Hydrology Prior to Wetland and Prairie Restoration in and around the Glacial Ridge National Wildlife Refuge, Northwestern Minnesota, 2002–5

By Timothy K. Cowdery and David L. Lorenz, with Allan D. Arntson

In cooperation with The Nature Conservancy and the Red Lake Watershed District

Scientific Investigations Report 2007–5200

U.S. Department of the Interior
U.S. Geological Survey

U.S. Department of the Interior
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Suggested citation:
Cowdery, T.K., and Lorenz, D.L, with Arntson, A.D., 2008, Hydrology prior to wetland and prairie restoration in and around the Glacial Ridge National Wildlife Refuge, northwestern Minnesota, 2002–5: U.S. Geological Survey Scientific Investigations Report 2007–5200, 68 p.

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Conversion Factors

Multiply	By	To obtain
	Length	
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
acre	0.4047	hectare (ha)
square mile (mi^2)	2.590	square kilometer (km^2)
	Volume	
gallon (gal)	3.785	liter (L)
milliliters (mL)	0.0.03382	ounce, fluid (fl. oz.)
million gallons (Mgal)	3,785	cubic meter (m^3)
acre-foot (acre-ft)	1,233	cubic meter (m^3)
cubic foot per second-day (CFS-day)	2,447	cubic meter (m^3)
	Flow, precipitation, evapotranspiration, and recharge rates	
acre-foot per year (acre-ft/yr)	1,233	cubic meter per year (m^3/yr)
cubic foot per second (ft3/s)	0.02832	cubic meter per second (m^3/s)
gallon per minute (gal/min)	3.785	liter per minute (L/min)
million gallons per year (Mgal/yr)	3,785	cubic meter per year (m^3/yr)
inch per day (in/d)	25.4	millimeter per day (mm/d)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
foot per second (ft/s)	0.3048	meter per second (m/s)
	Hydraulic conductivity	
foot per day (ft/d)	0.3048	meter per day (m/d)
	Hydraulic gradient	
foot per mile (ft/mi)	0.1894	meter per kilometer (m/km)
	Transmissivity*	
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)

Temperature in degrees Fahrenheit (°F) may be converted to degrees Celsius (°C) as follows:

$$°C=(°F-32)/1.8$$

Vertical coordinate information is referenced to the North American Vertical Datum of 1988 (NAVD 88). Elevation, as used in this report, refers to distance above the vertical datum.

Specific conductance is given in microsiemens per centimeter at 25 degrees Celsius (µS/cm at 25 °C).

Concentrations of chemical constituents in water are given either in milligrams per liter (mg/L) or micrograms per liter (µg/L).

Sample volumes are given milliliters (mL), and filter pore sizes are given in micrometers (µm).

Transmissivity: The standard unit for transmissivity is cubic foot per day per square foot times foot of aquifer thickness [(ft^3/d)/ft^2]ft. In this report, the mathematically reduced form, foot squared per day (ft^2/d), is used for convenience.

Water year: A water year in this report is from October 1st through September 30th.

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Abstract

The Nature Conservancy (TNC) owned and managed 24,795 acres of mixed wetland, native prairie, farmland and woods east of Crookston, in northwestern Minnesota. The original wetlands and prairies that once occupied this land are being restored by TNC in cooperation with many partners and are becoming part of the Glacial Ridge National Wildlife Refuge. Results of this study indicate that these restorations are likely to have a substantial effect on the local hydrology.

Water occurs within the study area on the land surface, in surficial aquifers, and in buried aquifers of various depths, the tops of which are 50 to several hundred feet below the land surface. Surficial aquifers are generally thin (about 20 feet), narrow (several hundred feet), and long (tens of miles). Estimates of the horizontal hydraulic conductivity of surficial aquifers were 2.7–300 feet per day. Buried aquifers underlie much of the study area, but interact with surficial aquifers only in isolated areas. In these areas, water flows directly from buried to surficial aquifers and forms a single aquifer as much as 78 feet thick. The surface-water channel network is modified by several manmade ditches that were installed to remove excess water seasonally and to drain wetlands. The channels of the network lie primarily parallel to the beach ridges but cut through them in places. Back-beach basin wetlands delay and reduce direct runoff to ditches.

Recharge to the surficial aquifers (10.97–25.08 inches per year during 2003–5) is from vertical infiltration of rainfall and snowmelt (areal recharge); from surface waters (particularly ephemeral wetlands); and from upward leakage of water from buried aquifers through till confining units (estimated at about 1 inch per year). Areal recharge is highly variable in space and time. Water leaves (discharges from) the surficial aquifers as flow to surface waters (closed basins and ditches), evapotranspiration, and withdrawals from wells. Unmeasured losses (primarily discharge to ungaged (closed) basins) were 53–115 percent of areal recharge during 2003–5, while discharge to ditches that leave the study area was 17–41 percent. Discharge over 100 percent of areal recharge indicates a loss in ground-water storage. During the dry year of 2003, substantial ground water (about one-third of annual areal recharge) was released from aquifer storage but was replenished quickly during the subsequent normal year. Shallow ground-water flow is complex, with water in surficial aquifers, ditches, and wetlands part of a single hydrologic system. The ages determined for surficial ground-water samples were less than 15 years old, and one-third (8 of 24) were less than 5 years old, substantiating the close connection of surficial ground water to the land surface.

During the study, 68–81 percent of water left the area through unmeasured surface-water losses (primarily evapotranspiration), which is 2- to 4-times that leaving through the ditch system. Base flow in ditches (ground-water discharge) was 30 to 71 percent of all ditch flow. Mean annual runoff in all gaged basins except SW3 (2.26 inches per year) was similar (3.69–4.12 inches per year).

The quality of water samples from surficial aquifers and surface water collected in the study area was generally suitable for most uses but was variable. Most ground- and surface-water samples were dominated by calcium, magnesium, and bicarbonate ions. About one-quarter of surficial ground-water samples contained nitrate at concentrations greater than the U.S. Environmental Protection Agency's (USEPA) Maximum Contaminant Level for human consumption. The median concentration of dissolved phosphorus ranged from 0.0108 milligrams per liter as phosphorus (mg/L-P) to 0.0293 mg/L-P. Nutrient concentrations in ditches were generally above the USEPA nutrient guidelines for reference streams in the area. Water samples contained detectable concentrations of atrazine, acetachlor, metolachlor, pendimethalin, prometon, and terbutryn and 11 of the 19 degradates analyzed. In general, degradates were found more frequently and at higher concentrations than were the parent herbicides. No herbicide or degradate was detected in water samples from buried aquifers, reflecting the protection that clay-rich confining units afford these aquifers.

The restoration of wetlands and prairies in the study area likely will result in more water retained on the land and improved water quality. Increased water retention could raise ground-water levels, but the rise likely would be very local

and short-lived. Restorations likely would substantially change ditch-flow characteristics in the study area, but the changes would be insubstantial further downstream. Reduction in agriculture should result in a net decrease in nutrient and pesticide load to the study area.

Effects of the wetland and prairie restorations could be measured in the future, when restorations are complete and the hydrologic system has had time to equilibrate. A comparison between a future assessment and the one documented in this report would quantify the hydrologic changes resulting from wetland and prairie restorations in the Glacial Ridge study area.

Introduction

In 2002, The Nature Conservancy (TNC) owned and managed 24,795 acres of mixed wetland, native prairie, farmland, and woods east of Crookston, in northwestern Minnesota (figs. 1 and 2). Before settlement, the land was a poorly drained part of the eastern edge of Pleistocene glacial Lake Agassiz composed of a series of sandy beach ridges separated by interbeach wetlands. In 2002, much of the land was artificially drained by ditches and under cultivation. TNC, in partnership with 11 national, state, local, and private organizations, is restoring much of this land to the original wetlands and native prairies with the aim of improving water quality, reducing flooding, and improving wildlife habitat. In 2004, the U.S. Fish and Wildlife Service established the Glacial Ridge National Wildlife Refuge and is incorporating TNC land with various state and national public lands into a planned 30,000-acre contiguous block of protected and managed natural lands. Wetland restoration will involve blocking, modifying, or removing many ditches; recreating original wetland basins; and reintroducing and managing original floral communities. There are concerns that these restorations may negatively affect nearby land or interests. Specific concerns include the following:

- The local water table will rise and inundate neighboring properties still in agricultural production.
- Unique natural features, such as calcareous fens, may be damaged.
- Rewetting of drained wetlands may degrade water quality by releasing accumulated mercury, pesticides, or pesticide degradates.
- The water quantity and quality of the City of Crookston well field, which is within the restorations boundaries, may be degraded.

To address the concerns above, guide restoration activities, and document the hydrologic changes resulting from these restorations, the U.S. Geological Survey (USGS), in cooperation with TNC and the Red Lake Watershed District, conducted a hydrologic investigation of a 124,000-acre study area (fig. 1) that included TNC and nearby public lands (fig. 2). The general objectives of this investigation were to

(1) describe the hydrology of the study area sufficiently to allow managers to make informed wetland and prairie restoration decisions, and

(2) document the pre-restoration hydrologic condition in the study area, against which to compare future hydrologic conditions resulting from restorations.

The study concentrated on near-surface ground water and surface water and their interactions because they are in direct contact with the land being changed. Investigation of buried aquifers in the area was limited to understanding how these deeper waters affect the surficial hydrology.

This report presents the results of the cooperative study to describe the hydrology prior to wetland and prairie restorations in and around the Glacial Ridge National Wildlife Refuge. It describes the conceptual basis of the study and its methods, how water moves in the study area, what the water quality is, and how flow and quality may change as the land is restored. Data were collected for this study during October 2002–September 2005. Some water-level, -flow, and -quality data continue to be collected (fall, 2007) by the USGS through a separate cooperative study; a discussion of these data are beyond the scope of this report.

Two previous major hydrologic reports discuss all or part of the study area. A USGS hydrologic atlas of the Red Lake River watershed (Bidwell and others, 1970) contains a map of the beach ridge aquifers and a cross section based on borings along U.S. Highway 2. The atlas describes the regional patterns of ground-water and surface-water flow and water quality, although it provides no details at the scale of the current study. A 1996 USGS report on water in glacial aquifers in northwestern Minnesota (Lindgren, 1996) provides some details on the hydrogeology of the study area. The report of this regional study contains maps of areal extent, thickness, and transmissivity of surficial and buried aquifers and water-quality results for six ground-water samples. The study also produced four detailed ground-water models, one of which (area C) simulated flow in the beach ridges and an underlying confined aquifer in the central north-south third of the study area. Several other reports by academicians and consultants contain hydrologic data in the study area. Noteworthy among these are a report by Svedarsky (1992) on the Burnham Creek flood-control impoundment and several unpublished consultant reports available from the Water Department of the City of Crookston from their exploration for new ground-water supplies.

Hydrologic Setting

The Glacial Ridge study area is on the eastern shore of what was glacial Lake Agassiz from about 11,600 to about 9,500 years ago. This lake formed as the Laurentide Ice Sheet retreated north of the continental divide at Browns Valley, Minn., about 30 mi southwest of the northeastern corner of

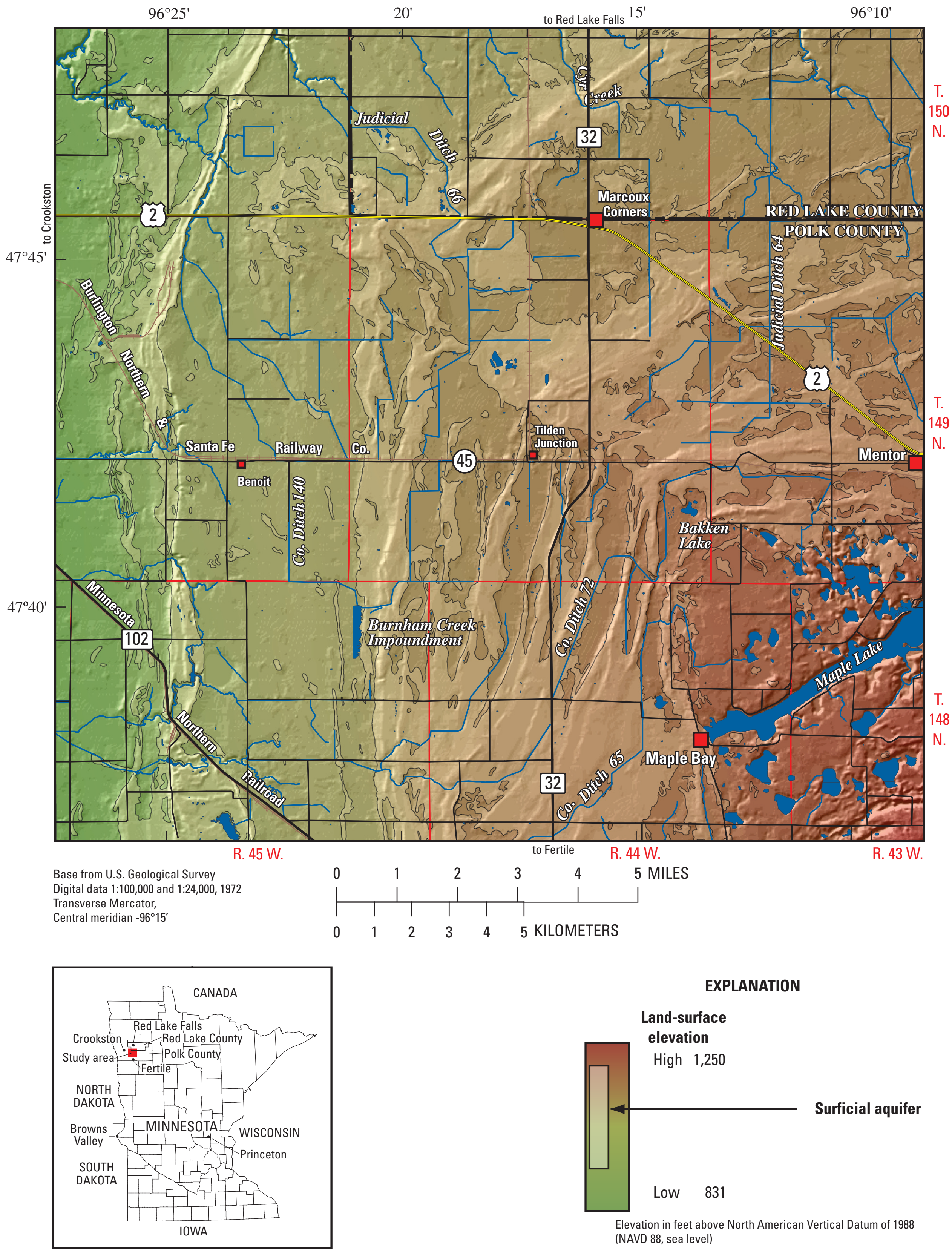
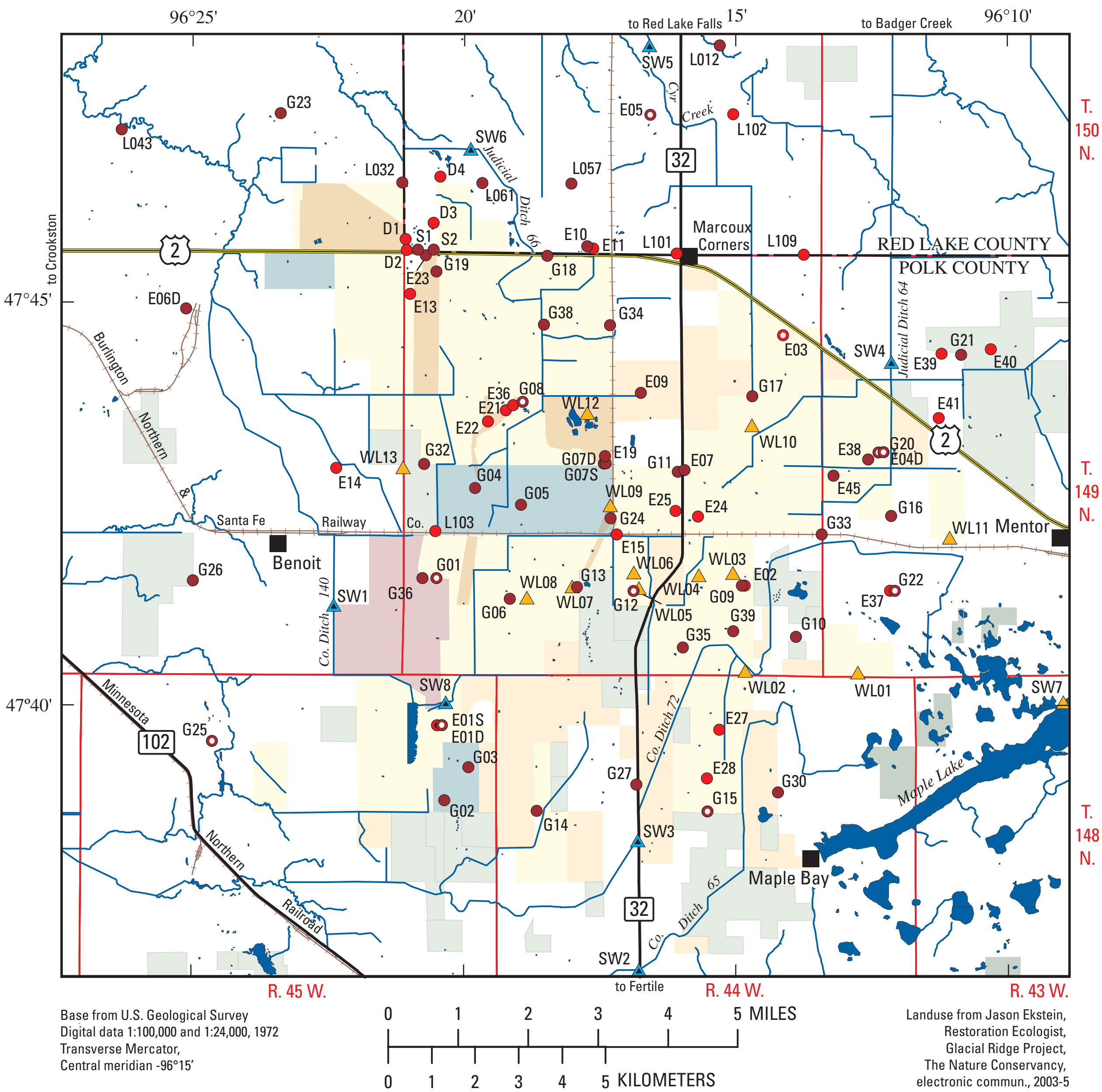


Figure 1. Topography, and surficial aquifer extent, Glacial Ridge study area, northwestern Minnesota.



EXPLANATION

Land ownership and use

- National Wildlife Refuge
- National Waterfowl Production Area
- State wildlife land
- Scientific and natural area
- The Nature Conservancy land
- Gravel lease
- Wetland Reserve Program
- Other, mostly private, agriculture

Water-level networks
(sites with site numbers)

- E01S Recorder, surficial aquifer
- E01D Recorder, buried aquifer
- G23 Synoptic, surficial aquifer
- D4 Synoptic, buried aquifer
- SW5 Recorder, ditch gage
- WL08 Synoptic, lake and wetland

Figure 2. Land-use and water-level networks, Glacial Ridge study area, northwestern Minnesota, 2005.

South Dakota (Teller, 1987). The position of the study area in relation to this glacial lake continues to be the main control on the flow of water within and from the study area. East and south of the study area, the land rises, built upon hundreds of feet of till and other sediments deposited by glaciers or their meltwaters (hereinafter called glacial sediments). In the study area, the lake curved around these morainal uplands, forming Lake Agassiz proper to the south and west and the Koochiching Arm to the north and east. After the lake drained with the deglaciation of Hudson Bay, the Red River of the North formed in the bottom of the Lake Agassiz Basin, and the Red Lake River formed in the bottom of the Koochiching Arm Basin. Ground- and surface-water flow radiates from the morainal uplands in the southeast quadrant of the study area toward the Red and Red Lake Rivers to the west and north, respectively.

Sands and gravels buried within fine-grained glacial sediments form confined (buried) aquifers within the study area. The areal extent and interconnectedness of these aquifers are poorly known. Sands and gravels at the surface were winnowed from and deposited on glacial tills as the waves of Lake Agassiz created beaches around the morainal uplands. These beaches form most of the surficial aquifers at Glacial Ridge, with the highest and oldest beaches to the southeast and the lowest and youngest to the west and north. Generally, surficial aquifers are as much as 35 ft thick, hundreds of feet wide and continuous along their length for tens or hundreds of miles. In several places, however, these surficial sands and gravels were deposited upon preexisting sands and gravels, creating localized surficial aquifers that are as thick as 78 ft and hydrologically connected to some buried aquifers.

Originally, a complex of wetlands and wet prairies developed on till or lake sediments in swales between sets of beach ridges. Surface-water flow was originally diffuse, flowing through these wetlands parallel to and behind the beach ridges until a low area allowed the flow to cut across a ridge and join the adjacent interbeach swale. These wetlands were partially drained by ditches in the early 20th century. Smaller ditches simply channelized the original drainage. Larger ditches flow perpendicular to the beach ridges and are deeply incised where they cross ridges. The resulting ditched drainage routes most of the flow at Glacial Ridge nearly at right angles to the original flow. Ditches in the study area are located at the top of their watersheds and transmit flow that is highly variable and that often ceases in the winter or late summer. A detailed land-use history of the study area is in appendix 1. This history helps put into perspective the current (2005) hydrologic conditions of the Glacial Ridge study area.

Climate

The climate of the study area is subhumid continental. During most of the year, the upper-level winds flow from west to east in the region, and surface winds have a predominantly westerly component. The study area has cold winters and moderately warm summers. Climate data from the High Plains Regional Climate Center (2006) for Crookston, about 16 mi west of the study area, show that the average January temperature is 4.3°F and the average July temperature is 69.5°F, and that most precipitation (13.9 in.) occurs during the growing season (May–Sept.) compared with 20.79 in. annually. This climate station has a long period of record (1890 to the present) and is useful for putting short-term climate data collected within the study area into historical perspective.

Extensive hourly climate data have been recorded since September 2001 at a Soil Climate Analysis Network (SCAN) station, operated by the U.S. Department of Agriculture (2006), Natural Resources Conservation Service (NRCS), near the center of the study area (adjacent to well G11, fig. 2). Real-time and historical data are available online (*http://www.wcc.nrcs.usda.gov/scan/site.pl?sitenum=2050&state=mn*). The precipitation sensor at the SCAN station did not function reliably from August 2005 through at least December 2005 (the latest climate data used in this report). For the purposes of the annual summary for 2005, precipitation recorded at well E03 (2 ½ mi northeast of well G11, fig. 2) was substituted for missing SCAN station data from August through December. Precipitation data also were recorded hourly at all other continuous ground-water level stations constructed for this study (10 wells, red circle with white center, fig. 2).

Precipitation varies dramatically between wet and dry periods within the study area. Multiyear droughts such as those during 1928–40 and 1984–93 have caused water shortages in the region, and wet periods such as those during 1968–74 and 1994–2005 have caused persistent flooding and drainage problems. The extreme annual precipitation totals during the years 1890–2005 for Crookston are 30.83 in. in 1941 and 9.99 in. in 1936.

Data for this study were collected during a period where precipitation was relatively dry in 2003 to relatively normal in 2004–5. The annual precipitation at Crookston during the 2003–5 was 20.46, 28.12, and 23.91 in., respectively. The SCAN station received considerably less precipitation than the station at Crookston during 2003–5 (15.86, 22.64, and 22.53 in., respectively). Assuming that long-term precipitation differences are small between these stations, the study area received about 75 percent (12th percentile) of average precipitation at Crookston in 2003 and more than average precipitation (65th percentile) in 2004 and 2005. Precipitation on a water year (WY) basis (October–September) was 14.87 in., 20.36 in., and 21.96 in. during WY 2003–5, respectively. The corresponding percentiles of Crookston precipitation are the 7th, 47th, and 65th. The maximum daily total precipitation measured at the SCAN station during 2003–2004 was 1.53 in. on June 10, 2003, and May 12, 2004. The maximum daily precipitation in 2005 for the SCAN station was 2.46 in. on August 17, 2005.

The mean potential evapotranspiration by the Thornthwaite method (1948) for the study area is about 23 in. The mean evapotranspiration by the precipitation minus runoff method (Baker and others, 1979) is about 20 in. Wisconsin-

Minnesota Cooperative Extension Service produces daily estimates of potential evapotranspiration (ET) during the growing season, which are calculated from satellite-derived measurements of solar radiation and air temperatures at regional airports (Wisconsin-Minnesota Cooperative Extension Service, 2006). The mean potential ET for the growing-season months (May–September) for 2004 and 2005 are 0.10, 0.15, 0.19, 0.13, and 0.10 in/d, respectively. July 2005 had the greatest average potential ET (0.205 in/d), with potential ET as much as 0.29 in/d on several days. Summing these values for the growing season produces a potential ET amount of 20.52 in., which is between the Thornthwaite and precipitation-minus-runoff method estimates for mean annual evapotranspiration. This comparison suggests that most of the ET in the basin occurs during the growing season, particularly during June–August. Estimates of growing-season potential ET during this study and mean annual ET are nearly equal to the mean annual precipitation at nearby Crookston (20.79 in.).

Methods

This study was designed to combine data from ground water, surface water, and water quality to produce a holistic assessment of hydrology at Glacial Ridge. Ground-water data were collected from water-well stratigraphic logs and from existing and new wells. Aquifer structure was determined by combining data from stratigraphic logs from water wells and test borings, soil surveys, geological mapping, geophysical logs, surface-water distribution and structure, water levels, and the glacial history of the study area. Surface-water data were collected at newly established gages on ditches near the study-area boundaries, located to integrate data from large parts of the study area. Data produced for this study included synoptic water-level measurements, continuously recorded water levels, water temperatures, and rainfall at wells; stratigraphy at well and test boring sites; stage, discharge, and rating at ditch gages; and elevations of measuring points at all sites. Water movement in the study area was determined from data recorded at the ditch gages and determined from water-level maps that were in turn compiled from synoptic and continuous water-level measurements made during summer 2002–fall 2005. Water quality was measured by a synoptic sampling of ground and surface waters in summer 2004. Measurements of water-quality variability began in October 2002 and continues (by the USGS, as of 2007) at a subset of the synoptic sampling sites for a subset of constituents (Tim Cowdery, project chief, U.S. Geological Survey, oral commun., 2007).

Data Sites

All data collected for this study came from sites in the networks listed in table 1. Locations of these sites are shown in figures 2 and 3, and identification information is listed in appendix 2. Water-level and water-quality networks were established in the study area. Each of these networks has a synoptic component and a variability (temporal) component. The synoptic components document the state of the water resources at a moment in time. The variability components measure variability at a subset of synoptic sites through time.

Ground-Water Sites

Thirty-six wells were installed in surficial aquifers for this study. The wells were located to provide a relatively even distribution of new and existing wells in surficial aquifers throughout the study area. Construction details of installed wells are in appendix 3. The ground-water part of the synoptic water-level network was composed of 79 new and existing wells screened in surficial (50 wells) and buried aquifers (29 wells) (fig. 2). Physical statistics of these wells are summarized in the boxplots of figure 4. A subset of these wells (39 in surficial and 9 in buried aquifers) was selected for synoptic water-quality sampling. The sampled wells were selected to give priority to water from shallow surficial aquifers, which are the more readily affected by land-use changes, while still sampling enough wells in buried aquifers to indicate the character and spatial variability of this water.

Twelve wells were selected from the synoptic water-quality network to examine the temporal variability of water

Table 1. Data networks for the Glacial Ridge study, northwestern Minnesota.

[WL, water level; WQ, water quality; WT, water temperature; ST, equipment shelter temperature (unventilated); RF, rainfall; —, none; min, minute]

Site type	Number of sites	Instruments	Continuous data collected	Synoptic WL network	Continuous WL network	Synoptic WQ network	Variability WQ network	Frequency of continuous data
Ditch gages	7	Pressure transducer	WL	7	7	6	7	15 min
Wetland gages	13	Staff gage	—	13	0	7	0	—
Lake gages	1	Staff gage	—	1	0	0	0	—
New wells	36	Pressure transducer, rain gage	WL, WT, ST, RF	36	7	34	7	60 min
Existing wells	43	Pressure transducer, rain gage	WL, WT, ST, RF	43	5	14	5	60 min

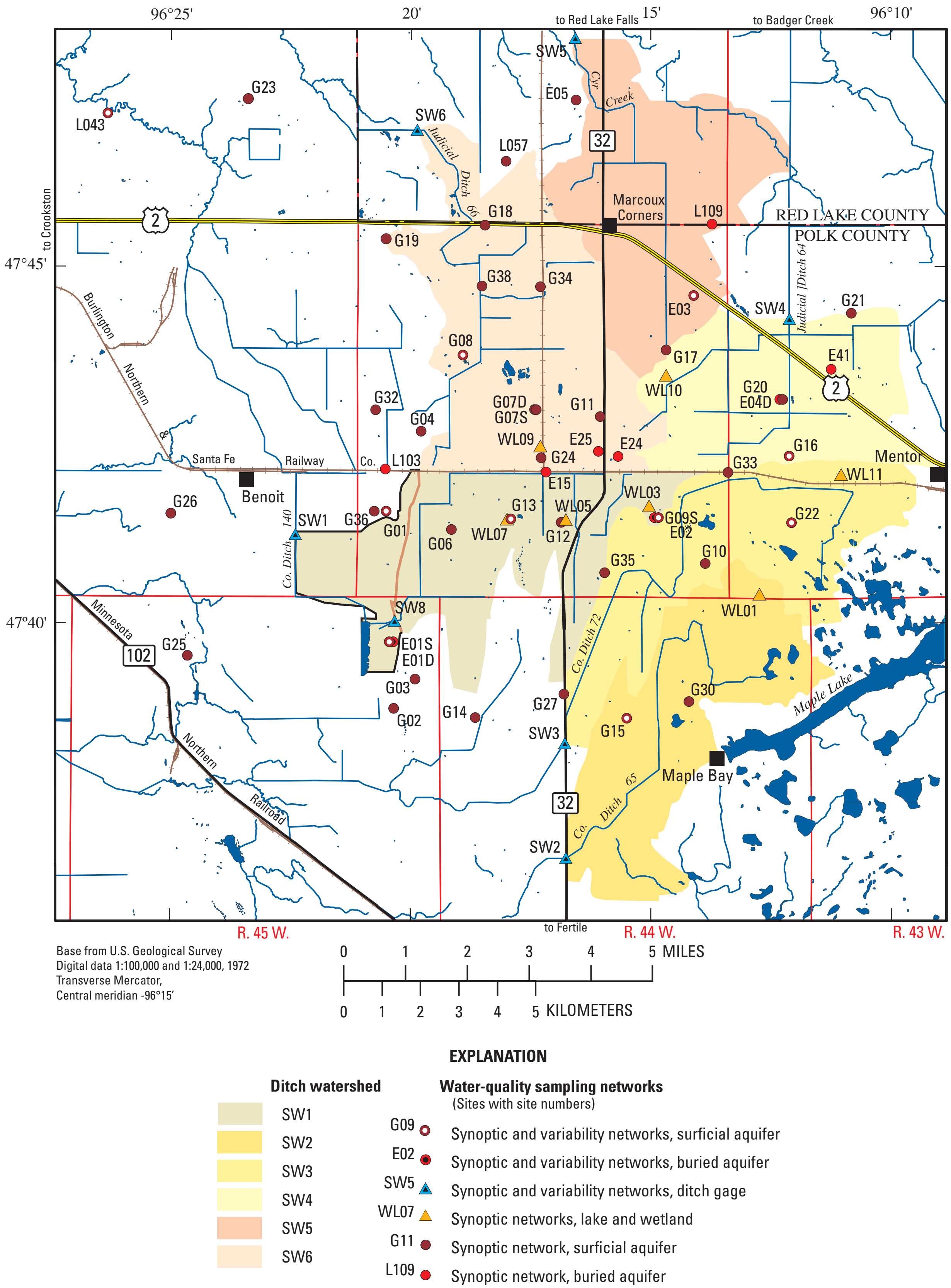


Figure 3. Ditch watersheds and water-quality networks, Glacial Ridge study area, northwestern Minnesota, 2005.

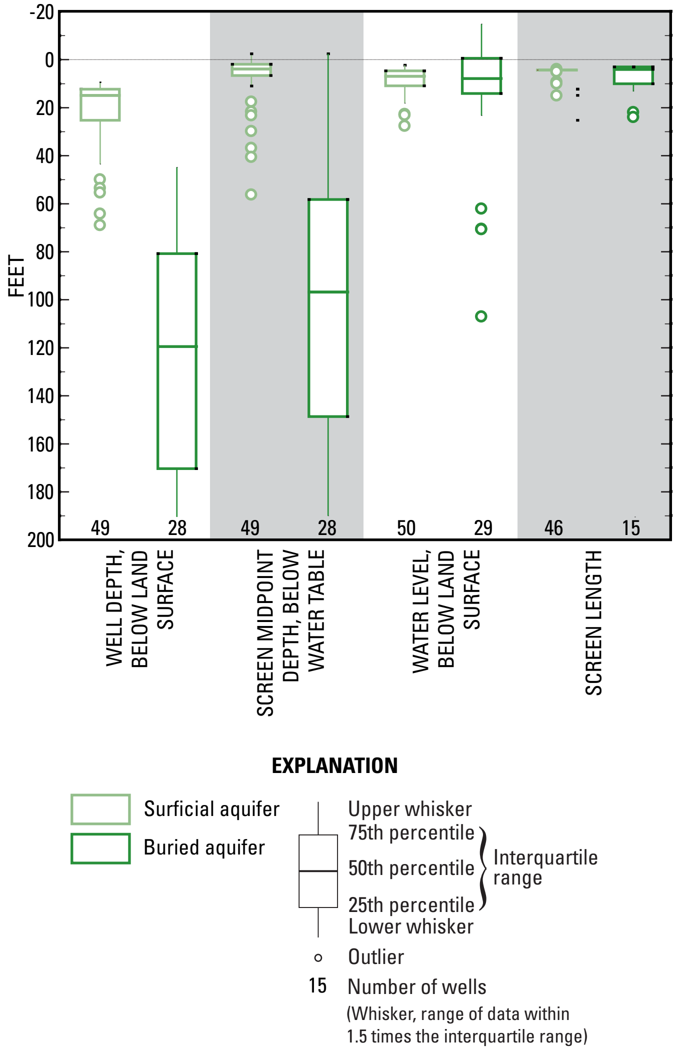


Figure 4. Characteristics of water-level network wells, Glacial Ridge study area, northwestern Minnesota.

levels and 12 wells were selected to examine temporal variability in water quality (figs. 2 and 3). Each of these variability networks contains 7 new and 5 existing wells, 10 of which were completed in surficial aquifers and 2 in buried aquifers. The variability networks have seven wells in common (six in surficial aquifers and one in a buried-aquifer). Wells for these networks were selected to cover the range of water levels and quality in the study area.

Continuous water-level wells were outfitted with submersible pressure transducers to measure water level and water temperature. These wells also had a thermister to measure instrument-shelter temperature and an unheated tipping-bucket rain gage to measure rainfall during thawed periods. Data were recorded at the well and uploaded to the USGS database daily through radio and telephone telemetry. These data are available online (*http://waterdata.usgs.gov/mn/nwis/current/?type=gw*) by way of the site numbers in appendix 2. Pressure transducers were calibrated at least every 2 months, and rain gages were calibrated yearly.

Surface-Water Sites

Seven ditch gages, thirteen wetland gages, and one lake gage were located in the study area (figs. 2 and 3). All surface-water sites were part of the synoptic water-level network. All ditch gages were part of the continuous water-level network. Six of the ditch gages and seven of the wetland gages were part of the synoptic water-quality network. The same six ditch gages were part of the variability water-quality network. Six of the ditch gages were on ditches that drain property owned by TNC. The other gage (SW2) was on a ditch adjacent to the property and was intended to gather data in a control basin that would undergo little land-use change during the project. Summary information about the ditch gages is in table 2. One ungaged ditch also drains TNC property. This ditch was not gaged because it drains very little of TNC property. Gage SW8 was installed in October 2004 above the Burnham Creek Impoundment (fig. 1) to assess the influence of that impoundment on the flow recorded at gage SW1. Data recorded at this gage are not included in this report because the period of record was less than 1 year, too short to analyze with the methods used in this report. The wetland gages were selected in deeper wetlands primarily for obtaining water levels, assuming that the wetlands are surface expressions of the ground-water surface. The lake gage (SW7) was established to measure the extreme upgradient elevation of the hydrologic system.

Selection of gage locations was guided by criteria presented by Carter and Davidian (1968). All ditch gages, except SW2, were located to measure the surface-water flow out of TNC property, with as little contribution from adjacent lands as possible. Gages were within 2 ditch mi of the edge of TNC property, except SW5, which was about 4.8 ditch mi from the edge of the property, and SW2, which was not designed to measure flow out of TNC property.

Ditch gages were instrumented with pressure transducers connected to a nitrogen-gas bubbling manifold, which senses the pressure required to force a bubble from a fixed orifice at the bed of the ditch. The pressure is recorded as feet of water representing the stage of the ditch. Stage, battery voltage, and equipment-shelter internal temperature were recorded every 15 minutes at the gage and transmitted to the USGS data base once per day through radio and telephone telemetry. In the process of data storage, stage is converted to discharge through a rating equation (Rantz and others, 1982) developed for each gage during the study. All data are available online (*http://waterdata.usgs.gov/mn/nwis/current/?type=flow*) by use of the site numbers in appendix 2. The pressure transducers were calibrated at least every 6 weeks. Ditch flow was measured with a current meter whenever flow conditions would improve the rating equations.

Ground-Water Data Analysis

A map of the areal extent and thickness of the surficial sand and gravel was produced to define the surficial aquifers

Table 2. Characteristics of ditch basins, Glacial Ridge study area, northwestern Minnesota.

[MC, main channel; mi², square miles; mi, miles; ft/mi, feet per mile; %, percent of total area; MN, Minnesota]

Short name	Gage number	Gage name	Drainage area (mi^2)	Length of ditches (mi)	MC length (mi)	MC slope (ft/mi)	Wetland and lake area (%)	Surficial aquifer area (%)
SW1	05078730	County Ditch 140 near Benoit, MN (SW1)	11.8	14.4	10.1	14.0	35	47
SW2	05079250	County Ditch 65 near Maple Bay, MN (SW2)	10.4	8.7	8.4	6.7	27	57
SW3	05079200	County Ditch 72 (Burnham Creek) near Maple Bay, MN (SW3)	10.7	11.7	11.2	5.3	38	52
SW4	05078470	Judicial Ditch 64 near Mentor, MN (SW4)	9.6	12.3	4.1	6.5	41	64
SW5	05078520	Cyr Creek near Marcoux Corners, MN (SW5)	11.4	10.7	7.1	10.0	14	38
SW6	05078770	Judicial Ditch 66 near Marcoux Corners, MN (SW6)	14.2	13.0	9.9	7.7	30	56
SW8	05078720	County Ditch 140 above BR-6 impoundment near Tilden Junction, MN (SW8)	9.0	12.6	8.3	15.3	36	46

for this study in greater detail than previously published. The extent of the aquifers was interpreted primarily from the areas of coarse-grained soils as described in NRCS digital soil surveys. Thickness was estimated primarily from well boring logs. The techniques and data sources used to produce the extent and thickness map are detailed in appendix 3.

Water-table and potentiometric-surface maps are based on water levels measured synoptically in June 2004 at 72 wells and 11 wetland, 1 lake, and 6 ditch gages. Details of data and assumptions used to construct these maps are presented in appendix 3. These ground-water surface maps were used to infer the areal distribution of ground-water flow direction and relative gradients (ground-water driving force) in surficial and buried aquifers. June 2004 represented average water-level conditions as indicated by continuous water-level data collected for this study. In areas where synoptic water levels were sparse, water levels collected from 47 boreholes constructed by the USGS in the early 1990s (Lindgren, 1996) were used as a rough guide in contouring. Land surface, wetland, and ditch topography were frequently considered during contouring. Wetlands adjacent to beach ridges and ditch reaches where ground-water discharge occurred were assumed to be the surficial expression of the water table. In June 2004, ditches in the study area contained only a few inches of water. Twenty-five wells in the study area were screened in buried artesian aquifers of about the same elevation. It is possible that water from these aquifers can discharge, in fractures or diffusely through the intervening tills, into the surficial aquifers from below. To evaluate the interconnectedness of the buried aquifers and to estimate the gradient driving water from buried to surficial aquifers on a regional scale, a potentiometric-surface map of the confined aquifers was produced from the water levels synoptically measured during June 2004.

Aquifer hydraulic conductivity was estimated from gas-displacement slug tests at 10 observation wells throughout the study area. Water was evacuated from the casing by forcing it out through the screen and into the aquifer with compressed nitrogen. The gas pressure was released instantly from the well, and the recovery of water in the casing was measured with a pressure transducer and recorded every second or whenever the water level rose by at least 0.02 ft. Recovery data were analyzed by means of the empirical Bower-Rice solution (Bouwer and Rice, 1976; Bouwer, 1989) as modified by Zlotnik (1994), using either the unconfined or confined solution as appropriate.

Recharge was estimated from hydrographs of the 10 wells completed in surficial aquifers and 1 well completed in a buried aquifer, at which continuous data were collected using the water-table-fluctuation (WTF) method (Rutledge, 1998; Healy and Cook, 2002). This method assumes that recharge can be estimated as the product of ground-water level rise and specific yield. Ground-water level rise was calculated by summing the rising portions of the 3-day running minima of the daily mean water levels. Three-day running minima were used to remove low-amplitude oscillations (generally less than 0.1 ft) of 1–2 days in length contained in many hydrographs. These oscillations are not diurnal, do not correspond to precipitation, and are synchronous in all hydrographs in surficial aquifers, although the amplitude varies among wells. The oscillations are not present throughout the year, may appear any time during the year, and are most prominent in the winter during times of low water levels. The cause of the oscillations is unknown, but they are not believed to be related to recharge. The monthly rise sums were multiplied by the specific yield of the aquifer material (assumed to be 0.25, average value for gravelly sand, Fetter, 1988) to produce monthly recharge at each well. Recharge rates (recharge per unit time) were computed by dividing summed recharge by summed precipitation during a period of time. Spring and winter recharge rates were not computed because recharge (recorded as water-level rises) during those seasons did not necessarily result from the precipitation recorded during the same period and, therefore, could be inaccurate.

The leakage estimate to a buried aquifer from the hydrograph at well E01D was calculated as above except that the rise sums were multiplied by the average storage coefficient (0.017), measured at seven wells during three aquifer tests, instead of the specific yield. This leakage estimate for a buried aquifer is the equivalent of a recharge estimate for surficial aquifer. The leakage could not be estimated at well E04D because pumping from a nearby production well severely influences its water levels. Technically, using the WTF method to estimate leakage to a buried aquifer is a misapplication, as it is only strictly defined for water-table aquifers. Other investigators (for example, Ruhl and others, 2002) have adapted the WTF method to estimate leakage in this way.

Mass-Balance and Ditch-Data Analysis

Annual mass balances were calculated for the surficial aquifers and ditch basins. Mass balances were calculated only for those parts of the study area that drain to gaged ditches because these are the only areas that have measurements of surface-water flow and estimates of ground-water discharge to ditches. Details of the balance equations and the data used to calculate the mass balances are in appendix 3.

Ditch hydrographs were analyzed to aid in understanding the characteristics of direct runoff and ground-water discharge to ditch flow with four different methods. Method one, a statistical description of daily flows, provided an overall view of the flow characteristics of each basin. Method two, a base-flow analysis determined from streamflow partitioning, quantified the contribution of ground-water discharge to each ditch. Streamflow was partitioned by means of the computer program PART (Rutledge, 1998). In partitioning, one assumes that all flow in a ditch is ground-water discharge at some fixed time after a hydrograph peak. This time is determined by an empirical formula based on the ditch-basin area. Method three, an analysis of the hydrograph recessions and slopes, described the change in ground-water discharge to each ditch over time. This analysis produces a recession index, which describes the ground-water discharge recession rate. The recession rate was calculated by means of the computer program RECESS (Rutledge, 1998). Method four, storm-runoff hydrograph modeling, quantified how quickly a stream reacts to rainfall and how quickly direct runoff flows out of the basin. Hydrograph models were produced with the HEC-HMS modeling system (U.S. Army Corps of Engineers, 2001) using the Clark unit-hydrograph method (Clark, 1945). Changes in land use in a basin result in changes in the variable values used in the Clark unit-hydrograph method. Details of these four hydrographic analysis methods are included in appendix 3.

Sample Collection, Analysis, and Quality Control

One water sample was collected from each of the 48 wells, 6 ditch gages and 7 wetland gages in the synoptic network during May–July 2004 (table 1, fig. 3). At that time, restoration activities were well under way in parts of the study area. Ideally, this initial characterization would have been made before any restoration had taken place. The lag time between land-use change and its effect on water quality, especially for ground water, may help ameliorate this problem. Sampling of the variability water-quality network began in October 2002 at 6 ditches and in May 2003 at 12 wells and continued through summer 2007. Some sampling by the USGS continues as of fall 2007. Ditches were sampled as many as 27 times (about monthly) through October 2005. Wells were sampled monthly from May to October 2003, irregularly three times in 2004, and again about monthly from April to September 2005. Variability water-quality data collected after WY 2005 will be used to help decide when another synoptic data set could be collected in the future in order to statistically attribute water-quality change to land-use change in the study area. Variability data collected during 2002–5 can be compared to future variability in the study area to help define a long-term water-quality change. However, the short time period (3 years) of these variability samples makes time-trend analysis impractical.

Synoptic samples were analyzed for physical properties and chemical constituents that characterize natural water quality and show agricultural land-use effects. Physical properties included temperature, pH, specific conductance, and dissolved oxygen (field measurements). Chemical constituents analyzed for include major ions, nutrients, corn and soybean herbicides and their degradates (hereinafter, herbicides), and water isotopes. Suspended sediment samples were collected at ditch gages. Dissolved-gas samples were collected at selected wells. One dissolved gas analysis, sulfur hexafluoride (SF_6), was used to estimate the ground-water recharge date, or age of ground water. Details of sample collection, equipment decontamination, and analytical methods are in the "Synoptic Sampling" section of appendix 3.

Ground-water and ditch variability networks were sampled to assess temporal variation in water quality. The ground-water and ditch networks were sampled on different time schedules. Ground-water network samples were analyzed for field measurements and nutrients. Ditch-network samples were analyzed for field measurements, nutrients, major ions, and suspended sediment. Separate whole-water and filtered samples were collected in ditches for some constituents. Particulate concentrations were calculated by subtracting the concentration of the filtered sample from the concentration of the whole-water sample. Major-ion analyses were discontinued from ditch-water samples in November 2004 because the variability of these constituents was adequately assessed. During a variability-sampling event, one sample was collected from each site in the network, if water could be sampled. Details of deviations from synoptic sample collection, sampling equipment decontamination, and analytical methods are in the "Variability Sampling" section of appendix 3.

The project water-quality-control program consisted of the comparison of field and laboratory measurements and the assessments of constituent totals, field blanks, and duplicate samples. The overall purpose of the program was to assess the accuracy and precision of project water-quality samples. The purpose of field and laboratory measurement comparison was to assess errors in sample labeling and to check field measurements. Assessments of constituent totals were made to determine whether any major ion was not analyzed for. Field blank results indicate whether decontamination procedures were successful. Finally, duplicate sample results gage the variability introduced by the sampling process. A complete description of the water-quality-control results is presented in appendix 4.

Hydrology

Water occurs within the study area on the land surface, in surficial aquifers that extend downward from the land surface, and in buried aquifers of various depths, the tops of which are 50 to several hundred feet below the land surface. Most water flow within the study area is shallow, occurring as surface water and as ground water within thin surficial aquifers. Surficial ground-water movement may extend a few feet into confining units (coarser-grained wave-modified till, fig. 5) but is slower than in aquifers. The uppermost buried aquifers (50–100 ft below land surface, hereinafter, the term "buried aquifers" only refers to these uppermost buried aquifers), underlie much of the study area, but ground-water flow in them does not substantially interact with the shallow-water system in most areas. In isolated areas however, buried aquifers directly underlie thin surficial aquifers where a confining unit is absent. In these areas, water flows directly from buried to surficial aquifers. The structure and characteristics of the aquifers and confining units within the study area are a product of the history of glacial advance, retreat, and lake formation that created them. The details of this history provide a geologic context for understanding the aquifer structure and composition and are presented in appendix 5.

The postglacial formation of the surface-water system and its subsequent substantial human modification (appendix 1) control the flow of nearly all water within and out of the study area. The flow in surficial aquifers and in ditch basins is a single hydrologic system. Within the study area, precipitation drives this hydrologic flow, moving through the ditch basins and surficial aquifers, and leaving the study area primarily as evapotranspiration and ditch outflow.

Hydrogeology

Ground water in the study area flows in surficial and buried aquifers. The surficial aquifers were formed from the former beaches of glacial Lake Agassiz, an enormous glacial meltwater lake that occupied the central part of the Red River of the North Basin during 13,800–8,440 years ago (Fenton and others, 1983, p. 69–70). (All dates in this report are given in calendar years before present (1950) and were calculated from radiocarbon years by use of a table in Teller and Leverington, 2004, p. 732 and a figure in Fisher, 2005, p. 1482.). The beach-ridge aquifers are thin, narrow, and long sand and gravel deposits. The aquifers are generally distinct but merge with each other in places, particularly in the southern part of the study area. The buried aquifers are generally separated from the surficial aquifers by clay-rich till that acts as a confining unit. The areal extent, thickness, and interconnectedness of the buried aquifers are unknown, but the uppermost of them are at a depth of 50 to 100 ft in much of the study area. They are composed of sand and gravel of glacial origin.

Horizontal flow in all aquifers is radial from the Maple Lake area in the southeastern part of the study area, following the downhill direction of the topography toward the Red Lake River to the north and northwest, and toward the Red River of the North in the center of the Lake Agassiz Basin to the west. These directions are generally perpendicular to the trend of the beach-ridge aquifers. Flow in the buried aquifers is primarily horizontal, but with a vertically upward component of flow resulting in leakage to the surficial aquifers and land surface. Flow in the surficial aquifers is closely connected to the adjoining wetlands upgradient and downgradient, to the beach ridges, and to ditches where they cut through the surficial aquifers.

Aquifer Descriptions

Surficial and buried aquifers in the study area are composed of relatively well-sorted, coarse-grained sediments deposited by many glacial and glaciolacustrine processes and events. These aquifers are separated from each other by fine-grained till, lake clay, and (or) organic-rich wetland deposits that have hydraulic conductivities several orders of magnitude lower than the aquifers. Some tills are somewhat coarser grained with higher hydraulic conductivities than other tills; but hydrologically, the tills, lake clays, and wetland deposits form nearly equally effective confining units. The hydraulic conductivity of the confining units has not been measured. The range of calibrated hydraulic conductivities of confining units in Lindgren's (1996; see the preceding "Introduction" section) ground-water flow model in the study area was 10–50 ft/d, which is about 10–20 times lower than the hydraulic conductivities of aquifers. The volume of confining-unit deposits is much greater than the volume of aquifer deposits.

Surficial Aquifers

Most surficial aquifers in the study area are beach-ridge sediments (figs. 1 and 5) winnowed from and deposited on till. Individual beach ridges form very small and variable aquifers. Each ridge is usually less than 20 ft thick but may be as much as 35 ft thick locally. Depth to water varies from zero to 20 ft. Ridges are narrow (250–1,000 ft) but are tens to hundreds of miles long, though usually hydraulically continuous for less than several miles. In many places, beach ridges coalesce into areas of wider surficial sands, particularly in the southern part of the study area. The extent of these wider surficial-sand areas can indicate a more substantial aquifer than really exists. In most cases, only a veneer of sand less than 10 ft thick lies between each beach ridge and does not form an areally extensive aquifer. This veneer is probably sand redistributed by storm waves behind the active beach ridge of Lake Agassiz.

Beach-ridge sediments range from fine sand to gravel but are generally well sorted and sandy. The base of a beach ridge is usually composed of gravels lying directly on wave-modified till. Beach sands rarely contain beds of well-sorted silts and lake clays. Based on gas-displacement slug tests of nine wells, the range of hydraulic conductivity of the surficial aquifers is 2.7–43.4 ft/d (transmissivity is 30–2,170 ft^2/d). These values agree well with hydraulic conductivities of 3–38 ft/d (transmissivity of 35–700 ft^2/d) measured by Lindgren (1996) with slug tests on seven wells in the study area. The median hydraulic conductivity of all 16 slug tests is 8.1 ft/d (transmissivity is 124 ft^2/d). Such slug tests only measure the hydraulic conductivity of a small part of the aquifer near the well screen. Lindgren also did single-well pumping tests, which measure hydraulic conductivity in a slightly larger part of the aquifer than do slug tests, at two wells. These pumping tests produced hydraulic conductivities that were 3.4 and 4.1 times the values produced from the slug tests at the same wells (23.4 and 63.5 ft/d for the pumping tests compared to 6.9 and 15.5 ft/d, respectively, for the slug tests). The horizontal hydraulic conductivity of surficial aquifers from Lindgren's calibrated ground-water model (1996) was 200–300 ft/d with a saturated thickness of 0–30 ft. Hydraulic-conductivity measurements in the study area follow the general pattern that the magnitude of hydraulic-conductivity estimates tend to increase with techniques that integrate larger parts of an aquifer.

In four parts of the study area, surficial sand and gravel thickness is much greater than that of the beach-ridge aquifers just described (figs. 6 and 7). Wells drilled in these areas have

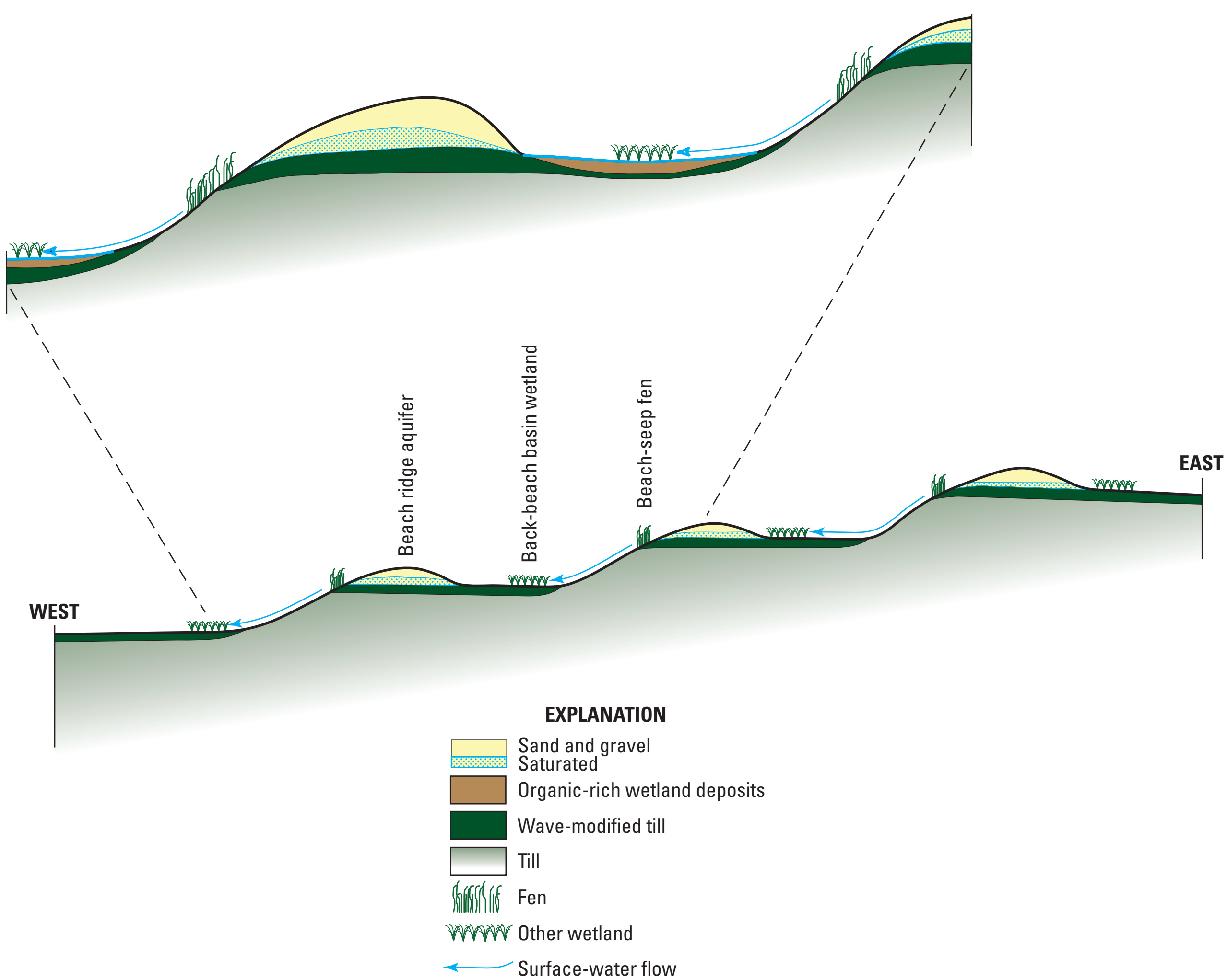


Figure 5. Conceptual hydrogeologic section through the Glacial Ridge study area, northwestern Minnesota.

penetrated more than 74 ft of sand and gravel. These areas form the only substantial surficial aquifers in the study area. The deposition of these thick aquifers appears to be unrelated to Lake Agassiz because, although the lake produced a land surface that is quite planar, these aquifers are incised into the land surface; through processes that the lake was incapable of. The thick aquifers are of limited extent, have steep lateral boundaries, and are adjacent to relatively large areas of thin sands. These thick aquifers trend along an east-southeast–west-northwest line from the ice-stagnation topography of the Itasca Moraine near Maple Lake to the northwest corner to the study area. This direction is along the trend of the ice margin that deposited the Itasca Moraine. These thick sand and gravel deposits are interpreted to be remnants of ice-contact stratified materials deposited within till during the wasting of one or more ice lobes. The last lobe to cover the study area exposed, or did not bury, these ice-contact materials. When Lake Agassiz formed, these exposed sands and gravels were locally distributed into adjoining thin sand plains, upon which beach ridges subsequently developed.

The thick surficial aquifers are stratigraphically more complex than the overlying beach-ridge deposits. The conductive parts of the aquifer are mainly well-sorted, medium-to-coarse grained sands with some gravel beds. The sediments composing these aquifers are generally better sorted and thicker than beach-ridge sediments. The thick aquifers can contain lenses of till, silt, and lake clay, as much as 20 ft thick, which are generally of small areal extent. Beach ridges generally overlie the thick aquifer sediments, but the contact between them cannot be distinguished geologically or hydrologically. In the north-central part of the study area ("Area A", fig. 6), a relatively continuous till and (or) clay layer is within the thick surficial aquifer just below the surface. Here, a 10 to 20-ft-thick fine-grained layer usually underlies 4 to 10 ft of surficial sand. This layer is underlain by sand and gravel that ends at till 47 to 70 ft below land surface. In area A, the fine-grained layer locally may be absent or may be at the surface. Although stratigraphy in area A was interpolated from 54 well and borehole logs, other areas of thick surficial aquifers were interpolated from far fewer logs. It is possible that these other areas are as stratigraphically complex as area A but that the data are insufficient to show such complexity.

In his 1996 hydrologic study, Lindgren described the thick surficial aquifer in area A as "partially confined," meaning that it "is predominantly under confined conditions but is under unconfined conditions in small isolated areas where sand and gravel are present at the land surface." However, because the thickness, continuity, and grain-size of the fine-grained layer are highly variable, the hydraulic influence of this layer on the surficial aquifer in area A is not well known. It is likely that this layer was deposited before Lake Agassiz beach formation as a till slump from a melting ice block or as shallow lake deposits. Subsequent winnowing and erosion of these deposits by lake waves likely would not produce an extensive confining unit, even on the scale of area A. Therefore, for the purposes of this report, the surficial aquifer in area A is considered to include all material from land surface to the bottom of the sand and gravel below the fine-grained layer. For this reason, surficial aquifer thickness contours may extend beyond the area delineated as surficial sand in figure 6, where the upper thin sand is absent and the fine-grained layer is at land surface.

Buried Aquifers

The buried aquifers are composed primarily of well-sorted sand and gravel separated from the surface or surficial aquifers by about 50–100 ft of clay-rich till and (or) lake clay. These aquifers are equivalent to the shallow confined aquifers of Lindgren (1996). The extent, thickness, and interconnectedness of these aquifers are unknown. Within the study area, the buried aquifers are ice-contact deposits and possibly remnants of outwash that are entirely surrounded by till or adjacent buried aquifers, with one exception. In the area of the thick surficial aquifers that is near several small lakes 2 mi southeast of area A ("Area B", fig. 6), the buried aquifers are hydraulically and probably physically connected to the surficial aquifers, as indicated by ground-water discharge, discussed in the "Ground-Water Flow" section that follows.

Individual buried aquifers can be traced horizontally at most several thousand feet and are 12 to more than 81 ft thick (Lindgren, 1996). Many buried aquifers are in physical contact with each other and most are in hydraulic contact, and they cover nearly all of the study area as indicated by the continuity of the potentiometric surface. From a hydraulic-head perspective, the buried aquifers in the study area function as a single, relatively continuous aquifer. Whether this is true from a ground-water flow perspective is unknown.

Lindgren (1996) did four slug tests in buried aquifers in the study area, of which one was repeated for this study (well L103; hydraulic conductivity from the new test was 69 percent greater than that from the original test). Based on these five tests, the range of hydraulic conductivity of buried aquifers is 10–20 ft/d (transmissivity range, 155–917 ft^2/d). Between 1991 and 2001, consulting firms did three multi-well aquifer tests in buried aquifers near U.S. Highway 2 in the northern and eastern parts of the study area. The hydraulic conductivities and storativities of the aquifers tested had ranges of 28–155 ft/d and 0.004–0.05, respectively (Lindgren, 1996). The horizontal hydraulic conductivity of buried aquifers from Lindgren's calibrated ground-water model (1996; see the preceding "Previous Studies" section) was 50–300 ft/d.

Recharge and Discharge

Recharge to surficial aquifers is from vertical infiltration of rainfall and snowmelt (areal recharge), from surface waters, (particularly ephemeral wetlands), and from upward leakage of ground water from buried aquifers through till confining units. Areal recharge is highly variable in space and time, depending on the amount and intensity of rainfall, amount of storage potential remaining in wetland basins, amount of

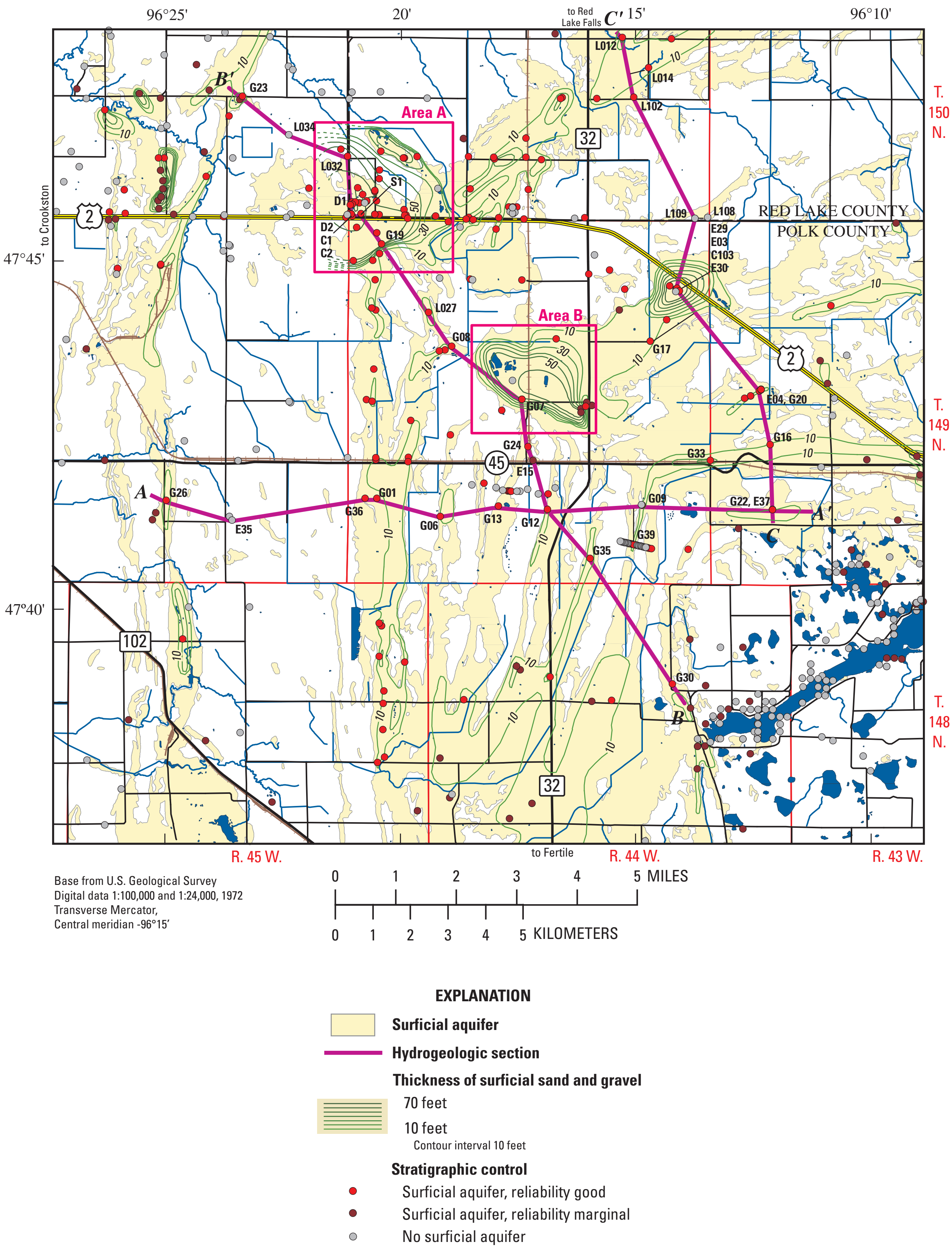


Figure 6. Surficial sand and gravel thickness, Glacial Ridge study area, northwestern Minnesota.

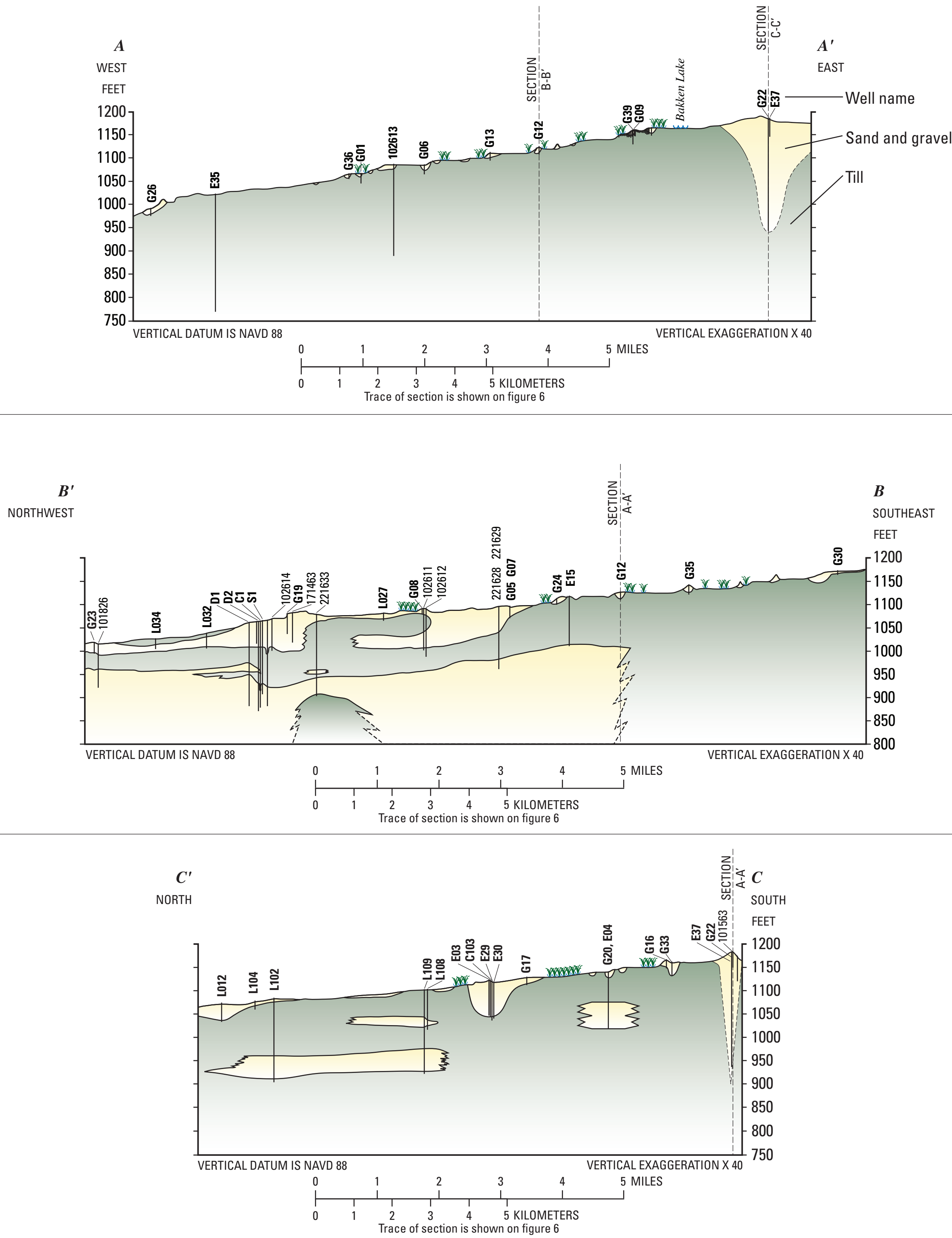


Figure 7. Hydrogeologic sections, Glacial Ridge study area, northwestern Minnesota.

Table 3. Precipitation and ground-water recharge, Glacial Ridge study area, northwestern Minnesota, 2003–5.

[high wells, wells with relatively high recharge rates (G01, G15), low wells, wells with relatively low recharge rates; WY, water year, S + F, summer and fall; grey shading, high-recharge wells; black shading, buried-aquifer well]

													Surficial aquifer average		
	SCAN	E01S	E03	E05	G01	G08	G12	G15	G20	G22	G25	E01D[a]	All wells	Low wells	High wells
							Precipitation, in inches								
Spring	4.58	5.96	5.71		3.77	4.85	4.94	6.39	5.10		4.61		5.10	5.11	5.08
Summer	9.40	11.34	11.31		10.19	9.98	9.13	12.02	6.95		6.15		9.61	9.18	11.10
Fall	6.32	7.58	7.06		7.23	6.81	6.93	7.61	4.91		9.01		7.05	6.95	7.42
Winter	.25	.42	.12		.13	.31	.47	.62	.21		.21		.30	.28	.38
WY2003	14.87														
WY2004	20.36	25.63	25.09		18.52	21.97	22.15	24.34	21.97		13.66		21.52	21.55	21.43
WY2005	21.96	27.63	25.90		25.34	22.61	22.72	29.33	20.84		19.24		23.95	22.99	27.34
Average	19.06	26.63	25.50		21.93	22.29	22.44	26.84	21.41		16.45		22.74	22.27	24.38
							Recharge, in inches								
Spring		5.84	6.77	10.33	23.23	4.96	6.80	13.93	4.36	5.28	7.63	0.30	8.91	6.50	18.58
Summer		2.96	3.73	4.77	18.79	3.32	3.99	10.23	1.74	3.62	4.58	.22	5.77	3.59	14.51
Fall		3.39	4.78	5.16	17.65	4.31	5.24	7.92	2.65	4.77	4.35	.34	6.02	4.33	12.78
Winter		.35	2.22	1.42	3.64	.42	1.34	3.43	.51	4.23	.42	.16	1.80	1.36	3.53
WY2003	7th[b]	9.41	12.60		62.10	6.61	12.96	33.98					22.94	10.39	48.04
WY2004	47th[b]	12.99	18.99	27.43	44.14	11.44	18.03	34.09	7.80	15.67	14.77	1.21	20.53	15.89	39.11
WY2005	65th[b]	14.70	19.11	22.73	68.97	17.31	18.99	35.70	14.14	25.20	18.70	.81	25.56	18.86	52.34
Average		12.37	16.90	25.08	58.40	11.79	16.66	34.59	10.97	20.44	16.73	1.01	22.39	16.37	46.50
							Recharge rate[c]								
Summer		.26	.33		1.84	.33	.44	.85	.25		.74		.63	.39	1.35
Fall		.45	.68		2.44	.63	.76	1.04	.54		.47		.87	.59	1.74
S + F		.34	.49		2.09	.57	.61	.93	.38		.59		.75	.50	1.51
WY2004		.51	.76		2.38	.52	.81	1.40	.36		1.08		.98	.67	1.89
WY2005		.53	.74		2.72	.77	.84	1.22	.68		.97		1.06	.75	1.97

[a]Well completed in a buried aquifer. Values in the recharge section are actually leakage estimates to the buried aquifer.
[b]Percentile of 1890–2005 precipitation at Crookston, Minnesota (High Plains Regional Climate Center, 2006).
[c]Recharge divided by rainfall during a water year or season.

snowpack, antecedent soil moisture, depth of frost, and the particular history of each spring thaw. Two wells (G01 and G15) were located adjacent to wet meadows and had water levels less than 3 ft below land surface. Based on ground-water hydrograph/precipitation relations, the aquifers near these wells were depleted locally by evapotranspiration from adjacent wet meadow plants, often on a daily cycle. At night, when evapotranspiration ceased, local water levels in the aquifers recovered by recharge from adjacent wetlands or ground-water flow from other parts of the aquifer. Such water-level rises do not represent areal recharge to the aquifer as a whole and areal recharge estimates based on hydrographs at these wells are too high. These wells will be termed “high-recharge wells” and will not be included in the areal recharge discussion that follows.

Average annual areal recharge (all annual sums in the following discussion are for water years (October 1st-September 30th)), estimated from the rise of water levels in hydrographs and assuming a specific yield of 0.25, ranged from 10.97 to 25.08 in/yr during 2003–5 (table 3). Recharge increased with increasing precipitation; it was lowest in 2003 (average, 10.39 in. when total precipitation was 14.87 in, at the SCAN station) and highest in 2005 (average, 18.86 in.). The aquifer near well E05 had consistently high recharge (2004–5 average, 25.08 in. when total precipitation averaged 21.16 in. at the SCAN station, located about 5 mi away), which is higher than the average precipitation for the period at all rain gages. This may indicate some recharge from surface water in the area, although no obvious source exists. No local precipitation record is available at wells E05 and G22. Recharge estimates at well G22 may be overestimated because the area around the well is irrigated and the recharge estimates contain rises from infiltration of irrigation water.

During this study, the most recharge occurred during the summer season, followed by the fall, and then the spring. Recharge was least in the winter. Precipitation amounts had a similar pattern during the study. Recharge during the spring is generally the highest of the year. During this study, however, spring recharge was no higher than recharge during summer storms or the fall because winter precipitation and snowpack were low. Recharge rates could be estimated at eight wells during 2004–5 using the WTF method. These rates ranged from 51 to 108 percent of precipitation and averaged 67 percent in 2004 and 75 percent in 2005.

Water enters the buried aquifers by leakage through confining units or by flow from adjacent aquifers. The hydrograph at well E01D provided the only estimate of the amount of water entering the buried aquifers. Annual leakage into the aquifer near E01D was 1.21 in. in 2004 and 0.81 in. in 2005, which is opposite to the relative precipitation amounts for those years. Most of this leakage entered the buried aquifers in the study area horizontally because they are overlain by thick clay-rich till, and heads in the buried aquifers are higher than those in the surficial aquifers. Water in the buried aquifers probably infiltrated from the surface somewhere to the east and southeast, where elevations are higher and adjacent aquifers may have surficial areas where recharge was greater.

Details of areal recharge can be seen by comparing hydrographs during a recharge event. In late October 2004, an average of 4.53 in. of rain fell on the study area, producing the hydrograph rises shown in figure 8. Assuming a specific yield of 0.25, if all the rain infiltrated, the water table would rise 1.51 ft, which is about the observed rise at all wells except G01. Assuming that the aquifer near G01 receives about as much areal recharge as other aquifers in the study area, only about one-half of the 3 ft of water-level rise in this well resulted from areal infiltration. The other one-half of the rise probably resulted from infiltration of water from adjacent wetlands as their water levels rose in response to the same rainfall. Notable in these rises is the decrease in the sharpness of the recharge peak with depth, which is an expression of the greater time needed for infiltrating water to reach the water table with greater depth. Well G12 is an exception to this trend. Here, the recharge peak occurred more quickly than at other aquifers, suggesting that the aquifer material there is hydraulically more conductive.

Recharge (leakage) from surface waters was not measured directly but can be inferred from some hydrographs, as in the case of well G01 above. Hydrographs from wells G15 and possibly G25 show similar recharge from surface water. Wetlands cover a large part of the study area, and many of them are ephemeral, having water at the surface for only a few weeks during the year. All ditches, except for the one measured at gage SW6, are seasonal, drying up completely most years in the summer and winter. Therefore, recharge from surface water is dynamic, both spatially and temporally. Generally, though, wetlands are a much more important source of recharge than are ditches because their surface area in contact with aquifers is much greater. In most cases, wetland basins are formed on till, adjacent to beach ridges. These wetlands only recharge aquifers when runoff raises their stage to the point where they expand to cover parts of an aquifer and wetland stage is higher than the water table. No quantitative estimate of annual recharge from surface water was possible with the data collected for this project.

Leakage from buried aquifers also was not directly measured, but likely occurs. The water levels in all wells open to buried aquifers were either above land surface or higher than the water level in surficial aquifers. About 1 in/yr of recharge to buried aquifers was estimated from the rise in the hydrograph from well E01D, which is assumed to be the case throughout the study area. Because very little change in storage can occur in saturated buried aquifers, discharge from buried aquifers by diffuse leakage upward through the till confining unit is assumed to be at a rate of 1 in/yr to the base of all surficial aquifers.

Isotope Evidence for Ground-Water Recharge and Discharge

The isotopic composition of water can yield information about sources, destinations, and seasonal timing of ground-water recharge and discharge (Gat, 1981). In settings such as wetlands, where water can evaporate from an open-water

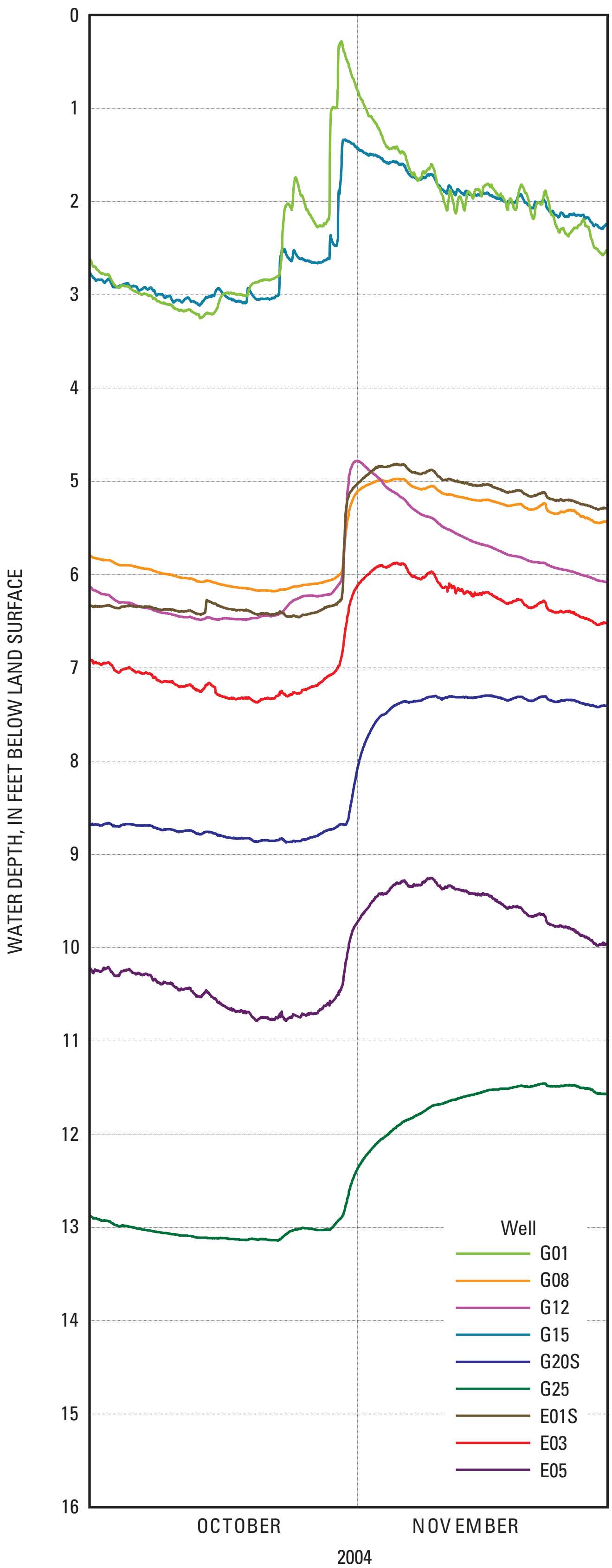


Figure 8. Ground-water levels, Glacial Ridge study area, northwestern Minnesota, October–November, 2004.

surface, the wetland water can acquire a heavier hydrogen and oxygen isotopic composition that identifies it as evaporated. If this water infiltrates and recharges ground water, the ground water will retain the heavier isotopic signature. Conversely, if isotopically light ground water discharges to heavier surface waters, those waters will become relatively lighter. The use of isotopes to identify a water source relies on the determination of a local relation between the isotopes of hydrogen (δ^2H) and oxygen (δ^{18}O) in precipitation. The relation is called the local meteoric water line (MWL). Surface waters generally deviate from the local MWL because water molecules that are composed of relatively light isotopes preferentially evaporate and leave behind water molecules that are composed of relatively heavy isotopes. The isotopic composition of ground water whose source is a combination of precipitation and surface water will fall on a line between the local MWL and the isotopic composition of the surface waters. The position on this line represents the proportion of ground water recharged from each source by simple mass balance.

The isotopic composition of samples collected for this study is shown in figure 9. Rainfall isotope samples were not collected for this study so no local MWL is available; however, a MWL measured at a hydrologic research site near Princeton, Minn. (250 mi southeast (fig.1), fig. 9; Landon and others, 2000) can be used as an approximation. The weighted mean precipitation (WMP) isotopic composition for a location is the isotopic composition of all the precipitation that falls at a site if it were collected for an average year and homogenized. This composition can be calculated from empirical equations developed with data from locations around Earth that have long precipitation-isotope records. The two large circles on figure 9 show the calculated WMP isotope composition for this study area and the Princeton, Minn., research site using an equation by Yurtsever and Gat (1981). These values are very close to each other and to the measured WMP at Princeton, Minn. indicating that the difference between the MWLs for this study area and Princeton is very small.

The isotopic composition of ground-water samples scatter along the Princeton MWL, showing that most ground water in the study area comes from precipitation that has not undergone evaporation. Isotopically very light samples are ground water recharged from cold winter precipitation (probably snowmelt) and isotopically heavy samples are ground water recharged from precipitation during warm summer storms. Many wetland samples and some ditch samples have isotopic compositions that are heavy, below and to the right of (hereinafter, below) the MWL. One ground-water sample (G13, near a wetland) also appears below the scatter of the MWL. These samples have undergone evaporation. The further a sample is below the MWL, the more evaporation it has undergone.

All but one surface-water sample (WL07, fig. 3) were well below the Princeton MWL, and as many as half were clearly beyond the scatter of ground-water samples about the MWL. These data show that substantial evaporation occurred in many surface waters by the time the samples were collected (mid-July 2004) but also that many surface waters had not undergone substantial evaporation. Because of this, the fact

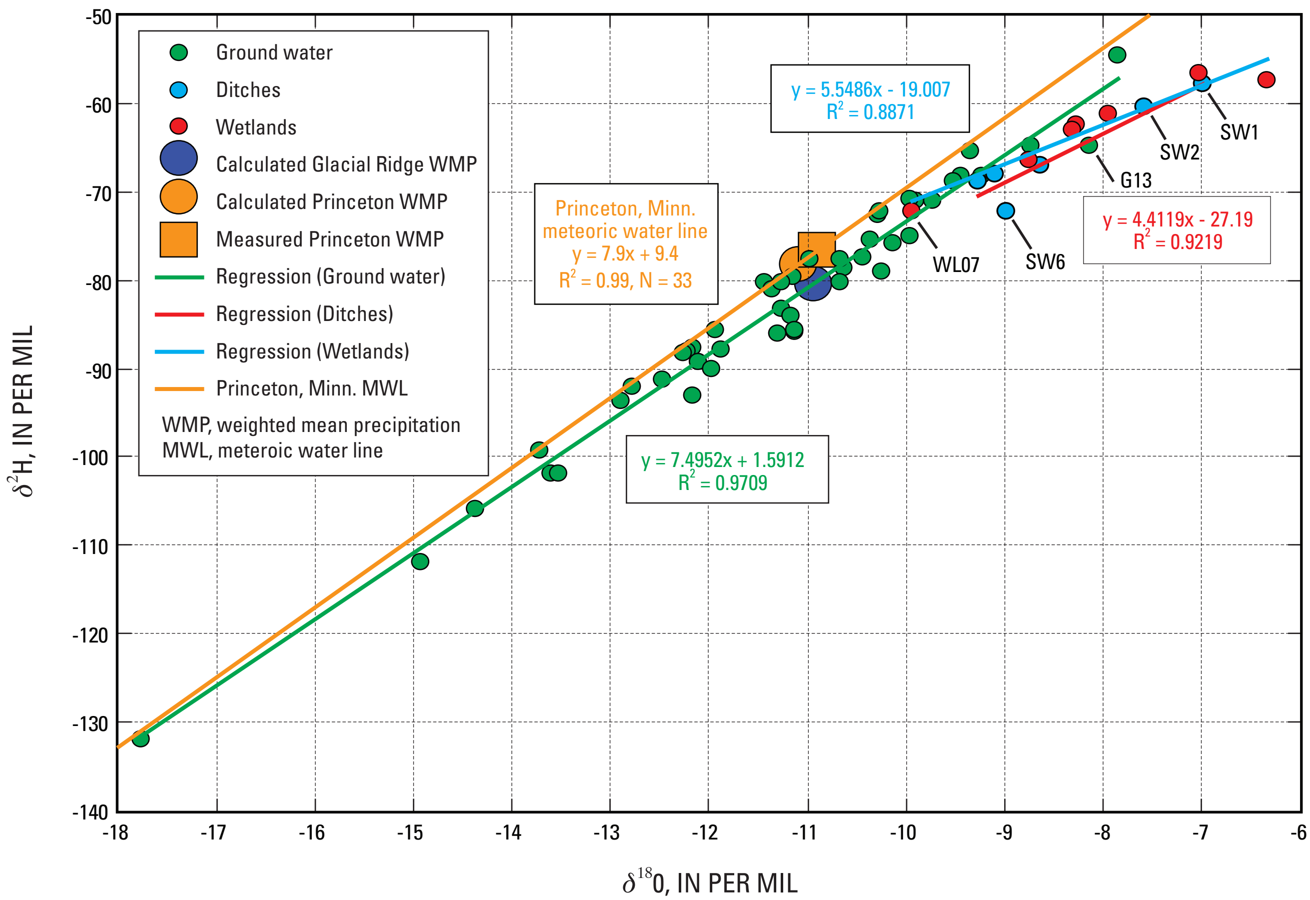


Figure 9. Water-isotope composition, Glacial Ridge study area, northwestern Minnesota, May–July 2004.

that almost no ground-water samples show evidence of evaporation does not mean that there is not substantial leakage from surface waters. On the contrary, unevaporated surface waters could recharge the aquifers and produce the isotopic composition found in the ground-water samples.

Flow in ditches with isotopically evaporated waters (for example, SW1, 2, and 6) is dominated by discharge from evaporated wetlands or evaporated ground water like that at well G13, or is evaporating as it flows. Because evaporated ground water was rarely sampled, the first and third possibilities are most likely. A likely explanation for unevaporated wetland waters (for example, WL07) is not that they evaporate less than other wetlands but rather that they receive substantial discharge from unevaporated ground waters. The most extreme example from the study area contradicts this hypothesis, however. Well G13 is immediately upgradient from wetland gage WL07. Although, water from well G13 is isotopically the most evaporated, the wetland that should receive its discharge is isotopically the most unevaporated. This example shows how dynamic and complicated are the recharge and discharge relations between ground water and surface water in the study area.

Discharge and Ground-Water Mass Balance

Water discharges from a surficial aquifer to surface waters that are closed basins, to ditches and surface waters that drain to them, to the atmosphere through evapotranspiration, and to pumped wells. Estimates of net discharge from surficial aquifers to closed surface-water basins were made by computing a net mass balance for water moving through the surficial aquifers (appendix 3). The net mass balance was computed as follows:

Net areal recharge to surficial aquifers =

Net ground-water discharge to ditches +

Net changes in ground-water storage +

Net unmeasured ground-water losses

Hereinafter, the net nature of these terms is implied. To simplify mass-balance-term estimates and to show spatial variability, balances were computed on ditch basin areas (referred to by the gage name, fig. 3), although the true ground-water basins may vary slightly from the ditch basins. The first three terms of the mass balance were estimated from measurements, and the equation was solved for the unmeasured-losses term.

This term includes at least the following losses from the mass balance:

- Surficial aquifer discharge to closed basins
- Evapotranspiration
- Well withdrawals
- Discharge to ungaged ditches and surface waters that drain to them
- Measurement errors

Estimates of these losses are difficult, and good estimates cannot be made with the data collected for this study. Relative to the amount of ground water that discharges to closed basins, discharge to the other components of unmeasured losses is small, with the exception of the measurement-errors component. The size of this error component is unknown.

An estimate of ground-water evapotranspiration was beyond the scope of this report. However, estimates of evapotranspiration rates and extinction depths were made from data collected for this study. Cowdery (2004) showed that diurnal water-level oscillations at well G15 during times of no precipitation were caused by evapotranspiration from an adjacent wetland. Extinction depth was estimated to be 4.6–4.9 ft below land surface on the basis of depth at which the water table had to fall before the diurnal oscillations ceased. The oscillations were analyzed similar to aquifer tests, with the wetland functioning as the pumping well and hydraulic conductivity of the aquifer material fixed at the value obtained from an independent slug test. Aquifer-test curves were matched to the oscillation data by adjusting the evapotranspiration discharge from the wetland, thereby producing the estimate. Re-analysis of these data produced a maximum evapotranspiration rate of 0.20 gal/min from the wetland during midday on July 29, 2003. Total evapotranspiration for that day was 90.6 gal, which equals 0.004 in. over the area of the wetland. An evapotranspiration estimate for the mass-balance basins would be possible by measuring the basin area where the water table is within 4.9 ft of land surface.

Withdrawals of ground water by wells were assumed to be negligible in this mass balance. During this study, substantial amounts of ground water were withdrawn from surficial aquifers in the basin by two City of Crookston municipal wells completed at depths of 50 to 54 ft in surficial aquifers. These wells produced an average of 107 Mgal/yr during 2002–4 (329 acre-ft/yr). The wells are outside any gaged surface-water basin but within 2,000 ft of the SW6 basin. The cone of depression of these wells had a diameter of about 1 mi, with less than one-half lying within the gaged basin (figs. 10 and 3). The well withdrawals amount to 7 percent of the unmeasured losses in the mass balance from the SW6 basin, and the amount of well withdrawals actually contained in the mass balances of SW6 basin is probably less than one-half of this. No other basin was affected by substantial well withdrawals.

In the study area, ground water and surface water function as components of one integrated hydrologic system. Therefore, it is reasonable to assume that ground-water basins are closely coincident with surface-water basins. Any error introduced into the mass balances from this assumption would be small. Mass balances computed for each gaged surface-water basin were summed to produce a total for all gaged basins, and this total is referred to as the "mass balance for the study area" even though this total represents only 33 percent of the study area.

Ground water mass-balance volumes, percentages, and yields are listed by basin in table 4. The computed unmeasured-loss terms in these mass balances are mostly composed of ground-water discharge to closed basins and evapotranspiration. Areal recharge is the total amount of water available to the ground-water system, excluding leakage from buried aquifers through till confining units. This leakage was ignored in the mass balance because the areal distribution and rates are not well known but are relatively small. Recharge to the buried aquifer in which well E01D is completed averaged 1 in/yr during 2004–5 based on rises in the ground-water hydrograph. An equal amount of water must leak out of the buried aquifers because changes in storage in confined aquifers is very small. Most of the water leaking out of buried aquifers probably moves vertically upward to the land surface or to the bottom of surficial aquifers, although a small amount is withdrawn by wells or may leak out to other adjacent, down-gradient buried aquifers. The average annual areal recharge (as aquifer area yield) to surficial aquifers was 16.34 in. during 2004–5 (table 4). Making the assumption that all water entering buried aquifers in the study area leaks to the surface in an areally even fashion, the amount of water entering the base of the surficial aquifers as leakage from buried aquifers amounted to 6.1 percent of the average areal recharge during that period.

Water available to the ground-water system (areal recharge) was 47 to 59 percent of the water available to the surface-water system (total precipitation - areal recharge to surficial aquifers + ground-water discharge to ditches) during water years 2003–5 (hereinafter in the mass-balance discussions, water year is implied). This amount ranged from 21,557 to 32,450 acre-ft (51 percent increase) between the driest year (2003) and the wettest year (2005). During the driest year of 2003 (when SCAN station precipitation was the 7th percentile of precipitation during 1890–2005 at Crookston, Minn., (High Plains Regional Climate Center, 2006)), ground-water storage decreased by 7,856 acre-ft as water was lost to surface waters, particularly closed-basin wetlands. This amount is more than one-third of the areal recharge received that year. Precipitation in year 2004 returned to near normal (47th percentile), and ground-water storage increased to more than make up for the deficit in 2003. Discharge to ditches, however, remained at 2003 levels as recharge increased aquifer storage. This storage was recorded as an annual increase in water levels in the hydrographs at all surficial aquifer wells. Unmeasured losses dropped to 59 percent of that computed for 2003. In 2005, precipitation was even higher (65th percentile) than in 2004. However, because aquifers were relatively full, ground-water discharge to ditches, closed surface-water basins, and

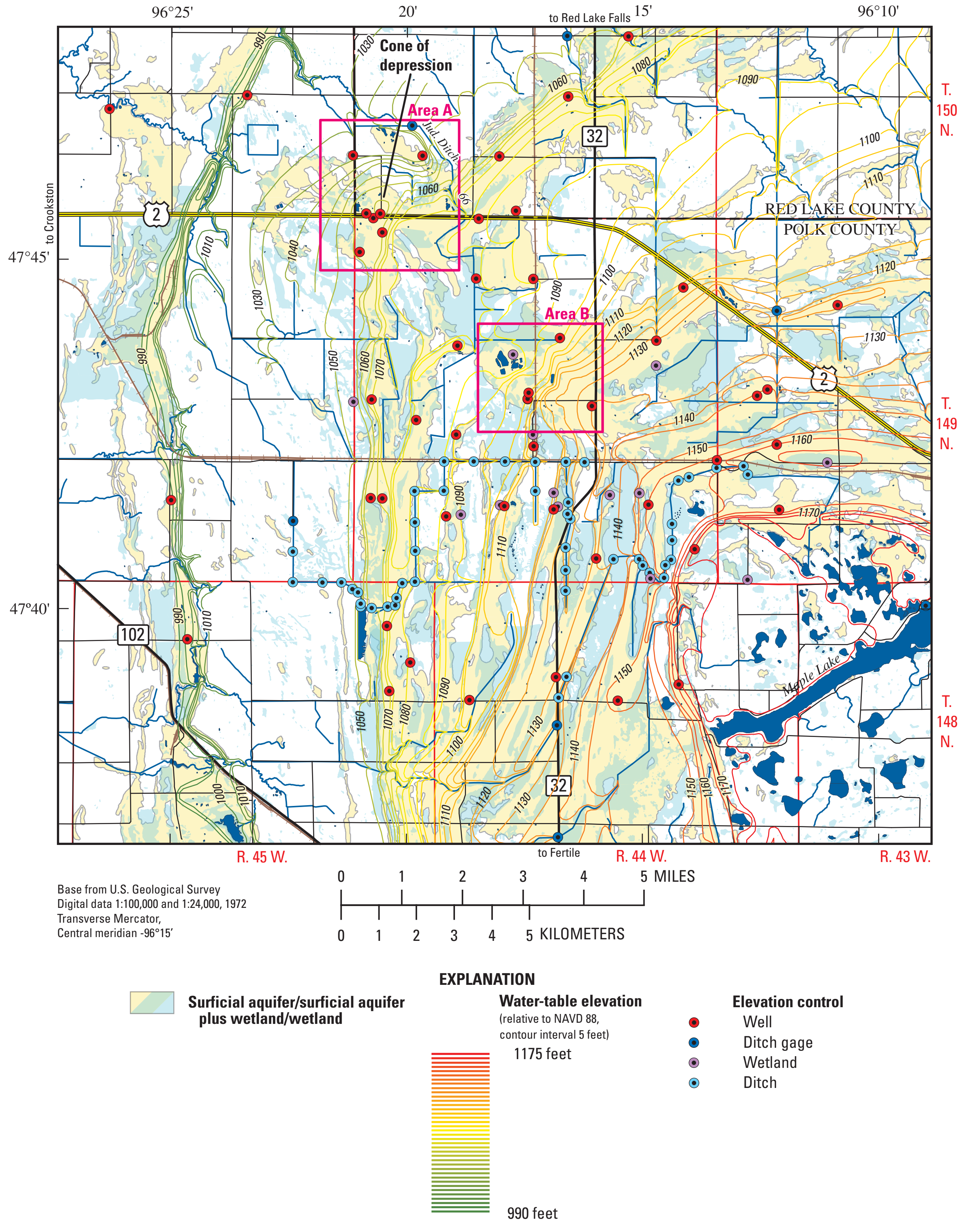


Figure 10. Water table, Glacial Ridge study area, northwestern Minnesota, June 2004.

Table 4. Net ground-water mass balance, Glacial Ridge study area, northwestern Minnesota, water years 2003–5.

[GW, ground water; Δ, change in]

Ditch basin	GW areal recharge[a]	- GW discharge to ditches	- Δ GW storage	= Unmeasured losses[b]
Volume of water per year, in acre-feet				
Water year 2003				
SW1	3,296	554	-1,053	3,795
SW2	3,385	736	-1,012	3,662
SW3	3,827	569	-1,034	4,292
SW4	4,241	1,210	-1,296	4,327
SW5	2,921	448	-1,496	3,968
SW6	3,886	1,182	-1,965	4,670
Total	21,557	4,699	-7,856	24,714
Water year 2004				
SW1	4,564	951	993	2,620
SW2	4,674	549	1,034	3,091
SW3	3,540	335	947	2,257
SW4	3,220	894	1,011	1,315
SW5	5,286	588	2,006	2,693
SW6	6,571	1,462	2,422	2,687
Total	27,854	4,779	8,413	14,662
Water year 2005				
SW1	5,000	3,805	-119	1,313
SW2	5,036	1,648	-227	3,615
SW3	4,768	1,501	-66	3,333
SW4	4,940	2,650	108	2,181
SW5	4,811	1,321	-644	4,134
SW6	7,895	2,394	-417	5,919
Total	32,450	13,319	-1,364	20,495
Percentage of total volume				
2003 total	100	22	-36	115
2004 total	100	17	30	53
2005 total	100	41	-4	63
Aquifer area yield, in inches				
Water year 2003				
SW1	11.11	1.87	-3.55	12.80
SW2	11.76	2.56	-3.51	12.72
SW3	12.86	1.91	-3.47	14.42
SW4	12.91	3.68	-3.94	13.17
SW5	12.60	1.93	-6.45	17.11
SW6	9.09	2.77	-4.60	10.93
Water year 2004				
SW1	15.39	3.21	3.35	8.83
SW2	16.24	1.91	3.59	10.74
SW3	11.90	1.13	3.18	7.58
SW4	9.80	2.72	3.08	4.00
SW5	22.79	2.53	8.65	11.61
SW6	15.38	3.42	5.67	6.29
Water year 2005				
SW1	16.86	12.83	-.40	4.43
SW2	17.49	5.72	-.79	12.56
SW3	16.02	5.05	-.22	11.20
SW4	15.03	8.06	.33	6.64
SW5	20.75	5.70	-2.78	17.83
SW6	18.48	5.60	-.98	13.85

[a] Excluding discharge from buried aquifers.

[b] Discharge to closed basins, evapotranspiration

evapotranspiration increased substantially as aquifer storage actually decreased slightly. In all years, unmeasured losses—that is, primarily discharge to closed surface-water basins and, to a lesser degree, evapotranspiration—was a greater loss of ground water than was loss to ditches (base flow out of the study area). This finding is related to the fact that the area of closed-basin wetlands adjacent to aquifers is much greater than is the area of ditches and basins that flow to ditches. The ratio of these areas is constantly changing, however, with the levels of surface waters. The ground-water mass balance estimates (table 4) show not only the dynamic interactions between ground and surface waters but also the importance of antecedent conditions on the movement of water.

Aquifer area yield is the amount of water per aquifer area that moves yearly between the terms of the mass balance (table 4). The yields show the variability of water movement in the study area in space and time. Differences in yield for areal recharge, change in ground-water storage, and discharge to closed basins for each basin are closely related to total precipitation. However, differences in discharge to ditches are not related to total precipitation and are probably more closely related to the number, area, and geometry of the aquifers crossed by the ditches. Generally, the year-to-year differences among basin yields show no consistent pattern because each basin is very sensitive to the individual hydrologic history affecting it.

Ground-water discharge is responsible for recessions in ground-water hydrographs (fig. 8). The rates of these recessions relate to the rates of ground-water discharge and vary among aquifers. Recession rates following recharge in late October 2004 fall into three categories. Wells G01, G12, and E05 had high recession rates; wells G20 and G25 had low recession rates; and all other wells had intermediate recession rates. Wells fell into the same recession-rate categories after most recharge events throughout the study period. It is difficult to explain why a particular well is in a particular category. However, some patterns exist. For example, high-recession-rate wells G01 and G12 are in basin SW1 (along with medium-recession-rate well E01S), implying that this basin has relatively more ground-water discharge to surface waters. Basin SW1 frequently had relatively high ground-water-to-ditch discharge volumes and relatively low unmeasured losses. This pattern indicates that ditches in basin SW1 are relatively well connected to aquifers. This deduction is supported by the low ditch discharge during the dry year of 2003. During that year, lack of precipitation left aquifer water levels low, with little remaining to sustain ditch flow. Therefore, high ground-water discharge to ditches quickly emptied aquifers at the beginning of a drought, resulting in low overall ditch discharge for the drought year as a whole. The mass balances in table 4 contain other details of aquifer function not detailed here.

Ground-Water Levels and Flow

Ground water flow is from areas of high elevation near Maple Lake to those of lower elevation on northern and western edges of the study area. This pattern is clear on maps of ground-water potentiometric surface (figs. 10 and 11), where flow is perpendicular to potentiometric contours. Shallow ground-water flow is complex, with water in surficial aquifers, ditches, and wetlands part of a single hydrologic system. Ground-water flow in buried aquifers does not interact with the surface directly except in areas A and B (fig. 11), where withdrawals from wells and hydraulic connection to surficial aquifers, respectively, affect flow locally.

Surficial Aquifers

Surficial aquifers contain and lie adjacent to wetlands of a variety of types within the study area. In many areas, wetlands underlain by poorly permeable till lie in back-beach basins, both upgradient from, downgradient from, and in physical contact with individual aquifers. In these situations, it is difficult to distinguish surface- and ground-water flow, and it is more accurate to think of both waters as flowing in one hydrologic system. For this reason, wetland areas are shown along with aquifer areas on figure 10, and water-table elevation contours are drawn across the entire study area. That said, there is essentially no water flow in till areas where there is neither aquifer nor wetland, even though a water-table contour shows the potential for such flow.

The water table was very shallow, lying between 0 and about 28 ft below land surface in late June 2004, a time of typical water levels during the study. The median measured water depth in wells completed in surficial aquifers was 6.76 ft at that time. The water-table depth is zero where an aquifer is in hydrologic contact with a wetland, lake, or ditch. Therefore, the water table mimics the topography in the study area. Although the land surface appears quite flat, there is a 250-ft drop in elevation from Maple Lake to the southwestern corner of the study area (a distance of about 11 mi). This gradient is substantial and is the force that drives the flow of water in the study area.

The basic radial pattern of ground-water flow is interrupted where ditches cut through aquifers, particularly those formed in beach ridges. Usually, ground water flows perpendicular to beach ridges, which lie parallel to lines of equal elevation because of their formational history. Where a ditch cuts through a beach-ridge aquifer, ground-water flow turns 90 degrees to flow toward the ditch, where it usually discharges. How far from a ditch this turn in ground-water flow occurs is variable. The greater the saturated thickness of the aquifer, the farther away from the ditch the turn will occur. If the saturated thickness of a beach ridge aquifer is great enough, a ground-water mound can occur in the aquifer in areas away from ditches, and ground water will flow locally both with and against the general regional flow direction. This situation occurs especially where a ditch drains an upgradient wetland, producing a locally low water table upgradient from a beach ridge.

Ground-water flow was affected by induced ground-water discharge in areas A and B (fig. 10). In area A, the City

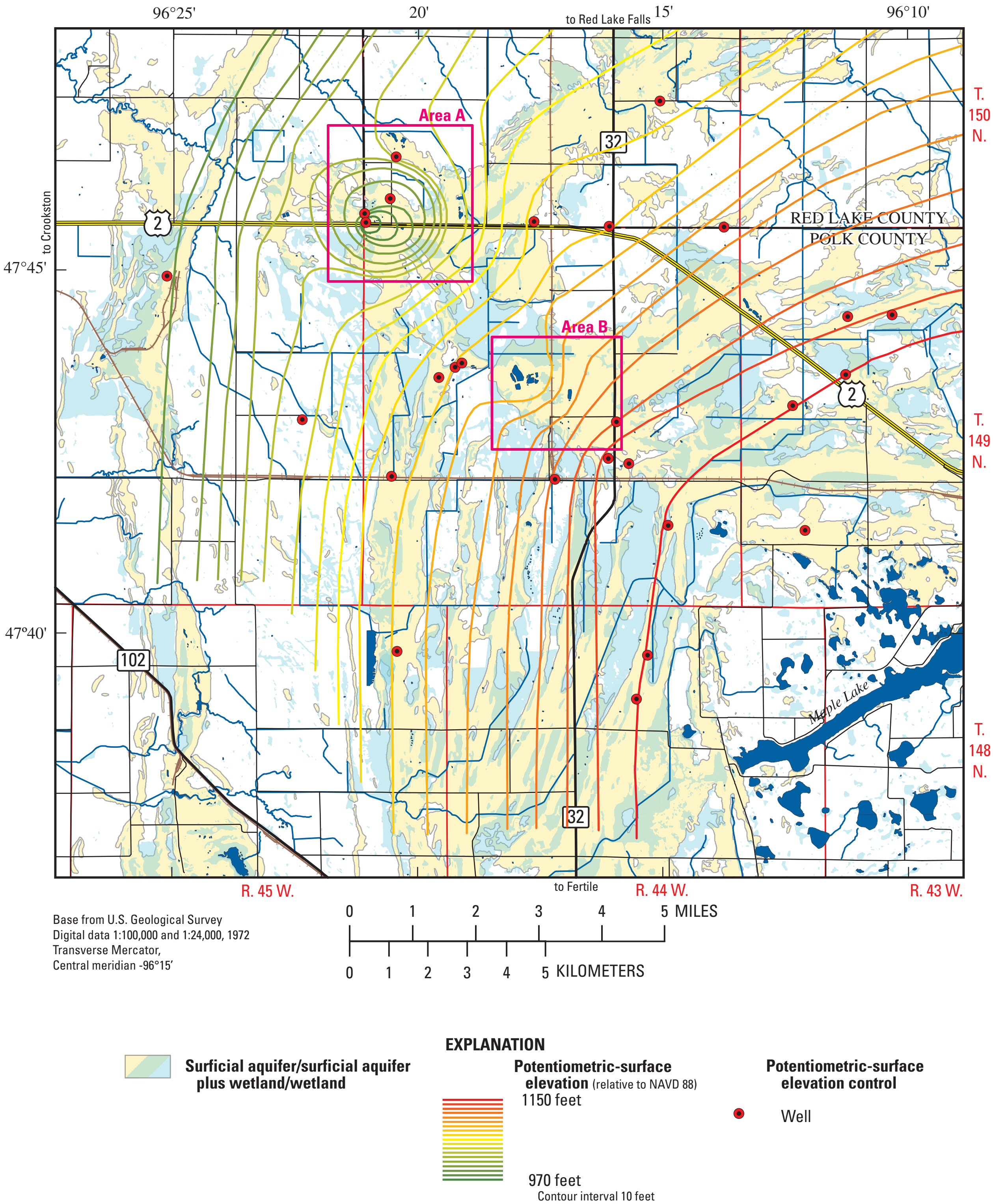


Figure 11. Potentiometric surface of buried aquifers, Glacial Ridge study area, northwestern Minnesota, June 2004.

of Crookston withdrew ground water with four high-capacity wells; two pump from a surficial aquifer and two pump from a buried aquifer. Two other wells, located east of area B in a new well field, were not used until July 2005. During 2002–4 the water department reported average withdrawals of 290 Mgal/yr (890 acre-ft/yr), of which 62 percent was withdrawn from buried aquifers and 37 percent was withdrawn from a surficial aquifer (Richard Normandin, Water Department, Crookston, Minnesota, written commun., 2004). These withdrawals caused a 20-ft-deep cone of depression in the water table in the area of the pumping wells. The cone was caused by withdrawals from the surficial aquifer and possibly by leakage to the buried aquifer through an intervening 70–85 ft-thick sandy till. Total annual recharge in the area of the cone of depression was estimated to be about 358 Mgal (1,100 acre-ft) (assuming an average recharge rate of 18.44 in. (the averages calculated from data at wells G08 and E05 recharge during 2003–5) and an area of 716 acres). The amount directly pumped from the surficial aquifers is 28 percent of this total. The amount of surficial ground water leaking to the underlying buried aquifers is unknown.

In area B, lakes in the bottom of a gravel pit receive constant ground-water discharge from surficial and buried aquifers that then flow through an outlet to Judicial Ditch 66 (the SW6 basin). This discharge has produced a cone of depression on the order of 10 ft deep in the surficial aquifer.

Buried Aquifers

The potentiometric surface of water in the buried aquifers formed a relatively smooth, sloping surface shaped much like one-quarter of an inverted shallow cone (fig.11). The degree to which these aquifers are physically or hydrologically interconnected is unknown. The water levels measured in these aquifers in June 2004 are consistent with aquifers that are in hydrologic connection. The general pattern of ground-water flow in buried aquifers was very similar to that in surficial aquifers except that the pattern is unaffected by surface waters (figs. 10 and 11). The potentiometric drop across the study area was 180 ft; nearly the same as the drop for surficial aquifers. Vertical head gradients in buried aquifers were upward. The 29 wells completed in buried aquifers that were measured in this study were always artesian unless being pumped and 8 wells had heads above land surface.

The buried-aquifer potentiometric surface had one cone of depression in area A and may have a flattening of the surface in area B (fig. 11). The depression in area A resulted from pumping from two Crookston high-capacity wells completed in the buried aquifer. During 2002–4, 180 Mgal/yr (551 acre-ft/yr) was pumped from the buried aquifer (Richard Normandin, Water Department, Crookston, Minnesota, written commun., 2004), resulting in a cone of depression more than 60 ft deep. The depth of this cone may have been lessened by leakage through the confining unit from the overlying surficial aquifer.

The flattening of the potentiometric surface in area B (fig. 11) is inferred, as is the discharge assumed to be producing it. The discharge from the gravel pit lakes in area B, estimated from minimum flows in Judicial Ditch 66, was 14, 43, and 290 acre-ft/yr in 2003, 2004, and 2005, respectively. The minimum flows in other ditches during 2003 and 2005 were 0–0.01 ft^3/s, indicating that discharge from surficial aquifers had virtually ceased. However, ditches SW2 and SW4 barely continued to flow during 2004 at a minimum of 0.02 ft^3/s and 0.20 ft^3/s, respectively. Judicial Ditch 66 is the only one in the study area that continually flows and this flow originates from the gravel pit lakes. No other surficial aquifers in the study area are capable of supplying year-round flow to ditches, so it is likely that the surficial aquifers around the gravel pit lake are not capable of supplying this discharge either. One explanation for the flow in Judicial Ditch 66 is that it is maintained by discharge from a buried aquifer that is being mined for gravel several tens of feet below the lake surface. In this case, a confining unit may not have separated the surficial and buried aquifers in this area or perhaps it was removed by mining. A conduit may exist from the buried aquifers to the gravel pit lakes through which ground water may be discharging under artesian pressure. If the buried aquifers are the source of water discharging to the gravel pit lakes, the head in the buried aquifer would be at the elevation of the lake surface at the bottom of the lake, producing the flattening in the potentiometric surface shown in area B (fig. 11).

The source of water in buried aquifers and other discharge areas is not well known. No obvious water source to the buried aquifers exists within the study area other than horizontal flow from adjacent buried aquifers in the southeastern part of the study area, because head gradients are upward, preventing direct recharge through interconnections with the surficial aquifers, and leakage through overlying till and lake clays. Likewise, the only plausible discharge areas aside from the point locations noted above is diffuse upward discharge to surficial aquifers or surface waters through thick confining units of till and lake clay. The fact that the potentiometric surface is generally uniform and planar but not horizontal (slope, 0.0038) indicates that horizontal flow occurs in the buried aquifers and that discharge is evenly distributed along a flow path. Assuming that

- the buried aquifers are in hydraulic connection with each other and cover most of the study area,
- ground water only discharges vertically upward through the till and clay confining unit,
- the hydraulic gradient is uniform over the study area (fig. 11, except areas A and B), and
- change in aquifer storage is small (valid for a confined aquifer),

then recharge will equal discharge and will be constant across the aquifer. Using the average recharge measured at well

E01D (1 in/yr = 0.00023 ft/d), and Darcy's Law, (0.00023 ft/yr = hydraulic conductivity (ft/d) x 0.0038) the vertical hydraulic conductivity of the buried aquifers is 0.06 ft/d. This conductivity falls near the middle of the range of hydraulic conductivity reported by Fetter (1988) for silt, sandy silts, clayey sands and till (0.003–0.3 ft/d). Hypothetically, the vertical discharge of buried ground water occurs evenly across the study area and at a rate of about 1 in/yr. Where buried ground water discharges beneath a surficial aquifer, this water enters as leakage and adds to surficial ground water. Elsewhere, discharge of buried ground water makes its way to surface water, adding to the mass balance of surface waters. Leakage from the buried aquifers was not accounted for in the mass balances presented above. This leakage is estimated to amount to 6.1 percent of areal recharge to surficial aquifers and would increase the unmeasured loss terms in the mass balances by the same percentages.

Ground-Water Age from Dissolved Gases

Ground-water age, which is the amount of time elapsed after water enters the ground as recharge, is useful for understanding the ground-water flow system, calibrating ground-water models, and delineating well contributing areas. Water recharged comparatively recently (within the past 50 years) is termed "young." Samples of water from 32 wells were analyzed for dissolved sulfur hexafluoride (SF_6) gas concentration in order to estimate ground-water age. The concentration in one of these samples could not be determined. Water from seven wells (including a well completed in a buried aquifer) was contaminated by local nonatmospheric sources of SF_6. Therefore, useable ground-water ages were determined at 24 wells. Busenberg and Plummer (2000) detail the technique of using SF_6 to estimate ground-water age.

In general, areal recharge infiltrates to the water table, builds up, and flows horizontally. Thus, ground water is progressively older with depth below the water table and along its flow path. The deeper water occurs beneath the water table, the farther upgradient the water entered as recharge and the older it is. The thickness of a water layer that contains recharge from a given year will decrease with depth because some water is lost to discharge each year. In all age dating, water is assumed to move like piston flow from recharge to discharge areas. However, leakage of water from surface waters also is important in the study area. Surface water entering the aquifer may be a complex mixture of runoff, buried-aquifer discharge, and direct precipitation. Therefore, the apparent ages determined with SF_6 may be from waters of many sources and ages, and not just water from areal recharge.

Ground-water ages are composites of all water that entered the well screen. In this study, most well screens were 4–5 ft long. The well screen of well E03 was 10 ft long. The age of water collected from wells that have shorter screens are more precise. Assuming that water is drawn into the well evenly along the length of the screen and completely mixed, young waters will provide exponentially more SF_6 to the composite sample because the concentration in the atmosphere increases exponentially with time (Busenberg and Plummer, 2000). Therefore, the resulting composite age from a well with a longer screen is skewed somewhat younger.

Seven of the 32 dissolved-gas samples contained concentrations of SF_6 higher than possible for water in equilibrium with the 2004 atmosphere. This is possible only if the water contains excess air trapped in the aquifer, usually during recharge, or if it is contaminated by SF_6. Concentrations of argon and nitrogen can quantify the amount of excess air present in a sample. Even after correcting SF_6 concentrations for excess air, these seven samples remain unrealistically high in SF_6, indicating contamination. The source of this contamination is unknown. Contaminated samples were not included in data analysis for this report.

Ground-water age determinations for this study are given in table 5. All sampled waters were less than 15 years old, and one-third (8 of 24 samples) were less than 5 years old. The general youth of the ground water represents the nature of the aquifers, which are thin, shallow, and dominated by areal recharge. Compounding this, samples were withdrawn from wells with short screens that intersected or were close to the water table, where the youngest water lies.

Surface-Water Hydrology

Direct runoff is that surface-water flow that moves over the ground surface, primarily from back-beach basins, into a network of channels without becoming ground-water flow. In the study area, the channels are dry most of the time and most are not visible except when carrying water. The time that water takes to drain from the land surface as direct runoff is affected by the shape, size, and slope of the drainage basin, vegetation type, antecedent moisture conditions, and temporary storage in and adjacent to the stream channel.

The channel network in the study area is modified by several manmade ditches. These ditches were installed to remove excess water during spring snowmelt so that fields could be planted earlier and to remove excess water from summer storms, reducing standing water. The ditches shorten the time that water takes to drain from the land surface and increase the volume of direct runoff. The ditches also drain ephemeral wetlands and reduce the size of permanent wetlands.

The back-beach basin wetlands are very important to direct runoff because they temporarily store water. These wetlands delay direct runoff and also can reduce ditch flow by permanently storing water when the level falls below the outlet elevation. Most of the water permanently retained is returned to the atmosphere as evapotranspiration.

Beach-seep fens are less important to direct runoff because they have very small capacity to retain water. During very wet conditions, ground water discharging into the beach-seep fens may flow overland into a back-beach basin wetland, thence into a channel flowing parallel to a beach ridge and eventually into a ditch. However, such wet conditions are rare and form a very small part of the ditch flow.

Table 5. Ground-water age and related well data, Glacial Ridge study area, northwestern Minnesota, 2004.

[Depths are below land surface; a negative value for "water-level to screen-top distance" indicates the screen top is above the water table; N, nitrogen; WL, water level; °C, degrees Celsius; mg/L, milligrams per liter; C, contaminated; <, less than; —, not available]

Well name	Recharge date[a]	Recharge temperature °C	Excess air mg/L	Excess N_2 gas mg/L	Nitrate as N mg/L	Methane mg/L	Oxygen mg/L	WL to screen-top distance feet	Screen length feet	Screen top depth feet	Screen bottom depth feet	Well depth feet	Water level depth feet
G01-R	C	9.6	3.5		< 0.03	0.36	0.29	0.79	4.30	5.58	9.88	10.42	4.79
G02	1996	12.7	7.4		3.89	.00	1.99	2.73	4.30	9.65	13.95	14.49	6.92
G04	2005[b]	7.6	1.8		19.50	.00	7.51	-4.00	4.30	5.01	9.31	9.85	9.00
G06	1999	9.1	.1	2.9[c]	19.46	.00	3.86	.63	4.30	8.10	12.40	12.94	7.47
G07D	C	9.6	4.0		4.80	.00	1.57	20.51	4.30	31.21	35.51	36.05	10.70
G07S	C	6.1	2.8		.35	.00	5.23	-.10	4.30	10.65	14.95	15.58	10.75
G08-R	C	12.3	6.2		42.98	.00	1.26	.58	4.30	6.97	11.27	11.81	6.39
G09	C	8.2	7.8		1.45	.00	.20	2.38	4.30	5.67	9.97	10.60	3.29
G10	2005[b]	1.9	.5		1.34	.00	5.17	-.57	4.30	3.01	7.31	10.23	3.58
G12-R	—	9.0	6.1		.33	.04	.11	2.80	4.30	8.76	13.06	14.94	5.96
G13	2005[b]	6.8	.4		.14	.00	7.90	-2.93	4.30	4.14	8.44	13.92	7.07
G14	2004	10.4	6.8	1.5	1.97	.00	.08	1.94	4.30	5.58	9.88	15.54	3.64
G15	1999	12.2	5.8	2.5[c]	15.09	.00	2.61	4.18	4.30	7.17	11.47	14.87	2.99
G16	2003	8.7	.9		.95	.00	4.67	3.02	4.30	6.70	11.00	14.41	3.68
G17	2005	16.7	.1	9.9	10.58	.00	.88	3.19	4.30	8.63	12.93	13.47	5.44
G18	1997	8.4	1.2	5.0	2.55	.00	.08	7.23	4.30	14.97	19.27	29.87	7.74
G19	1990	2.2	2.8		13.80	.00	6.27	15.38	4.30	38.60	42.90	43.53	23.22
G19	1991	1.9	3.0		13.80	.00	6.06	15.38	4.30	38.60	42.90	43.53	23.22
G20	2005[b]	10.8[d]	2.6		6.30	.00	3.80	-2.11	4.30	6.34	10.64	15.01	8.45
G21	1991	9.2	3.5		11.40	.00	4.92	.76	4.30	7.49	11.79	12.33	6.73
G23	1993	7.3	3.5		2.87	.00	4.48	4.37	4.30	13.86	18.16	24.56	9.49
G24	1995	6.3	3.2		.03	.00	1.53	.79	4.30	6.86	11.16	11.70	6.07
G25	1995	7.5	1.7	2.0	.50	.00	.08	9.03	4.30	20.38	24.68	30.35	11.35
G26	1993	5.3	.9		1.20	.00	7.55	1.35	4.30	10.05	14.35	14.89	8.70
G27	2003	9.0[e]	.0[e]		2.11	—	—	-.04	4.30	5.27	9.57	10.11	5.31
G33	1994	8.0	2.3		6.67	.00	6.84	3.97	4.30	20.32	24.62	25.25	16.35
G35	1991	12.6	6.1		22.18	.00	3.64	1.29	4.30	12.18	16.48	19.88	10.89
G36	1996	6.1	2.7		1.19	.00	3.24	.88	4.30	5.17	9.47	10.01	4.29
G38	1993	9.9	2.3		.25	.00	1.94	-.60	4.30	2.91	7.21	14.61	3.51
E03	C	7.6	4.1	9.0	5.305	.00	.05	51.13	10	59	69	69	7.87
E04	C	7.5	6.2		< .03	.02	.07	83.02	4	98	102	102	14.98
L043	1990	9.1	4.0		2.66	.00	6.06	3.02	5	18	23	24.26	14.98
L057	1993	6.5	2.1		4.71	.00	5.67	2.78	5	9.5	14.50	13.48	6.72

[a] Samples collected April–July 2004.
[b] Rounding to the nearest year produces recharge date after sampling date.
[c] Assumed denitrification despite non-zero oxygen concentration to reduce unrealistically high recharge temperature estimate.
[d] Used a recharge temperature 0.1° Celsius lower than the dissolved gas estimate to avoid contamination designation.
[e] No gas data for recharge temperature or excess air estimates. Estimate based on typical values for other sites.

Five ditches (SW1, SW3-SW6) that drain TNC property were gaged (table 2). One additional gage (SW2), on a ditch that drains land adjacent to TNC property, was installed to collect data in a control basin because its land use was not expected to change substantially. The physical characteristics of the basins in the study area are all similar, having relatively small ranges for most characteristics. Notable differences include SW4 has a shorter main-channel length (4.1 mi) than the other basins; SW1 has a greater main-channel slope (14.0 ft/mi) than the other basins; and SW5 has a smaller percentage of lakes and wetlands than the other basins. Only SW2 and SW3 had any lake area, and both areas were less than 1 percent of the total drainage.

Description of Daily Flows

Differences in daily flows among basins (table 6) result from differences in physical characteristics of those basins, including differences in land use. As land use changes in the future, the character of these daily flows also may change.

Except for SW3, which has a low mean annual runoff of 2.26 in/yr, the mean annual runoff is fairly consistent among the gaged basins, ranging from 3.69 to 4.12 in/yr (table 6). An examination of SW3 shortly after a rainstorm indicated that very little water leaves the upper part of the basin, even though the ditch is unobstructed. Water pools after rainfall or snowmelt near Bakken Lake, with a portion of the pooled water recharging ground water and the remainder flowing out of that basin into SW4 or SW6. In contrast to other basins, SW6 had flow (table 6), which was sustained by a relatively constant supply of ground-water discharge to gravel pit lakes located about 1.5 mi north-northwest of Tilden Junction.

Figure 12 shows hydrographs for a period with several rainfall events from August 20 to October 20, 2004. SW5 and SW6 show the greatest response to rainfall. SW5 has the highest main-channel slope, excluding SW1 which is much less responsive to rainfall because the ditch flows through the Burnham Creek Impoundment above the gage. The impoundment stores water and releases it slowly. SW6 has a relatively high main-channel slope and the largest drainage area, both of which would contribute to high rainfall responsiveness. SW3 consistently shows little response to rainfall, which is in agreement with observations of pooling in the upper part of the basin.

Ditch flow was measured at gages SW1–SW6 during water years 2003–5. The highest recorded daily flows at each gage are listed in table 7 for each year. The highest recorded daily flow, 116 ft^3/s, occurred at gage SW5 on October 30, 2004. In general, however, the highest flows at most gages were recorded in 2005, which also corresponded to the highest runoff volumes (in inches, table 7), whereas the lowest flows and runoff volumes occurred in 2003. This correlates directly to precipitation recorded at the SCAN station, which had the lowest total in 2003 and the highest in 2005. High flows at three gages in 2004 were estimated because of freezing conditions. Estimates were based on comparisons to nearby gages, discharge measurements, and climate data (Rantz and others, 1982). The lowest of the highest daily flows occurred at gage SW3 on June 25, 2003 with only 20 ft^3/s.

At five of the six gages, no flow occurred in most years, as shown in table 7. Periods of no flow occur when ground water no longer discharges from surficial aquifers to the ditches. SW1 maintains flow when the elevation of the water in the Burnham Creek impoundment is higher than the elevation of the outfall. The only gage where continuous flow occurred through all three years was gage SW6. The continuous flow at this gage is the result of ground-water flow from gravel pit lakes in the upper part of the basin.

Base Flow and Surface-Water Mass Balance

The flow hydrographs for the period of record of all gages were analyzed to determine the sustained contribution of ground water to flow in the ditches (base flow). The other component of flow in the ditches is direct runoff, which represents precipitation that runs overland to ditches. The amount of base flow is controlled by the elevation of the water table adjacent to the ditch.

Based on the drainage area, the recession period after the peak (N) was computed as 1.6 days for gages SW1–SW5 and 1.7 days for gage SW6 by the computer program PART (Rutledge, 1998). The program was used to calculate base flows

Table 6. Flow characteristics of ditch basins, Glacial Ridge study area, northwestern Minnesota, water years 2003–5.

[ft^3/s, cubic feet per second; in/yr, inches per year; d/log, days per log cycle; e, estimated]

Basin name	Minimum daily flow (ft^3/s)	Maximum daily flow (ft^3/s)	Maximum flow date (ft^3/s)	Mean flow	Mean runoff (in/yr)	Recession index (d/log)
SW1	0	88	10/30/2004	3.48	3.98	13.95
SW2	0	96	10/31/2004	3.06	3.97	10.17
SW3	0	e41	3/28/2004	1.79	2.26	11.32
SW4	0	91	10/30/2004	2.93	4.12	12.59
SW5	0	116	10/30/2004	3.12	3.69	7.86
SW6	.02	108	10/31/2004	4.18	3.97	17.64

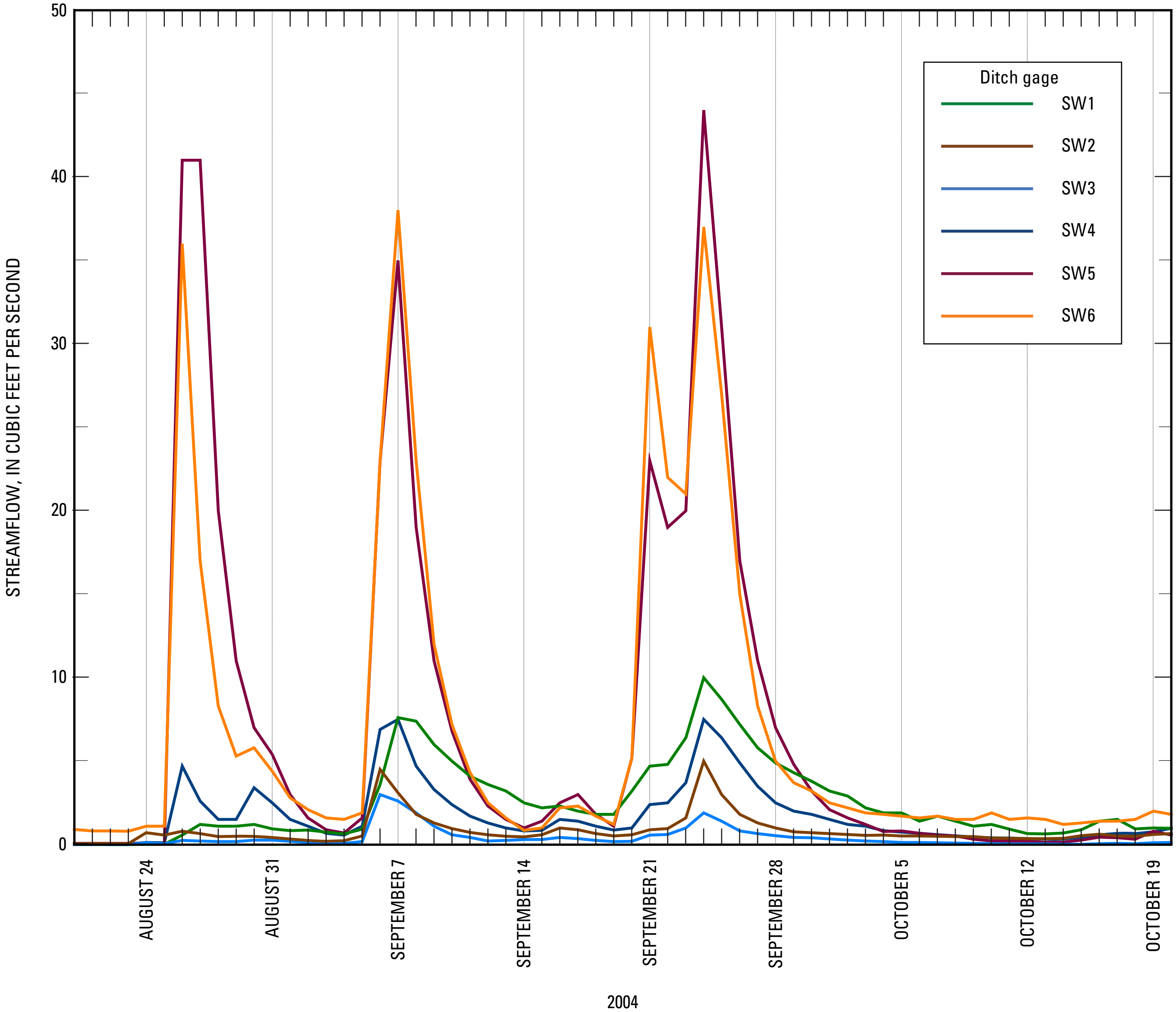


Figure 12. Ditch hydrographs, Glacial Ridge study area, northwestern Minnesota, August–October 2004.

for values of N equal to 1, 2, and 3 days to bracket the computed recession-period values. The three resultant base-flow component estimates were summed for the period of record for each of the six basins. Each of the three summed base-flow estimates was compared to the summed observed flow to compute the percentage of flow due to ground-water discharge (table 8). In addition, the average of the observed daily flows and base flows are listed in table 8 to highlight the variation among basins. An illustration of how base flow differs among the three values of N at SW4 is shown in figure 13. The example is from May–June 2004. As N increases from 1 to 3, the point at which the direct runoff ends and all ditch flow is base flow occurs later (farther to the right of the peak on the downward side of the recession) and at a lesser flow. Thus, higher values of N represent more direct runoff and less base flow. In the storm shown in figure 13, the PART-estimated base flow for N=2 and N=3 is exactly the same, although this is not always the case. Generally, the direct runoff computed with the 2-day recession period (N=2) agrees best with simulations of direct runoff from the Clark (1945) unit-hydrograph analysis considering all gages and all storms analyzed.

The estimated percentage of observed ditch flow that was base flow was highest for basin SW4 at 80, 71, and 64 percent for N=1–3, respectively. This estimated base flow was the lowest for basin SW5 at 44, 30, and 26 percent for N=1–3, respectively. This is consistent with earlier observations where flow was observed continuously at gage SW4 and where there were periods of zero flow at gage SW5. Wolock (2003)

Table 7. Ditch-flow statistics, Glacial Ridge study area, northwestern Minnesota, water years 2003–5.

[ft^3/s, cubic feet per second; e, estimated]

Water year	2003	2004	2005
Maximum daily flow (ft^3/s)			
SW1	64	e39	88
SW2	54	e70	96
SW3	20	e41	30
SW4	21	40	91
SW5	30	67	116
SW6	40	57	108
Minimum daily flow (ft^3/s)			
SW1	0	0	0
SW2	0	.02	0
SW3	0	0	0
SW4	0	.2	.01
SW5	0	0	0
SW6	.02	.06	.4
Runoff (inches)			
SW1	1.97	2.6	7.45
SW2	3.02	3.08	5.89
SW3	1.44	1.74	3.72
SW4	3.07	2.6	6.75
SW5	1.43	4	5.77
SW6	2.07	3.66	6.23

estimated base-flow percentages for the conterminous United States, which ranged from 41 to 45 percent of observed flow. Basins with substantially higher percentages than the Wolock estimates (SW1, SW3, and SW4) indicate higher sustained flow from aquifers and from impoundments and wetlands, which also can contribute to base flow. Basins with substantially smaller percentages (SW5) indicate lower sustained flow from aquifers and wetlands.

Surface-water mass balance can be estimated from total precipitation, ground-water recharge, and streamflow within a basin. Table 9 lists the surface-water mass balance for the six basins in the study area. The total precipitation for each basin was computed by allocating precipitation data at each precipitation gage to an area using the Theissen polygon method (Linsley Jr. and others, 1982). Areal recharge was estimated from WTF recharge estimates at six wells and allocated over the surficial-aquifer area in each basin using the Theissen polygon method. Available precipitation is equivalent to total precipitation minus areal recharge and represents the total of direct runoff; soil moisture (water retained in the soil and not recharged to aquifers); snow, which eventually melts and flows into one of the other components; and closed basin storage, including wetlands. Ground-water discharge to ditches was estimated from PART base-flow separation. Basin outflow is the measured flow at a ditch gage. The unmeasured losses represent the total precipitation lost as evapotranspiration from closed basins, ditches and other surface-water bodies, and soil moisture.

The bottom section of table 9 expresses the total amount of water in each basin for 2003–5 in terms of yield, which is useful for comparisons among basins. The total amount of water in the surface-water system is the sum of available precipitation and ground-water discharge to ditches. The percentage of ground-water discharge to ditches ranged from 8 percent in 2004 to 20 percent in 2005. That percentage varies from year to year because of differences in precipitation patterns and storage in the surficial aquifers. Storage in the surficial aquifers affects the gradient between the aquifer and the ditch, which drives flow from the aquifer into the ditch. The low percentage of ground-water discharge to ditches in 2004 is due to water going into storage in surficial aquifers from the relatively higher precipitation during 2004 after low precipitation in 2003. The unmeasured losses from the surface-water system, most of which is evapotranspiration, varied from 68 percent in 2005 to 81 percent in 2004. This is 2- to 4-times the amount of water leaving the basin through the ditch system. The relatively low percentage in 2005 was caused by water coming out of storage from surficial aquifers and the slightly greater total precipitation than in 2004. The actual amount of unmeasured losses is similar between 2004 and 2005.

The outflow from each basin varies considerably from year to year depending on the precipitation in that basin (table 9). For 2003 only data for the SCAN station were available, so these data were applied to the entire study area. The results for 2004 and 2005 reflect the actual variability of total precipitation among basins. This precipitation variability affects all of the yield values, so that normalizing for basin area still does not permit direct comparisons among basins.

Hydrograph-Recession Slope

Streamflow hydrograph-recession slopes are related to many physical characteristics of the basins. For example, wetlands near or adjacent to channels that can release water to a ditch slowly, cause lower recession slopes (higher recession indices). Consequently, as wetlands and prairies are restored in the study area, hydrograph-recession slopes may change (decrease).

Daily mean flow records for water years 2003–5 for basins SW1–6 were analyzed with the computer program RECESS (Rutledge, 1998) to compute the hydrograph-recession slope. A minimum recession period of 10 days was selected to obtain at least 7 days of linear recession after the antecedent recession (described in the base-flow determination part of appendix 3). Only recessions starting in May through September were used because the flow record is considered good throughout that period but poorer in other months.

Approximately 14 storm recession hydrographs were chosen for each basin and evaluated for linearity (a requirement of the analysis method), after which the recession rate

Table 8. Base-flow separation, Glacial Ridge study area, northwestern Minnesota, 2003–5.

[**Bold** indicates highest and *italic* indicates lowest percentages; *N*, number of days after the recession peak using the computer program PART (Rutledge,1998)]

Gage		Observed flow	Summation of daily flows (cubic feet per second-days) Base flow *N*=1	Base flow *N*=2	Base flow *N*=3	Number of days
SW1	TOTAL	3,836.03	3,007.77	2,509.46	2,141.32	1,117
	Percent observed		78	65	56	
	Average daily flow	3.43	2.69	2.25	1.92	
SW2	TOTAL	3,308.79	1,711.66	1,515.78	1,350.63	1,004
	Percent observed		52	46	41	
	Average daily flow	3.3	1.7	1.51	1.35	
SW3	TOTAL	2,043.74	1,462.11	1,175.62	1,035.57	1,117
	Percent observed		72	58	51	
	Average daily flow	1.83	1.31	1.05	.93	
SW4	TOTAL	3,262.01	2,620.52	2,325.68	2,096.13	1,116
	Percent observed		**80**	**71**	**64**	
	Average daily flow	2.92	2.35	2.08	1.88	
SW5	TOTAL	3,440.4	1,504	1,032.99	895.4	1,116
	Percent observed		*44*	*30*	*26*	
	Average daily flow	3.08	1.35	.93	.8	
SW6	TOTAL	4,637.14	2,958.46	2,433.46	1,961.5	1,116
	Percent observed		64	52	42	
	Average daily flow	4.16	2.65	2.18	1.76	

was determined for each storm. The median recession rate is called the recession index, expressed in days per common log cycle. The basin with the lowest recession index was SW5, with 7.86 days, and the basin with the highest recession index was SW6, with 17.64 days (more than 2 times that of SW5) (table 6). The high value for SW6 probably results from the sustained flow from ground-water discharge to the gravel pit lakes in that basin. The recession indices were generally higher for basins with greater percentage of wetland and lake area (table 2).

Storm-Runoff Hydrograph Modeling

The general shape of the crest and recession of the runoff hydrographs for a given basin should be consistent for different storms because this shape is a function of the constant physical characteristics of the basin, such as slope, channel length, drainage area, land cover, and infiltration (Linsley Jr. and others, 1982). The size and timing of the hydrograph peak are controlled by the variable characteristics of the storm and antecedent conditions. The Clark unit-hydrograph method (Clark, 1945) used in this study transforms excess rainfall into direct runoff. It is a method to define the general shape of a runoff hydrograph in a way to account for the differences in rainfall among storms and to model the actual runoff hydrograph for any storm. If the physical characteristics of a basin are modified, then the unit hydrograph of the basin also should change. The unit hydrograph is defined as the direct runoff hydrograph resulting from 1 in. of excess rainfall (rainfall available for direct runoff) for a specified period of time and uniformly distributed over a basin (Maidment, 1993). Clark unit-hydrograph models were constructed for this study to determine the variable values controlling storm runoff before wetlands and prairies were restored. Changes in these values after restoration will quantify the effects of restoration on how runoff behaves in the study area.

Clark unit-hydrograph models were constructed for Glacial Ridge basins SW1–6 for three storms in 2003–4 (table 10). Rainfall used in the storm-runoff hydrograph models were recorded every 60 minutes at four wells and at the SCAN station within the study area during June 9–10, 2003, and May 11–12 and 29–30, 2004. Total storm rainfall amounts ranged from 1.13 in. at well G01 on June 9–10, 2003, to 3.31 in. at well E03 on May 11–12, 2004 (table 10).

The precipitation for each storm in each basin was apportioned using the Theissen polygon method (Linsley Jr. and others, 1982) from the five precipitation gages (table 10). Precipitation from each storm in a basin was calculated from as many as four gages (SW1) and as few as one (SW2). The number of precipitation gages used to compile data for each storm varied because data at some gages were unavailable or unreliable. Where percentages are different between storms for a basin, the storms are listed separately.

The seven storm-runoff hydrograph model variables (table 11; appendix 3) were optimized for each Clark unit-hydrograph model in a two-step process. Initial variable

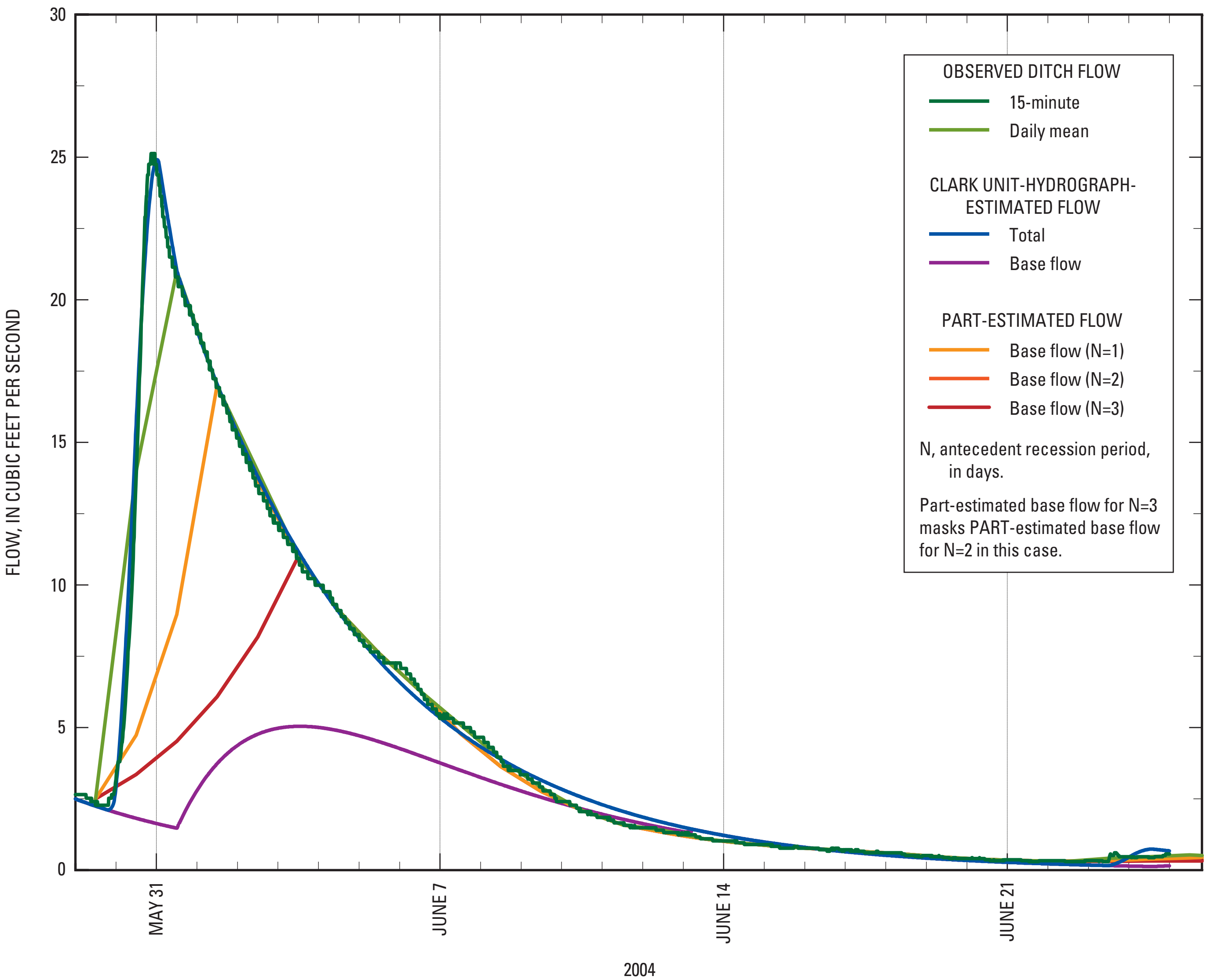


Figure 13. Base-flow separation of hydrograph from ditch gage SW4, Glacial Ridge study area, northwestern Minnesota, May 29–June 26, 2004.

values were estimated from the observed storm hydrographs and automatically optimized to minimize differences between simulated and observed hydrographs. After the variables were optimized, the values were used to produce a simulated storm hydrograph. Variables were further adjusted manually to produce a more realistic match between simulated and observed storm hydrographs.

Within each basin, the time of concentration varied more than did the Clark storage coefficient (table 11). Only SW2 and SW5 had time-of-concentration ranges that were within a factor of 2. For the Clark storage coefficient, SW1, SW2, SW4, and SW5 had ranges that were within 35 percent and SW3 and SW6 had ranges that were within a factor of 2. These relatively large time-of-concentration ranges could imply that the characteristics of the storm and antecedent conditions had an effect on the value of this variable. For example, the time of concentration for a storm in SW1 would depend on the status of the Burnham Creek Impoundment; if the impoundment were full, then the time of concentration would be shorter than if the impoundment were empty.

Medians for time of concentration and Clark storage coefficient were determined for each basin and can be used to compare basins. Within the Glacial Ridge study area, the median time of concentration for the observed storms ranged from 3 hours in SW3 to 29 hours in SW1. Short times of concentration indicate that a basin was more efficient at producing and removing direct runoff. SW3 had a relatively short and straight channel length in the narrow, lower part of the basin, which allows for faster runoff and a shorter time of concentration. The time of concentration in SW1 is longer than for other basins because the impoundment upstream from the gage will fill for some time before its peak discharge occurs. Basins with higher storage capacity or a greater drainage area typically have longer times of concentration. Within the study area, the Clark storage coefficient median ranged from a low of 28 hours in SW5 to a high of 89 hours in SW1 (table 11).

Table 9. Net surface-water mass-balance, Glacial Ridge study area, northwestern Minnesota, water years 2003–5.

[GW, ground water; SW, surface water]

Ditch basin	Total precipitation	- Areal GW recharge[a]	= Available precipitation	+ GW discharge to ditches	- Basin outflow	= Unmeasured losses[b]
Volume of water per year, in acre-feet						
Water year 2003						
SW1	9,369	3,296	6,072	554	1,240	5,387
SW2	7,493	3,385	4,108	736	1,680	3,164
SW3	8,456	3,827	4,629	569	821	4,377
SW4	7,593	4,241	3,352	1,210	1,570	2,992
SW5	9,009	2,921	6,088	448	869	5,667
SW6	11,264	3,886	7,378	1,182	1,570	6,990
Total	53,184	21,556	31,627	4,699	7,750	28,577
Water year 2004						
SW1	13,924	4,564	9,360	951	1,630	8,681
SW2	12,248	4,674	7,574	549	1,710	6,413
SW3	12,826	3,540	9,286	335	991	8,630
SW4	11,287	3,220	8,067	894	1,330	7,631
SW5	15,119	5,286	9,833	588	2,430	7,991
SW6	16,336	6,571	9,765	1,462	2,770	8,457
Total	81,740	27,855	53,885	4,779	10,861	47,803
Water year 2005						
SW1	15,701	5,000	10,701	3,805	4,680	9,826
SW2	14,721	5,036	9,685	1,648	3,270	8,063
SW3	13,244	4,768	8,476	1,501	2,080	7,897
SW4	10,952	4,940	6,012	2,650	3,470	5,192
SW5	15,623	4,811	10,812	1,321	3,490	8,643
SW6	17,160	7,895	9,265	2,394	4,750	6,909
Total	87,401	32,450	54,951	13,319	21,740	46,530
Percentage of total SW volume[c]						
2003 total	146	59	87	13	21	79
2004 total	139	47	92	8	19	81
2005 total	128	48	80	20	32	68
Basin area yield, in inches						
Water year 2003						
SW1	14.87	5.23	9.64	.88	1.97	8.55
SW2	14.87	6.72	8.15	1.46	3.33	6.28
SW3	14.87	6.73	8.14	1.00	1.44	7.70
SW4	14.87	8.31	6.56	2.37	3.07	5.86
SW5	14.87	4.82	10.05	.74	1.43	9.36
SW6	14.87	5.13	9.74	1.56	2.07	9.23
Water year 2004						
SW1	22.10	7.24	14.86	1.51	2.59	13.78
SW2	24.31	9.28	15.03	1.09	3.39	12.73
SW3	22.56	6.23	16.33	.59	1.74	15.18
SW4	22.10	6.31	15.79	1.75	2.60	14.94
SW5	24.96	8.73	16.23	.97	4.01	13.19
SW6	21.57	8.67	12.90	1.93	3.66	11.17
Water year 2005						
SW1	24.92	7.94	16.98	6.04	7.43	15.59
SW2	29.21	9.99	19.22	3.27	6.49	16.00
SW3	23.29	8.38	14.91	2.64	3.66	13.89
SW4	21.45	9.67	11.78	5.19	6.80	10.17
SW5	25.79	7.94	17.85	2.18	5.76	14.27
SW6	22.65	10.42	12.23	3.16	6.27	9.12

[a]Excluding discharge from buried aquifers.

[b]Runoff to closed basins, aquifer recharge from ditches, ditch evapotranspiration.

[c]100% = water available to SW = total precip. - GW areal recharge + GW discharge to ditches.

Table 10. Percentage of each basin receiving precipitation from each precipitation gage used in the Clark unit-hydrograph model, Glacial Ridge study area, northwestern Minnesota, June 2003 and May 2004.

[—, no record; *, no data available]

		Precipitation gage				
		SCAN	G01	G15	E01	E03
June 9–10, 2003	inches	1.66	1.13	2.15	—	1.63
May 11–12, 2004	inches	2.92	—	3.08	3.17	3.31
May 29–30, 2004	inches	1.72	1.86	2.05	2.27	1.79
SW1 storm 1	Percentage of SW basin	24	62	14	*	0
SW1 storm 2	Percentage of SW basin	30	*	10	60	0
SW1 storm 3	Percentage of SW basin	25	34	10	31	0
SW2[a]	Percentage of SW basin	0	0	100	0	0
SW3[a]	Percentage of SW basin	35	0	52	0	13
SW4[a]	Percentage of SW basin	25	0	0	0	75
SW5[a]	Percentage of SW basin	4	0	0	0	96
SW6 storms 1 and 3	Percentage of SW basin	71	6	0	0	23
SW6 storm 2	Percentage of SW basin	77	*	0	0	23

[a] Percentages the same for all storms.

Table 11. Storm-runoff hydrograph model-variable values for final simulated storm hydrographs, Glacial Ridge study area, northwestern Minnesota, June 2003 and May 2004.

[in, inches; ft^3/s, cubic feet per second]

Storm number	Initial loss (in)	Constant loss rate (in/hour)	Time of concentration (hours)	Clark storage coefficient (hours)	Initial base flow (ft^3/s)	Recession constant (unitless)	Recession threshold (ft^3/s)	Missing precipitation[a]
				SW1				
1	0.83	0.29	5	87	1	0.1	2	E01
2	.1	.27	31	104	.75	.89	5	G01
3	.04	.14	29	89	4	1	5	
				SW2				
1	.2	.54	10	50	4	.8	5	
2	.2	.28	14	51	1	1	3	
3	.24	.26	17	55	4	.91	4	
				SW3				
1	1.31	.25	2	50	2	1	5	
2	1.87	.24	3	55	1	.8	3	
3	.72	.1	17	94	1	.82	1.2	
				SW4				
1	.74	.47	8	47	3.5	.73	18	
2	.1	.29	5	52	1	.86	19	
3	.07	.19	27	62	2.5	.81	21	
				SW5				
1	1.19	.26	24	28	1	.04	11	
2	.1	.44	12	24	1	.61	22	
3	.1	.16	20	32	1	.63	47	
				SW6				
1	.85	.52	4	35	3	.76	14	
2	.2	.44	7	40	1	.89	10	G01
3	.02	.11	21	69	3	.95	.7	

[a] Precipitation missing from these gages during these storms.

Basin SW5 also has the lowest percentage of wetland and lake area (14), which lowers the coefficient. The high coefficient in basin SW1 is due, at least in part, to the impoundment.

Water Quality

The quality of surficial ground-water (water from surficial aquifers) and surface-water samples collected in the study area was generally suitable for most uses but was variable. All water samples were classified as hard. Water-quality data from this study are available at *http://nwis.waterdata.usgs.gov/mn/nwis/qwdata* using the USGS site numbers given in appendix 2. In general, no quality-control problems appear to affect the water-quality results of this study. The quality-control data collected during ground-water sampling demonstrate that values measured are reliable and meaningful. Laboratory alkalinity and the bicarbonate concentration calculated from the data are unreliable, however. Ion balances indicated that all major ions in samples were included for analysis. Sum-of-solids divided by specific conductance values were higher than usually reported. Ground-water blank-sample data show that the decontamination procedures employed between sample collections was sufficient to prevent cross-contamination that would affect water-quality data analysis. Reported values are meaningful as low as the long-term method detection level (hereinafter, detection limit) (Childress and others, 1999) with the exception of calcium, which appears to have a positive bias of less than 0.1 mg/L. In no case was a concentration detected in a blank sample near ambient concentrations. Ground-water duplicate sample data indicate that variability is low (less than 1.6 percent absolute relative differences (absolute RPD; see appendix 4 for definition)) except for one sulfate concentration (9.09 percent absolute RPD) and most alkalinity concentrations. Confidence (at the 99th percentile) in values is 6 percent for field measurements and major ions concentrations and 20 percent for nutrients. Herbicide concentrations varied by 0.04 µg/L or less.

Ground Water

Ground-water quality was characterized by 48 synoptic samples collected during May 18–July 22, 2004. Of these, 39 samples were from wells completed in surficial aquifers and 9 were from wells completed in buried aquifers. The average depth of the screened-interval midpoint below the water table for samples from surficial aquifers was 5.89 ft. Assuming a porosity of 0.25 and using the average recharge rate of 16.37 in/yr (table 3), the surficial-aquifer water sampled averaged about 1.1 years old based on Darcy's Law. This age may be slightly too old because water age integrated over a well screen is skewed slightly younger than the water at the screen midpoint. Thus, the resulting characterization of surficial ground-water quality represents water near the water table that was affected by land use during 2001–3.

Major Ions

Most ground-water samples in the study area were dominated by calcium, magnesium, and bicarbonate ions (fig. 14). Sum-of-solids concentrations computed from all ionic concentrations averaged 536 mg/L for surficial ground-water samples and 610 mg/L for buried ground-water samples. The corresponding standard deviations for sum-of-solids concentrations were 160 mg/L and 95 mg/L. The concentration of the six major ions in surficial ground-water samples varied by about 20 percent (in milliequivalents per liter), except for the anionic composition of five samples. These samples contained relatively more chloride or sulfate than other surficial ground-water samples. Two samples (from wells G18 and G27) with relatively high chloride concentrations also had the highest sodium percentages. These wells were the only ones sampled that were near large highways (fig. 3), so the relatively high sodium and chloride concentrations probably resulted from recharge of water containing salt used on the highways for deicing.

Samples from three wells (G26, G17, and G15) contained relatively high concentrations of sulfate. The source for this sulfate is unknown. Samples from buried aquifers contained neither high sulfate nor high chloride concentrations. Agricultural soil amendments could account for the excess sulfate, but the amount and kind of agriculture around these wells is not different from that around many wells with samples with relatively low-sulfate-concentration. However, a natural, spatially variable source for sulfate probably exists in the study area because samples from Polk County Ditch 140 at gage SW1 also contained relatively high sulfate concentrations at times. One such source may be recharge from wetlands, where decomposing organic matter can release sulfate. This could not be confirmed with the few wetlands samples analyzed for this study, however, because they did not contain water with high sulfate concentrations.

The major-ion composition for buried ground-water samples (samples from buried aquifers) was similar to that for most surficial ground-water samples except that sodium concentration varied widely, between 3 and 78 mg/L. The highest sodium concentration in surficial ground-water samples not affected by road salt was 23 mg/L. Sodium concentrations in buried ground-water samples do not correlate with concentrations of any major anion, including chloride (fig. 14).

Nutrients

Nutrient concentrations in surficial ground-water samples were spatially variable (fig. 15), reflecting the spatial variability of land use in the study area preceding the sampling period (May–July 2004). Surficial ground-water samples contained nitrogen nutrients, particularly the oxidized forms such as nitrate, at higher concentrations than phosphorus nutrients. These relations reflect the agricultural land use in the study area and the higher oxygen concentrations (median, 5.9 mg/L) in the surficial ground water. Nearly all surficial ground-water samples (36 of 38) contained nitrate at concentrations higher

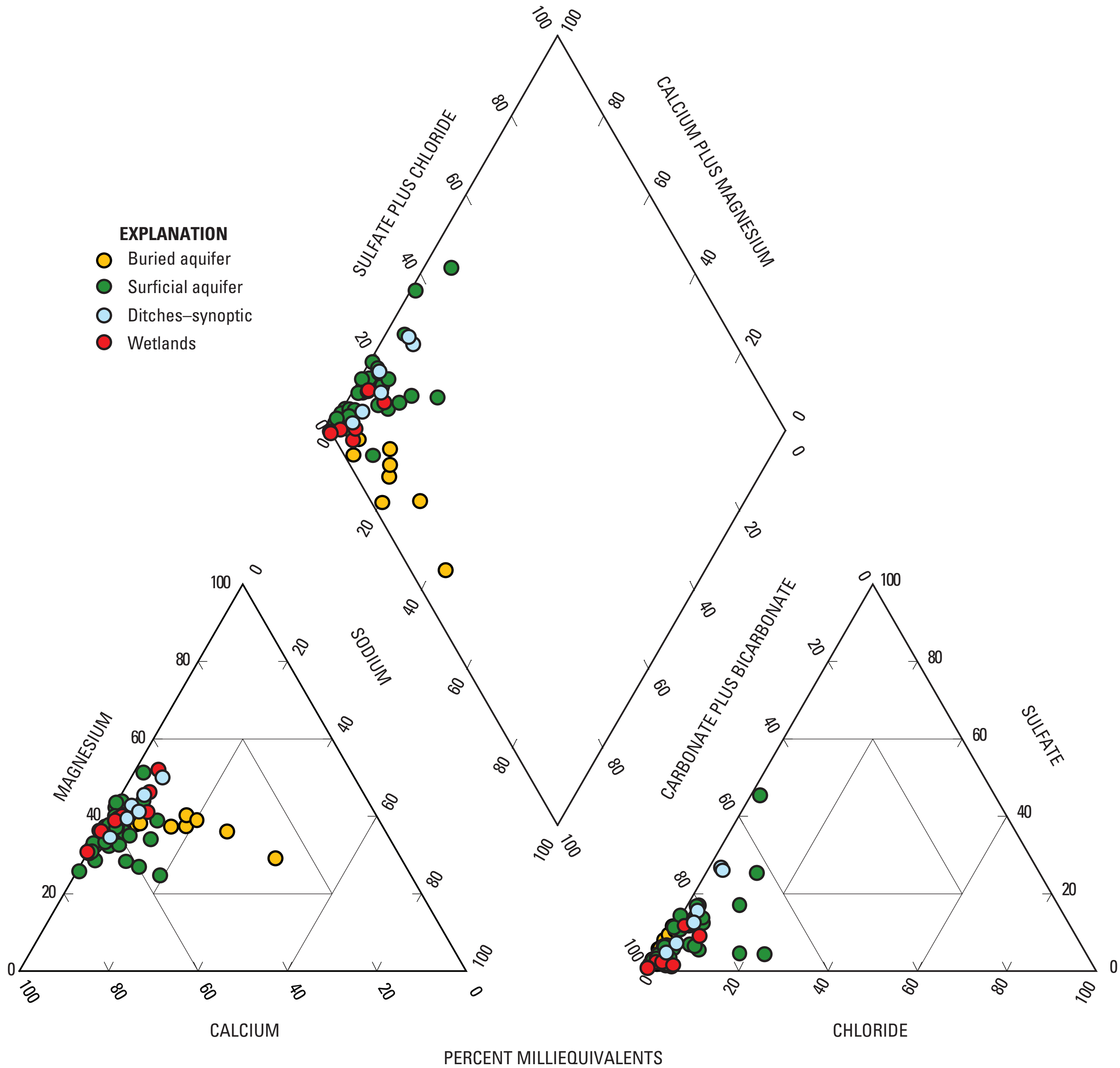


Figure 14. Water ionic composition, Glacial Ridge study area, northwestern Minnesota, May–July 2004.

than the detection limit of 0.03 mg/L as nitrogen (mg/L-N). Nearly one-half of the samples (47 percent) contained nitrate at concentrations higher than 3 mg/L-N, which Madison and Brunett (1984) concluded was the concentration above which some anthropogenic nitrate is present in natural waters. About one-quarter of the samples (26 percent) contained substantial amounts of anthropogenic nitrate (more than 10 mg/L-N; Madison and Brunett, 1984). The highest nitrate concentration (133 mg/L-N) came from water (well G22) beneath a field subject to intensive agriculture (irrigated corn and sunflowers). The U.S. Environmental Protection Agency (USEPA)(2006) has established three drinking-water standards for nutrients sampled in this study. These are a Lifetime Health Advisory of 30 mg/L-N for ammonia and Maximum Contaminant Levels (MCLs) for nitrate (10 mg/L-N) and nitrite (1 mg/L-N). Surficial ground-water samples exceeded only the nitrate MCL, with 10 samples (26 percent) higher than the limit. The highest concentrations were 2, 4 and 13 times the MCL.

Samples from 7 of 24 wells showed evidence for denitrification of nitrate or nitrite, in the form of excess nitrogen gas (see table 5.) This excess ranged from 1.5 to 9.9 mg/L, representing the equivalent additional concentration of nitrate or nitrite (as nitrogen) that was once in the samples. Samples with excess nitrogen gas were relatively low in dissolved oxygen (a requirement for anaerobic denitrification).

Concentrations of nitrogen species in buried ground-water samples were not as spatially variable as those in surficial samples, suggesting that the recharge area for buried aquifers has more homogeneous land use or that the longer flow paths and residence time of buried ground water permits hydrodynamic and geochemical processes to homogenize nutrient concentrations. Although buried ground-water samples contained some oxygen, the median oxygen concentration was low (2.0 mg/L, 19 percent saturation), and all detected nitrogen nutrients were in the reduced form of ammonia (median, 1.1 mg/L as N). Phosphorus concentrations were an order of magnitude higher in buried ground-water samples than in surficial ground-water samples and are comparable to concentrations found in surface waters. No buried ground-water samples exceeded USEPA drinking-water nutrient advisories or limits.

Herbicides and Their Degradates

The herbicides commonly used on corn and soybeans and the compounds into which these parent herbicides first degrade (degradates) were analyzed in water samples because these crops were commonly grown in the study area. These compounds fall into three broad chemical groups: triazines, acetamides, and others (fig. 16). A degradate may have been produced from one or more parent herbicides in the same chemical group. Thirty-nine surficial ground-water samples and nine buried ground-water samples were analyzed for herbicides and their degradates. Surficial ground-water samples contained detectable concentrations of atrazine, metolachlor, pendimethalin, prometon, and terbutryn (5 of 16 parent herbicides analyzed for) and 10 of the 19 degradates analyzed. In general, degradates were found more frequently and at higher concentrations than were the parent herbicides. The most commonly detected compound was 2-chloro-4-isopropylamino-6-amino-s-triazine (deisopropylatrazine), an atrazine degradate, which was detected in 28 percent of the surficial ground-water samples at concentrations as high as 0.46 µg/L. The compound measured at the highest concentration was 2-[(2-ethyl-6-methylphenyl)amino]-2-oxoethanesulfonic acid, an acetamide degradate, which was measured in one sample (well G22) at 39 µg/L. The sample from well G22 contained the highest number of quantified compounds (9 of the 35).

In all, 13 surficial ground-water samples (one-third) had a total of 46 quantified herbicide or degradate concentrations. The median concentration of quantified concentrations was 0.12 µg/L, and the interquartile range was 0.07–0.23 µg/L. The summation of the concentrations of all quantified herbicides and degradates is called the sum-of-herbicides (SH) concentration and is a gage of the degree to which herbicides have affected a water sample. Any sample with no detections has a SH concentration that is less than the sum of the detection limits of each analyte (less than 0.78 µg/L). All surficial ground-water samples had a SH concentration less than 1 µg/L except for the sample from well G22, where it was 82.35 µg/L. The median of quantified (concentrations above the detection limit) SH concentration was 0.30 µg/L and the interquartile range was 0.22–0.70 µg/L. As with nutrient concentrations, herbicide and degradate concentrations in surficial ground water from well G22 probably reflect the intensive agriculture in the area.

No herbicide or degradate was detected in buried ground-water samples, reflecting the protection that clay-rich confining units afford these aquifers. Not only do the confining

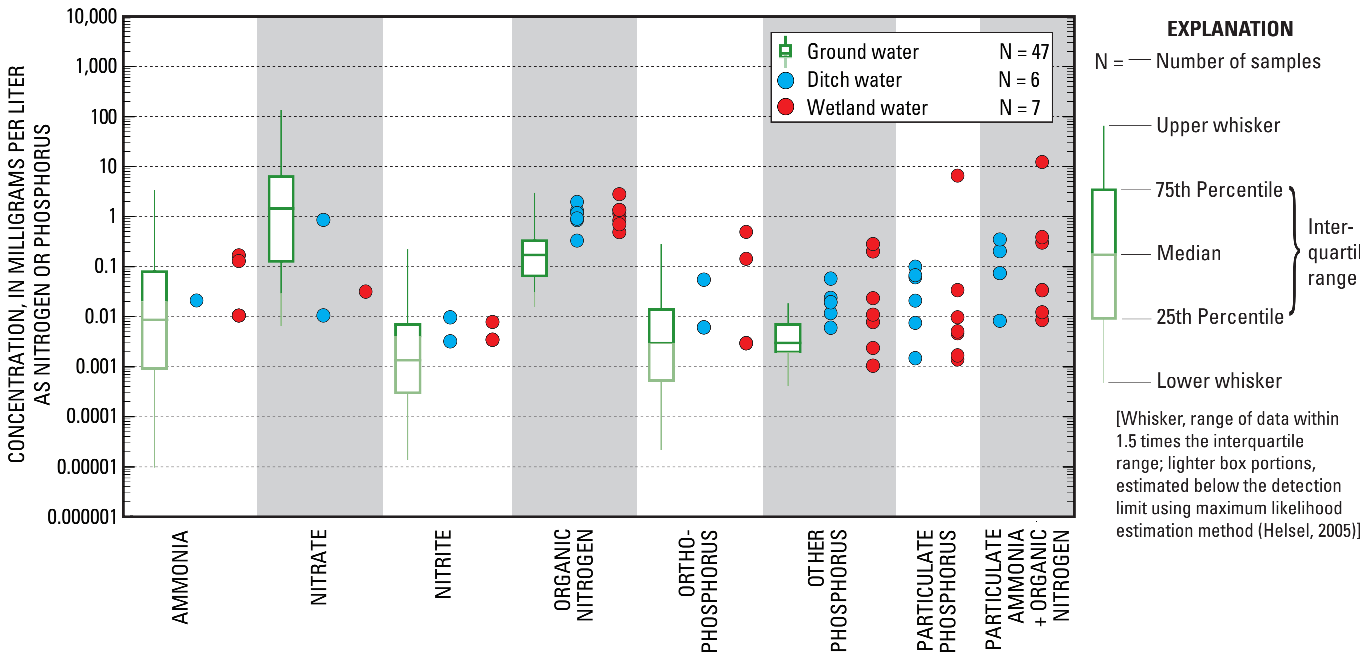


Figure 15. Nutrient concentrations in water, Glacial Ridge study area, northwestern Minnesota, May–July 2004.

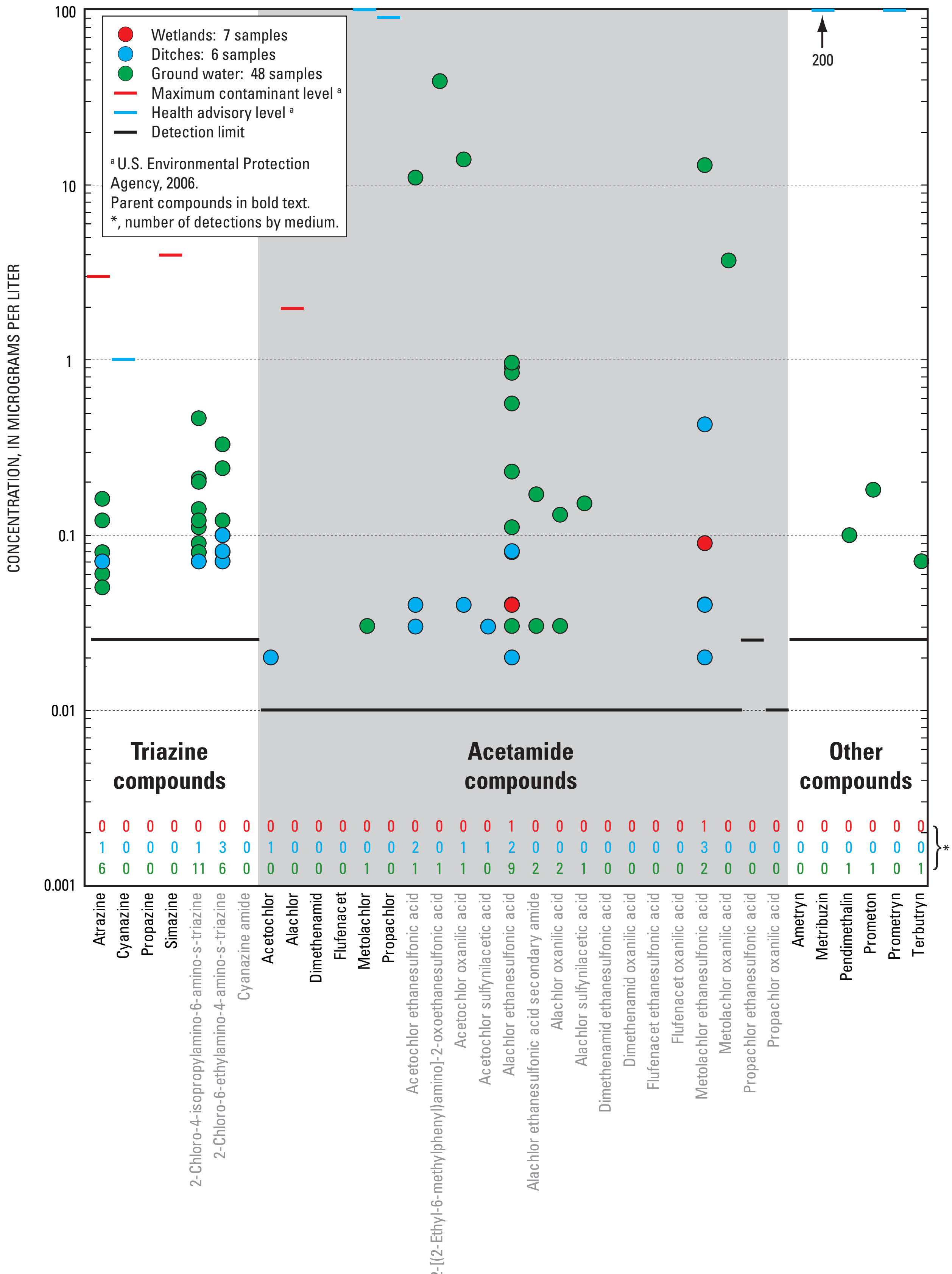


Figure 16. Herbicide and degradate concentrations in water, Glacial Ridge study area, northwestern Minnesota, May–July 2004.

units slow water movement from above, but the clay minerals can adsorb organic chemicals contained in water as it moves through the confining unit.

Variability

Short-term temporal and spatial water-quality variability was quantified using 14 samples collected at each of 10 surficial-aquifer and 2 buried-aquifer synoptic sampling network wells during May 2003–September 2005. These samples were analyzed for field measurements (alkalinity, and dissolved-oxygen concentration, pH, specific conductance, water level, and water temperature) and nutrient concentrations. Variability among the samples from each well was measured by calculating a coefficient of variation (CV, standard deviation divided by mean) for each measurement or concentration. Coefficients of variation were calculated only at wells where more than 15 percent of the samples were above the detection limit; regression-on-order statistics (Helsel, 2005) were used to estimate mean and standard deviation.

Samples from well E01S were more variable than samples from all other wells for most constituents. The concentration of most constituents at this well underwent a gradual transition in water quality that began in fall 2003 and ended between early summer 2004 and spring 2005, depending on the chemical measured. For example, figure 17 shows that the nitrate and ammonia concentrations in samples from well E01S had a wider range than other samples with comparable concentrations. Well E01S, which is completed in a surficial aquifer at 14.5–19.5 ft below land surface (BLS), is the shallow well in a nest with E01D, which is completed in a buried aquifer at 168–171 ft BLS. Well E01D is a flowing well whose hydraulic head is 6–8 ft above land surface. When this study began, well E01D had been allowed to flow constantly (at 0.13 gal/min in July 2002), presumably to prevent the steel well casing from cracking during the winter. Consequently, water flowing from well E01D infiltrated into the surficial aquifer and produced a zone of mixed buried ground water and surficial ground water. In late June 2003, well E01D was capped to allow the hydraulic head to be measured, thereby preventing water from the buried aquifer from further influencing either the water-table elevation or the water quality of the surficial aquifer. In all cases where a gradual water-quality transition exists in the samples from well E01S, the direction of change is from water quality more like that of the buried ground water from well E01D to that of other surficial ground waters. Therefore, the gradual transitions seen in water-quality of samples from well E01S are the result of mixing of buried ground water recharging the surficial aquifer from well E01D and of surficial ground water near the E01 nest. In the following discussion of ground-water quality variability, any mention of water quality of samples from well E01S will refer only to those collected after the gradual transition (for example, after June 2004 for nitrate).

No pattern in the spatial variability of measurements and concentrations was apparent among wells; that is, no well had consistently high or low variability among its measured values or concentrations. The only measurement or concentration whose variability could be easily explained by conditions near the well was water level. Water-level variability was highest at wells G01 and G15 (CV, 50 and 29 percent respectively, relative to the average CV of 14 percent) where the depth to water is shallowest. This relation is reasonable because shallow depth to water increases the response of the water table to both recharge and evapotranspiration.

Dissolved oxygen (DO) concentration was the most variable measurement or concentration (average CV, 85 percent) and pH was the least variable (average CV, 6 percent). With the exception of DO concentration, the average CVs of all of the field measurements were lower than the average CVs of nutrient concentrations. Nutrient concentrations, which may include inputs from the atmosphere and agriculture, should be more variable than are the field measurements, which are more reflective of the natural water itself. Also, many nutrient concentrations in samples from many wells were near the detection limit. The CV of any concentration near zero can appear exceptionally high because the statistic becomes high as the mean becomes low. This is an artifact of the statistic, not necessarily the nature of nutrient variability.

The temporal variability of nitrate and ammonia for each well for which CV could be calculated is shown in figure 17. Here, differences in concentration and differences in variability among wells can be seen. Although some wells like G22 and G08 had intensive agriculture near them and had a correspondingly high median nitrate concentration, no such land-use pattern explains the differences in nitrate variability among the wells. Contrastingly, high ammonia concentrations correspond with low variability and with buried ground water (E01D, E02D). Buried ground water has much longer flow paths and residence times than does surficial ground water. These characteristics allow dissolved oxygen to be consumed, producing reducing conditions that keep ammonia from oxidizing. The source of the ammonia in buried ground water is unknown, but its low variability suggests that the source may be widespread and constant, perhaps from air fall or from nutrients acquired during recharge, probably in the higher elevation forests to the southeast of the study area.

Surface Water

All ditch-water samples collected for this study were aggregated and analyzed together to form a single assessment of ditch-water quality and variability. The 146 ditch-water samples were collected at seven gages from October 16, 2002, through October 4, 2005, and were analyzed for field measurements, nutrient concentrations, and suspended-sediment concentration. The number of samples collected at a gage ranged from 19 (SW5) to 27 (SW2 and SW6) (table 12). Sampling at gage SW8 did not begin until October 2004, so data from that site are not included in the following analysis. Major-ion samples were collected at ditch gages until November 2004. One herbicide sample was collected at each of the six gages during mid-July 2004.

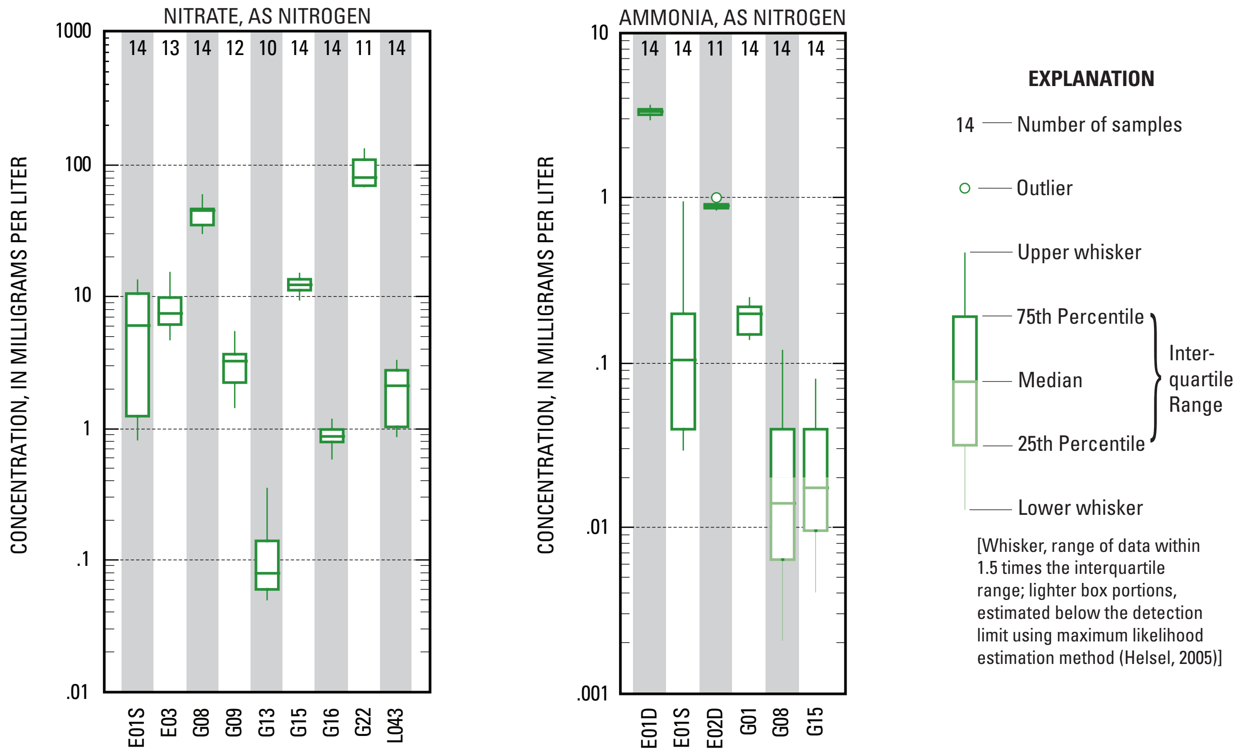


Figure 17. Ground-water nitrate- and ammonia-concentration variability, Glacial Ridge study area, northwestern Minnesota, 2003–5.

Physical Properties and Major Ions

As with surficial ground-water samples in the study area, ditch-water samples were dominated by calcium, magnesium, and bicarbonate ions (fig. 14). The mean pH of ditch-water samples ranged from 7.7 at gage SW2 to 8.1 at gage SW6 (table 12). The mean standard deviation was 0.29. The pH of ditch-water samples was generally higher than the pH of ground-water samples.

Specific conductance (SC) is a convenient measure of the ionic strength of water. In general, waters with higher specific conductance have higher concentrations of ions. The mean SC of ditch-water samples ranged from 505 µS/cm at gage SW2 to 631 µS/cm at gage SW1 (table 12). Variability of the SC, indicated by the standard deviation in the samples, ranged from 81 µS/cm at gage SW6 to 260 µS/cm at gage SW1. The relatively low variability in samples from gage SW6 can be explained by the discharge it receives from lakes in an active gravel pit in the basin. The high variability of SC in samples from gage SW1 can be attributed to the variability of sulfate concentrations in that ditch. Specific conductance in ground-water samples was typically higher than in ditch-water samples, and the lowest flows were dominated by ground-water discharge. The highest SC occurred at the lowest ditch flows. However, there was a very poor relation between flow and SC in samples from all gages because the difference in specific conductance between ground- and ditch-water samples is small.

The range of mean bicarbonate (as $CaCO_3$) in ditch-water samples was from 236 mg/L at gage SW6 to 308 mg/L at gage SW4, with a mean standard deviation of 92.0 mg/L (table 12). The alkalinity of ditch-water samples was generally lower than the alkalinity of ground-water samples.

Most major ions varied within a relatively small range and differed little from gage to gage (table 12). For example, the means of calcium concentration in samples from the ditch gages ranged from 60.2 mg/L at gage SW1 to 72.6 at gage SW4, which is small in comparison to the mean standard deviation of the concentrations (21.2 mg/L) for all samples. Chloride and sulfate concentrations varied more than other major ions. The range in mean chloride concentrations in samples was from 5.3 mg/L at gage SW2 to 14.2 at gage SW5, and the mean standard deviation was 4.0 mg/L. The range of mean sulfate concentrations in samples was much larger, from 14.0 mg/L at gage SW2 to 109 mg/L at gage SW1, and the mean standard deviation was 27 mg/L.

The most obvious anomaly in the data in table 12 is the elevated sulfate concentration in samples from gage SW1. There is no identifiable source for the sulfate in that ditch basin. Concentrations in surficial and buried ground waters are not high enough to produce concentrations as high as those observed in samples from gage SW1. It is possible that the source could be biochemical releases in wetlands. The sulfate concentrations are most highly correlated with sodium (Spearman's rank correlation (Helsel and Hirsch, 2002), $\rho = 0.95$, two-sided p-value = 0.0001) and magnesium (Spearman's rank correlation, $\rho = 0.86$, two-sided p-value = 0.0006) concentrations.

Table 12. Summary statistics of field measurements and major-ion concentrations in ditch-water samples, Glacial Ridge study area, northwestern Minnesota, 2003–5.

[All concentrations in milligrams per liter]

	SW1	SW2	SW3	SW4	SW5	SW6
pH						
Number of samples	22	27	25	26	19	27
Mean	8.0	7.7	7.9	8.0	7.8	8.1
Standard deviation	.3	.4	.3	.2	.2	.3
Coefficient of variation	.041	.047	.041	.030	.026	.034
Specific conductance in microsiemens per liter						
Number of samples	22	27	25	26	19	27
Mean	630.9	504.6	541.6	554.3	615.2	526.2
Standard deviation	260.4	125.0	160.4	131.0	202.1	81.2
Coefficient of variation	.413	.248	.296	.236	.329	.154
Calcium						
Number of samples	17	22	20	21	16	23
Mean	60.2	64.5	70.7	72.6	68.9	61.9
Standard deviation	22.9	18.9	25.8	21.3	26.3	12.0
Coefficient of variation	.380	.293	.364	.294	.381	.195
Magnesium						
Number of samples	17	22	20	21	16	23
Mean	38.9	27.1	25.3	30.3	33.2	28.1
Standard deviation	18.3	7.6	8.0	7.2	11.0	4.7
Coefficient of variation	.470	.282	.315	.236	.332	.168
Potassium						
Number of samples	17	22	20	21	16	23
Mean	7.4	3.7	3.7	4.2	7.5	4.8
Standard deviation	7.4	1.7	2.8	2.1	11.7	3.0
Coefficient of variation	.991	.447	.758	.498	1.559	.623
Sodium						
Number of samples	17	22	20	21	16	23
Mean	16.4	7.3	7.1	7.3	11.8	8.9
Standard deviation	14.1	3.3	2.9	1.7	3.9	2.7
Coefficient of variation	.863	.449	.408	.231	.334	.299
Bicarbonate, as calcium carbonate						
Number of samples	20	25	23	24	17	25
Mean	264.3	304.9	270.9	308.4	304.8	235.8
Standard deviation	102.7	90.9	95.1	89.9	133.8	54.2
Coefficient of variation	.389	.298	.351	.291	.439	.230
Chloride						
Number of samples	17	22	20	21	16	23
Mean	9.0	5.3	12.6	7.7	14.2	10.7
Standard deviation	4.0	2.0	5.4	1.8	5.4	5.7
Coefficient of variation	.444	.369	.431	.230	.378	.533
Sulfate						
Number of samples	17	22	20	21	16	23
Mean	109.0	14.0	43.2	30.1	48.6	60.8
Standard deviation	76.8	7.7	24.0	13.8	22.2	18.2
Coefficient of variation	.704	.555	.555	.459	.457	.298

Temporal trends in the concentration of major-ion constituents could not be assessed because of the relatively short time frame of the study. However, samples from gage SW6 indicate that a downward trend in bicarbonate concentration is likely.

Nutrients and Herbicides

Nutrient concentrations in ditch-water samples are comparable to those found in the Snake River above Alvarado (table 13) (Tornes and others, 1997), a stream located about 45 miles northwest of the study area, near the North Dakota border. Water-quality data for the Snake River above Alvarado are available at *http://nwis.waterdata.usgs.gov/mn/nwis/qwdata/?site_no=05085900*. The Snake River above Alvarado drains land with characteristics similar to those in the study area and was considered typical of streams in the Red River Valley Lake Plain. It is included herein for comparison because of its long-term water-quality record. In general, most nutrient concentrations in the study area ditches are lower than those in the Snake River, however.

Dissolved ammonia plus organic nitrogen is one form of nitrogen found in surface waters. In well-oxygenated systems, ammonia concentrations are low and most nitrogen is organic. The ditches in the study area were generally well oxygenated in the open-water season (April through October), and ammonia concentrations were low. Ammonia concentrations were slightly higher during the winter months than during the summer months. The median concentration of dissolved ammonia plus organic nitrogen ranged from 0.49 mg/L-N in samples from gage SW6 to 1.04 mg/L-N in samples from gage SW1. By comparison, the median concentration in samples from the Snake River above Alvarado was 1.20 mg/L-N.

Dissolved nitrate plus nitrite is another important form of nitrogen in surface waters. In well-oxygenated systems, nitrite is oxidized to nitrate, and nitrite concentrations are low. Median values of nitrite plus nitrate concentrations in ditch samples ranged from less than 0.03 mg/L-N at gage SW2 and gage SW3 to 0.635 mg/L-N at gage SW6 (table 13). By comparison, the median concentration of dissolved nitrite plus nitrate concentrations in samples from the Snake River above Alvarado was 0.175 mg/L-N.

Dissolved phosphorus is composed of several forms including orthophosphorus, which is readily used by plants. Orthophosphorus was the primary form of dissolved phosphorus in the ditches in the study area. The median concentration of dissolved phosphorus in samples ranged from 0.0108 mg/L as phosphorus (-P) at gage SW6 to 0.0293 mg/L-P at gage SW5 (table 13). By comparison, the median concentration of dissolved phosphorus in samples at the Snake River above Alvarado was 0.115 mg/L-P.

The USEPA (2006) has established aquatic-life criteria for ammonia in streams. Those criteria are dependent on the pH and temperature, the species present, and the life cycle of the fish in the stream. For ditches in the study area, which do not have salmon species, the acute criterion for streams was computed. This criterion is based on the effect of short-term ammonia exposure on aquatic life. That criterion was exceeded three times for SW2, twice for SW1 and SW3, once for SW4 and SW5, and no times for SW6. Most samples that exceeded the criterion were collected in the winter or early spring months, although samples from SW1 that exceeded the criterion were collected in the late summer and early fall.

The USEPA (2006) also has established nutrient criteria, which represent attainable background concentrations of nutrients in streams. These criteria are not enforceable regulations but are provided as guidance that can be used as starting points to set goals of nutrient concentrations. For the study area, the reference concentrations for total ammonia plus organic nitrogen is 0.816 mg/L-N, for nitrite plus nitrate is 0.034 mg/L-N, and for total phosphorus is 0.0875 mg/L-P. The criterion for total ammonia plus organic nitrogen was exceeded in about 60 percent of the samples collected during 2003–5. The criterion for nitrite plus nitrate was exceeded in about 58 percent of the samples collected during 2003–5. The criterion for total phosphorus was exceeded in about 17 percent of the samples collected during 2003–5. Because more samples exceeded the nitrogen criteria than the phosphorus criterion, nitrogen in ditches is the greater concern.

Herbicides and their degradates were sampled once, during July 12–14, 2004. Most herbicides in these samples had concentrations below the detection limit (fig. 16). The sample from gage SW1 had the most concentrations above the detection limits (seven), followed by samples from gages SW4 (four), SW3 (two), and SW5 and SW6 (one each). The sample from gage SW2 had no concentrations above the detection limit. Deethylatrazine (2-chloro-6-ethylamino-4-amino-s-triazine), a triazine degradate, was detected in 3 of 6 ditch-water samples.

Implications of Wetland and Prairie Restorations

As more land in the study area is restored to original wetland and prairies, the effects on its hydrologic system will increasingly become apparent. These restorations will retain more water in the study area and likely will reduce the application of agricultural chemicals to the land surface. Based on the information and understanding obtained from this study, some hydrologic changes in the ground- and surface-water flow systems and in the quality of water in the study area are expected.

Table 13. Summary statistics of nutrient concentrations in ditch-water samples, Glacial Ridge study area, northwestern Minnesota, 2003–5.

[All concentrations in milligrams per liter; nobs, number of observations; ncens, number of censored values; %, percentile; —, not available in Tornes and others, 1997.]

	SW1	SW2	SW3	SW4	SW5	SW6	Snake River
				Dissolved ammonia, as nitrogen			
nobs	22	27	25	26	18	27	32
ncens	12	15	15	20	8	11	—
25%	<.020	<.020	<.020	<.020	<.020	<.020	.04
50%	<.020	<.020	<.020	<.020	.021	.034	.08
75%	.088	.064	.044	<.020	.038	.095	.155
				Dissolved ammonia plus organic nitrogen, as nitrogen			
nobs	22	27	25	26	19	27	32
ncens	0	4	0	0	0	0	—
25%	.907	.159	.659	.616	.718	.382	1
50%	1.036	.635	.77	.787	.842	.494	1.2
75%	1.104	1.089	.928	1.081	1.031	.857	1.3
				Dissolved nitrite, as nitrogen			
nobs	22	27	25	26	18	27	32
ncens	11	17	17	16	7	3	—
25%	<.004	<.004	<.004	<.004	<.004	.005	<.010
50%	.004	<.004	<.004	<.004	.004	.01	.02
75%	.008	.006	.006	.005	.019	.014	.055
				Dissolved nitrite plus nitrate, as nitrogen			
nobs	22	27	25	26	19	27	32
ncens	11	17	14	10	9	4	—
25%	<.03	<.03	<.03	<.03	<.03	.159	.004
50%	.0485	<.03	<.03	.052	.128	.635	.175
75%	.207	.054	.198	.21	.433	1.089	.96
				Dissolved phosphorus			
nobs	22	27	25	26	19	27	32
ncens	0	0	1	0	0	0	—
25%	.0158	.0075	.0084	.0081	.0187	.0065	.06
50%	.0246	.015	.0142	.0134	.0293	.0108	.115
75%	.0371	.0265	.0241	.0157	.054	.0208	.19

Effects on Ground-Water Flow

As more water is retained on the land in the study areas, it is hypothesized that ground-water levels will rise in response to increased water levels in surrounding wetlands, resulting in increased ground-water storage. Water-level rise likely will not be uniform across the study area, however. The 2004 water table was depressed near areas where ditches cross beach ridges. If these ditches are abandoned, being filled with sand and gravel initially excavated from the beach ridge, the water table likely will rise near these restored beach ridges. The water table also could rise in areas where restored wetlands abut surficial aquifers and recharge them. These water-table rises likely will propagate downgradient with decreasing amounts of rise. Away from these areas, however, the water-table elevation likely will remain about the same as it was before wetland and prairie restoration. The amount of local water-table rise is difficult to predict. Generally, restored wetlands are only a few feet deep and many are ephemeral, some remaining wet only a few weeks in the spring. Most wetlands are in back-beach lowlands, which can be several feet lower than the downgradient beach ridge. Therefore, it is likely that not all of the wetlands will have higher water elevations that can recharge surficial aquifers. Those that do may raise the water table only a foot or two, and that only for a few weeks or months during the year. In any case, the water-table rise likely will occur only in the immediate downgradient beach-ridge aquifer and only initially after the ditch is abandoned. Storage in these aquifers likely will increase slightly with the increased water levels, but the amount of increased storage and the duration of the water-table rise probably will be small.

Substantial changes in the amounts of water flowing in each component of the ground-water system likely will result from wetland and prairie restoration. The net effect on the ground-water mass balance is harder to predict. The initial increase in wetland recharge to surficial aquifers likely will be offset by increased discharge because of higher hydraulic gradients. However, these gradient increases likely will be less than the ground-water-level increases because downgradient wetland water levels also may increase. Thus, flux through the aquifer likely would increase, but the net amount of discharge from the aquifers may not increase. Rise in the water table implies that evapotranspiration may increase in areas where water levels rise above the extinction depth. Both current and likely future evapotranspiration volumes are not well quantified in the study area.

Minor increases in ground-water storage and discharge to ditches are expected as water levels rise due to restoration. Storage increases likely will be small because water-level increases will be local, near new wetlands, and small for the reasons outlined above. Discharge of ground water directly to ditches may increase slightly because of increased head gradients caused by higher water levels. If substantial lengths of ditches are abandoned, however, especially where they cross surficial aquifers, total discharge to ditches may decrease. After restoration, the general features of the water-table surface likely will look similar to the 2004 water table (fig. 10) except in areas where ditches now cut through beach ridges. If these ditches are abandoned, contour lines should smooth out in the area of the abandoned ditch and become parallel with the beach ridges themselves. The change probably would not be easy to discern on a map at the scale of figure 10.

Most of the effects of ground-water-level rise and changes in the mass balance likely will be geographically limited, probably to land immediately adjacent to the restored areas. This is in part because the study area is in the headwaters area of the drainage basin. At a point where a ditch flows out of TNC property, restored wetland and prairie areas will form a substantial part of the drainage basins. However, at a point even a few miles downstream from TNC property, the restored areas would be only a relatively very small part of drainage basins because in these basins, area approximately increases as the square of the ditch distance.

The likely local nature of hydrologic changes resulting from restoration activities is illustrated by two scenarios in which ground-water levels could rise. One scenario is a wet meadow, which is immediately downgradient from a surficial aquifer that is itself immediately downgradient from a newly constructed wetland. The other scenario is a fen that receives its water from a surficial aquifer that is downgradient from a newly constructed wetland. In both scenarios, increased recharge to a surficial aquifer from a newly constructed wetland likely will cause increased discharge to a downgradient surface-water body.

If the wet meadow in the first scenario is within a restoration area, the wetter meadow likely will not be a problem because it would be returning to its natural hydrologic state. On the other hand, if the wet meadow is on a neighboring property, used perhaps for hay production, wetter conditions likely will reduce the usefulness of the property. However, wetter conditions in the already wet meadow are unlikely. In most cases, wet land (such as meadows) adjacent to the study area is drained by ditches. Any additional ground-water discharge likely will be intercepted by these ditches, leaving the wet meadow about as wet as it always was, although flow in the ditch draining it may increase.

Increased ground-water discharge to fen areas (fig. 5) could have two effects. If the discharge is diffuse, the area of the fen could increase as fen-plant communities colonize the newly wetter areas. This would be a return to the presumed hydrologic state before the upgradient wetlands were drained. If discharge to the fen is concentrated in a few areas, seepage velocities of the ground-water discharge could increase to the point that fen plants could no longer live in the immediate discharge area. Should this actually occur, the area of excessive ground-water discharge likely would be small because seepage velocities should fall quickly as the spring flow travels through the wetland. Any reduced fen area resulting from excessive discharge velocity should be more than offset by increases in the fen area elsewhere. In both cases, increases in discharge to fens will be small because the increase in ground-water levels likely will be small, local, and temporally short, as noted above.

The effects of wetland and prairie restorations on water levels and fluxes in the aquifers containing the Crookston well fields likely will be negligible. Of the six Crookston wells (four in the old well field and two in the new well field) only two, both in the old well field (central part of area A, fig. 10), withdraw water from surficial aquifers which could be affected by wetland and prairie restorations. However, two structural features near the old well field location make changes unlikely. One structural feature is the location of Judicial Ditch 66 one mile east of the well field. This ditch effectively intercepts ground-water flow to the well field from the northeast (fig. 10). This ditch likely will not be abandoned because it serves as the outflow for the gravel pit lakes about 1.5 mi north-northwest of Tilden Junction. Any restoration activities east of Judicial Ditch 66 likely will not influence surficial aquifers at the old well field. The other structural feature is an active pit in the beach ridge one-half mile south of the well field (south of U.S. Highway 2). This mining is removing surficial aquifer material through which water from the south would have to move to reach the old well field. Thus, restoration activities south of the well field could not affect the old well field with this aquifer material removed. This situation leaves a very small area (perhaps 160 acres) southeast of the old well field, where wetland and prairie restorations could affect two of the Crookston municipal wells (fig. 10). Restorations in this area could affect water in the surficial aquifer in the ways discussed in the beginning of this section but the effect may not be noticeable because of the small area involved.

Effects on Surface-Water Flow

Restoration activities in the study area will principally affect two streamflow variables quantified in Clark unit-hydrograph modeling: time-of-concentration and storage coefficient. The time of concentration quantifies how slowly water flows down the channel. The storage coefficient quantifies how slowly water flows overland to a stream. Three restoration activities occurring in the study area can affect these variables. Filling ditches and restoring prairies likely will increase the time of concentration because the filled ditches are less efficient at transmitting water downstream. Restoring wetland basins may increase the storage coefficient because wetlands and prairies have greater resistance to overland flow than does cropland. The greater water retention likely will also reduce the total volume of direct runoff, but this cannot be modeled with the Clark unit-hydrograph method. Filling ditches to produce wetlands could increase both the time of concentration and the storage coefficient. This restoration activity likely will occur at most ditches because they were usually constructed through wetlands. The combination of filling ditches, restoring prairies, and restoring wetlands probably will increase both the time of concentration and the storage coefficient.

The effect of filling ditches and restoring wetlands on direct runoff can be seen in the hypothetical scenarios in figure 18. The best-fit model hydrograph in figure 18 shows the modeled direct runoff of a rainstorm at ditch gage SW4 in May 2004 assuming calibrated median values for the time of concentration (8 hours) and storage coefficient (52 hours) variables. These calibrated values provide the best fit to measured storm hydrographs in the study area and quantify the variable values before restoration (current conditions). Scenario 1 shows how filling ditches to produce prairie may have affected the hydrograph at ditch gage SW4 from the May 2004 storm. By increasing the time of concentration from 8 hours to 16 hours, the peak occurs later and is about 6 percent lower (44 rather than 47 ft^3/s). Scenario 2 shows how restoring wetlands and prairies may have affected the storm hydrograph. By increasing the storage coefficient from 52 hours to 78 hours, the timing of the peak is not much different, but the peak is reduced by 32 percent, from 47 to 32 ft^3/s. Scenario 3 shows how filling ditches to produce wetlands or how filling ditches, and restoring wetlands and prairies may have affected the storm hydrograph. By increasing the time of concentration and the storage coefficient, the peak is later and 34 percent lower than current conditions.

If the ditches are filled in and flow is blocked across beach ridges, then the structure of the basin will be drastically altered, and very different reductions in direct runoff can be expected. Many of the peaks in the remaining ditches likely would be reduced substantially because the effective drainage area for direct runoff would be much smaller. Blocking flow across beach ridges also likely would induce more surface-water flow parallel to the beach ridges and greater storage in the back-beach basin meadows. This would extend the direct runoff recession in the receiving waters and decrease total flow because more water would be retained in the back-beach basins and lost to greater evapotranspiration. The effects shown in the example for SW4 (fig. 18) likely would be similar to those in the other basins in the study area and would be applicable near each gage.

The effect of the restorations should be most evident within the restored area where water will be retained longer. Farther downstream, the effects of the reduced peak flow and increased recession should be insubstantial because ditches on farmland near and adjacent to the restored area will continue to drain excess water. For example, the effects on peak flows in Badger Creek, into which Judicial Ditch number 64 flows (fig. 3), likely would be very small because of the large drainage area of Badger Creek that is outside the restoration area.

Restoring wetlands in the back-beach basins likely will reduce total flow to ditches near the study area and change the water balance in the surface-water system because more water likely will be retained in wetlands and lost through evapotranspiration. Some retained wetland water may recharge beach ridge aquifers when the elevation of the water surface is high enough to come into contact with the downgradient beach ridge.

A major concern throughout the Red River Basin is a flood resulting from snowmelt or high precipitation. Peak flow from direct runoff from such an event should be reduced in the study area because of the expected increase in storage (example of ditch gage SW4, fig. 18). The recession from such an event likely would be longer after restoration also because of the increased storage. This longer recession likely would be confined to ditches adjacent to or near the study area, however. Higher water elevation in back-beach basin wetlands could recharge ground water in the beach-ridge aquifers, which could result in increased flow in the beach-seep fens in and near to the study area. Ditches near these fens likely will drain the excess water away, reducing or eliminating the effects of increased spring flow away from the fen.

Effects on Water Quality

The restoration of wetlands and prairies in the study area likely will result in improved water quality. The most obvious and largest factor that could improve water quality is the planned reduction in the application of nutrients and pesticides for agriculture in the study area. Currently (2005), nutrients and pesticides enter the hydrologic system by being applied to agricultural fields and falling as drift from the atmosphere. Small amounts of pesticides also are used in restoration and management of the wetland and prairie restorations. The small increase in pesticide load from restoration activities probably will be more than offset by the large decrease in nutrient and pesticide load resulting from eliminated agricultural activities. The result should be a net decrease in nutrient and pesticide load to the study area.

Further reduction in the nutrient and pesticide load to the hydrologic system likely will result from the increase in

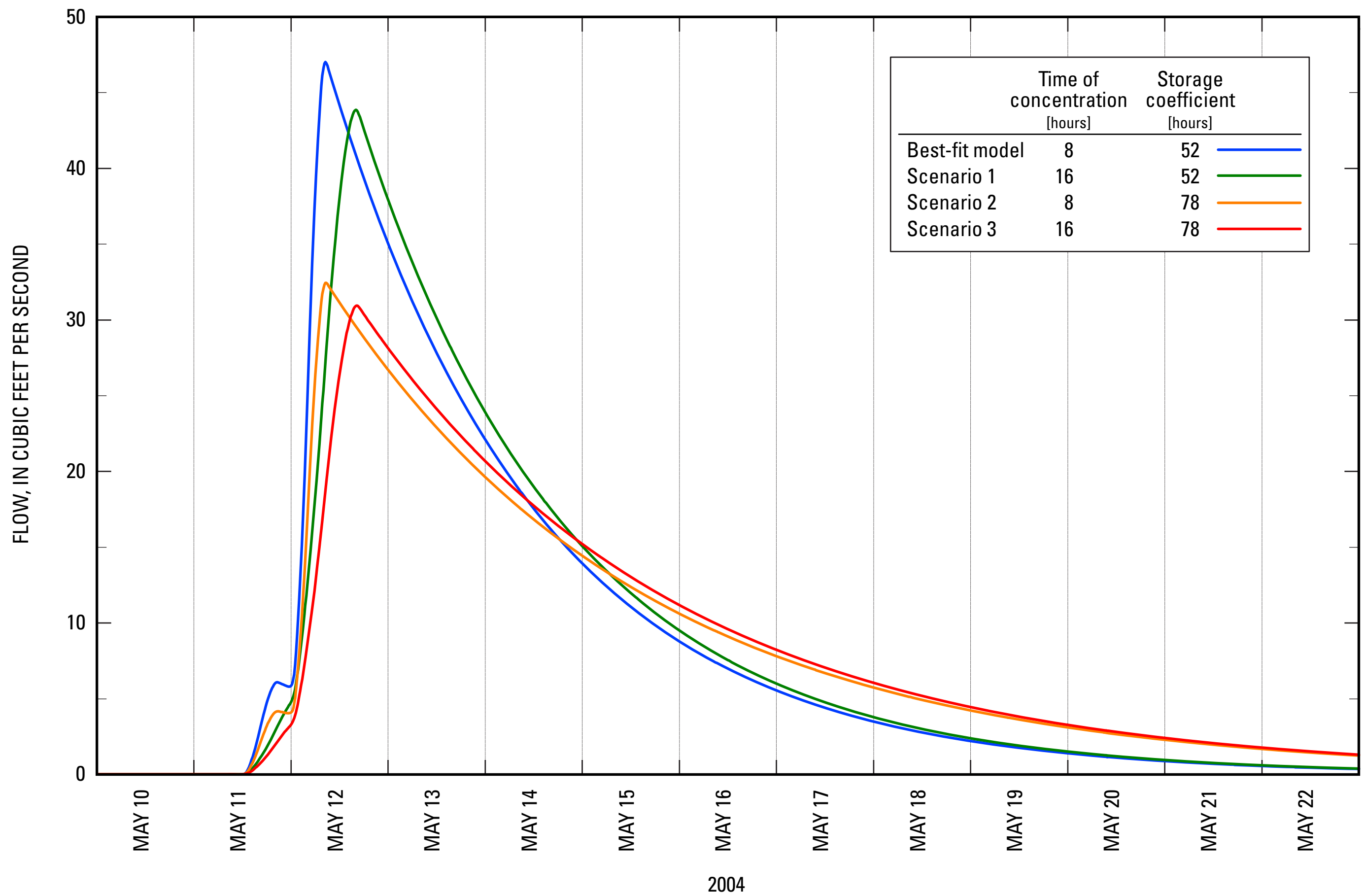


Figure 18. Modeled and hypothetical direct runoff using the Clark unit-hydrograph method at ditch gage SW4, Glacial Ridge study area, northwestern Minnesota, May 2004.

wetland area. Wetland plants and the organic-rich soils they produce can reduce chemical loads by sorption, degradation, and sequestration (Pierzynski and others, 1993; Mitsch and Gosselink, 2000). The organic-rich, fine-grained soils that collect and form in wetland basins also can adsorb nutrients and organic chemicals, removing them from the hydrologic system. As wetland plants grow, they use nutrients that would otherwise leach to ground water or runoff to ditches. The anaerobic water conditions commonly found in wetlands also are capable of supporting bacterial communities that can reduce nitrate and nitrite to forms of nitrogen gas and can degrade pesticides. As wetland sediments are buried beneath newly deposited sediments, these chemicals likely will become sequestered.

Surface runoff likely will be reduced because restoration will increase prairies with permanent ground cover and more water likely will be retained in wetland basins. The water that does run off or infiltrate likely will be improved in quality, containing smaller concentrations of nutrients, pesticides, and sediment. Gradually, if ground-water quality improves as expected, ditch-water quality likely will improve further because ground-water discharge forms ditch base flow. Eventually, the concentrations of nutrients in ditches could approach the reference concentrations stated in the USEPA nutrient criteria (attainable background concentrations).

The increase in wetland area resulting from restorations in the study area could increase methylmercury concentrations in surface water. High methylmercury concentrations are a concern in aquatic ecosystems because of the potentially toxic effects to fish, birds, and other higher organisms that eat food growing in waters with high-methylmercury-concentration (Wiener and others, 2002). Researchers have shown that flooding of terrestrial environments results in substantial increases in methylmercury production and methylmercury concentrations in both the physical (water and sediment) and biotic components of the resulting aquatic environment (Bodaly and others, 1997; Kelly and others, 1997; Snodgrass and others, 2000; St. Louis and others, 2004; Strange and Bodaly, 1999). Several researchers have observed methylmercury increases in small impoundments like those being restored in the study area (Bodaly and Fudge, 1999; Brigham and others, 2002; St. Louis and others, 2004) that undergo large changes in stage, which results in periodic drying and re-flooding of soils (Snodgrass and others, 2000). Although methylmercury was not analyzed in samples during this study, the conditions needed to methylate mercury probably will be produced by the wetland restorations in the study area. Additional studies could sample mercury and methylmercury concentrations in wetland soils and water to assess the degree to which methylmercury increases in restored wetlands in the study area. An effective design for such a study might pair similar wetland basins, sampling them both temporally as one is restored.

Measuring Restoration Effects

Effects of the wetland and prairie restorations could be measured in the future, when restorations are complete, after they have matured, and the hydrologic system has had time to adjust. The amount of time it will take for the restorations and hydrology to reach a new equilibrium is difficult to predict. One way to estimate when a new equilibrium will be established is to continue monitoring water levels and quality at key locations in the study area. Hydrologic changes should occur as restorations proceed and mature. When the rates of these changes slow, a new assessment of the area's hydrology may be appropriate. A comparison between a future assessment and the one documented in this report would quantify the hydrologic changes resulting from the wetland and prairie restorations. Comparisons of before-and-after areal recharge rates and seasonal timing, ditch base flow, ground- and surface-water mass balances, ditch hydrograph recession slopes, storm-runoff hydrograph model variables, and water quality would be particularly effective in quantifying the hydrologic changes resulting from wetland and prairie restorations in the Glacial Ridge study area.

Summary

The Nature Conservancy (TNC) owned and managed 24,795 acres of mixed wetland, native prairie, farmland, and woods east of Crookston, in northwestern Minnesota. The original wetlands and prairies that once occupied this land are being restored by TNC in cooperation with many partners and are becoming part of the Glacial Ridge National Wildlife Refuge. Results of this study indicate that these restorations are likely to have a substantial effect on the local hydrology.

Water occurs within the study area on the land surface, in surficial aquifers, and in buried aquifers of various depths, the tops of which are 50 to several hundred feet below the land surface. Surficial aquifers are generally thin (about 20 feet), narrow (several hundred feet) and long (tens of miles). Estimates of the horizontal hydraulic conductivity of surficial aquifers were 2.7–300 ft/d. The uppermost buried aquifers (50–100 feet below land surface) underlie much of the study area, but flow in them does not substantially interact with the shallow-water system in most areas. In isolated areas however, buried aquifers directly underlie thin surficial aquifers where a confining unit is absent. In these areas, water flows directly from buried to surficial aquifers and forms a single aquifer as much as 78 ft thick.

Recharge to the surficial aquifers (10.97–25.08 in/yr during 2003–5) is from vertical infiltration of rainfall and snowmelt (areal recharge); from surface waters (particularly ephemeral wetlands); and from upward leakage of ground water from buried aquifers through till confining units (estimated at about 1 in/yr). Areal recharge is highly variable in space and time. Water leaves (discharges from) the surficial aquifers as flow to surface waters (closed basins and ditches), evapotranspiration, and withdrawals from wells. Unmeasured losses (primarily discharge to ungaged (closed) basins) were 53–115 percent of areal recharge during 2003–5, while discharge to ditches that leave the study area was 17–41 percent. Discharge over 100 percent of areal recharge indicates a loss in ground-water storage. Mass-balance dynamics showed that substantial ground water (about one-third annual areal recharge in 2003) was released from aquifer storage during dry years and replenished quickly during wet ones.

Isotopic composition shows that most ground water in the study area comes from precipitation that has not undergone evaporation. Isotope samples show that substantial evaporation occurred in many surface waters by the time the samples were collected (mid-July 2004) but also that many surface waters had not undergone substantial evaporation. Analysis of isotopic data indicates that interactions between ground water and surface water are dynamic, complicated, and variable across the study area. Ground-water recessions indicate that ditches in basin SW1 are relatively well connected to aquifers.

The surface-water channel network is modified by several manmade ditches that were installed to remove excess water seasonally and to drain wetlands. The channels in the network lie primarily parallel to the beach ridges but cut through them in places. Back-beach basin wetlands delay and reduce direct runoff to ditches. During the study, 68–81 percent of water left the basin through unmeasured surface-water losses (primarily evapotranspiration), which is 2- to 4-times that leaving through the ditch system. Base flow in ditches (ground-water discharge) was 30 to 71 percent of all ditch flow. The main channel length of ditch SW4 (4.1 mi) and the main channel slope of ditch SW1 (14.0 ft/mi) are different from other ditches. The basin of ditch SW5 has a smaller area of lakes and wetlands (14 percent) than other basins. Mean annual runoff in all gaged basins except SW3 (2.26 in/yr) was similar (3.69–4.12 in/yr). The Clark storage coefficient median ranged from 28 hours (SW5) to 89 hours (SW1). The coefficient is lowered by a small percentage of lake and wetland area (SW5) and raised by impoundments (SW1).

The postglacial formation of the surface-water system and its subsequent substantial human modification control the flow of nearly all water within and out of the study area. Shallow ground-water flow is complex, with water in surficial aquifers, ditches, and wetlands part of a single hydrologic system. Ground-water and surface-water flow is from areas of high elevation near Maple Lake to those of lower elevation on northern and western edges of the study area. Ground-water flow in buried aquifers does not interact with the surface directly except in two areas where wells withdraw water and in a few areas where surficial aquifers are hydraulically connected. The ages of surficial ground-water samples were less than 15 years old, and one-third (8 of 24) were less than 5 years old, substantiating the close connection of surficial ground water to the land surface.

The quality of surficial ground-water (from surficial aquifers) and surface-water samples collected in the study area was generally suitable for most uses but was variable. Of the 48 ground-water samples collected during May18–July 22, 2004, 39 samples were from wells completed in surficial aquifers and 9 were from wells completed in buried aquifers. The surficial ground-water sampled represents water near the water table that was affected by land use during 2001–3. Most ground- and surface-water samples were dominated by calcium, magnesium, and bicarbonate ions. Samples from gage SW1 contained anomalously high sulfate concentrations that could be released biochemically from wetlands. The average sum-of-solids concentration was 536 mg/L for surficial ground-water samples and 610 mg/L for buried ground-water samples.

Nutrient concentrations in surficial ground-water samples were spatially variable, reflecting the spatial variability of land use in the study area during 2001–3. Nearly one-half of the samples (47 percent) contained nitrate at concentrations higher than 3 mg/L as nitrogen (-N), reflecting the agricultural land use in the study area. About one-quarter of surficial ground-water samples contained nitrate at concentrations greater than the USEPA's Maximum Contaminant Level for human consumption. Surficial ground-water samples contained detectable concentrations of atrazine, metolachlor, pendimethalin, prometon, and terbutryn and 10 of the 19 degradates analyzed. Six ditch samples contained atrazine and acetachlor and seven degradates. In general, herbicide degradates were found more frequently and at higher concentrations than were the parent herbicides. Deisopropylatrazine was detected in 28 percent of the surficial ground-water samples. No herbicide or degradate was detected in buried ground-water samples (from buried aquifers), reflecting the protection that clay-rich confining units afford these aquifers.

Median concentrations of nitrite plus nitrate in ditch samples ranged from less than 0.03 mg/L-N to 0.635 mg/L-N. The median concentration of dissolved phosphorus ranged from 0.0108 mg/L as phosphorus (-P) to 0.0293 mg/L-P. Nutrient concentrations in ditches were generally above the USEPA nutrient guidelines for reference streams in the area. Deethylatrazine was detected in 3 of 6 ditch-water samples.

The restoration of wetlands and prairies in the study area likely will result in more water retained on the land and improved water quality. Increased water retention could raise ground-water levels, but the rise likely would be very local and short-lived. Restorations likely would substantially change ditch-flow characteristics in the study area, but the changes would be insubstantial further downstream. Reduction in agriculture could result in a net decrease in nutrient and pesticide load to the study area.

A comparison between a future assessment and the one documented in this report would quantify the hydrologic changes resulting from the wetland and prairie restorations. Comparisons of before-and-after areal recharge rates and seasonal timing, ditch base flow, ground- and surface-water mass balances, ditch hydrograph recession slopes, storm-runoff hydrograph model variables, and water quality would be particularly effective in quantifying the hydrologic changes resulting from wetland and prairie restorations in the Glacial Ridge study area.

Acknowledgments

Many employees of The Nature Conservancy, both on-site at the Glacial Ridge project office and in their state office in Minneapolis, aided and supported this study. We particularly thank Jason Eckstein, the Glacial Ridge restoration ecologist, and Keith Mykleseth, Glacial Ridge project manager, for giving us important insights into the land they manage and a home base from which to work. We also thank Ron Nargang, former TNC state director, and Mr. Mykleseth for building and maintaining the coalition of Glacial Ridge neighbors and partners that supported this study. Other partners that substantially supported the study include employees of the Red Lake Watershed District, the City of Crookston Water Department, and Professor Phil Gerla of the University of North Dakota and of TNC. Many thanks go to Wally Larson of the USGS and to USGS interns Jody Hulne and Alyssa Boock for their dedication and diligence in collecting much of the field data for this study and keeping equipment running. Neighboring landowners were generous in providing permission to install wells and gages and giving access to their property.

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Appendix 1. Land-Use History

Agriculture has dramatically changed the land cover of the Glacial Ridge study area over the last 100 years. The following land-use history is based on a 2004 interview with Jason Eckstein, Restoration Ecologist for The Nature Conservancy (TNC) at their Glacial Ridge project office (Jason Eckstein, The Nature Conservancy, Glacial Ridge Project, oral commun., 2004).

Before European settlement, the study area was a treeless mixture of prairies occupying the beach ridges with various wetlands and wet meadows in the interbeach swales. The prairies were used for trails by native peoples, and European settlers did likewise, establishing the Pembina Trail across the study area well before 1846 (Gilman and others, 1979). When homesteaders began settling in the area in the last quarter of the 19th century, they grew crops (mainly small grains) on beach ridges and hayed the meadows wherever they could. The area made generally poor cropland because the coarse-grained beach ridges quickly dried out, whereas the wet meadows could not be planted early enough in the spring.

In an attempt to make some of the wet meadows tillable, the study area was extensive ditched during the early 1900s. Despite ditching, wetlands fed by ground-water discharge on the downgradient faces of the beach ridges and wetlands in deeper basins survived. Through the 1950s, small farms continued to operate in the area. Even with drainage improvements, however, the land was still agriculturally marginal, and farming was difficult. Beginning in the early 1960s a group of Texas investors began buying land in the study area, consolidating it for cattle grazing. This group eventually amassed a holding of about 25,000 acres of mostly contiguous land. Through the 1980s, the grazing operation was sold to one or two buyers about every 5 years. During this time two large feedlots were built on the consolidated property.

By the late 1980s, the ecological value of the property was beginning to become apparent. Although the land was drained, many wetlands and native prairie parcels remained that were important for wildlife, particularly waterfowl. When the property was put up for sale at this time, the Minnesota Department of Natural Resources, the U.S. Fish and Wildlife Service, and TNC considered acquiring the land for its ecological assets. However, a private investor interested in agriculture bought the land. The period 1986–92 was relatively dry. Many parts of the study area that had previously been impossible to farm were tilled for the first time during this period. Crops (wheat, barley, corn, soybeans, and sunflowers (alfalfa and edible beans on ridges)) were planted as closely as possible to land too wet to till, resulting in damage to substantial areas of temporarily dry wetlands and wet meadows. The row-crop agriculture practiced was at a scale unheard of in the former homestead-cropping period. Thousands of field boulder piles, initially gathered by homesteaders, were buried in place to facilitate industrial agriculture.

As precipitation returned to more normal levels in the later 1990s, large areas again became un-tillable, and the property was again offered for sale. TNC purchased the property in 2000 and began wetland and prairie restorations in 2002. To control noxious weeds, the land formerly under cultivation was rented for continued planting until restoration activities could begin. As of July 2007, about 11,000 acres of formerly cropped or grazed land had been restored to native prairie (10,000 acres) or wetland (1,000 acres), representing 44 percent of TNC-owned acreage in the study area. As of the same date, 58 miles of ditches were either filled or plugged. Within the study area, an additional 3,000 acres of other private land had been enrolled in permanent wetland easement and had been restored during 2000-5 (fig. 2).

Appendix 2. Site names, numbers, and types.

Table 2-1. Site names, numbers, and types.

[MUN, Minnesota unique well number (blank were unknown); GW-G, well drilled for the Glacial Ridge study; GW-E, existing well; GW-L, well drilled for an earlier U.S. Geological Survey project by Lindgren (1996); GW-C, Crookston Water Department observation well; SW, ditch gage; L, lake gage; WL, wetland gage; —, not applicable; S, surficial; B, buried; depths in feet below land surface]

Short name	Agency code	Site number	MUN	Type	Aquifer type	Well depth	Screened interval depth
G01	USGS	474135096203001	620661	GW-G	S	10.42	5.58–9.88
G02	USGS	473849096202101	620662	GW-G	S	14.49	9.65–13.95
G03	USGS	473914096195401	620663	GW-G	S	14.61	1.33–10.64
G04	USGS	474242096194701	620664	GW-G	S	9.85	5.01–9.31
G05	USGS	474229096185701	620665	GW-G	S	9.43	0.77–5.07
G06	USGS	474119096190901	620666	GW-G	S	12.94	8.1–12.4
G07D	USGS	474300096172602	620667	GW-G	S	36.05	31.21–35.51
G07S	USGS	474300096172601	620657	GW-G	S	15.58	10.65–14.95
G08	USGS	474346096185501	620668	GW-G	S	11.81	6.97–11.27
G09	USGS	474129096145202	620669	GW-G	S	10.60	5.67–9.97
G10	USGS	474109096133501	620670	GW-G	S	10.23	3.01–7.31
G11	USGS	474254096160401	620671	GW-G	S	16.34	3.58–7.88
G12	USGS	474126096165301	620672	GW-G	S	14.94	8.76–13.06
G13	USGS	474128096175501	620673	GW-G	S	13.92	4.14–8.44
G14	USGS	473842096183901	620674	GW-G	S	15.54	5.58–9.88
G15	USGS	473841096153101	620675	GW-G	S	14.87	7.17–11.47
G16	USGS	474221096120901	620676	GW-G	S	14.41	6.7–11
G17	USGS	474350096144101	620677	GW-G	S	13.47	8.63–12.93
G18	USGS	474534096182701	620678	GW-G	S	29.87	14.97–19.27
G19	USGS	474524096203101	620679	GW-G	S	43.53	38.6–42.9
G20	USGS	474310096121801	620680	GW-G	S	15.01	6.34–10.64
G21	USGS	474420096104901	620681	GW-G	S	12.33	7.49–11.79
G22	USGS	474125096120602	620682	GW-G	S	29.77	24.93–29.23
G23	USGS	474721096232201	620683	GW-G	S	24.56	13.86–18.16
G24	USGS	474220096171801	620684	GW-G	S	11.70	6.86–11.16
G25	USGS	473933096243701	620685	GW-G	S	30.35	20.38–24.68
G26	USGS	474133096245901	620686	GW-G	S	14.89	10.05–14.35
G27	USGS	473901096164901	620687	GW-G	S	10.11	5.27–9.57
G30	USGS	473855096141301	620690	GW-G	S	10.10	3.43–7.73
G32	USGS	474300096204901	620692	GW-G	S	11.66	4.96–9.26
G33	USGS	474201096132501	620693	GW-G	S	25.25	20.32–24.62
G34	USGS	474443096171801	620694	GW-G	S	12.64	7.71–12.01
G35	USGS	474043096155901	620695	GW-G	S	19.88	12.18–16.48
G36	USGS	474135096204501	620696	GW-G	S	10.01	5.17–9.47
G38	USGS	474444096183101	620698	GW-G	S	14.61	2.91–7.21
G39	USGS	474055096150301	620699	GW-G	S	14.08	7.41–11.71
E01S	USGS	473945096202402	249810	GW-E	S	19.77	14.5–19.5
E03	USGS	474436096140801	654754	GW-E	S	69.00	59–69
E05	USGS	474719096163100		GW-E	S	18.06	12.7–17.7
E09	USGS	474353096164401		GW-E	S		

Table 2-1 Site names, numbers, and types.—Continued

[MUN, Minnesota unique well number (blank were unknown); GW-G, well drilled for the Glacial Ridge study; GW-E, existing well; GW-L, well drilled for an earlier U.S. Geological Survey project by Lindgren (1996); GW-C, Crookston Water Department observation well; SW, ditch gage; L, lake gage; WL, wetland gage; —, not applicable; S, surficial; B, buried; depths in feet below land surface]

Short name	Agency code	Site number	MUN	Type	Aquifer type	Well depth	Screened interval depth
E13	USGS	474506096205901	221630	GW-E	S	55.42	52–56
E19	USGS	474305096172401		GW-E	S	39.16	
E23	USGS	474535096204201		GW-E	S	64.27	
L012	USGS	473042096151800	249806	GW-L	S	10.25	5–10
L032	USGS	474629096210801		GW-L	S	14.47	
L043	USGS	474708096261801		GW-L	S	24.26	18–23
L057	USGS	474628096180101		GW-L	S	13.48	9.5–14.5
L061	USGS	474629096193901		GW-L	S	29.84	24–29
S1	USGS	474539096205101	125721	GW-C	S	53.59	41–56
S2	USGS	474539096203302	105665	GW-C	S	50	46–50
E01D	USGS	473945096202401	516287	GW-E	B	171	168–171
E02	USGS	474129096145201		GW-E	B	173.58	
E04D	USGS	474309096122001	654760	GW-E	B	102	98–102
E06D	USGS	474455096250601	249807	GW-E	B	44.84	40–45
E07	USGS	474255096155601	107932	GW-E	B	80	67–80
E10	USGS	474541096174001	649189	GW-E	B	115.41	111–115
E14	USGS	474256096222001		GW-E	B	64.15	
E15	USGS	474207096171101	221063	GW-E	B	81.62	102–105
E21	USGS	474339096191301		GW-E	B	55.64	
E22	USGS	474331096193301		GW-E	B	169.64	
E24	USGS	474220096154101		GW-E	B	90.16	
E25	USGS	474224096160501		GW-E	B	173.80	
E27	USGS	473941096151801		GW-E	B	125.77	
E28	USGS	473905096153101		GW-E	B	123.53	
E36	USGS	474340096191301		GW-E	B	70.97	
E37	USGS	474125096120601		GW-E	B	65.15	
E38	USGS	474251096131201		GW-E	B	114.04	98–102
E39	USGS	474422096111301		GW-E	B	82.45	
E40	USGS	474424096101901		GW-E	B		
E41	USGS	474334096111601		GW-E	B	47.83	
E45	USGS	474251096131201		GW-E	B	162.24	
L101	USGS	474537096160300	513018	GW-L	B	171.77	169.6–172.6
L102	USGS	474720096150201	516274	GW-L	B	172.23	170–173
L103	USGS	474210096203101	516278	GW-L	B	190.43	187–190
L109	USGS	474536096134401	516273	GW-L	B	162.40	162–165
D1	USGS	474547096210501	105666	GW-C	B	147	123–147
D2	USGS	474540096210401	147234	GW-C	B	172	135–145
D3	USGS	474559096203302		GW-C	B	158.11	135–157
D4	USGS	474634096202601	147242	GW-C	B	96.97	87–97
SW1	USGS	05078730	—	SW	—	—	—
SW2	USGS	05079250	—	SW	—	—	—
SW3	USGS	05079200	—	SW	—	—	—
SW4	USGS	05078470	—	SW	—	—	—
SW5	USGS	05078520	—	SW	—	—	—
SW6	USGS	05078770	—	SW	—	—	—

Table 2-1 Site names, numbers, and types.—Continued

[MUN, Minnesota unique well number (blank were unknown); GW-G, well drilled for the Glacial Ridge study; GW-E, existing well; GW-L, well drilled for an earlier U.S. Geological Survey project by Lindgren (1996); GW-C, Crookston Water Department observation well; SW, ditch gage; L, lake gage; WL, wetland gage; —, not applicable; S, surficial; B, buried; depths in feet below land surface]

Short name	Agency code	Site number	MUN	Type	Aquifer type	Well depth	Screened interval depth
SW7	USGS	474003096085901	—	L	—	—	—
SW8	USGS	05078720	—	SW	—	—	—
WL01	USGS	474024096124601	—	WL	—	—	—
WL02	USGS	474026096145001	—	WL	—	—	—
WL03	USGS	474139096150301	—	WL	—	—	—
WL04	USGS	474137096154101	—	WL	—	—	—
WL05	USGS	474127096164701	—	WL	—	—	—
WL06	USGS	474139096165401	—	WL	—	—	—
WL07	USGS	474129096180001	—	WL	—	—	—
WL08	USGS	474120096185001	—	WL	—	—	—
WL09	USGS	474228096171901	—	WL	—	—	—
WL10	USGS	474328096144201	—	WL	—	—	—
WL11	USGS	474205096110401	—	WL	—	—	—
WL12	USGS	474330096175701	—	WL	—	—	—
WL13	USGS	474258096210702	—	WL	—	—	—

Appendix 3. Detailed Methods

This appendix provides details of well construction and the stratigraphic, water-level, mass-balance, and hydrographic analyses, outlined in the "Methods" section of this report. The appendix also specifies water-quality sampling methods.

Well construction

Wells installed for this study were constructed of 2-in.-inside-diameter, schedule 40, flush-threaded polyvinyl chloride (PVC) with 4.3-ft-long, 0.010-in.-slotted PVC screens set to intersect the water table. Screens were set at the water table to access ground water affected by the most recent land use. Each well is surrounded by a 6-in.-diameter, schedule 40 steel pipe set 3 ft into concrete, covered by a locking cap. The wells were installed with a hollow-stem power auger and the annulus was allowed to collapse around the screen to a height of 10 ft above the screen. If the aquifer material was finer than sand, washed sand was poured around the well screen during auger removal. If aquifer material did not collapse to a point 10 ft above the screen, sand was added to the annulus. Wells were then sealed to the surface with mixed bentonite grout. Wells were developed by pumping with a submersible, centrifugal, or hand pump. Augers and development pumps were used at subsequent wells without decontamination.

Stratigraphic and Water-Level Analysis

The extent of the surficial aquifers was interpreted primarily from the areas of coarse-grained soils as described in NRCS digital soil surveys. Aquifer thickness was interpreted from stratigraphic logs from many sources. Data from the Red Lake County and Polk County soil surveys (U.S. Department of Agriculture, various dates) were procured from the NRCS (*http://soildatamart.nrcs.usda.gov/*, accessed in 2002 (Polk County) and 2004 (Red Lake County)). Additional electronic information about soil parent material, depositional landform, and landform position of each soil-map unit in Polk County was provided by NRCS soil survey personnel in Crookston, Minn. (Soil survey personnel, U.S. Department of Agriculture, Natural Resources Conservation Service, Crookston, Minn., electronic commun., 2004). This information was extended to the same soil-map units that were used in Red Lake County. The additional information was inferred for soil-map units unique to Red Lake County by relating units with common "competing series" or that were "geographically associated soils." These series and associations are described online (U.S. Department of Agriculture, 2005).

Soil-map units with sand and gravel as a primary component of the soil horizons were considered surficial-aquifer areas. Soil-map units containing relatively coarse-grained, well-sorted deposits in the parent material and depositional landform descriptions were provisionally classified as surficial-aquifer areas. This classification was cross-checked for consistency with the descriptions of "typical pedon" and "geographical setting" descriptions to produce the final classification (table 3-1). All surficial-aquifer areas were aggregated and internal boundaries dissolved to produce the preliminary surficial-aquifer extent map. In effect, the result is the area where sand and gravel is found at the land surface.

The final surficial-aquifer extent map was produced by modifying the preliminary map to be consistent with other records of surficial aquifer material, particularly stratigraphic logs. Areas of conflict were addressed case by case. Typically, the locational uncertainty of the conflicting stratigraphic data was so high that no boundary adjustment was necessary. Several stratigraphic logs containing surficial sand were located in the middle of a soil map unit not designated as surficial-aquifer area. These areas were interpreted as isolated and not part of an extensive surficial aquifer, and no boundary adjustment was applied. Modifying the preliminary map to resolve other conflicts resulted in the final surficial-aquifer extent map.

Stratigraphic logs from many sources formed the 450-log database used to modify the surficial-aquifer extent map and to construct the surficial-aquifer thickness map. Only stratigraphic logs that contained complete location information and detailed stratigraphic descriptions and depths were included. The City of Crookston furnished consultant reports that contained 75 stratigraphic logs (Richard Normandin, Water Department, Crookston, Minnesota, written commun., 2002–4). Records of prior USGS studies in the area contained 77 stratigraphic logs. Stratigraphic logs of 36 new wells and 42 new boreholes constructed for this study supplemented existing data. Finally, the County Well Index (CWI), maintained by the Minnesota Geological Survey (Robert G. Tipping, Minnesota Geological Survey, electronic commun., 2002), and paper files maintained by the USGS made up the remaining 220 stratigraphic logs in the database.

The thickness of the surficial aquifers was inferred by use of stratigraphy from the stratigraphic-log data base and incorporating a conceptual understanding of aquifer shape based on aquifer depositional history. Stratigraphy from existing and new logs provided the general thickness of aquifers across the study area. The location of wells with logs was heavily skewed toward thick aquifer areas because most were drilled to supply water. New exploratory drilling in the area of well G39 (southeast part of the study area; fig. 2) confirmed the conceptual understanding of aquifer shape, which was then applied across the study area. Thickness contours of surficial aquifers included several assumptions: the thickest part of the aquifer is

Table 3-1. Soil map units with parent material interpreted as surficial aquifer deposits, Glacial Ridge study area, northwestern Minnesota.

Deposit type	Name	Code[a]	Soil texture	Dominant parent material	Landform	Position on landform
Beach deposits	Gravel Pits	1030	Sand	Beach deposits	none listed	none listed
	Sandberg loamy sand	258B	Loamy sand	Beach deposits	Beach ridges	Rises and backslopes
	Sandberg loamy sand	258C	Loamy sand	Beach deposits	Beach ridges	Summits and backslopes
	Radium loamy sand	1874	Loamy sand	Beach deposits	Beach ridges	Slight rises
	Hangaard sandy loam	111	Sandy loam	Beach deposits	Beach plains	Flats and swales
	Syrene sandy loam	435	Sandy loam	Beach deposits	Beach plains	Flats
	Wyrene sandy loam	704	Sandy loam	Beach deposits	Beach plains	Slight rises
Sandy and loamy glacial outwash and till on outwash plains and moraines	Arvilla sandy loam	341B	Sandy loam	Outwash	Beach plains and outwash plains	Rises and backslopes
	Halverson loamy fine sand	735B	Loamy fine sand	Glacial outwash over till	Moraines	Rises
	Halverson loamy fine sand	735C	Loamy fine sand	Glacial outwash over till	Moraines	Summits and backslopes
Sandy and loamy glaciolacustrine deposits on lake or outwash plains	Maddock loamy fine sand	45B	Loamy fine sand	Glaciolacustrine deposits	Lake and outwash plains	Rises and backslopes
	Flaming loamy fine sand	66	Loamy fine sand	Glaciolacustrine deposits	Lake plains	slight rises
	Ulen loamy fine sand	1264	Loamy fine sand	Glaciolacustrine deposits	Lake plains	slight rises
	Hamar loamy fine sand	372	Loamy fine sand	Glaciolacustrine deposits	Lake plains	Flats and swales
	Rosewood fine sandy loam	712	Fine sandy loam	Glaciolacustrine deposits	Lake plains	flats
	Rosewood-Venlo complex	1278	Fine sandy loam	Glaciolacustrine deposits	Lake plains	none listed
Depressions in sandy outwash or glaciolacustrine deposits	Deerwood muck	547	Muck	Organic materials over glaciolacustrine deposits	Lake plains	Depressions
	Markey muck	543	Muck	Organic materials over glacial outwash or glaciolacustrine deposits	Lake plains, outwash plains, and moraines	Depressions

Table 3-1. Soil map units with parent material interpreted as surficial aquifer deposits, Glacial Ridge study area, northwestern Minnesota.—Continued

Deposit type	Name	Soil Series (U.S. Department of Agriculture, 2005)		
		Description	Geographic Setting	
			Parent Material	Geomorphic Description
Beach deposits	Gravel Pits	none listed	none listed	none listed
	Sandberg loamy sand	excessively drained soils that formed in coarse or moderately coarse glacial outwash sediments or glacial beach deposits with or without a thin loamy mantle	formed in sandy and gravelly outwash sediments or glacial beach deposits with or without a thin loamy upper mantle	on outwash plains, glacial beach ridges, valley trains, stream terraces and glacial moraines
	Sandberg loamy sand	excessively drained soils that formed in coarse or moderately coarse glacial outwash sediments or glacial beach deposits with or without a thin loamy mantle	formed in sandy and gravelly outwash sediments or glacial beach deposits with or without a thin loamy upper mantle	on outwash plains, glacial beach ridges, valley trains, stream terraces and glacial moraines
	Radium loamy sand	moderately well-drained soils that formed in sandy glaciolacustrine and outwash sediments on glacial lake beaches and outwash plains	formed in sandy glaciolacustrine and outwash sediments	on convex areas on glacial lake plains and outwash plains
	Hangaard sandy loam	poorly drained soils formed in outwash sediments on interbeach areas of glacial lake plains, beach ridges and outwash plains	formed in a thin, discontinuous, moderately coarse-textured sediment overlying coarse textured outwash	on plane or slightly concave positions on glacial lake plains, beach ridges or outwash plains
	Syrene sandy loam	poorly and very poorly drained soils that formed in glaciolacustrine sediments consisting of a loamy mantle over sandy sediments on beaches of lake plains	formed in 12 to 24 inches of loamy sediments over sandy sediments	on lake plains mostly adjacent to beach ridges
	Wyrene sandy loam	somewhat poorly drained soils that formed in moderately coarse overlying coarse textured sediments	formed in moderately coarse overlying coarse-textured sediments	on level and nearly level slightly depressed areas on outwash plains and interbeach areas
Sandy and loamy glacial outwash and till on outwash plains and moraines	Arvilla sandy loam	somewhat excessively drained soils formed in moderately coarse textured glacial outwash and the underlying sand and gravel on glacial lake beaches, stream valley terraces and outwash plains	formed in a thin mantle of sandy loam alluvium underlain by thick beds of loose sand and gravel	on level to moderately steep outwash plains, beach areas of glacial lakes, and terraces of glacial stream valleys
	Halverson loamy fine sand	well-drained soils that formed in a mantle of sandy glacial outwash or eolian sands and in the underlying loamy till	formed in sandy glacial outwash or eolian sands and in underlying loamy till	on sand-capped till plains and moraines
	Halverson loamy fine sand	well-drained soils that formed in a mantle of sandy glacial outwash or eolian sands and in the underlying loamy till	formed in sandy glacial outwash or eolian sands and in underlying loamy till	on sand-capped till plains and moraines

Table 3-1. Soil map units with parent material interpreted as surficial aquifer deposits, Glacial Ridge study area, northwestern Minnesota.—Continued

Deposit type	Name	Soil Series (U.S. Department of Agriculture, 2005)		
		Description	Geographic Setting	
			Parent Material	Geomorphic Description
Sandy and loamy glaciolacustrine deposits on lake or outwash plains	Maddock loamy fine sand	well-drained or somewhat excessively drained, rapidly permeable soils that formed in fine sands deposited by wind or water	formed in fine sands deposited by wind or water	on level to steep sandy glaciolacustrine or glaciofluvial, outwash and delta plains, some of which have been wind worked
	Flaming loamy fine sand	moderately well-drained soils formed in sandy sediments on lake plains, lake plains and till plains	formed in deep sandy deposits of lacustrine and/or eolian origin	on level or nearly level glacial lake plains, lake plains, and till plains
	Ulen loamy fine sand	somewhat poorly drained soils that formed in sandy glaciolacustrine deposits on glacial lake plains	formed in thick glaciolacustrine deposits that are dominated by fine sand	on glacial lake plains
	Hamar loamy fine sand	poorly drained soils formed in eolian or lacustrine sands in upland swales and depressions	none listed	on sandy lacustrine and glacial outwash plains and till plains mantled by eolian sand
	Rosewood fine sandy loam	poorly and very poorly drained soils that formed in calcareous sandy lacustrine sediments on glacial lake plains	formed in calcareous sandy lacustrine or outwash sediments	on glacial lake plains and hillside seeps
	Rosewood-Venlo complex	very poorly drained, rapidly permeable soils that formed in glaciofluvial or glaciolacustrine deposits	formed in glaciofluvial or glaciolacustrine deposits	in low basins and swales on delta, outwash, and lake plains
Depressions in sandy outwash or glaciolacustrine deposits	Deerwood muck	very poorly drained soils formed in a thin organic mantle and sandy lacustrine or outwash sediments	formed in a thin layer of organic soil material over sandy glacial lacustrine or outwash sediments	on glacial lake plains and glacial outwash plains
	Markey muck	very poorly drained organic soils. They formed in herbaceous organic material 16 to 51 inches thick overlying sandy deposits in depressions on outwash plains, lake plains, flood plains, river terraces, valley trains, and moraines	in depressions within outwash plains, lake plains, flood plains, river terraces, valley trains, and moraines	within outwash plains, lake plains, flood plains, river terraces, valley trains, and moraines

[a] Polk County mapunit code from U.S. Department of Agriculture (various dates), Soil survey personnel, U.S. Department of Agriculture, Natural Resources Conservation Service, Crookston, Minn., written commun., 2004.

near the center of a beach ridge, a stratigraphic log is generally located at the thickest part of a beach ridge, and aquifer thickness gradually thins to zero at the edges of the beach ridge. These assumptions must be kept in mind when using the surficial-aquifer thickness map. In areas were several logs of differing stratigraphy were available, stratigraphy among them was compared and the best composite stratigraphic log was constructed, taking into account the detail contained in the original log and the usefulness of other logs produced by the same driller.

Surficial aquifer thickness in areas lacking stratigraphic logs was based on soil-map-unit descriptions and distribution, topography, and landform. Some soil-map units classified as aquifer areas are primarily in a transition zone between aquifer and nonaquifer areas. These areas were assumed to have aquifer thickness less than 10 ft. Features such as gravel pits and wetlands also were considered when contouring the surficial-aquifer thickness map. Gravel pits were assumed to be areas of relatively thick (greater than 35 feet) aquifer material. Wetlands are usually located in topographically low interbeach swales. Where an aquifer-area soil-map unit coincided with a wetland, the aquifer was assumed to be thin and overlain by a layer of fine-grained, organic-rich material.

Water levels measured synoptically in June 2004 at 72 wells, 11 wetlands, 1 lake, and 6 ditch gages were used to draw the water-table elevation and buried aquifer potentiometric-surface maps. These water levels were supplemented by water levels from 47 boreholes constructed by the USGS in the early 1990s (Lindgren, 1996), as were the elevations of surface waters likely to be a surface expression of the water table. The elevation of the land surface and measuring point at synoptic wells and the elevation of ditch beds were measured to within 4 in. using pairs of differential global-positioning-system (DGPS) receivers. The elevation of the bottom of ditches was estimated from USGS 7.5-minute topographic maps (accuracy ±2.5 ft) where DGPS data were not available.

Twenty-five wells in the study area were screened in buried artesian aquifers of about the same elevation. It is possible that water from these aquifers can discharge, in fractures or diffusely through the intervening tills, upward into the surficial aquifers. To evaluate the interconnectedness of the buried aquifers and to estimate the gradient driving water from buried to surficial aquifers on a regional scale, a potentiometric-surface map of the confined aquifers was interpolated from the water levels synoptically measured during the June 2004. Water levels for two wells measured in October and December of 2004 were included to increase the number of elevations with which to interpolate. The buried aquifers were assumed to be relatively hydraulically interconnected and a spatially smooth potentiometric surface resulted. Homogeneous leakage of water from buried aquifers upward into surficial aquifers or to the land surface was assumed. Data from stratigraphic logs available for 13 of these 25 wells helped confirm the interconnectedness of the buried aquifers.

Water levels from wetland gage WL12 were used to construct the potentiometric-surface map of the buried aquifers. This gage (fig. 2) is in a pool dug during gravel mining. This gravel pit is the largest active gravel mining operation in the study area. Gravel is removed from a pit whose bottom is well below the water table. This gravel deposit is probably formed of buried aquifer material overlain directly by surficial aquifer material. Therefore, locally, water from the buried aquifer likely discharges directly to the surficial aquifer and into the pit pool. The most compelling evidence for this connection and discharge is that water flows nearly continually from the pit lake into Judicial Ditch 66 (10-percent low flow, 0.08 ft^3/s, water years 2003–4). Other ditches in the study area do not receive water discharge directly from buried aquifer and are dry for large parts of the year. Water levels measured at gage WL12 are assumed to be those in the buried aquifer in this area.

Mass-Balance Analysis

Annual water mass balances were calculated separately for the surface-water and surficial ground-water systems. Mass balances were calculated only for those parts of the study area that drain to gaged ditches because these are the only areas that have measurements of ditch-water flow and estimates of ground-water discharge to ditches. The following equations were used to calculate the mass balances:

Surface-water mass balance

$$P + G = R + D + Ls$$

Ground-water mass balance

$$R = G + \Delta S + Lg$$

where:

- P Precipitation (measured at 8 wells)
- G Ground water discharge to ditches (calculated from hydrographs at 6 gages)
- R areal Recharge to surficial aquifers (calculated from hydrographs at 6 wells)
- D flow out of the basin in Ditches (measured at 6 gages)
- Ls unmeasured surface water Losses (calculated from mass balance)
- ΔS change in ground-water Storage (measured at 6 wells)
- Lg unmeasured ground-water Losses (calculated from mass balance)

Each of these terms was calculated for ditch basins SW1–SW6 and summed for the total gaged basin area. The following equations were used to calculate terms above for each basin:

annual Precipitation volume for basin

$$P_b = \sum_{w=1}^{8} (Ap_b \bullet Pt)_w$$

annual areal Recharge volume to surficial aquifers for basin

$$R_b = \sum_{w=1}^{6} (Ar_a \bullet Rt)_w$$

annual change in ground-water Storage for basin

$$\Delta S_b = \sum_{w=1}^{6} (Ar_a \bullet (ht_2 - ht_1))_w$$

G_b = annual Ground-water discharge to basin b
D_b = annual Ditch discharge out of basin b measured at gage b
Ls_b = annual surface water unmeasured Losses and measurement errors
Lg_b = annual ground water unmeasured Losses and measurement errors

where:

Ap_{bw}	area of the basin within the Thiessen polygon for precipitation site (measured)
Pt_w	annual precipitation at site (measured)
Ar_{aw}	surficial aquifer area of the basin within the Thiessen polygon for recharge well (measured)
Rt_w	annual recharge at well (measured)
ht_{1w}	ground-water level at time 1 at well
ht_{2w}	ground-water level at time 2 at well

Precipitation data recorded at 3 of 10 wells were not used in the mass-balance analyses. Precipitation data from well G25 was not used because its Thiessen polygon did not coincide with any basin area. Precipitation data from wells G22 and E05 were not used because the rain gages at these wells did not work reliably for long periods of time. Recharge estimates from hydrographs at 4 of 10 wells completed in surficial aquifers were not used in the areal recharge term of mass-balance analyses. Recharge at wells G01 and G15 included substantial amounts from adjacent wetlands and was not areal recharge. Recharge at well G25 was excluded because its Thiessen polygon did not coincide with any basin area. Recharge at well G22 was excluded because water levels at the well may have been affected by pumping and return flow from adjacent irrigation.

Buried aquifers with upward head gradients underlie most of the study area. Only one estimate of the leakage to, and hence leakage from, these aquifers has been made in this study (at well E01D). It is unknown how accurate and representative this leakage rate is or over what area it operates. Therefore, the leakage of ground water from buried aquifers to surface waters and to surficial aquifers was ignored in the mass balances calculated in this report. Any leakage from buried aquifers unaccounted for in these mass balances would have the effect of increasing the surface- or ground-water loss terms (Ls and Lg) above.

The loss terms Ls_b and Lg_b also contain all measurement error from all the other terms in the mass balances and are, therefore, the least well-known values. In physical terms, surface-water losses (Ls_b) are runoff to closed basins from aquifer and nonaquifer areas, aquifer recharge from ditches, evapotranspiration from ditches and adjacent wetlands, and so on. Ground-water losses (Lg_b) are runoff from aquifer areas (water that did not infiltrate) to closed basins, evapotranspiration from the water table, discharge to basins other than those gaged, and so on.

Hydrographic Analysis

Four different hydrographic analyses were used to aid in understanding the characteristics of direct runoff and ground-water discharge-to-ditch flow:

- a description of daily flows
- streamflow partitioning to determine ground-water base flow
- hydrograph recessions and slopes
- Storm-runoff hydrograph modeling by Clark unit-hydrograph analysis

The daily high- and low-flow analysis was made by compiling the highest and lowest recorded daily mean flow at each of six ditch gages for each water year, from 2003 to 2005. Stage was recorded every 15 minutes, converted to flow with a stage-flow relation equation, and stored in the USGS National Water Information System database. Daily mean flow is the average of recorded 15-minute interval flows during a day.

The USGS computer program PART (Rutledge, 1998) was used to determine the base-flow component of the streamflow hydrographs at six gages. The program uses streamflow partitioning to estimate a record of base flow from the streamflow hydrograph. Streamflow partitioning scans the streamflow record of daily means for flows after a specified number of days following a peak, designates base flow to be equal to streamflow on those days, and then linearly interpolates the record of base flow between the preceding recession and the specified number of days. The program is applied to a long period of record to give an estimate of the mean rate of ground-water discharge to a stream.

PART uses basin drainage area to compute the number of days after a peak when surface runoff ends and recession flow begins. The equation is $N = A^{0.2}$, where N is the number of days after a peak and A is the drainage area in square miles. PART computes three base-flow-separation estimates, one for the integer value of N and one each for that value plus 1 day and plus 2 days.

To describe the change over time of ground-water discharge to ditches, hydrograph recessions and slopes were analyzed to determine the recession index—the number of days per common log cycle of streamflow recession. The USGS computer program RECESS (Rutledge, 1998) was used to find hydrograph recessions of a minimum duration and to compute the recession slope. Recessions of at least 10 days were used in this study. The program relates the logarithm of streamflow to the time of recession (in days) to estimate the recession slope for each recession. The median value from all hydrograph recessions is the recession index for each gage.

Storm-runoff modeling was done with the U.S. Army Corps of Engineers HEC-HMS modeling system (U.S. Army Corps of Engineers, 1995, 2000, 2001). This modeling accounts only for rainfall loss, for direct runoff, and for ground-water discharge recession. Rainfall losses from each basin for three selected storms were estimated with the deficit-and-constant method. Excess rainfall was transformed into direct runoff using the Clark unit-hydrograph method (Clark, 1945). The variables of this method describe how water runs off of a basin with certain land-use or flow characteristics. The recession method was used to estimate ground-water base flow after direct runoff ended. Basins were assumed to have no impervious area in all models. The program's optimization routine used recorded runoff hydrographs and storm rainfall to determine seven model variable values. These variables are listed below:

Precipitation-loss variables for the deficit-and-constant method:

- Initial Loss—the initial interception and infiltration of precipitation that occurs prior to runoff.
- Constant Loss Rate—the rate at which precipitation infiltrates into the soils.

Transformation variables for the Clark-unit-hydrograph method:

- Time of Concentration—the maximum time required for water to travel as surface runoff from anywhere in the basin to the outlet of the basin, assuming no storage.
- Clark Storage Coefficient—a variable used to describe the effects of all storage within a basin.

Base-flow variables for recession method:

- Initial Base Flow—the flow prior to an increase in flow from rainfall runoff.
- Recession Constant—the rate of base-flow decrease, represented as the average ratio of base flow between subsequent days.
- Recession Threshold—the flow at which ground-water base flow replaces overland flow as the source of water leaving the basin.

A hydrograph model composed of a basin component, a rainfall component, and a time component was constructed for each storm in each basin. The basin component included the recorded ditch flow, drainage area, and model variables. The rainfall component is a weighted average of the recorded rainfall at each rain gage within or near a basin. The weights were the areas of the Theissen polygons of rainfall gages in each basin. The time component was the period of a storm and its subsequent runoff. Short time-period flow and rainfall data are required to model small basins like those in the study area. Flow recorded at 15-minute intervals and rainfall recorded at 60-minute intervals were used in these models. Flow was recorded at ditch gages, and rainfall was recorded at wells and the SCAN station. Rainfall data used in these models were preliminary but differed little from the final data.

Model variables were optimized after the inputs to the model were entered. Initial values of model variables for each storm and basin were estimated by examining the recorded hydrographs. The model then produced a model hydrograph, which was compared to the recorded hydrograph for accuracy. An automatic optimization process proceeded iteratively, by changing variables until differences between the model and recorded hydrographs were minimized. This process produced optimized model variable values.

Synoptic Sampling

Synoptic water samples were collected during May 18–July 22, 2004 from 39 surficial aquifer wells, 9 buried aquifer wells, 6 ditch gages, and 7 wetland gages (see appendix 2 for site details). Although interpreted as one data set, ground-water samples were not necessarily collected from a single hydrologically connected aquifer. Because surficial aquifers were generally subject to the same mix of hydrologic and land-use factors, the aggregate water quality of these 39 samples characterized the state and areal variability of the quality of all surficial ground water in the study area. This characterization is not statistically valid, however, because the sampling locations were not randomly chosen. Instead, they were selected to provide an even spatial coverage of the study area.

The areal extent and interconnectedness of buried aquifers in the study area is largely unknown. The depth of the nine synoptic wells completed in buried aquifers sampled ranged from 48 to 190 ft. Water in these aquifers was likely subject to many different hydrologic and land-use factors and cannot be considered one kind of water. Further, the nine samples collected are too few to adequately characterize the possibly various waters in buried aquifers. The buried-aquifer samples simply suggest a range of the quality of water that may be leaking to surficial aquifers.

All samples were analyzed for physical properties and the major chemical groups or schedules detailed in table 3-2. Details about sampling and decontamination procedures are in the USGS National Field Manual for the Collection of Water-Quality Data (U.S. Geological Survey, variously dated), as amended at the time of sampling.

For the purposes of synoptic description, concentrations below the detection limit were estimated using the maximum likelihood estimation method (Helsel, 2005) when at least 25 percent of the concentrations in the synoptic set were above the detection limit. This method assumes a log-normal distribution and uses the concentrations above the detection limits and the proportion of data below the detection limits to describe the distribution of the data set. This distribution is then used to estimate the concentration of samples below the detection limits. Individual estimated concentrations have no meaning, but the distribution of the synoptic sample set as a whole is estimated.

Dissolved gas samples only were collected from surficial aquifers, where recharge dates were likely to be recent, because SF_6 concentrations are below the detection concentration for waters recharged before about 1970. One well (G03) produced so little water that only herbicide and water isotope samples could be collected. Dissolved-gas samples were not collected at all surficial aquifer wells (only at 29 of 39 wells) because of sampling difficulties. Dissolved-gas samples must be collected with a positive-displacement pump to maintain the water at ambient pressures or greater to prevent gas exsolution from the water. Ten wells produced so little water that a positive-displacement pump emptied the well casing before a

Table 3-2. Analytical methods.

[USGS, U.S. Geological Survey; NWQL, National Water Quality Laboratory; OGRG, Organic Geochemistry Research Group Laboratory; RCL, Reston Chlorofluorocarbon Laboratory; GW, ground water; SW, surface water; GC/MS, gas chromatography/mass spectrometry; LC/MS, liquid chromatography/mass spectrometry]

Analyte group	USGS laboratory	Laboratory schedule	Analytical method reference
Major ions—GW	NWQL	1	Fishman, 1993
Major ions—SW	NWQL	573	Fishman, 1993
Nutrients—GW	NWQL	2752	Fishman, 1993
Nutrients—SW	NWQL	2702	Fishman, 1993
Herbicides: GC/MS	OGRG	GCS	Kish and others, 2000
Herbicides: LC/MS	OGRG	LCPD	Lee and Strahan, 2003
Water isotopes	NWQL	1142	Fishman and Friedman, 1989
Age dating: SF_6	RCL	None	Busenberg and Plummer, 2000
Dissolved gasses	RCL	None	Busenberg and Plummer, 1992

dissolved gas sample could be collected. One well completed in a buried aquifer (E04D) was sampled for dissolved gasses because pumping-test data indicated that the confining unit overlying the aquifer may leak or end nearby (Widseth Smith Nolting and Associates, written communs., 2001). One well was sampled twice as a quality-control duplicate.

Synoptic ditch-water samples were collected at six gages where continuous-record stream-stage gaging stations were located (SW1–SW6). During sampling, ditches were at low to medium flow, with all gages influenced by recent precipitation to various degrees. The flow at one gage (SW2) was substantially higher than base flow. Low stream velocities associated with the low to medium flow prevented isokinetic sampling. Isokinetic samples represent a flow-weighted composite of the streamflow as a whole (U.S. Geological Survey, 1998). The non-isokinetic samples collected for this study were composites of several vertically integrated dip samples across the ditch channel.

Variability Sampling

Procedures for sample collection were the same as those used for synoptic ground-water sampling with the following exceptions: a peristaltic pump was used in place of a positive-displacement pump to deliver well water to the sampling chamber because no gas samples were collected.

Ditch-water variability sampling was started in October 2002. Samples were collected monthly until winter, then about every 6 weeks until December 2004. No sample was collected when no water flowed in a ditch. Additional samples were collected during times of high runoff in June 2003, May 2004, and October 2004 to assess the variability of constituent concentrations with flow rate. Snowmelt-runoff samples were collected in March 2004 and March 2005. Suspended-sediment concentrations and major ions were not collected during initial sampling in October 2002. The total number of samples collected per year varied from 7 for gage SW5 in water year 2004 to 11 at gage SW6 for water years 2003 and 2004. Four or five samples at each gage were collected in water year 2005. A total of 146 ditch-water samples were collected for this study.

A churn splitter was not used and alkalinity concentrations were not determined during the first two samplings in October and November 2002. From January 2003 until October 2004 samples were collected for suspended sediment, major ions, and nutrients, and alkalinity was measured. After October 2004, major-ion samples were no longer collected because the variability of these concentrations was adequately characterized.

Procedures for sample collection were the same as those used for synoptic ditch sampling with the following exceptions: When stream velocities exceeded 1.5 ft/s, the equal-width-increment sampling method was employed, allowing for isokinetic samples to be collected. When stream discharges were extremely low, the single-vertical-dip sample method was used; and in one instance, a sample was collected by peristaltic pump directly from the ditch. A portable water-quality laboratory trailer was used to collect all samples except those collected during winter months. Instead, samples were processed and preserved in the cab of a field truck while the truck door remained open. During extremely cold conditions, it was necessary to run the truck engine and heater to prevent the samples from freezing. Churn splitting was omitted during freezing temperatures. Instead, a plastic compositing bottle was used to mix (by swirling) an unfiltered water sample, which was then poured into sample bottles. The remaining water was filtered into sample bottles. The plastic collection bottle was replaced by a 500-mL clear glass bottle during extremely low flows and freezing temperatures, after which samples were transferred to the compositing bottle for further processing.

Appendix 4. Water-Quality Assurance and Control

A quality-control (QC) program was used to assess the accuracy and precision of project water-quality samples. The program consisted of the comparison of field and laboratory measurements and the assessments of total constituents, field blanks, and duplicate samples. Analyses of field and laboratory values and of total constituents were made at every sample. A field blank or duplicate was collected for about every 10 ambient samples collected. The QC data generated by this program compliment those collected by the analyzing laboratories and quantify the reliability of the water-quality data produced.

Quality-control checks between the properties of water measured in the field and those measured in the laboratory are expressed as relative percent difference (RPD) [(field value minus laboratory value) divided by (field value plus laboratory value) times 100 percent]. The RPDs were evaluated for bias and temporal trends. Three properties of water were measured both in the field and at the lab: pH, specific conductance, and alkalinity. Alkalinity was not measured in the laboratory for ditch-water samples. Field and laboratory data pairs were available for 47 of the 48 ground-water synoptic samples collected, except for alkalinity. Well G05 yielded so little water that no field properties could be measured. Laboratory alkalinity was not determined for seven samples because the hold time was exceeded before analysis. The field alkalinity sample at well G16 was ruined, so no alkalinity comparison could be made. The laboratory alkalinity concentration was substituted for the missing field value in all other calculations. Field and laboratory data pairs were available for 119 of 146 ditch-water samples because laboratory values were not determined before November 2002 and after October 2004.

Differences between field and laboratory measurements for both pH and specific conductance were small for both ground- and ditch-water samples (fig. 4-1). The RPDs show no trends and indicate no sample labeling errors; however, alkalinity differences were substantial. In 19 of 40 samples with field/laboratory alkalinity pairs, absolute RPD (calculated the same as RPD, but with absolute values) were greater than 10 percent. Field alkalinity concentrations were determined by incremental titration of filtered samples within 30 minutes of collection, whereas laboratory concentrations were measured by fixed-endpoint titration of unfiltered samples (ditch water only, ground-water samples were filtered) 2 to 4 weeks after collection. Alkalinity concentrations can change after a sample is collected. If dissolved carbon dioxide degasses from a sample, raising its pH, calcite can precipitate. In a previous study comparing the field and laboratory alkalinity methods in samples from two surficial aquifers in the Red River of the North Basin (National Water-Quality Assessment (NAWQA); Cowdery, 1997), differences were about half those in this study. The larger alkalinity differences in samples from this study indicate that method difference alone cannot be accountable. It is possible that some characteristic (high dissolved carbon dioxide concentrations, for example) of the ground water in the study area may have resulted in precipitation of calcite and lower laboratory alkalinity measurements. In all but one sample where alkalinity RPDs exceeded 10 percent, the laboratory alkalinity concentration was less than the field alkalinity concentration, consistent with the possibility of calcite precipitation in laboratory samples. The relatively long holding times for samples from this study would promote such precipitation. The field alkalinity concentrations are more accurate than the laboratory concentration and were used in data analyses for this study.

Assessments of total constituents include ion balances (difference between the sums of cation and anion equivalents) and the ratio of the sum of solids to specific conductance. The RPDs of ion balances (cation sum minus anion sum divided by cation sum plus anion sum) were near zero in ground-water samples, indicating that the all major ions in the samples were accurately measured (fig. 4-1). Two ground-water sample ion balances exceeded 3.5 percent (samples from wells G16 and G08). The bicarbonate concentration of the sample from well G16 was determined in the laboratory and is likely too low, based on the preceding discussion of alkalinity (bicarbonate concentration is calculated from alkalinity concentration). The bicarbonate concentration of the sample from well G08 is too high, and there is a very large difference between field- and laboratory-determined alkalinity concentrations. It is unknown which concentration is correct. Assuming that all other ion concentrations are correct, an alkalinity value of 228 mg/L as calcium carbonate ($CaCO_3$) would produce a bicarbonate concentration of 279 mg/L-$CaCO_3$ and a perfect ion balance. This value of bicarbonate is about midway between the field and laboratory concentrations and is probably a better value to use for bicarbonate concentration for the sample from well G08. Most ion-balance RPDs in ditch-water samples were greater than zero indicating a slight cation bias, although most concentrations were within laboratory tolerances. There was a trend of increasing RPD over time, but it was not consistent. The reason for the two low values shown in the boxplot in figure 4-1 is unknown; all constituent concentrations are within typical values, and the relations between constituents are not unusual for those two low values.

The ratio of the sum of dissolved solids (SoS, in mg/L; fig. 4-1) and specific conductance (SC, in µS/cm) is generally within a small range for a given natural water. Hem (1989) notes that this value is usually in the 0.55–0.75 (mg/L)/(µS/cm) range, but may range up to 0.96. Surficial ground-water samples from the study area had a median of

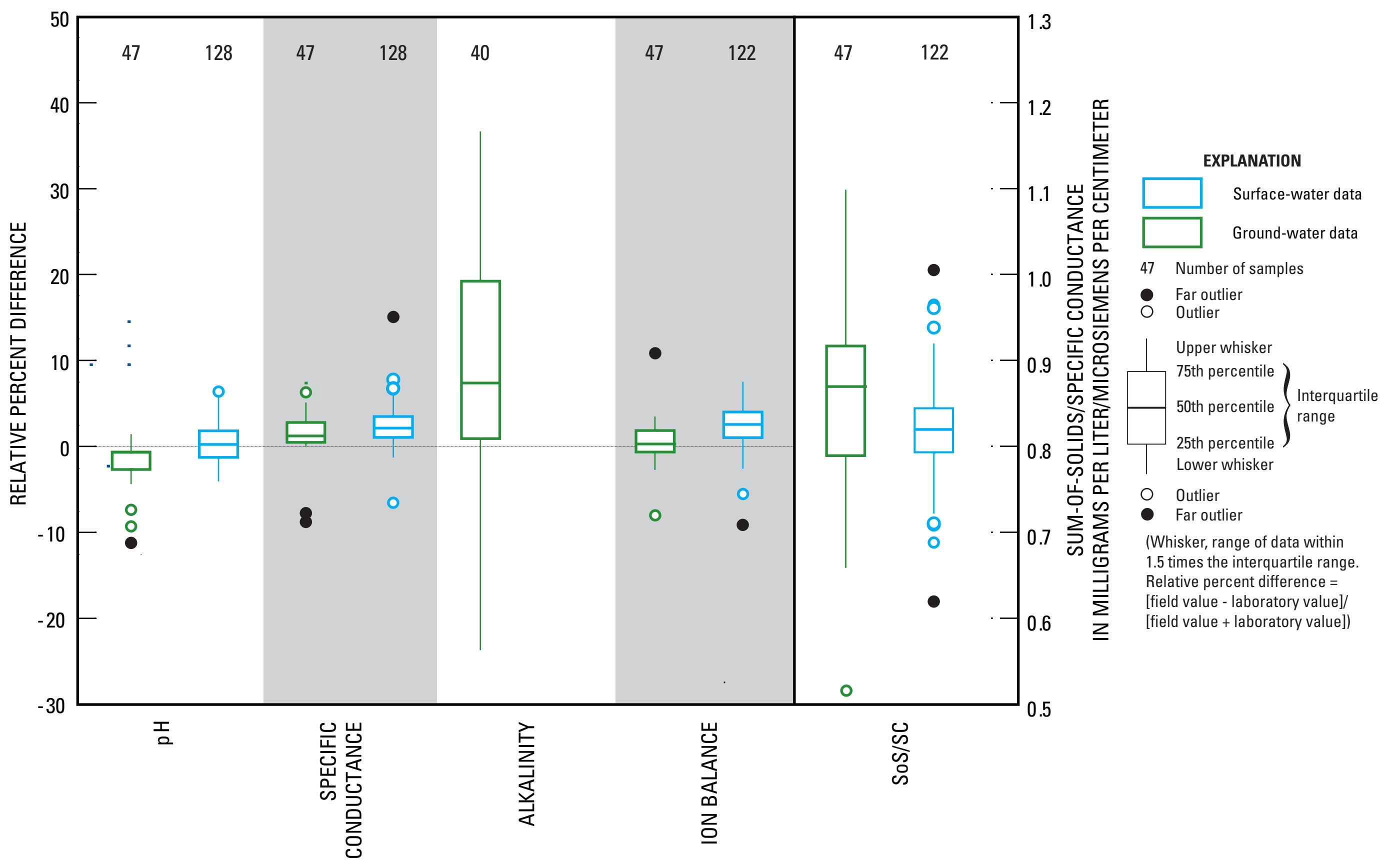

Figure 4-1. Water-quality-control data, Glacial Ridge study area, northwestern Minnesota, 2003–5.

0.85 (mg/L)/(µS/cm), high for Hem's range and 0.25 (mg/L)/(µS/cm) higher than the median for the previous NAWQA study samples (Cowdery, 1997). Four samples from this study (8.5 percent) were greater than 0.96, and one sample was 1.1 (mg/L)/(µS/cm). Hem reports that high SoS/SC values are associated with high sulfate concentrations, but sulfate concentrations in all samples were less than 155 mg/L and account for no more than 12 percent of the anions in high SoS/SC samples. The reason for the high SoS/SC values for study samples is unknown. Surface-water sample ratios were typically in the low range suggested by Hem (1989). The range of ratios were from 0.62 to 1.01 (mg/L)/(µS/cm) with a median of 0.82. Other water-quality characteristics of the sample with the lowest ratio show nothing unusual; the constituent concentrations are not unusually high or low, and the relations between different constituents do not show that it is an outlier.

Field-blank samples document possible contamination and gage the meaning of low concentrations. Blank samples were collected following collection of six ground-water, eight ditch, and one wetland samples. The wetland sample and two ground-water samples were synoptic network samples. Of the 15 blank samples, 2 ground-water and 6 ditch-water samples were analyzed for major ions, and all samples were analyzed for nutrients. Two ground-water and one wetland samples were analyzed for herbicides. Although results from some blank samples did have major-ion concentrations above the detection limit, the highest concentration for ground water was 0.33 mg/L of calcium, representing 0.38 percent of the ambient concentration for that sample, and the highest concentration for ditch water was 0.07 mg/L. Calcium is present at very low concentrations (less than 0.07 mg/L) in all other blank samples analyzed for this study. The laboratory specific conductance for these blank samples is also slightly above zero (3–5 µS/cm). This suggests some kind of systematic, very small calcium contamination in all samples, either in the field collection process or in laboratory analysis. This contamination is far smaller than ambient concentration and does not affect calcium characterization of water in this study. The highest proportion of blank-sample concentration to ambient-sample concentration of all major ions was 0.9 percent for magnesium. Three blank samples had concentrations at the detection limit (one for iron, one for particulate organic nitrogen, and one for ammonia), whereas the corresponding ambient samples had concentrations below the detection limit. No other nutrient concentrations were greater than the detection limits in any blanks. Of the three herbicide blank samples analyzed, only one had a corresponding ambient sample with a measurable concentration of herbicides. That ambient sample (well G01) contained 0.23 µg/L alachlor ethanesulfonic acid, whereas the blank-sample concentration was below the detection limit. These results demonstrate that decontamination procedures were adequate to prevent cross-contamination that could affect the interpretation of water-quality data.

Duplicate-sample data show the variability of the water sampled and that introduced in the sampling and analyzing processes. Ground-water duplicate samples were collected from the well sequentially and, therefore, potentially include variability in the source water as well as variability from sample processing and analysis. Most of the variability results from sample processing, however, because duplicate samples were collected immediately after one another and because laboratory QC results demonstrate that variability from the analyzing process is very low. Surface-water duplicate samples were obtained by collecting a larger sample volume and splitting that volume into two and, therefore, do not include source-water variability.

Duplicates samples were collected with 14 ground-water and 6 ditch-water samples, including the 3 ground-water synoptic samples. Of these, the three ground-water synoptic and four ditch-water samples were analyzed for major ions, and all samples were analyzed for nutrients. The three ground-water synoptic duplicate samples also were analyzed for herbicides. Field measurements and major-ion concentrations in ground-water duplicate samples were within 1.6 percent of the ambient concentrations with the following exceptions: The sulfate concentration for one sample had high variability (9.09 percent, absolute RPD) for unknown reasons. This sulfate concentration was about one-half the median concentration for all ground-water samples, meaning that a small difference in concentration produced a relatively large absolute RPD. Laboratory alkalinity, and the bicarbonate concentration calculated from it, had absolute RPDs of as much as 6.8 percent. This high variability probably is related to calcite precipitation during the relatively long period between collection and analysis, as noted previously. Field alkalinity had absolute RPDs of as much as 2.4 percent. Although field-alkalinity variability was much lower than the laboratory-alkalinity variability, it was relatively high and probably resulted from the fact that the titration environment in the field is harder to control. The median and standard deviation for absolute RPD for all field measurements and major-ion concentrations in all ground-water duplicate samples was 0.97 percent and 1.64 percent, respectively. This means that the true value of field measurements and major-ion concentrations of a sample is within about 5 percent of the reported value with 99 percent confidence.

The ground-water nutrient duplicate data are more complex. Of the 84 nutrient-concentration pairs from 14 samples, 41 (49 percent) had concentrations greater than the detection limit. Twenty-one concentration pairs (25 percent) had absolute RPDs greater than 3.5 percent. Of these, 62 percent (13 concentration pairs) were within 10 times the detection limit, indicating low-concentration samples. Small differences in low concentrations produce large absolute RPDs and do not represent high variability in nutrient analyses. Eight ground-water nutrient-concentration pairs with absolute RPDs greater than 3.5 percent had substantial concentrations. These pairs represent 29 percent of the 28 nutrient-concentration pairs with concentrations greater than 10 times the detection limit. Among the high nutrient-concentration pairs, the median and standard deviation for absolute RPD are 1.20 and 6.50 percent, respectively. This means that the true value of the nutrient concentration of a sample is within about 20 percent of the reported value with 99 percent confidence.

Three duplicate ground-water samples were analyzed for herbicides. Two of these samples contained very low, but detectable concentrations of 5 of the 35 compounds analyzed. In all cases, the duplicate samples had concentrations within 14 percent (absolute RPD) of the ambient sample. Concentrations ranged from 0.03 to 0.20 μg/L and absolute differences ranged from 0.00 to 0.04 μg/L. These same duplicate ground-water samples also were analyzed for isotopes. Composition differences ranged from zero to 0.5 percent.

Among major ions, the range of RPDs in ditch-water duplicate samples ranged from -0.9 percent for dissolved sodium to 1.2 percent for dissolved potassium. This indicates very little variability between the ambient and duplicate samples and confirms field collection methods.

Among nutrients, the range of RPDs was larger than for major ions; from -7.0 percent to 1.8 percent. Much of this higher variability was due to the smaller range of values in nutrients and the relatively higher variability of those low concentration values. For example, concentration pairs with the largest difference (-7.0 percent for dissolved orthophosphorus) were near the detection limit (0.01 mg/L), and seven were less than the detection limit. Very conservative estimates of the variability at these low concentrations are 0.01 mg/L (the detection limit), so the expected variability is high at these concentrations.

Aggregated blank and replicate data from variability and spatial samples and for ground-water and surface-water samples assess the cross-contamination and reproducibility of all the water samples for this study. In general, the above-described quality-control issues do not appear to affect the water-quality results of this study.

Appendix 5. Glacial History

For much of the past several hundred thousands of years, at least two lobes of the Laurentian Ice Sheet have covered the study area. These ice lobes and the water melted from them are responsible for the deposition of all of the geologic material found within 200–300 ft of land surface and for the natural geomorphology of the study area. In general, glaciers deposit two hydrologically different materials. Till is unsorted, containing grains that range in size from clay to boulders. It is deposited directly from melting ice without sorting by meltwater and is poorly permeable. Stratified deposits are generally coarse-grained sand and gravel sorted by glacier-fed meltwater streams. These deposits are sinuously linear where deposited on or within ice (ice-contact stratified deposits) or sheet-like where deposited in front of glacier ice (outwash). Most stratified deposits are permeable and form aquifers. Another kind of ice-contact stratified deposit, fine-grained silts and clays deposited in ice-walled or proglacial lakes, are relatively rare in the study area and are virtually impermeable.

Each time ice lobes retreated from the study area, a giant proglacial lake formed in the lowlands along the Red River of the North between the continental divide at Browns Valley, Minn., and the disintegrating ice. This lake formed because the natural drainages to Hudson Bay and Lake Superior were blocked by the ice lobes. The lake drained out through the Minnesota River valley except when lower outlets to the east or north were ice-free. The last of these lakes to form was glacial Lake Agassiz. At its largest, Lake Agassiz covered 135,000 mi^2 during 9,900–9,500 years BP (Teller and Clayton, 1983). Because the study area is at the eastern edge of this lake, a thin veneer of beach sands and lake clays covers most of the area.

Each time a glacial lobe advanced, it may have eroded previously deposited sediments, deposited till, produced ice-contact stratified deposits and outwash, and created a glacial lake in the Red River Lowlands that deposited coarse- and fine-grained sediments. With each advance, only some of these sediments were deposited in the study area. Each time a glacial lobe retreated, a large proglacial lake formed beaches and deposited in its deeper waters fine-grained sediments winnowed from the shores. Calving, sediment-laden icebergs, and sediment-choked outwash streams entered the lake, producing a range of mostly coarse-grained deposits in the study area. What remains today is a very complex set of sediments, none of which are areally extensive or homogeneous. The most continuous and homogeneous of these deposits are the beaches from the last lake because these were not overridden by subsequent glacial ice.

Throughout the study area, subsurface deposits are characterized by thick, poorly permeable tills containing isolated coarse-grained sediments that range in thickness from several feet to at least 69 ft. Lindgren (1996) identified coarse-grained sediment whose top was more than 400 ft below the land surface. Some buried coarse-grained sediments are traceable between boreholes over a mile, or so, but driller's logs are not numerous enough to delineate individual deposits. At least two tills are at the surface in the study area: the relatively sandy till of the Red Lake Falls Formation and the overlying, relatively clayey till of the Hout Formation (nomenclature of Harris and others, 1974). The intervening Wylie Formation (a lacustrine sediment) could not be recognized in the study area but likely is present. Some of the lacustrine sediments attributed to Lake Agassiz in this study may be exposures of the Wylie Formation. In the study area, tills of the Red Lake Falls Formation could not be distinguished from those of the Hout Formation in the field. The more clay-rich Hout Formation tills are nearly always modified by wave action of glacial Lake Agassiz, which winnows out some of the finer grained sediments, leaving it unusually sandy and appearing like Red Lake Falls Formation tills. In many areas, a sandy uppermost till is underlain by very dense, hard, clay-rich till. These tills were interpreted as wave-washed Hout Formation till overlying unwinnowed Hout Formation till.

On the study area surface, stagnation moraine deposits undisturbed by Lake Agassiz are present only in the southeastern quadrant, near Maple Lake. Though not topographically raised above the surrounding areas, this moraine is characterized by classic knobs, kettle lakes, and wetlands—remnants of glacial ice-blocks buried in ablation till as the last glacier disintegrated about 14,000 years BP. Most surface sediment here is till, presumably of the Hout Formation. The Maple Lake Basin appears to have formed differently. Early in the last episode of lake formation in the Red River Lowlands (Cass Phase, 13,800–13,680 years BP), ice still stood in the study area, but two lake basins formed against the melting ice: glacial Lake Climax to the southwest and glacial Lake Koochaching to the northeast of the study area (Hobbs, 1983). At this time, the lakes were connected by the McIntosh Channel. Water flowed through this channel in either direction, depending on the relative lake levels, and deposited deltas in each lake.

The Maple Lake Basin appears to be a channel similar to the McIntosh but formed slightly later, just as the two earlier glacial lakes merged into glacial Lake Agassiz. The elevation of the bottom of Maple Lake is about 1,160 ft, which is 65 ft below the lowest McIntosh Channel bottom near Fertile, Minn. when corrected for isostatic rebound (25 ft more at Maple Lake than at Fertile; Hobbs, 1983). Beach ridges at an elevation of 1,160 ft are visible northeast and south of the lake, showing that the glacial lake levels stood here for a time. The Maple Lake Basin probably was eroded as Lake Koochiching water found its way around the ice separating it from Lake Climax and abandoned the McIntosh Channel or other lower channels. Flow against the ice would have been very short lived as the lakes merged quickly into glacial Lake Agassiz proper, with southern and eastern basins. This was the beginning of the Lockhart Phase of the lake (13,680–12,880 years BP; Fisher, 2005).

Excepting the southeast quadrant, the study area surface is a sloping plain of till, covered with long, narrow, thin, sand and gravel beach ridges. The till plain is a ground moraine that was deposited at the base of one or more glacial lobes that crossed the area. The surface of this till is modified by Lake Agassiz waves, which produced the beach ridges when its stage was steady for some time. The lake waves removed fine-grained sediment from the till, leaving a lag deposit of sand and gravel at the beaches. At least 12 beach ridges are visible on topographic maps in the study area. Many more minor beach ridges are visible from a height when the sun is close to the horizon (see cover photograph). The highest and lowest of these beaches are the most well developed and easily identified. These are the Herman, which is the oldest, in the southeast, and the Campbell, which is the youngest, in the west. Three other beaches have names between the Herman and Campbell: the Norcross, Upham, and Tintah Beaches (Fisher, 2005). These beaches appear spread out across the study area forming sets of beach ridges. It is not known which ridges in the study area belong to these named beaches.

The sequence of the formation of these beaches is somewhat debated (Fisher, 2005; Teller and Leverington, 2004), but they all formed during the Lockhart, Moorhead, Emerson, and Nipigon phases of Lake Agassiz between 13,400 and 10,100 years BP. The Lake had three potential outlets. At any one time, two of these three outlets were either blocked by ice or higher than another ice-free outlet. As glacial lobes waxed and waned, outlets were opened or closed, changing the level of Lake Agassiz. During the time of Lake Agassiz, the southern outlet, which was ice free after 13,680 years BP, was periodically eroding when it was the active outlet. Each erosion event was probably triggered by unusually high flow out of the lake and resulted in a lower lake level. In this way, the level of Lake Agassiz was characterized by periods of stable lake level interspersed with rapid lake-level change. Each beach in the study area records a period of stable lake level. Larger beaches, like the Herman and Campbell, represent longer stable periods, and perhaps more stormy lake conditions.

The location of the study area at the junction of the southern and eastern basins of Lake Agassiz, and the low slope of the till plain there, probably explains the unusually large number of beach ridges deposited. The shore in the study area was exposed to the Lake on three sides, meaning that wave action produced by winds from most directions would form beach deposits. More than in other areas of Lake Agassiz, small changes in lake levels acting for relatively shorter periods could produce beach ridges that would be preserved in the study area.

Generally, waves at the shore of Lake Agassiz formed a beach by eroding scarps as high as 25 ft into the sloping till plain, winnowing the clay and silt out of the till, and leaving a sand and gravel lag on the side of the step toward the basin. This process created the characteristic beach-ridge landform seen across the study area (figs. 1 and 5). Hydrologically, each beach ridge is an aquifer. A ridge frequently dams a back-beach basin, which is now occupied by a wetland. Seeps sometimes occur downgradient from a beach ridge, creating an environment colonized by fen species. Discharge from these seeps flows downhill, sometimes feeding an adjacent back-beach basin wetland or sometimes simply evaporating. Drilling and coring of wells and boreholes showed that the upper few feet of till immediately beneath the beach ridges and wetland basin areas are coarser grained than underlying till. This till is a wave-washed version of the underlying, more clay-rich till.

As the lake level dropped, beaches were abandoned, and the original postlake drainage formed in the study area. By the time Lake Agassiz drained for the last time about 8,440 years BP, most of the study area was composed of sandy beach ridges covered by mesic prairie, separated by shallow wetlands of several types. Surface-water flow was diffuse, moving through wetlands and wet prairies along beach ridges until it could cross a ridge through a low area (Jason Eckstein, The Nature Conservancy, Glacial Ridge Project, oral commun., 2004). Flow would then proceed again along the next ridge.