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Subsurface Transport of Water and Phosphorus to Lakes in Central Alberta

by

Randall Dean Shaw ©

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ABSTRACT

In this study, I have evaluated methods to measure groundwater-lake flux of water and phosphorus (P), nearshore seepage patterns in ten lakes in central Alberta, and the contribution of groundwater compared to other sources of water and P to the study lakes. The lakes range in surface area from 0.1 to 84 km2; the drainage basin of most of the study lakes consists primarily of glacial till. Groundwater-lake flux was measured with seepage meters. There was an anomalous short-term influx of water to seepage meters. Thus, uncorrected seepage flux calculated from these data were biased; I discuss how this problem can be eliminated. In the nearshore region of the lakes, the direction of seepage was predominantly from the groundwater to the lakes. Similar to seepage patterns observed in other lakes, seepage velocities into the study lakes tended to decrease with distance from shore. However, measured deviations from this pattern were common and were a result of aspects of sampling design and/or the presence of coarse-grained material in the surficial deposits near the lakes. Groundwater may have contributed nearly 50 % of the total annual input of water to one study lake but was a relatively small component (~15 %) of the annual input of water to the other lakes. At Narrow Lake, seepage conditions were investigated with data from a drilling program, water chemistry, environmental isotopes, computer simulations, water budget, mini-piezometers, and seepage meters. Narrow Lake gains water through the nearshore region from a small, shallow groundwater flow system; at deeper offshore regions water moves from the lake to the groundwater system. Net seepage flux into Narrow Lake was about 30 % of the annual input of water to the lake, and

groundwater may be the single largest source of P to the lake. At five other study lakes, P input from groundwater averaged 176, 35, and 285 % of P inputs from molecular diffusion, surface runoff, and atmospheric deposition, respectively. Thus, groundwater can be an important source of P to lakes and should not be overlooked when nutrient budgets are prepared for lakes.

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1. GENERAL INTRODUCTION

Lakes are part of groundwater flow systems; they may be sites of recharge to and/or discharge from groundwater. Recent studies have indicated that lake hydrology and water chemistry can be greatly influenced by groundwater (Born et al. 1979; Loeb and Goldman 1979; LaBaugh 1986). Even so, the relationship between lakes and groundwater is ignored by most limnologists. This is vividly demonstrated by examination of a widely distributed book on limnology (Limnology, Wetzel 1983); only a few pages, of 767 pages of text, allude to groundwater.

There are probably many reasons for the lack of studies on groundwater-lake interactions. Historically, limnologists focussed on descriptive analysis of lake biota (Wetzel 1983). More recently, the focus has shifted towards attempts to understand functional relationships between biota and physical, chemical and environmental parameters. However, difficulties in quantifying seepage rates in lakes have limited the number of studies on groundwater-lake interactions.

I have evaluated methods to measure groundwater-lake flux of water and phosphorus (P), nearshore seepage patterns in ten lakes in central Alberta, and the relative contribution of groundwater compared to other sources of water and P to the study lakes. The following pages of this chapter include a brief description of (1) basic physics and mathematics describing groundwater flow, (2) aspects of groundwater flow in small drainage basins, (3) aspects of groundwater-lake interactions, and (4) methods to measure seepage to lakes. This chapter concludes with an outline of the framework of the thesis.

Groundwater Flow Equations

For my research, lakes situated in unconsolidated material (porous media) were investigated. These lakes seldom have distinct points of concentrated groundwater discharge ("springs"); instead, water may move into (or out of) the lake throughout the entire lakebed (Meyboom 1963). In this thesis, the term groundwater is used for water that occurs in saturated porous media below the water-table. The water-table is the surface at which fluid pressure in the pores of the porous medium is atmospheric.

Flow through porous media can be empirically described by Darcy's law:

$$Q = -K\,A\,\frac{dh}{dl} \qquad (1)$$

where Q is the volume discharge of water per unit length of time through a cross-sectional area, A, of porous medium, K is hydraulic conductivity, and dh/dl is hydraulic gradient (dh is the change in hydraulic head, h, over a distance, dl, between two points along a groundwater flow path).

Hydraulic conductivity is a function of both the porous medium and the fluid. Values of K tend to be low for clay and unfractured rocks (10^{-13} to 10^{-9} $m.s^{-1}$) and high for sand and gravel (10^{-5} to 1 $m.s^{-1}$; Freeze and Cherry 1979). In addition, the density (ρ) and viscosity (μ) of the fluid affect K. In shallow groundwaters that are subject to large seasonal fluctuations in temperature, μ can change sufficiently to affect K. For example, at 1 ^{o}C, $\mu = 1.8$ cP; at 20 ^{o}C, $\mu = 1$ cP

(Giancoli 1980). K is inversely proportional to μ; thus, all else being equal, the value of K at 20 ^{o}C would be 175% of the value at 1 ^{o}C.

The basic physics of groundwater flow was analyzed by Hubbert (1940). He showed that groundwater flows in response to changes in fluid potential, Φ:

$$\Phi = g\,z + \frac{p - p_o}{\rho} \qquad (2)$$

where g is acceleration of gravity, z is elevation, p is fluid pressure, and p_o is atmospheric pressure. Φ is a measure of the mechanical energy per unit mass of fluid, and it is related to hydraulic head by the simple relationship

$$\Phi = g\,h \qquad (3)$$

Φ and h are almost perfectly correlated, i.e., g can be considered a constant. Thus, h represents a measure of energy at a point in the groundwater flow domain, and Darcy's empirical law (Eq. 1) describes the movement of groundwater from a point of high energy to a point of low energy.

Hydraulic head and hydraulic conductivity can both be measured at points in a groundwater flow system by piezometers (water wells that are open at the top, sealed along their length, and open at the bottom). Hydraulic head is the elevation of water in the well with reference to a standard datum. Hydraulic conductivity can be determined from slug tests (Hvorslev 1951) or pumping tests (Cooper and Jacob 1946).

The process of groundwater flow can be described mathematically. If one assumes that K is independent of position within a porous medium (homogeneous), K does not vary with respect to the direction of measurement (isotropic), and Q is independent of time (steady-state), then Darcy's law can be generalized for the three space coordinates (x,y,z) as:

$$q_x = -K \frac{\partial h}{\partial x}, \qquad q_y = -K \frac{\partial h}{\partial y}, \qquad q_z = -K \frac{\partial h}{\partial z} \tag{4}$$

where $q = Q/A$ is seepage flux (i.e., the volume rate of flow per unit area). Darcy's law is the basic law of groundwater flow; it summarizes much of the physics of flow by relating groundwater flux to hydraulic potential (Wang and Anderson 1982).

A second law (conservation of mass) is required to develop a groundwater flow equation. If water is incompressible, there are no sources or sinks of water (i.e., no precipitation or evapotranspiration) to a unit volume of porous medium (elemental volume), and groundwater flow through the elemental volume does not vary with time, then the amount of water flowing into the volume must equal the amount of water flowing out of the volume, i.e.,

$$\frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} + \frac{\partial q_z}{\partial z} = 0 \tag{5}$$

A steady-state groundwater flow equation is obtained by combining Darcy's law (Eq. 4) with the equation of continuity (Eq. 5):

$$\frac{\partial}{\partial x}\left(-K \frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(-K \frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(-K \frac{\partial h}{\partial z}\right) = 0 \tag{6}$$

If K is homogeneous and isotropic, Eq. 6 reduces to Laplace's equation:

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = 0 \qquad (7)$$

More complex equations have been developed for transient flow and for flow through heterogeneous and anisotropic media (Freeze and Cherry 1979).

The solution of Laplace's equation is a boundary value problem; the region of groundwater flow must be defined and boundary conditions must be approximated (Fig. 1.1A). For relatively simple, two-dimensional problems, analytical solutions are possible. However, for more complex conditions numerical solutions are necessary. In either case, solution of Eq. (7) gives the hydraulic head distribution at points within the groundwater flow domain. From the hydraulic head distribution, flow nests can be constructed graphically to quantify the rate and direction of groundwater flow (Fig. 1.1B).

Groundwater Flow in Drainage Basins

Groundwater flow equations have been solved for a variety of conditions to elucidate features of groundwater flow in drainage basins. Groundwater flow systems are formed by the interactions of topographic, geologic and climatic factors.

Topography- The effect of topography on groundwater movement was investigated by Tóth (1962, 1963) with an analytical solution of a two-dimensional, homogeneous, isotropic groundwater flow equation (Laplace's equation). Tóth concluded that under extended flat areas, groundwater movement would be very low or nil. However, if there was a constant gentle slope in the water table (e.g., prairies), regional flow systems develop (Fig. 1.2A). As the topographic relief increases (e.g., hummocky moraine), local flow systems are superimposed on the regional system (Fig. 1.2B). If the hummocks are very large, local flow systems will reach the bottom boundary and create a series of small, independent flow systems. A direct consequence of Tóth's observations is that no large, unconfined regional flow system can extend across regional topographic highs or lows. Thus, regional topographic highs and lows may be considered as imaginary, impermeable boundaries for horizontal groundwater flow (Fig. 1.1A).

Geology- The position and extent of aquifers and aquitards within a flow system affect groundwater flow patterns (Fig. 1.2C). Aquifers are saturated permeable materials that are capable of transmitting significant quantities of groundwater; aquitards are less permeable materials that are not capable of transmitting significant quantities of groundwater (Freeze and Cherry 1979). Lithology, stratigraphy and structural features of geological deposits affect the distribution and extent of aquifers and aquitards.

Much of Alberta lies within the Interior Plains physiographic province (Bostock 1970). The uppermost bedrock units over much of this region are Upper Cretaceous, marine and non-marine sandstones and shales (Green 1972). In some areas, sandstones and fractured shales are aquifers (Borneuf 1972; Gabert 1975; Stein 1976). Unfractured shales have very low permeability and are a poor source of groundwater. The unfractured shales are aquitards and act as a vertical no-flow boundary for groundwater flow systems. Pre-glacial drainage carved channels into the bedrock. These channels are commonly filled with sand and gravel and can be important aquifers (Gabert 1975; Stein 1976; Ozoray et al. 1979).

In Alberta, the surficial deposits consist of three main groups: moraine, glaciolacustrine deposits, and outwash. Ground moraine and hummocky moraine are composed mainly of glacial till; till makes up about 70% of the surficial deposits in Alberta (Pawluk and Bayrock 1969). Till that overlies marine clays tends to be fine-textured and is an aquitard. However, till that overlies sandstone tends to be sandy and more permeable to groundwater flow. Moraine may also contain pockets of stratified material (e.g., sand and gravel lenses). These intertill lenses tend to be relatively permeable for groundwater movement and may be sources of water for rural usage (Gabert 1975). Glaciolacustrine deposits of silt and clay are relatively impermeable and are aquitards. In contrast, outwash deposits of sand and gravel, which were laid down by glacial-meltwater streams, tend to be relatively permeable for groundwater movement. In addition, aeolian deposits (sand) and recent alluvial deposits (sand and gravel) are highly permeable for groundwater flow, and these deposits are often aquifers (Fig. 1.3).

Climate- Climatic factors such as air temperature, precipitation and evapotranspiration affect groundwater flow systems by causing fluctuations in water-table elevations. Water-table elevations in Alberta tend to follow a pattern of (1) a rise in the spring due to snow-melt recharging the groundwater, (2) a decrease during the summer and fall due to evapotranspiration and groundwater discharge, which is interrupted by sporadic rises due to recharge from precipitation, and (3) a relatively constant decrease in water-table levels during the winter due to groundwater discharge (Gabert 1986). High rates of evapotranspiration by phraetophytic vegetation can cause diurnal fluctuations in water-table levels (Meyboom 1966).

General Aspects of Groundwater-Lake Interactions

In this thesis, lakes that gain water from groundwater are referred to as effluent lakes, and lakes that lose water to groundwater are referred to as influent lakes. This terminology is consistent with that used by hydrologists to categorize streams with respect to groundwater.

Factors which control groundwater-lake interactions have been examined by computer simulation models. Much of this work is a refinement of the analyses of regional groundwater flow systems in vertical sections carried out in the 1960's (e.g., Tóth 1963; Freeze and Witherspoon 1967). However, the more recent studies specifically address groundwater-lake interactions and provide details on flow conditions near lakes.

Winter (1976) provides the most detailed theoretical evaluation of groundwater-lake interactions. Winter contends that the stagnation point is a key to understanding the interaction of lakes and groundwater. The stagnation point is the point of least hydraulic head along a divide between groundwater flow systems of different orders of magnitude (Fig. 1.4). If a stagnation point exists, water cannot move out of the lake into the groundwater system. Under a wide variety of simulations of cross-sections through hypothetical lakes bounded by water table mounds, the stagnation point occurred under the lake shore on the downslope side of the lake (with respect to the regional flow pattern). The position and head of the stagnation point was strongly affected by the height of the water table on the downslope side of the lake relative to the lake level, position of aquifers near the lake, anisotropy, regional slope of the water-table and lake depth. When a stagnation point was not present, groundwater would flow into lakes through littoral sediments, and there would be a loss of water from the lake through the pelagic sediments. Even so, the quantity of water entering the lake would be generally much greater than the loss of water from the lake.

Three-dimensional analysis of steady-state groundwater flow near round lakes supports results of the two-dimensional analysis (Winter 1978). The stagnation point lies within a stagnation zone, a trough-shaped zone of low hydraulic head that underlies the downslope shore of the lake. The shape and head of the stagnation zone is related to the shape and head of the water-table mound on the downslope side of the lake. Similar to results from the two-dimensional analyses, the size and lateral position of aquifers within the groundwater system strongly affected the stagnation point.

Many lakes in permeable material (e.g., sandy outwash) are not bounded on both sides by water-table mounds, and groundwater may flow into the lake on one side and out of the lake on the other side (Winter 1983). These "flow-through" lakes are subject to seasonal reversals in the direction of groundwater flow. The reversals are due to seasonal fluctuations in water-table mounds and the formation of stagnation points following periods of major recharge, e.g., spring snow-melt or heavy summer rainfalls (Anderson and Munter 1981; Winter 1983). Computer simulation of transient flow conditions showed that in permeable material, small closed flow systems can develop and dissipate within weeks to months following major recharge events. In less permeable material these flow systems may last for years (Winter 1983).

The distribution of seepage within lakebeds is controlled by the geometry of the flow system and geology of porous media near the lake. In general, seepage to lakes that are bounded by water-table mounds is highest near the lake shore and decreases exponentially with distance from shore (McBride and Pfannkuch 1975). However, when the width ratio (the ratio of half the lake width to the thickness of the groundwater system) is less than 0.6, seepage tends to be uniformly distributed across the lakebed (Pfannkuch and Winter 1984). Lakes with width ratios greater than 2 tend to follow the general pattern of nearshore seepage concentration. In most natural lakes, the width ratio is probably greater than 2. The presence of highly permeable material intersecting the lakebed affects the nearshore seepage pattern. Groundwater is diverted into these highly permeable materials, which results in offshore zones of high seepage flux (Krabbenhoft and Anderson 1986). A list of some factors and their effects on groundwater-lake interactions is provided in Table 1.1.

Lake water quality can be influenced by the position of a lake in the groundwater flow system (Barica 1978; LaBaugh 1986; Swanson et al. 1988). In general, the local flow system most strongly affects the hydrology and water chemistry of prairie lakes (Sloan 1972; LaBaugh

1986; LaBaugh et al. 1987). Thus, there may be large variability in the water quality of lakes situated within a small geographic area. A striking example of this variability is given by Swanson et al. (1988); specific conductance of two lakes located 100 m apart from one another were 1,180 and 38,000 $uS.cm^{-1}$. Furthermore, lakes within the same region do not necessarily respond identically to seasonal and annual changes in climate. Instead, lakes respond according to their individual relationship with the groundwater flow system (LaBaugh 1988).

Eutrophication is a serious problem affecting the water quality of many lakes. Eutrophication is a natural process but can be enhanced by increased anthropogenic nutrient loading to lakes. In some areas, groundwater may contribute significant amounts of nutrients to lakes (Brock et al. 1982). However, nutrient inputs from groundwater are virtually always ignored in studies of lake eutrophication.

Groundwater can be a source of other contaminants to lakes. For example, fecal coliforms can enter groundwater from septic effluents (Hagedorn et al. 1981). The lake may become contaminated if the rate and pattern of groundwater flow and the geological deposits facilitate transport of the contaminated groundwater to the lake. Although the possibility for surface water contamination from groundwater exists, few data are available to actually quantify the extent of the interaction (Lee et al. 1980).

Methods to Investigate Groundwater-Lake Interactions

Computer simulation analysis of groundwater-lake systems are useful to isolate the factors that control groundwater-lake interactions.

However, field studies are necessary to determine whether hypotheses developed from theoretical investigations are applicable to real lakes. In addition, field studies are needed to quantify the effect of groundwater on lake hydrology and water chemistry.

The major obstacle confronting those who attempt to collect field data on groundwater-lake interactions is that there is no best method to measure seepage at lakes. Many methods are available; however, all are subject to error. The relationship between groundwater and lakes can be examined with (1) hydrogeological methods to investigate groundwater flow patterns, (2) indirect methods to determine the effect of groundwater on lake hydrology and/or lake chemistry, and (3) in situ methods to quantify groundwater-lake flux.

Hydrogeological methods- Hydrogeological investigations of groundwater flow near lakes typically involve a drilling program to determine the geology of the flow system. Water wells and piezometers are installed at selected points within the flow system to measure hydraulic gradients and hydraulic conductivity of the porous media. From these data, seepage flux can be estimated with Darcy's equation (e.g., Jaquet 1976; Loeb and Goldman 1979; Lee 1980). In addition, the evaluation of water chemistry or stable isotopes from lake water and groundwater can assist in the interpretation of groundwater flow patterns at the lake (e.g., Moran 1977; Karnauskas and Anderson 1978; Clare and Ko 1982).

Computer simulation models have been used to quantify seepage flux at some lakes (McBride and Pfannkuch 1975; Lee et al. 1980; Munter and Anderson 1981). However, detailed information about the hydrogeological environment near the lake is required: e.g., boundary conditions, geometry of the groundwater flow system, geology, hydraulic conductivity. These data can be difficult to obtain. The installation of wells to collect hydrogeological data can be prohibitively expensive, especially in areas of poor road access. In addition, wells should be correctly positioned to optimize information on groundwater-lake interactions (Winter 1976). Hydraulic conductivity can vary greatly, even in so-called homogeneous media (Freeze 1975). Regionalization of hydrogeological parameters can lead to errors in results from computer modelling of groundwater systems (Freeze 1972).

<u>Indirect methods</u>- Groundwater seepage is most often determined indirectly from the residual of a lake water balance:

$$\text{residual} = Pr + SI - E - SO - \Delta V \qquad (8)$$

where Pr is precipitation, SI is surface flow into the lake, SO is surface flow out of the lake, E is evaporation, and ΔV is the change in lakewater storage. A water balance can be calculated for any length of time, but it is usually calculated from data collected over a minimum of one water year. Precipitation, surface flow and the change in lakewater storage can be measured by standard methods (Church and Kellerhals 1970; McKay 1970; Winter 1981a). Evaporation is more difficult to quantify but can be determined from climatic factors (Morton 1979), isotopes (Welhan and Fritz 1977; Allison et al. 1979), evaporation pans (Winter 1981a) and energy transfer methods (Cray 1970).

There are errors associated with measuring each component of the water budget. Thus, the residual not only includes the net flux of unmeasured components of the water balance (net groundwater flux plus diffuse runoff to the lake), but, also the cumulative errors associated with measuring the other components of the water budget. In some cases, the errors can be greater than the value of the residual (Winter 1981a). Therefore, the residual may be a poor estimate of the groundwater component of the lake water balance.

A hydrological and hydrochemical model has been developed for lakes in the Canadian prairies (Crowe and Schwartz 1979a). This model is an extension of the watershed hydrology budget and routes water and mass through a lake-watershed system. The groundwater component of the water and mass balance can be determined by sensitivity analysis (Crowe and Schwartz 1979b). However, detailed information on hydrological and hydrogeological parameters are necessary inputs to the model. These data are not available for most lakes.

<u>In situ methods</u>- It was the introduction of a simple, <u>in situ</u> method to quantify seepage flux in lakes that is most likely responsible for the growing interest by limnologists in the relationships between groundwater and lakes. Seepage meters were developed and used by irrigation engineers in the 1940 to 1960's to measure the loss of water from irrigation canals (Israelsen and Reeve 1944; Rasmussen and Lauritzen 1953; Meerscheidt 1951; Robinson and Rowher 1959; Bouwer 1961; Bouwer et al. 1962; Bouwer and Rice 1963). More recently, seepage meters have been used to measure seepage flux in lakes, estuaries, and coral reefs (Lee 1977; Belanger and Mikutel 1985; Lewis 1987).

The most common type of seepage meter that is used in lakes is a bottomless drum constructed by cutting off the top or bottom 15 cm of a "45-gallon" drum. The meters are placed on the lake bottom and a collecting device, usually a plastic bag, is attached to the drum. After the plastic bag has been attached for a length of time, t (in s), the volume of water in the bag, V (in mL), is measured. For an effluent lake, seepage flux ($ml.m^{-2}.s^{-1}$) or equivalent units of velocity ($m.s^{-1}$) are estimated from the increase in V during t, corrected for the area of the seepage meter (e.g., 0.255 m^2 for "45-gallon" drums). For an influent lake, seepage flux is estimated from the decrease in V during t (from a bag that was filled with water before it was attached to the drum).

Since their introduction, seepage meters have been used to monitor groundwater-lake interactions at many lakes (e.g., Lee 1977; Brock et al. 1982; Lodge et al. 1989). The ability of seepage meters to accurately measure seepage flux has been investigated by various methods within lakes (Lee et al. 1980), irrigation canals (Israelsen and Reeve 1944; Robinson and Rohwer 1959), seepage rings (Robinson and Rohwer 1959), and laboratory tanks (Lee 1977; Erickson 1981). Results of these studies are summarized in Table 1.2.

Seepage meters provide point estimates of groundwater-lake flux. However, seepage can be highly variable, both spatially and temporally. The variability depends upon many hydrogeological features. Therefore, quantifying whole-lake seepage from point data can lead to considerable errors. An appropriate sampling design must be used to accurately quantify whole-lake seepage. However, the problem of sampling design has not been addressed (Winter 1981a).

In addition to seepage meters, other *in situ* devices have been developed to investigate groundwater-lake interactions. Piezometers have been modified and installed directly in lake sediments (Lee and Cherry 1978). These "mini-piezometers" can give estimates of hydraulic head and hydraulic conductivity similar to those from regular piezometers. Mini-piezometers measure groundwater conditions directly within the lakebed, and seepage flux can be calculated with Darcy's equation. However, mini-piezometers are not commonly used; they did not prove satisfactory at Lake Mead, Nevada because the well points became plugged with fine-grained lake sediments (Woessner and Sullivan 1984).

A lakebed sediment drag has been recently developed to identify springs in lakes (Lee 1985). The drag measures anomalies in sediment temperature and chemistry. However, no quantitative data or seepage flux are provided by this device.

Framework of Thesis

A number of methods are available to investigate groundwater-lake interactions. However, there is no best method to assess groundwater-lake flux. Each method has drawbacks. The accuracy of seepage estimates from a particular method is difficult to evaluate because all methods of measuring groundwater-lake flux are subject to error. Seepage meters are popular because they are simple and inexpensive; however, they have yet to be rigorously tested under a variety of field conditions. Therefore, a goal of this research was to assess whether seepage meters were an appropriate method to measure seepage flux in lakes in central Alberta (Chapter 2 and 3). In central Alberta, glacial till is the predominant surficial deposit, so seepage flux was expected to be low. For the most part, seepage meters have been used in lakes where seepage velocities were relatively high.

As noted earlier, seepage flux to lakes bounded by water-table mounds tends to be highest near shore and decrease exponentially with distance from shore. I examined nearshore seepage patterns at 10 lakes in central Alberta to determine whether this pattern is common to lakes in the study area (Chapter 4).

Groundwater is often cited as a potential source of phosphorus to lakes; however, there are few data to support or reject this perception. A detailed investigation of groundwater P transport was conducted at Narrow Lake, Alberta, a meso-eutrophic lake, 130 km north of Edmonton (Chapter 5). Narrow Lake was selected as the site for detailed investigations because (1) there is little public activity on the lake, so equipment could be left in the lake without fear of mischief, (2) the lake was close to laboratory facilities at the Meanook Biological Research Station, and (3) the lake was a site of other limnological studies. In addition, the potential importance of groundwater transport of phosphorus to five other lakes in central Alberta was investigated (Chapter 6).

For many lakes in central Alberta there is little surface runoff from the watershed to the lake. Therefore, in addition to groundwater, atmospheric deposition was measured to evaluate whether it was an important source of phosphorus to these lakes (Chapter 7).

Table 1.1 Effects of topography, geology, and flow system geometry on groundwater-lake interactions that have been determined from computer simulation studies of lakes in hypothetical groundwater flow systems.

Condition	Effect
Water-table Configuration:	
water-table mound surrounding lake	-nearshore seepage to lakes1
no mound surrounding the lake	-nearshore seepage from lake1
decrease in height of mound relative to lake	-stagnation point decreases1
mound between two lakes	-no exchange of groundwater between lakes1
slope from mound to regional discharge area < 0.01	-no seepage from lake2
slope from mound to regional discharge area > 0.01	-tendency for seepage from lake2
Aquifers:	
intersecting lakebed	-deviation from nearshore seepage concentration3
	-high seepage where aquifer intersects with lake3
limited size, beneath or upslope of lake	-little effect1
underlying water-table mound on downslope side	-stagnation point decreases1
Increased Anisotropy:	-stagnation point decreases1,4
	-more uniform seepage across the lake
Flow Pattern Geometry:	
increased lake depth	-stagnation point decreases1
	-more uniform seepage across the lake5
ratio of half the lake width to thickness of groundwater flow system $\leq$ 0.6	-uniform seepage across the lake5
ratio of half the lake width to thickness of groundwater flow system $\geq$ 2	-tendency for nearshore seepage concentration5,6
increase in thickness of groundwater system	-stagnation point decreases5,6

References: 1Winter 1976; 2Winter 1981b; 3Krabbenhoft and Anderson 1986; 4Winter and Pfannkuch 1984; 5Pfannkuch and Winter 1984; 6McBride and Pfannkuch 1975.

Table 1.2 Some factors and their effects on seepage meter measurements. The types of seepage meter are as follows: SCS - Soil Conservation Services meter (Robinson and Rohwer 1959); SL - Salinity Laboratory meter (Israelsen and Reeve 1944); LJ - Lock and John (1978) meter; L - Lee (1977) meter.

Factors	Effect	Type
Bottom cylinder:		
presence of light due to clear cylinder1	none	SCS
diameter (7 to 15 cm)2	none	SL
thickness of wall2	none	SL
size of opening to measuring device (<0.5-cm ID)3	decreased seepage flux	LJ
Collecting device (plastic bag):		
type4	none	L
not wetted before use4	erratic results	L
pre-wetted before use4	improves efficiency	L
deformed before use4	erratic results	L
Placement of cylinder:		
hammering into sediments2	erratic results	SL
gently pushing into sediments2,5	less variable results	SL,L
increased depth into sediment2,5	decreased seepage flux	SL,L
Allowing a few days before sampling1,5	less variable results	SCS,L

References: 1Robinson and Rohwer 1959; 2Meerscheidt 1951; 3Fellows and Brezonik 1980; 4Erickson 1981; 5Lee 1977.

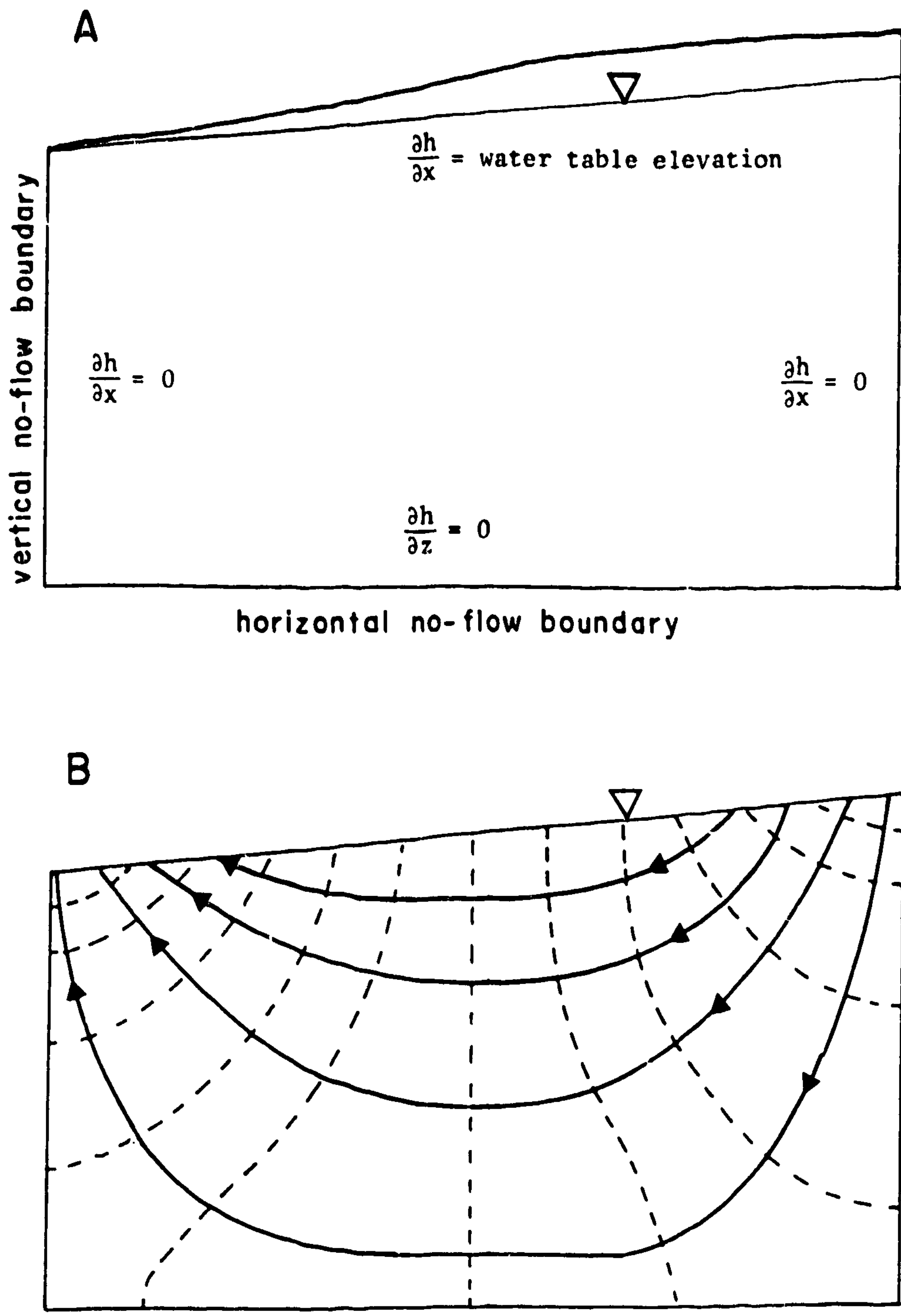

Figure 1.1. (A) Model of a vertical section of a regional groundwater flow system. (B) Flow net for cross-section shown in A. Equipotential lines (dashed) indicate lines of equal hydraulic head; flow lines (solid) indicate the groundwater flow path.

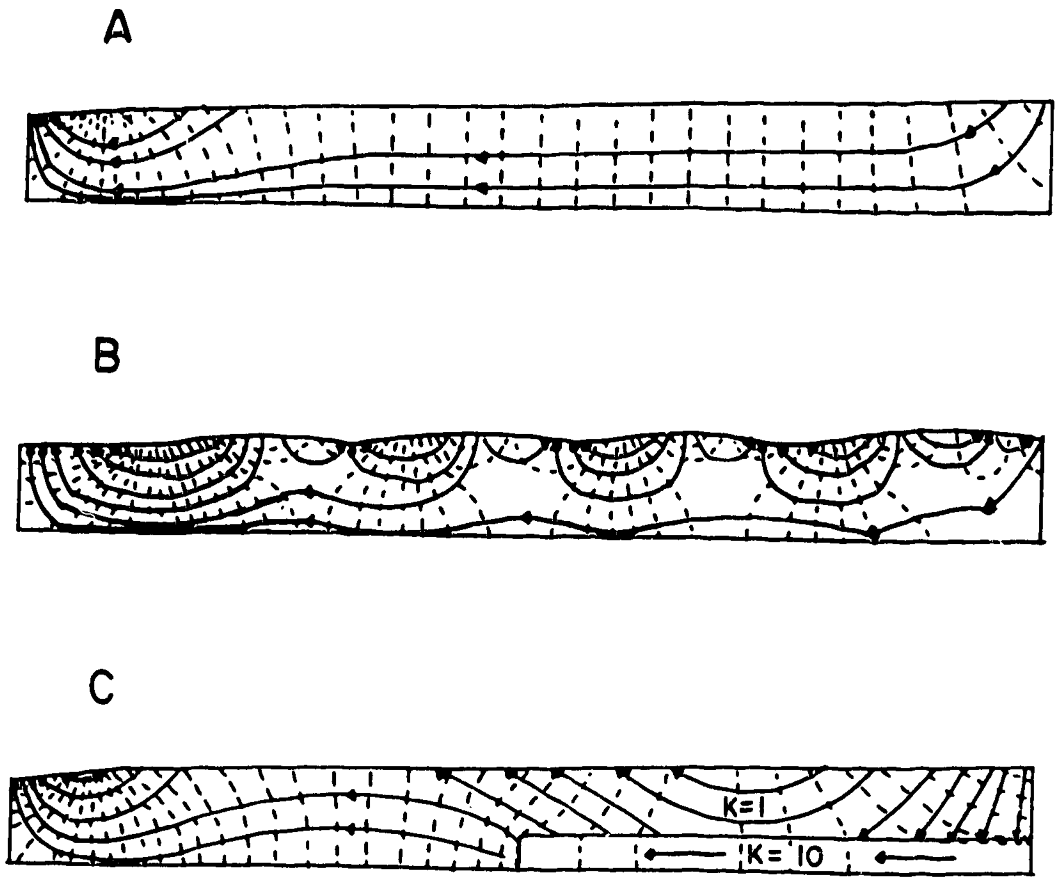

Figure 1.2. Effect of topography and geology on regional groundwater flow: (A) homogeneous, isotropic media with a constant, gentle sloping water-table, (B) homogeneous, isotropic media with rolling water-table, (C) as for (A) except for the inclusion of an aquifer; K indicates the relative value of hydraulic conductivity (after Freeze and Witherspoon 1967).

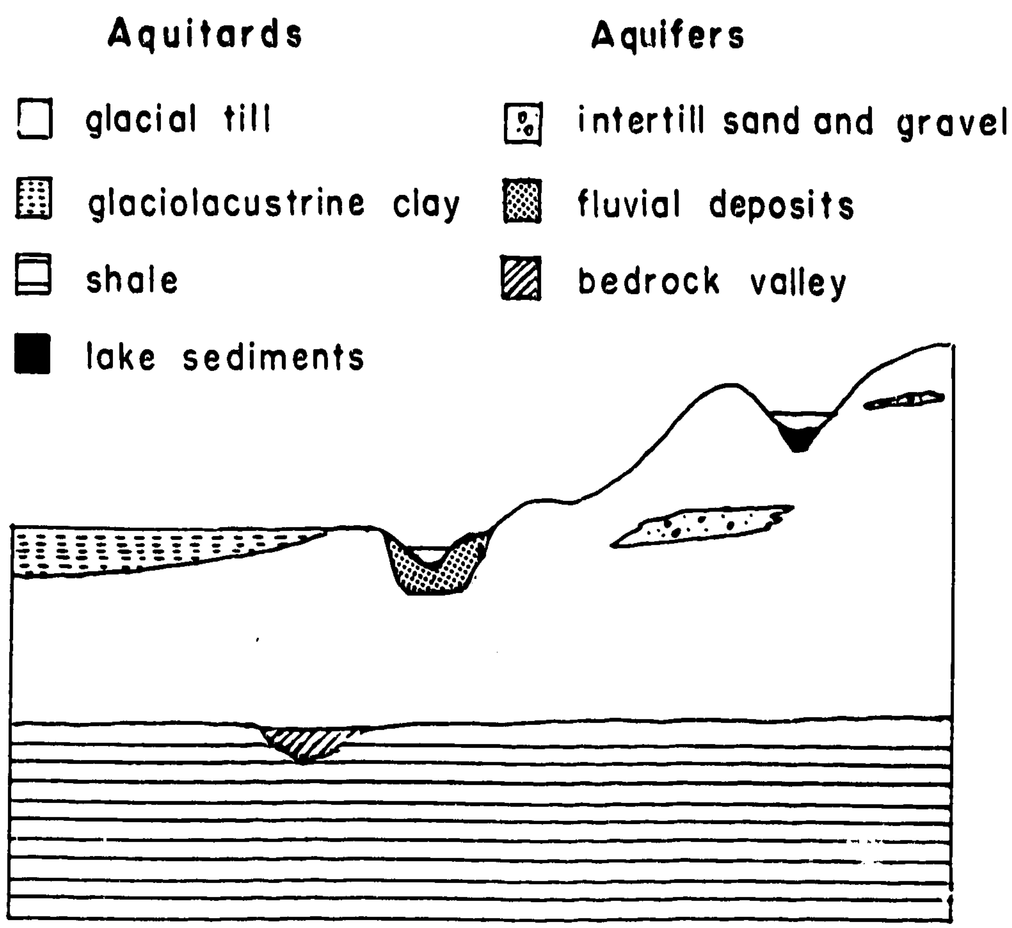

Figure 1.3. Schematic vertical section with some of the geological features that affect groundwater flow.

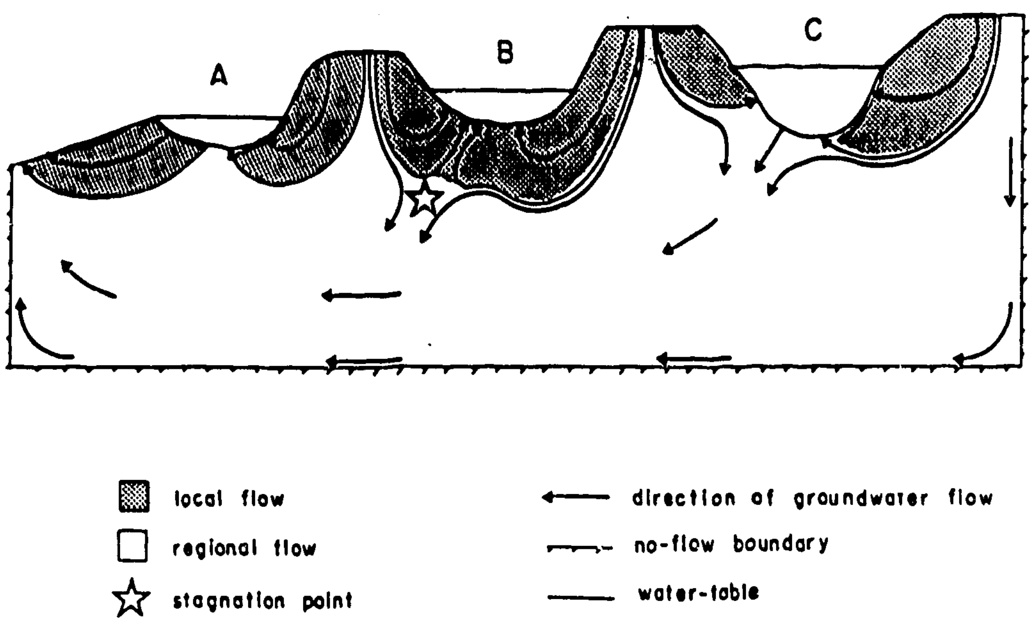

Figure 1.4. Vertical section through a groundwater drainage basin showing the general pattern of flow at lakes: (A) flow-through lakes (water table on one side slopes downward from the lake); (B) effluent lake and the general position of the stagnation point; (C) shallow-effluent, deep-influent lake (bounded by water table mounds).

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2. ANOMALOUS SHORT-TERM INFLUX OF WATER INTO SEEPAGE METERS[1]

2.1 ABSTRACT

Laboratory and field tests revealed that there was an anomalous, short-term influx of water into plastic bags after they were attached to seepage meters. Plastic bags (3.5-liter capacity) were submerged in an 830-liter tank of stagnant water; within 45 min, the volume of water in bags that initially were empty increased to 297 ml, bags prefilled with 1,000 and 2,000 ml of water increased by 160 ml, and bags prefilled with 3,000 ml decreased in volume. At Narrow Lake, Alberta, the anomalous, short-term (30-min) influx of water averaged 237 ml to bags that were initially empty, but the anomaly was effectively eliminated when bags were prefilled with 1,000-ml of water before they were attached to seepage meters. The impact of the anomaly on calculated seepage flux was greatest when seepage flux was low, e.g. 0.3 $ml.m^{-2}.min^{-1}$. The anomaly may be due to mechanical properties of the bag, and it may be alleviated by partially filling bags before they are attached to seepage meters.

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2.2 INTRODUCTION

Seepage meters were developed in the 1940s to measure loss of water from irrigation canals (Israelsen and Reeve 1944). More recently, seepage meters have been used to measure seepage into lakes, estuaries, and coral reefs (Lee 1977; Fellows and Brezonik 1980; Lewis 1987). Seepage meters are bottomless drums constructed by cutting off the top or bottom 15 cm of a "55-gallon" (204-liter) drum (Lee 1977). The meters are placed on lake sediments; a collecting device, usually a plastic bag, is attached to the drum. The rate of groundwater flow through the area of lake bottom enclosed by the seepage meter is calculated from the increase in the volume of water in the bag divided by the length of time the bag was attached to the meter.

During August 1984, I used seepage meters to investigate diurnal fluctuations of seepage flux in Narrow Lake, Alberta. I observed no diurnal fluctuations in seepage rates; instead, I observed that the measured seepage flux was a function of the time that the bag was attached to the meter. For example, seepage flux calculated with data collected over a 24-h interval were, on average, only 68 % of the values calculated with data collected over a 12-h interval. Preliminary investigations suggested that the anomalous, seepage velocities were related to methodology rather than to nearshore hydrological processes that can cause fluctuations in seepage flux (e.g. evapotranspiration by vegetation along the shoreline). In addition, these initial tests indicated that this problem with seepage meter data was due to an anomalous, short-term influx of water after plastic bags were attached to the meter. In this paper, I describe laboratory and field tests that document the anomalous, short-term influx of water to seepage meters, discuss the cause of this anomaly, and suggest how to alleviate this problem.

2.3 METHODS

Seepage meters were sampled in 1984 and 1988 at Narrow Lake (54°35'N, 113°37'W) and in 1987 at Buffalo Lake, Alberta (52°30'N, 112°58'W). Narrow Lake is relatively small and deep and is situated in a glacial meltwater channel of till and intertill sand and gravel lenses (surface area (A_o) 1.1 km^2; mean depth (z) 14 m; Prepas and Trimbee 1988). Hydraulic conductivity (K) of these surficial deposits ranges from $5x10^{-7}$ to $4x10^{-5}$ $m.s^{-1}$. Buffalo Lake is relatively large and shallow (A_o 84 km^2; z 3 m; Alberta Environment 1987). The surficial material near the site where seepage meters were sampled consists of sandy outwash; K ranges from 10^{-6} to 10^{-4} $m.s^{-1}$ (Clare and Ko 1982).

Seepage meters were constructed and installed in the lakebed according to Lee (1977; Fig. 2.1). A two- to three-day equilibration period was allowed before the meters were sampled. Low-density, polyethylene plastic bags (Alligator Baggies: 0.01-cm thick x 29 x 34 cm) were used to collect water samples. When full, the bags contained about 3.5 liters of water. Empty prewetted plastic bags were attached to seepage meters in 1984 and 1988 at Narrow Lake and in 1987 at Buffalo Lake. In addition, 1,000-ml prefilled bags were attached to meters in 1988 at Narrow Lake. To connect the bags to a seepage meter, air was forced out of the bag by slowly pulling the bag down, below the surface of the lake, until the amber-latex tubing was just above the water level; the amber-latex tubing was pinched; and the tubing was gently inserted over the vent tube (e in Fig. 2.1). About 5 ml of water inevitably entered the initially empty bags upon installation, because the rigid plastic tube (c in Fig. 2.1) filled with water. No water entered the 1,000-ml prefilled bags because the plastic tube already contained water. Bags were removed by pinching the amber-latex tubing and gently pulling the bag away from the bottom cylinder. The volume of water (V, in ml) in the bag was measured with 50- to 2,000-ml graduated cylinders. Seepage flux (q, in $ml.m^{-2}.min^{-1}$) was calculated from:

$$q = \frac{3.92\ \Delta V}{t} \qquad (1)$$

where ΔV is t change in volume of water in a plastic bag after the bag was attached to seepage meters for an interval of t min; the factor 3.92 converts the area covered by the seepage meter to the unit of area (i.e. 1 m^2). D. R. Lee (Atomic Energy of Canada Ltd., Chalk River, Ontario) observed the seepage meters at Narrow Lake. He confirmed that the seepage meters and procedure of attaching plastic bags to meters were similar to his (e.g. Lee 1977).

<u>Laboratory tests</u> - Experiments were conducted in an 830-liter, round fiberglass tank (diameter 1.22 m, depth of water 0.71 m) to test whether the plastic bags were the cause of the anomalous, short-term influx of water. In the tank, two 6-cm long, 0.64-cm i.d. plastic tubes were held upright, 40 cm below the water surface, by clamps fastened to a rigid pole (A in Fig. 2.2). The plastic tubes in the tank were identical to the vent tube that connected bags to a seepage meter (e in Fig. 2.1). Empty and prefilled (100-, 1,000-, 2,000-, 3,000-ml) plastic bags were

attached to the vent tubes in the tank. Water in the tank was stagnant except for a small amount of mixing when the bags were installed or removed. Two bags were tested concurrently, and the volume of water in the bags was measured after 0.2 to 3,960 min.

Hydraulic potential inside the bags relative to the tank was measured with a manometer. If hydraulic potential in the bag was lower than in the tank, water would flow from the tank to the bag until the hydraulic gradient was eliminated. The manometer was constructed from a meter stick and two, 1.5-m long, 0.64-cm i.d. flexible plastic tubes. At one end, the two tubes were connected to a meter stick that was fastened to the outside wall of the tank. The other end of each tube extended into the tank; one tube was connected to a vent tube, 40 cm below the water (B in Fig. 2.2). Empty and 1,000-ml prefilled bags were then attached to the vent tube. Hydraulic potential in the bag was measured as the height of water (in mm) in the manometer tube attached to the bag, relative to the height of water in the tube that extended into the tank.

Field tests - The relationship between measured seepage flux and time interval was investigated with data collected at Narrow Lake from 5 seepage meters in 1984 and 10 seepage meters in 1988. Seepage meters were spaced about 1 m apart, at lake depths of 0.5 to 1.0 m. Initially empty bags were sampled at intervals of 5 to 8,150 min, from 31 July to 16 August 1984. From 10 to 22 May 1988, initially empty and 1,000-ml prefilled bags were sampled at intervals of 1 to 4,380 min and 5 to 1,685 min, respectively. On 28 August 1987, plastic bags were attached to three seepage meters, spaced 1 m apart, at 1-m lake depth in Buffalo

Lake; the meters were sampled at time intervals of 16 to 318 min. Seepage flux in lakes may fluctuate seasonally and diurnally (Meyboom 1967). Therefore, each set of data was collected over as few days as possible to minimize the effect of seasonal fluctuations on calculated seepage flux. In addition, seepage meters were sampled at various time intervals during both day and night to minimize the effect of diurnal fluctuations on calculated seepage flux.

Hypothetical considerations - If seepage meters accurately measure seepage rates, a plot of the volume of water in the bag, V, on the time interval that the bag was attached to the meter, t, would be linear until V approached the capacity of the bag (Fig. 2.3A). Thus, the slope of a regression of V on t (from the linear portion of the plot) would equal the seepage rate (in $ml.min^{-1}$) through the seepage meter; the regression would pass through the origin. Consequently, seepage rates calculated over different time intervals would be the same, and measured rate and time interval would be independent (i.e. slope = 0; Fig. 2.3B). In contrast, if there were an anomalous, short-term influx of water, a plot of V on t would not be linear. Rather, initially V would increase rapidly (a in Fig. 2.3A), and seepage rates calculated from data collected from the initial period would be overestimates (a in Fig. 2.3B). After the short-term, V would increase in proportion to the actual seepage rate (b in Fig. 2.3A) until V approached the capacity of the bag (c in Fig. 2.3A). Seepage rates calculated from data integrated over this entire period would still be overestimates (b in Fig. 2.3B); but, the slope of a regression of V on t (from the linear portion of the plot) would equal the seepage rate, and the Y-intercept would equal the volume of the anomalous, short-term influx of water. As the bag reaches its capacity, back-pressure inside the bag would increase, the rate of flow to the bag would decrease, and eventually seepage rates determined from these samples would be underestimates (c in Fig. 2.3A and 2.3B.).

This hypothetical framework was used to evaluate the field data. For each data set (Narrow Lake: 1984, 1988; Buffalo Lake: 1987), V was plotted against t to determine the seepage rate, the volume of the anomalous, short-term influx of water, and whether the volume of water in the bag affected the rate of inflow to the bag.

Data preparation - Seepage rates measured on one date with closely-spaced seepage meters can vary nearly two-fold (Brock et al. 1982). Errors in seepage measurements are a relatively small source of variability. The high variability in measured seepage rates is probably due to variation in hydraulic conductivity of lake sediments (Brock et al. 1982). Seepage flux within a small area of lakebed is log-normally distributed (Chapter 3). Therefore, for field data, linear regressions of V on t were calculated from the geometric mean of V; geometric mean was determined from replicate seepage meter measurements. For laboratory data, linear regressions of V on t were calculated from raw data. Statistical tests are as outlined in Sokal and Rohlf (1981).

2.4 RESULTS

Laboratory tests - The laboratory tank tests indicated that the plastic bags used in this study were not perfectly passive collecting devices. Immediately after empty bags were attached to a vent tube in a water-filled tank, the collapsed bags started to open. Visually, the bags appeared to equilibrate after about 0.5 min. The volume of water in the bags increased significantly ($P < 0.01$) for 30 to 45 min after they were installed in the tank (Fig. 2.4A). After 0.5 min, they contained an average of 94 (SE 29) ml ; after 45 min they contained 286 (SE 48) ml. There was no significant change ($P > 0.05$) in V for intervals from 45 to 3960 min (mean 297, SE 24 ml). Short-term influx of water (0.17 to 45 min after the bags were attached) to 1,000- and 2,000-ml prefilled bags was lower than for initially empty (Fig. 2.4A) or 100-ml prefilled bags (Fig. 2.4B). Similar to initially empty bags, bags prefilled with 1,000- and 2,000-ml of water equilibrated within 30 to 45 min after they were installed in the tank. After 45 to 951 min, 1,000- and 2,000-ml prefilled bags contained an additional 156 (SE 54) and 163 (SE 59) ml of water, respectively. In contrast, the volume of water in 3,000-ml prefilled bag decreased after 5 min (Fig. 2.4B).

Plastic bags attached to a vent tube in a water-filled tank created a pressure gradient conducive to the flow of water from the tank into the bags. Before bags were attached, there was no difference in head between the submerged vent tube and the tank. Immediately after empty bags were attached, however, the height of water in the manometer tube connected to the bag decreased 20 to 40 mm relative to the height of water in the tank. After 1,000-ml prefilled bags were attached, head decreased only about 2 mm. These results suggest that an influx of water to plastic bags attached to seepage meters may be caused by a hydraulic gradient created during installation of the bags. The gradient was much lower with 1,000-ml prefilled bags compared to initially empty bags. Therefore, the influx of water should be lower to prefilled bags compared to initially empty bags. This conclusion is consistent with the data (i.e. Fig. 2.4B).

Field tests- During 1984 at Narrow Lake, there was an anomalous, short-term influx of water for at least 30 min after initially empty plastic bags were attached to seepage meters (Fig. 25A). The rate of inflow to the bags was not affected by V: maximum V (V_{max}) measured in 1984 was 1370 ml. Excluding data from the first 30 min, a regression of V on t for these 1984 data was highly significant:

$$V = 233 + 0.088t \qquad (2)$$

where df = 12, $r^2 = 0.80$, $P << 0.01$. Seepage flux calculated for these data according to Eq. 1 decreased exponentially with time (Fig. 2.5B). Over the range of time intervals tested, measured seepage flux varied more than two orders of magnitude. For example, initially empty bags contained 95 ml after only 15 min (seepage flux 25 $ml.m^{-2}.min^{-1}$); they contained 305 ml after one day (seepage flux 0.8 $ml.m^{-2}.min^{-1}$).

Similar to results from 1984, during 1988 at Narrow Lake there was an anomalous, short-term influx of water for about 30 min after initially empty bags were attached to seepage meters (Fig. 2.6A). The rate of inflow to the bags was not affected by V (V_{max} 950 ml). Excluding data from the first 30 min, a regression of V on t for these 1988 data was highly significant:

$$V = 240 + 0.073t \qquad (3)$$

where df = 5, $r^2 = 0.78$, $P < 0.01$. The two regression lines (Eq. 2 and 3) were not significantly different (ANCOVA: differences in slope:

df = 1,17, $F = 0.36$, $P > 0.5$; differences in adjusted Y-intercept: df = 1,18, $F = 0.12$, $P > 0.5$). The values of the Y-intercepts for both Eq. 2 and 3 were significantly different from 0 ml (Eq. 2: mean (SE) = 233 (92) ml, df = 12, $t = 2.53$, $P < 0.05$; Eq. 3: mean (SE) = 240 (63) ml, df = 5, $t = 3.81$, $P < 0.02$). The short-term influx of water, estimated from the Y-intercept, was about 20 % lower than the volume of water taken up by initially empty plastic bags in the laboratory tank tests (mean (SE) = 297 (24) ml).

In contrast to trials with initially empty bags, there was no obviously anomalous, short-term influx of water to 1,000-ml prefilled bags attached to seepage meters in Narrow Lake in 1988 (Fig. 2.6B). The rate of inflow to the bags was not affected by V (V_{max} 1170 ml). For intervals of 5 to 1,685 min, V increased significantly with t:

$$V = 9 + 0.061t \qquad (4)$$

where df = 5, $r^2 = 0.94$, $P < 0.01$. The slope of Eq. 4 was not significantly different from the linear regression of V on t for initially empty bags in 1988 (ANCOVA: df = 1,10, $F = 0.131$, $P > 0.5$). Adjusted Y-intercepts for Eq. 3 and 4 were significantly different (ANCOVA: df = 1,11, $F = 90$, $P << 0.001$). Furthermore, the Y-intercept from Eq. 4 was not significantly different from 0 (df = 5, $t = 0.8$, $P > 0.2$). These results indicate that, at least under the conditions tested, the anomalous, short-term influx of water to seepage meters can be eliminated by first prefilling bags with 1,000 ml of water. This conclusion is supported by results from the tank tests in which short-term inflow was reduced, though not eliminated, with prefilled bags.

At Buffalo Lake, the volume of water in initially empty bags increased linearly with t (Fig. 2.7):

$$V = 106 + 2.23t \tag{5}$$

where df = 2, $r^2 = 0.97$, $P < 0.05$. The seepage flux at Buffalo Lake, estimated from the slope of Eq. 5, was more than 25 times greater than at Narrow Lake (Eq. 2). The Y-intercept in Eq. 5 was positive (106 ml, SE 61 ml); but it was not significantly greater than 0 (df = 2, t = 1.8, $P > 0.1$). Therefore, at relatively high values of seepage flux (e.g. 8.7 $ml.m^{-2}.min^{-1}$), the anomalous, short-term influx of water to plastic bags may be less important than at low values of seepage flux. Similar to results from Narrow Lake, the rate of inflow to the bags was not affected by V (V_{max} 805 ml).

2.5 DISCUSSION

For initially empty bags, estimates of seepage flux can be corrected for the volume of the anomalous, short-term influx of water. Two seepage meter measurements are required for one corrected estimate. The corrected estimate, q_c (in $ml.m^2.min^{-1}$) is calculated from:

$$q_c = \frac{(V_2 - V_1)}{(t_2 - t_1)} 3.92 \tag{6}$$

where V_1 and V_2 (in ml) are the volumes of water collected after a short (t_1) and long (t_2) time interval, respectively; the factor 3.92 converts the area covered by the seepage meter to the unit of area. V_1 should include the anomalous, short-term influx of water, i.e. the time interval should be more than 30 min. An example based on data collected at $t_1 = 45$ min and $t_2 = 1,385$ min (about 1 d) with initially empty bags at Narrow Lake is given in Table 2.1. Uncorrected estimates of seepage flux for t_1 and t_2 were 19 and 1.7 times, respectively, the corrected flux. The variance of the corrected estimate was no greater than that of the uncorrected estimates (Table 2.1). At a higher seepage flux, a correction for the anomalous, short-term influx of water would be less critical. For example, at Buffalo Lake, uncorrected seepage flux at $t_1 = 27$ min and $t_2 = 319$ min were only 2 and 1.1 times, respectively, the corrected flux (9.1 $ml.m^{-2}.min^{-1}$).

The anomalous, short-term influx of water into plastic bags may be a result of the process used to make the bags. Bags are manufactured by blowing molten resin into a tube, to the form of a partially expanded bag, which is then cooled (G. White, pers. comm.; Dubois and John 1981). During installation of plastic bags to seepage meters (or to vent tubes in the laboratory tank), bags were deformed from their original, partially expanded states. After bags were attached to seepage meters (or vent tubes in the tanks) they appeared to regain their original states. Consequently, the expansion to their original states "pulls" water into the bag. Bags prefilled with 1,000 ml of water are partially expanded before they are attached to the meter.

It is difficult to evaluate whether seepage flux measured in other studies have been affected by an anomalous, short-term influx of water because sampling designs are rarely described in sufficient detail. A similar problem with seepage meter data was observed, however, by Erickson (1981) at Williams Lake, MN. Interestingly, he found a short-term (10 to 18 min) influx of water to seepage meters at sites of groundwater recharge from the lake (i.e. seepage from the lake to groundwater). This observation suggests that anomalous, seepage meter data can occur under conditions much different than those at the study lakes.

Results of this study do not necessarily imply that this problem with seepage meter data occurs under all conditions. : have shown that in areas of high seepage flux the anomaly may not be important. In addition, consider a seepage meter in a lake where sediments have a very low permeability to groundwater flow. The top and sides of a seepage meter drum are rigid, and the bottom sediments would act as a seal to prevent groundwater flow. If one assumes that water is incompressible, the law of conservation of mass requires that the amount of water flowing into an elemental volume of water (e.g. seepage meter drum) must equal the amount of water flowing out of the elemental volume. Thus, even if the hydraulic potential in the meter is greater than the hydraulic potential in an initially empty bag that is connected to the meter, water cannot flow from the meter to the bag. On the other hand, if the sediments are permeable to groundwater flow (or if the seepage meter is not properly installed), water can be "pulled" out of the sediments to the seepage meter; consequently, water can flow from meter to the bag.

2.6 CONCLUSION

In summary, I have shown that there may be an anomalous, short-term influx of water after plastic bags are attached to seepage meters. The anomaly was not caused by fluctuations in seepage due to hydrological processes; rather, it was likely caused by the plastic bags. The anomaly may vary greatly between (and within) lakes in response to seepage flux and permeability of the bottom sediments. To alleviate the problem, bags should be prefilled with 1,000 ml of water before they are attached to seepage meters or seepage flux should be corrected for the anomalous volume of water. Failing to do so may give estimates of seepage flux that are unrealistic and may lead to misconceptions about groundwater-surface water interactions.

Table 2.1. Geometric mean, coefficient of variation (C.V.) and 95 % confidence intervals for uncorrected (t_1 and t_2) and corrected (t_2 - t_1) seepage flux measured at one site in Narrow Lake, 11 May 1988. Meters were sampled at intervals of 45 (t_1) and 1385 (t_2) min. Corrected flux was calculated as in Eq. 6.

SM	Volume		
	t_1	t_2	(t_2 - t_1)
1	152	470	318
2	194	400	206
3	44	150	106
4	125	290	165
5	150	495	345
6	290	470	180
7	176	355	179
8	190	460	270
9	206	950	744
Mean Volume (ml)	155	407	239
Mean Flux ($ml.m^{-2}.min^{-1}$)	13.5	1.2	0.7
C.V. (%)	10.4	8.2	10.2
Upper 95% C.I.	19.0	1.6	1.0
Lower 95% C.I.	9.6	0.8	0.5

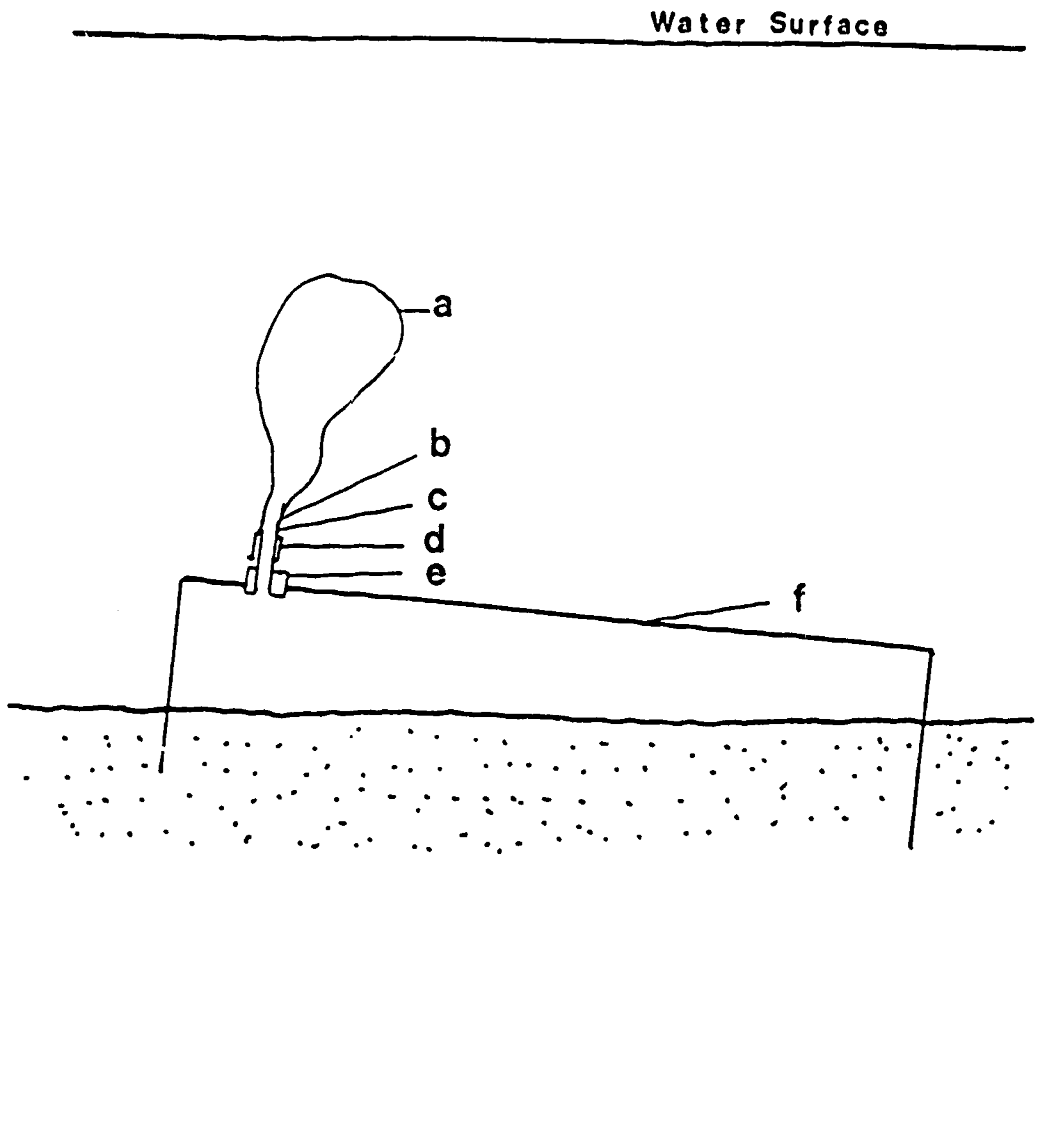


Figure 2.1. Seepage meter installed in lake. a - 0.01-cm thick x 26x34 cm Alligator Baggies plastic bag; b - rubber-band wrap; c - 0.64-cm i.d., 6-cm long plastic tube; d - 5-cm long amber-latex tube; e - No. 5 1/2 rubber stopper with 0.64-cm i.d., 6-cm long plastic vent tube; f - 15-cm high x 57-cm diameter drum. Modified from Lee (1977).

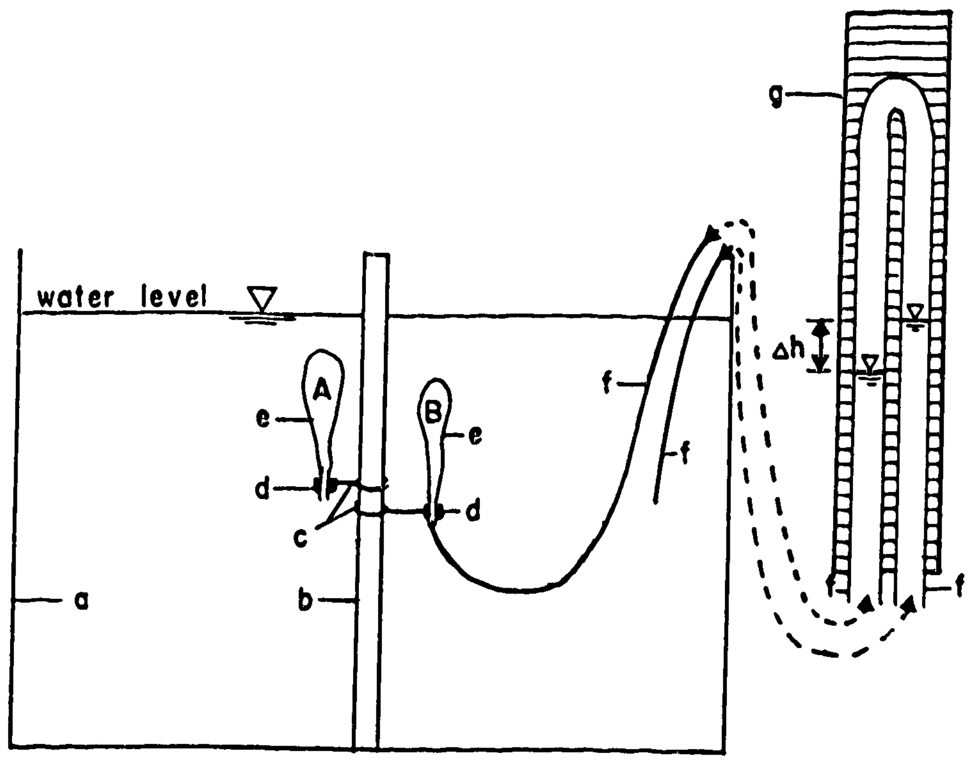


Figure 2.2. Laboratory tank set-up. A: bag attached to vent tube, B: manometer experiment. a - 1.0-m high x 1.22-m diameter fiberglass tank; b - support post; c - clamp; d - No. 5 1/2 rubber stopper with 0.64-cm i.d., 6-cm long plastic vent tube; e - seepage meter bag; f - 0.64-cm i.d. tubing; g - manometer (marked in mm) attached to outside wall of tank (not drawn to scale); h is the difference in head between the bag (B) and the water level in the tank.

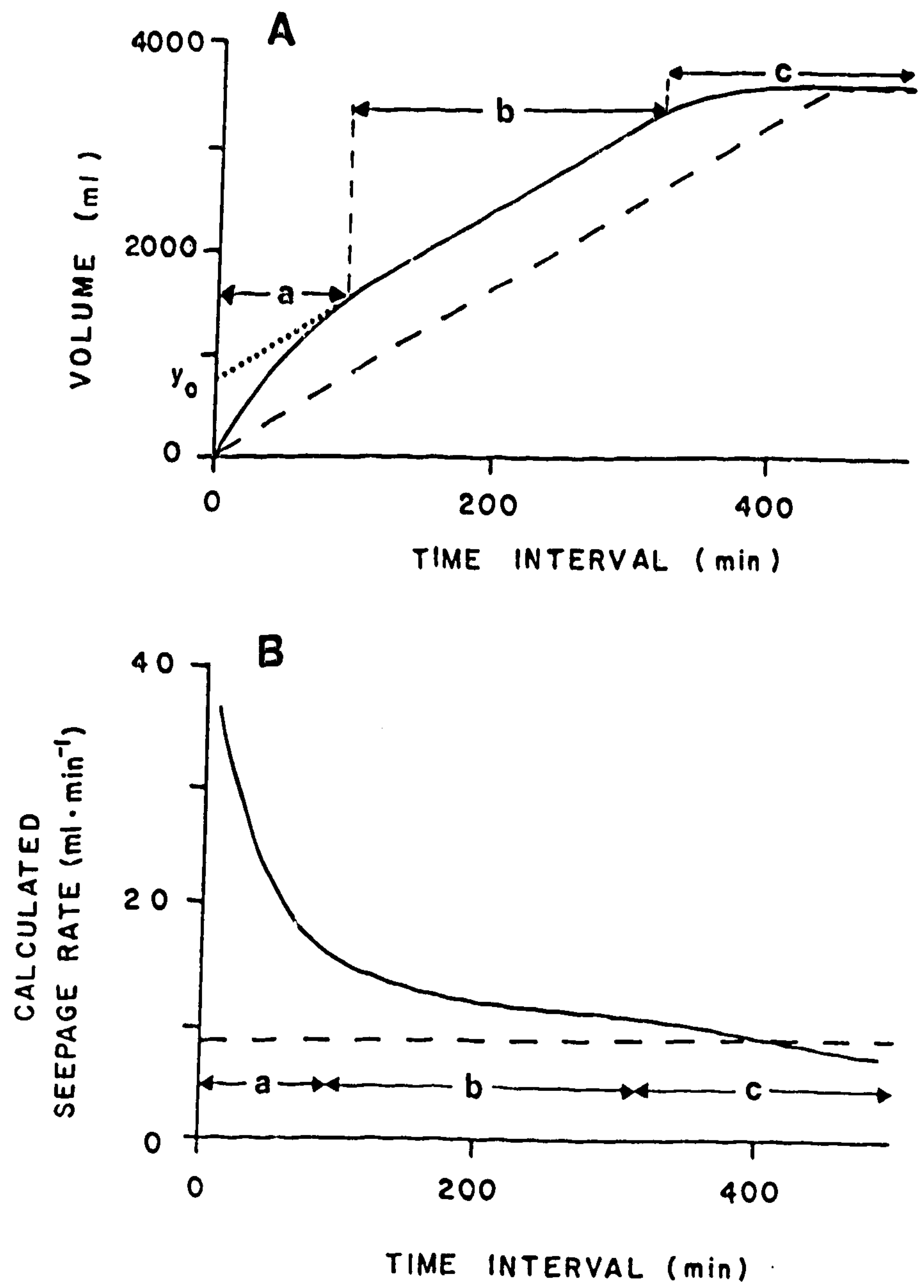


Figure 2.3. Hypothetical relationship between (A) the volume of water (V) collected from a time interval (t), and (B) seepage rate estimated from V on t. The dashed lines show the relationship that would occur if V were independent of t. The solid lines show the relationship if there was: a - an anomalous, short-term influx; b - inflow in proportion to seepage rate, and c - a decrease of inflow as the bag reaches its capacity (3.5 liter). y_0 is the y- intercept of a linear regression of V on t for data collected during the period b.

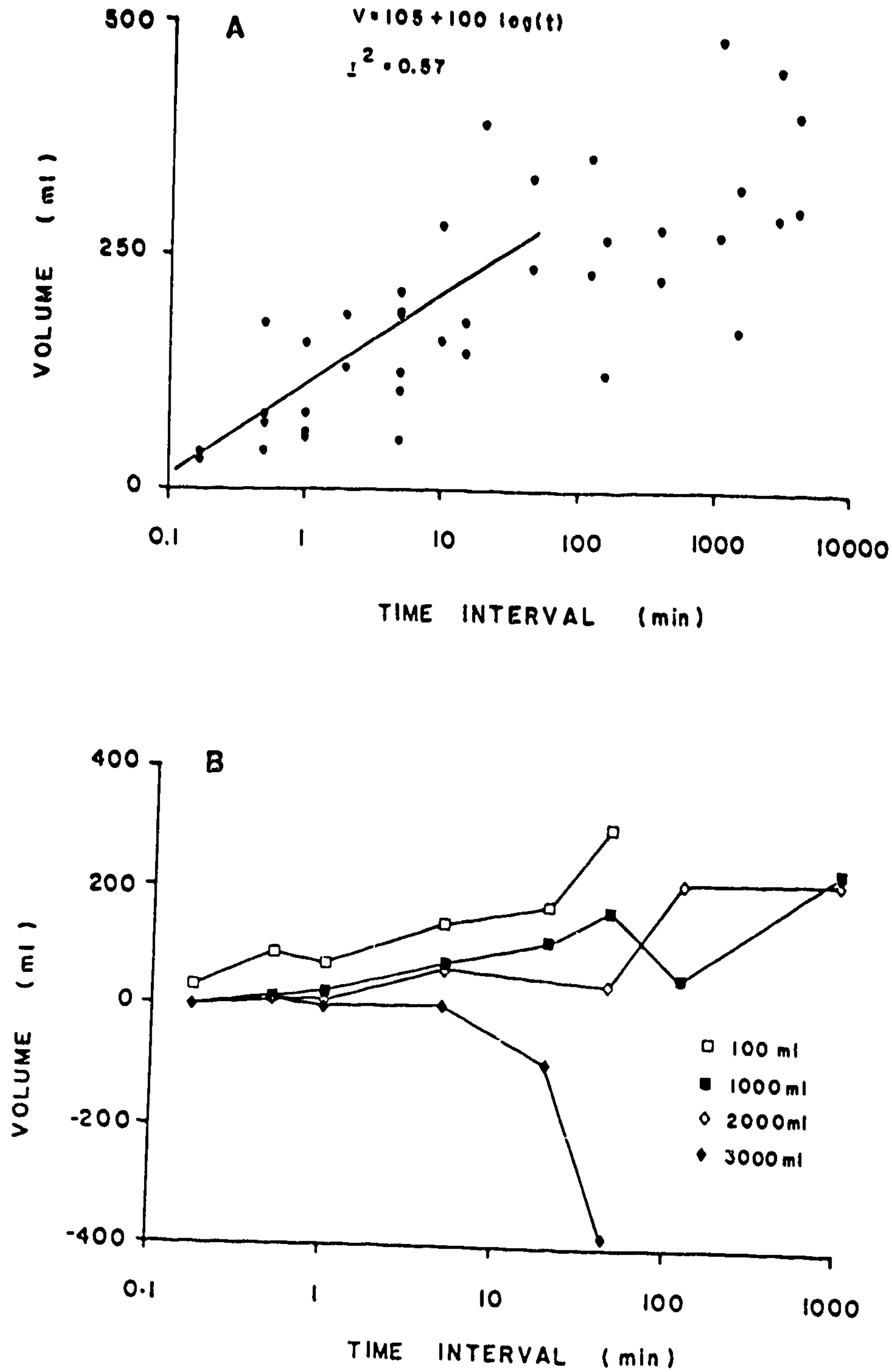
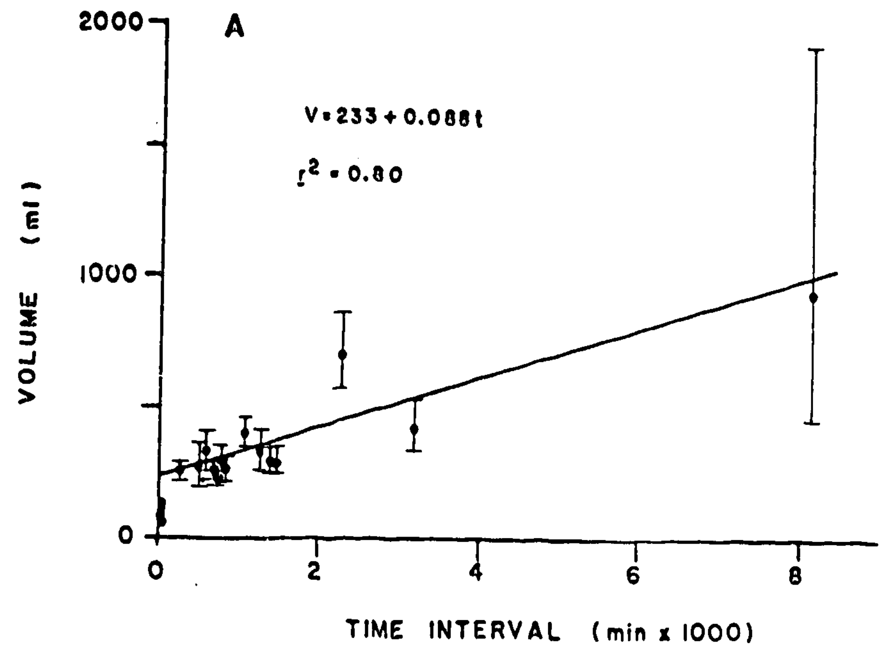


Figure 2.4. Volume of water collected by initially empty (A) and pre-filled (B) bags submerged in a water-filled tank versus time interval. The regression line (A) was determined from data collected from 0.2 to 45 min after the bags were submerged. For (B), each data point represents the mean of 1 to 3 replicates; average standard error was 20 ml (range 1 to 55 ml).



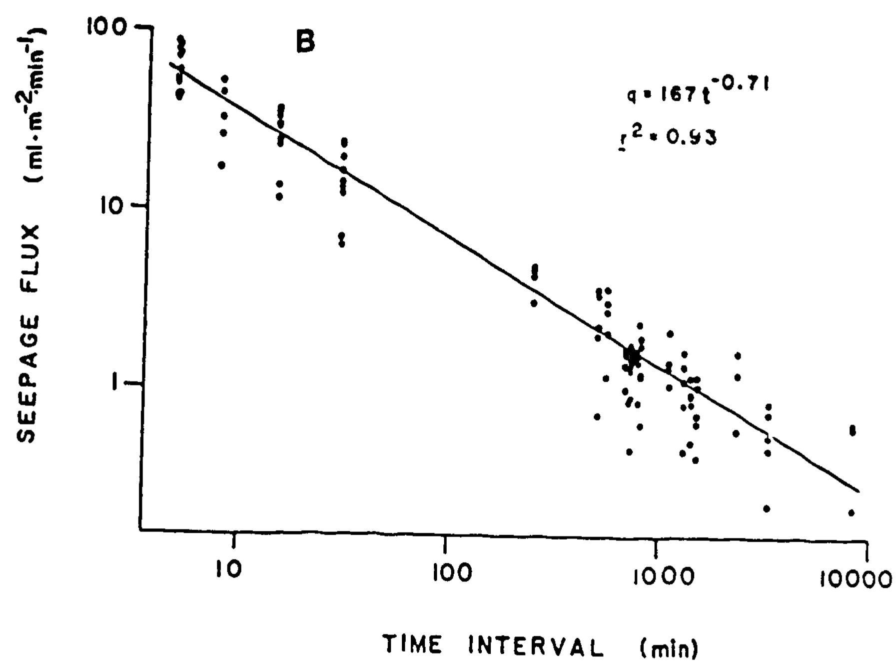


Figure 2.5. A: Volume of water (V) collected by seepage meters versus time interval (t) for Narrow Lake data in 1984. Geometric mean and 95 % confidence limits are shown for t greater than 30 min. B: Seepage flux versus t, calculated from data collected at Narrow Lake during 1984.

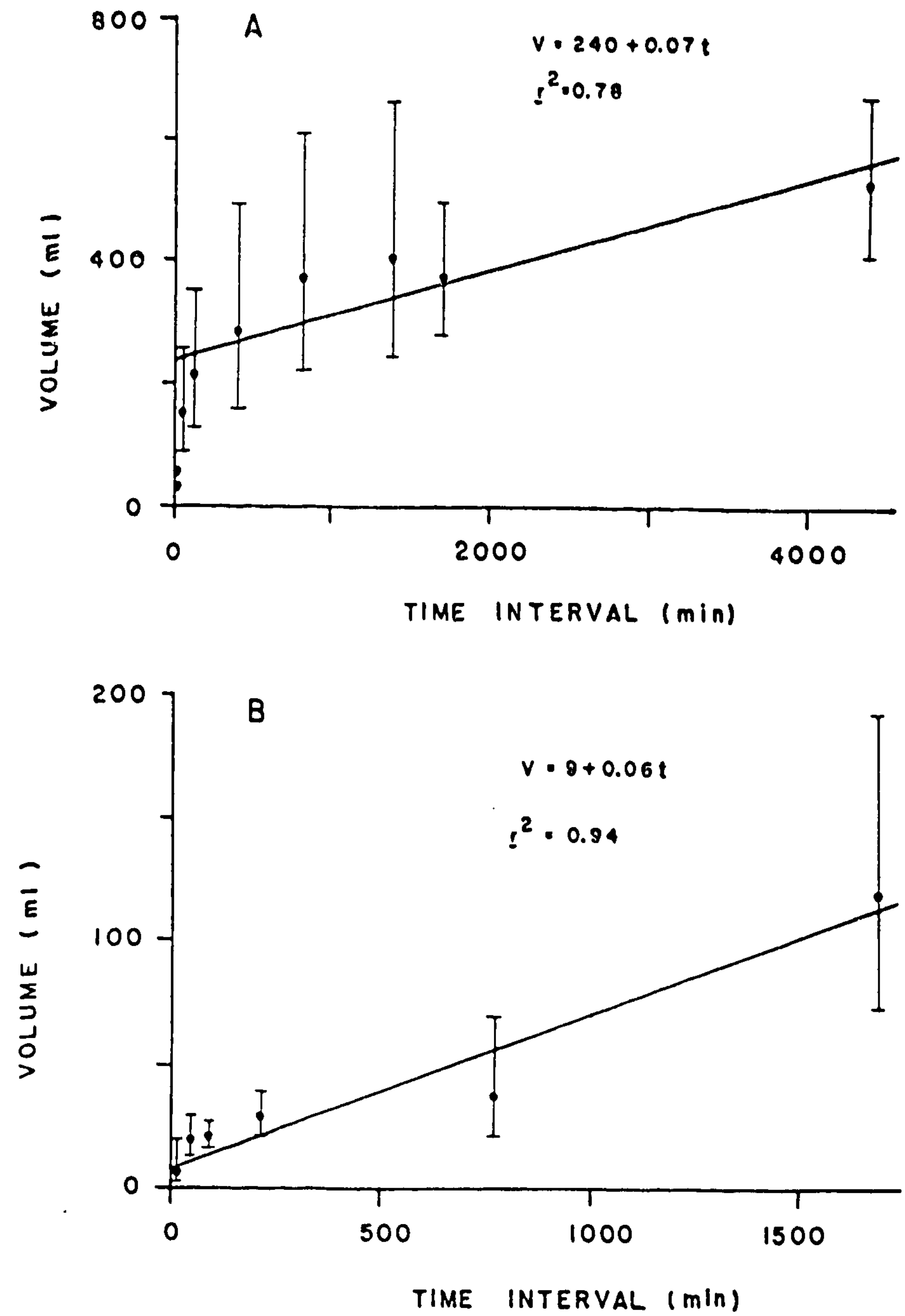


Figure 2.6. Geometric mean and 95 % confidence intervals for volume of water versus time interval for (A) initially empty and (B) 1,000-ml prefilled bags, at Narrow Lake in 1988. For (A), the regression line was determined from data collected at time intervals greater than 30-min; for (B), the regression was determined from data collected at all time intervals.

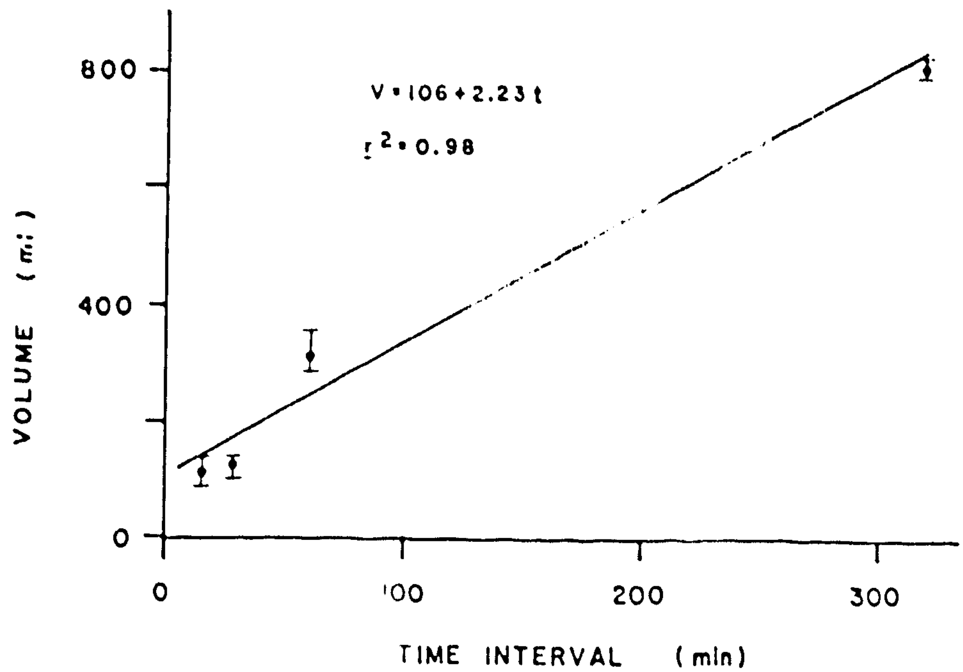


Figure 2.7. Geometric mean and 95 % confidence intervals for volume of water versus time interval for initially empty bags at Buffalo Lake in 1987.

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3. ACCURACY OF SEEPAGE METER ESTIMATES OF LAKE SEEPAGE[1]

3.1 ABSTRACT

The use of seepage meters to identify nearshore seepage patterns and to quantify seepage in lakes was evaluated with a Monte Carlo simulation model. The model simulated seepage flux, as would be derived from seepage meter measurements, along a transect extending from the shore of a hypothetical lake to 40 or 100 m off-shore. Along the transect, simulated seepage velocities decreased exponentially with distance from shore according to patterns measured at Narrow Lake, Alberta and Lake Sallie, Minnesota. To determine statistical parameters needed in the model, seepage flux was measured _in situ_ with closely-spaced seepage meters at four different sites in Narrow Lake. Seepage velocities within a small area of lakebed were log-normally distributed, and the variance was positively correlated to mean seepage flux. The modeling indicated that the most sensitive parameter affecting the accuracy of seepage meter estimates was the variability in the spatial distribution of seepage flux within a small area of lakebed. There was little improvement in the accuracy of estimates of seepage patterns or flux when more than 10 seepage meters were simulated along the transect, when the transect was "sampled" more than twice, or when seepage meters along the transect were simulated to follow a stratified rather than a systematic design.

[1]A version of this chapter has been accepted for publication. R.D. Shaw and E.E. Prepas. J. Hydrol.

3.2 INTRODUCTION

The number of studies of groundwater-lake interactions has increased dramatically over the past decade. This increased attention coincides with the introduction of seepage meters, a simple device to measure groundwater-lake flux. Seepage meters set in transects perpendicular to the shore of lakes have been used to identify nearshore seepage patterns (McBride and Pfannkuch, 1975; Brock _et al_., 1982; Chapter 4). Whole-lake seepage (i.e., groundwater component of lake water budget) has been quantified by extrapolating these transect data over the entire lake (Brock _et al_., 1982; Belanger and Mikutel, 1985; Chapter 4). However, little attention has been paid to factors that affect variability of seepage meter estimates of groundwater-lake flux.

In this paper, a stochastic approach was used to evaluate factors that affect the ability of seepage meters to accurately identify seepage patterns and quantify groundwater-lake flux. Seepage meters were sampled _in situ_ at Narrow Lake, Alberta, to evaluate statistical parameters of seepage flux and to quantify a nearshore seepage pattern in the lake. A computer model was developed to simulate seepage flux in lakes, as would be measured with seepage meters. Seepage velocities were simulated at sites along a transect in a hypothetical lake where seepage influx to the lake decreased with distance from the lake shore. With a Monte Carlo method, the effects of variability in spatial distribution of seepage flux within a small area of lakebed and the number and placement of seepage meters along the transect on estimates of seepage patterns and groundwater-lake flux were evaluated. I discuss how errors in seepage meter estimates of groundwater-lake flux can be reduced.

<u>Study Area</u> - Narrow Lake (54°35'N, 113°37'W) is a small, mesotrophic lake in the mixed-wood section of the Boreal forest of central Alberta (surface area 1.1 km^2, mean depth 14 m, mean summer chlorophyll *a* [illegible] $.m^{-3}$; Prepas and Trimbee, 1988). The lake is situated in a glacial meltwater channel. Glacial till is the predominant surficial deposit in the drainage basin; alluvial sand and gravel lenses are interbedded in the till. Hydraulic conductivity (K) of the surficial deposits ranges from 5×10^{-7} to 1×10^{-5} $m.s^{-1}$ (Chapter 5). The Wapiti Formation, an Upper Cretaceous, sandstone and siltstone unit, underlies the surficial deposits (Green, 1972). Precipitation and evaporation average 503 and 636 $mm.yr^{-1}$, respectively (Hydrology Branch, Alberta Environment, unpublished data).

3.3 FIELD STUDY

<u>Data Collection</u> - To simulate realistic seepage flux as would be measured with seepage meters, the frequency distribution and standard deviation of seepage flux within a small area of lakebed was required. In addition, the variance must be independent of mean seepage velocity.

Seepage meters were sampled at four sites in Narrow Lake to evaluate the preconditions for the simulation model (Fig. 3.1). At each site, three to five seepage meters were placed within an area of 2 m^2. Mean, standard deviation, and variance of seepage flux were determined four to six times per site from 9 to 13 August 1984 (Fig. 3.1). The frequency distribution of seepage flux was evaluated at sites 1 and 2 (Fig. 3.1); there were too few data to assess frequency distributions at the other sites. At sites 1 and 2, data collected from 9 to 13 August 1984 were pooled and then compared to a normal distribution (Kolmogorov-Smirnov test; Sokal and Rohlf, 1981). Homogeneity of variance was evaluated from data collected at all sites with the F_{max}-test (Hartley, 1950).

A nearshore groundwater flow pattern was measured at Narrow Lake on 7 June 1986. The flow pattern was determined with a transect of 10 seepage meters placed 1 to 40 m from shore (Fig. 3.1).

Seepage meters were constructed and sampled according to Lee (1977). SCUBA divers installed seepage meters at lake depths greater than 1 m. Alligator plastic bags (3.5-L capacity) were attached after a 2- to 5-d equilibration period. Seepage flux ($\underline{v}$, $m.s^{-1}$) was computed from the volume (m^3) of water in the bag after appropriate corrections for area (m^2) of bottom sediments enclosed by the seepage meter and length of time (s) the bag was attached to the meter (1984: 12 h; 1986: 30 h; Chapter 2).

<u>Field Results</u> - Seepage influx to Narrow Lake at the four sites sampled in 1984 ranged from $1.6x10^{-8}$ to $5.1x10^{-8}$ $m.s^{-1}$; variance ranged from $1.2x10^{-17}$ to $5.8x10^{-16}$ (Table 3.1). Seepage velocities measured with closely-spaced seepage meters in Narrow Lake were log-normally distributed (Site 1: $\underline{D}$=0.083, $\underline{n}$=25, $\underline{P}$>0.20; Site 2: $\underline{D}$=0.12, $\underline{n}$=27, $\underline{P}$>0.20). To my knowledge, the frequency distribution of seepage flux has not been statistically evaluated in other studies. However, a log-normal distribution for seepage flux is reasonable. Hydraulic conductivity of most porous media, even those that are considered homogeneous, has a log-normal distribution (Freeze, 1975). Thus, the

frequency distribution of seepage flux probably reflected variation of hydraulic conductivity in bottom sediments.

At Narrow Lake, variance of seepage flux was significantly correlated with mean seepage flux ($r=0.60$, $df=18$, $P<0.01$). The spatial variability of seepage flux, as measured with closely-spaced seepage meters, has not been reported for most studies. However, Brock *et al.* (1982) give two sets of seepage flux from Lake Mendota, WI. Each set was measured with three seepage meters placed within 3 m of each other; mean seepage velocities were 7.9×10^{-7} and 8.1×10^{-7} $m.s^{-1}$ and variance was 5.8×10^{-14} and 4.8×10^{-14}, respectively. The simultaneous increase of variance and mean seepage flux at Lake Mendota, relative to that at Narrow Lake, is consistent with my observation that mean and variance are highly correlated.

Variance and mean of log-transformed seepage flux were not correlated to one another ($r=0.36$, $df=18$, $P>0.05$), and variance of log seepage flux was homogeneous ($F_{max}=31$, $df=20,3$, $P>>0.05$) as required for the simulation model. The standard deviation of log-transformed seepage flux ($s_{(\log v)}$) at the four sites in Narrow Lake ranged from 0.09 to 0.28 log $m.s^{-1}$ (Table 3.1). Seepage flux at Lake Mendota was more than 15 times that at Narrow Lake; even so, the standard deviation of log seepage flux from Lake Mendota (0.12 and 0.14 log $m.s^{-1}$) were within the range measured at Narrow Lake. Thus, values of the standard deviation of seepage flux that were measured at Narrow Lake may be representative of those at other lakes.

Along the transect sampled in 1986 at Narrow Lake, seepage flux, v decreased exponentially with distance from shore, x (in m; $P<0.0001$):

$$\underline{v} = 4\text{x}10^{-8}\ 10^{-0.014\underline{x}} \quad (1)$$

A general pattern of decreasing seepage velocity with distance from shore has been predicted from theoretical investigations of groundwater-lake interactions and has been observed in many lakes (McBride and Pfannkuch, 1975; Brock et al., 1982; Chapter 4).

3.4 MODEL STUDY

A computer model was developed to simulate seepage flux along a transect in a hypothetical lake as would be measured by seepage meters along a transect in an actual lake. With a Monte Carlo method, I evaluated the effect of spatial variability of seepage within a small area of lakebed, the seepage pattern along the transect, and the number and placement of seepage meters along the transect on estimates of nearshore seepage patterns and groundwater-lake flux.

Model Development - Consider that along a transect in a hypothetical lake, seepage to the lake is constant over time, constant across the width of the transect, and decreases exponentially with distance from shore according to:

$$\underline{v} = \alpha 10^{\beta \underline{x}} \quad (2)$$

where $\alpha > 0$ and $\beta < 0$. The average seepage velocity ($\bar{\underline{v}}$, $m.s^{-1}$) along the transect is:

$$\bar{\underline{v}} = \frac{1}{\underline{z}} \int_{\underline{x}=0}^{\underline{x}=\underline{z}} \alpha 10^{\beta \underline{x}} d\underline{x} \quad (3)$$

where $\underline{z}$ is the length (in m) of the transect. Thus, $\dot{\underline{y}}$ is weighted for variation in seepage flux along the transect.

The field study indicated that seepage flux within a small area of lakebed is log-normally distributed. Therefore, a log-normally distributed seepage velocity, $v_{\underline{x}}$ (m.s^{-1}), can be generated for a site at a distance from shore, $\underline{x}$, along the transect by:

$$\log v_{\underline{x}} = \underline{s}_{(\log \underline{v})} R + \log \underline{v}_{\underline{x}} \tag{4}$$

where $\underline{v}_{\underline{x}}$ is the seepage flux at that site calculated from Eq. (2), $\underline{s}_{(\log \underline{v})}$ is the standard deviation of log seepage flux, and R is a normal random deviate from a normally distributed population with a mean of 0 and standard deviation of 1. The seepage flux generated with Eq. (4) is analogous to that measured with a seepage meter. Seepage flux can be generated from Eq. (4) for $\underline{n}$ sites along the transect. Each site corresponds to a seepage meter placed along a transect in a hypothetical lake. The seepage pattern along the transect can then be quantified by log-linear regression of v on $\underline{x}$:

$$v = \underline{a}10^{\underline{b}\underline{x}} \tag{5}$$

where $\underline{a}$ and $\underline{b}$ are estimates of α and β, respectively. Equation (5) represents a seepage pattern that might be derived from data collected with seepage meters along a transect where seepage flux decreased with distance from shore according to Eq. (2).

The accuracy of Eq. (5) depends on factors affecting the generation of $v_{\underline{x}}$ ($\underline{s}_{(\log \underline{v})}$ and R in Eq. 4) and the number of sites and distances from shore in which seepage flux is generated. If these factors (except for R) are held constant, and the simulation is repeated $\underline{i}$ times (where $\underline{i}$ is the number of Monte Carlo runs), then each regression of v on $\underline{x}$ (Eq. 5) will be different. Each of the $\underline{i}$ runs is analogous to sampling one transect of seepage meters. If seepage meters can accurately identify the nearshore seepage pattern indicated in Eq. (2) (and $\underline{i}$ is large), then mean $\underline{a}$ ($\bar{\underline{a}}$) and mean $\underline{b}$ ($\bar{\underline{b}}$) should converge to the values of α and β, respectively (Eq. 2).

For each of the $\underline{i}$ runs of the model, the average seepage flux, $\bar{v}$, along the transect was calculated by integrating Eq. (5), as shown for the regression of $\underline{v}$ on $\underline{x}$ in Eq. (3), i.e., after $\underline{a}$ and $\underline{b}$ were substituted for α and β, respectively. If seepage meters can accurately quantify the average seepage flux along the transect, then mean $\bar{v}$ (from a large number of Monte Carlo runs) should converge to $\bar{\underline{v}}$ (Eq. 3).

In reality, a transect of seepage meters is seldom sampled more than a few times, and $\bar{\underline{a}}$, $\bar{\underline{b}}$, and $\bar{v}$ from these few replicates may not converge to α, β, and $\bar{\underline{v}}$, respectively. The potential for a single replicate ($\underline{i}$=1) to accurately identify a nearshore seepage pattern was evaluated from the sampling distribution of $\underline{b}$, as determined from 500 Monte Carlo runs of the model. First, the percent of the 500 runs that correctly identified the flow pattern of decreasing seepage with distance from shore was quantified as the proportion of $\underline{b}$'s that were significantly less than 0. Second, the 95 % confidence interval (L), expressed as a percent of β, i.e., $L_\beta = 100L/\beta$, was evaluated, where:

$$L = \frac{1.96 \underline{s}_{\underline{b}}}{\sqrt{\underline{i}}} \tag{6}$$

and $\underline{s}_{\underline{b}}$ is the sample standard deviation of $\underline{b}$. With $\underline{i}$=500, $\underline{s}_{\underline{b}}$ would be an accurate estimator of the population standard deviation of β (σ_β). Accordingly, the 95 % confidence intervals would be β-L to β+L. As $\underline{i}$ increases, the 95 % confidence interval decreases and $\bar{\underline{b}}$ converges to β.

The accuracy of seepage meter estimates ($\check{v}$) of the average seepage flux along a transect ($\bar{\underline{v}}$) was evaluated in a similar manner to nearshore seepage patterns. The accuracy of $\check{v}$, where it was determined from a limited number of replicates, was evaluated with Eq. (6) after the standard deviation of $\check{v}$ was substituted for $\underline{s}_{\underline{b}}$.

The simulation model was programmed in FORTRAN and run on an AMDAHL 5870 computer.

Simulation Conditions - Seepage flux was simulated along a transect in a hypothetical lake where seepage flux decreased with distance from shore according to two different flow patterns; one was measured at Narrow Lake (Eq. 1) and the other at Lake Sallie ($\underline{v} = 6\mathrm{x}10^{-7}\ 10^{-0.017\underline{x}}$; Lee, 1972). Transects of 40 and 100 m in length were simulated for the patterns from Narrow and Sallie lakes, respectively. These lengths correspond to the distances from shore that seepage meter data were collected from the two lakes. The average seepage flux along the transect in Lake Sallie ($1.6\mathrm{x}10^{-7}$ m.s^{-1}) was seven-fold higher than in Narrow Lake ($2.3\mathrm{x}10^{-8}$ m.s^{-1}).

For both flow patterns, seepage velocities were generated with standard deviations ($\underline{s}_{(\log \underline{v})}$) of 0.05, 0.15, and 0.25 log m.s^{-1}; these values covered the range that were recorded in the field study (Table 3.1). The number ($\underline{n}$) of sites along each transect (3, 5, 10, 15 or 20) was selected to include the minimum number of seepage meters required to identify nearshore seepage patterns and the maximum that would likely be used. The position of the first site was randomly selected at a distance of 1 to 5 m from the shore of the hypothetical lake. The remaining sites were simulated to follow a systematic design; the sites were evenly spaced along the transect from the first site to the end of the transect. In addition, for two conditions ($\underline{s}_{(\log \underline{v})}$=0.15 and $\underline{n}$=10: at both flow patterns) a stratified design was simulated. For the stratified design, half of the sites were simulated within 14 and 18 m from shore at Narrow and Sallie lakes, respectively. Within those distances, one-half of the total discharge of nearshore seepage occurred.

In total, 32 conditions, which represented unique combinations of flow pattern, spatial variability of seepage flux within a small area, and number and placement of seepage meters along a transect were evaluated with the model. For each condition, 500 Monte Carlo runs were carried out; so for each condition 500 independent values of $\underline{a}$, $\underline{b}$, and $\bar{v}$ were obtained. In all cases, $\bar{\underline{a}}$ and $\bar{\underline{b}}$ converged to within 0.2 % of α and β, respectively, where α and β are known from actual field measurements in Narrow Lake and Lake Sallie. However, for all conditions tested, mean $\bar{v}$ was consistently higher, by 1 to 10 %, than $\bar{\underline{v}}$. Since the $\bar{v}$'s were determined with log-normally distributed seepage meter data (Eq. 4), they were also log-normally distributed. Therefore, I log-transformed the $\bar{v}$'s before determining their mean and standard deviation (mean(log $\bar{v}$) and $\underline{s}_{(\log \bar{v})}$, respectively). For all conditions, mean(log $\bar{v}$) was within 0.2 % of log $\bar{\underline{v}}$.

Model Results - The results from the different conditions were compared to the result from a standard condition. The standard condition was selected to simulate a realistic situation: (1) the nearshore seepage pattern measured on 7 June 1986 at Narrow Lake (Eq. 1), (2) the number of seepage meters that were used to identify that flow pattern (n=10), (3) the average standard deviation of seepage flux in Narrow Lake ($s_{(\log v)}$=0.15) and (4) a systematic sampling design along a transect 0 to 40 m from shore. For the standard condition, seepage flux decreased significantly with distance from shore in 94 % of the Monte Carlo runs (i=500; $\bar{b}$=-0.014, s_b=0.004, range = -0.026 to -0.003; Fig. 3.2). With i=1, the 95 % confidence limit (L) for the slope of the regression of seepage flux on distance from shore is 56 % of β, where β is the slope of the regression describing the Narrow Lake flow pattern (Eq. 1). Therefore, if a transect of 10 seepage meters were sampled once under the condition tested, there would be (1) a 94 % chance that the pattern of a decrease in seepage with distance from shore would be correctly identified and (2) a 95 % chance that the slope of the regression of seepage flux on distance from shore would be within ± 56 % of β. Increasing the number of times the transect is sampled, from 1 to 4, would halve the 95 % confidence limit from 56 to 28 % of β. However, further increases in sampling would reduce L_β only slightly (Fig. 3.2); e.g., 16 replicates are required to reduce L_β from 28 to 14 %.

Over the range of conditions tested, the spatial distribution of seepage flux within a small area of lakebed was the factor that had the largest impact on the accuracy of seepage meter estimates of flow patterns. With little spatial variability (0.05 log m.s^{-1}) and all other conditions the same as for the standard condition, a pattern of nearshore seepage concentration was correctly identified in all 500 runs (Fig. 3.2); with i=1, L_β was only 18 % (Fig. 3.3). Under that condition, patterns of seepage flux generated by individual runs of the model closely reflected the flow pattern that was tested (Fig. 3.4). In contrast, with high spatial variability of seepage flux (0.25 log $m.s^{-1}$), a pattern of nearshore seepage concentration was correctly identified in only 63 % of the runs; with i=1, L_β was 91 %. With high spatial variability, there was considerable difference between the patterns of seepage flux generated by individual runs of the model compared to the flow pattern that was tested (Fig. 3.4). Under conditions of high spatial variability, conclusions about seepage patterns at lakes based upon seepage meter data may be misleading, e.g., in some cases, the presence of off-shore zones of anomalously high seepage flux were suggested, even though such a pattern was not simulated (Fig. 3.4F).

The impact of the number of seepage meters per transect on the accuracy of estimates of nearshore patterns was less than for spatial variability of seepage flux. By doubling n from 10 to 20 (all other conditions the same as for the standard condition), a pattern of nearshore seepage concentration was identified in 99 % of the Monte Carlo runs (Fig. 3.2). With i=1, L_β decreased only slightly from 56 to 41 %, with n set to 10 and 20, respectively (Fig. 3.3). With n set at 3 (i=1), a pattern of nearshore seepage concentration was identified in only 33 % of the runs, and L_β was 82 %. A transect consisting of only three seepage meters would be of limited use for evaluating nearshore seepage patterns to lakes.

There was less variability in seepage patterns generated with the flow pattern from Lake Sallie compared to that from Narrow Lake (Fig. 3.4). A pattern of nearshore seepage concentration was identified in more than 99 % of the runs tested with the Lake Sallie flow pattern. With i-1, the Lake Sallie flow pattern and the other conditions as for the standard conditions, L_β was only 19 % compared to 56 % for the same conditions with the Narrow Lake flow pattern. For any matched condition, L_β was lower than those from Narrow Lake (Fig. 3.3). Differences in the accuracy of seepage patterns from Narrow and Sallie lakes were due to the relative decrease in seepage flux along the transects: i.e., at Narrow Lake, seepage flux at the end of the transect was 28 % of that at the lake shore; at Lake Sallie, seepage flux at the end of the transect was only 2 % of that at the shore.

Seepage meter estimates of nearshore seepage patterns did not improve with a stratified as compared to a systematic sampling design. For a stratified design and other conditions as for the standard condition, a pattern of nearshore seepage concentration was identified in 94 % of the runs, the same as for the systematic design. With i-1, L_β was 54 % for the stratified design compared to 56 % for the systematic design. With a stratified design and the Lake Sallie flow pattern, the seepage pattern was correctly identified in all runs, and with i-1, L_β was 18 % compared to 19 % for a systematic design.

Seepage meter estimates of the average seepage flux along a transect were affected by the number of replicates, spatial variability within a small area, and number of seepage meters per transect in a similar manner to that described above for estimates of the seepage pattern: (1) spatial variability had the greatest impact on the accuracy of estimates of the average seepage flux along the transect, (2) increases in the number of replicates and number of seepage meters per transect had less effect on the accuracy of estimates of average seepage flux (Fig. 3.5), and (3) a stratified sampling design did not improve the accuracy of estimates of average seepage flux. In contrast to results for seepage patterns, estimates of average seepage flux along a transect with the Narrow Lake flow pattern were slightly more accurate than estimates for the Lake Sallie flow pattern. With the standard condition and $\underline{i}$=1, the lower and upper 95 % confidence limits were 80 and 124 % of the average seepage flux, respectively; at Lake Sallie, they were 75 and 133 % of the average seepage flux, respectively. For any matched condition, the variability of estimates of average seepage flux ($\underline{s}_{(\log v)}$) from the Lake Sallie pattern was 15 to 40 % higher than that from the Narrow Lake pattern (Fig. 3.6).

3.5 APPLICATION TO FIELD STUDIES

Over the range of conditions tested, seepage meter estimates of nearshore seepage patterns and average seepage flux were most strongly influenced by the spatial variability of seepage within a small area of lakebed. Therefore, effort should be directed towards minimizing this variability when seepage meters are used to evaluate groundwater-lake interactions. Variability in seepage flux is caused by both real variation and measurement errors. The real variation is a result of spatial variability of lake sediments. For example, at Narrow Lake seepage flux measured with one seepage meter was consistently higher than that measured by another meter, even though the two seepage meters were only 1-m apart (Fig. 3.7). This variability in seepage probably reflects the stochastic distribution of hydraulic conductivity in lake sediments and cannot be reduced. However, there are errors associated with the use of seepage meters that can be minimized. For instance, seepage flux measured immediately after seepage meters are installed is more variable than when it is measured after the meters are allowed to equilibrate for a few days (Lee, 1977). In addition, the length of time that plastic bags are attached to meters can bias estimates of seepage flux (Chapter 2). However, to my knowledge, there has been no detailed evaluation of measurement errors associated with the use of seepage meters.

Whole-lake seepage has been quantified by extrapolating seepage velocities measured at one location over larger areas of a lake. For example, seepage flux along three transects was extrapolated over 30 km of shoreline to estimate the groundwater contribution to Lake Conway, FL (Fellows and Brezonik, 1980). In the next chapter (Ch. 4), average seepage flux measured along one or two transects per lake were used to estimate whole-lake seepage flux into 10 central Alberta lakes. The significance of errors caused by extrapolating average seepage flux measured at one transect over other areas of a lake is difficult to predict _a priori_; e.g., coarse-grained lenses may intersect the lakebed causing localized areas of high seepage. There was relatively little increase in the accuracy of estimates of nearshore seepage patterns or average seepage flux when transects at one location were repeatedly sampled. Therefore, rather than obtaining replicates at one location, effort should be redirected towards measuring seepage in other areas of the lake.

Based on results of this study, the potential accuracy of seepage meter estimates of nearshore seepage patterns can be assessed for predetermined sampling designs. For example, assume 10 seepage meters were to be used along a transect and seepage flux within a small area of the lakebed was highly variable, e.g., $s_{(\log v)} = 0.25 \log m.s^{-1}$. In a lake where seepage velocity decreases with distance from shore according to the Narrow Lake flow pattern, 70 % of the transects would correctly identify a pattern of nearshore seepage concentration (Fig. 3.2), and one could expect highly variable plots of seepage flux with distance from shore (Fig. 3.4).

The potential accuracy of seepage meter estimates of the average seepage flux along a transect can also be assessed with results from this study. For the conditions described above, the standard deviation of average seepage flux ($s_{(\log \bar{v})}$) was 0.086 (Fig. 3.6). With i=1, L is 0.169 (Eq. 6), and the lower and upper 95 % confidence intervals are 68 and 147 % of the average seepage flux along the transect, respectively. For corresponding conditions at Lake Sallie, $s_{(\log \bar{v})}$ was 0.106, and the lower and upper 95 % confidence intervals were 62 and 161 % of the average seepage flux, respectively. The differences between the 95 % confidence intervals are surprisingly low considering that between Narrow and Sallie lakes the average seepage flux along the transect varied almost an order of magnitude, nearshore seepage patterns differed between the lakes, and the length of the transect differed more than two-fold. This suggests that the modeling results are robust and may be used to design *in situ* sampling programs to monitor groundwater-lake flux at other lakes where seepage flux decreases with distance from shore.

3.6 CONCLUSIONS

I have presented a stochastic analysis of the use of seepage meters to quantify groundwater-lake flux along transects where the seepage flux decreases with distance from shore. The results indicated:

(1) Seepage flux within a small area of lakebed is log-normally distributed, and the variance of seepage flux increases with mean seepage flux.

(2) Seepage meters have the potential to accurately identify nearshore seepage patterns and accurately quantify the average seepage flux along a transect extending from shore to the edge of the nearshore sediments.

(3) The most important parameter that affected the accuracy of these estimates was the spatial variability of seepage flux within a small area of lakebed. The variability is largely due to stochastic properties of the lake sediments.

(4) Relatively little improvement in the accuracy of estimates of seepage patterns or seepage flux is obtained by using more than 10 meters per transect or sampling each transect more than twice.

(5) Results from this study can be used to assist in the design of *in situ* sampling programs for measurement of groundwater-lake interactions.

3.7 NOTATION

α	coefficient of $\underline{v}$ on $\underline{x}$
β	coefficient of $\underline{v}$ on $\underline{x}$
σ_β	standard deviation of β
$\underline{a}$	coefficient of v on $\underline{x}$
$\bar{\underline{a}}$	mean $\underline{a}$
$\underline{b}$	coefficient of v on $\underline{x}$
$\bar{\underline{b}}$	mean $\underline{b}$
$\underline{i}$	number of Monte Carlo runs
K	hydraulic conductivity, $m.s^{-1}$
L	95 % confidence limit
L_β	$100L/\beta$
$\underline{n}$	number of seepage meters per transect
R	normal random deviate from a population with mean=0 and standard deviation=1
$\underline{s}_{\underline{b}}$	standard deviation of $\underline{b}$
$\underline{s}_{(\log \underline{v})}$	standard deviation of log $\underline{v}$
$\underline{s}_{(\log \bar{v})}$	standard deviation of log $\bar{v}$
$\underline{v}$	seepage flux, $m.s^{-1}$
$\underline{v}_{\underline{x}}$	$\underline{v}$ at $\underline{x}$
$\bar{\underline{v}}$	average seepage flux along a transect where $\underline{v}$ decreases exponentially with distance from shore
v	seepage flux generated by the model, $m.s^{-1}$
$v_{\underline{x}}$	v at $\underline{x}$
$\bar{v}$	average seepage flux along a transect, determined by integration of log-linear regression of v on $\underline{x}$ (at $\underline{i}$=1)
$\underline{x}$	distance from lake shore in transect, m
$\underline{z}$	length of transect, m

Table 3.1. Mean and standard deviation of untransformed and $\log_{10}$-transformed seepage flux measured with seepage meters at four sites in Narrow Lake during 1984. Sites are as indicated in Fig. 3.1, and the date in August 1984 and the number (#) of seepage meters sampled on that date are also included.

Site	Date	#	Seepage Flux			
			mean ($\times 10^{-8}$ m.s^{-1})	std. dev.	mean (log m.s^{-1})	std. dev.
1	9	5	2.0	0.86	-7.74	0.23
	9	5	2.5	0.48	-7.61	0.09
	11	5	2.4	0.60	-7.63	0.12
	11	5	2.7	0.93	-7.59	0.16
	13	5	2.3	0.88	-7.67	0.19
2	9	4	2.7	0.34	-7.57	0.05
	9	4	1.8	1.02	-7.81	0.28
	11	5	2.5	1.05	-7.66	0.23
	11	5	1.9	1.01	-7.76	0.26
	13	4	1.6	0.53	-7.83	0.17
	13	4	1.9	0.58	-7.75	0.16
3	9	3	4.7	2.44	-7.36	0.21
	11	3	4.0	0.89	-7.41	0.09
	13	3	4.2	0.85	-7.38	0.08
	13	3	5.1	1.23	-7.30	0.10
4	9	3	3.8	1.52	-7.44	0.20
	11	3	2.4	1.28	-7.66	0.26
	11	3	3.4	1.00	-7.49	0.13
	13	3	3.9	2.20	-7.45	0.22
	13	3	3.7	1.28	-7.44	0.14

SITE	LD	DFS	NSM
1	1	2	5
2	1	2	5
3	1	2	3
4	2	5	3

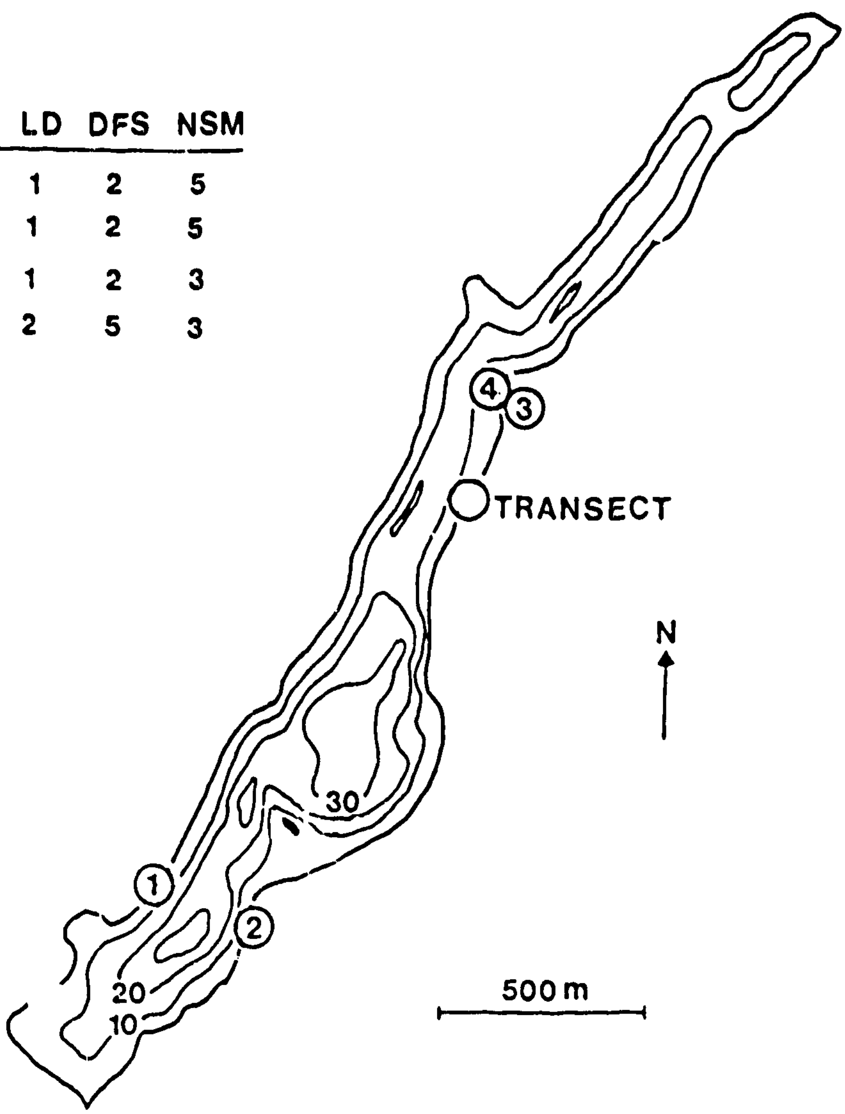

Figure 3.1. Lake depth (LD, in m), distance from shore (DFS, in m), number of seepage meters per site (NSM), and location of the four sites sampled during 1984 at Narrow Lake. The transect was sampled on 7 June 1986 with 10 seepage meters spaced at distances of 1, 5, 7.5, 10, 15, 20, 25, 30, 35, and 40 m from shore (0.5 to 10-m lake depth).

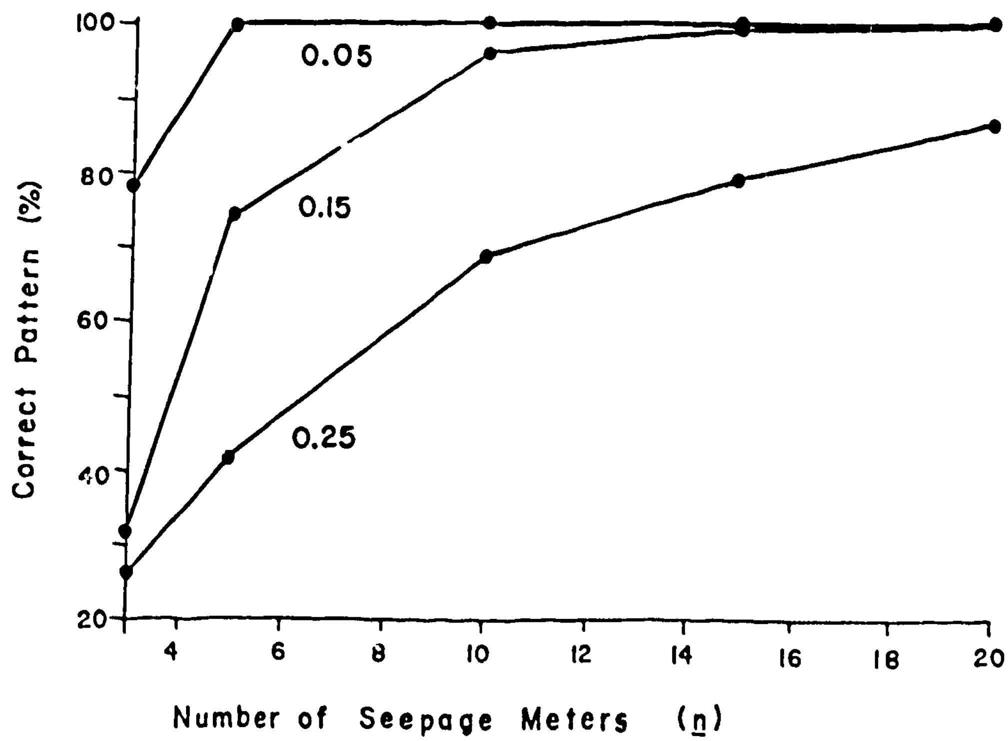

Figure 3.2. Effect of standard deviation of seepage flux (0.05, 0.15, and 0.25 log $m.s^{-1}$) and the number of seepage meters along a transect on the percent of Monte Carlo runs that correctly identified the pattern of decreasing seepage flux with distance from shore. For clarity, only results for the Narrow Lake flow pattern are shown.

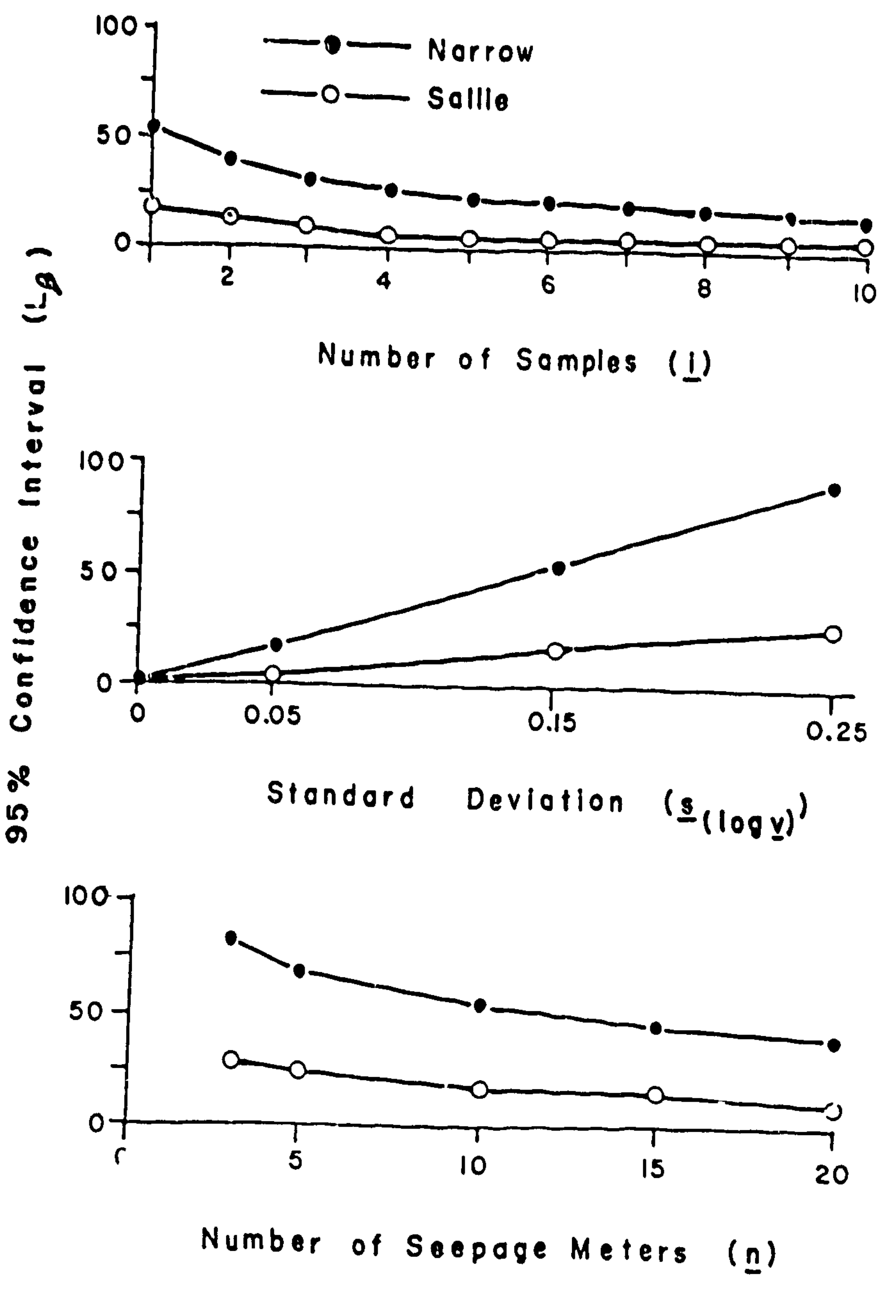

Figure 3.3. Effect of number of replicates (i), standard deviation of seepage flux ($s_{(\log v)}$), and number of seepage meters per transect (n) on L_β for both flow pattern; L_β is the 95 % confidence limit, expressed as a percent of the slope (β) of the log-linear regression of seepage flux versus distance from shore.

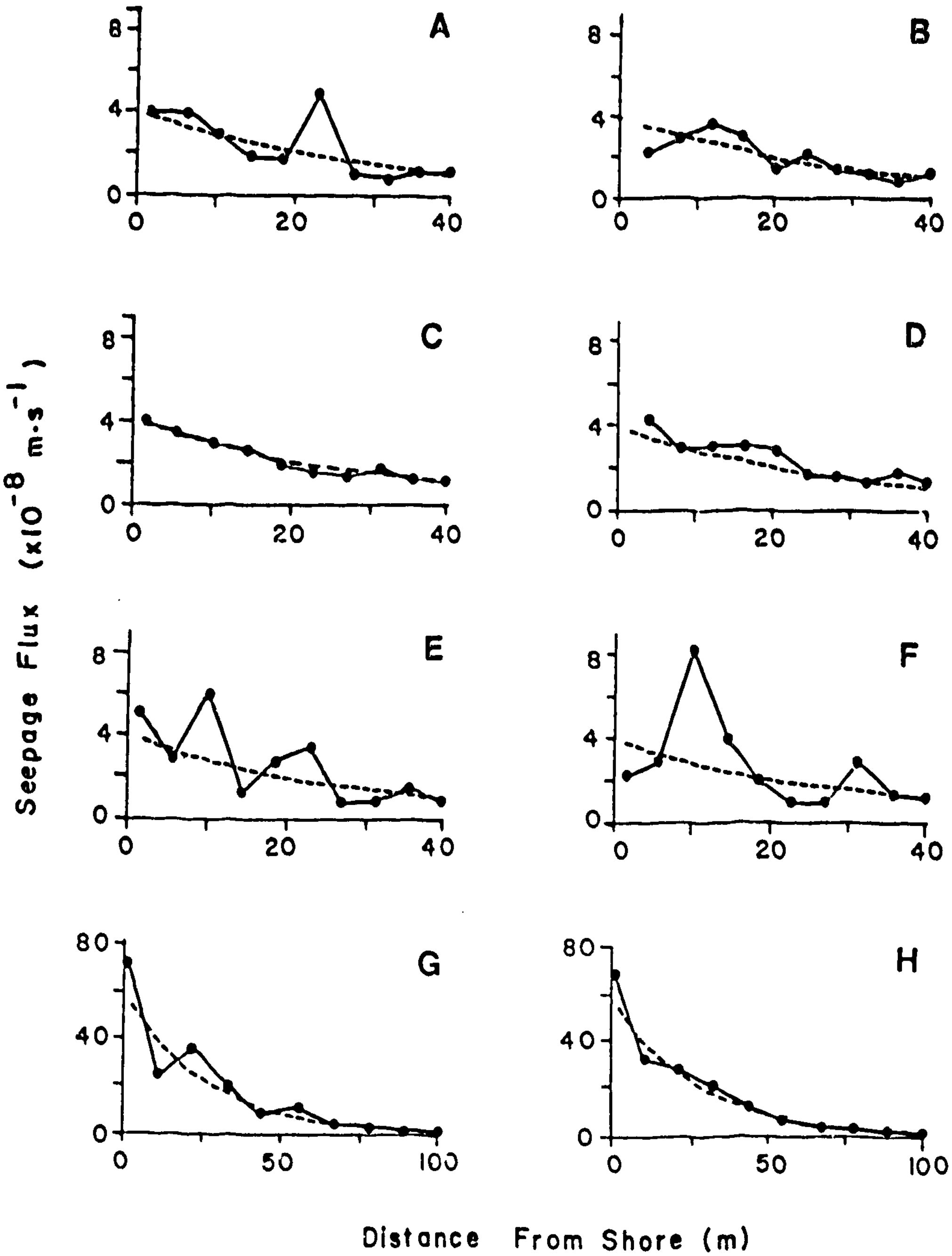

Figure 3.4 Examples of seepage patterns generated from individual runs of the simulation model. The solid dots indicate seepage flux as generated by the model for the seepage pattern shown by the dashed line. A and B: Narrow Lake flow pattern, $\underline{s}_{(\log \underline{v})}=0.15$, $\underline{n}=10$, systematic sampling; C and D: as above with $\underline{s}_{(\log \underline{v})}=0.05$; E and F: as above with $\underline{s}_{(\log \underline{v})}=0.25$; G and H: Lake Sallie flow pattern, other conditions as for A. Except for F, seepage flux generated by the model decreased exponentially with distance from shore ($\underline{P}<0.05$).

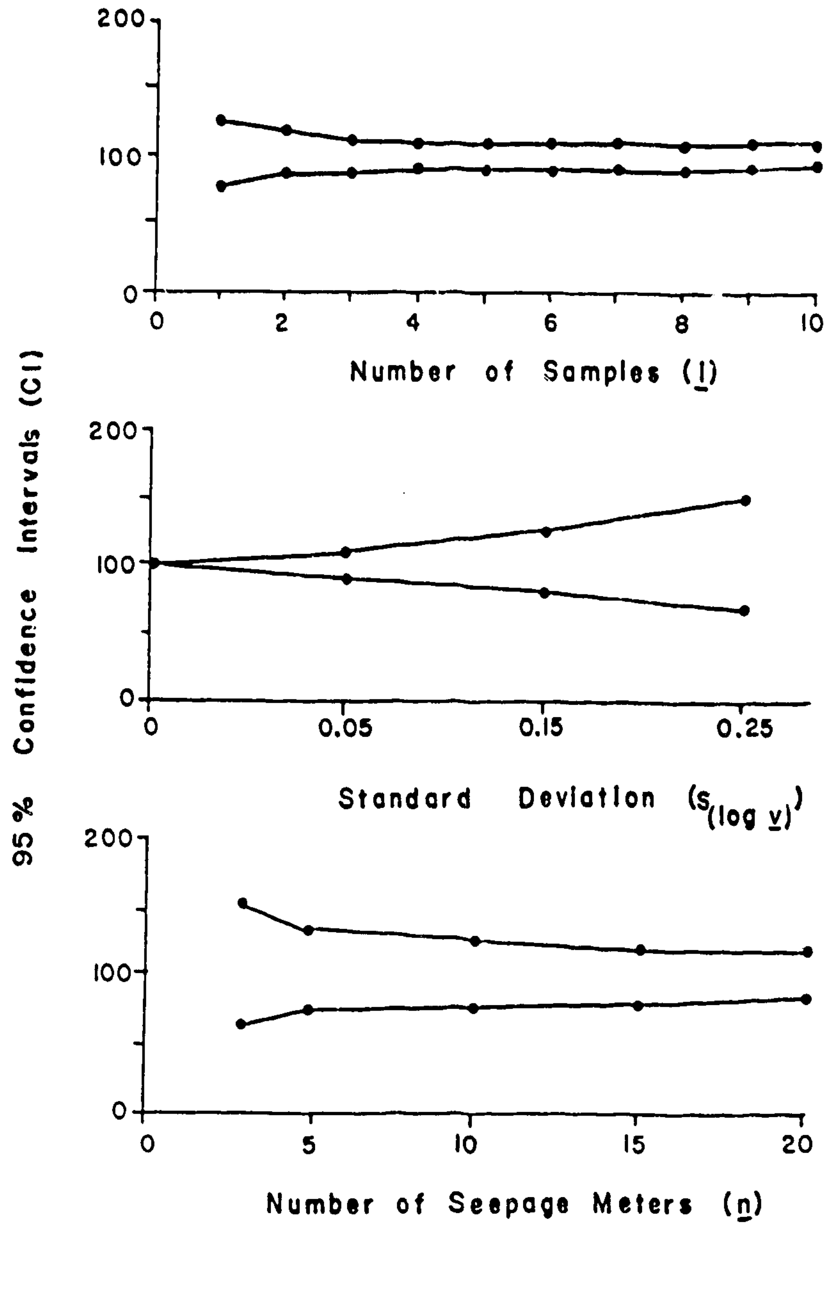

Figure 3.5 The effect of the number of replicates (*i*), standard deviation of seepage flux ($s_{(\log v)}$) and number of seepage meters per transect (*n*) on the lower and upper 95 % confidence intervals (CI) around the average seepage flux along the transect ($\bar{v}$); CI is expressed as a percent of $\bar{v}$. For clarity, only results for the Narrow Lake flow pattern are indicated.

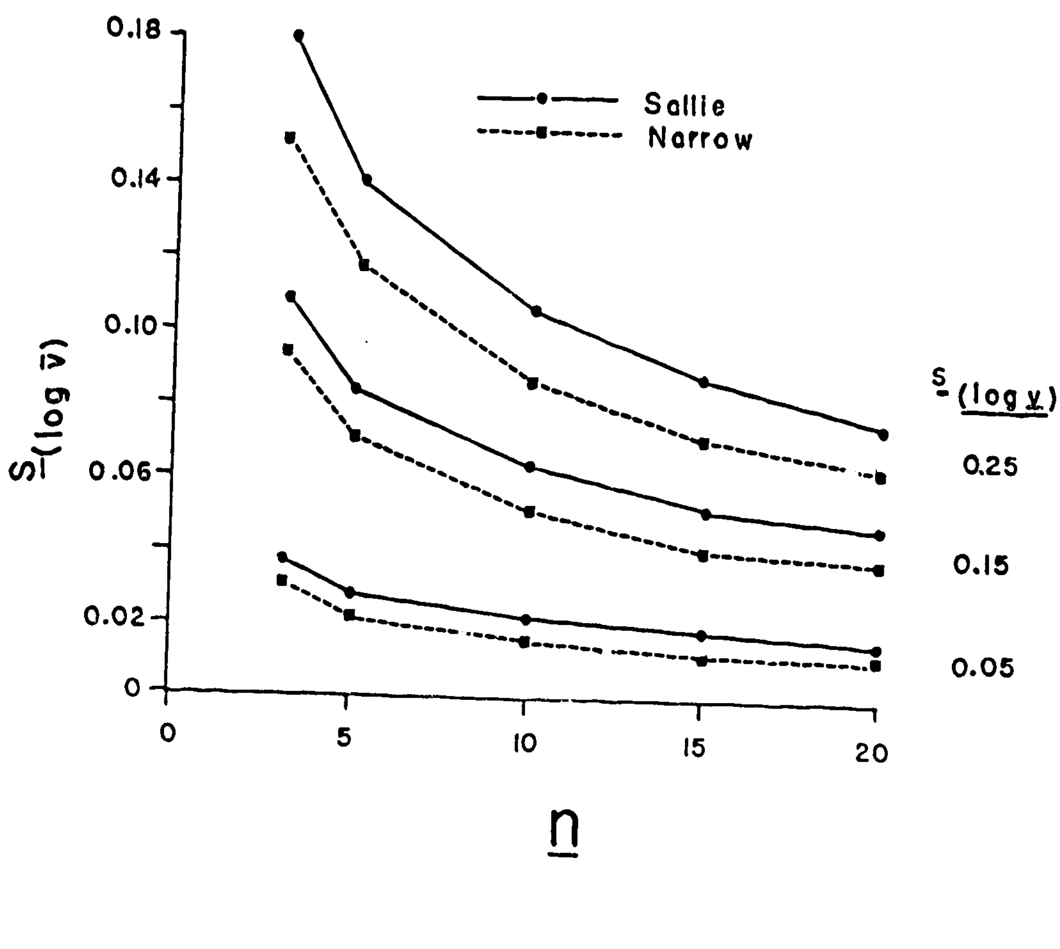

Figure 3.6. The impact of the standard deviation of seepage flux ($\underline{s}_{(\log\ \underline{v})}$) and the number of seepage meters per transect ($\underline{n}$) on the standard deviation of average seepage flux ($\underline{s}_{(\log\ \bar{v})}$) for both flow patterns.

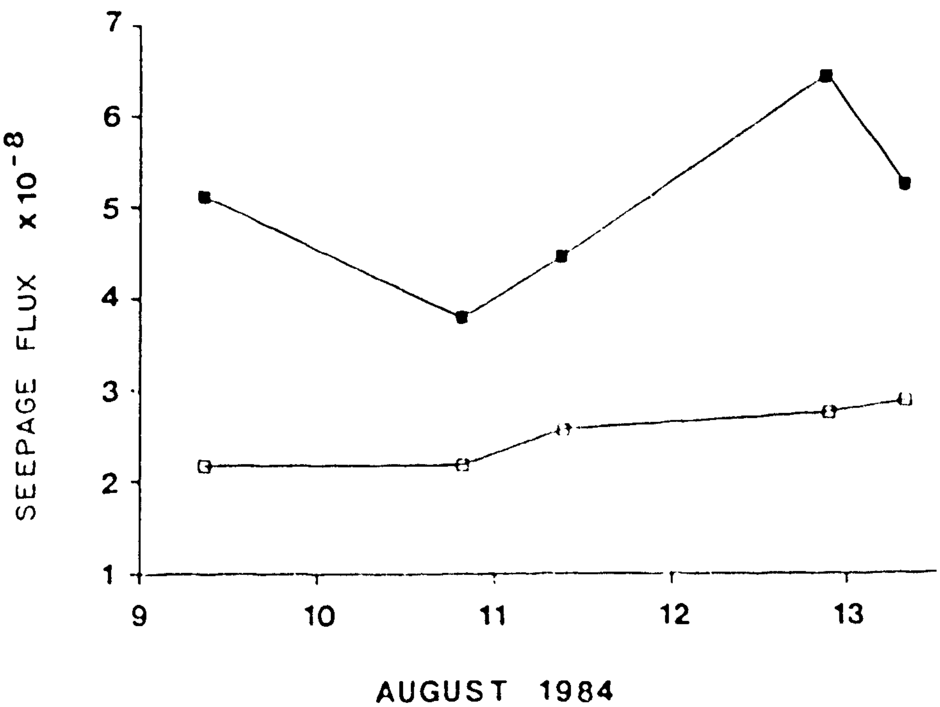

Figure 3.7. Seepage flux ($m.s^{-1}$) measured by two seepage meters spaced 1-m apart at site 4 (Fig. 3.1), in Narrow Lake (□ meter 1, ■ meter 2).

3.8 REFERENCES

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4. NEARSHORE SEEPAGE PATTERNS AND THE CONTRIBUTION OF GROUNDWATER TO LAKES IN ALBERTA[1]

4.1 ABSTRACT

Seepage flux was measured with seepage meters placed along transects from the lake shore to 30-110 m offshore in 10 lakes in central Alberta during May to August 1986. In the study area, the predominant surficial deposit is glacial till which is underlain by sedimentary bedrock. Seepage inflow to the lakes ranged from 3×10^{-10} to 2×10^{-7} $m.s^{-1}$. Seepage out of the lakes was recorded at only 1 of 92 seepage meter sites. At one lake, seepage was measured biweekly along transects at two locations, from May to August 1986; seepage patterns were consistent throughout that period. In the nearshore region of 6 of 10 lakes, seepage inflow to the lakes decreased with distance from shore. Deviations from that pattern were likely a result of: (1) spatial variability of seepage flux within a small area of lakebed, (2) intertill sand and gravel lenses near the lake, and (3) pre-glacial bedrock channels of sand and gravel underlying some of the lakes. Groundwater was the major source of water (49% of total inflow) to one lake; at the other lakes, groundwater was a relatively small component (~10 %) of total inflow.

[1]A version of this chapter has been accepted for publication. R.D. Shaw and E.E. Prepas. J. Hydrol.

4.2 INTRODUCTION

Groundwater may be an important source of water and inorganic and organic compounds to lakes. However, groundwater-lake interactions are rarely quantified, probably because of perceived difficulties in measuring seepage rates. In addition, many factors affect groundwater-lake interactions: e.g., fluctuations in hydraulic gradients near lakes, lake and watershed morphometry, and heterogeneity of porous media near lakes (Winter, 1976; Anderson and Munter, 1981; Winter, 1986). The hydrogeological setting of each lake is a unique combination of these factors, so there is potentially an infinite number of patterns of groundwater-lake flux. Nonetheless, both computer modeling and field investigations of groundwater-lake systems suggest that seepage inflow to lakes is generally highest near the lake shore and decreases with distance from shore, and seepage of lakewater to groundwater does not occur near the shore of lakes that are bounded by water-table mounds (McBride and Pfannkuch, 1975; Winter, 1976; Lee et al., 1980).

To date, most field studies of groundwater-lake interactions have focused on single lakes, and there is little information on the variation in seepage for lakes in similar hydrogeological settings. In this study, seepage flux was measured in situ at lakes in in central Alberta to compare spatial and temporal variability of seepage flux within a lake and to evaluate variation in nearshore seepage flux and seepage patterns for lakes located on sedimentary bedrock.

Hydrogeological Environment - The study lakes are within a 65 000 km^2 area in central Alberta (Fig. 4.1). The lakes range in surface area from 0.07 to 84 km^2, maximum depth from 5 to 36 m, and drainage area from 0.1 to 1530 km^2 (Table 4.1). The climate in the study area is continental; average annual precipitation, lake evaporation, and daily air temperatures are 466 mm, 631 mm, and 3.1 °C, respectively (Environment Canada, 1982). In general, lakes are ice-free from April to November. Most of the study lakes are in the Boreal Mixedwood Ecoregion; aspen (Populus tremuloides) and balsam poplar (Populus balsamifera) are common, and Luvisolic soils are predominant. The two most southerly lakes (Buffalo Lake and S-7) are in the Aspen Parkland Ecoregion; aspen and willows (Salix sp.) are common, and Chernozemic soils are predominant (Strong and Leggat, 1981). There is agricultural activity near most lakes: barley, canola, oats, and wheat are common crops, and beef cattle graze in areas of rough topography.

The study area is in the Interior Plains physiographic province (Bostock, 1970). Marine shales (LaBiche and Lea Park Formations) underlie three lakes (Green, 1972) and have little potential for significant groundwater yields (Borneuf, 1973; Crowe and Schwartz, 1981, Table 4.1). The Wapiti Formation, a non-marine sandstone and siltstone unit, which contains scattered coal beds, underlies five lakes; hydraulic conductivity (K) of this unit is about 10^{-5} $m.s^{-1}$ in areas of fractured rock or coal seams (Alberta Research Council, unpublished data). The Horseshoe Canyon Formation, a non-marine sandstone, mudstone, and shale unit underlies two lakes; K is about 10^{-5} $m.s^{-1}$ in areas of fractured rock (Clare and Ko, 1982). Sand and gravel deposits lie at the base of buried preglacial valleys underneath Baptiste, Buffalo, Minnie and Tucker lakes. Hydraulic conductivity of these deposits are in the order of 10^{-6} $m.s^{-1}$ near Baptiste Lake (Crowe and

Schwartz, 1981) to 10^{-4} $m.s^{-1}$ near Tucker Lake (Alberta Environment, 1983). Generally, only bedrock sandstones and pre-glacial buried channels of sand and gravel are important for municipal water supply (Lennox, 1965).

Surficial deposits in the watershed of nine study lakes are of glacial origin (Table 4.1); one (S-7) is an endcut pond formed in spoil from a reclaimed coal strip mining site. In the study area, glacial till is the predominant material, although intertill sand and gravel lenses are common. Hydraulic conductivity of the surficial material is highly variable and ranges from 10^{-10} to 10^{-4} $m.s^{-1}$ in clayey-till, and sand and gravel, respectively (Crowe and Schwartz, 1981; Clare and Ko, 1982; Chapter 5). Intertill sand and gravels can be sources of domestic and farm water supply; till yields only small amounts of water to shallow wells. Water table levels tend to reflect topographic elevations and are highest during spring snow-melt and after heavy summer rainfall. A cross-section of the hydrogeological setting of a hypothetical lake in the study area is shown in Fig 1.3.

4.3 MATERIALS AND METHODS

Nearshore seepage was measured in situ with seepage meters. Seepage meters were constructed by cutting off the top or bottom 15 cm from "45-gallon" drums (Lee, 1977). Meters were set, open-end down, about 8 cm into lake sediments, then allowed to equilibrate for 2 to 3 days before they were sampled. Most of the study lakes were bounded by water-table mounds, which suggested seepage into the lakes through the nearshore sediments (Crowe and Schwartz, 1981; Clare and Ko, 1982; MLM Ground-Water Engineering, 1985; Chapter 5). Therefore, in most lakes empty plastic bags (Alligator Baggies: 3.5-L capacity) were attached to seepage meters. At S-7 and Tucker lakes, hydraulic gradients near the lake suggested the potential for seepage from the lake to groundwater (Alberta Environment, 1983; Trudell et al., 1986). At those two lakes, bags were prefilled with 500 mL of water before they were attached to seepage meters.

At most lakes, seepage meters were sampled over 1 to 2 days, at two time intervals: (1) a short interval of approximately 1 h and (2) a long interval of 4 to 30 h. Seepage flux was corrected for the anomalous, short-term influx of water into the plastic bags after they are attached to seepage meters (Chapter 2). At Minnie Lake, seepage meters were only sampled after bags were attached for 16 h, so seepage flux was not corrected for the anomalous influx of water.

Seepage Patterns - From May to August 1986, 10 seepage meters were placed along one transect, perpendicular to shore, in the nearshore zone of each study lakes. At Baptiste and Narrow lakes, two transects were placed in each lake. The transect sites were selected to be away from areas of recreational activity and be representative of general slopes in the lake. In each transect, one seepage meter was placed as close to shore as possible given sediment and rooted plant conditions. The remaining nine seepage meters were evenly spaced along the transect to either the distance from shore where sediments were too soft for proper installation of seepage meters or a maximum distance of 110 m from shore.

At Narrow Lake, intralake variability of nearshore seepage flux was examined. Two transects were located 200-m apart, directly across the lake from each other, on the east and west shore (Narrow-East and Narrow-West, respectively). The seepage meter transects were installed in May 1986 and left in place for the remainder of the summer. At both transects, seepage flux was measured biweekly from 26 May to 5 August 1986. At Narrow Lake, seepage meters were sampled after 24 h at all dates, and also after 1 h on 10 July 1986; those 1 h data were used for correcting seepage flux for the other sampling dates.

For each lake, nearshore seepage patterns were evaluated graphically, by plotting seepage flux, $\underline{v}$ (in $m.s^{-1}$) against distance from shore, $\underline{x}$ (in m). In addition, except for S-7 (where the transect extended across the full width of the lake) patterns were evaluated statistically, by regression of $\underline{v}$ on $\underline{x}$; a slope ($\underline{b}$) significantly less than 0 indicated that seepage flux decreased along the transect.

In Chapter 3, I used computer simulations to assess the use of seepage meters to identify nearshore seepage patterns and average seepage flux along transects where seepage decreased with distance from shore, For the simulated conditions with 10 seepage meters per transect, the nearshore seepage pattern would be identified correctly by 90 % of the transects. The nearshore seepage pattern would not be correctly identified by 10 % of the transects because of inherent variation in seepage flux due to heterogeneity of sediments within small areas of the lakebed. In addition, the modeling indicated that the average seepage flux, measured along 95 % of the transects, should be within 80 to 124% of the "actual" average seepage flux. Even with five meters per transect, 80 % of transects sampled should correctly identify the pattern of nearshore seepage concentration, and for 95 % of the transects, the average seepage flux should be within 73 to 137% of the actual average nearshore seepage flux. Therefore, the sampling design used in this study should be adequate to identify nearshore seepage patterns in lakes where seepage flux decreases with distance from shore.

Average Seepage Flux - The average seepage flux ($\dot{\underline{v}}$, in m.s^{-1}) along each transect was calculated as the area under the curve of seepage flux ($\underline{v}$) on distance from shore ($\underline{x}$), as measured by planimetry, divided by the length of the transect (in m). Thus, average seepage flux was weighted for distance from shore. A jackknife method was used to reduce bias in the estimate of average seepage flux and to provide a standard error so confidence intervals could be computed (Sokal and Rohlf, 1981). For each transect, an average seepage flux ($\dot{\underline{v}}_{-i}$) was computed as for $\dot{\underline{v}}$, based on the data set with each of the $\underline{i}$ seepage meter sites left out in turn. Pseudovalues (v) were computed as:

$$v = \underline{n}\dot{\underline{v}} - (\underline{n} - 1)\dot{\underline{v}}_{-i}$$

where $\underline{n}$ is the total number of seepage meters per transect. Thus, for each transect, the jackknife procedure resulted in $\underline{n}$-1 estimates of average seepage flux. The jackknifed average seepage flux was computed from the geometric mean of the pseudovalues, and confidence intervals computed from standard errors of log-transformed pseudovalues

Whole-lake Seepage Flux - The rate of seepage (in m^{3} s^{-1}) through the nearshore zone of the study lakes was estimated by integrating the jackknifed estimate of average seepage flux over the area of lake covered by nearshore sediments. The nearshore region was arbitrarily defined as the distance from the lake shore to the end of the transect of seepage meters. Seepage meter transects extended from 4 to 100% of the distance from shore to the middle of the lakes; the nearshore zone, as defined in this paper, covered 5 to 100% of the surface area of the lakes (Table 4.2).

The contribution of nearshore seepage as a source of water to the study lakes was compared to precipitation falling directly on the lakes and to surface run-off from the watershed to the lakes. Average annual precipitation (in $m.yr^{-1}$) was based on long-term records from the meteorological station closest to the lake (Environment Canada, 1982). Average annual surface runoff to the lakes (in $m.yr^{-1}$) was estimated from studies in which streamflow to the lakes were gauged (Trew et al., 1981, 1987; Alberta Environment, 1987; Chapter 5), or from a regional analysis of hydrometric stations in the area (Hydrology Branch, Alberta Environment, unpublished data). Surface runoff was not available for S-7. To facilitate comparisons between groundwater, precipitation and surface runoff, the nearshore seepage rate was converted to a whole-lake seepage flux (in $m.yr^{-1}$). Whole-lake seepage flux was calculated by dividing the nearshore seepage rate ($m^3.s^{-1}$) by the surface area of the lake (m^2) and extrapolating that flux over a one year period.

4.4 RESULTS AND DISCUSSION

In total, seepage meters were installed at 120 sites in 10 lakes, and data were collected from, on average, 8 of every 10 seepage meters per transect. Missing data were from sites where meters were disturbed or plastic bags attached to meters were ripped. At nine sites, the volume of water collected after a short interval was greater than that collected after a long interval; at those sites seepage flux was assumed to be nil.

Uncorrected estimates of seepage flux that were calculated from data collected after bags were attached to seepage meters for 1 h (short interval) were as much as 22-fold higher than seepage flux measured from bags attached to the same meter for 4 to 30 h (long interval). These differences were most dramatic when seepage flux was relatively low; e.g., < $6x10^{-8}$ $m.s^{-1}$, as measured with seepage meters sampled after the long interval (Fig. 4.2A). These results are consistent with the hypothesis that there is an anomalous, short-term influx of water to bags after they were attached to seepage meters (Chapter 2). Seepage flux corrected for the anomalous influx of water varied linearly with uncorrected seepage flux that was measured with seepage meters sampled after the long interval (Fig. 4.2B). Thus, uncorrected seepage flux, as measured with seepage meters sampled after 4 to 30 h, could be used to identify seepage patterns in lakes. With the exception of seepage flux at Minnie Lake (where seepage meters were sampled at only one time interval); all values in the remainder of the paper are corrected for the short-term, anomalous inflow to seepage meter bags

Intralake Variation - During this study, seepage flux into Narrow Lake ranged from 0 to $6.2x10^{-8}$ $m.s^{-1}$. For individual seepage meter sites in Narrow Lake, the coefficient of variation (CV) of seepage flux measured from May to August 1986, ranged from 11 to 219%; the average CV was 61%. Over the study period, variation of seepage flux within Narrow-East (CV 50%) and Narrow-West (CV 70%) was similar to temporal variation (CV 61%). Even so, consistent trends in seepage flux along both transects in Narrow Lake were evident (Fig. 4.3). At Narrow-West, seepage flux decreased significantly (P<0.05) with distance from shore on 4 of 6 sampling dates (Table 4.3). On one other date (10 July 1986), seepage flux tended to decrease with distance from shore ($P<0.10$). At Narrow-West, there was, consistently, a zone of relatively high seepage flux approximately 20 to 30 m from shore (Fig. 4.3). At Narrow-East, seepage flux did not decrease with distance from shore ($P>>0.10$; Fig. 4.3). However, similar to Narrow-West, seepage flux increased 20 to 25 m from shore. Average seepage flux along the two transects were significantly different ($t = 5.3$, $df = 5$, $P<0.01$); average seepage flux at Narrow-East was 185 % higher than at Narrow-West.

In previous studies, temporal variation of seepage flux was correlated with mean daily rainfall (Downing and Peterka, 1978; Carignan, 1985). For most of the study period, neither the levels of Narrow Lake nor the elevations of nearby water tables fluctuated in synchrony with rainfall (Fig. 4.4). Therefore, hydraulic gradients would not have been affected by rainfall, and throughout most of the study period, seepage did not fluctuate with rainfall. However, on one date (18 July 1986), seepage may have been affected by rainfall. From 9 to 18 July, there were unusually heavy storms, in total 157 mm rain (31% of average annual precipitation). Total phosphorus concentrations in the epilimnion, and levels of Narrow Lake and water table near the lake increased in response to that heavy rainfall (Chapter 7; Fig. 4.4). Interestingly, the only time when seepage flux did not decrease with distance from shore was 18 July; that sample was probably impacted by the heavy rainfall. Water-table configuration is an important variable affecting groundwater-lake interactions (Winter, 1981); thus, a change in elevation may have caused the observed increases in seepage flux on 18 July.

Interlake Variation - Groundwater seepage into the study lakes was recorded at nearly all seepage meter sites (86 of 92 sites). Seepage from the lake to groundwater was recorded at only two sites (Tucker Lake: -2.4×10^{-10} and -1.5×10^{-9} $m.s^{-1}$). Seepage flux into the lakes varied three orders of magnitude from 3×10^{-10} to 2×10^{-7} $m.s^{-1}$, at S-7 and Baptiste lakes, respectively. In general, seepage tended to be low, relative to that measured with seepage meters at other lakes (Table 4.4). Low seepage flux at the study lakes probably reflects the predominance of glacial till in central Alberta; till is generally a poor medium for groundwater flow.

Seepage flux decreased significantly ($P<0.05$) along transects in five of nine lakes: Tucker, Island, Long, Minnie, and Narrow-West (Table 4.3). At S-7, the transect extended across the lake and seepage was highest near both shores and decreased towards the middle of the lake (Fig. 4.5). In all transects, there was considerable deviation from the pattern of seepage decreasing with distance from shore. Some of this variability may be due to the random placement of seepage meters along transects, since seepage flux measured by seepage meters 1-m apart can be affected by heterogeneity of lake sediments (Chapter 3). The offshore zones of relatively high seepage flux, which was observed at some lakes (e.g., Minnie, Jenkins, and Narrow-West; Fig. 4.5), may indicate connections between the lakebed and coarse-grained materials (e.g., Krabbenhoft and Anderson, 1986), or the presence of intertill sands and gravel lenses near the lake (Chapter 5).

At two transects, Baptiste-North and Buffalo Lake, seepage flux increased significantly with distance from shore (Table 4.3). At Baptiste-South, seepage decreased, then increased with distance from shore. The seepage patterns observed at Baptiste and Buffalo lakes may be a result of an off-shore hydraulic connection between the underlying aquifer (i.e., preglacial sand and gravels) and lake bottom sediments. Hydraulic gradients in the aquifers are upwards, towards these lakes; therefore, the aquifers may discharge groundwater into the lakes (Clare and Ko, 1982; Crowe and Schwartz, 1981). Interestingly, at Minnie and Tucker lakes, which also have underlying aquifers, seepage flux decreased with distance from shore. The contrast in seepage patterns at lakes with similar hydrogeological settings (e.g, aquifers under lake) illustrates the complexity of groundwater-lake interactions. At Minnie Lake, hydraulic gradients between the lake and aquifer were small (MLM Ground-Water Engineering, 1985); at Tucker lake, the gradients were downward towards the aquifer (Alberta Environment, 1986). At both lakes, there was groundwater seepage into the lake near shore, probably from local groundwater flow systems. At Tucker Lake, there was a loss of water from the lake to the groundwater, 90 m from shore, probably because of recharge to the aquifer. The seepage patterns observed at Tucker Lake are similar to those generated by computer simulations of hypothetical lakes with underlying aquifers (Winter, 1976).

Groundwater Component of Water Balance- Average seepage flux ($\bar{v}$) along transects at different locations within the same lake were much less variable than between lakes. For Baptiste and Narrow lakes, $\bar{v}$ ranged less than 2-fold between the two transects within each lake (Table 4.5). Whereas between the 10 study lakes, $\bar{v}$ ranged 24-fold (5×10^{-9} to 1.2×10^{-7}, at S-7 and Buffalo lakes, respectively). For the study lakes, the lower and upper 95% confidence limits averaged 61 and 311% of $\bar{v}$, respectively (Table 4.5).

Whole-lake seepage flux to the study lakes ranged from 0.04 to 0.94 $m.yr^{-1}$ (Table 4.5). Average annual flux of precipitation and surface runoff ranged from 0.39 to 0.50 $m.yr^{-1}$ and 0.05 to 1.86 $m.yr^{-1}$, respectively (Table 4. 5). At Spring Lake, groundwater was the major source of water. At the other lakes, groundwater ranged from 4 to 26 % of the annual water inflow (mean 12 %).

The estimates of whole-lake seepage flux are subject to a number of sources of error because (1) nearshore seepage flux was calculated from only one or two transects per lake, and seepage flux can vary greatly at different sites in the same lake, e.g., in flow-through lakes, groundwater enters the lake on one side and water is lost from the lake to groundwater at the other side of the lake, (2) seepage flux was not measured in the offshore regions of the lake, and (3) except for Narrow Lake, temporal variation in seepage was ignored; seepage flux during winter may be lower than during summer.

This study provides the first estimates of groundwater-lake flux for most of the study lakes, so it is difficult to evaluate the accuracy of most of the estimates of whole-lake seepage flux. However, at three lakes (Baptiste, Buffalo, and Narrow) the groundwater component of lake water budget has been previously estimated. For Baptiste Lake, groundwater was evaluated with a hydrologic simulation model (Crowe and Schwartz, 1981). Results from that study suggested groundwater contributed 13 % of the annual inflow of water to the lake; that value was remarkably similar to my estimate (11 %). For Buffalo Lake, groundwater was evaluated with hydrogeological methods (Clare and Ko, 1982) and estimated at 8 % of the annual inflow of water to the lake

(Alberta Environment 1987); that value was about one-third of my estimate of 26 %. My value for Buffalo Lake may be overestimated because seepage meters were sampled at a site near outwash deposits. Other area of the lake watershed are composed of less permeable glacial till and glacio-lacustrine deposits. For Narrow Lake, groundwater was evaluated with several different methods (Chapter 5) and estimated at 30 % of the annual inflow of water; that value was about 1.6 times higher than my estimate of 19 %. The generally close agreement between my estimates and other estimates of whole-lake seepage flux indicate that the sampling design used in this study is useful for providing a preliminary indication of the relative importance of groundwater-lake flux.

4.5 CONCLUSIONS

This study of groundwater-lake interactions in central Alberta indicated:

(1) At Narrow Lake, seepage flux derived from seepage meter measurements varied both temporally and spatially. Even so, nearshore seepage patterns remained fairly constant throughout the study period. Except for a period of unusually heavy rainfall, seepage flux did not fluctuate measurably with rainfall.

(2) Seepage flux varied greatly between the 10 study lakes. Groundwater seepage into the lakes was measured at nearly all seepage meter sites. In the nearshore region, seepage from lakes to groundwater was rare.

(3) In most cases, groundwater seepage into lakes was highest near shore and tended to decrease with distance from shore. However,

deviations from that pattern were frequent and may have been caused by variability of seepage flux measured by closely-spaced seepage meters, stochastic properties of lake sediments, coarse-grained material either intersecting the lakebed or positioned near the lake, or the presence of preglacial valleys underlying lakes.

(4) Seepage through the nearshore sediments contributed, on average, 15 % of the total water input to 10 lakes in an area where till is the predominant surficial deposit in the watershed.

(5) The sampling design used in this study is useful for evaluating groundwater-lake interactions. Nearshore seepage patterns and whole-lake seepage estimates (including confidence intervals for those estimates) can be obtained.

Table 4.1. Location, surface area (Ao), maximum depth (z), watershed area (Ad), type and lithology of the surficial deposits in the watershed, and the underlying bedrock formations (LaBiche = LB; Lea Park = LP; Horseshoe Canyon = HC; Wapiti = WP) of the study lakes.

Lake	Location		Ao	z	Ad	Surficial Geology		Bedrock
	N	W	km2	m	km2	Deposits	Lithology	
Baptiste[1,2]	54o45'	113o33'	9.2	28	309	ground moraine	clayey- to silty-till	LB
Buffalo[3,4]	52o03'	112o55'	84	7	1530	outwash	sand, gravel	HC
Island[1,5,6]	54o53'	113o33'	7.3	15	71	ground moraine	till	WP
Jenkins[1,5,6,7]	54o55'	113o36'	1.8	18	98	rough-broken land of variable origin	till, sand, gravel	WP
Long[6,7]	54o35'	113o38'	1.6	28	27	till and alluvium	till, sand, gravel	WP
Minnie[8,9]	54o17'	111o07'	0.9	25	4	glacial-fluvial	till, sand	LP
Narrow[5,6]	54o37'	114o37'	1.1	36	8	till and alluvium	till, sand and gravel	WP
S-7[10]	52o30'	112o10'	0.07	5	0.1	spoil	clayey-till	HC
Spring[11,12]	53o31'	114o06'	0.8	10	10	glacial-lacustrine	sand	WP
Tucker[13,14]	54o32'	110o36'	7.2	7	15	glacial-fluvial	till, sand	LP

References: [1]Trew et al., 1987; [2]Crowe and Schwartz, 1981; [3]Alberta Environment, 1987; [4]Clare and Ko, 1982; [5]Borneuf, 1973; [6]Kjearsgaard, 1972; [7]Prepas et al., 1988; [8]MLM Goundwater-Engineering, 1985; [9]Alberta Environment, 1986; [10]Trudell and Moran, 1982; [11]Ozary, 1972; [12]Hydrology Branch, Alberta Environment, unpublished data; [13]Trew et al., 1981; [14]Alberta Environment, 1983.

Table 4.2. Length of seepage meter transect (L, in m), distance from lake shore to middle of lake (MID, in m), and area of nearshore zone (NS), expressed as a percent of lake surface area.

Lake	L m	MID m	NS %
Baptiste-South	65	1500	28
Baptiste-North	100	1500	20
Buffalo	110	2500	5
Island	47	500	14
Jenkins	50	250	21
Long	30	125	22
Narrow-East	40	95	29
Narrow-West	40	95	29
S-7	25	25	100
Spring	50	190	48
Tucker	100	750	24

Table 4.3 Summary of statistical evaluation of nearshore seepage patterns. At transects where seepage flux ($\underline{v}$, in $m.s^{-1}$) were ≤ 0 $m.s^{-1}$, linear regressions were determined; others were analyzed by log-linear regression of $\underline{v}$ versus distance from shore ($\underline{x}$ in m). Only those relationships where slopes were significantly ($\underline{P}<0.05$) different from 0 are indicated.

Lake	Date (1986)	Regression	$\underline{P}$
Narrow-West	26 May	$\underline{v} = 3x10^{-8} - 9x10^{-10}\underline{x}$	<0.01
Narrow-West	7 June	$\underline{v} = 3x10^{-8} - 5x10^{-10}\underline{x}$	<0.0005
Narrow-West	23 June	$\underline{v} = 4x10^{-8}10^{-0.04x}$	<0.01
Narrow-West	5 August	$\underline{v} = 2x10^{-8}10^{-0.03x}$	<0.025
Baptiste-North	11 May	$\underline{v} = 1x10^{-9}10^{0.02x}$	<0.005
Buffalo	3 July	$\underline{v} = 1x10^{-7}10^{0.002x}$	<0.025
Island	6 June	$\underline{v} = 2x10^{-8}10^{-0.02x}$	<0.05
Long	26 May	$\underline{v} = 2x10^{-8}10^{-0.03x}$	<0.05
Minnie	21 June	$\underline{v} = 5x10^{-8}10^{-0.01x}$	<0.025
Tucker	20 June	$\underline{v} = 1x10^{-8} - 2x10^{-10}\underline{x}$	<0.01

Table 4.4 Range of seepage flux measured by seepage meters at other lakes.

Reference	Lake	Seepage Flux ($m.s^{-1}$)
Cherkauer and McBride (1988)	Michigan, WS	$4x10^{-10}$
Connor and Belanger (1981)	Washington, FL	$-4x10^{-8}$ to $5x10^{-7}$
Downing and Peterka (1978)	Metigoshe, ND	$6x10^{-8}$ to $2x10^{-7}$
Fellows and Brezonik (1980)	Conway and Apopka, FL	0 to $1x10^{-6}$
Krabbenhoft and Anderson (1986)	Trout, WS	$1x10^{-7}$ to $5x10^{-7}$
Lee (1977)	Mendota, WS	$3x10^{-7}$ to $5x10^{-7}$
Lee (1977)	Movil, MN	$8x10^{-7}$
Lee (1977)	Sallie, MN	$1x10^{-8}$ to $3x10^{-6}$
Lock and John (1978)	Taupo, New Zealand	$2x10^{-8}$ to $6x10^{-6}$
This study		$-2x10^{-9}$ to $2x10^{-7}$

Table 4.5. The contribution of groundwater to the study lakes. Geometric mean, lower, and upper 95% CI of the jackknifed estimate of average seepage flux along the transects and whole-lake seepage flux (GW) for the study lakes. For Narrow-East and Narrow-West, average transect seepage was calculated as the average of the biweekly samples. The contribution of groundwater to total inflow of water is expressed as a percent of whole-lake seepage flux to total inflow, where total inflow is precipitation (P) plus surface runoff (SR) plus GW; values for P and SW are were from the studies indicated in footnotes.

Lake	Average seepage flux			GW	P	SR	Percent GW
	mean	lower	upper				
	$x10^{-8}$ $m.s^{-1}$				$m.yr^{-1}$		%
Baptiste-S	2.2	1.7	2.8	0.20	0.49	1.49	9
Baptiste-N	4.4	4.0	9.4	0.28	0.49	1.49	12
Baptiste-Mean				0.24	0.49	1.49	11
Buffalo	11. 8	8.7	16.4	0.19	0.43	0.12	26
Island	0.9	0.6	1.4	0.04	0.49	0.50	4
Jenkins	1.5	0.3	2.5	0.11	0.49	0.45	11
Long	0.9	0.3	2.5	0.07	0.50	0.40	7
Narrow-E	1.9	1.4	2.5	0.17	0.50	0.05	24
Narrow-W	1.0	0.8	1.3	0.09	0.50	0.05	14
Narrow-Mean				0.13	0.50	0.05	19
S-7	0.5	0.2	1.3	0.16	0.39	-	-
Spring	6.2	0.6	57.9	0.94	0.50	0.47	49
Tucker	0.7	0.3	1.4	0.21	0.46	1.86	8

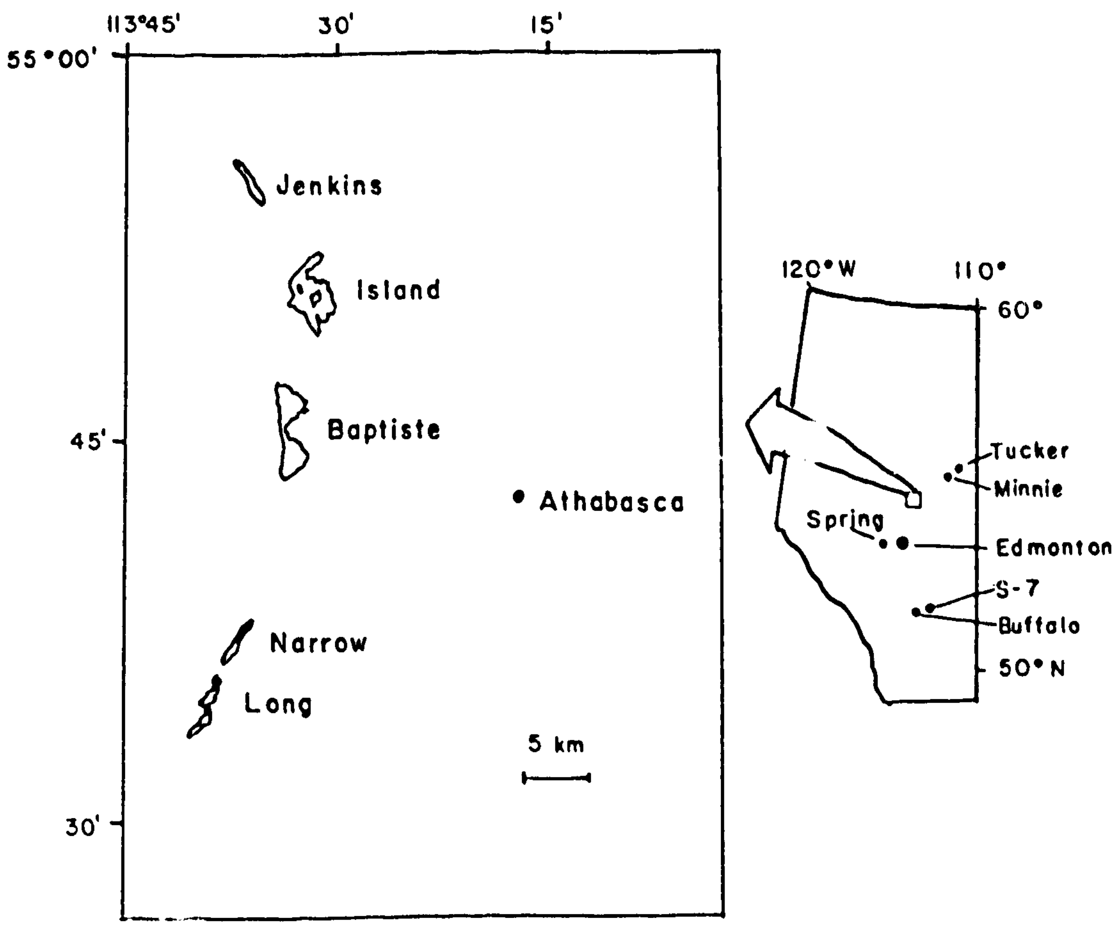

Figure 4.1. Location of study lakes in Alberta.

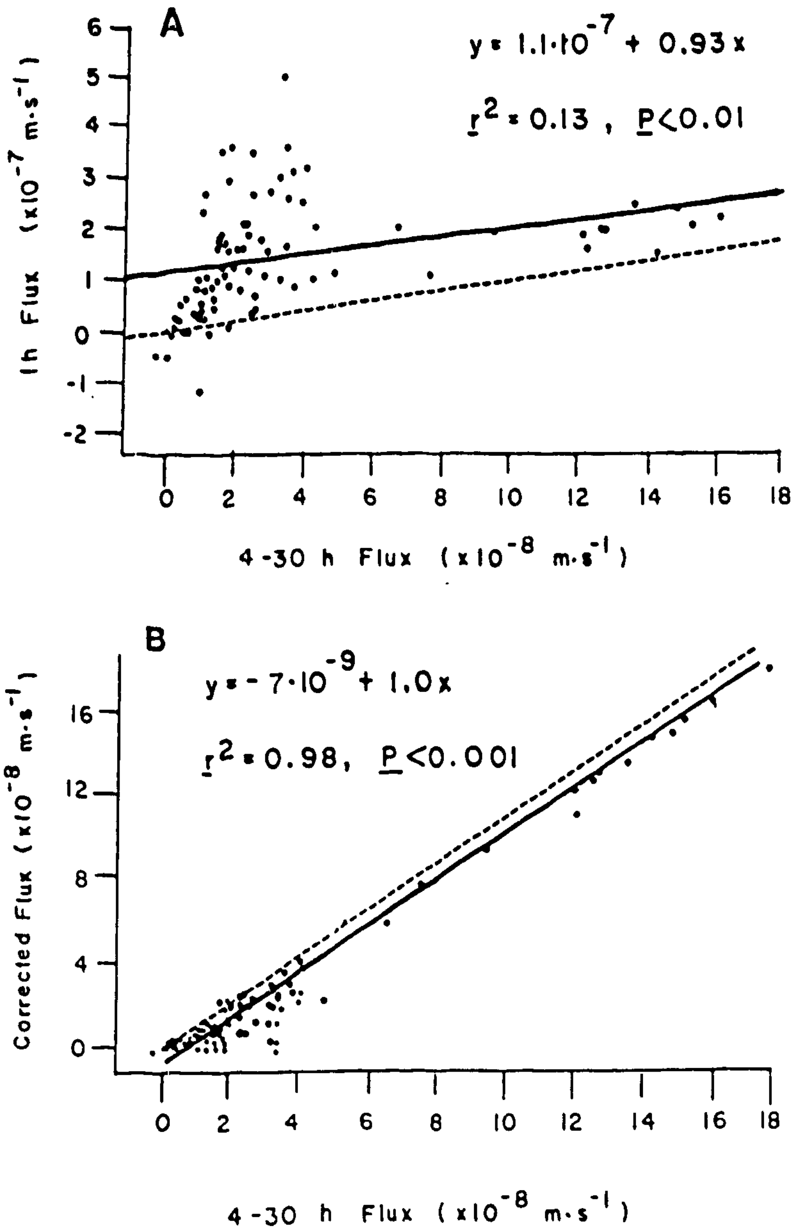

Figure 4.2. A: Uncorrected seepage flux estimated from data collected from bags attached to seepage meters for 1 h versus flux estimated at the same site from bags attached for 4 to 30 h. B: Seepage flux corrected for anomalous volume of water in bag versus uncorrected flux from meters sampled after 4 to 30 h. The dashed lines indicate the relationship that would exist if there were no short-term, anomalous inflow of water to the bags.

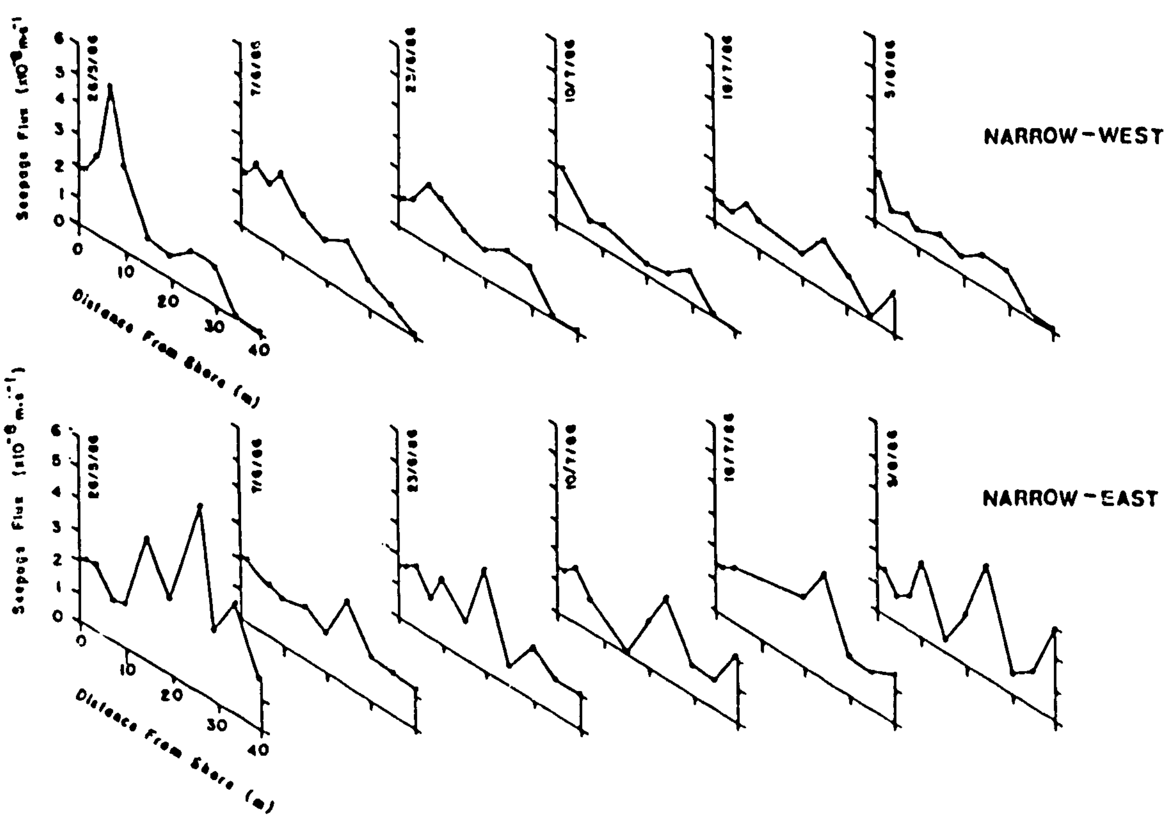

Figure 4.3. Nearshore seepage patterns along Narrow-West and Narrow-East.

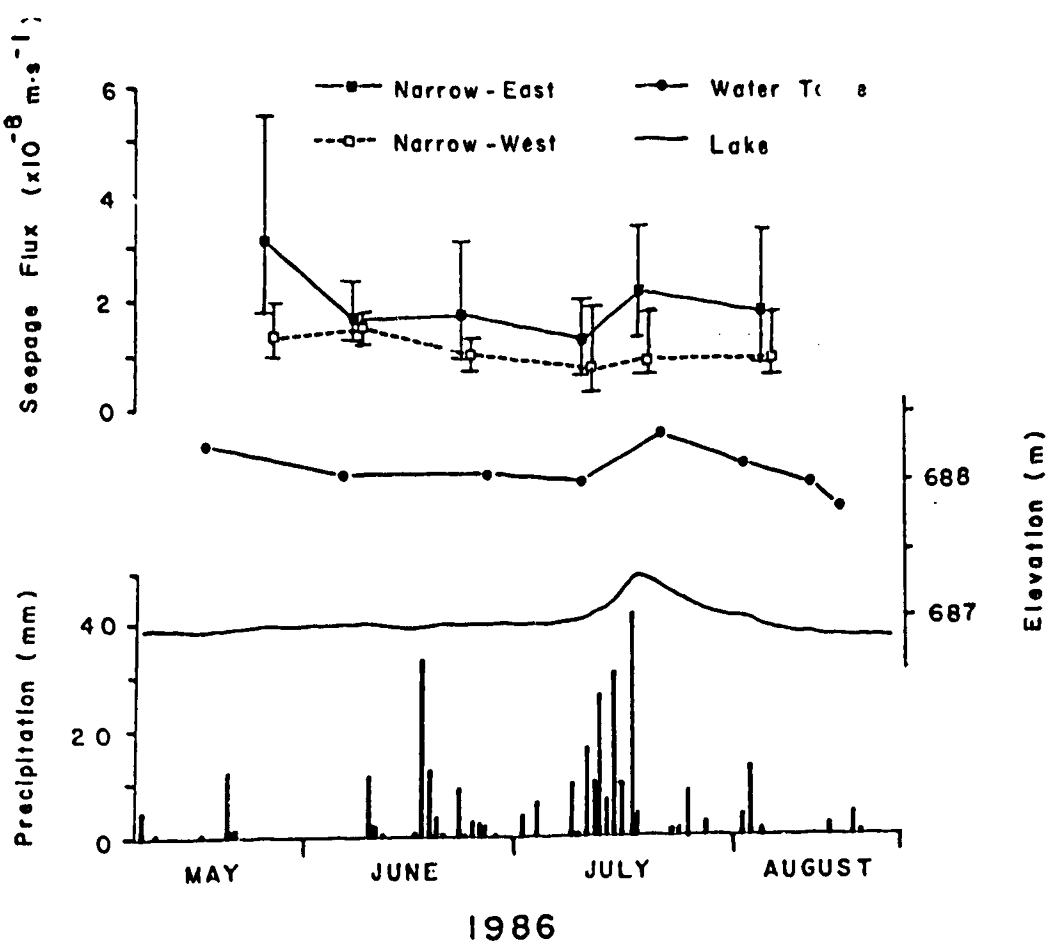

Figure 4.4. Average seepage flux along Narrow-East and Narrow-West compared to rainfall, and elevation of the lake and nearby water table. Vertical bars indicate lower and upper 95% CI.

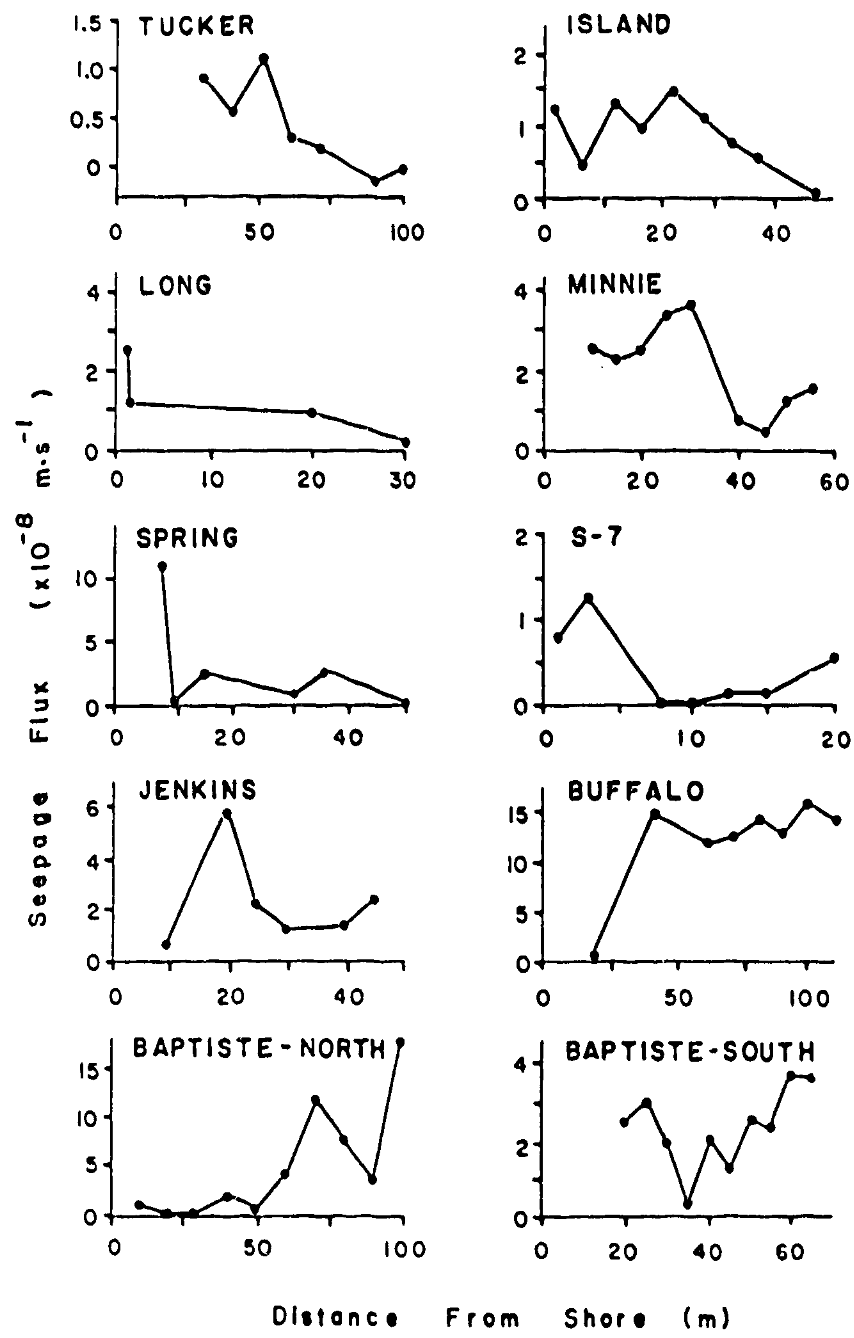

Figure 4.5. Nearshore seepage patterns in the study lakes.

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5. AN INTEGRATED APPROACH TO QUANTIFY GROUNDWATER TRANSPORT OF PHOSPHORUS TO NARROW LAKE, ALBERTA[1]

5.1 ABSTRACT

An integrated approach was used to quantify groundwater P flux to Narrow Lake, a small glacial-terrain lake in central Alberta (lake surface area 1.1 km^2; mean depth 14 m). Data from a drilling program, major ions, environmental isotopes, and computer simulations indicated that the lake gains water through the nearshore region from a small, shallow groundwater flow system; at deep offshore regions, water moves from the lake to the groundwater flow system. Seepage flux was quantified by four methods: (1) water budget (2) Darcy's equation with data from wells near the lake, (3) Darcy's Equation with data from mini-piezometers in the lake, and (4) seepage meters. Whole-lake seepage flux determined from mini-piezometer data (33 $mm.yr^{-1}$) was only 10 to 25 % of the other estimates (mean 246 $mm.yr^{-1}$; range 133 to 332 $mm.yr^{-1}$, from seepage meter and water budget data, respectively). Groundwater contributed about 30 % of the annual water load to the lake. The concentration of P ([P]) in porewater from lake sediments (mean 175 $mg.m^{-3}$) was eight times higher than in water from wells near the lake (mean 21 $mg.m^{-3}$). Thus, if well water was used to estimate [P] in seepage water, groundwater P loading rates would be underestimated. Based on porewater [P], the groundwater P loading rate to Narrow Lake was 43 $mg.m^{-2}.yr^{-1}$, and groundwater may be the largest single source of P to epilimnetic water in Narrow Lake.

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5.2 INTRODUCTION

Groundwater is an important source of phosphorus (P) to some lakes (Uttormark et al. 1974). However, transport of P from groundwater to lakewater is rarely measured, probably because of difficulties in methodology. In general, groundwater loading rates of P to lakes have been determined from two different approaches: (1) the residual of a lake P budget (e.g., Whitfield et al. 1982) or (2) the product of independent estimates of seepage flux into the lake times the concentration of P ([P]) in the seepage water (e.g., Brock et al. 1982). However, there are problems with both of these approaches. As for (1), phosphorus budgets are difficult to accurately quantify and the residual is subject to large errors. As for (2), there is no ideal method to measure groundwater inputs of water to lakes.

Even though there is no best method to measure groundwater inputs of water to lakes, a wide variety of methods are available. Groundwater-lake flux has estimated from the residual of water or ion balances (Meyboom 1967; Rinaldo-Lee and Anderson 1980; Crowe and Schwartz 1981), Darcy flux from wells (Karnauskas and Anderson 1978; Loeb and Goldman 1979; Rinaldo-Lee and Anderson 1980) or mini-piezometer data (Woessner and Sullivan 1984), seepage meters (Fellows and Brezonik 1980; Brock et al. 1982; Belanger and Mikutel 1985), tracers (Payne 1970; Zuber 1970; Lee et al. 1980), and groundwater flow models (Munter and Anderson 1981). Furthermore, [P] in seepage water have been estimated from groundwater sampled in wells near lakes (Gibbs 1979; Brock et al. 1982; Belanger and Mikutel 1985), porewater of lake sediments (Brock et al. 1982) and water collected by seepage meters (Lee

1977; Belanger and Connor 1980; Fellows and Brezonik 1981). However, there are few studies in which groundwater-lake flux of water and/or materials have been measured with more than one method.

In this study, I used an integrated approach to evaluate groundwater transport of P to Narrow Lake, a glacial-terrain lake in central Alberta. Qualitative aspects of groundwater flow to the lake were examined with hydrogeological, major ion and stable isotope data, and groundwater flow models. Groundwater-lake flux of water was quantified with four methods: water balance, Darcy's equation with data collected from wells near the lake, Darcy's equation with data collected from mini-piezometers in the lake, and seepage meters. Concentrations of P in the seepage water were measured in well water and lake sediment porewater. From these data, I calculated the rate of groundwater P loading to the lake and assessed the importance of groundwater relative to other sources of water and P to the lake.

Study Area

Narrow Lake (54°37'N, 114°37'W) is a small mesotrophic lake located in the mixed-wood section of the boreal forest of north-central Alberta (average euphotic total P 10.5 $mg.m^{-3}$, lake surface area 1.1 km^2, mean depth 14.4 m). The lake watershed is small, 7.0 km^2, and is completely forested except for two private camps on the west and south shore of the lake. The climate in the study area is continental; at Narrow Lake, average annual precipitation and lake evaporation are 503 and 636 mm, respectively (Hydrology Branch, Alberta Environment, unpublished data). Most of the annual precipitation (73 %) falls during the open-water season; 48 % of the annual precipitation falls during June, July and

August. From 1983 to 1986, the open-water season at Narrow Lake extended, on average, from 29 April to 11 November.

Geology

Narrow Lake is in the Interior Plains physiographic province (Bostock 1970). The Wapiti Formation, an Upper Cretaceous, non-marine, sandstone and siltstone unit underlies the lake (Green 1972). Melting ice masses carved steep-sided channels into the bedrock north, east and west of the lake (Fig. 5.1; Kjearsgaard 1972). The alluvium at the floor of the bedrock channels ranges in size from clay to gravel. Ground moraine covers much of the study area, and luvisolic soils are predominant (Strong and Leggat 1981). Small morainic ridges, which probably originated from material deposited from the melting ice as it stagnated, are associated with ground moraine east of Narrow Lake (Kjearsgaard 1972). Narrow Lake lies in a glacial meltwater channel bounded by large morainic ridges. The thickness of the surficial deposits ranges from only a few meters under Narrow Lake to more than 65 m under the morainic ridges.

5.3 MATERIALS AND METHODS

Drilling Program

Geology of the surficial deposits near Narrow Lake was investigated with 19 test-holes drilled during 1983 and 1984. Road access was limited to the southern half of the lake, so the drilling program was concentrated around the southern half of the lake. The test-holes ranged in depth from 3 to 60 m, and the deepest hole penetrated 30 m into bedrock. Particle size distribution was determined on till and sand collected from the test-holes.

To investigate groundwater flow conditions near Narrow Lake, 10 water-table wells and nine piezometers were installed near the lake (Fig. 5.2). At five piezometer nests, a water-table well and one or two piezometers were spaced 1-m apart to monitor vertical hydraulic gradients. Water-table wells and seven piezometers were constructed from 5-cm diameter PVC pipe and ranged in depth from 3 to 35 m. Two piezometers were completed in till (N1, N4), one in sand (N2) and four in sand and gravel (N1, N3, N4, N5). At two sites, 13-cm diameter steel-cased wells were also completed in sand and gravel (N1, N4). Water levels (i.e., hydraulic head, h) in the 5-cm diameter wells were monitored from July 1983 to October 1986. Each year, water levels were measured daily to biweekly from the beginning of May to the end of August. During September 1984 to April 1985, wells were sampled biweekly to monthly. A water-table map was prepared from water levels measured in the wells, topographic maps and aerial photos.

Hydraulic conductivity (K) of the surficial deposits was measured at all piezometer sites with slug tests (Hvorslev 1951), and at N1 and N4 with pumping tests (Cooper and Jacob 1946). Hydraulic conductivity of the bedrock was previously measured at a site 3 km south of Narrow Lake (Groundwater Resources Information Services (G.R.I.S.) files, Alberta Environment, Edmonton, Alberta).

Ion and Isotope Study

Water was sampled four times from August to October 1983 from the wells, weekly to biweekly from May to September 1983 from the lake, and on 20 May 1983 from a temporary surface inflow to the lake. The water samples were analyzed for pH, conductivity and major ions. Water chemistry of groundwater in the bedrock was obtained from a site 3 km south of Narrow Lake from records in the G.R.I.S. files.

Conductivity and pH were measured with a YSI Model 31 Conductivity meter and a Fisher Accumet digital pH meter, respectively. Water samples were filtered through pre-washed Whatman GF/C filters and stored at 4 ^{o}C until they were analyzed. Sulphate and Cl were determined by the turbidimetric and mercuric-nitrate methods, respectively (American Public Health Association 1975). Na, K, Ca and Mg were analyzed by flame emission on an atomic absorption spectrophotometer (American Public Health Association 1975).

In addition to major ions, stable isotopes were measured to examine the groundwater flow pattern near Narrow Lake. Water from four wells, the lake (depths: 2, 4, 6, and 15 m below surface), and bottom sediments were sampled on 7 June 1985 and analyzed for deuterium (D) and ^{18}O. Lake sediments were collected with a 4-barrel corer at a lake depth of 15 m (Fig. 5.2). The sediment core was sectioned at 10-cm intervals; Porewater was extracted from the top 0 to 10, 10 to 20, and 20 to 30 cm of the cores by centrifuging the mud. ^{18}O was analyzed by Epstein and Mayeda's (1953) CO_2 equilibration method on a Micromass 602D mass spectrometer (as outlined by Freeman 1986). D was analyzed on a Micromass 602C mass spectrometer (Friedman, 1953). Isotope ratios are reported in delta units (δ) as per mille differences relative to Standard Mean Ocean Water (SMOW).

Groundwater Flow Model

The goal of the modeling study was to determine whether the flow pattern predicted from the major ion and isotope study was realistic based on piezometer data obtained from the drilling program. The hydraulic head distribution in groundwater near Narrow Lake was simulated by a finite-element model developed by Schwartz and Crowe (1980). The model solves a two-dimensional, steady-state, anisotropic and heterogeneous equation of flow to calculate hydraulic head distribution. The computer code was modified so that the linear triangular elements could be arranged in an irregular array of rows and columns. The irregular spacing allowed nodes to be concentrated around the lake so that flow patterns near the lake could be better delineated.

Preliminary simulations indicated that regional groundwater divides occur at a topographic high that marks the western boundary of the Narrow Lake watershed and a topographic low 7 km east of Narrow Lake (Muskeg Creek). Therefore, flow conditions were simulated along the transect A-A' (Fig. 5.1). At the base of the Wapiti Formation, it wasassumed that the underlying Lea Park Formation (Upper Cretaceous gray shales) created a vertical no-flow boundary. Constant head nodes at the water table defined the upper boundary of the model.

Four different hydrogeological units were included in the cross-section: drift, bedrock, intertill sand and gravel, and lake sediments. Values of hydraulic conductivity were assigned to each unit based on values measured in the drilling program and from records in the G.R.I.S. files. In the study area, drift is a heterogeneous mixture of glacial till and interbedded sand and gravel lenses. However, on a regional scale, groundwater flow through drift can be approximated by treating drift as a homogeneous, anisotropic unit. Anisotropic ratios of drift are highly variable; values as high as 1000 have been reported (Bennet and Giusti 1971). Therefore, the cross-section was simulated with ratios ranging from 1 to 1000. In contrast to drift, anisotropy of clastic sedimentary rocks is low; e.g., values averaged 2 for sandstone and siltstone (Davis 1969). Anisotropy of sand and gravel also tends to be low: e.g., mean 6 (Davis 1969), 10 (Lee et al. 1980), 3 and 15 (Barwell and Lee 1981); a value of 10 was assumed for the intertill sand and gravels near Narrow Lake. Lake sediments were simulated as isotropic.

The position and extent of the sand and gravel layers are the most poorly defined of the four units. Initially, the cross-section was simulated with sand and gravel lenses in contact with both sides of the lake. The importance of sand and gravel to seepage conditions at Narrow lake. The importance of sand and gravel to seepage conditions at Narrow Lake was evaluated by repeating the simulation with sand and gravel layers at different positions near the lake. The model was calibrated against (1) water levels measured in piezometers near the lake, and (2) the groundwater flow pattern predicted from the major ion and isotope study.

Water Balance

Groundwater seepage to Narrow Lake was estimated from the residual of the water budget equation:

$$\text{residual} = Pr + SI - E - SO - \Delta V \qquad (1)$$

where Pr is the amount of precipitation falling directly on the lake

($m.yr^{-1}$), E is evaporation from the lake surface ($m.yr^{-1}$), SI and SO are surface inflow and outflow, respectively (in $m.yr^{-1}$; i.e., volume of discharge to or from the lake, m^3, divided by the lake surface area, m^2), and ΔV is the change in lake stage from the beginning to the end of a water-year ($m.yr^{-1}$). Water budgets were calculated on an annual basis for a water year from 1 June to 31 May, for 1983-1984 to 1986-1987.

In addition to net groundwater flux, the residual includes the amount of water from ungauged sources (e.g., diffuse runoff) and the compounded errors associated with measuring the other components of the water budget. In some cases, the errors can comprise a considerable portion of the residual; consequently, the residual may be a poor estimate of groundwater flux (Winter 1981). Therefore, it can be instructive to assess the potential magnitude of the error as a proportion of the residual. Assume that (1) the variance of the measured components is independent, and (2) there is some compensation for measured values that are too high or too low relative to the actual values of the components. Total variance of the measured components of the water budget (s^2_T) can then be estimated from:

$$s^2_T = s^2_{PR} + s^2_{SI} + s^2_E + s^2_{SO} + s^2_{\Delta V} \qquad (2)$$

where s^2_i is the variance of the measured component i. s_T is an estimate of the total standard deviation (SD) of the measured components of the water budget. Variance associated with each component of the water budget was not measured directly; instead, variance was estimated from studies of error analysis of hydrological components (e.g., Winter

1981).

Precipitation- Precipitation was measured with a Type B rain gauge (diameter 11.4 cm; height of orifice above ground 40 cm; no wind shield) at the Atmospheric Environment Services (A.E.S., Environment Canada), Athabasca 2 meteorological station, 20 km north-east of Narrow Lake. Variance of 25 % may be associated with annual precipitation because precipitation was not measured at the lake (Winter 1981). An additional 25 % variance may be caused by the type of gauge used to measure precipitation (Neff 1977). Thus, total variance (s^2_{PR}) is 50 % of the annual precipitation rate.

Lake Evaporation- Lake evaporation was estimated with a computer simulation model (WEVAP) that is based on a complementary relationship between areal and potential evapotranspiration (Morton et al. 1980). The model utilizes monthly air temperature, dew point temperature, and sunshine duration records to calculate annual rates of lake evaporation. Variance of 25 % may be associated with evaporation because input data for the model were not measured at the lake; instead, data were collected at three meteorological stations, 150 km north-west, 120 km south and 200 km east of Narrow Lake. In addition, variance of 100 % may be associated with even the "best" method of measuring lake evaporation (Winter 1981). Thus, s^2_E is 125 % of the annual evaporation rate.

Surface Runoff to the Lake- Drainage patterns to many north-temperate lakes in western Canada are poorly defined; drainage to Narrow Lake is no exception. Visual inspection of the watershed revealed that, for most of the study period, only a small portion of the drainage basin contributed runoff to Narrow Lake (via diffuse or overland flow). Surface runoff over most of the watershed drains to shallow depressions that are filled by peat or water. There are no distinct channelized streams from the depressions to the lake. Instead, these potential areas of surface flow into the lake are blocked by extensive networks of beaver dams.

Twice during the study, dams were removed from one of the beaver ponds to prevent flooding of a nearby road (12 May 1985 and 15 July 1986). There was considerable flow of water from the ponds to the lake for a few days after the dams were removed; flow from the impounded area stopped completely by 21 May 1985 and 2 August 1986, respectively. From 12 to 21 May 1985 and from 15 to 20 July 1986, there was no surface discharge of water from the lake. For these periods, the amount of runoff into the lake was estimated from short-term water balance: i.e., the change in lake stage corrected for precipitation. Lake evaporation was not estimated for short-time periods because WEVAP does not take into account the effects of seasonal changes in subsurface heat storage in lake water. However, some of the errors caused by the lack of evaporation measurement would be compensated by groundwater seepage into the lake. Lake stage was measured with a continuous recording gauge, so variance associated with these estimates is small. Thus, variance associated with values of surface runoff determined from the short-term water balance would be largely due to precipitation measurement. For short-term data, variance associated with precipitation measurements are much higher then for annual data (type of gauge 25 %, precipitation measured 20 km from lake 625 %, total variance 650 %, Winter 1981).

During the remainder of the study period, the only source of surface inflow to the lake was diffuse runoff. Discharge rates from diffuse flow are difficult to quantify (Ayers 1970); I could not directly measure this source of water. However, if diffuse runoff was an important source of water to the lake, one would expect that following heavy rainfalls the increase in lake level would be greater than the amount of rainfall; i.e., the slope of the regression of change in lake level (mm) versus rainfall (mm) would be greater than 1. To assess the magnitude of diffuse runoff I examined the relationship between lake stage and rainfall when lake stage increased $\geq$ 5 $mm.d^{-1}$ or when daily rainfall was $\geq$ 5 $mm.d^{-1}$ (excluding those periods of surface runoff to the lake from the beaver ponds or surface discharge from the lake).

Surface Discharge from the Lake- The single stream that flows from the lake meanders through a large marshy area at the south end of the lake and then through a culvert, under a road that separates Narrow and Long lakes (Fig. 5.1). For most of the study period, a beaver dam across the culvert prevented surface flow from Narrow Lake; however, on three occasions the dam was removed (23 August 1983, 3 May 1985, 20 July 1986). Following removal of the dam in 1983, discharge of water from the lake was measured at a weir constructed across the culvert; variance associated with these measurements are about 100 % of total water discharge (Winter 1981). On the two other occasions that the dam was removed, water flowed out of the lake from 3 to 12 May 1985 and 20 July to 10 August 1986; during these time periods, discharge of water from the lake was not directly measured. Instead, discharge was estimated by a short-term water balance, as outlined above for surface runoff to the lake. The estimate of discharge from 20 July to 10 August 1986 represents net surface discharge from the lake because there was also a small amount of surface runoff into the lake from 20 July to 2 August 1986.

Darcy Flux From Well Data

Visual observation from SCUBA diving and cores collected from the lake bottom indicate that gyttja is found at lake depths greater than about 8 m. Presumably, gyttja is relatively impermeable to groundwater flow; thus, seepage flux at lake depths greater than 8 m would be very low. In addition, seepage into lakes tends to decrease as a functions of distance from shore (McBride and Pfannkuch 1975). Therefore, I focussed on estimating rates of seepage in the nearshore region of Narrow Lake; i.e. at lake depths of 8 m or less.

Nearshore seepage to Narrow Lake was estimated from Darcy's Equation:

$$q = K\ I \qquad (3)$$

where q ($m.s^{-1}$) is seepage flux, K ($m.s^{-1}$) is hydraulic conductivity of porous media and I is hydraulic gradient ($\Delta h/\Delta l$), Δh is the change in hydraulic head between two points, Δl-m apart, along a groundwater flow path. I assumed that shallow groundwater flows laterally towards Narrow Lake. Therefore, within the shallow drift, hydraulic head (h) would be constant in the vertical direction, and the elevation of the water table would be representative of h in shallow groundwater (Fig. 5.4). Average values of K and h were determined from data obtained from the drilling program. Nearshore seepage flux to the lake was calculated from the average flux measured for lake depths of 0, 4 and 8 m (Fig. 5.3). Annual whole-lake seepage flux was calculated by correcting the average nearshore seepage flux for time (i.e., s to yr) and the proportion of nearshore sediments (i.e., < 8 m lake depth) to total lake surface area (29 %). This estimate of whole-lake seepage flux was compared to a value calculated by Darcy's Equation, based on data obtained at piezometer nest N1.

The error in the estimate of whole-lake seepage flux is potentially large, but difficult to quantify. For example, error in K alone may approach 100 % of the measured value (Winter 1981), and in most porous media K follows a log-normal distribution (Freeze 1975). In addition, hydraulic gradients were estimated from a water-table map developed with relatively few control points; I could not assess the magnitude of this source of error. 95 % confidence limits for this estimate of seepage were assumed to be proportional to the 95 % confidence limits for K in till (as measured in the drilling program).

Darcy Flux From Mini-Piezometer Data

During August 1985, six mini-piezometers were installed by SCUBA divers in Narrow Lake (Fig. 5.2). The sites ranged in lake depths from 1 to 15 m. Mini-piezometers were constructed from 1.27-cm I.D. PVC pipe, and hydraulic head was measured with a manometer (Fig. 5.4). Under ideal conditions (i.e., no waves near manometer) readings of h were accurate to about 1 mm; under wavy condition, h was accurate to about 10 mm. During installation, water was pumped through the hose into the mini-piezometer to prevent clogging of the tip. Hydraulic conductivity (K) was measured by falling head tests (Lee and Cherry, 1978). Seepage flux was calculated by Darcy's Equation (Eq. 3).

Seepage Meters

During August 1984, seepage meters were sampled in Narrow Lake to examine diurnal fluctuations in nearshore seepage flux and to evaluate the variability of seepage flux within the lake. Three to five seepage meters were placed at each of four sites in the lake (Fig. 5.2; SM1 to SM4). At each site, seepage meters were 1-m apart, at lake depths of 0.5 to 1 m (SM1 to SM3) and 2 m (SM4). In addition, eight seepage meters were sampled at SM1 during May 1988. In 1986, I measured spatial and temporal variability of nearshore seepage patterns at two sites within Narrow Lake (Chapter 4). At each site, ten seepage meters were placed along a transect, perpendicular to shore (Fig. 5.2: SM5 and SM6); the transects extended 40 m from shore to lake depths of 7 and 10 m, respectively. I assumed that the average seepage flux measured at SM5 and SM6 were representative of the flux at other nearshore regions in the lake. The whole-lake seepage flux was calculated by correcting the average nearshore seepage flux for the proportion of nearshore sediments to total lake surface area (29 %). Seepage meters were constructed from "45-gallon" drums (Lee, 1977), and seepage flux was corrected for the short-term anomalous influx of water that enters the plastic bags after they are attached to seepage meters (Chapter 2).

Phosphorus Concentrations

Phosphorus concentrations ([P]) in groundwater were determined from well water and lake sediment porewater. From 1983 and 1984, wells around Narrow Lake were sampled a total of three to four times each and analyzed for total dissolved P (TDP). Samples for TDP analyses were filtered through a prewashed 0.45-μm HAWP Millipore membrane filter. The filtered samples were digested and analyzed by Menzel and Corwin's (1965) potassium persulfate method.

Sediment porewater was collected in situ with Plexiglass peepers fitted with Gelman HT-450 membrane filters (Shaw and Prepas 1989a). Porewater from the top 0 to 10 cm in the sediment was analyzed for soluble reactive P (SRP) (Murphy and Riley 1962). In porewater of Narrow Lake, SRP and TDP concentrations were similar (Shaw and Prepas 1989); therefore porewater [SRP] and well water [TDP] were likely directly comparable. For the remainder of the paper, the term [P] is used for porewater [SRP] and well water [TDP].

During 1985 and 1986, spatial and temporal variability of porewater [P] in Narrow Lake was examined (Shaw and Prepas 1989). Additional information on spatial and temporal variability of porewater [P] was obtained from sporadic sampling of nine sites in 1985 (lake depths 3 and 10 m) and three sites in 1986 (lake depths 1 and 3m).

5.4 RESULTS AND DISCUSSION

Drilling Program

The morainic ridges next to Narrow Lake are primarily sandy-clay till. The particle size distribution of the till is similar to that at other locations on the Wapiti Formation (sand:silt:clay was 45:24:31 %, respectively; Pawluk and Bayrock 1969). Interbedded layers of sand and gravel were commonly encountered during the drilling program. The sand and gravel layers ranged from 0.5- to 10-m thick, though they were usually less than 2-m thick. In general, the sand and gravel layers were of limited lateral extent; e.g., at N1, there was poor continuity of the layers between holes only 25 m apart (Fig. 5.5).

Water-table elevations near Narrow Lake were a subdued reflection of the ground surface (Fig. 5.2). At surface elevations less than 700 m, the water table was 0.5 to 2 m below the ground; at higher elevations, the water table was 1.5 to 8 m below the ground. During the study period, water-table elevations fluctuated 1 to 4 m. The water table increased in response to spring snow-melt and heavy summer rainfalls. Changes in the level of Narrow Lake corresponded to changes in the water table and hydraulic head measured at the piezometers around the lake. At all wells, the water-table elevation was higher than the lake level; average hydraulic gradients from the water-table to the lake were about 0.03. Hydraulic head decreased with depth at the piezometer nests on the ridges; vertical gradients ranged from 0.03 to 0.1. In contrast, at the one piezometer nest between the ridges (N4), vertical gradients were upward (mean 0.005).

Hydraulic conductivity of the surficial deposits were highly variable: K of sandy-clay till ranged from $5x10^{-7}$ and $2x10^{-6}$ $m.s^{-1}$ (geometric mean $1x10^{-6}$ $m.s^{-1}$) and K of sand and gravel ranged from $2x10^{-6}$ to $4x10^{-5}$ $m.s^{-1}$ (geometric mean $1x10^{-5}$ $m.s^{-1}$). The hydraulic conductivity of the bedrock near Narrow Lake ($1x10^{-6}$ $m.s^{-1}$) was within the range measured for till.

Major Ion and Isotope Study

Chemical analysis of major ions, pH and conductivity from surface water and groundwater is summarized in Table 5.1. Conductivity and major ion concentrations of the surface inflow and lake were much lower than in groundwater; pH of all samples was slightly alkaline. There was little difference between the major ion composition of Narrow Lake, surface runoff and drift (recharge sites). Ca and Mg were the major cations and HCO_3 was the major anion. The dominant cation in groundwater from the discharge site (piezometer nest N4) was Mg rather than Ca. The change from dominance by Ca to Mg probably indicates increased groundwater residence time and increased contact with clay minerals. Water from bedrock was more mineralized than shallow groundwater; Na was the major cation, and HCO_3 and SO_4 were the major anions. The dissimilarity of water chemistry from the lake and drift, versus bedrock suggests that Narrow Lake lies within a shallow, local flow system. The similarity between water from the lake and drift suggest that, like other prairie lakes (Schwartz and Gallup 1978), the major ion chemistry may be controlled by shallow groundwater seepage to the lake from glacial drift. The low ionic concentrations in Narrow Lake relative to groundwater could be caused by dilution from (1) precipitation falling directly on the lake surface, and/or (2) ephemeral surface inflows.

Values of $\delta^{18}O$ and δD measured from wells near Narrow Lake ranged from -17.7 to -18.4 (mean -18.1) $^{o}/_{oo}$ and -138.8 to -145.6 (mean -142.8) $^{o}/_{oo}$, respectively). These values were close to the average meteoric water values listed for Edmonton, 130 km south of Narrow Lake (Hage et al. 1975: $\delta^{18}O$ -17.9 and δD -137.3 $^{o}/_{oo}$). However, lakewater was highly enriched with respect to the average meteoric water values; $\delta^{18}O$: -10.1 to -10.4 (mean 10.2) $^{o}/_{oo}$, and δD: -101.4 to -105.2 (mean -103.4) $^{o}/_{oo}$. Enriched lakewater is caused by evaporation; the lighter ^{16}O and ^{1}H

atoms are selectively removed so the water becomes heavier (i.e., δ values become less negative). Like lakewater, porewater was also enriched with respect to $\delta^{18}O$ and δD: ^{18}O -10.4 to -10.7 (mean -10.5) $^{o}/_{oo}$, and δD -105.4 to -107.7 (mean -106.5) $^{o}/_{oo}$. The similarity between values of $\delta^{18}O$ and δD from porewater and lakewater suggests that there was seepage out of the lake at the site where porewater was collected (lake depth 15 m; Fig. 5.2).

Groundwater Flow Model

The model study indicated that the pattern of shallow groundwater discharge to the lake and deep recharge of groundwater from the lake is consistent with the hydraulic head measured in groundwater near the lake. To calibrate the model to the observed head distribution, a value of 600 was required for anisotropy of till. Under that condition, there would be discharge of groundwater to the lake throughout most of the lake bottom from a small, shallow groundwater flow system (Fig. 5.6).; through 16 % of the lakebed, there would be recharge to groundwater from the lake (Fig. 5.7A). There is no guarantee that the calibrated data set provides a unique hydraulic head distribution. For instance, different hydraulic conductivities (and different anisotropy) could produce similar flow patterns (Gillham and Farvolden 1974). Nonetheless, the results from the model study indicate that there is certainly a potential for shallow discharge and deep recharge of groundwater at Narrow Lake.

Intertill sand and gravels near the lake strongly affected seepage patterns in the lake. Without the sand and gravel layers, seepage patterns at the lake would be symmetrical on both sides of the lake (i.e., hydraulic head distribution is symmetrical), and seepage would decrease with distance from shore (Fig. 5.7B). However, high zones of offshore seepage flux in Narrow Lake could be caused by intertill lenses near the lake, even if the lenses do not extend to the lakebed (Fig. 5.7C).

Water Balance

Precipitation and evaporation were the major input and output, respectively, at Narrow Lake (Table 5.2). The residual was, on average, 37 % of the total input of water to the lake (Pr + SI + Residual; Eq. 1). The total standard deviation of the measured components of the water budget was, on average, 30 % of the residual. Thus, there was likely an unmeasured net flux of water into the lake. The unmeasured components included diffuse runoff from the drainage basin and groundwater. However, the change in lake level following heavy rainfalls increased directly proportionally to the amount of rain; i.e., the slope of the regression of change in lake level on rainfall was not significantly different from 1 ($\underline{P}<0.01$). This supports my visual observations that rates of diffuse runoff into the lake were low. Thus, the residual would be comprised largely of groundwater discharge to the lake. The water balance study indicates an average annual net flux of 332 $mm.yr^{-1}$ of groundwater to the lake; the lower and upper 95 % confidence limits (CI) around this estimate are 228 and 436 $mm.yr^{-1}$, respectively.

Darcy Flux From Well Data

The water table slopes downward, from the ridges along Narrow Lake towards the lake; thus, shallow groundwater flows towards the lake (Fig. 5.2). At an elevation of 690 m, the water table is on average, 90 m from the lake shore. Therefore, given the lake elevation of 687 m, or a difference of 3 m head over 90 m distance from the lake, the hydraulic gradient (I) to the lake shore is 0.033 (Fig. 5.3). At 8 m lake depth, the distance to the lake is 140 m, so I is 0.021, and the average gradient to the nearshore region of the lake (i.e., 0 to 8 m lake depth) is 0.027. Average K in till was 1×10^{-6}, so average nearshore seepage flux (Eq. 3) to Narrow Lake would be 2.7×10^{-8} $m.s^{-1}$ (CI 8×10^{-9} and 8×10^{-8} $m.s^{-1}$). Expressed over the entire surface area of the lake, whole-lake seepage flux was 272 $mm.yr^{-1}$ (CI 81 and 807 $mm.yr^{-1}$, respectively). This value of whole-lake seepage flux is slightly lower than that estimated from the water balance.

To calculate the average seepage flux, I assumed that the elevation of the water table was representative of h in shallow drift. However, water levels measured at the piezometer nests showed that h varied with depth (Fig. 5.5). Even so, the whole-lake seepage flux calculated from actual data collected at piezometer nest N1 (237 $mm.yr^{-1}$) was only 9 % higher than that calculated with h estimated from the water-table elevation (272 $mm.yr^{-1}$).

Darcy Flux from Mini-Piezometers

Hydraulic heads measured by mini-piezometers in Narrow Lake tended to be low; they ranged from -2 to 70 mm, relative to lake level (Table 5.3). The negative head was recorded at the deepest site (Fig. 5.2: MP5). This supports the stable isotope study that indicates recharge to groundwater may occur at deep areas of the lake. The high values of h measured at MP6 could not be replicated and were probably due to problems with the instrument. Similar problems with false readings have been reported and were attributed to plugging of the well point by bottom sediments (Cartwright et al., 1979). Discounting MP6, vertical hydraulic gradients in the sediments of Narrow Lake ranged from -0.001 to 0.004. The hydraulic gradients were of similar magnitude to that measured at the one piezometer nest at a site of groundwater discharge near the lake (i.e., N4: 0.005).

Hydraulic conductivity (K) in bottom sediments ranged from $3x10^{-6}$ to $2x10^{-7}$ $m.s^{-1}$ (Table 5.3). At one site (MP1), replicate falling head tests were conducted, and in both cases K was $2x10^{-6}$ $m.s^{-1}$. Hydraulic conductivity measured at two mini-piezometers (MP2 and MP3), only 1-m apart, varied 3-fold. The range of K of the nearshore sediments was similar to that measured in the drift near Narrow Lake. At the deepest site (MP5), K was higher than expected for gyttja. This high K may indicate that (1) the mini-piezometer was not completed in gyttja (i.e., the well point penetrated completely through the layer of gyttja), or (2) during installation, a permeable zone around the well tip was created by pumping water through the mini-piezometer.

Average seepage flux at the mini-piezometer sites ranged from $-6x10^{-10}$ to $8x10^{-9}$ $m.s^{-1}$ (Table 5.3). Flux was highest at relatively shallow lake depths. I calculated the average nearshore seepage flux (0- to 8-m lake depth) from the geometric mean of data collected by MP1 to MP4: $3x10^{-9}$ $m.s^{-1}$. At the deepest site (MP-5), there was seepage out of the lake. However, the flux was very low; if one assumed seepage from the lake through 16 % of the lake bed (as indicated from the modeling study), at the flux measured at MP5, only 3 $mm.yr^{-1}$ of water would be lost from the lake. Therefore, I assumed seepage at depths greater than 8 m was insignificant and expressed the nearshore seepage flux over the entire lake to get an estimate of whole-lake seepage flux to Narrow Lake: geometric mean 33 $mm.yr^{-1}$ (CI 11 and 101 $mm.yr^{-1}$). This value is much lower than those estimated by either the water balance or Darcy flux from well data.

Seepage Meters

At the four sites in Narrow Lake sampled during 9 to 13 August 1984, there were no consistent differences between seepage flux measured during the day or night (Fig. 5.8). However, there were significant differences between seepage flux at the four sites ($F=20.1$, $df=3,74$, $P<<0.001$); flux at SM3 were 3-fold higher than at SM1. Seepage flux measured with seepage meters less than 5 m from shore during 1986 (SM5, SM6) were within the range measured in 1984 (Table 5.4).

Relative to spatial differences in seepage flux, temporal variability tended to be low. Measured seepage fluxat SM1 in May 1988 was 50 % greater than that measured in August 1984 ($t=2.08$, $df=31$, $P<0.05$). During 1986, consistent trends in seepage flux along the transects in Narrow Lake were evident (Chapter 4). At SM5, seepage flux decreased significantly ($P<0.05$) with distance from shore on 4 of 6 sampling dates. On one other date (10 July 1986), seepage tended to decrease with distance from shore ($P<0.10$). Interestingly, the only time when seepage flux did not decrease with distance from shore was 18 July; that sample was probably impacted by heavy rainfall. At SM6, seepage flux did not decrease with distance from shore ($P>>0.10$). Instead, seepage flux was highest 20 to 25 m from shore. The offshore zones of high seepage flux may be due to the intertill sand and gravel lenses near the lake.

Average seepage flux along the transect at SM6 were 185 % of the rate at SM5 (t=5.3 df=5, P<0.01). Expressed over the entire surface area of the lake, nearshore seepage contributed 133 mm.yr^{-1} of water to Narrow Lake (CI 103 and 163 mm.yr^{-1}; Chapter 4). Whole-lake seepage flux estimated from seepage meters was 40 and 49 percent of the values estimated by water balance and well data, respectively; but it was more than 4-fold higher than the value estimated with mini-piezometer data.

Groundwater Phosphorus Concentrations

Phosphorus concentrations in groundwater near Narrow Lake ranged from 3.6 to 84.6 mg.m^{-3}. There were no trends in [P] with depth of groundwater, and no differences in [P] between sites of groundwater discharge or recharge. The average [P] measured in groundwater sampled from the wells was 20.6 mg.m^{-3} (CI 16.5 and 24.7 mg.m^{-3}).

At 15 m lake depth, porewater [P] tended to be higher than values in the nearshore zone (Shaw and Prepas 1989). However, within the nearshore region at Narrow Lake ($\leq$ 8 m lake depth), there were no consistent trends in porewater [P]. At 5 m depth, the variance of porewater [P] measured at one site was as great as the variance between 10 other sites within the lake (Shaw and Prepas 1989). Since there were no trends in porewater [P], an average concentration for the nearshore region was calculated from the average of all data collected at lake depths $\leq$ 8 m: mean 175 mg.m^{-3}, CI 134 and 216 mg.m^{-3}.

Groundwater P Loading Rates

There was a wide range in estimates of whole-lake seepage flux to Narrow Lake determined by the four different methods (Table 5.5). The value estimated with mini-piezometer data was much lower than the others. In addition, the mini-piezometer estimate was based on a smaller data set than any of the other estimates. Therefore, it was excluded from calculation of an overall average seepage flux. The overall average seepage flux, estimated from the average of the other three methods, was 246 $mm.yr^{-1}$ (CI 130 and 362 $mm.yr^{-1}$). At this flux, groundwater would contribute about 30 % of the total annual water load to Narrow Lake.

The difference between [P] in well water and porewater was almost as great as for seepage flux; porewater [P] was 9-fold higher than well water [P]. Thus, the amount of P transported by groundwater from the drainage basin to the lake is small relative to the amount that is transported by groundwater moving through P-rich lake sediments into the lake. High [P] in porewater is a function of biogeochemical processes within lake sediments (e.g., adsorption/desorption on or from particulate matter in sediments). Thus, groundwater could be considered as a mechanism for enhancing nutrient recycling from lake sediments to lakewater, rather than a new source of P to the lake. Regardless whether groundwater is considered as an external or internal source of P to lakes, porewater [P] should be used to estimate P loading rates from groundwater. Based on average annual seepage flux and average porewater [P], the average groundwater P input to Narrow Lake was 43 $mg.m^{-2}.yr^{-1}$ (CI 17 and 78 $mg.m^{-2}.yr^{-1}$).

Groundwater P loading rates were higher than other external sources of P to Narrow Lake. The annual rate of P loading by groundwater was more than twice that of average annual atmospheric deposition to the lake (Chapter 7; average atmospheric TP load 20.3 $mg.m^{-2}.yr^{-1}$). Concentrations of total P in impounded water near Narrow Lake ranged from 21 to 686 $mg.m^{-3}$ (mean 177 $mg.m^{-3}$; R.D. Shaw, unpublished data). Based on the rate of surface runoff to the lake, estimated in the water balance study (45 $mm.yr^{-1}$), surface loading of total P to the lake would only be about 8.0 $mg.m^{-2}.yr^{-1}$. Thus, the annual rate of P loading by groundwater was more than five times that of surface runoff.

5.5 CONCLUSIONS

An integrated approach was useful to investigate groundwater P loading to Narrow Lake. Data from the drilling program, major ion chemistry, environmental isotopes, and computer simulations indicated that groundwater recharges in the lake watershed and flows through the shallow drift before discharging into the nearshore regions of the lake. At deeper, offshore regions there may be seepage out of the lake to the groundwater system. Data collected from the four different methods used to quantify seepage flux were consistent with these observations. Even though groundwater contributed less than one-third of the total annual water load to Narrow Lake, groundwater may be the major single source of P to epilimnetic lakewater. Groundwater influx occurs near shore, so that P transported by groundwater enters the epilimnion, where it could be directly utilized by lake biota. Groundwater should not be overlooked when preparing P budgets for lakes.

Table 5.1. Mean and standard deviation (in brackets) of specific conductance (S.C.), pH and major ions measured during 1983 from Narrow Lake, a temporary surface inflow to the lake, groundwater in drift (at recharge and discharge sites) and groundwater from bedrock; n is the number of samples analyzed.

Type	n	S.C.	pH	Ca	Mg	Na+K	HCO3	SO4	Cl
		$uS.cm^{-1}$					$Eq.m^{-3}$		
Lake	29	295(38)	7.96(0.30)	1.74(0.68)	1.23(0.22)	0.41(0.10)	3.11(0.38)	0.02(0.01)	0.002(0.001)
Inflow	1	210	7.40	1.78	1.11	0.23	1.98	0.07	0.001
Recharge	27-43	780(286)	7.57(0.16)	4.94(1.17)	3.68(1.67)	0.71(0.72)	9.49(4.04)	0.63(0.60)	0.004(0.002)
Discharge	4-6	922(99)	7.93(0.12)	3.96(3.09)	7.15(0.32)	1.19(0.14)	11.06(0.43)	0.56(0.11)	0.003(0.001)
Bedrock	1	1393	8.20	1.20	0.58	12.9	11.5	3.64	0.08

Table 5.2. Annual water budget for Narrow Lake from 1 June to 31 May, 1983-1984 to 1986-1987, standard deviations are given in brackets. A negative value for the residual indicates the net input of water to the lake that was not accounted for by the other parameters in the water budget. All values are in mm.yr^{-1}.

Water Year	Precipitation	Surface Inflow	Evaporation	Surface Outflow	Change in Lake Stage	Residual
83-84	652 (46)	0	622 (70)	230 (23)	-16	-184 (87)
84-85	590 (42)	22 (6)	671 (75)	61 (16)	236	-356 (88)
85-86	386 (27)	0	657 (73)	0	82	-353 (78)
86-87	480 (34)	158 (40)	618 (69)	430 (110)	25	-435 (140)
Mean	527	45	642	180	82	-332 (98)
SD	118	76	26	193	110	106

Table 5.3. Lake depth, depth of well point into sediments, hydraulic conductivity (K), hydraulic head, and seepage flux from six mini-piezometers (MP) in Narrow Lake 22 to 28 August 1985. Negative values of seepage indicate flux of water from the lake to groundwater: '-' indicates data were not collected.

	MP1	MP2	MP3	MP4	MP5	MP6
Lake depth (m)	1	4	4	7	15	15
Well point (m)	0.5	1.0	1.1	2.5	5.6	3.0
K ($m.s^{-1}$)	$2x10^{-6}$	$3x10^{-6}$	$1x10^{-6}$	$2x10^{-7}$	$2x10^{-6}$	-
Hydraulic head (mm)						
August 22	2	2	2	-	-	-
August 26	2	2	3	5	-2	10
August 27	b.d.*	3	3	10	b.d.	70
August 28	1	3	3	10	-1	20
Seepage flux ($m.s^{-1}$)						
Mean	$7x10^{-9}$	$8x10^{-9}$	$3x10^{-9}$	$7x10^{-10}$	$6x10^{-10}$	-
SD	$2x10^{-9}$	$2x10^{-9}$	$5x10^{-10}$	$2x10^{-10}$	$3x10^{-10}$	-

* below detection limit ($\pm$ 1 mm).

Table 5.4. Geometric mean and 95 % confidence intervals for seepage flux measured by 3 to 10 (n) seepage meters at six sites (SM) in Narrow Lake; N is total number of samples obtained from each site. Values given for SM5 and SM6 are only from seepage meters less than 5 m from shore.

				Seepage flux ($x10^{-8}$ $m.s^{-3}$)		
Site	Year	n	N	Mean	Lower	Upper
SM1	1984	5	25	1.2	1.0	1.5
SM2	1984	5	26	1.5	1.3	1.8
SM3	1984	3	12	3.6	3.0	4.3
SM4	1984	3	15	2.5	2.0	3.1
SM5	1986	2	11	1.4	1.0	1.9
SM6	1986	2	11	1.7	1.4	2.0
SM1	1988	8	8	1.8	1.3	2.6

Table 5.5. Mean and 95 % confidence intervals for seepage flux and [P] at Narrow Lake, 1983 to 1986.

Method	Mean	Lower	Upper
Seepage Flux ($mm.yr^{-1}$)			
Residual	332	228	436
Wells	272	81	807
Mini-piezometers	33	11	101
Seepage Meters	133	103	163
[P] ($mg.m^{-3}$)			
Wells	21	17	25
Porewater	183	149	227

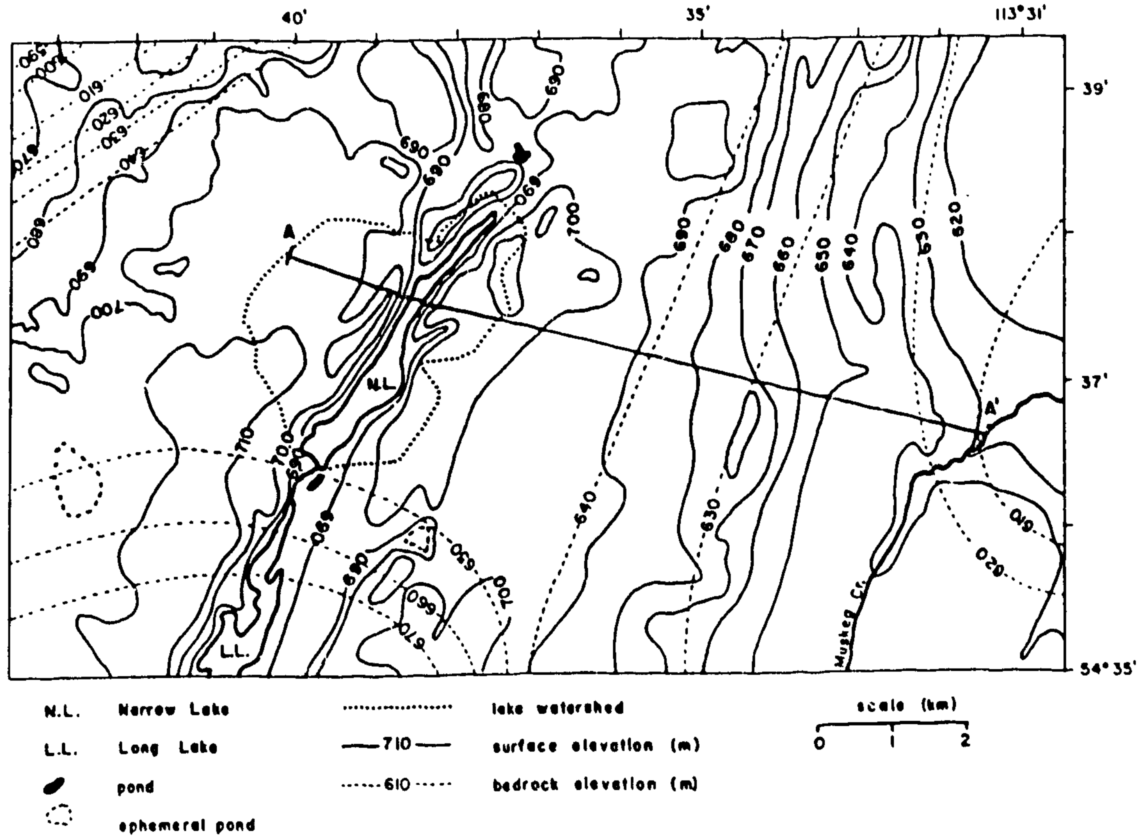

Figure 5.1. Surface and bedrock elevations near Narrow Lake. A-A' is the cross-section simulated with the groundwater flow model.

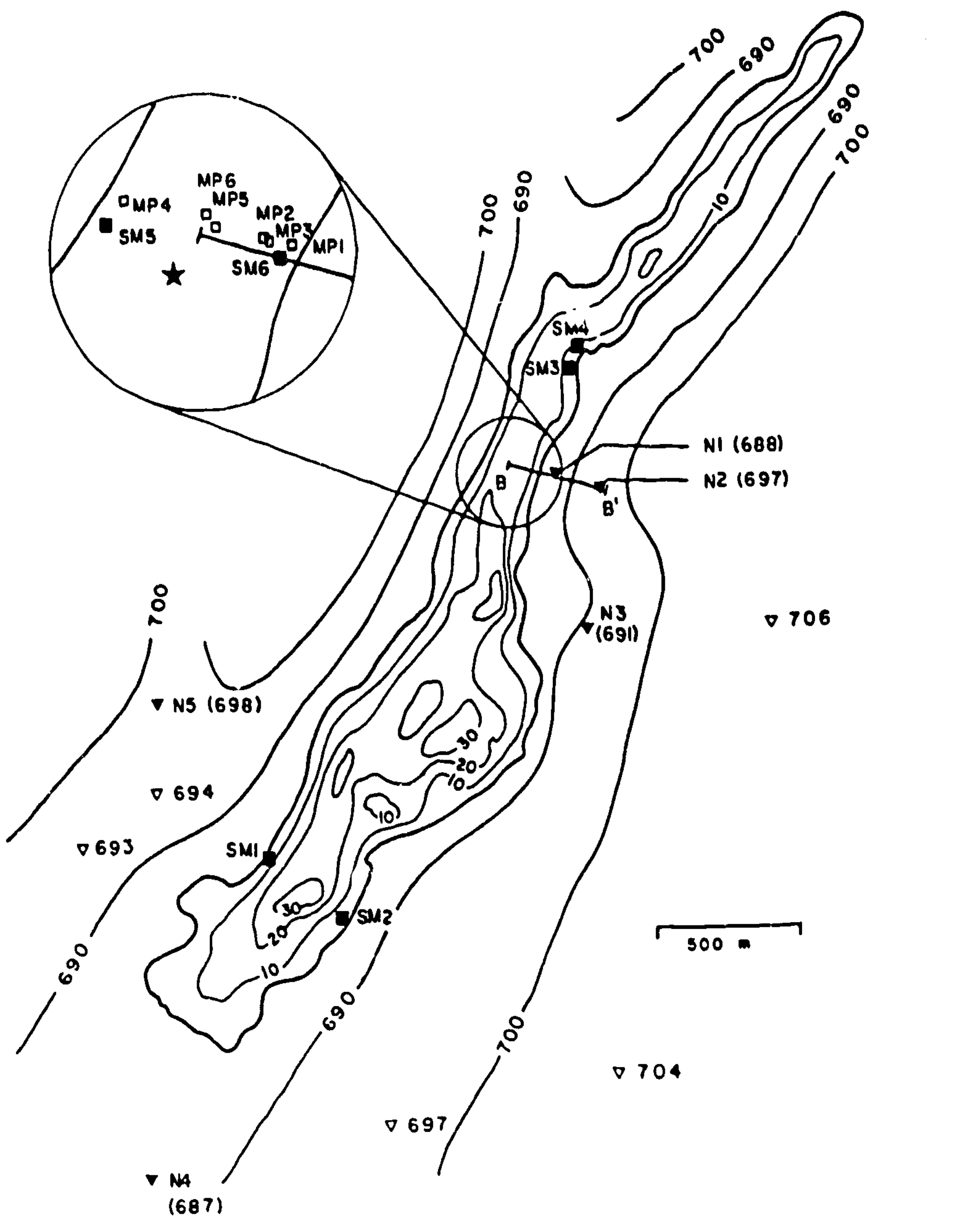

Figure 5.2. Location of instrumentation within and around Narrow Lake, and water-table elevations (in m) near Narrow Lake. ■ SM1 - seepage meter site; □ MP1 - mini-piezometer site; ★ porewater sampling site for D and ^{18}O; ▽704 - water-table well and average water-table elevation (m); ▼N1 (691) - piezometer nest and average water-table elevation (m). Water-table elevation and lake depth are shown at 10-m intervals.

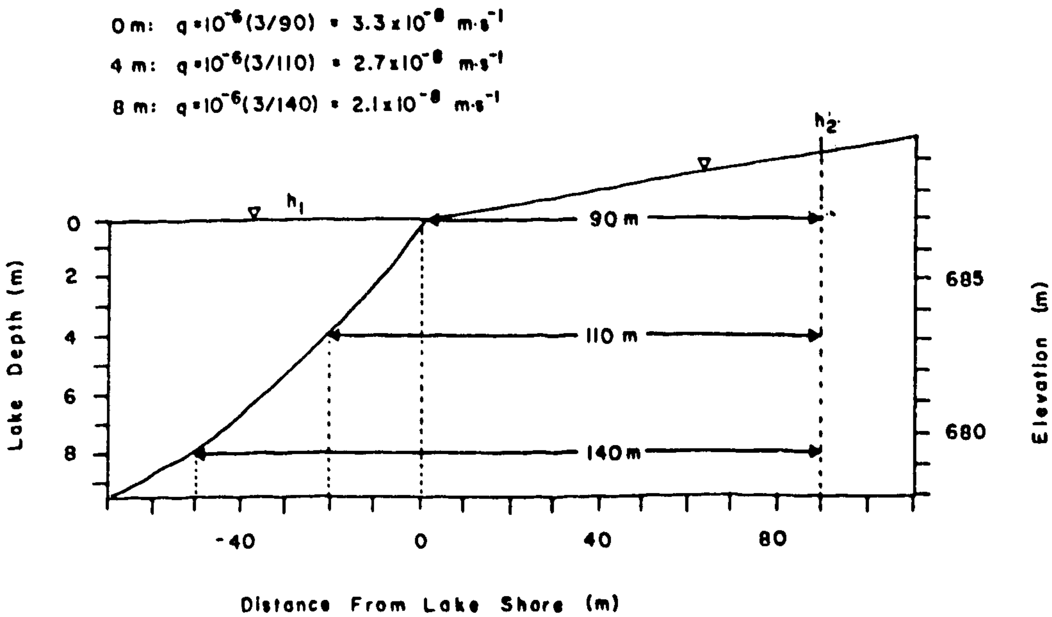

Figure 5.3. Example calculation of nearshore seepage flux (q, $m.s^{-1}$) from Darcy's equation $q=KI$. Hydraulic conductivity (K) is 10^{-6} $m.s^{-1}$; hydraulic gradient is determined from $\Delta h/\Delta l$, where Δh (3 m) is the change in elevation of the water-table between the lake (h_1) and a point (h_2), near the lake; Δl is the horizontal distance between the lakebed, at depths of 0-, 4-, and 8-m below the surface of the lake, and points underlying h_2.

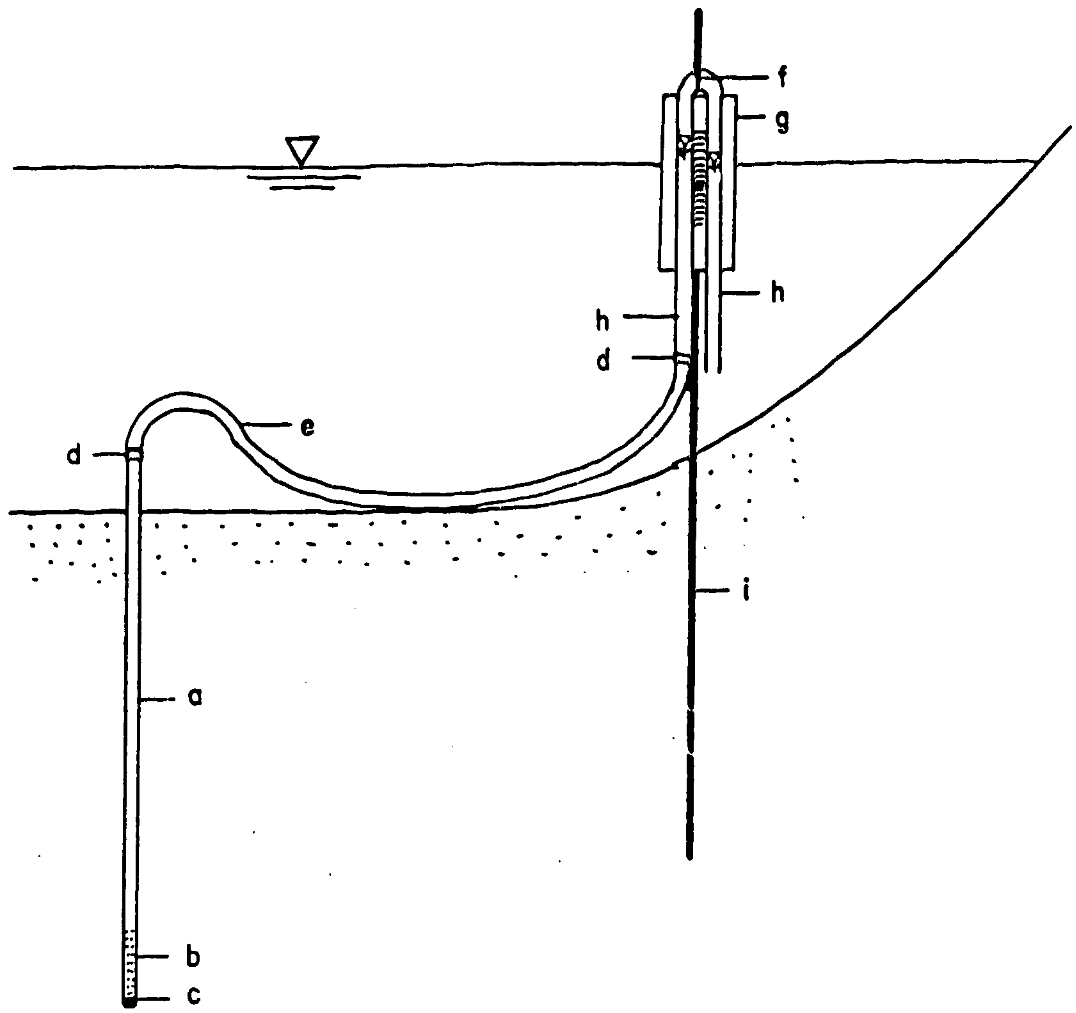

Figure 5.4. Mini-piezometers installed in lake bottom. a - 1.27-cm ID, 1.91-cm OD PVC pipe (2- to 6-m long), b - 50-cm long well point, c - rubber plug, d - 1.27-cm ID threaded connector, e - 1.27-cm ID hose, f - 1.27 cm ID connector, g - manometer board, marked in mm gradation, h - 1.27-cm ID clear polyethylene tubing, i - metal post.

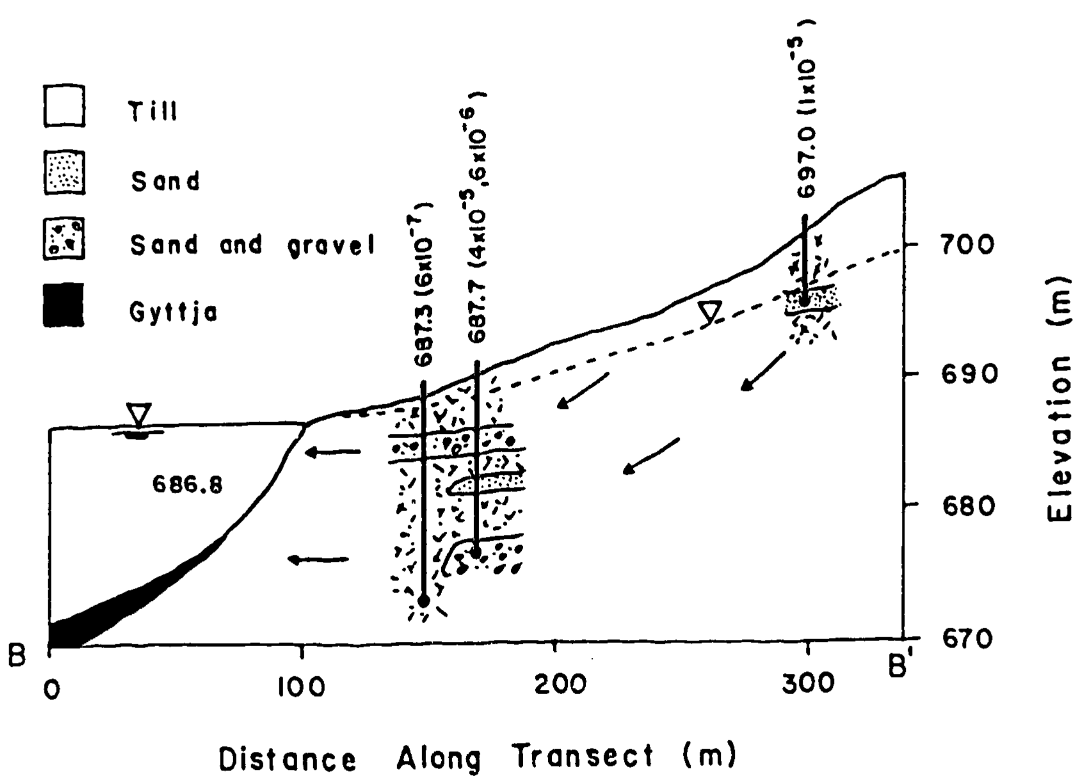

Figure 5.5. Lithology, hydraulic head (m), and hydraulic conductivity ($m.s^{-1}$, in brackets) along cross-section B-B' (Fig. 5.2). The two values of K for the shallow piezometer at site N were estimated from slug tests ($4x10^{-5}$ $m.s^{-1}$) and pumping tests ($6x10^{-6}$ $m.s^{-1}$). The arrows indicate the general direction of groundwater flow.

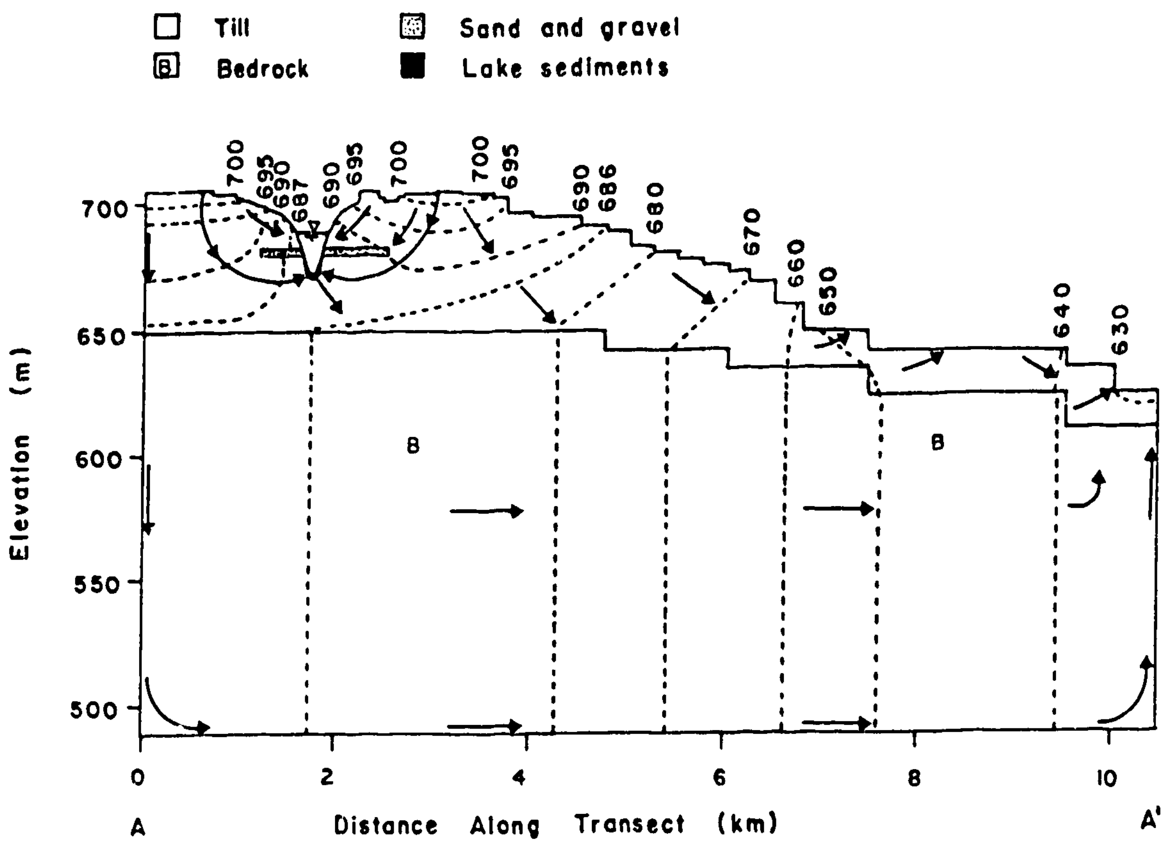

Figure 5.6. Regional flow conditions along the cross-section A-A' (Fig. 5.1). Lines of equal hydraulic head (m) are indicated by dashed lines. The arrows indicate the general direction of groundwater flow. Values of K_h ($m.s^{-1}$) and $K_h:K_v$ (in brackets) are 10^{-6} (600), 10^{-6} (2), $5x10^{-5}$ (10), and 10^{-8} (1) for till, bedrock, sand and gravel, and lake sediments, respectively.

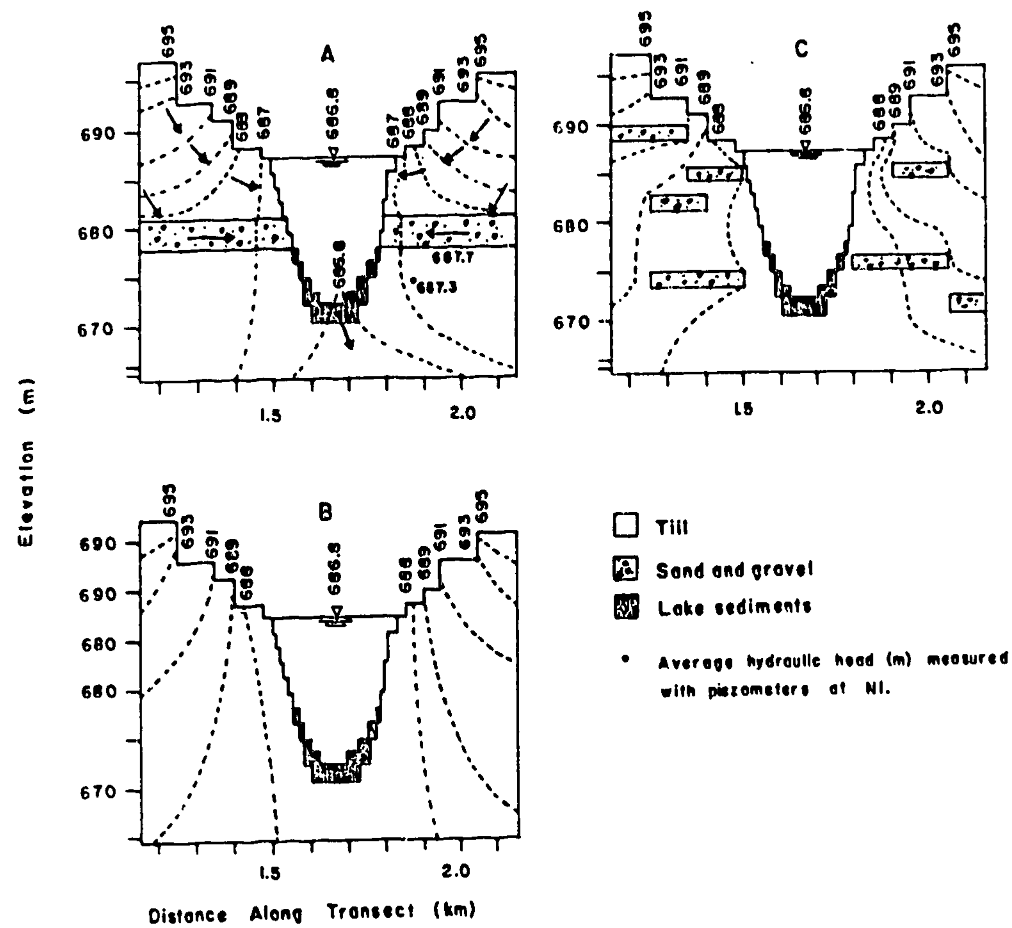

Figure 5.7. Local flow conditions near Narrow Lake. A - shallow discharge- deep recharge; B - no intertill sand and gravel lenses. C - sand and gravel layers scattered throughout till. Values of K_h ($m.s^{-3}$) and $K_h:K_v$ (in brackets) are 10^{-6} (600), $5x10^{-5}$ (10), and 10^{-8} (1) for till, sand and gravel, and lake sediments, respectively. Lines of equal hydraulic head (m) are indicated by dashed lines.

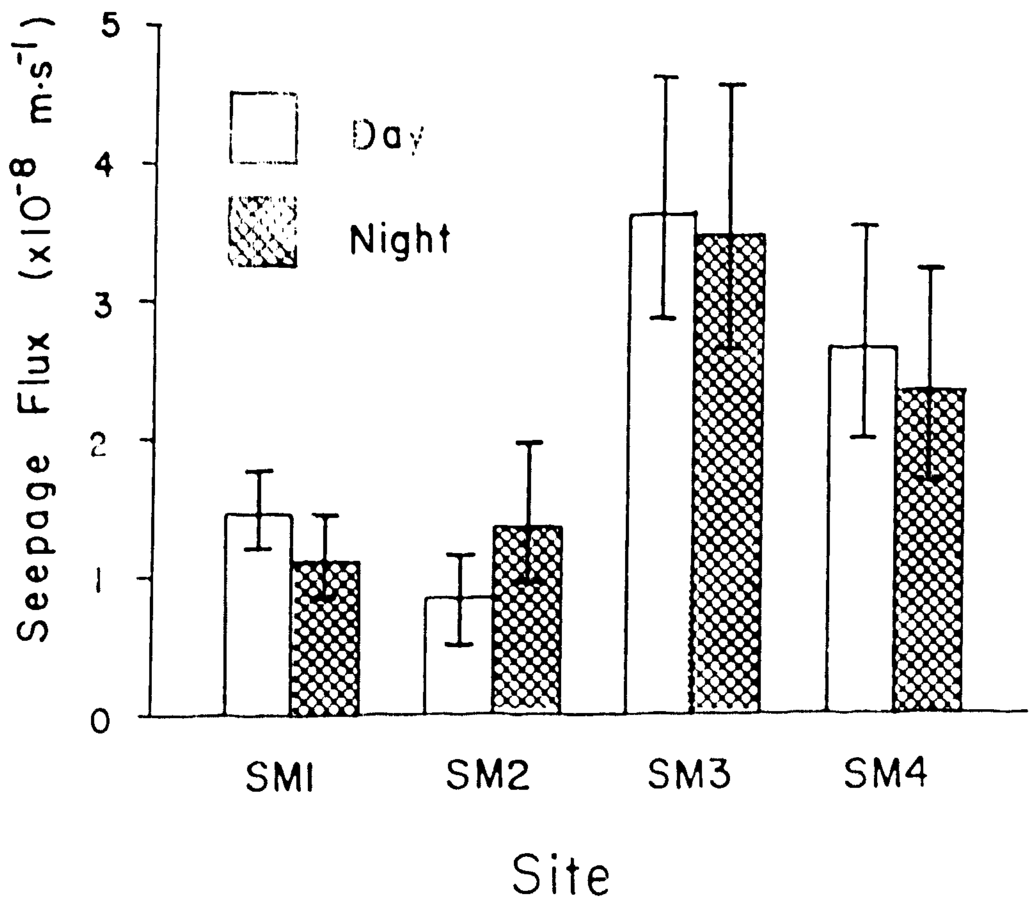

Figure 5.8. Geometric mean and 95 % confidence intervals of seepage flux measured during the day and night at four sites in Narrow Lake from 9 to 13 August, 1984.

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6. GROUNDWATER TRANSPORT OF PHOSPHORUS FROM LAKE BOTTOM SEDIMENTS INTO LAKEWATER[1]

6.1 ABSTRACT

Phosphorus (P) is the nutrient that limits algal biomass in most lakes. Lake bottom sediments are rich in P; even so, advective transport of P by groundwater to lakewater is rarely measured. In this study, simple methods were used to estimate *in situ* rates of advective transport of P to six lakes in western Canada. In general, input of P from anaerobic sediment porewater to aerobic lakewater was not inhibited by sorption onto iron- or calcium-complexes. Advective transport of P to the lakes averaged 176, 35, and 285% of P inputs from molecular diffusion, surface runoff and atmospheric deposition, respectively.

[1] Porewater P data were collected in conjunction with J.F.H. Shaw

6.2 INTRODUCTION

Phosphorus (P) is the nutrient that limits algal production in most lakes (Schindler 1977); thus, much effort has been directed to quantify sources of P to lakes. Lake bottom sediments contain a vast pool of P, and groundwater flowing into lakes may transport P from sediment porewater into the overlying lakewater. Groundwater seepage into lakes is generally concentrated near shore (Winter 1976). Thus, materials transported by groundwater to lakes enters surface water directly and could be utilized by algae. However, advective transport of P by groundwater into lakes has been largely ignored, probably because of perceived difficulties in measuring groundwater-lake flux (Brock et al. 1982). In addition, P release from anaerobic porewater into aerobic lakewater is thought to be unlikely due to sorption of P onto iron- or calcium-complexes (Otsuki and Wetzel 1972; Tessenow 1974). In this study, I demonstrate that (1) simple methods can be used to quantify advective transport of P from sediments to lakewater, (2) P influx from anaerobic porewater to aerobic lakewater is possible, and (3) advective transport of P could be important relative to other P inputs to six north-temperate lakes in western Canada.

The six study lakes are of glacial origin and are located in central Alberta (Table 6.1). Glacial till is the predominant surficial deposit; it is underlain by sedimentary bedrock (Chapter 4). Surface area and maximum depth of the study lakes varied from 1.1 to 9.2 km^2 and from [illegible] to 36 m, respectively; trophic status ranged from meso-eutrophic to eutrophic (Chapter 4; Shaw 1989; Table 6.1). Morphometry and trophic status of the study lakes are representative of other lakes in the region (Prepas and Trew 1983).

6.3 MATERIALS AND METHODS

Groundwater seepage was measured *in situ* with seepage meters at 4 to 10 sites along one transect in each lake (Chapter 4). Seepage flux was computed from the volume of water collected in the bag after appropriate corrections for anomalous short-term influx of water in bag and area of bottom enclosed by the cylinder (Chapter 2). The transects extended perpendicular from shore to 30 to 100 m from shore (Table 6.1). The location of the transect within each lake was selected to represent the average slope of the bottom sediments (i.e., ratio of distance from shore to lake depth). The average seepage flux measured along each transect was weighted for distance from shore (Chapter 4).

Within 24 h of sampling seepage meters, sediment porewater was collected with peepers at two to nine sites along each transect (Shaw 1989). Peepers are dialysis chambers which collected porewater 0-10 cm below the sediment-water (Hesslein 1976). Porewater was analyzed for soluble reactive P (SRP), Fe^{2+}, and pH (Murphy and Riley 1962; American Public Health Association 1980). Porewater SRP and Fe^{2+} were assumed to be the P and Fe fractions, respectively, in sediment porewater that could be transported by groundwater to overlying lakewater. Oxygen penetration into sediments is limited to < 4mm below the sediment-water interface (Carlton and Wetzel 1988). Therefore, I assumed porewater 0-10 cm below the sediment-water interface was anaerobic; thus, all Fe in porewater would be Fe^{2+}.

At five lakes, seepage flux and porewater SRP concentrations were measured once from 20 May to 21 July, 1986; at Narrow Lake they were measured twice (24 May and 23 June, 1986). The rate of advective transport of P along each transect was computed from the average seepage flux and average porewater SRP concentration. Average seepage flux ($\bar{v}$, log-m.d^{-1}) and porewater P concentration ([P], log-mg.m^{-3}) along each transect were calculated from $\log_{10}$-transformed data (Chapter 4; Shaw 1989). Therefore, the average rate of advective transport of P along each transect (L, in mg.m^{-2}.d^{-1}) was calculated as $L = 10^{L}$, where $L = \bar{v} + [P]$. The variance of L (s^2_L) was computed from $s^2_L = s^2_v + s^2_{[P]}$.

It is instructive to compare the potential importance of advective transport of P to other sources of P to the study lakes. Atmospheric deposition of P to lakes in the study area was measured during 1983 to 1986 (Chapter 7). Surface loading rates of P to the study lakes waspreviously measured by Shaw (1989), Yonge and Trew (1989), and Chapter 5. Rates of molecular diffusion of P from the sediments to the overlying lakewater were previously calculated for the transects (Shaw 1989); thus, rates of advective transport and molecular diffusion could be compared directly. However, to compare groundwater to external sources of P to the lakes, the rate of advective transport of P measured along the transects had to be extrapolated over the entire surface area of the lake and over the entire year (i.e., atmospheric deposition and surface runoff were expressed as areal loading rates, in units of mg P per m^2 lake surface area per year).

6.4 RESULTS AND DISCUSSION

Seepage flux along the transects (10^{-10} to 10^{-8} m.s^{-1}) was within the range measured with seepage meters at other lakes in North America

(Table 4.4). At two seepage meter sites in Narrow Lake (24 May 1986), there was no seepage flux; at one site in Tucker Lake, seepage was from the lake to the groundwater system. Average seepage flux along the transects varied 3-fold between the study lakes (Table 6.2). Except for transects at Baptiste and Jenkins lakes, seepage flux decreased significantly ($P<0.05$) with distance from shore. The predominance of nearshore groundwater discharge into the lakes and patterns of decreased seepage flux with distance from shore are consistent with hypotheses generated from computer simulation models of hypothetical groundwater-lake systems (Winter 1976).

Porewater SRP concentrations along the transects (29 to 2274 $mg.m^{-3}$) were within the range measured in porewater of shallow sediments at other lakes in North America and Europe (Holdren et al. 1977; Carignan 1984; Drake and Heany 1987). The range of porewater Fe^{2+} concentrations along the transects (65 to 1954 $mg.m^{-3}$) was of similar magnitude as SRP concentrations, and porewater pH ranged from 7.5 to 9.0 (Table 6.2). Average porewater SRP concentrations along the transects varied 31-fold between lakes. Unlike seepage flux, there was no distinct pattern of porewater SRP concentrations with distance from shore.

Phosphorus is reactive. Under aerobic conditions, and if a minimum molar ratio of Fe:P of 1.8 is exceeded, most P sorbs onto iron oxy-hydroxides (Tessenow 1977). The formation of these iron-phosphorus complexes would prevent influx of P from anaerobic porewater to aerobic lakewater. However, porewater from the study lakes had an average Fe:P molar ratio of 0.9, and the critical value (1.8) was exceeded at only 4 of 20 peeper sites. At pH greater than 8, P can sorb onto calciumcarbonates, and sorption increases with increasing pH (Otsuki and Wetzel 1972). However, porewater pH at all lakes but Tucker Lake was 8 or less; at Tucker it was 9.0 (Table 6.2). Therefore, at most sites in the study lakes influx of P from porewater to lakewater was not inhibited by sorption of P onto Fe- or Ca-complexes.

Average rates of advective transport of P along the transects varied nearly 50-fold, from 0.06 to 2.8 $mg.m^{-2}.d^{-1}$ (Table 6.2). There was no significant difference between rates of advective transport along the transects at Narrow Lake for May and June 1986 ($\underline{t}$=0.85, $\underline{df}$=13, $\underline{P}$>0.40). For the study lakes, rates of advective transport of P along the transects were, on average, 206% of rates of molecular diffusion of P along the transects (Table 6.2).

Expressed over the entire surface area of the lake, P inputs from nearshore groundwater discharge ranged from 9 to 215 $mg.m^{-2}.yr^{-1}$ (Table 6.3). Narrow Lake is the only study lake where groundwater P loading has been previously estimated (Chapter 5). In that study, groundwater P loads were determined from data collected from 1983 to 1988; the rate of groundwater P loading was four times greater than the value estimated in this study.

Rates of groundwater P loading to the study lakes should be considered preliminary because of errors that may be introduced from extrapolating groundwater and porewater data collected from one site in the lake over the entire surface area of the lake. Phosphorus inputs from groundwater to Lake Mendota, WI, were estimated with comparable methods to those in this study (Brock et al. 1982). Interestingly, the rate of advective P transport to Lake Mendota (113 $mg.m^{-2}.yr^{-1}$) was within the range of values for my study lakes (10 to 215 $mg.m^{-2}.yr^{-1}$;

Table 6.3). Other studies have estimated groundwater P inputs to lakes based on P concentrations measured in nearby water wells (Chapter 5; Trew et al. 1987). Those water samples may be adequate for estimating inputs of conservative solutes into lakes (Hurley et al. 1985), but are likely not representative of non-conservative solutes. For example, P concentrations in porewater at Narrow Lake were, on average, 8-fold higher than P concentrations measured in wells near the lake (Chapter 5). To accurately estimate advective transport of P (or other non-conservative solutes) to lakes, samples should be collected as close as possible to the point of discharge into the lake (i.e., sediment-water interface).

Estimates of annual influx of P from groundwater to the study lakes was, on average, 36 and 285% of P input from surface runoff and atmospheric deposition, respectively (Table 6.3). Thus, advective transport was a relatively important source of P to the surface waters of the six study lakes.

Methods used in this study can be applied to measure advective transport of both conservative and non-conservative material to lakes. Although groundwater flow rates into lakes situated in glacial till are relatively low, this study demonstrates that groundwater can potentially be a major source of P to these lakes. Groundwater inputs should not be overlooked when preparing nutrient budgets for lakes.

Table 6.1. Location, surface area (A_o, km^2), maximum depth (z_{max}, m), transect length (TL, m), ratio of shallow sediments (A_s) to lake surface area, transect length (TL, m), and mean summer trophogenic total phosphorus (TP, $mg.m^{-3}$) of the six study lakes.

Lake	Location N	Location W	TL	A_o	z_{max}	$A_s:A_0$	TP
Baptiste	54° 45′	113° 33′	65	9.2	268	0.28	38
Island	54° 52′	113° 31′	47	7.3	15	0.14	20
Jenkins	54° 55′	113° 36′	45	1.8	18	0.21	25
Long	54° 34′	113° 38′	30	1.6	28	0.22	13
Narrow	54° 35′	113° 37′	40	1.1	36	0.29	11
Tucker	54° 32′	110° 37′	100	6.6	8	0.24	42

Table 6.2. Seepage flux, porewater soluble reactive phosphorus (SRP) concentrations and rates of advective transport and molecular diffusion of P along the transects: geometric mean and 95 % confidence intervals (in brackets). Also indicated are average pH and ferrous iron concentrations in porewater along the transects.

Lake	Flux $\times10^{-8}$ m.s^{-1}	pH	Fe mg.m^{-3}	SRP mg.m^{-3}	Advective Transport mg.m^{-2}.d^{-1}	Molecular Diffusion mg.m^{-2}.d^{-1}
Baptiste	2.18 (1.69-2.81)	7.9	247	415 (250-690)	0.78 (0.6-1.0)	0.14
Island	0.93 (0.62-1.41)	8.0	232	225 (115-442)	0.18 (0.1-0.7)	0.14
Jenkins	1.54 (0.29-8.10)	8.0	379	2113 (824-5420)	2.8 (0.7-11 1)	1.02
Long	0.89 (0.32-1.95)	7.9	945	230 (65-817)	0.18 (0.08-0.4)	0.51
Narrow May	1.31 (0.88-1.95)	7.8	250	57 (19-175)	0.06 (0.04-0.11)	0.06
June	1.00 (0.78-2.28)	-	-	144 (68-303)	0.12 (0.08-0.20)	0.19
Tucker	0.69 (0.34-1.42)	9.0	168	348 (168-720)	0.21 (0.11-0.38)	0.30

Table 6.3. Comparison of phosphorus loading rates by advective transport and external (runoff and atmospheric deposition) sources for the six study lakes. Rates are expressed as $mg.m^{-2}.yr^{-1}$ over the entire lake surface.

Lake	Advective Transport	External	
		Runoff	Atmospheric Deposition
Baptiste	79.7	480	20.3
Island	9.2	74	20.3
Jenkins	214.6	560	20.3
Long	14 5	144	20.3
Narrow	9.5	9	20.3
Tucker	18.4	88	20.3

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7. ATMOSPHERIC DEPOSITION OF PHOSPHORUS AND NITROGEN IN CENTRAL ALBERTA WITH EMPHASIS ON NARROW LAKE[1]

7.1 ABSTRACT

Average rates of atmospheric deposition of total phosphorus (TP) and total nitrogen (TN) to Narrow Lake, located on sedimentary bedrock in the boreal forest of central Alberta, were 20 and 424 $mg.m^{-2}.yr^{-1}$, respectively, between 1983-1986. There were no significant differences ($\underline{P}$>0.05) in deposition rates between sites on Narrow Lake, on the lake shore, and on land 18 km away. Deposition of TP, but not TN, followed a distinct pattern during the open-water season; TP was highest just after ice-off (May) and decreased throughout the remainder of the open-water season. Deposition during the winter accounted for only 4 and 12% of the annual TP and TN loads, respectively. Dry fallout contributed 50 and 33% of atmospheric deposition of TP and TN, respectively. In both dry and wet fallout, dissolved P (<0.45 μm) and organic N were the predominant fractions of TP and TN, respectively. During July 1986, unusually heavy rainfalls caused an increase in TP, but not TN, concentrations in the epilimnion of Narrow Lake. Wet fallout accounted for only 9% of the observed increase of epilimnetic TP; the rest was from surface runoff from the drainage basin. The design of sampling programs to measure atmospheric deposition of nutrients to lakes is discussed.

[1] A version of this chapter has been published. R.D. Shaw, A.M. Trimbee, A. Minty, H. Fricker, and E.E. Prepas. 1989. Water, Air, and Soil Pollut. 43:119-134.

7.2 INTRODUCTION

Atmospheric deposition can be a significant source of P and N to lakes. Globally, bulk loading rates (wet plus dry deposition) of total P (TP) and total N (TN) range from 1 to 800 $mg.m^{-2}.yr^{-1}$ and from 8 to over 3000 $mg.m^{-2}.yr^{-1}$, respectively (Uttormark *et al.*, 1974; Welch and Legault, 1986). The high loading rates far surpass the permissible loads for these nutrients to many lakes (Vollenweider, 1968).

From 1983 to 1986, I measured atmospheric deposition to evaluate: (1) spatial variability of P and N loads over the surface of a medium-sized lake in central Alberta (Narrow Lake: surface area 1.1 km^2), (2) differences in P and N loads to the lake and the lake shore, and to the lake and a site on land 18 km away, (3) temporal variability of TP and TN loads between seasons and years, and (4) components of TP and TN loads including wet and dry fallout, dissolved and particulate P, and inorganic and organic N. In addition, unusually heavy rains fell in July 1986, which enabled me to evaluate the contribution of P and N from storms to the nutrient budget of the epilimnion in Narrow Lake, Alberta. This paper describes the results of data, collected over four years, to evaluate atmospheric deposition of P and N to lakes in central Alberta. Results from this study are compared to other atmospheric deposition studies and used to evaluate the design of sampling programs to measure atmospheric loading rates of nutrients to lakes.

7.3 MATERIALS AND METHODS

Description of study area- Narrow Lake (54°35'N, 113°37'W) is a deep, medium-sized, mesotrophic lake in the mixed wood section of the boreal forest of central Alberta (mean depth 14.4 m; surface area 1.1 km^2; mean summer chlorophyll *a* 3 $mg.m^{-3}$; Prepas and Trimbee, 1988). The drainage basin (area 7.0 km^2) is almost completely forested. There are two small camps on the lake shore, one on the east and one on the south shore. Glacial till is the major surficial deposit in the drainage basin. The soil is predominantly Orthic Gray Luvisol (Kjearsgaard, 1972). Although there is no agricultural or industrial activity in the drainage basin, there are farms within 5 km of the lake. Edmonton, 140 km south of Narrow Lake, is the closest major urban or industrial site.

The climate in the study area is continental; average annual precipitation and lake evaporation are 503 mm and 636 mm, respectively (Hydrology Branch, Alberta Environment, unpublished data). From 1983 to 1986, the open-water season at Narrow lake extended, on average, from April 29 to November 11 (54% of the year). Most (73%) precipitation falls during the open-water season and nearly half (48%) falls during three summer months (June, July and August; Fig. 7.1).

Data collection- During the open-water season (1983-1986), I obtained samples of wet and dry fallout using collectors constructed with two types of funnels; one was plastic (surface area 0.04 m^2) and the other was metal, lined with teflon (surface area 0.25 m^2). The funnels were continuously exposed to the atmosphere and they drained into polyethylene bottles (reservoirs). The collectors with metal funnels were used exclusively on land-based sites and were similar to those used by Mitchell (1985). The collectors with plastic funnels were a variant of the Hubbard Brook Rain Collector (Galloway and Likens, 1976). Inner-tubes from truck tires were used to float collectors on Narrow Lake.

The top of the funnel was at least 50 cm above the water level. Even when the waves were large, lake water did not splash into these collectors. The reservoirs were emptied at intervals of 0.7 to 21 d, depending on rainfall and logistics. Dry fallout was estimated after periods of no rain by washing 1 to 4 L of double-distilled water (DDW) down the funnel and collecting the water from the reservoir. Rain water was collected from the reservoir 1 to 48 hr after the rain ended.

Samples were analyzed for TP by the potassium persulphate digestion method (Menzel and Corwin, 1965). Water for TDP analysis was filtered through a prewashed (DDW) 0.45 μm Millipore HAWP membrane filter and analyzed as for TP. Particulate P (PP) was calculated as the difference between TP and TDP, thus, representing P greater than 0.45 μm. Total Kjeldahl N (TKN) was analyzed by Solorzano's (1969) phenolhypochlorite method. NH_4-N and NO_2-N+NO_3-N were determined with a Technicon-Autoanalyzer II (Stainton et al., 1977). Total N (TN) was calculated as the sum of TKN and NO_2-N+NO_3-N, inorganic N the sum of NH_4-N and NO_2-N+NO_3-N, and organic N was TKN minus NH_4-N.

All statistical tests were from Sokal and Rohlf (1981).

Spatial variability- In 1984, collectors with plastic funnels were placed at three sites on Narrow Lake and in 1986 at two sites to evaluate spatial variability in atmospheric deposition over the surface of the lake. A single collector was placed on Narrow Lake in 1985. To evaluate spatial variability between the lake, the lake shore, and a site on land, 18 km away, collectors with metal funnels were also placed on the east shore of Narrow Lake and at the Meanook Biological Research Station (MBRS) in 1985 and 1986. In addition, a collector with a plastic funnel was placed at MBRS in 1986. For each year (1984 to 1986), differences in rainfall and loading rates between sites were examined by two-way ANOVA; collectors were treatments and sampling periods were blocks. Separate ANOVA's were carried out for each nutrient parameter (i.e., TP, TDP, TKN, NH_4-N, and NO_2-N+NO_3-N) and condition (bulk, wet and dry fallout). The treatment sum of squares was partitioned into orthogonal components; differences in loading rates between specific sites were tested (i.e., within Narrow Lake, between the lake and lake shore, and between the lake and MBRS).

Temporal variability- Atmospheric nutrient loading rates ($mg.m^{-2}.d^{-1}$), for the open-water season, were calculated from the nutrient concentration in the reservoir ($mg.m^{-3}$), multiplied by the volume of water collected (m^3), and divided by the surface area of the funnel (m^2) and the time period the sample represented (d). The daily rates were used to estimate monthly loading rates of TP (1983: September and October; 1984: May to September; 1985 and 1986: May to August) and TN (1985 and 1986: May to August). If no data were collected for a particular month, the rate was assumed to equal the average measured for that month in other years.

Atmospheric deposition to Narrow Lake was estimated for the ice-covered period in 1985 to 1986. TP concentrations were determined from samples of fresh snow, scooped from four to six sites on the surface of Narrow Lake on 25 February and 18 March 1986; TN concentration was determined from snow collected 18 March 1986. Nutrient loading rates ($mg.m^{-2}$) were then calculated from average TP or TN concentrations ($mg.m^{-3}$) in the snow multiplied by total precipitation (in m) from 9

November 1985 to 7 May 1986 (i.e., period of ice cover).

Partitioning nutrient loads- The contribution of wet and dry fallout to atmospheric deposition was evaluated by comparing loading rates measured during periods of rain and no rain, respectively (TP: May to August, 1985 and 1986; TN: May to August, 1986). There were not enough data to compare wet and dry fallout rates for other years of the study. The contribution of TDP and PP to TP loads was assessed with samples collected from May to August 1984 to 1986. The contribution of inorganic and organic N, and NH_4-N and NO_2-N+NO_3-N, to TN and inorganic N loads, respectively, was assessed with samples collected from May to August 1986.

Summer storm- From 8 to 18 July 1986, there were heavy rains which caused flooding over much of central Alberta. During this period, wet fallout was sampled; integrated water samples were collected from the epilimnion (0 to 6 m) of Narrow Lake by Trimbee et al. (1988). Epilimnetic water was sampled before, during, and after the storm (July 7, 14, 21, respectively) and analyzed for TP and TN. A mass balance method was used to determine the increase of TP and TN in the epilimnion of Narrow Lake from the beginning to the end of the storm. The increase of TP and TN in the epilimnion, caused by rain falling directly onto the lake surface, was estimated from the measured atmospheric load for that time period multiplied by the surface area of Narrow Lake. Atmospheric inputs of TP and TN were then compared to the total change in mass of these nutrients in the epilimnion over that period.

Initial data analysis- The surface area of the plastic funnel was 16% that of the metal funnel. Subsequently, less rain was obtained from collectors with plastic vs metal funnels. To test whether the funnels caused bias in estimates of loading rates, collectors with plastic and metal funnels were placed next to each other at MBRS, from May to August 1986. There were no significant differences (Paired t-tests; $P>0.10$) between the paired collectors at MBRS in rainfall or loading rates of any of the nutrient fractions (TDP, TP, TKN, NH_4-N, or NO_2-N+NO_3-N) under any conditions (wet, dry or bulk fallout) that were tested (Table 7.1).

7.4 RESULTS

Spatial variability- Spatial variability in atmospheric deposition of nutrients between Narrow Lake, the lake shore, and MBRS was examined with data collected from 1984 to 1986. There was no significant variation in the amount of rainfall received at each site ($P>0.10$). Loading rates at most sites were highly variable; coefficients of variation (CV) ranged from 9 to 300% (Table 7.2). Overall, there were no significant differences between loading rates of all nutrient fractions and conditions that were tested ($P>0.05$; Table 7.3). The treatment sum of squares from each of the 21 ANOVA's was partitioned into orthogonal components to test for differences in loading rates between specific sites. Of the 72 orthogonal components, there were significant differences in only two: TDP and NH_4-N loading rates in wet fallout were higher during 1986 at MBRS than at Narrow Lake (TDP: $df=1,8$, $F=6.18$, $P<0.05$; NH_4-N: $df=1,32$, $F=8.95$, $P<0.01$)

<u>Temporal variability</u>- During the open-water season (May-October), Loading rates for TP (1983 to 1986) and TN (1985 and 1986) ranged from).005 to 0.98 and 0.3 to 14.0 mg.m^{-2}.d^{-1}, respectively. Monthly loading :ates (May to September) of TP were consistent between years; they were ıighest in May, just after ice-off, and decreased for the remainder of :he open-water season (Table 7.4). TP loading rates in September (1983 ınd 1984: mean=1.1 mg.m^{-2}) and October (1983: 0.2 mg.m^{-2}) were only 15 ınd 3%, respectively, of the average for May. Average loading rates ıeasured in September 1983 and 1984, and in October 1983, were used as 'alues for these months in 1984 (October only), 1985 and 1986. TP oading rates during the open-water season ranged from 17.1 mg.m^{-2} in 985 to 22.9 mg.m^{-2} in 1986; the average was 19.6 mg.m^{-2} (Table 7.4).

TN loading rates (May to August) in 1985 were only a third of the N loading rates for the same period in 1986 (Table 7.4). The lower TN oads measured in 1985 <u>vs</u> 1986 may be due to lower rainfall in 1985 <u>vs</u> 986. I had no estimates of TN loading rates for September or October, o I used the values measured for this period at other lakes in Alberta September to October: mean loading rate 48.5 mg.m^{-2}; Pollution Control ivision, Alberta Environment, unpublished data). The atmospheric eposition rate of TN during the open-water season in 1986 was 534 g.m^{2}, 245% higher than in 1985.

In 1986, TP concentrations in the snow were low in February and arch: 5.3 and 5.0 mg.m^{-3}, respectively. Similarly, TN concentrations ere low in March: 306 mg.m^{-3}. From 9 November 1985, to 7 May 1986, i.e., period of ice cover) there was 148 mm of precipitation. The alculated loading rates of TP and TN for the ice-covered season were .8 and 45 mg.m^{-2}, respectively, and were only 4 and 12% of the open-water loads, respectively. Average annual atmospheric loads (1984 to 1986) of TP and TN were 20.3 and 424 $mg.m^{-2}$, respectively (Table 7.4).

Partitioning nutrient loads- From May to August 1985 and 1986, wet and dry fallout both contributed, on average, 50% of bulk fallout of TP. From May to August 1986, wet fallout contributed, on average, 67% of bulk fallout of TN. Although the concentrations of all nutrient parameters decreased with rainfall (e.g., Fig. 7.2), the relative contribution of wet to bulk fallout of TP or TN was greatest during months of heavy precipitation (Fig. 7.3). In July 1986, when the highest monthly precipitation was recorded (180 mm), wet fallout contributed 94 and 90% of bulk fallout of TP and TN, respectively. Conversely, during May 1986, when the lowest monthly precipitation was recorded (23 mm), wet fallout contributed only 31 and 35% of bulk fallout of TP and TN, respectively. From May to August, 1984 to 1986, 51 to 84% of monthly bulk fallout of TP was in the dissolved fraction (Fig. 7.4A). From May to August 1986, 11 to 23% of TN was in the inorganic fraction; NH_4-N was the major inorganic N input contributing 53 to 78% of inorganic-N (Fig. 7.4B).

Summer storm- During July 1986, there was 180 mm of precipitation (long-term average is 90 mm), and 87% of it (157 mm) fell from 9 to 18 July (Fig. 7.5). At Narrow Lake, inflows and outflows are dammed by beavers. Therefore, surface runoff from the watershed to the lake is restricted to periods after heavy rainfall when water levels in the beaver impoundments increase and the dams overflow. The water then flows through marshy areas into the lake; however, there are no distinct channels to gauge the discharge of surface runoff into Narrow Lake. On

15 July, following 87 mm of rain (8 to 15 July), the dams were breached. Surface water flowed to and from Narrow Lake throughout the remainder of July. From the day before the storm started (7 July) to the day before the beaver dams overflowed (14 July), precipitation falling directly on the lake surface was the major source of water to Narrow Lake. During this period, there was an increase of 69 mm in lake level and a comparable amount of precipitation (68 mm). The importance of precipitation as a source of water to the lake decreased after this period (e.g., 14 to 20 July: lake level increased 287 mm and there was 89 mm of rain).

During the period when rain from the storm was the major source of water to Narrow Lake (i.e., 7 to 14 July), there was no change in TP concentration in the epilimnion. However, TP concentration in the epilimnion increased by 21 July and remained high for the balance of the summer (Fig. 7.5). That increase of TP concentration in the epilimnion significantly affected water clarity, phytoplankton biomass, vertical stratification, species composition and nutrient status (Trimbee et al., 1988). From 7 to 14 July, TP concentration in wet fallout (10.3 $mg.m^{-3}$) was similar to that in the epilimnion of Narrow Lake; however, the total load of TP from the atmosphere (0.8 kg) was only 1% of epilimnetic TP on 7 July. From 15 to 21 July, TP concentrations in wet fallout were lower (6.1 $mg.m^{-3}$) than in the previous week, and the atmospheric load was only 0.7 kg. However, TP in the epilimnion increased during that period by 17 kg (28%), suggesting that sources other than atmospheric deposition (e.g., surface runoff) were responsible for the observed increase in epilimnetic TP. In contrast to TP, there was no change in TN concentration in the epilimnion (Fig. 7.5). From 7 to 21 July, the atmospheric load of TN was 119 kg; this was only 3% of epilimnetic TN on 7 July (3501 kg).

7.5 DISCUSSION

Atmospheric deposition of nutrients to north-temperate lakes was the primary focus of many studies (e.g., Barica and Armstrong, 1971; Gomolka, 1975; Nicholls and Cox, 1977; Peters, 1977; Caiazza et al., 1978; Jeffries et al., 1978; Scheider et al., 1979; Linsey et al., 1987). With one exception, these studies were at sites in eastern Canada on igneous bedrock. In Caiazza et al.'s (1978) study, atmospheric deposition of P and N was measured at a site in central Alberta on sedimentary bedrock. However, their results are not realistic because of problems with their analytical methods to determine P (G. Hutchinson, Department of Zoology, University of Alberta, personal communication). This study is the first to examine in detail, spatial and temporal patterns of atmospheric deposition of P and N to north-temperate lakes on sedimentary bedrock.

Spatial variability- Spatial variability of atmospheric deposition of TP has also been examined for two lakes on the Precambrian Shield near Dorset, Ontario: Lake St. Nora (surface area 290 ha; atmospheric TP load 37 $mg.m^{-2}.yr^{-1}$; Gomolka, 1975) and Red Chalk Lake (surface area 75 ha; atmospheric TP load 16 $mg.m^{-2}$ from June to October; Jeffries et al., 1978). As for Narrow Lake, there were no consistent differences in bulk fallout between sites on these lakes. However, at both Lake St. Nora and Red Chalk Lake, TP loads were significantly higher at sites off the lake (in the drainage basin) than on the lake. Gomolka (1975) attributed the difference to the relatively higher amounts of particulate dry fallout that were collected at sites on land.

Annual rates of atmospheric deposition of TP and TN to Narrow Lake (20.3 and 424 $mg.m^{-2}.yr^{-1}$ for TP and TN, respectively) were similar to those measured at other lakes on sedimentary bedrock in central Alberta (TP: mean (SE) = 20.6 (2.2) $mg.m^{-2}.yr^{-1}$, n=12; TN mean (SE) = 358 (34) $mg.m^{-2}.yr^{-1}$, n=7; Mitchell, 1985; Trew et al., 1987; Pollution Control Division, Alberta Environment, unpublished data). Atmospheric loading rates of TP to lakes in central Alberta were within the lower range of values measured at sites on the Precambrian Shield in Ontario (6 to 77 $mg.m^{-2}.yr^{-1}$; Scheider et al., 1979). Annual loads of TN to lakes in central Alberta (264 to 500 $mg.m^{-2}.yr^{-1}$) are lower than most rates reported from sites on the Precambrian Shield in Ontario (495 to 1600 $mg.m^{-2}.yr^{-1}$; Scheider et al., 1979). The low deposition rates of TP and TN observed in central Alberta may reflect the limited impact of anthropogenic sources compared to Ontario, which is more populated and industrialized.

Temporal variability- Daily atmospheric loading rates of TP at Narrow Lake ranged over two orders of magnitude. A wide range in loading rates at a single site has been observed at other locations (e.g., Jeffries, 1984) and can be explained by processes such as sedimentation and washout which remove P-aerosols from the atmosphere (Brezonik, 1975). Daily loading rates of TN at Narrow Lake were less variable than for TP, but still ranged over one order of magnitude.

At Narrow Lake, monthly TP loads were highest in May and decreased over the remainder of the summer (Table 7.4). A similar pattern was observed at sites in forested areas on the Precambrian Shield in Ontario (Linsey et al. (1987): Experimental Lakes Area (ELA); Gomolka (1975): Lake St. Nora). This pattern was attributed to spring pollen inputs (Gomolka, 1975) and soil cultivation on the prairies (Linsey et al. 1987). At Narrow Lake, it is unlikely that pollen was responsible for the high May input, because for three consecutive years the dissolved fraction of P (<0.45 μm) made up most of the TP load (Fig. 7.4A). The average size of pollen is about 10 μm (Junge, 1963); pollen would have been included in my PP fraction. High TP loading rates during May may have been caused by very small particles (i.e., <0.45 μm) of mineral dust that had accumulated in the atmosphere over the winter and early spring (perhaps from soil cultivation in April and May). In May, changes in meteorological conditions (e.g., increased relative humidity and precipitation) may have facilitated the deposition of these small particles.

Very low rates of atmospheric deposition of TP to Narrow Lake were measured for Septembe .1 $mg.m^{-2}$) and October (0.2 $mg.m^{-2}$). This pattern is consistent with loads measured at other lakes in Alberta (e.g., September-1.0 $mg.m^{-2}$; October-0.9 $mg.m^{-2}$; Pollution Control Division, Alberta Environment, unpublished data). These low rates indicate that, in this region, soil disturbances related to agricultural activity in the fall (e.g., harvesting) are not a significant source of atmospheric nutrients. The low rates of atmospheric deposition in September and October may reflect: (1) changes in meteorological (e.g., temperature) and biological conditions that decrease the release of mineral dust to the atmosphere, and/or (2) a decrease in rainfall and subsequently, in deposition of P-containing aerosols to the lake.

Unlike TP, monthly TN loading rates showed no distinct patterns. This lack of a pattern may reflect differences in the sources of atmospheric N as compared to P. Nitrogen is derived from atmospheric (i.e., gaseous forms of N) and terrestrial (i.e., dust) sources, while P is derived solely from terrestrial sources.

At Narrow Lake, the amount of TP and TN that accumulated on the ice was very low (0.8 and 45 $mg.m^{-2}$, respectively). Low rates of TP and TN deposition during the ice-covered season have also been reported at other sites in Alberta (e.g., Wabamun Lake: 3.5 and 87 $mg.m^{-2}$, respectively; Mitchell, 1985) and northwest Ontario (e.g, 4.0 and 59 $mg.m^{-2}$, respectively; Barica and Armstrong, 1971). Even though nutrient loads during the winter are low, they may be of significance to lake biology since infusion of nutrients into the lake water during spring thaw may be rapidly taken up by phytoplankton and incorporated in the food chain (Barica and Armstrong, 1971).

Partitioning nutrient loads- Dry fallout has received less attention than wet fallout as a source of atmospheric deposition of TP and TN. One reason for the focus on wet fallout is that it is more difficult to obtain accurate estimates of dry fallout. For particles less than 20 μm, turbulence (e.g., changes in wind patterns created by a forest) and impact onto surfaces, are the dominant mechanisms for dry deposition (White and Turner, 1970). Thus, to accurately measure dry fallout to a lake, the collector should simulate the lake surface. However, there is contradictory evidence about whether estimates of dry fallout of P are affected by the collector surface. One study (Gomolka, 1975) showed that 80% more TP was collected on a wet surface as compared with a dry surface. In contrast, Lewis (1983) reported no differences in rates of soluble reactive or dissolved organic P between collectors with wet or dry surfaces. If dry fallout measured on Narrow Lake was increa...d by 8 $mg.m^{-2}$, as Gomolka's study would indicate, the annual TP load to Narrow Lake would be 28 $mg.m^{-2}$, or 40% higher. Regardless, my measured rates indicate dry fallout was an important source of TP and TN to Narrow lake. Dry vs wet fallout ratios were 1 to 1 and 1 to 3, for TP and TN, respectively. Similar wet to dry ratios for both TP and TN have been reported at other north temperate lakes (e.g., Scheider et al., 1979).

Most (80%) of the atmospheric N input to Narrow Lake was organic. At other lakes in central Alberta, organic N was 33 to 62% of TN (Pollution Control Division, Alberta Environment, unpublished data). Organic N was less important as a source of atmospheric N to lakes in Ontario, on the Precambrian Shield; organic N was 20 to 37% of TN (TN loading rates at these lakes ranged from 1010 to 1270 $mg.m^{-2}$; Jeffries, 1984; Nicholls and Cox, 1977; Scheider et al., 1979). The importance of atmospheric organic N as a source of available nitrogen to lakes has not been evaluated. Of the inorganic N fraction in wet fallout at Narrow Lake, NH_4-N input was slightly larger than the NO_2-N+NO_3-N input: NH_4-N was 53 to 78% of the inorganic N load during 1986. A similar proportion (53%) was measured at ELA (Linsey et al., 1987).

Sampling Design- In general, atmospheric deposition is an important source of the annual nutrient load to a lake when the drainage area (A_d) is small and the lake surface area (A_o) is large. In my study, atmospheric deposition rates of TP and TN to lakes were 20.3 and 424 $mg.m^{-2}.yr^{-1}$, respectively. In comparison, TP and TN loading from surface runoff to lakes from forested watersheds in central Alberta were 10 and 100 $mg.m^{-2}.yr^{-1}$, respectively (Mitchell, 1985; Munn and Prepas, 1986; Trew et al., 1987). Therefore, in forested regions in central Alberta, atmospheric deposition contributes 50% of the annual surface load of TP and TN to lakes where $A_d:A_o$ are 2:1 and 4:1, respectively, and 25% where $A_d:A_o$, are 6:1 and 13:1, respectively. At forested lakes in central Alberta, where $A_d:A_o$ is greater than 30, atmospheric deposition is relatively unimportant (6 and 12% of total surface load of TP and TN, respectively). My rates could be used as estimates of atmospheric loading to those lakes. Export coefficients in central Alberta from cultivated watersheds are 16 and 186 $mg.m^{-2}.yr^{-1}$ so atmospheric deposition is relatively less important when the drainage basin is cultivated rather than forested.

In central Alberta, there is little accumulation of snow over the winter because precipitation is low and snow-melt extends over a long period. Subsequently, nutrient inputs to lakes from runoff from the drainage basin are low in May. In contrast, the highest monthly atmospheric loads are in May. In addition, many lakes in central Alberta do not mix completely following break-up of ice cover (late April to early May), and there is little or no nutrient input from deep to shallow lake water. I did not evaluate the bioavailability of atmospheric nutrients; however, estimates from atmospheric fallout in eastern Canada were as much 100% for TP in rain (Peters, 1977), 24% for TP in snow (Peters, 1977), and 57% for TP in dry fallout (Gomolka, 1975). Thus, during May, atmospheric deposition may be the most important source of nutrients for primary producers, even though annual rates of fallout may be low relative to other sources.

I found no significant difference between atmospheric loading rates measured at sites on Narrow Lake, the lake shore, or land 18 km away. Therefore, collectors on land, at a convenient location (i.e., MBRS) can be used to accurately measure atmospheric deposition to Narrow Lake. Further investigations are required to determine if the similarity of nutrient loading rates on and off Narrow Lake are common to other north-temperate lakes on sedimentary bedrock in western Canada.

7.7 CONCLUSIONS

This study of spatial and temporal patterns of atmospheric deposition of P and N to north-temperate lakes indicates that:

(1) Increases in epilimnetic TP, related to heavy summer rainfalls, are probably from increased loading rates from surface inflows, rather than wet fallout directly to the lake surface,

(2) There is little spatial variability in atmospheric loading rates of P and N,

(3) Atmospheric deposition rates of TP and TN to forested lakes in central Alberta are generally lower than those measured at forested lakes in Ontario,

(4) Atmospheric loading rates of TP and TN are temporally variable; rates are highest from May to August and are very low during the winter,

(5) Dry fallout is an important component of bulk fallout of TP and TN,

(6) At forested lakes in central Alberta, where A_d:A_o is less than 20, atmospheric deposition may be an important source of nutrients to the lake,

(7) During May, atmospheric deposition may be the most important external source of nutrients to lakes in central Alberta.

Table 7.1 A comparison of precipitation (mm) and nutrient parameters ($mg.m^{-2}.d^{-1}$) for wet, dry and bulk fallout measured during 1986 with plastic and metal collectors at MBRS.

Parameter	Condition	df	*t*	*P*
Precipitation	-	11	0.86	> 0.25
TDP	Dry	[illegible]	1.36	> 0.25
TDP	Wet	10	1.28	> 0.20
TDP	Bulk	19	1.71	> 0.10
TP	Dry	7	1.17	> 0.25
TP	Wet	10	1.14	> 0.25
TP	Bulk	19	1.59	> 0.10
TKN	Bulk	5	0.50	> 0.50
NH_4-N	Wet	10	1.12	> 0.25
NO_2-N+NO_3-N	Wet	11	1.22	> 0.20

Table 7.2. Mean and standard deviation (in mg.m-2.d-1) for parameters (Par.) and conditions (Cond.) measured at collector sites after common sampling periods (n). In 1984, all collectors were on the lake (L84-1, L84-2, L84-3); in 1985, single collectors were on the lake (L85), the lake shore (S85), and at MBRS (M85); in 1986, two collectors were on the lake (L86-1, L86-2), one on the lake shore (S86), and two at MBRS (M86-1, M86-2, plastic and metal collectors, respectively).

Year	Par	Cond	Collector sites			n
1984			L84-1	L84-2	L84-3	
	TP	Bulk	0.132 (0.042)	0.150 (0.071)	0.135 (0.084)	13
		Wet	0.174 (0.030)	0.213 (0.088)	0.188 (0.048)	5
		Dry	0.105 (0.023)	0.110 (0.039)	0.102 (0.052)	8
	TDP	Bulk	0.093 (0.041)	0.099 (0.068)	0.100 (0.063)	13
		Wet	0.132 (0.030)	0.151 (0.074)	0.143 (0.045)	[illegible]
		Dry	0.069 (0.023)	0.067 (0.042)	0.074 (0.060)	8
1985			L85	S85	M85	
	TP	Bulk	0.139 (0.133)	0.135 (0.124)	0.143 (0.138)	22
		Wet	0.188 (0.151)	0.182 (0.190)	0.192 (0.182)	10
		Dry	0.099 (0.104)	0.113 (0.087)	0.102 (0.098)	12
	TDP	Bulk	0.111 (0.114)	0.094 (0.154)	0.082 (0.111)	22
		Wet	0.148 (0.137)	0.124 (0.227)	0.098 (0.139)	[illegible]
		Dry	0.070 (0.085)	0.069 (0.039)	0.081 (0.081)	12
	TKN	Bulk	1.43 (2.00)	1.91 ([illegible])	1.83 ([illegible])	[illegible]

Year	Par	Cond	L86-1	L86-2	S86	M86-1	M86-2	n
1986	TP	Bulk	0.115 (0.083)	0.133 (0.100)	0.119 ([illegible])	[illegible]	[illegible]	
		Wet	0.086 (0.047)	[illegible] (0.040)	0.081 (0.028)	[illegible]	[illegible]	[illegible]
		Dry	0.137 ([illegible])	[illegible]	0.183 ([illegible])	[illegible]	[illegible]	[illegible]
	TDP	Bulk	0.103 ([illegible])	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]
		Wet	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]
		Dry	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]
	NH4-N	Wet	[illegible]	[illegible]	0.80 ([illegible])	[illegible]	[illegible]	[illegible]
	NO2-N+NO3-N	Wet	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]

Table 7.3 Results of two-way ANOVA (F) of nutrient loads between sites that were sampled after common time periods. Separate analyses were done for each year of the study (1984-1986), and time period is treated as a block in the analysis.

Year	Parameter	Condition	df	F	P
1984	TDP	Dry	2,14	0.20	> 0.75
		Wet	2,8	0.50	> 0.50
		Bulk	2,24	0.33	> 0.50
	TP	Dry	2,14	0.14	> 0.75
		Wet	2,8	2.50	> 0.10
		Bulk	2,24	1.57	> 0.25
1985	TDP	Dry	2,22	0.15	> 0.75
		Wet	2,18	0.33	> 0.75
		Bulk	2,42	0.45	> 0.50
	TP	Dry	2,22	0.12	> 0.75
		Wet	2,18	0.30	> 0.75
		Bulk	2,42	0.05	> 0.75
	TKN	Bulk	2,24	0.20	> 0.75
1986	TDP	Dry	4,12	0.47	> 0.75
		Wet	4,8	2.02	> 0.10
		Bulk	4,24	0.62	> 0.50
	TP	Dry	4,12	0.26	> 0.75
		Wet	4,8	1.24	> 0.25
		Bulk	4,24	0.50	> 0.50
	NH_4-N	Wet	4,32	2.36	> 0.05
	NO_2-N+NO_3-N	Wet	4,32	1.36	> 0.25

Table 7.4 **Summer and annual precipitation (mm), and monthly (May-October), open-water, ice-on and annual atmospheric loads (mg.m^{-2}) of TP and TN in central Alberta from 1983 to 1986.**

	Time Period	Year				Mean	SE
		1983	1984	1985	1986		
Precip	May-Aug.	416	368	152	309	311	57
	Annual	593	659	360	499	528	65
TP	May	-	7.1	6.5	8.7	7.4	0.7
	June	-	3.3	3.5	6.1	4.3	0.9
	July	-	4.0	3.1	3.2	3.4	0.3
	August	-	2.6	2.7	3.6	3.0	0.3
	September	0.8	1.4	1.1*	1.1*	1.1	0.3
	October	0.2	0.2*	0.2*	0.2*	0.22	-
	Open-Water	-	18.7	17.1	22.9	19.6	1.8
	Ice-on	-	0.8*	0.8*	0.8	0.8	-
	Annual	-	19.5	17.8	23.7	20.3	1.8
TN	May	-	-	58	83	70	12
	June	-	-	61	141	101	10
	July	-	-	30	181	105	76
	August	-	-	21	86	54	32
	Sept.-Oct.	-	-	49**	49**	49	-
	Open-water	-	-	218	534	379	160
	Ice-on	-	-	45*	45	45	
	Annual	-	-	264	584	424	160

*Average load measured during that time period in other years of the study.

**Average load measured during those months at other lakes in Alberta (Pollution Control Division, Alberta Environment, unpublished data).

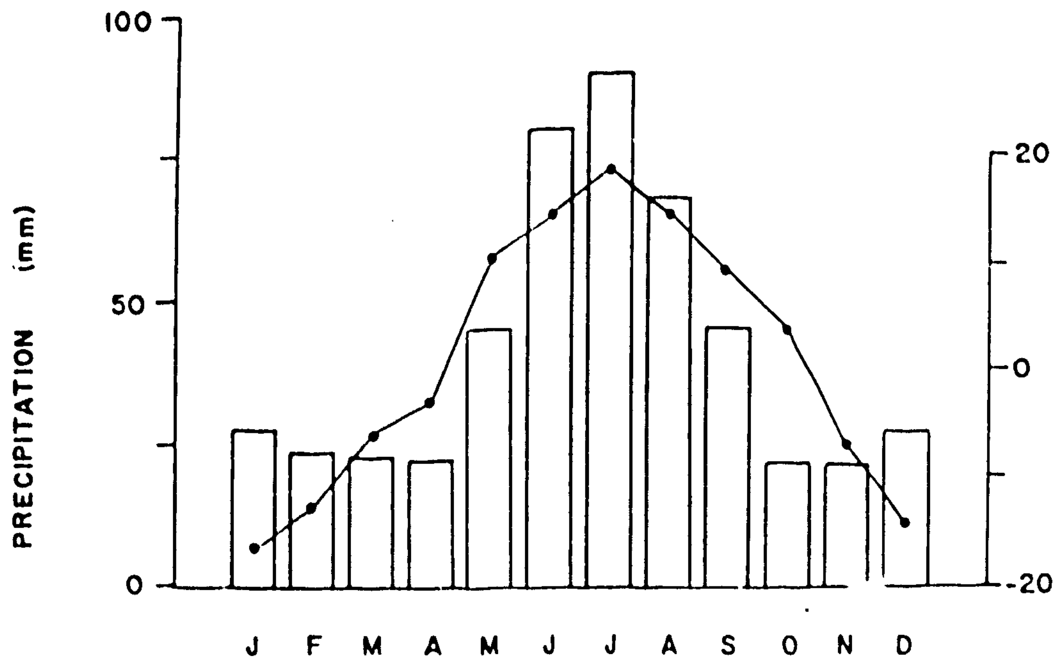
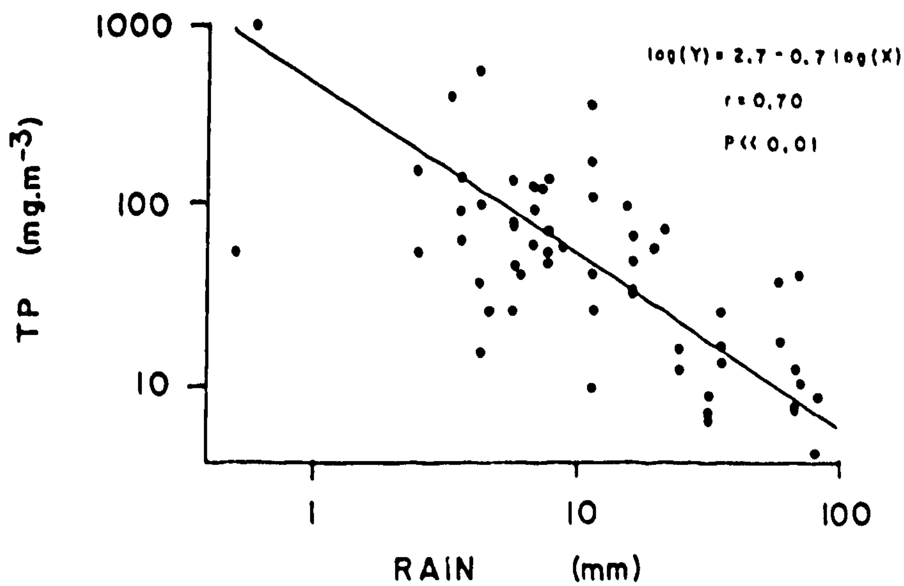


Figure 7.1. Average monthly precipitation and mean daily temperatures (1951-1980) at Athabasca, Alberta, 20 km northeast of Narrow Lake (Environment Canada, 1982).



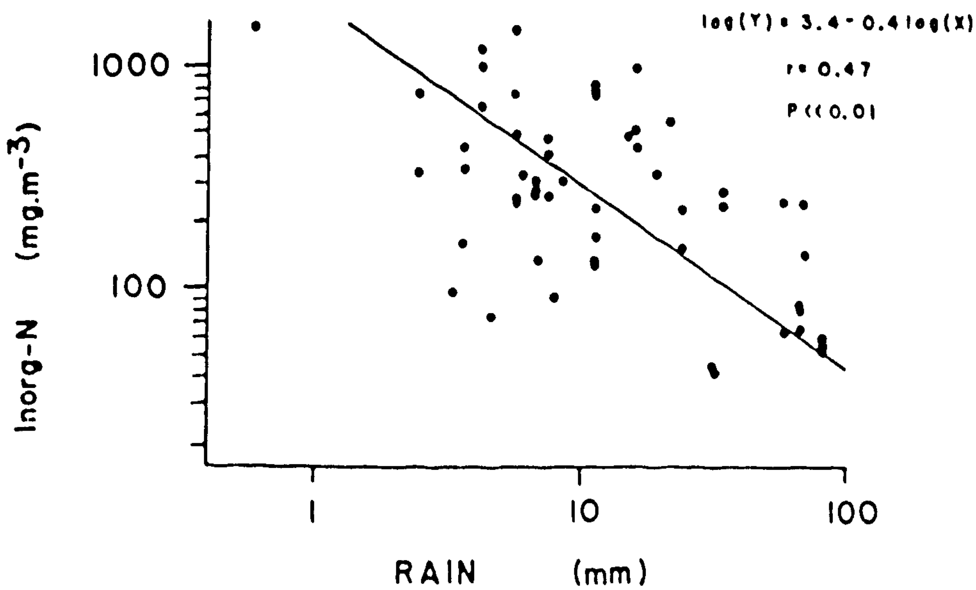


Figure 7.2. Total P and Inorg-N concentrations (mg.m^{-3}) vs rainfall (mm) from May to August 1986.

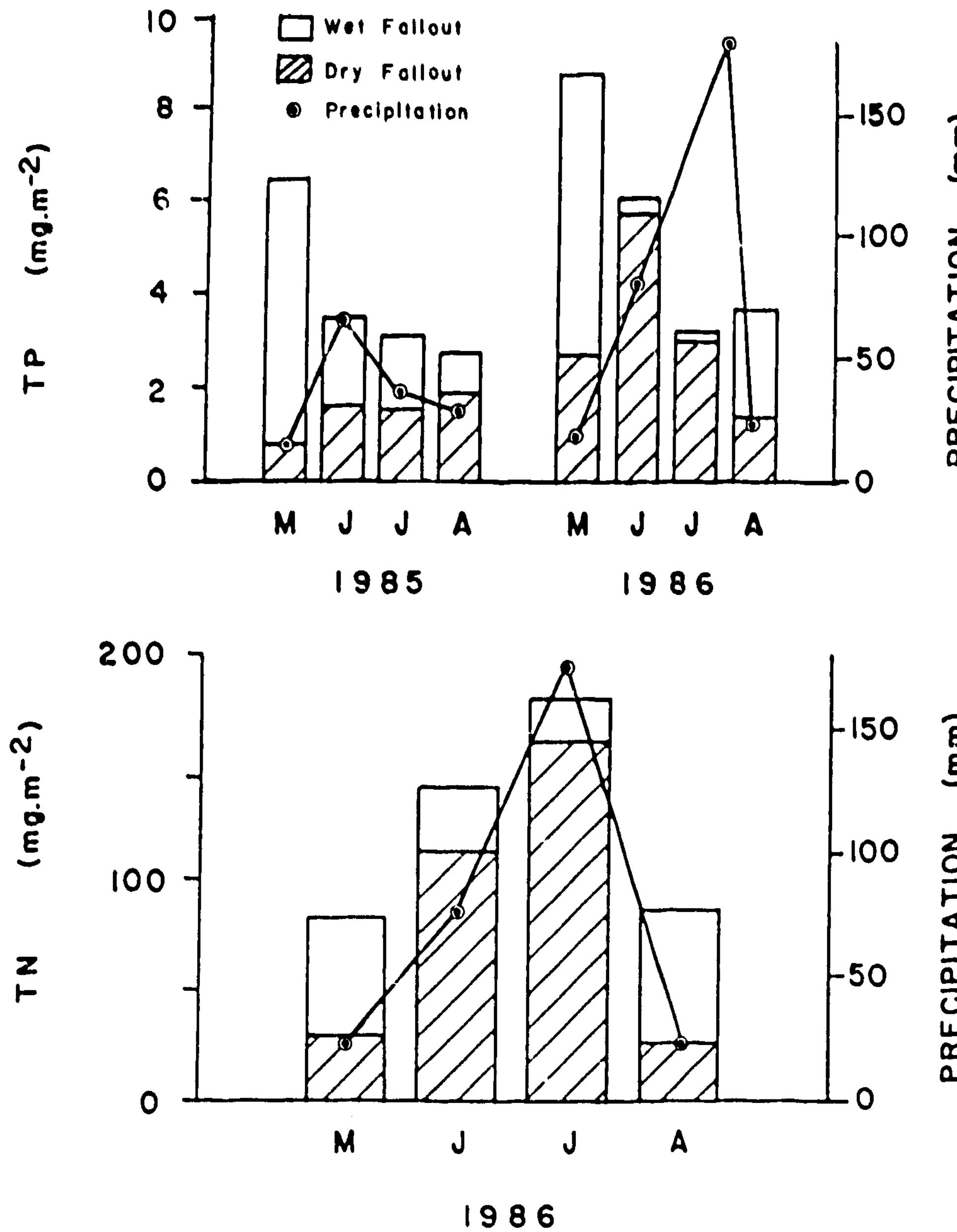
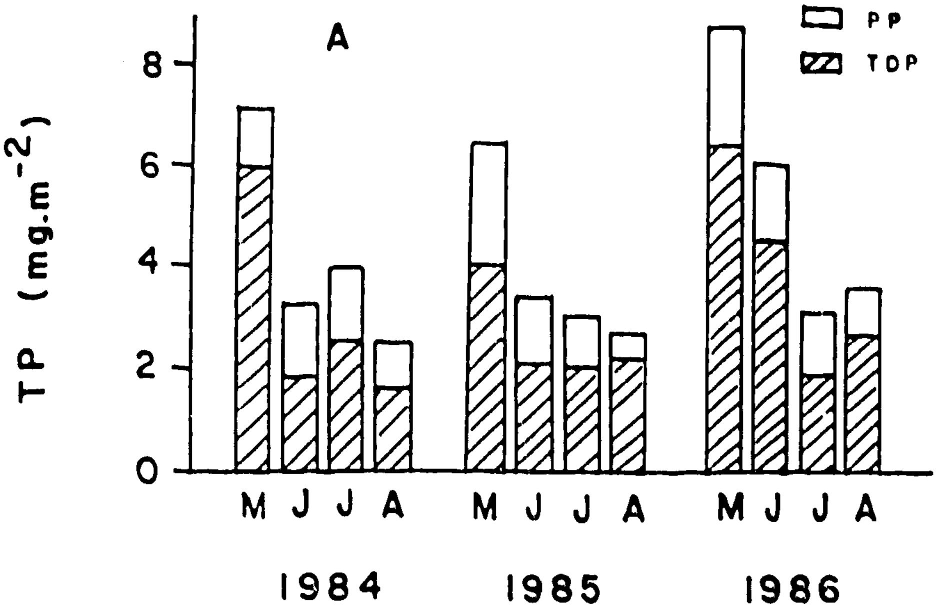


Figure 7.3. Total monthly precipitation and contribution of wet and dry fallout from May to August, 1985 and 1986 for TP and May to August 1986 for TN.



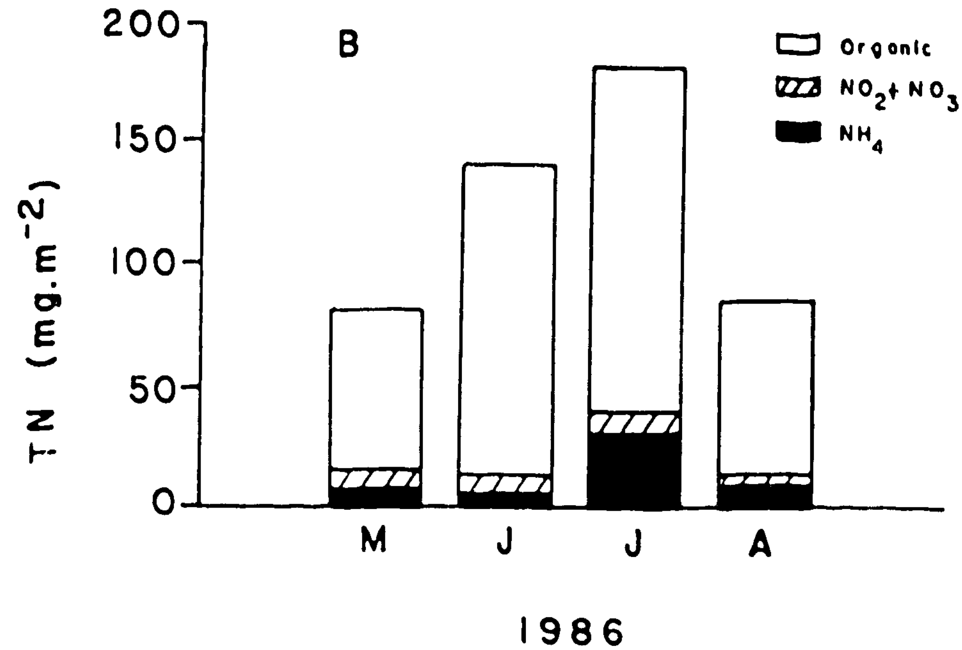


Figure 7.4. A: Contribution of total dissolved P (TDP) and particulate P (PP) to TP loads from May to August, 1984 to 1986. B: Contribution of inorg-N and org-N to TN loads from May to August 1986.

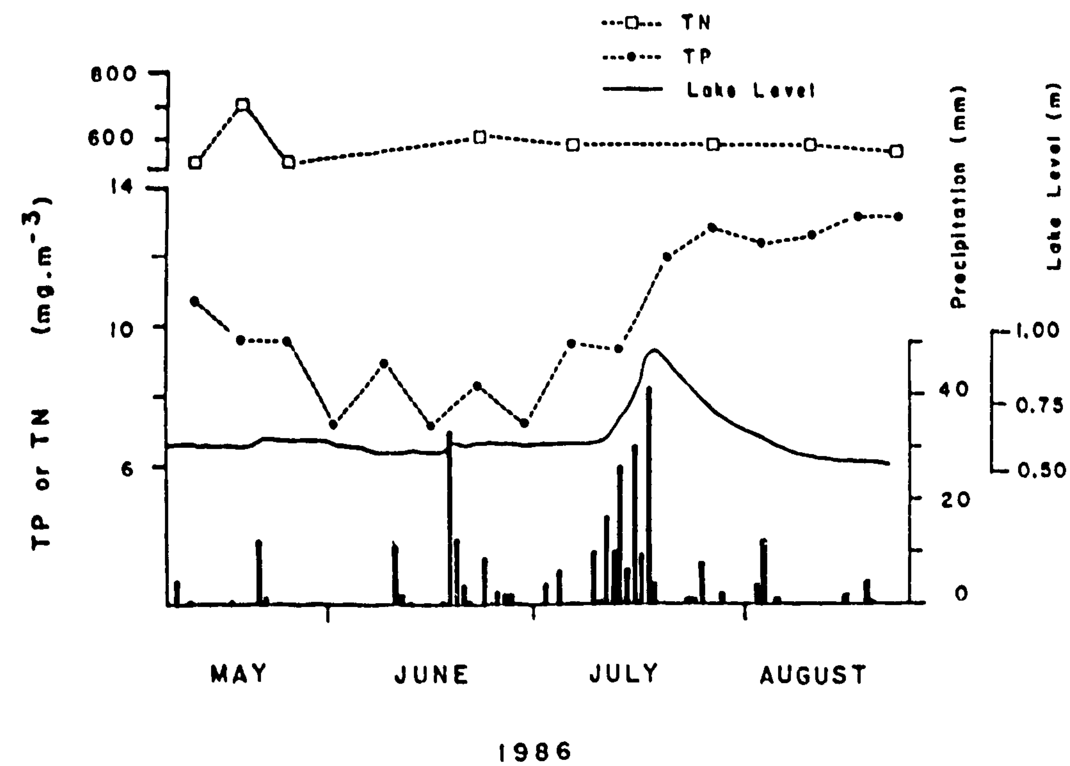


Figure 7.5. Daily precipitation (mm) at Athabasca (Environment Canada, 1986), Narrow Lake elevation (in m above 686.34 m), and TP and TN concentrations ($mg.m^{-3}$) in the epilimnion (0-6 m) in Narrow Lake from 8 May to 23 August 1986.

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8. GENERAL DISCUSSION

Groundwater inputs of water, nutrients and other materials may be important to lakes (e.g., Uttormark et al. 1974). However, the input of groundwater to lakes is difficult to quantify (Chapter 1). My research focussed on (1) the evaluation and comparison of methods to quantify seepage rates in lakes, (2) the distribution of seepage within lakes, and (3) the impact of groundwater input of water and phosphorus to lakes. This chapter discusses the contribution of my research in these three areas and concludes with the implications of this research to lake management.

Methods of Monitoring Groundwater-Lake Interactions

In general, an integrated approach would be desirable to investigate groundwater seepage conditions in a lake. Data from a drilling program, water chemistry, environmental isotopes and computer simulations can be useful to determine the position of a lake in the groundwater flow system and the distribution of seepage within a lake (Chapter 5). However, for most lakes these data are not available; the cost of initiating these programs may be prohibitively expensive. Furthermore, interpretation of such data may be difficult (with or without training in hydrogeology).

Knowledge of the position of a lake within the groundwater flow system is not essential if the goal is to estimate the magnitude of groundwater-lake flux of water and/or materials. A "quick-and-dirty" approach to evaluate the potential importance of groundwater is to evaluate the lake water balance (Chapter 5). In many cases, a water balance can be calculated from published data on precipitation, lake evaporation and surface runoff rates for the study area. However, these rates can be highly variable from lake to lake, and the residual (unmeasured components of the water balance, e.g., groundwater) is subject to large errors. Thus, one must use extreme caution when the residual of a water balance is used to estimate groundwater-lake flux.

In situ methods of measuring groundwater-lake flux offer another alternative. The concept of placing mini-piezometers directly into the lake bottom to measure hydraulic gradient and hydraulic conductivity is sound in theory, but not in practice (Chapter 5). Seepage meters were much more useful than mini-piezometers. However, seepage flux in the study lakes tended to be low relative to those measured with seepage meters at other lakes (Chapter 4). The low seepage flux created a problem for collecting data with seepage meters; there was an anomalous, short-term influx of water to the bag (Chapter 2). However, this problem was eliminated by filling bags with 1 L of water before the bags were attached to seepage meters.

During the course of my research, hundreds of seepage meter samples were obtained under many different conditions. Seepage meters were most easily used when they could be placed close to shore (0.5 to 2 m of water) so SCUBA was not required; SCUBA greatly increased the time required to sample the meters. It was best to place the meters away from areas where they could be tampered with by the public. Seepage meters worked best in sandy sediments. When the porosity of sediments was greater than about 70%, the meters would slowly sink into the sediment. In shallow eutrophic lakes, macrophyte beds, and at water depths greater than about 10 m, sediments were generally too soft for seepage meters. On average, about 25% of the bags attached to seepage meters would either be dislodged or punctured by the following day. This percentage greatly increased (1) due to storms which enhanced turbulence in shallow lake water (especially for those seepage meters in less than 1 m of water) and (2) when bags were left attached to seepage meters for more than a few days.

One should expect seepage flux measured by closely-spaced seepage meters to be highly variable (Chapter 3). Consistent differences in seepage flux measured over several days would indicate that the variability is caused by actual differences in seepage flux between the different seepage meter sites (Chapter 3). The accuracy of seepage meter data is difficult to evaluate, especially for conditions of low seepage flux. Laboratory tank tests have been carried out to investigate the accuracy of seepage meters (Erickson 1980). However, these tests were conducted at a flux much higher than that observed at most lakes in my study. It is very difficult to simulate low flow conditions in laboratory tanks because the water does not flow homogeneously throughout the tank. The alternative to tank test calibrations is to test seepage meters against other methods of measuring seepage flux (which are also subject to error). At three lakes, I compared seepage meters to other methods; results based on seepage meter data were comparable to other values (Chapter 4 and 5).

In summary, an ideal study would include several different methods to investigate groundwater-lake interactions. However, for most cases such an approach is not be feasible. The use of seepage meters offers an attractive alternative. These instruments are simple to construct and inexpensive (about $10 each). In addition, they can provide information on seepage distribution within a lake and quantify seepage rates to (or from) the lake. However, the limitations of these instruments must be recognized. They are labour intensive, measure seepage flux at a single point in space and in time within the lake, and must be corrected for the anomalous, short-term inflow of water.

Distribution of Seepage in Lakes

In the study area, till was the predominant surficial deposit; most other studies of groundwater-lake interactions have been conducted at lakes situated in deposits which are more permeable to groundwater movement (Chapter 4). In general, seepage flux to the study lakes followed a pattern observed at many other lakes: flux tended to be highest close to shore and decreased with distance from shore (Chapter 4). However, deviations from this pattern were also observed. Some of the observed deviations could have been caused by the random placement of seepage meters within the nearshore zone (Chapter 3). In addition, the geology near the lakes strongly affected the seepage distribution within the lakes. Intertill sand and gravel lenses near the lakes could cause the presence of offshore zones of high seepage rates (Chapter 5). Pre-glacial channels of sand and gravel underlying some of the lakes affected the seepage distribution; however, the effect was not consistent between lakes (Chapter 4). Without detailed information on hydrogeological conditions near a lake, it would be difficult to predict the seepage distribution to a lake.

Groundwater Component of Lake Water and Material Budgets

In general, groundwater was a relatively small component (15 %) of the total annual input of water to the study lakes (Chapter 4). This was expected because most of the lakes were situated in glacial till. However, there were exceptions. Groundwater contributed about one-third of the total inflow of water to Narrow Lake (Chapter 5), and about one-half of the total inflow of water to Spring Lake (Chapter 4).

Even though groundwater was not necessarily an important source of water, it may be an important source of phosphorus to most of the study lakes (Chapters 5 and 6). Rather than transporting significant amounts of P from the drainage basin to the lake, groundwater flushes dissolved P from the porewater in lake sediments into the overlying lake water. Thus, groundwater enhances P recycling from sediments to lake water. At most of the study lakes, phosphorus, flushed from the anaerobic lake sediments, would not be sorbed onto iron- or calcium-complexes in aerobic lakewater (Chapter 6; Shaw 1989). Thus, porewater phosphorus could enter the water column of a lake. High porewater phosphorus concentrations are maintained because of desorption of phosphorus from the particulate sediments to porewater (Shaw 1989). The amount of P in particulate sediments is much greater than in porewater and is maintained by sedimentation of particulate phosphorus from the lake water. Therefore, flushing of P from sediments to lake water by groundwater can be maintained at a relatively constant level throughout the year.

It is possible that for some lakes, the importance of seepage meters as a source of P (and other materials) to lakes has been underestimated because of sampling design. Water chemistry from wells near the lake (especially non-conservative materials) may not reflect that which enters the lake (Chapter 5). Thus, porewater from lake sediments should be sampled to obtain more accurate estimates of the chemical composition of groundwater entering lakes.

Implications to Lake Management

Groundwater seepage to lakes is generally concentrated near the lake shore. The nearshore zone of lakes is a habitat for much of the lake's flora and fauna. In addition, groundwater may be structuring biological communities; e.g., at one lake, the distribution of macrophytes was related to seepage flux (Lodge et al. 1989).

Seepage water may be a particularly important source of P (and other nutrients) to lake biota in the nearshore region of the lake. Groundwater can provide a relatively constant supply of P to lakes; whereas other sources of P, such as atmospheric deposition and surface runoff, can be much more seasonal (Chapters 5 and 7). Furthermore, phosphorus transported to the lake by groundwater would enter epilimnetic lake water, where it could be directly utilized by epilimnetic algae (Chapter 6). On the other hand, contaminants in groundwater will most strongly impact the nearshore zone of lakes. Agricultural activity and sewage disposal on land can cause elevated levels of contaminants in groundwater (e.g., N, P, trace metals). Increased groundwater loading of P may enhance lake eutrophication.

My research provides additional support to the growing body of evidence that suggests groundwater can be a significant source of water and/or materials to lakes. This is true not only for lakes in highly permeable materials, but also for lakes in central Alberta that are situated in glacial till. Lakes and groundwater should not be considered separate entities; management strategies that affect groundwater can impact surface water, and vice versa. Further investigations of the relationship between groundwater and lakes will undoubtedly enhance our understanding of chemical, physical and biological processes that occur within lakes.

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APPENDIX A

SEEPAGE METER DATA FOR NARROW AND BUFFALO LAKES (CHAPTER 2)

Table A.1. Seepage meter data from 1984 at Narrow Lake. Seepage flux is corrected (Corr.) for the average volume of water measured after a 30 min. sampling interval.

Seepage Meter	Date	Time	Interval (min)	Volume (mL)	Seepage Flux (m/s) Uncorr.	Corr.
6	31-Jul-84	14:50	1086	375	2.26E-08	1.21E-08
	02-Aug-84	20:30	3197	690	1.41E-08	1.05E-08
	08-Aug-84	12:20	8150	1370	1.10E-08	9.59E-09
	08-Aug-84	20:45	505	444	5.74E-08	3.64E-08
	09-Aug-84	08:40	715	286	2.61E-08	1.02E-08
	09-Aug-84	20:00	680	294	2.83E-08	1.16E-08
	10-Aug-84	19:15	1395	420	1.97E-08	1.15E-08
	11-Aug-84	07:15	720	278	2.52E-08	9.38E-09
	11-Aug-84	20:15	780	284	2.38E-08	9.14E-09
	12-Aug-84	21:00	1483	450	1.98E-08	1.22E-08
	13-Aug-84	10:30	811	370	2.98E-08	1.60E-08
	13-Aug-84	19:40	550	438	5.20E-08	3.25E-08
	15-Aug-84	09:55	2290	720	2.06E-08	1.56E-08
	15-Aug-84	14:05	251	300	7.83E-08	3.59E-08
	15-Aug-84	14:35	30	180	3.93E-07	-
	15-Aug-84	14:50	15	134	5.85E-07	-
	15-Aug-84	14:58	8	93	7.54E-07	-
	15-Aug-84	15:03	5	96	1.24E-06	-
	15-Aug-84	15:18	15	134	5.85E-07	-
	15-Aug-84	15:33	15	142	6.20E-07	-
	15-Aug-84	16:03	30	178	3.88E-07	-
	16-Aug-84	13:35	1290	455	2.31E-08	1.43E-08
	16-Aug-84	13:45	5	92	1.23E-06	-
	16-Aug-84	13:55	5	66	8.81E-07	-
7	02-Aug-84	20:30	3197	380	7.77E-09	5.68E-09
	08-Aug-84	20:45	505	295	3.81E-08	2.61E-08
	09-Aug-84	08:40	713	303	2.78E-08	1.90E-08
	09-Aug-84	19:55	675	280	2.71E-08	1.77E-08
	10-Aug-84	19:15	1400	340	1.59E-08	1.12E-08
	11-Aug-84	07:15	720	240	2.18E-08	1.28E-08
	11-Aug-84	20:15	780	325	2.72E-08	1.92E-08
	12-Aug-84	21:00	1483	396	1.75E-08	1.31E-08
	13-Aug-84	10:30	811	252	2.03E-08	1.23E-08
	13-Aug-84	19:40	550	294	3.49E-08	2.38E-08
	15-Aug-84	09:55	2290	350	9.99E-09	7.09E-09
	15-Aug-84	14:05	251	272	7.10E-08	4.95E-08
	15-Aug-84	14:35	30	108	2.36E-07	-
	15-Aug-84	14:50	15	88	3.84E-07	-
	15-Aug-84	14:58	8	54	4.38E-07	-
	15-Aug-84	15:03	5	56	7.26E-07	-
	15-Aug-84	15:18	15	88	3.84E-07	-
	15-Aug-84	15:33	15	95	4.15E-07	-
	15-Aug-84	16:03	30	102	2.23E-07	-
	16-Aug-84	13:35	1290	373	1.89E-08	1.39E-08
	16-Aug-84	13:45	5	68	9.08E-07	-
	16-Aug-84	13:55	5	55	7.34E-07	-
	16-Aug-84	14:05	5	53	7.08E-07	-
8	31-Jul-84	14:50	1086	295	1.78E-08	1.51E-08
	02-Aug-84	20:30	3197	186	3.80E-09	2.80E-09
	08-Aug-84	20:45	505	94	1.22E-08	5.98E-09
	09-Aug-84	08:40	716	86	7.85E-09	3.38E-09
	09-Aug-84	20:00	680	178	1.71E-08	1.28E-08
	10-Aug-84	19:15	1395	184	8.62E-09	6.39E-09
	11-Aug-84	07:15	720	168	1.53E-08	1.11E-08
	11-Aug-84	20:15	780	170	1.42E-08	1.04E-08
	12-Aug-84	21:00	1483	160	7.05E-09	4.92E-09

Table A.1. Continued.

Seepage Meter	Date	Time	Interval (min)	Volume (mL)	Seepage Flux (m/s) Uncorr.	Corr.
8	13-Aug-84	10:30	811	132	1.06E-08	6.82E-09
	13-Aug-84	19:40	550	166	1.97E-08	1.45E-08
	15-Aug-84	14:35	30	48	1.05E-07	-
	15-Aug-84	14:50	15	52	2.27E-07	-
	15-Aug-84	14:58	8	36	2.92E-07	-
	15-Aug-84	15:18	15	44	1.92E-07	-
	15-Aug-84	15:33	15	44	1.92E-07	-
	15-Aug-84	16:03	30	53	1.16E-07	-
	16-Aug-84	13:35	1290	152	7.70E-09	5.27E-09
9	31-Jul-84	14:50	1086	400	2.41E-08	1.79E-08
	02-Aug-84	20:30	3197	450	9.20E-09	7.01E-09
	08-Aug-84	12:20	8150	460	3.69E-09	2.81E-09
	08-Aug-84	20:45	505	260	3.36E-08	2.06E-08
	09-Aug-84	08:35	710	160	1.47E-08	4.76E-09
	09-Aug-84	19:55	680	246	2.37E-08	1.36E-08
	10-Aug-84	19:15	1400	310	1.45E-08	9.52E-09
	11-Aug-84	07:15	720	330	3.00E-08	2.08E-08
	11-Aug-84	20:15	780	475	3.98E-08	3.17E-08
	12-Aug-84	21:00	1483	248	1.09E-08	6.18E-09
	13-Aug-84	10:30	811	255	2.06E-08	1.21E-08
	13-Aug-84	19:40	550	385	4.57E-08	3.45E-08
	15-Aug-84	09:55	2290	960	2.74E-08	2.46E-08
	15-Aug-84	14:05	251	196	5.11E-08	2.53E-08
	15-Aug-84	14:35	30	94	2.05E-07	-
	15-Aug-84	14:50	15	114	4.98E-07	-
	15-Aug-84	14:58	8	67	5.43E-07	-
	15-Aug-84	15:03	5	76	9.86E-07	-
	15-Aug-84	15:18	15	88	3.84E-07	-
	15-Aug-84	15:33	15	112	4.89E-07	-
	15-Aug-84	16:03	30	127	2.77E-07	-
	16-Aug-84	13:35	1290	275	1.39E-08	8.53E-09
10	31-Jul-84	14:50	1086	595	3.58E-08	2.66E-08
	02-Aug-84	20:30	3197	610	1.25E-08	9.17E-09
	08-Aug-84	12:20	8150	1320	1.06E-08	9.29E-09
	08-Aug-84	20:45	505	476	6.16E-08	4.27E-08
	09-Aug-84	08:35	710	270	2.49E-08	1.00E-08
	09-Aug-84	19:55	680	300	2.88E-08	1.35E-08
	11-Aug-84	07:15	720	320	2.90E-08	1.46E-08
	11-Aug-84	20:15	780	350	2.93E-08	1.61E-08
	12-Aug-84	21:00	1483	278	1.23E-08	5.06E-09
	13-Aug-84	10:30	811	408	3.29E-08	2.03E-08
	13-Aug-84	19:40	550	518	6.15E-08	4.43E-08
	15-Aug-84	09:55	2290	960	2.74E-08	2.30E-08
	15-Aug-84	14:05	251	308	8.03E-08	4.22E-08
	15-Aug-84	14:35	30	149	3.25E-07	-
	15-Aug-84	14:50	15	130	5.67E-07	-
	15-Aug-84	14:58	8	107	8.67E-07	-
	15-Aug-84	15:03	5	107	1.39E-06	-
	15-Aug-84	15:18	15	140	6.11E-07	-
	15-Aug-84	15:33	15	129	5.63E-07	-
	15-Aug-84	16:03	30	182	3.97E-07	-
	16-Aug-84	13:35	1290	550	2.79E-08	1.99E-08
	16-Aug-84	13:45	5	112	1.50E-06	-
	16-Aug-84	13:55	5	100	1.33E-06	-
	16-Aug-84	14:05	5	98	1.31E-06	-

Table A.2. Seepage meter data from 1988 at Narrow Lake. Seepage flux is corrected (Corr.) for the average volume of water measured after a 45 min. sampling interval.

Seepage Meter	Date	Time	Interval (min)	Volume (mL)	Seepage Flux (m/s) Uncorr.	Corr.
1	10-May-88	16:10	5	58	7.59E-07	-
	11-May-88	15:50	1378	470	2.23E-08	1.56E-08
	11-May-88	18:00	124	196	1.03E-07	3.64E-08
	11-May-88	18:15	18	90	3.27E-07	-
	12-May-88	07:50	815	385	3.09E-08	1.98E-08
	12-May-88	15:15	400	280	4.58E-08	2.36E-08
	12-May-88	16:20	1	37	2.42E-06	-
	12-May-88	20:37	45	152	2.21E-07	-
	25-May-88	11:00	4380	600	8.95E-09	6.75E-09
2	10-May-88	16:15	5	35	4.58E-07	-
	11-May-88	16:00	1382	400	1.89E-08	1.01E-08
	11-May-88	18:00	119	180	9.89E-08	-
	11-May-88	18:20	18	71	2.58E-07	-
	12-May-88	07:50	814	455	3.65E-08	2.22E-08
	12-May-88	15:15	400	330	5.39E-08	2.50E-08
	12-May-88	16:30	1	49	3.20E-06	-
	12-May-88	20:38	45	194	2.82E-07	-
	25-May-88	11:00	4380	400	5.97E-09	3.11E-09
	26-May-88	15:20	1690	283	1.09E-08	3.54E-09
3	10-May-88	16:15	5	40	5.23E-07	-
	11-May-88	16:00	1383	150	7.09E-09	5.18E-09
	11-May-88	18:00	121	69	3.73E-08	2.15E-08
	11-May-88	18:20	18	30	1.09E-07	-
	12-May-88	07:55	814	152	1.22E-08	9.18E-09
	12-May-88	15:20	400	74	1.21E-08	5.52E-09
	12-May-88	16:30	1	9	5.89E-07	-
	12-May-88	20:40	45	44	6.39E-08	-
	25-May-88	11:00	4380	380	5.67E-09	5.07E-09
4	10-May-88	16:15	5	34	4.45E-07	-
	11-May-88	16:05	1388	290	1.37E-08	8.03E-09
	11-May-88	18:05	111	150	8.83E-08	2.48E-08
	11-May-88	18:25	18	64	2.32E-07	-
	12-May-88	08:00	815	270	2.17E-08	1.23E-08
	12-May-88	15:25	399	178	2.92E-08	9.79E-09
	12-May-88	16:30	1	18	1.18E-06	-
	12-May-88	20:40	45	125	1.82E-07	-
	25-May-88	11:00	4380	440	6.57E-09	4.75E-09
	26-May-88	15:20	1693	386	1.49E-08	1.04E-08
5	10-May-88	16:20	5	62	8.11E-07	-
	11-May-88	16:10	1389	495	2.33E-08	1.68E-08
	11-May-88	18:05	111	213	1.25E-07	6.15E-08
	11-May-88	18:25	18	78	2.83E-07	-
	12-May-88	08:00	816	345	2.76E-08	1.65E-08
	12-May-88	15:25	398	320	5.26E-08	3.14E-08
	12-May-88	16:35	1	43	2.81E-06	-
	12-May-88	20:45	44	150	2.23E-07	-
	25-May-88	11:00	4380	540	8.06E-09	5.88E-09

Table A.2. Continued.

Seepage Meter	Date	Time	Interval (min)	Volume (mL)	Seepage Flux (m/s) Uncorr.	Corr.
6	10-May-88	16:30	5	52	6.80E-07	-
	11-May-88	16:10	1390	470	2.21E-08	8.74E-09
	11-May-88	18:10	111	350	2.06E-07	5.85E-08
	11-May-88	18:25	18	200	7.26E-07	-
	12-May-88	08:00	816	790	6.33E-08	4.23E-08
	12-May-88	15:25	398	420	6.90E-08	2.40E-08
	12-May-88	16:40	1	86	5.62E-06	-
	12-May-88	20:46	44	290	4.31E-07	-
	25-May-88	11:00	4380	460	6.86E-09	2.56E-09
	26-May-88	15:30	1696	330	1.27E-08	1.58E-09
7	10-May-88	16:25	5	114	1.49E-06	-
	11-May-88	16:10	1389	355	1.67E-08	8.70E-09
	11-May-88	18:10	112	250	1.46E-07	7.11E-08
	11-May-88	18:25	18	88	3.20E-07	-
	12-May-88	08:04	817	330	2.64E-08	1.30E-08
	12-May-88	15:25	398	320	5.26E-08	2.66E-08
	12-May-88	16:50	1	32	2.09E-06	-
	12-May-88	20:48	44	176	2.61E-07	-
	25-May-88	11:00	4380	550	8.21E-09	5.64E-09
8	10-May-88	16:30	5	57	7.46E-07	-
	11-May-88	16:20	1397	460	2.15E-08	1.30E-08
	11-May-88	18:10	103	225	1.43E-07	3.88E-08
	11-May-88	18:30	18	112	4.07E-07	-
	12-May-88	15:30	397	315	5.19E-08	2.31E-08
	12-May-88	16:50	1	43	2.81E-06	-
	12-May-88	20:51	44	190	2.82E-07	-
	25-May-88	11:05	4380	875	1.31E-08	1.03E-08
	26-May-88	15:40	1696	545	2.10E-08	1.40E-08
9	10-May-88	16:30	5	236	3.09E-06	-
	11-May-88	16:25	1399	950	4.44E-08	3.59E-08
	11-May-88	18:15	103	360	2.28E-07	1.68E-07
	11-May-88	18:30	18	250	9.08E-07	-
	12-May-88	08:10	818	560	4.47E-08	2.99E-08
	12-May-88	15:30	397	400	6.59E-08	3.58E-08
	12-May-88	17:00	1	28	1.83E-06	-
	12-May-88	20:52	43	206	3.13E-07	-
	25-May-88	11:05	4380	520	7.76E-09	4.73E-09
10	10-May-88	16:00	5	63	8.24E-07	-
	11-May-88	18:00	121	364	1.97E-07	1.69E-07
	11-May-88	18:15	19	214	7.36E-07	-
	12-May-88	15:15	401	525	8.56E-08	6.55E-08
	12-May-88	16:20	1	36	2.35E-06	-
	12-May-88	20:33	45	168	2.44E-07	-
	25-May-88	11:00	4380	600	8.95E-09	6.51E-09

Table A.3. Seepage meter data from 1988 at Narrow Lake. Bags were prefilled with 1000 mL of water before they were attached to the meter.

Seepage Meter	Date	Time	Interval (min)	Volume (mL)	Seepage Flux (m/s)
1	10-May-88	18:00	45	12	1.74E-08
	21-May-88	10:50	5	0	0.00E+00
	21-May-88	12:30	15	1	4.36E-09
	21-May-88	15:10	92	19	1.35E-08
	26-May-88	15:20	1685	143	5.55E-09
2	10-May-88	18:05	45	20	2.90E-08
	21-May-88	12:30	15	17	7.41E-08
	21-May-88	15:20	92	24	1.71E-08
	21-May-88	20:15	218	25	7.50E-09
	22-May-88	09:30	770	27	2.29E-09
3	10-May-88	18:10	45	21	3.05E-08
	21-May-88	11:05	5	2	2.61E-08
	21-May-88	13:09	15	30	1.31E-07
	21-May-88	20:20	218	36	1.08E-08
	22-May-88	09:30	770	21	1.78E-09
	26-May-88	15:20	1685	69	2.68E-09
4	21-May-88	11:20	5	5	6.54E-08
	21-May-88	13:00	15	9	3.92E-08
	21-May-88	15:30	92	22	1.56E-08
5	21-May-88	10:50	5	0	0.00E+00
	21-May-88	12:30	15	4	1.74E-08
	21-May-88	15:15	92	19	1.35E-08
	21-May-88	20:15	218	38	1.14E-08
	22-May-88	09:30	770	74	6.28E-09
6	21-May-88	11:20	5	4	5.23E-08
	21-May-88	13:05	15	18	7.84E-08
	21-May-88	15:30	92	26	1.85E-08
	22-May-88	09:30	770	27	2.29E-09
7	21-May-88	11:30	5	0	0.00E+00
	21-May-88	13:10	15	5	2.18E-08
	21-May-88	15:30	92	26	1.85E-08
	21-May-88	20:20	218	31	9.29E-09
8	21-May-88	11:40	5	3	3.92E-08
	21-May-88	13:30	15	15	6.54E-08
	21-May-88	15:30	92	26	1.85E-08
	22-May-88	09:35	770	44	3.73E-09
9	21-May-88	11:50	5	1	1.31E-08
	21-May-88	13:30	15	5	2.18E-08
	21-May-88	15:30	92	22	1.56E-08
	21-May-88	20:30	218	35	1.05E-08
	22-May-88	09:35	770	95	8.06E-09
	26-May-88	15:40	1685	170	6.59E-09
10	10-May-88	15:50	45	32	4.65E-08
	21-May-88	10:50	5	4	5.23E-08
	21-May-88	12:25	15	3	1.31E-08
	21-May-88	15:10	92	12	8.53E-09
	21-May-88	20:15	218	17	5.10E-09
	22-May-88	09:30	770	24	2.04E-09

Table A 4. Seepage meter data from 1987 at Buffalo Lake.

Seepage Meter	ate	Time	Interval (min)	Volume (mL)	Seepage Flux (m/s)
13	28-Aug-87	12:35	61	285	3.05E-07
	28-Aug-87	13:05	28	130	3.03E-07
	28-Aug-87	13:25	16	80	3.27E-07
	28-Aun-87	18:45	318	805	1.65E-07
	28-Aug-87	19:05	16	100	4.08E-07
17	28-Aug-87	12:35	59	350	3.88E-07
	28-Aug-87	13:05	27	110	2.66E-07
	28-Aug-87	13:25	16	120	4.90E-07
	28-Aug-87	19:05	16	120	4.90E-07
19	28-Aug-87	12:35	59	310	3.43E-07
	28-Aug-87	13:05	27	130	3.15E-07
	28-Aug-87	13:25	16	145	5.92E-07
	28-Aug-87	18:45	319	810	1.66E-07
	28-Aug-87	19:05	16	150	6.13E-07

APPENDIX B

SEEPAGE METER DATA FOR SITES 1 TO 4, AT NARROW LAKE (CHAPTER 3)

Table B.1. Seepage meter data from the four sites sampled during 1984 at Narrow Lake. Seepage flux for sites 1 and 2 were corrected (Corr.) for the volume of water measured after a 30 min sampling interval. Seepage flux for sites 3 and 4 were corrected for the average volume of water measured after 30 min at seepage meters 1 to 10 (mean 96 mL)

Site	Seepage Meter	Date	Time	Interval (min)	Volume (mL)	Seepage Flux (m/s) Uncorr.	Corr.
1	6	15-Aug-84	14:35	30	180	3.93E-07	-
	7	15-Aug-84	14:35	30	108	2.36E-07	-
	8	15-Aug-84	14:35	30	48	1.05E-07	-
	9	15-Aug-84	14:35	30	94	2.05E-07	-
	10	15-Aug-84	14:35	30	149	3.25E-07	-
	6	15-Aug-84	16:03	30	178	3.88E-07	-
	7	15-Aug-84	16:03	30	102	2.23E-07	-
	8	15-Aug-84	16:03	30	53	1.16E-07	-
	9	15-Aug-84	16:03	30	127	2.77E-07	-
	10	15-Aug-84	16:03	30	182	3.97E-07	-
	6	09-Aug-84	08:40	715	286	2.61E-08	1.02E-08
	7	09-Aug-84	08:40	713	303	2.77E-08	1.90E-08
	8	09-Aug-84	08:40	716	86	7.86E-09	3.38E-09
	9	09-Aug-84	08:35	713	160	1.47E-08	4.74E-09
	10	09-Aug-84	08:35	710	270	2.49E-08	1.00E-08
	6	09-Aug-84	20:00	680	294	2.83E-08	1.16E-08
	7	09-Aug-84	19:55	675	280	2.71E-08	1.77E-08
	8	09-Aug-84	20:00	680	178	1.71E-08	1.28E-08
	9	09-Aug-84	19:55	680	246	2.36E-08	1.36E-08
	10	09-Aug-84	19:55	680	300	2.88E-08	1.35E-08
	6	11-Aug-84	07:15	720	278	2.52E-08	9.38E-09
	7	11-Aug-84	07:15	720	240	2.18E-08	1.28E-08
	8	11-Aug-84	07:15	720	168	1.53E-08	1.11E-08
	9	11-Aug-84	07:15	720	330	3.00E-08	2.08E-08
	10	11-Aug-84	07:15	720	320	2.91E-08	1.46E-08
	6	11-Aug-84	20:15	780	284	2.38E-08	9.14E-09
	7	11-Aug-84	20:15	780	325	2.72E-08	1.92E-08
	8	11-Aug-84	20:15	780	170	1.43E-08	1.04E-08
	9	11-Aug-84	20:15	780	475	3.98E-08	3.17E-08
	10	11-Aug-84	20:15	780	350	2.93E-08	1.61E-08
	6	13-Aug-84	10:30	811	370	2.99E-08	1.60E-08
	7	13-Aug-84	10:30	811	252	2.03E-08	1.23E-08
	8	13-Aug-84	10:30	811	132	1.07E-08	6.82E-09
	9	13-Aug-84	10:30	811	255	2.06E-08	1.21E-08
	10	13-Aug-84	10:30	811	408	3.29E-08	2.03E-08

Table B.1. Continued.

Site	Seepage Meter	Date	Time	Interval (min)	Volume (mL)	Seepage Flux (m/s) Uncorr.	Corr.
2	1	15-Aug-84	16:15	30	70	1.53E-07	-
	2	15-Aug-84	16:15	30	78	1.70E-07	-
	3	15-Aug-84	16:15	30	56	1.22E-07	-
	4	15-Aug-84	16:15	30	74	1.61E-07	-
	5	15-Aug-84	16:15	30	70	1.53E-07	-
	1	09-Aug-84	08:45	705	308	2.86E-08	2.30E-08
	2	09-Aug-84	08:50	706	260	2.41E-08	1.76E-08
	3	09-Aug-84	08:55	705	340	3.15E-08	2.75E-08
	4	09-Aug-84	08:50	710	270	2.49E-08	1.88E-08
	2	09-Aug-84	20:10	680	266	2.56E-08	1.89E-08
	3	09-Aug-84	20:15	680	272	6.92E-09	2.17E-08
	4	09-Aug-84	20:05	675	288	2.79E-08	2.17E-08
	5	09-Aug-84	20:15	685	125	1.19E-08	5.48E-09
	1	11-Aug-84	07:20	720	380	4.36E-08	2.94E-08
	2	11-Aug-84	07:20	714	220	2.01E-08	1.36E-08
	3	11-Aug-84	07:20	714	198	1.82E-08	1.36E-08
	4	11-Aug-84	07:20	715	380	3.47E-08	2.92E-08
	5	11-Aug-84	07:20	710	172	9.39E-09	9.80E-09
	1	11-Aug-84	20:30	790	375	3.10E-08	2.62E-08
	2	11-Aug-84	20:30	790	260	2.15E-08	1.57E-08
	3	11-Aug-84	20:45	805	193	9.09E-09	1.16E-08
	4	11-Aug-84	20:30	790	325	2.69E-08	2.16E-08
	5	11-Aug-84	20:30	790	132	9.27E-09	5.33E-09
	2	13-Aug-84	09:10	763	164	1.40E-08	7.67E-09
	3	13-Aug-84	09:10	760	204	8.94E-09	1.33E-08
	4	13-Aug-84	09:10	770	238	2.02E-08	1.45E-08
	5	13-Aug-84	09:10	760	228	1.96E-08	1.41E-08
	2	13-Aug-84	19:50	639	200	2.04E-08	1.31E-08
	3	13-Aug-84	19:50	639	204	1.06E-08	1.59E-08
	4	13-Aug-84	19:50	639	240	2.45E-08	1.78E-08
	5	13-Aug-84	19:50	639	186	1.90E-08	1.24E-08

Table B.1. Continued.

Site	Seepage Meter	Date	Time	Interval (min)	Volume (mL)	Seepage Flux (m/s) Uncorr.	Corr.
3	14	09-Aug-84	20:20	660	337	3.34E-08	2.50E-08
	15	09-Aug-84	20:20	660	330	3.27E-08	2.43E-08
	16	09-Aug-84	20:20	660	760	7.53E-08	6.89E-08
	14	11-Aug-84	21:00	810	455	3.67E-08	3.01E-08
	15	11-Aug-84	21:00	810	410	3.31E-08	2.63E-08
	16	11-Aug-84	21:00	810	620	5.00E-08	4.39E-08
	14	13-Aug-84	09:20	755	425	3.68E-08	2.97E-08
	15	13-Aug-84	09:20	755	435	3.77E-08	3.06E-08
	16	13-Aug-84	09:20	755	600	5.19E-08	4.54E-08
	14	13-Aug-84	19:55	635	410	4.22E-08	3.39E-08
	15	13-Aug-84	19:55	635	440	4.53E-08	3.72E-08
	16	13-Aug-84	19:55	635	630	6.48E-08	5.77E-08
4	11	09-Aug-84	20:35	655	217	2.17E-08	1.27E-08
	12	09-Aug-84	20:35	655	419	4.18E-08	3.38E-08
	13	09-Aug-84	20:35	655	515	5.14E-08	4.38E-08
	11	11-Aug-84	07:45	720	240	2.18E-08	1.36E-08
	12	11-Aug-84	07:45	720	342	3.10E-08	2.33E-08
	13	11-Aug-84	07:45	720	420	3.81E-08	3.07E-08
	11	11-Aug-84	21:15	810	315	2.54E-08	1.84E-08
	12	11-Aug-84	21:15	810	380	3.07E-08	2.38E-08
	13	11-Aug-84	21:15	810	555	4.48E-08	3.85E-08
	11	13-Aug-84	09:25	805	340	2.76E-08	2.06E-08
	12	13-Aug-84	09:25	805	313	2.54E-08	1.83E-08
	13	13-Aug-84	09:25	805	596	4.84E-08	4.22E-08
	11	13-Aug-84	20:00	635	278	2.86E-08	1.97E-08
	12	13-Aug-84	20:00	635	304	3.13E-08	2.25E-08
	13	13-Aug-84	20:00	635	506	5.21E-08	4.43E-08

APPENDIX C

SEEPAGE METER DATA FOR NARROW-EAST, NARROW-WEST AND THE NINE OTHER SURVEY LAKES (CHAPTER 4)

EXAMPLE CALCULATION OF JACKKNIFE METHOD

Table C.1. Seepage meter data from 1986 for Narrow-East. Seepage flux is corrected for the average volume of water measured after a 59 min sampling interval on 10 July, 1986. The distance from shore (DFS) and lake depth (z) of the seepage meters are indicated.

Date	Seepage Meter	DFS m	z m	Interval min	Volume mL	Flux m/s
10-Jul-86	1	1	0.5	59	130	-
	2	5	0.6	59	200	-
	3	8	0.9	59	60	-
	4	10	1.2	59	171	-
	5	15	1.5	59	244	-
	6	20	3	59	133	-
	7	25	4.6	59	204	-
	8	30	5.5	59	90	-
	9	35	6.4	59	80	-
	10	40	7.3	59	73	-
26-May-86	1	1	0.5	1418	425	2.13E-08
	2	5	0.6	1419	485	2.05E-08
	3	8	0.9	1419	245	1.33E-08
	4	10	1.2	1420	360	1.36E-08
	5	15	1.5	1414	790	3.95E-08
	6	20	3	1414	470	2.44E-08
	7	25	4.6	1414	1060	6.19E-08
	8	30	5.5	1414	410	2.32E-08
	9	35	6.4	1414	580	3.62E-08
	10	40	7.3	1413	290	1.57E-08
7-Jun-86	1	1	0.5	1810	490	2.02E-08
	2	5	0.6	-	-	-
	3	8	0.9	1809	365	1.71E-08
	4	10	1.2	1809	445	1.54E-08
	5	15	1.5	1809	550	1.71E-08
	6	20	3	1809	360	1.27E-08
	7	25	4.6	1809	705	2.81E-08
	8	30	5.5	1809	350	1.46E-08
	9	35	6.4	1809	325	1.37E-08
	10	40	7.3	1810	270	1.10E-08
23-Jun-86	1	1	0.5	1383	370	1.78E-08
	2	5	0.6	1383	485	2.11E-08
	3	8	0.9	1383	240	1.33E-08
	4	10	1.2	1383	465	2.18E-08
	5	15	1.5	1384	405	1.19E-08
	6	20	3	1383	575	3.27E-08
	7	25	4.6	1384	290	6.36E-09
	8	30	5.5	1385	335	1.81E-08
	9	35	6.4	1384	220	1.04E-08
	10	40	7.3	1385	215	1.05E-08

Table C.1. Continued.

Date	Seepage Meter	DFS m	z m	Interval min	Volume mL	Flux m/s
10-Jul-86						
	1	1	0.5	1294	320	1.51E-08
	2	5	0.6	1295	460	1.87E-08
	3	8	0.9	1294	200	1.11E-08
	4	10	1.2	-	-	-
	5	15	1.5	1294	205	0.00E+00
	6	20	3	1294	480	1.39E-08
	7	25	4.6	1294	440	2.66E-08
	8	30	5.5	1294	215	9.14E-09
	9	35	6.4	1294	195	7.95E-09
	10	40	7.3	1295	270	2.15E-08
18-Jul-86	1	1	0.5	1359	325	1.47E-08
	2	5	0.6	1359	445	1.85E-08
	3	8	0.9	-	-	-
	4	10	1.2	-	-	-
	5	15	1.5	-	-	-
	6	20	3	1361	500	2.23E-08
	7	25	4.6	1360	565	3.58E-08
	8	30	5.5	1361	240	1.21E-08
	9	35	6.4	1360	230	1.18E-08
	10	40	7.3	1360	200	1.51E-08
5-Aug-86	1	1	0.5	1430	340	1.50E-08
	2	5	0.6	1427	320	8.60E-09
	3	8	0.9	1428	230	1.22E-08
	4	10	1.2	1428	505	2.39E-08
	5	15	1.5	1428	305	4.37E-09
	6	20	3	1429	425	1.58E-08
	7	25	4.6	1428	590	3.58E-08
	8	30	5.5	1429	170	6.44E-09
	9	35	6.4	1429	225	1.09E-08
	10	40	7.3	1429	415	2.97E-08

Table C.2. Seepage meter data from 1986 for Narrow-West. Seepage flux is corrected for the average volume of water measured after a 59 min sampling interval on 10 July, 1986. The distance from shore (DFS) and lake depth (z) of the seepage meters are indicated.

Date	Seepage Meter	DFS m	z m	Interval min	Volume mL	Flux m/s
10-Jul-86	1	1	0.6	60	185	-
	2	5	0.9	60	115	-
	3	8	1.2	60	120	-
	4	10	2.1	60	225	-
	5	15	3.4	60	205	-
	6	20	4.6	60	70	-
	7	25	6.1	60	70	-
	8	30	7.6	60	75	-
	9	35	9.1	60	180	-
	10	40	10.7	60	138	-
26-May-86	1	1	0.6	1417	462	2.00E-08
	2	5	0.9	1418	467	2.55E-08
	3	8	1.2	1417	846	5.25E-08
	4	10	2.1	1419	607	2.76E-08
	5	15	3.4	1418	334	9.34E-09
	6	20	4.6	1417	185	8.34E-09
	7	25	6.1	1418	245	1.27E-08
	8	30	7.6	1418	240	1.20E-08
	9	35	9.1	1418	150	0.00E+00
	10	40	10.7	1417	100	0.00E+00
7-Jun-86	1	1	0.6	1818	510	1.81E-08
	2	5	0.9	1817	545	2.40E-08
	3	8	1.2	1817	475	1.98E-08
	4	10	2.1	1816	700	2.66E-08
	5	15	3.4	1812	505	1.68E-08
	6	20	4.6	1812	300	1.29E-08
	7	25	6.1	1811	370	1.68E-08
	8	30	7.6	1812	240	9.26E-09
	9	35	9.1	1812	280	5.61E-09
	10	40	10.7	1812	130	0.00E+00
23-Jun-86	1	1	0.6	1384	320	1.00E-08
	2	5	0.9	1384	280	1.22E-08
	3	8	1.2	1384	385	1.97E-08
	4	10	2.1	1385	455	1.71E-08
	5	15	3.4	1384	365	1.19E-08
	6	20	4.6	1384	195	9.29E-09
	7	25	6.1	1384	240	1.26E-08
	8	30	7.6	1383	240	1.23E-08
	9	35	9.1	1383	185	3.72E-10
	10	40	10.7	1383	140	1.50E-10

Table C.2. Continued.

Date	Seepage Meter	DFS m	z m	Interval min	Volume mL	Flux m/s
10-Jul-86	1	1	0.6	1273	430	1.98E-08
	2	5	0.9	-	-	-
	3	8	1.2	1280	195	6.04E-09
	4	10	2.1	1280	320	7.65E-09
	5	15	3.4	-	-	-
	6	20	4.6	1281	125	4.43E-09
	7	25	6.1	1281	140	5.64E-09
	8	30	7.6	1281	220	1.17E-08
	9	35	9.1	1281	140	0.00E+00
	10	40	10.7	-	-	-
18-Jul-86	1	1	0.6	1350	280	7.22E-09
	2	5	0.9	1352	215	7.61E-09
	3	8	1.2	1352	275	1.18E-08
	4	10	2.1	1353	350	9.50E-09
	5	15	3.4	-	-	-
	6	20	4.6	1353	165	7.23E-09
	7	25	6.1	1353	280	1.60E-08
	8	30	7.6	1353	205	9.89E-09
	9	35	9.1	1354	155	0.00E+00
	10	40	10.7	1353	375	1.82E-08
5-Aug-86	1	1	0.6	1430	435	1.79E-08
	2	5	0.9	1430	205	6.46E-09
	3	8	1.2	1429	235	8.25E-09
	4	10	2.1	1430	310	6.10E-09
	5	15	3.4	1430	325	8.61E-09
	6	20	4.6	1429	145	5.39E-09
	7	25	6.1	1429	210	1.01E-08
	8	30	7.6	1428	220	1.04E-08
	9	35	9.1	1429	210	2.16E-09
	10	40	10.7	1428	140	1.45E-10

Table C.3. Seepage meter data from 1986 at nine lakes in central Alberta. The calculated seepage flux was corrected for the volume of water collected after a short (1-H) sampling interval (min). At S-7 and Tucker lakes, bags were prefilled with 500 mL of water before they were attached to the meters; the values given here are corrected for the 500 mL.

Lake (Date)	DFS m	z m	Sampling Inter 1-H	Sampling Inter 1-D	Volume (mL) 1-H	Volume (mL) 1-D	Flux m/s
Baptiste- (10-May-8	10	0.5	55	1225	130	325	1.09E-08
	20	0.9	55	1225	190	198	4.47E-10
	30	1.05	60	1241	62	112	2.77E-09
	40	1.4	60	1241	164	525	2.00E-08
	50	2.1	60	1248	50	195	7.98E-09
	60	2.7	60	1253	95	820	3.97E-08
	70	3	60	1253	173	2340	1.19E-07
	80	3.4	60	1253	105	1490	7.59E-08
	90	3.4	60	1261	150	670	2.83E-08
	100	3.7	60	1261	250	3460	1.75E-07
Baptsite- (10-May-8	20	0.6	60	1266	240	695	2.47E-08
	25	0.6	58	1269	220	770	2.97E-08
	30	0.7	58	1270	70	430	1.94E-08
	35	0.8	55	1272	300	360	3.22E-09
	40	0.9	49	1274	90	480	2.08E-08
	45	1.1	59	1204	40	250	1.20E-08
	50	1.5	58	1202	35	480	2.54E-08
	55	1.7	56	1204	130	530	2.28E-08
	60	3.7	55	1204	170	800	3.58E-08
	65	5.2	60	1202	80	690	3.49E-08
Buffalo (2-Jul-86	20	0.2	58	1089	170	1610	9.13E-08
	30	0.2	58	1089	145	-	-
	40	0.3	58	1089	180	2570	1.52E-07
	50	0.4	58	1089	195	-	-
	60	0.4	58	1089	180	2125	1.23E-07
	70	0.5	56	1089	170	2160	1.26E-07
	80	0.7	56	1092	210	2510	1.45E-07
	90	0.7	56	1092	215	2300	1.32E-07
	100	0.9	58	1092	195	2720	1.60E-07
	110	1.1	56	1092	130	2410	1.44E-07
Island (5-Jun-86	2	0.9	82	1392	155	410	1.27E-08
	7	1.5	81	1392	225	320	4.73E-09
	12	2.1	84	1390	210	475	1.33E-08
	17	3	78	1390	150	345	9.72E-09
	22	4.6	78	1390	215	510	1.47E-08
	27	5.2	78	1392	75	300	1.12E-08
	32	5.8	78	1392	115	270	7.71E-09
	37	6.1	78	1392	100	215	5.72E-09
	42	6.7	78	1392	-	100	-
	47	7.3	78	1392	65	85	9.94E-10
Jenkins (9-Jun-86	5	0.9	97	1436	205	-	-
	10	1.5	96	1436	395	550	7.56E-09
	15	2.4	96	1435	200	-	-
	20	3.7	95	1434	295	1490	5.83E-08
	25	4.9	96	1434	155	620	2.27E-08
	30	6.1	96	1431	395	650	1.25E-08
	35	7.3	95	1430	-	-	-
	40	8.5	87	1431	110	405	1.43E-08
	45	9.4	96	1429	50	550	2.45E-08
	50	9.8	87	1430	-	-	-

Table C.3. Continued.

Lake (Date)	DFS m	z m	Sampling Inter 1-H	Sampling Inter 1-D	Volume (mL) 1-H	Volume (mL) 1-D	Flux m/s
Long (26-May-8	1	0.5	70	1335	340	835	2.56E-08
	1.5	0.6	69	1332	220	465	-
	2	0.8	70	1326	180	-	1.27E-08
	13	3.4	67	1327	-	415	-
	15	4	67	1332	-	450	-
	20	4	65	1325	30	220	9.86E-09
	20	4.6	64	1325	155	-	-
	25	5.2	65	1321	-	190	-
	30	5.5	65	1321	20	75	2.86E-09
Minnie (21-Jun-8	10	0.9		976	-	395	2.64E-08
	15	1.5		975	-	340	2.28E-08
	20	2.4		976	-	395	2.64E-08
	25	3.4		971	-	505	3.39E-08
	30	4.9		978	-	550	3.68E-08
	35	5.5		975	-	-	-
	40	6.7		975	-	130	8.73E-09
	45	6.7		975	-	70	4.69E-09
	50	7.6		975	-	170	1.14E-08
	55	10.1		975	-	175	1.17E-08
Spring (30-Jul-8	8	0.3	71	238	175	450	1.08E-07
	10	0.6	76	244	120	125	1.95E-09
	15	0.9	73	245	125	185	2.28E-08
	20	1	72	244	-	315	-
	25	1.2	82	238	90	-	-
	30	1.3	82	237	80	95	6.32E-09
	35	1.8	84	234	15	65	2.18E-08
	40	2.1	85	234	-	-	-
	45	2.4	75	234	-	-	-
	50	3.4	85	233	60	45	0.00E+00
S-7 (2-Jul-86	1	0.2	81	1171	65	200	8.09E-09
	3	0.9	79	1171	0	215	1.29E-08
	6	2.0	79	1172	40	-	-
	8	2.5	79	1171	320	315	0.00E+00
	10	3.5	75	1174	20	25	2.97E-10
	13	4.0	76	1174	30	50	1.19E-09
	15	3.8	78	1174	25	50	1.49E-09
	18	3.3	80	1174	50	-	-
	20	2.8	82	1174	45	140	5.69E-09
	23	2.2	82	1174	-	-	-
Tucker (20-Jun-8	20	1	49	1385	25	-	-
	30	2.1	49	1385	15	205	9.29E-09
	40	4.3	49	1387	0	115	5.62E-09
	50	4.3	48	1387	60	290	1.12E-08
	60	4.9	48	1387	10	70	2.93E-09
	70	4.9	48	1387	0	45	2.20E-09
	80	5.5	18	1387	5	-	-
	90	6.1	48	1387	-30	-60	-1.46E-09
	100	6.1	48	1387	-85	-90	-2.43E-10
	110	6.1	48	1387	-	-	-

Table C.4. Example of calculation of average seepage flux along a transect. A is the area (cm^2) under a plot of seepage flux vs distance from shore, based on the data set with each of the i seepage meter sites left out in turn, v_{-i} ($m.s^{-1}$) is the average seepage flux along the transect computed from A (see text for details), log v is the log-transformed psuedovalue computed as:

$$\log v = 10(-7.83) - 9\log(v_{-i})$$

where -7.83 ($\log m.s^{-1}$) is the average seepage flux based on the data set with all of the seepage meter sites included.

i	A	v_{-i}	$\log(v_{-i})$	log v
1	63.4	1.52×10^{-8}	-7.82	-7.93
2	66.6	1.60×10^{-8}	-7.80	-8.12
3	55.0	1.32×10^{-8}	-7.88	-7.37
4	66.1	1.59×10^{-8}	-7.80	-8.09
5	66.1	1.59×10^{-8}	-7.80	-8.09
6	62.8	1.51×10^{-8}	-7.82	-7.89
7	60.3	1.45×10^{-8}	-7.84	-7.73
8	58.7	1.41×10^{-8}	-7.85	-7.63
9	64.2	1.54×10^{-8}	-7.81	-7.98
10	61.8	1.48×10^{-8}	-7.83	-7.83
Mean				-7.87
SD				0.24

APPENDIX D

LITHOLOGICAL LOGS FROM TEST-HOLES NEAR NARROW LAKE

WATER CHEMISTRY DATA

PHOSPHORUS DATA

Test hole	Depth (m)	Texture and Lithology	Moisture Content*
N1-W	0.0-0.9	gravel fill	m
	0.9-1.8	muskeg	s
	1.8-2.7	grey-brown sandy clay some organics	vm
	2.7-4.5	grey coarse sand & gravel	s
	4.5-8.8	grey sandy clay till, firm, high plastic	vm
	5.1 m of slough at completion Completed: 5-cm PVC water-table well slotted 3.7-5.7 m		
N1-P1	0.0-8.8	same as N1-W	
	8.8-12.5	grey sandy till, a few pebbles, stiff, medium plastic	sm
	12.5-14.0	a/a, softer	sm
	14.0-15.8	a/a, a few medium-grained sand lenses	m
	0.75 m of slough inside casing prior to installation Completed: 5-cm PVC, 0.95-m piezometer tip at 15.05 m		
N1-P2	0-3.5	till, yellow-brown sandy clay, firm, some stones	
	3.5-12	till, grey sandy clay, firm, more silty inbedded coarse sand	
	12-15	sand and gravel, poorly sorted, well rounded to angular, mostly quartz and feldspar	
	Completed: 5-cm PVC, 1-m piezometer tip at 13.5 m		
N2-W	0.0-0.6	clay fill	m
	0.6-0.9	topsoil	m

* Moisture content: d(dry); sm(slightly moist); m(moist); vm(very moist); s(saturated).

Test hole	Depth (m)	Texture and Lithology	Moisture Content*
	1.5-4.5	a/a, darker, trace white deposits	m
	0.9-1.5	brown sandy clay till, pebbles medium, firm plastic	m
	4.5-5.0	brown medium-grained sand	s
	5.0-6.1	brown clay sand, soft, low plastic	s
	6.1-9.1	slate grey sandy clay till, pebbles stiff	m
	4.5 m of water and slough at completion Completed: 5-cm PVC watertable well slotted 3.4-6.4 m		
N2-P	0.0-6.1	same as N2-W	
	1.5 m of water & slough at completion Completed: 5-cm PVC, 0.75-m piezometer tip at 5.05 m		
N3-W	0-6	till, yellow-brown sandy clay, firm, some stones	
	6-10	gravel, very coarse	
	10-15	sand and gravel, poorly sorted, less coarse	
	Completed: 5-cm PVC water-table well slotted 3-6 m		
N3-P	0-15	same as N3-W	
	Completed: 5-cm PVC, 2-m piezometer tip at 15 m		
N4	0.0-2.4	brown sandy clay till, a few thin sand lenses (d), stiff, medium plastic	sm
	2.4-4.0	brown coarse sand, some clay	s
	4.0-4.5	gray sandy clay till, soft	vm
	4.5-5.8	grey coarse sand, trace of gravel	s
	5.8-13.7	slate grey sandy clay till, a few pebbles, stiff, medium plastic	sm - m
	13.7-15.2	grey coarse sand	s

Test hole	Depth (m)	Texture and Lithology	Moisture Content*
	15.2-16.5	grey fine sand, some clay & silt	vm
	Water & slough at 3 0 m at completion No well installed.		
N4-W	0.0-3.0	same as N4	
	Trace of water at completion Completed: 5-cm PVC water-table well, slotted 1.4-2.9 m		
N4-P1	0.0-7.6	same as N4	
	Water & slough at 2.0 m at completion Completed: 5-cm PVC, 0.45-m piezometer tip at 4.15 m		
N4-P2	0-4.9	lt. brown clay till, sandy and gravelly, thin sand layers	
	4.9-5.5	md. grey clay, sandy and gravelly	
	5.5-8.2	md. grey clay till, sandy and gravelly	
	8.2-11.6	as above, some silty sections	
	11.6-14.3	as above, more sandy and gravelly.	
	14.3-15.8	md. grey sand and gravel, fine- to medium-grains	
	15.8-17.7	md. grey clay, silty, sandy and gravelly	
	17.7-23.8	md. grey/brown clay, very silty and sticky	
	13.8-26.8	lt. brown sand, poor return, cuttings balling	
	26.8-30.4	lt. brown sand, bright feldspars: fine- to coarse-grained	
	30.4-36.9	md. grey siltstone, possible bedrock- soft and silty (almost lacustrinelike)	
	36.9-43.0	md. grey siltstone, very slippery and silty, increased pump pressure	
	43.0-60 4	lt. gr/brn/grn siltstone, slightly bentonitic, interbedded thin sandstone lenses	

Test hole	Depth (m)	Texture and Lithology	Moisture Content*
		Completed: 5-cm PVC, at 25 m	
N5-P	0.0-0.3	brown silty clay	m
	0.3-0.9	brown silty sand, medium-grained loose	m
	0.9-1.8	brown coarse sand & gravel, some silt & clay, dense	sm - m
	1.8-2.4	light brown medium sand	s
	2.4-3.4	grey-brown sandy clay till, stiff, medium plastic	m - vm
	3.4-8.5	a/a, very stiff	sm - m
	8.5-13.7	a/a, grey	sm - m
	13.7-14.3	grey, very coarse sand	s
	14.3-15.2	slate grey sandy clay till, stiff	m - vm
		Water & slough at 7.6 m at completion Completed: 5-cm PVC, 0.65-m piezometer tip at 12.3 m	
N5-W	0.0-4.5	same as N5-P	
		Dry at completion Completed: 5-cm PVC water-table well, slotted 1.2-4.2 m	
WT1	0.0-4.5	brown sandy clay till, pebbles, stiff, medium plastic	sm - m
	4.5-6.1	a/a, grey-brown	m
	6.1-6.3	brown medium-grained sand	s
	6.3-7.6	grey sandy clay till, a few thin sandy partings, stiff	m
		1.5 m of water & slough at completion Completed: 5-cm PVC water-table well, slotted 2.8-7.3 m	

Test hole	Depth (m)	Texture and Lithology	Moisture Content*
WT2	0.0-0.6	brown sandy clay till, a few thin sand lenses (sm), stiff medium plastic	sm
	0.6-1.3	a/a, grey-brown	sm
	1.3-4.0	a/a, a few thin fine sand lenses (s)	m
	4.0-5.2	a/a, stiffer, darker	sm - m
	5.2-7.6	a/a, slate grey, numerous thin silty fine sand lenses (s)	m
	Trace of water at completion Completed: 5-cm PVC water-table well, slotted 2.8-7.3 m		
WT3	0.0-1.5	brown sandy clay till, a few thin coarse sand lenses (sm), stiff, medium plastic	sm
	1.5-4.3	a/a, darker, numerous fine sand partings	sm
	4.3-4.9	a/a, a few thin medium sand lenses (s)	vm
	4.9-5.5	a/a, uniform	m
	5.5-7.6	a/a, slate grey, very stiff	m
	Trace of water at completion Completed: 5-cm PVC water-table well, slotted 4.3-7.3 m		
WT4	0.0-2.1	brown sand, very coarse, saturated at 0.6 m	s
	2.1-5.2	gravel & coarse sand	s
	5.2-5.6	grey sandy clay till (refusal on a boulder at 5.6 m)	vm
	Water & slough at 0.6 m at completion Completed: 5-cm PVC water-table well, slotted 1.8-2.8 m		

Test hole	Depth (m)	Texture and Lithology	Moisture Content*
Wt-5	0.0-3.0	brown medium sand, some silt & clay (saturated at 0.5 m)	s
	3.0-3.7	grey-brown sandy clay till, pebbles, firm, medium plastic	vm
	3.7-4.5	a/a, stiff	m

Dry at completion
Completed: 5-cm PVC water-table well, slotted 1.3-4.3 m

Table D.1. Water chemistry data from groundwater near Narrow Lake.

WELL	DATE	Spec. Cond. uS/cm	pH	Na	K	Ca	Mg	HCO3	Cl	SO4
						mEq/L				
N1-W	03-Aug-83	661	7.42	0.26	0.07	4.62	2.98	6.93	0.006	1.80
	16-Aug-83	678	7.45	0.24	0.07	4.84	3.04	6.80	0.006	0.71
	13-Sep-83	655	7.35					7.28		
	19-Oct-83	642	7.21	0.24	0.07	5.43	3.20	7.00	0.004	0.78
N1-P1	03-Aug-83	948	7.70	2.22	0.12	5.31	3.31	9.27	0.010	2.61
	16-Aug-83	841	7.84	1.29	0.12	5.22	3.35	8.43	0.008	1.24
	13-Sep-83	978	7.62			5.44		8.81		
	19-Oct-83	862	7.65	3.33	0.10		2.80	9.25	0.006	1.83
N1-P2	27-Feb-85	880	7.70	0.83	0.11	5.59	3.70	9.79		0.46
N2-W	03-Aug-83	882	7.50	0.64	0.12	6.09	3.61	13.89	0.003	0.90
	16-Aug-83	920	7.30	0.54	0.11	3.38	3.47	19.95	0.004	1.33
	13-Sep-83	905	7.59					10.16		
	19-Oct-83	870	7.43	0.40	0.12	5.57	4.71	10.02	0.005	0.74
N2-P	03-Aug-83	880	7.49	1.17	0.11	6.08	3.54	11.44	0.003	0.47
	16-Aug-83	860	7.59	0.89	0.12	5.78	3.82	11.03	0.004	0.41
	13-Sep-83	903	7.64					10.60		
	19-Oct-83	910	7.47			5.84	4.75	10.76	0.006	0.60
N5-W	03-Aug-83	548	7.52	0.14	0.05	4.32	2.06	7.22	0.003	0.09
	16-Aug-83	547	7.72	0.11	0.05	4.41	2.06	5.90	0.003	0.06
	13-Sep-83	572	7.67							
	19-Oct-83	566						6.81		
N5-P1	03-Aug-83	425	7.64	0.13	0.08	3.41	1.91	5.24	0.003	0.08
	16-Aug-83	600	7.94	0.38	0.09	4.38	2.97	6.54	0.003	0.56
	13-Sep-83	639								
	19-Oct-83	625	7.71					6.95	0.005	0.94
WT1	03-Aug-83	1010	7.69	0.50	0.07	5.56	4.89	12.59	0.002	0.60
	16-Aug-83	1040	7.65	0.39	0.03	5.23	4.95	11.01	0.002	0.58
	03-Nov-02	1041	7.69					15.48		
	19-Oct-83	1038	7.52	0.41	0.05	6.95	5.21	12.19	0.004	0.71
WT2	03-Aug-83	1420	7.43	0.60	0.09	7.30	7.48	17.83	0.004	0.27
	16-Aug-83	1470	7.37	0.57	0.09	4.37	7.38	15.48	0.004	0.21
	13-Sep-83	1430	7.42					15.90		
	19-Oct-83	1360	7.63	0.56	0.08	6.66	7.85	16.68	0.006	0.36
WT3	03-Aug-83	600	7.86	0.27	0.06	4.35	3.34	7.00	0.003	0.15
	16-Aug-83	660								
WT4	03-Aug-83	578	7.55	0.37	0.09	4.42	2.94	6.33	0.002	0.40
	16-Aug-83	430	7.70	0.12	0.07	3.19	1.86	4.50	0.004	0.07
	13-Sep-83	448								
	19-Oct-83	425	7.59			2.90	1.90	4.75	0.003	0.23
WT5	03-Aug-83	410	7.38	0.15	0.05	3.40	1.80	4.72	0.002	0.11
	16-Aug-83	436	7.57	0.14	0.06	3.16	1.92	4.58	0.004	0.07
	13-Sep-83	427	7.35					5.04		
	19-Oct-83	540	7.59					5.82	0.003	0.17
N4-W	03-Aug-83	845	7.95	0.94	0.10	3.26	7.16	10.63	0.004	0.44
	16-Aug-83	760	8.07	0.99	0.11	3.09	7.59	10.55	0.004	0.45
	13-Sep-83									
	19-Oct-83									
N4-P1	03-Aug-83	983	7.82	1.20	0.14	5.83	7.02	11.34	0.003	0.62
	16-Aug-83	1003	8.00	1.15	0.12	3.65	6.83	11.34	0.003	0.64
	13-Sep-83	998								
	19-Oct-83	940	7.80					11.44	0.003	0.67

Table D.2. Total dissolved phosphorus concentrations (mg.m^{-3}) in groundwater collected from water wells near Narrow Lake.

Well	Date			
	15/8/83	13/9/83	13/10/83	22/6/84
N1-W	16.2	17.7	38.7	84.6
N1-P1	17.4	16.1	38.1	31.4
N2-W	14.8	14.3	23.1	27.6
N2-P	9.2	8.6	47.1	12.9
N4-W	9.5			31.9
N4-P1	6.5	16.1	42.8	7.5
N4-P2				15.4
N5-W	11.5	4.8	18.3	10.4
N5-P	12.4	16.5	46.6	24.1
WT1	7.2	24.6	27.5	16.7
WT2	8.9	21.4	50.1	8.1
WT3	3.6	16.2		14.8
WT4	17.2	7.3	31.4	15.4
WT5	31.2	10.7	23.8	13.8