounts for the large number of seeps that discharge directly through the peat into small pools. It is likely that the upward ground-water gradient was not observed in the midslope piezometers because they were completed in the upper one-fifth of the peat and were responding to local surficial flow paths in the peat.

It is hypothesized that the difference in saturated hydraulic conductivity of the peat and the coarser sandy material below is sufficient to develop the observed upward gradients. The seepage pools scattered throughout the fen might occur where the peat is breached by a column of ground water following a path to the surface that has existed since the formation of peat in the fen began at the end of the last glaciation. In areas where the peat is thin, such as along the upland border, the difference in hydraulic conductivity may not be large enough to significantly impede the upward movement of ground water.

Because water chemistry varied more between sample dates and sites than between sample depths, the data for all depths at each piezometer nest were combined and medians of all depths at each nest are given in table 25 and are shown in figure 84. Primary solutes in water sampled from the piezometers and surface waters of the fen were calcium, magnesium, and bicarbonate; the pH (pH 7.3-8.4) was in the range expected of water in equilibrium with calcite and (or) dolomite (table 25). Results of the WATEQFP equilibrium calculations indicate that fen waters were slightly oversaturated with respect to calcite and dolomite and are consistent with the hypothesis that water chemistry is dominated by equlibrium reactions between ground water and the carbonate that comprises a portion of the Itasca moraine (Norton, 1982).

As stated previously, the waters entering the fen are predominantly calcium and magnesium bicarbonate, with smaller concentrations of other ions generated through geochemical weathering reactions (table 25). Water temperatures were generally less than 10°C in the fen piezometers and greater than 15°C in the Shingobee River, Little Shingobee Lake, and the stream draining the fen (fig. 84A). A decreasing trend in pH was observed in piezometer nests 2-7 (Sites 3-8); values were largest in the small stream and Little Shingobee Lake (Sites 9-10) (fig. 84B), probably as a result of carbon dioxide degassing. With the exception of piezometer 2 (Site 3), calcium concentrations were near 3,500 μeq/L throughout the fen and were smallest in the Shingobee River and Little Shingobee Lake (fig. 84C). Bicarbonate concentrations were greatest in fen waters. Calcium and bicarbonate concentrations are probably controlled by calcite precipitation in these slightly oversaturated waters.

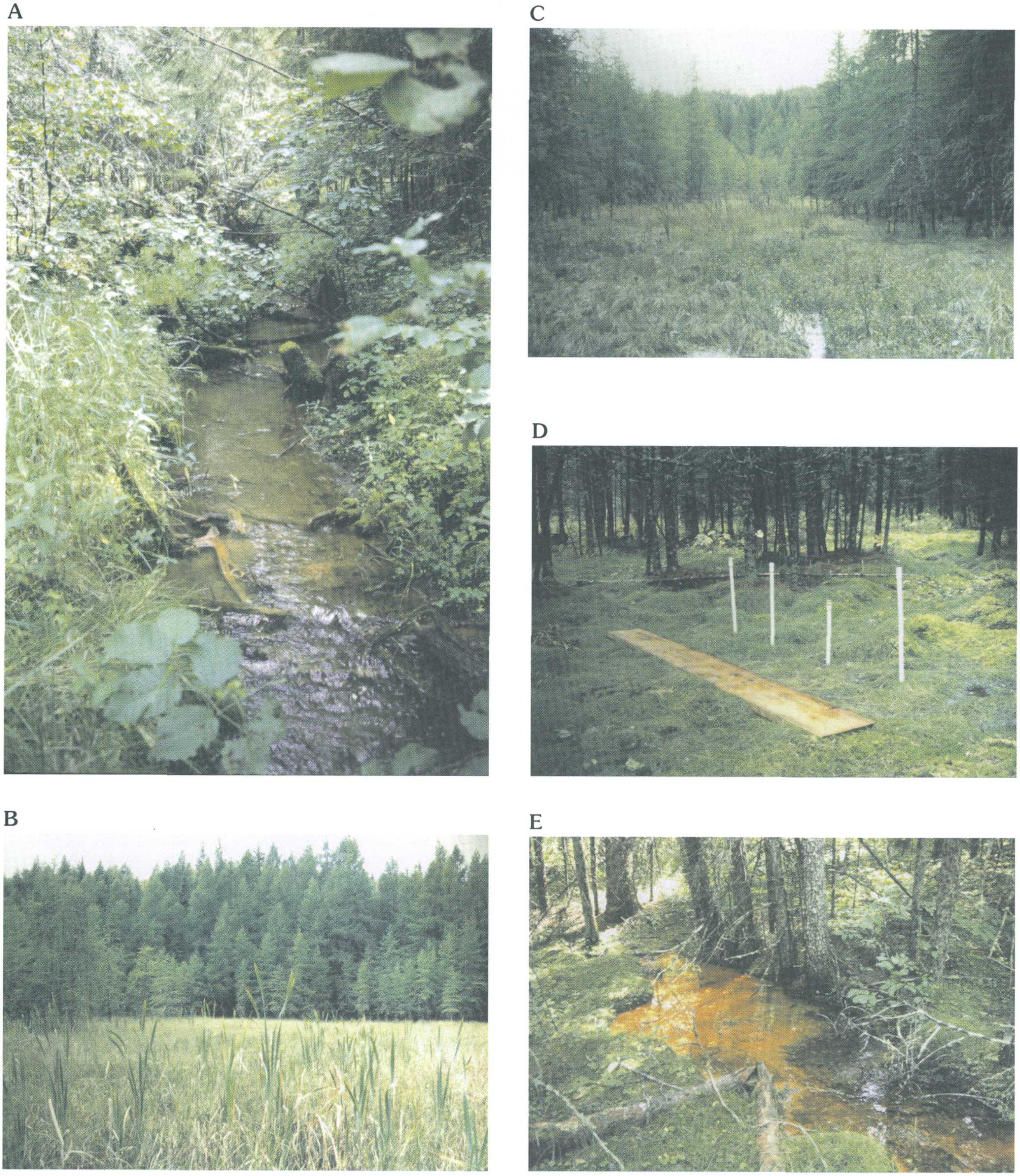


Figure 81. (A) The small stream draining the fen; (B) the grass and sedge zone around Little Shingobee Lake with the tamarack-spruce forest in the background; (C) the open area in the fen; (D) a piezometer nest in the tamarack-spruce forest; (E) one of the seepage pools in the fen.

Magnesium concentrations were largest at piezometer nest 3 (Site 4) then generally decreased in downslope piezometer nests (fig. 84E). Concentrations of silica were essentially the same in piezometer nests 1-5 (Sites 2-6) but then increased markedly in nests 6 and 7 (Sites 7-8), suggesting an influx of ground water with more silica from geochemical weathering reactions (fig. 84F). Potassium concentrations were highly variable and were smallest at Site 2, the upslope piezometer (fig. 84G).

Sodium and chloride concentrations were smallest in the upslope piezometer (Site 2) and more than doubled in the downslope piezometer nests (Sites 5 and 6). This pattern in sodium and chloride concentrations suggested evaporative concentration along the flow path through the peat layer followed by an influx of ground water near piezometer nest 7 and the lake (Sites 8-10) that had not undergone as much evaporation (fig. 84H-I). Nitrate concentrations were small and variable and were not significantly different among any of the piezometer nests (fig. 84J). Sulfate concentrations in piezometer nests 3-7 (Sites 4-8) were negligible relative to nests 1-2 (Sites 2-3), suggesting sulfate reduction may have occurred along the flowpath (fig. 84K).

Of the hypothetical flow-path models tested using the NETPATH program, the most realistic is based on the assumption that infiltration of precipitation in the surrounding uplands and seepage losses from the Shingobee River or a similar regional ground-water flow path are the primary sources of ground water. This model required approximately 34 percent Shingobee River water and 66 percent precipitation; dissolution of 219 μmol of albite, 569 μmol of

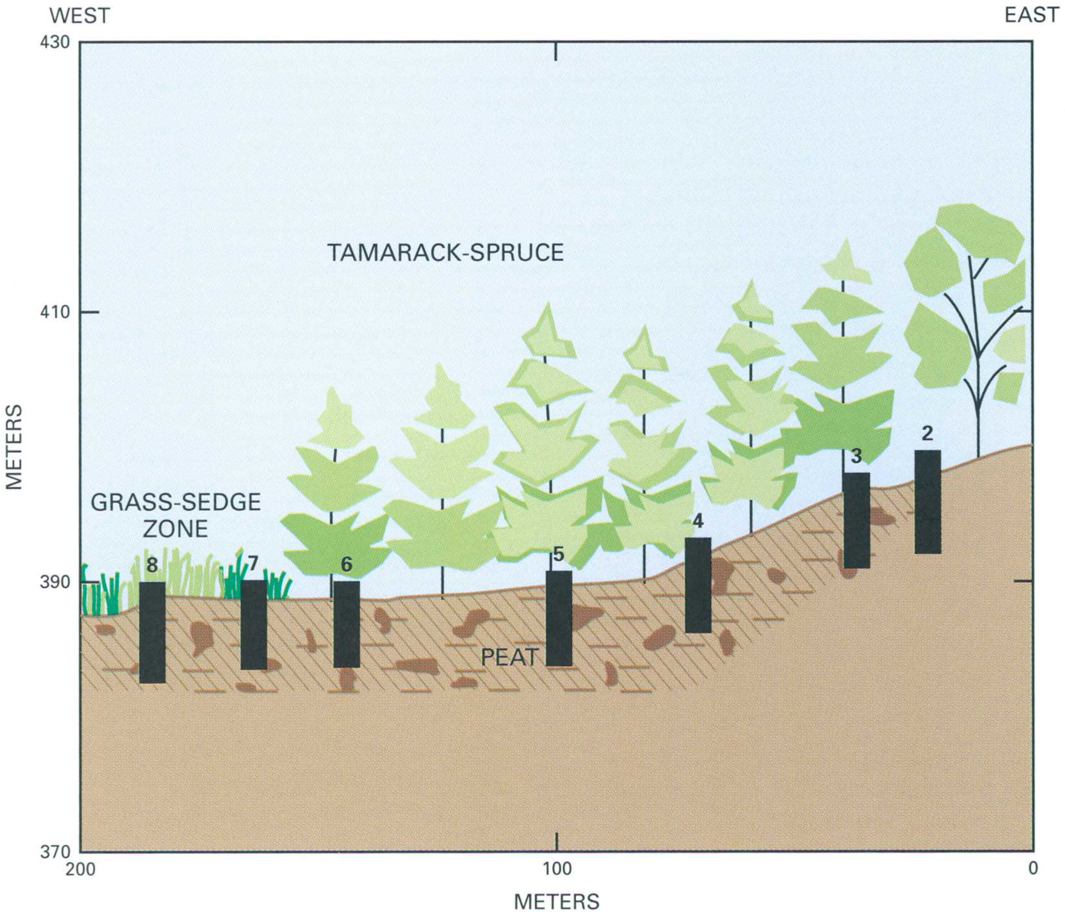


Figure 82. Little Shingobee Fen cross section with approximate location of piezometer nests (black vertical bars) along the study transect.

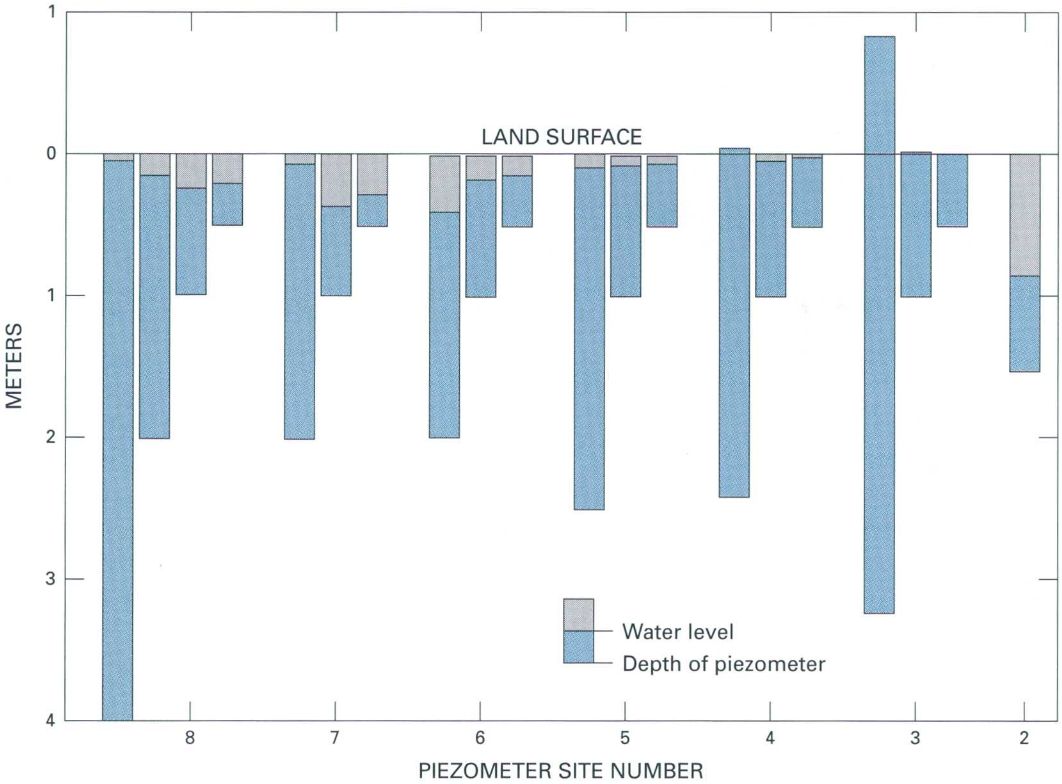


Figure 83. Water levels and depths of piezometers along the study transect during 1991. Site numbers are the same as in table 25 and figure 82.

calcite, and 39 µmoles of dolomite; formation of 288 µmol of kaolinite; consumption of 874 µmol of carbon dioxide or organic matter (CH_2O); production of 39 µmol of hydrogen sulfide; and evaporative concentration at a ratio of 2.49:1 per liter of water discharged into the fen. Although this model seems realistic based on the chemical data, there is no unique solution, and other combinations may provide similar results. Given the uncertainty about the mineral composition of the glacial sediments and lack of deep ground-water chemistry data in the local area, this model can only be considered a preliminary step in understanding of the processes controlling water chemistry of the fen.

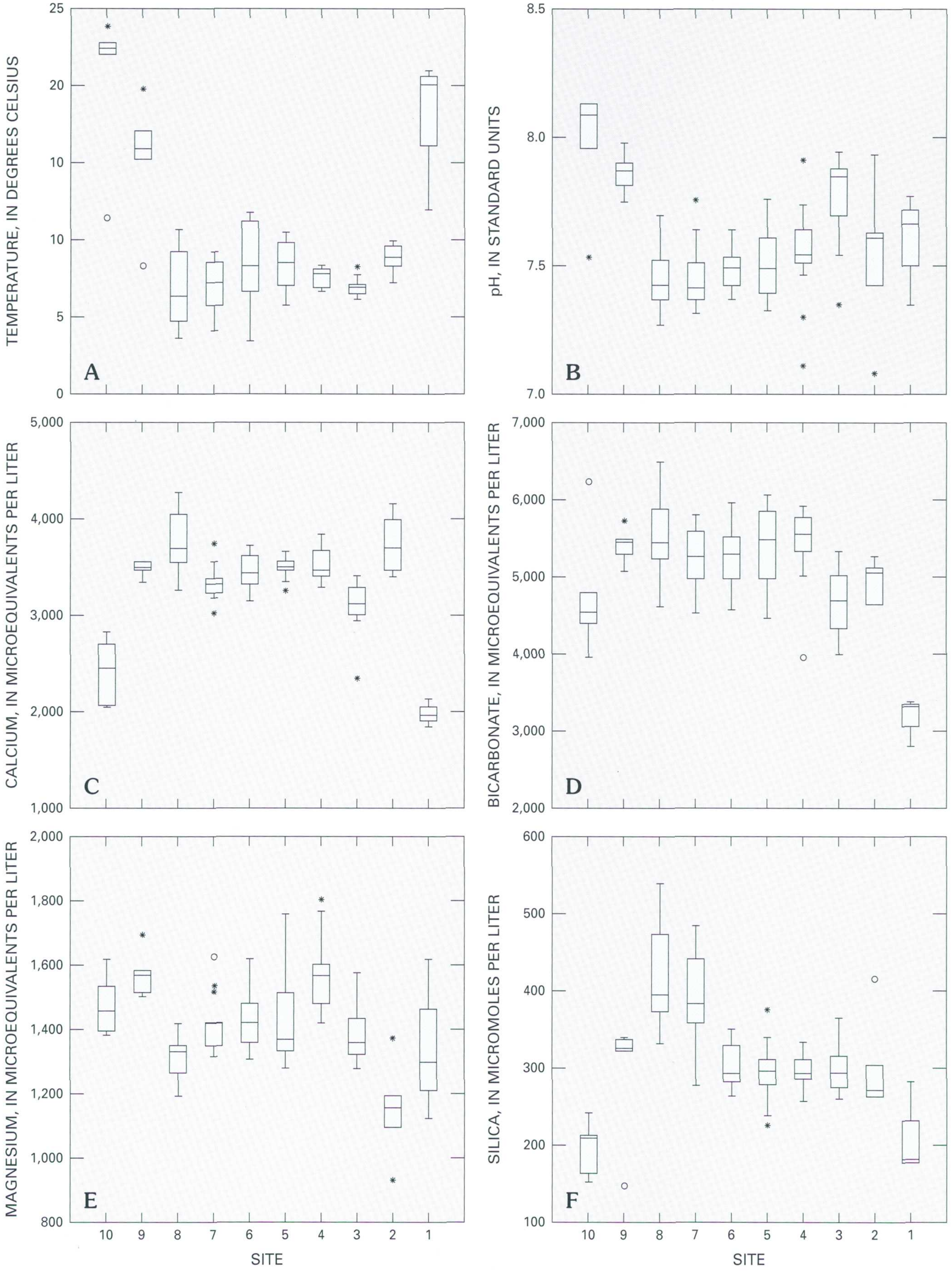
Based on a combination of the water chemistry and hydrology, the following conceptual flow-system model is proposed. Ground water, originating either as infiltrating precipitation, seepage from the Shingobee River, or a similar ground-water source, or some combination of these, is discharged at the contact between the sandy outwash material and glacial till forming the seep zone at the upslope border of the fen. The discharged ground water flows downslope toward Little Shingobee Lake through the accumulated peat and marl layers undergoing evaporation along the way. Ground water also flows under the peat, possibly through a coarser sand layer with larger hydraulic conductivity. The difference in hydraulic conductivity of the confining peat-marl layers and the coarser sand layer results in the observed ground-water vertical gradients. Near the lake, ground water flows to the surface, possibly through zones of preferential flow.

This conceptual model poses several questions requiring further research. Deep piezometers are needed in the fen, on the upland ridge, and in the local area to test the hypotheses about ground-water chemistry and gradients along the conceptual flow path. Flow studies of the Shingobee River could determine whether losing reaches potentially contribute to the ground water flowing into the fen. Water budgets need to be refined to estimate whether precipitation inputs in the upland area are sufficient to account for the amount of water being discharged in the fen. Stable isotope analyses would provide more information on the possible sources of the water discharging into the fen and Little Shingobee Lake. In-depth studies of the mineralogy and stratigraphy of the glacial deposits are required to provide more confidence in the geochemical modeling and the conceptual flow path. Hydraulic conductivities of the peat, marl, sand and till materials need to be determined.

Table 25. Median temperature, pH, and concentrations of major ions in water samples from the Shingobee River, piezometers in the Little Shingobee Fen, the small stream draining the fen, and Little Shingobee Lake during June through October 1991.

[Temperature (T°C) is in degrees Celsius, pH is in standard units, and concentrations are expressed as microequivalents per liter except for SiO_2 which is in micromoles per liter. Ca^{2+}=calcium, HCO_3= bicarbonate, Mg^{2+}=magnesium, SiO_2=silica, K^+=potassium, Na^+=sodium, Cl^-=chloride, NO_3^-=nitrate, and SO_4^{2-}=sulfate. Values for the piezometers nests represent samples collected at all depths]

Site	Description	T°C	pH	Ca^{2+}	HCO_3^-	Mg^{2+}	SiO_2	K^+	Na^+	Cl^-	NO_3^-	SO_4^{2-}
1	Shingobee River	20.2	7.71	2,050	3,370	1,460	229.0	32.9	74.5	38.8	4.8	93.3
2	Piezometer Nest 1	8.9	7.60	3,703	5,049	1,155	270.2	19.2	169.2	30.2	4.6	103.2
3	Piezometer Nest 2	6.8	7.84	3,130	4,700	1,360	290.0	40.8	287.0	46.9	7.6	43.2
4	Piezometer Nest 3	7.8	7.53	3,493	5,580	1,570	292.0	48.3	261.0	57.4	4.1	5.4
5	Piezometer Nest 4	8.5	7.48	3,520	5,504	1,365	294.8	41.9	303.0	83.1	3.5	6.3
6	Piezometer Nest 5	8.4	7.49	3,464	5,326	1,417	292.7	45.0	370.3	87.7	1.8	6.5
7	Piezometer Nest 6	7.3	7.40	3,340	5,268	1,376	384.5	57.0	417.1	83.2	3.0	6.3
8	Piezometer Nest 7	6.4	7.42	3,710	5,459	1,330	394.5	49.2	229.5	60.7	6.2	8.8
9	Stream	15.9	7.86	3,510	5,460	1,571	326.0	33.3	243.0	44.8	5.4	11.5
10	Little Shingobee Lake	22.4	8.08	2,470	4,570	1,456	207.5	35.3	260.1	40.7	2.2	42.8


A
TEMPERATURE, IN DEGREES CELSIUS
25
20
10
10
5
0
B
pH, IN STANDARD UNITS
8.5
8.0
7.5
7.0
C
CALCIUM, IN MICROEQUIVALENTS PER LITER
5,000
4,000
3,000
2,000
1,000
D
BICARBONATE, IN MICROEQUIVALENTS PER LITER
7,000
6,000
5,000
4,000
3,000
2,000
E
MAGNESIUM, IN MICROEQUIVALENTS PER LITER
2,000
1,800
1,600
1,400
1,200
1,000
800
F
SILICA, IN MICROMOLES PER LITER
600
500
400
300
200
100
10
9
8
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SITE

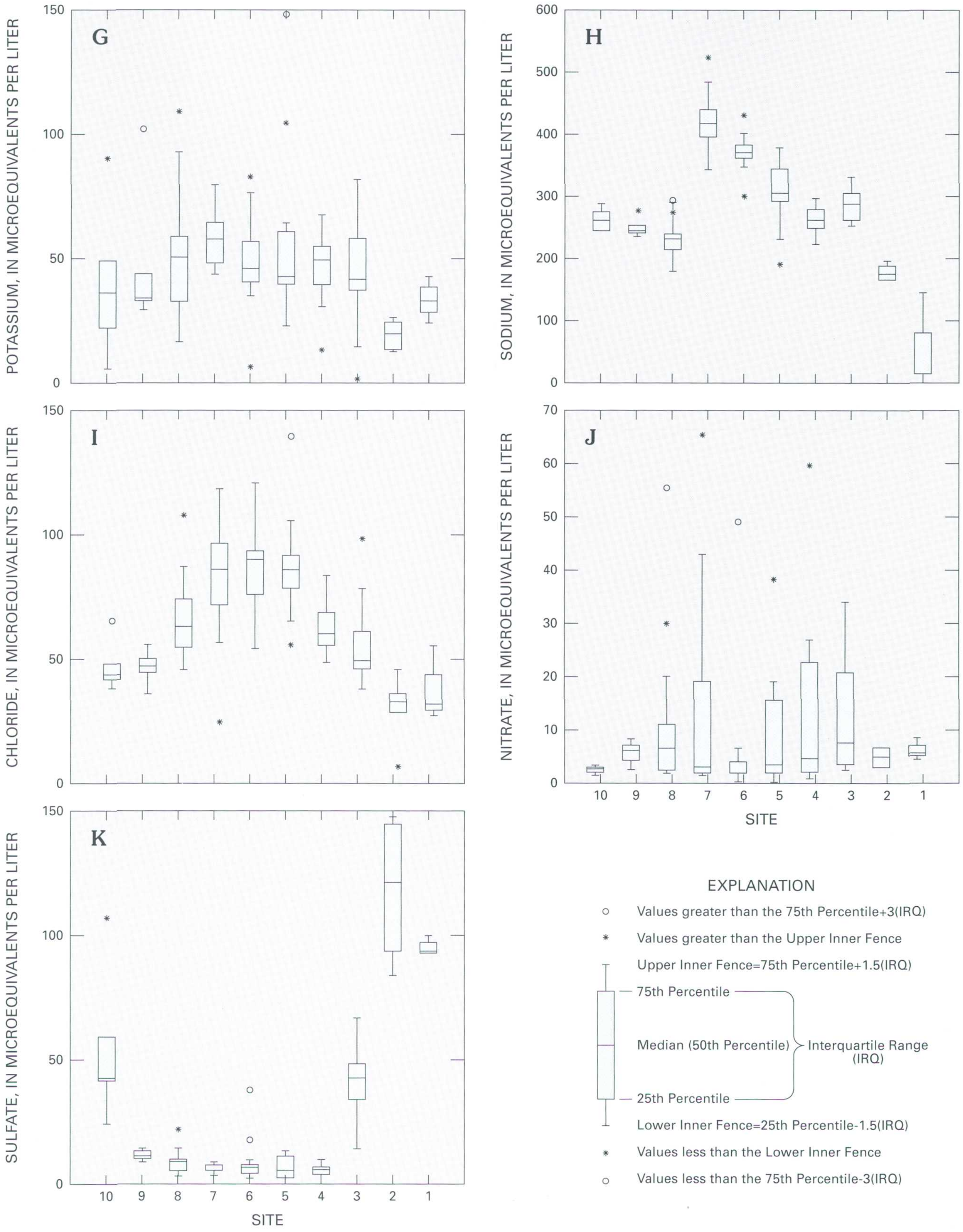

Figure 84. Box plots of (A) temperature, (B) pH, and concentrations of (C) calcium, (D) bicarbonate, (E) magnesium, (F) silica, (G) potassium, (H) sodium, (I) chloride, (J) nitrate, and (K) sulfate in water samples from the Shingobee River, piezometers in Little Shingobee Fen, the small stream draining the fen, and Little Shingobee Lake. Site numbers are the same as in table 25.

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Bacterial Indicators of Ground-Water Discharge - Iron Seeps in the Shingobee River and Crow Wing Watersheds, Northern Minnesota

By Eleanora I. Robbins, James W. LaBaugh, Dennis A. Merk, Renee S. Parkhurst, Larry J. Puckett, Donald O. Rosenberry, Paul F. Schuster, and Patricia A. Shelito

INTRODUCTION

At seep sites in watersheds having high concentrations of iron and manganese, iron bacteria typically produce oil-like biofilms, loose red/orange flocculates (flocs), attached red/orange precipitates, and brown-black coatings on manganese-coated cobbles in riffles (Robbins and others, 1992; Robbins and Norden, 1994). Iron seeps ("chalybeate springs") are sites where anoxic ground water containing dissolved iron discharges into oxygenated water or air, thereby "oxidizing" the iron. Manganese is often present at such sites also. Iron bacteria are typically abundant at such iron seeps (Harder, 1919; Pringsheim, 1949; Ghiorse and Ehrlich, 1992).

The iron bacteria are a polyglot group or consortium that includes taxa that either actively oxidize iron for energy (i.e. *Gallionella ferruginea*) or passively precipitate iron oxide (i.e. *Leptothrix ochracea*) (Nealson, 1983; Ehrlich, 1990). Many iron bacteria accumulate both iron and manganese (Balows and others, 1992; Ghiorse and Ehrlich, 1992; Staley and others, 1989). *Leptothrix discophora* is among the rod-shaped (rods) bacteria that produce oil-like biofilms at the air-water interface (Ghiorse, 1984; Robbins and others, 1992). Microanalysis of such biofilms has shown the presence of microbially-precipitated protoferrihydrite and siderite (Sawicki and others, 1995). Many iron bacteria have distinct morphologies and are therefore considered easy to identify in samples from natural environments (Hanert, 1991; Pringsheim, 1949). However, taxonomic confusion is prevalent, so physiological and genetic analyses are important additional techniques for identity confirmation.

Typically, the diversity pattern for iron bacteria in a seep microbial community has one dominant species and a few individuals of most of the consortium (Robbins and Norden, 1994). Dominance may be a function of season (Hanert, 1991) and therefore of temperature.

In the "iron country" of northern Minnesota, red or red-orange iron seeps are common (Harder, 1919). Ground water in northern Minnesota and southern Ontario contains iron and manganese (Myette, 1982; Stark and others, 1991) due to the presence of iron- and manganese-bearing Precambrian basement rocks including the iron deposits of the Lake Superior iron ranges (James, 1954; Schmidt, 1963).

As part of the studies on interactions between ground and surface water being undertaken in the IRI effort, bacteria and chemistry in iron seeps were analyzed in the Shingobee River watershed (fig. 85). Sites along Eleventh Crow Wing Lake (fig. 85) were also studied in detail to test the relationship between the oil-like biofilms and ground-water chemistry.

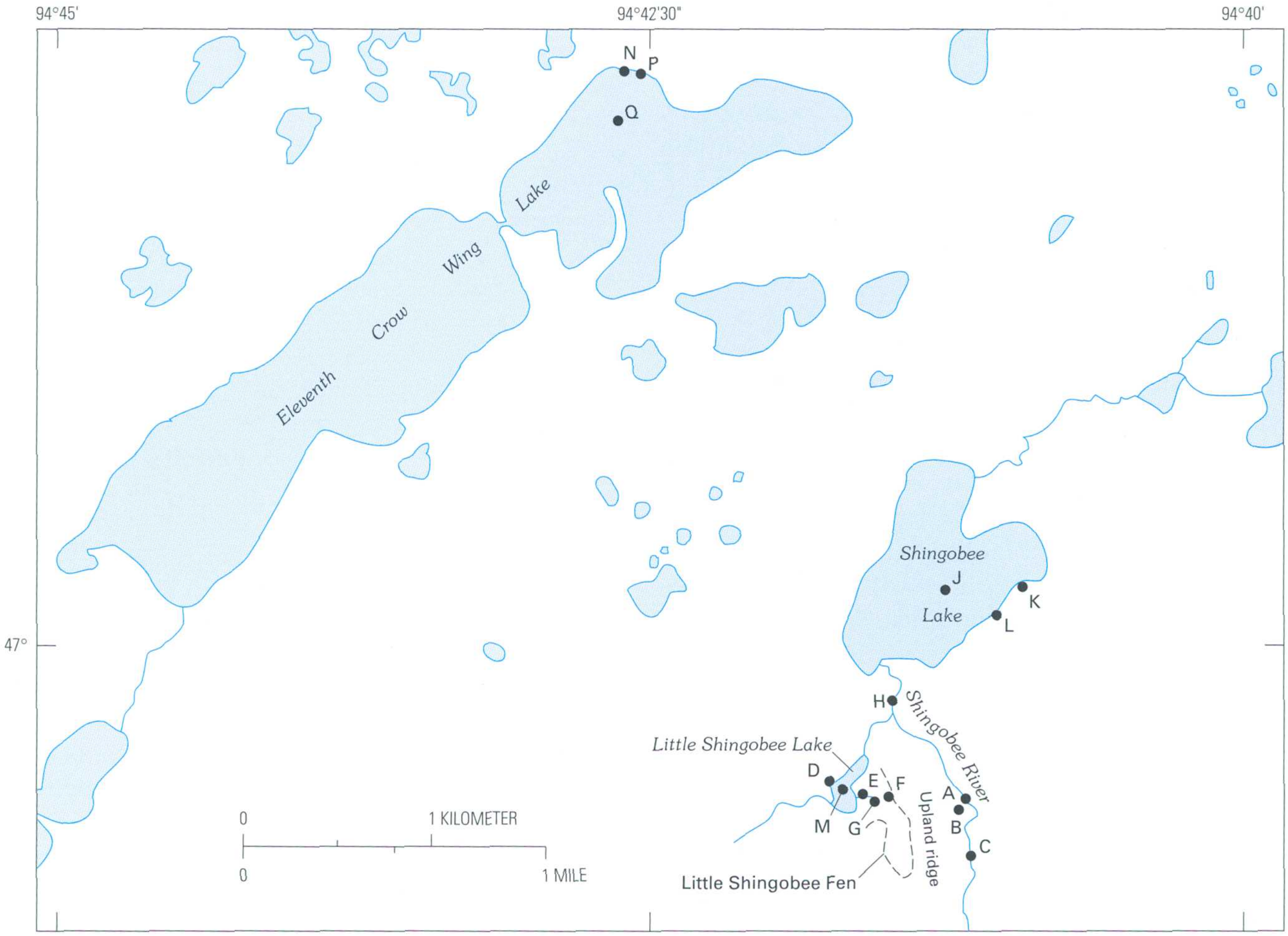
METHODS

For microbial sampling and analysis, loose red flocs and precipitates attached to surfaces were extracted with eyedroppers and stored in vials. Oil-like biofilms were lifted from the water surface with single microscope slides. Bacteria that would attach to glass were collected using paired microscope slides affixed to stakes left in the water for 3 weeks. Samples for this study were collected in May, June, and July 1995. Bacteria were identified on the basis of morphology (fig. 86). The presence of oxidized manganese on the bacteria was tested using ortho-toluidine (Morgan and Stumm, 1965).

Iron bacteria were collected at selected seeps in the Shingobee River watershed and along the north shore of Eleventh Crow Wing Lake in the Crow Wing watershed (table 26, fig. 85).

In the Shingobee River watershed, water for chemical analysis was not collected at the time of microbial sampling. Instead, relevant chemical data were selected from the Shingobee River watershed data base (table 27). Iron and manganese were determined by inductively coupled plasma emission spectroscopy according to the method described by Garbarino and Taylor (1979). Values for pH were determined in the field or laboratory by the electrometric, glass-electrode method (Fishman and Friedman, 1985).

At the Eleventh Crow Wing sites, hydraulic head of ground water was determined using a portable piezometer, having a retractable point with a fine stainless-steel screen. The piezometer was driven 39 to 58 cm into the sediment. Ground water was pumped 5 minutes before collection of samples for chemical analyses. Temperature, pH, and specific conductance were taken at the time of water sampling. Chemical characteristics of Crow Wing Lake surface and ground water (tables 27 and 28) were analyzed using standard methods (Fishman and Friedman, 1985) by A.W. Research Labs, Brainerd, Minnesota. The 1,10-phenanthroline/ferrous iron method was used for total iron, the periodate oxidation method was used for total manganese, and the methylene blue method was used for sulfide.



Base from Akeley and Walker, Minn.,
7.5' topographic quadrangle maps

Figure 85. Sampling sites in the Shingobee River and Crow Wing watersheds.

RESULTS AND DISCUSSION

Bacteria of Ground Water Seeps

The iron bacteria of these sites in northern Minnesota are typical of those in iron seeps. *Leptothrix ochracea*, rods that accumulate iron on empty, drinking-strawlike sheaths (fig. 86 A), dominate flocs at almost every locality and test positive for the presence of manganese oxide. At the upland ridge on the west side of the Shingobee River, the sheaths are adherent, a phenomenon that has been ascribed to production of excessive amounts of capsular substances intermixed with iron hydroxides (Mulder and Deinema, 1992). Rods that stay in drinking-strawlike sheaths, tentatively identified here as *Leptothrix* cf. *cholodnii* (fig. 86 D), are universally present but never dominant in flocs. These bacteria test negative for manganese. *Gallionella ferruginea*, kidney-shaped rods that oxidize iron on braid-like holdfasts with which they attach to surfaces (fig. 86 A), and an unidentified spiral form (fig. 86 G), are present but rare in all localities along the Shingobee River watershed. *Siderocapsa* cf. *treubii*, which are cocci (spherical forms) that accumulate iron outside globular capsules (fig. 86 A), are most common on organic matter, including the sheaths of other iron bacteria. *Siderocystis confervarum* is the dominant iron bacterium colonizing green algae (fig. 86 C). *Leptothrix discophora* holdfasts are present, along with colorless rods, on most oil-like biofilms (fig. 86 H); the holdfasts form thick deposits of brown-black manganese oxide on microscope slides left in riffles (fig. 86 F and G). Brown (manganese-coated) or red (iron-coated) rods are ubiquitous in almost all Shingobee River watershed sites (fig. 86 E and F). Although red cocci (cf. *Siderococcus* sp.) are noted only where abundant, they probably are present universally. *Toxothrix trichogenes*, a common associate with other the iron bacteria, is not present in any sample.

Oscillatoria sp. produces distinctive black or very dark green mats at many of the seeps. This dark color may be a function of a specific pigment or the reflection of light off the bundled sheaths (R.C. Jones, George Mason University, oral commun., 1995). At seep sites in Little Shingobee Fen and Shingobee River, this cyanobacterium is colonized by iron bacteria including *L. ochracea* and *L.* cf. *cholodnii*.

Colonization of the *Oscillatoria* mats by iron bacteria produces features seen in thin sections of Precambrian banded iron formation (BIF), where thin organic bands of filamentous organisms are sandwiched between layers of hollow hematites (LaBerge and others, 1987; Robbins and others, 1987). The *Oscillatoria* mats are thin and coated with hollow sheaths of *L. ochracea*. Sheath-forming, iron-oxide-accumulating bacteria such as *Leptothrix* are coated with the hydrated hematite precursor, ferrihydrite (Chukhrov and others, 1973). Postdepositional dehydration reactions could produce hollow hematites from ferrihydrite-coated empty sheaths (Robbins and others, 1987). Compression after burial of the Little Shingobee Fen mats could easily produce the thin organic-rich bands between hollow hematites that are seen in the banded iron formation.

Chemistry of Ground and Surface Waters

Dissolved concentrations in surface and ground water ranged from below detection limit to 3,100 μg/L for iron and from below detection limit to 1,400 μg/L for manganese (table 27), pH values ranged from 6.8 to 8.8. Because water samples were not collected simultaneously with bacterial samples, the water-chemistry analyses provide only general data about the ideal chemical environments for iron bacteria. The data show that high iron and manganese concentrations are being provided at the bankside seeps. Iron bacteria test positive for manganese oxide in some of the Shingobee River seeps, the upland ridge, Shingobee Lake, and Little Shingobee Lake, but not in Little Shingobee Fen or Crow Wing Lake seeps, even though manganese is available. These data indicate that microbial interactions with specific chemical components are not solely governed by availability.

Iron and manganese concentrations of the Shingobee River and Crow Wing watersheds can be compared to drinking-water standards to assess the natural chemical environment in which iron bacteria proliferate. According to Office of Water-U.S. Environmental Protection Agency (1994), the maximum allowable concentration in drinking water for iron is 300 μg/L and for manganese is 50 μg/L. Therefore, the iron bacteria in this study were collected from chemical environments that exceed drinking-water standards for iron and manganese.

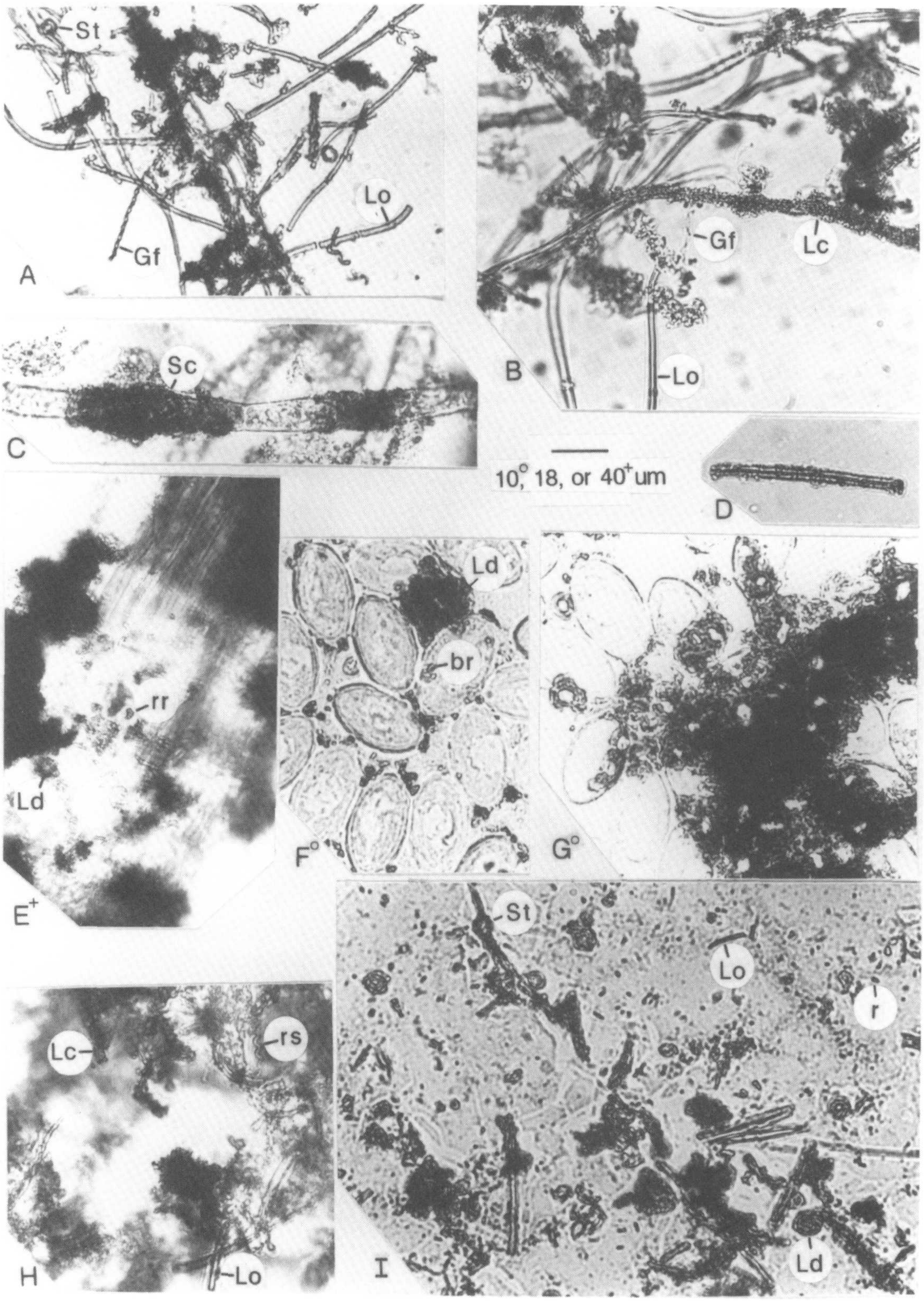


Figure 86. Bacteria of the Shingobee River and Crow Wing watersheds.

A. Mass of *L. ochracea* (Lo) with *Siderocapsa* cf. *treubii* (St) and *Gallionella ferruginea* (Gf) colonizing biofilm from seep along margin of Shingobee Lake (Sample SF-4). B. Mass of *Leptothrix ochracea* (Lo) with *L.* cf. *cholodnii* (Lc) and *Gallionella ferruginea* (Gf) from Little Shingobee Fen spring pool (Sample 21). C. *Siderocystis confervarum* (Sc) colonizing unidentified green alga in backwater of Shingobee River (Sample 1) D. *L.* cf. *cholodnii* in seep at channel margin of Shingobee River (Sample 5). E. *Leptothrix discophora* (Ld) and red rods (rr) colonizing roots along lake margin seep of Shingobee Lake (Sample 12). F and G. *L. discophora* (Ld) and brown rods (br) colonizing cells of the diatom *Cocconeis* sp. and microscope slide left in riffle for 20 days in Shingobee River (Sample SS-1). H. Mass composed of *Leptothrix ochracea* (Lo), *L.* cf. *cholodnii* (Lc), and red spiral form (rs) at toe of seep along creek between Shingobee Fen and Little Shingobee Lake (Sample 18). I. *Leptothrix* cf. *ochracea* (Lo), *L. discophora* (Ld), *Siderocapsa treubii* (St), diatoms, and rods (r) colonizing biofilm along margin of Eleventh Crow Wing Lake (Sample SF-6)

Table 26. Iron bacteria in the Shingobee River and Crow Wing watersheds

[*, positive for manganese oxide; +, present; --, not found; +++, dominant; ?, questionable identification]

Location [in fig. 85]	Sample number	*Gallionella ferruginea*	*Leptothrix discophora*		*Leptothri x ochracea*	*Sidero-capsa cf. treubii*	*Sidero-cystis confervarum*	*Leptothrix cf. cholodnii*	Red cocci	Brown or red rods	Red rod chains	Red spiral -type	Specific location	Comments
			Holdfast	*Biofilm*										
Shingobee River [A]	1	--	+	--	+	+	+	--	+	+	--	--	backwater	on *Elodea* and green algae
Shingobee River [A]	2	--	+	+	--	+	+	--	--	+	--	--	backwater	attached to stream sediment
Shingobee River [A]	3	--	+	+	+	?	--	+	--	+++	--	--	channel margin seep	iron-cemented sand grains
Shingobee River [A]	4	+	--	--	+++*	--	--	+	--	+*	--	+	channel floor	attached to organics
Shingobee River [A]	5	+	--	--	+++	--	+	+	--	+*	--	+	channel margin seep	attached to organics
Shingobee River [A]	SF-5	--	+	+	+	--	--	+	--	--	--	--	flood plain seep	oil-like biofilm
Shingobee River [C]	SS-1	--	+*	--	--	--	--	--	--	+	--	--	stream riffle	slide in water for 20 days
Upland ridge on west side of Shingobee River [B]	6, 7, 24	+*	+*	--	+++*	+*	+	+	+*	+*	--	+	bank seep	attached to sediment
Shingobee Lake [K]	11	--	--	--	--	--	+	+	+	--	--	--	lake-margin seep	attached to algae
Shingobee Lake [L]	12	+	+	--	+	--	+*	+	--	+*	--	--	"power pole" lake margin seep	attached to roots
Shingobee Lake [L]	SF-4	+	+	--	+	+	--	+	--	+	--	--	"power pole" lake margin seep	oil-like biofilm
Little Shingobee Lake [D]	16	--	+	+*	+++	+++	--	+	--	+	--	--	in beaver ditch	loose flocculate
Little Shingobee Lake [D]	SF-1	--	+	+	+	--	--	+	--	+	--	--	pool on flood plain	oil-like biofilm
Little Shingobee Lake [E]	SF-2	--	+	+	+	--	--	--	--	+	+	--	pool on flood plain	oil-like biofilm
L. Shingobee Fen creek [E]	17	+	--	--	+++	+	--	+	--	+	--	+	in creek	loose floc
L. Shingobee Fen creek [F]	18	+	--	--	+++	--	--	+	--	+	--	+	"black seep" toe	attached to sediment
L. Shingobee Fen creek [F]	19	+	--	--	+++	--	?	+	--	+	--	--	"black seep" head	attached to sediment
L. Shingobee Fen creek [F]	20	--	--	--	+	--	+++	+	+	+	--	--	small tributary	attached to algae
Little Shingobee Fen [G]	21	+	+	+	+++	?	+	+	--	+	--	+	spring pool	attached to sediment
Little Shingobee Fen [G]	22	--	+	--	+++	+	+	+	--	+	--	+	spring pool	part of microbial mat
Little Shingobee Fen [G]	23	--	--	--	+	+	--	--	--	+	--	--	spring pool	part of microbial mat
Eleventh Crow Wing Lake [N]	SF-6	--	+++ +		+++	+	--	+	--	--	--	--	edge seep	oil-like biofilm

Other Indicators of Seeps

Accumulations of iron oxides are not the only indicators of ground-water seeps. Absence or presence of vegetation provides additional information. For example, areas around seeps within the Shingobee Lake are almost bare of emergent and floating-leaf vegetation, whereas the lake bottom is otherwise densely vegetated. Many of these bare areas are ringed by dense growths of the alga *Chara*. On the lake shore and along the creeks and in the fen, marsh marigold (*Caltha palustris*) grows at nearly all of the seeps; furthermore, watercress is commonly present and provides a good visual indicator of seeps because it remains green throughout the year and provides good contrast with snow cover during winter.

Temperature differences are also useful for identification of seeps. Ground water is cooler than lake surface water in the summer (table 28).

Oil-like films are being tested as indicators of ground-water seepage (Robbins and others, 1992; Sawicki and others, 1995). Discrete patches of oil-like films emanate from the north shore of Eleventh Crow Wing Lake. Some of these sites are accompanied by the presence of loose red flocs along the banks and at the shore. Four sites along the lake shore were chosen to test the presence or absence of ground-water seepage. These sites either had or lacked oil-like films; hydraulic head, temperature, and chemistry were analyzed (tables 27 and 28). The sites having biofilms also have sandy substrates, ground water colder than the lake, and the distinct odor of hydrogen sulfide during water extraction. The sites lacking such biofilms did not meet these three conditions. Where these three conditions are present, oil-like biofilms indicate the seepage of ground water.

CONCLUSIONS

The iron bacteria provide useful visual signals of anoxic, ground-water, point-source discharge in the Shingobee River and Crow Wing watersheds in Hubbard County, Minnesota. All the iron seeps contain a variety of iron bacteria and are dominated by *Leptothrix ochracea*. Sheaths and capsular material of many test positive for manganese. Brown-black coatings on cobbles in riffles are composed of manganese oxide that is precipitated around the holdfasts of *L. discophora* and short brown rods. Ground water in the Shingobee River watershed contains dissolved iron and manganese, attesting to its availability to support populations of iron- and manganese-accumulating and oxidizing bacteria.

The test for the presence and absence of oil-like biofilms in Eleventh Crow Wing Lake provides useful information about ground-water discharge locations. Oil-like biofilms correlate with sandy substrates, ground water colder than the lake, and the distinct odor of hydrogen sulfide during water extraction.

REMAINING PROBLEMS

The bacterium tentatively identified in this study as *L.* cf. *cholodnii* presents a special problem. Mulder and Deinema (1992) stated that *L. cholodnii* accumulates manganese, whereas the form in the Shingobee River watershed tests negative for manganese. An identification method more stringent than morphology, such as 16S rRNA sequence analysis, is required to confirm the identity of *L. cholodnii* in the Shingobee River and Crow Wing watersheds.

The seep along the upland ridge has features that make it a potential precursor of bog iron ore. Its spongy accumulation of sediments, plant tissues, and iron-oxide precipitates is unusual because intense rainfall usually flushes out any accumulated precipitates. More mineralogy and precipitate and ground-water chemistry might provide useful information about the formation of older organic-rich, point-source, economic deposits.

ACKNOWLEDGEMENTS

We thank Connie Converse and Ted Soteroplos for field assistance.

Table 27. Summary statistics of selected chemical constituents in surface and ground water in the Shingobee River and Crow Wing watersheds

[Iron is in the form of Fe^{2+}; μg/L, micrograms per liter; bd, below detection; m, meter; <, less than ; n, number of observations; Data from PCRM-Puckett and others (this volume), PFS-P.F. Schuster (unpub. data), PAS-P.A. Shelito (unpub. data)]

Location [in fig. 85]	Sample date	pH (pH units)		Iron (μg/L)		Manganese (μg/L)		n
		Median	Range	Median	Range	Median	Range	
Shingobee River near PCRM8 [B]	1991	7.7	7.7-7.8	210	bd-430	77	bd-150	5
Shingobee River below beaver dam [A]	1991	7.6	--	530	--	41	--	5
Ground water from well along Shingobee River at 5.79 m depth [A]	1992-93	7.2	7.1-7.8	54	7-200	6	3-10	7
Seep in upland ridge west of Shingobee River-"Seep 410" [B]	1991	7.8	--	2,700	--	61	--	5
Shingobee Lake inlet [H]	1989-93	7.9	6.8-8.8	68	<3-350	62	16-290	52
Shingobee Lake at 2 m below surface [J]	1989-92	8.6	7.4-8.7	9	<3-1,700	25	1-310	46
Shingobee Lake edge seep PFS1, furthest from lake [L]	July 1993	8.1	--	390	--	210	--	1
Shingobee Lake edge seep PFS2 [L]	July 1993	8.2	--	100	--	230	--	1
Shingobee Lake edge seep PFS3 [L]	July 1993	8.4	--	30	--	160	--	1
Shingobee Lake edge seep PFS4, closest to lake [L]	July 1993	8.4	--	390	--	210	--	1
Little Shingobee Lake at 1 m below surface PCRM [M]	1991	8.1	7.5-8.1	370	10-510	9	3-22	53
Little Shingobee Fen creek-"black seep" PCRM [F]	1991	7.9	7.7-8.0	670	200-960	170	88-260	5
Little Shingobee Fen creek-PCRM8 [F]	1991	7.1	7.0-7.2	2,100	530-4,900	330	240-370	5
Little Shingobee Fen, 2 spring pools [G]	October 1992	--	7.4-7.8	--	60-3,100	--	47-1,400	2
Little Shingobee Fen, PCRM1 on hill slope at top of water table [G]	1991	7.6	7.1-7.9	350	81-3,000	38	11-410	5
Little Shingobee Fen, PCRM2 discharge zone [G]	1991	7.8	7.4-7.9	420	30-1,100	13	5-230	15
Little Shingobee Fen, PCRM4 middle of slope [F]	1991	7.5	7.3-7.8	660	bd-2,300	500	10-1,300	15
Little Shingobee Fen, PCRM7 edge of fen in sedge community [F]	1991	7.4	7.3-7.7	1,700	bd-4,800	200	3-290	20
Eleventh Crow Wing Lake edge, PAS1 west of Crow Wing Crest Resort dock [N]	July 23, 1995	6.8	--	1,900	--	220	--	1
Eleventh Crow Wing Lake edge, PAS8 east of Crow Wing Crest Resort dock [P]	July 23, 1995	7.1	--	390	--	90	--	1
Eleventh Crow Wing Lake, PAS average of top 1.8 m [Q]	July 23, 1995	7.1	--	<1	--	<1	--	1

Table 28. Summary statistics of selected chemical and physical parameters from ground water along northern edge of Eleventh Crow Wing Lake

[°C, degrees Celsius; H_2S, hydrogen sulfide; mm, millimeters; nd, not determined; µS/cm, microsiemens per centimeter; µg/L, micrograms per liter; Data from P.A. Shelito, unpub. data]

Location [in fig. 85]	Lithology	Oil-like biofilm	Temperature (°C)		pH (pH units)		Head difference ground water>lake (mm)	Specific conductance (µS/cm)		H_2S (µg/L)	Other
			Lake	Ground water	Lake	Ground water		Lake	Ground water		
East of dock [P]	tight clay	absent	17	13	nd	6.8	+2	nd	470	nd	No smell
East of dock [P]	sandy	absent	22	21	7.1	7.1	0	190	450	119	Strong H_2S smell
West of dock PAS1 [N]	loose sand	present	17	12	nd	6.9	+22	nd	330	nd	H_2S smell
West of dock PAS8 [N]	sandy	present	20	18	7.1	6.8	+19	260	360	20	H_2S smell

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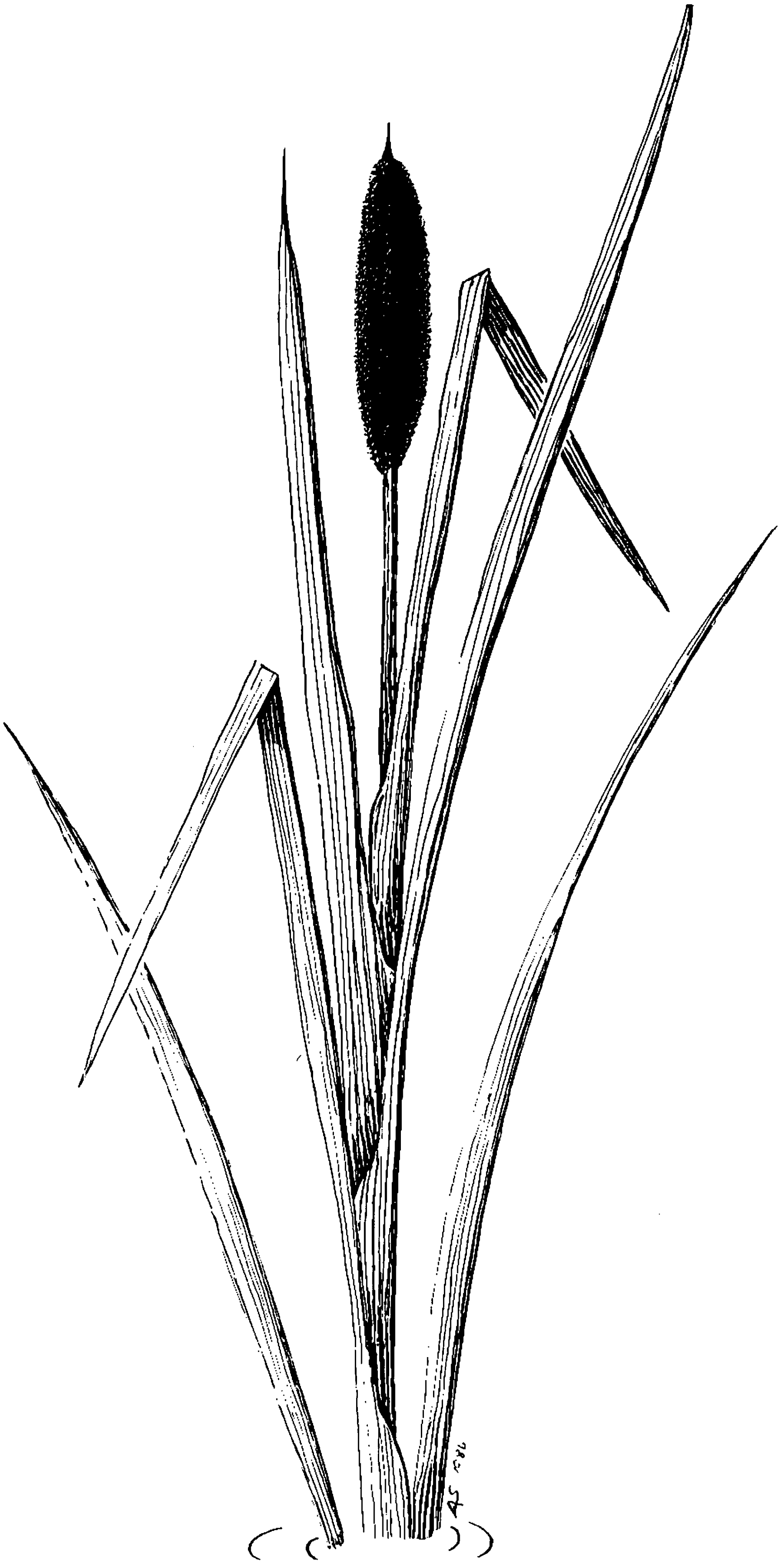
AS '86

Sediment Stratigraphy and Paleolimnological Characteristics of Williams and Shingobee Lakes

By Sharon M. Locke and Antje Schwalb

INTRODUCTION

Sedimentary profiles from Williams and Shingobee Lakes contain a detailed record of significant shifts in lake level and chemical deposition during the past 10,000 years. Were these shifts driven by major changes in the regional climate, geomorphic evolution of the local drainage network, or internal processes related to basin infilling? The long-term role of these environmental drivers should be discernible by detailed analysis of the sediments in these two lakes because of their different hydrological setting. For example, Williams Lake should experience larger lake-level fluctuations in response to climatic change because it is a closed basin located farther from a river than Shingobee Lake (Almendinger, 1990). This hypothesis is tested by two related projects. First, the magnitude and chronology of paleolake-level fluctuations are reconstructed from the sediment of the two lakes within the context of regional vegetation changes. Second, the isotopic signature of carbonate is determined from the sediments of Williams and Shingobee Lakes to evaluate changes in residence times for lake water.

METHODS

Sedimentological methods of Digerfeldt (1986) were used to identify past lake-level changes at the two sites. In this method, changes in sediment type and aquatic seed assemblages in sediment cores are used to trace changes in the position of the lake shoreline relative to the coring site, and thus give a measure of past lake levels relative the present. Transects of sediment cores were taken from shoreline towards the lake centers from the south shore of Williams Lake and the southeast shore of Shingobee Lake using a modified Livingston piston corer (Wright, 1967). Core depths were measured from the ice surface at the time of coring. The 1-m core segments were extruded onto the ice surface and wrapped in plastic wrap and aluminum foil for cold storage in the laboratory at about 5°C. Loss-on-ignition (Dean, 1974) and whole-core magnetic susceptibility analyses were used to determine downcore changes in minerogenic, organic, and carbonate contents. Plant macrofossils and mollusks were identified in intervals of key sediment changes, and ostracode valves were picked from samples at 10-cm intervals for species identification and stable isotope analysis. A chronology for sediment deposition was determined by pollen stratigraphy and radiocarbon dates on wood.

RESULTS

Regional Vegetation History

The pollen stratigraphy of Williams Lake records four major changes in regional vegetation and inferred climate during the Holocene (fig. 87). The climatic changes must have altered the water balance of the area and produced distinct changes in lake levels. Five pollen assemblage zones were identified according to changes in the percentage of dominant pollen types: spruce zone [more than 9,800 yr B.P. (years Before Present)], jack/red pine zone (9,800-7,700 yr B.P.), prairie zone (7,700-4,000 yr B.P.), hardwood zone (4,000-3,200 yr B.P.), and white pine zone (3,200 yr B.P. to present). Sedimentation in the lake began when the vegetation was spruce forest with openings of herbaceous plants. Dominant arboreal pollen types include spruce and birch, implying a cold, dry climate. Later, spruce was replaced by jack/red pine, birch, and elm in response to a slightly warmer and wetter climate. At about 7,700 yr B.P. the vegetation shifted from forest to a more open oak savanna and prairie. This mid-Holocene vegetation change is observed in cores from western Minnesota to Illinois, and indicates expansion of the prairie eastward in response to a warmer, drier climate. At 4,000 yr B.P. the climate again became cooler and wetter, allowing the growth of a transitional hardwood forest dominated by oak, birch, alder, and elm. A mixed conifer/hardwood forest, including both jack and white pine, was established in the area by about 3,200 yr B.P.

Sediment Stratigraphy

Total thickness of accumulated lake sediment in the deep core from Williams Lake is 7 m (fig. 88). Basal sediments consist of dark gray sands and gravels overlain by an organic-rich layer containing wood pieces and spruce needles. The transition from spruce to jack pine forest occurred during deposition of a tan marl (carbonate content of 80-95 percent) containing the aquatic moss *Warnstorfia* (*Drepanocladus*) *exannulatus*. The jack pine zone is characterized by silty marl. Carbonate and silt deposition remained high during the prairie period; in shallow cores, prairie sediments also contain mollusk shells and are interrupted by sand layers 5-20 cm thick. The deep core contains a thin sand layer deposited at the boundary of the jack pine and prairie zones. Above the silty marl there is a short interval of color banding due to alternating layers of marl and marly gyttja, followed by a rapid transition to organic-rich, gelatinous gyttja coincident with the establishment of the modern conifer-hardwood forest.

Total accumulated thickness of sediment at Shingobee Lake is 9.5 m for a site in 6-m water depth (fig. 89). The basal sequence is identical to that at Williams Lake. Basal sands and gravels are overlain by an organic layer and a tan marl 3-7 cm thick. The beginning of the jack pine zone is characterized by a 0.5- to 1.5-m thick section of marl (carbonate content of 60 percent) with alternating light and dark layers that appear to be annual varves. The sediment is black

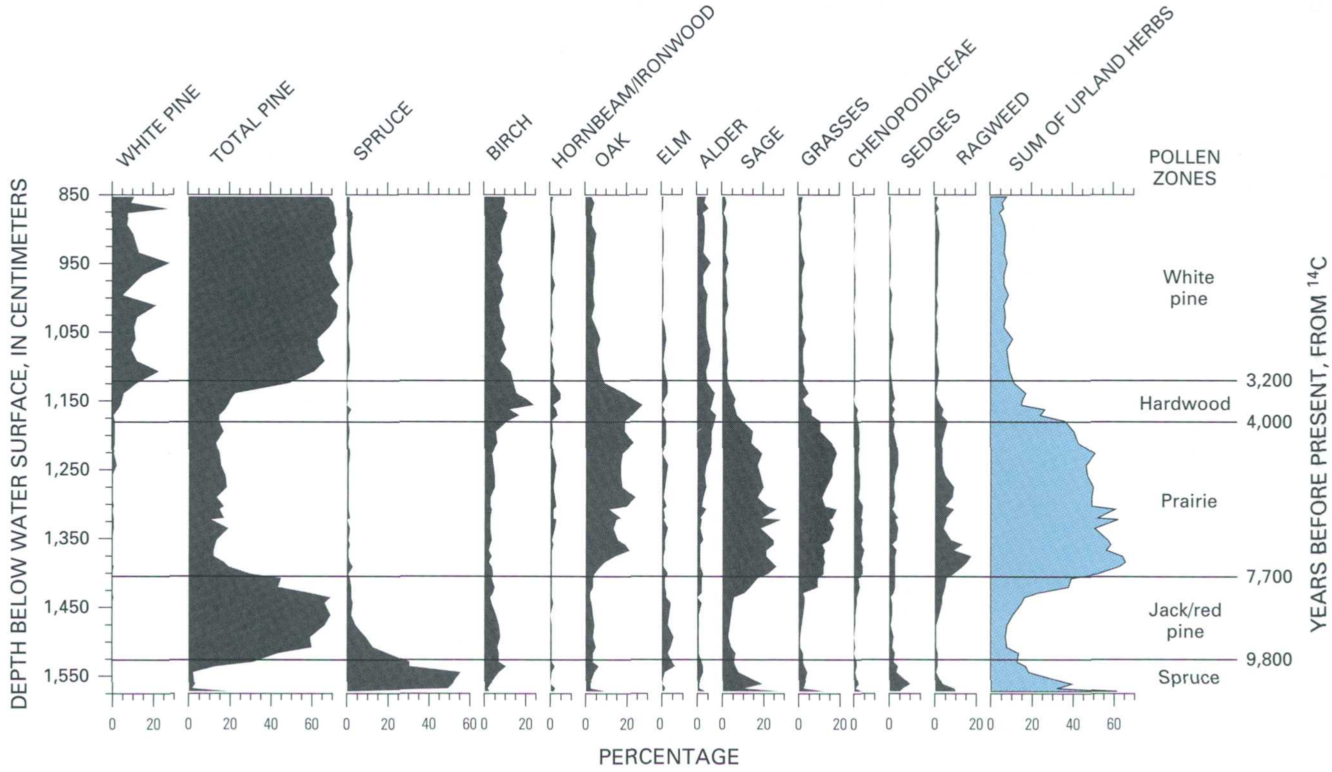

Figure 87. Pollen stratigraphy in core from Williams Lake. Depths are relative to the water surface at the time of coring. Dates for zone boundaries are based on a radiocarbon date on wood and correlation with the stratigraphy of Portage Lake, 10 km east of Williams Lake (J.H. McAndrews, unpublished pollen diagram, 1973).

when fresh, but an orange oxidation crust forms in storage, suggesting that this part of the core has a significant iron content. The uppermost unit, which constitutes the remainder of the sedimentary sequence at Shingobee Lake, is a nonlaminated marl with a carbonate content of 60-80 percent, inorganic content of 10-25 percent, and organic content of 6-8 percent. Sand layers containing abundant shells and Najas seeds were deposited at the jack pine/prairie zone boundary and within the prairie zone in a shallow core.

Paleolimnology and Past Lake Levels

Whereas Williams Lake sediments are characterized by high silt and carbonate content in the early to mid-Holocene and high organic content in the late Holocene, Shingobee Lake sediments are characterized by a consistently high carbonate content and low organic content throughout the Holocene. At Williams Lake, a major sediment change from minerogenic to organic is coincident with the prairie/white pine vegetation-zone boundary. At Shingobee Lake external factors such as climate and catchment vegetation have had less influence on sediment composition and preservation compared to Williams Lake. Varves in the early Holocene sediments at Shingobee Lake indicate bottom circulation was sufficiently weak to exclude bioturbating organisms and allow accumulation of iron oxides.

The sediments of Williams Lake record a substantial lowering of lake level during the mid-Holocene (prairie zone). At Williams Lake a lake level of as much as 4.5 meters below present is indicated by a sand layer containing *Scirpus* and *Potamogeton gramineus* seeds, which is interpreted to be a buried shoreline. A minimum lowering of 2.5 m is indicated by the shallowest cores, in which sediments deposited prior to those of the white pine zone have been eroded due to subaerial exposure. Sandy layers in the Shingobee Lake cores may be a consequence of a lower lake level, but their seed content indicates they cannot necessarily be interpreted as buried shorelines (Locke, 1995). Therefore at Shingobee Lake the lake-level lowering cannot be precisely determined solely by examining the lithology and animal and plant remains of the cores.

These results suggest that Williams and Shingobee Lakes responded differently to the mid-Holocene climate change. It may be that Shingobee Lake dropped below the level of its inlet and/or outlet, causing it to become a closed basin with an increased sensitivity to evaporation, or Shingobee Lake may have maintained a relatively constant level because it is located directly on the Shingobee River. This hypothesis can be tested by examining the oxygen isotopic records of the two lakes, which should be similar for the mid-Holocene if both basins were closed.

Ostracode $\delta^{18}O$ and $\delta^{13}C$ Signatures as Indicators of Past Lake/Ground-Water Interactions and Climate Change

A data set of the current oxygen isotopic composition of lake water, lake ice, precipitation, and ground water in the catchment area shows significant differences between Williams Lake and Shingobee Lake (Kendall and others, this volume). These isotopic differences are related to the hydrologic differences between the lakes. Williams Lake is a surficially closed basin, and its water budget is dominated by atmospheric and ground-water components. Shingobee Lake is an open basin, and its water budget is dominated by the inlet and outlet streams.

The residence time of water in Williams Lake is much greater than that of Shingobee Lake; therefore, the water of Williams Lake shows evaporative concentration of ^{18}O compared to the water of Shingobee Lake, which has an isotopic composition similar to that of precipitation. The authigenic carbonate component of sediments reflects, at least to some degree, the isotopic composition of the lake water at the time of deposition. Thus, sediments from this pair of lakes should have registered a continuous record of both the isotopic composition of precipitation (Shingobee Lake sediments) and how evaporation has changed during the Holocene (difference between Williams Lake and Shingobee Lake sediments). Furthermore, past changes in evaporation may be correlated with past changes in lake level. The current data set can be used as a key for the interpretation of past signatures.

Sediment cores with continuous depositional sequences since the early Holocene were sub-sampled for both lakes. The stable oxygen and carbon isotope analyses were carried out on the authigenic carbonate of ostracode valves. Their isotopic composition was in

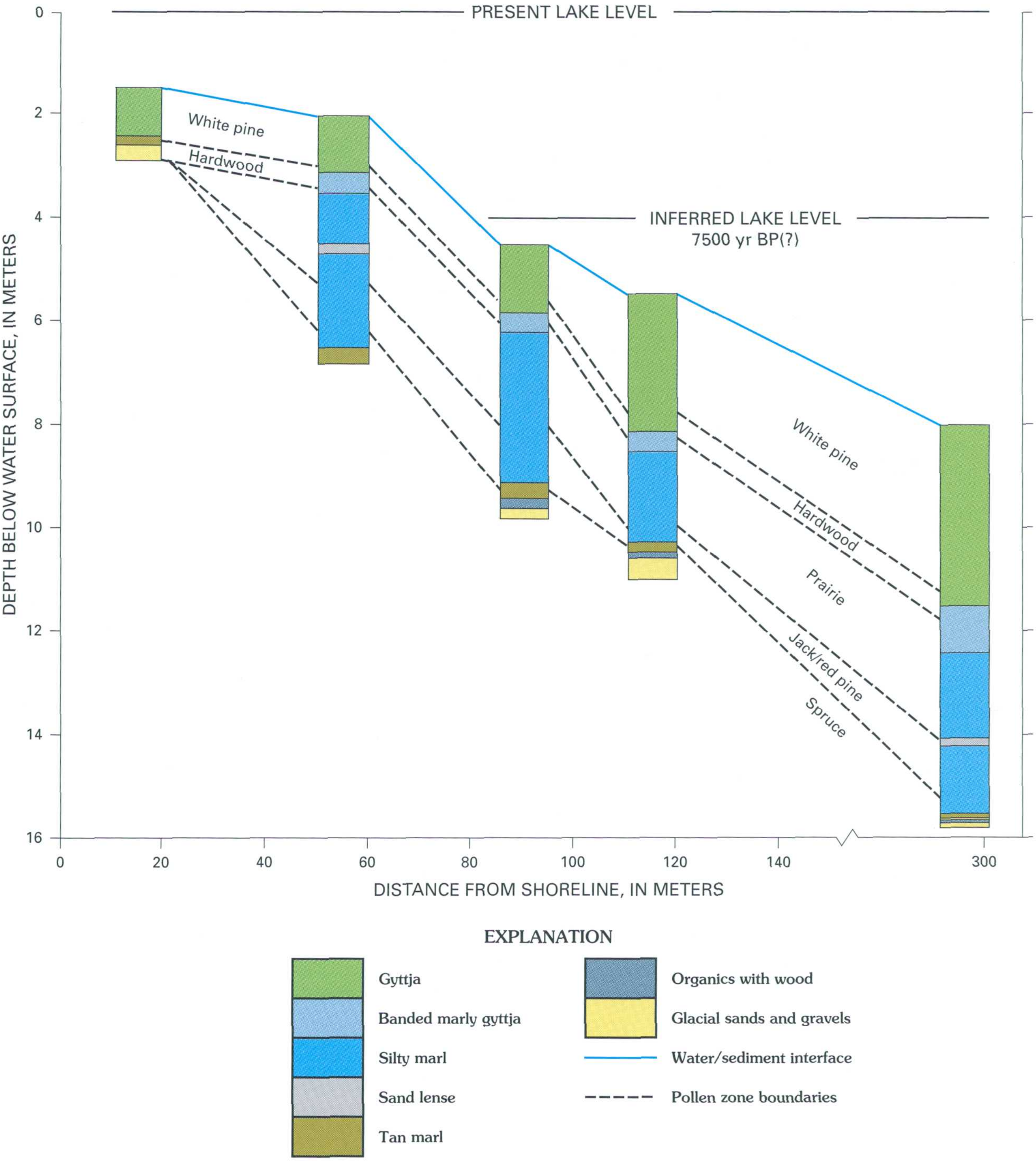

Figure 88. Sediment stratigraphy of Williams Lake for a transect taken from the south shore of the lake.

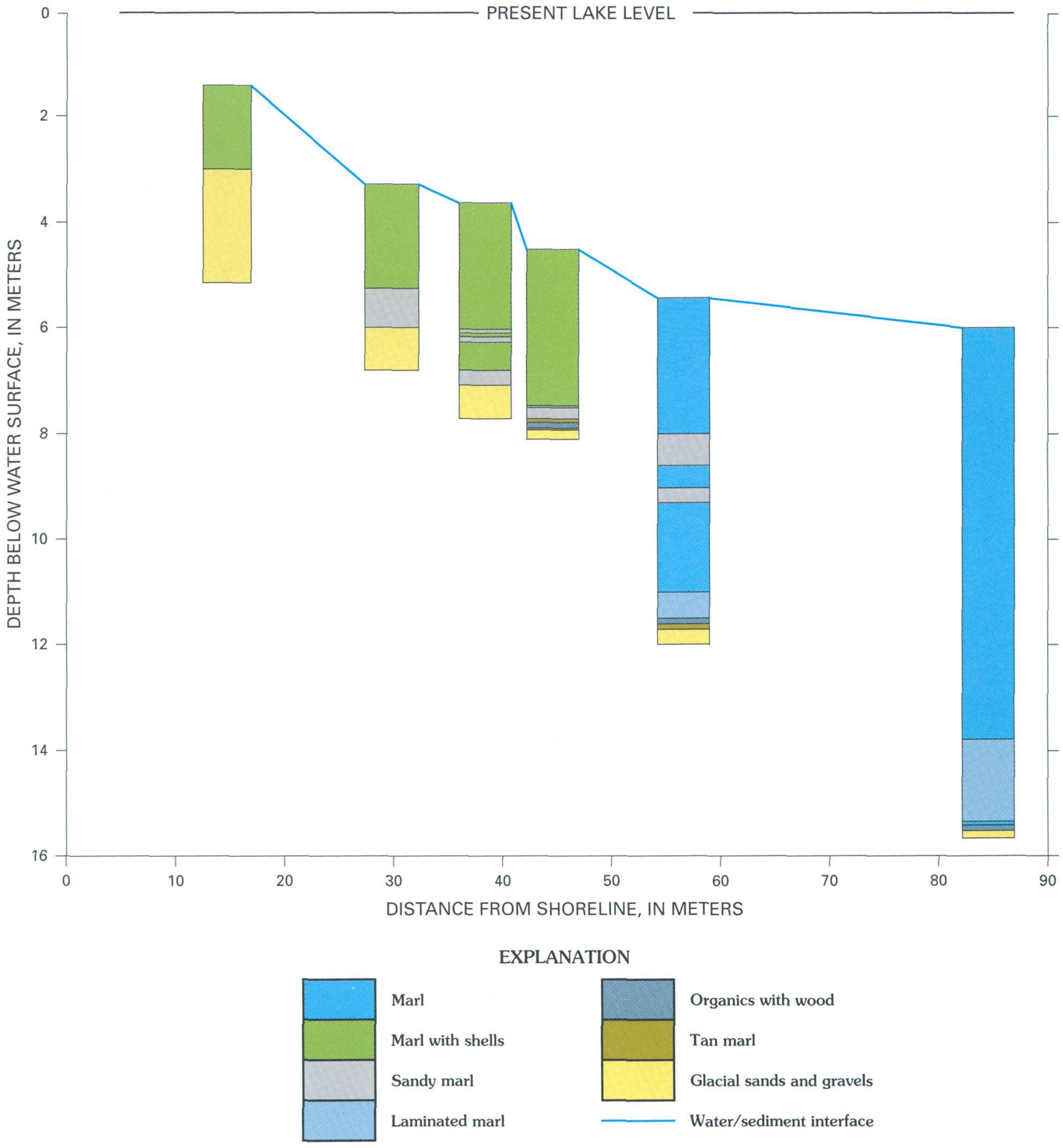

Figure 89. Sediment stratigraphy of Shingobee Lake for a transect taken from the southeast shore of the lake.

equilibrium with the isotopic composition of the lake water during the time of precipitation of the carbonate. Analysis of detrital carbonates from the catchment, which are often incorporated in bulk sediment, is thus avoided. Ostracodes are also valuable bioindicators, and their abundances and combined assemblage throughout a sediment core are indicators of some environmental parameters, such as the oxygen supply or the organic carbon content in the bottom waters.

In the Williams Lake core, ostracodes are limited to minerogenic sediments from the spruce and lower prairie zones. The main species are *Candona* spp., mostly *Candona rawsoni*, and *Cytherissa lacustris*. The loss of ostracodes in the prairie zone is coincident with the abrupt sediment change from minerogenic to organic. Probably the lake reached a threshold depth at which time it began to stratify in the summer, creating an anoxic hypolimnion allowing organic accumulation and carbonate dissolution. Ostacodes in the Shingobee Lake core are present throughout the entire Holocene section, but in low abundance in the lower and middle Holocene section. *Candona ohioensis* and *Candona acuta*, both indicators of freshwater lakes with a low ionic concentration, are the dominant ostracodes in Shingobee Lake.

The ostracode carbonate in sediments of Williams and Shingobee Lakes have distinct isotopic populations (Schwalb and others, 1995). The isotopic compositions also show a very distinct evolution through time that is closely linked to the different pollen zones. The isotope geochemistry indicates that before 9,800 yr B.P. the two lakes were connected. Increased evaporation through the jack/red pine period, as indicated by an increase in $\delta^{18}O$ values in Williams Lake ostracodes, led to lower lake levels, leaving small separated basins. The oxygen isotopic composition of Shingobee Lake ostracodes remains relatively unchanged through the prairie period; thus, there is no evidence for a significant drop in the level of Shingobee Lake.

CONCLUSIONS

At Williams Lake, long-term changes in lake level and chemical deposition are primarily the result of regional climate change and basin infilling. At Shingobee Lake, the lake level and sediment deposition has been more consistent, a result of the consistent open-basin hydrology that has tempered the effects of long-term climate changes.

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Stratigraphy and Biological Characteristics of Sediments in Little Shingobee Fen

By Virginia Carter, Sharon M. Locke, Ted A. McConnaughey, and Thomas C. Winter

INTRODUCTION

Wetlands in glaciated terrain commonly are formed by the gradual filling of lakes and shallow ponds with organic sediment and marl. As the lakes become shallower, plants encroach on the littoral zone, often forming floating mats, and add to the accumulation of organic matter. Eventually vegetation may fill the major part of the basin. Wetlands also form where geologic strata or breaks in slope force ground water to discharge to land surface continuously; organic soils accumulate in these wetlands and the vegetation resembles that in wetlands formed by lake filling. A fen is a minerotrophic wetland deriving its water from relatively mineral-rich ground and(or) surface water and typified by a pH greater than 6 and calcium concentrations greater than 350 µeq/L (Gorham and others, 1985).

Little Shingobee Fen surrounds Little Shingobee Lake (fig. 90), which drains northward into Shingobee Lake (pl. 1). A small stream originates at the south end of the open fen and flows north and west through the forested fen to form the major surface drain of the fen into Little Shingobee Lake. A smaller stream drains to the lake from the north part of the forested fen. Little Shingobee Fen is a typical northern fen with organic soils and characteristic fen vegetation (Carter and others, this volume). Calcium and bicarbonate constitute a large percentage of the dissolved constituents in Shingobee and Little Shingobee Lakes and adjacent ground-water systems (LaBaugh, this volume). In this high carbonate environment, littoral zone or shallow lake sediments are primarily marls, whereas deep lake organic sediments consist of loose organic flocculates and gyttja. Large numbers of shelled invertebrates and precipitation of calcium carbonate by plants such as *Chara* sp. during photosynthesis contribute to the accumulation of marl. These organic and marl deposits are reflected in the stratigraphy of the fen.

One of the major objectives of the research in the Shingobee River Headwaters Area is to understand and quantify carbon cycling. Part of the process for developing carbon budgets is identification and quantification of carbon reservoirs, one of which is the organic or carbonate-rich soils in watershed wetlands. Cores taken of wetland sediments and underlying sands or clays provide a biological, chemical, and physical record of wetland development. If wood is present toward the base of the core, it may be possible to date the peat deposit. In addition, these cores provide information on the depth of peat and marl underlying the wetland and can serve as the basis for estimates of peat and marl volumes.

Ground water moving through wetland sediments may undergo changes in constituent chemistry; the water chemistry of the incoming ground water may differ significantly from that of the water discharging from the fen to Little Shingobee Lake. Stratigraphic profiles of the wetland sediments can be compared with location and depths of wells and piezometers to improve understanding of the hydrology and water chemistry of the sampled strata.

The major hypothesis for this research is that ground-water hydrology and biogeochemistry of the fen and Little Shingobee Lake are functions of the stratigraphy of the organic and marl sediments underlying the vegetation. In addition, the wetland soils are one of the important carbon reservoirs in the watershed. The purpose of this research is to document the stratigraphy of the fen soils and estimate the carbon storage.

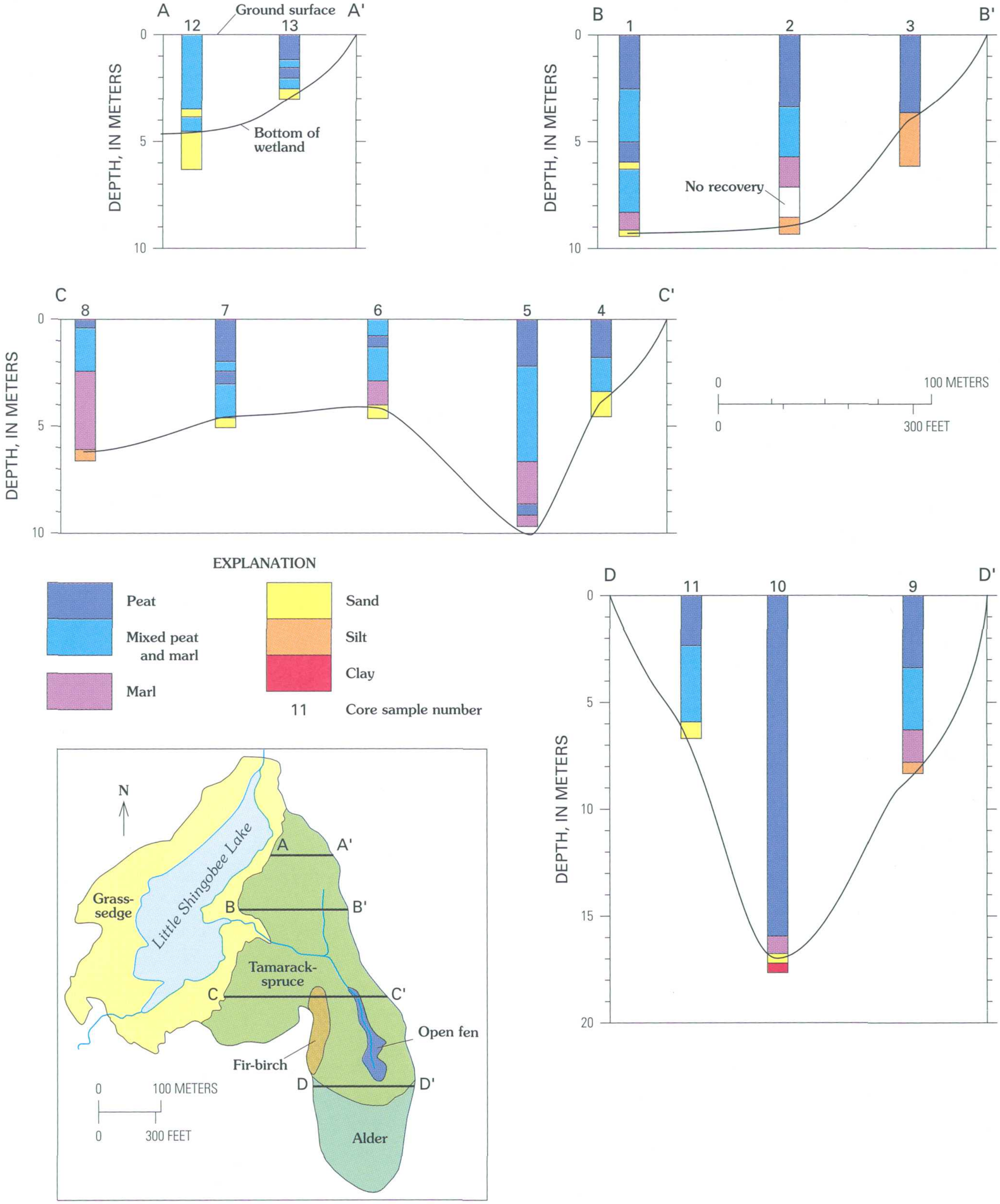

Figure 90. Major vegetation communities of Little Shingobee Lake and Little Shingobee Fen, locations of core transects, and stratigraphy for each core site.

METHODS

Four east-west transects were made across the forested part of the fen from upland to the edge of emergent wetland or from upland to upland (fig. 90). A piston corer was used to take 1-m cores from the surface of the fen down to the underlying sand, silt, or clay (fig. 91). Each core was measured and described in the field (fig. 92). Selected samples of soil and wood fragments were taken for further examination and stored in labeled plastic bags in a cold room at 4°C. Eleven samples from a single core and basal samples from five other cores were prepared for pollen analysis using standard methods modified from Faegri and Iversen (1975). Pollen analysis was used to determine a general chronology for fen sediments and timing of the onset of peat formation.

RESULTS

The stratigraphic profiles for each transect are shown in figure 90. The transects extend only to the edge of the emergent wetland, so the preliminary interpretation is for the forested area. All cores, except for cores 6 and 12, had peat at the top of the profile, which was underlain by alternating layers of peat and marl or mixed peat and marl and, near the bottom of the profile in some cases, by marl with very little organic content. The base of the profile was sand, silt, or clay. The wetland basin is widest at transect C-C', but the deepest core (17.5 m) was taken in the center of transect D-D' at the south edge of the open part of the fen. Little Shingobee Lake is 5-m deep at the south end and shallows rapidly toward the north. The fen is narrowest and the organic soils are thinnest at the north end (transect A-A'). The underlying basin soils slope toward the lake at sections A-A' and B-B'. They are deepest in the center of transects C-C' and D-D'. This suggests that the fen resulted from filling of an ancient lake basin and that the southern end of the basin was intersected by a north-south ridge just west of the fir-birch area of the fen. The deepest part of the lake basin may be the present site of Little Shingobee Lake.

The pollen stratigraphy for the westernmost core of transect C-C' (fig. 93) shows regional vegetation changes associated with long-term climate change and the local vegetational succession resulting from development of the fen. Four pollen assemblage zones were identified on the basis of percentage changes of the dominant regional taxa. A piece of wood retrieved from core 11 from a depth of 7.6 to 8 m was radiocarbon-age dated at 10,345 ± 115 years before present (BP) (Herbert Haas, Southern Methodist University, written commun., 1994).

Figure 91. Extraction of core sample with piston corer.

An approximate chronology for sedimentation can be determined by correlating the vegetation zone boundaries at the fen (fig. 93) to those determined for Williams Lake (Locke and Schwalb, this volume). A *Chara* marl with abundant shells was deposited throughout the early Holocene (10,350-7,600 years BP), when the described core site was located in a relatively shallow portion of the lake basin. Pollen data show that the surrounding upland was forested throughout the early Holocene but shifted from spruce to jack/red pine forest at about 9,800 years BP as the climate became slightly warmer and wetter. During the mid-Holocene (7,700-4,000 years BP), the climate was warm and dry, and the regional upland vegetation consisted of prairie and oak savanna. Mid-Holocene sediments are more organic rich but still contain shells. Water depth at the site was probable shallower than during the early Holocene, which was the result of both sediment infilling and a lowering of the regional water table. At about 4,000 years BP, climate again became cooler and wetter and upland vegetation shifted to the modern white pine forest. At 110-cm depth in core 8, the pollen assemblage is characterized by high percentages of sedge and aquatic pollen, indicating that this part of the basin was shallow enough to allow the development of a marsh. The beginning of peat accumulation (50-cm depth in the core) is marked by an increase in tamarack, spruce, and *Sphagnum* percentages, indicating the transition to the current forested fen environment.

Figure 92. Examination of extracted core sample.

Previous studies have shown that the development of peat mats around pond margins occurred after 4,000 years BP in northern and central Minnesota (Janssen, 1967; Wright and Watts, 1969). A rise in water levels following the end of the warm, dry, mid-Holocene climatic period apparently allowed the spread of wetland species over previously exposed mudflats in shallow areas of lake basins (Cushing, 1963). Further stratigraphic analyses are needed to determine if Little Shingobee Fen fits this model. Sediment chronology and the timing of peat formation should be determined for other parts of the basis. In particular, the location of the thickest accumulation of peat needs to be identified, and woody material from the base of the peat needs to be dated. In addition, the sediments of Little Shingobee Lake and Little Shingobee Fen should be closely examined for evidence of a water-level lowering in the mid-Holocene. A sand layer in the westernmost cores of transects A-A' and B-B' is possible evidence for a water-level change, but the pollen stratigraphy is unknown.

More information on present geochemical processes would clarify the effects of organic sediments on surface-water chemistry in Little Shingobee Fen. Paleohydrology studies would be useful for understanding the evolution of this ecosystem. To complete the information on the size of the carbon reservoir, it will be necessary to determine the exact boundary and volumetric carbon content of the fen. Additional cores need to taken on the west side of Little Shingobee Lake.

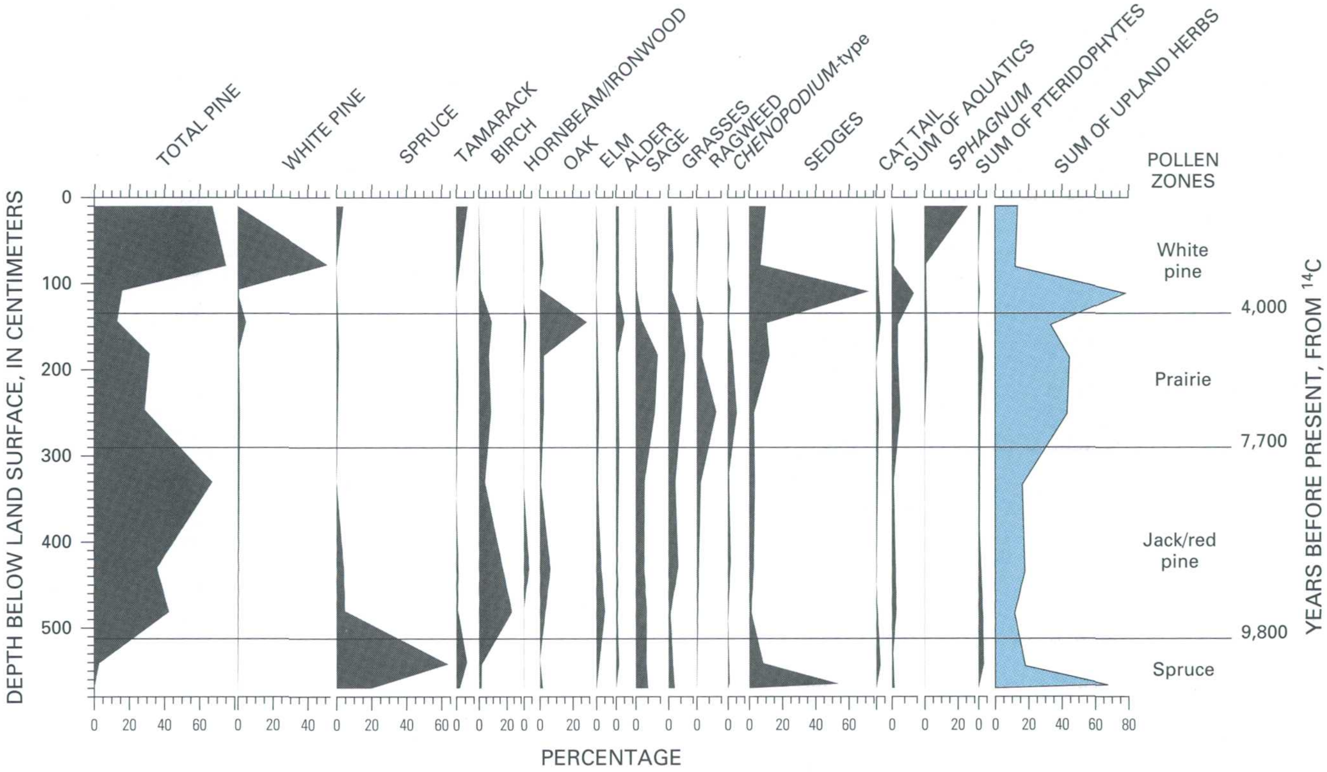

Figure 93. Pollen stratigraphy of selected taxa in core from site 8 on transect C-C'. Dates for vegetation zone boundaries are based on correlation with Williams Lake (Locke and Schwalb, this volume) and a radiocarbon date on wood obtained from basal sediments at Little Shingobee Fen.

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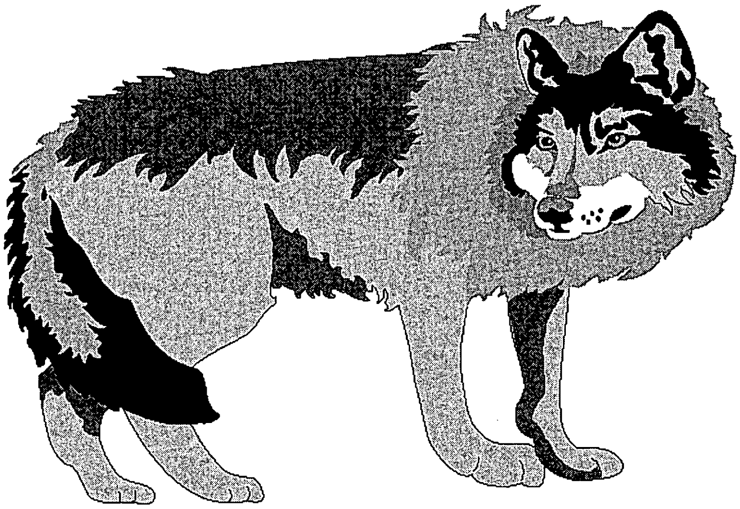

Summary and opportunities for Additional Research in the Shingobee River Headwaters Area

By Thomas C. Winter

INTRODUCTION

The Interdisciplinary Research Initiative was established in 1989 to address issues related to aquatic ecosystems. The underlying premise was that unbiased information on the fundamental physical, chemical, and biological interactions within ecosystems is needed by managers of aquatic resources. Furthermore, the most efficient and effective way to focus on the interactions among the many components of ecosystems is to conduct research at a common field site. The Shingobee River Headwaters Area (SRHA) was selected because it has many different types of aquatic environments within a single compact watershed, and it contains lakes having greatly different residence times. The aquatic environments in the watershed include streams, lakes that interact with streams, lakes that do not interact with streams, riparian wetlands, wetlands that occupy depressions, and wetlands formed and maintained by ground-water discharge. The research conducted thus far has focused partly on carbon budgets of Williams and Shingobee Lakes, the Little Shingobee Fen, and hyporheic exchange with the Shingobee River. However, the research has evolved to encompass studies of how the integration of atmospheric, surface, and ground water sets the stage for chemical and biological processes in a variety of aquatic systems within the watershed. The purpose of this chapter is to provide an overview of the findings to date and note where opportunities exist to address questions that have arisen or remain unanswered.

PHYSIOGRAPHIC SETTING

The SRHA is underlain by more than 120 m of glacio-fluvial sediments consisting of clay, silt, sand, and gravel and glacial till. Glaciers that shaped the present landscape moved into the area from at least two different glacial ice sources more than 10,000 years ago (Mooers and Norton, this volume). Surficial deposits range from sand and gravel in the southern part of the watershed, near Crystal and Williams Lakes, to silt and clay in the northern part, near Shingobee Lake. Throughout the watershed, the surficial deposits gradually become coarser with depth (Winter and Rosenberry, this volume).

Geologic mapping and studies of paleoecology indicate that Williams Lake may have been about 6 m higher than its present level (Seaburg, 1985) and 4.5 m lower than its present level (Locke and Schwalb, this volume) during the past 10,000 years. Glacial lakes were common south of the Itasca Moraine as the glacial ice receded (Mooers and Norton, this volume), and much of the SRHA may have been covered by one of these lakes prior to about 9,000 years ago.

Test holes drilled to depths greater than 100 m at 6 locations in the SRHA indicate the complexity of the geologic framework typical of glacial terrain. More test drilling and geophysical surveys are needed to define the three-dimensional distribution of the various glacial deposits so ground-water flow paths and geochemical reactions along those flowpaths can be determined. In addition, more mapping of surficial geology and soils is needed to understand the evolution of the landscape, such as the conditions that caused the formation and drainage of the glacial lake that covered the SRHA before 9,000 years ago.

INTERACTIONS BETWEEN ATMOSPHERIC WATER AND SURFACE WATER

From 1982 to 1986, evaporation from Williams Lake was determined by the energy-budget method.

That study indicated that evaporation can vary by as much as 30 percent from year to year (Sturrock and others, 1992). The evaporation research at Williams Lake included evaluation of surrogate data to determine energy budgets (Rosenberry and others, 1993) and evaluation of eleven other methods of estimating evaporation (Winter and others, 1995). Data presented herein extend the research on evaporation, including the exchange of other gases between Williams Lake and the atmosphere. For example, Stannard and others (this volume) made eddy-correlation measurements of evaporation for comparison with energy-budget measurements over short time periods. Results of this comparison indicated that one of the problems with using the energy budget for periods of a day or less is related to accurate determination of the heat content of the lake. A few examples of research needed for increased understanding of evaporation include: (1) methods need to be developed to more accurately measure heat storage automatically so that it could be calculated for short time periods, (2) sensors for measuring net-radiation need to be evaluated and compared to the models used for calculating outgoing radiation, and (3) areal variability of the energy fluxes of a lake needs to be investigated using multiple eddy-correlation and/or Bowen ratio installations coupled with a lake circulation model for the storage term.

Hostetler (this volume) described the utility of a model that uses climate data as input for predicting the seasonal and annual temperature distribution within Williams Lake. These investigations need to be extended to include modeling of limnological characteristics, such as the interactions of dissolved oxygen with the atmosphere and with biogeochemical processes within lakes.

To address the carbon budget of lakes, one of the primary goals of the IRI effort, Anderson and others (this volume) measured a large flux of carbon dioxide from the lake immediately following ice melt. Measurements made several times over a 3-year period indicated considerable inter- and intra-year variation in flux. They also compared several models of predicting carbon dioxide flux with direct measurements. Striegl and others (this volume) indicated that methane emissions from Williams and Shingobee Lakes were high immediately following ice melt and were greatest during summer. Although Williams Lake has received the greatest attention, comparative studies between the closed lake (Williams) and the open lake (Shingobee) indicate that on an annual basis Shingobee Lake is a net source of carbon dioxide to the atmosphere, while Williams Lake is approximately in balance (Striegl, personal communication). Exact reasons for the differences are not known but appear to be related to the relative importance of hydrologic inputs and outputs of carbon at the two lakes. In addition, the interactions between dissolved gases and organisms in these two types of lakes need better description to understand and predict the role of lakes in carbon fluxes to and from the atmosphere. Exchange between the atmosphere and surface waters has focused on the exchange of water vapor and gases for Williams and Shingobee Lakes, but little has been done on atmospheric exchange in wetlands or streams.

Input of chemicals from the atmosphere to the SRHA provide an important component of the chemical budgets of Williams and Shingobee Lakes (Reddy and others, this volume). Chloride, nitrogen, and sulfur contribute as much as 10 percent, and calcium and magnesium as much as 20 percent, of the total amount of chemical input from the atmosphere. These studies need to be extended to include transformations of atmospheric input of chemicals before they reach the surface-water and ground-water systems. In addition, details about the role of dry deposition in the input of chemicals from the atmosphere to the watershed are uncertain.

INTERACTIONS BETWEEN GROUND WATER AND SURFACE WATER

The complex geologic framework of the SRHA provides for a number of research opportunities related to understanding infiltration of precipitation, movement of water through the unsaturated zone, ground-water recharge in a wide variety of soils and depths to the water table, and discharge into surface water. Furthermore, these processes could be studied in different types of land use areas within the watershed, such as coniferous forest, deciduous forest, pasture, irrigated cropland, and a small urban area. The great thickness of glacial deposits provides the opportunity to investigate ground-water flow systems of different magnitude. For example, the wide variety of particle sizes and mineral types and the complex interlayering of fluvial and till units within these thick deposits provide the opportunity to investigate flow paths, geochemical-reaction processes, and chemical transport.

Lakes

Williams Lake was selected for study because it lies between a higher lake (Crystal) and a lower lake (Mary); therefore, it was expected to be a flow-through lake with respect to ground water. Subsequent field work documented that the lake received ground water on the side toward Crystal Lake and lost water to ground water on the side toward Mary Lake. Furthermore, numerical modeling studies indicated that the ground-water flow system interacting with Williams Lake is restricted to the upper part of the surficial aquifer (Siegel and Winter, 1980). Models of regional ground-water flow need to be extended to include the entire area from Crystal Lake to the Itasca Moraine, north of Shingobee Lake.

Recent studies of the interaction of lakes and ground water have concentrated on the highly dynamic flow conditions near the edges of lakes. For example, ground water is recharged quickly where the unsaturated zone is thin near the shore of a lake. This focused recharge, which can cause temporary increases of ground-water flow on the inflow sides of lakes or temporary inflow of ground water on the outflow sides of lakes, was documented at Williams Lake (Rosenberry, 1985). In this same near-shore environment, roots of terrestrial plants commonly extend into the ground-water system, which can result in transpiration rates great enough to draw down the water table and induce seepage out of lakes. The effect of these two processes can result in highly variable seepage rates and directions in the littoral zone of lakebeds on seasonal and annual time scales.

It is hypothesized that biological communities and biogeochemical processes in the near-shore littoral zone reflect these variable seepage conditions. Therefore, many opportunities exist for detailed biogeochemical studies of the littoral zones of lakes in the SRHA, especially because of the greatly different ground water and lake interactions between Williams and Shingobee Lakes. The data from pore-water peepers presented by Kendall and others (this volume) indicate the complex biogeochemical environment in the upper part of sediments in the near-shore zone of lakes. This environment in lakebeds is analogous to the highly variable biogeochemical environment in the hyporheic zone of rivers, such as presented by Duff and others (this volume) for the Shingobee River. The SRHA provides an ideal field setting to develop understanding of the way flow affects the chemical and biological processes in sediments of flowing versus ponded surface-water bodies. Processes occurring in lake sediments as water moves to and from lakes can be compared with processes occurring in stream sediments as water moves to and from the streambed.

Wetlands

The SRHA has a wide variety of wetland types, including those in land-surface depressions, such as Chase Marsh; riparian, such as along the Shingobee River; fen, such as Little Shingobee Fen; and those occupying the perimeter of lakes and former bays of lakes. The altitudes of Chase Marsh and the contiguous water table have been monitored using continuous recorders for many years to evaluate focused ground-water recharge associated with land-surface depressions. The opportunity exists for studies of water and chemical transport through the unsaturated zone as well as biogeochemical studies of a depressional wetland.

Little Shingobee Fen is a particularly intriguing wetland because of the large quantity of water flowing from it relative to its small drainage area. The fen also has a number of rare plants (Carter and others, this volume), the ecology of which need to be better defined. The preliminary work reported by Puckett and others (this volume) indicates that the fen may be receiving a large part of its inflow from a regional ground-water flow system. The fen is one of the three principal focal points of the IRI effort; therefore, a number of hypotheses are being tested with respect to the source of water to the fen. Additional work in recent years has included constructing piezometer nests along two transects across the fen that extend much deeper than those used by Puckett and others. The deepest piezometer at each site is completed in the mineral sediments underlying the fen. Although samples for chemical analyses are being taken from these piezometers, much additional work is needed to complement the studies of the fen that are in progress. For example, hydraulic characteristics, flow paths through the peat, and geochemical processes within the peat need to be determined. In addition, the flora

(Carter and others, this volume), fauna, and bacteria (Robbins and others, this volume) need to be related to the unusual hydrologic environment of the Little Shingobee Fen.

The entire Little Shingobee Lake area, including the Little Shingobee Fen, is of major interest in the carbon budget of the SRHA because of the large amount of organic sediments that are present -- essentially carbon in storage. Determination of the volume of carbon in storage through coring needs to be extended from the fen (Carter and others, this volume) across the entire SRHA.

How the geologic framework and ground-water flow systems control the chemical and biological processes in the variety of surface waters found in the SRHA is one of the more interesting questions arising from the IRI studies. The original concept of comparing lakes with different residence times has expanded to include wider examination of the role of hydrogeologic setting in the control of chemical and biological processes throughout the watershed. Insights gained from the focus on the carbon budgets of Williams and Shingobee Lakes may enhance understanding of why so much organic material has accumulated in the Little Shingobee Fen.

Streams

The Shingobee River has received considerable attention for studies of the hyporheic zone, which is the zone in streambeds and banks where surface water moves repeatedly into and out of the sediments along relatively short flow paths. The Shingobee River was selected for study because (1) its setting is different from the high-gradient, cobble-bed mountain streams that have been the focus of most research on the hyporheic zone, (2) the channel is in an area of large upward ground-water gradients, and (3) dissolved solutes in ground-water discharge are modified as a result of biotic and abiotic processes in the hyporheic zone, which may affect the seasonal water chemistry of the stream inflows to Shingobee Lake. Hyporheic exchange along the Shingobee River penetrates to depths ranging from 0 to 30 cm, depending on streambed characteristics and magnitude of ground-water discharge (Duff and others, this volume; Jackman and others, this volume).

Emphasis on biogeochemical processes in both the channel and hyporheic zone have been directed toward nitrogen uptake and transformation. Whole-stream injections of nitrate and ammonium in conjunction with a conservative tracer have been made to address dissolved inorganic nitrogen processes at the reach scale. Enzyme assays for organic nitrogen mineralization, nitrification, and denitrification potentials and closely-spaced porewater solute profiles in the streambed have been measured seasonally at sites having different vertical hydraulic gradients to identify and quantify inorganic nitrogen cycling processes in the hyporheic zone. Results of these efforts have provided an extensive seasonal database of surface-water and porewater chemistry. The database on porewater chemistry of the Shingobee River provides a foundation for future studies of other solutes that are interactive in geochemical processes, both biotic and abiotic, in the hyporheic zone. For example, the hyporheic zone appears to be a sink for stream-derived sulfate during all seasons; however, sulfate reduction along the study reach has not been quantified. The mechanism for retention of sulfate, whether biotic, abiotic, or both, remains to be investigated. The hyporheic zone also is a zone of iron and methane oxidation -- current data indicate that soluble reactive phosphorus (SRP) in stream water may be controlled by interaction with oxidation of ferrous iron.

Submerged macrophytes seem to dominate biological production during summer. However, primary production has not been estimated for any season, and the relative role of macrophytes versus epilithic periphyton is yet to be determined. Particulate organic matter (POM) seems to be an important source of organic carbon to Shingobee Lake; however, its concentration, and how it may vary within daily or seasonal cycles, is unknown. Dissolved organic carbon (DOC) has a different concentration and molecular structure in surface water than in ground water, and little is known about DOC sources. The composition of the invertebrate community and its relationship to available organic matter and nutrients has never been examined.

Bankside ground-water seeps are a common characteristic of the Shingobee River valley (Triska and others, this volume). The source of water to the seeps is related directly to the question of the source of water to the Little Shingobee Fen. Preliminary chlorofluorocarbon data indicate that both may be receiving ground water from the same regional flow system. Therefore, integrated studies are needed to define the ground-water flow paths and the geochemical processes along those flow paths for the entire lower (northern) part of the SRHA. Triska and others (this volume) have described the geochemical characteristics of one of the bankside seeps in the Shingobee River valley and its relationship to the Shingobee River. Recently this data set has been expanded to seasonal resolution. Robbins and others (this volume) found unique bacteria in one of the seeps that may have implications for understanding the formation of Precambrian iron ore. Further study of these unique deposits and environmental conditions could have far-reaching implications for understanding aquatic ecosystems as well as for economic geology.

COMPARISON OF SURFACE WATERS IN DIFFERENT HYDROLOGIC SETTINGS

The original impetus of the Interdisciplinary Research Initiative in the SRHA was to do comparative studies of lakes that have no streamflow in or out, such as Williams Lake with lakes that have stream inflow and outflow, such as Shingobee Lake (Averett and Winter, this volume). One result of the studies is the recognition that the geologic framework and hydrologic systems, which result from the integration of atmospheric, surface and ground waters, play an important role in chemical and biological processes in surface waters in the watershed (Rosenberry and Winter, this volume).

The greatest hydrological distinction between the two lakes is the fact that one interacts with streams and the other does not. The inorganic characteristics of water in Williams Lake reflect input from ground water and rainfall, whereas the greatest input to Shingobee Lake is from the Shingobee River. Concentrations of the major ions such as calcium, magnesium, and bicarbonate in Shingobee Lake are approximately twice the concentrations in Williams Lake, but they are similar to concentrations in shallow ground water (LaBaugh, this volume). Although a limited effort has gone into studies of trace metals these data indicate that much more needs to be done to characterize the relationship of trace metals to hydrogeologic setting (van Geen, this volume).

Shingobee Lake also has nearly twice the concentration of total phosphorus than in Williams Lake, which reflects the input of phosphorus from the Shingobee River. In contrast, concentration of total nitrogen is similar in the two lakes. The difference in plant nutrients, such as phosphorus, may account for the fact that the algal biomass in Shingobee Lake is nearly double the amount in Williams Lake (LaBaugh, this volume). Primary production, which is the uptake of inorganic carbon by algae, also is greater in Shingobee Lake than in Williams Lake (Cole and others, this volume), but loss rates in terms of particulate carbon are also greater in Shingobee Lake than in Williams Lake (Cloern and others, this volume). More research is needed to determine if the increased algal biomass and primary production in Shingobee Lake is related to supply of nutrients to the lake by the Shingobee River or if other factors are involved.

One of the goals of the IRI studies is to understand how hydrological conditions and processes are related to differences in the carbon dynamics of surface waters in the watershed. The importance of hydrological processes in the cycling of carbon is evident in the preliminary $\delta^{13}C$ data (Kendall and others, this volume). The amounts of carbon in storage as well as rates of transfer of carbon have begun to be documented. Although the lakes differ in algal biomass, the amount of carbon in storage in the form of rooted aquatic plants were similar in the middle of summer (Carter and others, this volume). Losses of particulate carbon from the middle of the lake to the sediments were greater in Shinogbee Lake than in Williams Lake (Cloern and others, this volume). The deep-water sediments in Shingobee Lake contain abundant calcium carbonate, which is absent from the deep-water sediments of Williams Lake (Dean and Bradbury, this volume). Much calcite is produced by rooted aquatic plants in Williams Lake (McConnaughey and others, this volume), which retains most of the annual input of calcium (LaBaugh, this volume). Carbon can also be lost to the atmosphere because of emission of carbon dioxide, as in Williams Lake (Anderson and others, this volume); however, loss of carbon in the form of methane is similar in the two lakes (Striegl and Michmerhuizen, this volume). Decompositon of material in the sediments is one source of carbon-containing gases that escape from the lakes, which makes the contrast between the relatively abundant organic matter in the sediments of Williams Lake and the relative lack of organic matter in the sediments of Shingobee Lake a particularly intriguing question. This difference in the content of the sediments is also interesting in light of the fact that the concentration of dissolved organic carbon in lake water is much greater in Williams Lake than in Shingobee Lake (Aiken and McKnight, this volume).

While the main thrust of the initial effort in the SHRA has been on the relation of hydrological, chemical, and biological processes, the identity of the biota doing the processing is also important. The phytoplankton communities of the lake are quite different (LaBaugh, this volume; Cole and others, this volume), and these differences are reflected in algal remains in the sediments (Dean and Bradbury, this volume). The fact that Williams Lake retains much of the annual input of silica is evident in the assemblage of diatoms in its sediments, which include planktonic as well as benthic, attached, and motile species. The benthic, attached, and motile diatoms are most common in intermediate to shallow depths, especially on the ground-water inflow side of the lake where they must have an effect on the cycling of silica, carbon, phosphorus and other elements (Dean and Bradbury, this volume). Planktonic diatoms are the most abundant of the diatom assemblages in the sediments of Shingobee Lake. The importance of biota in processing chemicals moving into and from sediments is also evident from studies in the Shingobee River where the nitrogen cycle is affected by the presence of rooted aquatic vegetation (Triska and others, this volume).

Despite the differences in algae between the lakes, the species of zooplankton that consume the algae are similar. The fish communities that use these food resources also are somewhat similar (Striegl and others, this volume). All of these organisms are involved in the transfer of carbon within the lakes. The fact that fish growth is faster in Shingobee Lake than in Williams Lake may simply reflect differences in primary productivity. However, results presented herein are for a short period of time. The seasonal and annual dynamics of carbon transfer and the relation of that transfer to the size distribution of the plankton and the food habits and composition of the fish community need to be addressed. In addition, studies of carbon and energy transfer need to include evaluation of the effects of physical factors, such as wind and temperature. The material that ultimately is deposited in the sediments is a by product of production and transfers within the biota and, to a lesser extent, geochemical processes in the lake water. Only additional study can determine if the seasonal patterns and magnitudes of sedimentation in the two lakes are preserved in their sedimentary record.

The present-day sediment characteristics of Williams and Shingobee Lakes are distinctly different, to the extent that they are end-members among lakes in the upper Midwest with respect to the form of carbon in their sediments (Dean and Bradbury, this volume). The sediments of Williams Lake contain high concentrations of organic matter and almost no calcium carbonate. In contrast, the sediments of Shingobee Lake contain much lower concentrations of organic matter but abundant calcium carbonate. Water-chemistry budgets for Williams Lake (LaBaugh, this volume) indicate that calcium and silica are both retained in the lake through biologically mediated removal of calcium carbonate and biogenic opal. However, study of the sediments indicate that although silica is retained in Williams Lake as remains of planktonic and benthic diatoms, the recent sediments are almost completely free of calcium carbonate (Dean and Bradbury, this volume).

The carbon budgets of Shingobee Lake and especially of Williams Lake remain one of the largest unsolved problems in the IRI effort. More closely coordinated investigations of the water and sediment columns in the lakes and wetlands in the area are needed. Sediment- and water-flux studies using sediment traps and benthic-flux chambers would be particularly useful in these investigations. Growth-rate studies of benthic diatoms and other algal communities are necessary to track the removal of Si, C, P, and other elements at intermediate depths in the lakes.

Studies to date have concentrated on Williams and Shingobee Lakes, Shingobee River, and Little Shingobee Fen. Other lakes and wetlands in the SRHA and reaches of the Shingobee River upstream of Steel Lake have had only a few samples taken for chemical analyses. These surface water features, as well as much of the terrestrial landscape, provide potential research sites for studies that could complement the studies underway.

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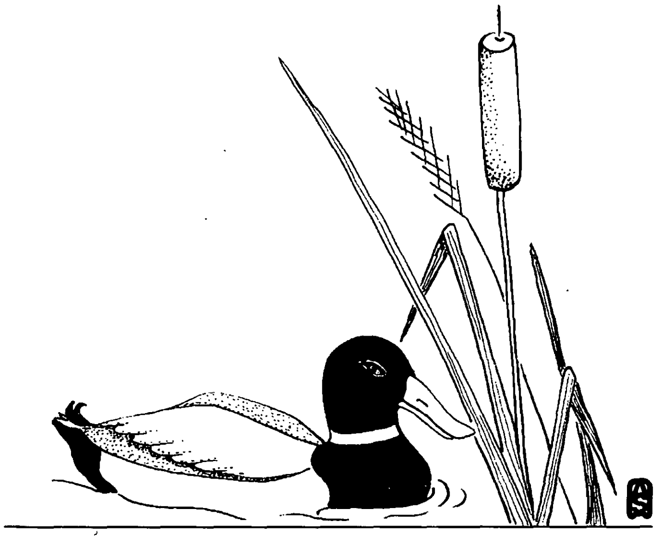

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CONVERSION FACTORS

Multiply	By	To obtain
	Length	
inch (in.)	2.54	centimeter
inch (in.)	25.4	millimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
mile, nautical (nmi)	1.852	kilometer
yard (yd)	0.9144	meter
	Area	
acre	4,047	square meter
acre	0.4047	hectare
acre	0.4047	square hectometer
acre	0.004047	square kilometer
square foot (ft^2)	929.0	square centimeter
square foot (ft^2)	0.09290	square meter
square inch (in^2)	6.452	square centimeter
section (640 acres or 1 square mile)	259.0	square hectometer
square mile (mi^2)	259.0	hectare
square mile (mi^2)	2.590	square kilometer
	Volume	
barrel (bbl), (petroleum, 1 barrel = 42 gal)	0.1590	cubic meter
ounce, fluid (fl. oz)	0.02957	liter
pint (pt)	0.4732	liter
quart (qt)	0.9464	liter
gallon (gal)	3.785	liter
gallon (gal)	0.003785	cubic meter
gallon (gal)	3.785	cubic decimeter
million gallons (Mgal)	3,785	cubic meter
cubic inch (in^3)	16.39	cubic centimeter
cubic inch (in^3)	0.01639	cubic decimeter
cubic inch (in^3)	0.01639	liter
cubic foot (ft^3)	28.32	cubic decimeter
cubic foot (ft^3)	0.02832	cubic meter
cubic yard (yd^3)	0.7646	cubic meter
cubic mile (mi^3)	4.168	cubic kilometer
acre-foot (acre-ft)	1,233	cubic meter
acre-foot (acre-ft)	0.001233	cubic hectometer
	Flow rate	

Multiply	By	To obtain
acre-foot per day (acre-ft/d)	0.01427	cubic meter per second
acre-foot per year (acre-ft/yr)	1,233	cubic meter per year
acre-foot per year (acre-ft/yr)	0.001233	cubic hectometer per year
foot per second (ft/s)	0.3048	meter per second
foot per minute (ft/min)	0.3048	meter per minute
foot per hour (ft/hr)	0.3048	meter per hour
foot per day (ft/d)	0.3048	meter per day
foot per year (ft/yr)	0.3048	meter per year
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
cubic foot per second per square mile [(ft^3/s)/mi^2]	0.01093	cubic meter per second per square kilometer
cubic foot per day (ft^3/d)	0.02832	cubic meter per day
gallon per minute (gal/min)	0.06309	liter per second
gallon per day (gal/d)	0.003785	cubic meter per day
gallon per day per square mile [(gal/d)/mi^2]	0.001461	cubic meter per day per square kilometer
million gallons per day (Mgal/d)	0.04381	cubic meter per second
million gallons per day per square mile [(Mgal/d)/mi^2]	1,461	cubic meter per day per square kilometer
inch per hour (in/h)	0.0254	meter per hour
inch per year (in/yr)	25.4	millimeter per year
mile per hour (mi/h)	1.609	kilometer per hour
	Mass	
ounce, avoirdupois (oz)	28.35	gram
pound, avoirdupois (lb)	0.4536	kilogram
ton, short (2,000 lb)	0.9072	megagram
ton, long (2,240 lb)	1.016	megagram
ton per day (ton/d)	0.9072	metric ton per day
ton per day (ton/d)	0.9072	megagram per day
ton per day per square mile [(ton/d)/mi^2]	0.3503	megagram per day per square kilometer
ton per year (ton/yr)	0.9072	megagram per year
ton per year (ton/yr)	0.9072	metric ton per year
	Pressure	
atmosphere, standard (atm)	101.3	kilopascal
bar	100	kilopascal
inch of mercury at 60 °F (in Hg)	3.377	kilopascal
pound-force per square inch (lbf/in^2)	6.895	kilopascal
pound per square foot (lb/ft^2)	0.04788	kilopascal
pound per square inch (lb/in^2)	6.895	kilopascal

Multiply	By	To obtain
Density		
pound per cubic foot (lb/ft^3)	16.02	kilogram per cubic meter
pound per cubic foot (lb/ft^3)	0.01602	gram per cubic centimeter
Energy		
kilowatthour (kWh)	3,600,000	joule
Radioactivity		
picocurie per liter (pCi/L)	0.037	becquerel per liter
Specific capacity		
gallon per minute per foot [(gal/min)/ft)]	0.2070	liter per second per meter
Hydraulic conductivity		
foot per day (ft/d)	0.3048	meter per day
Hydraulic gradient		
foot per mile (ft/mi)	0.1894	meter per kilometer
Transmissivity*		
foot squared per day (ft^2/d)	0.09290	meter squared per day
Application rate		
pounds per acre per year [(lb/acre)/yr]	1.121	kilograms per hectare per year
Leakance		
foot per day per foot [(ft/d)/ft]	1	meter per day per meter
inch per year per foot [(in/yr)/ft]	83.33	millimeter per year per meter

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