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Groundwater Interactions with Holland Lake, MN

by

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Abstract

Holland Lake, a small but deep mesotrophic lake in the Twin Cities Metropolitan Area, has been considered by the Minnesota Department of Natural Resources, Division of Fisheries, for stocking with brown trout. Holland Lake, with a surface area of 0.14 km^2 (35 acres) and a maximum depth of about 18.8 m (61 ft) consists of two shallow bays covered with rooted macrophytes and a deep main basin. The deep basin is thermally suitable for brown trout. However, due to a high oxygen depletion rate in summer, the lake becomes anoxic below the surface mixed layer. The field study conducted in the summer of 1999 by the authors concluded that several mechanisms, all regarding some sort of horizontal advection process, could explain the observed high dissolved oxygen (DO) depletion rates: transport of detrital material from the shallow bays, density currents combined with sediment oxygen demand in the shallow bays and flushing effect by groundwater flow through the lake. Density currents from the shallow bays were attributed to the temperature regimes of the shallow bays.

To aid in the design of an aeration system for the lake, a new field study was conducted in the summer of 2000 to quantify the potential groundwater flow through the lake, especially through the shallow bays. The field study included the measurement of groundwater piezometric heads underneath the lake bed using a potentiomanometer and DO concentrations and temperatures of groundwater. In addition, the water temperature profiles were measured at several locations in the shallow bays to investigate the potential for density currents.

The results showed that the shoreline sediments were quite heterogeneous. The groundwater inflow to the shallow bays was weak and localized. In the winter of 2000, at the beginning of the ice cover period two springs were identified in the western shallow bay. However, the sediment temperatures and the piezometric heads at the periphery of the springs did not indicate any significant groundwater inflow.

By keeping track of the drop in lake stage after Holland Lake was flooded in July 2000, the groundwater throughflow was estimated from a lake water budget to be about 8,500 $m^3 day^{-1}$. It appears that most of the groundwater flow enters the shallow bays of Holland Lake, and is either intermittent near surface flow (caused after rainfall events) or/and due to a few springs, which do not significantly affect the thermal characteristics of the shallow bays. It also appears that the aquifer is continuously interacting with the lakes in the region.

Table of Contents

List of Figures

I. Introduction/Objectives

Holland Lake (Figure 1) located in Dakota County, in the City of Eagan, has been considered for stocking with brown trout by the Division of Fisheries of the Minnesota Department of Natural Resources (MNDNR). Holland Lake is considered suitable for this purpose because it is exceptionally deep and cold in comparison to other lakes in the Metro Area and has relatively good water quality because of a small drainage area. However, there are frequent problems with low concentrations of dissolved oxygen (DO) in the lake metalimnion and hypolimnion, which adversely affect the stocking of the lake with brown trout. In summer, water temperature in the surface mixed layer (epilimnion) exceeds the maximum temperature tolerance of brown trout (21 °C [*MNDNR*, 1978], 23 to 25 °C [*Lee and Rinne*, 1980]), which forces brown trout to find suitable habitat (between 10 °C to 20 °C) in deeper layers of the lake. However, thermocline anoxia prevents fish from migrating to the lower cooler lake strata.

Water temperature and dissolved oxygen measurements by the Minnesota Department of Natural Resources (MNDNR), the Twin Cities Metropolitan Council, and the St. Anthony Falls Laboratory showed that an anoxic layer develops in the upper metalimnion of Holland Lake in late June or early July and progresses downward (Figure 2) until the entire metalimnion and hypolimnion become anoxic in mid- to late August. Beginning in late September (Figure 3), the aerated surface mixed layer starts deepening until a complete turnover occurs in November.

The field study conducted by the authors in the summer of 1999 showed that the shallow bays exhibited significant temperature stratification in July and August. The bottom temperatures in the shallow bays were low, ranging from 16 °C to 19 °C, which implied that there was not enough energy available to warm up the water at the bed level of the bay. Water at the bottom of the eastern shallow bay was about 3 to 5 °C colder and hence denser than water at the same depth in the deep main basin. This temperature difference can be expected to cause a continuously intruding density current from the deeper parts of the shallow bays into the deep main basin, which can transport oxygen depleted water and substantial amounts of suspended solids and detritus into the upper stratum of the metalimnion in the deep main basin. This material ultimately acts as a major oxygen sink with a high DO depletion rate in the metalimnion.

The thermal stratification in the shallow bays can be attributed to several potential causes: (1) The small wind fetch across the lake and the tall wind sheltering trees around it, which decrease wind mixing in the shallow bays, (2) the presence of dense macrophyte beds, which dampen the turbulent kinetic energy, produced by wind blowing over the surface, and prevent short wave radiation reaching deeper layers by self shading, and (3) a source of cool water at the bottom of the shallow bay, which has to be groundwater. Groundwater flow into and out of Holland Lake is investigated in this report.

The groundwater inflow to Holland Lake may be connected to several lakes with significantly higher lake stages located to the west and south of Holland Lake. In particular, O'Brien Lake, which is only 350 m away, has an elevation of 277 m a.m.s.l. compared to Holland Lake's 265 m. It appears that Holland Lake is at the focal point of these lakes and may be connected to them by groundwater. Holland Lake intercepts a

Quaternary aquifer. According to the *Minnesota Geological Survey* [1990] the aquifer is confined. Estimating very roughly the inflow from the piezometric head gradient upstream of the lake, assuming a capture width of 900 m, and a hydraulic conductivity of 40.7 $mday^{-1}$ gives an approximate groundwater throughflow rate of 9570 m^3 day^{-1}. The preliminary results obtained by *Mohseni and Stefan* [2000] regarding the groundwater flow through Holland lake are reproduced in this report in Appendix A.

The 9570 m^3day^{-1} is a rough estimate and represents a substantial amount of groundwater flow through Holland Lake. To verify the magnitude of this flow, its entrance region to the lake, and its temperatures and dissolved oxygen concentrations, it was necessary to conduct a field study. This report describes the field study conducted in the summer of 2000, and the analysis of the field data to better characterize the groundwater flow through Holland Lake.

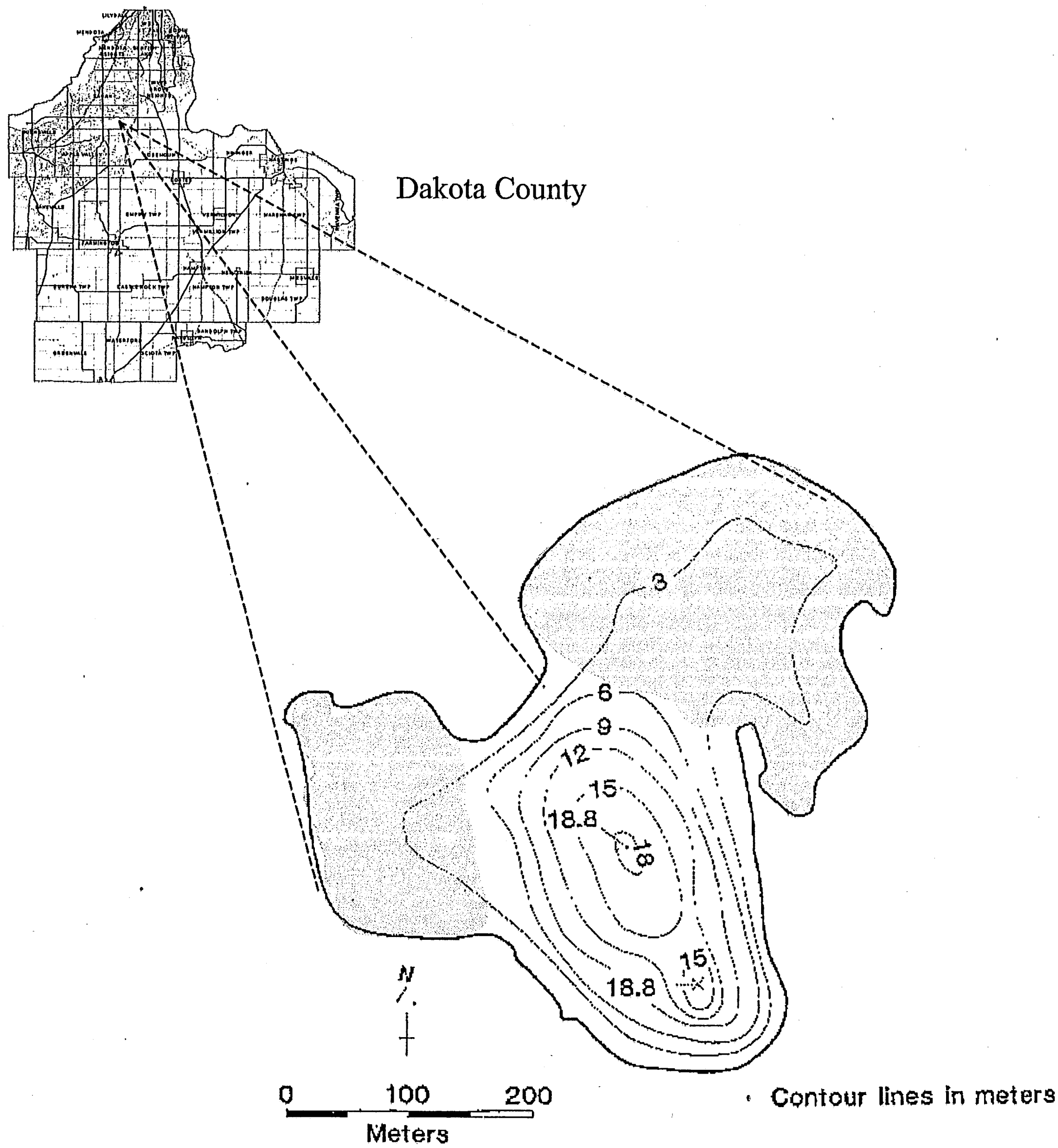

Figure 1. Location and plan view of Holland Lake, MN. The gray areas are macrophyte beds.

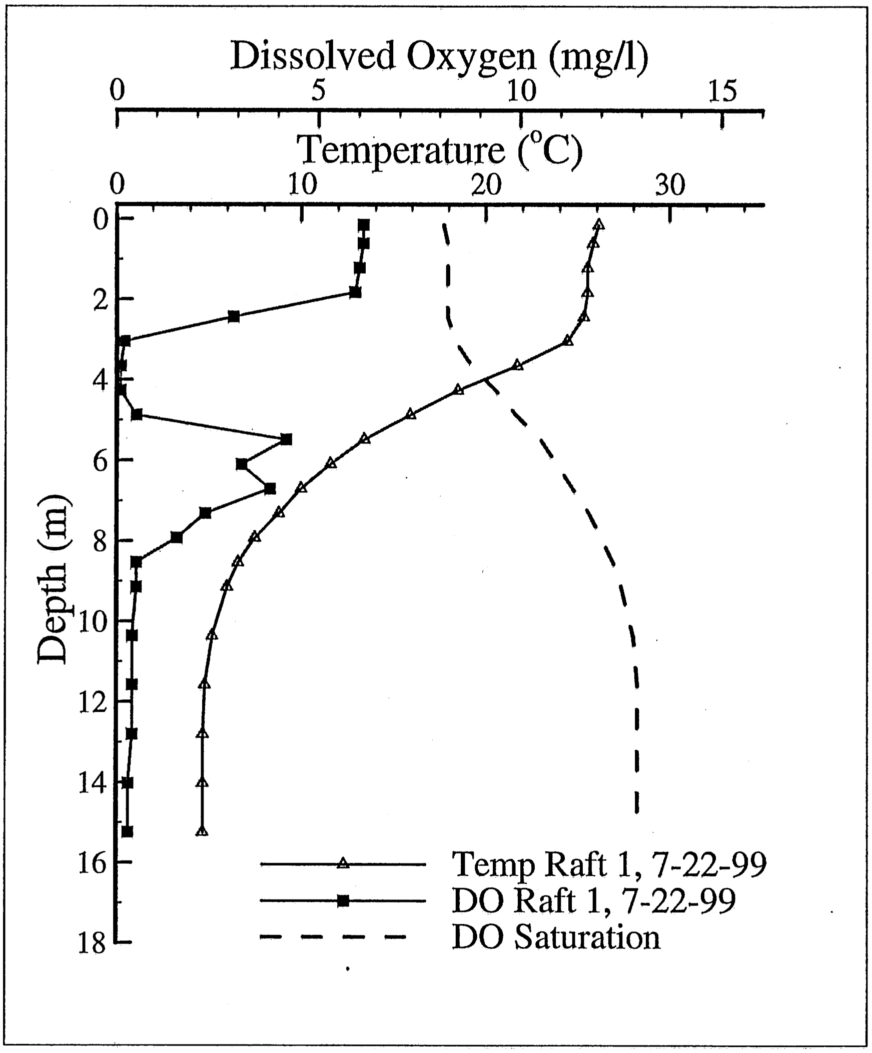

Figure 2. Temperature and DO profile measured in the deep main basin of Holland Lake in the summer of 1999.

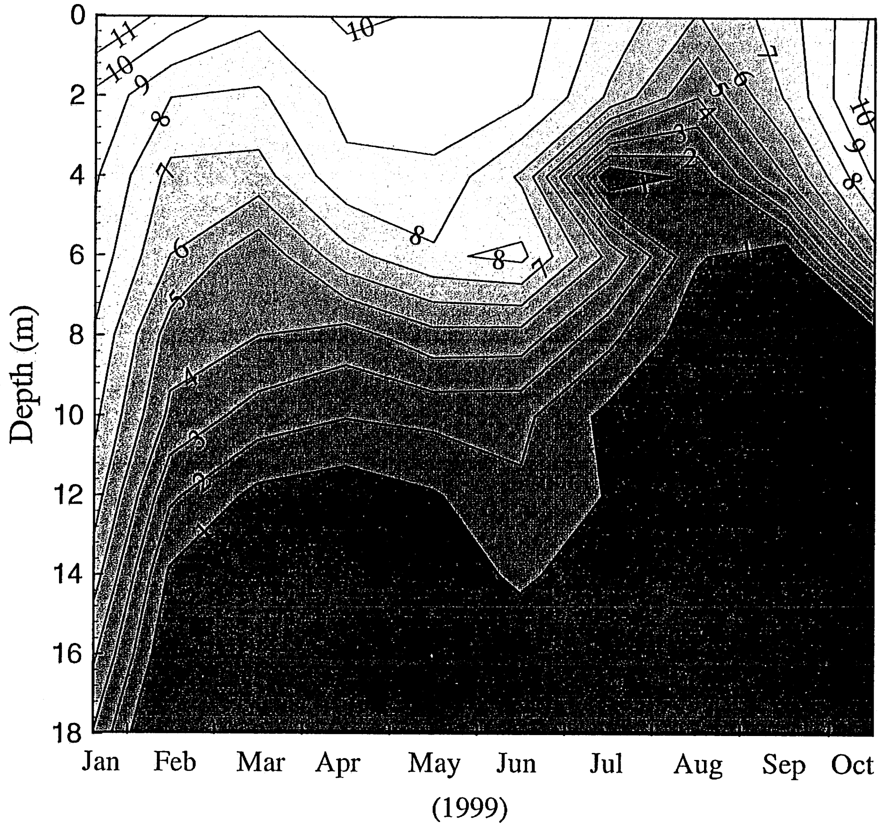

Figure 3. DO isopleths of the deep basin of Holland Lake. The figure is plotted using the monthly, biweekly and weekly DO profiles measured by the MNDNR, the Met Council and the St. Anthony Falls Laboratory.

II. Field Data Collection

II.1. Lake Setting

Holland Lake has a surface area of 0.14 km^2, with mean and maximum depths of 5.1 m (15 ft) and 18.8 m (61 ft), respectively (N44.7887° and W93.1429°). The lake basin consists of two shallow bays covered with rooted macrophytes and a deep main basin (Figure 1). According to the MNDNR, the lake basin is 69% littoral and 31% profundal*. The drainage area of the lake is heavily forested (hardwood). The two bays have maximum depths of 4.5 m (14.8 ft) and surface areas of 0.05 and 0.02 km^2.

II.2. Storm Event of July 9, 2000

In 1986, the MNDNR reported [*Department of Natural Resources*, 1986] that a major storm sewer of the City of Eagan, which had been draining into Holland Lake, had drastically reduced the water quality of the lake since the 1975 survey. According to the MNDNR, no storm runoff has been allowed to drain into the lake, since 1988. However, in July 9, 2000, a storm occurred in the City of Eagan with an average rainfall of 180 mm (7 inches) over the entire region and a maximum of 300 mm (12 inches) at some points in the city (see Appendix B). The region was flooded, and the Dakota County Park diverted some of the storm water from the adjacent parks to Holland Lake. As a result, the lake stage raised by more than 1.5 m (5 ft) and lake clarity decreased substantially for more than four weeks (based on site observations). The DO and temperature profiles measured by the MNDNR and the St. Anthony Falls Laboratory after the storm are displayed in Figure 4. The seasonal surface mixed layer after the storm was deeper than normal, presumably due to enhanced mixing between the surface mixed layer and the layer immediately below it due to the surface water inflow. The DO anoxic layer was not yet formed by July 27, but a DO deficient layer was developing from the 4.2 m (14 ft) depth to 5.4 m (18 ft) depth (Figure 4a) in a deeper stratum than in previous years presumably due to the increase of the lake stage. The entire water column, even the hypolimnion had a minimum of 2 *mg* DO l^{-1}. By early August, the upper stratum (from 3.6 m to 6 m) became anoxic (Figure 4b).

There was no DO below the 2.4 m depth (8 ft) in the shallow bays. The western shallow bay was not thermally stratified (Figure 4d), while the eastern shallow bay was weakly temperature stratified (Figure 4c).

II.3. Data Collected

Data related to groundwater flow through Holland Lake, were collected as follows:

(1) In the summer of 1999, temperatures measured by a sonde with an attached thermistor in lake sediments of the littoral region were also assembled and plotted for analysis.

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* According to the MNDNR's definition, littoral areas have a water depth less than or equal to 15 ft.

(2) Since the lake stage rose substantially in July 2000, it was expected that the lake would exhibit an increase in groundwater outflow until it would reach again near an equilibrium stage. Therefore, the lake stage was measured at every trip to the lake, with a minimum time interval of 3 days and a maximum of one month.

(3) At the beginning of the summer season, water temperature profiles were measured at several locations in the shallow bays, and at one location in the deep main basin. These profiles were taken to examine the spatial variability of water temperature in the shallow bays in comparison to the deep main basin.

(4) The piezometric head difference between the lake surface and the lake sediments was measured at more than 50 points in the shallow bays and the littoral zone of the deep main basin using a potentiomanometer (see section II.4 for more information on the instrument). The potentiomanometer was positioned at depths from 0.6 m to 4.5 m below the lake bed (sediment/water interface). The piezometric heads were mapped through the shallow bays. Any time in September and October 2000, that positive piezometric heads were detected, indicating that a groundwater flow into the lake existed, a groundwater sample was withdrawn and DO concentration was measured using a portable YSI Model 58 Dissolved Oxygen Meter. The water samples were withdrawn using the potentiomanometer and were stored in the vacuum bottle. The inflow tube to the vacuum bottle was extended to the bottom of the bottle to prevent any possible reaeration of water when entering the bottle. The bottle and its opening were large enough for an easy insertion of the DO probe and its attached stirrer. The vacuum bottle was void of air, therefore, it was unlikely to aerate the samples.

(5) During the ice cover period, ice thicknesses, sediment and water temperatures along the shores of the shallow bays were measured to detect the presence of springs in the lake.

II.4. Potentiomanometer

To quantify the groundwater flow through Holland Lake, a potentiomanometer [*Winter et al.*, 1988; *Mitchell et al.*, 1989;*Ruddy*, 1989; *Magner and Regan*, 1994] was used to measure the piezometric head difference between the groundwater and the lake water, as shown in Figure 5. The probe consists of a conic tip, a retractable shield which could cover a fine screen over the inlet holes, and four 1.5 m (5 ft) ½" steel pipes, which could be mounted on top of each other for deeper parts of the lake. Since the probe was manually pounded into the sediment, the maximum workable length of the pipe was 6 m (20 ft). The probe was driven into the sediment to depths between 0.6 to 4.5 m, depending on the frictional resistance between the tip and the sediment. In order to drive the probe into the lake sediment, the boat was anchored by concrete block from three different directions. After the probe was driven to the maximum attainable depth in the sediment, the 5 cm long screen was exposed by retracting the shield, and a vacuum was applied to the potentioprobe and to the lake tube, to lift the head above the lake surface for observation. The difference in water elevations in the two stems of the manometer gave the piezometric head and the gradient was the ratio of the head to the depth beneath the lake bed.

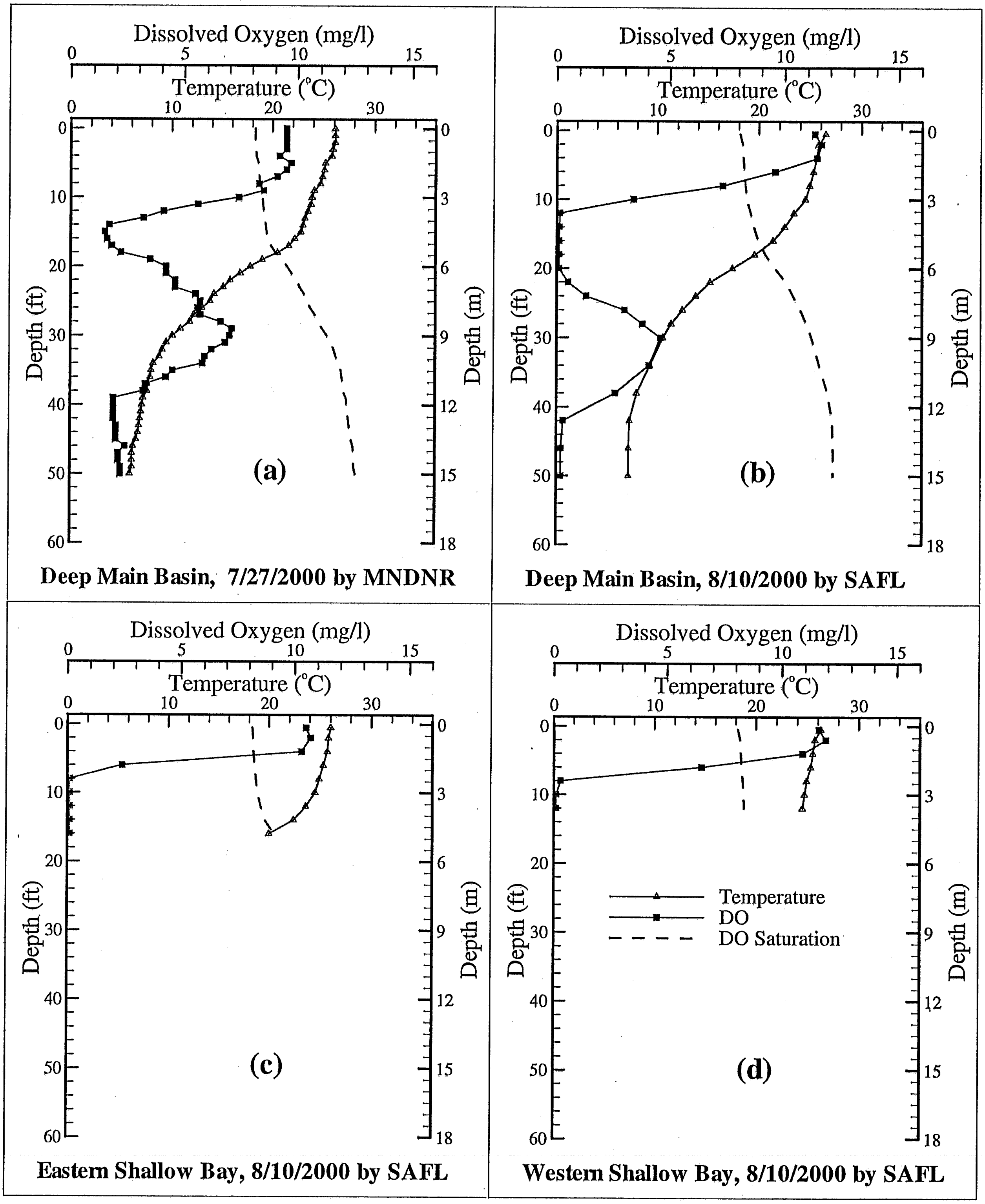


Figure 4. Temperature and DO profiles measured (a) on July 7, 2000 in the deep basin, (b) on August 10, 2000 in the deep basin, (c) on August 10, 2000 in the eastern shallow bay, and (d) on August 10, 2000 in the western shallow bay.

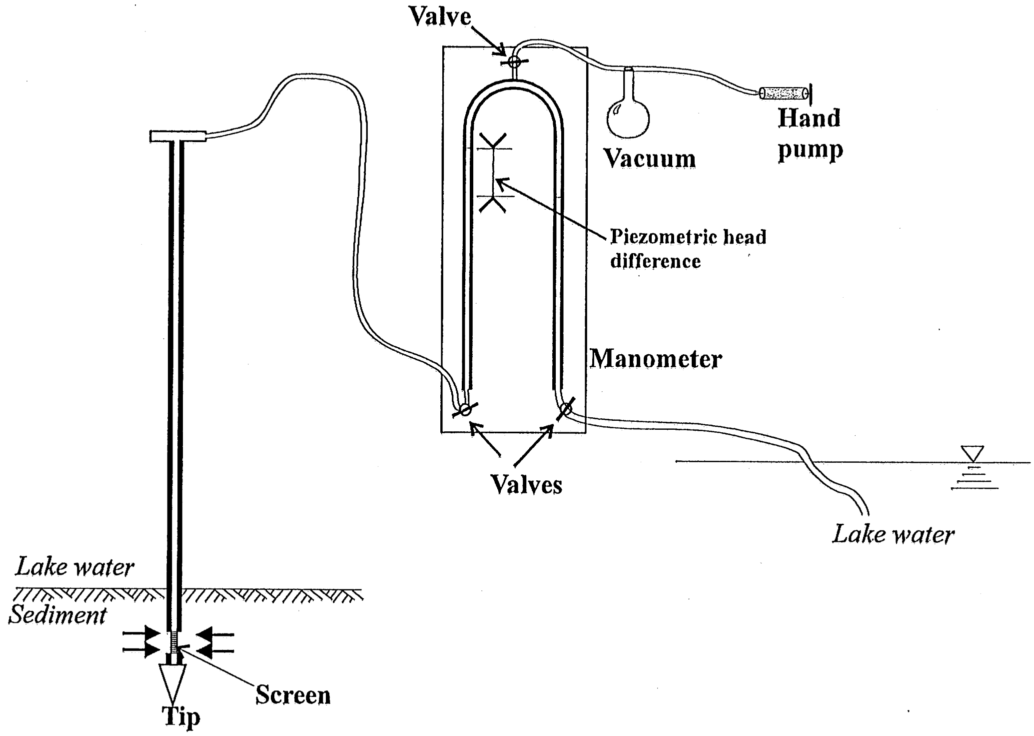


Figure 5. Schematic illustration of the potentiomanometer used to measure the groundwater piezometric heads.

III. Data Analysis

III.1. Sediment Temperatures as Indicators of Groundwater Inflow

During the field study conducted by *Mohseni and Stefan* [2000] in 1999, the potential effect of groundwater on the DO dynamics of Holland Lake was given some consideration. In order to detect the groundwater inflow to Holland Lake, sediment temperatures in the littoral zone were measured on two occasions (Figure 6). Groundwater inflow temperatures are expected to be lower than ambient temperatures in summer; groundwater outflow temperatures would equal lake temperatures near the bed. The first measurements were on August 12, 1999, when a YSI probe was inserted into the lake sediments by 2 to 2.5 cm (1 inch), and the second sediment temperature set was measured on September 3, 1999; at that time the probe was inserted more than 5 cm (2 inches) into the sediments. On September 3 the water temperatures above the sediment were also measured.

During this field work, no sediment temperature measurement fell below 21 °C. On August 12, the minimum temperature measured was 22.8 °C at the southernmost point of the deep basin, and the maximum sediment temperature was 25.8 °C at the northern end of the eastern shallow bay. On September 3, the minimum sediment temperature measured was 21.7 °C, in the western part of the eastern shallow bay. The maximum sediment temperature measured was 24.4 °C, in the eastern shallow bay. The maximum temperature difference between sediment and the ambient water was -4.3 °C, in the northwest corner of the western shallow bay.

If there was a groundwater flow through the system, the observations would indicate that groundwater was most likely entering the deep basin from the southernmost point, and the western shallow bay from northwest and southwest. However, the observed sediment temperatures near the shoreline were too warm to be representative for typical groundwater flows. Even the 4.3 °C temperature difference in the western shallow bay does not necessarily indicate the presence of groundwater. Even lakes which are not groundwater fed, can exhibit temperature gradients in the sediment below the water/sediment interface. Therefore, the 1999 sediment temperature survey did not lead to positive identification of any groundwater inflow along the shorelines.

The highest temperatures were measured in the north of the eastern shallow bay (Figure 6). If there is groundwater outflow through the shoreline, this will be the most likely location.

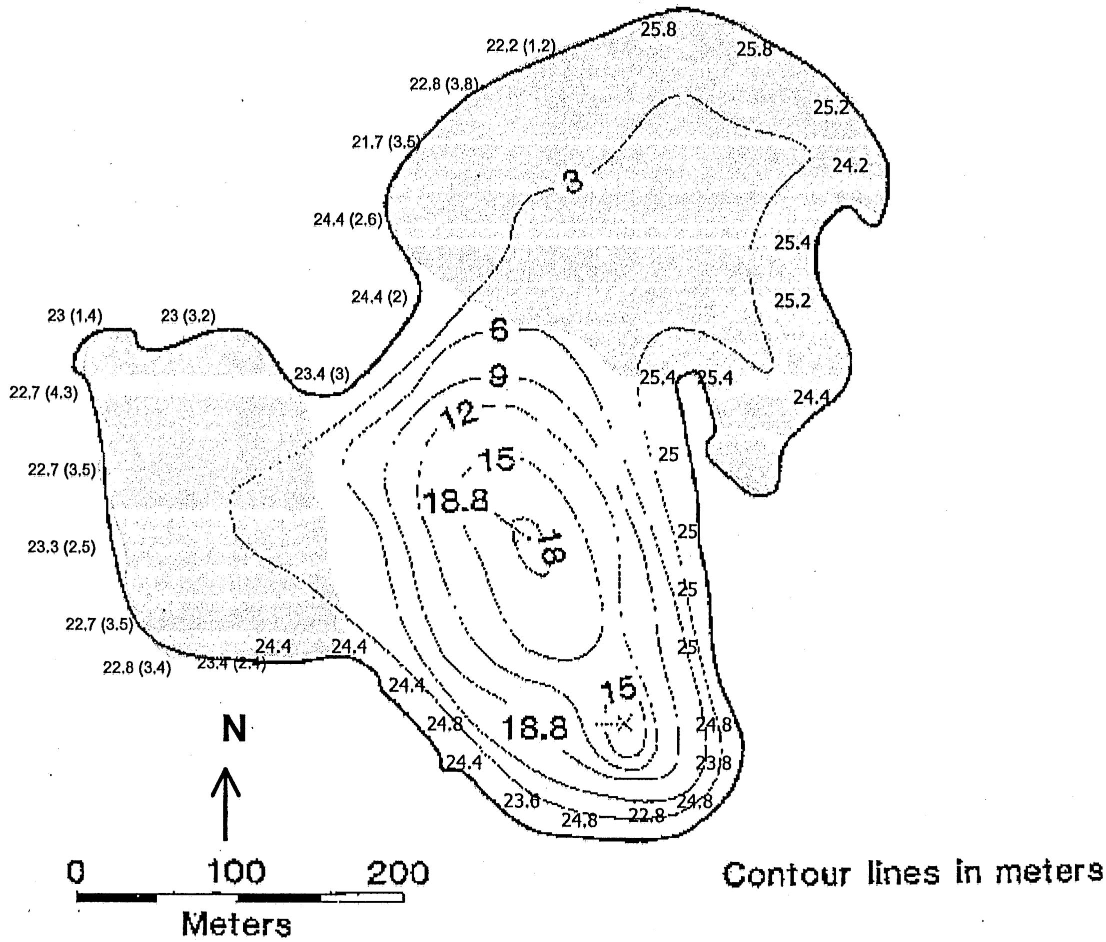


Figure 6. Sediment temperatures measured in summer 1999 in the littoral zone of Holland Lake. The numbers inside the lake surface area are sediment temperatures measured on 8/12/99, and the numbers outside the lake surface area are sediment temperatures measured on 9/3/99. The numbers in parentheses give the temperature difference between the ambient water above the sediment and the sediment. Positive values indicate that ambient water is warmer than the sediment.

III.2. Changes in Lake Stage and Net Groundwater Outflow in 2000

Since groundwater/surface water interactions are highly complex, no single method can provide definite information about groundwater flow [Winter et al., 1999]. In order to estimate the groundwater flow into the lake, the authors took advantage of the raised stage of Holland Lake due to the flood of July 2000, using the water budget of the lake. Holland Lake is a closed basin lake, i.e. it has no surface water outflow. The water budget of Holland Lake can therefore be stated in a simplified form as

$$\frac{\Delta S}{\Delta t} = P + q_s - E + q_{G,net} \tag{1}$$

where S is the lake stage (m), P is precipitation (m day^{-1}), q_s is the surface runoff per unit lake surface area (m day^{-1}), E is the evaporation rate (m day^{-1}), and $q_{G,net}$ is the net groundwater flow per unit lake surface area (m day^{-1}). Net groundwater flow is the difference between groundwater inflow and outflow. If the calculation of $q_{G,net}$ from equation 1 gives a negative value, it implies a groundwater outflow. Equation 1 was applied to the period from August 19 to December 1, 2000.

Since the Holland Lake drainage basin is small (less than 0.3 km^2) and heavily wooded, interception by the canopy and the canopy trash on the ground, does not allow any significant surface runoff, except during very intensive rainfall events. Therefore, surface runoff q_s into Holland Lake can often be assumed negligible with respect to other terms in equation 1. Consequently, the net groundwater flow for Holland Lake after the storm of July 2000 can be estimated from measurements of the lake stage and precipitation, and estimates of evaporation from weather parameters. Precipitation was not measured over the lake, but there were rain gages in the metro area, which recorded rainfall, e.g. at the St. Anthony Falls Laboratory (SAFL), in South Minneapolis on Highway 35W, at the Minneapolis-St. Paul International (MSP) Airport, at the Rosemount Agricultural Experiment Station (RAES), at Hastings Dam No. 2 (HD2) and in Farmington (F) (Figure 7). Rainfall varies significantly both spatially and temporally during storm events. Therefore, none of these rain gages represents the true rainfall over Holland Lake. However, the data from these stations provide estimates of precipitation over Holland Lake. Table 1 gives the distances of rain gage locations from Holland Lake and their available lengths of record.

Figures 8a and 8b show the precipitation data from August 15th (when the lake stage measurements started) until December 1st (when ice covered the lake). For each rainfall event three values are shown in Figure 8: (1) the combined SAFL/35W data, (2) the "Dakota County precipitation estimate" by averaging of the MSP, F, RAES and HD2 data, and (3) the RAES data as the closest rain gage. Since rainfall data were not available at the SAFL weather station prior to October 2000, and the data recorded on Highway 35W were available until November 1, 2000, the rainfall record from Highway 35W until November 1 were combined with the SAFL record of November 2000 and displayed as SAFL/35W in Figure 8. The data from MSP, RAES, HD2 and F, as the surrounding rain gages, were averaged to represent rainfall for the region in the proximity of Holland Lake. For the groundwater outflow estimation in the summer and fall of 2000 all three data sets were used.

Table 1. Rainfall gages in the metro area

Weather Station	Distance from Holland Lake (km)	Record
St. Anthony Falls Laboratory (SAFL)	22	Until 12/1/00
South Minneapolis on Highway 35W	19	Until 11/1/00
Minneapolis-St. Paul International Airport (MSP)	13	Until 9/30/00
Rosemount Agricultural Experiment Station (RAES)	9	Until 8/31/00
Hastings Dam no. 2 (HD2)	22	Until 8/31/00
Farmington (F)	14	Until 8/31/00

Lake evaporation was estimated using the Penman equation which requires weather data for solar radiation, air temperature, dew point temperature and wind speed [*Shuttleworth*, 1993]. Daily weather data were obtained from the SAFL weather station. The computations are given in Appendix C.

Figure 9 shows the components of equation 1. The rainfall data in Figure 9 are the Dakota County precipitation estimate. The net groundwater flow shown in Figure 9 is the average of computed groundwater flows using the three rainfall data sets.

The lake stage decreased significantly in August and September with an average of about 10 mm (0.4 inches) per day. The computations of groundwater flow from equation 1 consistently give a net outflow, however, some discrepancies are evident (Figure 10). They can be attributed to errors associated with (1) the representative precipitation over the lake, (2) neglecting surface runoff for all rainfall events, and (3) overestimating lake evaporation (see Appendix C). The net groundwater flow was an outflow and decreased from approximately 2000 $m^3 day^{-1}$ to 400 $m^3 day^{-1}$ in magnitude, as the lake stage approached equilibrium. A net groundwater outflow of this magnitude implies that Holland Lake is strongly interacting with the shallow aquifer. The net groundwater flow value does not indicate whether the interactions occur in the shallow bays, in the deep basin or in both areas.

Figures 9 and 10 show a net groundwater outflow of about 3 $mmday^{-1}$ or 400 $m^3 day^{-1}$ on December 1, 2000 when the lake stage appears to have reached equilibrium. This would reflect a net excess water input (from precipitation and surface runoff minus evaporation) of equal magnitude , i.e. 3 $mmday^{-1}$, which coincides with an increase of rainfall and a significant decrease in evaporation in early November (Figure 9).

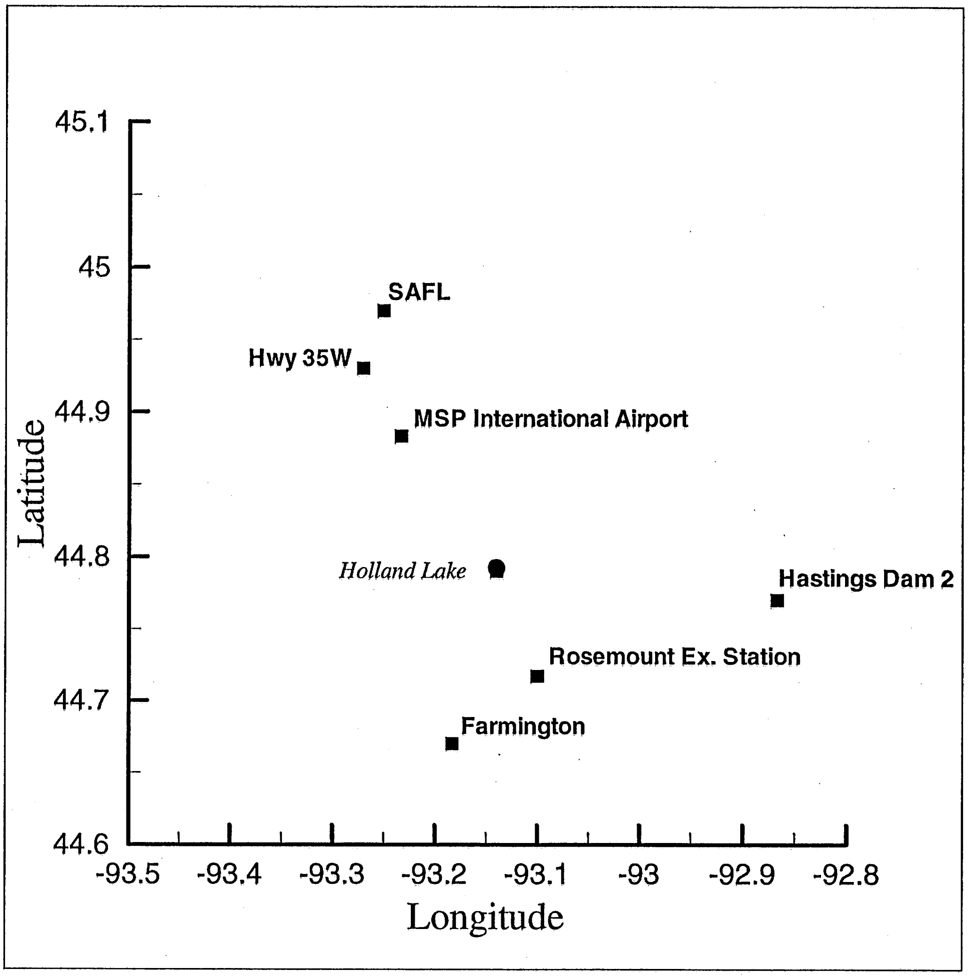
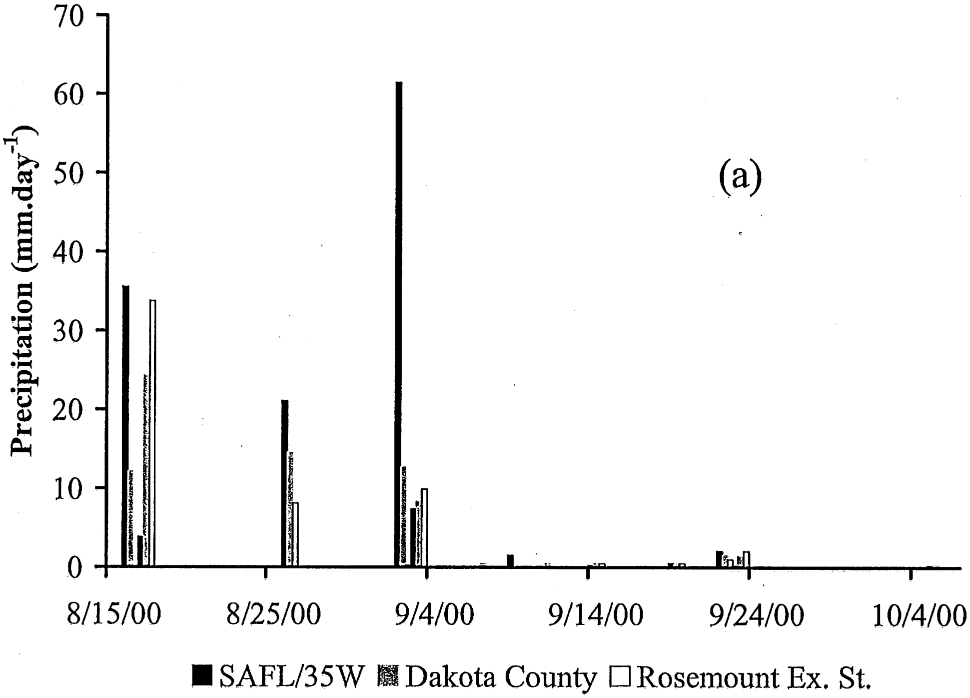


Figure 7. Locations of weather stations in the metro area with rainfall data for the summer of 2000.



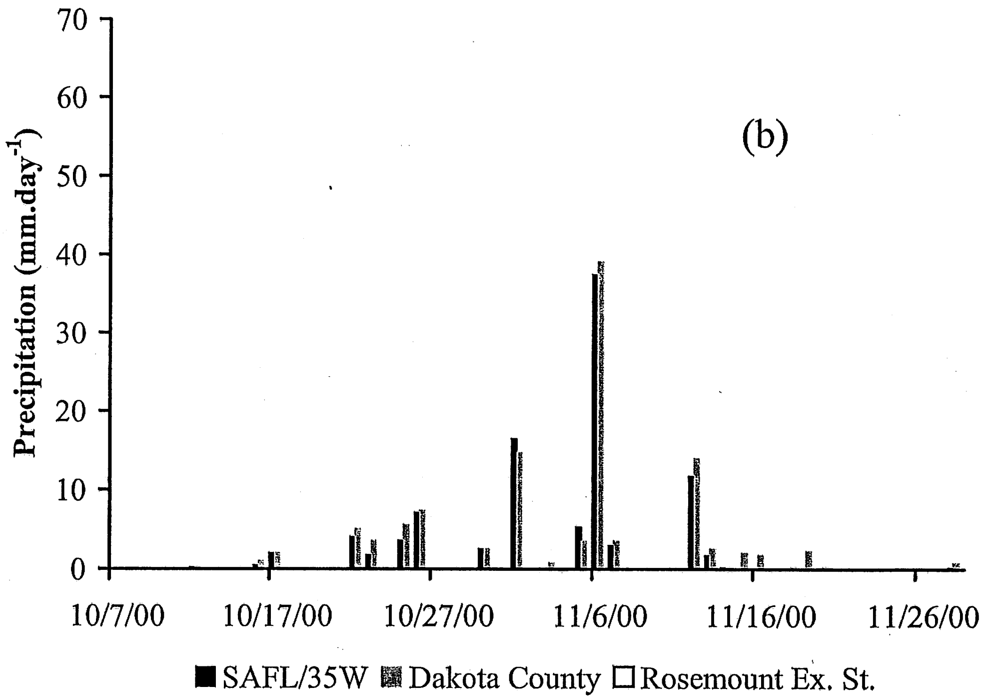


Figure 8. Precipitation data from 8/15/2000 to 12/1/2000 (1) recorded at the St. Anthony Falls Laboratory (SAFL) and Highway 35W combined, (2) estimated for Dakota County using the arithmetic average of the MSP International Airport, Rosemount Agricultural Experiment Station, Farmington and the Hastings Dam No. 2 data, and (3) recorded at the Rosemount Agricultural Experiment Station.

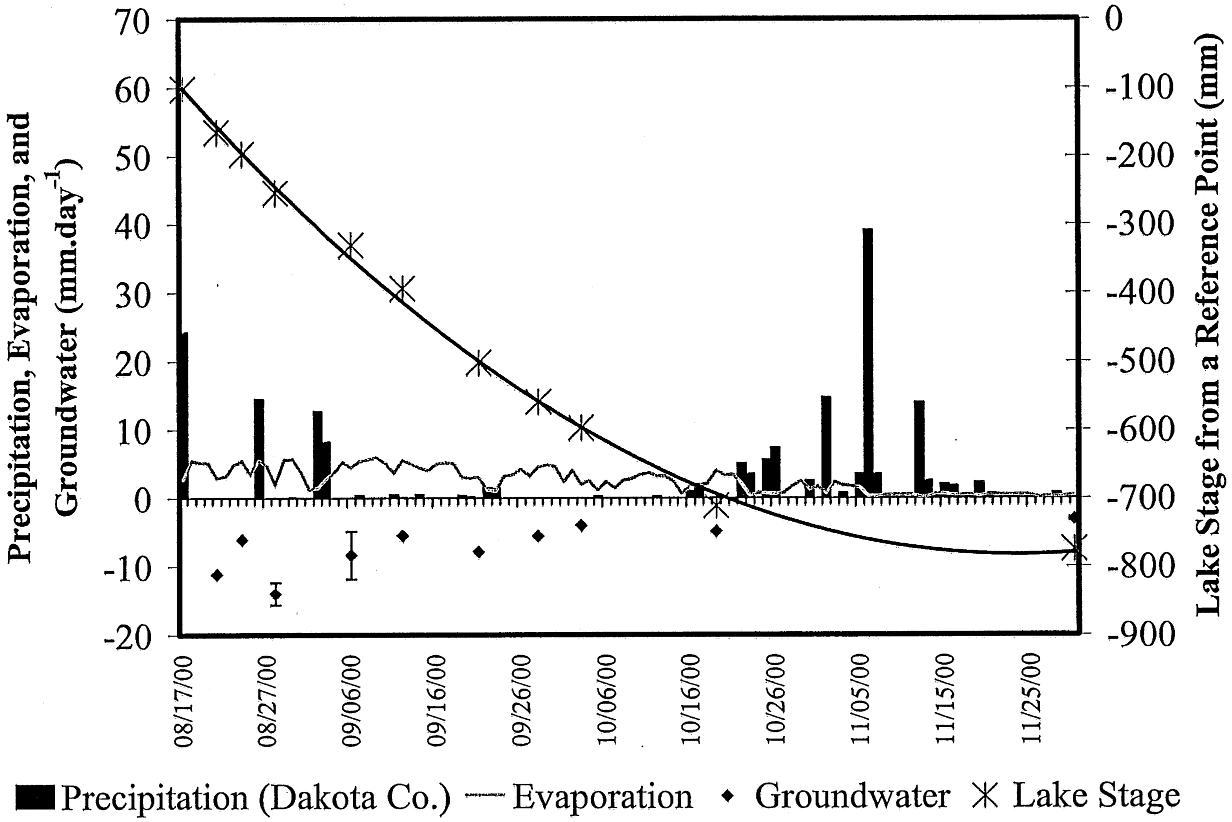


Figure 9. Daily values of the water budget components of Holland Lake from mid-August to early December, 2000. The solid line represents the quadratic function fitted to the lake stage data. The error bars for the estimated groundwater flows represent one standard deviation, when using the three rainfall data sets shown in Figure 8.

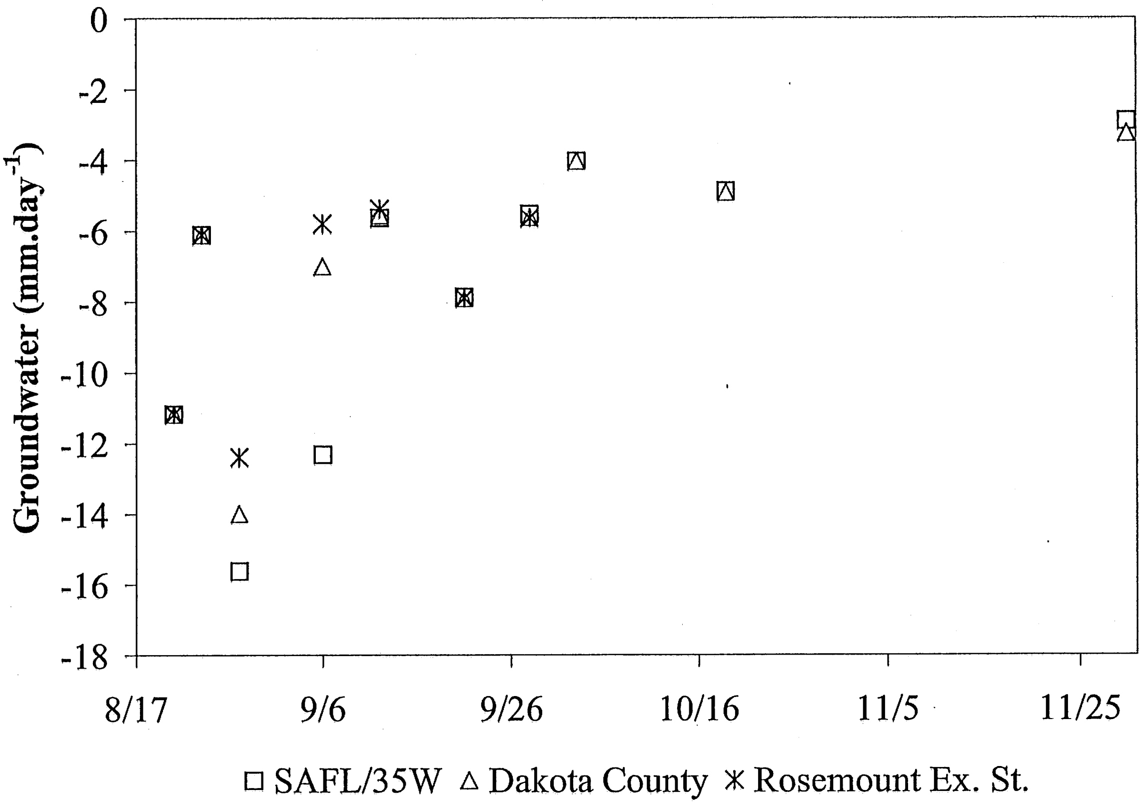


Figure 10. Net groundwater outflow from Holland Lake in the fall of 2000 for three precipitation data sets.

III.3. Piezometric Heads in the Lake Sediment

At 53 locations in the lake, shown in Figure 11, more than 80 attempts were made to measure the piezometric heads in the lake sediments. The map shown in Figure 11 was prepared by Osgood for 1985 conditions, but the lake surface area expanded after the flood of July 2000. Therefore, those locations (numbers) in Figure 11 that seem outside of the lake surface area were actually in the lake. The latitudes and longitudes of the points were measured by a Magellan 315 GPS with an accuracy of 17 m (50 ft). Since most locations were close to the shoreline, they were marked by some sort of landmark (e.g. trees) hence, one could return to the same location with an error of about 10 m. The measurements started in August 21; about a month and a half after the flood. In August, the probe was pounded to depths of 0.6 to 0.7 m below the sediment surface; in September and October it was pounded deeper (from 1.2 to 2.7 m).

During the fieldwork, it became evident that the lake sediment close to the shoreline was quite heterogeneous. In 35% of all attempts, no water entered the plastic tube of the potentiomanometer due to the presence of very fine materials such as silty loam and sandy loam that clogged the screen. The numeric labels in circular markers identify these locations in Figure 11. The numeric labels with light shaded square backgrounds show where the piezometric heads were zero or less than 1 cm (0.03 ft). Therefore, all shaded labels represent areas where no groundwater is likely to enter the lake (68% of all points). The numeric labels with no markers show the points where some positive or negative piezometric head difference between the groundwater and lake water was detected. Hence, in a clockwise direction in the eastern shallow bay, points 1, 13, 11, 6, 9, 7, 24 and 22, in the western shallow bay points 43, 37, 35, 34, 33 and 39, and in the deep basin 52, 48 and 49 represent areas with potential groundwater inflow/outflow. Only 10 attempts were made in the deep basin to measure the piezometric heads of groundwater, because the goal was to find any groundwater inflow/outflow through the shallow bays.

III.3.1. Eastern Shallow Bay

Point 1 (north of the larger bay) showed a significant outflow. The piezometric head at point 1 was measured four times from August 21 to October 12, and except in the third attempt, the piezometric head was at least 1.2 m (-3.9 ft), representing a significant outflow. Point 13 also showed a significant outflow. Cliff Road passes through the northern side of the lake. There is a pond on the other side of the road (northeast of Holland Lake), which had a lower stage than Holland Lake in summer 2000. The outflow through points 1 and 13 was most likely towards the pond.

The piezometric heads at the remaining 6 points in the eastern shallow bay are plotted versus time in Figure 12a. There was only a single measurement at point 9. Initially, all points, except point 9, showed a high groundwater inflow potential that diminished to zero rapidly. On August 22, point 11 showed less than 0.1 m groundwater inflow potential, while the surrounding points (4, 5 and 28) did not show any piezometric head. Therefore, it is quite likely that either there was a transient groundwater flow at this location, especially after the storm of 8/15 and 8/17, or there was a very weak localized groundwater inflow as a spring, which was not discernable later in September and October. The open symbols in Figure 12a represent attempted measurements in very fine materials, but no piezometric head could be recorded, e.g. point 22 on August 24th, points 6 and 7 on October 12th and point 24 on October 19th.

It is difficult to conclude from the data that there were groundwater inflows into the eastern shallow bay near points 22 and 24, because the measurements taken close to them (points 15 and 20) did not show any flow. Points 6 and 7 showed significant piezometric heads in August, but zero by early October. For points 6 and 7, the GPS reading on October 12th also showed that the probe was at least 10 m away from its original locations. Thus, it is possible that there was a localized groundwater inflow at point 6 and possibly point 7. In conclusion, the two long arrows with solid line shown in Figure 11, represent the most probable groundwater inflow/outflow points through the eastern shallow bay.

III.3.2. Western Shallow Bay

The piezometric heads measured in the western shallow bay (Figure 12b) follow the same pattern as in the eastern shallow bay (Figure 12a), i.e. they show significant groundwater flow potential in August, but diminished potential in September and October. Only point 37 shows an increase in the piezometric head with time, but the September 12th and 21st readings were at least 12 and 7 m away from the original point. Excluding points 37 and 43, no significant groundwater flow was detected in September and October in the western shallow bay. Consequently, if there was a groundwater inflow to the western shallow bay, it was a very weak localized inflow.

III.3.3. Deep Main Basin

Since the measured groundwater potential in the shallow bays was localized and weak, a survey of the groundwater flow potential in the littoral zone of the deep main basin was conducted. The piezometric heads were only measured at eight points (points 46 to 53). In late August, a +0.07 m piezometric head difference between the sediment and lake water was observed at point 48 (Figure 12c), and a lower value, i.e. +0.03 m, at point 49. A very weak outflow potential was also detected in the northeast of the deep basin, i.e. point 52. Points 48 and 49 were revisited in September 12, 2000. Almost no piezometric head was recorded at point 48 and +0.01 m at point 49. It was concluded that if there was any inflow to the littoral zone of the deep basin, it would be a transient weak flow.

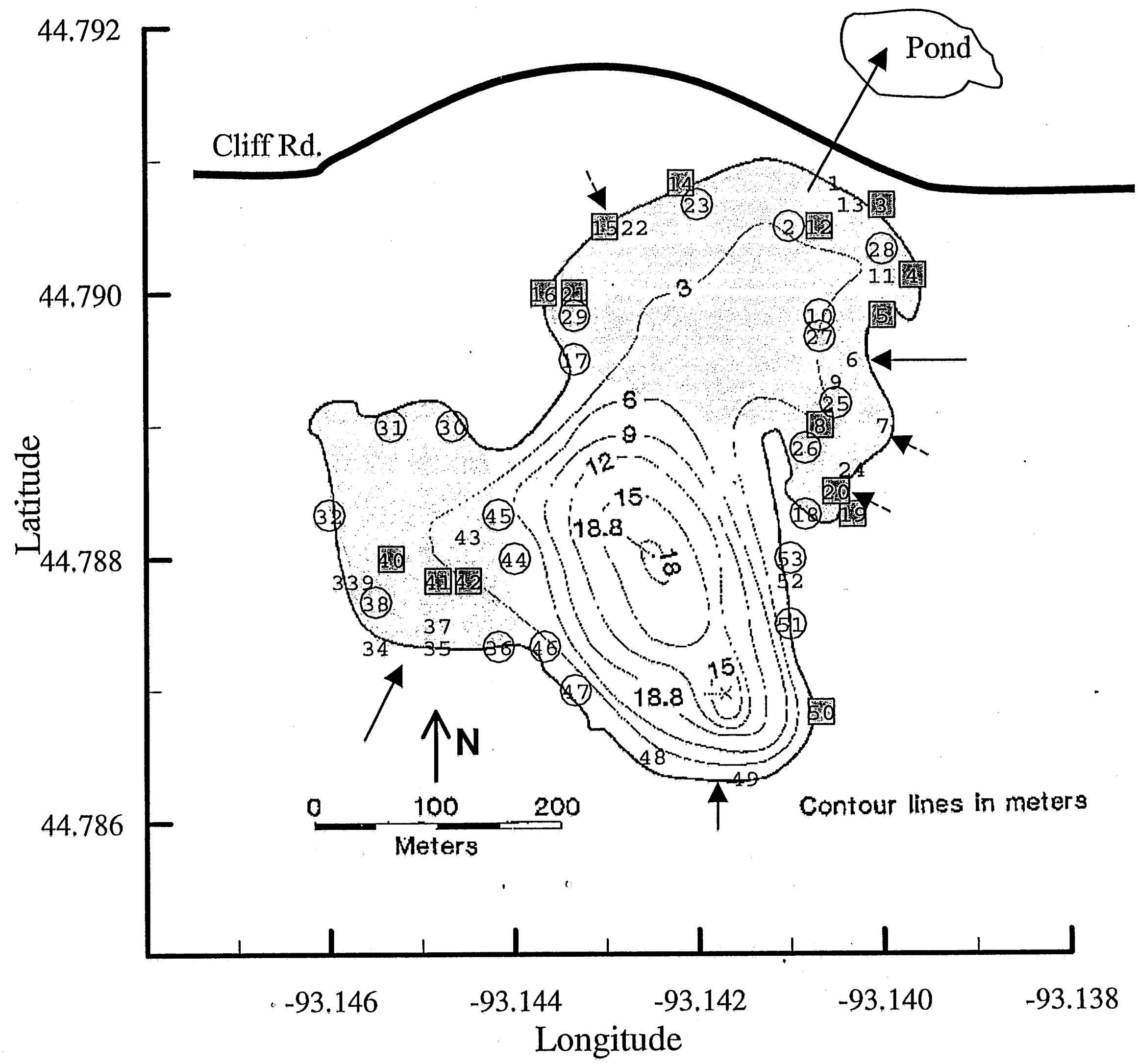


Figure 11. Locations where piezometric heads were measured in Holland Lake in the summer and fall of 2000. The circles display the points where no readings were obtained due to presence of very fine materials (clayey soils, silty loam and organic sediments). The shaded squares display areas where the piezometric heads were zero. Arrows with solid lines illustrate where there is a potential groundwater inflow or outflow, and the arrows with dashed lines show the points where the outflow/inflow is not certain due to discrepancies in measurements.

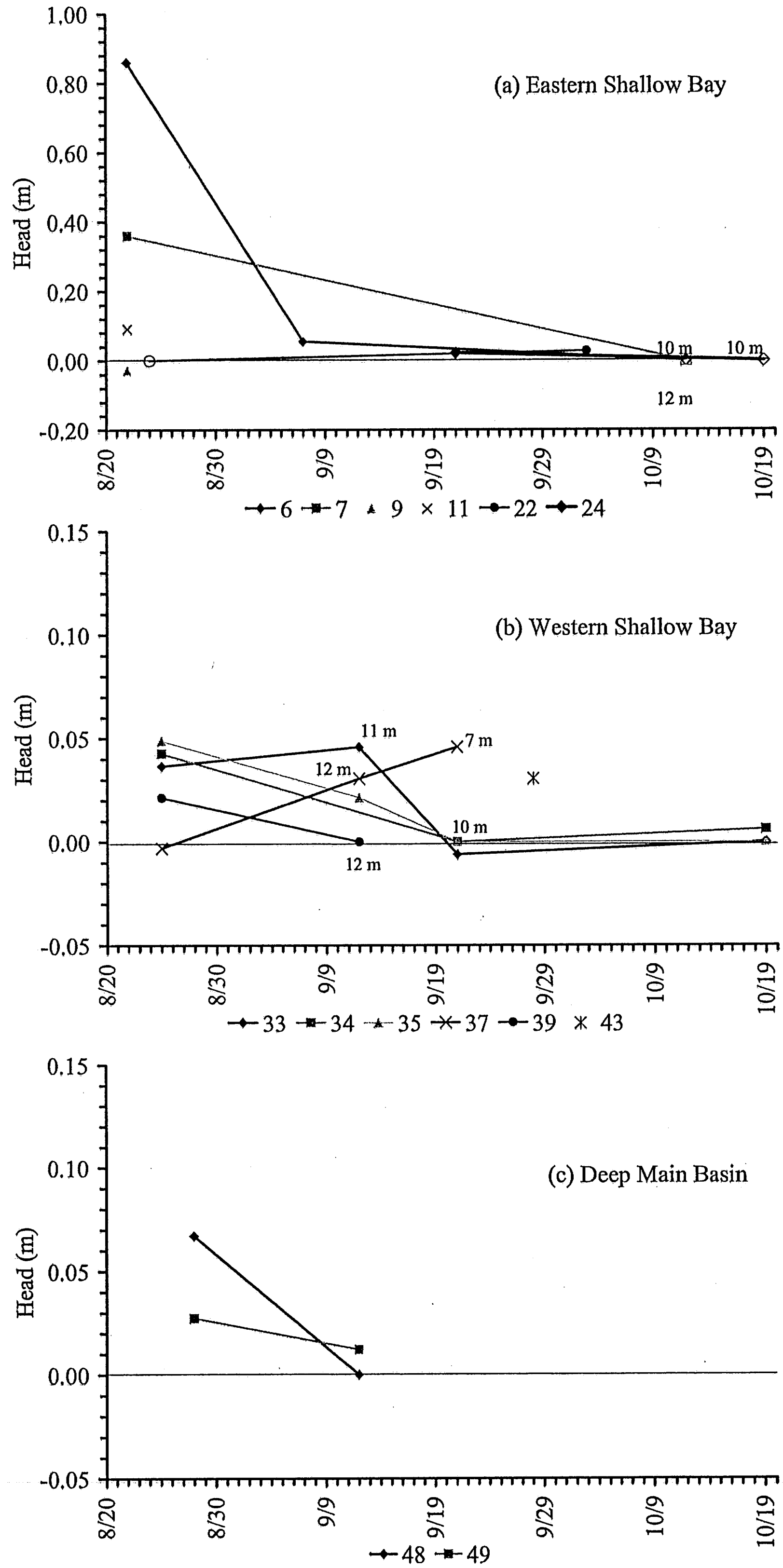


Figure 12. Piezometric head changes with time in (a) the eastern shallow bay, (b) the western shallow bay, and (c) the littoral zone of the deep main basin. Distances in meters indicate horizontal distances from the first measurement.

III.4. Dissolved Oxygen in the Sediment Water

From September 12 to September 28, 2000, points at which previously potential for groundwater inflow had been determined were revisited to measure the DO content of the pore water (groundwater inflow). During the revisiting, most points did not show any significant piezometric head difference between the sediment and the lake water. DO was measured in water samples withdrawn from six of those points: 5 points in the western shallow bay had DO concentrations ranging from 0.9 to 1.0 mg.l^{-1} and one point in the main basin (point 49) had a DO concentration of 0.75 mg.l^{-1}. Such DO concentrations are not typical of groundwater which has usually zero DO, and are more likely due to transient subsurface flow resulting from rainfall events or seepage from surface waters.

III.5. Ice Thicknesses As Indicators of Springs

Since the piezometric head measurements in the shallow bays during the summer of 2000 did not ascertain significant groundwater inflows to Holland Lake, and there were uncertainties associated with those points, which exhibited some flow into the lake, it was decided to revisit the shallow bays in winter during the ice cover period. Ice thickness is a function of heat transfer processes between lake water and the atmosphere. Ice thickness is therefore affected by air temperatures, solar radiation and snow cover. Ice thickness is also affected by the presence of groundwater inflow to a lake. Groundwater inflow temperature would exceed the water column temperature under the ice cover in most Minnesota lakes. Deep groundwater, therefore, has a lower density than water of 0 °C. This will cause the upwelling of groundwater, resulting in a reduction of ice thickness through convective heat transfer. This ice thinning occurs especially in shallow parts of lakes, where the groundwater is less diluted as it rises to the ice cover.

On December 15, 2000, the ice thickness was measured where it was expected to be susceptible to groundwater inflow. Figure 13 shows the points where 2" holes were drilled to measure the ice thickness. In addition to ice thickness, sediment temperature was measured at every location except point 19, the center of the eastern shallow bay (Table 2).

Snow depth where the ice thickness was measured, i.e. close to shoreline, was often about 7 to 8 cm (3 inches) and in the middle of the lake, it was less than 5 cm (2 inches). The ice thickness variability was partly due to snow cover variability, i.e. ice thickness was thinner where snow cover was thicker and vice versa. However, there were two spots (Figure 13) on the lake (points 9 and 14), where the ice was covered with slush over areas of less than 20 m^2 (210 ft^2). Those spots could be easily identified from 50 m away. The ice thicknesses at those locations were the smallest (13.3 and 14.6 cm, respectively) among those measured. Since the measurements taken at points 9 and 14 were on the periphery of the wet spots to avoid any ice break, it is possible that the actual ice thickness was even smaller in the center of the spots. The slush was most likely due to springs, even though the sediment temperature measurements at points 9 and 14 did not show any additional evidence (i.e. a temperature of about 8 °C) of deep groundwater inflow. The sediment temperatures ranged from 3.2 to 4 °C indicating a shallow aquifer connection possibly to other lakes of higher elevation in the area.

Based on these observations, groundwater mostly enters the western shallow bay and mainly through two spots in the north and south of the bay. The southern spot was also found by the piezometric head measurements (Figure 11), the northern spot was not.

Table 2. Ice thicknesses and sediment temperatures measured in Holland Lake, 12/15/2000.

Point No.	Longitude	Latitude	Ice Thickness (cm)	Sediment Temperature (°C)
1	93.14033	44.78983	19.1	3.2
2	93.14033	44.78950	16.5	3.4
3	93.14033	44.78933	15.9	3.6
4	93.14017	44.78900	16.2	3.3
5	93.14033	44.78883	15.9	3.5
6	93.14067	44.78850	18.1	3.3
7	93.14400	44.78883	16.5	3.4
8	93.14450	44.78900	13.3	3.6
9	93.14517	44.78900	13.3	3.8
10	93.14583	44.78900	15.9	4
11	93.14583	44.78850	18.4	3.7
12	93.14583	44.78800	17.8	3.6
13	93.14550	44.78767	15.9	3.4
14	93.14533	44.78750	14.6	3.6
15	93.14500	44.78733	16.5	3.3
16	93.14417	44.78733	20.3	3.9
17	93.14183	44.78650	20.3	3.6
18	93.14067	44.78667	18.7	3.2
19	93.14067	44.79017	24.1	N.M.

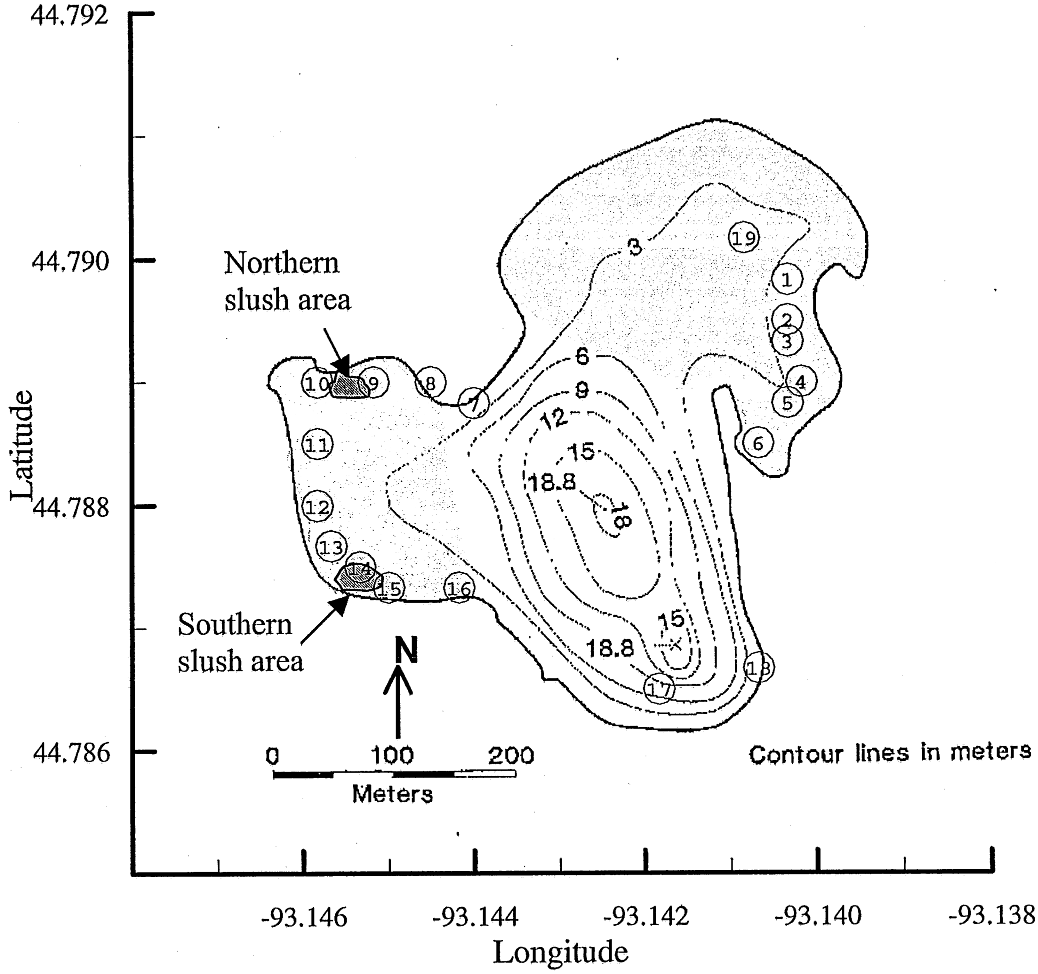


Figure 13. Points where ice thickness and sediment temperature were measured on 12/15/2000. The lake was covered by dry snow except at the two small shaded areas which were covered with slush due to presence of springs.

IV. Estimates of the Groundwater Flow Through Holland Lake

IV.1. Total Groundwater Throughflow

The summer and winter 2000 field studies showed that there was localized groundwater flow into and from the shallow bays. The flow rates could not be determined from the field data. The flood of July 2000 and the resulting lake stage rise caused a significant net groundwater outflow. Net groundwater outflow rates from Holland Lake were determined from a water budget to be from 2000 to 400 $m^3 day^{-1}$ (Section III.2). Groundwater flow through Holland Lake was still unknown but could be determined from lake stage data during the summer of 2000 as follows:

Groundwater inflow from a confined aquifer to a lake can be estimated [*Strack*, 1988] from

$$Q = KBL\frac{\phi_2 - \phi_1}{X} = \frac{KBL}{X}h \tag{2}$$

where Q is discharge (L^3/t), K is the bulk saturation hydraulic conductivity of the aquifer and the lake sediment (L/t), B is the thickness of the aquifer (L), L is the width of the capture zone (L), ϕ_1 and ϕ_2 are the piezometric heads of the lake and the aquifer at a distance X upstream of the lake (L), respectively, and h is the piezometric head difference between the aquifer and the lake (L). Since there is not enough information on the width of the capture zone, and the hydraulic conductivity presented by the *Minnesota Geological Survey* [1990] seems to be overestimated (40.7 m. day^{-1}), one can denote the entire fraction before the parameter h by a single factor C. Thus, equation 2 becomes

$$Q = Ch \quad . \tag{3}$$

Under steady state conditions, groundwater inflow $Q_{i,s}$ and outflow $Q_{o,s}$ are related by the water budget equation for a closed basin lake

$$Q_{i,s} + (P - E)A_s + Q_s = Q_{o,s} \tag{4}$$

which simplifies to $Q_{i,s} = Q_{o,s}$ because precipitation P is approximately equal to evaporation E and overland surface runoff Q_s can be assumed to be negligible in the water budget of Holland Lake. This equality can be rewritten as

$$C_2h_2 = C_1h_1 \tag{5}$$

where C_1 and C_2 are physical characteristics of groundwater inflow and outflow, as expressed by equation 3, respectively, and h_1 and h_2 are the piezometric heads of groundwater flow upstream and downstream of Holland Lake under steady state conditions with respect to the lake stage (Figure 14 and Figure A-2).

Under unsteady state conditions, i.e. when the lake stage changes with time, there would be a net groundwater inflow or outflow as was the case for the entire summer of 2000. If the lake stage increases by Δh, the departure from the steady state condition, then the net outflow becomes

$$Q_n = Q_i - Q_o = C_1(h_1 - \Delta h) - C_2(h_2 + \Delta h) \tag{6}$$

where Q_i and Q_o are groundwater inflow and outflow, respectively; under unsteady state conditions they are not equal. With a known net flow Q_n and the associated lake stage change Δh, one can solve equations 4 and 5 for C_1 and C_2 as functions of h_1 and h_2

$$C_1 = \frac{-Q_n}{\Delta h}\left(\frac{h_1}{h_2} + 1\right)^{-1} \tag{7}$$

$$C_2 = \frac{-Q_n}{\Delta h}\left(\frac{h_2}{h_1} + 1\right)^{-1} \tag{8}$$

According to equations 7 and 8, if Δh is positive, then Q_n, net outflow, is negative.

Using the hydrogeological map of the region (Figure A-1), the piezometric heads h_1 and h_2 can be estimated as 10.1 m (33 ft) and 20.4 m (67 ft), respectively (Figure A-2). From the lake stages measured in the summer of 2000 (Figure 9) and the estimated net outflows (Figure 10), ten pairs of C_1 and C_2 values were estimated from each rain data set described in section III.2 (Figure 8). Accordingly, the steady state groundwater flow through Holland Lake varies between 6,000 to 14,000 m^3 day^{-1} with an average of 8,500 $m^3 day^{-1}$. The results of these groundwater throughflow calculations are in reasonable agreement with an earlier and cruder estimate of 9,750 m^3 day^{-1} by *Mohseni and Stefan* [2000].

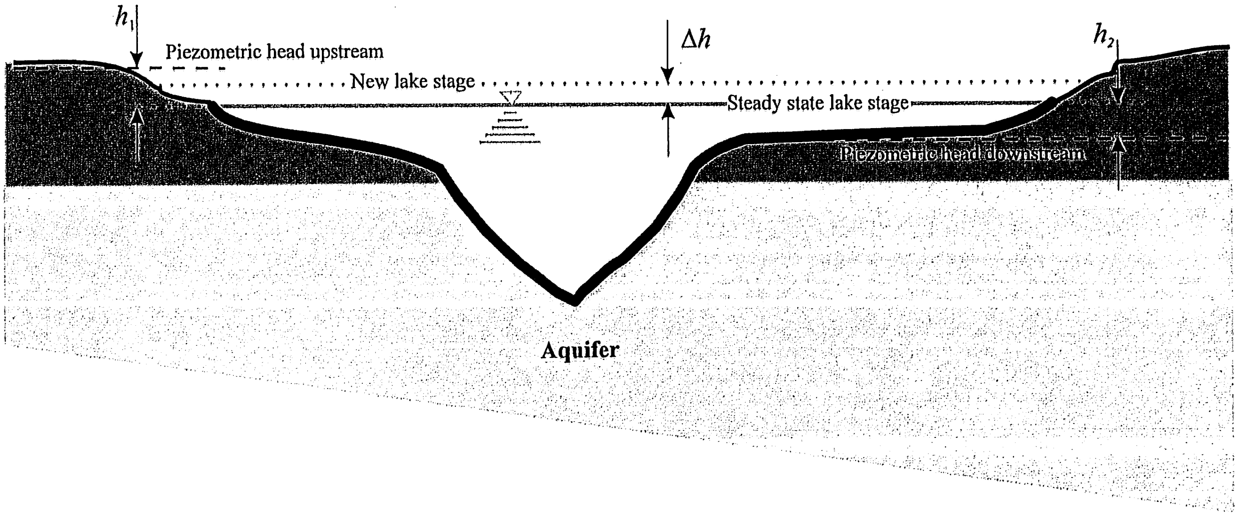


Figure 14. Schematic illustration of the piezometric heads (dashed lines) upstream and downstream of Holland Lake under steady-state conditions. Δh represents a change in the lake stage under unsteady-state conditions.

IV.2. Partitioning of Groundwater Flow

IV.2.1. Piezometric Heads and Sediment Temperatures as Indicators of Groundwater Flow

The analysis of the sediment temperatures measured in 1999 and the piezometric heads measured in 2000 leads to the following interpretation. Holland Lake intercepts a confined Quaternary aquifer. No information is available on the depth of the aquifer in the vicinity of Holland Lake. The lake is expected to intercept the aquifer as shown schematically in either Figure 15a or Figure 15b. Figure 15a shows a confined aquifer, which connects at both shallow bays and the deep basin, while Figure 15b shows a connection at the deep basin only. The piezometric heads measured in the summer of 2000 suggest that Figure 15b better represents the shallow aquifer underneath the Lake.

Sediment temperatures measured in the summer of 1999 (Figure 6) give information about groundwater inflow/outflow more or less similar to the piezometric heads measured in 2000 (Figure 11): groundwater enters the deep basin from the south, and the western shallow bay from the southwest, and leaves the eastern shallow bay from the north. However, the sediment temperatures measured in 1999, and the significant changes in the piezometric heads from August to October 2000, suggest that any subsurface inflows to the shallow bays of Holland Lake are either intermittent near-surface flows (caused by rainfall events) or due to a few springs, which do not significantly affect the thermal characteristics of the shallow bays. Both processes may act simultaneously.

IV.2.2. Lake Water Temperatures as Indicators of Groundwater Flow

The relatively low water temperatures measured in July and August of 1999 near the bottom of the eastern shallow bay could indicate groundwater inflow. The hydrogeological map of the Minnesota Geological Survey [1990] shows that groundwater enters the lake from the southwest (Figure 16). Thus, if groundwater is the source of cold temperatures at the bottom of the shallow bays, according to the general direction of groundwater flow and considering the small volume and surface of the western shallow bay, then the western shallow bay should show stronger temperature stratification with relatively colder bottom temperatures than the eastern shallow bay.

On August 4, 2000, about a month after the lake stage rose about 1.2 m (4 ft), water temperature profiles were measured at 6 locations in the shallow bays and one location in the deep basin (Figure 17). To include temporal differences due to diurnal heating and cooling, water temperature profiles were measured twice at each location on the same day (Figure 18). The measurements showed that the shallow bays were thermally stratified. It is noteworthy that the bays were deeper in the summer of 2000 than in the summer of 1999 due to the flood of July 2000. Consequently, stratification was stronger in 2000 than in 1999. The dense macrophytes had not yet covered a significant portion of the bay surface (Figure 17) when the measurements were taken. Therefore, not the shading by macrophytes but the water depth controlled the stratification in the bays. Figure 19 shows the average water temperatures (both spatially and temporally averaged) in both shallow bays, and in the deep basin (measured at 12:30 pm). Down to 4 m (13.1 ft) depth, there was no significant temperature difference among the bays and the deep basin, except at the surface. At 4.1 m (13.4 ft) depth, the western shallow bay (bed temperature) was about 1 °C colder than the eastern shallow bay and the deep basin. This information is not sufficient to ascertain that a density current from the bottom of the western shallow bay into the deep basin had to occur. The eastern shallow bay, however, showed a colder temperature (1 to 1.8 °C) than the deep main basin at the 4.5 m to the 4.8 m depth. The error bars in Figure 19 indicate one standard deviation from the mean and show that these temperature differences were significant. It is therefore quite likely that a density current from the eastern shallow bay into the deep main basin occurred in the summer of 2000. In the summer of 1999, when the temperature differences ranged from 3 to 5 °C, the density current probably occurred as well.

The water temperatures measured near the bed in the shallow bays were about 20 °C or more in the summer of 2000 warmer than those measured in the summer of 1999 (15 to 16 °C). This range of water temperatures is not indicative of any groundwater inflow from a typical, deep aquifer. Therefore, if there is groundwater intruding the shallow bays, these warm and fluctuating water temperatures suggest a connection to nearby lakes such as O'Brien Lake (Figure 16) or transient subsurface flows.

IV.2.3. Dissolved Oxygen as an Indicator of Groundwater Flow

If Figure 15b represents the groundwater/lake interactions, then the DO depletion rates measured in the lower metalimnion (6-9 m depth) in 1999 should be related to the groundwater wash out effects on this layer. Groundwater flow through the lower metalimnion of Holland Lake can be estimated using the DO budget (equation 9) of the strata from 6 to 9 m (20 to 30 ft), assuming diffusion is negligible with respect to the groundwater wash out effects.

$$\forall \frac{dC}{dt} = Q(C_g - C) + \forall P_s - \forall R \pm S \tag{9}$$

In equation 9, $\forall$ is the volume of the stratum, C_g is the DO concentration in groundwater inflow (if the groundwater flow is from the aquifer shown in Figure 15, then $C_g \approx 0$), C is the DO concentration of the stratum, Q is groundwater flow through the stratum, P_s is the rate of photosynthetic oxygen production, R is the rate of respiration, and S is the sum of all other sources and sinks of DO including sedimentary oxygen demand. Neglecting the kinetics in equation 9 (the last three terms on the right hand side of equation 9), the differential equation would give an analytical solution for groundwater throughflow Q as follows

$$Q = \frac{\forall}{t} ln\left(\frac{C_0}{C}\right) \tag{10}$$

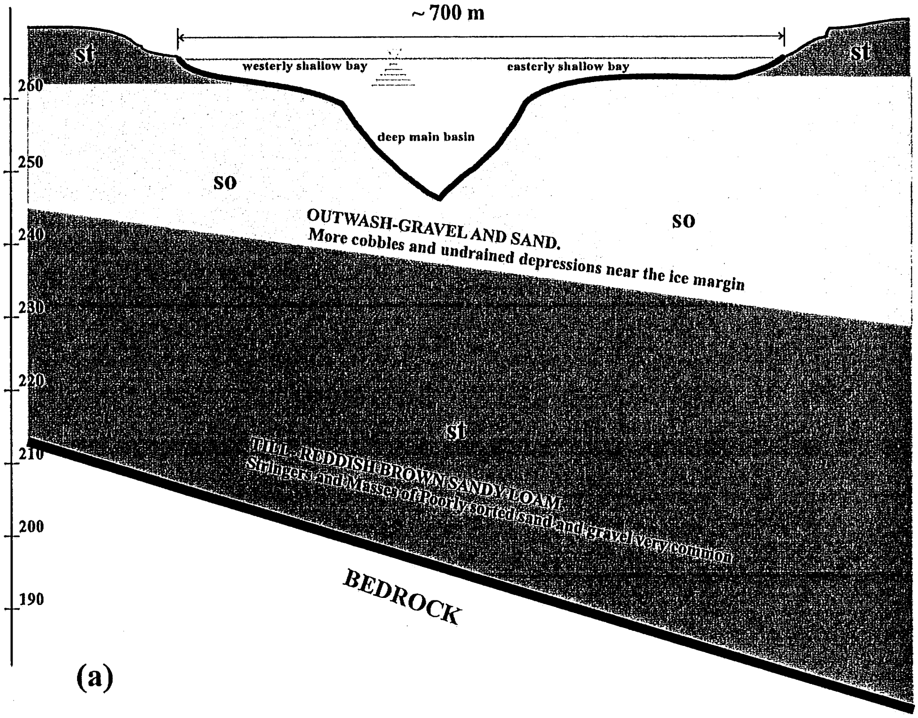
where C_0 is the DO concentration at the beginning of the period, and t is a time period. With 104,000 m^3 as the volume of Holland Lake between 6 to 9 m, and utilizing the DO concentrations measured in June and July [*Mohseni and Stefan*, 2000], Q is estimated from equation 10 to be about 4,300 m^3 day^{-1}. Therefore, assuming no biochemical contribution, only about half of the total estimated groundwater flow (8,500 m^3 day^{-1}) would be flowing through the lower metalimnion of Holland Lake.

Since the photic depth was at 5.5 m (18 ft), it is likely that there was no photosynthetic activity in the lower metalimnion. Therefore, the second term on the right hand side of the equation is actually zero. The respiration rate was measured at the ambient temperature by incubation of 4 liter bottles in September 1999, and was about 0.15 mg l^{-1} day^{-1} [*Mohseni and Stefan*, 2000]. Incorporating this rate and neglecting other sinks of oxygen, i.e. $S = 0$, equation 9 can be discretized and solved numerically for the groundwater flow by trial and error. Since the net DO depletion rates varied in June and July, equation 9 gave variable groundwater flows through the lower metalimnion (from 600 to 1,100 m^3 day^{-1}) depending upon the selection of the time period from June 23 to July 29, 1999. The results are therefore sensitive to the biochemical reactions. The 0.15 mg l^{-1}day^{-1} respiration rate used in the computation was obtained in September, which could be different from the actual respiration rates in the same layers in June and July. The groundwater flow of 4,300 m^3 day^{-1} is the upper limit of the groundwater flow through the lower metalimnion, and by incorporating biochemical oxygen demand in the computation, the groundwater flow should be significantly smaller than 4,300 m^3 day^{-1}. Consequently, most of the groundwater flow through Holland Lake should enter the lake through the shallow bays and the upper metalimnion. This is in general agreement with the high DO depletion rates observed in the upper metalimnion in 1999.

IV.2.4. Synthesis of Information on Groundwater Partitioning

Winter observations of springs (Figure 13), the piezometric head measurements, and the sediment temperatures, suggest that most of the groundwater entering the lake is either intermittent near-surface flow or due to a few springs, which do not significantly affect the thermal characteristics of the shallow bays. It is likely that the groundwater entering the western shallow bay mixes with the epilimnion during the summer months, except when the bay is strongly stratified due to abundance of rooted macrophytes. In the eastern shallow bay which is deeper and thus has a higher thermal inertia, i.e. more likely to stratify in summer, it is expected that groundwater flows from the easterly shores along the bed level and intrudes the deep basin at the upper metalimnion.

Overall, most of the groundwater flow enters the shallow bays. During the thermal stratification of the bays, starting in June and July, it may intrude the deep basin as a low DO density current carrying high BOD waters from the near bed areas into the upper metalimnion of the deep basin. Groundwater most likely leaves the lake from the upper metalimnion.



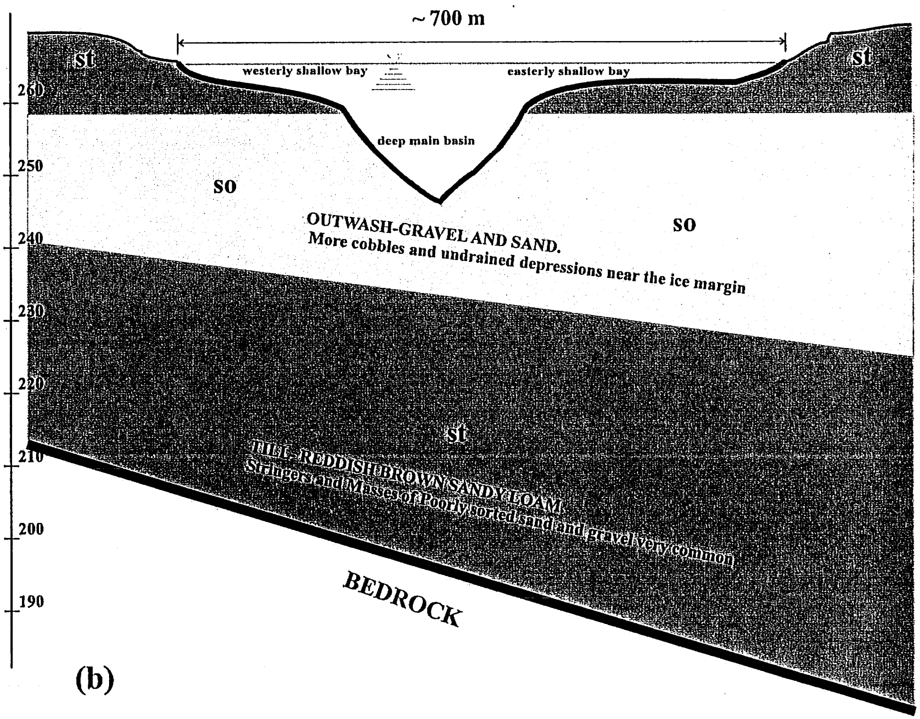


Figure 15. Two possible scenarios for the shallow aquifer intercepting Holland Lake. The shallow aquifer connects to Holland Lake (a) at the shallow bays, (b) at the deep main basin. The data are extracted from the Geological Atlas of Dakota County, MN, prepared by the *Minnesota Geological Survey* [1990].

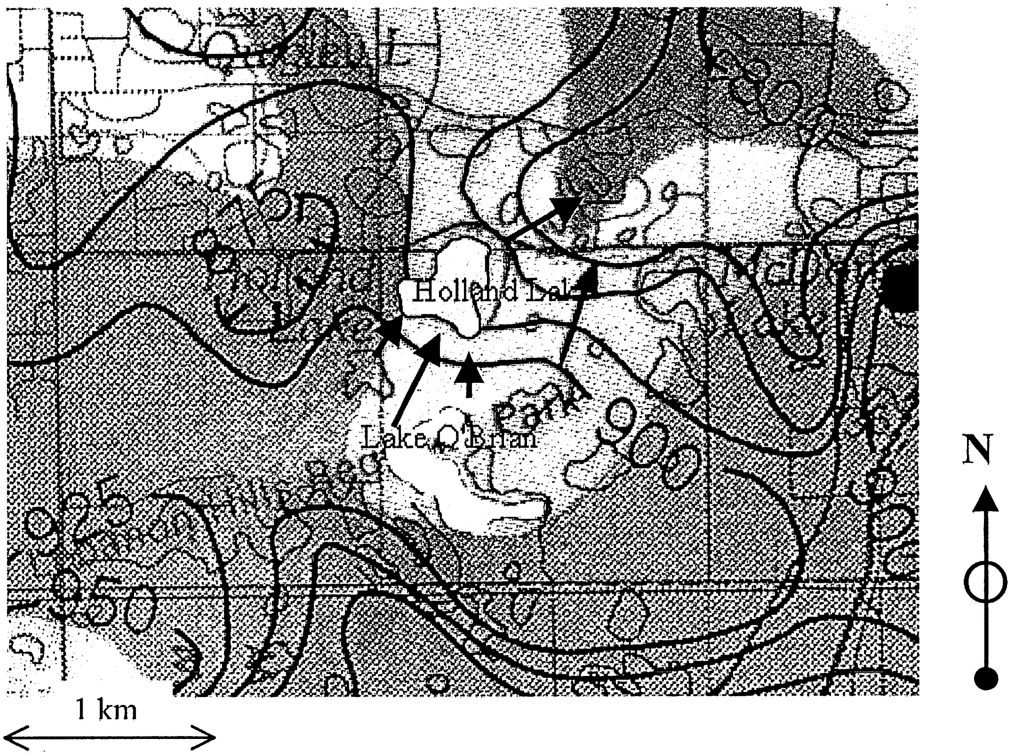


Figure 16. Hydrogeological map of the region around Holland Lake [after *Minnesota Geological Survey*, 1990]. The lake is between the dark arrows.

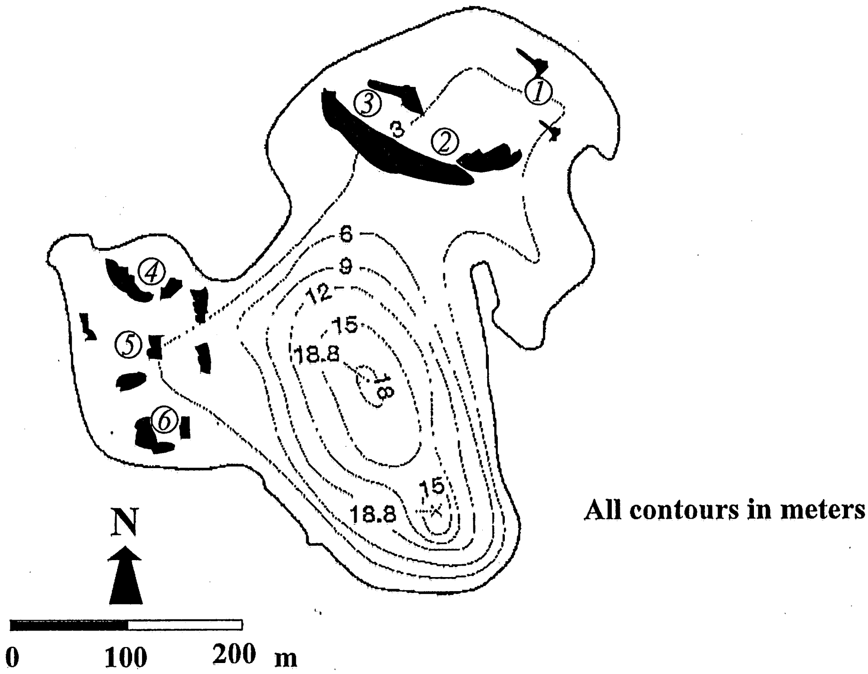


Figure 17. Locations 1 to 6 where temperature profiles were measured on August 4, 2000 in the shallow bays. The dark areas show where dense rooted macrophytes had reached the lake surface.

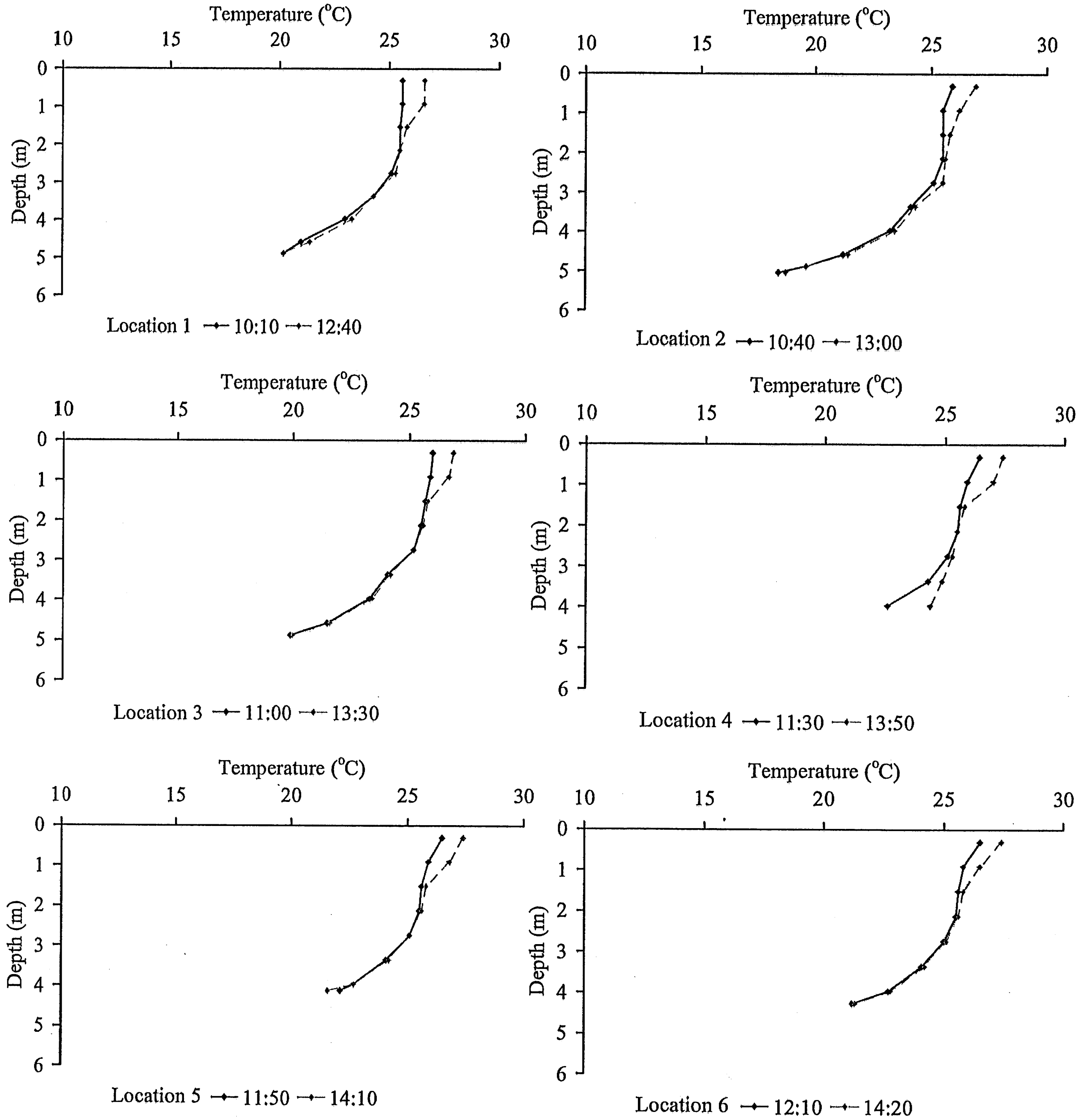


Figure 18. Temperature profiles measured on August 4, 2000 in the shallow bays.

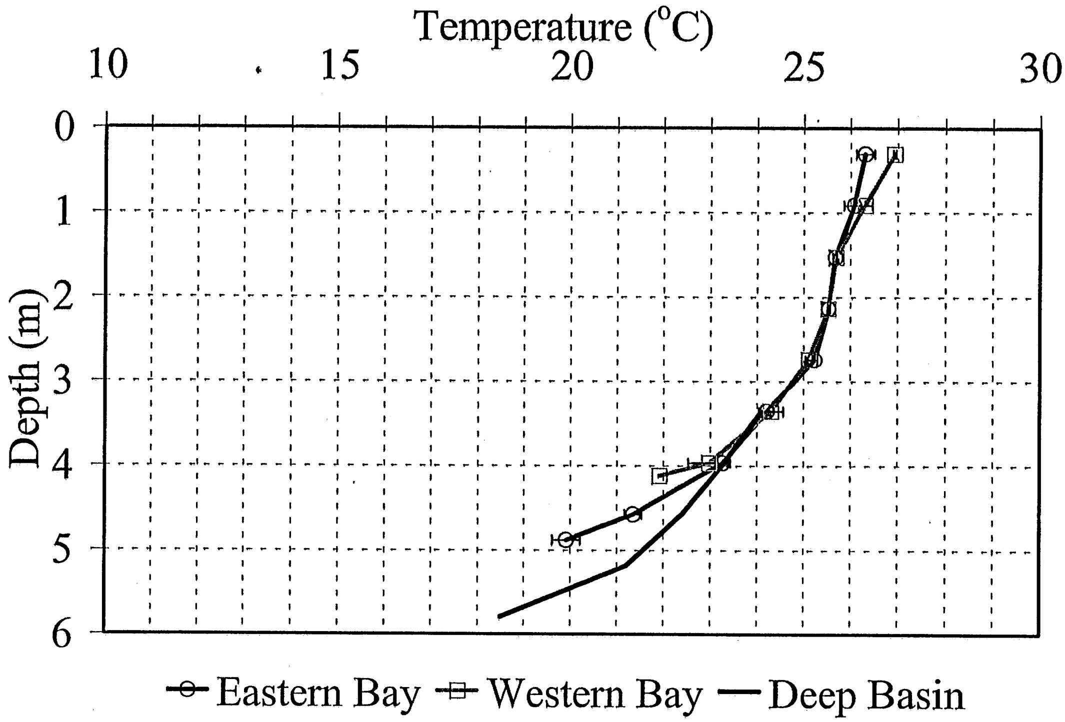


Figure 19. Average water temperature profiles measured in the basins of Holland Lake on August 4, 2000. The error bars show one standard deviation from the mean.

V. Summary and Conclusions

Holland Lake in Dakota County has been considered for stocking with brown trout. Holland Lake is suitable for this purpose because it is exceptionally deep in comparison to other lakes in the Metro Area and has relatively good water quality due to a limited drainage area. However, for a period of two to three months the strata below the surface mixed layer become anoxic. Consequently, midsummer high water temperature at the lake surface and anoxia below the surface would exert exceptional stresses on brown trout endangering their survival.

The field study conducted in the summer of 1999 concluded from lake temperature profiles, that the bottom water of the shallow bays intruded into the deep main basin as a density current. The density current transports low DO water, detritus and plant material from the shallow bays into the deep main basin. The density current could also be influenced by groundwater flow through Holland Lake.

In the summer and winter of 2000, a field study was conducted to explore the interactions between Holland Lake and groundwater. The measurements of piezometric heads in the lake sediments identified two potential groundwater inflow areas, one in the southern portion of the western shallow bay, another near the easterly shore of the eastern shallow bay (Figure 11). An additional potential inflow area to the deep basin was found near the southern shore of the lake. A strong outflow potential was seen near the northernmost point of the lake. Measurements of ice thicknesses in December 2000 confirmed the inflow area near the southern shore of the western shallow bay, and added another near the northern shore of the same bay. Sediment temperatures measured in summer 1999 gave information about potential groundwater inflow/outflow more or less similar to the piezometric heads measured in 2000. However, the warm sediment temperatures measured in 1999, and the significant changes in the piezometric heads from August to October 2000, suggest that any subsurface inflows to the shallow bays of Holland Lake are either intermittent near-surface flows (caused by rainfall events) or from a shallow aquifer possibly connected to other lakes in the area, or from both.

Water budget computations based on the observed rise and subsequent drop of the lake stage after a major rainfall/flood event in July 2000, gave an average estimate of 8,500 $m^3.day^{-1}$ for groundwater flow through Holland Lake. That figure is not very different from the 9,750 $m^3.day^{-1}$ value estimated previously by another cruder method (see Appendix A).

The DO budget of the lower metalimnion showed that only a small portion of the groundwater may enter the lower metalimnion. Most of the groundwater flow appears to enter the shallow bays. During the thermal stratification of these bays due to the presence of dense rooted macrophytes (June to August), groundwater intrudes into the deep basin by low DO density currents carrying high BOD waters from the near bed areas into the upper metalimnion of the deep basin. Groundwater leaves the lake most likely from the upper metalimnion.

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Appendix A: Advection by Groundwater (Reproduced from a Report by *Mohseni and Stefan*, 2000)

Since the low water temperatures measured at the bottom of the shallow bays relative to the deep main basin are hard to explain by shading alone, we have made an assessment of the potential groundwater flow into and out of Holland Lake. This investigation was beyond the scope of the work, but has provided important information that explains some of the inlake observations.

Holland Lake is located in glacial deposits of sand and gravel of the Rosemount outwash. It intercepts a Quaternary aquifer. Figure 32 gives elevation of the groundwater table and general groundwater flow directions in the proximity of Holland Lake [*Minnesota Geological Survey*, 1990] . The water table was determined from the static water levels reported in driller's records and observation wells. The aquifer is confined and has an average hydraulic conductivity of 40.7 m. day^{-1} (1000 gallons.ft^{-2}.day^{-1}). The thickness of the saturated layer is about 24 m (80 ft) around the lake. The potentiometric surface of the aquifer along the arrows of Figure A-1 is shown in Figure A-2. The piezometric head of the aquifer drops by 15 m (50 ft) from the southwest side of the lake to its northeast side.

An attempt was made to identify areas with groundwater inflow along the periphery of the shallow bays of the lake by measuring temperatures in the surfacial sediments and in the overlying water. About 45 temperatures were measured around the lake periphery on two occasions. Measurements 5 cm (2 inches) into sediments were from 1.5 to 3.5 °C colder than lake water, which was between 22 to 27 °C. This may be due to groundwater intrusion, but is not proof.

Another observation suggesting groundwater inflow into Holland Lake is the presence of several lakes to the west and south of Holland Lake with significantly higher lake stages (Figure A-3). In particular, O'Brien Lake has an elevation of 277 m (908 ft) a.m.s.l. compared to Holland Lake's 265 m (867 ft), and it is only 500 m away. It appears that groundwater flow from the vicinity of these lakes may converge towards Holland Lake as a focal point. Based on these characteristics a substantial amount of groundwater flow into and out of the lake can be expected.

In the Twin Cities, the mean annual precipitation and lake evaporation are about 710 mm (28 inches) and 760 mm (30 inches), respectively. Precipitation over the lake more or less compensates for evaporation. Provided no change in the lake stage occurs on an annual basis, the groundwater must be compensated by surface water and groundwater outflow. To roughly estimate the average groundwater flow through Holland Lake, the following equation for confined aquifers can be utilized

$$q = \frac{K(\phi_1 - \phi_2)}{L} \qquad \text{(A-1)}$$

where q is flow per unit area, K is hydraulic conductivity, and L is the distance between two points with known piezometric heads ϕ_1 and ϕ_2. Estimating the inflow from the piezometric head gradient of 125/12000 = 0.01 and a K value of 40.7 m. day^{-1} (133.5 ft.day^{-1}) (Figure A-2), and assuming that the groundwater from the 24 m (80 ft) thick aquifer flows all through Holland Lake gives a groundwater inflow of 11 m^3 m^{-1} day^{-1} (120 ft^3 ft^{-1} day^{-1}). Figure A-4 illustrates the approximate capture zone of Holland Lake based on the hydrogeological map of the region [*Minnesota Geological Survey*, 1990]. The flow gradient varies within the lake capture zone; it is smaller in the west and south, and reaches a maximum in the southwest. Assigning an average gradient of 0.01 for the entire capture zone, and assuming a width of 870 m (2900 ft) for the capture zone, total groundwater inflow becomes 9570 m^3 day^{-1} (348,000 ft^3 day^{-1}).

Groundwater inflow into Holland Lake is quite significant with respect to the lake volume, which is 725,000 m^3 (25,380,000 ft^3). With the above values, groundwater has a hydraulic residence time on the order of 73 days in Holland Lake. Since groundwater usually contains little or no DO, the effect of groundwater flow through a lake would be to remove DO from the lake.

A whole lake mass balance equation gives

$$C_2 = \frac{C_1\left(\forall/\Delta t\right) + C_i Q}{Q + \forall/\Delta t} \tag{A-2}$$

where C_1 and C_2 are initial and final DO concentrations of the Lake, C_i is inflow concentration, $\forall$ is the volume of the lake, Q is the groundwater inflow (which herein is assumed to be equal to the groundwater outflow), and Δt is the time interval. For an inflow concentration $C_i = 0$, the relative concentration change in the lake becomes

$$\frac{\Delta C}{C_1} = -\frac{Q\,\Delta t}{\forall} \tag{A-3}$$

For a one day time step, this value is about 1.4%. But a 1.4% decrease in the DO concentration, or 73 days of residence time is only true when groundwater mixes with the entire water of the lake, as may be the case during the fall turnover. In summer, when the lake is temperature stratified, groundwater enters only into some of the layers. The groundwater temperature is slightly above the mean annual air temperature of the region, which is about 10 °C in the Twin Cities area. Consequently, in summer, when the deep basin is stratified, groundwater enters only layers, which have a temperature greater than 10 °C. According to Figure 28 (in *Mohseni and Stefan* [2000]), groundwater must enter above a depth of 6.6 to 7.2 m (22 to 24 ft) in the deep basin. In the shallow bays, it enters the bottom of the basins and mixes with the bed layers. Low temperatures at a depth of 3.6 m (16 to 18 °C) in the shallow bay are likely associated with groundwater inflow in addition to shading (Figure 30 in *Mohseni and Stefan* [2000]). In summer, the groundwater residence times must be estimated using the thickness of the layers, with which the groundwater is mixing. Table A-1 gives estimates of the groundwater residence times using the capture zone in Figure A-4. For each bay, an appropriate width of the capture zone is assigned based on the figure.

Table A-1. Estimates of the groundwater residence times in bays of Holland Lake in summer.

Bay	Inflow Width (m)	Inflow (m^3. day^{-1})	Inflow depths	Volume (m^3)	Residence time (day)
Western shallow bay	375	45,000	Bottom 0.9 m	11,100	0.3
Eastern shallow bay	135	16,200	Bottom 0.9 m	47,500	2.9
Deep basin	375	45,000	5.4-7.2 m depth	367,000	8.2

The estimates of the groundwater residence times in Table A-1 are only crude estimates, however, they indicate that the groundwater inflow may have a significant impact on thermal stratification of the shallow bays as well as DO dynamics. It also implies that, it is most likely that the western shallow bay has a colder bottom temperature than the eastern shallow bay. However, there are no measurements from the western shallow bay to verify this potential difference.

One can summarize the estimates presented in Table A-1 as follows: Groundwater enters the bottom of the western shallow bay with a residence time of less than a day and decreases the bed water temperatures. The bottom water in the shallow bays is cooler and therefore denser than the surface water and enters the deep basin as density current. The density current intrudes into a region with the same temperature (4.2 m to 4.5 m (14 to 15 ft) depth in July). High concentrations of TSS and TVSS were measured at those depths in the deep basin. They may be attributed to plant material carried by the density current from the shallow bays into the deep basin. Since the groundwater usually carries little or no DO it may gradually displace oxygen richer water and replace it by water with a high oxygen demand from the shallow bays. If the groundwater comes from the upper lakes it will have some DO in it.

Overall it is believed that groundwater flow into Holland lake contributes to the observed DO dynamics especially the DO depletion in the upper metalimnion by July as shown in section V.1 (in *Mohseni and Stefan* [2000]). The second spike in the TSS and the TVSS profiles at a depth of 7.2 m (24 ft) may be associated with direct groundwater inflow to the deep basin. Resuspension of sediments at those depths in the deep basin would increase the TSS.

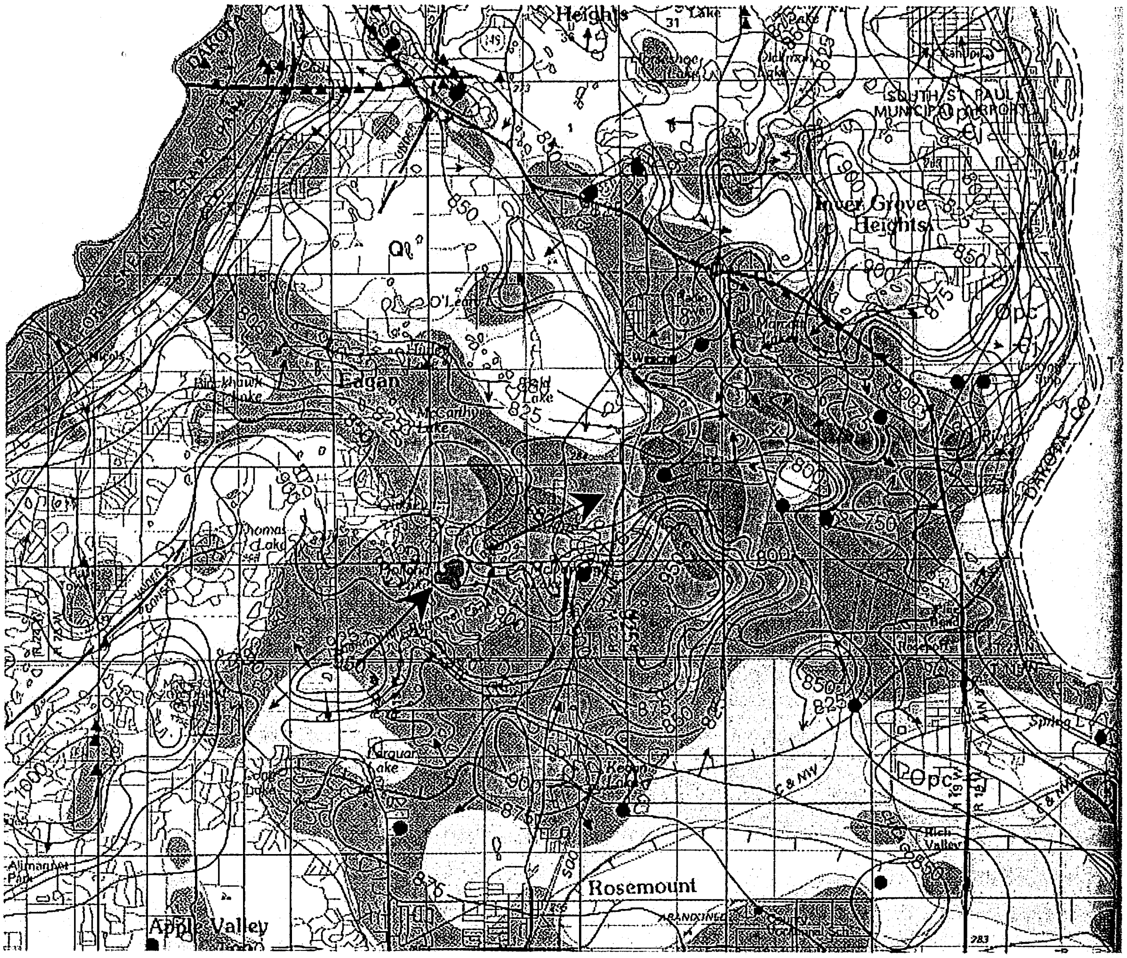

Figure A-1. Hydrogeological map of the region around Holland Lake [after *Minnesota Geological Survey*, 1990]. The lake is between the black arrows.

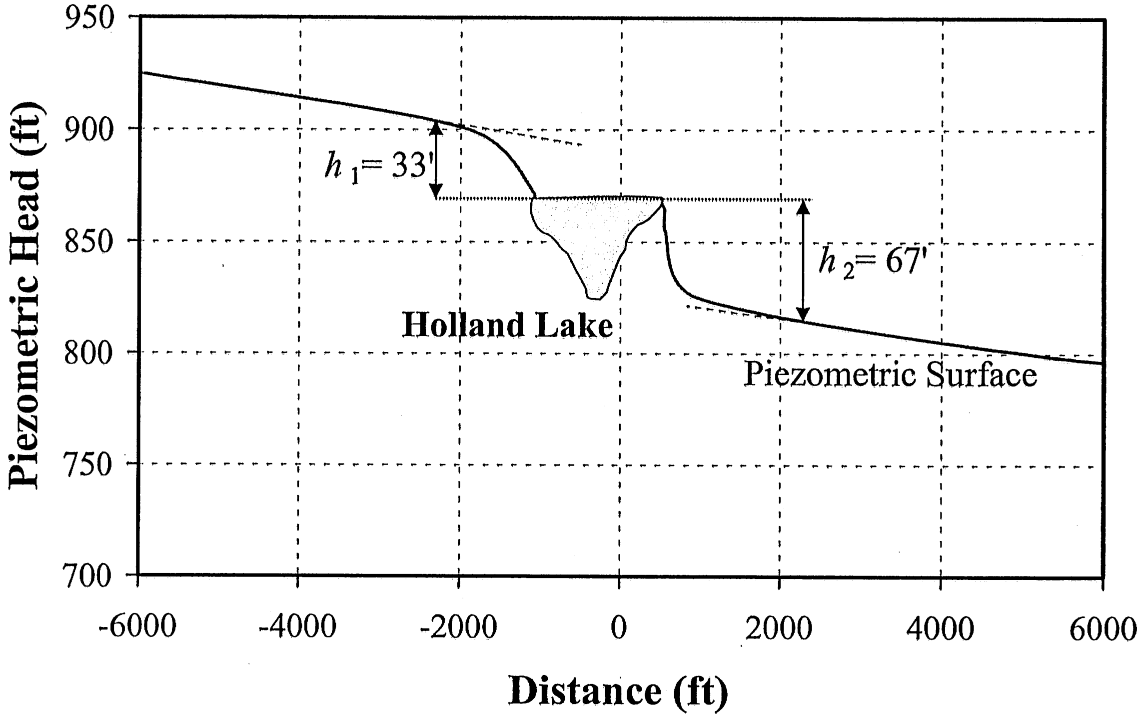

Figure A-2. Piezometric groundwater surface of the Rosemount outwash in a cross section through Holland Lake (elevation data are from the *Minnesota Geological Survey*, 1990).

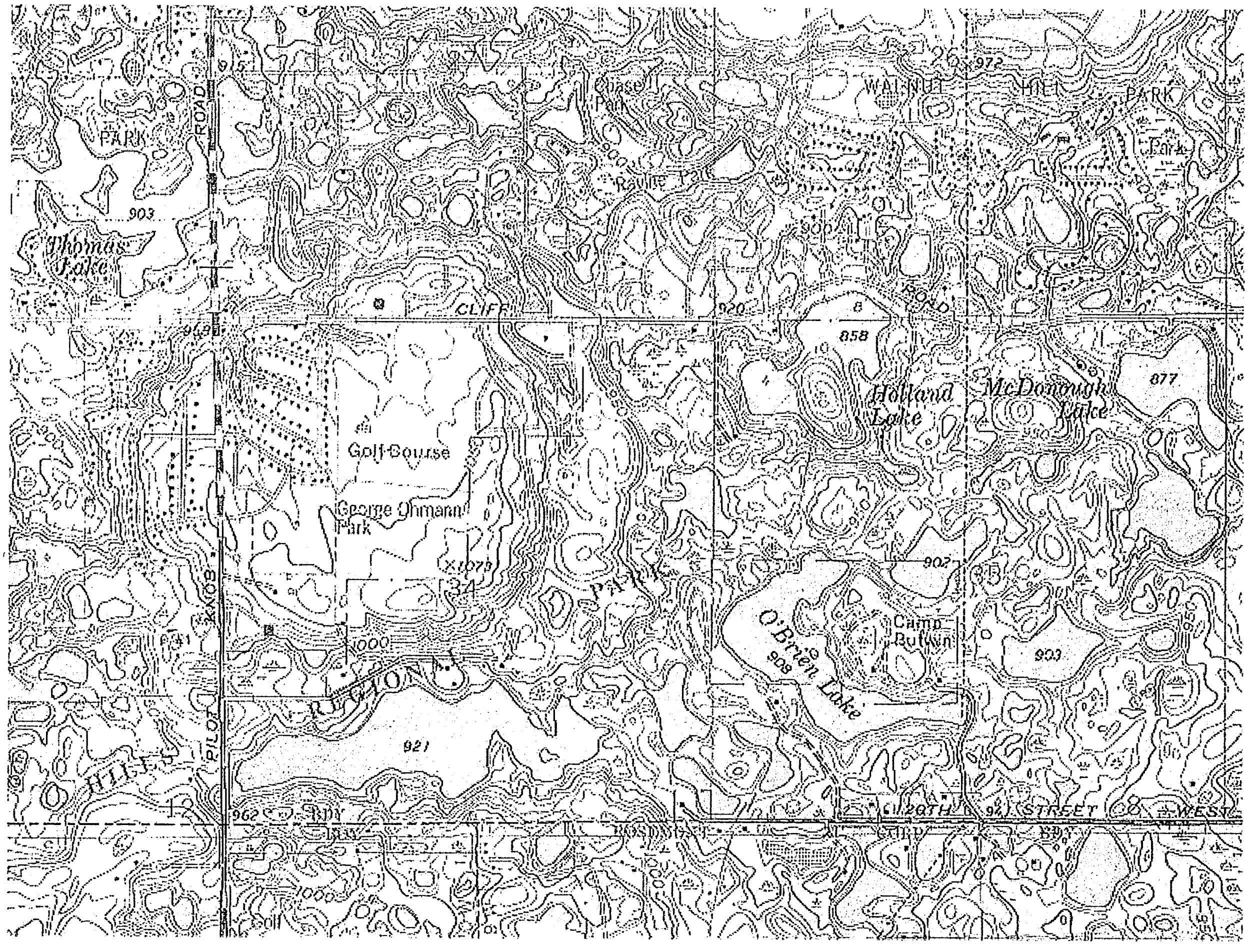

Figure A-3. Topographical map of the vicinity of Holland Lake with lake surface elevations.

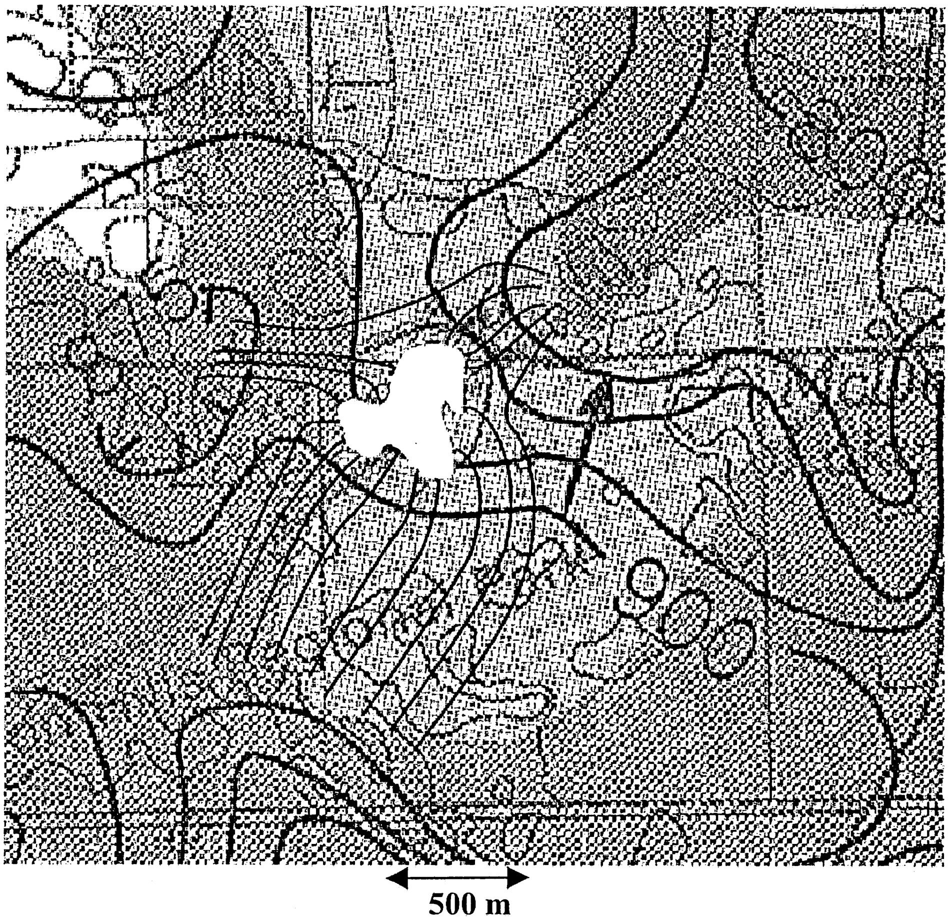

Figure A-4. The Holland Lake capture zone of the aquifer. Approximate groundwater streamlines and equal piezometric head lines around Holland Lake showing the capture zone of groundwater.

Appendix B. Flood of July 2000

Figure B-1 shows Holland Lake and the surrounding lakes managed by the Dakota County Parks Department. During the flood of July 2000, excess water in Cattail Lake and McDonough Lake was pumped and diverted towards Holland Lake, as indicated by the arrows.

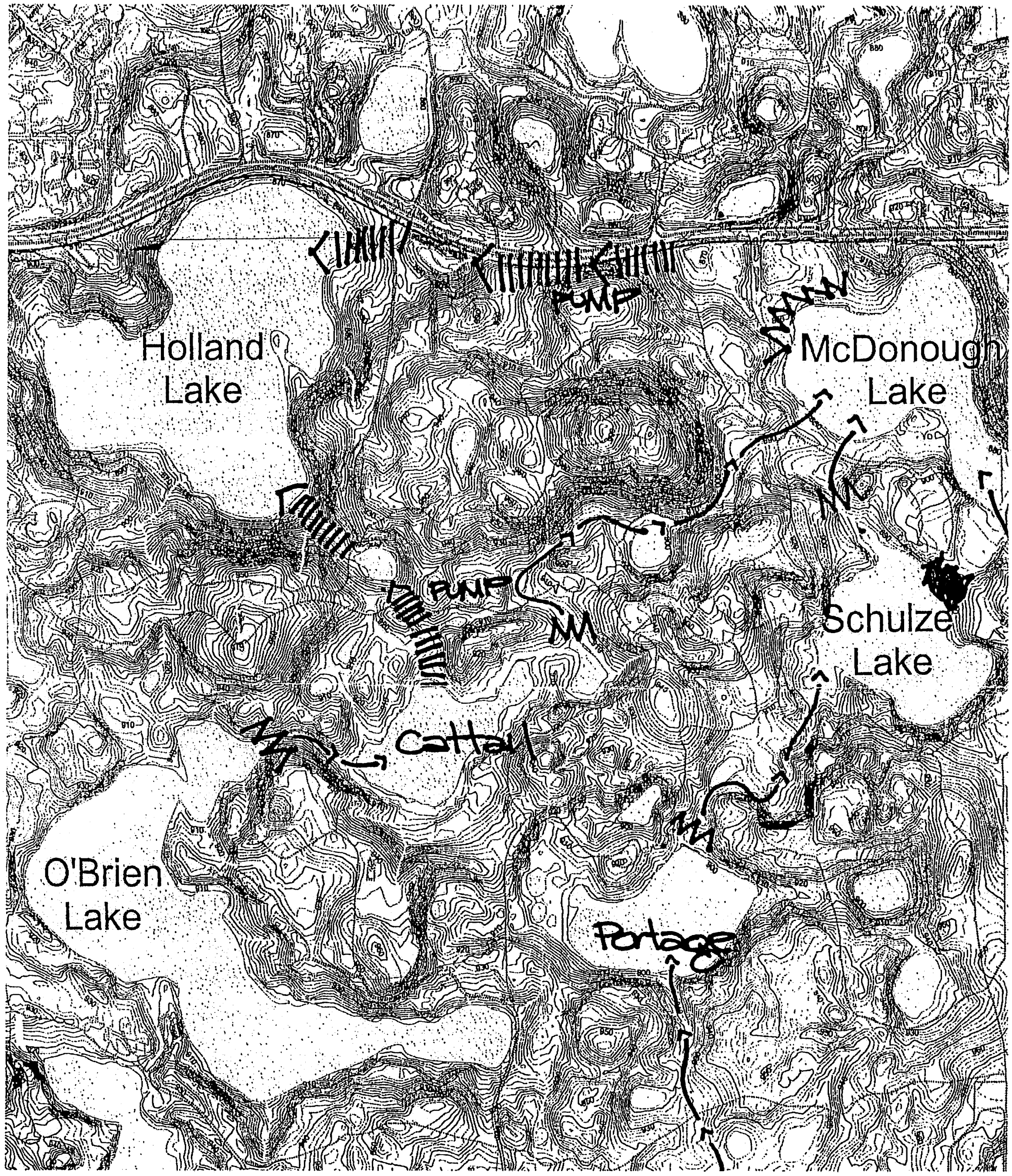

Figure B-1. Diversion of the excess water from the surrounding lakes to Holland Lake, July 2000 (prepared by a Consultant).

Appendix C. Estimating Lake Evaporation

Lake evaporation can be estimated using the Penman equation [*Shuttleworth*, 1993]

$$PE = \frac{1}{lv}\left[\frac{\Delta}{\Delta+\gamma}Rn + \frac{\gamma}{\Delta+\gamma}6.43(1+0.536U)(e_s^* - e_d^*)\right] \qquad \text{(C-1)}$$

where PE is lake evaporation in mm day^{-1}, lv is the latent heat of evaporation (MJ kg^{-1}), Δ is the gradient of saturation vapor pressure (kPa °C^{-1}), γ is the psychrometric constant (kPa °C^{-1}), Rn is net radiation (MJ m^{-2} day^{-1}), U is the wind velocity (m sec^{-1}) measured at 2 m above the surface, e_s^* is the saturation vapor pressure at the surface (kPa), and e_d^* is the atmospheric vapor pressure or the saturation vapor pressure at dew point temperature (kPa). The parameters in equation C-1 were estimated using the weather data collected at the SAFL weather station.

Latent heat of vaporization is a function of temperature and was estimated using the following equation [*Harrison*, 1963]

$$lv = 2.501\text{-}0.002361\ T_s \qquad \text{(MJ kg}^{-1}\text{)} \qquad \text{(C-2)}$$

where T_s is surface temperature in °C. Since the surface temperatures were not measured on a daily basis in Holland Lake, air temperature T_a was used to replace T_s.

Net radiation Rn is the sum of shortwave and longwave radiation. Shortwave or solar radiation was measured at the SAFL weather station, and net longwave radiation H_l was estimated using the following equation [*Shuttleworth*, 1993]

$$H_l = -cl \in' \sigma (T_a + 273.2)^4 \qquad \text{(MJ m}^{-2}\text{ day}^{-1}\text{)} \qquad \text{(C-3)}$$

where cl was the cloudiness factor and was estimated from [*Wright and Jensen*, 1972]

$$cl = a_c \frac{H_s}{H_{ext}} + b_c \qquad \text{(C-4)}$$

where a_c and b_c are regression values to be determined in local studies of longwave radiation. For humid areas, a_c and b_c are estimated to be 1.35 and –0.35, respectively. H_s is the measured solar radiation, and H_{ext} is the extraterrestrial solar radiation, which can be estimated from the Julian day and the latitude of the region. The parameter $\in'$ is the net emissivity and can be estimated from [*Allen et al.*, 1989]

$$\in' = a_e + b_e\sqrt{e_d^*} \qquad \text{(C-5)}$$

where a_e and b_e are correlation coefficients with average values of 0.34 and –0.14, respectively [*Doorenbos and Pruitt*, 1975].

e_s^* and e_d^* were calculated using air temperature and dew point temperature, respectively, and the following equation [*Tetens*, 1932]

$$e^{*} = 0.611\exp\left(\frac{17.27T}{237.3+T}\right) \qquad \text{(C-6)}$$

Wind speed varies with the topography and land use, i.e. it varies as surface roughness changes. There is also a significant spatial variability of wind speed over a lake due to the sheltering effects. An example of the spatial variability of wind speed is shown in Figure C-1, which gives wind speed over Lake McCarrons in Roseville, MN, at 2 m above the surface and at the MSP International Airport at 10 m above the surface. In Figure C-1, the wind speeds at the MSP International Airport are adjusted for the height of 2 m above the surface using the logarithmic velocity distribution law. The two sites are only about 20 km apart. As is evident, wind speeds at the airport are about an order of magnitude larger than those over Lake McCarrons. Since sheltering is due to both the topography and the trees around a lake, it is quite difficult to estimate the wind speeds over Holland Lake. In this report, the wind speeds measured at the MSP International Airport were taken for Holland Lake evaporation calculations. Therefore, it is likely that the Holland Lake evaporation is overestimated.

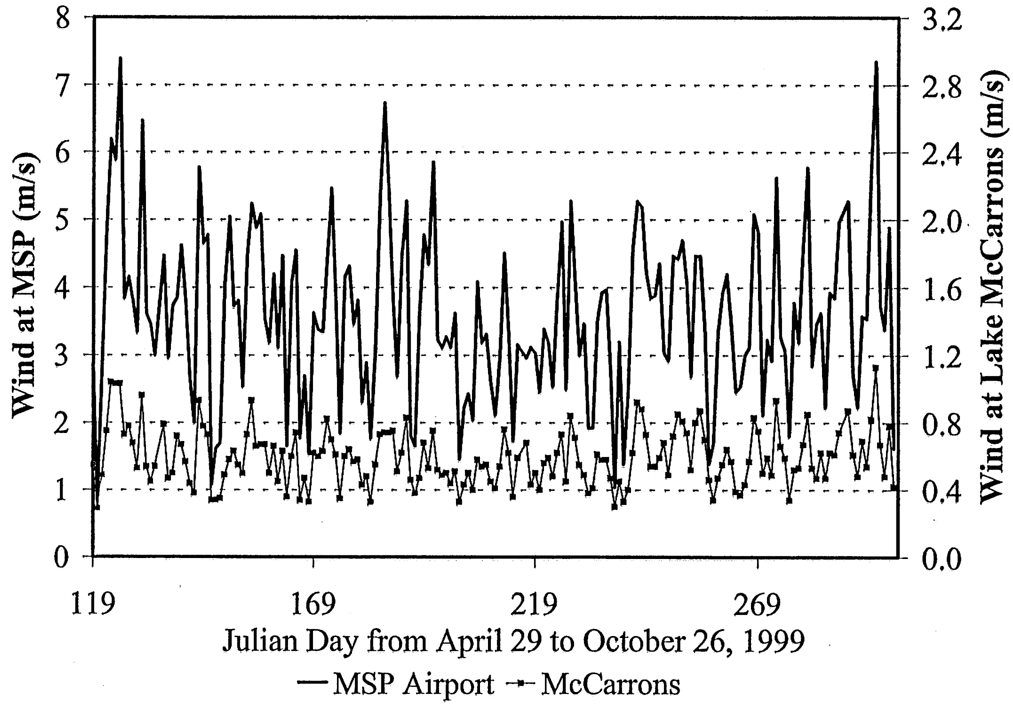

Figure C-1. Measured wind speeds over Lake McCarrons and at the Minneapolis/St. Paul International Airport at 2 m above the surface.